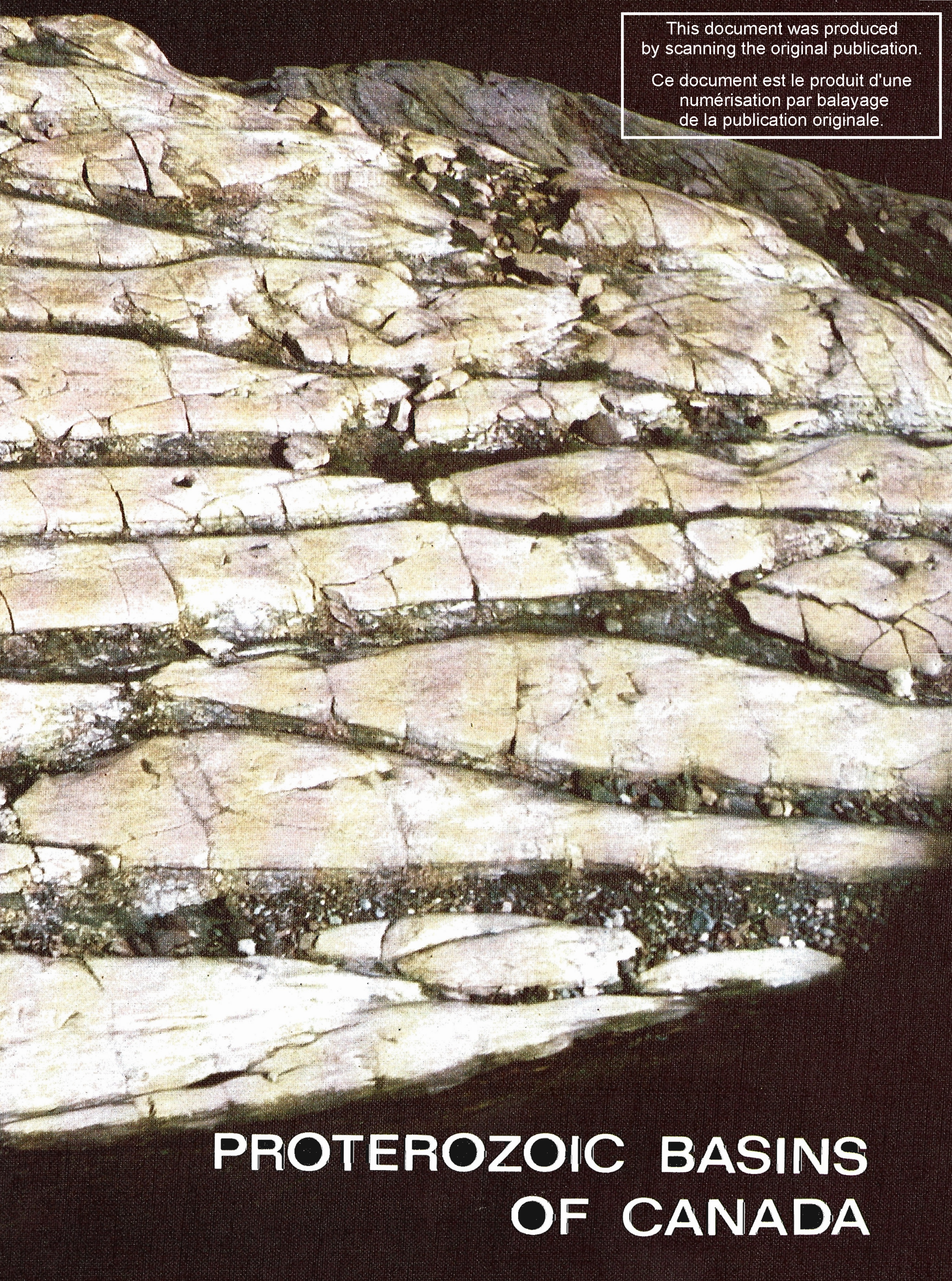

PROTEROZOIC BASINS OF CANADA

Geological Survey
Paper 81-10

PROTEROZOIC BASINS OF CANADA

edited by F.H.A. Campbell

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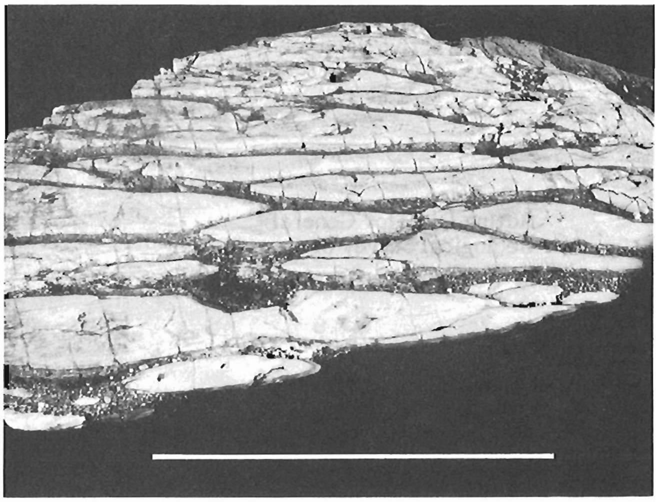
Preface

From time to time the Geological Survey has published the results of symposia of broad, general interest. This allows a wide distribution of the results of these meetings and also makes a much larger audience aware of advances being made in certain fields.

In 1970 we published the proceedings of a symposium on "Basins and Geosynclines of the Canadian Shield". This was probably the first attempt to bring together the results of the reconnaissance studies that had been initiated in the Shield after World War II. In the past decade more detailed mapping projects have been undertaken and our understanding of the Shield has undergone considerable refinement as a result of the application of facies models and plate tectonic theory. It thus seemed appropriate for another symposium to be held, which would allow these advances to be described and discussed. Because much of the work in the past decade has been directed to the study of basinal evolution, particularly in the less highly altered successions, the organizers decided to limit the topic to "Proterozoic Basins of Canada". This highly successful symposium sponsored by the Geological Association of Canada, was held in Halifax in May 1980. Early in the planning for the symposium the Geological Survey was asked to undertake the publication of the proceedings. I am pleased that this was possible and I congratulate the symposium editor and the staff of the Survey's Geological Information Division for producing such an informative and attractive volume.

Ottawa, October 1981

J.G. Fyles
Acting Director General
Geological Survey of Canada



COVER

Aerial view of the mounds of the Reef facies of the Beechey Platform, Western River Formation, Kilohigok Basin. The island outcrop shown in the photograph is in the southeastern part of Bathurst Inlet, N.W.T.

The bar scale represents approximately 100 m.

GSC photograph 203059-H by F.H.A. Campbell.

The funds for the printing of the covers for this volume were generously provided by the Robinson Fund of the Geological Association of Canada.

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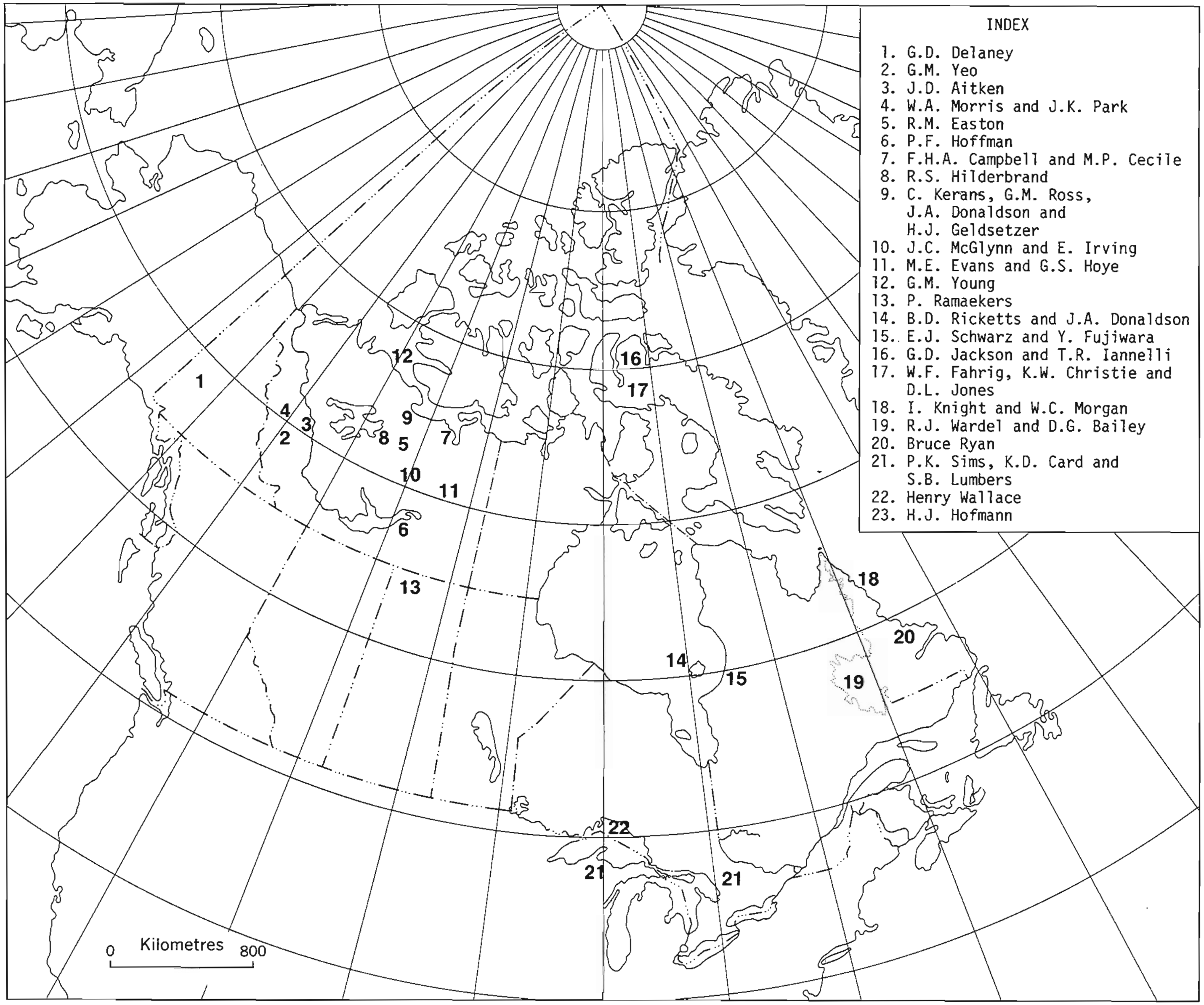
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INTRODUCTION

In 1970 the Geological Survey of Canada published the results of a symposium on "Basins and Geosynclines of the Canadian Shield" (GSC Paper 70-40). The studies on which this symposium was based were largely of a reconnaissance nature but provided the foundation upon which Precambrian geologists have been able to establish local, regional, and continental frameworks. Advances over the past decade have aided in deciphering the basinal evolution of individual successions and relating these to other, typically widely separated sequences. Thus in 1980, a symposium could be devoted solely to Proterozoic sequences in the Canadian Shield and the Cordillera which provided scientists with a forum to discuss and develop their ideas and interpretations. This symposium on "Proterozoic Basins of Canada", was held in Halifax at the Annual Meeting of the Geological Association of Canada and Mineralogical Association of Canada.

The most striking difference between the 1970 and 1980 symposia is the degree to which interpretations of basinal evolution have been affected by the advent of plate tectonics. During the coming decade, as in the past, I am sure the interpretations contained herein will be modified and refined resulting in a comprehensive synthesis of the evolution of Proterozoic Basins in Canada.

The papers in this volume are arranged from west to east (see index map) and in stratigraphic order where papers are from the same region. With the exception of the paper by H.J. Hofmann on Precambrian fossils, the papers cover specific basins and individual authors have gone to considerable lengths to integrate their models into a regional picture. The papers cover specific basins in Canada, the exception being the paper by H.J. Hofmann on Precambrian fossils, and individual authors have gone to considerable lengths to integrate their models into a regional picture.

Acknowledgments

R.T. Haworth, Program Chairman, and organizers of the GAC-MAC Annual Meeting are gratefully acknowledged for their co-operation and assistance in making the symposium a success.

In the two years prior to the meeting Dr. D.J. McLaren, then Director-General of the Geological Survey of Canada, gave his whole-hearted support to the venture.

While I accept full responsibility for the editorial content of the volume, many others aided in reviewing the papers. In alphabetical order these were: J.D. Aitken, W.R.A. Baragar, R.T. Bell, K.D. Card, M.E. Evans, W.F. Fahrig, J.M. Franklin, M.J. Frarey, P.F. Hoffman, M.B. Lambert, J.C. McGlynn, W.C. Morgan, H.C. Palmer, E.J. Schwarz, F.C. Taylor and G.M. Young.

Mrs. B. Cox of the Precambrian Geology Division (GSC) conscientiously typed and retyped versions of the manuscripts. The Photographic and Photomechanical sections of the Geological Survey supplied their customary excellent service in reproducing the illustrations.

Finally, I thank all authors for their perseverance in dealing with a sometimes capricious editor.

En 1970, la Commission géologique du Canada a publié les résultats d'un symposium sur les "Bassins et géosynclinaux du Bouclier canadien" (Étude 70-40 de la C.G.C.). Les études sur lesquelles reposait ce symposium consistaient en grande partie en des travaux de reconnaissance, mais elles ont tout de même permis à des géologues du Précambrien d'établir des structures locales, régionales et continentales. Les progrès accomplis depuis dix ans ont contribué à retracer l'évolution de certaines successions, qui sont ordinairement des séquences largement séparées, et de les mettre en relation les unes avec les autres. Ainsi, en 1980, un symposium a-t-il pu être consacré entièrement aux séquences protérozoiques du Bouclier canadien et de la Cordillère, permettant à des scientifiques de se réunir pour discuter et développer leurs idées et leurs interprétations. Ce symposium, tenu sous le thème de "Bassins protérozoiques du Canada", a eu lieu à Halifax dans le cadre de l'assemblée annuelle de l'Association des géologues du Canada et de l'Association canadienne de minéralogie.

Ce qui frappe le plus dans les différences entre les symposiums de 1970 et de 1980, c'est de voir jusqu'à quel point la tectonique des plaques a modifié les interprétations de l'évolution des bassins. Je suis sûr qu'au cours des dix prochaines années, comme par le passé, nos interprétations d'aujourd'hui seront modifiées et précisées et qu'elles déboucheront sur une synthèse complète de l'évolution des bassins protérozoiques du Canada.

Les études composant ce volume sont groupées géographiquement d'ouest en est (voir la carte-index) et ceux qui s'appliquent à la même région sont classées par ordre stratigraphique. Ces études portent sur des bassins précis du Canada, à l'exception de celle de H.J. Hofmann, qui a pour objet les fossiles du Précambrien; certains auteurs ont dû déployé des efforts considérables pour intégrer leurs modèles dans un tableau régional.

Remerciements

Je remercie chaleureusement R.T. Haworth, président du Programme, ainsi que les organisateurs de l'assemblée annuelle des deux associations pour avoir contribué à faire de ce symposium un succès.

Durant les deux années qui ont précédé la réunion, M. D.J. McLaren, alors directeur général de la Commission géologique du Canada, a soutenu l'entreprise du fond du coeur, et je lui en suis reconnaissant.

Bien que j'assume l'entière responsabilité de la révision du volume, beaucoup d'autres personnes ont contribué à revoir les documents. Ce sont, par ordre alphabétique, J.D. Aitken, W.R.A. Baragar, R.T. Bell, K.D. Card, M.E. Evans, W.F. Fahrig, J.M. Franklin, M.J. Frarey, P.F. Hoffman, M.B. Lambert, J.C. McGlynn, W.C. Morgan, H.C. Palmer, E.J. Schwarz, F.C. Taylor, et G.M. Young.

Je remercie également Mme B. Cox, de la Division de la géologie du Précambrien (C.G.C.), qui a consciencieusement tapé et retapé les diverses versions des manuscrits. J'adresse aussi mes voeux de remerciement aux sections de la photographie et de la photomécanique de la Commission géologique, qui ont reproduit les illustrations avec le brio qu'on leur connaît.

Finalement, je remercie tous les auteurs de la patience dont ils ont fait preuve à l'endroit d'un réviseur parfois capricieux.

F.H.A. Campbell

NOTE

"For purposes of comparison the Rb-Sr ages in this volume have been recalculated where necessary using $\lambda^{87}Rb$ of $1.42 \times 10^{-11}\ yr^{-1}$."

1.

THE MID-PROTEROZOIC WERNECKE SUPERGROUP, WERNECKE MOUNTAINS, YUKON TERRITORY

G.D. Delaney
Department of Geology, University of Western Ontario, London, Ontario, N6A 5B7

Delaney, G.D., The mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 1-23, 1981.

Abstract

The Wernecke Supergroup is subdivided, from oldest to youngest into the Fairchild Lake Group, the Quartet Group and the Gillespie Lake Group. The Fairchild Lake Group consists of at least 4 km of light-grey weathering, thin bedded to laminated siltstone, mudstone and fine sandstone with some intercalated carbonate. Locally these rocks have been transformed into phyllites and schists. The Fairchild Lake Group is tentatively subdivided into five formations. Carbonate members near the middle and at the top of this group are important stratigraphic markers. Although most of the Fairchild Lake Group was deposited in deeper water, by predominantly southerly flowing currents, a shallow marginal marine regime characterized the depositional setting of the second youngest formation.

The Quartet Group, which generally conformably overlies the Fairchild Lake Group, consists of at least 5 km of monotonous dark-grey-weathering, wavy to lenticular to flaser bedded siltstone, carbonaceous siltstone, mudstone and fine sandstone; two subdivisions are recognized in this group. Although the basal part of the Quartet Group probably accumulated in a sediment-starved stagnant basin, the greater part of the group was deposited in a shallow marine environment.

The contact between the Quartet Group and the overlying Gillespie Lake Group is transitional. The Gillespie Lake Group consists of more than 4 km of buff- to orange- to grey- to locally maroon-weathering dolomite, silty and clayey dolomite, siltstone, mudstone and fine sandstone. In the type area, seven subdivisions are defined in the Gillespie Lake Group. The lower part of this group is fine grained terrigenous sediments and terrigenous-carbonate admixtures. The upper part of the Gillespie Lake Group consists of shallow water dolomite characterized by stromatolites, cryptalgal laminates, oolites, and pisolites.

The Wernecke Supergroup hosts numerous spectacular breccia complexes which locally contain significant enrichments of U, Cu, Ba, Co or Fe. These bodies are of variable size and shape; their most spectacular development occurs in the upper part of the Fairchild Lake Group. Individual breccia complexes are commonly composed of three genetic varieties of breccia: fault breccia, stope breccia and channel-way breccia. Several types of alteration are associated with the breccia complexes including Na-feldspathization, silicification, hematitization and carbonatization. The breccia complexes are located on or near splays of the Richardson Fault Array, a system of west- to northwest-trending faults of great antiquity. The breccia complexes are envisaged as having developed in a dilational tectonic regime.

Dykes and sills of diorite, gabbro and peridotite intrude the Wernecke Supergroup. These intrusive bodies are probably of several ages.

Strata of the Wernecke Supergroup have been deformed into west-northwest trending open folds, although the structural style is locally more complex, involving faults and the effects of intrusive breccia complexes. Some of this deformation is related to the Racklan orogeny, which probably culminated between 1.2 and 1.3 Ga ago.

The Wernecke Supergroup is correlated with the lower three groups of the Belt-Purcell Supergroup of the southern Cordillera.

Paleocurrent analyses coupled with limited facies information indicate that in the Middle Proterozoic, an east-west trending depositional margin lay to the north of the present day Wernecke Mountains. It is hypothesized that the formation of this margin is related to a continent-continent collision event recorded in the Aphebian Coronation Geosyncline.

Résumé

Le supergroupe de Wernecke comprend les divisions suivantes, données en ordre ascendant: groupe de Fairchild Lake, groupe de Quartet et groupe de Gillespie Lake. Le groupe de Fairchild Lake, provisoirement subdivisé en cinq formations, est composé d'au moins 4 km de couches minces ou de lamelles altérées de siltstone, de mudstone et de grès fin, gris pâle, avec des intercalations de carbonate. Par endroits, ces roches se sont transformées en phyllades et en schistes. Des carbonates près du milieu et au sommet du groupe sont d'importants repères stratigraphiques. Bien que le groupe ait été presque entièrement déposé en eau profonde, surtout par des courants en direction sud, la deuxième formation la plus récente s'est accumulée dans un environnement marin côtier, peu profond.

Le groupe de Quartet comprend deux subdivisions et en général a été déposé en concordance sur le groupe de Fairchild Lake; il est composé d'au moins 5 km de siltstone, de siltstone carbonée, de mudstone et de grès fin altérés, de couleur gris foncé monotone, en couches onduleuses à lenticulaires, parfois à stratification confuse. Bien que la base du groupe se serait accumulée dans un bassin stagnant dépourvu de sédiments, la majeure partie du groupe a été déposée dans un environnement marin peu profond.

Une surface de contact transitional existe entre le groupe de Quartet et le groupe de Gillespie Lake susjacent. Le groupe de Gillespie Lake comprend plus de 4 km de dolomie, de dolomie silteuse et argileuse, de siltstone, de mudstone et de grès fin altérés, dont la couleur varie du chamois à l'orangé, au gris et par endroits, au bordeaux. Sept subdivisions sont identifiées dans la région type. La partie inférieure du groupe est composée de sédiments terrigènes fins et de mélanges de sédiments terrigènes et de carbonates. La partie supérieure est une dolomie d'eau peu profonde, caractérisée par la présence de stromatolites, de lamelles de carbonates construites, d'oolites et de pisolites.

Le supergroupe de Wernecke contient un grand nombre de complexes de brèches spectaculaires, par endroits enrichis en U, en Cu, en Ba, en Co, ou en Fe. Ces masses ont une forme et des dimensions variées et sont le mieux développées dans la partie supérieure du groupe de Fairchild Lake. Les complexes de brèches particuliers sont en général divisés en trois variétés génériques: les brèches de faille, les brèches de pente et les brèches de chenal. Plusieurs types d'altération y sont associés, y compris la feldspathisation sodique, la silicification, l'hématitisation et la carbonatation. Les complexes de brèches sont situés près des failles mineures de l'alignement de failles de Richardson, réseau de failles très ancien orienté ouest-nord-ouest. Ces complexes se seraient formés dans un régime tectonique de dilatation.

Le supergroupe de Wernecke est traversé par des filons et des filons-couches de diorite, de gabbro et de péridotite, sans doute d'âges différents.

La déformation des couches du supergroupe de Wernecke a produit des plis ouverts orientés ouest-nord-ouest et par endroits, un style structural plus complexe comprenant des failles et les effets des complexes intrusifs de brèches. Cette déformation se rapporte en partie à l'orogène de Racklan, qui se serait terminé il y a 1,2 ou 1,3 Ga.

Une corrélation a été établie entre le supergroupe de Wernecke et trois groupes inférieurs du supergroupe de Belt-Purcell dans la Cordillère du Sud.

L'analyse de paléocourants et les renseignements limités fournis par les faciès suggèrent l'existence, au Paléozoïque moyen, d'une marge de sédimentation au nord des monts Wernecke actuels. La création de cette marge serait reliée à une collision entre deux continents enregistrée dans le géosynclinal de Coronation de l'Aphébien.

INTRODUCTION

Parts of the northwestern Canadian Cordillera are underlain by thick successions of Proterozoic sediments (Fig. 1.1). These strata are important not only in documenting the early geological evolution of the Cordillera, but also because they host a variety of types of mineralization (Delaney et al., in press).

Recently, on the basis of lithostratigraphic correlations, Young et al. (1979) subdivided the Proterozoic successions of northwestern North America into three sequences, separated by significant regional unconformities. These sequences are named A, B, and C and were deposited from about 1.7 to 1.2 Ga, 1.2 to 0.8 Ga, and 0.8 to 0.57 Ga respectively (Fig. 1.1).

This paper documents the geology of the Proterozoic Wernecke Supergroup (Delaney, 1978) in the northern Wernecke Mountains of the north-central Yukon Territory. The Wernecke Supergroup constitutes the oldest recognized sequence (A of Young et al., 1979) in the northern Canadian Cordillera.

Other than a preliminary open file report (Delaney, 1978), only brief summary descriptions of the Wernecke Supergroup have been published (e.g. Young et al., 1979; Yeo et al., 1978). Therefore a somewhat detailed description of these rocks is presented here, together with preliminary interpretations of environments of deposition.

Prior to the present investigation, the Proterozoic strata of the Wernecke Mountains had only been studied during reconnaissance mapping projects (e.g. Wheeler, 1954; Green, 1972; Blusson, 1974; Norris, 1976). As a result of this early work several different informal stratigraphic classifications were proposed. Subsequent revisions (Bell and Delaney, 1977; Delaney, 1978) led to the proposal of some formal stratigraphic names.

In addition to the stratigraphic studies summarized in this paper, other aspects of the Wernecke Supergroup have recently been investigated. Morin (1976), Archer and Schmidt (1977), Bell and Delaney (1977), Bell (1978) and Delaney et al. (in press) examined the mineralized intrusive breccia complexes which cut the Wernecke Supergroup.

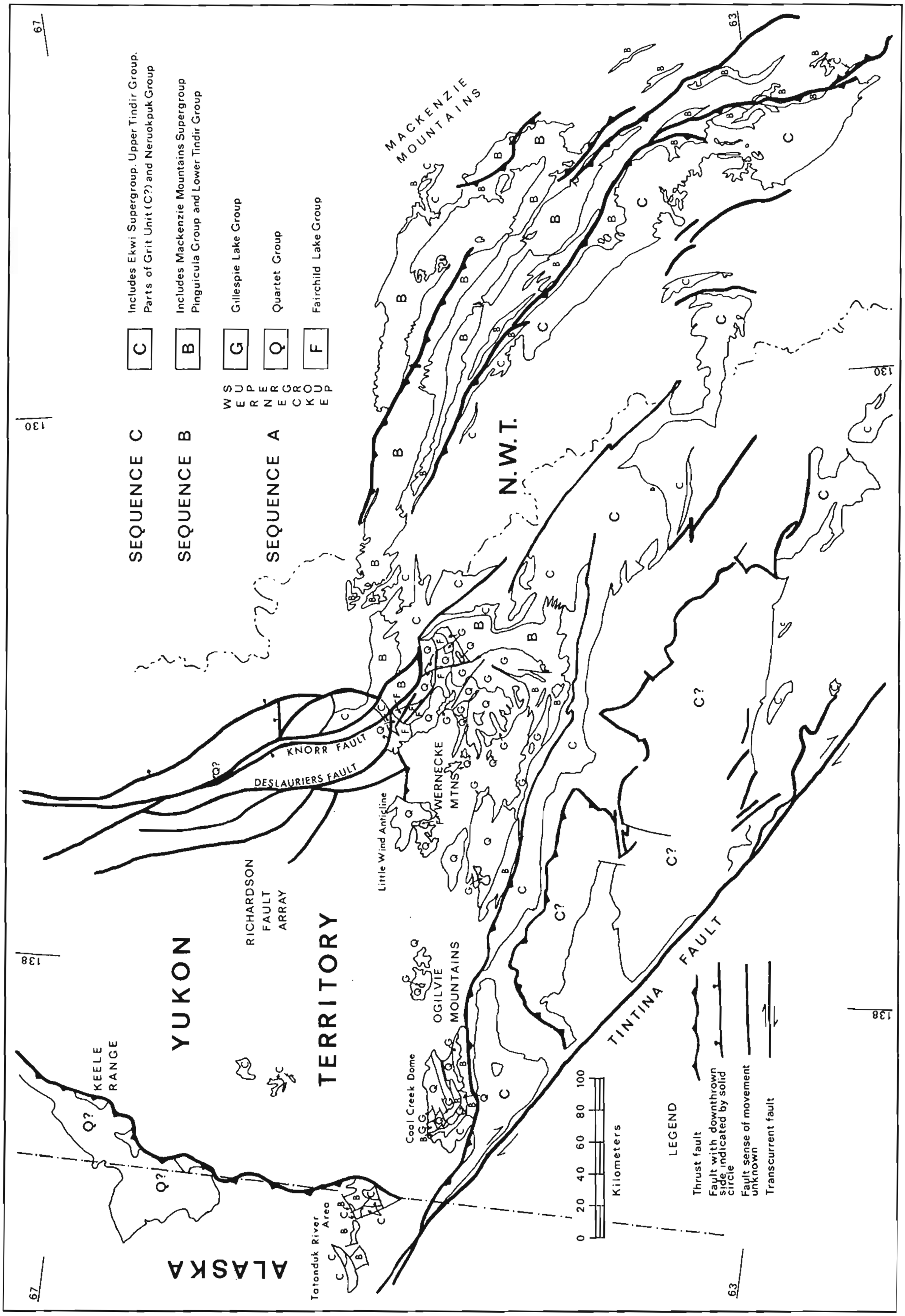

Figure 1.1. *Tectonic assemblage map of Proterozoic units, Northwestern Canadian Cordillera (modified after Tipper, 1978; personal communication, Norris, 1980).*

During the same years, W. Goodfellow of the Geological Survey of Canada conducted reconnaissance and detailed stream sediment geochemical investigations in the northern Wernecke Mountains (Goodfellow et al., 1976; Jonasson and Goodfellow, 1976; Geological Survey of Canada, 1978; Goodfellow, 1979). Laznicka (1977a, b, 1976) and Laznicka and Edwards (1979) studied the geology and mineralization in the Delores Creek area.

The oldest Proterozoic rocks exposed in the Wernecke Mountains consist of at least 14 km of generally fine grained terrigenous and carbonate rocks. This sequence has been subdivided into three groups which are named, from oldest to youngest, the Fairchild Lake Group, the Quartet Group and the Gillespie Lake Group (Delaney, 1978). This entire succession was named the Wernecke Supergroup (ibid.) as the type area is in the Wernecke Mountains and because these rocks had earlier been referred to as either the "Wernecke type Proterozoics" (Archer and Schmidt, 1977) or the "Wernecke Assemblage" (Eisbacher, 1978).

Spectacular mineralized breccia complexes are developed within strata of the Wernecke Supergroup. These sediments are also cut by dykes and sills of gabbro, diorite and peridotite.

FAIRCHILD LAKE GROUP

The Fairchild Lake Group[1] consists of at least 4 km of light grey-, grey-, and greenish grey-weathering, generally thin bedded[2], commonly laminated siltstone[3], mudstone, claystone and fine sandstone with minor intercalated carbonate rocks. These rocks are regionally metamorphosed to greenschist facies (Turner, 1968), and are locally converted to schists, slates and phyllites.

The Fairchild Lake Group outcrops in narrow, irregular, commonly deformed strips along a corridor which parallels the Bonnet Plume River (Fig. 1.2). East and southwest of Fairchild Lake however, there are thick continuous sequences of this group. The linear distribution of the Fairchild Lake Group along the Bonnet Plume River suggests that it forms the core of a complex anticlinal structure. Outside the Bonnet Plume River corridor, the only other known exposures of Fairchild Lake Group strata occur at the headwaters of Bond Creek (Fig. 1.2) and near the southern margin of the Little Wind River anticlinorium (Fig. 1.1). Nowhere has the base of this group been observed.

The Fairchild Lake Group is tentatively subdivided into five units of formational status (Fig. 1.3; c.f. Delaney, 1978). From oldest to youngest these units have been informally designated F-1, F-2, F-3, F-4, and F-TR.

F-1

The only known exposures of the F-1 occur southwest of Fairchild Lake in a thick overturned sequence (Fig. 1.2). This unit consists of at least 1800 m of generally thin to medium beds of light grey-, grey-, and greenish grey-weathering siltstone, mudstone and fine sandstone with a few thin beds of limestone. These rocks are fairly resistant to weathering, except in zones where they are dissected by closely spaced fractures or where they contain a high carbonate component. The F-1 is mostly even and parallel-laminated to lenticular bedded, with local development of wavy bedding. Small scale crossbeds, small scale asymmetrical ripple marks and load structures produced from loaded ripples are common throughout the F-1. It is characterized by a unimodal southeasterly paleocurrent pattern (Fig. 1.1).

The F-1 is conformably overlain by the F-2. This contact is marked by the first appearance of the ribbed-weathering siltstone-limestone member of the F-2.

F-2

The F-2 consists of about 400 m of light grey-, light greenish grey-, and locally buff-weathering, generally thin bedded siltstone, mudstone, fine sandstone and silty to sandy limestone. The most diagnostic feature of this unit is the presence of several ribbed-weathering members (Fig. 1.4) characterized by rhythmically alternating thin beds of siltstone and thin beds to thick laminae of silty to sandy limestone. Both these rock types contain small scale cross-beds (Fig. 1.5). Between these carbonate rhythmite members are sequences of wavy-bedded, planar-bedded and lenticular-laminated siltstone. Locally intercalated within these sequences are medium to thick beds of coarse siltstone to fine sandstone. Crossbedding and load structures are common.

Small scale folds are locally developed and as in most other units of the Fairchild Lake Group, there are zones of closely spaced fractures which mask all primary features. The mineral assemblages and textures are typical of greenschist facies regional metamorphism.

Crossbeds from the F-2 southwest of Fairchild Lake indicate a unimodal south-southwesterly trending dispersal pattern (Fig. 1.10).

The F-2 is conformably overlain by the F-3. This contact is placed where the uppermost member of ribbed-weathering siltstone and limestone of the F-2 is succeeded by a sequence of greenish grey-weathering, even parallel-laminated to lenticular bedded siltstone.

F-3

The F-3 of the Fairchild Lake Group consists of about 2000 m of thin to locally medium beds of light grey-, grey- and drab greenish grey-weathering siltstone, mudstone and fine sandstone, with a few intercalated thin beds of silty limestone. In the F-3 the bedding style varies from even parallel laminations to lenticular bedding (Fig. 1.6). Sole structures, in particular flutes are present in parts of this unit; ripple marks and load structures are abundant throughout.

The F-3 mudstones are composed of medium to coarse silt size grains of quartz and plagioclase in a matrix of sericite and chlorite. These rocks are commonly recrystallized. Many of the mudstone beds are characterized by very thin, even parallel laminations delimited by variations in the proportions of the constituent mineral grains. In the vicinity of the breccia complexes (see below), mudstones have commonly been converted to schists and phyllites. Some of these metamorphosed rocks contain garnet porphyroblasts, biotite and chlorite; other varieties are characterized by medium-grained chloritoid porphyroblasts. Whole rock geochemical studies suggest that the metamorphic assemblage present in a particular rock is in part a function of the rock's chemical composition. Chloritoid for example, occurs only in schists and phyllites with a high alumina content.

[1] The designation and descriptions of some of the type and typical sections of the various subdivisions of the Wernecke Supergroup are contained in Delaney, 1978.

[2] The terminology which is used in this paper to describe the thickness of beds and laminae is that of Ingram (1954) as adapted by Campbell (1967).

[3] The classification of both the terrigenous and carbonate rock types described in this paper is based on Folk (1974).

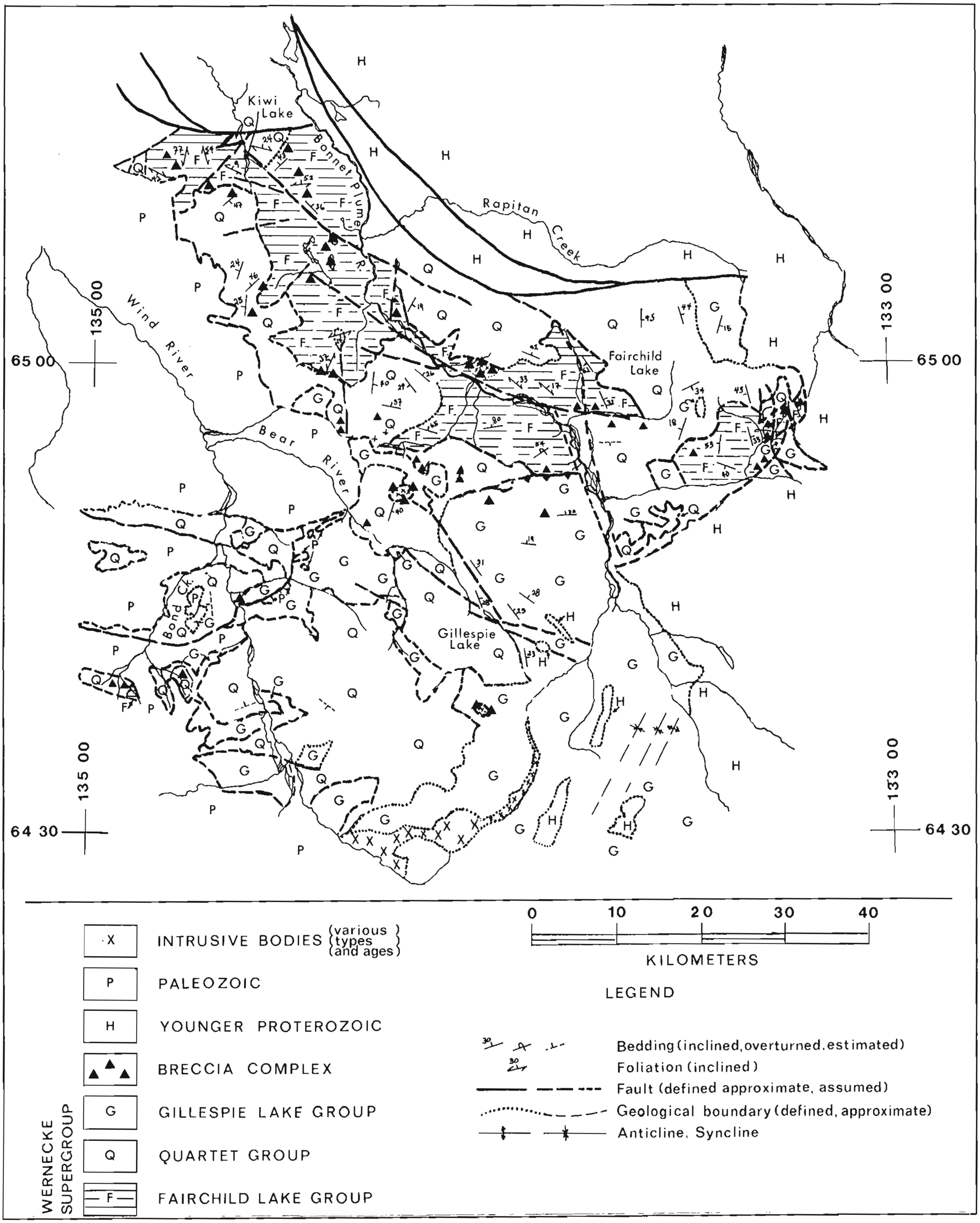

Figure 1.2. *Simplified geology of Wernecke Supergroup, northern Wernecke Mountains, Yukon Territory (modified after Delaney, in preparation).*

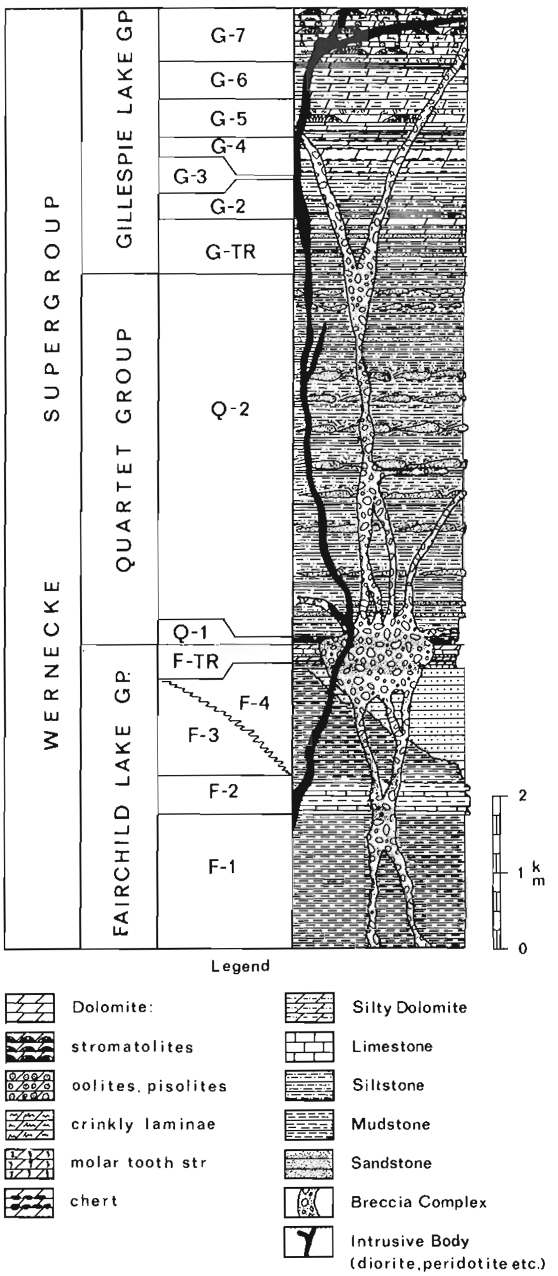

Figure 1.3. *Stratigraphic cross-section, Wernecke Supergroup.*

Figure 1.4. *Outcrop of a ribbed-weathering limestone-siltstone member of the F-2 of the Fairchild Lake Group. GSC 203062-V*

Figure 1.5. *Close-up view of ribbed-weathering limestone-siltstone member of the F-2 of the Fairchild Lake Group. Note that resistant weathering siltstone beds are characterized by either even-parallel laminations or lenticular crossbeds. The recessive weathering, silty limestone beds are typically cross laminated.*

Siltstone beds of the F-3 are composed of an equigranular mosaic of recrystallized medium to coarse silt-size grains of quartz and plagioclase with scattered grains of chlorite and sericite.

Although a unimodal south-southwesterly-trending dispersal pattern characterizes the F-3 in some exposures southwest of Fairchild Lake (Fig. 1.10), at other localities in this area there is a polymodal dispersal pattern.

The contact between the F-3 and the F-TR is conformable and may be in part transitional. The F-3 is probably a facies equivalent of the F-4 (Fig. 1.3).

F-4

The F-4 consists of at least 500 m of grey- to greenish grey-weathering siltstone, fine sandstone and mudstone. The coarse siltstone to very fine sandstone occurs in medium to thick, even, parallel-laminated to cross-laminated beds (Fig. 1.25). Thin to locally medium beds of dark grey- to greenish grey-weathering laminated to wavy bedded mudstone are interbedded within this siltstone-sandstone sequence. To the northeast of Fairchild Lake, the F-4 is commonly characterized by parallel, asymmetrical wave formed ripple marks.

The mudstone beds of the F-4 are composed of scattered grains of silt size quartz and plagioclase in a matrix of biotite, muscovite and sericite. Very thin laminations are defined by variations in mineral percentages.

Figure 1.6. *Overturned sequence of thin beds of mudstone and siltstone of the F-3 of the Fairchild Lake Group. Beds of even-parallel laminated mudstone alternate with lenticular beds of siltstone. Load structures are ubiquitous throughout the F-3.*

In some exposures the mudstones are characterized by a spotted alteration texture defined by spherical bodies of silica and, in some instances, chlorite.

The F-4 siltstone beds consist of an equigranular mosaic of quartz and minor plagioclase with biotite, chlorite and sericite ± calcite ± dolomite ± epidote ± margarite. In some instances laminations in the siltstone beds are defined by linear trains of opaque minerals; in others thin to medium laminae of siltstone alternate with laminae of mudstone.

The fine sandstone beds of the F-4 are composed of an equigranular mosaic of quartz and plagioclase with scattered grains of muscovite and in some instances ferroan dolomite and minor heavy minerals. Some F-4 sediments contain clots of brown amorphous carbonaceous dust.

At the northeast end of Fairchild Lake, crossbeds and ripple crest trends (Fig. 1.10) indicate that the F-4 sediments were deposited by southeasterly-flowing currents. In the Delores Creek area, this unit is characterized by a polymodal paleocurrent pattern (Fig. 1.10) typical of a tidally-influenced shallow marine environment (Klein, 1970).

The stratigraphic position of the F-4 suggests that it may be a facies equivalent of the F-3 (Fig. 1.3). The contact between the F-4 and the F-TR has not been observed. In the Delores Creek area, the F-4 appears to be unconformably overlain by unit Q-1 of the Quartet Group.

F-TR

The F-TR consists of up to 365 m of grey-, dark grey-, brown-, and locally white-weathering slate, mudstone, siltstone, dolomitic mudstone, silty dolomite and limestone. The lower part of the F-TR is a sequence of thin beds of grey- to dark grey-weathering slates and mudstones which contains members characterized by thin beds of grey to generally orange brown-weathering, even parallel laminated to cross-laminated silty dolomite (Fig. 1.7). Rusty-weathering slate beds commonly contain small cubic crystals of pyrite. The mudstone beds are characterized by thin to thick, rhythmically alternating, light grey and dark grey laminae. Crossbeds and molar tooth structure (O'Connor, 1972) occur in some silty dolomite beds.

Approximately two thirds of the way above the base of the F-TR there is a distinctive 7 to 14 m-thick white-weathering limestone member which constitutes one of the most important markers in the Wernecke Supergroup (Fig. 1.8). This member, which has gradational contacts, consists of thin to medium beds of generally coarsely crystalline calcite separated by dark grey carbonaceous laminations. A few scattered grey-weathering chert nodules are locally present.

Above the white limestone marker is a sequence of thin, commonly rust-stained, beds of grey- to dark grey-weathering, carbonaceous claystone and mudstone, which in its lower part is interbedded with thin beds of grey- to orange-brown-weathering silty dolomite, silty limestone and dolomitic or calcareous siltstone and claystone. The number and proportion of carbonate beds gradually decrease up section until the sequence consists entirely of dark grey, locally rust-stained carbonaceous claystone.

In the northern part of the study area, near Kiwi Lake, a larger proportion of the F-TR beds are carbonate. Almost all are dolomite. To the southeast, in the Fairchild Lake area, the thickness and number of carbonate beds decrease, and many of the beds are composed of calcite.

The white limestone marker is composed of a mosaic of coarse to very coarsely crystalline calcite, with minor silt size quartz, plagioclase, and scattered grains of muscovite and sericite, with some specks of carbonaceous dust. The beds of silty dolomite-dolomitic siltstone are composed of varying amounts of clastic dolomite, quartz, plagioclase, sericite, muscovite, chlorite, carbonaceous dust and heavy minerals. In some instances this lithology is characterized by a homogeneous mosaic of very finely to finely crystalline dolomite (senso-stricto) to ferroan dolomite with up to 30% silt size quartz, minor silt size plagioclase, and carbonaceous dust with scattered grains of sericite and muscovite. Most silty dolomite beds however, are laminated. Although most

Figure 1.7. *Even-parallel laminated and cross laminated silty dolomite beds of the F-TR of the Fairchild Lake Group.*

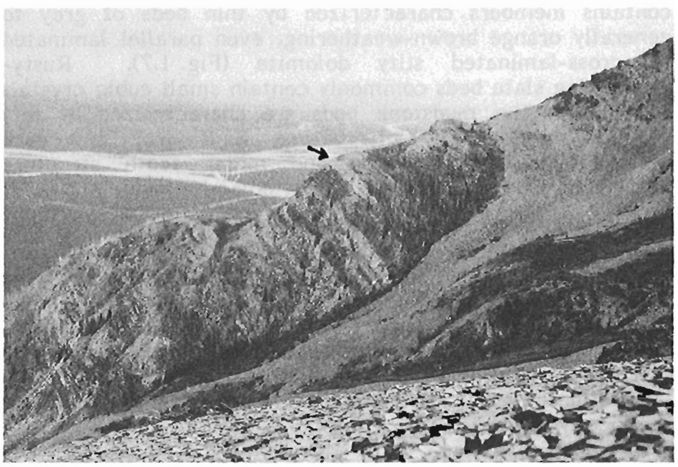

Figure 1.8. *Panoramic view of a portion of the F-TR in an exposure along the Bonnet Plume River. White-weathering limestone marker bed (10 m), which is offset by a fault, is in the centre of the photograph (arrow). The ridge below the saddle on the left side of the photograph is underlain by a breccia complex.*

Figure 1.9. *Unconformity in the Little Wind River area between near flat lying light grey Ordovician dolomites and steeply dipping strata of the F-TR of the Fairchild Lake Group.*

of the laminae are defined by minor to significant variations in the percentages and/or size of the constituent minerals, some are defined by planar trains of carbonaceous dust. Microscopic primary structures include graded bedding and syn-sedimentary folds; stylolites are a common secondary structure.

A pervasive closely-spaced fracture system dissects many of the outcrops of the F-TR, obliterating primary structures. Widespread development of breccia bodies also occurs at this stratigraphic level. At some localities the breccias completely encompass the F-TR.

The only paleocurrent data obtained for the F-TR were collected from a crossbedded silty dolomite member (Fig. 1.7) exposed on a hill east of Kiwi Lake. Although the rose diagram (Fig. 1.10) of these data exhibits some scatter, the main dispersal pattern appears to have been from east to west.

The contact between the F-TR and the Q-1 of the overlying Quartet Group is transitional. The contact is placed at the top of the uppermost orange brown-weathering dolomitic siltstone in the F-TR. At many places, however, the F-TR and strata of the Quartet Group are in fault contact.

Summary

The five formations of the Fairchild Lake Group comprise a thick sequence of fine grained siliciclastic sediments, with distinctive carbonate units near the middle and at the top of the group. Although most of the group was deposited in deeper water, by predominantly southerly flowing currents, the F-4 was deposited in a shallow marginal marine environment. The carbonate members in the F-2 and the F-TR are probably distal representatives of areas of carbonates which may have fringed the margin of the Wernecke Basin during deposition of parts of the Fairchild Lake Group. Facies changes within the F-TR, characterized by thicker dominantly dolomitic carbonate beds in the north and thinner dominantly calcitic beds in the south, suggest that the depositional margin of the Wernecke Basin may have lain north of the present day Wernecke Mountains. This association of limestone beds in the deeper parts of the basin and dolomite beds near the shallow margin is typical of some of the thick miogeoclinal terrigenous-carbonate successions of the Middle Proterozoic (e.g. the Belt-Purcell Supergroup; Smith and Barnes, 1966).

QUARTET GROUP

The Quartet Group consists of a monotonous succession, at least 5 km thick, of dark grey-weathering siltstone, fine sandstone, mudstone and claystone with minor intercalated silty dolomite. Locally these rocks have been transformed into slates, phyllites and schists.

There are extensive exposures of rocks of the Quartet Group throughout the Wernecke and Olgivie mountains and in the Keele Range of the Porcupine Plateau (Norris, personal communication, 1980; Fig. 1.1). A fault wedge of the Quartet Group outcrops in the Richardson Mountains (Norris, 1976).

The Quartet Group has been tentatively subdivided into two units of formational rank. From older to younger these have been informally designated Q-1 and Q-2.

Q-1

The Q-1 consists of about 200 m of dark grey-weathering, thin bedded, carbonaceous claystone and silty carbonaceous mudstone. The Q-1 is characterized by two sequences, each consisting of carbonaceous claystone overlain by clayey siltstone. The upper of these two sequences contains abundant pyrite and usually weathers a rusty colour. Generally, the Q-1 strata have been metamorphosed to slate.

The contact between the Q-1 and the conformably overlying Q-2 is marked by the first appearance of even parallel laminated mudstone beds.

Q-2

The Q-2 consists of up to 5000 m of dark grey- to grey-weathering siltstone, mudstone, fine sandstone and claystone (Fig. 1.11). The lower several hundred metres of the Q-2 is rhythmically alternating, even parallel laminated to thin bedded, light grey-weathering siltstone and dark grey-weathering mudstone.

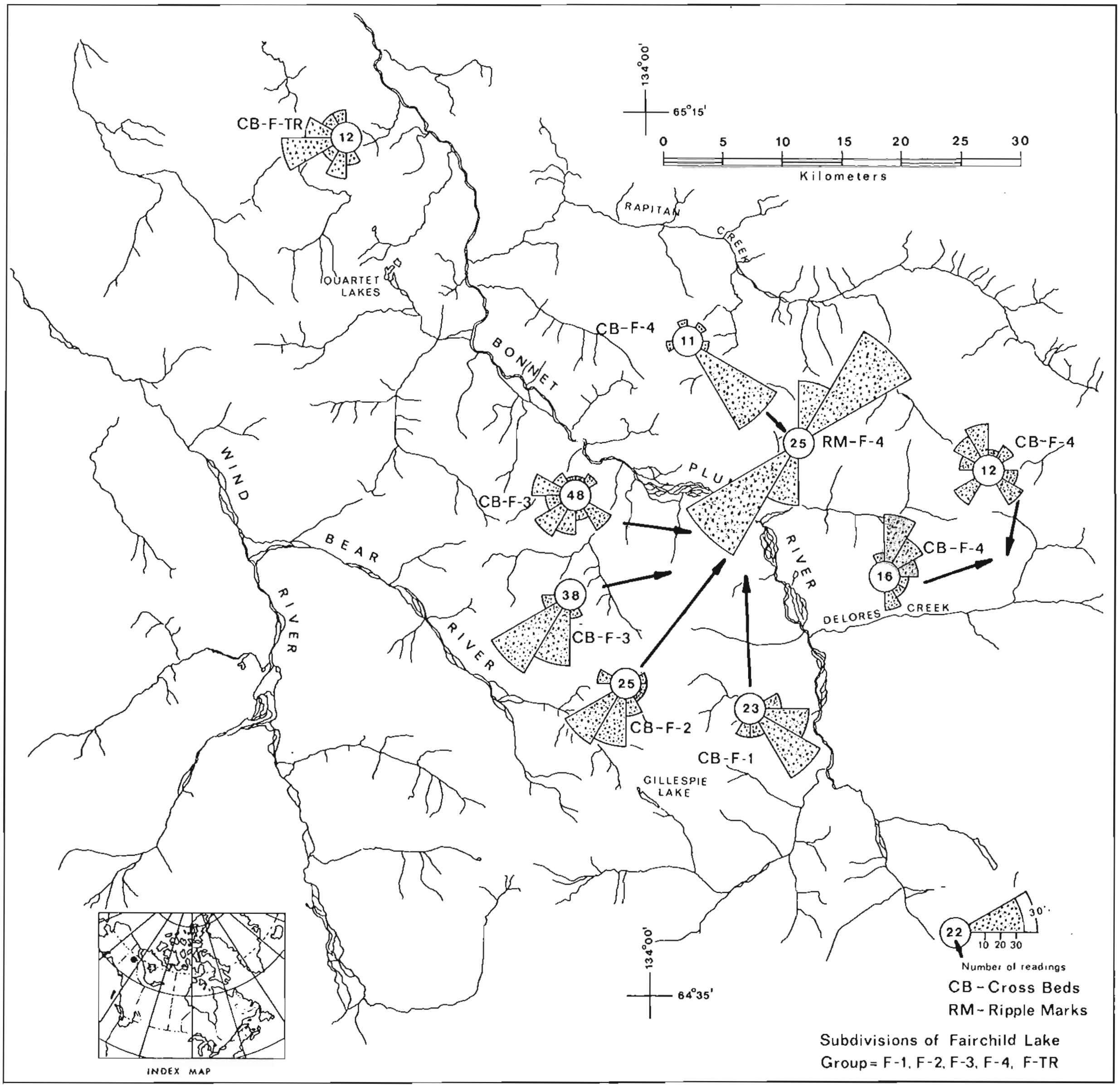

Figure 1.10. *Paleocurrent data from the Fairchild Lake Group of the northern Wernecke Mountains.*

Above this lower succession, the Q-2 is characterized by medium to thick beds of wavy, lenticular and flaser-bedded siltstone (Fig. 1.12, 1.13); thin, medium and locally thick beds of cross-laminated coarse siltstone to very fine sandstone; medium to thick beds of "chaotic" carbonaceous siltstone and laminae to thin beds of mudstone and claystone.

Both hexagonal and rectangular varieties of interference ripples (Fig. 1.15; Bucher, 1919) are a ubiquitous feature throughout the Q-2. These structures are best developed in the wavy, flaser and lenticular-bedded sequences. Other common primary sedimentary structures in the Q-2 include both asymmetrical and symmetrical ripple marks, load structures produced by piled, loaded ripples, flame structures, and ball and pillow structures. Shrinkage cracks (Fig. 1.14) are abundant throughout the Q-2, particularly in the lenticular bedded sequences. Although a few of these structures are obviously desiccation cracks, most are interpreted to have formed by synaeresis (White, 1961; Burst, 1965). Beds of coarse siltstone-fine sandstone commonly pinch and swell and terminate in recumbently folded masses (Fig. 1.16), interpreted as slump structures. Wavy, lenticular and flaser bedded sequences within the Q-2 are texturally inhomogeneous, compositionally variable rocks characterized by very thin laminations. These consist of various amounts of quartz, plagioclase, micaceous minerals and carbonaceous dust and debris. The framework is variable proportions of fine silt and very fine sand. Recrystallization has commonly masked the original microscopic textures. Microscopic primary structures include graded bedding, laminations, crossbedding, load structures, scour surfaces and shrinkage cracks.

The medium to thick, commonly slump-folded, beds of coarse siltstone-to-very fine sandstone are submature to mature arkose.

Figure 1.11. *Panoramic view of the Q-2 looking east near the headwaters of a tributary of Rapitan Creek. Note the monotonous dark grey-weathering colour of the Q-2.*

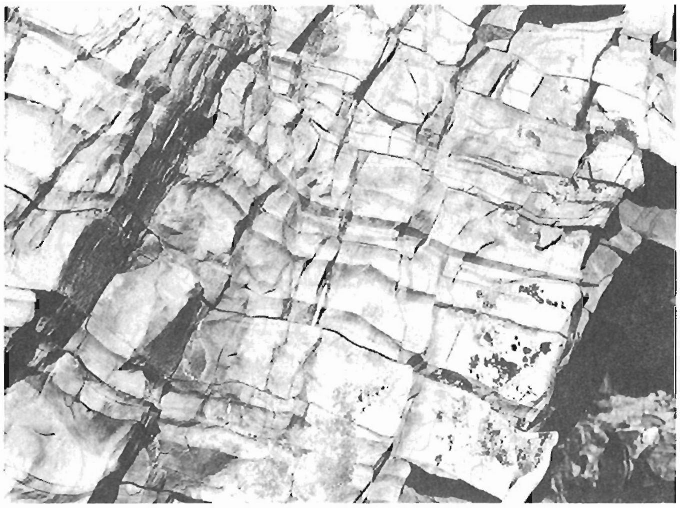

Figure 1.13. *Wavy-bedded siltstone-claystone in the Q-2 of the Quartet Group. Note grading in some beds; match stick (centre of photo) is about 40 mm long.*

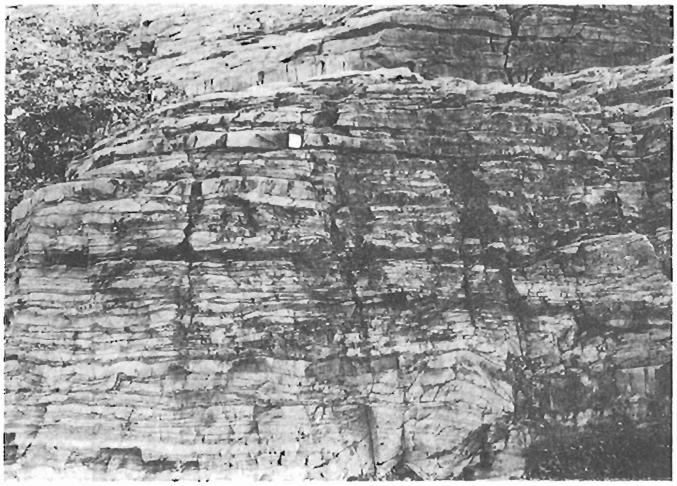

Figure 1.12. *Sequence of wavy to lenticular bedded siltstone of the Q-2 of the Quartet Group. The tape measure is 50 mm wide.*

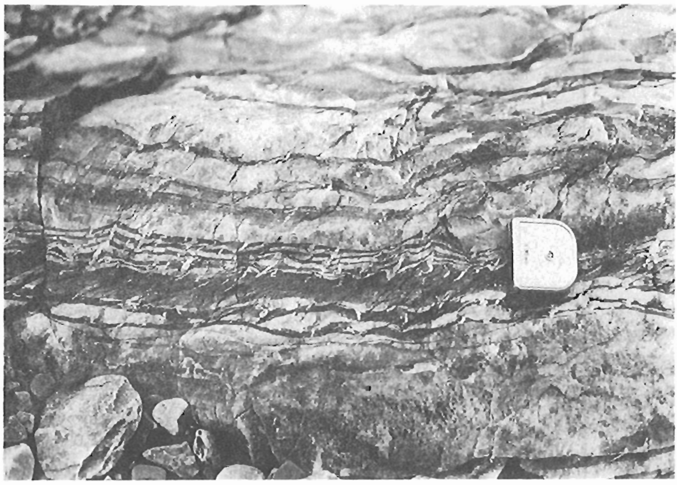

Figure 1.14. *Synaeresis cracks in dark grey laminated mudstone bed of the Q-2 of the Quartet Group. The tape measure case is 50 mm wide.*

The chaotic carbonaceous mudstone-to-siltstone beds of the Q-2 are characterized by thin, parallel-aligned, wispy to bulbous lenses of coarse silt-to-fine sand size quartz and plagioclase, forming an intact-to-disrupted framework in a matrix of micaceous minerals. Locally distinctive pebble-size lithic fragments are present in these rocks.

Cleavage, which commonly intersects bedding at a low angle, is ubiquitous in the Q-2. The Q-2 is commonly dissected by normal, reverse, and thrust faults which locally cause repetition of the succession. The faulting, coupled with the monotonous character of the Q-2, greatly complicates stratigraphic reconstruction.

The Q-2 is characterized by bimodal or polymodal paleocurrent dispersal patterns (Fig. 1.17). For the bimodal dispersal patterns, the modes are either 180° or 90° apart. Although some of the polymodal distribution patterns are characterized by modes at 90°, in most cases the pattern is much more complex.

The contact between the Q-2 and the overlying G-TR of the Gillespie Lake Group is transitional. The contact is marked by the first appearance of a bed of orange-weathering dolomitic siltstone.

Summary

The homogenous, pyritic, carbonaceous claystone and mudstone at the base of the Quartet Group (Q-1) probably accumulated in a sediment-starved basin. A gradually increasing influx of sediment into this basin is recorded by the siltstone-mudstone rhythmites at the base of the Q-2. The thick fine grained dark grey-weathering siliciclastic sediments comprising the bulk of the Q-2 are characterized by facies associations and paleocurrent dispersal patterns typical of shallow marine environments.

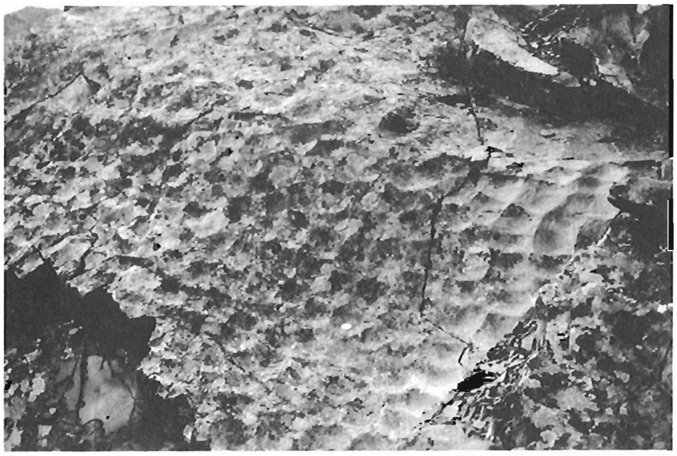

Figure 1.15. *Interference ripple marks on a siltstone bed of the Q-2 of the Quartet Group. Coin near bottom of photo is 18 mm in diameter. GSC 203062-T*

Figure 1.16. *Bulbous termination of a slump-folded bed of coarse siltstone in the Q-2 of the Quartet Group.*

Bimodal, bipolar paleocurrent dispersal patterns are interpreted to reflect tidal currents (Klein, 1967). Bimodal and polymodal dispersal patterns with modal classes 90° apart or in more complex distribution patterns characterize a shallow marine, tidal flat environment (Tanner, 1955; Klein, 1967; Picard and High, 1968). This regime, is subjected to the combined effects of oceanic circulation currents, tidal currents, meteorological currents and density currents (Johnson, 1978).

In addition to the paleocurrent dispersal trends, other factors supporting a subtidal to intertidal origin for the Q-2 are the wavy, lenticular and flaser bedded facies (Reineck and Wunderlich, 1968; Reineck and Singh, 1973) and the presence of interference ripple marks. The slumped siltstone beds of the Q-2 (Fig. 1.16) may be related to locally unstable slopes, perhaps created by rapid sediment deposition. An initial analysis of the orientation of these slump structures indicates that they formed on a southeasterly-trending paleoslope.

GILLESPIE LAKE GROUP

In the Wernecke Mountains, the Gillespie Lake Group consists of at least 4 km of buff-, orange-, and locally grey- or maroon-weathering dolomite, clayey, silty or sandy dolomite, limestone, dolomitic claystone, mudstone and sandstone, and claystone, mudstone and sandstone. Strata of the Gillespie Lake Group are widely exposed in both the Wernecke and Olgivie mountains (Fig. 1.1).

At the type locality the Gillespie Lake Group has been tentatively subdivided into 7 units of probable formational status (Fig. 1.3). These subdivisions have been informally designated G-TR, G-2, G-3, G-4, G-5, G-6, and G-7 in ascending order. At present, some of these subdivisions probably cannot be traced outside the type area because of dramatic facies changes and structural complications. The nature of these facies relationships will only be resolved following further detailed stratigraphic studies in the region surrounding Gillespie Lake.

G-TR

The contact between Quartet Group and Gillespie Lake Group is transitional. The lowermost subdivision of the Gillespie Lake Group, the G-TR, consists of up to 700 m of grey-weathering siltstone, fine sandstone and mudstone, and orange-to-locally brown-weathering silty dolomite and dolomitic siltstone (Fig. 1.3). The thickness of this unit is highly variable, ranging from a maximum of 700 m to as little as 25 m. Locally the G-TR is dissected by faults which cut out or repeat part of the section.

The lower part of the G-TR is characterized by the same facies components as the Q-2 of the Quartet Group. Scattered throughout the lower part of the G-TR are a few orange-to-brown, slightly recessive weathering, thin, cross-laminated beds and laminae of dolomitic siltstone to silty dolomite. In some cases, where these dolomitic beds are traced along strike, their distinctive orange-weathering colouration pinches out and disappears only to reappear farther along strike. The number and proportion of these dolomitic beds gradually increase up section so that in its upper part this unit assumes a distinctive striped weathering colouration. Stripes consist of alternating grey-weathering beds of wavy bedded siltstone and fine sandstone and orange-weathering beds of silty dolomite. The G-TR contains the same sedimentary structures as unit Q-2 of the Quartet Group.

Microscopic examination of siltstone, sandstone and mudstone beds of the G-TR reveals similar composition to those from the Q-2. The composition of the recessive-weathering dolomite beds and laminae of the G-TR is also highly variable. Most are characterized by thin laminae and lenticles of coarse silt-size quartz and plagioclase which alternate with laminae of finely crystalline ferroan dolomite containing medium to coarse silt size grains of quartz, minor plagioclase and micaceous minerals. The calcareous siltstone beds of G-TR are composed of a framework of coarse silt size quartz and plagioclase with scattered flakes of muscovite in medium crystalline calcite.

The presence of pockets of primary limestone, together with the crystalline nature of the dolomite, suggests that the entire unit has been dolomitized. Very weak regional metamorphism is indicated by the occurrence of secondary muscovite and chlorite.

Paleocurrent patterns derived from crossbeds and ripple marks are commonly unimodal (Fig. 1.18), with transport to the south or southeast. Unimodal trends appear to predominate in the upper part of the G-TR. In the lower part

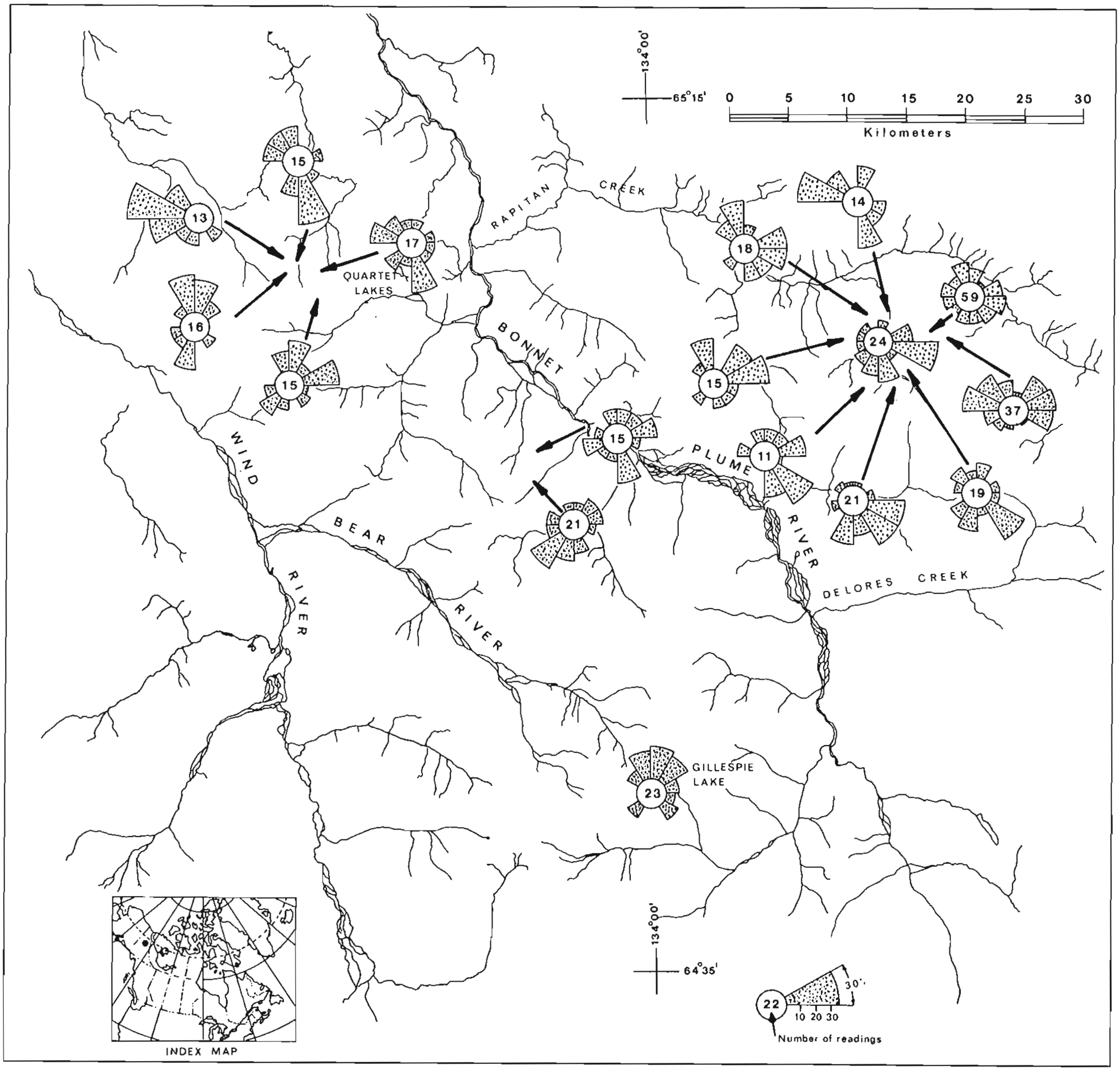

Figure 1.17. *Paleocurrent data from crossbedding in unit Q-2 of the Quartet Group of the northern Wernecke Mountains.*

of the G-TR, however, bimodal or polymodal patterns similar to those found in the Q-2 are more common. These complex paleocurrent dispersal patterns and the various facies present, suggest that the lower part of the G-TR was deposited in a shallow marine environment.

G-2

Conformably above the G-TR, the unit called G-2 consists of 400 to 600 m of tan-, brownish grey-, and orange-weathering dolomitic siltstone, dolomitic mudstone, siltstone and silty dolomite. The lower part is mostly thin lenticular beds of dolomitic siltstone, but the greater part is a homogeneous sequence of brownish grey-weathering, slightly dolomitic mudstone with some lenticular beds of siltstone. Locally, the dolomitic mudstones weather to a distinctive greenish grey. In the upper part of the G-2 a few of the beds contain granule size, elliptically-shaped, lenses of black chert. Load structures are locally developed.

The dolomitic mudstone beds of the G-2 are composed of scattered, finely crystalline dolomite to ferroan dolomite, silt size quartz, micaceous minerals (mostly sericite and chlorite with scattered grains of biotite) and carbonaceous dust. Laminations in this lithology are typically defined by either variations in constituent components, or by linear trains of carbonaceous dust.

The silty crystalline dolomite is composed of a framework of aphanocrystalline to very finely crystalline dolomite to ferroan dolomite with silt size grains of quartz, minor plagioclase, sericite, muscovite, chlorite, carbonaceous dust (usually in thin laminae and lenses), pyrite (as isolated octahedra), iron-free calcite and elliptical lenses of chert. Many dolomitic siltstone beds consist of even parallel, very thin laminae of various proportions of silt size quartz and plagioclase, sericite, crystalline ferroan dolomite and carbonaceous dust.

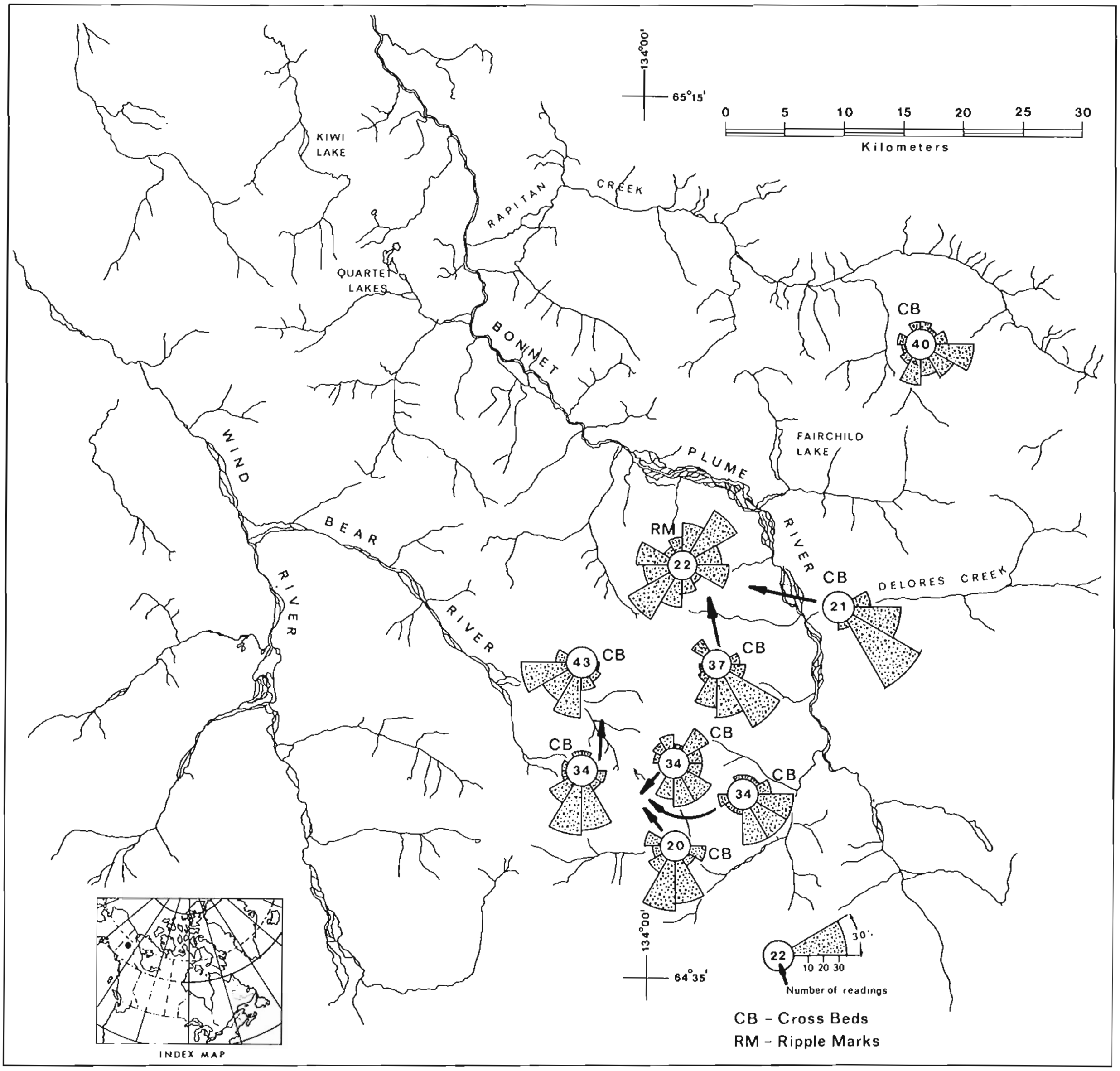

Figure 1.18. *Paleocurrent data from unit G-TR of the Gillespie Lake Group of the northern Wernecke Mountains.*

Paleocurrents show that the sediments of G-2 were deposited by southerly-flowing currents (Fig. 1.26).

G-3

The G-3, which conformably overlies the G-2, consists of medium to thick beds of buff-weathering silty dolomite. Generally, these beds are characterized by even parallel laminations. Some beds have a distinctive ribbed-weathering appearance, resulting from alternation of recessive-weathering laminae of silty dolomite and more resistant laminae of dolomitic siltstone. Some of the ribbed-weathering units contain lenses of elongate intraformational fragments aligned perpendicular to bedding to form tent-like structures. No paleocurrent data were collected from the G-3.

Although a fault contact separates G-3 and G-4 in the type area, it appears that they are conformable.

G-4

The G-4 (Fig. 1.3) consists of about 450 m of brown-, buff-, grey- and locally orange- or maroon-, recessive-weathering silty dolomite, dolomite, dolomitic mudstone, dolomitic siltstone and chert. It is characterized by thin beds of even parallel laminated to locally wavy laminated dolomitic siltstone with discontinuous lenses and thin beds of dark grey chert, which constitutes up to 20 per cent of the section.

The middle subdivision of the G-4 consists of about 125 m of buff-, grey-, and locally orange-weathering silty dolomite and dolomite with minor dolomitic siltstone, dolomitic mudstone and chert. The basal part of this subdivision is medium to thick beds of buff- and grey-buff-weathering, generally even parallel laminated silty to clayey dolomite. This is overlain by cyclically thickening-upward, medium to thick beds of silty dolomite. The upper part of

Figure 1.19. *Inclined columnar stromatolites in the G-7 of the Gillespie Lake Group. Note reversal in direction of inclination of columns. Bedding parallels base of photograph. Each division of pole is 10 cm.*

the middle subdivision is thin- to medium-bedded dolomite and silty dolomite. The middle subdivision of the G-4 is characterized by a paucity of chert compared to the upper and lower subdivisions.

The 100 m thick upper subdivision consists of thin beds of buff-weathering silty dolomite to dolomite. Throughout this succession are thin beds, discontinuous lenses and irregular-shaped masses of grey and dark grey chert. Concretions are locally present in the upper part. Near the top is a sequence of thin to medium beds of silty dolomite with interbeds of dolomitic claystone.

Although the contact between the G-4 and the G-5 is covered at the type locality, it appears conformable.

G-5

A threefold subdivision of this unit was also recognized. The lowest 150 to 175 m consists of buff- and locally maroon-weathering, even parallel laminated and cross-laminated thin to medium beds of silty dolomite with thin interbeds of grey- and buff-weathering dolomitic claystone.

The middle subdivision of the G-5 is a sequence of buff-weathering, thin to locally medium beds of even parallel laminated and ripple cross-laminated dolomite and silty dolomite which alternate in a regular fashion with grey-weathering laminae and thin beds of dolomitic claystone. Interbedded within this sequence are buff-, grey-, and pinkish grey-weathering, thin to thick beds of stromatolitic dolomite pebble conglomerate. Locally some intact columnar and columnar branching stromatolites occur within these conglomerate beds. Thin beds of flat chip conglomerate are also associated with these beds of algal debris.

The lower part of the middle subdivision of the G-5 contains isolated breccia mounds up to 3 m high. These consist of angular clasts of stromatolitic dolomite with a few isolated intact columnar stromatolites. There are a few beds of inclined columnar stromatolites in which the stromatolites exhibit alternations in the direction of their inclination (Fig. 1.19). The morphology of these stromatolites probably reflects alternations in current directions.

The upper 100 m subdivision is a sequence of thin beds of buff- and grey-, locally orange-weathering, wavy and flaser-bedded, even parallel-laminated and cross laminated dolomite together with silty and clayey dolomite and minor dolomitic siltstone and mudstone. Large parts of this unit weather with a distinctive striped colouration. This is defined by alternations of laminated and bedded slightly silty, buff-weathering dolomite with laminated and thinly bedded grey- to dark-grey-weathering carbonaceous dolomite.

Various types of conglomerate in this sequence include subangular to subrounded granule- to pebble-size dolomite fragments in a disrupted matrix, pebble-size flat chips of dolomite in lenses and angular pebble- to cobble-size dolomite fragments. Pods and lenses of dark grey chert are also present. slump structures are common in the upper part.

The total thickness and stratigraphic characteristics of portions of the upper part of the G-5 are unknown as it commonly occupies steep inaccessible terrain in the type area.

Paleocurrent measurements were obtained from only two localities in the G-5 (Fig. 1.26). Although a polymodal dispersal pattern characterizes each of these sites, in both cases there is a strong south trending mode.

G-6

The G-6, which conformably overlies the G-5, is a 500- to 800 m-thick sequence of buff-, grey-, green-, and maroon-weathering silty dolomite, silty limestone, dolomitic mudstone and dolomitic siltstone.

This unit is dissected by several faults of unknown displacement. This factor, coupled with its monotonous character, make the thickness figures for the unit only estimates.

The lower part of the G-6 consists of thin to medium beds of buff-weathering, even parallel-laminated, locally cross-laminated silty dolomite and dolomitic mudstone. Intercalated within this sequence are some thin to medium beds of grey weathering clayey carbonaceous dolomite. This succession is overlain by medium and thick beds of buff-weathering silty dolomite which vary from massive and homogeneous to interbedded, cross-laminated silty dolomite and fine grained silty dolomite. This unit weathers in a distinctive ribbed fashion. Load structures are abundant in the G-6.

About 200 m above the base of the G-6 are greenish grey-weathering, even parallel-laminated beds of very finely crystalline limestone. The weathering colour of this part of the sequence ranges from green to maroon. The maroon units locally contain lenses of breccia and conglomerate of granule- to pebble-size fragments of silty hematitic limestone. These conglomerate lenses, commonly coated with an irregular apple-green malachite stain, are cut by very thin calcite veins containing pyrite and chalcopyrite. These distinctive drab green- and maroon-weathering silty limestone beds occur within a sequence of buff-weathering and rhythmically alternating thin beds of dolomitic siltstone and silty dolomite which weather in a distinctive ribbed fashion.

The upper 300 to 400 m of the G-6 consists of grey- and buff-weathering lenticular and wavy-bedded silty dolomite and silty limestone. In general, this sequence is characterized by beds of cross-laminated silty dolomite and silty limestone which alternate with thick laminae of clayey silty dolomite.

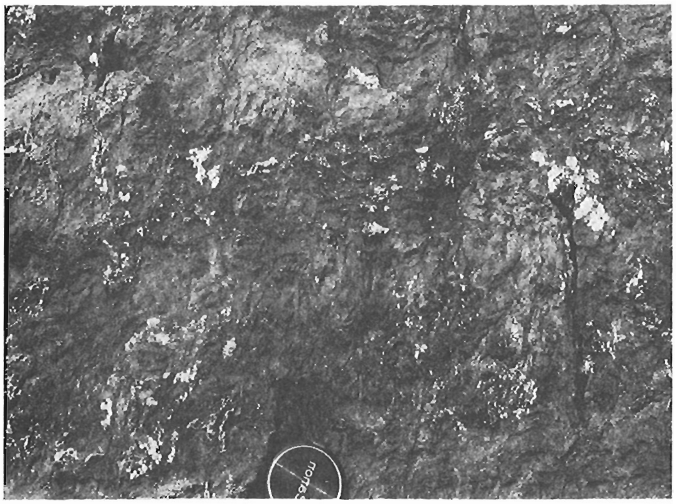

Figure 1.20. Bed of crinkly-laminated algal dolomite in the G-7 of the Gillespie Lake Group.

Figure 1.21. Large domal stromatolite in the G-7 of the Gillespie Lake Group.

The beds of even parallel-laminated silty dolomite and dolomitic mudstone are composed of quartz, micaceous minerals, finely crystalline dolomite, ferroan dolomite, and carbonaceous dust. This lithology is characterized by two varieties of laminations: those defined by parallel discontinuous carbonaceous streaks and those defined by variations in the percentages of the constituent minerals.

The distinctive ribbed-weathering, silty dolomite beds in the G-6 are composed of alternating thin beds and thick laminae of finely crystalline ferroan dolomite and thick laminae to thin beds composed of a matrix of aphanocrystalline to very finely crystalline dolomite containing scattered silt-size quartz grains, minor muscovite and parallel planar trains of carbonaceous dust. Microstylolites are preserved in many of these rocks.

Although bimodal paleocurrent patterns characterize the G-6, in each case the dominant mode trends south (Fig. 1.26).

G-7

Conformably overlying the G-6 is the G-7, a 400- to 700 m-thick sequence of orange-, buff-, and grey-, locally pink-weathering carbonates. The main components of this formation are thin, generally medium and occasionally thick beds of crinkly laminated dolomite (Fig. 1.20), oolitic and pisolitic dolomite, stromatolitic dolomite, clayey carbonaceous dolomite, and even parallel-laminated to wavy and lenticular bedded dolomite. Molar-tooth structure (O'Connor, 1972; Smith, 1968) is a ubiquitous feature.

Cryptalgal crinkly-laminated dolomite beds commonly pass laterally into hemispheroidal stromatolite mounds (Fig. 1.21). The stromatolitic facies of the G-7 is characterized by a wide variety of forms (Fig. 1.19, 1.21), including oncolites. Individual dolomite beds of the G-7 locally include patches of light grey limestone. Lenses and nodules of grey and dark grey chert are present throughout the G-7, but appear to be preferentially developed at the tops of pisolite and oolite beds.

Most of the G-7 is composed of crystalline dolomite, with minor amounts of silt size quartz, muscovite, sericite, chlorite and carbonaceous dust.

Polymodal paleocurrent patterns are typical (Fig. 1.26).

Summary

The Gillespie Lake Group is a 4 km thick sequence of buff-, grey-, and orange-weathering, fine grained terrigenous and carbonate sediments. Although seven subdivisions of formational rank are defined in the type area, structural complications and dramatic facies changes hinder correlation with surrounding areas.

The basal part (G-TR) of the Gillespie Lake Group is shallow marine siliclastic sediments with a gradual increase in the proportion of dolomite occurring up sequence. These silty dolomite beds are characterized by southerly directed paleocurrent dispersal patterns and reflect the initiation of a major inundation of carbonate detritus from a source to the north.

The lower formations of the Gillespie Lake Group (G-2 to G-4) are fine grained siliclastic-dolomite admixtures. Unimodal southerly directed paleocurrent dispersal patterns characterize the G-2. Distinctive cherty units occur in parts of the G-4. These formations are interpreted to reflect the more distal facies of a major area of carbonate buildup which lay to the north. This basin also received a reduced but still significant input of siliciclastic sediments.

The G-5 is also fine grained dolomitic-siliclastic admixtures, but contains beds of stromatolitic conglomerate and breccia as well as beds of flat chip conglomerate. Again southerly directed paleocurrent dispersal patterns predominate. The G-5 probably reflects a shallowing of the depositional environment.

The G-6 is a thick sequence of dolomite, limestone and terrigenous-carbonate admixtures. In addition to the typical buff- and grey-weathering colouration of the Gillespie Lake Group, this formation has distinctive green- and maroon-weathering zones which locally contain copper mineralization. Southerly directed paleocurrent dispersal patterns typify the G-6.

At the type area, the uppermost part of the Gillespie Lake Group (G-7) is shallow marine platformal carbonates characterized by beds of crinkly laminated dolomite, oolitic and pisolitic dolomite, stromatolitic dolomite and wavy and lenticular beds of dolomite. Polymodal paleocurrent patterns are typical.

The presence of patches of primary limestone throughout the Gillespie Lake Group indicate that much of this group was probably originally limestone.

BRECCIA COMPLEXES

Strata of the Wernecke Supergroup are cut by numerous breccia complexes which locally are significantly enriched in U, Cu, Ba, Co, or Fe (Morin, 1976; Bell and Delaney, 1977; Bell, 1978; Archer et al., 1977; Archer and Schmidt, 1977; Laznicka, 1977a, b, 1978; Laznicka and Edwards, 1979; Delaney et al., in press).

These breccia complexes are characterized by both variable shape and size. In plan view, individual breccia bodies are circular, elliptical or have an elongate, commonly irregular, outline. In vertical section, they are clearly discordant, pinch and swell (Fig. 1.25), and are characterized by numerous offshoots. Some breccia apophyses are parallel to bedding; these stratiform breccias were previously mistaken for volcanic breccias (c.f. Bell and Delaney, 1977). The widest breccia bodies occur in units F-3, F-4 and F-TR of the Fairchild Lake Group (Fig. 1.3).

The contacts between breccias and surrounding country rock are variable. They range from those characterized by sharp margins, with upward dragging of host beds, to those in which the degree of brecciation gradually decreases outward, grading into weakly-fractured country rock. The variable nature of these boundaries is related to the complex genetic origin of the breccia bodies (see below). Nowhere have the breccias been observed cutting strata which overlie the Wernecke Supergroup.

The breccia bodies exhibit a considerable range in size. For example, the large breccia complex in F-3, F-4 and F-TR of the Slab Mountain area is estimated to be about 4 km by 0.5 km (Bell and Delaney, 1977), whereas some breccias are only a few metres wide. The average dimension of these bodies ranges between 100 and 800 m (Archer and Schmidt, 1977). It should be noted, though, that the perceived size variations are in part a function of the level of stratigraphic exposure.

Most breccia complexes are developed on, or in close proximity to, faults (Fig. 1.2). In particular, splays of west- to northwesterly-trending faults of the Richardson Fault Array (Norris and Hopkins, 1977) appear to have served as loci for the emplacement of these bodies.

Three genetically distinct varieties of breccia can be differentiated (Delaney et al., in press). These are: fault breccia, stope breccia and channelway breccia. Fault breccia typically contains pebble and cobble size material or larger fragments. These fragments occur in a granulated rock matrix. This variety of breccia usually occupies a linear zone. The contacts between fault breccias and the surrounding country rock are usually diffuse, and characterized by a gradual decrease in both the number of fractures and the degree of rotation of clasts. The stope breccias (Fig. 1.24) consist of varied sized, generally angular fragments (maximum 30 m) which occur either as isolated blocks in a granulated matrix or as an irregular, jumbled-intact framework. The margins of this variety of breccia are usually sharp and locally the contact with the host lithology is steeply inclined. The channelway breccias are thin (1-5 m) pipe-shaped bodies which are commonly at a high angle to the horizontal. They are characterized by sharp margins, contain well-rounded pebble-size clasts, and locally have flow structures preserved in their matrix.

A variety of types of alteration are associated with the breccia complexes including Na-feldspathization, silicification, hematitization and carbonatization (Bell and Delaney, 1977; Laznicka and Edwards, 1979; Delaney et al., in press). Locally, phases of this alteration have permeated outward from the breccias into the surrounding country rock. The intensity of alteration can be quite variable. Near the headwaters of Delores Creek for example sodic metasomatism has been so pervasive as to create medium to coarsely crystalline albitic "syenites" (Laznicka and Edwards, 1979). These bodies were originally interpreted to be porphyry copper syenites.

A metamorphic aureole is commonly developed around the breccia bodies, particularly at lower stratigraphic levels. In these aureoles, the fine grained sediments of the Quartet and Fairchild Lake groups have been transformed into light greenish grey phyllites and schists which commonly contain porphyroblasts of chloritoid or garnet.

A review of the various types of mineralization associated with the breccia complexes is beyond the scope of this paper (see Morin, 1976; Archer and Schmidt, 1977; Bell and Delaney, 1977; Bell, 1978; Laznicka and Edwards, 1979; Delaney et al., in press). It is interesting to note, however, that all known occurrences of uranium are confined to an interval near the contact of the Fairchild Lake and Quartet groups.

Bell (1978) and Delaney et al. (in press) have presented models for the genesis of the Wernecke breccia bodies. The structures are believed to have developed in a dilatational tectonic regime (Sawkins, 1976; Gabelman, 1977a, b). The initiation of breccia development is related to activity of the Richardson Fault Array (Norris and Hopkins, 1977), a series of deep-seated, long active structures.

Breccia development related to fault activity was probably followed by hydraulic stoping (Gilmour, 1977). Finally, gas-charged fluids in small pipes dissected the fault-stope breccia complexes. Each phase in the development of the breccia complexes was probably multistaged, and commonly overlapped other stages. Mineralizing fluids associated with the breccia bodies were probably derived from a variety of sources including those tapped from the

Figure 1.22. *Panorama near Algae Mountain of diorite dyke (25 m wide) cross-cutting a sequence of dolomites of the Gillespie Lake Group. Note the white dedolomitization alteration zone developed around the dyke.*

mantle (Gabelman, 1977b), those associated with intrusive bodies (Laznicka and Edwards, 1979) and those enriched by leaching of sediments of the Wernecke Supergroup (Bell, 1978).

INTRUSIVE BODIES

Green- to brown-weathering dykes and sills of diorite, gabbro, and at least one body of peridotite, intrude strata of the Wernecke Supergroup (Green, 1972). In the northern part of the study area, these rare bodies are generally 1 to 3 m wide. To the south, in the Rackla Range, both the size and number of dykes and sills increase. Where dykes cross-cut Fairchild Lake and Quartet groups strata, their presence is commonly masked by a similar weathering. This subdued expression contrasts with the spectacular contact aureoles developed around these bodies where they intrude strata of the Gillespie Lake Group (Fig. 1.22).

Stratigraphic relationships suggest that the intrusive bodies may be of several ages. Fragments of diorite occur locally in the breccia complexes; at other places however, dykes intrude the breccias. Some of these bodies are cut off at the unconformity separating strata of the Wernecke Supergroup from younger rocks, whereas others penetrate this boundary. Green (1972) suggested that some dykes may be as young as Cretaceous.

STRUCTURAL OVERVIEW

Structural relationships of rocks of the Wernecke Supergroup are complex and not yet completely understood. These rocks were folded, faulted and cut by intrusive breccia complexes prior to deposition of the Proterozoic Pinguicula Group (Eisbacher, 1978).

In a regional sense, the Wernecke Supergroup appears to have been deformed into open folds whose axes trend northwest. In the southeastern part of the map area however, strata of the Gillespie Lake Group are deformed into tight northeasterly-trending folds. This simple structural style is not present everywhere; locally rocks of all three groups are overturned.

The Wernecke Supergroup in the northern Wernecke Mountains is dissected by a group of west- to northwest-trending vertical curvilinear faults (Fig. 1.1, 1.2) which represent the southern extension of the Richardson Fault Array (Norris and Hopkins, 1977). This system of faults was interpreted (ibid.) as deep-seated, long-lived structures which have undergone both vertical and lateral movements. The large anticline-like structure which parallels the Bonnet Plume River may be related to differential vertical movements along the Richardson Fault Array (Fig. 1.2).

In addition to the Richardson Fault Array and its related splays, strata of the Wernecke Supergroup are cut by numerous steep-to-shallow normal and reverse faults with varying amounts of offset. These structures hinder stratigraphic reconstruction.

Besides the effects of folding and faulting, distinct structural aberrations are also encountered in the vicinity of the breccia complexes. Near these bodies the surrounding country rocks are contorted, fractured, metamorphosed and locally metasomatized.

The northern boundary of the Wernecke Supergroup is marked by the Knorr Fault (of the Richardson Fault Array), a steeply dipping normal fault which juxtaposes younger Proterozoic rocks against strata of the Quartet Group (Fig. 1.2). In the western part of the study area, Lower Paleozoic rocks unconformably overlie the Wernecke

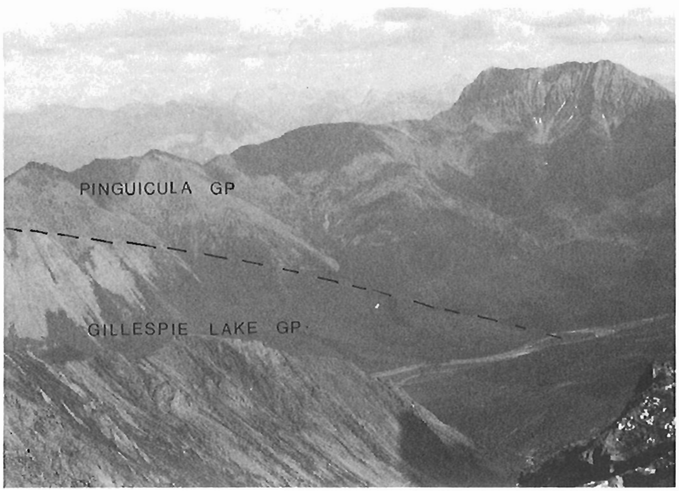

Figure 1.23. *Panoramic view to east from near Algae Mountain of contact between Gillespie Lake Group and Pinguicula Group. This site is near the type locality of the Racklan Orogen.*

Supergroup commonly forming spectacular angular unconformities (Fig. 1.9). To the east, strata of the Wernecke Supergroup are unconformably overlain by those of either the Pinguicula Group (Fig. 1.23) or those of the Ekwi Supergroup.

Regional Tectonics

The deformational events affecting strata of the Wernecke Supergroup prior to the deposition of rocks of sequence B are attributed to the Racklan orogeny (Gabrielse, 1967; Eisbacher, 1978; Yeo et al., 1978; Young et al., 1979). These events, which probably culminated about 1.2 to 1.3 Ga ago, included folding, faulting (both vertical and lateral movements along the Richardson Fault Array) and probably the development of the breccia complexes.

Recently it has been recognized (Eisbacher, 1978; Yeo et al., 1978; Young et al., 1979) that deformation attributed to the Racklan orogeny (Gabrielse, 1967) occurred much earlier (1.2 vs 0.8 Ga) than events in the Mackenzie Mountains to which it was formerly equated. The younger Proterozoic deformational event which affected the Mackenzie Mountains Supergroup is now referred to as the Hayhook orogeny (Young et al., 1979) and it is considered to be the temporal equivalent of the enigmatic East Kootenay orogeny of the southern Cordillera (White, 1959).

GEOCHRONOLOGY

Few geochronometric data are available for the Wernecke Supergroup. Archer et al. (1977) reported a K-Ar age determination of 1.5 Ga on biotite from an intrusive breccia complex which cuts strata of the Upper Fairchild Lake Group at Quartet Mountain. A ^{207}Pb-^{235}U age of 1153 Ma and a ^{207}Pb-^{206}Pb age of 1249 Ma were obtained on a pitchblende sample collected from a fracture system peripheral to a breccia complex developed in the lower Quartet Group west of Quartet Lakes (Archer and Schmidt, 1977). Godwin et al. (1978) reported a Stacey and Kramers model lead age of about 1.44 Ga for two lead deposits hosted in strata of the Gillespie Lake Group in the Coal Creek Dome of the Olgivie Mountains. Morin (1978) reported a Stacey and Kramers model lead isotope age of 1288 Ma for layered galena in rocks of the Gillespie Lake Group from the Hart River area.

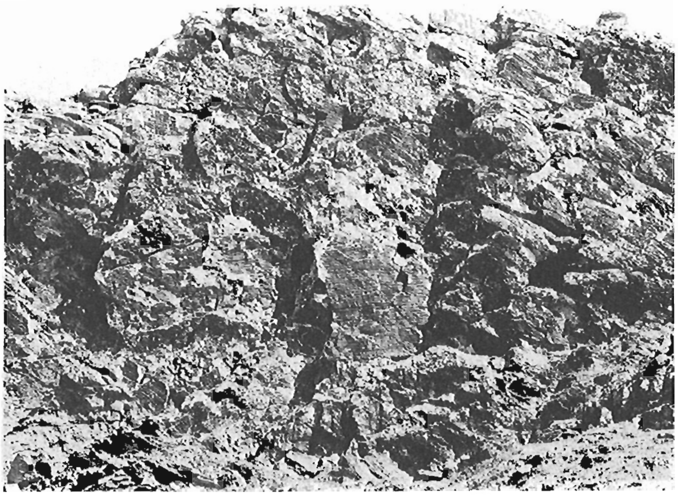

Figure 1.24. Stope breccia phase of a breccia body at Quartet Mountain. Note variable orientation of blocks. Large block near centre of photo is about 1 m wide.

Recently, K-Ar age dates of 613 ± 15 Ma (GSC K-Ar No. 3118) and 552 ± 13 Ma (GSC K-Ar No. 3115) were determined on biotite from two petrographically different lamprophyres which cut strata of the upper Fairchild Lake Group in the northern part of the area. These intrusive bodies show chemical affinities to carbonatites (Gabelman, 1977a). Both dykes have a spatial association with splays of the Richardson Fault Array. It has subsequently been reported (G.M. Yeo, personal communication) that dykes petrographically similar to one of those dated (GSC K-Ar No. 3118) cut strata of the Rapitan Group in the Knorr Range northeast of Kiwi Lake. The lamprophyre dykes at this locality are intimately associated with the Knorr Fault, an easterly splay of the Richardson Fault Array. It is concluded that the dates obtained may provide a minimum age for the deposition of the Rapitan Group (Yeo, 1978).

STRATIGRAPHIC CORRELATIONS

More detailed stratigraphic information on the middle and upper Proterozoic successions of the northern Cordillera and Shield has permitted more detailed lithostratigraphic correlations in this region (Young, 1977; Aitken et al., 1978; Young et al., 1979). These correlations have also been extended to encompass northern Canada and northern Greenland (Young, 1979) with the view of gaining a comprehensive understanding of middle to late Proterozoic evolution of the region surrounding the Canadian craton.

Strata of the Wernecke Supergroup are correlated with the lower three groups of the Belt-Purcell Supergroup of the south-central Cordillera (Young et al., 1979; Young, 1979; Young et al., in press). The successions of these two areas are characterized by a thick sequence of fine grained siliclastic rocks with some intercalated carbonate units, capped by thick stromatolitic carbonates (i.e. Wallace-Siyeh and Gillespie Lake groups). Although the distinct angular unconformity at the top of the Gillespie Lake Group is not found within the Belt-Purcell Supergroup, there are several lines of evidence suggesting a period of major tectonic adjustment prior to the deposition of the fine grained terrigenous rocks of the Missoula Group which overlie the stromatolitic carbonates of the Wallace-Siyeh Formation (Harrison, 1972; Young et al., 1979).

DEPOSITIONAL SETTING OF THE WERNECKE SUPERGROUP

Many hypotheses have been published on the nature and origin of the western continental margin of North America during the Middle Proterozoic. Most workers have characterized the thick fine grained terrigenous-carbonate sequences of the Belt-Purcell Supergroup and correlative rocks as a miogeoclinal succession akin to those formed as a prograding terrace wedge (Price, 1964; Gabrielse, 1972; Harrison and Reynolds, 1976; Monger and Price, 1979). Some "Beltologists" have suggested that this margin was created by a major rifting event in the middle Proterozoic (Stewart, 1977; Monger et al., 1972). Others have carried this concept further, suggesting that counterparts formed during the rifting event may be found in Siberia (Sears and Price, 1978) or Australia (Jefferson, 1978). Badham (1978) however, has disputed this mid-Proterozoic rifting event, suggesting that there has been a continental margin to the west of the North American craton since the Archean.

Available information on the depositional basin in which the Belt-Purcell sediments were deposited suggests a somewhat complex margin characterized by re-entrants (Harrison and Reynolds, 1976; Harrison et al., 1974; Harrison, 1972). The nature of this same margin in northern Canada however, is for the most part conjecture.

In the northern Cordillera, Proterozoic strata occur in an arcuate pattern which follows the trend of the Mackenzie, Wernecke, Ogilvie and Barn mountains (Fig. 1.1). Gabrielse (1972), in an analysis of the nature and distribution of Proterozoic strata in the Mackenzie Mountains, favoured the model of Jeletzky (1962) who suggested that, north of the Mackenzie Mountains, the trend of the depositional margin was east-west (Fig. 1.27). In an interpretation of the Laramide structural style of the eastern margin of the northern Cordillera, Norris (1972, p. 638) theorized that the arcuate structural plan of this region (Fig. 1.1, 1.27) may have in part been due to the "ancient, persistent and fundamental shape of the eastern margin of the miogeocline". Aitken and Long (1978) provided some corroborating evidence for this hypothesis when they suggested on the basis of isopach analysis, that the arcuate shape of the Mackenzie Mountains existed during deposition of rocks of the Mackenzie Mountains Supergroup (sequence B).

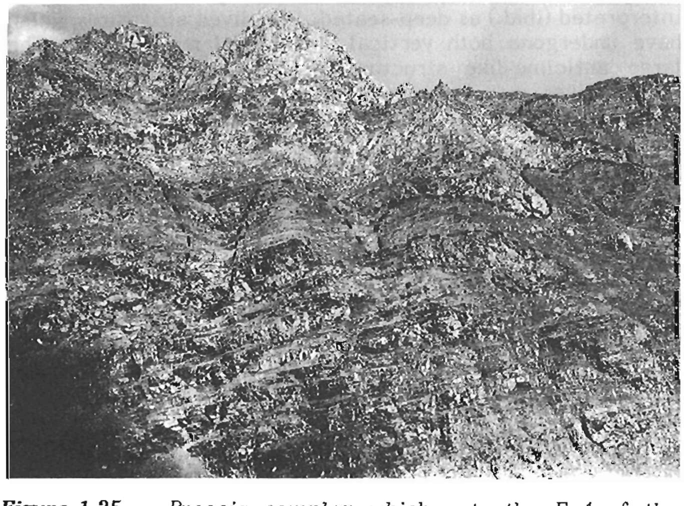

Figure 1.25. Breccia complex which cuts the F-4 of the Fairchild Lake Group in the vicinity of Delores Creek. Note irregular configuration of breccia.

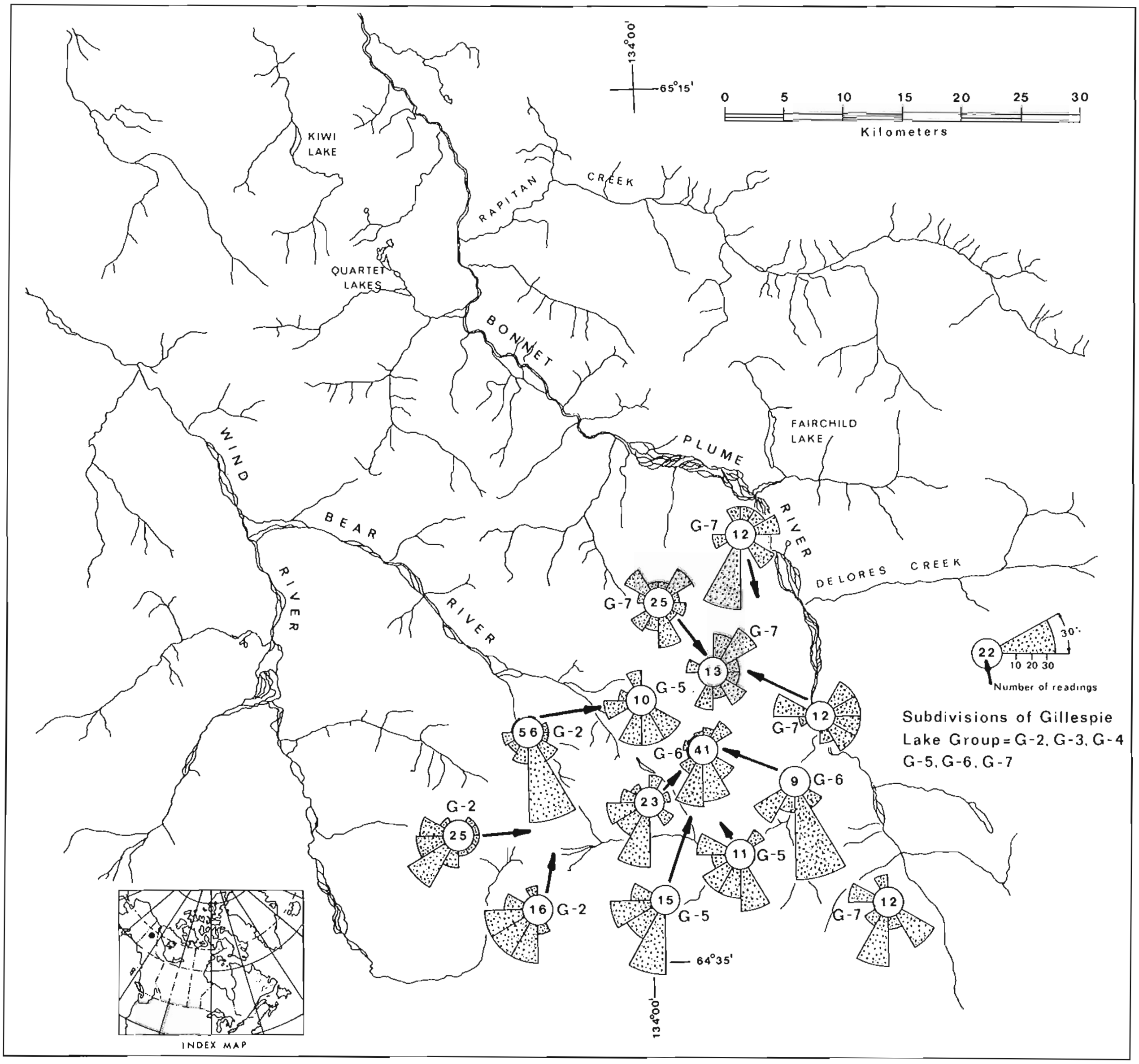

Figure 1.26. *Paleocurrent data from cross-bedding in the Gillespie Lake Group of the northern Wernecke Mountains.*

Another factor which would seem to support the arcuate nature of the Proterozoic depositional margin of the northern Cordillera, at least during the deposition of rocks of sequences B and C, is the relative position of these strata (Fig. 1.1). Rocks of sequence B generally lie "outboard" of those of sequence A and they in turn are flanked "outboard" by rocks of sequence C.

Detailed facies and paleocurrent analyses will be required to determine whether or not the arcuate configuration of the eastern margin of the northern Cordillera existed during deposition of the sediments of the Wernecke Supergroup. Future investigations in the Olgivie Mountains and in the Keele Range of the Porcupine Plateau (Fig. 1.1) will be of particular importance in resolving this problem.

There is however, some evidence supporting the existence of an east-west-trending depositional margin for strata of the Wernecke Supergroup in the Wernecke Mountains. This evidence includes: 1- the predominantly unimodal southerly-directed paleocurrent dispersal trends which characterize the great thickness of the Fairchild Lake Group, as well as some formations of the Gillespie Lake Group (Fig. 1.10, 1.18, 1.26); 2- the orientation of slump structures in the Quartet Group which indicate a south-easterly trending paleoslope; 3- facies changes in some of the formations of the Fairchild Lake Group which indicate that thicker, dominantly dolomitic carbonate rocks lie to the north with thinner dominantly limestone units lying to the south.

If in fact the eastern margin of the northern Canadian Cordillera was characterized by the arcuate trend suggested by Norris (1972), then what was the origin of this margin and what was the provenance of the thick sequence of miogeoclinal sediments comprising the Wernecke Supergroup? Clues to the genesis of this depositional regime may lie to

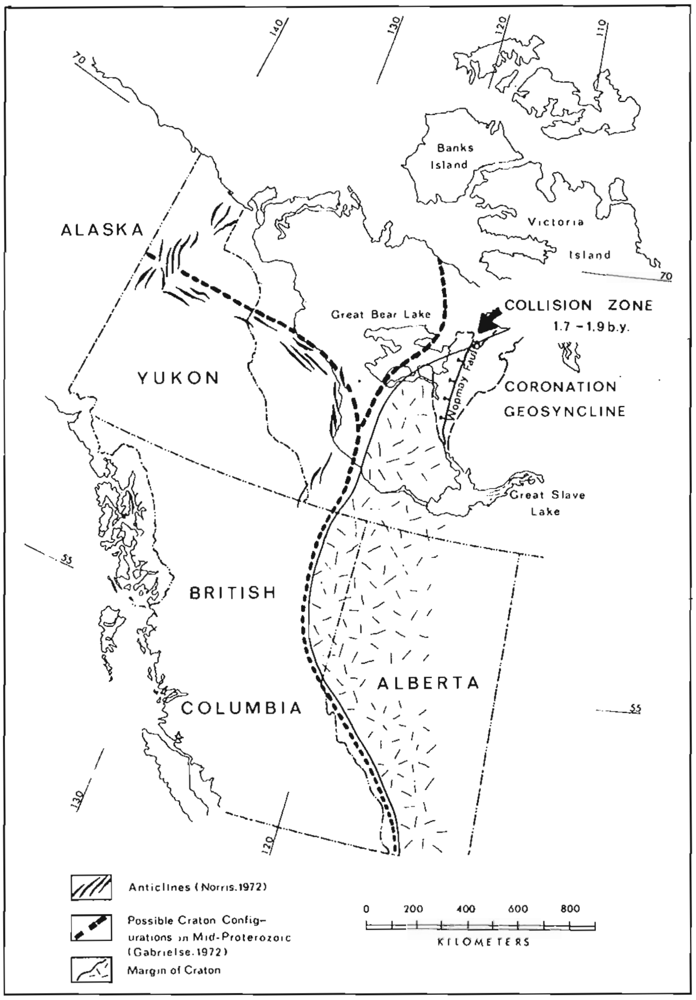

Figure 1.27. *Tectonic elements which have a bearing on the nature of the mid-Proterozoic continental margin of northwestern North America.*

the east of the Cordillera, in the Aphebian Coronation Geosyncline (Hoffman, 1973; Hoffman and McGlynn, 1977). Hoffman (1979) has suggested that a recently recognized younger orogenic event in the Coronation Geosyncline may record a continental collision which juxtaposed the Great Bear batholith complex against the Hepburn Belt (Fig. 1.27). Hoffman has further speculated that this collisional event probably postdates the development of the Great Bear Batholith (1.7-1.9 Ga). The plate involved in this collisional event may be akin to one of the microplates which Churkin and Eberlein (1977) have hypothesized were colliding with the western margin of the North American continent during the Proterozoic and Paleozoic.

CONCLUSIONS

The mid-Proterozoic Wernecke Supergroup is a 15 km-thick sequence of fine grained terrigenous and carbonate sediments which records a shoaling-upward miogeoclinal depositional regime. These strata constitute the oldest succession in the northern Cordillera and thus provide clues to the early geological evolution of this region. Analyses of paleocurrent data suggest that in the mid-Proterozoic, an east-west depositional margin lay to the north of the present day Wernecke Mountains. The origin of this depositional margin and the provenance of the sediments of the Wernecke Supergroup may be related to a continental collision event which occurred in the Coronation Geosyncline about 1.7 to 1.9 Ga ago. The development and refinement of this hypothesis awaits further studies of the Wernecke Supergroup, particularly in the Olgivie Mountains and in the Keele Range of the Porcupine Plateau.

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2.

THE LATE PROTEROZOIC RAPITAN GLACIATION IN THE NORTHERN CORDILLERA

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Yeo, G.M., The late Proterozoic Rapitan Glaciation in the northern Cordillera; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 25-46, 1981.

Abstract

The Rapitan Group comprises a sequence of marine, glaciomarine, and possible glacial sediments outcropping in the Mackenzie and Wernecke mountains. The lowest unit, the Mount Berg Formation, is a grey-green mixtite seen only in the southern Mackenzie Mountains. Conformably above this, the Sayunei Formation is dominated by maroon to red, silty rhythmites with abundant dropstones. Conformably overlying this, the Shezal Formation successively comprises maroon and greenish grey mixtites. These two formations outcrop discontinuously throughout the southern and central Mackenzie Mountains. A homologous succession of rhythmites and mixtites is present in the Upper Tindir Group in Alaska. In the northern Mackenzie Mountains, however, only red and greenish grey mixtites occur. Jasper-hematite iron formation of probable rift-related, volcanogenic hydrothermal origin was found in the red mixtites of the northern Mackenzies as well as in the upper Sayunei and Tindir rhythmites. The Rapitan Group lies not far below the Precambrian-Cambrian boundary and overlies rocks judged to be about 0.8 Ga old. Except locally, the basal contact is distinctly unconformable. This unconformity marks the last major faulting event before the Paleozoic Era. An early, local(?), glacial advance (Mount Berg) was separated from a later one (Shezal) by an interstadial period (Sayunei).

On nearly every continent late Proterozoic glacial and glacial marine sequences, commonly bearing iron formations of hydrothermal origin, were deposited following major faulting episodes. This suggests that global(?) glaciation(s) followed widespread continental breakup. The development of new seaways in combination with extensive continental uplift resulted in new global circulatory patterns and weather systems. Glaciation in the late Proterozoic was the likely consequence of these factors.

Résumé

Le groupe de Rapitan se compose d'une série de roches sédimentaires marines, glaciomarines et peut être glaciaires affleurant dans les monts Mackenzie et Wernecke. L'unité la plus profonde, la formation du mont Berg, est une mixtite gris-vert qui n'apparaît que dans le sud des monts Mackenzie. La formation de Sayunei repose en conformité en dessus; elle est constituée surtout par des rythmites limoneuses maron à rouges, avec d'abondantes stagmalites; en conformité au-dessus de cette dernière, la formation de Shezal se compose successivement de mixtites maron et gris-vert. Les deux formations affleurent de façon discontinue dans le sud et le centre des monts Mackenzie. On trouve une succession homologue de rythmites et mixtites dans la partie supérieure du groupe de Tindir en Alaska. Dans le nord des monts Mackenzie cependant, on ne retrouve que des mixtites rouges et gris-vert. Une formation ferrifère à jaspe-hématite d'origine probablement hydrothermale, volcanogène et reliée à une fissure, a été trouvée dans les mixtites rouges du nord de la chaîne de Mackenzie ainsi que dans les rythmites supérieures de Sayunei et celles de Tindir. Le groupe de Rapitan se trouve à peu de distance au-dessous de la limite Précambrien-Cambrien, et repose sur des roches qu'on estime âgées de 0,8 Ga. Sauf par endroits, le contact de base est nettement discordant. Cette discordance marque le dernier important évènement de failles avant le Paléozoïque. Une ancienne avancée glaciaire localisée (?) (mont Berg) a été séparée d'une autre plus récente (Shezal) par une période interstadiaire (Sayunei).

Sur presque tous les continents, les séries glaciaires et glaciomarines de la fin du Protérozoïque portant souvent des formations ferrifères d'origine hydrothermales, ont été déposées après des évènements de failles importants. Cette observation donne à penser qu'une ou plusieurs glaciations mondiales (?) ont suivi le fractionnement majeur des continents. L'apparition de mers nouvelles et la remontée considérable des continents ont donné lieu à de nouveaux modes de circulation planétaire et de nouveaux systèmes météorologiques. La glaciation de la fin du Protérozoïque résulte vraisemblablement de ces fracteurs.

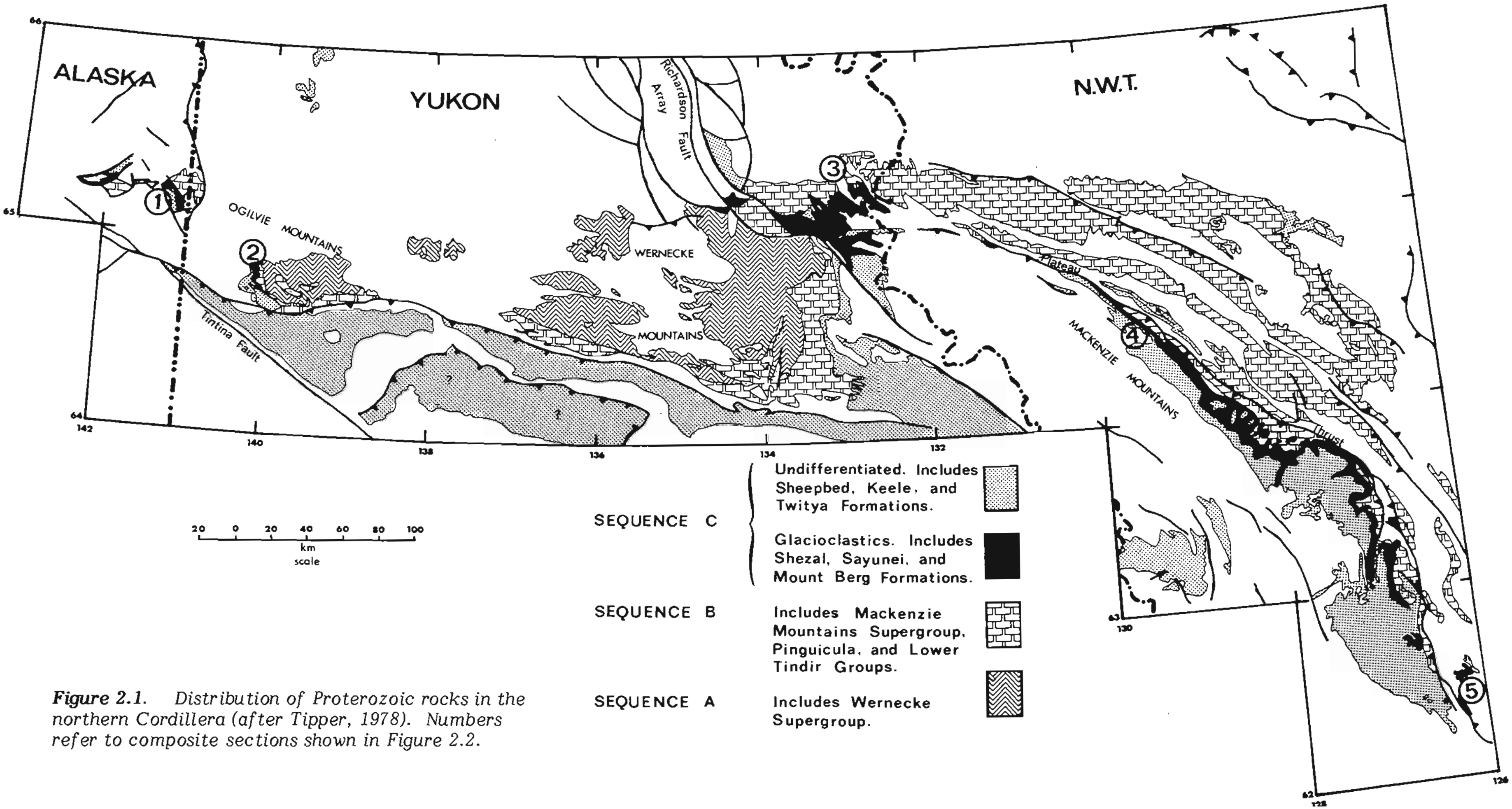
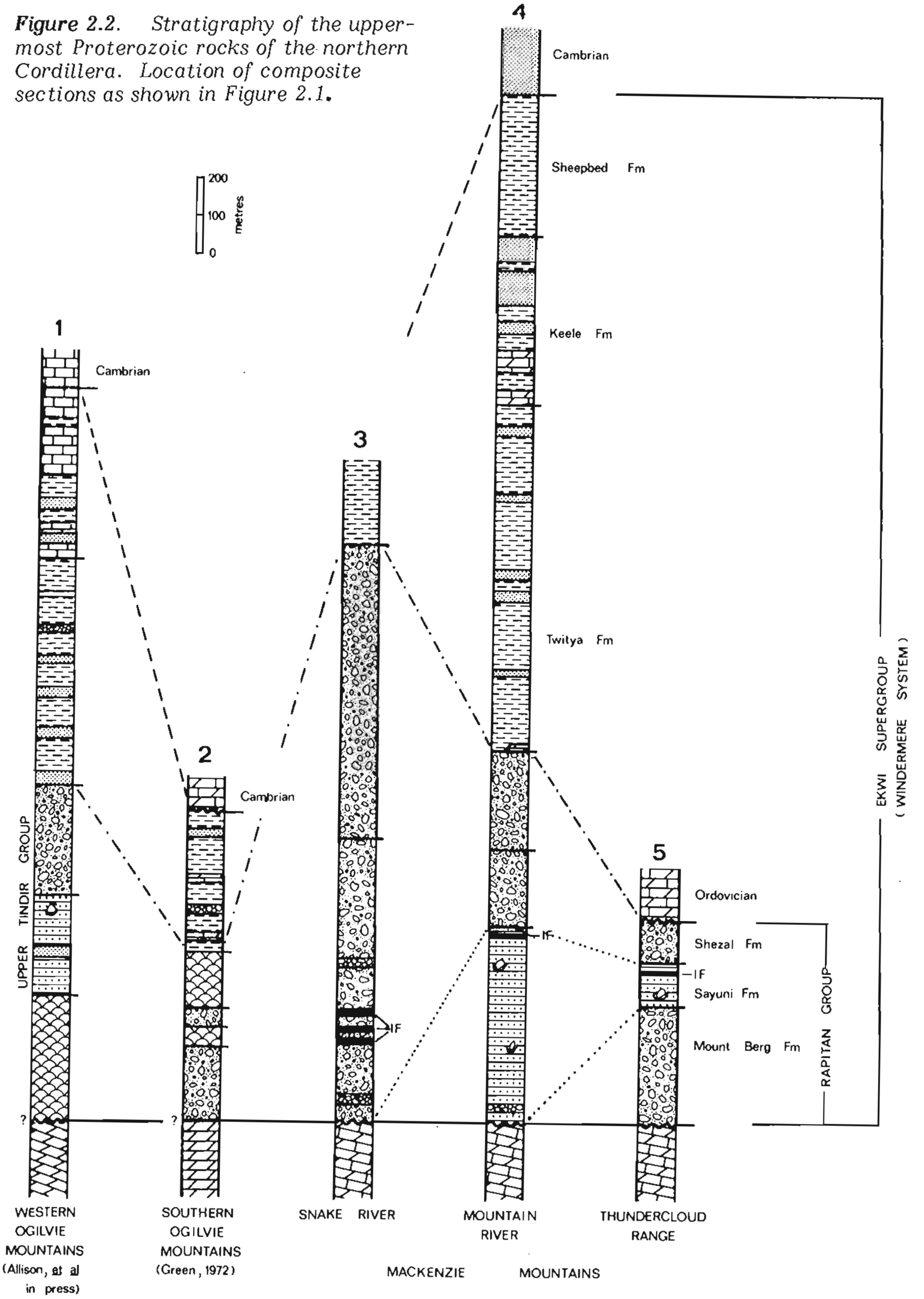


Figure 2.1. *Distribution of Proterozoic rocks in the northern Cordillera (after Tipper, 1978). Numbers refer to composite sections shown in Figure 2.2.*

INTRODUCTION

Among the most striking features of the late Proterozoic is the occurrence, on nearly every continent, of deposits of probable glacial origin. Less well-known is the widespread occurrence of iron formation, often intimately associated with supposed glacial sediments. Such an association occurs in the Cordillera of northwestern Canada and adjacent Alaska. This paper reviews the stratigraphy and geochemistry of the Rapitan glacial sediments and iron formation and speculates on the significance of these phenomena in the late Proterozoic.

Figure 2.2. Stratigraphy of the uppermost Proterozoic rocks of the northern Cordillera. Location of composite sections as shown in Figure 2.1.



Stratigraphic nomenclature formalized by Eisbacher (1978a) is followed here, except for recognition of a distinct mixtite unit, the Mount Berg Formation, below the Sayunei Formation, and inclusion of Ziegler's (1959) Snake River Tillite (maroon mixtites, conglomerates, and sandstones in Snake River area) within the Shezal Formation rather than the Sayunei. Eisbacher (1978a) expanded the Rapitan Group to include the Keele and Twitya formations, presumably to emphasize a major clastic to carbonate cycle within a sequence of such megacycles (Eisbacher, 1976). It has also been proposed that the Rapitan Group be restricted to its original sense (Green and Godwin, 1963) to include only the mixtites and closely related rocks and that a new unit, the Hay Creek Group, be established to include the overlying Proterozoic rocks (Yeo, 1978; Young et al., 1979). Because the emphasis of this paper is on the supposed glacial units, this little, restricted, usage is continued here.

STRATIGRAPHIC AND STRUCTURAL FRAMEWORK OF THE RAPITAN GROUP

The late Proterozoic sedimentary record of northwestern North America can be subdivided into a triad of major sequences separated by significant tectonic events at approximately 1.2 Ga and 0.8 Ga (Young et al., 1979). The youngest of these is recognized only in the Cordillera (Fig. 2.1). The Rapitan Group and its correlatives lie at the base of this sequence (Fig. 2.2).

Stratigraphic Framework

The unmetamorphosed Rapitan Group unconformably overlies the stable platform assemblage of carbonates and relatively mature clastics of the Mackenzie Mountains Supergroup, which forms the middle sequence of the upper Proterozoic triad in Mackenzie Mountains. Synsedimentary faults, intrusive and sedimentary breccias, conglomerates, redbeds, and evaporites indicate restricted basin development and tectonic instability towards the end of deposition of the Mackenzie Mountains Supergroup (Eisbacher, 1978b; Jefferson, 1978).

An angular unconformity generally marks the base of the Rapitan Group. In eastern Wernecke Mountains, the Rapitan Group unconformably overlies the Pinguicula Group (Eisbacher, 1978b), a poorly known platform sequence thought to be at least partly correlative with the Mackenzie Mountains Supergroup (Eisbacher, 1978b; Yeo et al., 1978; Young et al., 1979). Blusson (1976) reported Rapitan equivalents in the southern Wernecke Mountains as well.

The Rapitan Group (sensu stricto) includes maroon and grey mixtites, silty rhythmites, conglomerates, sandstones, and iron formation (Fig. 2.2). Maroon rhythmites of the Sayunei Formation are the basal unit over most of the Mountain River and Redstone River areas. In Thundercloud Range, however, mixtites underlie the Sayunei. Maroon mixtites of the Snake River Tillite member of the Shezal Formation take the place of the Sayunei rhythmites in Snake River area. Maroon mixtites of the Shezal Formation also locally overlie the Sayunei Formation. The upper Shezal Formation comprises mainly greenish grey mixtites. Jasper-hematite iron formation occurs in both the Shezal and Sayunei formations.

Overyling the Rapitan Group are recessive, friable to fissile, dark grey shale and minor sandstone of the Twitya Formation. The Twitya interfingers with and is overlain by characteristically resistant carbonates, sandstones, and siltstones of the Keele Formation. The Proterozoic succession is capped by recessive mudstones of the Sheepbed Formation. To emphasize their distinctness from the underlying clastics these three units have been collectively called the Hay Creek Group (Yeo, 1978; Young et al., 1979). Conformably above this lie resistant sandstones and carbonates of the late Lower Cambrian Backbone Ranges Formation.

Unmetamorphosed Precambrian strata in the western Ogilvie Mountains of east-central Alaska make up the Tindir Group (Cairnes, 1914; Mertie, 1932; Churkin, 1973). In the upper part of the Tindir Group a sequence of iron formation-bearing maroon rhythmites and mixtites associated with basalts is remarkably like the Rapitan (Fig. 2.2). Shales, sandstones, and carbonates above strongly resemble the Twitya and Keele formations. The Upper Tindir Group is also overlain by Lower Cambrian carbonate and clastic strata.

South and west of Coal Creek Dome in the southern Ogilvie Mountains (116B and 116C/E½) thick dolomite boulder conglomerates (Unit 2d of Green, 1972) may be correlative with the Upper Tindir mixtites or the Shezal Formation (Fig. 2.2). Locally these rocks are interbedded with and overlain by calcareous, vesicular lava flows, breccias, and agglomerates. They overlie possible Pinguicula Group equivalents (grey dolostones with minor black shale and quartzite) and are overlain by possible Hay Creek Group equivalents (dark shales with minor dolostone).

Geochronology and Paleomagnetism

Geochronology

Stromatolites, macrofossils and algal microfossils described from the Little Dal Group suggest a middle to late Riphean age (1100-800 Ma) (Hofmann and Aitken, 1979; Aitken, 1978). The middle to upper Riphean stromatolite **Baicalia** is reported in carbonate rocks of the Lower Tindir Group in Alaska (Churkin in Allison and Moorman, 1973). A lower age limit of about 800 Ma for the onset of Rapitan deposition is suggested by an age of 790 Ma (Rb-Sr) from gabbros intruding the Mackenzie Mountains Supergroup (Aitken, 1979).

Nodular and tuberose, parallel branching, columnar stromatolites occur in the Keele Formation and resemble the Late Riphean to Cambrian form, **Acaciella** (Walter, 1972).

A lower Cambrian (Waucoban) upper age limit is indicated by the Ollenellid-Archaeocyathid fauna reported in rocks conformably overlying the Proterozoic succession in Mackenzie Mountains (Gabrielse et al., 1973). K-Ar age determinations on biotite from lamprophyre dykes south of Margaret Lake (106 E) yielded dates of 552 ± 13 Ma and 5613 ± 15 Ma (Delaney, personal communication). Similar dykes cut the Twitya Formation north of Margaret Lake.

A Lower Cambrian archaeocyathid-trilobite fauna is also present in rocks conformably overlying the Tindir Group (Brabb, 1967). Microfossils similar to those in the late Proterozoic Hector Formation at the top of the Windermere Supergroup occur in shales and limestones overlying the Basalt-Redbed unit of the Upper Tindir Group (Allison and Moorman, 1973).

Paleomagnetism

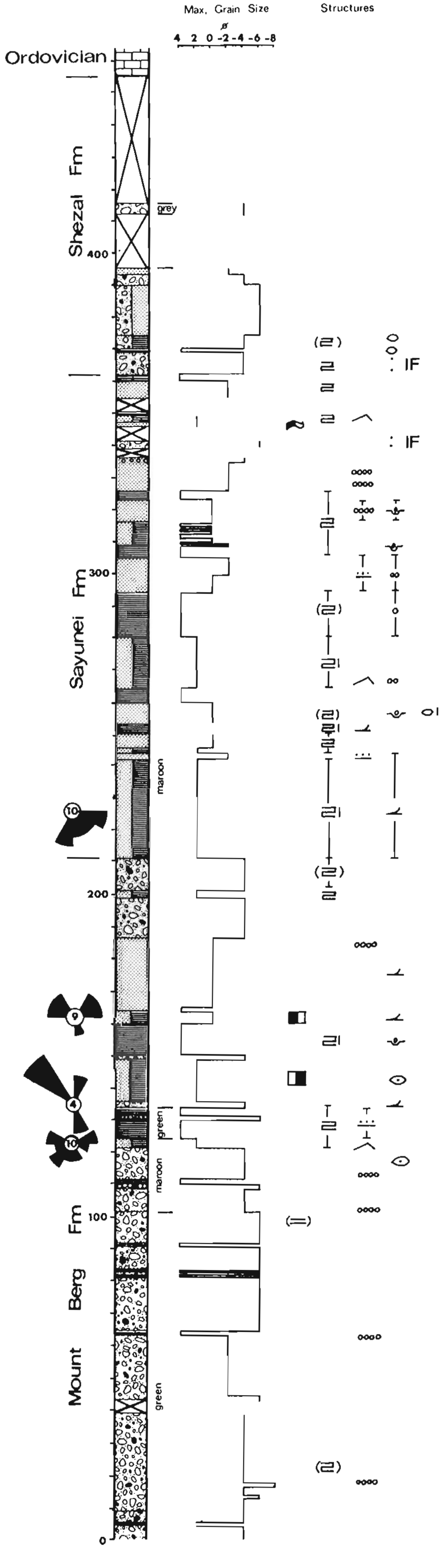
Paleomagnetic data from the late Proterozoic are still relatively sparse and open to numerous interpretations (e.g. Morris and Roy, 1977; Nairn and Ressetar, 1978). Paleomagnetic investigations of the lower Rapitan by Morris (1977) and Morris and Park (1981) yielded three directions of remnant magnetization. One of these is interpreted to be due to thermal overprinting on account of its lower magnetic stability. The most stable direction gave a low latitude pole, while the remaining direction gave a high latitude pole. Morris (1977) tentatively interpreted the low latitude pole to represent the original depositional remnance of the lower Rapitan and the high latitude pole to represent later diagenetic overprinting. Paleomagnetic studies of other supposed late Proterozoic tillites in east Greenland, Norway, and Scotland have also revealed two remnance directions representing both high and low latitude poles (Morris, 1977). The ambiguity of paleolatitude has yet to be resolved.

Structural Geology and Metamorphism

Deposition of the Rapitan was preceded by and controlled by widespread block faulting and uplift of underlying strata to the east. Jefferson (1978) showed that the northwest trending pre-Rapitan miogeocline in the southern Mackenzie Mountains was probably complexly embayed. These embayments were presumably controlled by normal faulting along north-northwesterly trends as indicated by Eisbacher (1977, Fig. 46.9). This faulting may have accompanied northwesterly displacement along a dextral transform system ancestral to the Richardson Fault Array (Norris, 1977; Norris and Hopkins, 1977). Paleoscarps exposed at a few localities show that pre-Rapitan relief of hundreds of metres was developed. Consequently, an angular, onlapping unconformity is widely developed at the base of the Rapitan, particularly towards the margin of the Rapitan basin (cf. Eisbacher, 1978a, Fig. 3). Locally, as at Coates Lake, the Rapitan lies conformably on rhythmites of the Coppercap Formation. This suggests that no great hiatus separated deposition of the Rapitan and the underlying Mackenzie Mountains Supergroup.

Eisbacher (1978b) and Helmstaedt et al. (1979) have suggested that this basal unconformity occurs locally within the Sayunei Formation. They described folded and faulted Coppercap Formation carbonates and basal Sayunei siltstones overlying Redstone River evaporites in Sekwi Mountains (105 P) between the Keele and Ekwi rivers. There, fold axes and faults trend north-northwest parallel to the trend of Laramide folding and thrusting. Although maroon mixtites are reported to truncate the folds and faults (Helmstaedt et al., 1979, Fig. 4) this is not clear in the field. The supposed unconformity is obscured by talus and it is unclear whether it postdates the major reverse fault shown cutting the folds. The maroon mixtite itself is gently folded about the same axes as the underlying Sayunei siltstones and Coppercap limestones. Therefore the folding event postdates the apparent unconformity. An alternative explanation for the observed angular discordance is that it is a product of Laramide thrusting (Yeo, 1978; Yeo et al., 1978).

Local faulting continued during Rapitan deposition. Small scale synsedimentary faults are common, as are minor slump folds (Young, 1976; Plate 1b; Eisbacher, 1978a, Fig. 9).



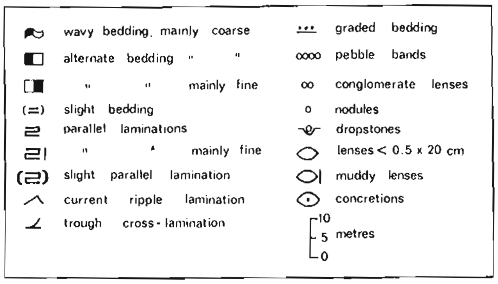


Figure 2.3. Type section of the Mount Berg Formation, on the south side of Mount Berg in the Thundercloud Range of the Mackenzie Mountains (62°31'N; 126°15'W).

Continued local tectonic instability following Rapitan deposition is demonstrated by the spectacular carbonate olistoliths in the Twitya Formation in Mountain River area (Eisbacher, 1978a, Fig. 28). Norris and Hopkins (1977) postulated at least 40 km of pre-Cretaceous, dextral displacement of the Shezal outlier south of Margaret Lake (106 E) along the Knorr Fault, part of the Richardson Fault Array. The latter was probably active from late Proterozoic time (Norris, 1977; Delaney et al., in press).

Major deformation affecting the Rapitan probably did not occur until the Laramide orogeny. Thrusting and en echelon folding at this time produced tectonic shortening of 12 to 14 per cent (Norris, 1972; Aitken and Long, 1978). Aitken and Long (1978) have suggested that the thrust faults themselves were controlled by the shape of the youngest Proterozoic clastic wedge (i.e. Sequence C of Young et al., 1979).

STRATIGRAPHY OF THE RAPITAN GROUP AND RELATED ROCKS

The Rapitan Group outcrops discontinuously for about 630 km in an arcuate north and northwest trending belt (Fig. 2.1). Two major basins are recognizable in Mackenzie Mountains: the Snake River basin in the northwest and the Mountain River-Redstone River basin in the south and east. The latter may be subdivided into at least three subbasins. In the southern basin three possible glacioclastic formations occur: the Mount Berg, Sayunei, and Shezal formations. In the northern basin only the Shezal Formation outcrops.

Two possible glacioclastic formations occur in the Basalt-Redbed unit of the Tindir Group of east-central Alaska. As mentioned earlier, other rocks probably correlative with the Rapitan, but poorly known, outcrop in the southern Ogilvie and Wernecke mountains.

Mount Berg Formation

The Mount Berg Formation (formerly Mount Berg Member, Yeo, 1978) outcrops only in the southeast in Thundercloud Range (95 L). It was first described briefly by Condon (1964). Predominantly greenish grey mixtites and siltstones are locally more than 313 m thick. Besides these two dominant lithologies, conglomerate and sandstone beds are also present. The base of the formation is defined as the first appearance of mixtite above carbonate and clastic rocks of the Mackenzie Mountains Supergroup. The lower contact is not exposed at the type section on the south side of Mount Berg (Fig. 2.3), but underlying Little Dal carbonates outcrop to the east. The Mount Berg interfingers above with the maroon rhythmites characteristic of the Sayunei Formation. The Mount Berg Formation rocks are typically greenish grey, but may be red to brown, particularly in the upper part. The strong resemblance between the Mount Berg and Shezal formations suggests a similar origin.

Three principal lithofacies, described below, are present in the Mount Berg.

Mixtite Lithofacies

The mixtite and mudstone lithofacies are crudely interbedded. Beds are up to 7 m thick but are typically less than 1 m. In the mixtites, the megaclast content varies up to about 60 per cent. Clasts are angular to subrounded and up to 25 cm in diameter; pebble sizes predominate. Angular and subangular intraformational siltstone fragments are most abundant, but better rounded carbonate and greenstone clasts are common. In one section, subrounded quartzite and red jasper pebbles are present. With this exception, no extrabasinal clasts were found. Other than a vague orientation subparallel to bedding no clast orientation was recognized. The mixtite matrix is silty to sandy and weathering is blocky to friable, and moderately recessive.

Mudstone Lithofacies

Predominantly silty mudstone beds are up to 2 m thick, but generally thinner than the interbedded mixtites. The mudstones are massive to faintly laminated. Contacts appear to be sharp and even. The mudstones are friable and more recessive than the mixtite.

Sandstone and Conglomerate Lithofacies

Blocky and resistant, fine- to medium-grained laminated sandstone beds up to 0.3 m thick are interbedded with the mixtites and siltstones. Lower contacts are generally sharp and even. Pebble-cobble conglomerate beds up to 0.6 m thick are interbedded with and transitional into the mixtite. No structures were noted in the conglomerates. Subrounded carbonate and greenstone clasts predominate in a sandy to silty matrix.

Paleocurrents

Too few paleocurrent data were collected to clearly indicate the dispersal of the Mount Berg Formation. However, its absence below the Sayunei to the west in Backbone Ranges suggests derivation from the east.

Sayunei Formation

The Sayunei Formation is characterized by "numerous, monotonously laminated siltstone or sandstone beds, commonly intercalated with lenses of coarse, angular material..." (Eisbacher, 1978a). It is typically red, reddish brown, or maroon. Green or greenish grey colours are common, however, especially at the base and top of the formation and in coarse grained beds (cf. Eisbacher, 1978a, Fig. 10). It is less resistant than the underlying carbonates, but more resistant than the overlying mixtites and shales. It lies unconformably above the Mackenzie Mountains Supergroup, except locally, as in Thundercloud Range where it interfingers with the upper part of the Mount Berg Formation, or at Coates Lake where it appears to be transitional from greenish grey carbonate rhythmites of the Coppercap Formation. The Sayunei outcrops above the Plateau Thrust as far north as the western edge of Mount Eduni map area (106 A). Regional thickness variations indicates that it was deposited in three or more subbasins (Fig. 2.4).

Five principal lithofacies are distinguished in the Sayunei Formation: siltstone-argillite rhythmites, sandstone, conglomerate, mixtite, and iron formation. The iron formation is described in a later section.

Siltstone-argillite Rhythmite Lithofacies

The predominant lithofacies of the Sayunei Formation comprises millimetre to centimetre thick graded and laminated couplets of siltstone and argillite. Contacts between rhythmites are typically knife-sharp and planar while contacts between the two size phases may be sharp or gradational. Recessive, calcareous siltstone and sandstone lenses and beds up to a few centimetres thick occur in the rhythmites. Small-scale crosslamination may be preserved in such beds.

The siliceous argillite phase of the rhythmite lithofacies is commonly so enriched in hematite towards the upper part of the Sayunei Formation as to be iron formation. Beds of jasper and hematite are commonly associated with such beds.

Flame structures and load casts are common at the base of thick sandy rhythmites. Ripple crosslaminated sand beds and lenses up to several centimetres thick may be intercalated in the rhythmites. The rhythmites include Bouma AE, BCDE, CDE, and DE sequences. Complete Bouma sequences are rare. Convolute bedding is locally well developed (Young, 1976, Plate 1b). Coarse grained, sandy, clastic dykes are also found locally. Rare dolomite pseudomorphs may be present (Young, 1976, Plate 2b).

Subangular to rounded, rarely striated, lonestones are common throughout the Sayunei Formation. These commonly pierce and depress underlying laminae, suggesting that they were dropped into place (Young, 1976, Plates 1c, d). Aggregate sediment pellets are also reported (Young, 1976). Outsize stones commonly are basal to, or just below, conglomerate beds. One such stone, a subrounded carbonate clast 2.5 m in diameter, is associated with a pebble conglomerate bed only 10 cm thick.

Sandstone Lithofacies

The sandstone lithofacies is transitional into the siltstone phase of the rhythmites and into the conglomerate lithofacies. Petrographically, the sandstones are mainly lithic (mudstone) wackes and feldspathic lithic (mudstone) wackes. Beds up to 15 cm thick may be amalgamated into units up to 1 m. Basal contacts are typically scoured with ripped-up argillite fragments, flame structures, and load casts. Internal structures include normal, reverse, or partial grading, preferred grain orientation, and recumbent flow-folding (Eisbacher, 1978a, Fig. 6). Crossbedding may be accentuated by muddy laminae (cf. Bouma and Hollister, 1973, Fig. 13).

Laterally continuous beds up to 3 cm thick, composed of 1 to 2 mm siltstone nodules, occur north of North Redstone River. Contacts of these beds are sharp and even.

Conglomerate Lithofacies

Sayunei conglomerates range from stringers of scattered clasts to beds more than 1.5 m thick. The conglomerates are characteristically organized (Walker and Mutti, 1973) with sharp, irregular contacts, lenticular bedding, framework support, grading, and imbrication. Conglomerate beds are more abundant in the lower part of the Sayunei where they commonly occur in closely spaced sets.

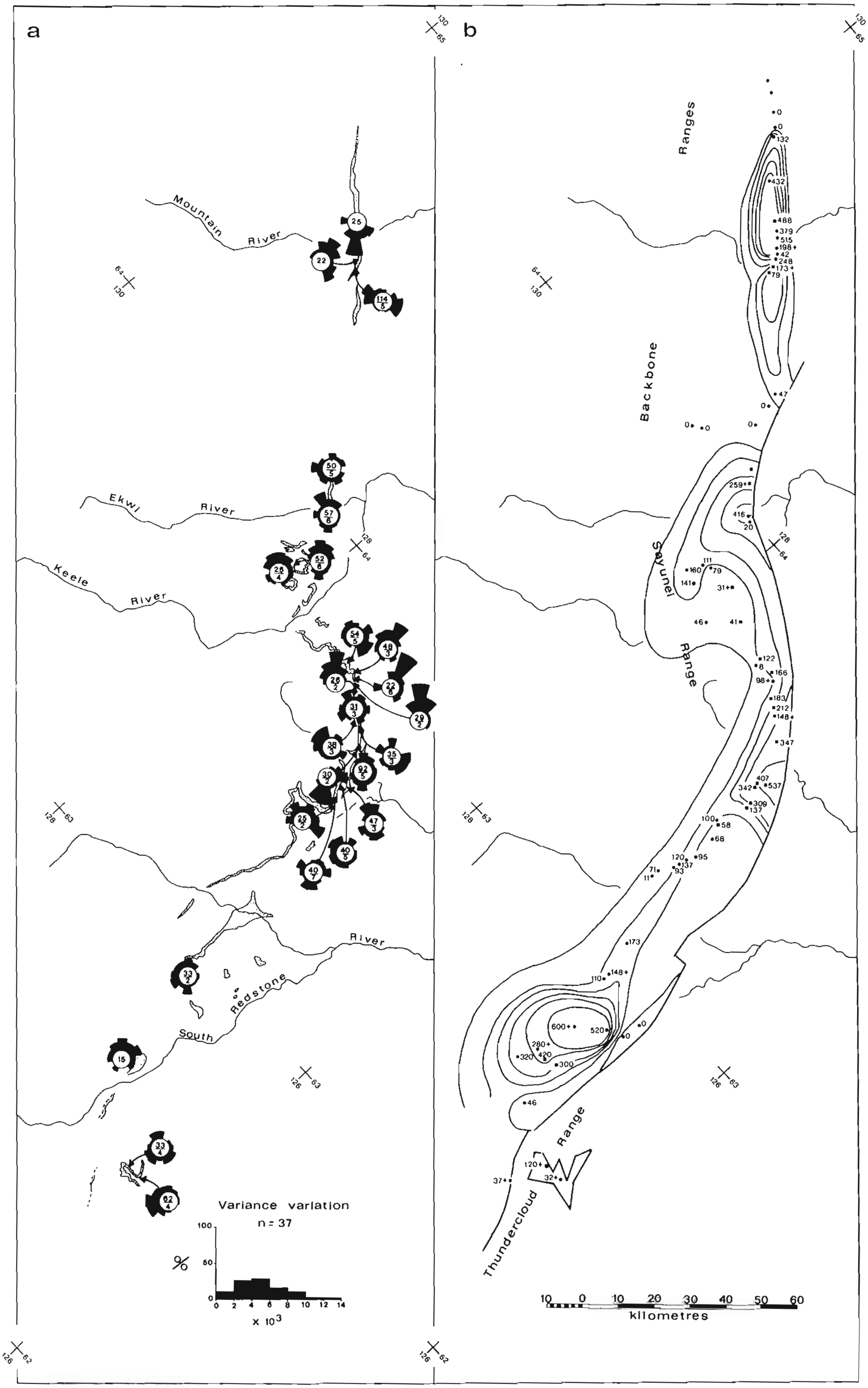
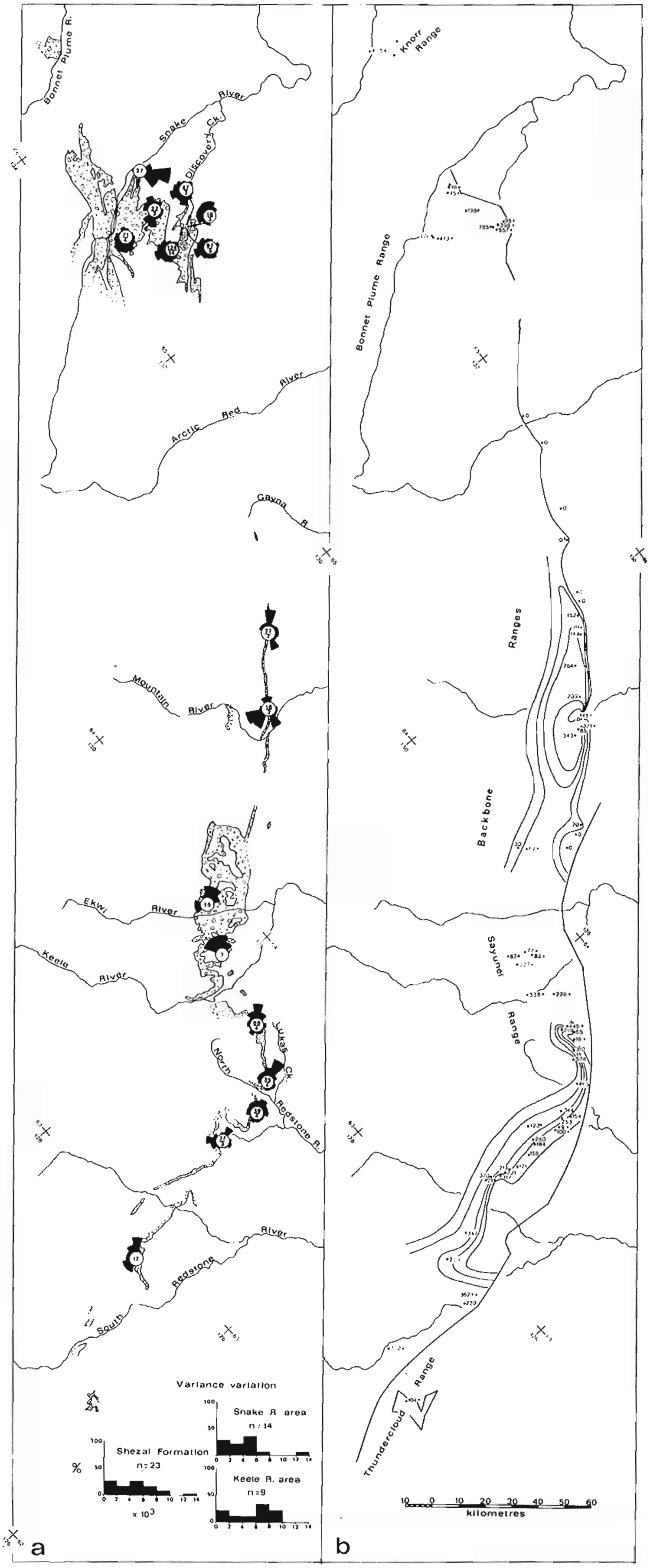


Figure 2.4a

Distribution and paleocurrent summary of the Sayunei Formation in Mackenzie Mountains. Crossbed data are grouped. The lower number indicates the number of stations at which the measurements (upper number) were made.

Figure 2.4b

Preliminary isopach map (non-palinspastic) of the Sayunei Formation. The erosive eastern limit is shown as a wavy line.



The most conspicuous clasts are carbonate. These range from angular to subrounded, but tend to be angular. Volcanic and quartzite fragments are less common and tend to be better rounded. No extrabasinal clasts have been reported.

Mixtite Lithofacies

Minor mixtites are interbedded throughout the Sayunei and include the sharpclast-siltstones of Eisbacher (1978b, Fig. 8). Pebbles to boulders are suspended in a silty matrix. Beds up to a few metres thick are particularly well developed in the upper part of the Sayunei. A thin mudstone layer is present at the base of some mixtite beds. Preferred clast orientation subparallel to bedding may be developed. Contacts are typically sharp and even. Most clasts tend to be subangular or angular, but subrounded clasts do occur.

Paleocurrents

Paleocurrent data, mainly from ripple crosslamination, are quite variable (Fig. 2.4a). Transport is generally westerly. Unimodal, bimodal, and bipolar transport patterns suggest that several transport mechanisms were active. The wide range of variance values (Fig. 2.4a) is typical of marine environments (Long and Young, 1978).

Shezal Formation

The Shezal Formation (Eisbacher, 1978a) is dominated by massive and crudely bedded buff, grey-green, and reddish mixtites. It outcrops discontinuously throughout the Mackenzie arc and in the northeastern Wernecke Mountains (Fig. 2.5). Like the underlying Sayunei, the Shezal Formation pinches out to the southeast and northwest over a few tens of kilometres. In the Keele-Twitya rivers area the depoaxis of the Shezal lies to the west of that of the Sayunei. The Shezal Formation appears to rest conformably on the Sayunei and unconformably on older strata. Locally, Sayunei rhythmites interfinger with basal Shezal mixtites. The upper contact is sharp and conformable beneath dark grey shales and calclithites of the Twitya Formation (Eisbacher, 1978a, Fig. 12). This contact is generally obscured beneath Twitya talus.

Three members, distinguished by colour and structures, are recognized regionally. The lower, hematitic Mountain River Member (Yeo, 1978) was formerly included in the Lower Rapitan (Gabrielse et al., 1973; Aitken et al., 1978). In Snake River area the lower red member is called the Snake River Tillite (Ziegler, 1959). Grey-green to buff mixtites above the red mixtites comprise the Bonnet Plume River Member (Ziegler, 1959; Yeo, 1978). The significance of the colour change is discussed elsewhere in this paper.

__Figure 2.5a.__ Distribution and paleocurrent pattern of the Shezal Formation in Mackenzie Mountains. Crossbed data are grouped as in Figure 2.4.

__Figure 2.5b.__ Preliminary isopach map (non-palinspastic) of the Shezal Formation. The erosive eastern limit is shown as a wavy line.

In addition to mixtite, important lithofacies include mudstone, sandstone, and conglomerate. Jasper-hematite iron formation of economic potential is found in the Snake River Tillite. Minor iron formation occurs locally in the lower part of the Shezal in the southern basin.

Mountain River-Redstone River Area

The reddish lower member of the Shezal Formation, the Mountain River Member, varies in thickness up to 362 m. It is characteristically slightly more recessive than the Sayunei Formation, but less recessive than the overlying grey mixtites. Thick-bedded, generally massive mixtites characterize this member. The scaly weathering characteristic of the grey mixtites is not well developed. Minor lithofacies are conglomerate, sandstone, and mudstone.

The grey upper member of the Shezal, the Bonnet Plume River Member, is reported to be up to 824 m thick (Upitis, 1966). Widespread scaly weathering (Eisbacher, 1978a, Fig. 13) results in recessive sections in which features are obscured by talus. Clast-rich beds tend to be most resistant. Interfingering with, and reworking of, underlying red sediment occurs locally. Massive mixtites predominate but stratified mixtite, conglomerate, sandstone, and mudstone also occur.

Snake River Area

The Snake River Tillite was the name given by Ziegler (1959) to the red mixtites near Snake River. These were subsequently called the Rapitan Group by Green and Godwin (1963). It unconformably overlies carbonates, shales, and sandstones of the Mackenzie Mountains Supergroup and Pinguicula Group (Eisbacher, 1978a, Fig. 12.4). It is transitional both upwards and laterally into grey-green mixtites of the Bonnet Plume River Member. Maximum thickness is greater than 750 m. It is moderately resistant, but becomes recessive upwards. This unit contains significant amounts of jasper-hematite iron formation (Stuart, 1963; Gross, 1965a).

Eisbacher (1978a, b) included the Snake River Tillite in the Sayunei Formation, but, as discussed elsewhere (Yeo, 1978; Yeo et al., 1978) these units are lithologically distinct, although they may be homotaxial equivalents. Mixtites bearing rounded clasts are the characteristic lithofacies of the Snake River Tillite. The rhythmites typical of the Sayunei are absent. Following the reasoning implicit in Eisbacher's (1978a) inclusion of the former Lower Rapitan maroon mixtites in the Shezal Formation, the Snake River Tillite is included here as well.

The major lithofacies of the Snake River Tillite are massive and stratified mixtite, mudstone, sandstone, conglomerate, and jasper-hematite iron formation.

The Bonnet Plume River Tillite described by Ziegler (1959) outcrops south of Margaret Lake (106 E). Farther east, on South Iron Creek, it lies above and lateral to the Snake River Tillite Member. The grey mixtites comprising the bulk of the Shezal farther south are also included in this member. It is conformably overlain by black shales and sandstones of the Twitya Formation. Maximum thickness exceeds 775 m on South Iron Creek (106 F).

Major lithofacies include greenish grey to buff, massive and stratified mixtite, conglomerate, sandstone, and mudstone.

Mixtite Lithofacies

From a distance the Shezal mixtites appear crudely bedded, especially in the lower part. Beds may exceed 10 m in thickness. Bedding is generally difficult to see in outcrop, however. Weathering of the Bonnet Plume River mixtite is typically scaly and recessive, but the red mixtites are more resistant and weather blocky to friable. Two kinds of mixtite occur: massive and stratified.

The bulk of the Shezal comprises massive mixtite. It is characterized by lack of internal structure except for preferred subparallel clast orientation. At several sites in the Snake River Tillite at Discovery Creek (106 F), preferred clast orientation in planes parallel to bedding was observed. Scarcity of well-defined bedding surfaces precluded a regional study of clast orientation. Locally, irregular slickenside surfaces are preserved in the mixtite matrix. The striae were formed by relative movement of small clasts during plastic deformation of the matrix (Eisbacher, 1976, 1978a).

The stratified mixtite is characterized by crude, centimetre-scale bedding due to variable muddy matrix content, wispy mudstone streaks, thin beds of mudstone chip breccia, and occasional laminated, fine grained sandy beds. The stratification is commonly emphasized by colour variation. The scaly weathering and internal deformation characteristic of much of the Shezal (Eisbacher, 1976, 1978a) may obscure stratification. Stratified mixtite is especially common in the Snake River Tillite. Thin, relatively well-bedded mixtite beds here may be traced for hundreds of metres.

The matrix of the mixtites is locally calcareous or dolomitic siltstone. Matrix composition generally reflects megaclast abundance (Eisbacher, 1978a). Hematite cement characterizes the Mountain River and Snake River Tillite members. These are relatively better indurated than the Bonnet Plume River Member.

Stones in the mixtites range from subangular to rounded. Clasts are up to 5 m in diameter (Eisbacher, 1978a). Large clasts tend to be better rounded. Faceted and striated clasts are locally common. Buff to greenish grey carbonate clasts are the most abundant extraformational stones. Greenstone and quartzite clasts are locally common. Reworked jasper-hematite clasts occur locally. Eisbacher (1978a) reported an increase in clast size, density, and diversity towards the top of the Shezal in Redstone River area. However, the reverse occurs in the Snake River Tillite in Discovery Creek area where most stones are probably intrabasinal. Rare rhyolite porphyry and metamorphic clasts locally occur. The nearest known source for such rocks lies east of Great Bear Lake (e.g. Hoffman, 1978).

Mudstone Lithofacies

Structureless mudstone beds up to several metres thick are interbedded with the mixtites. They are identical to the matrix of the massive mixtite, but lack megaclasts. They may grade laterally into mixtite. Contacts are sharp or transitional.

Sandstone Lithofacies

Sandstone beds range up to 3 m in thickness, but are generally less than 40 cm thick. They are commonly massive with sharp even contacts. Basal contacts may show load casts. Less commonly, graded bedding and ripple cross-bedding is developed. Slump folding is common. Pockets of two or more sandstone beds may occur within a short interval. Stuart (1963) reported that graded, fine- to medium-grained, moderately well-sorted feldspathic arenite beds were useful marker beds in the Snake River Tillite. Such lateral persistence has not been observed in Mountain River-Redstone River area.

Minor, finely laminated siltstone beds up to 30 cm thick with sharp even contacts are more akin to the sandstone beds than to the massive mudstones.

Conglomerate Lithofacies

Conglomerate beds in the Shezal are generally less than a metre thick, but locally may be much thicker, particularly in the Snake River Tillite. Commonly the conglomerates form boulder or cobble beds with irregular contacts. Locally, irregular lenses of conglomerate with uneven channeled or load casted bases (Eisbacher, 1978a) are present. Conglomerate beds typically pinch out laterally over a few tens or hundreds of metres. Wedge-shaped massive and graded conglomerates are conspicuous in the Snake River Tillite at Discovery Creek. Clasts are typically subrounded to rounded. Carbonate clasts are most abundant, but greenstone, quartzite, and chert are common.

Paleocurrents

Paleocurrent data from crossbedded sandstones in Mountain River-Redstone River area are variable, but generally suggest westerly transport (Fig. 2.5a).

Paleocurrents in the Snake River area are directed westerly and southerly in the eastern part of Snake River basin (i.e. Discovery Creek, 106 F), but are directed easterly and southerly elsewhere (Fig. 2.5a). No crossbedded sands were found in the type area (Ziegler, 1959) of the Bonnet Plume River Member near Margaret Lake (106 E).

Upper Tindir Group

Cairnes (1914) used the name, Tindir Group, for the thick, unmetamorphosed sedimentary sequence underlying Cambrian limestones in the Tatonduk River-Yukon River area of east-central Alaska. Subsequent studies were undertaken by Mertie (1932), Brabb and Churkin (1965, 1969 in Churkin, 1973), and others. Several authors (e.g. Gabrielse, 1967, 1973; Churkin, 1973; Eisbacher, 1978a) have suggested correlation of part of the upper Tindir Group with the Rapitan Group. Detailed field investigations (Allison et al., in press) support this.

In ascending stratigraphic order, the Upper Tindir Group comprises mafic volcanics, maroon rhythmites with iron formation, maroon mixtite, shales, and resedimented carbonates, and carbonaceous magnesian limestones (Allison et al., in press). This sequence is conformably overlain by Lower Cambrian carbonates and argillites (Brabb, 1967; Churkin, 1973). Due to complex faulting, a complete section through the Upper Tindir is nowhere exposed. The best section outcrops along the north bank of the Tatonduk River. The aggregate thickness of the Basalt-Redbed Member (Unit C of Mertie, 1932), the suggested Rapitan correlative, is estimated at 760 m (Brabb and Churkin, 1969 in Churkin, 1973). This member includes three subunits: amygdaloidal volcanics, maroon siltstone rhythmites with sandstone, conglomerate, and mixtite interbeds, and maroon mixtites. These various subunits are described below.

Volcanic Subunit

Mertie (1932) estimated the lavas to be up to 300 m thick. The lavas vary from massive to pillowed to agglomeratic. Individual flows are generally 1 to 2 m thick, with pillows up to 1 m in diameter, but flows up to 22 m thick with pillows as much as 4 m in diameter are reported (Mertie, 1932). The volcanics are greenish and weather reddish grey to brown. Copper staining is locally conspicuous.

Although the volcanics appear to overlie Lower Tindir carbonates with angular discordance, the actual contact is not exposed. Friable maroon mixtites with greenstone and carbonate megaclasts apparently overlie the lavas conformably. The relationship of this mixtite to the rest of the overlying clastics is obscured by faulting.

Rhythmite Subunit

Maroon rhythmites are well exposed along the Tatonduk River and on the mountain to the north; their thickness is greater than 300 m (Allison et al., in press). Three lithofacies occur: siltstone rhythmites, sandstone, and polymictic conglomerate. The lower part of the subunit is folded and faulted and the base is not exposed. Along the river the rhythmites interfinger conformably with the overlying mixtites but to the north they are overlain by mixtite in angular discordance. Petrographically, the rhythmites comprise lithic (carbonate) wackes and mudstones. The matrix is hematite-rich.

The siltstone rhythmites are laminated and parallel-bedded. Graded bedding, ripple crosslamination, flame structures, load casts, slump structures, and rip-up structures are present. In some beds classic Bouma sequences are present.

Isolated pebbles and cobbles are ubiquitous in the rhythmites. Commonly these lonestones are faceted and striated and appear to have been dropped into the laminated beds. Carbonate lonestones predominate, but greenstone and jasper clasts are also common. The finest grained mudstones are commonly so hematitic as to be iron formation and contain thin bands of red jasper.

Sandy beds in the rhythmite subunit are up to 0.5 m thick and vary from massive to graded, with typically planar or load casted basal contacts. Such coarser grained beds are most common towards the middle part of the rhythmite subunit.

The conglomerates of the rhythmite subunit range from pebble-rich zones to discontinuous pebble-cobble orthoconglomerate beds up to 15 cm thick. The conglomerates are polymictic with predominantly subangular to subrounded carbonate pebbles in a silty matrix. Some clasts appear to have depressed the underlying rhythmites, while others poke up into overlying sediment. A weak preferred orientation of clast-long axes is developed subparallel to bedding. Conglomerate beds become more abundant towards the top of the rhythmite subunit.

Mixtite Subunit

Maroon mixtites near Tatonduk River are more than 300 m thick and contain rare discontinuous orthoconglomerate beds. Poorly exposed mixtites at the top of the rhythmite subunit are greenish grey and are conformably overlain by grey shales and calcareous turbidites. The mixtites are crudely stratified. Rounded to angular, predominantly dolostone clasts occur throughout the clast-rich mixtites; minor greenstone fragments are also present. The clasts range up to boulders, but pebble-sized ones are most common. Faceted and striated surfaces are common on the clasts, which are oriented with their long axes subparallel to bedding. The gross matrix composition reflects the megaclast abundance. The matrix is hematite-rich in red mixtite; chlorite-rich in grey mixtite (Farfan, 1979).

Paleocurrents

Measurements of ripple crosslamination in the rhythmite subunit indicate strongly unimodal, west-northwesterly transport (Farfan, 1979). A southwesterly mean azimuth from crossbedding in turbidites above the mixtite suggests that the mixtite subunit was also derived from the east.

IRON FORMATION

Iron formation in the Rapitan Group was first reported by prospectors at the time of the Klondike gold rush (Keele, 1906). The presence of iron formation in Snake River area was noted by Camsell (1906). Keele (1910) described iron formation in the Sayunei Formation where the Keele River enters the Backbone Range (95 L). The Crest deposit east of Snake River was discovered in 1961 and an intensive program of exploration and evaluation was undertaken (Stuart, 1963; Green and Godwin, 1963). Exploration was undertaken in the southern Mackenzie Mountains as well (Condon, 1964). Reserves in the Crest deposit were estimated to be in excess of twenty billion tons of which six billion tons averaging 47.2 per cent iron might be extracted by open pit methods within a single ten square mile area (Stuart, 1963).

Sayunei Iron Formation

Laminated, concordant, jasper-hematite iron formation beds commonly occur in the upper part of the Sayunei Formation. Up to three separate beds may occur in one section. Beds may exceed a metre in thickness and persist laterally for several kilometres. Four forms of iron formation occur in the Sayunei Formation: laminated and nodular jasper-hematite, hematitic argillite, and locally, jasper-magnetite iron formation. All forms occur in close association.

Shezal Iron Formation

Iron formation occurs locally in the Keele River area near the base of the Shezal Formation. On Nite Mountain, west of Keele River (105 P), jasper-magnetite iron formation occurs in the lower part of the Bonnet Plume River mixtite. Lenses of nodular and irregular jasper-hematite iron formation occur occasionally in the Mountain River mixtite. Reworked fragments of iron formation are also locally common.

The principal iron deposits of the Rapitan are in the Snake River Tillite east of Snake River itself (106 F). The main iron-bearing zone at Snake River is about 150 m thick and lies about 150 m above the base of the Shezal. More than 10 iron-rich subzones up to 24 m thick can be recognized within the main zone (Stuart, 1963). These zones interfinger with mixtite and pinch out east and west of North Iron Creek (106 F). The iron formation is eroded to the north and appears to thin downdip to the south. In addition to the three forms of iron formation Stuart (1963) recognized at Snake River (laminated, nodular, and irregular), mixtite iron formation also occurs. Centimetre-scale banding in laminated iron formation could be traced 1.5 km at North Iron Creek (106 F). At Discovery Creek and on North Iron Creek thick clastic sills and dykes cut the iron formation.

Upper Tindir Iron Formation

Massive hematite iron formation was found in talus, but not in place, from volcanics which underlie hematitic rhythmites and mixtites. Laminated argillite iron formation is common in the rhythmite succession on Tatonduk River. The overlying mixtites may also be sufficiently ferruginous to be classified as iron formation.

Iron Formation Lithofacies

Laminated Iron Formation

This is the predominant form of iron formation in Sayunei Formation. It is characterized by alternating laminae of hematite and jasper. Either component may dominate to form jasper-rich or hematite-rich bands and/or lenticular patches up to several centimetres thick. Contacts between the bands are generally sharp and even, coincident with lamination, but contacts between patchy areas are typically diffuse and irregular. Laminae may be traced continuously across their lateral contacts. Irregular hematite-rich patches within jasper-rich areas suggest local diagenetic iron enrichment. Effects of hematite depletion in laminae and haloes adjacent to hematite-rich bands and patches appear as brighter, reddish orange, iron-poor jasper.

Authigenic calcite crystals up to 5 mm occur singly, in patches, and in laminations. These are concentrated in maroon, iron-rich, jasper bands, and are less common in the red, iron-poor jasper or in the hematite bands. Fine grained calcareous laminae are common in the hematitic bands and patches. Hematite pseudomorphs after dolomite are also present.

Locally, small-scale synsedimentary faults developed after partial segregation of banded silica and hematite. Silica structures preserved as intergrown, flattened jasper discs up to 1 cm thick and more than 20 cm in diameter were found at one locality.

Nodular Iron Formation

Nodular iron formation is less common in the Sayunei Formation than in the Snake River Tillite. Uniform and irregular red jasper nodules, typically less than 5 mm thick, occur scattered or clustered in laminated or massive hematite in beds from 1 to 20 cm thick.

The nodules have sharp margins and commonly contain two or three concentric zones which reflect internal variation in silica and iron content. Some nodules have hematite or calcite nucleii. Matrix laminae terminate abruptly against the nodules and enveloping laminae are commonly deflected around them. Flattening and smearing of the nodules was observed above and below lonestones. Nodules range from spheroids to discoids flattened parallel to bedding. Less commonly, dark grey chert nodules are developed in laminated red (iron-poor) jasper. Hematite-enriched pressure shadows occur adjacent to many nodules. Nodules may be intergrown in grape-like clusters which grade into lenticular structures.

Less resistant, laminated, maroon, calcareous, hematitic lenses and beds, generally less than 2 cm thick, are commonly interbedded in the nodular iron formation. Laminae may be traced across the sharp, commonly irregular contacts of these features into adjacent hematite or jasper-rich beds. Laminae are slightly deflected around many of the calcareous lenses. Such beds are much less common in the laminated iron formation.

Irregular Iron Formation

Irregularly-shaped jasper lenticles have many features in common with the nodules. The lenticles commonly form irregular masses and intergrowths of jasper in a massive or laminated hematitic matrix. Compositional zoning is common. Matrix laminae may be traced into the lenticles while enveloping laminae "flow" around them. The lenticles may be broken or bent. Less commonly, the hematite structures occur in a jasper matrix. Both nodules and lenticles may occur in the same bed. The size of a nodular or lenticular structure tends to be similar to other associated structures in the same bed.

Hematitic Argillite Iron Formation

The argillitic phase of the rhythmite couplets typical of the Sayunei Formation and Upper Tindir Group are locally so enriched in hematite as to be iron formation. This is the most common type of iron formation in the Upper Tindir Group. Similar iron formation has been described from Archean turbidite sequences (Dunbar and McCall, 1969; Shegelski and Scott, 1974). Such hematite-rich beds are commonly associated with other forms of iron formation, and are transitional to the laminated iron formation.

Hematitic Mixtite Iron Formation

Mixtite may also be so iron-rich as to be iron formation. This type of iron formation is relatively uncommon in the Rapitan Group. Mixtite iron formation has been described from late Proterozoic rocks of South Australia (Whitten, 1970).

GEOCHEMISTRY

Average compositions of Rapitan clastics and related rocks are shown in Table 2.1. Their variability and high degree of differentiation compared to normal sedimentary rocks can be shown by the wide range of alkali ratios and high silica-alumina ratios (cf. Garrels and Mackenzie, 1971, Fig. 9.1).

The minor transition metals, cobalt, nickel, copper, and zinc were analyzed as these are useful indicators of the influence of hydrothermal systems (Bonatti, 1975; Calvert, 1978; Toth, 1980).

The mean composition of shales from the Twitya Formation may be used as a norm to which the composition of the underlying sediments and iron formation might be compared. Except for low calcium, the major element composition of the Twitya shale is similar to widely quoted average mudstone values (e.g. Blatt et al., 1972, Table II-2). Trace element data for mudstones are relatively limited. However, the cobalt, copper, and nickel values for the Twitya shales fall within the range of values reported by Shaw (1954) for various mudrocks.

The silica-titanium correlation reflects the presence of silica in detrital form in the grey mixtites. The alumina-iron association reflects the presence of detrital magnetite as well. Significant negative correlations were found between silica and volatiles, alumina and manganese, iron and manganese, titanium and volatiles, and sodium and potassium. These negative correlations reflect an inverse relationship between abundance of carbonate and noncarbonate detritus. The sodium-potassium antithesis is more difficult to explain. Some sodium may be associated with carbonates in the form of authigenic albite.

Trace metal abundances were low. Compared to the Twitya shale, the grey mixtites are slightly depleted in copper and enriched in the other three trace elements. Correlation coefficients (Yeo, manuscript in preparation) suggest a significant positive association between zinc and both iron and nickel. A negative association between copper and phosphorus is indicated.

Geochemistry of the Red Clastics

Mean chemical analyses of red clastic sediments from the Rapitan and Tindir groups are also given in Table 2.1.

Except for higher manganese, the mean composition of red clastics from the Rapitan is not strikingly different from that of the grey clastics. Compared to the Twitya shale, the underlying clastics are strongly enriched in calcium and depleted in potassium. Analyses of Upper Tindir clastics suggest similar tendencies.

Mean analyses of iron formation from the Rapitan and Upper Tindir groups are also reported in Table 2.1. Compared to the red clastics, the Rapitan iron formation is enriched in iron and phosphorus and depleted in everything else but silica and manganese. Trace metal abundances are also lower. The Upper Tindir iron formation also shows iron and phosphorus enrichment compared to associated clastics. It is depleted in everything but titanium. Except for zinc, trace metal concentrations are also lower than in the associated clastics.

The Rapitan iron formation is notably depleted in nickel and zinc compared to the Twitya shale. It is slightly depleted in copper and slightly enriched in cobalt. The Tindir iron formation is increasingly depleted in zinc, nickel, and copper, and slightly enriched in cobalt.

Correlation coefficients (Yeo, manuscript in preparation) show that cobalt, nickel, and zinc are associated with one another and with alumina, magnesium, potassium, and titanium. Cobalt and nickel also show a positive correlation with sodium. They show a negative correlation with iron, however.

Unlike Algoman and Superior type iron formations (Gross, 1965b), but like Phanerozoic iron-rich chemical sediments, rare earth element patterns from the iron formation show europium depletion and resemble the modern seawater pattern (Fryer, 1977). However, the relatively low abundance of REEs is characteristic of earlier Proterozoic iron formation and unlike Phanerozoic chemical sediments.

ORIGIN OF THE CLASTIC SEDIMENTS

Although many earlier workers preferred a mass flow origin for the Rapitan mixtites, recent opinion (Young, 1976; Eisbacher, 1976, 1977, 1978a; and others) has favoured Ziegler's (1959) hypothesis that they are of glacial marine origin. Controversy over the deposition of the Sayunei rhythmites persists, however. Young (1976) suggested that they are glacial-marine turbidites with interbedded flow tillites and ice-rafted debris. A similar origin is suggested for the Upper Tindir redbed unit (Allison et al., in press). Eisbacher (1978a) supported a turbidity current origin for the rhythmites but recognized a glacial influence only in the upper part. He suggested that the abundant angular clasts, including lonestones, were derived from synsedimentary fault scarps and reworked by mass flow.

Evidence for Glaciation

The extensive distribution of the Rapitan and stratigraphic continuity upwards and locally downwards into marine sediments make a marine environment for Rapitan

Table 2.1. Average compositions of Rapitan clastics and related rocks

	Twitya		Shezal		Sayunei		Mount Berg		Grey Clastics		Red Clastics	
		C.V.%		C.V.%		C.V.%		C.V.%		C.V.%		C.V.%
SiO_2	55.9	(5.6)	46.1	(11.1)	47.8	(24.7)	51.3	(19.9)	46.4	(14.5)	48.1	(19.0)
Al_2O_3	17.7	(13.9)	9.7	(13.6)	10.4	(36.9)	13.2	(13.5)	11.3	(18.9)	10.0	(28.6)
CaO	0.4	(137.8)	9.3	(31.5)	9.0	(135.6)	4.3	(94.4)	7.8	(59.5)	8.8	(96.3)
MgO	3.4	(23.8)	7.2	(18.9)	5.2	(41.7)	6.7	(40.5)	7.5	(19.5)	5.9	(34.4)
Na_2O	0.8	(81.3)	0.9	(59.6)	2.1	(67.7)	0.2	(90.7)	0.9	(58.7)	1.3	(100.2)
K_2O	3.8	(25.8)	1.2	(51.6)	1.0	(66.0)	1.9	(42.1)	1.2	(44.7)	1.2	(63.6)
Fe_2O_3	9.1	(12.9)	10.3	(18.9)	12.4	(39.5)	12.1	(29.1)	10.9	(23.4)	11.5	(33.6)
MnO	0.1	(57.5)	0.2	(49.5)	1.0	(143.9)	0.1	(51.6)	0.1	(46.6)	0.6	(170.4)
TiO_2	0.9	(10.2)	1.2	(24.3)	0.9	(39.7)	1.4	(41.3)	1.3	(35.2)	1.0	(31.6)
P_2O_5	0.9	(27.0)	0.1	(22.7)	0.2	(61.3)	0.1	(28.6)	0.1	(22.0)	0.2	(65.7)
L.O.I.	7.0	(45.4)	12.7	(28.9)	9.8	(87.6)	8.6	(53.9)	11.4	(42.8)	11.0	(57.8)
Total	99.1		98.9		99.8		99.9		98.9		99.6	
Co	21	(32)	24	(32)	63	(130)	19	(13)	23	(33)	43	(138)
Ni	37	(29)	39	(31)	48	(99)	47	(43)	45	(37)	42	(80)
Cu	80	(47)	64	(42)	61	(133)	8	(13)	47	(71)	58	(103)
Zn	80	(11)	88	(30)	95	(74)	111	(29)	103	(28)	89	(57)
No. of Samples	4		10		6		3		7		12	

	Rapitan Clastics		Rapitan Chemical IF		Rapitan Clastic IF		Rapitan IF		Tindir Clastics		Tindir IF	
		C.V.%		C.V.%		C.V.%		C.V.%		C.V.%		C.V.%
SiO_2	47.5	(17.2)	60.6	(35.6)	42.7	(31.7)	45.3	(28.6)	41.7		38.4	(49.8)
Al_2O_3	10.5	(25.1)	0.5	(73.7)	5.2	(57.6)	3.4	(95.2)	5.0		3.8	(77.8)
CaO	8.4	(85.0)	0.9	(48.0)	3.2	(109.3)	2.3	(127.6)	17.3		3.8	(114.5)
MgO	6.5	(30.3)	tr	(125.8)	2.4	(64.7)	1.5	(113.5)	5.6		2.2	(75.7)
Na_2O	1.2	(94.8)	0.0	(0.0)	0.3	(158.2)	0.2	(212.9)	0.6		0.1	(200.0)
K_2O	1.2	(56.5)	tr	(99.0)	0.8	(138.4)	0.5	(181.7)	0.4		0.3	(116.8)
Fe_2O_3	11.3	(29.9)	33.7	(61.6)	39.9	(40.4)	41.5	(35.3)	6.2		45.9	(59.1)
MnO	0.4	(196.9)	0.1	(103.6)	0.1	(66.1)	0.1	(77.3)	1.5		0.7	(148.0)
TiO_2	1.1	(33.8)	tr	(106.3)	0.5	(49.7)	0.3	(100.0)	0.4		0.4	(70.8)
P_2O_5	0.2	(65.7)	0.2	(51.9)	0.5	(102.2)	0.4	(109.1)	0.1		0.5	(80.1)
L.O.I.	11.1	(51.3)	2.8	(139.1)	4.1	(59.7)	3.8	(88.3)	19.9		4.8	(91.7)
Total	99.4		98.8		99.7		99.3		98.7		100.9	
Co	35	(134)	4	(50)	43	(195)	28	(237)	33		28	(125)
Ni	43	(65)	7	(19)	17	(66)	13	(80)	18		12	(117)
Cu	54	(94)	48	(113)	82	(109)	62	(118)	9		5	(67)
Zn	94	(47)	13	(61)	40	(59)	29	(80)	30		31	(85)
No. of Samples	19		7		8		13		2		4	

Group deposition certain. Genetic interpretation of supposed ancient glacial deposits must be viewed with caution since criteria for distinguishing among Quaternary glacial deposits are still uncertain (Dreimanis, 1976). In fact, many concepts about modern glacial-marine sedimentation come from ancient deposits (i.e. Carey and Ahmad, 1960; Reading and Walker, 1966; and others).

Massive Mixtites

The prevalance of rounded, commonly faceted and striated stones in thick, crudely bedded, massive mixtites is the strongest argument against a simple mudflow origin. The lateral distribution and thickness of the Shezal mixtites as a lensoid ribbon up to a few tens of kilometres wide and hundreds of metres thick, as well as intimate association with other sediments having possible glaciogene features, support this. This distribution pattern resembles that of Quaternary glacial-marine deposits off the coast of Alaska (Molnia and Carlson, 1978). Thick-bedded, massive tills found here resemble the massive mixtites. They were attributed to ice-rafting by Miller (1953). The apparent scarcity of glaciodynamic structures in the massive mixtites favours a melt-out origin (Dreimanis, 1976), although the presence of slickenside surfaces locally in the matrix suggests some lodgment till deposition beneath grounded ice (Sugden and John, 1976; Krüger and Marcussen, 1976). The peculiar fissility characteristic of silty basal tills (Dreimanis, 1976) may be preserved in the scaly weathering typical of grey Shezal mixtites. Local development of strong clast fabrics also suggests basal till deposition (Marcussen, 1975; Krüger and Marcussen, 1976), but these can also be interpreted as transverse fabrics developed during mass flows.

Stratified Mixtite

The stratified mixtites may also be produced by more than one mechanism. The well-stratified kind typical of parts of the Snake River Tillite resemble stratified tills ascribed to subaqueous till flow by Marcussen (1973), May (1977), Evenson et al. (1977) and Hicock et al. (in press). The more crudely stratified kind may be formed by iceberg rafting (Lavrushin, 1968; Ovenshine, 1970) or, less likely, smearing of soft material beneath overriding ice (Krüger, 1979, Fig. 10).

Rhythmites

Rhythmite deposits containing ice-rafted material are common in Quaternary (Banerjee, 1973) and ancient glacial deposits (Banerjee, 1966; Rattigan, 1967; Reading and Walker, 1966; McCann and Kennedy, 1974; Nystuen, 1976). The Sayunei rhythmites have features typical of turbidites (Young, 1976; Eisbacher, 1976, 1978a). The predominance of base-cut-out Bouma sequences indicates that they are distal turbidites, corresponding to Facies D of Walker and Mutti (1973; cf. Reading and Walker, 1966, Fig. 12). The occurrence of mixtites below the rhythmites in the southern Mackenzie Mountains (i.e. Mount Berg Formation), the ubiquitous lonestones, including occasionally striated and faceted dropstones, and the occurrence of possible till pellets in the rhythmites suggest that glaciation began before, and persisted during rhythmite deposition. The sediment source for the rhythmites was likely at or near the glacier terminii.

Eisbacher (1978a) suggested that lonestones in the rhythmites, particularly in the lower part, might be individual clasts isolated from low-viscosity slurry flows which gave rise to the massive sharp clast siltstone beds. This, however, does not explain the common occurrence of lonestones piercing underlying rhythmites (Young, 1976, Plate 1C, D). Such lonestones were most likely dropped in from floating ice.

Conglomerates and Sandstones

Currents and flow mechanisms which deposited the sandstones and conglomerates must have ranged in competence widely, reflecting variation in sediment supply and fluctuations of the ice front.

Sandstone beds with scoured bases, rip-up clasts, planar and crosslamination are interpreted as products of turbulent traction currents. Sandstone beds with non-erosive bases, planar tops, and lack of internal structure except for grading were probably produced by grain flow.

Highly organized conglomerates (e.g. imbricate Sayunei sharpstones of Eisbacher, 1976) with channeled lower contacts were transported as bed load by strong bottom currents. Less well-organized massive or graded conglomerates were probably deposited by inertia flow.

In the Rapitan Group, well-stratified mixtites, conglomerates, sandstones, and rhythmites form a spectrum of resedimented lithofacies (Middleton and Hampton, 1973, Fig. 10). Inertia flow mechanisms predominated close to the sediment source while turbulent flow was increasingly important distally.

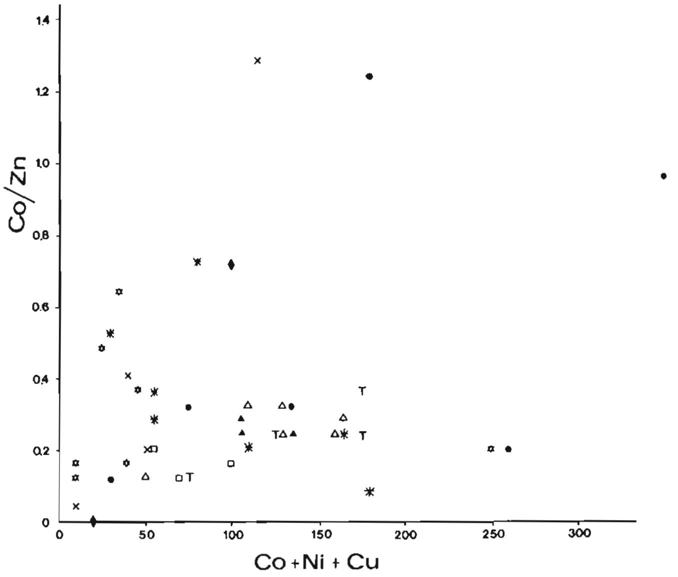
ORIGIN OF THE IRON FORMATION

There is little doubt that iron formation is a chemical precipitate. Both hydrogenous and hydrothermal precipitation have been proposed to explain the Rapitan Group iron formations. Condon (1964) suggested that iron weathered from penecontemporaneous volcanic rocks was transported in suspension as colloidal ferric oxide in freshwater streams, and precipitated on entering the sea. The presence of pseudomorphs after evaporite minerals (i.e. dolomite and glauberite) suggested an evaporitic mechanism to Young (1976), who speculated that iron and silica might have been precipitated from brine solutions formed when large quantities of seawater froze to the base of a cold-based ice shelf (Carey and Ahmad, 1960). Gross (1965a, 1973) proposed a hydrothermal origin by which iron and silica precipitated from fumarolic waters discharged along submarine fault zones. This last model seems most likely.

Iron-rich sediments of hydrothermal origin may be distinguished from those of hydrogenous origin by numerous geochemical features. Hydrothermal iron-silica precipitates are characterized by high SiO_2/Al_2O_3 and high Fe/Ni+Co+Cu (Bonatti, 1975); high Fe/Al+Mn and Ni/Cu < 1 (Calvert, 1978); low Co/Zn and Co+Ni+Cu abundance (Toth, 1980). The Rapitan and Upper Tindir iron formations exhibit all of these features. The low REE abundance reported by Fryer (1977) is also typical of hydrothermal precipitates (Calvert, 1978; Toth, 1980). For brevity only the minor transition metal relationships are shown here (Fig. 2.6).

Source of the Iron

The great volume of the iron formation in association with sediments that must have been relatively rapidly deposited requires a system capable of quickly producing an immense quantity of iron and silica. Recent experimental studies (e.g. Bischoff and Dickson, 1975; Seyfried and Bischoff, 1977) and investigations of oceanic hydrothermal deposits associated with active ridges (Dymond et al., 1973; Dymond and Veeh, 1975; Corliss et al., 1978, 1979; Toth, 1980; Bischoff, 1980), volcanic buildups (Butusova, 1966; Puchelt, 1976), and non-volcanic fracture systems (Castellarin and Sartori, 1978) confirm and refine models for hydrothermal metal deposits proposed by Corliss (1971), Bonatti (1975), and others. Such a model best explains the source of the Rapitan iron.



T - Twitya shale
Δ - Shezal mixtite
o - Sayunei mudstone
□ - Mount Berg mixtite
** - Sayueni iron formation*
✡ - Shezal iron formation
x - upper Tindir iron formation
Solid symbols - red rocks
Open symbols - grey-coloured sedimentary rocks

Figure 2.6. *Relative abundances of minor transition metals in clastic and chemical sedimentary rocks of the Rapitan Group and related rocks. The low Co/Zn ratios and/or low Co+Cu+Ni abundances shown here suggest a hydrothermal source (c.f.: Toth, 1980, Fig. 6).*

Gross (1965, 1973) suggested that iron-bearing hydrothermal solutions were discharged along fractures possibly related to contemporaneous volcanism. He proposed that hematitic stockworks and veins west of Snake River might be remnants of the fumarolic systems. Bell (1978) echoed this hypothesis and showed that these ferruginous breccia complexes were fault controlled. However, brecciation appears to pre-date Rapitan sedimentation (Delaney, 1981). Locally abundant volcaniclastic material reported in the Rapitan sediments (Condon, 1964; Gross, 1973) suggests contemporaneous volcanic activity. Pillow lavas interfinger with possible Shezal equivalents in the southern Ogilvie Mountains (Green, 1972), but no iron formation was reported there. However, volcanics in the Upper Tindir Group in western Ogilvie Mountains are associated with ferruginous sediments and iron formation. This, plus the evidence for a dilational tectonic regime proposed by Eisbacher (1977), support the probable existence of a hydrothermal system related to volcanism and rifting (Fig. 2.7).

Development of the Redbeds

Until recently it was generally thought that the abundance of organic material and consequent reducing conditions in most marine sediments made marine redbed formation unlikely (Berner, 1971; Van Houten, 1973; Walker, 1974; and others). Many examples of marine redbeds are now known. They may be formed by resedimentation and rapid burial of previously oxidized sediment (Lajoie and Chagnon, 1973; Ziegler and McKerrow, 1975; Francke and Paul, 1980), submarine weathering of ferruginous rocks (Muller, 1967; Fairbridge, 1967; Scott and Hajash, 1976), or by hydrothermal activity (Davies and Supko, 1973). Association with chemical iron formation indicates that enrichment in iron and silica in the red sediments was due to ongoing precipitation during clastic deposition. The early development of red pigmentation is indicated by restriction of zones of diagenetic reduction within the redbeds to coarse grained, permeable beds and adjacent to carbonate clasts. It is likely that the red coloration is a primary feature produced by the same process that formed the iron formation.

DEPOSITIONAL HISTORY

Basin Configuration and Tectonics

In Mackenzie Mountains glacioclastic sediment formed a lensoid belt generally less than 30 km wide, parallel to rising eastern source areas. This uplift is indicated by eastward-increasing unconformity as well as preservation of ancient fault scarps downthrown to the west. Thick sediment accumulations in subbasins nearly coincident with embayments developed in Mackenzie Mountains Supergroup time (Jefferson, oral communication, 1980) support evidence for continued fault-controlled, basin subsidence during deposition and suggest that major outlet glaciers debouched into these subbasins. In the northern Mackenzie Mountains a centripetal transport pattern suggests a major embayment open to the south and bounded to the west by a probable high. In the western Ogilvie Mountains, generally westerly transport suggests a rising source area to the east for the Upper Tindir Group clastics.

Mount Berg Formation

The Mount Berg mixtites were probably deposited in front of or beneath floating ice. At this time mountain glaciers may have existed locally in the rising Mackenzie uplift area or, if a cratonic ice sheet was already developed, only isolated outlet glaciers had pierced the marginal barrier uplift.

Sayunei Formation

Sayunei rhythmites interfinger with the underlying Mount Berg mixtites and overlying Shezal Formation. The rhythmites are at least partly the distal equivalents of the mixtites. In addition to material carried away from the ice margin by turbidity currents and mass flows, considerable detritus was deposited by melting and overturning icebergs. The thickness and distribution of the rhythmites suggest that much deposition took place in locally subsiding subbasins, perhaps at the terminii of outlet glaciers. The relative angularity of many Sayunei megaclasts may result from transport as high level load rather than as basal load within the ice. This might be due to extensive exposed uplands shedding debris onto the surface of the glaciers or cold-base glacial conditions inhibiting plucking and abrasion of clasts at the glacier sole (Boulton, 1978).

Iron and silica, derived from nearby hydrothermal systems in an active marginal basin must have been carried in slightly acid and relatively warm countercurrents into the area of clastic deposition where they were precipitated as ferric hydroxyoxides(?) and silica gel upon mixing with cold, oxygenated, weakly alkaline water. This hypothesis accounts for the iron-silica enrichment of the redbeds. During periods of relatively low clastic influx iron formation could accumulate. Manganese, much more soluble than iron and silica, must have been carried out of this environment and deposited elsewhere.

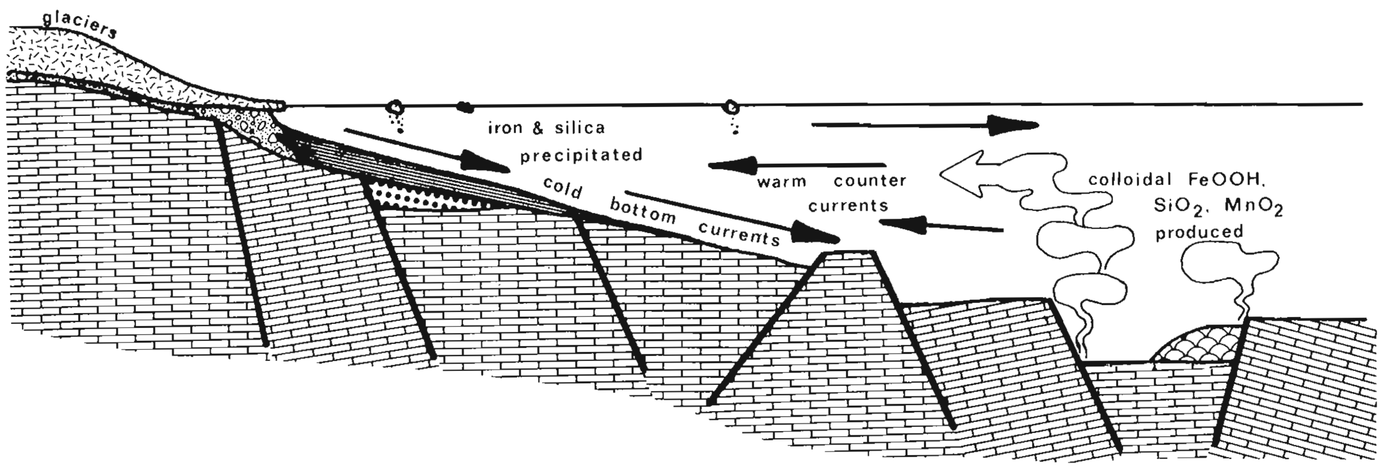


Figure 2.7. Hypothetical model for precipitation of iron and silica from a hydrothermal seawater system with glacioclastic sediments.

Shezal Formation

The Shezal mixtites are interpreted to be principally aquatillite (Schermerhorn, 1966) and flow tillite (Boulton, 1968) laid down in front of or beneath floating ice. Extensive ice shelves probably existed. The largest of these floated in the Snake River area, anchored to the east and west of the Mackenzie and Wernecke uplifts. Sea level must have fallen during Shezal time as the depoaxis of the Shezal mixtites lies somewhat basinward of that of the distal facies equivalent Sayunei rhythmites and because Sayunei material, including rounded clasts of iron formation, was locally reworked into the Shezal. The great quantity of iron formation in the lower Shezal near Snake River was deposited in the same manner and probably at the same time as the Sayunei iron formation. The transition from redbed mixtites and rhythmites to grey mixtites may reflect either cessation of hydrothermal activity or change in the circulatory system. This also may have been a synchronous event throughout the Rapitan basin.

Upper Tindir Redbeds

Deposition of the Upper Tindir rhythmite and mixtite redbeds and iron formation was analogous with that of the Rapitan. Deposition of the Rapitan and Upper Tindir redbeds was probably contemporaneous.

CONCLUSIONS

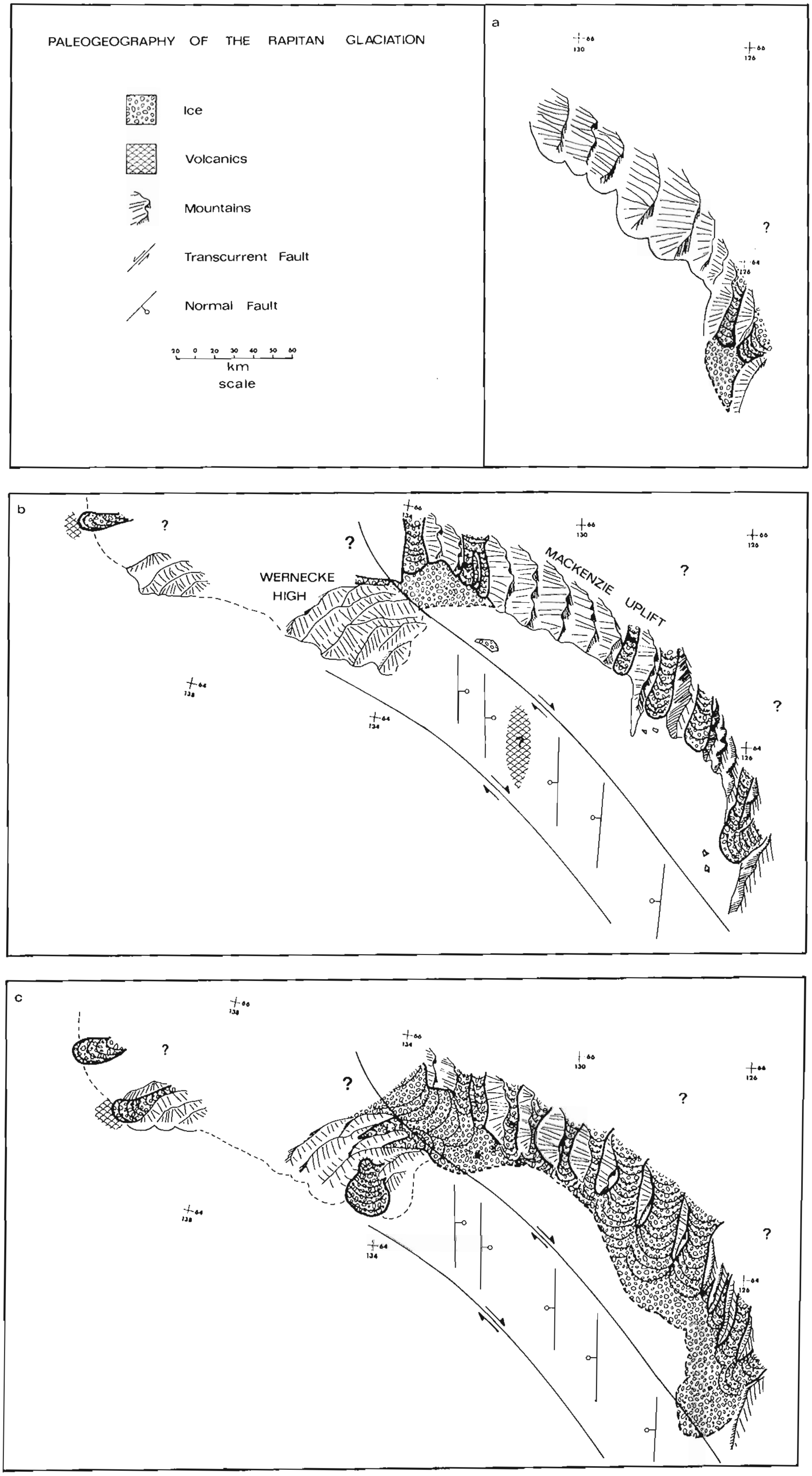
In the northern Cordillera, thick, widespread mixtites and rhythmites were deposited under glacial marine conditions along an evolving continental margin in late Proterozoic time. The mixtites were deposited mainly by basal meltout from floating ice and by mass flow either directly from the ice or by remobilization of meltout till. The rhythmites are distal facies equivalents of the mixtites laid down by turbidity currents. Together with iceberg-rafted material, other sediment-gravity flow deposits are intercalated with the rhythmites. These sediments were deposited from the calving terminii of outlet glaciers and ice shelves as a lensoid ribbon along a rising paleocoast subparallel to and cratonward of the subsequently developed Silurian platform-basin transition line (Tipper, 1978).

The geochemistry of the iron formation suggests a hydrothermal source, probably controlled by a regional fracture system which controlled the shape of the basin margin. Volcanics in the Upper Tindir Group may also be a product of rifting. The red colour of much of the Rapitan and equivalents must result from the same mechanism that produced the iron formation.

All along the eastern edge of the North American Cordillera late Proterozoic rocks lap onto strata of middle and late Proterozoic age. The contact is generally a major unconformity which may reflect considerable uplift (e.g. Lis and Price, 1976). Mixtites commonly occur at the base of the youngest Proterozoic succession. They overlie a thick stable platform succession of carbonates and relatively mature clastic rocks and are overlain by mudstones and carbonates. In the Upper Tindir and Rapitan groups, and in the Kingston Peak Formation of eastern California, banded iron formation is intimately associated with the mixtites. The overlying rocks are interpreted as a passive continental margin clastic wedge deposited consequent to an extensive rifting event (Stewart, 1972; Young et al., 1979; and others).

The upper Proterozoic tectono-stratigraphic assemblage outlined above is not unique to western North America. It may be found along the margins of Precambrian cratons on nearly every continent (Anonymous, 1976; Dunn et al., 1971; Kröner, 1979; Mendes, 1971; Salop, 1977; Schuller and Szu-Haui, 1959; Spencer, 1975). This indicates widespread similarity of tectonic and climatic conditions in the late Proterozoic although the relative ages of these deposits are still poorly known.

The arguments over glacial or tectonic control of late Proterozoic mixtite deposits can be reconciled. Uplift and reduced continentality are among the factors suggested as causes of widespread glaciation (Harland and Herod, 1975). They are known to be extremely sensitive parameters of glaciation (Sugden and John, 1976). Unlike more exotic hypotheses concerning ancient glaciations, a hypsographic model can easily be confirmed. Development of fault controlled intracractonic and marginal basis accompanied widespread uplift along basin margins before and during deposition of the upper Proterozoic mixtites (Schermerhorn, 1974; and others). Rising highlands and development of new seaways near the end of the Precambrian favoured widespread glaciation. Other factors may also be involved, but their effect is less certain.



a. Deposition of the Mount Berg mixtites was apparently restricted. These sediments may be remnant products of localized mountain glaciers or an early outlet glacier breaching the barrier of an uplifted craton margin.

b. Numerous outlet glaciers must have developed during Sayunei time. Glacioclastic rhythmites were deposited outward from these. Proximal mixtites were probably eroded during subsequent relative uplift, except in the Snake River area where an ice shelf developed in a major embayment. Deposition of iron and silica at this time indicates an active hydrothermal system with possible submarine volcanism. Pillow lavas are associated with Rapitan correlatives in the upper Tindir Group of eastern Alaska.

c. The relative fall of sea level in Shezal time allowed the mixtite facies to extend basinward over the rhythmites. Widespread ice shelves probably developed. With the end of this glacial period sea level rose and passive craton margin sedimentation resumed (i.e., Twitya shale and equivalents).

Figure 2.8. *Tentative paleogeography of the northern Cordillera during the late Proterozoic Rapitan glaciation. Basin tectonics after Eisbacher (1977).*

Global development of dilational fault systems, commonly accompanied by volcanism, also helps explain the appearance of widespread iron formation near the end of the Precambrian. These iron formations are commonly associated with mixtites. The geochemistry of many of these is consistent with a hydrothermal origin like that proposed for the Rapitan and Tindir iron formations (Yeo, manuscript in preparation).

Extensive glacial-marine deposits and iron formation in the late Proterozoic, typified by the Rapitan Group and the correlative Upper Tindir redbeds, can be viewed as manifestations of the transition in crustal evolution from the Proterozoic stage, characterized, with a few important exceptions, by extensive stable platform conditions, to the Phanerozoic stage, characterized by widespread interplate activity.

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3. STRATIGRAPHY AND SEDIMENTOLOGY OF THE UPPER PROTEROZOIC LITTLE DAL GROUP, MACKENZIE MOUNTAINS, NORTHWEST TERRITORIES

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Aitken, J.D., Stratigraphy and sedimentology of the Upper Proterozoic Little Dal Group, Mackenzie Mountains, Northwest Territories; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 47-71, 1981.

Abstract

The Little Dal Group (0.7-1.0 Ga), consisting mainly of non-clastic strata, accumulated uninterruptedly in an epicratonic setting. Depositional strike follows the concave-southwestward tectonic arc of Mackenzie Mountains. Southwestward thickening is demonstrable for most formations.

Early in the history of the group, facies differentiation was pronounced, between a high-energy carbonate platform and a basin (epicratonic), with the facies boundary cutting across the long-term isopach trend. Once depositional relief was smoothed by sedimentation, it did not recur to any marked degree.

Lithofacies demonstrate control by: (a) pulses of terrigenous detritus from a southeastern source, that define crudley chronostratigraphic units; (b) dominance of longshore transport of detritus in the littoral zone; (c) repeated cutoff of detrital input, coincident with marine transgressions.

Refinement of Little Dal stratigraphy permits delineation of an unconformity within strata formerly assigned to the group, here recognized as the base of the overlying "copper cycle".

Résumé

Le groupe de Little Dal (0,7-1,0 Ga) comprend principalement des couches non-clastiques qui se sont accumulées de façon ininterrompue dans un contexte épicratonique. La direction de la sédimentation suit l'arc tectonique dont la concavité est orientée vers le sud-ouest des monts Mackenzie. L'épaississement vers le sud-ouest de la plupart des formations peut être démontré.

Depuis le début de l'histoire du groupe, les faciès étaient bien marqués, allant d'une plate-forme carbonatée de haute énergie à un bassin (épicratonique), dans lequel la limite des faciès coupe la direction des isopaques à long terme. Une fois que le relief de dépôt a été adouci par la sédimentation, il n'a pas réapparu sous une forme marquée.

La limite des lithofaciès a été démontrée par: (a) des apports de roches détritiques terrigènes du sud-ouest, qui définissent approximativement des unités chronostratigraphiques; (b) la dominance d'un transport littoral de roches détritiques dans la zone littorale; (c) des arrêts répétés de l'apport de roches détritiques correspondant à des transgressions marines.

Le raffinement de la stratigraphie de Little Dal permet de délimiter une discordance dans les couches attribuées antérieurement au groupe; qui a été reconnu ici comme étant la base du "cycle de cuivre" sus-jacent.

INTRODUCTION

The structural culmination of the Mackenzie Fold Belt, extending from north Nahanni River northward and westward around the Mackenzie Arc nearly to Snake River (Fig. 3.1) provides widespread exposures of the "Mackenzie Mountains supergroup"[1], a thick and virtually conformable succession that includes the oldest strata exposed in the fold belt (Fig. 3.2). The upper part of the supergroup forms a natural subdivision. It consists mainly of nonclastic strata, and is bounded beneath by a thick, clastic-dominated succession (Katherine Group, Tsezotene Formation) and above by unconformities on which lie various formations of Hadrynian to Devonian age. This easily recognized, upper subdivision of the "Mackenzie Mountains supergroup" is the Little Dal Group (Gabrielse et al., 1978).

The writer's continuing studies of Mackenzie Mountains geology, commenced in 1969, have repeatedly focused on the Little Dal, for the following reasons:

a. Widespread misunderstanding of the relationships between isolated sections and the type section, and the lack of effective criteria for distinguishing between the lower and upper divisions of the group, have led to mismapping of a number of major structures.

b. Because of the widespread exposures of the Little Dal, its lithology and thickness and their variations provide an ideal opportunity to assess the tectonic regime in late Proterozoic time.

[1] "Mackenzie Mountains supergroup" lacks formal status at this date, although it already has been widely employed. As informally proposed by Young et al., (1979, in press), it encompasses at its base the oldest strata exposed in Mackenzie Mountains (map unit H1), and at its top the Redstone River and Coppercap formations. Recognition in this paper of an unconformity at the base of the "copper cycle" (base of Redstone in some sections, lower in many) suggests that when "Mackenzie Mountains supergroup" is formalized, it may be advisable to exclude strata of the "copper cycle".

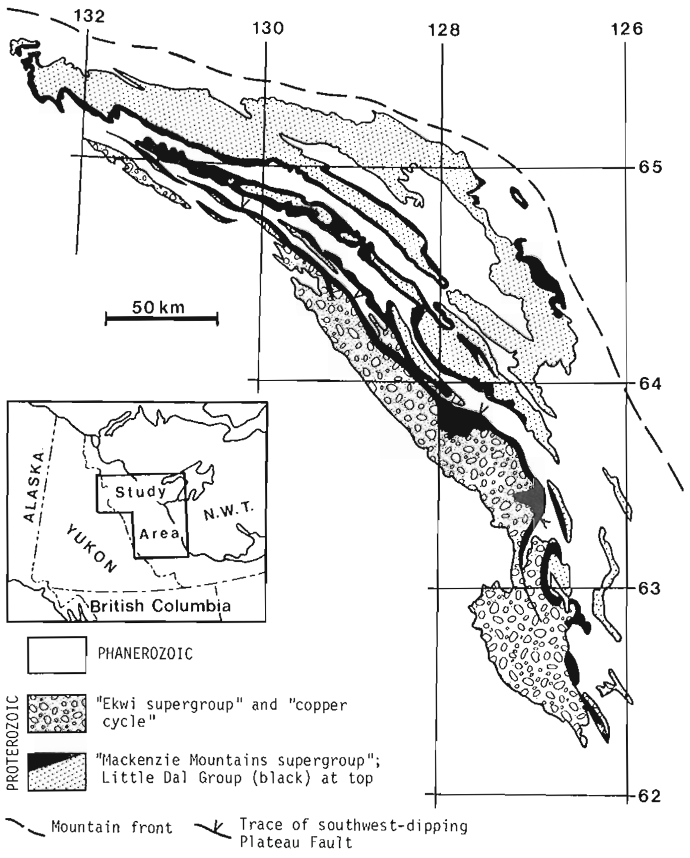


Figure 3.1. *Index map, showing distribution of Proterozoic rocks.*

c. The Little Dal is of economic importance in that it contains important showings of zinc, and forms the floor of the Redstone Copper Belt.

d. Many of the individual rock types found in the Little Dal, and some of the algal stromatolites in particular, are easily recognized as clasts in younger conglomeratic formations, and are thus useful as provenance indicators.

e. The Little Dal carbonates are important parts of the structural skeleton of the Mackenzie Mountains, and the Gypsum formation has acted as a level of major structural detachments.

f. The Little Dal contains abundant microfossils, and some of the oldest known macrofossils; analysis of the life-environment of both may contribute to evolutionary biology.

g. Its varied and abundant stromatolites, coupled with increasing knowledge of its absolute age, will provide another test of the validity (or otherwise) of the use of stromatolite assemblages in intercontinental biostratigraphic correlation (work in progress by M.A. Semikhatov).

Thus, the Little Dal has been singled out for analysis in this symposium volume. In view of the limitations of space, the treatment is long on interpretation and short on data; even so, only the major highlights of depositional history can be dealt with.

STRATIGRAPHIC RELATIONS AND AGE OF THE LITTLE DAL GROUP

The Little Dal is either an upper, or the uppermost major subdivision of the "Mackenzie Mountains supergroup" (Fig. 3.2), depending on the choice of upper limit for the latter, as discussed later. In its northwestern exposures, it rests with apparent conformity on the uppermost unit of the Katherine Group (Aitken et al., 1978), whereas in the southeast, the contact with the same unit is obviously erosional.

The upper contact of the Little Dal, as amended here, is a newly recognized unconformity at the base of the proposed, informal "copper cycle". The "copper cycle" encompasses the Redstone River and Coppercap formations, and with them, an underlying cyclical unit with sandstones, formerly assigned to the Little Dal. The basaltic lavas at the top of the Little Dal Group are also tentatively transferred to the "copper cycle".

Virtually all who have written about the Proterozoic rocks of Mackenzie Mountains have recognized the parallels between the northern and southern Proterozoic successions of the Cordillera. In both regions, a platform assemblage rich in carbonate rocks (Purcell Supergroup in the south, "Mackenzie Mountains supergroup" in the north) is unconformably overlain by a clastic-dominated assemblage, partly of slope and deepwater origin and containing glacial deposits (Windermere Supergroup in the south, "Ekwi supergroup"[1] in the north). Making due allowance for the dearth of radiometric dates, virtually all authors, including the writer, have assumed broad stratigraphic equivalence between the lower and upper assemblages of each region (Gabrielse, 1972; Stewart, 1972; Eisbacher, 1978). A dissenting view was taken by Young et al. (in press), who suggested limiting ages of 1.2 and 0.7 Ga for the "Mackenzie Mountains supergroup", although the younger limit was based on K-Ar dates that may be open to question, and on assumed relationships between intrusive and extrusive basic rocks. Evidence is accumulating that the upper part, at least, of the "Mackenzie Mountains supergroup" may indeed be younger than the youngest Purcell.

Paleontological evidence suggests an age of 0.9-1.0 Ga for the Little Dal Group. A change in stromatolite assemblages that occurs low in the Upper Carbonate formation corresponds to a change at the base of the Upper Riphaean of the USSR, dated at 950 ± 50 Ma (see Aitken et al., 1978). Hofmann and Aitken (1979) concluded that the micro- and macroflora of the Basinal assemblage ("Basinal sequence") suggested an age of 1.1-0.8 Ga. That estimate was strongly influenced by the presence of **Chuaria**, now known to range back into much older rocks in China, and must be reconsidered in that light. Nevertheless, the macrofossil **Tawuia** Hofmann and Aitken (1979), which has recently been found in the Rusty Shale and Upper Carbonate formations as well as in the Basinal and Platform formations, is now known from strata in China dated at 0.9 Ga (Duan, ms). The implied correlation is consistent with other evidence for the age of the Little Dal, although the total range of **Tawuia** is known neither in North America nor in Asia.

The implications of the extremely complex paleomagnetism of some of the Little Dal strata and the igneous rocks associated with them are discussed by Morris and Park (1981). A paleopole for strata low in the Little Dal, determined by Park (in press), agrees closely with poles from the Grenville Province dated at around 0.95 Ga, but also falls on a younger, tentative, polar-wandering curve at about 0.77 Ga. Paleopoles from diabase sills intruded into the Tsezotene Formation are close to the Little Dal poles, and similarly ambiguous as to implications of age. The younger age, however, is consistent with the Rb-Sr isochron age of

[1] Term informally proposed by Young et al. (1979, in press), and already widely used.

about 0.77 Ga reported for the sills (Armstrong et al., in press). Diabase dykes assumed to be related to the sills cut the Little Dal Group as high as the Grainstone formation, and are believed to be the source of diabase clasts in the Shezal (middle Rapitan) Formation.

A further, important outcome of the paleomagnetic work is that the magnetization of the lavas at the top of the Little Dal Group is significantly different from that of the diabase sills (Morris and Park, 1981). The intrusive and extrusive rocks record two different events, not one.

A final implication of the new, younger, apparent dates is that the onset of Rapitan glaciation in Mackenzie Mountains, post-0.77 Ga, is significantly younger than the estimates of 0.8-0.85 Ga for the onset of deposition of the Toby diamictite at the base of the classical Windermere Supergroup of the southern Cordillera (Gabrielse, 1972; Stewart, 1972; Miller, 1973).

STRATIGRAPHY AND SEDIMENTOLOGY

Overview

The strata of the Little Dal Group are generally well exposed, fresh and unmetamorphosed. The widespread dolomitization has commonly caused little loss of the elements of depositional fabric critical to sedimentological interpretation. Destruction of depositional fabric has occurred mainly in strata subjacent to the sub-Rapitan unconformity; however, the distribution of this alteration (weathering) is in itself significant to the interpretation of the Proterozoic history of the region.

A persistent problem is the geographic distribution of exposures (Fig. 3.1, 3.3). With local exceptions, the upper Little Dal has been stripped from the region north and east of Plateau Fault by erosion at the sub-Rapitan, sub-Cambrian, sub-Upper Cambrian, and possibly other unconformities.

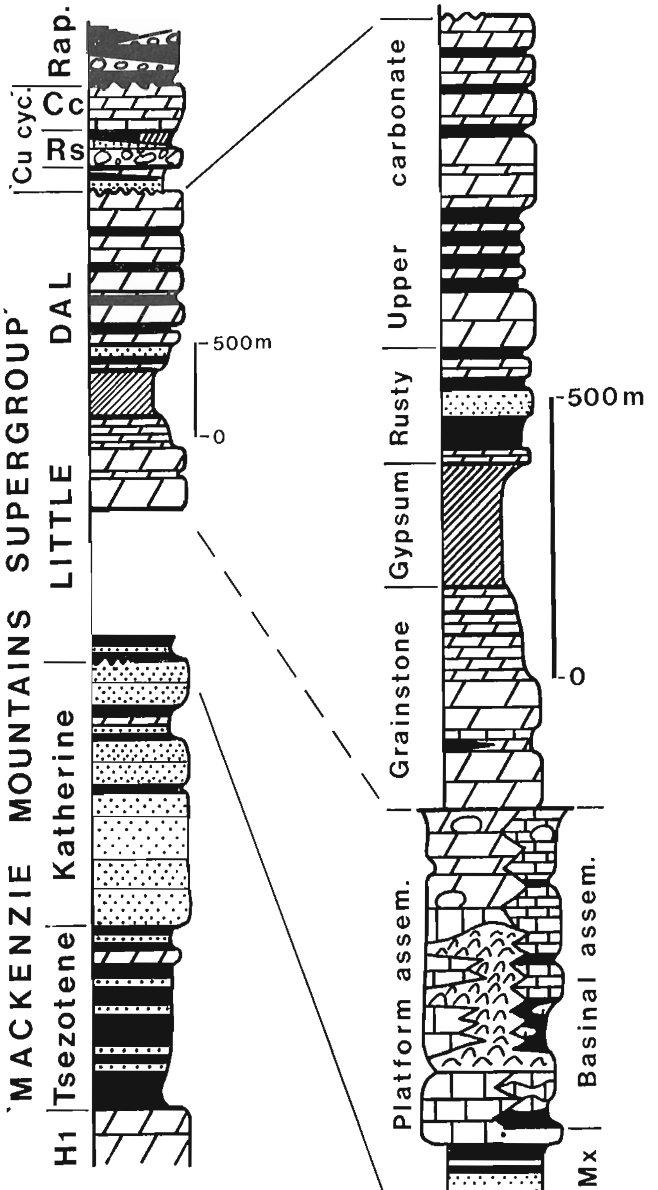


Figure 3.2

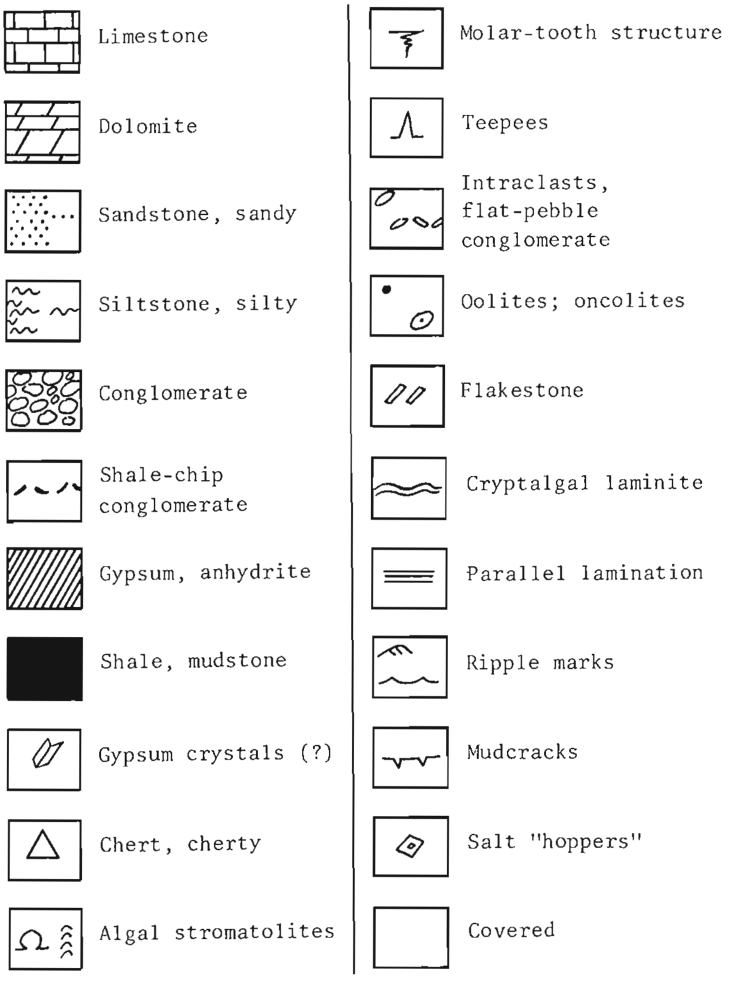


Figure 3.2a

J.D.A. 80

Figure 3.2a. *Key to lithologic symbols.*

Figure 3.2. *'Mackenzie Mountains supergroup' and Little Dal Group; nomenclature and gross lithology. For key to lithologic symbols, see Figure 3.2a.*

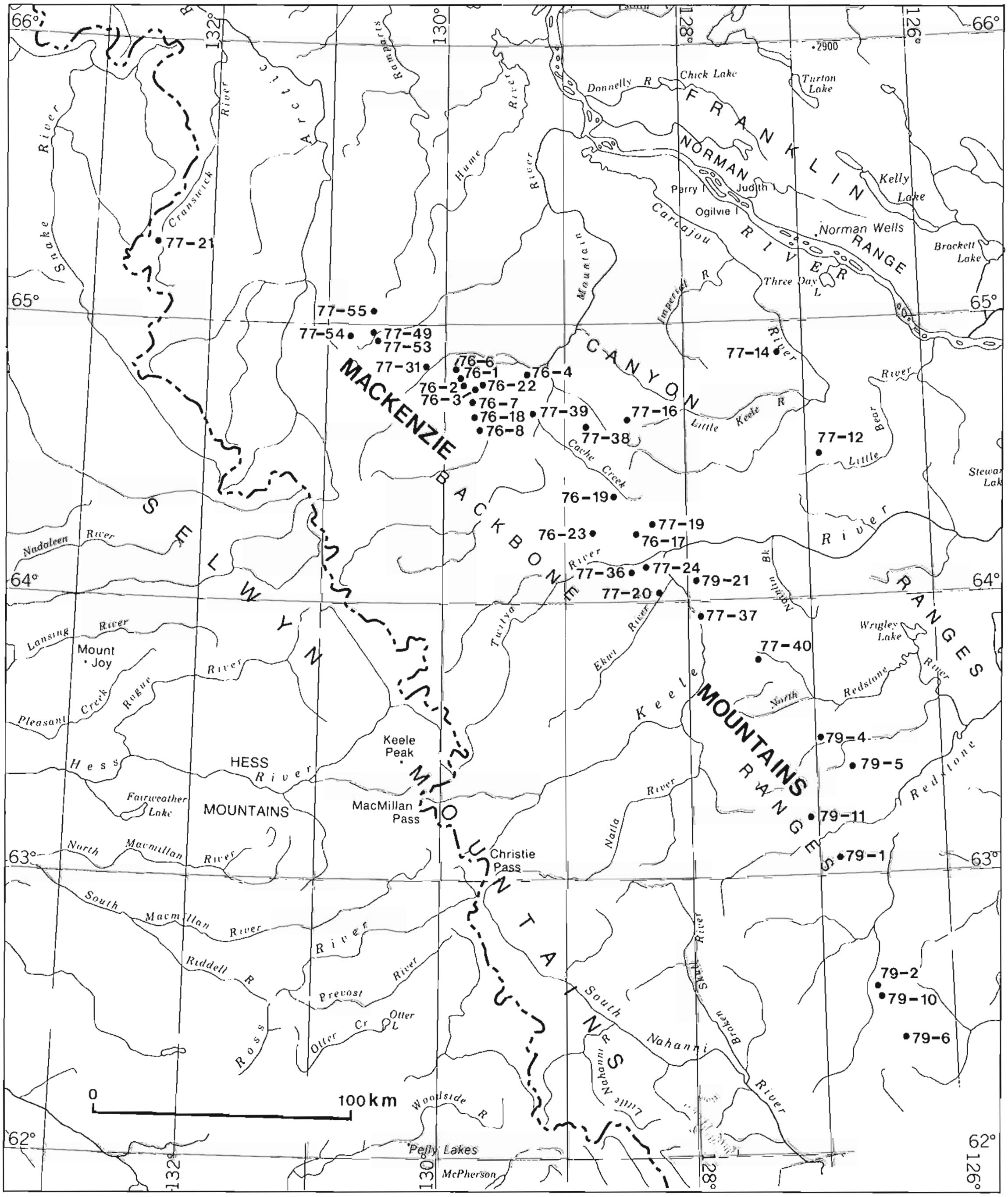


Figure 3.3. *Index to localities*

A further effect of this stripping is that the lower Little Dal reaches the mountain front only within the transverse structural sag between Keele and Carcajou rivers. On the other hand, the Plateau Thrust Sheet (Fig. 3.1) is structurally detached from older strata at the level of the Gypsum formation (upper Little Dal), and thrust northeastward over younger strata. The Plateau Thrust thus marks the southwestward limit of observation for the lower Little Dal. In consequence, the stratigraphic sections that are the data base for this report lie along a band about 350 km long and at most 100 km wide (much narrower for the upper part). Isopachs based on data points so distributed can be drawn "objectively" or mechanically in many different ways. By drawing on data other than thickness alone, some units can be "isopached" subjectively so as to present a reasonable and harmonious configuration. For other units, however, it is not even certain whether depositional thickening is away from or toward the craton.

Subdivision of the Little Dal Group

The Little Dal Group is divided here into six informal lithostratigraphic units of formational rank as follows (see Fig. 3.2):

'Copper Cycle'		Coppercap Formation Redstone River Formation Basaltic lava and unnamed member
Unconformity		
Little Dal Group	Upper[1]	Upper Carbonate formation Rusty Shale formation Gypsum formation
	Lower[1]	Grainstone formation Basinal and Platform assemblages, jointly Mudcracked formation
Unconformity		
KATHERINE GROUP		

Most of these units have been referred to previously as subunits of "Map unit H5" (now obsolete), with the same descriptive names (Aitken, 1977; Aitken et al., 1978).

The Basinal and Platform assemblages are lithologically and geographically distinct, mutually exclusive facies of a stratigraphic interval bounded by roughly time-parallel contacts. Consequently, their future formalization as equivalent formations will be appropriate.

Type Section of the Little Dal

The type section of the Little Dal Group is composite (Gabrielse et al., 1973). It was measured by its authors and remeasured by the writer and R. Booker on westward-extending spurs from a dominant north-south ridge, the lower part 10 km, the upper part 15 km south-southeast of Coates (Little Dal) Lake, in Glacier Lake map area (95L).

The type section contains the following units:

Sayunei Formation (Rapitan Group)
  unconformity
LITTLE DAL GROUP:
  Upper Carbonate formation
  Rusty Shale formation
  Gypsum formation (covered; gypsum outcrops along strike N and S of section)
  Grainstone formation
  Platform assemblage (basal, shaly, recessive unit is thin equivalent of Mudcracked fm)
  structurally concordant, erosional contact
KATHERINE GROUP (Unit K7)

By a happy historical accident, two units assigned to the Little Dal are missing from the type section. The first is the basaltic lava that locally overlies member D of the Upper Carbonate. The second is a cyclical unit of sandstone, dolomite, siltstone and mudstone that overlies either the lava, or members D and C. For reasons given at the end of this paper, these two units are transferred to the overlying "copper cycle". Emendation of the type section of the Little Dal Group is not required.

Mudcracked formation

The Mudcracked formation, 20 to 60 m thick, is the thinnest unit here accorded formational rank. It is so ranked because it is the significant initial deposit of the Little Dal Group and differs markedly from the overlying Basinal assemblage. Its base is drawn at the base of the lowest shale above the monolithic quartzites of the uppermost member (K7) of the Katherine Group (see Aitken et al., 1978). Its top is drawn at the top of a member of dolomite derived from ooid-intraclast grainstone that occurs below the first nodular lithology characteristic of the lower Basinal assemblage, and near the highest occurrence of quartz sand in this part of the column. It is informally and temporarily named for the predominant development of wide, orthogonal, sand-filled mudcracks.

Lithology

Lower member. The lower, clastic-dominated member that forms most of the Mudcracked formation consists mainly of grey, brown, black, green and red mudstone, with subordinate fine and very fine grained sandstone in very thin to rare thick beds. Much of the sandstone, and all of it in cratonward (presumably shoreward) sections, appears in very thin to medium beds with rippled tops, and in chains of starved ripples (Fig. 3.4). The ripples are mainly of current type, but in the common interference-rippled beds both current- and oscillation-type cross-sections occur, and the terminal laminations consistently drape across the underlying cross-laminae. Hummocky bedding also occurs. Flutes and grooves occur sporadically. Basinward, sandstone and mudstone are commonly arranged into coarsening-upward, thickening-upward cycles (Fig. 3.5), which locally at least are separated by an erosion surface. The thick, upper sandstones of these cycles are in some cases structurally cryptic; elsewhere they are cross-laminated in both trough and tabular accretion styles, and plane-parallel laminated. Both straight-crested and lunate current ripples are also present. Prominently mudcracked horizons are recurrent in the mudstones, and salt "hoppers" occur at many, if not all sections. Shale-pebble conglomerates occur throughout units of ripple cross-laminated sandstone, but occur only at the bases of thick sandstone beds.

In the lower member, carbonate rocks decrease basinward from as much as a third to near-vanishing. They are microcrystalline, argillaceous, locally nodular, yellow-weathering dolomites, commonly associated with dolomite-pebble rudstones.

Oolite member. The Oolite member, nowhere more than 14 m thick, forms the top of the Mudcracked formation and is an important regional marker (Fig. 3.6, 3.7). It consists mainly of dark grey, dolomitized, ooid and intraclast grainstone, weathering dull orange or orange-brown, generally in thick beds. Minor flakestone and oncoid grainstone occur locally.

[1] The informal upper and lower divisions of the Little Dal Group, as presented here, do not correspond with the upper and lower members of the Little Dal Formation, as proposed by Gabrielse et al. (1973, p. 17). The original members were distinguished on criteria that are not applicable regionally; attempts to extend them beyond the vicinity of the type section have led to mapping errors.

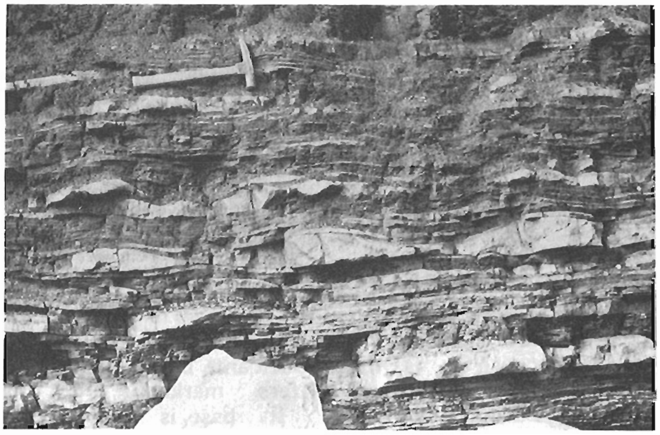


Figure 3.4. *Mudcracked formation: shale and sandstone, with sandstone in characteristic "starved" ripple forms.*

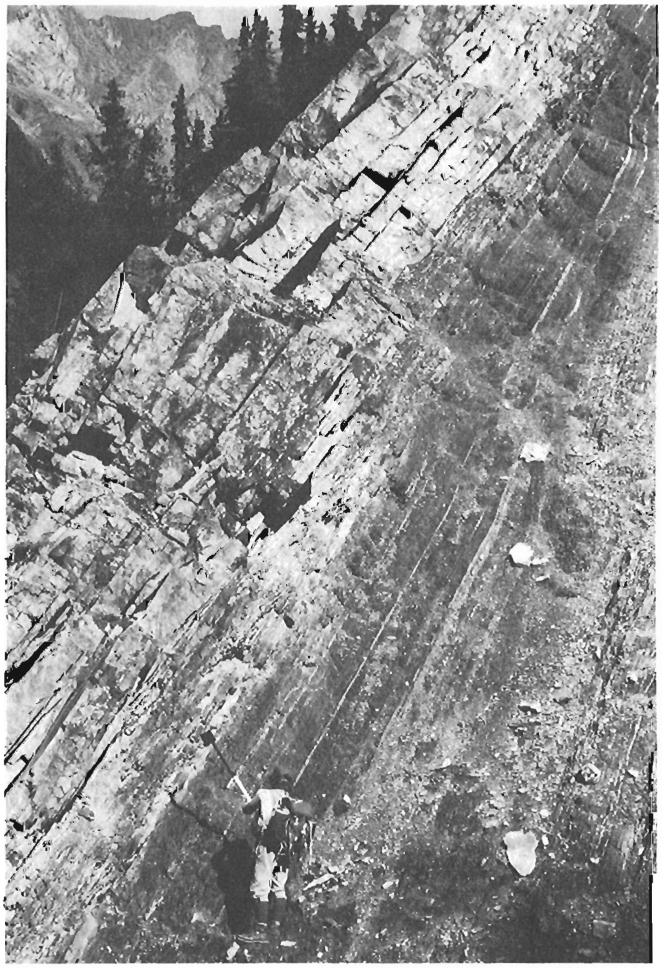


Figure 3.5. *Mudcracked formation: one of three coarsening-upward cycles, shale below, sandstone above, at section 77-20.*

Regional Variation

The uniformity of the formation as seen in correlation panels drawn along the arcuate tectonic trend (Fig. 3.6) suggests that the local tectonic strike approximates the depositional strike, while panels drawn perpendicular to that trend (Fig. 3.7) show basinward (southwestward) thickening. These considerations allow a subjective choice to be made among the many possible isopach configurations, two of which are illustrated in Figure 3.8. The dashed version is an attempt to accommodate the trend of the facies change in the succeeding Basinal/Platform package (see Fig. 3.14). Notable and somewhat surprising is the basinward increase in sandstone content (Fig. 3.7).

The Mudcracked unit is regarded as having formational status only in the region that became basinal during the succeeding depositional episode. Traced into the platformal region (Fig. 3.6), it is only 10-30 m thick, and contains little sandstone; dolomite beds are relatively prominent, the mudrocks tend to be reddish, and salt hoppers are present. These are characteristics of the shoreward, upper and later aspects of the Mudcracked formation. In the platformal region, where problems of thickness, exposure and mappability are encountered, it is best treated as a member of the Platform assemblage, rather than as a formation.

Interpretation

Sedimentological interpretation of the lower, clastic-dominated member of the Mudcracked formation begins with the coarsening-upward, thickening-upward, regressive mudstone-sandstone cycles that are obviously littoral and appear to be of beach, bar, tidal inlet and tidal delta origin. A progradational, alluvial origin is rejected because of the lack of such cycles and the low sandstone content at "shoreward" sites. Assuming a littoral environment for the sandstones capping the cycles, the underlying association of mudstone with thin sandstones, starved current ripples, interference ripples, flutes and grooves invites a sublittoral interpretation. Each thin sandstone in this association may be the deposit of a single storm surge, reworked by waning storm currents and normal currents. Mudcracked mudstones, grouped in some sections with this facies association, contradict this interpretation; focused re-examination may resolve the problem. Mudcracked, partly red mudstones with salt hoppers that occur above the littoral sandstones can be accommodated in a muddy tidal environment, behind inferred offshore bars. The thin argillaceous dolomites and dolomite-pebble rudstones also fit a lagoonal or tidal-flat interpretation.

This apparently straightforward interpretation is complicated by recognition of transgressive onlap within the Mudcracked formation. The basal contact appears conformable in the region that became basinal during the ensuing depositional episode; but in the region that developed instead as a carbonate platform, clear evidence of an erosional contact is widespread, and the strata equivalent to the Mudcracked formation are thin and contain little quartz sand. Southeastward transgressive onlap is strongly implied. Following this view, the presence of thick sandstones in littoral cycles only at basinward sites may be interpreted as a temporal change; a sand supply sufficient to build an offshore bar system may have existed only early in the history of the formation, before marine transgression had flooded the broad subaerial exposures of the Katherine sandstones that were the apparent sand source.

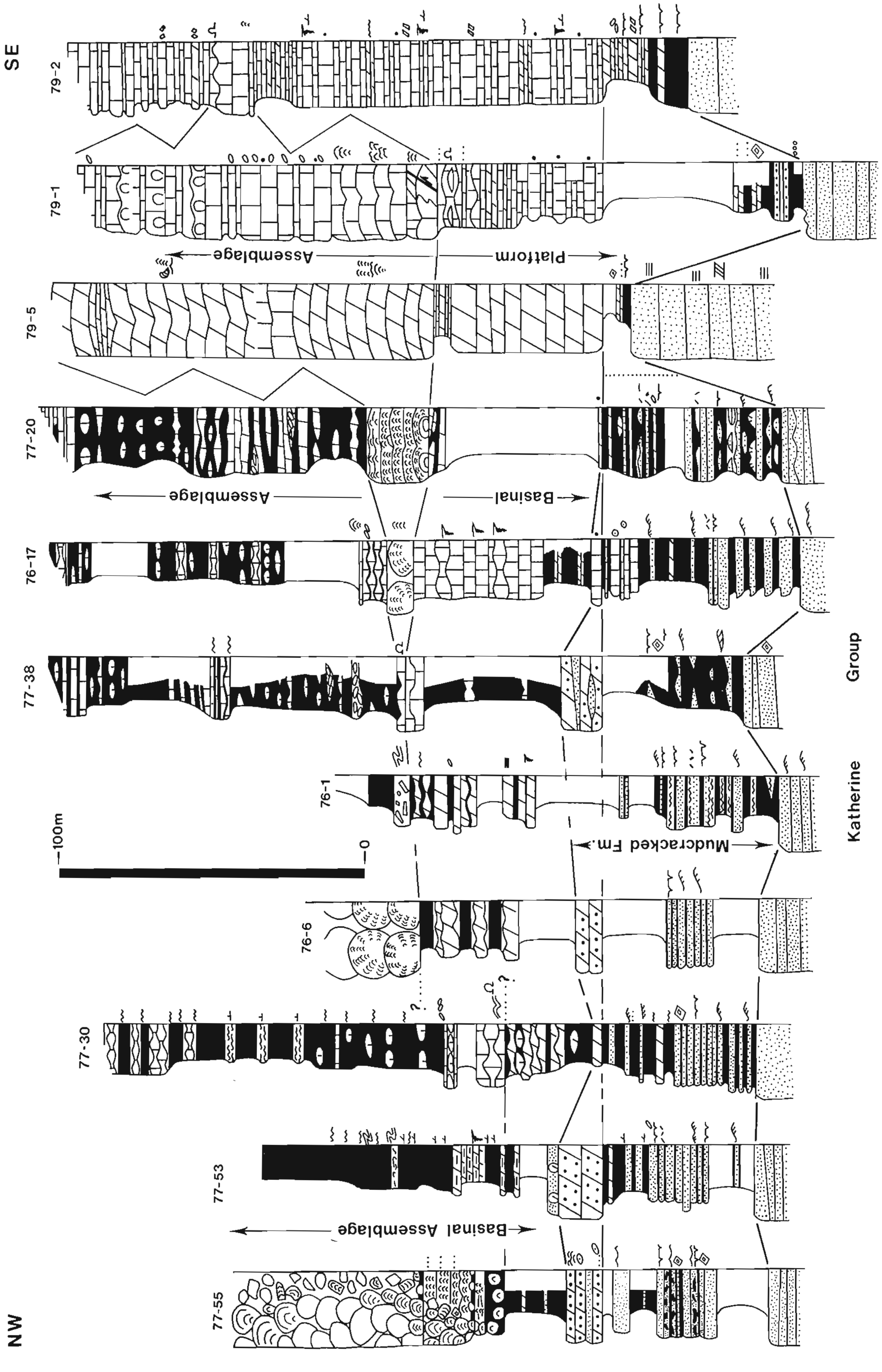


Figure 3.6. *Mudcracked formation: correlation panel parallel to depositional and tectonic strike. For key to lithologic symbols, see Figure 3.2a.*

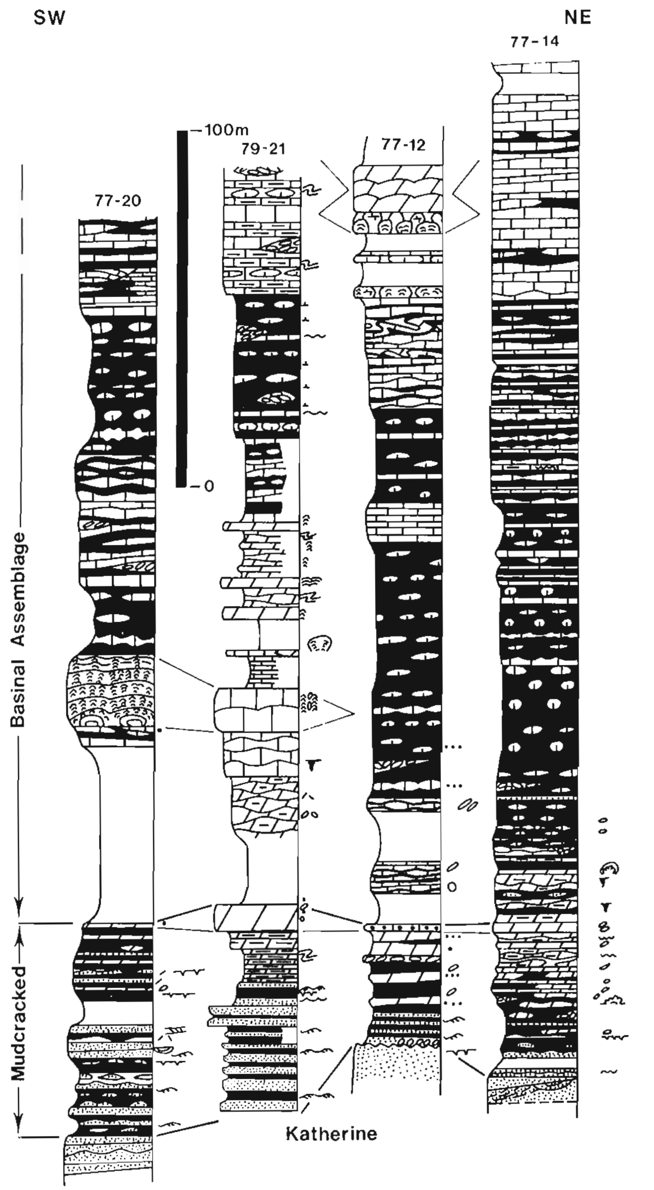


Figure 3.7. *Mudcracked formation: correlation panel across depositional and tectonic strike.*

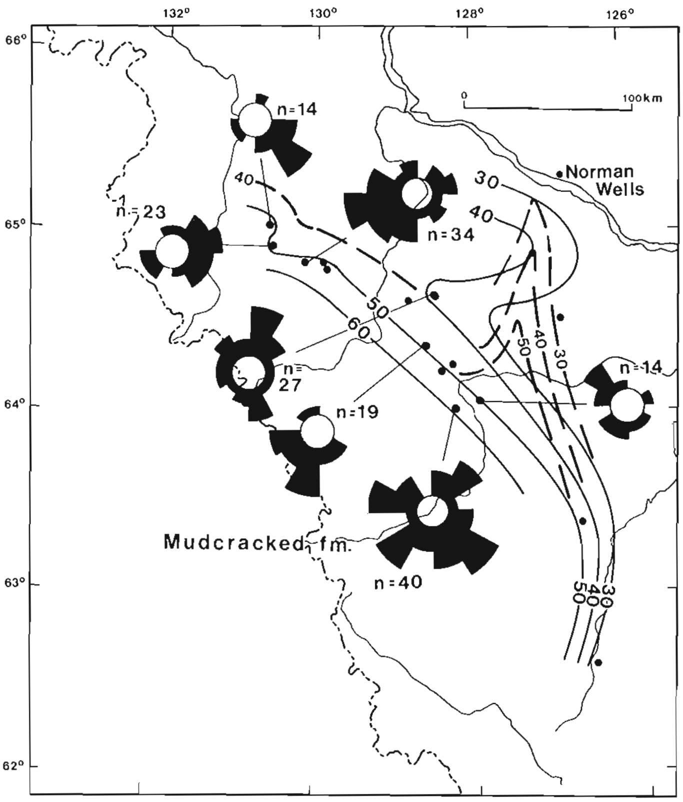


Figure 3.8. *Mudcracked formation: tentative isopachs and paleocurrents. Broken-line isopachs conform to trend of boundary between platform and basin in succeeding formations.*

Paleocurrent patterns determined for the Mudcracked formation lack consistency (Fig. 3.8), but tend to be polymodal. The principal mode for current lamination dip-directions at different sites may be offshore (SW) or longshore (NW or SE). At Section 79-21, the mode of a small sample of flutes and grooves (n=6) is perpendicular and directed offshore relative to a northwest-directed longshore maximum determined from ripple cross-lamination. This is the pattern predicted by the model presented, but it is not consistently repeated. Onshore-directed crossbedding occurs mainly in the thick sandstone beds interpreted as littoral.

The upper, oolite member of the Mudcracked formation separates a littoral assemblage below from strata of subtidal to deepwater origin above. It is interpreted as a transgressive, high-energy shoreline deposit. The cutoff of important sand supply at roughly the time of deposition of the oolite may record entrapment of quartz sand in estuaries by rising sea level.

"Basinal assemblage"

The Basinal and Platform assemblages were deposited contemporaneously during a period recorded by the stratigraphic interval between two approximately time-parallel markers, the oolite member of the Mudcracked formation and the base of the Grainstone formation.

Pronounced regional variation is present in the Basinal assemblage, however, its overall unity is expressed in the following characteristics:

Intimate mixing of carbonate rocks and siliciclastic mudrocks;

A tendency to nodular character, especially in the lower part;

Prominence of decimetre-scale depositional rhythms;

General (though not complete) absence of indicators of shallow-water deposition; and

Abundance of limestone, as opposed to the generally dolomitized nature of most Little Dal carbonate strata.

Lithology

Nodular lithofacies. Diagenetic nodules and nodular bedding occur throughout the Basinal assemblage, but are most prominent in the lower part. At most localities, the unit commences with interbedded calcareous black shale and microcrystalline, argillaceous limesone (locally dolomitized). The decimetre-scale carbonate beds pinch and swell abruptly, in nodular rather than lenticular fashion, and are usually accompanied by a few discrete nodules tens of centimetres in largest dimension. Together with local preservation of bedforms, for example, large-scale current ripples, that do not conform to the external outline of the carbonate body, this geometry demonstrates that the nodular beds are largely diagenetic in origin, although many may be nucleated on a bed of particulate lime sediment. Molar-tooth structure is common in association with the black shale - nodular limestone lithofacies, occurring not only in limestone, but also in beds described in the field as black calcareous mudstone. The lowest, massive, conglomeratic, matrix-supported debris-flow deposits and slide masses are associated with the nodular black shale lithofacies.

In all but the most basinward sections, a highly characteristic red nodular lithofacies (Fig. 3.9) appears near the base of the Basinal assemblage, and in some sections near the transition to the Platform assemblage, and persists as recurrent tongues well into the upper half. The dusky-red matrix rock is consistently described in field notes as calcareous mudstone or shale, and will be referred to here as mudstone. In stained thin section, however, it is in fact a very argillaceous limestone. The discoid to spheroidal nodules are grey to pink microcrystalline limestone. Their dimensions rarely exceed 10 cm as measured perpendicular to bedding, but commonly are much greater along the bedding. At many levels, the nodules coalesce laterally into persistent limestone beds that are nodular in some instances and bounded by planar surfaces in others; it is difficult to judge whether or not any primary lime-mudstone beds are present. The size, shape, and concentration of nodules varies stratigraphically, so that outcrops display distinct bedding based on nodule distribution. The nodules, and in compressed form, the enclosing mudstone display corrugated, sub-millimetre lamination of uncertain origin. In many sections, the red nodular mudstone lithofacies contains, near the base and diminishing upward, laminae and thin lenses of very fine grained, quartz sandstone. Sandstone is missing in basinward sections; shoreward, sandstone is more abundant and persists higher in the section. Flat-pebble rudstone composed of tabular rounded clasts of limestone like that forming the nodules is common near the base of the red nodular lithofacies and dies out upward. These conglomerates rarely exceed 25 cm in thickness, and usually occur as depositional lenses. No sedimentary structures other than those already described are known from the red nodular mudstones; many nodules have been split in the search for uncompressed structures such as mudcracks, but without result.

Locally, the nodular facies described above is grey to greyish-olive. Red intervals pass laterally to olive over a distance of 100 m. In variegated outcrops, the red is cut by the green; hence, red is apparently the primary colour.

Rhythmite lithofacies. Rhythmites of one kind or another are the most characteristic and ubiquitous of the basinal lithofacies. Easily perceived as rhythmites (Fig. 3.10), they are nevertheless difficult to describe, because of the intergradation between silty (quartz silt) limestone and calcareous (quartz) siltstone, between silty shale and shaly siltstone, etcetera. The most common rhythmites involve a simple ABAB alteration between dominant, mainly thin beds of calcisiltite and thinner beds or laminae of siltstone or silty shale. Less commonly, an ABCABC pattern, with near-black shale added as "C", occurs. The limestone and siltstone are laminated at a submillimetre scale, and normally graded laminae, commonly with micro-load casts, are widespread (Fig. 3.11).

Figure 3.9. *Basinal assemblage: red nodular facies.*

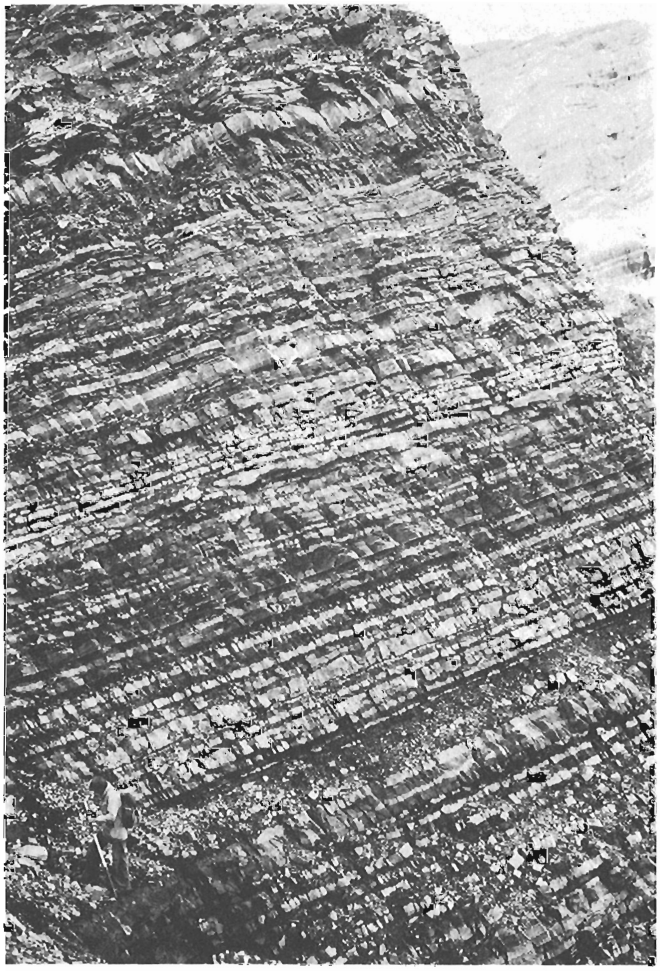


Figure 3.10. *Basinal assemblage: deepwater rhythmite, section 76-2.*

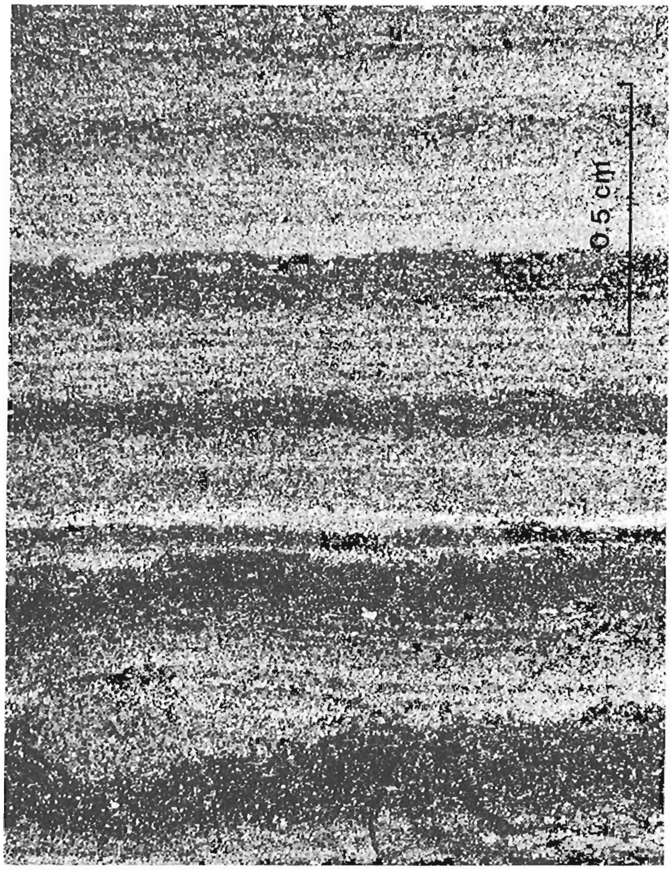


Figure 3.11. Basinal assemblage: micro-graded laminae from argillaceous limestone bed of rhythmite facies. Scale bar is 0.5 cm long.

Many of the calcisiltite beds contain low-amplitude, long-wavelength (25 cm) current-ripple forms in which the terminal laminae drape across the underlying ones. In stained thin section, silt-sized clasts of both calcite and dolomtie are recognizable. Paleocurrents determined from the foreset lamination generally show no distinct maxima, and lack consistency between stations.

Nodules and nodular thin beds of microcrystalline limestone, often incorrectly described as lime-mudstone, are commonly centred on horizons of current ripples. Some of these outgrowths are undoubtedly diagenetic, but given the setting, it is difficult to judge whether some of the nodular bedding may be due to downslope creep.

Siltstone intergrades completely with both calcisiltite and brown to black, calcareous carbonaceous shale. In addition to its part in the decimeter-scale rhythms, siltstone with or without limestone nodules forms units up to 20 m thick in some of the more basinward sections.

At many localities fragments of carbonaceous mats or films (**Beltina, Morania**), flattened in the bedding plane, are abundant in both limestone and siltstone beds. Careful search of the carbonaceous beds can yield specimens of the millimetre-scale disc **Chuaria** and the centimetre-scale, strap-like compression **Tawuia** (Hofmann and Aitken, 1979); a well-preserved microflora is also present (Hofmann and Aitken, 1979).

Pale-pellet grainstone. One or more units of limestone consisting of pale cream to grey pellets of microcrystalline calcite in a sparry cement occur at the top of the Basinal assemblage at a number of sections, usually higher than the first unit of small stromatolites that indicates the transition into the Grainstone formation. This distinctive, pale grey-weathering rock is medium- and thick-bedded, and usually faintly cross laminated.

Stromatolitic lithofacies. Quite apart from the giant stromatolitic reefs, which will be discussed separately, algal stromatolites occur in the Basinal assemblage as isolated small bioherms, extensive biostromes, or what will be termed here differentiated biostromes, that is, "biostromes of bioherms". Small bioherms and thin (less than 1 m) biostromes occur here and there in the lower sixth, more or less, of the formation. The stromatolite at this level is a small, unwalled, rarely branching and highly consistent form. One or more horizons of small hemispherical biostromes commonly occur in the uppermost tenth of the formation; the stromatolite here is locally identifiable as **Baicalia.**

Two especially prominent differentiated biostromes also occur. The lower, up to 20 m thick but generally much less, is a near-regional marker extending far into the basin and apparently providing a foundation for the giant stromatolite reefs (Fig. 3.12). Its base is about 50 m above the base of the formation. The upper one, some 330-350 m above the base, forms a tongue 55 m thick at Ekwi River that splits into successively thinner biostromes northwestward and disappears, over a distance of 30 km. The stromatolites contained therein are not well preserved. At one locality, the form is centimetre-scale, erect, walled, and rarely branching, with cylindrical and both moderately upward-expanding and -contracting intervals, but larger, strongly expanding forms have also been observed. Both of these biostromes may be regarded as tongues extending from the stromatolitic buttress that defines the edge of the platform (see below, under "Platform assemblage").

Giant reefs. Giant stromatolitic reefs up to 300 m high, unknown elsewhere in the world, occur only in a fairly well-defined belt in the western part of the exposed Basinal assemblage (Fig. 3.13, 3.14). These reefs are not merely variants of the tens-of-metres thick differentiated biostromes described above. The reefs are critically distinguished from the biostromes and their constituent bioherms by the following criteria:

a. Height-to-breadth ratios greater than 1:50, as compared to one to thousands for the biostromes.

b. Thick mantles of bouldery debris (talus) on their flanks, a consequence of their great topographic relief during growth.

c. Lack of decimetre-scale growth increments that can be traced for hundreds of metres through biostromes or correlated from bioherm to bioherm within differentiated biostromes.

d. Dominance in the reefs of a sub-centimetre-scale, unwalled, digitate, commonly recumbent stromatolite that occurs only locally and rarely elsewhere in the Little Dal.

Field study of the reefs is not complete, yet all of what is presently known would constitute a paper in itself. Briefly, the reef cores are compound; they consist of a pile of spheroidal bioherms greater than 1 m and less than 10 m in diameter. Gently inclined, horizontal and even downward growth of the small digitate stromatolites that built the

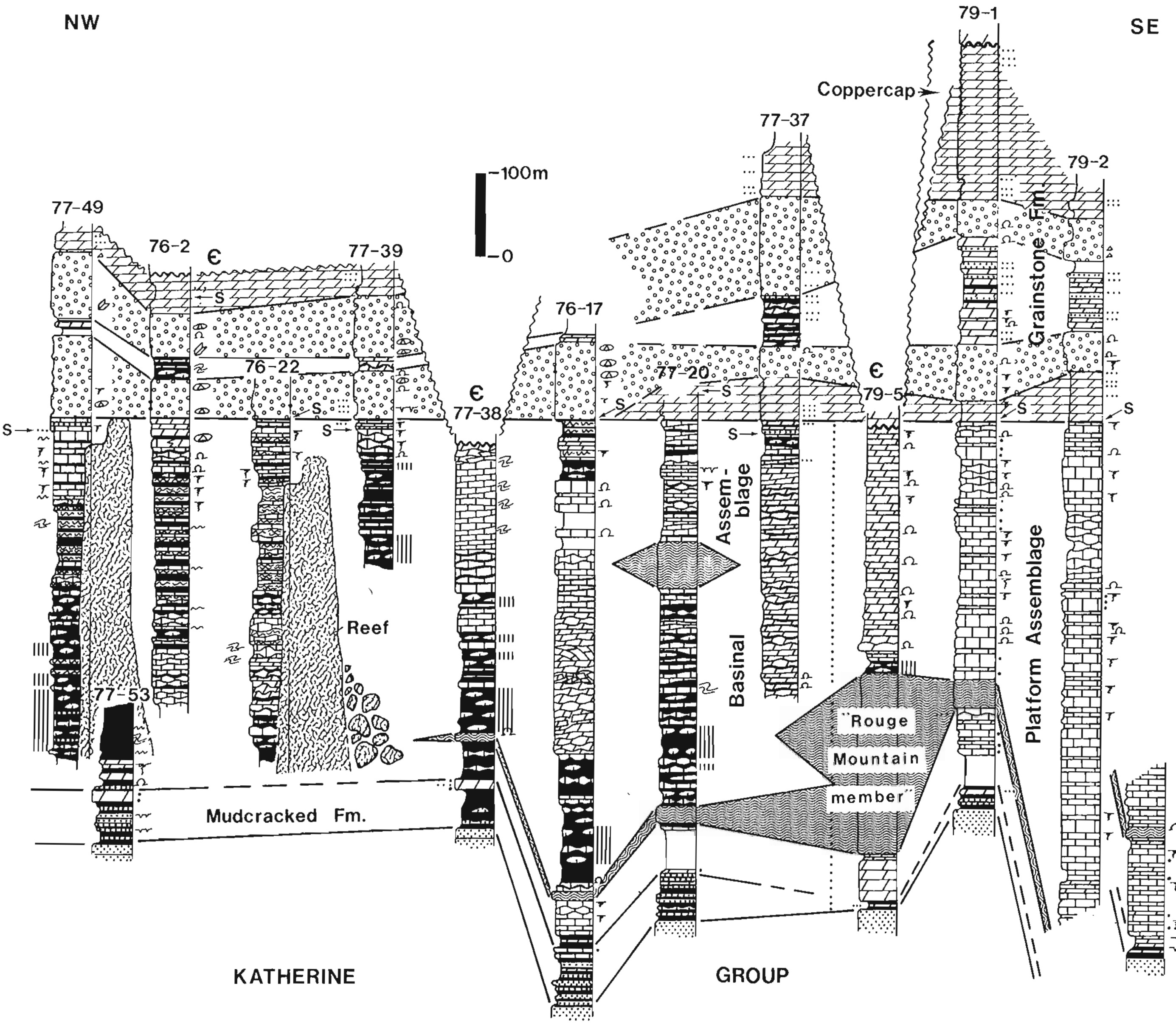


Figure 3.12. Basinal and Platform assemblages and Grainstone formation: correlation panel. For key to lithologic symbols, see Figure 3.2a. Vertical lining denotes red coloration. "S" denotes lowest occurrence of quartz sand near base of Grainstone formation.

bioherms is the norm. Development of each individual bioherm generally ended with an envelope of **Stratifera** (essentially, undifferentiated cryptalgal laminite) that commonly displays vertical relief of several metres. The reef cores are mantled on their flanks by up to 100 m of bouldery reef talus. Rotated and inverted blocks up to 10 m are common. The relief on talus lobes that tongue downward into basinal rhythmite demonstrates minimum relief of at least tens of metres above the basin floor. Although many of the erosionally stripped reefs appear as pinnacles, all reefs with intact tops are flat-topped and commonly stepped. Only within the uppermost 1 to 10 m of intact reef tops is there any evidence of hydraulic fragmentation (stromatolite-chip breccias and conglomerates, oncolites, etc.). Some of the reefs, especially the most easterly, are dolomitized, while others remain pale grey limestone. Paradoxically, primary fabric detail is generally better preserved in the dolomitized rocks.

Regional Variation

Regional variation in the Basinal assemblage between basin-flank (proximal) and more basinward (distal) sites is clear-cut (Fig. 3.12-3.14).

Proximal sites are characterized by:

a. A high carbonate to clastic ratio in rhythmite lithofacies.

b. A tendency to dolomitization of the rhythmite limestones.

c. Prominence of penecontemporaneous slump-folds, slide masses and debris flows.

d. Relatively strong development of stromatolite biostromes near the top and base.

e. Prominence of the red nodular lithofacies in the lower half.

f. Presence and persistence to high levels of quartz sand in the rhythmite and red nodular lithofacies.

Distal sites are characterized by:

a. Relatively low carbonate to clastic ratio in rhythmites.

b. Prominence of siltstone, especially in metres-thick units.

c. Prominence of thick units of black mudrocks.

d. Presence of stromatolitic reefs.

Interpretation

Interpretation of the rhythmite facies as of deeper-water, subtidal origin does not appear open to serious challenge. The limestone and siltstone beds are at least in part very fine grained turbidites (units C, D, E of the Bouma sequence). Evidence of depositional slope is widespread, and most prominent in areas judged on other criteria to be at the basin margin. The talus tongues flanking the giant reefs give direct evidence of at least tens of metres of water depth.

Consideration of the limestone-rich basin-flank assemblage in relation to the siltstone- and shale-rich distal assemblage suggests more than one source for the basinal sediments. The mainly silt-grade carbonate sediment (including silt-grade dolomite) probably poured off the "carbonate factory" of the shallow-water platform, either along a broad front or through a number of passes through the platform rim, during and immediately following major storms (McIlreath and James, 1979). The amount of clay and quartz silt occurring in the rocks of the platform assemblage is insignificant, despite the presence there of many quiet-water sediment traps, as attested by units of lime mudstone, hence, a separate source or sources for detrital mud appears probable. The notion of multiple sources accounts for the observed variability in proportions of rock-types between basinal sections a few kilometres apart, especially within the reef belt. Turbiditic pulses of detrital mud, if from a point source, would be expected to form a fan-like deposit, while pulses of fine carbonate from the platform would be expected to form either fans or sheets, depending on whether they were introduced through passes in the platform rim or along a broad front. The effect of the giant reefs in deflecting currents must have been profound; one flank of a reef would have been a "shadow" area for one type of sediment or another, depending on the direction of transport. This may account for the observed lack of consistency in paleocurrent directions determined for the rhythmites. In areas shielded from both carbonate and siliciclastic flows, only hemipelagic mud would be deposited.

The conclusion that the red nodular lithofacies is also subtidal in origin may be more contentious. Some would point to the flat-pebble rudstones as evidence for intertidal/supratidal origin. The following facts appear to stand against the latter possibility:

a. Absence of mudcracks and ripple marks.

b. Absence of algal stromatolites (in a carbonate-depositing environment).

c. Absence of evidence for the former presence of evaporitic minerals, despite their presence in overlying and underlying formations.

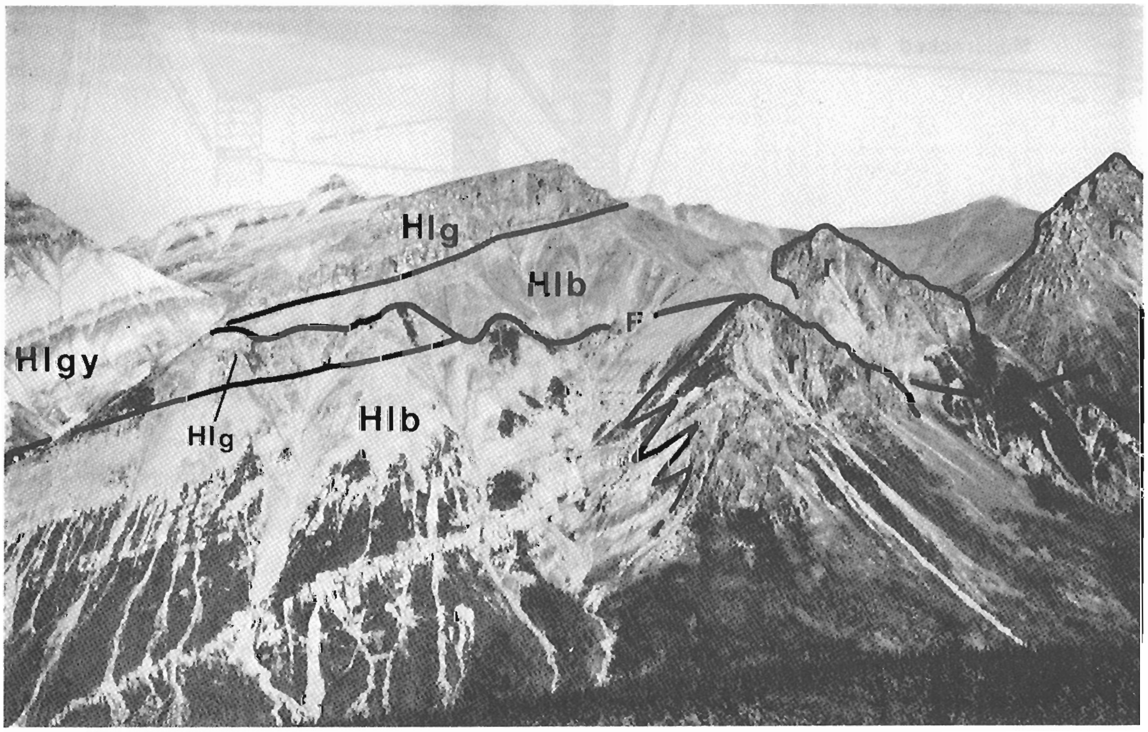


Figure 3.13. *Basinal assemblage, Grainstone and Gypsum formations at locality 76-22. Stromatolite reef at right is split by a fault occupied by a diabase dyke. Note drape of rhythmite beds against reef. Hlb-basinal rhythmites; r-reef; Hlg-Grainstone formation; Hlgy-Gypsum formation.*

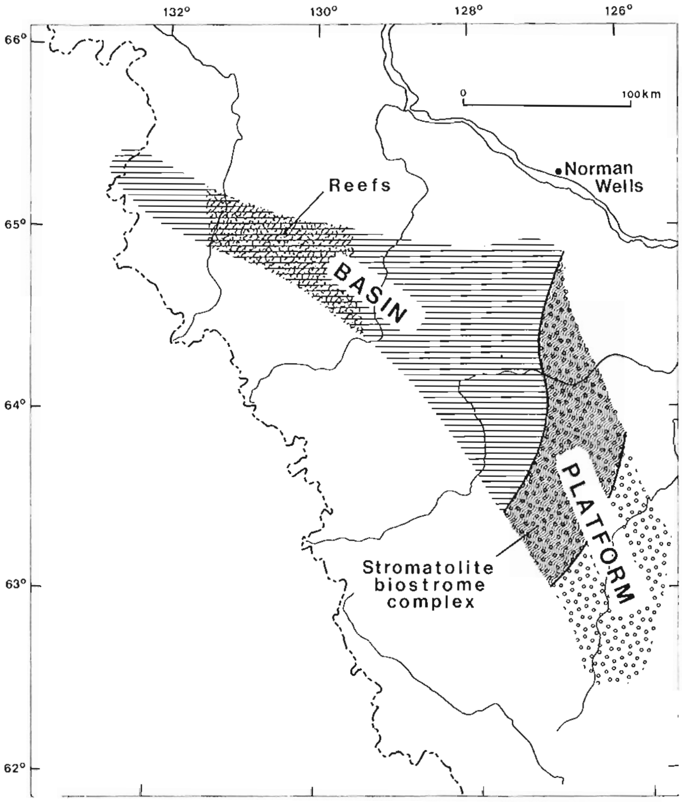


Figure 3.14. Facies differentiation of the Platform and Basinal assemblages.

d. Persistence of thin (10 cm) beds of limestone, and thin nodule layers, for hundreds of metres.

e. Transition to, and intertonguing with rhythmite lithosomes (Fig. 3.12) without the intervention of any lithofacies identifiable as littoral.

f. Similarity to pelagic redbeds, (commonly with calcareous nodules and beds of diagenetic origin) that are widespread in the Devonian of Germany (Franke and Paul, 1980). The red coloration in the Devonian example is attributed to an oxygenated water column due to a paucity of organic growth that in turn is due to a lack of nutrients.

Although there is no reason to deny the possibility that the stromatolite biostromes may record temporary shallowing, neither is such a postulate necessary. A subtidal origin for some stromatolites (especially in the Precambrian) has gained wide acceptance, and needs no detailed defence here. To be sure, origin within the photic zone must be assumed, and the presence of stromatolite-derived flakes suggests a depth within the reach of storm waves (the absence of elongate, oriented stromatolites suggests the absence of persistent currents). Continuing this thought, is it not plausible that, rather than shallowing, the biostromes record temporary diminution in turbidity, simultaneously encouraging the growth of algal films and deepening the photic zone? Significantly, neither an erosion surface nor a conglomerate at the bases of these particular biostromes was noted in the field.

Although the mode of growth of the giant reefs is reasonably clear from the description given above, the reason for their localization in what on other criteria is a distal part of the basin remains largely an enigma. At two localities (76-6, 77-55), reef growth commenced on a foundation provided by the lower differentiated biostrome, and this may be assumed to be the rule. (Because the reefs expanded outward as they grew upward, the lowest reef-rock encountered in a random section is by no means necessarily the earliest reef-rock.) On the other hand, as the extent of the foundation biostrome is vastly greater than that of the reef belt, another localizing influence must be sought. Existing stratigraphic control does not support a hypothesis of reef-growth on paleostructural highs, nor is consistent orientation to elongate reefs apparent (a greater volume of data might change these tentative observations). The observation that, within the basinal rhythmites, terrigenous sediment is most abundant within the reef belt, suggests that vigorous reef growth may have been dependent on land-derived nutrients. This hypothesis is consistent with the observation that the red nodular lithofacies, attributed ultimately to lack of nutrients, is to a large degree antipathetic to the reefs.

The Basinal assemblage records at its base fairly rapid deepening from the peritidal zone, and at its top, gradual shallowing to peritidal conditions once more. It is of considerable interest that molar-tooth structure is prominently developed in these transition zones (Fig. 3.12). This observation, which can be repeated in other Little Dal formations, combined with several observations of post-emplacement molar-tooth structure developed in the matrix of debris flows, negates the often-made identification of molar-tooth as a "shallow-water indicator". The shallow subtidal environment (1-10m?) is clearly at least one of the environments in which molar-tooth structure formed, and may be the only one.

By the onset of deposition of the Grainstone formation, sedimentation had filled the basin to the peritidal zone, and there is no evidence on the shallow-water carbonate platform for a sea level drop at this time. At first glance, therefore, it might be expected that the Basinal assemblage would be thicker than the equivalent platform assemblage. Not so; the total thickness measured from the base of the Little Dal Group to the base of the Grainstone formation varies from 450 to 635 m in the basinal region, with the thicker sections lying along the basin flank, while the equivalent interval in the platform region varies from 450 to 815 m. The answer lies in differential compaction; with a much higher content of silicate mudrocks, the Basinal assemblage, with up to 40% mudrocks, has compacted more than the Platform assemblage, with less than 5%.

The sparse and poorly distributed thickness data for the combined Basinal and Platform assemblages permit isopachs to be drawn to show cratonward thickening, basinward thickening, or - an interesting possibility to be recalled later - basinward thickening to a depositional axis, then thinning. In any event, the boundary between Basinal and Platform assemblages cross-cuts isopachs, both those for the combined Basinal and Platform assemblages and those for the underlying Mudcracked formation, an indication that differentiation of facies at this level was a response to factors other than differential subsidence. It appears likely that oceanic factors were responsible for the creation of the carbonate platform, and the maintenance thereon of shallow conditions through rapid carbonate sedimentation, while the adjoining basin deepened as a consequence of "starvation".

"Platform assemblage"

The Platform assemblage is correlative and continuous with the Basinal assemblage, but lithologically distinct. It includes the sub-regional, multistorey stromatolite biostrome complex (Rouge Mountain member) that largely separates the platform and basinal lithofacies.

The earmarks of the Platform assemblage are:

a. Dearth of siliciclastic sediments,

b. Prominence of intraclast and ooid grainstones,

c. Prominence of lime mudstone with molar-tooth structure,

d. Many small stromatolite bioherms and biostromes, and

e. Distinctly thicker bedding, on average, than the Basinal assemblage.

Lithology

The limestones that form the Platform assemblage at most localities are characteristically grey to dark grey to brown-grey, and weather largely to rubbly brownish grey outcrops of sombre appearance. This monotony is broken by a few units of pale yellow-weathering "platy dolomite" lithofacies, as described below, under "Grainstone formation", and by thin, usually "one-storey" stromatolite biostromes and differentiated biostromes which, where undolomitized, weather very pale grey.

Intraclast grainstone is the most characteristic of major lithofacies. It consists of well rounded, millimetre-scale clasts of near-black lime mudstone in a white calcite matrix. The somewhat less common ooid grainstone is similar in appearance, as are the frequently recurring beds of flakestone (mat-chip grainstone, packstone and floatstone). All these rocks are usually medium- and thick-bedded, and weather blocky or rubbly.

Lime mudstone is variably medium- and thick-bedded, indistinctly bedded, or nodular thin- and medium-bedded. In different units it weathers platy, blocky, or rubbly. Molar-tooth structure is very common. Units of current-laminated calcisiltite are inconspicuous.

Isolated stromatolite bioherms in the Platform assemblage are usually small, less than 1 m in height. Biostromes and differentiated biostromes ("biostromes of bioherms") locally exceed 10 m in thickness. The constituent stromatolites are mainly simple domal and short columnar types lacking the showy complexity and variety of the stromatolites of the upper Little Dal. Small bioherms of a centimetre-scale digitate stromatolite like the unnamed reef-builder of the Basinal assemblage are rare.

An important unit of laminated, platy, silty dolomitic limestone, low in the type section of the Little Dal Group, contains abundant **Tawuia** and a few **Chuaria.**

The endless repetition of a small number of rock-types in the Platform assemblage suggests cyclicity. The writer has not had the opportunity to study this very thick formation in sufficient detail to identify the cycle, if one exists.

A prominent, resistant, many-storied stromatolite biostrome complex, here termed the Rouge Mountain member, marks the boundary between Platform and Basinal assemblages (Fig. 3.12, 3.14). This member is commonly altered to deep rusty-orange weathering dolomite. Excellent sections are exposed on the west flank of Tsezotene Range and near the head of Rouge Mountain River, which takes its name from the spectacularly "red" stromatolitic dolomite.

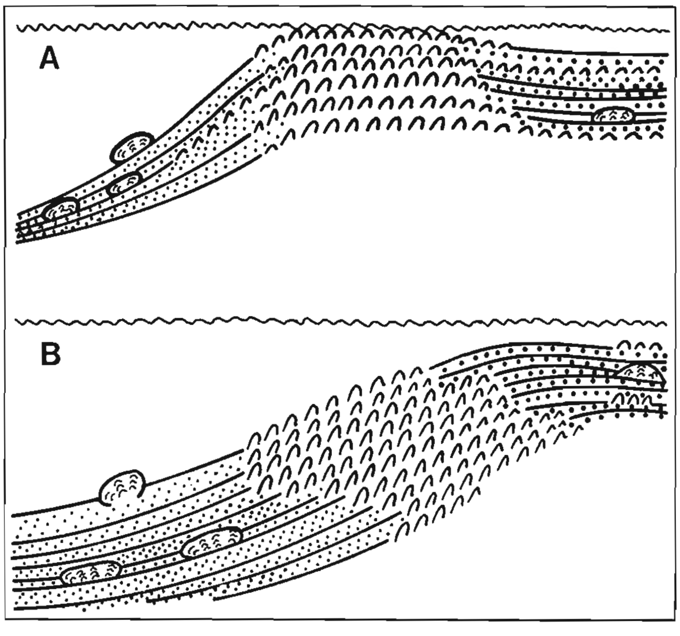
The barrier biostrome complex is up to 203 m thick. It is built of 15 to 25 cm biostromal growth-increments of a variety of intergrading stromatolites, giving rise to a medium-bedded appearance. The dominant stromatolite form consists of short, unwalled, rarely branching, regular columns. This dominant form intergrades with rapidly expanding, actively branching forms assignable to **Baicalia.** Other common forms are a divergent, "bushy" form with numerous semi-prostrate columns; a very irregular, unwalled form with many "bridges" between columns, and a uniform, slim, cylindrical form, apparently walled. The general prevalence of columnar forms is broken at intervals by sheets of laterally linked hemispheroids ("**Stratifera**"). Interruptions of biostrome development by beds of non-stromatolitic sediment are rare, and beds recording high turbulence, such as grainstones or rudstones, are not present.

Only two complete sections of the Platform assemblage have been measured. At the type section of the Little Dal (79-2) the assemblage is 812.8 m thick, as measured by R. Booker and the writer. At section 79-1, regionally basinward of the type section, the assemblage is only 453.8 m thick.

Interpretation

The Platform assemblage is the product of nonclastic sedimentation on a fairly shallow, fairly high-energy platform. Interruptions of the prevailing grainstone deposition by lime-mud deposition took place either sequentially, during episodes of deeper water or lessened turbulence, or spatially, in sheltered areas behind grainstone banks. Subaqueous deposition appears to have been the rule, in view of the paucity of indications of subaerial exposure and the general lack of dolomitization. The stromatolites appear to be largely of shallow subtidal origin (0-10 m?).

It might appear, at first glance, that the Rouge Mountain member represents a slightly elevated, perhaps peritidal rim to the platform, analogous to modern barrier reefs (Fig. 3.12, 3.15A). There are several difficulties with this interpretation. Such an elevated rim would be exposed to the attack of deep-sea waves generated in the basin, and



A - interpreted as a shallow rim to the platform
B - interpreted as a deeper water buttress to the platform

Coarse stipple – limestone, mainly grainstone
Fine stipple – limestone, turbiditic calcisiltite and lime mudstone

Figure 3.15. *Two possible interpretations of the 'Rouge Mountain member' (regional, multi-storey biostrome).*

the expected consequences are three: a. hydraulic fragmentation of stromatolites, or at least their youngest laminae; b. elongation and orientation of individual stromatolites; c. a low-energy regime inside the rim. None of these consequences is recorded.

More attractive is a model in which the biostrome complex formed a gently sloping, stabilizing buttress in fairly deep water, (10 to 80 m?) flanking the platform (Fig. 3.15B). Such a configuration would permit open-sea wave energy to reach the platform (tides may of course have contributed to the energy regime of the platform in either model). Because of the dearth of particulate, non-stromatolitic sediment within the biostrome complex, the buttress model does not admit the earlier-mentioned possibility of "feeding" the basinal rhythmite facies with turbid waters that flowed off the platform on a broad front during storms. It is necessary to assume instead that the turbidity currents originated at channels or passes through the complex, although no such channels have been observed.

Grainstone formation

The Grainstone formation records a period during which facies variation was reduced, and "shallow-water" sedimentation was re-established across the entire region of Little Dal exposures. Little is known of the upper part of the formation, missing at many localities and covered at others. In fact, only two complete sections of the formation, both with covered upper contacts, have been obtained. On the other hand, regionally valid members can be differentiated within the lower part of the formation, and thus this lower division is amenable to stratigraphic and sedimentological analysis.

The base of the Grainstone formation is gradational. It is drawn at the lowest appearance of either the massive oolite or the platy dolomite lithofacies, described below. Its top, though little observed, is apparently conformable, and drawn at the lowest appearance of continuous beds of gypsum or anhydrite.

Maximum thickness determined for the Grainstone formation is 425 m, in an incomplete section. The lower division ranges in thickness from 139 to 265 m; the poor distribution of data points allows isopachs of this division to be drawn showing basinward thinning, basinward thickening, or basinward thickening to a maximum, followed by basinward thinning.

The Grainstone formation is economically significant as host to the major Mississippi Valley-type zinc showings at Gayna River.

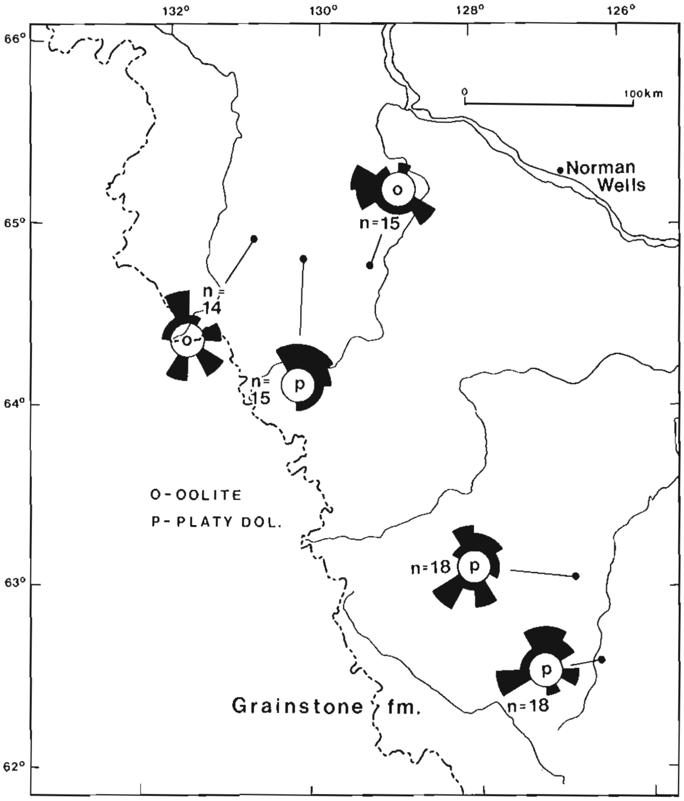
Lithology

The Grainstone formation is a group of extensive lithosomes, or members, composed of one or other of two lithofacies, termed massive oolite and platy dolomite.

Massive oolite lithofacies. The massive oolite lithofacies is dominated by thick, massive, resistant beds of brownish grey, mainly fine equicrystalline dolomite. Accessory quartz sand is noted only at a few, relatively shoreward sites. Rarely is the depositional texture visible as dolomite, but the scattered blebs and nodules of pale grey or pale brown chert preserve an ooid grainstone texture in great detail. Primary structures are planar lamination, low-angle cross-lamination, low-angle tabular accretion sets and megaripple or small dune forms. Paleocurrents determined from the dip directions of ripple crosslamination and low-angle accretion sets weakly define bimodal - bipolar patterns, with the maxima oriented subparallel to depositional strike (Fig. 3.16).

Subordinate rock-types associated and "lumped" with the massive oolite beds include cryptalgal laminite and associated flakestone (mat-chip packstone), small stromatolite bioherms of simple domal and short columnar forms, and units of ripple cross-laminated dolomite interpreted as ex-calcisiltite.

Platy dolomite lithofacies. The distinctive, platy dolomite lithofacies is dominated by thin, platy and flaggy beds of yellow-weathering, grey, microcrystalline dolomite. Quartz sand is common, either as scattered accessory grains or as lenses, laminae and very thin beds of dolomitic sandstone, as sand-filled mudcracks, and as starved ripples of sandstone. A consistent and significant tendency in units of platy dolomite is that the basal beds are sand-free, the first quartz grains appearing some metres above the base and increasing in amount upward. Scour-and-fill structures, cryptalgal laminite, flakestone, low domal algal stromatolites and pale-pellet grainstone are present locally, but are not as prominent as the environmental model (below) might predict. Thin units of ooid grainstone are rare. Slumped (broken and unbroken) beds appearing here and there probably record slumping into tidal creeks, but no definite channel profiles have been seen. Fenestral (birdseye) fabric, that would be expected in this association according to the model of James (1979) has not been recorded.



O-foreset dips of crossbeds in massive oolite
P-foreset dips of current ripples in platy dolomite

Figure 3.16. *Grainstone formation: paleocurrents.*

Grey dolomitic shale is present in variable amounts, as partings and thin interbeds. Shale tends to be sparse or absent low in the Grainstone formation, and becomes a major component towards the top. This shale is very susceptible to red weathering at, and for a considerable distance below, unconformities. Care is required in defining contacts with overlying red formations.

Minor rock types. Minor rock types of the Grainstone formation include lime mudstone with molar-tooth structure that, together with grey shale, forms a thin, recessive-weathering member (the "argillaceous marker" of the geologists at Gayna River) in the western half of the study area. Millimetre-scale tabular vugs after gypsum occur at the top of the member. Salt-hoppers in sandstone and shale occur in the upper division, increasing in abundance upward. Also in the upper divison, pockets and beds of brecciated platy dolomite are fairly common. Gypsum in bodies interpreted as lenses is encountered within the platy dolomite lithofacies by drilling at Gayna River (R. Hewton, personal communication), but has not been seen in outcrop. Random masses of dark grey chert occur in the platy dolomites, especially near the top of the formation.

Regional Variation

In the southeast, at the type section of the Little Dal, the sequence of informal members in the lower division of the Grainstone formation is as follows:

platy dolomite III	
massive oolite II	Contacts between oolite and overlying platy members are usually interbedded.
platy dolomite II	
massive oolite I	
platy dolomite I	

This arrangement persists northwest around the tectonic arc to Godlin River, beyond which platy dolomites I and II pinch out (Fig. 3.12). Farther northwest (retaining the same designations for persistent units), the arrangement is:

	Nomenclature, Gayna River prospect
platy dolomite III	"platy dolostone"
massive oolite II	"upper host"
shale, molar-tooth lime mudstone	"argillaceous marker"
massive oolite I	"lower host"

Parallel changes are noted in the distribution of quartz sand. Sand is most abundant in the southeast and diminishes markedly northwestward. Nevertheless, the first, or lowest appearance of quartz sand defines a horizon essentially parallel to other markers (Fig. 3.12).

Interpretation

The massive oolite lithofacies provides a clear record of a mobile, high-energy ooid shoal complex. Widespread planar lamination suggests that the swash zone on shallow banks may have been the principal "ooid factory". Algal mats and stromatolites became established on inactive banks, and silt-grade carbonate accumulated in sheltered areas behind banks. The gradationally overlying platy dolomite facies then must represent the lagoonal, intertidal and supratidal deposits of James' (1979, p. 111) model; if so, the lagoonal deposits differ little in basic lithology from the subaerial deposits. The lagoon is recorded in the non-mudcracked, non-sandy beds that form the lower parts of many units of platy dolomite, and in the pale-pellet grainstones. Apparently, the lagoon was shallow, narrow, transitory, or all of these.

The amount of spectacularly mudcracked platy dolomite suggests an intertidal - supratidal zone of great breadth, perhaps tens of kilometres. Even so, all modern models predict that dry land lay not far cratonward of today's exposures of the Grainstone formation. The relatively sparse development of evaporites on the extensive flats suggests a climate only slightly arid.

The distribution of quartz sand in the platy dolomite tests the imagination. Because the massive littoral oolite is not in general sandy, it is unreasonable to appeal to a model applicable to other situations, namely, one in which sand has been blown inland from the shoreline onto the supratidal flats. In the Grainstone formation, the quartz sand must be an aeolian contribution from the landmass. The northwestward diminution of quartz sand, and dispersed but northward-directed modes of sandy current-ripple cross-lamination (Fig. 3.16), support northward dispersion of quartz from a southern or southeastern source, as for other Little Dal formations.

A sandy zone that follows closely the gradational, interbedded Basinal-Grainstone contact in the western part of the study area, is not associated with platy dolomite lithofacies, and is directly overlain by massive oolite, requires a separate interpretation. This sand was presumably transported alongshore in the surf zone on the seaward side of the oolite shoals.

Figure 3.12 demonstrates complex marine strandline movements. In two cycles, "platy dolomite I" - "massive oolite I" and "platy dolomite II" - "massive oolite II", the lagoonal to supratidal platy dolomite thickens southeastward at the expense of the overlying littoral oolite. This is a clear record of two marine transgressions following the regression that initiated deposition of the Grainstone formation. The second transgression advanced further, bringing a thin member of subtidal facies, the "argillaceous marker", into the western half of he study area. "Platy dolomite III" is regressive relative to underlying members, and increasingly evaporitic upward. The correlation panel must be read with the understanding that the line of section is oblique to depositional trends, and undoubtedly closer to depositional strike than to regional dip; it is <u>not</u> a "right section".

Gypsum formation

The Gypsum formation is the least studied of the Little Dal formations. It is generally poorly exposed or covered, as at the type section of the Little Dal. In addition, structural involvement of the formation limits, even in theory, the number of complete sections that might be obtained.

Along the trace of Plateau Fault, the hanging wall is the Gypsum formation; thus, in addition to the expected disturbance in detail, an unknown thickness of the formation is faulted out. North and east of Plateau Fault, only five areas are known in which the Gypsum formation has escaped erosional removal at one or other of the post-Little Dal unconformities.

The basal contact is abruptly gradational and apparently conformable. The upper contact, drawn at the top of the highest gypsum bed, is gradational by interbedding.

The greatest known thickness is 530 m, at a section (76-7) in which tens of metres have been eroded at the sub-Upper Cambrian unconformity.

Lithology

Gypsum is the essential rock-type making up the formation in outcrop; all sections contain units tens of metres thick that are described as 98-100% gypsum (X-ray diffraction reveals the presence of anhydrite, as well). The microcrystalline gypsum is predominantly white, dirty-white and pale grey, rarely pale green and pink, the non-white varieties having a small clay content. In the Mountain River drainage area, dark grey, argillaceous, carbonaceous gypsum occurs near the base. Most sections contain one or more reddish zones near the top. Planar lamination predominates, but most sections contain subordinate intervals of mosaic or "chicken wire" nodular fabric, usually associated with enterolithic folds. Gypsum occurs rarely in a nodular displacive relationship in red shale. Salt hoppers are locally present in the transition beds at the top of the formation.

Excepting the regional "carbonate member" and the possible shoreward facies described below, non-sulphate beds nowhere amount to as much as 10 per cent of the formation. They include laminae of argillaceous, gypseous, microcrystalline dolomite, laminae and metres-thick units of red and green gypseous shale, laminae of quartz siltstone, and metres-thick units of sandy (quartz) gypsum and gypsiferous sandstone, usually pink or red. Beds with quartz sand are limited to a thin zone close to the base and another up to 20 m thick at the top of the formation. The total content of fine grained siliciclastic material (clay, silt) shows a distinct increase eastward and southward around the tectonic arc.

Carbonate member. Well-exposed sections in front of the Plateau Fault from Godlin River northeastward reveal a remarkably uniform carbonate member, 88 to 120 m below the top. The member, 7.4 to 20.9 m thick, is everywhere composed of two beds subequal in thickness. The lower bed consists of decimetre-scale nodules of grey microcrystalline limestone in a matrix of dolomite that is brownish grey, microcrystalline, calcareous and argillaceous. At three out of four sections, the nodules are jumbled and bent, and angular fragments of nodule material are present, suggesting that the bed has undergone mass movement. The upper bed consists of grey microcrystalline limestone, finely laminated and massive. At one section, the entire member is dolomitized and contains nodules of grey chert.

Shoreward(?) facies. On the west flank of Tsezotene Range, a unit not known from other areas underlies the Rusty Shale formation. It consists largely of thin-bedded, red, pink and yellow, silty, earthy dolomite, interbedded dolomite and maroon dolomitic shale, and pink dolomitic siltstone. Breccia of yellow dolomite fragments, in part with red mdustone matrix, is common. Yellow and red, sandy, laminated dolomite occurs here and there. Most exposures reveal slump-folds and slump-breccias. By its stratigraphic position, this unit appears to be a shoreward facies of the Gypsum formation. Alternatively, it would be a shoreward facies of the upper Grainstone formation; in the latter case, the Gypsum pinches out at Tsezotene Range.

Interpretation

In view of the limited amount of data available, any attempt at an interpretation of the Gypsum formation must be tentative. The principal evidence is negative; the complete lack of indications of subaerial exposure, coupled with the general planar lamination (showing that depositional structure has not been destroyed by diagenesis) suggests that gypsum deposition was mainly subaqueous. The local, basal carbonaceous gypsum is consistent with such an interpretation. The confinement of quartz sand and most of the quartz silt to the base and top of the formation (and hence, by application of Walther's Law, to the margins of the gypsum body) supports a subaqueous origin; there was no extensive dry surface on which aeolian transport of sand could occur. The environment appears, then, to have been lagoonal, and for much of the extent of the formation, not a complex of many small lagoons but (in view of the continuity of the carbonate member) a single lagoon at least 160 km long and of unknown width.

The carbonate member appears to record a period of reduced salinity in the lagoon, when calcium carbonate, instead of calcium sulphate, was precipitated. The lower, nodular bed is possibly a bed of calcitized anhydrite nodules. If such is the case, the observed distortion may be due, not to mass translation, but to volume changes undergone during complex diagenesis. Such an interpretation avoids the necessity of imagining a single mass flow taking place along a 160 km front, in a lagoonal environment.

Subaqueous precipitation of marine evaporites can take place only in an environment having restricted exchange with normal seawater; a barrier or sill of some kind must have existed to isolate the lagoon from the open sea. As for all of the Little Dal Group, the cratonal shoreline lay to the northeast. Because the Gypsum formation is present along the trace of Plateau Fault, the barrier lies somewhere downdip, southwest of that trace, and is inaccessible. Available evidence supports either of two distinct speculations as to the origin of the barrier, the one appealing to depositional, the other to tectonic processes.

Depositional barrier. The hypothesis of a depositional barrier arises from the close and apparently conformable relationship between Grainstone and Gypsum formations. It has been noted above that in two cycles that make up the lower division of the Grainstone formation, a facies relationship exists between basinward massive oolite of littoral origin and cratonward platy dolomite of lagoonal, intertidal and supratidal origin. A third cycle (the upper division) is observed only in a platy dolomite lithofacies that is partly evaporitic. It is reasonable to infer, therefore, that downdip to the southwest, beyond observation, the upper Grainstone formation passes into a largely oolitic, littoral shoal complex. Under conditions only slightly changed from those obtaining during the lower two cycles, such as stronger winds, or winds blowing more directly onshore, the shoal complex could develop so as to restrict seawater exchange with the lagoon more severely than during earlier cycles, leading to evaporative precipitation of gypsum in the lagoon. Under this hypothesis, the Gypsum formation would be a lagoonal facies equivalent of an upper, younger, basinward development of the Grainstone formation.

Tectonic barrier. The hypothesis of a tectonic barrier isolating the evaporitic lagoon is suggested by thinning of the Grainstone formation at sites that are seaward in terms of depositional facies. In addition, thickness data for the combined Basinal and Platform asemblages can be "isopached" to show southwestward thickening to a depositional axis, followed by southwestward thinning (although numerous other possible isopach configurations can be drawn). Basinward thickening followed by thinning, if real, suggests a model involving an outer-shelf uplift or structural "high" of the sort that has been documented widely, both in the geological record and in the Recent (see Schuepbach and Vail, 1980).

According to this hypothesis, the shelf-edge "high" would have begun to develop at some time during deposition of the Basinal and Platform assemblages and continued to develop during Grainstone deposition, culminating as the "barrier" sill of the gypsum-depositing lagoon.

Rusty Shale formation

The Rusty Shale formation is an easily recognized, recessive-weathering, shaly interval that lies between the very recessive, usually covered, Gypsum formation beneath and the impressive cliffs of the Upper Carbonate formation above. Its generally rusty aspect and ribbed topographic expression are distinctive. The formation is generally confined to the Plateau thrust plate, but is locally preserved in the footwall region.

The basal contact of the Rusty Shale formation is gradational and interbedded. The upper contact is also interbedded (note the presence of identical carbonate rock-types above and below the contact). It is drawn at the top of the highest shale beneath the distinctive member A of the Upper Carbonate formation. This definition of the contact requires the "marker quartzite" to be placed in the Upper Carbonate formation at section 79-2.

Lithology

The varied rocks of the Rusty Shale formation are best discussed in terms of five superposed, informal members, some of which display informative regional variations (Fig. 3.17). Each member would be mappable at a sufficiently large scale.

First carbonate member. The basal, first carbonate member is up to 40 m thick. It is dominated by cryptalgal laminite, mainly dolomitized but locally preserved as limestone. Local differentiation into linked hemispheroids (**Stratifera**) is common. Minor flakestone, intraclast grainstone, and flat-pebble rudstone normally accompany the cryptalgal laminite. A little accessory quartz sand is usually present. Partings and thin interbeds of dark grey to black shale separate the carbonate beds. Minor secondary chert is present locally. The first carbonate grades into the overlying member by interbedding.

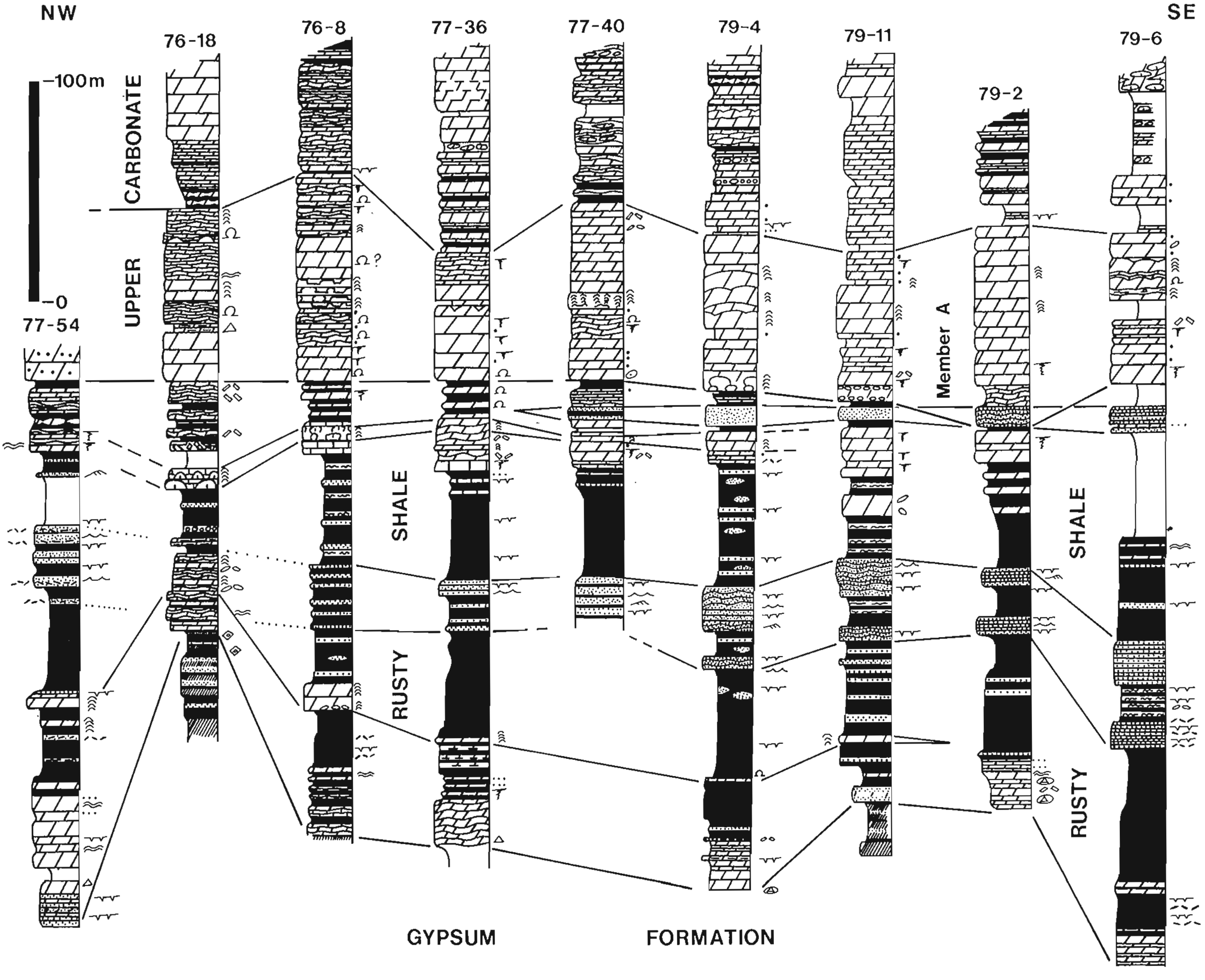


Figure 3.17. *Rusty Shale formation: correlation panel parallel to depositional and tectonic strike. For key to lithologic symbols, see Figure 3.2a.*

First shale member. The first shale member, up to 84 m thick, is dominated by mudrocks. Fissile shale at the base is grey and green, with abundant carbonaceous flakes (**Beltina**). Upward in the member, the mudrocks are green and/or reddish purple, in part fissile and in part hard and nonfissile, inviting the name "argillite". The green phase is characterized in part by zones containing up to 20 per cent of sub-millimetre-sized spheres of ankeritic carbonate. The dense, cleavable, splendently-lustered brown carbonate is easily mistaken for sphalerite; its weathering is largely responsible for the rusty aspect of the formation. In some sections, the member contains laminae, pods and mudcrack fillings of grey, very fine grained and pyritic quartzite. About one third of the way up from the base of the member is a regional, differentiated biostrome characterized by the stromatolite **Baicalia**, preserved mainly as dolomite but locally as limestone. Parts of the biostrome are replaced locally by coarsely crystalline ankerite. The biostrome is thickest in the northwest, and thins to extinction toward the southeast (Fig. 3.17).

Middle sandstone member. The middle sandstone is up to 48 m thick. It is dominated by quartzite that is white, green, purple and very fine grained. Wavy lamination due to low-relief ripple forms of both current and oscillation type is general. Shale-chip conglomerates are abundant. Mudcracks are spectacularly developed, especially the "organic looking" pseudofossil **Manchuriophycus** (Fig. 3.18), an earmark of the formation. In the southeast, where thickness and sandstone content are greatest, the member consists of two sandstone units separated by a unit of red and green mudcracked shale, with or without laminae and lenses of sandstone. To the northwest, the big sandstone units split up, and the member becomes interbedded sandstone and shale (Fig. 3.17).

Second shale member. The second shale member, up to 54 m thick, consists mainly of shale that is fissile, grey to black, carbonaceous, pyritic, and partly interlaminated with grey siltstone. Laminae and pods of quartzite as described above also occur. Mudcracks are present, and outcrops commonly display a "sulphur bloom". The shales contain abundant carbonaceous flakes (**Beltina**), and locally at least, **Tawuia** and **Chuaria** in outstandingly detailed preservation.

Second carbonate member. The second carbonate member is the uppermost of the formation. It is up to 48 m thick and consists mainly of varied dolomites interbedded with subordinate black shale in a cyclical arrangement. The dolomites include ex-lime mudstone, in part with molar-tooth structure, cryptalgal laminite that makes up most of the cycle, flakestone and intraclast and ooid grainstone. The cycle, where complete, is similar to the characteristic cycle of Member B of the Little Dal Upper Carbonate, described below.

Two regional marker beds occur within the second carbonate member. The lower of these, which dies out to the southeast, is a differentiated stromatolite biostrome characterized by **Baicalia.** The upper marker bed, up to 11.8 m thick, is termed the "marker quartzite". Best developed in the southeast, it dies out to the northwest. It consists of quartz sandstone that is grey, white, pale green, and locally purple. Predominantly fine grained, and thus distinct from lower sandstones, it is partly medium grained only in the extreme southeast, at section 79-6. The proportion of the marker having different primary structure varies greatly between localities. Much of it is in thick massive beds that are cryptic as to fine structure. Elsewhere, thin planar bedding or slightly wavy parallel lamination may dominate.

Figure 3.18. *The pseudofossil **Manchuriophycus**, characteristic of the Rusty Shale (actually mudcracks).*

Ripple cross-lamination and shallow trough cross-lamination appear locally at the top. Mudracks and oscillation ripple marks are rare. The marker quartzite includes minor dolomite and sandy dolomite in some sections.

Interpretation

The remarkable persistence of thin members of the Rusty Shale along the tectonic arc (Fig. 3.17) shows the orientation of depositional strike in at least a general way. The anomalous, shortened section at 76-18 may have beds faulted out, although no fault was recognized.

Isopachs for the Rusty Shale formation are shown in Figure 3.19. The overall pattern superimposed on southwestward thickening is arcuate and congruent with the tectonic arc. Contouring is biased, first, by knowledge of depositional strike, and second, by observed harmonious thickness variations in the Rusty Shale and in members A, B and C of the Upper Carbonate. That is, where one is thick, the others are thick, and vice versa. Although simpler, less inflected isopachs could be drawn for each of these four units over part of their extent, certain inflections for each unit are unavoidable, and it is possible to draw isopachs for each unit that are congruent with those for each of the others, without violating available data. The significance of this will be evident later.

The entire formation is peritidal in origin. The transition from Gypsum formation to the first carbonate member mainly records filling of the lagoon, and the transgressive advance of peritidal carbonate sediments laid down at and near a rather sheltered, gently sloping shoreline. The supply of siliciclastic detritus at this time was small; it is notable that it is the most basinward sections that contain more quartz sand.

The first shale and middle sandstone members, with their mudcracks, shale-chip conglomerates and ripple marks (including small-scale oscillation ripples) are the deposits of extensive tidal flats. They record a manifold increase in the supply of siliciclastic mud and sand that smothered carbonate-producing mechanisms, except for the **Baicalia** biostrome that seems to record a brief pause in the supply of detritus.

The second shale member, carbonaceous and lacking the oxidized colours of the lower shales, was apparently laid down subaqueously. The local mudcracks may record temporary exposure or may be of subaqueous origin. The overall "low-energy" aspect of the member suggests a

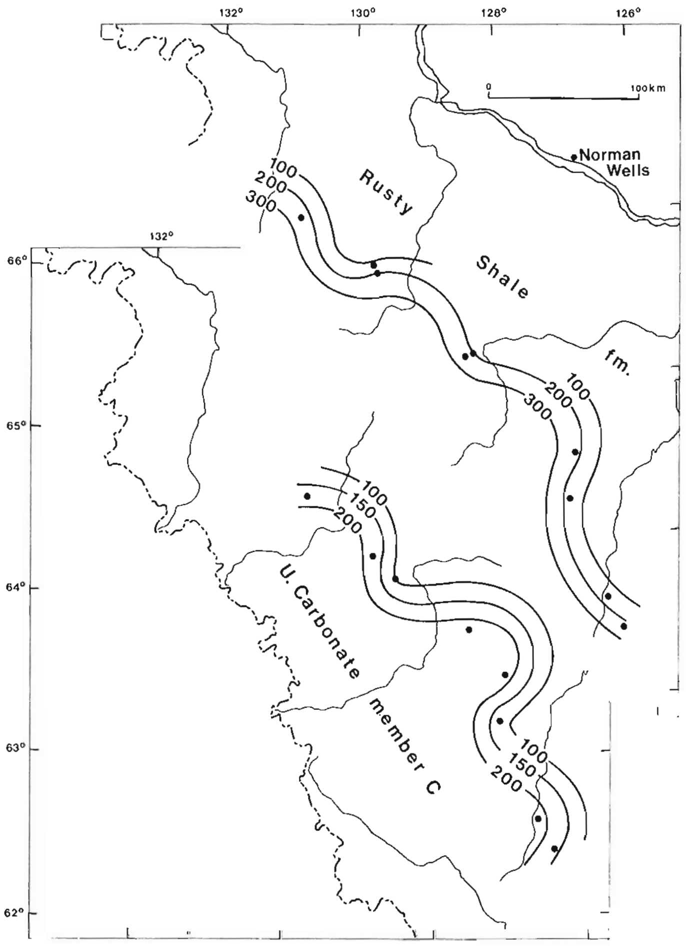


Figure 3.19. *Tentative isopachs, Rusty Shale formation and Member C of Upper Carbonate formation.*

lagoonal origin, the lagoon forming behind a stromatolitic littoral complex represented by the lower part of the second carbonate member. The presence here of **Tawuia** and **Chuaria**, in a sedimentary association totally distinct from that of the Basinal assemblage, raises the question as to whether these primitive plants lived in the open sea or in the lagoons. It seems unlikely that they could be adapted to both environments.

The upper part of the second carbonate member consists largely of cycles involving a grainstone unit. Such shallowing-upward cycles with a basal grainstone are reasonably attributed to migrations of a littoral zone fronting the sea (James, 1979), and are so viewed here.

The "marker quartzite" is an extraordinarily informative unit. Its internal structure and relationship to enclosing strata indicate a beach deposit. Its distribution, consistent with what is known of grain-size variation, records movement of quartz sand northward. When the Rusty Shale formation as a whole is considered, it is clear that from southeast to northwest along the tectonic arc, quartz sand decreases and carbonate strata increase in every level, a distribution mimicking that observed for the Grainstone formation.

Clearly, we have here a record of a persistent supply of siliciclastic detritus, from somewhere southeast of the last exposures of the Little Dal. It seems equally clear that, as for the Mudcracked formation, mud and sand were moved northwestward mainly in the high-energy littoral zone. Sand that reached the tidal flats was mainly washed or blown inland from the littoral zone. Finally, the question of the cutoff of sand supply arises. In view of the interpretation of the base of the Upper Carbonate as transgressive (see below), it appears to be yet another case of the ponding of detritus in the estuaries by relative sea level rise.

Upper Carbonate formation

The Upper Carbonate formation, as amended here, comprises the thick, cliff-forming, stromatolite-rich carbonate rocks that cap the Little Dal Group (Aitken et al., 1978). It has been erroneously referred to as "Little Dal Formation, sensu stricto" (Aitken, 1977).

The base of the formation is drawn arbitrarily at the top of the highest shale, in the zone of transition with the Rusty Shale formation. Relationships at the top of the formation are discussed in the section following.

Lithology

The Upper Carbonate is characterized by the dominance of carbonate rocks (almost entirely dolomites), cyclicity, abundance of columnar branching stromatolites, and colourful outcrops striped in shades of grey, yellow, orange, and red. In the course of fieldwork, four rather imprecisely defined members were widely recognized. Subsequent, detailed comparison of measured sections established firm definitions and the regional validity of the four members (Fig. 3.20).

Member A. Member A, 57 to 92.5 m thick, consists of dolomite in thick, massive, resistant beds generally weathering dark grey. Ex-lime mudstone with prominent to spectacular molar-tooth structure is present everywhere, displaced to varying degrees by thick stromatolite biostromes and bioherms. Ex-grainstone, notably ooid grainstone, occurs in nearly every section. Cryptalgal laminite is usually present; chert is rare.

Member B. Member B, 30 to 150 m thick, is recognized by its slightly recessive, ledgy outcrop that reflects its cyclicity, and by the reddish and yellowish weathering colours of its mudrocks and argillaceous dolomites. Much of the member consists of small-scale cycles, 1 to 6 m thick. Although the cyclicity is obvious, high variability makes it difficult to designate an "ideal" cycle, without resort to an approach more detailed than has been possible. Where ex-grainstone, notably ooid grainstone, is present, it tends to rest on an erosional base, and is taken as unit A in a shallowing-upward, ABC cycle. In broad terms, the rest of the cycle consists of either mudstone/shale or locally mudcracked muddy dolomite, or ex-lime mudstone with molar-tooth structure (B unit), gradationally overlain by cryptalgal laminite that is argillaceous, silty or sandy in some cycles and generally mudcracked, with local, small-scale teepee structures (C unit). However, oolite also occurs interbedded with laminite in the C units, in violation of the stereotyped cycle.

Algal stromatolites are relatively inconspicuous. Where columnar and columnar-branching stromatolites occur, they are embedded in the A and B units. Stromatolites associated with the C laminites are of simple domal form. Beds of ripple cross-laminated microcrystalline dolomite, assumed to be ex-calcisiltite, occur in the cyclical B units.

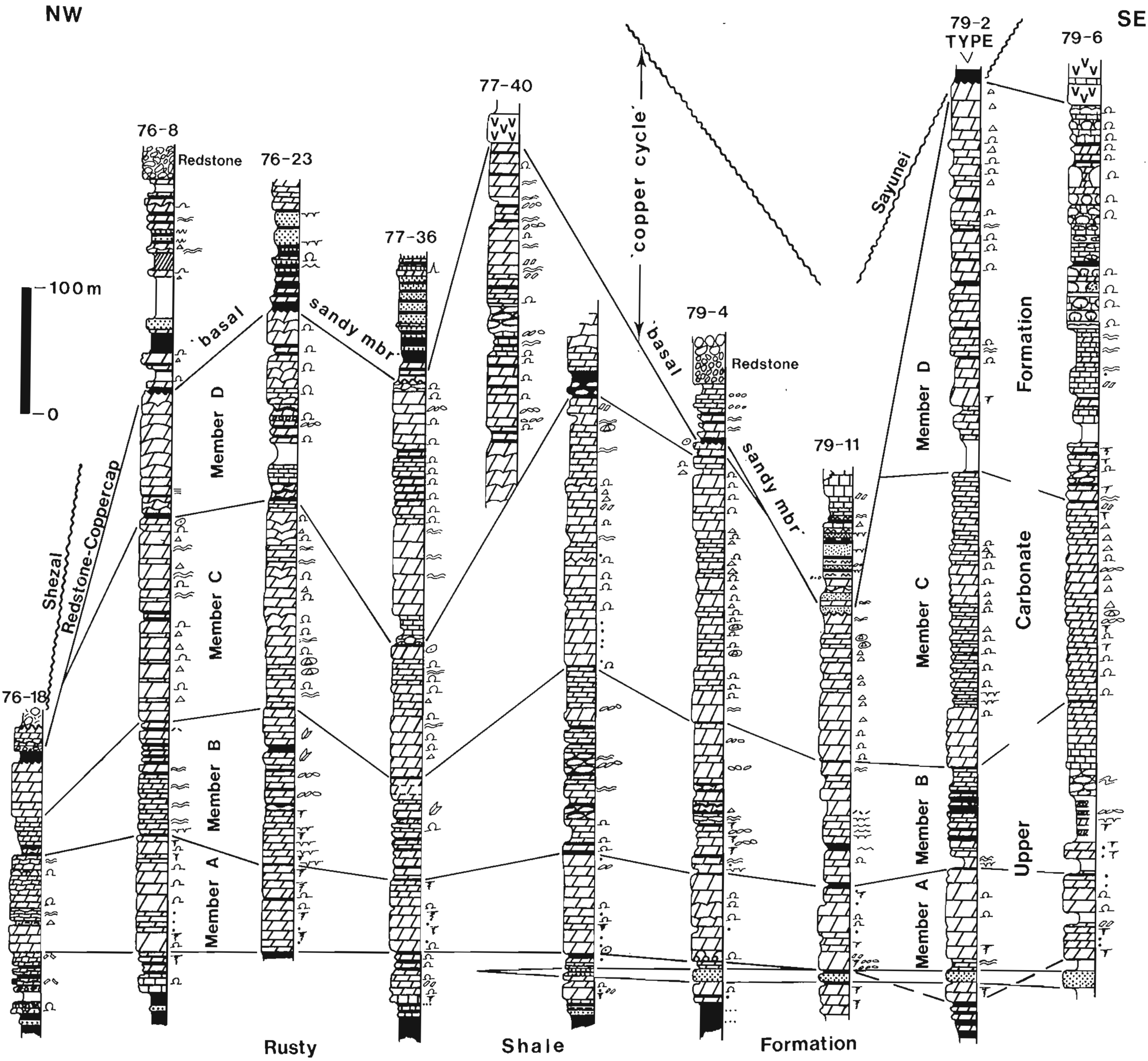


Figure 3.20. *Upper Carbonate formation: correlation panel, mainly along the tectonic and depositional strike. For key to lithologic symbols, see Figure 3.2a.*

Minor constituents of the member are flat-pebble rudstones, flakestones, and beds with millimetre-scale, tabular, silicified objects that may be replaced gypsum crystals, and vugs of similar form. Brecciated dolomite beds occur here and there. Chert is rare. Local, thick beds of molar-tooth dolomite and ex-oolite near the base suggest an intertongued contact with Member A.

Member C. Member C, 102 to 212 m thick, consists essentially of dolomite with much secondary chert. Thick, massive, resistant units dominated by bioherms and biostromes with varied and well preserved branching stromatolites are characteristic. At most localities, much cryptalgal laminite is present. Units of ex-calcisiltite(?) as described above are fairly prominent. Small tabular vugs, apparently after gypsum, occur here and there. Mudcracks were observed at one section only.

Member D. Member D, up to 342 m thick, consists largely of stromatolitic carbonate units like those that dominate Member C, interrupted by widely separated beds of dark grey, green-grey, brown and purple siliciclastic mudrocks. Small-scale, shallowing-upward cycles like those of Member B occur, as do cyclical alternations of ex-grainstone with stromatolite biostromes and of ex-mudstone with biostromes. All sections contain units of stromatolite bioherms several metres high. These are built in part by a small branching stromatolite similar to the reef-builder of the Basinal assemblage. Chert is rare. Identification of the lower contact on overall lithologic character is difficult. Fortunately, the basal beds of Member D are a marker zone consisting of a unit of dolomite nodules in a matrix of reddish to purplish argillaceous dolomite or mudstone, overlain by grey to lavender, laminated, shaly limestone (locally dolomite), grading to calcareous shale (the "clinky shale" of local parlance). At several localities, the base of the marker zone is a thin bed (rarely more than 25 cm) of oncolites in a sandy, glauconitic matrix. This bed strongly suggests a depositional break at the contact.

Interpretation

Figure 3.19 presents isopachs for Member C of the Upper Carbonate formation, drawn so as to be congruent with isopachs for the Rusty Shale formation without violating any thickness data. Isopachs for Members A and B can be drawn to a similar configuration. The procedure is justified by the observation that the Upper Carbonate members and the Rusty Shale formation change thickness in concert (Fig. 3.17, 3.20).

The gradational, interbedded, basal contact of the Upper Carbonate records two changes from conditions prevailing during terminal Rusty Shale deposition, first, cutoff of supply of terrigenous sediment, and second, assumption of dominance by the high-energy littoral (grainstone) and low-energy subtidal (lime mudstone) carbonates that form parts of the underlying cycles. The latter change is a record of transgression, which accounts for the cutoff of siliciclastic sediment. Although Member A may be cryptically cyclical, the overall effect is that of a transgression of greater amplitude (though probably slower) than the multiple transgressions recorded by the terminal cycles of the Rusty Shale formation.

The Member A transgression resulted in the building of an extensive, shallow-water, low-energy, almost mud-free platform. Terrigenous mud in small amounts gained access to this platform only at times of deepest water cover, as the Member B shallowing-upward cycles were laid down. Many cycles ended with buildup of largely cryptalgal carbonate strata to high intertidal and supratidal levels, reaching a static condition until renewed and apparently rapid deepening caused the invasion of a high-energy shoreline and subsequent deposition of subtidal to low intertidal lime and/or siliciclastic muds.

The lower two members of the Upper Carbonate, and their relationship to the Rusty Shale, bear a remarkable resemblance to the Upper Cambrian Lyell Formation of the southern Canadian Rockies and its relationship to the underlying Sullivan Formation. The paleogeography and sedimentology of the latter case were recently outlined by the author (Aitken, 1978), and may have application to the interpretation of the Little Dal units.

The correlation panel (Fig. 3.20) is drawn along the depositional strike, and shows only slight evidence for a facies relationship between Members A and B. Application of the Upper Cambrian model to the Upper Proterozoic case suggests that if data were obtainable to construct a correlation panel aligned with the depositional dip, it would show both the Rusty Shale-Member A and the Member A – Member B contacts rising stratigraphically northeastward. In this context, Member A would be a marginal facies, and Member B an interior facies of an expanding carbonate shoal complex advancing cratonward over the deposits of a muddy, shallow inshore "basin".

Member C of the Upper Carbonate formation records continued development of the carbonate platform, again in the absence of significant terrigenous clastics. Stromatolite complexes fringing the platform may have provided a screen against mud invasion.

As noted above, Member D commences with a sandy, glauconitic oncolite bed that may record a depositional hiatus. This is succeeded immediately by first, nodular, and second, finely laminated argillaceous carbonate strata, suggesting lagoonal deposition. The remainder of the member is fundamentally cyclic and somewhat similar to member B, although incursions of siliciclastic mud were less common. Given essentially similar paleogeographic settings, the reasons for the weak development of stromatolites in Member B and their outstanding development in member D are unclear.

Upper Contact, and Strata Overlying the Little Dal Group

Where the Upper Carbonate formation is overlain by pillowed basaltic lava, the contact is conformable. No conglomerate is present, no truncation of bedding is seen. The base of the lava has developed load casts into underlying lime mud at several localities, with attendant injection of the soft carbonate sediment upward into the base of the lava flow. In the Thundercloud Range, limestone deposition was briefly resumed during a pause between eruptions.

Subdivision of the Upper Carbonate into members, and detailed correlation of these members (Fig. 3.20, 3.21) demonstrate that the overlying member of sandstone, dolomite, siltstone, and mudstone with local gypsum and conglomerate, that succeeds Member D, is unconformable on underlying units. This member, here termed Basal Sandy member of the "copper cycle", is recognized by the appearance of sandstone above a thick interval of carbonate strata and minor mudrocks containing no detrital material coarser than silt, except for a little accessory quartz sand in rare carbonate beds. This is accompanied by a marked decrease in the ratio of carbonate to non-carbonate strata. Marked cyclicity is evident at a number of localities. In common with the rest of the "copper cycle" (see below), and in marked contrast with the members of the underlying Upper Carbonate formation, the Basal Sandy member displays pronounced lateral variation in thickness and in the proportions of its different rock types, although the character of each rock-type is reasonably constant.

Cyclicity in the Basal Sandy member is best developed at section 79-4. There, cycles 0.3 to 4 m thick are as follows:

1. Shale, purple, etc.
3. Dolomite, grey, pink, microcrystalline, cryptalgal laminite; commonly sandy, or with sandy laminae.
2. Sandstone, red, pink, purple, grey, fine and medium grained, thin and medium bedded, flaggy to massive; generally planar-laminated.
1. Shale, purple, sandy, hackly; oscillation ripple-marks. A thin pebbly lag deposit at the base in some cycles.

At section 79-4, two thin pebble conglomerates are interbedded with the mixed sequence below the thick, coarse conglomerate that most geologists would take as "base of Redstone River Formation".

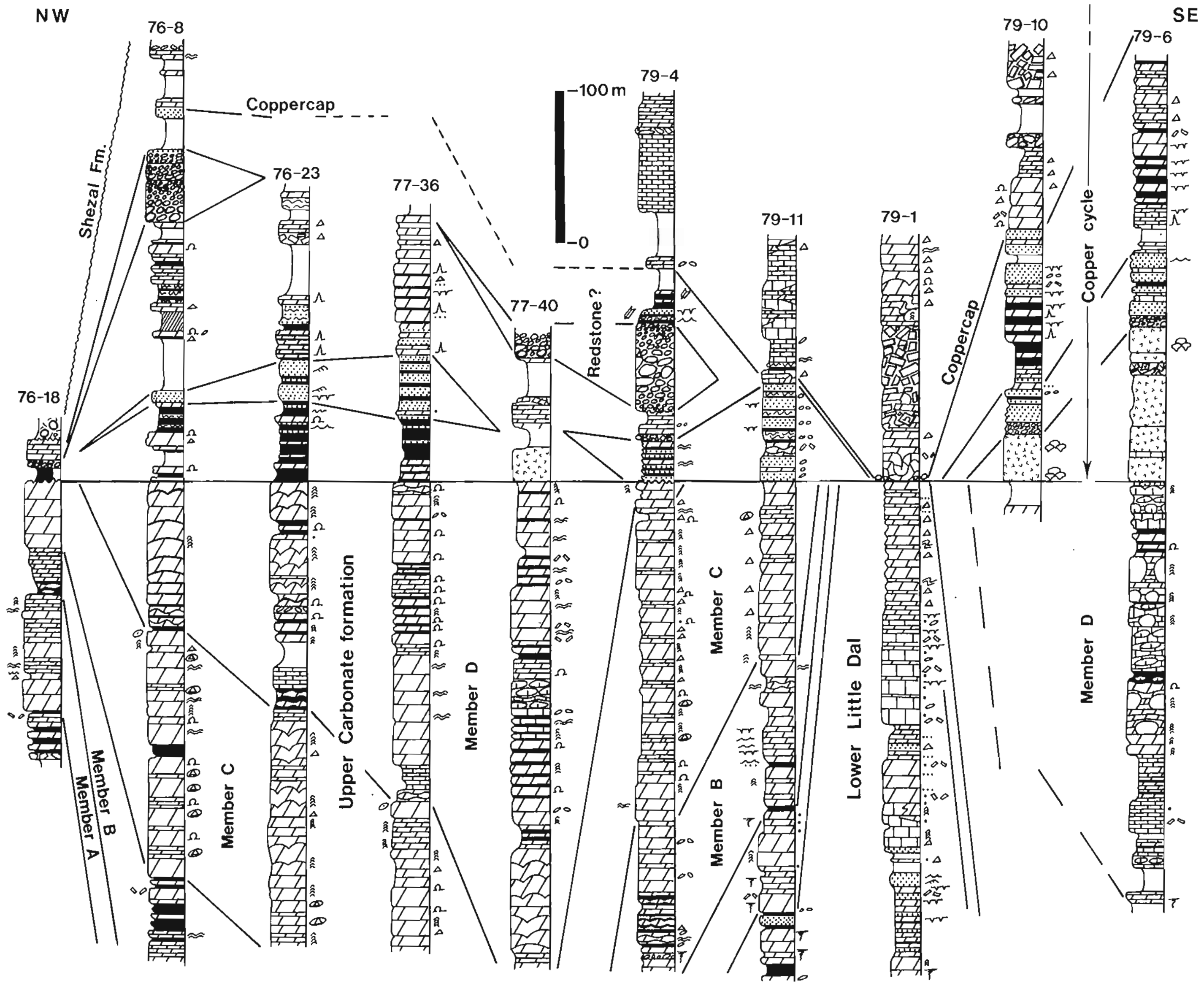


Figure 3.21. *Correlation panel: relationships at the base of the 'copper cycle'. For key to lithologic symbols, see Figure 3.2a.*

Where the basaltic lavas are present, but not elsewhere, the overlying Basal Sandy member strongly reflects basaltic provenance at its base; higher sands are progressively more quartzose and locally pyritic. Dominant primary structures are planar lamination and low-angle planar and tangential cross-sets 10 cm and less thick. Mudcracks are present locally.

Thin beds of matrix-supported, sharpstone and roundstone dolomite-pebble conglomerate occur erratically in the member; they interrupt both sandsone and dolomite units.

Units of grey, green, pink and purple, ripple cross-laminated siltstone, or interlaminated siltstone and shale, commonly pyritic, occur in some sections.

At sections in the most southerly extent of 'copper cycle' exposures (e.g. 79-6, 79-10), sandstone diminishes upward, and the member becomes an alternation of red or purple, dolomitic, commonly silty mudstones with units of laminated, cherty dolomite or limestone in which mudcracks, dewatering structures and teepee structures are conspicuous and beds of flakestone prominent. This sequence is directly overlain by carbonates of the Coppercap Formation, and given the absence of polymict conglomerate and evaporites, would be mapped as atypical Redstone River Formation by most.

At section 76-8, a unit of pale grey shaly dolomitic gypsum is exposed within the Basal Sandy member. Elsewhere, brecciated carbonate units suggest that evaporites have been removed by solution.

At a number of sections, the highest beds are conspicuous, pale grey-weathering beds of microcrystalline, laminated dolomite with large-scale mudcracks and teepee structures, much replaced by dark grey to white chert. Chert replacements of gypsum rosettes occur in this unit at section 77-36.

Where coarse framework conglomerates mark the base of the Redstone River Formation, the sub-conglomerate beds, that is the Basal Sandy member, have been placed in the Little Dal (e.g., Jefferson, 1978; Ruelle, in press). Elsewhere, the contact has been placed at the base of the evaporites (ibid.). Problems are brought to light at such localities as 76-8, where gypsum occurs well below the lowest conglomerate.

The present study provides a solution to the problem, in recognizing that the onset of the depositional cycle largely represented by the Redstone River and Coppercap formations is recorded at a horizon lower than the accepted base of the Redstone River, namely, at the newly-discovered unconformity below the Basal Sandy member. Communication will be improved by uniting the unnamed, basal, sandy member with the Redstone River and Coppercap formations as the informal "copper cycle". Resolution of the problem of formal nomenclature is left to those involved in detailed study of the post-Little Dal succession.

The best nomenclatural solution for the volcanic rocks is not yet determined. To erect a new formation for the thin discontinuous lavas would be inappropriate at the present stage and scale of mapping in the region. Reasonable arguments may be made for either retaining the lavas within the Little Dal, or transferring them to the "copper cycle". The former option is favoured by the conformable base of the lavas, and the evidence (volcanic-derived conglomerates) of erosion at their top. Furthermore, the paleomagnetism of the lavas (Morris and Park, 1981) records weathering of the upper part of the lavas in a magnetic field of different orientation from that in which the lavas cooled originally; a significant hiatus is undoubtedly present. On the other hand, localization of the lavas above the thickest sections, member-by-member, of the Little Dal Upper Carbonate shows that the lavas were erupted into paleostructural lows, where the sub- "copper cycle" erosion recorded elsewhere need not have occurred.

In this report, the lavas are tentatively assigned to the "copper cycle", for the following reasons:

a. In terms of stratigraphic level, the base of the lavas correlates with the base of the Basal Sandy member of the "copper cycle".

b. Both eruption of lava and the renewed influx of sandy detritus are viewed as consequences of the onset of faulting that is more strongly expressed in Redstone River sedimentation and apparently reached a climax immediately prior to the onset of Rapitan deposition.

c. The chemistry of the Redstone River evaporites suggests continued vulcanism (Ruelle, in press); that vulcanism or the copper-rich lavas are probably primary sources of the stratabound copper of the Redstone Copper Belt (Ruelle, in press).

Figures 3.20 and 3.21 reveal a degree of correspondence between depositional "thicks" and "thins" in the "copper cycle" and similar features in the Little Dal, downward at least to the level of the Rusty Shale Formation. This correspondence suggests that the differential vertical movements ultimately expressed by faulting during the "copper cycle" had a long history, beginning in a subtle way at least as far back as Rusty Shale time. This theme is examined in greater detail by C.W. Jefferson, personal communication, 1980.

The Redstone River-Coppercap cycle has previously been included within the Mackenzie Mountains supergroup, presumably because the sub-Rapitan unconformity provided a natural upper boundary, and because the base of the Redstone River was only locally unconformable. Recognition of an unconformity beneath the "copper cycle" suggests that the top of the supergroup should perhaps be drawn to exclude the "copper cycle".

SUMMARY AND CONCLUSIONS

Deposition of the two kilometre-thick Little Dal Group commenced with deposition, during southeastward onlap, of the Mudcracked formation - a complex of littoral, subtidal, intertidal and lagoonal, mixed detrital and carbonate strata. Siliciclastic sediment was probably derived from a source to the southeast, as for all sandy formations of the Little Dal.

Deposition of the Mudcracked formation ended with the transgressive passage of a high-energy shoreline that laid down ooid and intraclast grainstones. In the southeast, these were the first deposits of a persistent, fairly high-energy carbonate platform, recorded as the Platform assemblage. In the northwest, the grainstones were succeeded by subtidal deposits, dominated by rhythmites, partly turbiditic carbonates, of the Basinal assemblage. A red, nodular mudstone facies deposited proximally to the platform may record a dearth of plant nutrients, while giant stromatolitic reefs of distal position may record supply of nutrients from a cratonal source.

The boundary between Basinal and Platform assemblages is marked by an extensive, thick, multi-story stromatolite biostrome. The stromatolites appear to have buttressed the platform-edge slope, rather than forming a peritidal rim to the platform.

With depositional filling of the basinal region, shallow-water deposition was re-established over the entire region of Little Dal exposures, and regional differentiation of facies greatly subdued. The Grainstone formation, a complex of littoral grainstones and platy dolomites (ex-mudstones) of lagoonal, intertidal and supratidal origin, records two transgressions followed by regression and the onset of evaporitic conditions.

At the end of Grainstone deposition, a sill or barrier of unknown character developed somewhere downdip, to the southwest, and formed a lagoon against the cratonal shoreline. In this lagoon, thick, fairly pure gypsum and anhydrite accumulated as the Gypsum formation.

At the end of Gypsum deposition, relative sea level rise drowned the sill, and a carbonate-depositing shoreline advanced across the former lagoon, initiating deposition of the Rusty Shale formation. The basal carbonates were overwhelmed by a new pulse of siliciclastic sediments from the southeast; these were mainly laid down in tidal-flat and lagoonal settings. As the supply of detritus waned toward the end of Rusty Shale time, carbonate deposition was renewed in mixed clastic-carbonate, shallowing-upward cycles, and a (quartz) sandy beach was briefly established in the south.

With the onset of deposition of the Upper Carbonate formation, a regional, frequently stromatolitic carbonate platform was again established; minor amounts of clay reached the middle of the platform only following pulses of relative sea level rise. Thickness variations in the members of the Upper Carbonate are harmonious with those of the Rusty shale, and record slight differences in subsidence rate that may be premonitory to faulting manifested during the ensuing "copper cycle". The base of a cyclical member of sandstone, dolomite and mudrocks, formerly placed in the top of the Little Dal, unconformably truncates underlying members. It is accordingly removed from the Little Dal and grouped with the Redstone River and Coppercap formations as the informal "copper cycle". Locally underlying, basaltic lavas are also tentatively transferred from the top of the Little Dal to the "copper cycle", although an erosional hiatus above the lavas is recognized.

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4.

CORRELATION OF UPPER PROTEROZOIC STRATA IN THE CORDILLERA: PALEOMAGNETISM OF THE TSEZOTENE SILLS AND THE LITTLE DAL LAVAS

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Morris, W.A. and Park, J.K., Correlation of upper Proterozoic strata in the Cordillera: paleomagnetism of the Tsezotene sills and the Little Dal lavas; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 73-78, 1981.

Abstract

Two possible correlations have been suggested between the Proterozoic sequences of the Mackenzie and Wernecke mountains in the northern Cordillera of Canada. Crucial to these correlations is the possible existence of an unconformity at some point immediately above, or near the top of, the Little Dal Group in the Mackenzie Mountains Supergroup. Paleomagnetic results from the Little Dal Group sediments, the overlying Little Dal lavas, and diabase sills in the older Tsezotene Formation suggest that there is a major discontinuity at some point below the lavas. Comparing poles from these rock units with others from Laurentia provides an estimate of 700-780 Ma for the age of this unconformity in the Mackenzie Mountains. In the absence of equivalent data from the Wernecke Mountains it is not certain to which unconformity this should be related.

Résumé

On a suggéré deux corrélations possibles entre les successions protérozoïques des monts Mackenzie et des monts Wernecke, situés dans le nord de la Cordillère canadienne. Un point crucial pour l'établissement de ces corrélations, est l'existence possible d'une discordance stratigraphique en un point situé immédiatement au-dessus, ou près du sommet du groupe de Little Dal, dans le supergrope des monts Mackenzie. Les résultats paléomagnétiques obtenus sur les sédiments du groupe de Little Dal, les alves sus-jacentes de Little Dall et les sills de diabase de la formation plus anciennes de Tsezotene, semblent indiquer qu'il existe une importante discontinuité en un certain niveau au-dessous des laves. En comparant les pôles de ces unités rocheuses aux pôles d'autres unités de la région laurentienne, on a estimé l'âge de la discordance stratigraphique des monts Mackenzie à 700-780 Ma. En l'absence de données équivalentes pour les monts Wernecke, on ne peut dire avec certitude à quelle discordance stratigraphique on pourrait rattacher la discontinuité observée.

INTRODUCTION

A number of possible lithostratigraphic correlations have been proposed for the post-Hudsonian rocks of the North American Cordillera, and the northwest margin of the Canadian Shield. Criteria employed in the erection of these correlations include stratigraphic position, lithologic similarity, structural and metamorphic state, and the presence of particular stromatolite types. Most crucial to these correlations, however, is the recognition of major unconformities. These are usually adopted as the boundaries of a particular rock sequence. Correlation of sequences from one region to the next is based on the assumption that the bounding unconformities in the regions under comparison are at least partially contemporaneous. Unfortunately, in the absence of diagnostic fossils by which the age of rock strata can be established, it is difficult to prove that Precambrian unconformities are indeed synchronous. Further, in cratonic regimes where tectonism is limited, it is possible that a number of paraconformities may be totally missed, and hence erroneous correlations may result.

In resolving this problem of age control there are two possible approaches: (1) geochronology - to establish absolute ages for particular rock units; and (2) paleomagnetism - to establish and compare pole positions characteristic of certain rock units. (Clearly both methods have limitations mainly related to the rock types present, and also to the possible presence of later postdepositional alterations.) In this paper new paleomagnetic data from two rock units in the Mackenzie Mountains, and a review of the paleomagnetic and radiometric evidence relating to the possible significance of one of the Mackenzie Mountain unconformities are presented.

PALEOMAGNETIC RESULTS

The Tsezotene Sills

The stratigraphic succession in the Mackenzie Mountains Supergroup is shown in Figure 4.1. The upper part of the Tsezotene Formation contains a major, regionally extensive diabase sill from which samples were collected at 11 sites over a distance of about 200 km (Fig. 4.2). Material was collected from an additional site (site 1) in a lower sill. The general paleomagnetic results from these sites are presented here with a more detailed account given elsewhere (Park, in press).

As many as four separate magnetizations were uncovered in single specimens using a two-stage method of treatment involving alternating field (AF) treatment (generally 20 to 45 mT) followed by thermal demagnetization (up to 650°C). Under AF, directions generally migrated to a shallow westerly direction. With increase in temperature during subsequent thermal treatment, successive magnetizations associated with probable maghemite and magnetite (?) (R_R, west), magnetite (?) (R_N, east), and reverse (B_R, west) and normal (B_N, east) hematite (?) were recovered. The magnetization directions as depicted in Figure 4.3 were resolved using vector diagrams (Zijderveld 1967) and vector subtraction techniques (Park and Roy, 1979; Park, in press). R_R is a low inclination westerly magnetization with unblocking temperatures (T_{UB}) usually ranging from about 200°C up to between 350° and 500°C, and R_N a more easterly magnetization of shallow to steep inclination having T_{UB}'s from 500°C up to near the 575°C Curie point of magnetite. The R_R direction is well defined

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Contribution of the Earth Physics Branch No. 937.

(38 specimens; mean direction: D=272°, I=0°; α_{95}=6°), but R_N is very scattered (29 specimens; mean direction: D=94°, I=+35°; α_{95}=13°). Site 1 from the lower sill yields a very steep direction (crosses in Fig. 4.3a).

The higher T_{UB} magnetizations above 575°C were probably acquired during the alteration of magnetite to hematite over a period of one or many field reversals. (A small amount of hematite probably formed by the inversion of maghemite above 350°C during the heating experiments, but it would account for far too small a proportion of the measured hematite intensity to significantly bias the B directions.) This magnetite alteration probably occurred relatively soon after the acquisition of R, since the predominant B component (B_R) has a direction (18 specimens; mean direction: D=269°, I=01°; α_{95}=9°) within 3° of R_R and the less well-defined B_N (7 specimens; mean direction: D=094°, I=-10°; α_{95}=22°) is antiparallel to B_R within error.

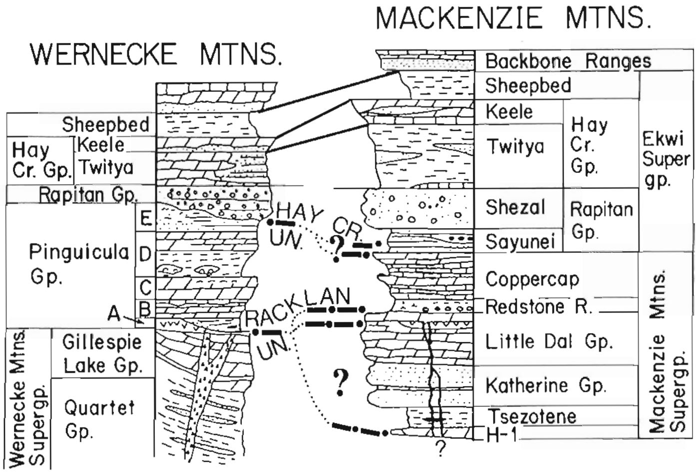

Figure 4.1. *Proterozoic stratigraphy in the Mackenzie Mountains (Northwest Territories), and the possible options for its correlation with the Wernecke Mountains sequences (Yukon) by means of the inferred unconformities (un.) present. The figure is in part after Eisbacher 1978b (Fig. 12.1).*

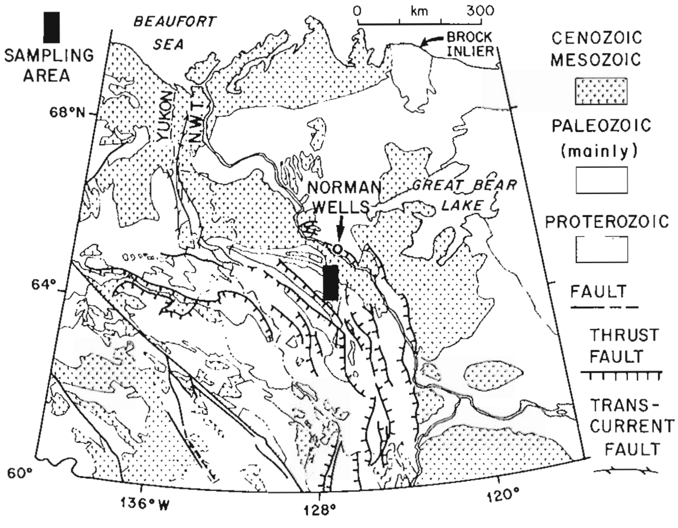

Figure 4.2. *Location map showing the sampling area (solid rectangle).*

Because of the close agreement between the mean R_R and B_R (and also B_N) directions (<3° apart) we believe that R_R is in fact the initial magnetization direction. This agreement between magnetite (and maghemite) and hemetite components precludes a remanence acquisition through later geological processes. In fact, the predominant maghemite portion of R_R would be less susceptible to later geological processes other than heat. The less well defined and quite scattered R_N does not lie in the same direction (irrespective of sign), and this could be attributed to a number of causes. R_N could be a resultant magnetization composed of the initial R_N component (antiparalled to R_R) and a chemical remanent magnetization of varying proportion developed during the Laramide uplift and folding, thus producing the smeared distribution of directions ranging to the Cretaceous field direction. (It is of note that compared to other sites most of the directions of the more extensively altered site 1 are significantly near the Cretaceous direction.) The apparent lack of change in R_R would then be owing to its being predominatly carried by maghemite. R_N could also be explained as the resultant of an initial R_N (east) component and R_R. A third alternative is that the individual R_N directions could represent actual directions obtained during the course of a field reversal either by cooling by a pure thermal acquisition process or soon after by a chemical acquisition process. At present it is not possible to pinpoint the true nature of R_N, but it is probable that R_R represents the initial magnetization of the sill.

The Little Dal Lavas

At a number of localities in the Mackenzie Mountains an amygdaloidal lava occurs in the upper part of the Little Dal Group (Fig. 4.1, 4.2). Ten sites were sampled for paleomagnetic analysis from two sections through the lava outcropping at 63° 48'N, 231° 30'E. At each of these sections the lava could be subdivided into two distinct portions: a lower portion of clean unaltered lava, and an upper portion exhibiting extensive subaerial weathering. Paleomagnetically these two portions have quite different remanence directions. Here, we present a general discussion of the acquisition history of the three magnetizations observed from the two sections (Fig. 4.4). A more detailed discussion will be given elsewhere (Morris, in preparation).

Direction LD-L (4 sites; mean direction: D=304°, I=20°; α_{95}=7°; pole 24°N, 115°E) is considered to represent the earliest phase of magnetization at this locality, and possibly dates from the initial extrusion of the lava. LD-L is found only in the lowermost parts of the lava, never in the uppermost weathered part. It is always unidirectional and in any one specimen never coexists with another remanence direction. At site 10, one half of the specimens carry LD-L and the other half a mixture of LD-M and LD-N. The LD-L remanence resides in both magnetite and hematite with no significant differences between the directions. Thermal demagnetization experiments on specimens with this direction indicate that it has a restricted unblocking window of only about 25°-50°C for both the hematite (at c. 660°C) and magnetite (at c. 570°C) components.

LD-M and LD-N (Fig. 4.4) are found only in the upper weathered part of the lava. These two directions commonly coexist in the same specimen. On the basis of their differing remanence characteristics the preferred acquisition sequence is LD-M → LD-N. LD-M (5 sites; mean direction: D=266°, I=48°; α_{95}=11°; pole 24°N, 159°E) is always found as the higher temperature phase relative to LD-N. Specimens whose T_{UB} spectra are dominated by LD-N (8 sites; mean direction: D=270°, I=71°; α_{95}=4°; pole 47°N, 175°E) often show a transition to the LD-M direction upon thermal treatment. In many cases, companion specimens to those which yield two directions upon thermal demagnetization do not show the same two components upon AF treatment; the coercivity spectra being dominated by the LD-N remanence. From this evidence it is suggested that the magnetic mineral carrier of the LD-N remanence is probably geothite. LD-M, on the other hand, is most commonly observed as a hematite residing remanence.

There are no conclusive tests to establish the absolute ages of any of these magnetizations. In the absence of any evidence to the contrary, we assume that the association of LD-L with the unaltered lava may be taken as indicating that LD-L represents the initial extrusion of the lava. Certainly the fact that the magnetite and hematite components of LD-L are in close directional agreement points to an initial magnetization. A fold test verification of this assumption would be most useful. LD-M appears intimately associated with the hematization of the upper part of the lava. This event may have taken place quite early in the evolution of the Mackenzie Mountains, since lava pebbles in a conglomerate below the immediately overlying Redstone River Formation show well-defined hematized rims which appear to have been formed prior to the incorporation of the pebbles in the conglomerate. It is probable that it is this hematization event that is recorded by the hematitic magnetizations of the upper lavas. Hence LD-M could predate deposition of the Redstone River Formation.

The age of the LD-N direction is the least well constrained. Comparing a similar pole from the Little Dal Group to the Phanerozoic pole path for North America, Park (in press) preferred a Cretaceous age for this remanence, associating its acquisition with the Cretaceous orogenesis in the Mackenzie Mountains. Hence the possible age range for this component is Late Precambrian to Cretacous. We will not consider the LD-N direction further in this paper.

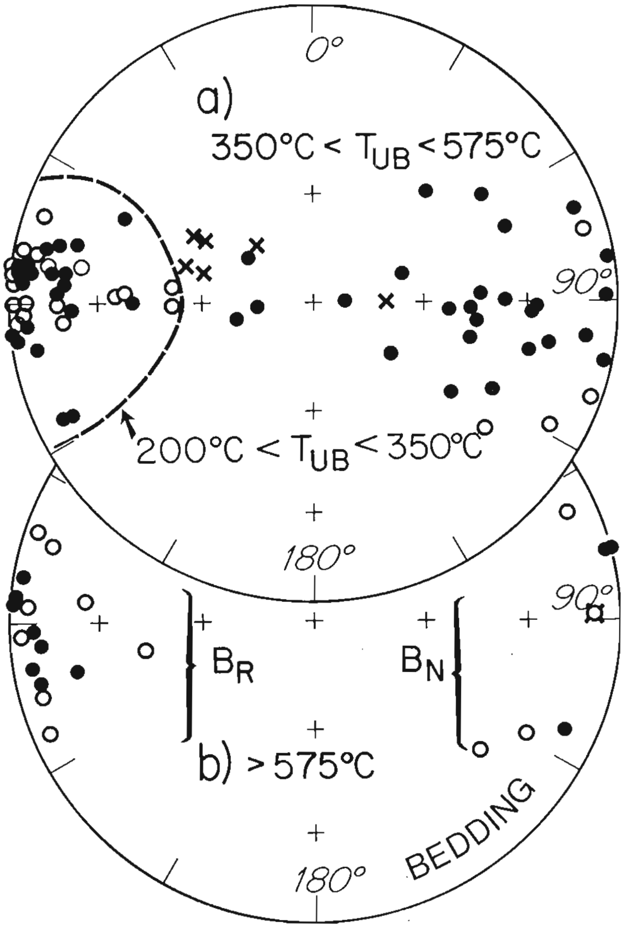

Figure 4.3. *Thermal directions from individual specimens previously treated by alternating fields. In (a), R_R directions (maghemite/magnetite ?) are to the west, R_N (magnetite) to the east. (b) represents the hematite directions. Directions have been resolved by using vector subtraction and Zijderveld diagrams. Crosses depict directions from site 1. Solid (open) symbols denote directions with positive (negative) inclinations in this and succeeding stereographic projections.*

DISCUSSION

Regional Geology

In the Mackenzie Mountains the number, position, and magnitude of major unconformities (Fig. 4.1) within the succession is very much open to question (Eisbacher, 1978a, 1978b; Yeo et al., 1978; Young et al., 1979). Two groups of unconformities have been discussed: a) unconformities at the base and/or within the Rapitan Group, and b) unconformities within the upper part of the Mackenzie Mountains Supergroup. An unconformity at the base of the Rapitan Group has been accepted without question. The intra-Sayunei Formation unconformity claimed by Helmstaedt et al. (1979) and Eisbacher (1978b), however, has been questioned by Yeo et al. (1978) who claimed that it is merely an apparent unconformity, produced either by a much later décollement event, or by very early large-scale slump folds.[1]

Verification of, and the significance of, the lower group of unconformities is crucial to the two correlation schemes proposed between the Proterozoic sequences of the Mackenzie and Wernecke mountains. Figure 4.1 (in part after Yeo et al., 1978) summarizes the schematic stratigraphy of the Proterozoic units in the eastern Wernecke Mountains and the Mackenzie Mountains. Subdivision of the lower portion of the Wernecke Mountains section is closely demarcated into three distinct sequences. In decreasing age these are the Wernecke Supergroup, the Pinguicula Group, and the Rapitan Group. The boundary between each sequence is marked by a pronounced regional unconformity (Eisbacher, 1978b).

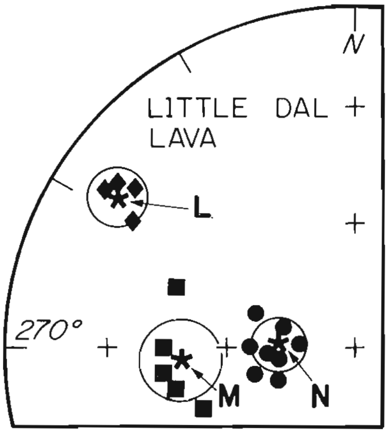

Figure 4.4. *Site directions (corrected for tilt) of the magnetic phases present in the unaltered (L) and altered portions (M and N) of the lava. Circles depict the α_{95} errors about the site means.*

[1]See also Aitken et al., Helmstaedt et al. (1981), Canadian Journal of Earth Sciences, v. 18, p. 410-418.

Assuming that the Redstone River and Coppercap formations unconformably overlie the Little Dal Group, Eisbacher (1978b) suggested the correlation:

Rapitan: Rapitan

Pinguicula: "Copper Cycle" (Redstone River + Coppercap)

Wernecke: Mackenzie Mountain (less Redstone River and Coppercap)

Yeo et al. (1978), in contrast, did not recognize the sub-Redstone River unconformity and therefore suggested that the correlation should be:

Rapitan: Rapitan

Pinguicula: Mackenzie Mountain

Wernecke: ?

Much clearly depends on the significance of the sub-Redstone River unconformity. Yeo et al. (1978) stated that "the contact between the Redstone River Formation and the underlying Little Dal Formation [Group] does not appear to be a major stratigraphic break". They admit, however, that the contact is locally abrupt but prefer not to designate this as an unconformity as has Eisbacher (1978b, Fig. 12.1). More recently Aitken (1980; 1981) has proposed that this unconformity is not located at the base of the Redstone River, but rather within the Little Dal Group at the base of the Little Dal lavas. The Redstone River, Coppercap, and Little Dal lava sequence is then identified as a separate "Copper Cycle" sequence. As we now have paleomagnetic data from the Little Dal sediments (Park, in press), the Little Dal lavas (herein), and the Tsezotene sills (herein), it is possible to examine the possible temporal extent of this latter unconformity.

Paleomagnetic Poles

Figures 4.5a and 4.5b compare the paleomagnetic pole positions obtained from the Mackenzie Mountains with two segments of the late Precambrian pole path proposed for Laurentia (Morris and Roy, 1977); a) covers the interval approximately 900-1050 Ma, and b) covers the period from approximately 625-800 Ma.

First we discuss the possible fit of these new poles to the pole segment for 900-1050 Ma (Fig. 4.5a). The two poles LD-A (15°S, 142°E) and LD-B (03°S, 138°E) from the Little Dal sediments ("basinal sequence") closely agree with poles from the Grenville Province dated at around 950 Ma, and if we adopt the preferred acquisition sequence A → B, we find that these poles plot in the inferred temporal order of acquisition on the pole path. From these pole positions and the stromatolitic evidence (Hofmann and Aitken, 1979) we infer an age for the deposition of the Little Dal of around 950 Ma. (These poles are sufficiently divergent from Belt Series poles as to preclude a possible older (c. 1100 Ma) age.) The new pole T from the Tsezotene sills is not significantly different from pole LD-B. It could be suggested, then, that the sills were also formed around 950 Ma. This, however, conflicts with the reported 766 and 769 Ma Rb-Sr isochron ages ($\lambda=1.42 \times 10^{-11} \alpha^{-1}$) on samples from two of these sills in the Mackenzie Mountains (R.L. Armstrong, G.H. Eisbacher and P.D. Evans, in preparation). We have two options. Either, this radiometric age is not truly representative of the sills sampled in this paleomagnetic study, or pole T is being compared to the wrong pole path segment in Figure 4.5a.

Poles LD-L and LD-M from the lavas also appear to be at variance with this segment of pole path. Assuming the Little Dal lava conformably overlies Little Dal sediments, and assuming that the correct original magnetization has been identified in both cases, then it should follow that the pole positions from these two units should be somewhat similar. As shown in Figure 4.5a poles LD-A and LD-L are quite dissimilar, and moreover, pole LD-L does not plot anywhere near the generally accepted pole path for the period 900-1050 Ma. One of our assumptions therefore must be false. Without conclusive tests of remanence age it is not possible to examine the second assumption (i.e. LD-L is the initial magnetization), but it is possible to test the first (accepting the second) by comparing these new poles to other segments of pole path.

For the period 900 Ma to Middle Cambrian there are few poles available from Laurentia, and any pole path for this interval is somewhat speculative. Figure 4.5b presents a pole path which may be considered speculatively representative of the interval 800 to 650 Ma. The younger portion of the Hadrynian Track proposed by Morris and Roy (1977) is tied by poles from the Upper Torridonian sediments of northern Scotland. In Figure 4.5b these 777 Ma (Rb-Sr isochron, $\lambda=1.42 \times 10^{-11} \alpha^{-1}$) sediments are indicated by poles 6-12 (Piper and Smith, in press). Of particular interest is the sharp bend in the pole path suggested by the polar movement 11 → 12. Further detailing of this possible change in pole path trend is quite important. The second portion of the path is based on only three results from the Tudor gabbro (pole 458) and the Franklin diabase (poles 450 and FR). If we accept that the magnetization age in both these studies is directly equivalent to their reported K-Ar isochron ages then the suggested pole sequence would be from west to east with an approximate age of around 650 to 670 Ma. Our new poles from the Mackenzie Mountains compare more favourably with these younger segments of pole path. First, comparing pole T from the sills to the pole segment of Figure 4.5b we find that it plots at the younger end of the Torridonian segment (as defined above). The Rb-Sr age for these sediments is around 770 Ma, not greatly different from the 766 and 769 Ma Rb-Sr isochron ages reported from the intrusions. It is also possible to plot the Little Dal sediment poles (LD-A and LD-B) on this younger segment of pole path (Fig. 4.5a). Therefore from the paleomagnetic evidence alone, we cannot differentiate between the two age estimates for the sediments of c. 770 Ma and c. 950 Ma, although a 950 Ma age is preferred.

Poles LD-L and LD-M are supportive of the younger segment of the pole path. The two poles plot in the correct location and also conform to the pole sequence suggested by petrology and the demagnetization data. This path also suggests that pole LD-M was acquired relatively soon after formation of the lava, again closely agreeing with the geological evidence. Further, poles T, LD-L and LD-M can now be joined in a simple single path that does not conflict with any other known paleomagnetic data.

The format of the pole path between the segments shown in Figure 4.5b and between the younger segment and Paleozoic pole paths is presently unknown. The present format suggests the lava has an age of around 680 Ma. This is quite different from either of the two ages proposed for the underlying Little Dal sediments, or Tsezotene sills. Should future data confirm this analysis then there must be a major unconformity between the Little Dal lavas, and the underlying "basinal sequence" of Little Dal sediments (cf. Aitken, 1980) with an age around 700-750 Ma. Alternatively, it may later be shown that LD-L does not represent the initial magnetization in the lava (i.e. our second assumption was wrong). If then the correct pole sequence should be LD-A → LD-B → LD-M there would be no supporting evidence for any major discontinuity within the Little Dal Group.

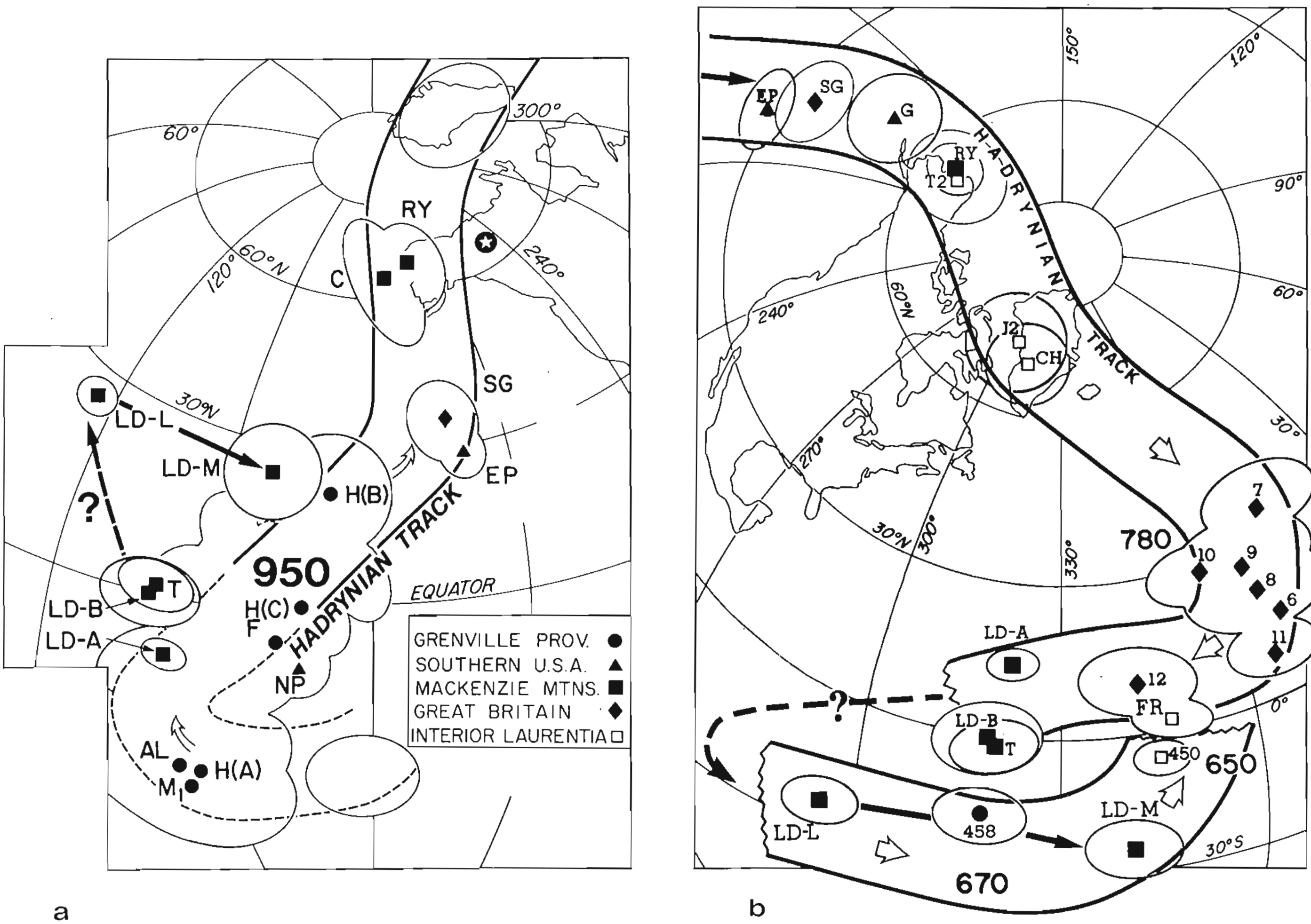

Figure 4.5. The Hadrynian polar track relative to North America after Morris and Roy (1977) as defined by poles in the periods from about 1050 to 900 Ma (a) and 950 to 650 Ma (b). Only a few of the poles which define this track are shown, the rest are depicted by error circles. For a fuller discussion of this path interested readers are referred to the afore-mentioned reference. In this paper our discussion is limited to the following pertinent poles:

LD-A and LD-B: Little Dal Group "basinal sequence" A and B components (Park, 1980);
LD-L and LD-M: Little Dal lava L and M components (this paper);
T: sills in Tsezotene Formation (this paper);
FR and 450: Franklin diabases (averages calculated by Palmer and Hayatsu, 1975);
poles 6-12: Upper Torridonian sediments (Piper and Smith, in press);
458: Tudor gabbro (Palmer and Carmichael, 1973).

Other poles shown are:

AL: Allard Lake anorthosite (Hargraves and Burt, 1967);
C: Little Dal Group sediments C pole (Park, 1980);
CH: Chequamegon Sandstone (Dubois, 1962);
EP: rocks of El Paso, Texas (Spall, 1971);
F: Frontenac Axis dykes (Park and Irving, 1972);
G: Grand Canyon Suite (Elston and Gromme, 1974);
H(A), H(B), H(C): Haliburton intrusions (Buchan and Dunlop, 1976);
J_2: Jacobsville Formation (Roy and Robertson, 1978);
M_1: Morin Complex (Irving et al., 1974);
NP: Nankoweap Formation (Elston and Grommé, 1974);
RY: Rapitan Group Y pole (Morris, 1977);
SG: Stoer Group (Stewart and Irving, 1974);
T_2: Orienta 2 (Dubois, 1962).

The approximate error ovals of 95% confidence are plotted about the mean site (or sample) poles. The star represents the present sampling area.

CONCLUSIONS

The data presented above suggest that there is a significant unconformity within the Little Dal Group as it is presently defined. Perhaps, as Aitken (1980) has proposed the upper part of the Little Dal should be included in the "Copper Cycle". A best estimate for the age of this unconformity appears to be around 700 to 750 Ma. In itself this information does not pass judgment on either of the two correlations proposed between the Wernecke and Mackenzie mountains (see above). It merely confirms the existence of one unconformity within the Mackenzie Mountains section, and provides an age estimate of this event. In order to test either correlation it would be necessary to establish and compare the complete paleomagnetic signatures for each unconformity in both the Wernecke and Mackenzie mountains.

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5. STRATIGRAPHY OF THE AKAITCHO GROUP AND THE DEVELOPMENT OF AN EARLY PROTEROZOIC CONTINENTAL MARGIN, WOPMAY OROGEN, NORTHWEST TERRITORIES

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Easton, R.M., Stratigraphy of the Akaitcho Group and the development of an early Proterozoic continental margin, Wopmay Orogen, Northwest Territories; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 79-95, 1981.

Abstract

The early Proterozoic Akaitcho Group consists of 8 to 10 km of metasedimentary and metavolcanic rocks located in the central metamorphic zone of the Wopmay Orogen (Bear Province). In the northern part of the orogen (north of 66° N), the following generalized stratigraphic sequence has been recognized: (1) a lower basaltic unit of unknown thickness (Ipiutak formation), (2) 3-4 km of arkosic turbidites (Zephyr formation) intruded by sills of rhyolite porphyry (Okrark sills), (3) basalt and rhyolite volcanic complexes (Nasittok subgroup), and (4) 1 to 3 km of pelite and tuffaceous sedimentary rocks (Aglerok formation), locally with abundant basaltic and gabbroic rocks (Tallerk formation). The upper pelite is conformably overlain by a different pelite belonging to the Odjick Formation (lower Epworth Group).

Two types of REE patterns are present in the Akaitcho Group basalts: a sloping pattern (Ce/Yb ratio 10 to 15) similar to patterns for recent continental tholeiite suites such as the Columbia River basalts; and a flat REE pattern (Ce/Yb ratio 2 to 5) similar to ocean tholeiite (Group II ocean basalts and marginal basin basalt). The Akaitcho Group records a temporal evolution in basalt geochemistry from older evolved continental tholeiites, through continental tholeiites, to younger, ocean tholeiites. An evolution in basalt geochemistry from older, transitional tholeiites to ocean tholeiites has been documented from recent rift areas, such as the Afar. Akaitcho Group rhyolites have total REE abundances lower than the most evolved Akaitcho Group basalts, and Ce/Yb ratios (20 to 25) different from the basalts, and are probably crustal-derived.

Bimodal volcanism (subalkaline basalt-rhyolite) in association with continent-derived sediments, the temporal evolution of the basalt geochemistry, and the similarity of the Akaitcho Group to recent rift sequences indicates that the Akaitcho Group was deposited in a rift. The stratigraphic position of the Akaitcho Group beneath the lower Epworth Group preserves products of the initial rifting in the Wopmay Orogen, and that ocean crust probably did exist west of the Slave Craton in the early Proterozoic.

Résumé

Le groupe d'Akaitcho, d'âge protérozoïque inférieur, consiste en 8 à 10 km de roches métavolcaniques à métasédimentaires, situées dans la zone métamorphique centrale de l'orogène de Wopmay (province du Grand lac de l'Ours). Dans la partie nord de l'orogène (au nord de 66° N), on a identifié la succession stratigraphique générale suivante: (1) une unité basaltique inférieure de puissance inconnue (formation d'Ipiutak), (2) 3 à 4 km de turbidites arkosiques (formation de Zéphyr) traversées par des sills de porphyre rhyolitique (sills d'Okrark), (3) des complexes volcaniques basaltiques et rhyolitiques (sous-groupe de Nasittok) et (4) 1 à 3 km de pélites et roches sédimentaires tufacées (formation d'Aglerok) localement accompagnées d'abondantes roches basaltiques et gabbroïques (formation de Tallerk). Les pélites supérieures sont recouvertes en conformité par des pélites différentes appartenant à la formation d'Odjick (groupe inférieur d'Epworth).

On rencontre deux types de tendances REE dans les basaltes du groupe d'Akaitcho: une courbe descendante (rapport Ce/Yb de 10 à 15), semblable à celle des suites tholéiitiques continentales récentes telles que les basaltes du fleuve Colombia; et une courbe REE en palier (rapport Ce/Yb de 2 à 5) comme pour les tholéiites océaniques (basaltes océaniques du groupe II et basaltes de bassin marginal). Le groupe d'Akaitcho montre une évolution dans le temps de la géochimie des basaltes allant de tholéiites continentales anciennes modifiées à des tholéiites continentales, et enfin à des tholéiites océaniques plus récentes. Grâce à l'étude de récentes zones de fracture (rifts) comme les Afars, on a pu documenter l'évolution de la géochimie des basaltes depuis les tholéiites anciennes et celles de transition aux tholéiites océaniques. Les rhyolites du groupe d'Akaitcho présentent des quantités totales de REE inférieures à celles des basaltes les plus évolués du groupe d'Akaitcho, et les rapports Ce/Yb (20 à 25) diffèrent de ceux des basaltes; ces roches sont probablement issues de la croûte.

Le volcanisme bimodal (basaltes subalcalins-rhyolites) associé aux sédiments dérivés des continents, l'évolution dans le temps de la géochimie des basaltes, et la ressemblance entre le groupe d'Akaitcho et certaines successions rencontrées dans les rifts indiquent que le groupe d'Akaitcho s'est déposé dans un tel rift. La situation stratigraphique du groupe d'Akaitcho au-dessous de la partie inférieure du groupe d'Epworth, qui est une succession de marge continentale passive et la présence de tholéiites océaniques dans la partie supérieure du groupe d'Akaitcho, indiquent que le groupe en question contient encore des produits de la période initiale de fracturation qui a eu lieu dans l'orogène de Wopmay, et que la croûte océanique était probablement présente à l'ouest du craton du Grand lac des Esclaves pendant le Protérozoïque inférieur.

INTRODUCTION

Present understanding of the early development of rifted continental margins is poor because rift structures of modern continental margins are commonly covered by thick sedimentary sequences. This difficulty can be overcome by studying sediment starved continental margins, such as North Biscay and Galicia (Montadert et al., 1979), or by studying very young rifts, such as the Afar (Mohr, 1978) or the Gulf of California (Moore, 1973; Einsele et al., 1980) where the continental margins are not yet fully developed. Alternatively, one can examine the products of rifting preserved in deformed continental margins. In these areas, the opportunity to observe the products of rifting in cross-section compensates for the metamorphism and deformation these rocks are likely to have undergone.

The early Proterozoic Wopmay Orogen (McGlynn, 1970; Hoffman, 1980) located in the Bear Structural Province (Fig. 5.1) contains the Akaitcho Group: a rock sequence consisting of bimodal (basalt-rhyolite) volcanics and subarkosic to arkosic sediments believed to have been deposited in a rift (Easton, 1980; Hoffman et al., 1978). The Akaitcho Group is associated with a younger passive continental margin sedimentary sequence (Hoffman, 1973, 1980), providing an opportunity to study a rift sequence which preceded the development of such a continental margin.

This paper is aimed at a review of the present knowledge of the Akaitcho Group (first defined in 1978) in terms of its significance with respect to the evolution of the Wopmay Orogen, and outlines further research needed on the Akaitcho Group, particularly in the southern part of the Wopmay Orogen.

The Akaitcho Group has been well studied only in the northern half of the Wopmay Orogen above 66°N (Hoffman et al., 1978; Easton, 1980), and is discussed in detail here whereas the remainder of the Akaitcho Group is discussed in more general terms.

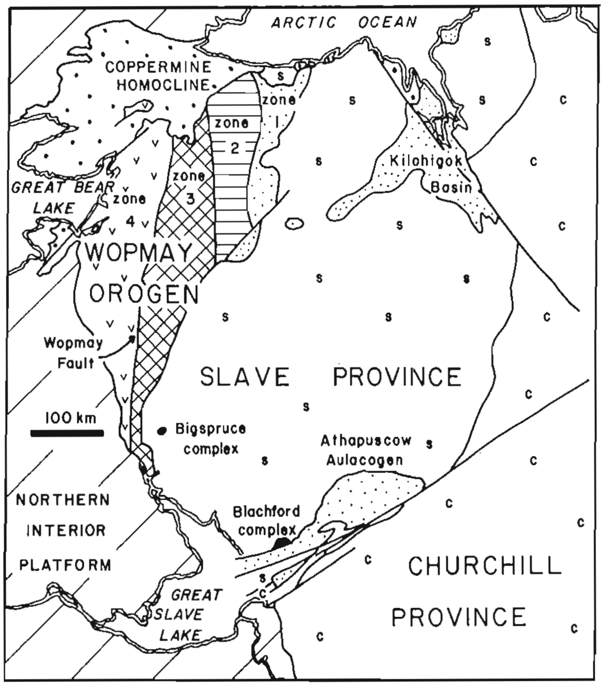


Figure 5.1. Tectonic subdivisions of the northwestern Canadian Shield (after Hoffman, 1980).

REGIONAL SETTING

The Wopmay Orogen is a north-south trending mobile belt which developed along the western margin of the Slave Structural Province in early Proterozoic time (Fig. 5.1). The peak of orogenic activity, as determined by K-Ar radiometric dates, was about 1800 Ma (Stockwell, 1970). The orogen can be divided into 4 tectonic zones (Fig. 5.1; Hoffman, 1980). Zone 1 comprises thin, autochthonous platformal sedimentary rocks overlain by distal flysch and molasse. The Kirkerk Thrust Fault separates Zone 1 from the more westerly Zone 2, which contains thicker, platformal (miogeosynclinal) rocks overlain by northerly-derived flysch, both of which have been thrust towards the craton. Cloos anticline separates Zone 3 from Zone 2, and also marks the position of the facies change between platformal rocks (outer continental shelf) and deeper water equivalents (continental rise and slope deposits). Zone 3 is a metamorphic and batholithic terrane, divided into 3 subzones by the Marceau and Okrark thrust faults (Fig. 5.2). The eastern subzone contains continental slope and rise deposits of the Epworth Group. The central subzone contains intrusive rocks of the Hepburn Batholith, volcanic and sedimentary rocks of the Akaitcho Group, and continental rise pelites of the Epworth Group which conformably overlie the Akaitcho Group (Easton, 1980). The western subzone comprises sedimentary and volcanic rocks of the Akaitcho Group, and intrusive rocks of the Wentzel Batholith. Metamorphic grade in Zone 3 ranges from chlorite (lower greenschist) to above muscovite breakdown (upper amphibolite to granulite facies), and is spatially related to the Hepburn and Wentzel batholiths (St-Onge and Carmichael, 1979; Hoffman et al., 1980).

The Wopmay Fault separates Zone 3 from Zone 4. Zone 4 contains weakly metamorphosed calc-alkaline volcanic and sedimentary rocks and epizonal plutons of the Great Bear volcano-plutonic belt (Hoffman and McGlynn, 1977; Hildebrand, 1981). Conglomerates in the easternmost Zone 4 contain detritus derived from the Wentzel Batholith, and locally, sedimentary and volcanic rocks of Zone 4 lie unconformably on metamorphosed Akaitcho Group rocks and the Wentzel Batholith (Hoffman and McGlynn, 1977), indicating that at least by the later stages of volcanism in eastern Zone 4, Zone 3 rocks had been deformed, metamorphosed, uplifted, and exposed.

STRATIGRAPHY

The distribution of the three main groups of supracrustal rocks present in Zones 1, 2 and 3 of the Wopmay Orogen – the Epworth Group, the Akaitcho Group and the Snare Group – is shown in Figure 5.2. The Akaitcho Group is preserved at upper greenschist to lower amphibolite metamorphic grade in 5 major areas in the Wopmay Orogen: the Hepburn Lake area, the Calder River belt, the Grant Lake area, the Rebesca Lake area, and the Redrock Lake area (Fig. 5.2). The Hepburn Lake area is the best studied (Easton, 1980), and the bulk of the data presented in this paper concerns this area. The Calder River belt (McGlynn, 1974, 1975, 1976; Lord and Parsons, 1952) and the Grant Lake area (McGlynn, 1964) are a continuation of the Hepburn Lake area (Easton, 1981). The Rebesca Lake (Lord, 1942) and the Redrock Lake area (Fraser et al., 1960; Mursky et al., 1970) are poorly known.

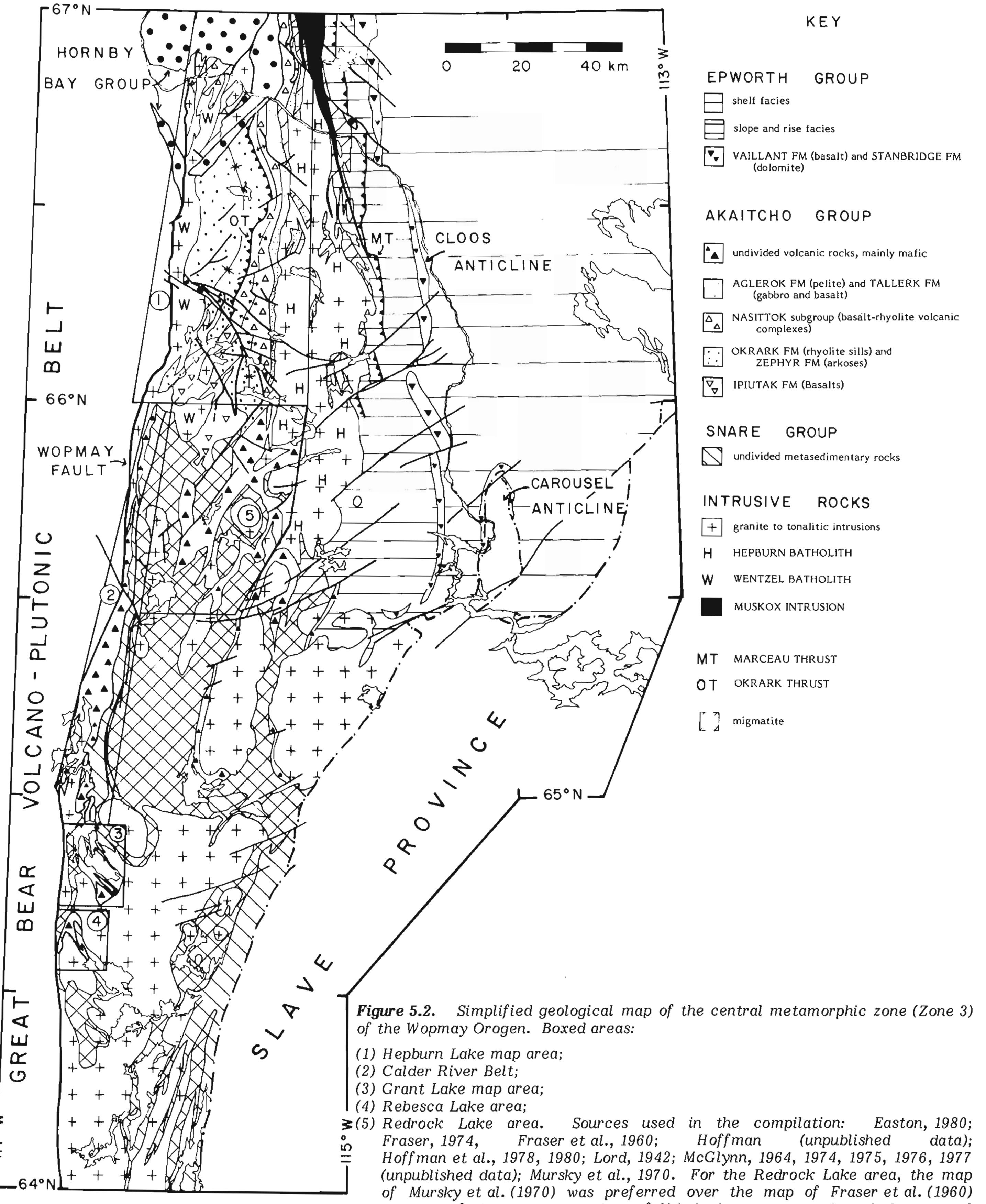


Figure 5.2. *Simplified geological map of the central metamorphic zone (Zone 3) of the Wopmay Orogen. Boxed areas:*

(1) Hepburn Lake map area;
(2) Calder River Belt;
(3) Grant Lake map area;
(4) Rebesca Lake area;
(5) Redrock Lake area. Sources used in the compilation: Easton, 1980; Fraser, 1974, Fraser et al., 1960; Hoffman (unpublished data); Hoffman et al., 1978, 1980; Lord, 1942; McGlynn, 1964, 1974, 1975, 1976, 1977 (unpublished data); Mursky et al., 1970. For the Redrock Lake area, the map of Mursky et al. (1970) was preferred over the map of Fraser et al. (1960) because of the greater number of lithologies represented, and the general reliability of the map of Mursky et al. (1970) in the Hepburn Lake area.

AKAITCHO GROUP

The Akaitcho Group consists of 8 to 10 km of metasedimentary and metavolcanic rocks. The term Akaitcho Group was introduced by Hoffman et al. (1978) for metamorphic rocks east of the Wopmay Fault and west of the Hepburn Batholith which could not be correlated with Epworth Group strata. Easton (1980) subdivided the Akaitcho Group into 5 unnamed formations upon completion of 1:50 000 scale mapping of the Akaitcho Group in the Hepburn Lake map area (86 J). These formations, in addition to a new formation and a subgroup, are described below, and informal names are applied to the formations. A generalized stratigraphic column for the Akaitcho Group, which also shows the facies relations between the main Akaitcho Group lithologies, is shown in Figure 5.3.

Ipiutak formation amphibolites

The oldest exposed rocks of the Akaitcho Group constitute the Ipiutak formation which consists of a minimum of 500 m of basalt flows and tuffs now preserved as migmatized amphibolite. Rhyolite has not been found in association with the amphibolites, and this could be a distinguishing characteristic of the formation. Pelitic rocks, now migmatite, interfinger with the upper amphibolites, and are considered to be part of the Ipiutak formation. The pelitic rocks reach a maximum thickness of 300 to 400 m. The base of the Ipiutak formation has not been observed. The type area is 5 km south of Wentzel Lake (Fig. 5.4).

Zephyr formation

Three to four kilometres of subarkosic to arkosic turbidite overlie the Ipiutak formation. Turbidite beds typically range from 10 to 100 cm thick, and typically have arkosic, medium sand bases with green or grey pelite tops (Fig. 5.5). Current structures are rarely observed in the turbidites. Rounded grains of epidote, tourmaline and zircon are common in the basal turbidite beds. Upper and lower contacts of the formation are sharp. Typical outcrops of the Zephyr formation can be observed between Zephyr Lake and Akaitcho Lake, and on the northwest shore of Akaitcho Lake (Fig. 5.4).

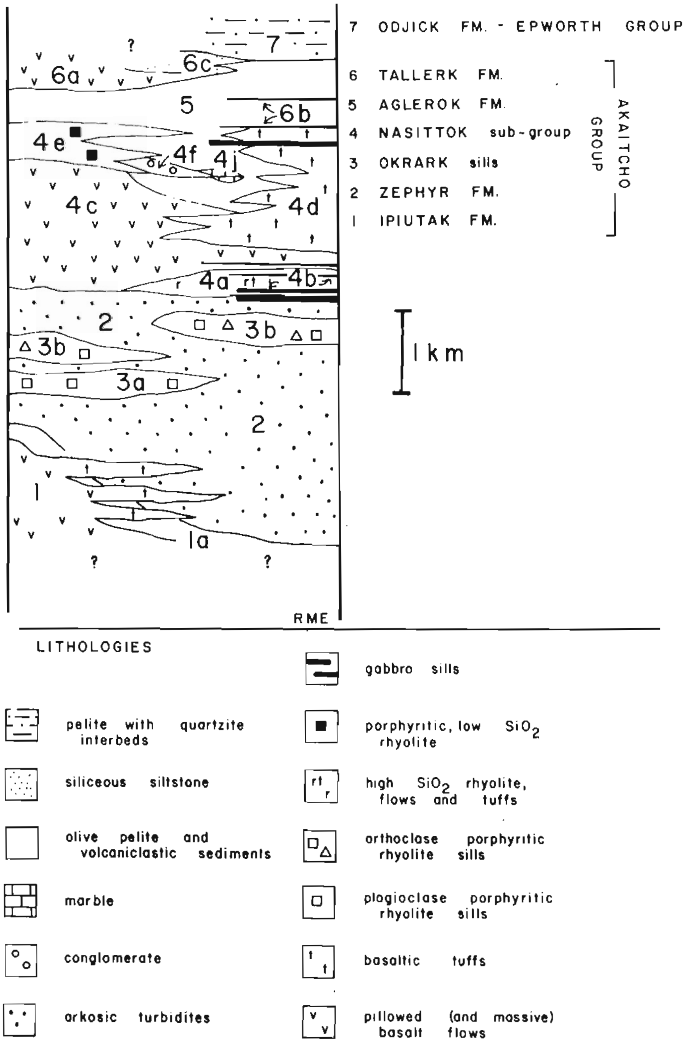
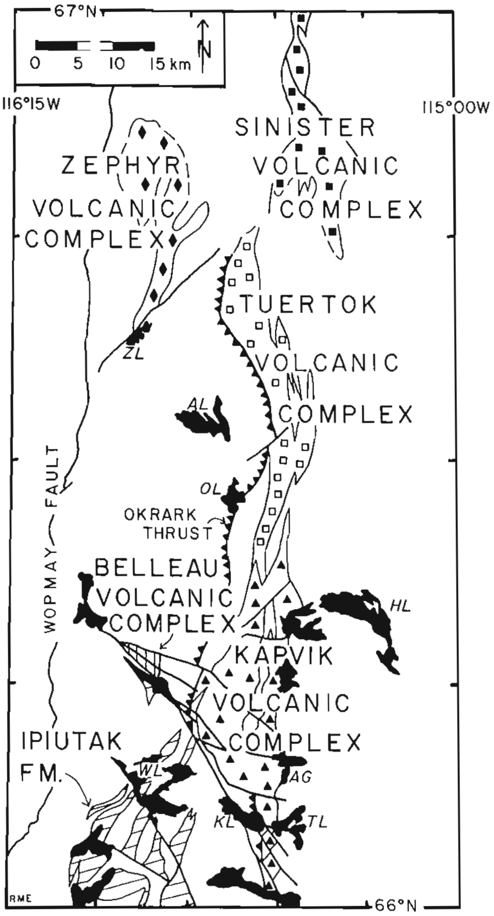


Figure 5.3. *Generalized stratigraphic column showing main lithologies and facies relations between formations of the Akaitcho Group.*



AL	-	*Akaitcho Lake*	*OL*	-	*Okrark Lake*
AG	-	*Aglerok Lake*	*TL*	-	*Tallerk Lake*
HL	-	*Hepburn Lake*	*WL*	-	*Wentzel Lake*
KL	-	*Kapvik Lake*	*ZL*	-	*Zephyr Lake*

Figure 5.4. *Distribution of Akaitcho Group volcanic complexes in the northern Wopmay Orogen (Hepburn Lake map area, west half).*

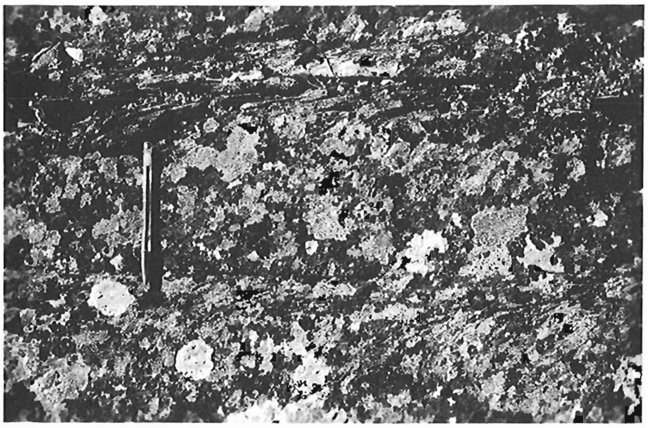


Figure 5.5. *Arkosic turbidite of the Zephyr formation. The central bed shows refraction of cleavage in the upper pelitic zone of the turbidite. Hammer is 30 cm long. Lichen cover is typical of most Akaitcho Group outcrops. GSC 203061-X*

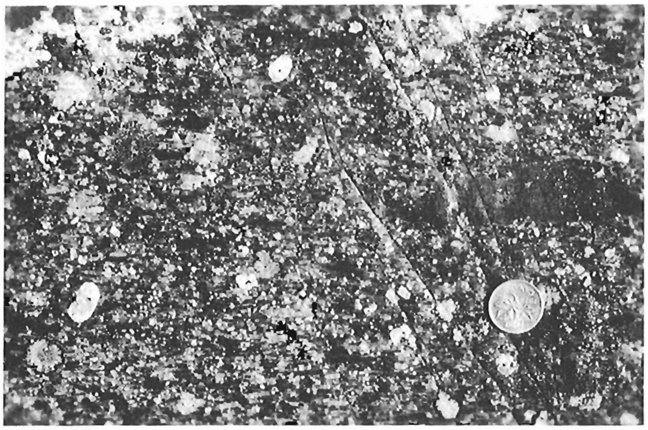


Figure 5.6. *Orthoclase porphyritic sill of the Okrark sills. GSC 203061-Y*

Okrark sills

Plagioclase porphyritic and orthoclase porphyritic, 300 to 600 m thick, rhyolite porphyry sills intrusive into the Zephyr formation are termed the Okrark sills (Fig. 5.6). The plagioclase porphyritic and orthoclase porphyritic sills are indistinguishable chemically. Contacts between the sills and the turbidites are generally sharp, and the sills have contact metamorphosed the adjacent sedimentary rocks. The Okrark sills are correlative with low SiO_2, high Zr porphyritic rhyolite found in the upper parts of the Tuertok, Sinister and Belleau volcanic complexes of the Nasittok subgroup (Easton, 1980). Both types of sills are exposed on Okrark Lake (Fig. 5.4).

Nasittok subgroup

The subgroup consists of five volcanic complexes, the Sinister, Tuertok, Kapvik, Belleau and Zephyr complexes (Easton, 1980; Fig. 5.4). The volcanic complexes typically have a 2 km thick basal basalt unit containing both pillowed flows (Fig. 5.8) and tuffs (Fig. 5.9) overlain by porphyritic, low SiO_2, high Zr rhyolite domes and flows. The only exception to this typical stratigraphy is the Kapvik Volcanic Complex (Fig. 5.7), where high SiO_2, low Zr rhyolite is present below the basal basalt sequence. The stratigraphy of the volcanic complexes has been outlined in Easton (1980) and will not be repeated here. Figure 5.7 is, however, a simplified diagram showing the complex stratigraphic relationship between the Kapvik and Tuertok volcanic complexes. In addition to the volcanic units, conglomerates, which commonly contain granite and granite gneiss cobbles (Fig. 5.10), breccias (Fig. 5.11), marbles and quartzites, are closely associated with the volcanic complexes, and are considered to be part of the Nasittok subgroup. The Tuertok Volcanic Complex (Fig. 5.5) is regarded as typical. The contact of the Tuertok Volcanic Complex with the Zephyr formation is sharp, in the few places where it has been observed. The contact between the Kapvik Volcanic Complex and the Zephyr formation is complex, because a gabbro sill swarm, intrusive into the Zephyr formation, underlies the basal rhyolites and basalts of the Kapvik Complex (Fig. 5.7). The bases of the other complexes are not exposed. The volcanic complexes are overlain, and interfinger with, the pelites of the overlying Aglerok formation. Volcanism in the Nasittok subgroup was mainly subaqueous, with pillowed basalt and basalt tuff being predominant. The rhyolites were deposited as water lain tuffs, flows, and domes. A few possible welded ash-flow tuffs are present in the upper parts of the volcanic complexes (Easton, 1980), indicating that locally, volcanic complexes were subaerial.

Aglerok formation

Olive pelites, 3-4 km thick overlie and interfinger with the Nasittok subgroup. Bedding in the pelites is on a millimetre to a centimetre scale. Included with the pelites are volcaniclastic sedimentary rocks and minor basaltic and rhyolitic tuffs. The Aglerok formation was derived primarily from weathering of the Nasittok subgroup volcanic complexes, with a minor contribution from an adjacent continental source area (Easton, 1980). Contacts with the sills and volcanics of the overlying Tallerk formation are sharp. Typical exposures of Aglerok formation lithologies can be observed west of Aglerok Lake.

Tallerk formation

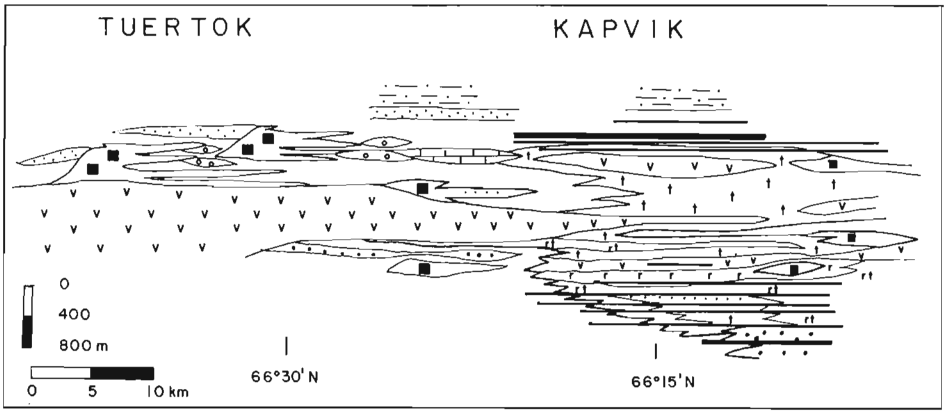
On the north shore of the Coppermine River, pillow basalts overlie pelites of the Aglerok formation. From Hepburn Lake south to 66°N (Fig. 5.4), 50 to 200 m thick gabbro sills intrude the Aglerok formation. These mafic rocks are considered to be part of the Tallerk formation. A titaniferous siliceous siltstone overlying the Aglerok formation and underlying the Odjick Formation 8 km northwest of Hepburn Lake (Easton, 1980) is included with the Tallerk formation. The upper part of the Tallerk formation is intruded by tonalitic, granitic and gabbroic intrusions of the Hepburn Batholith, and hence the thickness of the formation is not well known. In the two areas where Odjick Formation pelites were observed to overlie the Akaitcho Group (Easton, 1980), the Tallerk formation is 300 to 400 m thick.

GEOCHEMISTRY

On the basis of major and trace element geochemistry, in conjunction with field observations, Easton (1980) recognized two types of rhyolite in the Akaitcho Group: (1) volumetrically minor, generally aphanatic, high SiO_2 (73 to 76%), low Zr (100 to 200 ppm) rhyolite; and

Figure 5.7

Stratigraphic relations between the Tuertok and Kapvik volcanic complexes. Lithologies as in Figure 5.3.



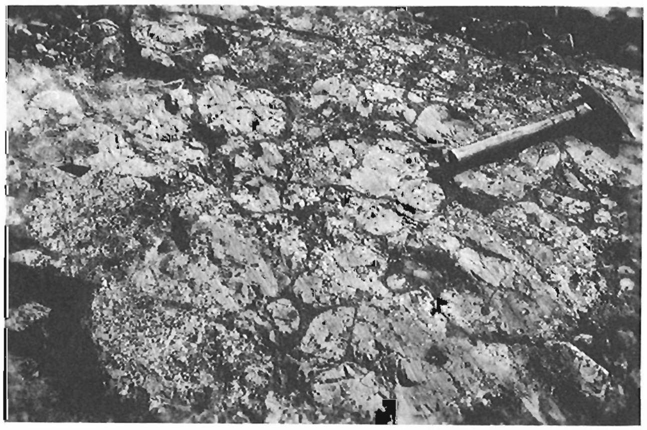


Figure 5.8. *Pillow lavas, upper Belleau Volcanic Complex, Nasittok subgroup. GSC 203061-W*

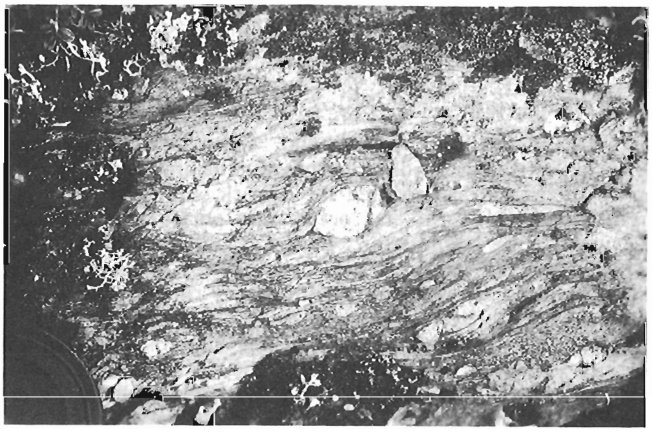


Figure 5.10. *Granitoid gneiss pebbles in conglomerate in the upper part of the Tuertok Volcanic Complex. GSC 203062-E*

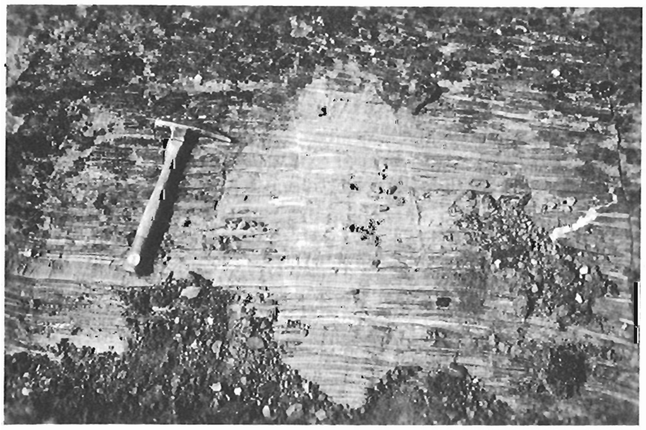


Figure 5.9. *Finely bedded basaltic tuffs of the Kapvik Volcanic Complex. GSC 203061-U*

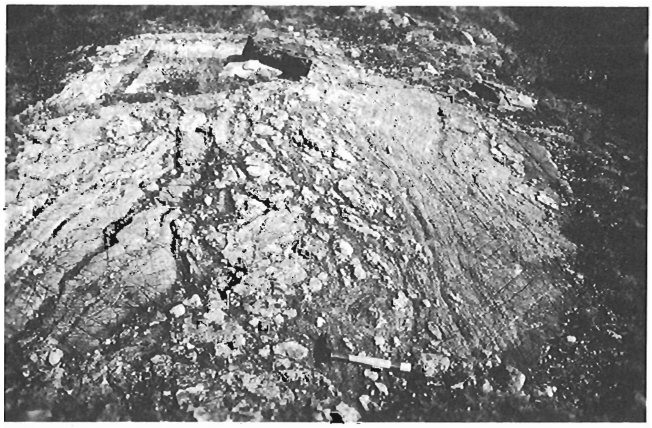


Figure 5.11. *Carbonate-cemented rhyolite volcanic breccia present in the upper part of the Kapvik Volcanic Complex. GSC 203061-Z*

(2) abundant, generally orthoclase porphyritic low SiO_2 (68 to 71%), high Zr (250 to 350 ppm) rhyolite (Table 5.1). Both types of rhyolite may have been derived from melting of lower crustal rocks. Two types of tholeiitic basalt are also present: Type 1 basalts, which are stratigraphically highest, have trace element contents similar to marginal basin oceanic basalts, and Type 2 basalt is similar to continental tholeiites (Easton, 1980; Table 5.1). Newly obtained REE data presented in Figures 5.12 and 5.13 support these subdivisions. In addition, the REE data have a bearing on two important questions, namely: (1) the temporal evolution of the Akaitcho Group basalts, and (2) the origin of the rhyolites.

BASALTS

Two types of REE patterns are observed in the Akaitcho Group basalts (Fig. 5.12). The sloping pattern is characteristic of continental tholeiites (e.g. BCR-1, Columbia River basalt, Fig. 5.12). The flat pattern is characteristic of marginal basin and some ocean floor basalts (Fig. 5.12). The distribution of the REE pattern in basalts of the Belleau Volcanic Complex of the Nasittok subgroup (Fig. 5.12) is particularly significant. The Belleau Volcanic Complex (Fig. 5.12) is one of the most westerly of the Akaitcho Group volcanic complexes (Fig. 5.4), and is preserved at lower greenschist grade (chlorite grade in pelitic rocks). The Belleau Complex consists of a 1 to 1.5 km thick lower unit of predominantly massive basalt with minor pillowed basalt, both cut by gabbro sills. The lower basalt sequence is capped by a 300 m thick compositionally zoned (rhyolitic base, dacite top) lithic lapilli tuff, possibly an ash-flow tuff, overlain by 100 to 200 m of volcaniclastic sedimentary rocks. The sedimentary rocks are overlain by 1 to 1.5 km of pillowed basalt. Several 300 to 500 m thick gabbro sills are present near the base of this upper basalt pile. The lower basalts are all continental tholeiites (Fig. 5.12). The upper basalts have flat REE patterns characteristic of Group II oceanic basalts (Frey et al., 1977; Fig. 5.12). Thus, the basalts in the Belleau Complex evolved through time from early, continental basalts to later, oceanic basalts. A similar trend is observed in the Kapvik and Tuertok volcanic complexes.

The temporal evolution of the basalts of the Belleau Volcanic Complex also occurs through the whole Akaitcho Group (Fig. 5.12; Table 5.1). The stratigraphically lowest basalts (Ipiutak formation) are continental tholeiites, and are somewhat transitional towards alkaline basalts. The Kapvik, Tuertok and lower Belleau basalts are less evolved continental tholeiites, and have lower REE abundances than the Ipiutak basalts, but they still have sloping REE patterns characteristic of continental tholeiites. The stratigraphically highest basalts of the Belleau, Tuertok, and the Kapvik volcanic complexes have flat REE patterns. The major and trace element contents of the basalts also show this temporal evolution (Table 5.1), but less dramatically.

RHYOLITES

Average REE patterns of Akaitcho Group rhyolites are shown in Figure 5.13. The REE patterns for the high and low Si rhyolites are similar, the high Si rhyolites having slightly greater LREE abundances, and slightly lower HREE contents than the low Si rhyolites. The lower HREE contents of the high Si rhyolites is consistent with their observed low Zr content.

The absolute abundances of REE in both types of rhyolites are lower than the REE abundances observed in the most fractionated Akaitcho Group basalts (e.g. Ipiutak formation, Fig. 5.13), and the Ce/Yb ratios of the rhyolites (20 to 25) are different than the Ce/Yb ratios for the basalts (flat pattern, 5 to 7; sloping pattern 10 to 15). Neither of these effects would be expected if the rhyolites were related to the basalts by differentiation. In addition, the high Fe, Ti, Mn, K, Ba and radiogenic Sr contents (Easton, 1980) in the rhyolite, plus the abundance of rhyolite in the Nasittok subgroup (felsic/mafic ratio .2 to .3), indicate that the rhyolites are most likely derived by crustal melting. The REE contents of the rhyolites are similar to estimates of the REE content of the continental crust (Fig. 5.13) as indicated by direct sampling (Shaw et al., 1976; Taylor, 1964), or in continent-derived pelitic rocks (Nance and Taylor, 1976, 1977). The slightly higher HREE patterns of the rhyolites could reflect the more mafic character of the lower crust (c.f. Eichelberger, 1978). The difference between the high and low Si rhyolites could be related to slight differences in the degree of melting. The lower Fe, Mn, and Ti contents (Table 5.1) in the high Si rhyolites, plus the low Zr, and LREE contents, and the small quantities and lower stratigraphic position of the high Si rhyolites (base of the Kapvik Volcanic Complex, Nasittok subgroup) is consistent with the high Si rhyolites representing early melts. Greater degrees of partial melting would lower the SiO_2 content, and increase the Fe, Mn, Ti and Zr contents of the low Si rhyolites. The larger volumes and higher stratigraphic position of the low Si rhyolites is consistent with this interpretation.

DEPOSITIONAL ENVIRONMENT

Bimodal volcanism with a high felsic/mafic ratio in the Nasittok subgroup, in association with subarkosic to arkosic sedimentary rocks suggests that the Akaitcho basin probably was a rift valley. The linear distribution of the volcanic rocks may also indicate volcanism in a rift valley, although it could reflect later tectonism. The temporal evolution shown by the REE patterns of the Akaitcho Group basalt from older, continental tholeiites to younger oceanic tholeiite (Fig. 5.12) is similar to that observed in recent rift terranes such as the Afar (Barberi and Varet, 1978).

The geochemical evolution of the basalt geochemistry from continental to oceanic basalts, and subsequent burial of the Akaitcho Group by deepwater pelites of the Odjick Formation suggest that the size of the Akaitcho basin (rift) increased with time.

REGIONAL CORRELATION AND AGE

Correlation of the main stratigraphic units in the northern Wopmay Orogen is shown in Figure 5.14. Data for the Epworth Group are from Hoffman (1973) and Hoffman et al. (1978). The Akaitcho Group is known to conformably underlie pelites of the Odjick Formation (Easton, 1980). This suggests that the Vaillant Formation basalts and the Stanbridge Formation dolomites, which are also overlain by the Odjick Formation, are correlative with the Akaitcho Group. This seems reasonable on lithological grounds also, because both the marbles present in the Nasittok subgroup and the Stanbridge dolomite contain abundant quartz grit, and are associated with basaltic rocks. The eastward thinning of the Vaillant basalt, and the change from pillow basalts in the west to tuffs interbedded with mudstone in the east (Hoffman, 1973), is consistent with the Vaillant Formation being correlative with the Akaitcho Group.

Two periods of rift-related volcanism and sedimentation are preserved in the Athapuscow Aulacogen (Hoffman et al., 1977). The older, metamorphosed Wilson Island Group has been interpreted as an early, failed rifting

Table 5.1. Representative chemical analyses of Akaitcho Group volcanic rocks

			older		Nasittok sub-group				younger			
	low Si, high Zr rhyolite	high Si, low Zr rhyolite	Ipiutak Fm. basalt, amphibolite	Kapvik Dykes	Kapvik Volcanic Complex	Tuertok Complex	Zephyr Complex	Sinister Complex	Belleau lower	Complex upper	Tallerk gabbro Dykes	Vaillant Fm. basalts
no. of Analyses	(7)	(4)	(1)	(2)	(7)	(2)	-	-	(8)	(4)	-	(2)
SiO_2	70.0	76.3	47.7	49.8	49.6	50.6	-	-	52.0	49.1	-	49.6
Al_2O_3	13.7	12.8	15.2	13.6	14.3	16.6	-	-	14.2	16.6	-	14.3
TiO_2	.8	.2	1.2	1.0	1.3	1.2	-	-	1.4	1.0	-	1.3
Fe_2O_3	5.6	2.1	14.6	14.3	14.2	11.9	-	-	13.7	11.9	-	12.7
MnO	.05	.01	.25	.22	.19	.15	-	-	.18	.17	-	.18
MgO	1.2	.50	7.3	7.5	6.6	6.8	-	-	6.5	7.9	-	7.3
CaO	1.5	.35	9.4	10.4	10.0	9.8	-	-	9.0	10.8	-	12.0
Na_2O	2.0	2.5	2.6	2.6	2.3	2.0	-	-	2.4	2.1	-	2.4
K_2O	5.2	5.3	1.6	1.5	1.4	1.1	-	-	.52	.36	-	.35
P_2O_5	.12	.01	.14	.05	.15	.11	-	-	.13	.08	-	.12
no. of Analyses	(20)	(10)	(15)	(9)	(13)	(13)	(4)	(1)	(9)	(7)	(8)	(2)
Nb	16	21	7	7	8	7	5	22	8	24	4	6
Y	41	58	63	48	45	34	45	52	36	74	40	26
Zr	264	191	164	141	128	113	99	238	123	191	105	75
Sr	146	48	116	163	141	174	128	153	170	9	170	153
Rb	202	167	57	39	30	20	37	15	18	22	50	15
Ga	25	22	11	18	164	20	10	19	24	68	12	21
Ba	1088	594	176	173	310	173	168	154	131	246	160	270
V	49	20	339	368	101	269	429	342	296	133	377	325
Cr	17	8	88	38		103	50	72	69		82	100

All analyses performed by XRF at M.U.N., Errors as reported by Strong et al. (1978)

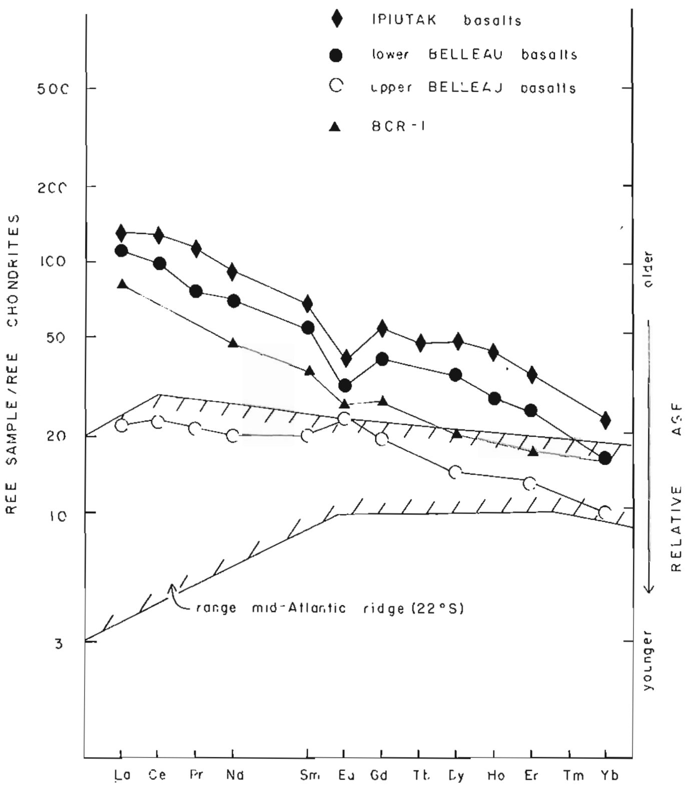


Figure 5.12. *Chrondrite normalized REE contents of Akaitcho Group basalts. The Ipiutak basalt plotted is representative of 15 Ipiutak basalts analyses. The lower Belleau basalt is representative of 15 lower Belleau basalts. Lower Kapvik basalts and Tuertok basalts are also similar to the lower Belleau basalts. The upper Belleau basalt is representative of 10 upper Belleau basalts. The upper Kapvik and Tuertok basalts are similar to the upper Belleau basalt. Field of mid-Atlantic ridge lavas from Bryan et al. (1976). REE were determined using the thin film technique of Eby (1972) as modified by Fryer (1977). Data normalized to Leedy chrondrite values of Masuda et al. (1973) divided by 1.20.*

event in the aulacogen (Hoffman et al., 1977). The Union Island Group, which forms the basal part of the Great Slave Supergroup (Hoffman, 1968, 1973) is presumed to date from the initial formation of the aulacogen (Hoffman et al., 1977). Only a limited amount of geochemical data is available for the Union Island Group basalts. Goff and Scarfe (1978) found that basalts in a marginal volcanic complex had alkaline affinities, whereas those of a later, medial volcanic complex are continental tholeiites. The data reported by Goff and Scarfe (1978) for the Union Island Group are unlike the data available for the Akaitcho Group (Easton, 1980; Fig. 5.15). The reason for this difference is unknown. One possible explanation is that the Union Island Group basalts had clinopyroxene as a residual phase, or underwent shallow level fractionation of clinopyroxene, whereas the Akaitcho Group basalts did not. Although the Union Island Group represents a rifting event in roughly the same stratigraphic position as the Akaitcho Group, the two groups may not be contemporaneous, and hence, not strictly correlative.

The term Snare Group was introduced by Lord (1942) for metasedimentary and metavolcanic rocks preserved west of the Bear-Slave boundary, and included rocks located on both sides of the Wopmay Fault. Although he did not designate a type area, it is evident that Lord (1942) considered metasedimentary rocks in the vicinity of Ingray and Mattberry lakes to represent typical Snare Group lithologies.

Since 1942, in various parts of the Wopmay orogen, particularly in the north, rocks which previously would have been assigned to the Snare Group have been assigned to one of the Akaitcho, Epworth, Labine (Hildebrand, 1981) or Sloan (Hoffman, 1978) groups. Thus, at present, there is some confusion over stratigraphic nomenclature within Zone 3 of the Wopmay Orogen. Workers in the type area of the Snare Group have suggested that the Snare Group may be correlative with the Epworth Group (e.g. McGlynn and Fraser, 1972). Metavolcanic rocks in the western Snare Group are probably correlative with the Akaitcho Group especially in the Calder River, Grant Lake and northern Bedrock Lake areas (Fig. 5.2). Unfortunately, there is very little geological information in the area between 65° and 66°N. Until this area is mapped, and Akaitcho and Epworth Group lithologies mapped southward to the Ingray and Mattberry Lake areas, firm correlations between the Snare, Epworth and Akaitcho groups cannot be made, although it is suspected that the Snare Group includes rocks of both the Akaitcho and Epworth groups. It is recommended that the term Snare Group be used for metasedimentary (and metavolcanics) rocks west of the Bear-Slave boundary and east of the Wopmay Fault which cannot, or have not been assigned to the Akaitcho or Epworth groups. This usage differs from Lord's (1942) usage in that it excludes rocks west of the Wopmay Fault.

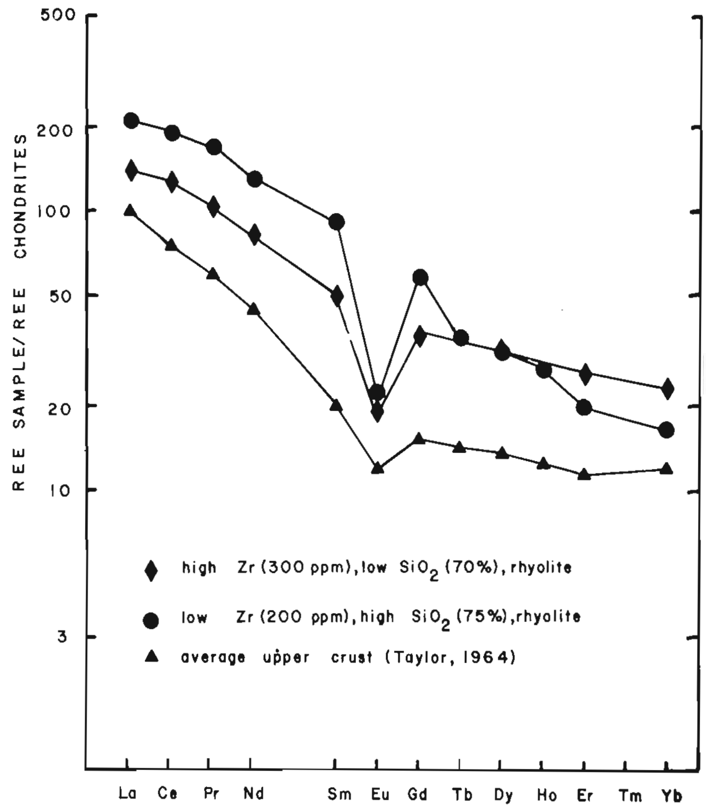


Figure 5.13. *Chrondrite normalized REE contents of Akaitcho Group rhyolites. High Zr rhyolite is representative of 5 Nasittok subgroup rhyolites and 6 Okrark sills. Low Zr rhyolite plotted is representative of 6 analyses. Average crust from Taylor (1964).*

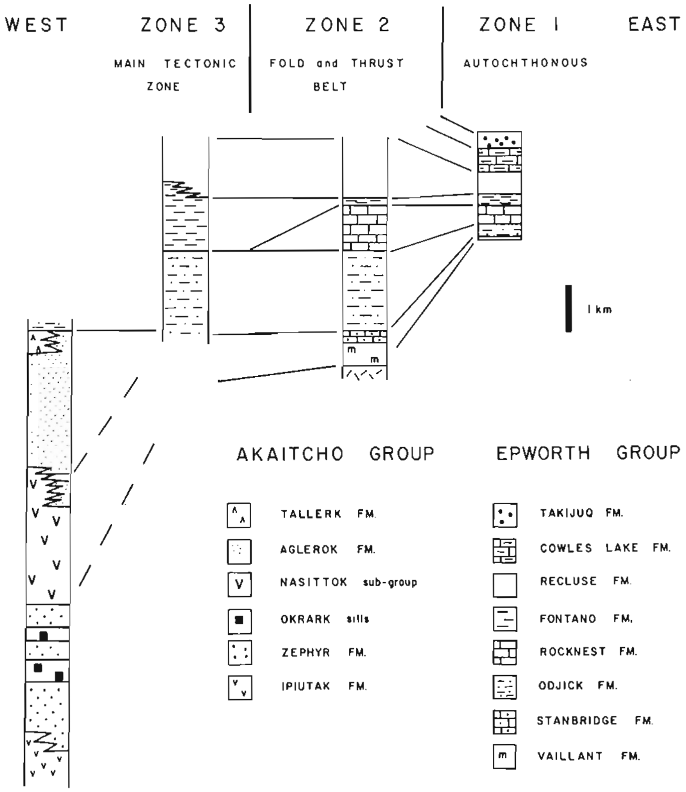


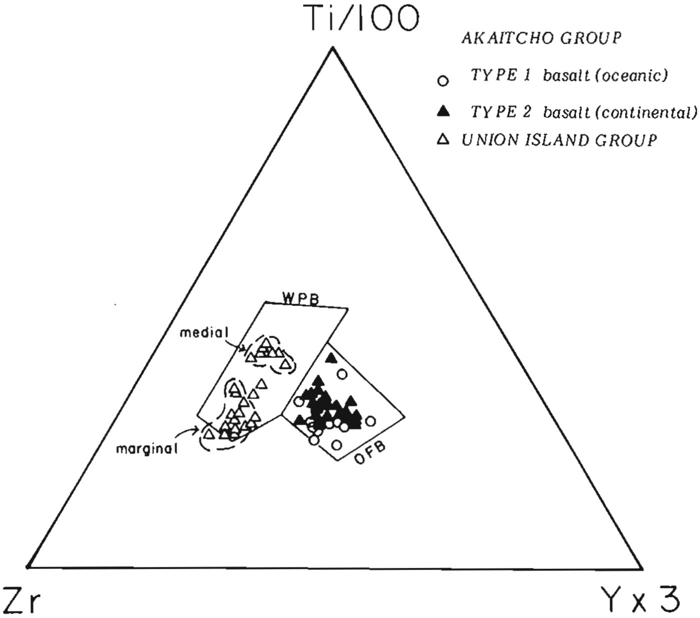
Figure 5.14. *Correlation of strata in the northern Wopmay Orogen, showing position of the Akaitcho Group relative to the Epworth Group. Epworth Group data from Hoffman (1973), and Hoffman et al. (1978).*

The age of the Akaitcho Group remains uncertain. A minimum age is 1800 Ma based on K-Ar cooling ages for the eastern Bear Province (Stockwell, 1970), and a maximum age is about 2100 Ma, the age of alkaline complexes and diabase dyke swarms presumably related to early rifting along the western margin of the Slave Craton (Hoffman, 1980). A Rb-Sr whole rock age of 1881 ± 82 Ma (2σ) has been obtained using 6 samples from an Okrark sill located 15 km south of Okrark Lake. The geological significance of this date, namely is it an age of metamorphism or an age of emplacement, has not yet been ascertained.

DEPOSITIONAL HISTORY OF THE AKAITCHO GROUP

Between 1.9 and 2.1 Ga, rifting began in the western part of the Slave Craton. Although basement to the Akaitcho Group has not been observed, the presence of probable crust-derived rhyolites suggests that the Akaitcho Group was initially deposited on continental crust. The thickness of the crust is unknown, but by analogy with recent rift terranes, probably thinned to the west towards the centre of the rift, by rotation along fault blocks in the upper crust (Morton and Black, 1975; Zanettin and Justin-Visentin, 1975) and by ductile spreading of the lower crust (Bott, 1976; Montadert et al., 1979).

The oldest rocks observed in the Akaitcho Group in the northern Wopmay are continental tholeiitic basalts; hence, rifting may have been well advanced by the time these basalts were erupted. Nevertheless, continental terranes were nearby, and shed voluminous arkosic turbidites into the Akaitcho basin. As rifting progressed, large basalt volcanoes were constructed. The tholeiitic chemistry of the Akaitcho



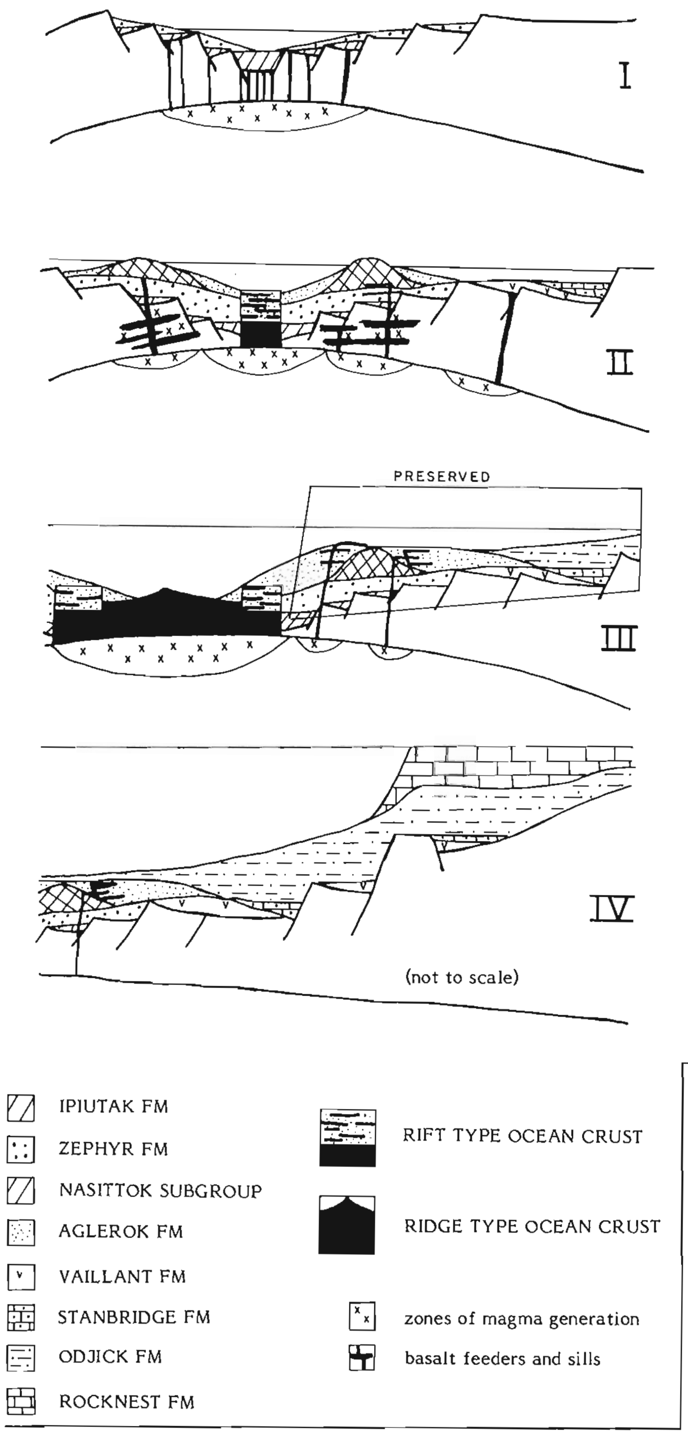
WPB - within plate basalt field
OFB - ocean floor basalt field boundaries after Pearce and Cann (1973)

Figure 5.15. *Ti-Y-Zr plot for Akaitcho Group and Union Island Group basalts. Union Island Group data from Goff and Scarfe (1978). Plot is not considered to have significance regarding the tectonic environment of the basalts.*

Group basalts in these volcanoes, suggests significant degrees of partial melting in the mantle source area. The large volumes of basaltic magma introduced sufficient heat into the crust to cause crustal melting and production of the abundant rhyolites of the Akaitcho Group. The abundance of basalt tuffs suggests that some of the complexes formed in shallow water. Probable welded ash-flow tuffs indicate that the upper parts of the volcanoes were probably emergent as islands. The basin probably deepened and widened with time, as finer grained sedimentary rocks become dominant in the upper part of the Akaitcho Group. Volcanism appears to wane, probably due to westward migration of the centre of volcanism. Eventually, the continent-derived clastic rocks of the westward prograding Odjick Formation buried the Akaitcho Group. The tectonic evolution of the Akaitcho Group is shown schematically in Figure 5.16. The nature of the crust which was present on the western margin of the craton during the later stages of Akaitcho Group deposition is discussed in the next section. The subsequent history of Epworth Group deposition has been summarized by Hoffman (1973, 1980).

DISCUSSION

The discussion is divided into two sections. The first section compares the Akaitcho Group to modern rift sequences. The purpose of this discussion is to point out similarities between the Akaitcho Group and modern rifts, to show that the absence of peralkaline rhyolite in the Akaitcho Group is not of great import, and to suggest that oceanic crust is not observed in rift areas because the processes involved in the generation of oceanic crust in rifts is different than the process in operation at spreading ridges. In the second section, the significance of the Akaitcho Group to the development of the Wopmay Orogen is discussed.



I - post-Ipiutak formation deposition and syn-Zephyr formation deposition. Akaitcho Group units older than the Ipiutak formation, if they exist, are considered part of the basement.

II - late Nasittok subgroup deposition. Rift-type ocean crust (see text for discussion) has developed west of the continental crust underlying the Nasittok subgroup.

III - late Tallerk formation deposition. Centre of volcanism has migrated to the west, development of ridge-type ocean crust is inferred to have begun to the west.

IV - late Rocknest Formation deposition. Akaitcho Group is buried beneath a passive continental margin sedimentary succession.

Figure 5.16. *Cartoon illustrating the evolution of the Coronation continental margin during the early Proterozoic.*

Comparison

In the Afar triangle north of the lake region of the Ethiopian rift, Barberi et al. (1972a, 1975) and Barberi and Varet (1977, 1978) have recognized three types of volcanism: oceanic volcanism of the axial ranges, volcanic centres near the rift margins underlain by continental crust, and the stratoid series of southern Afar. The stratoid series contains transitional flood basalts with moderate LREE enrichment interlayered in the upper part with silicic flows and ash-flow tuffs. Locally, central cumulo-volcanoes, without summit calderas, formed by lava domes with scanty pyroclastics developed. Porphyritic lavas are subalkaline rhyolites. Peralkaline lavas are glassy, and only occur high in the stratigraphic sequence of the cumulo-volcanoes. The stratoid series is underlain by an unknown thickness of clastic sedimentary rocks (Chessex et al., 1975; Mohr, 1978). The overall stratigraphy of the stratoid series, in particular the presence of cumulo-volcanoes with few pyroclastic rocks constructed on large basaltic plains, and the basalt geochemistry is similar to that observed in the Nasittok subgroup, in particular, the Tuertok and Kapvik volcanic complexes. The major difference is that the Nasittok subgroup was mainly deposited subaqueously. The absence of peralkaline rhyolites in the Nasittok subgroup does not appear to be a significant difference, because they are only preserved at high stratigraphic levels in the stratoid series, and hence would be easily eroded. Barberi et al. (1971, 1975), Barberi and Varet (1977), and Coombs (1963) have attributed the origin of some peralkaline rhyolites to crystal fractionation from a mildly alkaline basalt parent, probably under conditions of low pH_2O-pO_2 (Barberi et al., 1975). Thus, the absence of peralkaline rhyolites in the Nasittok subgroup can be attributed to the lack of a suitable parent (no alkali basalts have been found in the Akaitcho Group) and possibly high pH_2O-pO_2 conditions in the basaltic source region. In addition, peralkaline rhyolites, commonly ignimbritic, are more common in the marginal volcanic centres and the axial ranges of Afar (Barberi et al., 1972a, b, 1975; Barberi and Varet, 1977, 1978) where high level magma chambers are present. The high level magma chambers are probably necessary for crystal fractionation, and volatile accumulation (as in calc-alkaline magma chambers; Smith, 1978; Hildreth, 1979) responsible for the generation of peralkaline rhyolites. Such large chambers appear to be absent beneath the stratoid series (Barberi et al., 1975; Barberi and Varet, 1977, 1978). This could also be the case for the Nasittok volcanic complexes. Finally, because the Akaitcho Group rhyolites are derived from crustal fusion, and are not related to the basalts through differentiation, the absence of peralkaline rhyolites is not surprising.

The presence of oceanic basement beneath the Afar region is in dispute by geophysicists and geologists (see discussion by Barberi and Varet, 1975a, b; Mohr, 1978). The geophysicists note that the crust is abnormally thick for oceanic basement (Fig. 5.17) and have interpreted the 6.0-6.3 km/sec layer as representing attenuated upper continental crust (Mohr, 1978).

The geologists interpret the layer as a metamorphosed dyke complex, similar to layer '2B' of oceanic crust (Mohr, 1978). In the latter case, the oceanic crust beneath the Afar is unusually thick compared to normal ridge-type ocean crust (Fig. 5.17), and the greater thickness of this rift-type ocean crust has been attributed to a slow extension rate that permits mantle derived magmas to accumulate, and differentiate in the crust (Barberi and Varet, 1975a). The net effect produces oceanic crust different from ridge-type crust. The presence of ridge-type oceanic crust in the Red Sea and the Gulf of Aden (Fig. 5.17) indicates that the thick rift type ocean crust either evolves into ridge-type ocean crust, or that it is a transitional stage between oceanic and continental crust.

The Guayamas Basin is part of the developing Gulf of California rift. Preliminary results of DSDP Leg 64 indicate that in the axial rift areas, and between rifts, basalt flows have been intruded into rapidly deposited soft sediments. Younger sills overlie the older sills because the sills can penetrate soft sediments more easily than older, compacted, intruded and hydrothermally altered sediments (Einsele et al., 1980; Curray et al., 1979). The important fact is that this process of sill injection produces an oceanic basement quite different in stratigraphy (Einsele et al., 1980) and in thickness (Moore, 1973; Fig. 5.17) from normal ophiolites. If the type of oceanic basement described from the Gulf of California and the Afar is typical of rifts, this could explain the absence of recognizable ophiolite stratigraphy in rifting sequences preserved in orogens (see discussion by Rankin, 1975).

Extent of rifting

The remaining question is whether the Akaitcho Group formed in an ensialic rift, of which only the east margin is now exposed, or if rifting continued, with the development of an Atlantic type continental margin and generation of ridge-type oceanic crust.

Hoffman (1973, 1980) and Hoffman et al. (1978) have presented convincing stratigraphic and sedimentological evidence for the development of an Atlantic type continental margin along the west margin of the Slave Craton in the early Proterozoic. This continental margin developed after deposition of the Akaitcho Group as pelites of the Odjick Formation overlie the Akaitcho Group.

Hoffman (1980; see Table 5.2) has summarized the evidence for rifting along the western edge of the Slave Craton in the early Proterozoic. This evidence includes the presence of three, 2.1 ± .1 Ga alkaline-peralkaline complexes, four diabase dyke swarms all about 2.1 Ga in age, and the presence of two aulacogens spaced 800 km apart which strike at high angles to the trend of the Wopmay Orogen (Fig. 5.18). This close association of alkaline complexes, four diabase dyke swarms, aulacogens and dyke swarms is similar to magmatic and tectonic activity observed during the breakup of continental areas during the Cenozoic (Burke and Dewey, 1973; Burke, 1976, 1977).

In addition to the above evidence, there is the rift related Akaitcho Group which is overlain by a passive continental terrace-rise sedimentary sequence. The evolution of the Akaitcho Group tholeiites from the older,

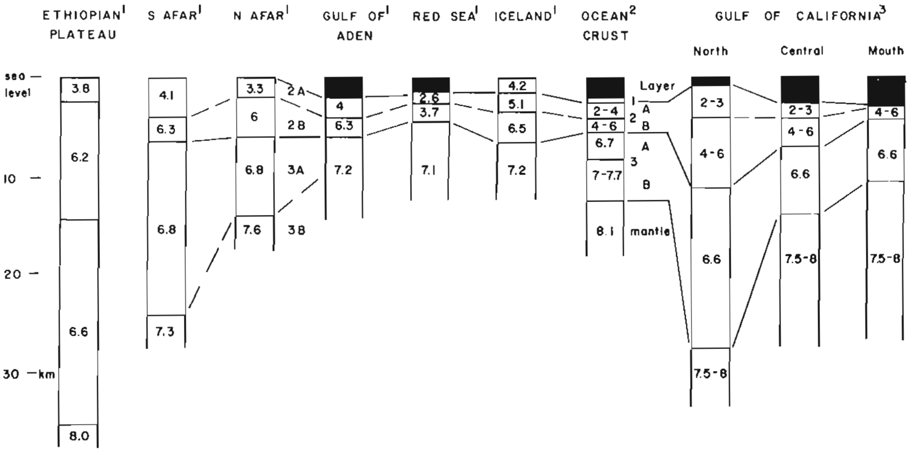


Figure 5.17. Crustal sections for the Afar rift (Mohr, 1978), ridge type ocean crust (Salisbury et al., 1979), and the Gulf of California (Moore, 1973). Note the greater thickness of oceanic crust in the developing rifts, particularly north and central Gulf of California. Layer with 4-6 km/sec velocities in the Gulf of California is a mixed layer of basalt and sediment (Moore, 1973; Einsele et al., 1980).

Table 5.2. Comparison of the evidence of rifting in the Wopmay Orogen and the Appalachian Orogen

	Wopmay Orogen (early Proterozoic 2.1 to 1.8 Ga)	Appalachian Orogen (Eocambrian to Permian)
Volcanism	Akaitcho Group (this paper) bimodal (i) basalt - continental tholeiite (early) - ocean tholeiite (late) (ii) rhyolite - subalkaline, crustal derived	- Catoctin, Tibbet Hill, Mt. Rodgers and Ashe formations - tholeiitic basalt (decrease in alkalinity towards promontories? (Rankin, 1976) - peralkaline rhyolites and granites associated with promontories (Rankin, 1976)
Sedimentary Fill	- arkosic turbidites, continent derived (early) - pelites and volcaniclastic sediments (late) overlain by continental shelf clastics	- variable, coarse clastics and conglomerates in some areas, pelite and volcaniclastic sediments elsewhere, sediment thickness increased towards promontories; sediments and volcanics overlain by continental shelf clastics derived from the west (Rodgers, 1972; Rankin, 1975; Thomas, 1977)
Basement	- volcanics rest on basement in fold and thrust belt (Zone 2) - basement not observed in Zone 3, but continental in part because of abundant crustal derived rhyolites	- continental crust (Grenville) overlain by volcanics, or conglomerates
Associated Magmatism	- Simpson Island Dyke (Burwash and Cavell, 1978; Badham, 1979; Reinhardt, 1972) - Blachford Alkaline Complex (Davidson, 1978) - Bigspruce nepheline syenite carbonate complex (Martineau and Lambert, 1975) - Indin, Dogrib, 'X' and Hearne diabase dykes (Hoffman, 1980)	- dykes in northern Newfoundland and on Belle Island cutting Greenville basement (Strong and Williams, 1972; Strong, 1975) - alkaline complexes (Doig, 1970) generally along St. Lawrence-Ottawa graben
Associated Rifts	- Athapuscow aulacogen (Hoffman, 1973, this volume; Hoffman et al., 1980) - Taktu failed arm (Campbell and Cecile, this volume) - aulacogen spaced 800 km apart	- southern Oklahoma aulacogen (Hoffman et al., 1974; Ham, 1969) - Ottawa-St. Lawrence-Bonchere Graben - promontories spaced 600-800 km apart

continental Ipiutak basalts to the younger, oceanic, upper Belleau basalts (Fig. 5.12) indicates that products of rifting preserved by the Akaitcho Group progressed to the stage where rift-type ocean crust was being generated. The presence of the overlying continental terrace-rise sequence indicates that rifting was probably successful, and that ridge-type ocean crust was generated. This is unlike the late Proterozoic Burin Group in the Avalon Zone of the Appalachians (Strong et al., 1978; Strong and Dostal, 1978, 1980), where only rift-type ocean crust was generated and where there is no overlying continental terrace-rise sequence.

Thus, it seems probable that an ocean basin floored by ridge-type ocean crust did develop off of the western margin of the Slave Craton in the early Proterozoic.

Hoffman et al. (1970) and Hoffman (1973, 1980) have previously emphasized the similarity between the Wopmay Orogen and Paleozoic and Cenozoic orogens. As a final point, I submit that the evidence for rifting in the Wopmay Orogen is as compelling as that for the Appalachian Orogen (Table 5.2; Fig. 5.18). The difference between the two orogens is that ophiolite nappes present in the northern Appalachians provide independent evidence that ridge-type ocean crust did exist, whereas in the Wopmay Orogen, this definitive evidence is absent.

SUMMARY

In summary, the early Proterozoic Akaitcho Group preserves products of rifting similar to those found in modern rift valleys. The Akaitcho Group basalts evolved upward from continental tholeiites to younger oceanic tholeiites, a geochemical evolutionary pattern observed in modern rifts such as the Afar. The geochemical evolution of the Akaitcho Group basalts, and the subsequent burial of the Akaitcho Group by an early Proterozoic continental terrace sequence indicates that ridge-type ocean crust most probably developed off the western margin of the Slave Craton during the early Proterozoic. Subsequent deformation of the Akaitcho Group and other supracrustal rocks of the Wopmay Orogen probably reflects the destruction of this oceanic crust.

Some important, outstanding problems in our understanding of the Akaitcho Group are:

1. The nature, composition and thickness of the Akaitcho basement.
2. If preserved, what are the lower parts of the Akaitcho Group like? If volcanic rocks are present, are they alkaline or subalkaline?

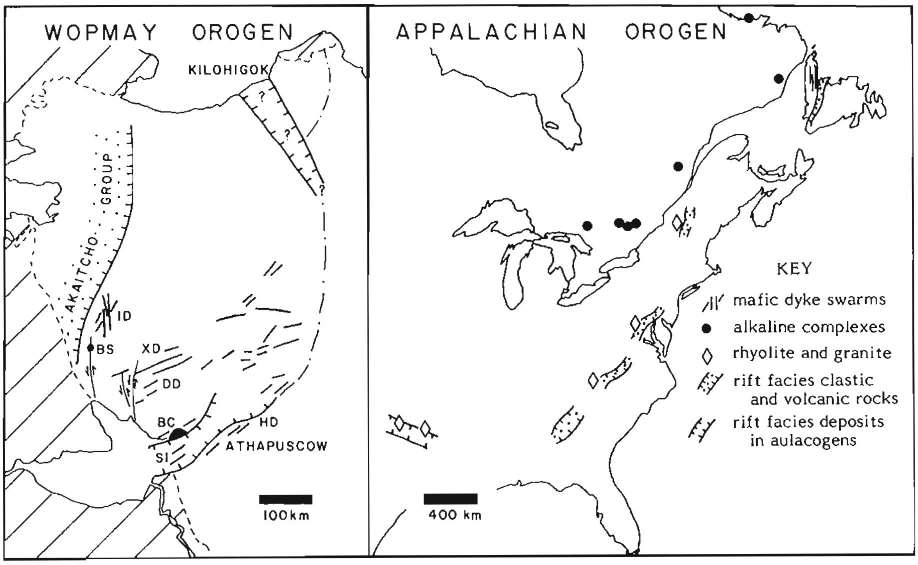


Figure 5.18

Comparison of rocks related to the early Proterozoic rifting of the Wopmay Orogen (after Hoffman, 1980) and Eocambrian rifting of the Appalachian Orogen (after Rodgers, 1972; Rankin, 1976; and Thomas, 1977). Note difference in scale between the two orogens. For a more detailed map of the Eocambrian rift facies rocks consult Williams (1978).

3. Does the character of the Akaitcho Group change to the south?
4. What is the age of the Akaitcho Group? An answer to this question might provide some constraint on the width of the ocean basin.

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6.

AUTOPSY OF ATHAPUSCOW AULACOGEN: A FAILED ARM AFFECTED BY THREE COLLISIONS

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SUMMARY

Athapuscow Aulacogen is a deformed east-northeast-trending basin, 270 km long by at most 80 km wide, of little-metamorphosed early Proterozoic sedimentary and magmatic rocks exposed in and around the East Arm[1] of Great Slave Lake, Northwest Territories. To the northwest is the Slave Province, a 2.6 Ga craton mostly stripped of its Proterozoic cover. Southeast of the aulacogen is the "Northwest Churchill Province", consisting mostly of high-grade Archean (?) rocks, strongly retrograded, and relicts of a low- to medium-grade volcanic-sedimentary cover sequence (Wilson Island Group), locally intruded by granite, that predates the aulacogen. These metamorphic and basement rocks are involved in a 20-km-wide east-northeast-trending zone of severe mylonitization (Slave-Chantrey Mylonite Zone) that skirts the southeast border of the aulacogen. The western end of the aulacogen is onlapped by Phanerozoic cover, beneath which it probably merges with the foreland sedimentary prism (Coronation Supergroup) of Wopmay Orogen, which evolved contemporaneously with the aulacogen. Zircon chronology suggests that the major orogen (Trans-Hudson Orogen) between the northwest Churchill Province and the Superior Province closed within the same time interval (1.8-1.9 Ga) as Wopmay Orogen and deformation of the aulacogen.

Excluding the aforementioned Wilson Island Group, the stratigraphy of the aulacogen consists of two relatively conformable sequences separated by a regional angular unconformity. The younger sequence (Et-then Group) is virtually unmetamorphosed and consists of alluvial fanglomerate, locally with basalt flows, and pebbly fluvial sandstone. The older sequence (Great Slave Supergroup) is subgreenschist in grade and records a grand transgressive-regressive cycle that has been correlated in detail with the Coronation Supergroup of Wopmay Orogen and the Goulburn Group of Kilohigok Basin, a less deformed basin on the Slave Craton. Basal conglomerates of the Great Slave Supergroup contain clasts of Wilson Island Group and basement rocks derived in part from the Slave-Chantrey Mylonite Zone.

The Great Slave Supergroup comprises five formal groups, the Union Island, Sosan, Kahochella, Pethei and Christie Bay groups, listed in ascending order. The Union Island Group is preserved only in a narrow (15 km) east-northeast-trending zone and comprises a sequence of dolomitic, carbonaceous and argillaceous lake (?) sediments that contain buildups of both tholeiitic and alkalic basalt flows and related gabbro sills. The Sosan Group disconformably overlies the Union Island Group and extensively onlaps the basement. Its lower part (Hornby Channel and Duhamel formations) is mostly devoid of volcanics and exhibits considerable local variation in thickness and facies. It is generally thickest, about 3 km, in a zone coincident with but broader and longer than the preserved Union Island Group, hereafter referred to as the "axial zone". The sequence

ABRÉGÉ

L'aulacogène d'Athapuscow est un bassin déformé orienté est-nord-est, d'une longueur de 270 km et d'une largeur maximale de 80 km. Il se compose de roches magmatiques et sédimentaires du Protérozoique inférieur, peu métamorphisées, qui sont exposées dans le bras Est (East Arm)[1] du Grand lac des Esclaves et à proximité. Au nord-est se trouve la province des Esclaves, craton de 2,6 Ga et débarrassé en bonne partie aujourd'hui de sa couverture protérozoïque. Au sud-est de l'aulacogène, se trouve la portion nord-ouest de la province de Churchill, composée surtout de roches archéennes (?) fortement métamorphisées, et présentant les effets d'un intense métamorphisme régressif, ainsi que les restes d'une succession volcano-sédimentaire de couverture, faiblement à moyennement métamorphisée (groupe de Wilson Island), traversée localement par des intrusions de granite, et plus récente que l'aulacogène. Ces roches métamorphiques et roches du soubassement font partie d'une zone d'intense mylonitisation de 20 km de large, orientée est-nord-est (zone mylonitisée de Slave-Chantrey) qui se trouve à la bordure sud-est de l'aulacogène. La partie ouest de l'aulacogène est chevauchée par une couche du Phanérozoïque, et au-dessous de celle-ci, elle se confond probablement avec le prisme sédimentaire de la plate-forme continentale (supergroupe de Coronation) de l'orogène de Wopmay, qui a évolué en même temps que l'aulacogène. D'après la datation réalisée sur le zircon, le principal orogène (orogène Trans-Hudsonien) situé entre le nord-ouest de la province de Churchill et la province du lac Supérieur s'est fermé en même temps (1,8-1,0 Ga) que l'orogène de Wopmay et pendant la déformation de l'aulacogène.

Exception faite du groupe de Wilson Island susmentionné, la stratigraphie de l'aulacogène comporte deux successions relativement concordantes, séparées par une discordance angulaire régionale. La succession la plus jeune (groupe de Et-then) n'est pratiquement pas métamorphisée; elle se compose d'un fanglomérat localement recouvert de coulées basaltiques et de grès fluviatiles caillouteux. La succession la plus ancienne (supergroupe de Great Slave) a été métamorphisée dans le sous-faciès des schistes verts et contient les vestiges d'un important cycle de transgression et régression, qui a été corrélé étroitement avec le supergroupe de Coronation dans l'orogène de Wopmay et le groupe de Goulburn dans le bassin de Kilohigok, moins déformé et situé sur le craton du Grand lac des Esclaves. Les conglomérats de base du supergroupe de Great Slave contiennent des roches clastiques du groupe de Wilson Island et des roches de soubassement provenant en partie de la zone de mylonite de Slave-Chantrey.

Le supergroupe de Great Slave se compose de cinq groupes reconnus, appelés groupes de Union Island, de Sosan, de Kahochella, de Pethei et de Christie Bay, dans l'ordre ascendant. Le groupe de Union Island n'est conservé que dans

[1] "East Arm" is not an officially recognized place name and does not appear on topo maps, although it is in regular use by residents of Great Slave Lake.

Le toponyme "East Arm" n'est pas un toponyme officiellement reconnu, par contre, il est utilisé couramment dans la région du Grand lac des Esclaves.

consists of a lower pebbly subfeldspathic arenite (Hornby Channel Formation), deposited by braided rivers that flowed uniformly to the west-southwest, overlain gradationally by peritidal stromatolitic dolomite (Duhamel Formation). The dolomite is commonly missing beneath a disconformity at the base of the upper Sosan Group, a laterally more uniform sandstone blanket about 1 km thick, again with west-southwesterly directed paleocurrents. This blanket consists of a lower white to pink quartzose sandstone (Kluziai Formation) that grades upward into red micaceous sand-siltstone (Akaitcho River Formation). Tholeiitic basalt centres, locally with rhyolite cauldrons and domes, are developed especially near the top of the sandstone blanket (e.g. Seton Formation). The top of the blanket marks a major deepening of the water column and deposition of non-mudcracked red and green marine (?) shale (Kahochella Group). The shale increases smoothly in thickness from less than 0.5 km on the northwest margin of the aulacogen to more than 2.5 km toward the axial zone. Siltstone turbidites occur in the axial zone and were derived at least in part from the southwest. Basaltic and bimodal volcanic centres flared up intermittently during shale deposition along northeast and east-northeast-trending lines. Volcanism ceased during deposition of the succeeding 0.5 km thickness of carbonates (Pethei Group), which undergo a marked facies change from shallow water cryptalgal platforms and ramps along the northwest margin of the aulacogen, through a relatively thin transitional slope facies, to deep-water limestone and marlstone that intertongue in the axial zone with greywacke turbidites derived from the southwest. Halite casts at the top of the basinal facies Pethei Group signal an impending salinity crisis. The succeeding unit is a broken formation (Stark Megabreccia), comprising a pervasively brecciated consanguineous association of peritidal dolomite-limestone and red halite-casted silt-mudstone that becomes sandy toward the west-southwest. The megabreccia sharply overlies all facies of the Pethei Group and is thought to have resulted from withdrawal of salt originally deposited between the deeper water facies of the Pethei Group and the preserved shallow water lithologies in the megabreccia. The megabreccia was further disrupted during subsequent tectonic overthrusting. The megabreccia and the succeeding nonbrecciated redbed sequence, mostly nonmarine, together comprise the Christie Bay Group, about 1.5 km thick. The immature fluvial and deltaic sandstones (Tochatwi Formation) overlying the megabreccia are derived from the southwest, as is the terrigenous component of the megabreccia itself. The fluvial sandstones are succeeded by playa lake (?) sediments (Portage Inlet Formation) capped by subaerial basalt flows (Pearson Formation).

In addition to the volcanism already mentioned, there were at least four significant episodes of intrusive magmatism of contrasting character. Alkaline and peralkaline intrusions, 2.1-2.2 Ga in age, are recognized on the north (Blachford Intrusive Suite) and south (Simpson Island Dyke) sides of the aulacogen. Both are intruded by east-northeast-trending swarms of altered diabase dykes (Hearne Dykes) that appear to predate the Union Island and Sosan groups. The alkaline-peralkaline complexes, Hearne Dykes and Union Island Group alkalic-tholeiitic volcanism could all be closely related but, as age control is lacking, the unlikely possibility that the alkaline-peralkaline complexes, or even the Hearne Dykes, predate the Wilson Island Group cannot be completely excluded.

The Jackson Gabbros are a suite of minor intrusives commonly associated with the upper Sosan Group and Kahochella Group volcanism (Seton volcanism sensu lato) but more important are the calc-alkaline Compton Laccoliths, which postdate the Great Slave Supergroup and predate the

une étroite (15 km) bande orientée est-nord-est et comprend une succession de sédiments lacustres (?) dolomitiques, carbonacés et argileux qui contiennent des accumulations de coulées basaltiques tholéiitiques et alcalines, ainsi que les sills gabbroïques associés. Le groupe de Sosan repose en discordance d'érosion sur le groupe d'Union Island et recouvre une grande partie du soubassement. Sa partie inférieure (formations de Hornby Channel et de Duhamel) est presque entièrement dépourvue de roches volcaniques et présente localement de grandes variations d'épaisseur et de faciès. Son épaisseur est généralement maximale, 3 km environ, dans une zone qui coincide avec le groupe encore existant d'Union Island, plus large et plus long, appelé plus loin "zone axiale". La succession se compose d'une arénite inférieure, de caractère subfeldpathique et caillouteux (formation de Hornby Channel), déposée par des cours d'eaux anastomosés qui coulaient uniformément en direction ouest sud-ouest; ce dépôt se transforme graduellement en une dolomie stromatolitique péritidale (formation de Duhamel). La dolomie est souvent absente au-dessous d'une disconcordance d'érosion située à la base de la portion supérieure du groupe de Sosan, qui forme une couverture latéralement plus uniforme de 1 km d'épaisseur environ, et où apparaît l'orientation ouest-sud-ouest des paléocourants. Cette couche est composée des grès quartzeux inférieurs blancs à roses, (formation de Kluziai) qui passe vers le haut à un siltstone sableux et micacé rouge (formation d'Akaitcho River). Il existe des cheminées de basaltes tholéiitiques, accompagnés localement de caldeiras et de dômes rhyolitiques, surtout près du sommet de la couverture de grès (par exemple, la formation de Seton). Celui-ci montre un important accroissement de la colonne d'eau et le dépôt de schiste argileux marin (?) rouge et vert dépourvu de fentes de dessiccation (groupe de Kahochella). L'épaisseur du schiste argileux augmente très progressivement, de moins de 0,5 km sur la marge nord-ouest de l'aulacogène à plus de 2,5 km près de la zone axiale. On rencontre des turbidites de siltstone dans la zone axiale, partiellement issues du sud-ouest. Des cheminées basaltiques et de caractère bimodal sont sporadiquement entrées en éruption durant la sédimentation des schistes argileux, le long de lignes d'orientation nord-est et est-nord-est. Le volcanisme a cessé durant la sédimentation de la couche carbonatée suivante (groupe de Pethei), d'une épaisseur de 0,5 km, dont le faciès varie fortement d'un faciès de plates-formes et talus cryptalgaux le long de la marge nord-ouest de l'aulacogène, à un faciès de transition incliné relativement mince, puis à des calcaires et marnes d'eau profonde s'interdigitant dans la zone axiale avec des turbidites de grauwacke provenant du sud-ouest. Des empreintes de halite au sommet du groupe de Pethei, dont le faciès est celui d'un bassin, signalent un changement brutal de la salinité. La couche suivante est une formation fragmentée (mégabrèche de Stark), qui comprend des brèches, associées et simultanément formées, de dolomies et calcaires et de mudstones et silts contenant des empreintes de halite qui deviennent sableuses dans la direction est-sud-ouest. La mégabrèche recouvre nettement tous les faciès du groupe de Pethei; on suppose qu'elle s'est formée pendant le retrait du sel déposé initialement entre le faciès d'eau profonde du groupe de Pethei et les sédiments d'eau peu profonde conservés dans la mégabrèche. Celle-ci a été perturbée à nouveau pendant le charriage tectonique ultérieur. La mégabrèche et la succession suivante à red beds non bréchifiée et en majorité non marine, forment ensemble le groupe de Christie Bay, d'à peu près 1,5 km d'épaisseur. Les grès fluviatiles et deltaïques immatures (formation de Tochatwi) qui recouvrent la mégabrèche proviennent du sud-ouest, comme la composante terrigène de la mégabrèche elle-même. Les grès fluviatiles sont suivis de

Et-then Group. They are hornblende or hornblende-biotite diorites (in the west half of the aulacogen) and quartz monzonites (in the east half), using IUGS definitions. The laccoliths are generally floored by the top of the Pethei Group and bulge discordantly upward into the Stark Megabreccia. The laccoliths definitely transgress and postdate the thrusting of the Great Slave Supergroup. Except for contrasting host rocks, the laccoliths have much in common with certain intrusions in the 1.87 Ga calc-alkaline Great Bear Magmatic Zone of Wopmay Orogen. Finally, after more than 600 Ma of tectonic inactivity, the aulacogen was intruded by a 125 km-long cone sheet (Fortress Gabbro) and throughgoing north-northwest-trending diabase dykes during the Mackenzie Igneous Event at about 1.22 Ga.

Deformation of the aulacogen occurred in two distinct phases. The first occurred after Great Slave Supergroup deposition and before intrusion of the calc-alkaline Compton Laccoliths. It involves thick-skinned northwest-directed overthrusting of the northwest Churchill Province toward the Slave Craton, resulting in the stacking of pellicular allochthonous sheets of axial zone facies rocks onto the tectonically steepened northwest margin of the aulacogen. The southeast margin of the aulacogen was destroyed by erosion, probably at this time. In this interpretation, the basement rocks along the present-day southeast margin of the aulacogen (e.g. Simpson Islands-Hornby Channel-Union Island area and south of the McDonald Fault) and the axial facies Union Island and Sosan group rocks in depositional contact with that basement are allochthonous with respect to the Slave Craton. They tectonically overlie the sedimentary allochthons between the two basement blocks. To what extent the sedimentary allochthons are thrust sheets rooted beneath the southeast basement allochthon or gravity slides derived from the basement allochthon itself is uncertain. Resolution of this problem could have far-reaching paleogeographic implications.

The allochthons can be readily correlated along the length of the aulacogen and are classified into six groups or tiers on the basis of stratigraphic interval and sedimentary facies involved, structural style and position relative to other allochthons or nappes within the stack. The lowest tier consists of multiply imbricated and strongly refolded slope facies Pethei Group limestone and uppermost Kahochella Group shale. These are parautochthonous nappes in the sense that they lie above nearly the same slope facies in the autochthon, or have at most been translated only across the outer edge of the shallow-shelf facies zone. The overlying "Bunting Nappes" are composed of thick axial basin-facies Kahochella and Pethei group rocks, detached near the base of the Kahochella Group. Structurally, these nappes preserve the nose and/or one or the other limb of major northwest-facing recumbent anticlines, the overturned limbs of which are strongly sheared. All of the Christie Bay Group is probably allochthonous, having moved laterally without stratigraphic dislocation by slip within the Stark Megabreccia, which behaved as a ductile unit during overthrusting. The Christie Bay Group is preserved mainly in front of the Bunting Nappes, tectonically prograding onto the autochthon, but the Stark Megabreccia locally encloses the leading edges of the Bunting, Meridian and parautochthonous nappes and occurs as tectonic lenses along major nappe-bounding thrusts. The "Meridian Nappes" are composed of thick axial-zone arenites of the Sosan Group, very similar lithologically to outliers preserved on the southeastern basement allochthon. The Meridian Nappes are well developed only in the northeast part of the aulacogen, where the then adjacent basement allochthon (i.e. before strike-slip faulting) is almost devoid of such outliers. The reverse is true to the southwest – the outliers are extensive and the

sédiments de lac de playa (?) (formation de Portage Inlet) coiffés de coulées basaltiques subaériennes (formation de Pearson).

En plus du volcanisme déjà mentionné, il y a eu au moins quatre importantes périodes de magmatisme intrusif, de caractère contrasté. Des intrusions alcalines et peralcalines, âgées de 2,1 à 2,2 Ga apparaissent sur les côtés nord (série intrusive de Blachford) et sud (dyke de Simpson Island) de l'aulacogène, qui sont traversés par un essaim de dykes de diabase altérés (dykes de Hearne) d'orientation est-nord-est et sont nettement ultérieurs aux groupes d'Union Island et de Sosan. Les complexes alcalins et peralcalins, les dykes de Hearn et le volcanisme alcalin et tholéiitique du groupe d'Union Island pourraient tous être étroitement liés mais, comme la datation fait défaut, on ne peut exclure totalement la mince possibilité que les complexes alcalins et peralcalins, voire les dykes de Hearne, aient précédé le groupe de Wilson Island.

Les gabbros de Jackson constituent une série intrusive mineure fréquemment associée au volcanisme du groupe supérieur de Sosan et à celui du groupe de Kahochella (au sens large, volcanisme de Seton); mais plus importantes sont les laccolites calco-alcalines de Compton, plus jeunes que le supergroupe de Great Slave mais plus anciennes que le groupe de Et-then. D'après la classification de l'Union internationale des sciences géologiques, il s'agit de diorites à hornblende ou à hornblende et biotite (dans la moitié ouest de l'aulacogène) et de monzonites quartziques (dans la moitié est). Les laccolites reposent dans l'ensemble sur le sommet du groupe de Pethei et forment un bombement au-dessous de la mégabrèche de Stark. Elles dépassent nettement le charriage du supergroupe de Great Slave et sont ultérieures à celui-ci. Excepté la roche encaissante contrastée, les laccolites ont beaucoup en commun avec certaines intrusions de la zone magmatique calco-alcaline de Great Bear de l'orogène de Wopmay (âgée de 1,87 Ga). Enfin, après plus de 600 Ma d'inactivité tectonique, l'aulacogène a été pénétré par un complexe annulaire (cone sheet) d'une longueur de 125 km (gabbro de Fortress) et traversée par des dykes de diabase d'orientation nord-nord-ouest durant l'intrusion éruptive de Mackenzie (1,22 Ga).

La déformation de l'aulacogène a eu lieu en deux phases distinctes. La première phase est ultérieure au dépôt du supergroupe du Grand lac des Esclaves et antérieure à l'intrusion des laccolites calco-alcalines de Compton. Elle est due à un important charriage du nord-ouest de la province de Churchill vers le craton du Grand lac des Esclaves, puis à l'empilement de minces nappes allochtones issues des faciès de la zone axiale sur la bordure nord-ouest de l'aulacogène, tectoniquement accentuée. La bordure sud-ouest de l'aulacogène a été détruite par l'érosion, probablement à la même époque. Dans cette interprétation, les roches du soubassement qui bordent la marge sud-est actuelle de l'aulacogène (par exemple la zone des îles Simpson, du chenal Hornby et de l'île Union et le sud de la faille McDonald), et les roches des groupes de Union Island et de Sosan à faciès axial directement déposées sur ce soubassement sont allochtones par rapport au craton du Grand lac des Esclaves. Elles reposent tectoniquement sur les roches sédimentaires allochtones situées entre les deux blocs du soubassement. Dans quelle mesure les roches allochtones sédimentaires sont des nappes de charriage immobilisée sous la portion sud-est de la couche allochtone basale, ou encore des portions de cette couche allochtone entraînée par gravité n'est pas clair. La solution de ce problème pourrait permettre de tirer des conclusions paléogéographiques particulièrement importantes.

nappes rare – and this gives additional support to the idea that the Meridian Nappes are derived from the basement allochthon. The Meridian Nappes locally override the trailing edges of the Bunting Nappes, and their complementary stratigraphies and similar structural styles suggest a common origin in the same zone.

In the southwest half of the aulacogen is a single(?) large nappe, the "Basile Nappe", composed exclusively of Wilson Island Group rocks and a small high-level intrusion (Butte Granite). The Basile Nappe appears to override the trailing edges of the Bunting and Meridian nappes, and elsewhere the autochthon. Its metamorphism and much of its internal deformation probably predate its emplacement as a nappe. Because this nappe comprises the bulk of the known Wilson Island Group, the original configuration of the Wilson Island Group and its relation to the Archean basement before thrusting remains obscure.

The southeastern basement allochthon and its indigenous Union Island and Sosan Group outliers appear to override all other structural units. The basement allochthon is believed to have moved on "Inconnu Thrust", named after Inconnu Channel which separates the leading edge of the basement allochthon from the Basile Nappe. A southeast-plunging tectonic lineation is developed in the Wilson Island Group beneath the thrust. The basement allochthon is intruded by diatreme-like breccia dykes that contain blocks of Stark Megabreccia, possibly derived from beneath the allochthon.

The existence of overthrusting, not recognized in reconnaissance mapping, is consistent with the increased crustal thickness beneath the aulacogen known from seismic refraction experiments. A genetically important peculiarity of the nappes is their apparent retrogradational order of emplacement.

The second major deformation affecting the aulacogen is associated with the McDonald Fault Zone, an east-northeast-trending zone of right-lateral strike-slip faulting. Precise fault separations are not easily obtained because the zone nearly parallels the tectonic strike of the earlier thrusts and folds, and the sedimentary axial trend of the aulacogen. Nevertheless, consistent separations of 65-80 km across the main strand, the McDonald-Wilson Fault, of rock units as old as Archean and structures as young as the nappes demonstrates that, contrary to the views of some, virtually all of the strike-slip movement occurred late. The calc-alkaline Compton Laccoliths are displaced by subsidiary strike-slip faults (they do not contact the main strand) and there is abundant structural as well as sedimentological evidence that strike-slip faulting was accompanied by deposition of the Et-then Group, which unconformably overlies the laccoliths. The distribution of laccoliths is not related to major strike-slip faults. Adjacent to and within the strike-slip zone are northeast-trending right-handed en echelon folds of low amplitude and large wavelength. This folding affects the Et-then Group and has also refolded the nappes and the underlying allochthon. That is why the once rather continuous allochthons are now exposed in such a complicated pattern of discontinuous nappes.

In the light of plate tectonic models recently proposed for the Wopmay and Trans-Hudson orogens, a six stage evolutionary model is proposed for Athapuscow Aulacogen. Stage 1 includes the Union Island Group and possibly also the alkaline-peralkaline complexes and Hearne Dykes. This activity is interpreted as an expression of crustal and lithospheric stretching across an east-northeast-trending failed arm connected and more-or-less contemporaneous with initial rifting and breakup along the Coronation continental margin

Les terrains allochtones sont faciles à corréler entre eux sur toute la longueur de l'aulacogène et sont classés en six groupes ou étages, d'après l'intervalle stratigraphique et le faciès sédimentaire concernés, le style structural et la position par rapport aux autres terrains allochtones ou aux nappes à l'intérieur de la colonne stratigraphique. L'étage de base se compose du sommet des schistes argileux du groupe de Kahochella et des calcaires du groupe de Pethei à faciès de pente, qui sont fortement imbriqués et remodelés par de nouveaux plissements. Ce sont des nappes parautochthones, puisqu'elles dominent presque le même faciès de pente que la roche autochtone, ou ont seulement été transportées par-dessus la bordure extérieure de la zone à faciès de plate-forme peu profonde. Les "nappes de Bunting" sus-jacentes sont composées des strates épaisses des groupes de Pethei et de Kahochella à faciès d'axe de bassin, près de la base du groupe de Kahochella. Sur le plan structural, ces nappes de charriage portent encore le nez et parfois aussi l'un ou l'autre des flancs des grands anticlinaux couchés, exposés au nord-ouest, dont les flancs déversés sont fortement ci-saillés. Tout le groupe de Christie Bay est probablement allochtone, ayant latéralement glissée sans dérangement stratigraphique à l'intérieur de la mégabrèche de Stark, qui s'est comportée comme une unité malléable durant la poussée. Le groupe de Christie Bay est conservé surtout devant les nappes de Bunting, et recouvre tectoniquement une partie de la roche autochtone, mais la brèche de Stark englobe localement le front de charriage des nappes parautochtones et des nappes de Bunting et de Meridian, et se présente sous forme de lentilles tectoniques le long des grandes failles limitant les nappes. Les "nappes de Meridian" sont composées des arénites axiales épaisses du groupe de Sosan, très semblable lithologiquement aux klippes conservées sur l'allochtone basale au sud-est. Les nappes de Meridian ne sont bien développées que dans le nord-est de l'aulacogène, où le soubassement allochtone adjacent à l'époque (c.-à-d. avant les décrochements horizontaux) est presque dépourvu de tels témoins. L'inverse est vrai au sud-ouest: les témoins sont étendus et les nappes de charriage sont rares, ce qui donne un poids supplémentaire à l'hypothèse selon laquelle les nappes de charriage de Meridian proviennent de la roche allochtone du soubassement. Les nappes de charriage de Meridian recouvrent localement la bordure arrière des nappes de Bunting, et leurs stratigraphies complémentaires ainsi que leurs styles structuraux semblables laissent croire à une origine commune dans la même zone.

Dans la moitié sud-ouest de l'aulacogène se trouve une seule (?) grande nappe de charriage, la "nappe de Basile", composée exclusivement de roches du groupe de Wilson Island et d'une petite intrusion de niveau élevé (granite de Butte). La nappe de Basile semble recouvrir les bords arrières des nappes de charriage de Bunting et de Meridian, et, ailleurs, le terrain autochtone. Son métamorphisme et une grande partie de ses déformations internes sont probablement antérieurs à sa mise en place. Étant donné que cette nappe comprend la majeure partie de ce qu'on connaît du groupe de Wilson Island, la configuration initiale de ce même groupe et sa situation par rapport au soubassement archéen avant le charriage demeurent obscure.

L'allochtone basale du sud-est et ses klippes indigènes des groupes de Union Island et de Sosan semblent recouvrir toutes les autres unités structurales. On croit que l'allochtone basale s'est déplacé au-dessus du charriage d' "Inconnu", ainsi dénommé d'après le chenal Inconnu qui sépare le front de charriage de l'allochtone basale de la nappe de Basile. Une linéation tectonique plongeant vers le sud-est est présente dans le groupe de Wilson Island au-dessous du charriage. L'allochtone basal a subi l'intrusion de dykes to the west. This correlation is in need of radiometric control. Stage 2 includes the sandstone-dolomite sequence of the lower Sosan Group. This is readily interpreted as an embayment of the sedimentary prism (Epworth Group) developed along the Coronation passive margin as a result of thermal contraction and thickening of the lithosphere following rifting. Stage 3 includes all the rest of the Great Slave Supergroup, and was probably deposited in less than 20 million years. It is interpreted as part of the foreland basin resulting from attempted west-dipping subduction of the Coronation margin during its collision with the Hottah Terrane between 1.92 and 1.89 Ga. Stage 4 is the overthrust deformation of aulacogen, the timing and orientation of which are inexplicable by events in Wopmay Orogen. This deformation may be a distant intracontinental response to collision in the Trans-Hudson Orogen, 650 km to the southeast, analogous to the Neogene thrusting in the Tien Shan of central Asia related to the India-Eurasia collision. Stage 5 includes the calc-alkaline Compton Laccoliths which, like the Great Bear Magmatic Zone, are interpreted as the product of an oblique east-dipping subduction zone initiated by arc-polarity reversal following the Hottah-Coronation collision. This continental arc was active between about 1.88 and 1.86 Ga. Stage 6 includes the major strike-slip faulting and related en echelon folding, and Et-then Group sedimentation. This activity is correlated with the conjugate transcurrent faulting in Wopmay Orogen and left-lateral strike-slip faulting on the Bathurst Fault Zone affecting the Kilohigok Basin. This is believed to result from terminal collision in Wopmay Orogen, probably occurring soon after the cessation of calc-alkaline magmatism at about 1.86 Ga.

In conclusion, Athapuscow Aulacogen is an outstanding example of a multistage intracratonic basin, controlled at virtually every stage by events occurring at nearby continental margins. When compared with Phanerozoic basins, the following very tentative and qualitative differences seem to emerge:

1. Thermally driven subsidence (e.g. passive margin sequence) seems less;
2. Tectonically driven subsidence (e.g. foreland basin) seems more; and
3. Collisional cracking of the subducting plate to permit tholeiitic magmatism in the foreland basin (e.g. Seton Formation and Jackson Gabbro) is more common.

These conclusions are consistent with the idea of a somewhat hotter mantle and consequently thinner lithosphere in the early Proterozoic, and have obvious implications for Archean tectonics.

bréchitiques de type diatrème qui contiennent des blocs de la mégabrèche de Stark, provenant peut-être de dessous la roche allochtone.

Le charriage, qui n'est pas identifié sur les cartes de reconnaissance, est en accord avec l'accroissement de l'épaisseur de la croûte sous l'aulacogène, déterminée par des expériences de réfraction sismique. L'ordre de mise en place apparemment rétrograde des nappes de charriage est une particularité importante de leur genèse.

La deuxième grande déformation qui touche l'aulacogène est associée à la zone de failles de McDonald, zone orientée est-nord-est de failles à décrochement latéral droit. La séparation précise des failles est difficile à déterminer étant donné que la zone est presque parallèle à la direction tectonique des poussées et plis antérieurs, et à l'orientation axiale des sédiments de l'aulacogène. Néanmoins, des décrochements persistants de 65 à 80 km le long de la faille la plus importante, celle de McDonald-Wilson, d'unités rocheuses pouvant dater de l'Archéen et des structures aussi jeunes que les nappes de charriage démontrent que, contrairement à l'opinion de certains, pratiquement tous les décrochements horizontaux ont été tardifs. Les laccolites calco-alcalines de Compton sont déplacées par des failles secondaires à rejet vertical (qui ne touchent pas l'ensemble principal) et il existe d'abondants indices structuraux et sédimentologiques prouvant que le décrochement horizontal a été accompagné du dépôt du groupe Et-then, qui repose en discordance sur les laccolites. La distribution des laccolites n'est pas liée aux grandes failles de décrochement horizontal. À proximité et à l'intérieur de la zone de décrochement horizontal se trouvent des plis orientés nord-est, dextres, en échelon de faible amplitude et de grande longueur d'onde. Ce plissement a touché le groupe d'Et-then et plissé à nouveau les nappes de charriage et la roche autochtone sous-jacente. Voilà pourquoi les terrains allochtones autrefois assez continus se présentent aujourd'hui comme un ensemble complexe de nappes de charriage discontinues.

À la lumière des modèles de tectonique des plaques proposés récemment pour les orogènes de Wopmay et Trans-Hudsonien, on propose un modèle évolutionnaire en six étapes pour l'aulacogène d'Athapuscow. Le stade 1 comprend le groupe d'Union Island et peut être également les complexes alcalins et peralcalins et les dykes de Hearne. On interprète cette activité comme une expression de l'étirement de la croûte et de la lithosphère le long d'un bras est-nord-est avorté, associé aux fractures et au rift longeant la bordure continentale du Couronnement à l'ouest et à peu près contemporain de ces structures. Cette corrélation nécessite une vérification par datation radiométrique. Le deuxième stade comprend la succession à grès et dolomies de la partie inférieure du groupe de Sosan. On l'interprète d'emblée comme une indentation du prisme sédimentaire (groupe d'Epworth) qui s'est développée le long de la marge passive du Couronnement par suite de la contraction thermique et de l'épaississement de la lithosphère après l'apparition du rift. Le troisième stade comprend tout le reste du supergroupe de Great Slave et s'est probablement déposé en moins de 20 Ga. On l'interprète comme une partie du bassin de l'avant-pays résultant d'un début de subduction, avec plongement vers l'ouest, de la bordure du Couronnement pendant la collision de celle-ci avec la zone de Hottah, il y a 1,92 à 1,89 Ga. Le quatrième stade est la déformation par charriage de l'aulacogène, dont la chronologie et l'orientation ne peuvent s'expliquer par les événements de l'orogène de Wopmay. Cette déformation pourrait être une réaction intracontinentale distante à la collision qui s'est produite dans l'orogène Trans-Hudsonien à 600 km au sud-est, comme la poussée datant du Néogène qui a eu lieu dans le Tien Shan au centre de l'Asie, par suite de la collision entre l'Inde et l'Eurasie. Le cinquième stade comprend les laccolites calco-alcalins de Compton qui, comme la zone magmatique du Grand lac de l'Ours, sont interprétées comme le produit d'une zone de subduction oblique inclinée vers l'est, et initiée par une inversion de la polarité de l'arc à la suite de la collision entre la zone d'Hottah et la zone de Coronation. Cet arc continental était actif il y a environ 1,82 à 1,86 Ga. Le stade 6 comprend la formation de grandes failles à décrochement horizontal et les plissements associés en échelon, ainsi que la sédimentation du groupe d'Et-then. Cette activité est corrélée avec la formation des failles conjuguées à décrochement horizontal de l'orogène de Wopmay et de failles à décrochement latéral gauche de la zone de faille de Bathurst, dans le bassin de Kilohigok. Ce phénomène résulterait de la dernière collision affectant l'orogène de Wopmay, probablement survenue peu de temps après la fin du magnétisme calco-alcalin, il y a environ 1,86 Ga.

En conclusion, l'aulacogène d'Athapuscow est un exemple remarquable d'un bassin intracratonique formé en plusieurs étapes, et soumis durant presque chaque étape de son développement à l'influence d'événements affectant les marges continentales voisines. Lorsqu'on le compare aux bassins phanérozoïques, on peut essayer de dégager les différences qualitatives suivantes:

1. La subsidence induite thermiquement (c.-à-d. la succession de la marge passive) semble moindre;
2. La subsidence induite tectoniquement (c.-à-d. le bassin d'avant-pays) semble plus importante;
3. La fracturation par collision de la plaque en subduction, accompagnée d'un magmatisme tholéiitique dans le bassin d'avant-pays (c.-à-d. le volcanisme de Seton et de Pearson) est beaucoup plus fréquent.

Ces conclusions sont conformes à l'idée d'un manteau un peu plus chaud, et par conséquent d'une lithosphère plus mince au début du Protérozoïque, et ont des implications évidentes du point de vue de la tectonique de l'Archéen.

7.

EVOLUTION OF THE EARLY PROTEROZOIC KILOHIGOK BASIN, BATHURST INLET – VICTORIA ISLAND, NORTHWEST TERRITORIES

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Abstract

The Kilohigok Basin is a large intracratonic feature covering more than 7000 km² and hosting up to 7000 m of Goulburn Group strata. The Kilohigok Basin is correlative in age with the Coronation Supergroup of the Wopmay Orogen and Great Slave Supergroup of the Athapuscow Aulacogen. Subsidence in the Kilohigok Basin commenced as an Axial Zone developed as a splay off the Taktu Aulacogen of Wopmay Orogen. Initial shallow marine to nonmarine sedimentation, centred about the Axial Zone, was accompanied by development of extensive stromatolite reef complexes on paleotopographic highs. Following a period of minor(?) uplift and erosion, increased uplift in the source areas to the south and east supplied sands and gravels to extensive braided rivers which spread these clastics down the Axial Zone and across the flanking platforms. Periodic shallow marine deposition on the floodplain was overwhelmed by renewed fluvial sedimentation during this phase of accumulation, which gradually decreased as the basin and source areas stabilized.

Waning uplift in the source areas, coupled with stabilization of the Axial Zone and platforms, resulted in the deposition of shallow marine to deltaic sands. Regional stabilization, and subsequent emergence of the platforms, culminated with formation of regionally extensive pisolitic calcrete paleosoils. Coeval with emergence, continuous subsidence of the Axial Zone was accompanied by deposition of deep- to shallow-water stromatolite complexes adjacent to an intertidal delta at the southern end of the zone.

Extrabasinal-derived, perhaps collision-generated, calcareous mudstone turbidites rapidly buried this entire sequence. These sediments may record the initial clastic influx generated by the first stages of continental collision in the northwestern extension of the Wopmay Orogen. The mudstone basin was in turn buried by a northward-prograding stromatolite reef complex, which supplied some, if not most, of the clastic carbonate to the shoaling basin.

The emergent stromatolite reef complex was in turn covered by southward-prograding supratidal to intertidal evaporitic mudstones. These fine grained clastics were the precursors of the coarsening-upward "molasse stage" clastics, which ended with deposition of coarse fluvial sands and conglomerates as the coarse clastics generated in the collision zone advanced into the basin.

Résumé

Le bassin Kilohigok est un vaste élément intracratonique couvrant plus de 7 000 km² et contenant jusqu'à 7 000 m de couches du groupe de Goulbourn. Ce bassin est aussi ancien que le super-groupe de Coronation de l'orogène de Wopmay et que le super-groupe Great Slave de l'aulacogène d'Athapuscow. La subsidence du bassin Kilohigok a commencé au moment où une zone axiale s'est évasée et s'est dissociée de l'aulacogène de Taktu de l'orogène de Wopmay. Les premiers sédiments marins à non marins, qui se sont accumulés dans la zone axiale, étaient accompagnés par la formation de vastes complexes récifaux de stromatolites sur des crêtes paléontopographiques. Après une période de soulèvement et d'érosion mineures (?), le soulèvement accru dans les zones originales au sud et à l'est a alimenté en sables et graviers les grandes rivières anastomosées qui s'étendaient sur ces roches clastiques le long de la zone axiale et d'un bout à l'autre des plates-formes latérales. Les dépôts marins qui se sont accumulés périodiquement à faible profondeur sur les lits majeurs ont été recouverts par des sédiments fluviatiles renouvelés pendant cette phase d'accumulation, qui a progressivement diminué à mesure que le bassin et les sources se stabilisaient.

Le soulèvement décroissant dans les régions originales, associé avec la stabilisation de la zone axiale et des plates-formes, a entraîné le dépôt de sables marins à deltaiques peu profonds. La stabilisation régionale, et l'apparition ultérieure des plates-formes, ont entraîné la formation de paléosols à croûtes calcaires pisolithiques qui ont recouvert de grandes étendues régionales. En même temps, la subsidence continue de la zone axiale s'est accompagnée de dépôts de complexes de stromatolites en eau profonde à peu profonde à proximité d'un delta intertidal à l'extrémité méridionale de la zone.

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Des argiles calcaires déposées par des courants de turbidité et provenant de l'extérieur du bassin, peut-être dues à une collision, ont rapidement enfoui toute cette séquence. Ces sédiments indiquent peut-être l'apport initial de matières clastiques provoqué par les premières étapes d'une collision continentale dans le prolongement du nord-ouest de l'orogène de Wopmay. Ce bassin d'argiles a, lui aussi, été enfoui par un complexe récifal de stromatolites qui s'avançait vers le nord; le complexe a fourni une partie, sinon la plupart des carbonates clastiques au bassin peu profond.

Ce complexe récifal de stromatolites naissant a, à son tour, été recouvert par des mudstones évaporitiques intertidales. Ces roches clastiques à grain fin sont apparues avant les roches clastiques arrivées au stade de plasticité et dont les grains vont en grossissant vers le haut; cette période s'est terminée avec le dépôt de sables fluviatiles à gros grain et de conglomérats au moment où les roches clastiques à gros grain qui se sont formées dans la zone de collision se sont avancées dans le bassin.

INTRODUCTION

The northwestern Canadian Shield is unique in its preservation of several large correlative early Proterozoic sedimentary basins, all part of an early continental margin in northwestern Canada. These basins preserve large areas of relatively undeformed strata containing spectacular sedimentary structures, diverse microscopic to gigantic stromatolites, olistostromes, and paleosols.

Kilohigok Basin is one of these basins located well within the craton but connected to the continental margin by a paleorift system. The basin covers more than 7000 km^2 and preserves up to 7000 m of strata, representing numerous sedimentary environments.

PREVIOUS WORK

O'Neill (1924) first introduced the term Goulburn to describe 4000 feet of quartzites exposed on Goulburn Peninsula in north-central Bathurst Inlet. Wright (1957) extended the use of the term Goulburn to include two map units in the Beechey Lake area south of the inlet. He also elevated the Goulburn to group status by adding a lower unit of quartzites, slates and limestones to O'Neill's quartzites.

Fraser (1964) produced the first comprehensive map of the Goulburn Group, correlated O'Neill's and Wright's map areas, and added four additional units to the Goulburn Group. Tremblay (1967) formally named four of the map units of Fraser and Wright, and later named another of Fraser's units (Tremblay, 1968).

Fraser and Tremblay (1969) first correlated strata of the Epworth and Goulburn groups. In addition, they also indirectly identified one of the major intra-basinal tectonic features which separates the two regions – the Rockinghorse Arch. This fundamental correlation and their subsequent work provided the basis on which much of the later work in both regions has been based.

Tremblay (1971) produced a detailed account of the stratigraphy and sedimentology of the Western River and Burnside River formations in the Beechey Lake area, and also named the sandstones and conglomerates which unconformably rest on the Goulburn Group, the Ellice and Tinney Cove formations.

Campbell and Cecile (1975a, 1976a, b) further subdivided the Goulburn Group, with the addition of the Quadyuk and Amagok formations. These, and previously-defined formations, were also subdivided into members and submembers (Campbell and Cecile, op. cit.; Cecile and Campbell, 1978). The only major revision to this previous terminology made here is the elevation of the B_R member of the Burnside River Formation to the Mara Formation, and the subdivision of this new formation into two members (see Table 7.1).

Thorsteinsson and Tozer (1962) mapped quartzites in the Wellington High area of Victoria Island as equivalents of the Glenelg Formation of the Shaler Group (see Young, 1981). Subsequent examination of these sediments (Campbell and Cecile, 1979) demonstrated that they are correlative with the Burnside River Formation to the south. Sediments in the Hadley Bay area of northern Victoria Island were interpreted to be Archean in age (Thorsteinsson and Tozer, 1962), but have recently been correlated with the Western River Formation (Campbell, 1981).

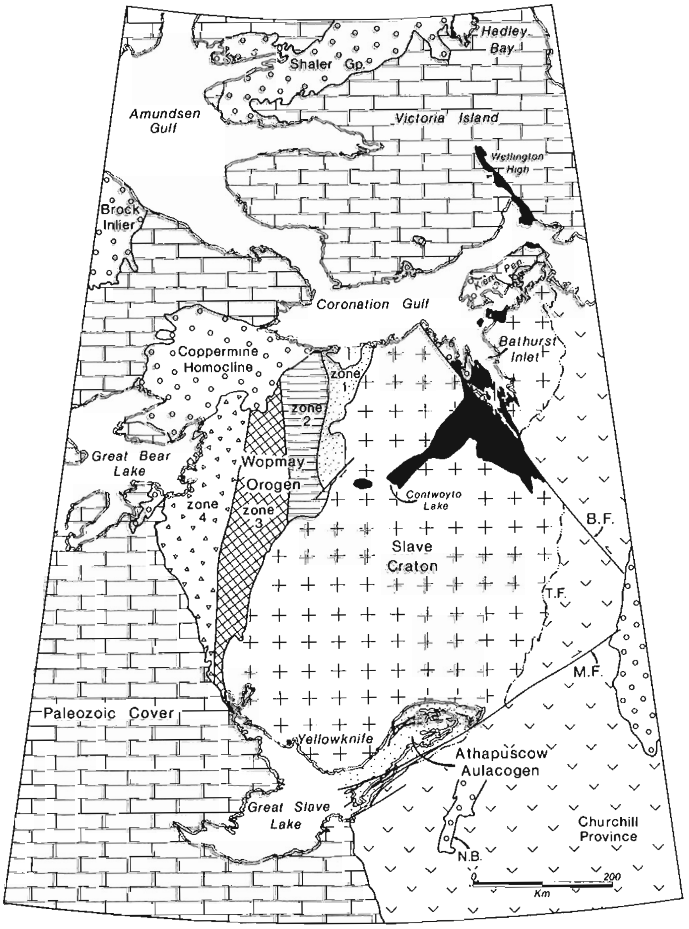


Figure 7.1. Distribution of the Goulbourn Group and correlative Hadley Formation (west side Hadley Bay) shown in solid black; correlative successions in the Wopmay and Athapuscow as shown. Younger Proterozoic sequences shown as open circles. Letter designationas as follows:

B.F. - Bathurst Fault System	*T.F. - Thelon Front*
M.F. - MacDonald-Wilson Fault	*N.B. - Nonacho Basin*

Table 7.1. Table of Formations, Goulburn Group

Group	Formation	Unit	Description
GOULBURN GROUP	AMAGOK FORMATION	A	- white to mauve coarse-grained, moderately indurated lithic and arkosic sandstones; minor conglomerate
	BROWN SOUND FORMATION	$\underline{B}_3C$	- red, well indurated lithic and arkosic sandstones interstratified with white and mauve coarse-grained, moderately indurated sandstones
		$\underline{B}_3B$	- thin vesicular basalt flows interstratified with red sandstones
		$\underline{B}_3A$	- red, medium- to fine-grained, well indurated lithic and arkosic sandstones
		$\underline{B}_2$	- ferruginous, calcareous muddy siltstones
	Omingmaktook Member		- allochthonous sheets of brecciated and chaotically folded carbonates surrounded by carbonate-mudstone breccia
		$\underline{B}_1S$	- buff-brown, medium- to coarse-grained immature sandstone
		$\underline{B}_1$	- ferruginous, calcareous mudstone, salt casts locally abundant near the base of the succession
	KUUVIK FORMATION	K_4	- stromatolitic carbonate, clastic carbonate; abundant edge-wise conglomerate, oncoliths (western equivalent of K_1-K_3)
		K_3	- stromatolitic carbonate, clastic carbonate; abundant intraformational conglomerate, minor mudstone
		K_2	- very thick units of alternating carbonate-rich and mudstone-rich beds
		K_1	- thin-bedded carbonate-mudstone rhythmites (more than 50% carbonate)
	PEACOCK HILLS FORMATION	P_5	- red and green mudstones and siltstones with minor carbonate (western equivalent of P_1-P_3)
		P_4	- thin-bedded mostly red mudstone-carbonate rhythmites with carbonate concretions (eastern equivalent of P_1-P_3)
		P_3	- thin-bedded mostly green carbonate-mudstone rhythmites; minor concretionary mudstone
		P_2	- thin-bedded green, red, and red-brown mudstone rhythmites; massive, thick-bedded siltstones with rare concretions or lenses of carbonate
		P_1	- thin-bedded red and green mudstone rhythmites; minor concretionary mudstone and carbonate beds
	QUADYUK FORMATION	Q	- stromatolitic carbonate, clastic carbonate; minor clacareous quartzite, mudstone, and rare intraformational breccia
	MARA FORMATION	M_P	- pisolitic ferruginous dolomite; granular hematite ironstone; minor ferruginous dolomitic quartzite
		M_S	- red fine grained sandstone and siltstone; minor red quartzite
	BURNSIDE RIVER FORMATION	B	- pink, white, red quartzite and minor subarkose; quartz-pebble conglomerate; intraformational conglomerate; conglomerate; rare shaly or muddy partings
		B_D	- arenaceous dolomite; doloarenite
		B_M	- red mudstone, minor dolomite and stromatolitic dolomite
			Disconformity (?)
	WESTERN RIVER FORMATION	Upper Argillite Member	
		W_5	- grey, buff, and red argillite and mudstone; minor quartzite and subarkose
		Quartzite Member	
		W_4	- white, pink and red quartzite and subarkose; red mudstone and argillite; minor grey-green quartzose turbidites. Stromatolitic and clastic carbonate, doloarenite and dolosiltite; minor pisolitic ferruginous dolomite (Beechey Platform).
		Red Siltstone Member	
		W_3	- red siltstone, mudstone, and argillite; minor clastic and quartose carbonate and quartzite; rare stromatolitic carbonate
		Lower Member	
		W_2	- interbedded siltstone, quartzite, argillite, mudstone; minor thin-bedded quartzose turbidites. Stromatolitic and clastic carbonate, calcareous quartzite and minor quartzite (Kimerot Platform).
		Basal Conglomerate and Regolith Member	
		W_1	- quartzite, quartz-pebble conglomerate, argillite, regolith; minor clastic carbonate
			UNCONFORMITY
		$\underline{A}$	- undifferentiated granitoid, gneissic, metasedimentary and metavolcanic rocks

TECTONIC SETTING

Hoffman et al. (1970) first suggested that the basin which contained the Goulburn Group was related to the basins containing the Great Slave Supergroup and Epworth Group (Fig. 7.1). Subsequently, Hoffman (1973) suggested that the Goulburn Group was deposited in an aulacogen related to the Coronation Geosyncline (since re-named the Wopmay Orogen).

Campbell and Cecile (1975a) introduced the term Kilohigok Basin for the depositional site of the Goulburn Group, and interpreted it as an intracratonic basin intimately related to the development of the Coronation Geosyncline and Athapuscow Aulacogen (Campbell and Cecile, 1975b,; 1976c). These same authors (1980) delineated and defined intra- and interbasinal tectonic elements of the Kilohigok Basin.

The Western and Eastern platforms are areas of relatively thin Goulburn Group, and their contained facies are indicative of platformal depositional environments. The Rockinghorse Arch is a linear-type hinge area, separating Coronation Supergroup rocks from Goulburn Group strata. The Axial Zone is a north-northwest trending linear feature which contains the thickest accumulations of Goulburn Group sediments. Facies of the Western River, Burnside River, and Quadyuk formations effectively delineate the margins of the zone. The Hanimok High is an intrabasinal linear(?) feature which affects only the thickness of the Western River Formation.

Paleocurrent data from the Burnside River Formation suggested that there was a major east-west depositional trough in the Coronation Gulf area during deposition of the Goulburn Group (Campbell and Cecile, 1979). Independent work by Hoffman (1980) suggested that there was a triple point at the western end of Coronation Gulf during formation of Wopmay Orogen, and that the failed arm from this junction trended to the east. Paleocurrent data from the Hadley Formation of Northern Victoria Island supported this interpretation, and the failed arm was named the Taktu Aulacogen[1] (Campbell, 1981).

Tectono-depositional analysis of the Kilohigok Basin and correlative successions in the Wopmay Orogen (Hoffman, 1980), indicate that the Taktu Aulacogen was a major controlling factor during the evolution of these basins.

GENERAL GEOLOGY

The Goulburn Group is an unmetamorphosed succession of sandstones, mudstones, siltstones, and carbonates contained within the Kilohigok Basin, and resting unconformably on Archean gneisses, metasediments, metavolcanics, and plutonic rocks of the Slave Structural Province (Fig. 7.1). In and around the Axial Zone, the group consists of a 5000+ m thick succession of quartzite, conglomerate, stromatolitic carbonate, carbonate, siltstone and mudstone, overlain by 2000+ m of arkosic and lithic sandstone, siltstone, and minor basalt (Table 7.1).

On the Western Platform (Fig. 7.2), the Goulburn Group consists of quartzite, subarkose, stromatolitic carbonate, carbonate, siltstone and mudstone. The younger arkosic and lithic sandstone, siltstone and basalts of the Axial Zone have either been eroded or were not deposited on the Western Platform. At the western extremity of the Goulburn Group, in the Contwoyto Lake area (Fig. 7.1), the group consists of 350+ m of quartzite, subarkose, stromatolitic carbonate, carbonate, siltstone and mudstone. Again, the arkosic and lithic sandstones are absent.

The Goulburn Group has been subdivided into eight formations and numerous members and submembers (Table 7.1), as described briefly below, from the base to the top:

1. Western River Formation: a thick succession of interstratified, varicoloured mudstone, siltstone, quartzite, carbonate, stromatolitic carbonate, pebble conglomerate and minor pisolitic dolomite and/or ironstone.
2. Burnside River Formation: a thick succession of quartzite, conglomerate, subarkose and minor siltstone, shale, mudstone, and arenaceous carbonate.
3. Mara Formation: a thin succession of red siltstone, fine grained quartzite and pisolitic dolomite and granular hematite-ironstone.
4. Quadyuk Formation: a thin succession of stromatolitic carbonate, carbonate, dolomitic quartzite, and minor mudstone.

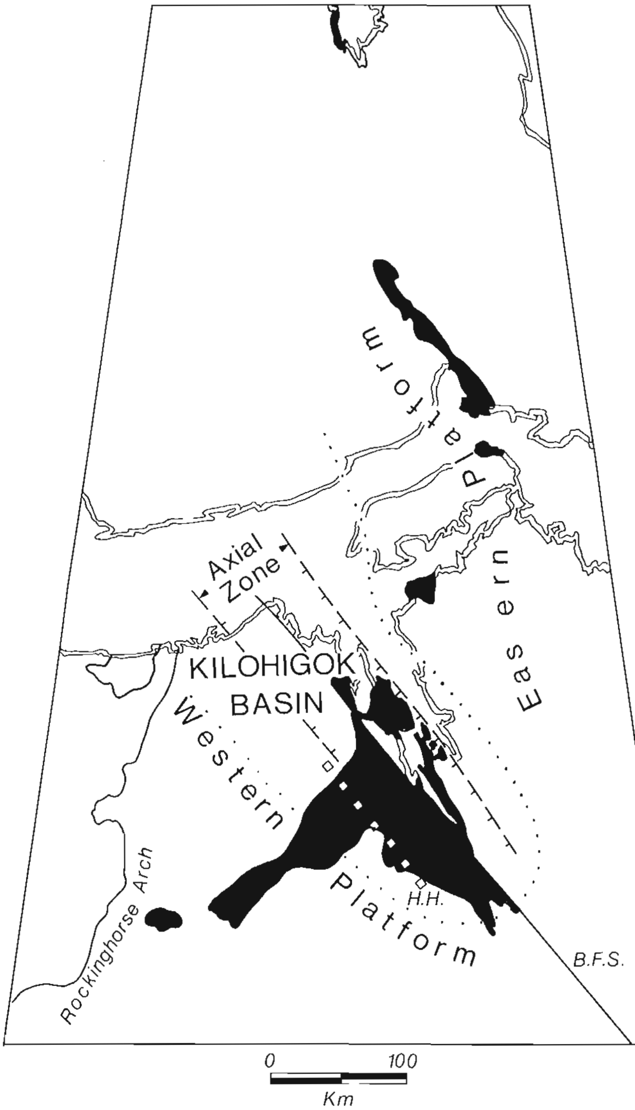


Figure 7.2. Tectonic elements of the Kilohigok Basin, shown in their prefault position. The Goulburn Group and Hadley Formation are solid black. H.H. is the designation for the intrabasinal Hamimok High.

[1] Taktu is the Inuit word for fog, thus an appropriate name for this inferred structure.

5. Peacock Hills Formation: a thin succession of calcareous mudstone, mudstone, siltstone, and detrital carbonate.

6. Kuuvik Formation: a thin succession of carbonate, stromatolitic carbonate, and minor mudstone.

7. Brown Sound Formation: a thick succession of red calcareous mudstones, slump breccia, red siltstone, red arkosic and lithic sandstone and conglomerates, and minor basalt.

8. Amagok Formation: a thick succession of arkosic and lithic sandstones and minor conglomerate.

The Goulburn Group is unconformably overlain by middle Proterozoic polymictic conglomerates, arkose, quartzite and quartzitic conglomerates of the Tinney Cove and Ellice formations in the Bathurst Inlet area, as well as to the northeast on Victoria Island and Kent Peninsula (Fig. 7.1) (Campbell and Cecile, 1976a; Campbell, 1978, 1979, 1981).

The various formations in the Goulburn Group are described in ascending stratigraphic order below. Unless otherwise stated, the formations and other subdivisions of the Group were defined either by Tremblay (1968, 1971), Campbell and Cecile (1975a, 1976a, b), or Cecile and Campbell (1977, 1978). The various tectonic elements referred to in the formation descriptions and discussions are shown in Figure 7.2.

The Western River Formation

The Western River Formation, although highly variable in thickness and contained lithologies, occurs throughout the southern part of the basin, and across the Western Platform. It is present only in the southern part of the Eastern Platform.

The formation has been subdivided into five informal members (Table 7.1), which are described in ascending stratigraphic order below:

Regolith and Basal Conglomerate Member

This member is developed locally at the base of the Goulburn Group at the periphery of the Kilohigok Basin. In the Bathurst Inlet area, it comprises pebble to boulder dolomitic conglomerate, with minor quartz-pebble conglomerate, up to 6 m in total thickness. The conglomerate is commonly transitional both laterally and vertically into coarse grained white quartzite, or rarely beachrock (Donaldson and Ricketts, 1979) in the southern part of the Axial Zone, and into argillite in the remainder of the basin. The regolithic part of the member consists of a thin zone (0.5-3.0 m) of saprolitic-weathered Archean granitoid, gneissic, or rarely Yellowknife Supergroup-type metasediments.

Lower Member

Initially defined by Tremblay (1971) as the Lower Argillite member, this usage was continued by others (Campbell and Cecile, 1975a; 1976a, b). However, due to the diverse lithologies and facies present in the member, the name is here abbreviated to the Lower member of the Western River Formation.

We here subdivide the member into two laterally-equivalent depositional facies, described briefly below:

W_1 Facies: consists of grey and grey-green argillites, with minor fine grained quartzite and thin beds of stromatolitic ferruginous dolomite. It forms the basal part of the Goulburn Group where the underlying member is absent. In the southernmost part of the Goulburn Group east of Bathurst Inlet, the facies consists of greywacke turbidites and thin-bedded clastic carbonate. These facies are laterally transitional into the Kimerot Facies.

Kimerot Facies: occurs predominantly east of Bathurst Inlet, and in a few outcrops in the southernmost Goulburn Group in the Beechey Lake area (Fig. 7.3, 7.4). It comprises a thick succession of stromatolitic and clastic carbonate, quartzite, quartz-pebble conglomerate, and minor thin red siltstones and mudstones. The stromatolitic platform assemblage is symmetrical about the paleotopographically highest intertidal to supratidal zone, and is flanked on both the north and south sides by relatively deep-water troughs fringed by reef complexes (Fig. 7.5).

Red Siltstone Member

The Red Siltstone member is a thick, monotonous succession of red siltstone or mudstone, with several locally-developed, discontinuous, units of grey siltstone, and argillite. Thin, discontinuous stromatolite beds, generally thin but locally thick wedge-shaped quartzite and dolomite beds occur in the uppermost part of the member in the southernmost part of the Axial Zone. Fine lamination, mudcracks, flaser bedding, slump balls, concretions, and ripples occur throughout the member. The concretions are particularly well developed in the westernmost parts of the Western Platform.

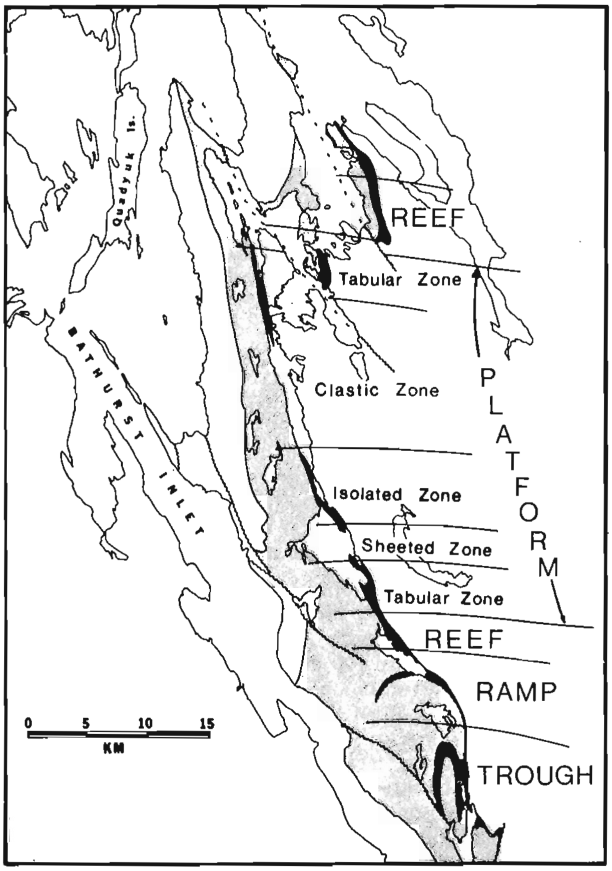


Figure 7.3. Distribution of the facies (capitals) and subfacies (lower case) of the Kimerot Platform, east of Bathurst Inlet. The north-south cross-section shown in Figure 7.4 is reconstructed from these facies.

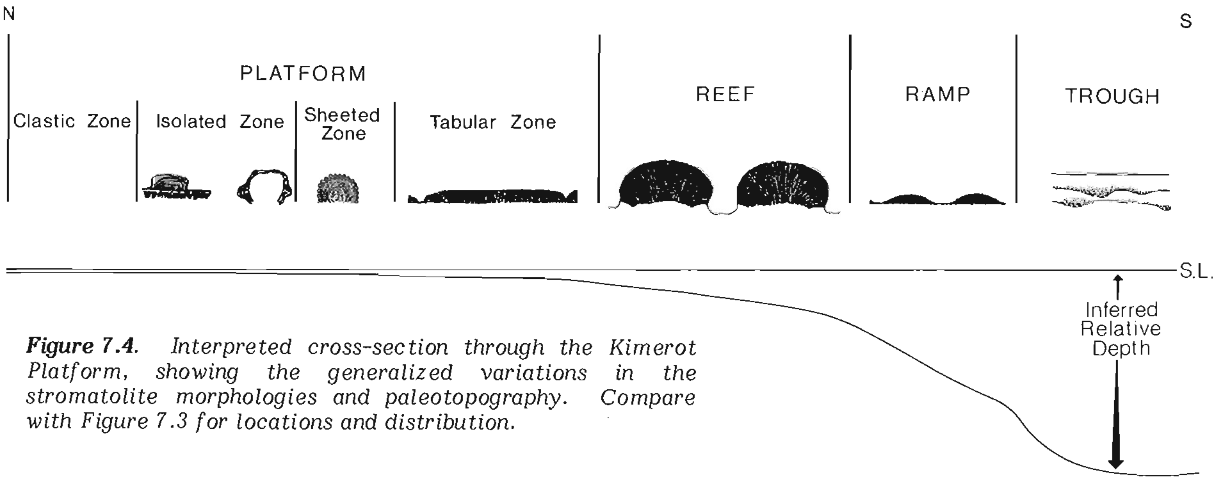


Figure 7.4. *Interpreted cross-section through the Kimerot Platform, showing the generalized variations in the stromatolite morphologies and paleotopography. Compare with Figure 7.3 for locations and distribution.*

Quartzite Member

The Quartzite member is divided into two depositional facies, based primarily on the presence of stromatolitic and/or clastic dolomite. These laterally-equivalent facies occupy well-defined geographic positions within the basin, and are described below.

Quartzite Facies: consists of quartzite, minor quartz-pebble conglomerate and red siltstone, and rare carbonate. The rocks are predominantly massive, with bedding defined by slight variations in colour and/or grain size. Rare trough crossbeds and ripple marks are the only primary sedimentary structures. Along the northern margin of the Goulburn Group, northern distal(?) equivalents are grey-green immature sands with rare sole marks.

Apart from a slight overall decrease in grain size of the sands of this facies, there appears to be little variation in their character from the Axial Zone to the western extremity of the Western Platform. Calcareous quartzites on south-central Victoria Island (Campbell and Cecile, 1979) may be the northern lateral equivalent of this facies.

Beechey Platform Facies: occurs only in the southern part of the Axial Zone, as an east-west trending extensive stromatolitic carbonate belt, containing shallow marine terrigenous clastics. A maximum of four stromatolite sub-units occur within the facies. All thin and pinch out to the west, where the facies is transitional into the Quartzite Facies. The Beechey has been subdivided into two regional subfacies, both of which occur on either side of the Bathurst Fault System, and are correlative across it.

The Reef subfacies, the most northerly of the two, consists of vast, elongate, high-relief bioherms of branching columnar stromatolites with deep inter-bioherm channels. The bioherms are up to 100 m long, and from 2-20 m wide (Fig. 7.6A, B). Intervening channels, commonly with overhanging margins, and filled with calcareous siltstone and fine sandstone, are usually less than 4 m deep. The Reef subfacies is capped by a thin ferruginous pisolitic dolomite and minor red shale interpreted as a paleosol or caliche. In easternmost Goulburn Group exposures, there is a complete transition from the Reef subfacies into fine grained terrigenous clastics; this is not exposed elsewhere in the basin.

The Platform subfacies, which occurs south of the Reef subfacies, comprises laterally and vertically highly variable stromatolite types. There are three interrelated stromatolite groups in the subfacies, namely parallel- and undulatory laminated algal mats, and tabular bioherms of branching columns (Fig. 7.6C). The latter appear to mark the transition zone between the two subfacies. In addition to the clastic and biogenic carbonate, minor red siltstone or shale, white dolomitic quartzite and reddish pisolitic dolomite (caliche) occur throughout the subfacies, but nowhere form the dominant lithology.

Upper Argillite Member

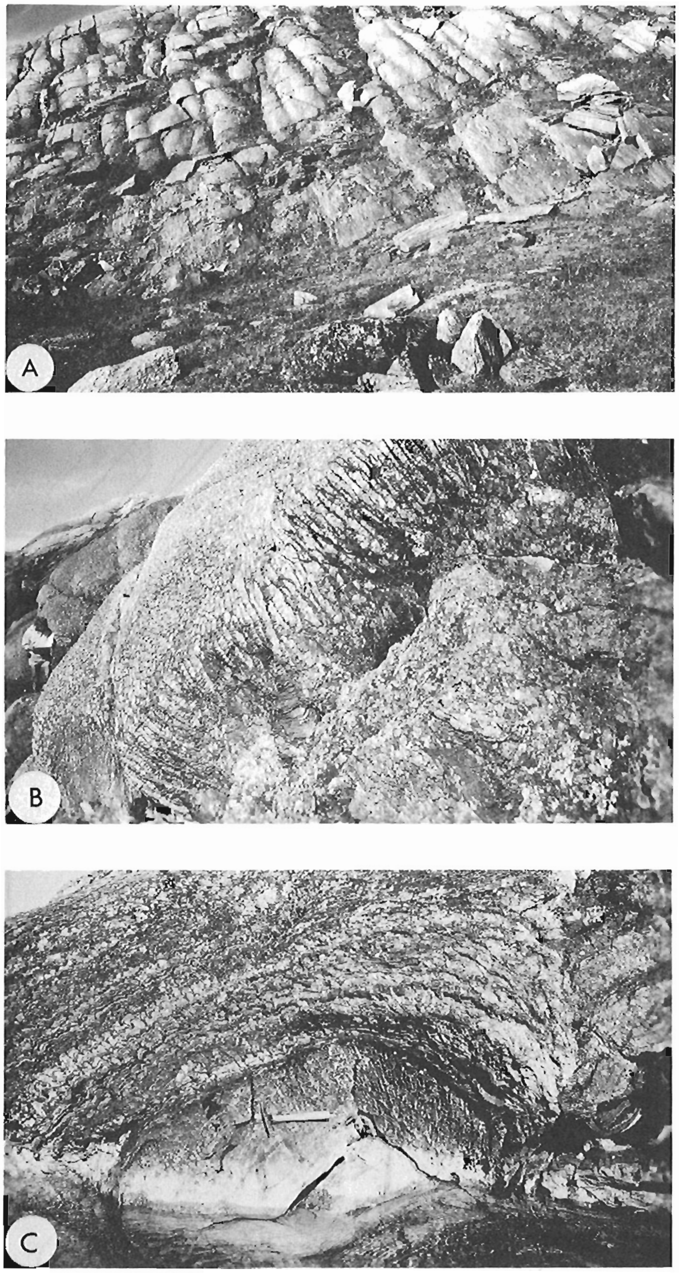
The Upper Argillite member consists predominantly of grey or grey-green, thin-bedded argillite, siltstone, and minor mudstone over most of the basin. In the southernmost Axial Zone, thin grey or white quartzite is present, and thin beds of red siltstone occur locally west of the Bathurst Fault System. East of the fault system, the member is dominated by red to grey lithic and feldspathic sandstones and quartzites, and contains only minor grey-green siltstones and argillite. In this area east of Bathurst Inlet, there appears to be a complete gradation from the underlying Quartzite member into the overlying Burnside River Formation. The Upper Argillite member is missing near the western margin of the Axial Zone, in the vicinity of the Hanimok High (Fig. 7.2), where the Quartzite member and the Burnside River Formation are in conformable contact.

The argillites, siltstones, and mudstones of the Upper Argillite member display rare flutes, grooves and graded bedding; on the whole they resemble distal turbidites. The relatively coarser sediments on the easternmost part of the member are thicker bedded, rarely crossbedded and rippled; parallel lamination is locally common.

Paleocurrents of the Western River Formation

Paleocurrent indicators are not so abundant in this formation as in others of the Goulburn Group, and are insufficient to demonstrate regional transport trends. However, trough and planar crossbeds, primarily from the Lower and Quartzite members, show dispersal patterns from the southeast to the northwest and west (Fig. 7.7A). Current and wave ripples, primarily from the Axial Zone of the basin, indicate this same dispersal pattern (Fig. 7.7B).

Stromatolites have been interpreted to develop elongations parallel to the predominant direction of wave and current attack (Hoffman, 1969, 1974; Donaldson, 1976; Gebelein, 1969). Kimerot Platform stromatolites are all elongate approximately north-to-south (Fig. 7.8A), but this is apparently unrelated to the tectonic evolution of the basin.

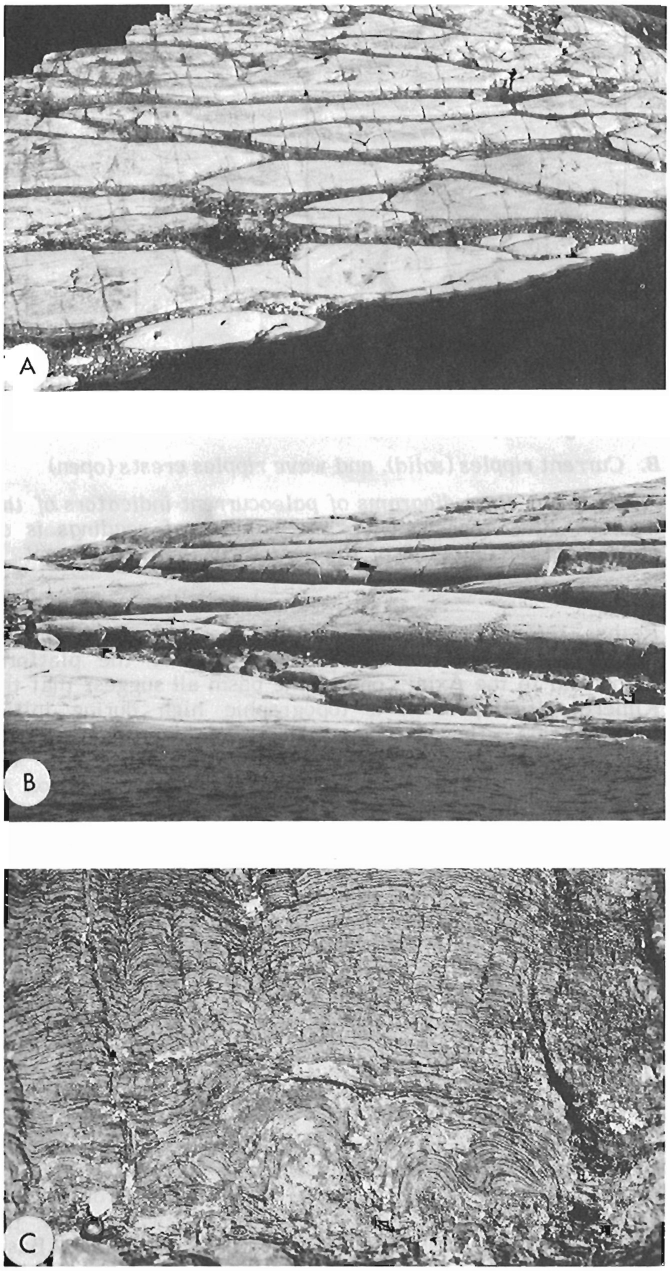


A. *Ramp facies elongate stromatolite bioherms. The person for scale is at the left margin. GSC 202670-F*

B. *Reef facies elongate stromatolite bioherms. GSC 202667-0*

C. *Isolated mound from the Platform facies. These locally have well-developed molar tooth structure in their cores. GSC 202670-C*

Figure 7.5

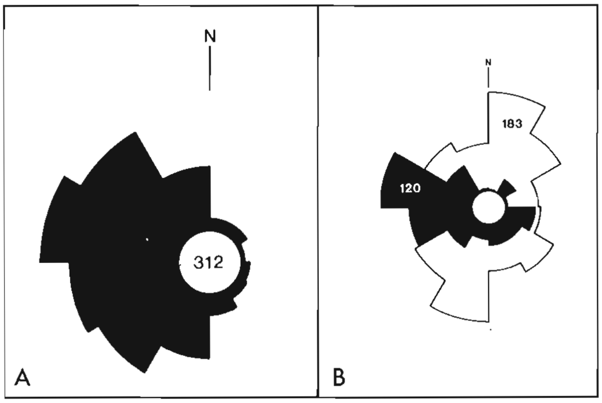


A. *Aerial view of reef-mounds from the Reef facies, Beechey Platform. The largest mound is approximately 100 m. Compare with Figure 7.6B, taken from ground level of the same exposure. GSC 203059-H*

B. *Reefs facies, Beechey Platform. The person for scale is at the left margin of the photograph. GSC 203059-J*

C. *Undulatory stromatolite mat from the Platform facies of the Beechey Platform. Pseudo-columnar stromatolites are at the base. GSC 202671*

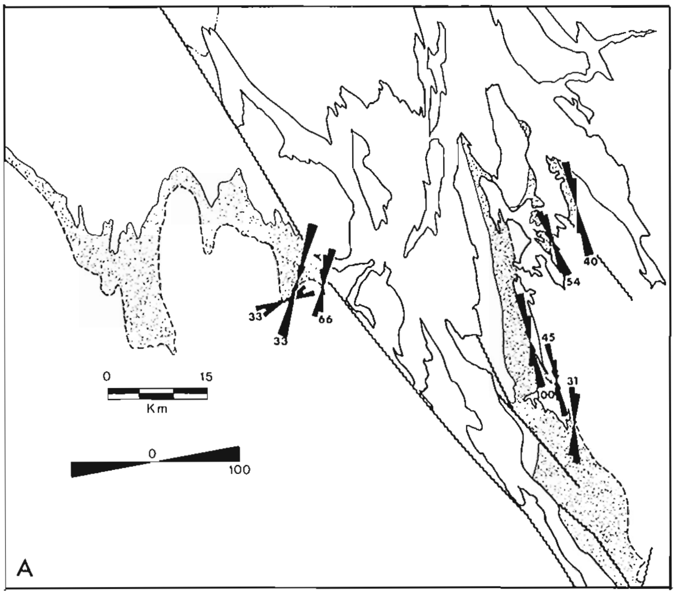
Figure 7.6. *Western River Formation.*



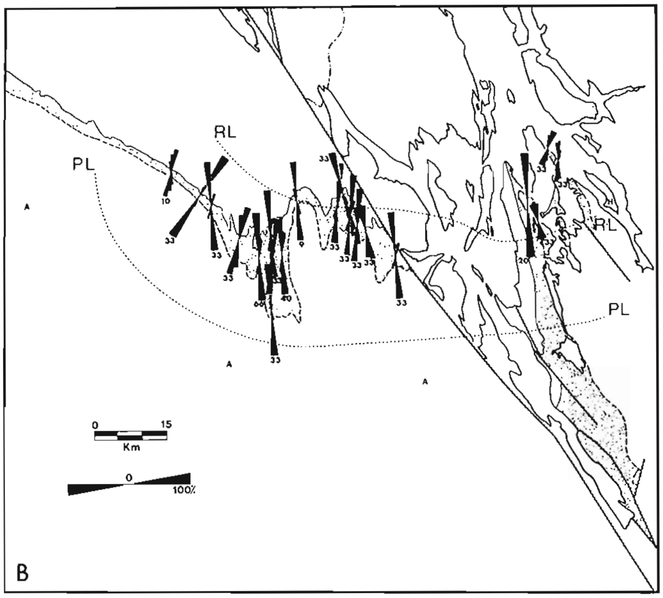
A. Trough and planar crossbeds

B. Current ripples (solid), and wave ripples crests (open)

Figure 7.7. *Rose diagrams of paleocurrent indicators of the Western River Formation. The number of readings is as shown, and the diameter of the centre circle is 20 per cent.*



A. Rose diagrams of stromatolite elongations from the Kimerot Platform. The basin is shown in its restored position.



B. Rose diagrams of stromatolite elongations from the Beechey Platform. The basin is shown in its restored position. MS is the designation for the interpreted southern limit of the Reef facies, and PL for the Platform facies.

Figure 7.8

Rather, the variation in stromatolite morphologies (Fig. 7.4), lateral facies variations, and the trend of the platform subnormal to the Axial Zone of the basin all suggest that the Kimerot developed on a topographic high during initial transgression. The absence of debris breccias, together with the gradual north-to-south change in relief of the stromatolites, suggests that the platform developed on an area of low slope (Playford et al., 1976; Hoffman, 1974).

Beechey Platform stromatolites are also elongate approximately north-to-south (Fig. 7.8B). However, in contrast to those of the Kimerot, they are restricted to the southern end of the Axial Zone, and show a marked south-to-north transition from supratidal/intertidal to subtidal facies (Fig. 7.8B). The distribution of the Beechey Platform, and relative position of its facies is the first indication that the Axial Zone of the basin was behaving as an active tectonic element.

Discussion

Deposition in the Kilohigok Basin commenced with the initial onlap of the Archean basement complex during the subsiding passive margin stage of the Wopmay Orogen. The developing Taktu Aulacogen spread eastward into the continental margin, and the Axial Zone of the Kilohigok Basin developed off it as a southeast-trending splay.

During initial transgression, an early Proterozoic regolith was stripped from high-relief areas, and buried in regions which were covered relatively rapidly by Western River strata. Terrigenous clastics of the Lower member initially accumulated in pre-existing valleys, while the intervening topographic highs were the sites of extensive stromatolite platforms. Thin pisolitic paleosols formed during emergent periods of Lower member deposition on the western edge of the proto-Axial Zone. Ferruginous stromatolites accumulated in ephemeral lagoons in the same region during periods of minimal clastic sedimentation. As the pre-Goulburn Group topographic depressions were filled, and Axial Zone subsidence decreased, the sediments of the Lower member spread across the Western Platform.

With increasing sediment supply relative to subsidence of the basin, the prograding shallow marine sediments of the Red Siltstone member buried the Lower member rocks.

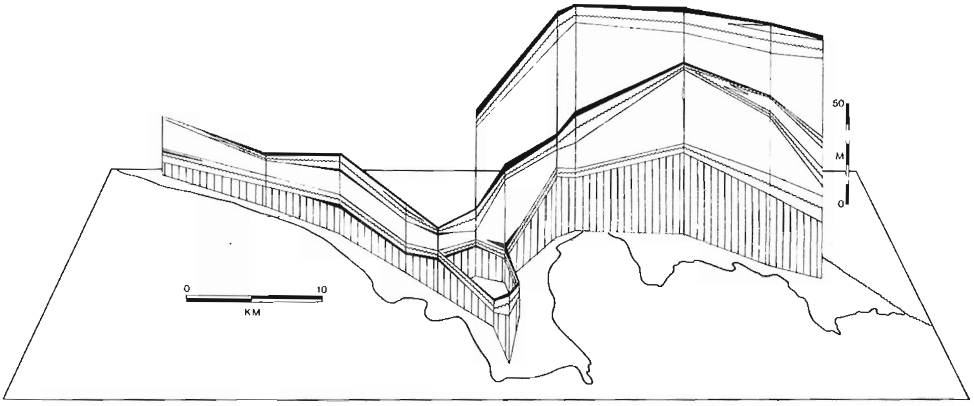


Figure 7.9

Fence diagram, looking north, of measured sections through the Beechey Platform, in the southernmost part of the Axial Zone, adjacent to the Bathurst Fault. The base of the Western River Formation is shown as a solid line. Paleosols are solid black in the sections, and the stromatolitic segments are patterned. The unpatterned area south of the Western River is Archean.

Sedimentary structures suggest that this member was deposited in a muddy, tidal-flat dominated regime, with the grey siltstones deposited during periodic marine incursions (Reif and Slatt, 1979). The interstratified, relatively coarse clastics in the southern part of the Axial Zone suggest that this was probably the distributary locus for the remainder of the sediments of the member.

Lithologies, sedimentary structures, stromatolites, and intimate association with the Red Siltstone member suggest that a shallow marine depositional environment was continuously maintained throughout deposition of the Quartzite member. Subsidence of the Axial Zone was continuously greater than that of the flanking Western Platform, resulting in the east-to-west thinning characteristic of the entire formation. The Beechey Platform accumulated during periodic minor transgressions in the southern Axial Zone. The Reef subfacies formed the seaward-facing frontal barrier, while the Platform subfacies accumulated in the protected "back-reef" region to the south. Platform subfacies development was intermittently curtailed during emergent periods, which culminated with paleosol formation and possible erosion as the Axial Zone filled with biogenic and terrigenous sediments. This repeated transgression-filling-emergence sequence produced the characteristic repetitive alternation in the Platform Subfacies (Fig. 7.9). With relative stability in the Axial Zone, and increased supply of terrigenous clastics, the transgressive-filling cycles ended, and the quartz sands spread down the Axial Zone and across the Western Platform.

Lithologies, sedimentary structures, and complete lack of shallow-water features suggest that the Upper Argillite member accumulated in a relatively deep-water environment with little traction current activity. The member was deposited after a rapid transgression which buried the shelf quartzites beneath their possible distal equivalents. The coarser clastics in the member in the southern Axial Zone suggest that this region was continuously maintained as the primary distributary locus.

The Burnside River Formation

The Burnside River Formation consists of a thick (2150 m maximum) sequence of pink, white, and red quartzite, quartz-pebble conglomerate, conglomerate, and minor dolomite, mudstone, and shale. The formation is the most extensive in the Goulburn Group, and outcrops throughout the basin from central Victoria Island in the north to the outlier west of Contwoyto Lake in the west (Fig. 7.1). The formation as redefined here (see Table 7.1) includes only those coarse grained sediments which overlie either the

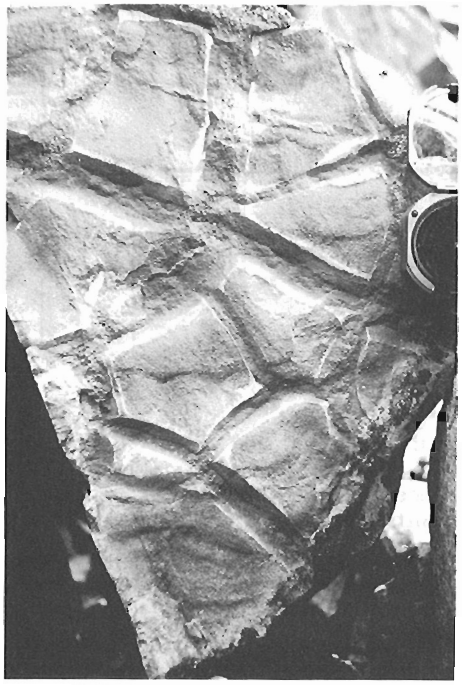
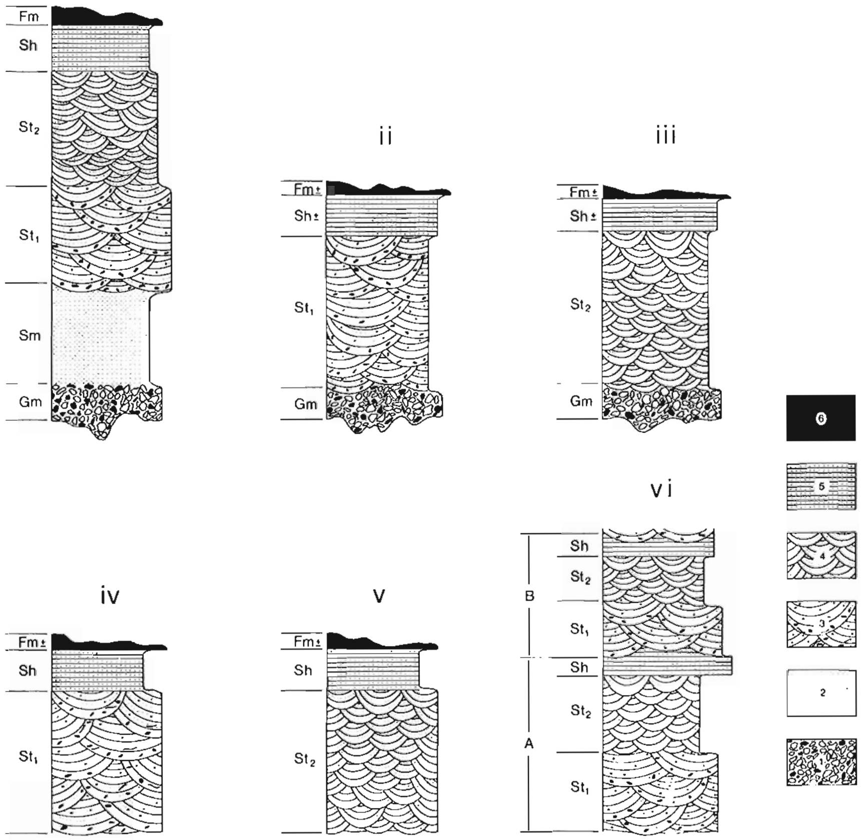


***Figure 7.10**. Dessication cracks in Burnside River Formation fine grained quartzites, in the southern part of the basin. GSC 202668-W*

Western River Formation or the Archean basement (in the northeast). The fine grained sediments previously mapped as the B_R member are here assigned to the Mara Formation (see below).

The Burnside River Formation apparently conformably overlies the Upper Argillite member of the Western River Formation at almost all localities in the southern part of the basin and on the Western Platform. In one area, near the western margin of the Axial Zone, the Burnside River rests directly on the Quartzite member of the Western River Formation. The significance of this relationship will be discussed further below. On the central and northern parts of the Eastern Platform, the Burnside River rests unconformably on Archean basement rocks. Although the base of the formation is not exposed on southern Victoria Island, possible Western River Formation equivalents may



LEGEND

1. *conglomerate*
2. *massive, structureless sands*
3. *pebbly trough crossbedded sands*
4. *trough crossbedded sands*
5. *parallel-bedded or laminated sands*
6. *mud, silt, or rare shale*

Figure 7.11. *Typical fining-upward cycles from the Burnside River Formation. (i) is the ideal, complete cycle, and the remainder are typical variations from within the basin. The divisions of the fining-upward cycles are after Miall (1977, 1978).*

occur beneath the Burnside River (see Campbell and Cecile, 1979). The top of the formation is defined as the first appearance of greater than 50 per cent red fine grained sandstone or siltstone of the overlying Mara Formation.

Two members are defined within the coarse quartzose clastics of the Burnside River Formation. Both members occur only in the western part of the basin and on the Western Platform.

The B_D Member

The B_D member consists of up to 35 m of laminated to thick bedded doloarenite, quartzose doloarenite, and minor dolomitic quartzite. Other than ripple marks, poorly-developed planar crossbedding is the sole primary sedimentary structure.

The B_M Member

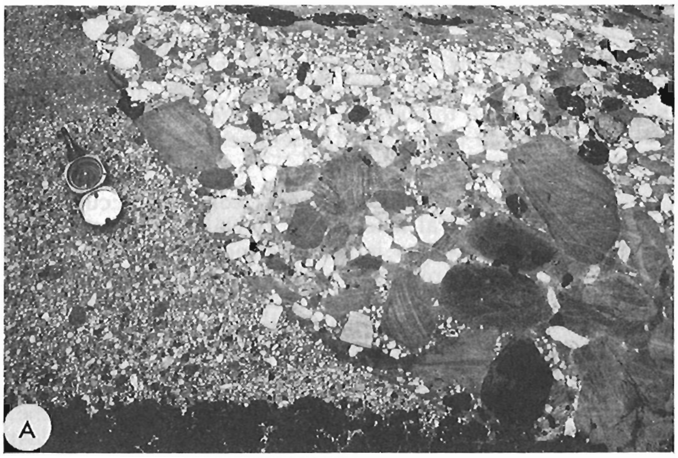
The B_M member consists of laminated mudstone, shale, fine grained red sandstone, minor fine grained doloarenite and stromatolitic dolomite, and, at the top, dolomitic quartz-pebble conglomerate. The member, only approximately 15 m thick, occurs only in the southern part of the Axial Zone, and pinches out in both directions along strike (Campbell and Cecile, 1976b). The rare, isolated, biscuit-type stromatolites occur only in the uppermost part of the member. Desiccation cracks (Fig. 7.10) in both fine and coarse clastics are common throughout the member; poorly-developed planar crossbeds occur in the uppermost pebbly part of the unit.

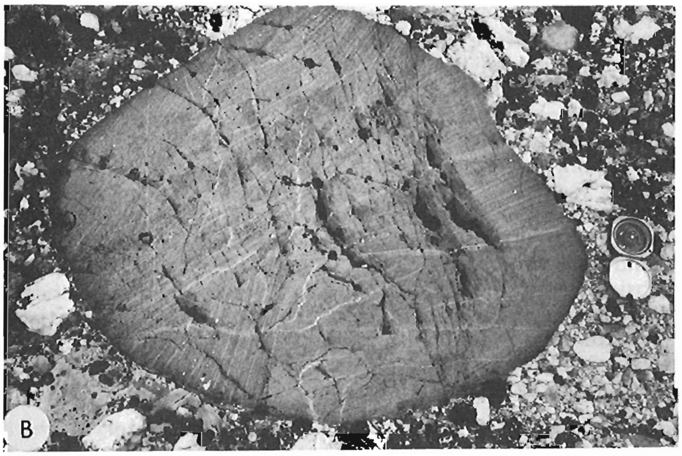
Siliceous Clastics of the Burnside River Formation

Approximately 90 per cent of the Burnside River consists of variously coloured quartzite, pebble and boulder conglomerate, and minor shale and/or mudstone. The quartzites are ubiquitously trough crossbedded, and typically form segments of fining-upward cycles (0.5-2.5 m thick). The typical facies of these fining-upward cylces (see Fig. 7.11), following the terminology of Miall (1977; 1978) and Rust (1978), are:

Gm Facies: occurs at the base and consists of usually structureless orthoconglomerate, with a coarse sand or grit matrix, and rare thin lenses of coarse grained quartzite. The Gm Facies commonly incises the upper 0.5-2.0 m of the underlying unit (Fig. 7.12A), and the facies is best developed in the southern part of the Axial Zone of the Basin. The large channels commonly contain white vein quartz pebbles 10-20 cm in diameter (maximum 45 cm), and large intraformational quartzite clasts (Fig. 7.12B). The upper contact of the Gm Facies is commonly gradational into one of the two overlying facies of the cycle.

Sm Facies: consists of massive, coarse grained, near-structureless sandstone, with only a faint internal lamination and small isolated quartz pebbles. The contact with the underlying Gm Facies is marked by an upward increase in the amount of coarse sand, and a sympathetic decrease in both the number and the size of the boulders. The Sm Facies is rare, and occurs only in the southern and eastern parts of the Axial Zone. Its thickness is variable (up to 2.0 m), and this appears directly related to the thickness of the underlying, basal, facies.





St Facies: is the most common, and consists of trough crossbedded, locally pebbly quartzite and grit. Where the St Facies rests on the Gm, the contact is usually abrupt, with the basal St sediments containing only small scattered pebbles. The St Facies is here subdivided into a lower (St_1) and an upper (St_2) segment. St_1 consists of large-scale (0.4-1.7 m) trough crossbedded, commonly pebbly quartzite, which passes upward into crossbedded, less pebbly quartzite (Fig. 7.12C). The base of the facies is transitional into either the Sm, or rests abruptly on the Gm. Where the St_1 forms the basal division of the fining-upward cycle, the lowermost troughs commonly have a basal pebble lag which delineates the base of the cycle. The top of the St_1 is transitional into the overlying St_2, with the transition marked by a decrease in the size of the troughs, percentage of pebbles, and grain size of the sands.

The St_2 consists of smaller-scale (0.3-1.2 m) trough crossbedded, generally medium grained quartzite. The sands usually lack pebbles, except in the southernmost part of the formation. The St_2 typically forms the basal division of the fining-upward cycle in the western, central, and some northern parts of the formation on the Western Platform. The transition from the top of the St_2 into the overlying Sh Facies usually occurs over less than 20 cm.

Sh Facies: consists predominantly of reddish to purplish, fine grained, parallel-bedded or laminated quartzite, with rare planar crossbedding and locally well-developed ripple marks. Where preserved, the contact with the overlying Fm Facies is sharp, abrupt, and parallel to the internal lamination in the Sh sands.

Fm Facies: is characterized by deep red to purplish red, 1-5 cm mudstones and shales which form the topmost veneer or "skin" of the cycle. Most commonly, though, the only evidence of the facies are the scattered mud chips and flakes in the basal parts of the overlying cycle. Mudcracks and small-scale ripples are common where the upper surface of the facies is preserved.

Conglomerates

Conglomerate dominated successions, unrelated to the Gm Facies of the fining-upward cycles, are best developed in the southern part of the Axial Zone and the southern part of Victoria Island, on the Eastern Platform. In these areas, framework and clast-supported conglomerates up to 1.5 m thick, with parallel upper and lower surfaces, are interstratified with coarse grained quartzites and grits. Most pebbles and boulders in the conglomerates are white vein quartz and intraformational quartzite (together commonly over 80%). The remainder of the clasts are quartzite of unknown origin, and granitoid of gneissic rocks. Quartz clasts are generally 20-30 cm, while the intraformational quartzite boulders are somewhat larger (30-80 cm).

A. *Conglomerate-filled (Gm facies) channel at the base of a fining-upward cycle in the Burnside River Formation, from the southern part of the Axial Zone. GSC 202667-W*

B. *Vein quartz boulders (white), and large, trough crossbedded intraformational quartzite megaboulders filling a channel at the base of the fining-upward cycle in the Burnside River Formation, southern Axial Zone. GSC 202667-J*

C. *Large-scale trough crossbeds of the St_1 subfacies of the Burnside River fining-upward cycles. The scale is at the approximate centre of the photograph. GSC 202917-C*

Figure 7.12

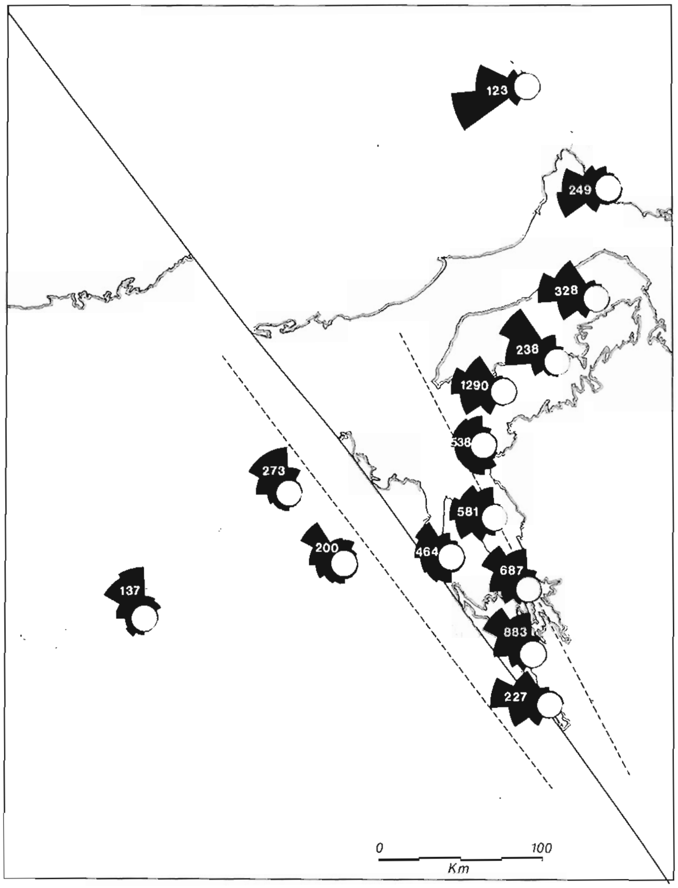


Figure 7.13. *Paleocurrent roses of trough crossbed axes from the Burnside River Formation. The data have been grouped from several localities in the vicinity of each rose, for clarity. The number of readings is as shown, and the diameter of the centre circle is 10 per cent.*

Paleocurrents of the Burnside River Formation

Ubiquitous trough crossbeds in the St Facies show almost unimodal paleocurrent patterns, generally to the northwest, over wide areas and thick sequences of the formation (Fig. 7.13). There is, however, a significant variation in the paleocurrent trends along the Eastern Platform of the basin (from south to north), showing a well-defined shift in the orientation form north-northwest to westerly and then a southwesterly direction.

Current and wave ripples, while not so abundant as trough crossbeds, show the same overall dispersal pattern (Fig. 7.14). However, insufficient ripples were recorded to reconstruct regional paleocurrent maps.

Depositional Environment of the Burnside River Formation

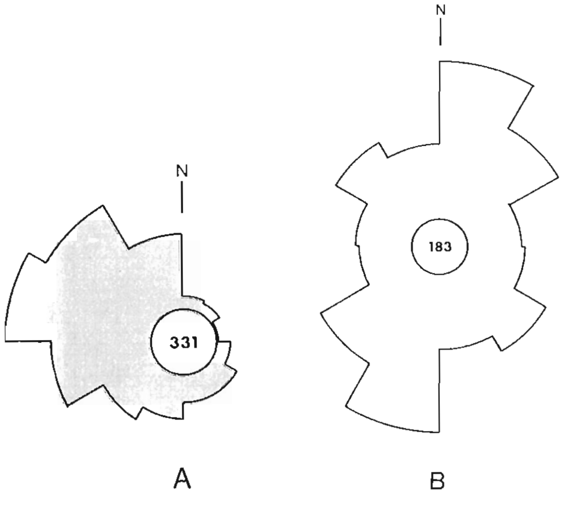
Fining-upward cycles in coarse clastic successions, which are commonly devoid of fine grained detritus, have been interpreted elsewhere as the products of fluvial depositonal systems (see Rust, 1978; Cant, 1978; Miall, 1978; Long, 1978a; Morey, 1974; Smith, 1971). Taken together, the unimodal paleocurrent patterns, characteristic uneven bedding surfaces, scour and fill structures, absence of fine grained detritus, and rapid lateral and vertical grain size variations all suggest rapidly changing variable flow conditions, characteristic of a braided river system.

The B_D member of the Burnside River Formation was deposited during a minor marine incursion, possibly associated with fluvial avulsion, as suggested by Allen (1974) and Friend and Moody-Stewart (1970). The member was probably deposited throughout the southern part of the basin, but with renewed coarse clastic sedimentation, the member was eroded from the Axial Zone.

The B_M (or mudstone) member sediments accumulated in a shallow ephemeral lake on the fluvial distributary plain. Stromatolitic carbonates initially formed at the rim of the lake, and gradually prograded across the filling depression to form the resistant carbonate-silica cemented "cap" which protected the underlying fine grained clastics from later fluvial erosion (see Clemmenson, 1978; Wheeler and Textoris, 1978).

The intraformational quartzite boulders, with their contained quartz veinlets and associated quartz clasts (Fig. 7.12B), indicate that cementation was coeval with sedimentation throughout deposition of the Burnside River Formation. Cementation and/or silcrete formation, elsewhere interpreted as a consequence of subaerial exposure in a semiarid environment (Williamson, 1957; Selleck, 1978; Smale, 1973) is consistent with deposition by braided rivers across an extensive floodplain. Sequential repeated periodic fluvial avulsion would have exposed large tracts of the floodplain, with consequent precipitation of silica in the porous sands. This may have been accompanied by dewatering and desiccation, which could have produced the deep, relatively narrow, fractures in which the "vein" quartz could have been precipitated. Rapid, possibly fault-initiated, subsidence of the Axial Zone of the basin, with attendant fracturing of the contained sediments and essentially instantaneous fluvial capture, would have produced the large "mega-boulders" of intraformational quartzite and quartz to fill the rapidly downcutting channels.

In summary, then, following an erosional interval which locally removed some of the Western Formation from within the basin, and probably stripped it completely from the



A. *Rose diagram of current ripples from the Burnside River Formation. The number of readings is as shown, and the diameter of the centre circle is 10 per cent.*

B. *Rose diagram of wave ripple crest orientations from the Burnside River Formation. The number of readings is as shown and the diameter of the centre circle is 10 per cent.*

Figure 7.14

Eastern Platform, the braided rivers carried coarse sands and gravels from rising source areas to the south and east. As the Taktu Aulacogen progressively subsided more in the east, causing a southwestward swing in dispersal (Fig. 7.13), the rivers carried their detritus across the slowly subsiding Eastern Platform. With continued subsidence, a thin, relatively proximal, fluvial succession was deposited directly on the denuded Archean basement complex.

The Axial Zone of the basin, first weakly developed during Western River sedimentation, was a major tectono-depositional feature during deposition of the Burnside River Formation. Coincident with elevation of the source areas and subsidence of the Taktu, the Axial Zone began a period of continuous, though intermittent, subsidence. Initially, the distributing rivers were channeled into the zone, but during stable periods the excess clastics spread across the Western Platform. The repetitive alternations of subsidence-filling-spillover produced the thick sequence in the Axial Zone, and the relatively thin sequence on the Western Platform.

Prolonged periods of nondeposition in the Axial Zone, and possibly on the Eastern Platform as well, produced the characteristic cemented intraformational quartzites of the Burnside River Formation conglomerates. Equivalent fine grained clastics, generated in and bypassed through the braided floodplain, accumulated downslope in the Taktu as the Mara Formation fine sands and silts. With waning uplift in the source areas, progressively finer grained sediments were supplied to the basin, as Burnside River deposition ended.

The Mara Formation

The Mara Formation was initially mapped as the B_R member of the Burnside River Formation (Campbell and Cecile, 1975a). However, its thickness, lateral extent, and distinct lithologies merit formational rank.

The formation consists of red siltstone, fine grained reddish or purplish sandstone, mudstone, grey siltstone, and ferruginous pisolitic dolomite and/or granular hematite-ironstone at the top.

The formation occurs throughout the southern part of the basin and across the Western Platform, but not on Victoria Island or on the southern part of the Eastern Platform. The type area of the formation is near the junction of the Mara and Burnside Rivers, where it is approximately 200 m thick. The base of the Mara Formation is transitional into the underlying Burnside River Formation, and the contact is marked by upward-increasing amounts of fine grained red sandstone or siltstone at the expense of the coarse and medium grained quartzites below. The top of the formation is defined as the first appearance of either the stromatolitic carbonate (continuous upward) of the Quadyuk Formation, or calcareous mudstones of the Peacock Hills Formation.

The Mara Formation has been subdivided into two members, a lower Siltstone member (M_S) and an upper, Pisolite member (M_P).

Siltstone Member (M_S)

The siltstone member consists of thin-bedded to parallel-laminated units of red siltstone or fine grained quartzite, locally capped by veneers of shale or mudstone. In the southern part of the Axial Zone, Sh-Fm fining-upward cycles (terminology as above) with abundant mudcracks and laddered ripples predominate. Rare, discontinuous thin stromatolite units (30-40 cm) locally occur intimately interstratified with the red siltstones near the top of the member in the south-central part of the Axial Zone. Beyond the southern Axial Zone, the member is mostly Sh Facies sands, with minor Fm Facies muds or silts. Thin (20-70 cm) fining-upward cycles are locally present throughout the member, but are difficult to discern due to lack of colour contrast and narrow range.

Pisolite Member (Mp)

The Pisolite member of the Mara Formation consists of thin to medium beds of carbonate pisoliths in a matrix of fine grained hematitic quartzite. Minor thin beds of granular hematite ironstone occur in the eastern part of the member. The pisolitic beds typically weather a rusty brown; the ironstone beds a deep purple.

The member caps the Mara Formation in the southern and western parts of the basin, but is absent in the south-central part of the Axial Zone, where the M_S member is conformably overlain by the Quadyuk Formation. The member is nowhere greater than 12 m thick, but forms a continuous, very distinct, unit throughout its entire extent. Its base is defined as the first appearance of dolomite pisoliths or ironstone in thin beds. The top is defined by the first appearance of the stromatolitic carbonates of the Quadyuk Formation or calcareous mudstones of the Peacock Hills Formation. The contact with the Peacock Hills is always abrupt, but thin beds of red siltstone or dolomite may occur beneath the Quadyuk Formation.

Throughout the Western Platform, the pisolitic carbonate occurs in well-defined beds, locally intercalated with thinner units of fine grained hematitic quartzite. The pisoliths range in size from 0.25 to 4.0 cm, are mixed, and show no evidence of sorting. Rarely, a crude inverse grading is present in some beds (Fig. 7.15). Locally, individual pisoliths are incorporated into the overlying sandstone beds. Although partially to completely recrystallized, the pisoliths sometimes show an original concentric lamination. Pisolith cores are generally a small fragment of quartz, or rarely, hematite. The pisoliths are little broken, and are rarely amalgamated to form larger, composite, pisoliths.

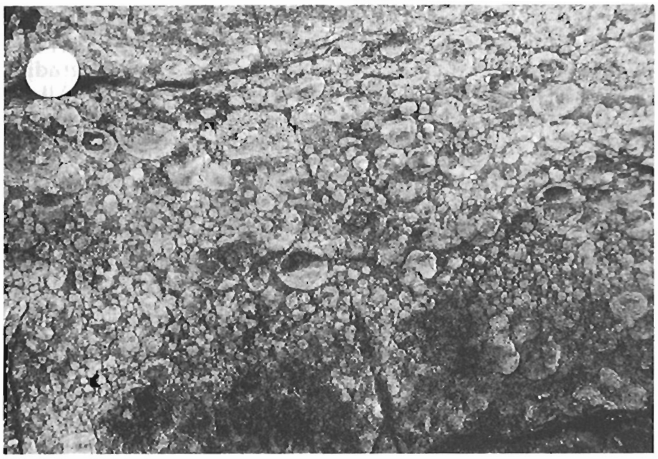
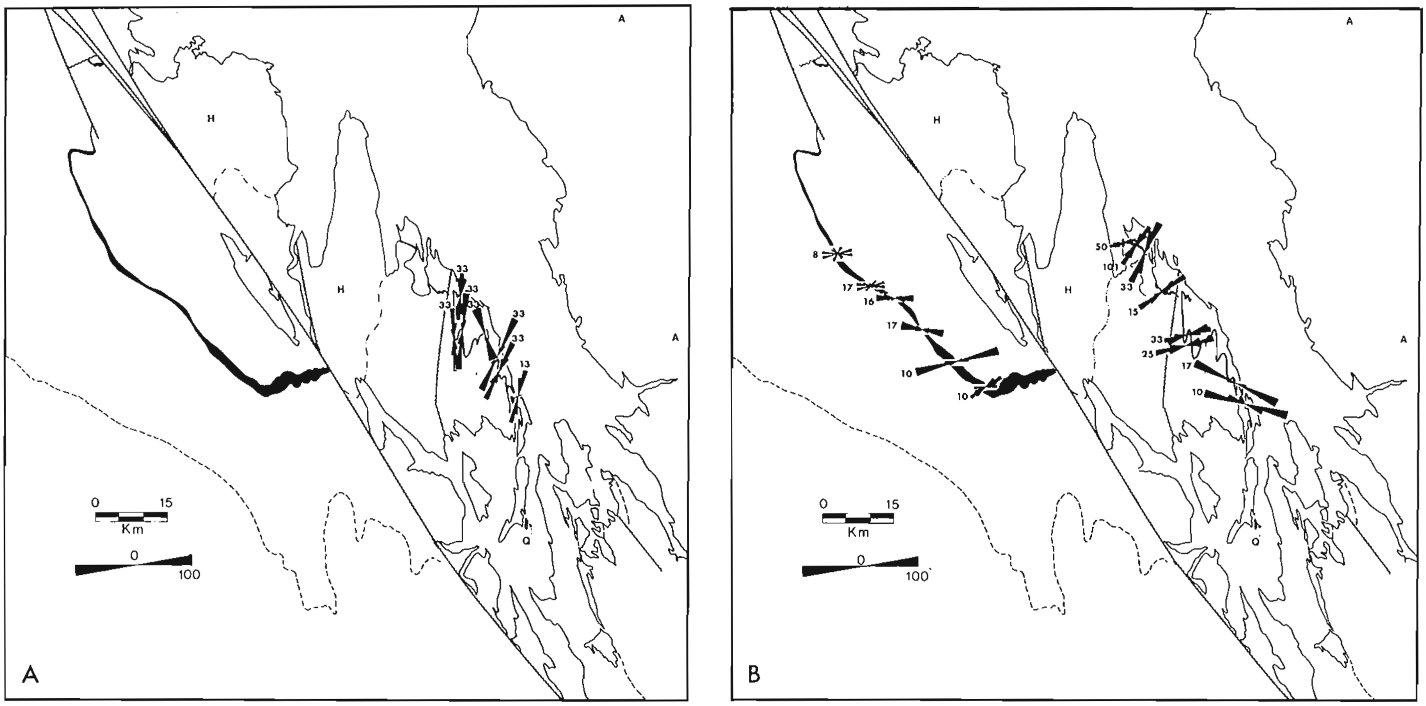


Figure 7.15. *Carbonate pisoliths of the Mp member of the Mara Formation, near the western margin of the Axial Zone. The pisoliths are set in a matrix of fine grained hematitic quartzite. Note the poorly developed inverse grading. GSC 202667-U*



A. *Rose diagrams of stromatolite elongations from the Sheeted facies of the Quadyuk Formation (solid black). The basin is shown in its restored position, and the base of the Goulburn Group is shown as a dashed line. 'A' designates the Archean basement, and 'H' is the Helikian cover.*

B. *Rose diagram of stromatolite elongations from the Reef-Mound facies of the Quadyuk Formation (solid black). Designations as in Figure 7.16 A.*

Figure 7.16

Depositional Environment of the Mara Formation

The thin fining-upward cycles, fine grained sediments, ripples, mudcracks, and rare discontinuous stromatolitic units all suggest that the Siltstone member was deposited in a muddy tidal-flat type of environment.

Most authors presently consider large pisoliths of the Mp member type to have formed in the vadose zone, rather than shallow marine environments (see Blatt et al., 1972). Definitive criteria for this interpretation are lithologic association, degree of reworking by tractive currents, preferential accretion, geopetal structures, reverse grading, and polygonal "fitting" of the pisoliths in the rock. All but the last of these characteristics were observed in the Mp member, strongly suggesting a caliche-type origin.

The granular ironstone typical of the western and central parts of the Axial Zone is also interpreted as a variety of paleosol, but formed in an environment which was not as continuously exposed as the Western Platform (Wanless, 1975).

The carbonate in the member was probably formed at the periphery of the basin, perhaps in an intertidal environment in the southern part of the Axial Zone, laterally equivalent to the Quadyuk Formation (see discussion below). If the bulk of the carbonate was aeolian-transported, the amount of carbonate supplied as dust would depend solely on the available carbonate in the source area, degree of induration, runoff, and total precipitation. In present-day, arid, aeolian-dominated areas, vegetation largely controls the amount of dust, runoff, and water loss through evapotranspiration. However, with the total absence of stabilizing vegetative cover during the Proterozoic, both dust and runoff would have been considerably higher. Thus, although the present-day recorded dust maximum is some 450 kg/ha/month (Goudie, 1973), the accumulation on the Western Platform would have been far greater and easily sufficient to produce the carbonate of the Pisolite member.

Regionally, similar pisolitic paleosols formed at equivalent stratigraphic positions in the foreland basin of Wopmay Orogen (upper Tree River Formation) and the north margin of Athapuscow Aulacogen (Akaitcho River Formation) (Hoffman, personal communication, 1980).

Summary

A decreasing supply of clastics following the end of Burnside River sedimentation, coupled with basin-wide stability and/or slow subsidence, resulted in near sea-level conditions being maintained throughout deposition of the Siltstone member of the Mara Formation. Apparently, though, the differential in the relative rates of subsidence beween the Axial Zone and the Western Platform was continuously maintained throughout deposition of the member, resulting in a greater accumulation of the member in the zone than elsewhere.

Near the end of deposition of the Mara Formation, subsidence of the basin terminated, except in the Axial Zone. This emergent period, periodically interrupted by local stream reworking, resulted in the formation of the caliche and/or calcrete of the Pisolite member. However, sedimentation in the Axial Zone continued, albeit frequently with large supralittoral areas, until marine transgression of both the Axial Zone and the remainder of the basin resulted in deposition of the Quadyuk and Peacock Hills formations.

The Quadyuk[1] Formation

The Quadyuk Formation conformably overlies the Mara Formation and comprises stromatolitic and nonstromatolitic carbonate, calcareous fine grained quartzite near the base, minor calcareous siltstone and mudstone turbidites with rare quartz-pebble conglomerate or grit in the southern Axial Zone at the base of the unit. The Quadyuk varies in thickness up to 60 m, in the area southwest of Bathurst Inlet, but pinches out south of Quadyuk Island, and is absent from the Western Platform (see Campbell and Cecile, 1976b).

The base of the formation is defined as the first stromatolitic carbonate or calcareous sandstone overlying the red siltstones or Pisolite member rocks of the underlying Mara Formation. The top of the formation is defined as the first appearance of grey-green calcareous mudstones (rarely reddish) of the Peacock Hills Formation.

The Quadyuk Formation has been subdivided into two regionally extensive, mappable, depositional facies, based on the character of the contained stromatolites.

Sheeted Facies: consists of a succession of laterally-linked crude columns arranged in a sheeted pattern. Where best developed, the facies occurs directly above a fine grained dolomite-quartzite succession which rests on or is transitional into the Siltstone member of the Mara Formation. The facies consists of low-amplitude, crudely columnar stromatolites which are elongate ovals in plan, and occur in thin sheets of varying thickness (0.5-2.0 m). The individual stromatolites from this facies are characteristically elongate north-south (Fig. 7.16A). The facies outcrops nearly continuously in the southern part of the Axial Zone, but occurs only in thin units in the southwest. It extends to the southern limit of the formation in the Axial Zone, between Quadyuk Island and Tinney Cove (Campbell and Cecile, 1976b).

Mound Facies: is the thickest and most extensive of the facies of the Quadyuk. It always overlies the Sheeted facies, and ranges in thickness from approximately 50 m southwest of Bathurst Inlet to less than 0.5 m south of Quadyuk Island.

The Mound facies consists of isolated and nested, elongate to sub-circular mounds and small bioherms composed of narrow (1.0-2.0 cm), laterally-linked, crudely branching columns, and these characteristically weather a rust-brown to rust-red. The mounds and bioherms have a rounded, convex-upward top surface, with relatively abrupt lateral terminations. They range in size from the "megamounds" 20-30 m long in the southwest to the "minimounds" only some 1-2 m long in the north. They are consistent, however, in their form, orientation, and development. They appear to have formed initially as relatively flat sheets, with little or no vertical relief, and these rapidly developed strongly oriented east-west flanks with deep inter-mound channels (Fig. 7.16B).

Locally within this facies, nearly triangular mounds (in cross-section) apparently developed as isolated individuals, in areas of little or no clastic deposition, and these rapidly expanded upward until they became laterally linked.

Depositional Environment of the Quadyuk Formation

Stromatolites composed predominantly of ferruginous carbonate have been interpreted elsewhere as subtidal, with the mound size increasing with depth, while iron-poor stromatolites have been interpreted as intertidal (Truswell and Eriksson, 1973; Eriksson et al., 1976). The restricted distribution, thickness, and elongations of the stromatolites from the Sheeted facies suggest that they were deposited in a shallow north-south trending depression, with an east-west trending shoreline (Fig. 7.16A).

The distribution of the Mound facies, together with their intimate association with probable deep-water sediments, east-west elongations on the flanks of the Axial Zone (Fig. 7.16B), and their superposed relationship to the Sheeted facies, suggests that they were deposited during a transgressive episode produced by a marked increase in the rate of subsidence of the Axial Zone.

The southward pinchout of both facies within the Axial Zone, into laterally-equivalent(?) periodically desiccated shales and mudstones of the upper part of the Siltstone member of the Mara Formation indicates that deposition of terrigenous clastics was continuous throughout formation of the Quadyuk stromatolites, and that the transgressive limit during deposition of the upper-most Quadyuk was between Quadyuk Island and Tinney Cove, in the southern Axial Zone (see Fig. 7.17).

Coeval with calcrete and ferricrete formation on the Western Platform and western margin of the Axial Zone, subsidence of the zone resulted in the channeling of the remaining amounts of terrigenous detritus (Mara Formation, Siltstone member) into the southern part of the zone. These shales, siltstones, and mudstones accumulated on a small intertidal-supratidal delta at the southern end of the Axial Zone (Fig. 7.17).

At the same time, the relatively shallow-water stromatolites of the Sheeted facies developed in areas of little terrigenous sedimentation in the offshore regions of the delta complex. As well, the first deep-water stromatolites of the Mound facies formed along the flanks of the delta. As the Axial Zone began to subside at a more rapid rate, the shallow subtidal to intertidal(?) stromatolites of the Sheeted facies regraded southward across the deltaic sediments of the Mara. In addition, the deepwater Mound facies stromatolites also regraded over their shallow-water equivalents.

With continued transgression (Fig. 7.17), both facies continued to migrate shoreward, up the slope of the delta complex. The consistently maintained elongation orientation of the stromatolites of both facies suggests that subsidence was rapidly increasing near the end of Quadyuk deposition. The increasing amount of terrigenous clastics in the uppermost part of the Quadyuk supports this hypothesis.

This same detritus (lowermost Peacock Hills Formation) rests conformably on the top of the Pisolite member of the Mara Formation on the Western Platform, indicating that the entire basin foundered as Quadyuk deposition ended.

The Peacock Hills Formation

The Peacock Hills Formation consists predominantly of varicoloured calcareous and noncalcareous mudstones and siltstones. The formation everywhere conformably overlies either the Quadyuk or Mara formations. It is missing (eroded) from the entire Eastern Platform, is thickest in the north-central part of the Axial Zone (300+ m), and thins dramatically to less than 70 m in the southern Axial Zone and the western periphery of the Western Platform.

The formation has been subdivided into three dominant members, which occur throughout the central part of the basin:

P_1 member: consists of thin-bedded red and green to greenish grey mudstone rhythmites and concretionary mudstone.

[1] Quadyuk is the Inuit word for backbone or spine.

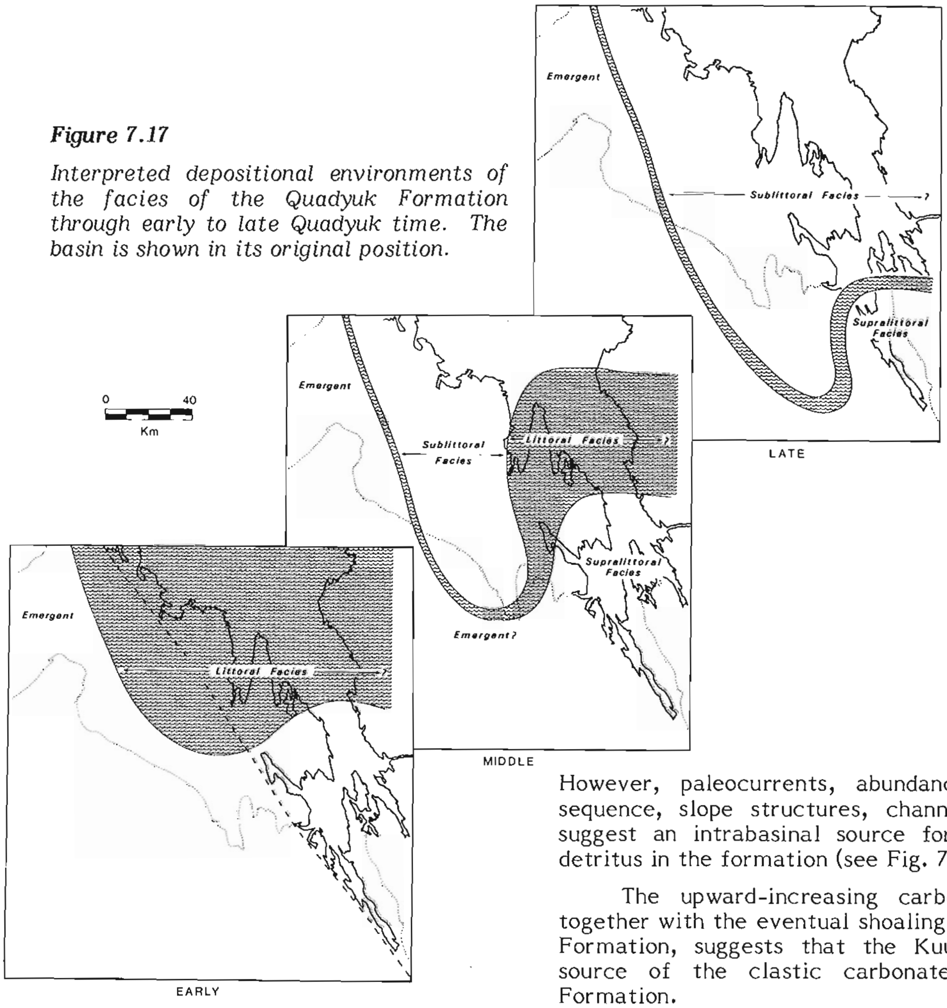


Figure 7.17

Interpreted depositional environments of the facies of the Quadyuk Formation through early to late Quadyuk time. The basin is shown in its original position.

P_2 member: consists of green, red, and red-brown mudstone rhythmites and massive siltstone.

P_3 member: consists of thin-bedded carbonate-mudstone rhythmites.

There are two facies that are lateral equivalents to these three members, and which are considerably condensed when compared to typical sections of the members. These are:

P_4 member: thin-bedded red carbonate-mudstone rhythmites (eastern facies, in the southern part of the Axial Zone).

P_5 member: red and green mudstones and siltstones, minor carbonate (western facies, western extremity of the Western Platform).

In addition to graded bedding present throughout the formation, climbing ripples, flutes, grooves, scours, slump folds, convolute bedding, and local in situ breccias are also present (Fig. 7.18A, B,C). The formation has been described and discussed in some detail elsewhere, and thus detailed descriptions of the internal variations are not included here (see Cecile, 1976; Cecile and Campbell, 1978).

Paleocurrents of the Peacock Hills Formation

Climbing ripples and rare crossbeds in the Peacock Hills indicate that one source of the mudstone rhythmites in the central Kilohigok Basin was to the north and east (Fig. 7.19). However, paleocurrents, abundance of carbonate in the sequence, slope structures, channels, and in situ breccias suggest an intrabasinal source for much of the carbonate detritus in the formation (see Fig. 7.19).

The upward-increasing carbonate in the formation, together with the eventual shoaling into the overlying Kuuvik Formation, suggests that the Kuuvik may have been the source of the clastic carbonate in the Peacock Hills Formation.

Depositional Environment of the Peacock Hills Formation

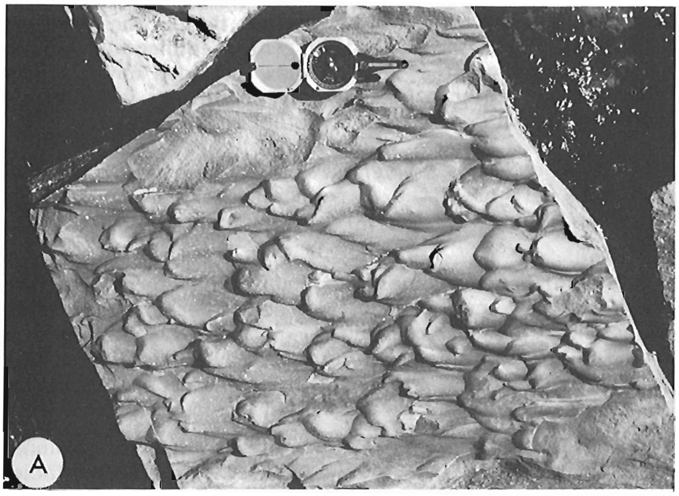
As the Axial Zone of the basin continued to subside following deposition of the Quadyuk Formation, the mudstone basin facies of the Peacock Hills buried the Quadyuk stromatolites. The continued subsidence of the basin may have been accompanied by significant sediment loading as the prograding clastic wedge of the Peacock Hills gradually filled the basin. The absence of shallow-water structures and textures, together with structures normally associated with turbidites, suggest that the mudstone basin facies was deposited in relatively deep water by repeated strong currents.

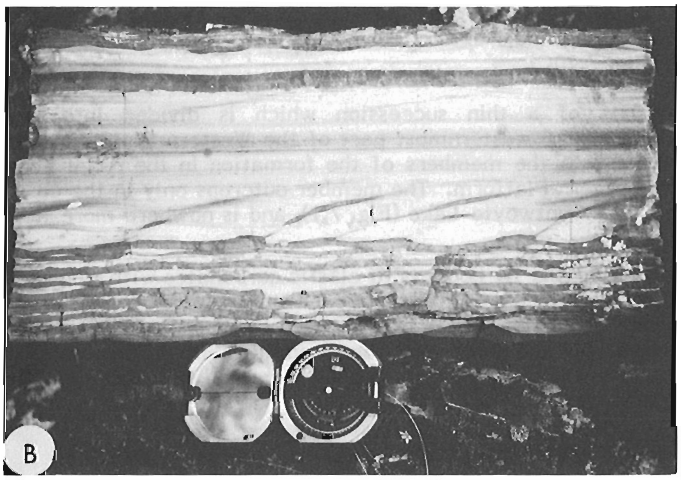
The terrigenous muds, and carbonate muds and silts, were deposited in relatively rapidly subsiding area of the basin across a significant depositional slope, as suggested by the slump folds, channels, and breccias. The rhythmites accumulated during periods of high sediment supply, which interrupted normal pelagic sedimentation represented by the thin, continuous, structureless mudstones.

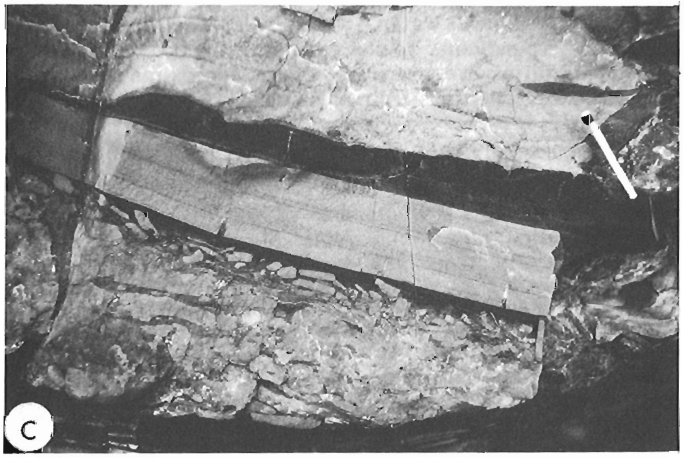
Paleocurrents and thickness variations of the formation suggest that deposition commenced in the northern part of the Axial Zone and gradually spread across the remainder of the basin. The very thin (70 m) section of the formation in the southernmost part of the Axial Zone suggests that this area was continuously a topographic high during deposition of much of the Peacock Hills Formation.

The Kuuvik[1] Formation

The Kuuvik Formation is composed predominantly of clastic and stromatolitic carbonate, with varying amounts of mudstone, siltstone, or calcareous mudstone or siltstone. The formation conformably overlies the Peacock Hills at all localities in the basin and on the Western Platform. It was initially subdivided into four members, and then subdivided further into submembers (see Cecile and Campbell, 1978).







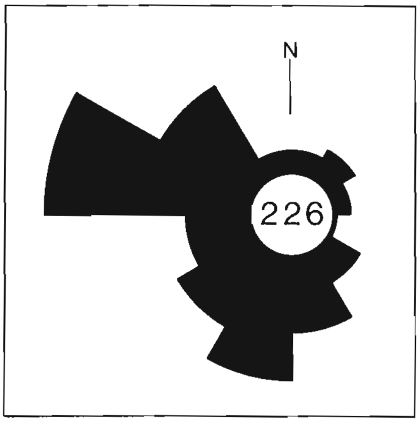


Figure 7.19. *Rose diagram of climbing ripples from the Peacock Hills Formation. The number of readings is as shown, and the diameter of the centre circle is 10 per cent.*

K_1 member (Carbonate Basin facies): the basal unit, consists of thinly-bedded (2-4 cm) carbonate-mudstone rhythmites, with greater than 50% carbonate by volume. The sequence conformably overlies the Peacock Hills, and the contact is generally transitional upward, defined by increasing carbonate. Carbonate siltstone or mudstone which grades upward into grey-green or red argillaceous mudstones are interstratified with beds of uniform mudstone, carbonate, or laminated carbonate. All sedimentary structures present in the underlying Mudstone Basin facies of the Peacock Hills are present in this member.

K_2 member (Shelf facies): consists of a sequence of interstratified thick units of clastic carbonate and mudstone. The prime distinguishing characteristic of this member is the presence of two alternating lithologies – carbonate and mudstone. Thick carbonate-mudstone rhythmites are common near the base of the member, while continuous or discontinuous mudstones are common in the upper part. Abundant, thick, intraclast-bearing, locally stromatolitic carbonate beds are also common in the upper part of the succession. Mudstones of the member change from grey, green and red in the lower part to red or mauve in the upper part. The carbonates commonly show evidence of traction-current deposition (scouring, parallel lamination, crossbedding) in the rhythmites, while the mudstones are either massive or very poorly sorted.

K_3 member (Stromatolitic Carbonate facies): consists almost exclusively of clean carbonate with abundant stromatolites. The contact with the underlying Shelf facies is normally transitional over a few metres, and is marked by

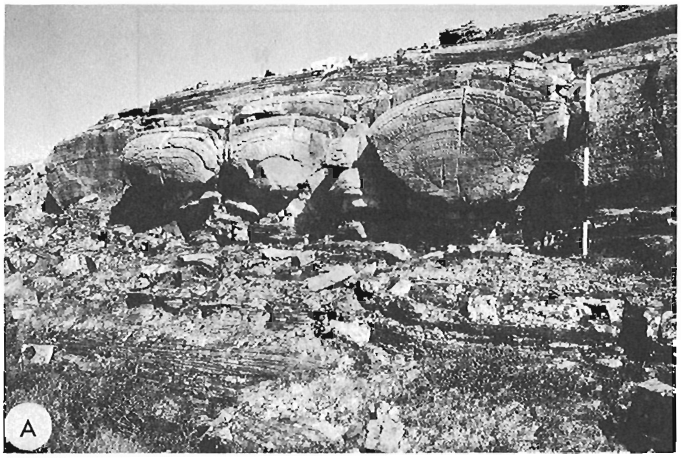
A. Flutes on the base of a muddy siltstone bed in the Peacock Hills Formation. Current flow from left to right. GSC 202670-J

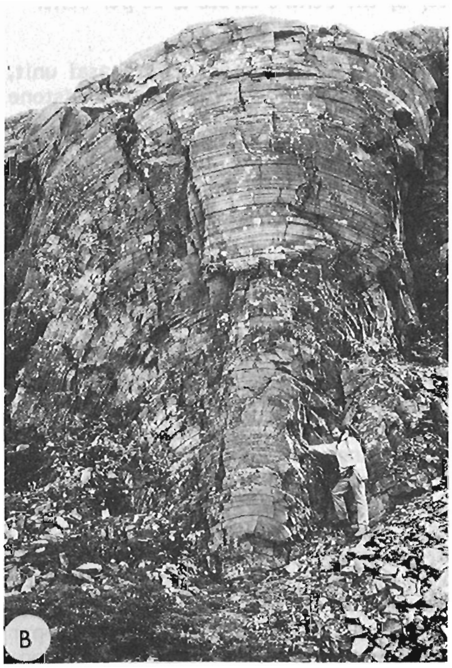
B. Cross-section through climbing ripples in calcareous siltstone, Peacock Hills Formation. GSC 202670-P

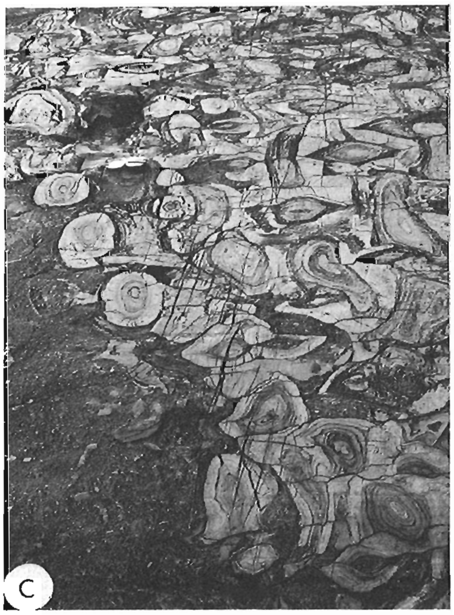
C. Block carbonate breccia from the uppermost part of the Peacock Hills Formation, in the southern part of the Axial Zone. GSC 202667-Y

Figure 7.18

[1] Kuuvik is possibly the Inuit word for muskox underfur.







a rapid gradation into nearly pure carbonate sediments. Cecile and Campbell (1978) subdivided this member into five submembers, which are described briefly below:

K_{3c} and K_{3m} submembers (Subtidal Carbonates with Stromatolite Mounds): the lower part is characterized by scoured, channeled, and crossbedded fine sand to silt-sized carbonates with isolated stromatolite mounds. The upper part is similar, but contains large, well developed, isolated stromatolite mounds (Fig. 7.20A). All mounds are moderately elongate, and their elongation parallels that in the overlying subfacies. The clastic and biogenic sediments of these subfacies accumulated in a shallow subtidal environment, with active preferential currents, resulting in abundant traction-current depositional structures.

K_{3rm} and K_{3r} submembers (Reef Mounds and Reef Columns): are preserved as vertically-stacked stratigraphic units, and the lower (K_{3rm}) is transitional into the upper (K_{3r}).

Large (greater than 1 m) upward-coalescing mounds that form extensive sheets, vertically separated by crossbedded, intraclast-bearing carbonate siltstones and fine sandstones characterize the K_{3rm} subfacies. The mounds range in height from 2-3 m to greater than 10 m, strongly suggesting a high depositional relief (Fig. 7.20B). Channels commonly incise the upper surfaces of the sheets, which have developed directly from the individual mounds.

The K_{3r} subfacies consists of large sheets of laterally-linked, columnar stromatolites which have developed on the individual mounds. The subfacies contains four distinct columnar stromatolite types (see Cecile and Campbell, 1978). The sheets of columns are commonly incised by intraclast and debris-filled channels which are oriented parallel to the elongation of the individual columns. These channels extend downward up to 2 m into the upper surfaces of the sheets (Fig. 7.20C).

K_4 member: consists of a thin succession which is divided into two subfacies. It occurs only on the westernmost part of the Western Platform, and is interpreted as equivalent to the members of the formation in the Axial Zone and eastern part of the Western Platform. The member outcrops only in the cores of two small synclines near Contwoyto Lake (Fig. 7.1), and is nowhere more than 30 m thick.

The basal submember of the K_4 consists of clastic carbonates with thin beds of red or green mudstone with abundant intraclasts and small ripple crosslaminae. The upper submember consists of intraclast-rich carbonates, edgewise conglomerates, thin beds of red and green mudstone, and minor linked and isolated stromatolite biscuits and oncoliths. Rarely, the fine grained carbonate in the uppermost part of the subfacies is mudcracked.

Paleocurrents of the Kuuvik Formation

Stromatolite elongations for all carbonate facies and subfacies of the formation are identical for any particular locality. As many of these forms are interpreted to have formed in shallow water to intertidal environments, the elongations are interpreted as perpendicular to the paleoshoreline(s). In the Axial Zone of the basin, the stromatolites are mostly elongate approximately north-south, while there are significant variations from this orientation on the Western

A. *Well-developed, closely spaced individual mounds from the K_{3m} submember of the Kuuvik Formation. The scale is divided into 30 cm units. GSC 202670-E*

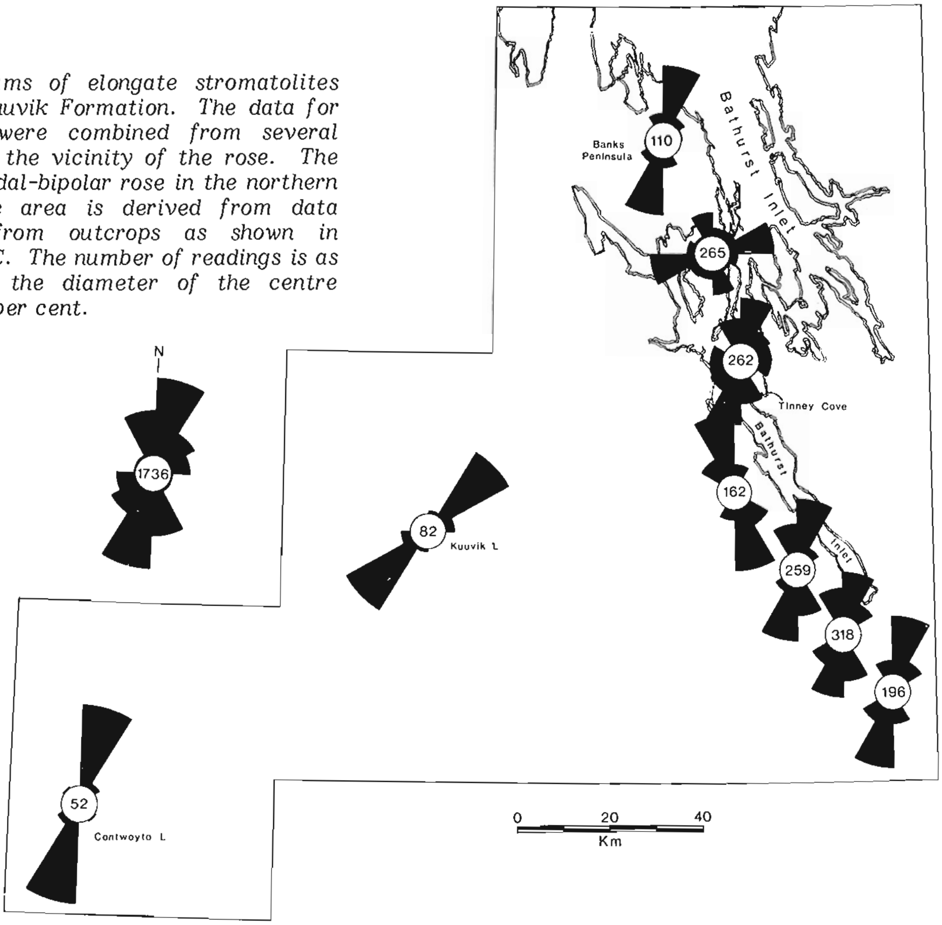
B. *Very large reef-mound from the K_{3m} submember of the Kuuvik Formation, perpendicular to the elongation direction. Note the upward decrease in synoptic relief of the laminae sets. GSC 202918*

C. *Bedding plane-parallel erosion surface showing two elongate stromatolite forms that are mutually perpendicular, from the uppermost part of the Kuuvik Formation, in the southern part of the Axial Zone. The stromatolites are bordered by an intraclast-filled channel (lower left). The scale is given by the compass in the top left-centre of the photograph. GSC 202465-W*

Figure 7.20

Figure 7.21

Rose diagrams of elongate stromatolites from the Kuuvik Formation. The data for the roses were combined from several locations in the vicinity of the rose. The nearly bimodal-bipolar rose in the northern part of the area is derived from data combined from outcrops as shown in Figure 7.20C. The number of readings is as shown, and the diameter of the centre circle is 10 per cent.



Platform, and southeastern part of the Axial Zone (Fig. 7.21). This departure is consistent with the interpretation that the Axial Zone was initially filled by the Mudstone and Carbonate Basin facies, and the stromatolitic part of the Kuuvik prograded over its own fore-reef shelf and rhythmite detritus. This is in part confirmed by the paleocurrent data from the underlying Peacock Hills Formation (Fig. 7.19; see also discussion below).

The stromatolite elongations from the Kuuvik also show that the Axial Zone of the basin maintained its orientation relative to the Western Platform during Kuuvik deposition. The thickness variations, together with minor facies changes, suggest that the paleoslope on the Western Platform was generally to the east-northeast during deposition of the Kuuvik.

Depositional Environment of the Kuuvik Formation

Vertically stacked members of the Kuuvik Formation all show evidence of once being laterally juxtaposed facies (see Walther's Law, in Middleton, 1973). During deposition, members of the Kuuvik Formation accumulated as a lateral succession of basin, to shelf, to fore-reef clastics, to reef and back reef facies (see Cecile and Campbell, 1978).

As do the underlying Mudstone Basin rhythmites of the Peacock Hills, the K_1 Carbonate Basin sediments show evidence of deep water deposition by turbidity currents, which interrupted normal pelagic sedimentation. The overall upward increase in the percentage of carbonate through the uppermost P_3 and into the overlying K_1 strongly suggests that the detrital carbonate source was prograding into the basin.

The overlying K_2 Shelf facies was deposited by periodic strong currents carrying abundant detrital carbonate, as well as by gravitational settling of large volumes of mudstone. The sympathetic upward increase in thickness of both the carbonate and the mudstone beds suggests a more proximal source than for the underlying K_1 member. The abundant intraclasts, red colour, and variegated nature of the uppermost K_2 mudstones suggests progressively more shoaling environment. The K_2 sediments were initially deposited on a submerged carbonate platform, which received storm-generated and transported carbonate detritus from a proximal, prograding, possibly emergent reef tract. The mudstones may have been bypassed through gaps in the reef tract during these storms.

In the K_3 member, abundant intraclast-rich sediments, channels, crossbedding, lack of mudstone, and extreme elongations of individual stromatolites and mounds indicate that the clastic and biogenic sediments accumulated in a high-energy environment. The size of the mounds, vertical and lateral spacing, variations in relief, and their extreme elongation indicate that they had the potential of wave resistance, and of strongly influencing the local environment. The possible drainage, wave-generated, or swash channels suggest that they formed in the surf zone, and thus probably continuously supplied detritus to the Mudstone and Carbonate Basin facies which they overlie.

The K_4 Platform facies accumulated in a markedly different environment than its lateral equivalents to the east. In contrast to the stromatolites of the Axial Zone, those of the Platform facies show no diversity or reef development, consistent with development in shallow water. This interpretation is consistent with the very thin gross character of the K_4, suggesting that it is the vestige of a once-continuous basin-rimming carbonate veneer that may have extended as far south as the Athapuscow Aulacogen.

The Brown Sound Formation

The Brown Sound Formation conformably overlies the Kuuvik, with the dominant lithologies interfingering in the contact area (10-50 m). The contact between the two formations is defined as the first appearance of greater than 50 per cent mudstone, in thick successions, and this contact can generally be mapped to within a few metres.

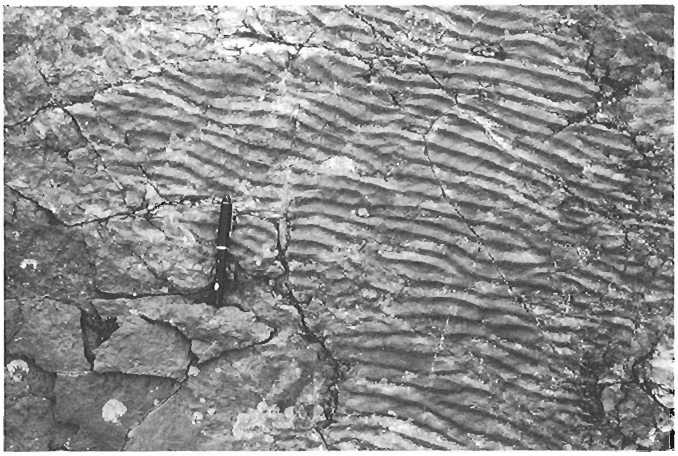
The Brown Sound is conformably overlain by the Amagok Formation. In the southern part of the Axial Zone, the upper part of the Brown Sound (10-30 m) becomes coarser grained and lighter red in colour, until the sediments are only moderately indurated and cream to mauve in colour. In the north-central part of the Axial Zone, the upper part of the Brown Sound consists of thick alternations of fine and medium grained spotted red arkoses, and coarse grained cream to mauve sandstones over about 400 m of section. In both areas, the contact between the Brown Sound and Amagok Formations is defined by the last thick succession of fine to medium grained well-indurated red or reddish sandstones. The Brown Sound has been subdivided into three members, as well as numerous submembers (see Campbell and Cecile, 1976a, b; Cecile and Campbell, 1977).

$\underline{B}_1$ member: consists of, from the base to the top, red mudstones (200 m), sandstones (60 m), and a slump breccia named the Omingmaktook[1] olistostrome (70-200 m where present). The lower mudstones are gradational into the Kuuvik Formation, and contain thin beds of carbonate and mudstone with ripple marks, mudcracks, and abundant salt casts (Fig. 7.22A, B). The mudstones are commonly calcareous, display distinctive red-green mottling, and are conformably overlain by thick units of coarse grained buff immature sandstones interstratified with thin units of mudstone. The sandstones are crossbedded and locally contain abundant carbonate rock fragments.

The sandstone submember is apparently thickest on Banks Peninsula, in the south-central part of the Axial Zone. However, exposures of this unit are complicated by faulting in the southern area of the zone.

The sandstones are directly overlain by a laterally discontinuous slump breccia (Omingmaktook olistostrome) of allochthonous brecciated and chaotically folded sheets of carbonate set in a carbonate-mudstone breccia matrix. The allochthonous carbonate sheets apparently accumulated essentially in the same stratigraphic position as that in which they were deposited. The largest slumped sheets occur in south-central Goulburn Group exposures, (see Cecile and Campbell, 1977).

$\underline{B}_2$ member: conformably overlies $\underline{B}_1$, is approximately 300 m thick, and occurs throughout the southern part of the basin without noticable variation in thickness. The base of the member is defined as the first appearance of red muddy siltstones immediately above the Omingmaktook olistostrome. The siltstones, which become progressively coarser upward through the member, are thinly bedded, texturally homogenous, laminated, red, muddy, and calcareous. They typically have flaggy partings, specular hematite, and abundant detrital muscovite. Approximately 100 m above the base of the member, thin mud-chip conglomerates first appear, with small, low-amplitude ripples on the upper surfaces of the siltstone beds. These continue sporadically throughout the remainder of the member.



A. Small-scale current ripples on mudstones of the lowermost Brown Sound Formation. GSC 202666-I

B. Salt casts on the upper bedding surface of mudstones of the lowermost Brown Sound Formation. GSC 203062-R

Figure 7.22

$\underline{B}_3$ member: consists of well-indurated red arkoses and minor thin basalt flows. The conformable contact with the $\underline{B}_2$ member is gradational over 10-40 m, and is defined by the first appearance of thick successions of fine grained sandstones. On Banks Peninsula, in the south-central Axial Zone, the $\underline{B}_3$ has been subdivided into three submembers. Thinner, fault-disrupted equivalents of these occur throughout eastern Goulburn Group exposures, but were not mapped at the scale of this study (1:250 000).

$\underline{B}_{3a}$ submember: the lowest, consists of 200-400 m of fine to medium grained well-indurated sandstones, locally interstratified with red mudstones and rare muddy siltstones.

[1] Omingmaktook is the Inuit word for muskox place.

B_{3b} submember: is a succession of white-spotted red sandstones interstratified with very thin (3-8 m) amygdaloidal basalt flows. Each flow is characterized by a chilled base, fine to medium grained sub-ophitic centre, and a chilled, locally blocky top with quartz- and carbonate-filled vesicles. Thicker, more massive flows occur at the southern end of Bathurst Lake, but disruption of the sequence by faulting has obliterated much of the stratigraphic detail. In this area, though, they appear to be approximately 50 m below the base of the Amagok Formation.

B_{3c} submember: is a transitional unit between the Brown Sound and Amagok formations. It consists of thick, white-spotted red sandstones interstratified with thick cream-coloured pebbly sandstones similar in texture and composition to those of the Amagok.

The Amagok[1] Formation

The Amagok Formation consists of a succession of cream to mauve, poorly indurated immature sandstones which range from 800 m in the north to 1000 m in southern exposures. The base of the formation is defined as the last appearance of thick, fine to medium grained, well-indurated sandstones of the Brown Sound Formation. The top of the formation is marked by an angular unconformity with younger Proterozoic sandstones and conglomerates of the Tinney Cove and Ellice formations.

Bedding units in the Amagok consist of 1-2 m thick trough crossbedded sandstones, capped by abundantly rippled and massive ferruginous sandstones. On Banks Peninsula, in the south-central Axial Zone, the sandstones contain abundant dispersed pebbles, and one thin orthoconglomerate bed. Of the pebbles in the ortho-conglomerate, approximately 50 per cent are white quartz and minor quartzite; the remainder are granitoid and gneissic rocks (30-40%), mylonite (10-20%) and fine grained sediments and minor volcanics. The volcanic clasts, which comprise only approximately 1 per cent of the clasts, include fine grained acid to intermediate tuffs, red crystal-rich acid to intermediate tuff, and sub-ophitic and porphyritic fresh intermediate to basic rocks.

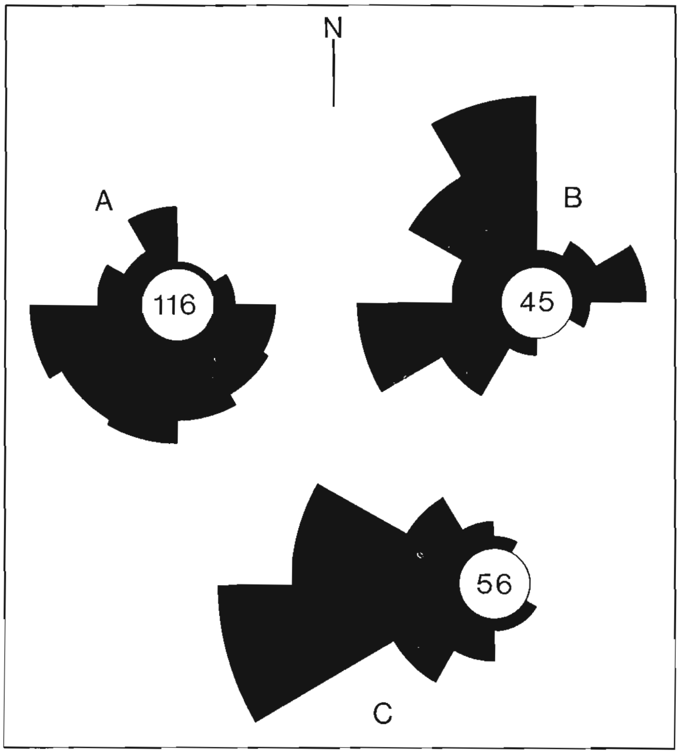
Paleocurrents of the Brown Sound and Amagok Formations

Approximately 275 paleocurrent indicators were recorded from these two formations throughout the area examined (Fig. 7.23). Paleocurrent indicators from the B_3 sandstones (Fig. 7.23A) show a general but diffuse southward dispersal pattern. Current ripples in the lower Brown Sound (Fig. 7.23B), associated with mudcracks and salt casts, are consistent with the orientation of the paleostrandline interpreted from stromatolite elongations from the underlying Kuuvik Formation (Fig. 7.21).

Trough and planar crossbeds from the Amagok Formation in both the northern and southern parts of the Axial Zone show that these sands and conglomerates were transported to the west and southwest as well (Fig. 7.23C). Thus, the paleocurrent and facies data together show that these sediments were transported across the basin by currents flowing from the north, northeast, and east.

Depositional Environments of the Brown Sound and Amagok Formations

The Brown Sound and Amagok formations are considered to be a diachronous vertical succession formed by lateral progradation of adjacent facies into the Kilohigok Basin (Walther's Law, in Middleton, 1973).



A. Trough and planar crossbeds from the Brown Sound

B. Current ripples from the Brown Sound

C. Trough and planar crossbeds from the Amagok

Figure 7.23. *Rose diagrams of paleocurrent indicators from the Brown Sound and Amagok formations. Number of readings is as shown, and the diameter of the centre circle is 20 per cent.*

The lowermost (B_1) facies of the Brown Sound records a pause or possibly a reversal in the major regression of the seas from the Kilohigok Basin recorded in the underlying carbonates and mudstones (Kuuvik and Peacock Hills). The overlying facies of the Brown Sound demonstrate a continuation of this regression, resulting in a widespread influx of terrigenous detritus into the basin (B_2 and B_3 members), and culminating with deposition of the braided fluvial coarse sands and conglomerates of the Amagok Formation.

KILOHIGOK BASIN – WOPMAY OROGEN TECTONO-DEPOSITIONAL RELATIONS

Rocks of the Kilohigok Basin have been lithostratigraphically correlated with the Coronation Supergroup (Wopmay Orogen) and the Great Slave Supergroup (Athapuscow Aulacogen) by Campbell and Cecile (1976c) and most recently by Hoffman (1981a). In addition to the gross correlation of formations and groups, each sequence contains unique common elements at precisely the same stratigraphic position which reinforce an interpreted common evolution. The interdependant depositional relationships of the Wopmay Orogen and Kilohigok Basin are interpreted here using the Wilson Cycle model proposed by Hoffman (1980, 1981b).

Hoffman (1980) suggested that two aulacogens (Athapuscow and the since-named Taktu) developed as failed arms of triple rifts which formed over two hot spots, while

[1] Amagok is the Inuit word for wolf.

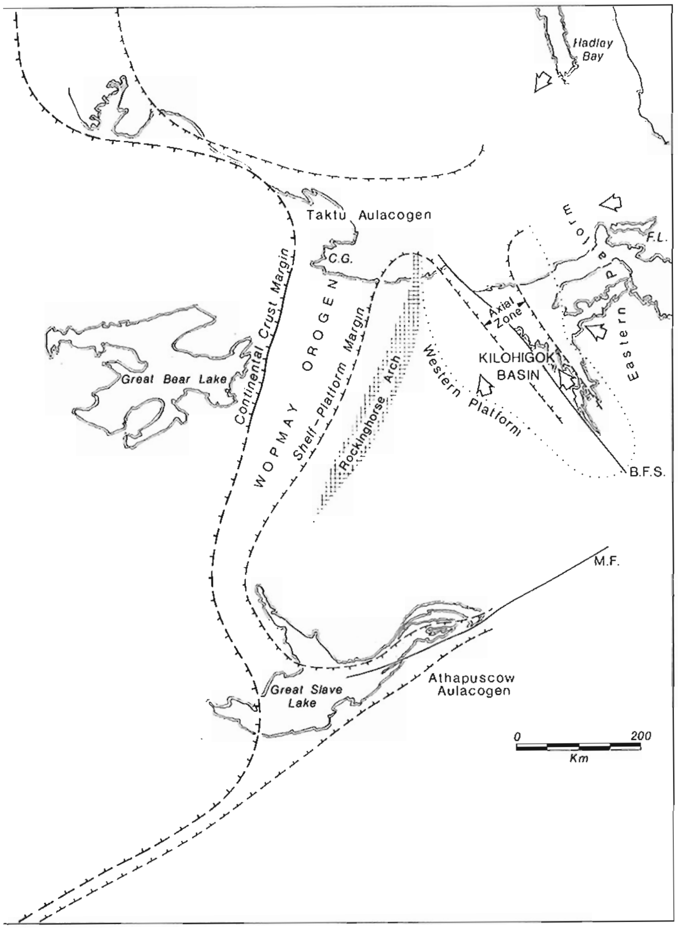


Figure 7.24. *Inter-related elements of the Wopmay Orogen, Taktu Aulacogen and Kilohigok Basin. Modified from Campbell (1981) and Hoffman (1980).*

the active arms in the intervening area linked to form the Wopmay Orogen (see Fig. 7.24). The initial protracted period of continental rifting, terrace-rise sedimentation, and early volcanism was followed much later by oblique subduction of the Slave Plate.

Sedimentation in the Kilohigok and eastern Wopmay Orogen can be subdivided into five main tectofacies[1], based on the character of the contained sediments, and their relative positions within the Wilson Cycle model as follows:

A. Initial Subsidence tectofacies: Western River; Odjick, Rocknest formations.

B. Secondary Subsidence tectofacies: Burnside River, Mara; Tree River and Fontano formations.

C. Early Tertiary Subsidence tectofacies: (top of Mara), Quadyuk, Peacock Hills; Asiak, Fontano formations.

D. Middle Tertiary Subsidence tectofacies: Kuuvik, lower Brown Sound; Cowles Lake formations.

E. Terminal Subsidence tectofacies: mid-upper Brown Sound, Amagok; Takiyuak formations.

Initial Subsidence Tectofacies

With eastward propogation of the Taktu Aulacogen from the epicentral hot spot at the western end of Coronation Gulf with earliest ocean opening, sedimentation commenced in the Wopmay as the Akaitcho Group was deposited. With increasing eastward-spreading subsidence, the earliest clastics of the Odjick Formation accumulated on the Slave Craton margin, supplied from source areas to the east and south (paleocurrent data from Hoffman, personal communication, 1980).

The earliest sediments of the Western River accumulated in the Kilohigok Basin, as intracontinental equivalents of the initial Odjick clastics (see Fig. 7.25). Initially, the Western River clastics and biogenic carbonates were essentially restricted to the proto-Axial Zone of the basin, as well as minor areas of the Western Platform.

With increasing subsidence of the eastern Taktu, equivalent sediments of the Western River Formation may have been deposited on the northern part of the Eastern Platform as the Hadley Formation (Campbell, 1981). In addition, although not extensively preserved, Western River equivalents were probably deposited on the remainder of the Eastern Platform, in much the same fashion as they accumulated on the Western Platform.

With depression of the flanking marginal areas of the aulacogen and Axial Zone of the Kilohigok Basin, easterly-derived terrigenous clastics were increasingly restricted to the Kilohigok Basin, with little detritus reaching the eastern margin of the Wopmay. This predominantly marine period of Western River sedimentation is equivalent to the formation of the Rocknest stromatolite carbonate platform, which commenced following a period of accumulation of fine, pelagic, clastics on the subsiding passive margin (Fig. 7.26; Hoffman, 1975). During final stages of initial rifting, the Taktu and Axial Zone reached their subsidence maxima. During this period of maximum transgression, only pelagic

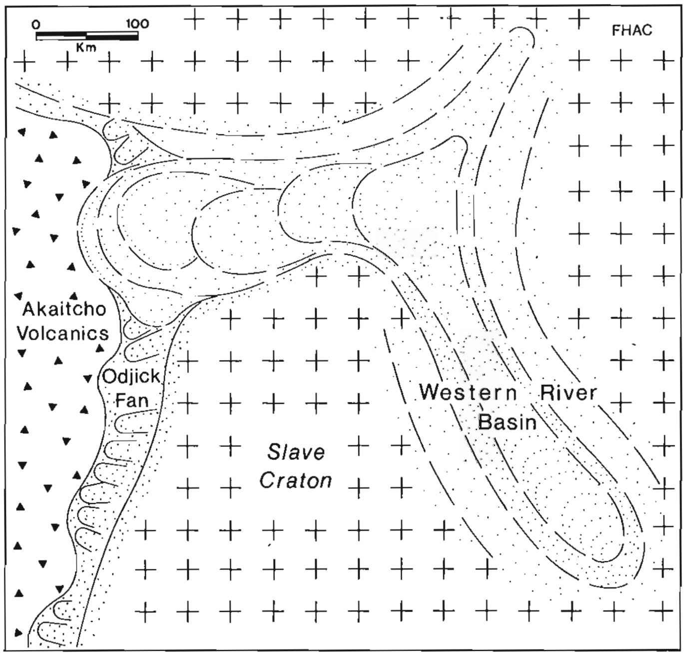


Figure 7.25. *Distribution of the units of the earliest Initial Subsidence tectofacies. This tectofacies is related to the passive margin subsidence to the west, at the continental margin. The Odjick sediments are beginning to cover the Akaitcho Group, and are gradually regrading onto the Slave Craton. The earliest sediments of the Western River Formation are accumulating in the Axial Zone of the Kilohigok Basin, and advancing onto the marginal areas of the Eastern and Western platforms.*

[1] Tectofacies: "a lithofacies that is interpreted tectonically" (Glossary of Geology, R.L. Bates and J.A. Jackson, ed., American Geological Institute, 1980).

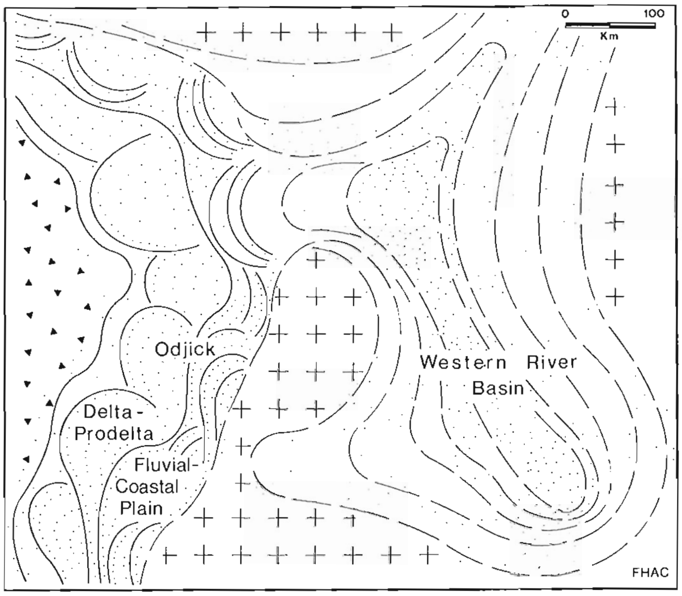


Figure 7.26. Distribution of the units of the middle part of the Initial Subsidence tectofacies. This tectofacies is related to the continued subsidence of the passive margin of the continent to the west. Southeasterly-derived Odjick delta-prodelta clastics are deposited on the slowly subsiding margin, and have buried the Akaitcho Group. Western River sediments are deposited farther east and west onto the flanking platforms adjacent to the Axial Zone of the basin.

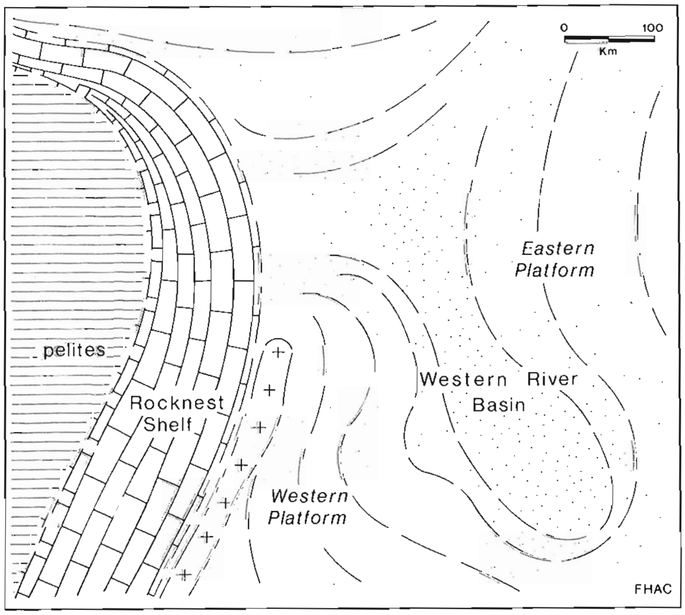


Figure 7.27. Distribution of the latest units of the Initial Subsidence tectofacies. This is near the terminal stages of the passive margin subsidence. The Western River marine basin is now at its maximum extent, and only minimal terrigenous clastics reach the subsiding passive craton margin. The lack of clastic sediment supply permitted the development of the extensive Rocknest stromatolite platform.

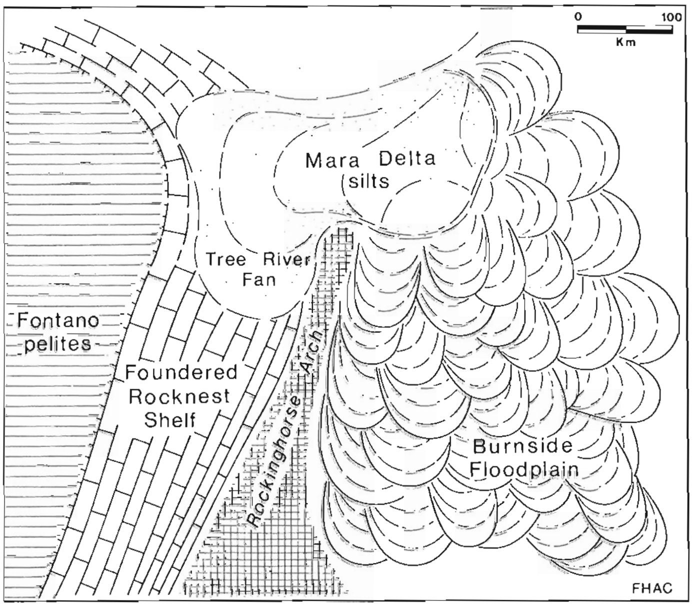


Figure 7.28. Distribution of units of the early Secondary Subsidence tectofacies. This phase in the Kilohigok Basin is intimately related to the initiation of west-dipping subduction of the Wopmay Orogen (see Hoffman, 1980). As the distal edge of the Slave Plate descends into the west-dipping subduction zone, the Rocknest Shelf founders, and the Rockinghorse Arch develops as a broad swell that crudely parallels the subduction zone. Fluvial clastics of the Burnside River spread down the Axial Zone and into the Taktu Aulacogen, and are also deposited on the Eastern Platform, now denuded of older Western River sediments. The distal equivalents of the Burnside River which reach the sea at the mouth of the Taktu accumulate as the Mara silts. These silts are also swept southward parallel to the coast, to be deposited as part of the Tree River Fan.

and hemipelagic sediments reached the passive margin, and the Rocknest Platform reached its maximum development (Fig. 7.27).

The near-complete absence of the Western River Formation from the Eastern Platform, the locally missing Upper Argillite member within the basin, and Archean boulders in the Burnside River Formation suggest that there was a protracted temporal break before deposition of the Burnside River. This was first indicated by the aberrant paleomagnetic pole position for the Western River as shown by Evans and Hoye (1981). A similar erosional interval also occurs between the Duhamel and Kluziai in the Athapuscow Aulacogen (Hoffman, 1968) and possibly also between the Rocknest Formation and the Recluse Group in the eastern Wopmay (Hoffman, 1981a).

With the onset of west-dipping subduction of the Slave Plate, the peritidal Rocknest shelf foundered (Fig. 7.28). Tensional stress within the Slave Plate produced the Lupin Fault-type structures, and also extended the Axial Zone of the Kilohigok Basin. Subduction of the Slave Plate also produced an upward "bulge" within the craton, subparallel to the subduction zone – the Rockinghorse Arch – which effectively dammed the westward spread of fluvial clastics.

Secondary Subsidence Tectofacies

Subduction and associated distension of the Axial Zone apparently peaked during deposition of the Burnside River fluvial clastics. Distension resulted in periodic, perhaps

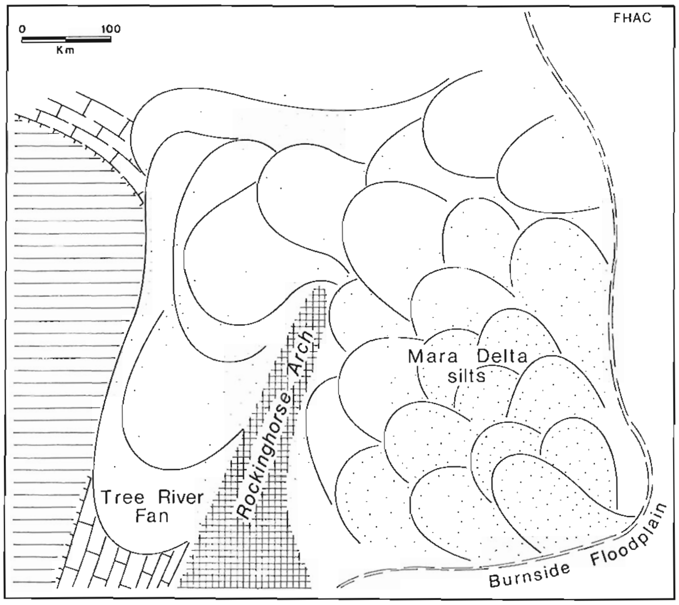


Figure 7.29. *Distribution of the units of the late Secondary Subsidence tectofacies. With decreasing supply of clastics, and accompanying marine transgression, the Mara silts regraded into the Kilohigok Basin, and the Burnside River floodplain all but disappeared. With dimunition of the positive topographic effects of the Rockinghorse Arch, the Mara silts may have spread across the arch directly into the Tree River Fan complex.*

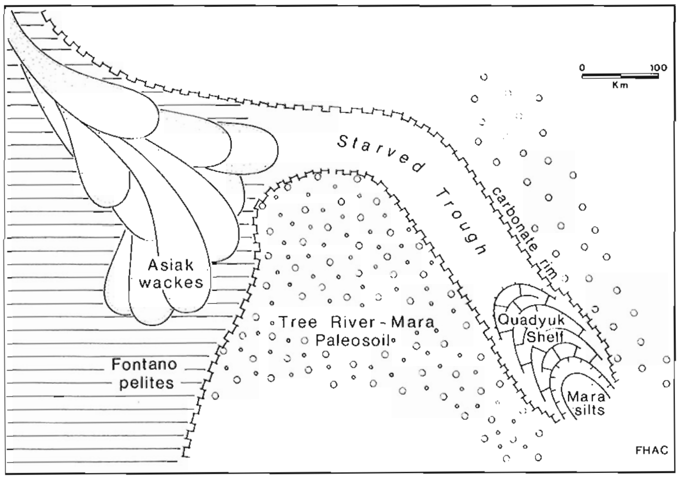


Figure 7.30. *Distribution of the units of the early Tertiary Subsidence tectofacies. This phase is intimately related to the continent-microcontinent collision (see Hoffman, 1980). This collision, initially north of the Taktu, produced uplifted zones of earlier clastics, which were cannibalized and shed southward as the Asiak wackes into the trough generated by continuing subduction in the south. During this same period, regional uplift of most of the Kilohigok Basin and Wopmay shelf produced the Mara-Tree River paleosols, while Quadyuk stromatolites formed in, and probably rimmed, the sediment-starved trough.*

catastrophic, episodes of faulting/subsidence which channeled Burnside River clastics into the Axial Zone, and distal equivalents into the Taktu Aulacogen. With continued extension, perhaps generated through clockwise rotation of the craton northwest of the Taktu during oblique subduction, the Taktu continued its eastward propogation, and proximal fluvial clastics of the Burnside River accumulated on the Eastern Platform.

A minor splay may have developed off the eastern end of the Taktu at this time, as suggested by paleocurrents in the northernmost Burnside River on the Eastern Platform. This splay, or depression, acted as a channel down which the Burnside River-Mara prograded into the Taktu, and merged with their southern-derived equivalents (Fig. 7.28).

During deposition of the coarse clastics of the Burnside River, the distal equivalent fine sands and silts accumulated as the Mara Formation fluvio-deltaic succession at the prograding braided floodplain front. Initially deposited directly on the Western River Formation in the southern part of the basin, these fine grained sediments were removed and redeposited during progradation of the Burnside River fluvial plain. However, as the Mara silts advanced into the shallow marine facies at the mouth of the Taktu, they were swept along the shallow continental margin and accumulated as the Tree River Formation (Fig. 7.29).

Early Tertiary Subsidence Tectofacies

Regional subsidence and marine transgression which accompanied deposition of the Mara and Tree River may have been initiated by reactivation of the Taktu and Axial Zone during terminal stages of subduction and initial stages of collision (see Sengor, 1976). As collision continued, the entire area became emergent, and the characteristic regional paleosols formed at the top of the Mara and Tree River.

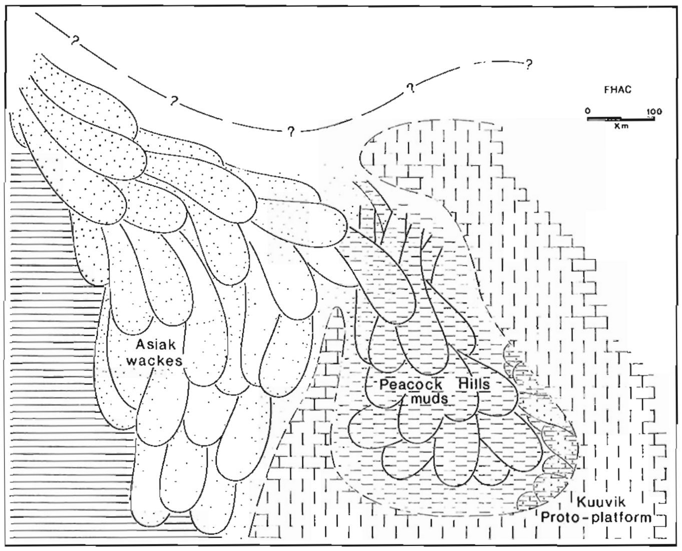


Figure 7.31. *Distribution of the units of the middle Tertiary Subsidence tectofacies. As the collision zone continued its southward migration from north of the Taktu Aulacogen, progressively coarser and more cratonic clastics were supplied to the Asiak wackes, following denudation of the basement north of the Taktu. Coeval with continuing collision, the Taktu Aulacogen was reactivated, distal muddy equivalents of the Asiak wackes were deposited as the Peacock Hills in the Kilohigok Basin, and the Kuuvik proto-platform developed on the relatively stable, shallow marine continental hinterland.*

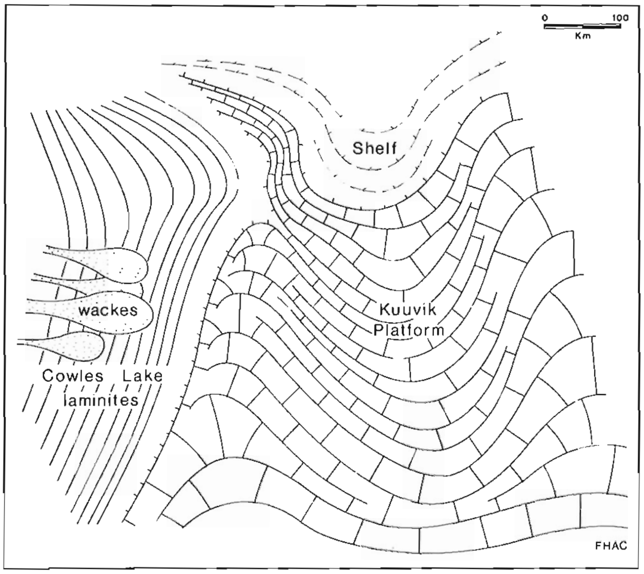


Figure 7.32. Distribution of the units of the late-middle Tertiary Subsidence tectofacies. Following erosion of the uplifted, collision-generated highlands, and prior to rapid post-collision isostatic uplift, Cowles Lake laminites and westerly-derived wackes were the only terrigenous clastics supplied to the region. The paucity of terrigenous clastics supplied to the Kilohigok Basin resulted in the rapid(?) progradation of the Kuuvik Platform into the basin and possibly down the aulacogen. Southward-prograding muds and silts of the Brown Sound Formation were still far to the north of the Taktu at this time, but were gradually advancing southward.

At the same time, though, the Axial Zone of the Kilohigok, and probably the Taktu and Wopmay axes were maintained as static or slowly subsiding depocentres.

The paucity of terrigenous detritus supplied to the Axial Zone and the remainder of the region at this time permitted the fringing stromatolitic carbonates of the Quadyuk to develop. These apparently rimmed the Axial Zone, and may have extended around the rim of the Taktu (Fig. 7.30). Coeval with Quadyuk platform deposition, wackes derived from the collision-generated highlands north of the Taktu spread southward into the Wopmay axis and accumulated as the Asiak Formation. The Wopmay axis, as indicated by paleocurrents (Hoffman, personal communication, 1980) in the Asiak, was parallel to subparallel to the trace of the subduction zone, similar to the present Timor Basin (Shanmugam and Walker, 1980).

Middle Tertiary Subsidence Tectofacies

As collision continued, both the Taktu and Axial Zone continued to subside. Coarse detritus from the orogenic source areas to the north continued to pour into the Wopmay (Hoffman, 1980), while their fine grained distal equivalents accumulated in the Kilohigok as the Peacock Hills Formation (Fig. 7.31). The Kuuvik proto-Platform developed in the shallow epeiric sea which covered the continental interior during this period of minor, though extensive, subsidence.

As the tectonic highlands were denuded, and supplied less and less clastic detritus to the basins, the Kuuvik prograded into the basin over the Peacock Hills and its own fore-reef detritus (Fig. 7.32). Waning sediment supply from the north is reflected in the Wopmay by deposition of the Cowles Lake laminites and subordinate westerly-derived Cowles Lake wackes. The Wopmay was continuously a relatively deep-water basin during Cowles Lake deposition and equivalent Kuuvik progradation, as the sequence is stromatolite-free. However, the uppermost part of the Cowles Lake contains fragments of the characteristic "ridged stromatolite" biscuits typical of the top of the Kuuvik (Cecile and Campbell, 1978), suggesting that the Kuuvik prograded relatively close to the Tree and Takijuq Basins (Fig. 7.32).

Terminal Subsidence Tectofacies

Following separation of the subducted slab from the west-dipping Slave Plate, the isostatically imbalanced, overthickened sedimentary wedge in the collision zone began to rise and shed detritus back onto the craton and basins. The initial fine grained clastics of the Brown Sound Formation were the penultimate deposits of the coarsening-upward "molasse phase" in the Wopmay and Kilohigok (Fig. 7.33).

As the rate of isostatic uplift in the collision zone increased, progressively coarser detritus was supplied to the basin(s) as vast muddy fluvio-deltaic systems spread south and east into the Kilohigok and Wopmay respectively. Takiyuak muds, silts and sands were derived from the same eastward-advancing source as the earlier westerly-derived Cowles Lake wackes.

As the fluvial distributaries advanced over the basin, minor subcrustal fractures breached the surface and formed conduits for the thin subaerial basalts of the uppermost Brown Sound. This early fracturing may have been a precursor of the later Bathurst Fault System. If this is correct, then considerable volumes of volcanics may have been deposited in the northern Kilohigok and Taktu (Fig. 7.34).

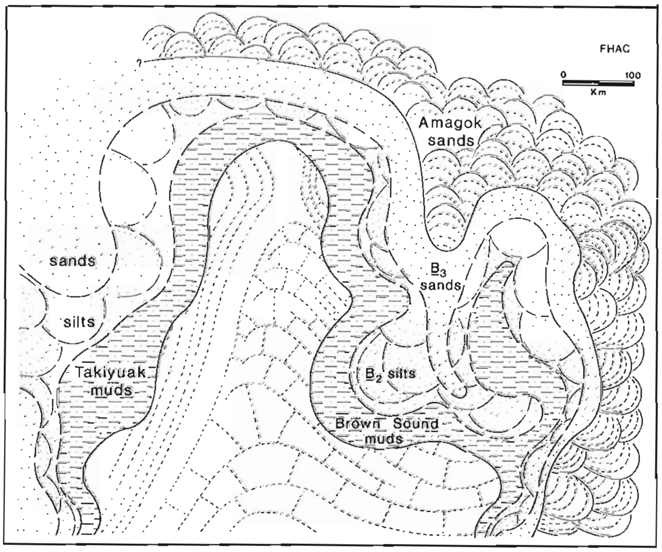


Figure 7.33. Distribution of the units of the early Terminal Subsidence tectofacies. Following, or coeval with, detachment of the west-dipping subducted slab, possibly related to the effects of a distal second collision, isostatic uplift commenced in the region north of the Taktu Aulacogen. The southward-prograding uplift supplied progressively coarser terrigenous clastics to fluvio-deltaic distributaries, which spread them across the Kuuvik Platform as the Brown Sound, and over the Cowles Lake as the Takiyuak.

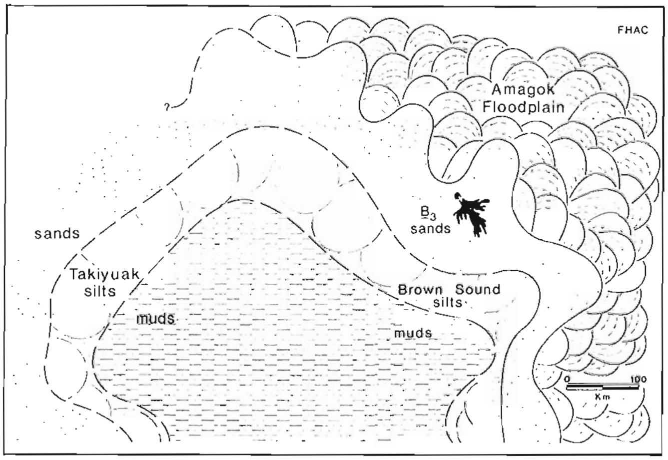


Figure 7.34. *Distribution of units of the late Terminal Subsidence tectofacies. With an increased rate of isostatic uplift, progressively coarser westerly, northerly and easterly-derived terrigenous sediments spread across the region as the Brown Sound and Takiyuak sands. Basalts (solid black) in the uppermost Brown Sound may have been extruded during periods of minor faulting. The southward-prograding sediments ultimately covered both basins, as the braided fluvial grits and gravels of the Amagok and Takiyuak(?) were deposited.*

During terminal deposition in the Kilohigok and Wopmay, braided fluvial systems spread coarse sands and gravels from the north, east and west into both regions. The boulders in the Amagok Formation conglomerates suggest that the earlier sedimentary cover in and north of the Taktu had been completely cannibalized, and the Archean basement was exposed around the Taktu and Kilohigok Basin.

SUMMARY AND CONCLUSIONS

The spatially related Wopmay Orogen and Kilohigok Basin both developed in response to initial rifting of the continental margin over the Coronation Hot Spot (Hoffman, 1980). The Taktu Aulacogen, propogated as a failed rift arm from the triple junction over the Coronation Hot Spot, developed a subsidiary splay off its eastern termination which became the Axial Zone of the Kilohigok Basin.

Marine transgression of the dissected Archean basement in the Kilohigok Basin region resulted in deposition of the shallow marine-dominated Western River Formation, with its attendant stromatolite platforms, shallow shelf quartz sands, and subordinate deltaic muds and silts. These sediments are temporally equivalent to the initial passive margin succession accumulating on the eastern border of the Wopmay Orogen.

With relatively(?) rapid subsidence, the shallow shelf clastics were covered by a thin succession of their deeper-water equivalents as the Upper Argillite member regraded into the basin.

At some time after deposition of the Western River Formation, the basin was locally and/or regionally uplifted, and parts or all of the formation were stripped from the Eastern Platform and local areas within the southern part of the basin. This pre-Burnside River period of uplift and erosion may be equivalent to similar episodes at the same stratigraphic position in the passive margin succession of the Wopmay and the Athapuscow Aulacogen.

With increased uplift in the source area(s) to the south and east of the basin, together with coeval subsidence of the Axial Zone of the basin, braided rivers spread coarse prograding sands and gravels throughout the southern Axial Zone and across the Western Platform. The distal equivalents of these clastics accumulated at the apex of the Axial Zone in the Taktu Aulacogen as the Mara Formation prograding fluvio-deltaic shallow silts and muds. With increasing lateral subsidence of the flanks of the Axial Zone of the basin, the sediments of the Burnside River Formation were deposited on the flanking Eastern Platform, and onto the northern margins of the Taktu Aulacogen on the present Victoria Island.

With decreasing sediment supply to the basin, and stabilization of the Kilohigok and Taktu, the fine grained equivalents of the fluvial sediments of the Burnside River – the Mara Formation – regraded into the basin as the shallow marine or fluvio-deltaic deposits accumulated in the basin and across the passive margin of the Wopmay Orogen to the west.

As subsidence of the basin ended, the entire Wopmay and attendant aulacogens were uplifted, and the regional paleosols formed on the top of the Mara, Tree River and sediments in the Athapuscow Aulacogen at the equivalent stratigraphic level. However, the Axial Zone of the basin continued to be a depocentre during paleosol formation, and the Quadyuk Formation stromatolitic carbonates were deposited in the trough and presumably around the rim of the sediment-starved trough.

Uplift north of the Taktu, as well as the Wopmay Orogen, possibly related to the initial stages of oblique(?) collision of the continent-microcontinent over the subduction zone, spread cannibalized detritus back into the remnant subduction trough as the Asiak wackes were deposited. The distal equivalents of these clastics accumulated in the Kilohigok as the precursors of the Peacock Hills Formation calcareous mudstones. With continued, perhaps collision-generated subsidence of the Kilohigok Basin and the Taktu, increasing amounts of these clastics were spilled into the basin and onto the slowly-subsiding carbonate platform. The Kuuvik stromatolitic carbonates were spreading across this intracratonic platform, and may have reached as far as the Athapuscow Aulacogen to the south. This carbonate "veneer" across the craton may have been the major source of the clastic carbonate which was intimately admixed with the terrigenous clastics of the Peacock Hills to form the upward-increasing carbonates of the formation as the Kuuvik prograded across its own fore-reef detritus.

With diminution of terrigenous clastics supplied to the basin, the Kuuvik Proto-Platform prograded into the Axial Zone of the basin and across the Western and Eastern(?) Platforms. Continuous rapid subsidence was greater than the accumulation/growth rate of the Kuuvik stromatolites in the northernmost part of the Wopmay shelf, and the Asiak wackes were continuously deposited in this area.

With rapid isostatic uplift, following detachment of the west-dipping subducted slab, the region north of the Taktu Aulacogen rapidly rose, and shed terrigenous detritus southward across the distal reaches of the Kuuvik Platform. These sediments were deposited as the silts and muds of the basal Brown Sound Formation.

With continuing southward-prograding and coarsening clastics supplied to the Kilohigok Basin, the entire region was buried beneath the fluvial sands, grits, and gravels of the Amagok Formation as deposition in the Kilohigok Basin ended.

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8. EARLY PROTEROZOIC LABINE GROUP OF WOPMAY OROGEN: REMNANT OF A CONTINENTAL VOLCANIC ARC DEVELOPED DURING OBLIQUE CONVERGENCE

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Abstract

The 1.87 Ga LaBine Group outcrops along the western margin of Wopmay orogen at Great Bear Lake and rests on a deformed and metamorphosed 1.92 Ga sialic basement complex. It is overlain by rocks of the mainly rhyodacitic Sloan Group. Syn- to post-volcanic plutons of the Great Bear batholith intrude both groups.

From Echo Bay northward to Hornby Bay the oldest rocks of the LaBine Group are mainly andesitic lavas, breccias, and pyroclastic rocks at least 3000 m thick, interpreted to be the remains of a number of large stratovolcanoes. Overlying, and in part interfingering with the stratovolcanoes, are seven major ash-flow tuff sheets, which are locally intercalated with andesite, dacite, rhyolite flows and domes, and a diverse assemblage of fluvial and lacustrine sedimentary rocks.

Sheets of ash-flow tuff include units thicker than 800 m deposited within cauldrons and thin cooling units deposited outside the cauldrons. Intercalated with intracauldron tuff are wedges of breccia and megabreccia, presumably derived from the walls of the cauldrons during subsidence.

Facies relations and the overall evolution of the field from early gas-poor andesitic eruptions to gas-rich eruptions of ash-flow tuff are closely comparable to Oligocene volcanic fields of the United States believed related to subduction. Rocks of the LaBine field were hydrothermally altered by high-level geothermal processes but on the basis of SiO_2, TiO_2, REE, and phenocryst mineralogy they can be classified as calc-alkaline. Therefore, it is concluded that the LaBine Group represents an early Proterozoic volcanic arc developed upon continental crust. Preserved stratovolcanoes and other high-level volcanic strata indicate that the LaBine Group was erupted into a basin which was subsiding concomitant with eruptions. The basin was probably generated in a wrench zone related to oblique convergence.

Laccoliths in Athapuscow Aulacogen together with recent geochronological and field data suggest that the LaBine Group postdates continent-microcontinent collision in Wopmay Orogen and was probably generated above an eastward-dipping Benioff zone which was either segmented or became shallower with time.

Résumé

Le groupe de Labine, qui date de 1.87 Ga, affleure le long de la marge ouest de l'orogène de Wopmay, dans la région du Grand lac de l'Ours, et repose sur un complexe rocheux de caractère sialique, déformé et métamorphisé, âgé de 1.92 Ga. Il est recouvert par les roches principalement rhyodacitiques du groupe de Sloan. Des plutons synvolcaniques à postvolcaniques du batholite du Grand lac de l'Ours traversent les deux groupes.

De la baie d'Echo au nord à la baie Hornby, les plus anciennes roches du groupe de Labine sont principalement des laves andésitiques, des brèches et des roches pyroclastiques d'au moins 3 000 m d'épaisseur, qui d'après certaines interprétations, seraient les restes de quelques grands stratovolcans. Sept vastes nappes de tufs répandues sous forme de coulées de cendres, qui recouvrent les stratovolcans et parfois présentent des interdigitations avec ceux-ci, sont localement intercalées dans des andésites, dacites, coulées et dômes rhyolitiques, et divers assemblages de roches sédimentaires d'origine fluviatile et lacustre.

Les nappes de tufs répandues sous forme de coulées de cendres contiennent des unités de puissance supérieure à 800 m, déposées à l'intérieur des caldeiras, et de minces coulées de lave, déposées à l'extérieur de celles-ci. Des formations en biseau, formées de brèches et mégabrèches, probablement dérivées des parois des caldeiras pendant la subsidence de celles-ci, sont intercalées avec les tufs de l'intérieur des caldeiras.

Les relations de faciès et l'évolution globale du terrain, d'abord soumis à des éruptions andésitiques pauvres en émanations gazeuses, puis aux éruptions gazeuses de tufs répandus en coulées de cendres rappellent fortement les secteurs volcaniques oligocènes des Etats-Unis, que l'on estime associés aux phénomènes de subduction. Les roches du secteur de Labine ont été altérées par des réactions hydrothermales intenses, mais en raison de leur teneur en SiO_2, en TiO_2 et REE et de la minéralogie des phénocristaux, on peut les classer dans les roches calcoalcalines. On en conclut donc que le groupe de Labine correspond à un arc volcanique d'âge protérozoïque inférieur, formé *au-dessus de la croûte continentale. Les stratovolcans et autres niveaux volcaniques créés par une activité volcanique intense, et encore conservés, indiquent que le groupe de Labine s'est écoulé lors d'éruptions dans un bassin, dont l'affaissement a accompagné les éruptions. Le bassin s'est probablement formé dans une zone de déchirement résultant d'une convergence oblique.*

Dans l'aulacogène d'Athapuscow, l'existence de laccolites et les récentes données géochronologiques et données obtenues sur le terrain semblent indiquer que le groupe de Labine est ultérieur à la collision entre continent et microcontinent qui a eu lieu lors de l'orogène de Wopmay, et a probablement été formé au-dessus d'une zone de Benioff plongeant vers l'est, qui s'est fragmentée ou est devenue moins profonde au cours des temps.

INTRODUCTION

In recent years the concept of plate tectonics has provided an actualistic framework to interpret the geology of continents, which contain the vast majority of recorded earth history. Geologists now examine pre-Mesozoic terranes with an eye toward determining an evolutionary scheme for the development of the earth based on similarities with, and dissimilarities to, post-Mesozoic terranes.

Studies of post-Mesozoic volcanic and plutonic rocks suggest it is possible to relate magmatic belts to specific tectonic settings. In fact, studies in Cenozoic volcano-plutonic terranes indicate that most can be related to lithospheric plate motions. Thus, igneous rocks may provide valuable insight into tectonic processes in older, cratonic regions where there is no sea floor record.

The purpose of this paper is to present stratigraphic and geochemical data from the LaBine Group, a 1.87 Ga volcanic field located along the eastern shore of Great Bear Lake, and to discuss the tectonic setting and regional implications.

PREVIOUS WORK

Bell (1901) first investigated the geology in the region of Great Bear Lake as part of a lengthy canoe reconnaissance for the Geological Survey of Canada in 1899. He noted "cobalt bloom and copper stain" along the east shore of the lake but was unable to investigate his discovery as he was under great pressure to reach civilization before freezeup.

The area received only minor attention from prospectors and trappers until 1930 when Gilbert LaBine, then president of Eldorado Mining Company, discovered high grade silver-pitchblende veins at the present townsite of Port Radium. The veins were mined for radium and silver until 1940 when the mine was shut down due to World War II and the resulting disruption of the radium market.

Kidd (1932) examined the mineral deposit for the Geological Survey of Canada in 1931. Kidd subsequently mapped much of the region at a scale of 1:250 000 (1933) and also made a broad reconnaissance of a 20 mile wide strip from Great Bear Lake to Great Slave Lake (1936). Smaller areas near Port Radium were mapped by Robinson (1933), Riley (1935), and Furnival (1939).

In 1941 Eldorado gave Enrico Fermi and associates at Columbia University 5 tons of uranium oxide for their experiments to generate a chain reaction and the mine was reopened to supply the strategic metal uranium to the United States Government. In 1944 the Canadian Government obtained ownership of the property and a program of 1 inch to 400 foot mapping in the vicinity of Port Radium was initiated by the Geological Survey of Canada (Joliffe and Bateman, 1944; Thurber, 1946; Feniak, 1947; Fortier, 1948). Later, Feniak (1952) mapped the MacAlpine Channel area at a scale of 1:50 000 while Lord and Parsons (1947) mapped the Camsell River region.

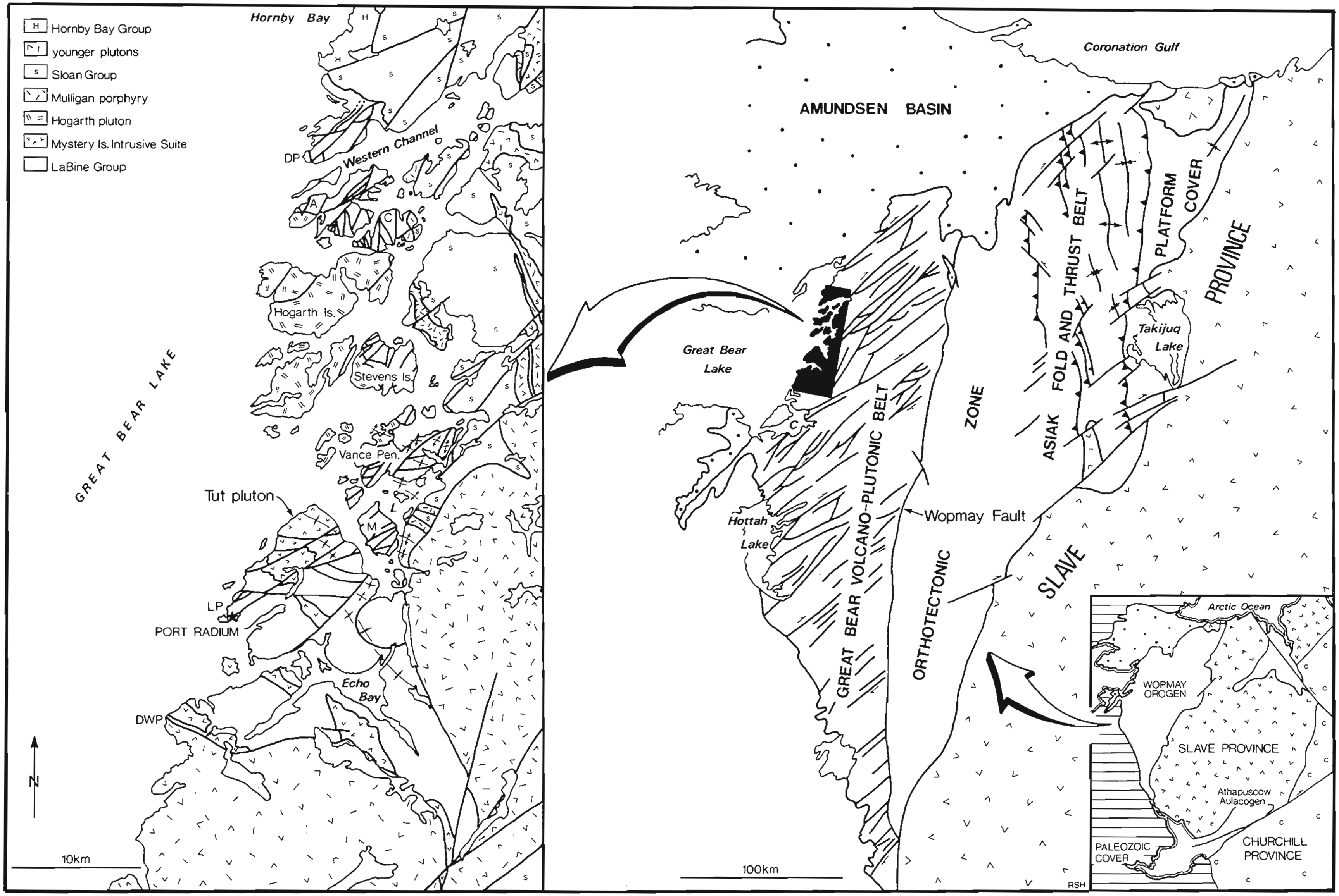
During the next 25 years geological work was mainly confined to detailed studies of the mineral deposits at LaBine Point (Campbell, 1955, 1957; Jory, 1964; Robinson, 1971; Robinson and Morton, 1972; Robinson and Badham, 1974) and in the Conjuror Bay-Camsell River region (Badham, 1972, 1973a, b, 1975; Badham et al., 1972; Shegelski, 1973; Badham and Morton, 1976). Mursky (1973) compiled much of the previously restricted data collected by the Geological Survey of Canada during the war.

Hoffman (1978) made the first comprehensive reconnaissance maps of the area during the middle 1970s and established the regional stratigraphy (Hoffman and McGlynn, 1977). Hoffman (1972) and Badham (1973a) first alluded to a subduction origin for the Great Bear Volcano-Plutonic Belt by pointing out the similarities of the Great Bear batholith with batholiths of the Andes.

REGIONAL SETTING

The LaBine volcanic field lies along the western margin of the Bear Structural Province in the northwestern Canadian Shield (Fig. 8.1). It is part of the Great Bear Volcano-Plutonic Belt (Fig. 8.1) which forms the western part of the early Proterozoic Wopmay Orogen (Fraser et al., 1972; Hoffman, 1973; Hoffman and McGlynn, 1977). The orogen developed along the western side of an Archean craton (Slave Province) between about 2.1 and 1.8 Ga. Hoffman (1980a) divided Wopmay Orogen into 4 tectonic zones: 1) a thin autochthonous cratonic cover and foreland basin sequence that unconformably overlies Archean basement; 2) a fold and thrust belt where rocks of the continental shelf and foredeep are thrust eastward relative to the craton; 3) an orthotectonic zone in which deformed initial rift clastic and volcanic rocks, passive continental slope-rise and foredeep rocks are metamorphosed and intruded by a multitude of syn- to post-tectonic, mesozonal S-type plutons; and 4) the little-deformed Great Bear Volcano-Plutonic Belt, consisting of subgreenschist facies volcanic and sedimentary rocks (McTavish Supergroup) which unconformably overlie a deformed and metamorphosed basement complex, and are intruded by tabular to sheetlike, epizonal I-type plutons of granitoid composition (Great Bear batholith).

High level plutonic rocks of the Great Bear batholith and the deformed basement rocks dominate the southern half of the Great Bear Volcano-Plutonic Belt while their consanguinous volcanic and sedimentary roof (McTavish Supergroup) is widely exposed in the northern half. The McTavish Supergroup is folded about gently-plunging, northwest-trending axes. These folds are asymmetric with the northeasterly dipping limbs generally being much larger (Hoffman, et al., 1976). Thus, the supracrustal rocks become progressively younger to the northeast, with the oldest rocks of the belt exposed only in the southwest. These relations suggest that the Great Bear Volcano-Plutonic Belt has a slight northeastward plunge.



C =	Conjuror Bay-Camsell River area	DP =	Doghead Point
DWP =	Dowdell Point	L =	Lindsley Bay
LP =	LaBine Point	A =	Achook Island
M =	Mackenzie Island	C =	Cornwall Island

Figure 8.1. *Maps showing study area (blackened), tectonic subdivisions of Wopmay Orogen (after Hoffman, 1980a), and northwestern Canadian Shield.*

The McTavish Supergroup is divided into 3 groups separated by unconformities: the LaBine Group, Sloan Group and the Dumas Group, in ascending order. The Sloan Group, exposed in the central part of the belt, consists mostly of thick sequences of densely-welded dacite and rhyodacite ash-flow tuff (Hoffman and McGlynn, 1977), perhaps cauldron fill. The Dumas Group exposed on the east side of the belt, comprises mafic lavas and mudstones cut by siliceous sills (S. Bowring, personal communication). The LaBine Group is a diverse assemblage of tholeiitic (Wilson, 1979) and calc-alkaline lava flows and pyroclastic rocks, plus a variety of sedimentary and high-level porphyritic intrusive rocks. The LaBine Group outcrops in 3 areas, all in the western part of the volcano-plutonic belt: one around Conjuror Bay and Camsell River; another at Hottah Lake; and along the eastern shore of Great Bear Lake from Echo Bay northward to Hornby Bay (Fig. 8.1). This last area is the principal subject of this paper.

An ash-flow tuff, high in the LaBine Group stratigraphy at Conjuror Bay, is 1.87 ± .01 Ga (Van Schmus and Bowring, 1980; and personal communication). U-Pb ages of plutons at Port Radium, at least one of which is demonstrably synvolcanic, are indistinguishable at present from the ash-flow at Conjuror Bay (Van Schmus and Bowring, 1980; and personal communication).

There are many exposures near Hottah Lake and at Conjuror Bay where the LaBine Group can be seen to unconformably overlie metamorphic rocks, here informally termed the metamorphic suite of Holly Lake. The suite is cut by many foliated granitoid plutons, one of which has yielded zircons dated as 1.92 ± .01 Ga (Van Schmus and Bowring, 1980). Together the metamorphic suite and the plutons collectively form what is known as the Hottah Terrane. This terrane was intensely deformed prior to deposition of the LaBine Group and has a prominent north-northeast – south-southwest penetrative fabric.

The Great Bear Volcano-Plutonic Belt is generally separated from the orthotectonic zone by the Wopmay fault (Fig. 8.1), although volcanic and sedimentary rocks of the Dumas Group locally overstep the fault and onlap the high grade rocks of the orthotectonic zone unconformably (Hoffman et al., 1976; Hoffman and McGlynn, 1977).

The entire Great Bear Volcano-Plutonic Belt is cut by numerous, nearly vertical, northeast-trending strike-slip faults (McGlynn, 1977) that postdate magmatism in the belt and that bend, splay, and die out towards the Wopmay fault (Fig. 8.1). These faults are but one part of a regional conjugate set of transcurrent faults that occurs throughout Wopmay Orogen (Hoffman, 1980b). Separations of units across these faults are typically hundreds of metres to several kilometers. Many of the fault zones are filled with quartz stockworks up to several hundred metres across (Furnival, 1935).

The Great Bear Volcano-Plutonic Belt is unconformably overlain by middle Proterozoic rocks of Amundsen basin (Hornby Bay Group) to the north (Fig. 8.1) and by Paleozoic sedimentary rocks of the Northern Interior Platform to the west.

STRATIGRAPHY

Introduction

The LaBine Group formed as a composite volcanic field upon sialic crust. Complex facies relations, tremendous variations in topographic relief, and long, varied eruptive histories are characteristics of such fields. The LaBine volcanic field is no exception, but discussion of all rock types and their structural relations is beyond the scope of this paper. Consequently, only the major stratigraphic units from Echo Bay to Hornby Bay will be discussed here, in an attempt to characterize the general volcanic evolution of this region and provide a broad overview of the volcano-tectonic environment.

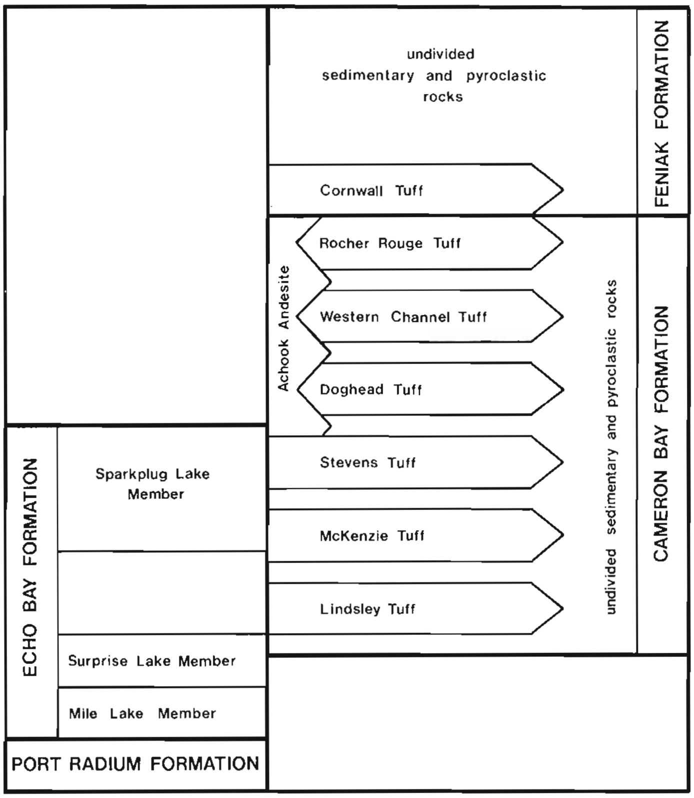


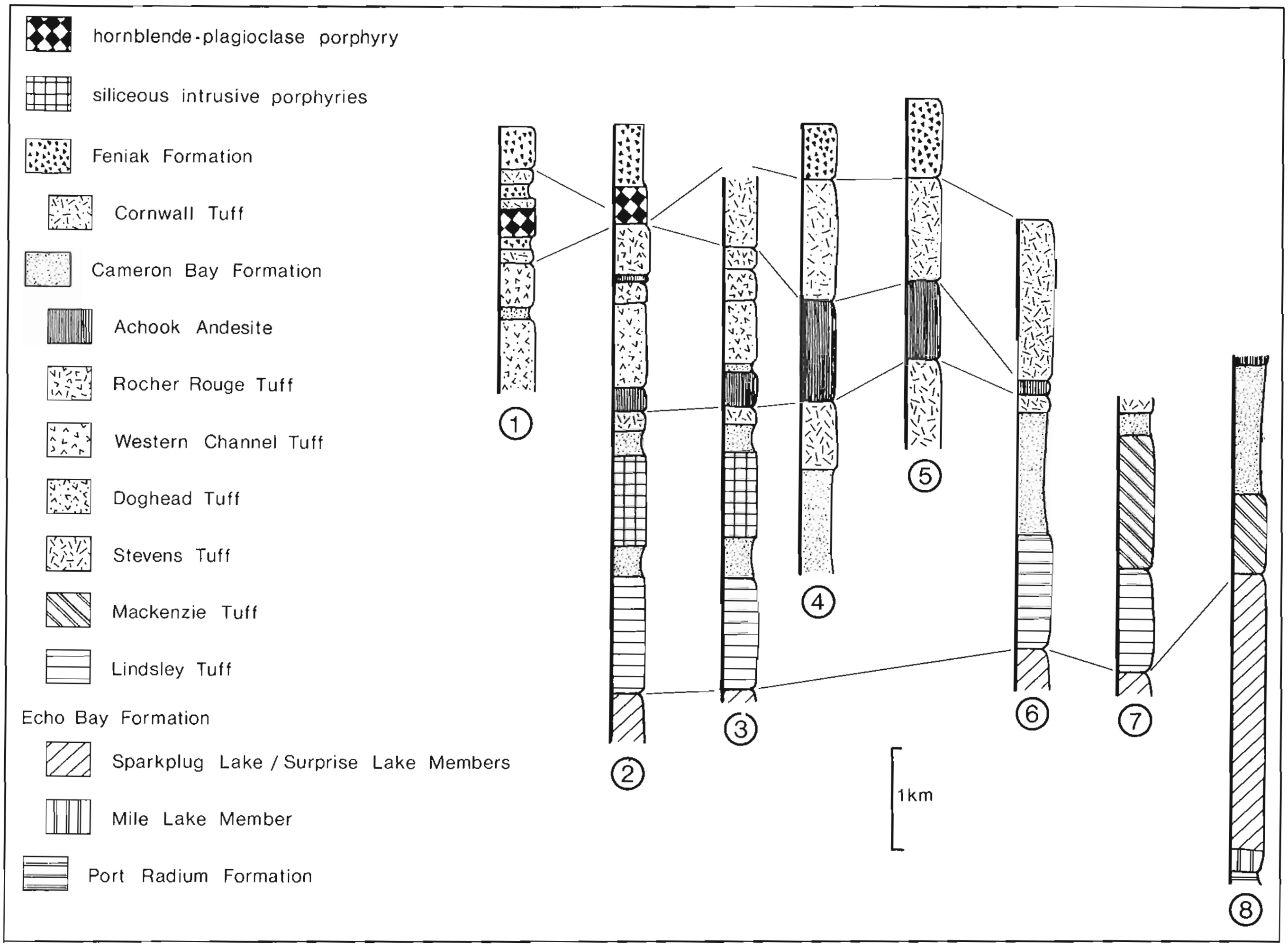
Figure 8.2. *Major stratigraphic subdivisions and nomenclature of the LaBine Group in the study area.*

In a crude way, the stratigraphy of the map area can be subdivided into two main eruptive phases: an early phase characterized by relatively gas-poor eruptions of andesitic lava represented by the Port Radium and Echo Bay formations, and a younger, more gas-charged, phase typified by voluminous eruptions of ash-flow tuff. The younger, siliceous volcanics are divided into the Cameron Bay and Feniak formations. The stratigraphic nomenclature used in this paper is shown in Figures 8.2 and 8.3.

Port Radium Formation

The Port Radium Formation is the oldest unit in the succession and is exposed only near Dowdell Point and on LaBine Peninsula (Fig. 8.4). The base of the formation is everywhere truncated by younger plutonic rocks of the Great Bear Batholith. These intrusions plus intense folding, brecciation, hydrothermal alteration, and intrusion by at least two additional suites of high-level plutons, make thickness estimates unreliable. The contact with the overlying Echo Bay Formation is placed at the base of the lowest lava flow.

Where undisturbed and relatively unaltered, Port Radium Formation consists predominantly of laminated to thinly bedded siltstone, sandstone, ashstone, and minor conglomerate of andesitic provenance. Relict sedimentary



1 = western Doghead Point
2 = eastern Doghead Point
3 = western Achook Island
4 = central Achook Island
5 = Cornwall Island
6 = Stevens Island
7 = Mackenzie Island
8 = Dowdell Point-Echo Bay

Figure 8.3. *Generalized columnar sections*

structures, such as ripple laminations, graded and convoluted bedding, and low-angle cross stratification, are common. Mudcracks were reported from this unit by Campbell (1955) but none were seen during the present investigation. Rocks initially described by Jory (1964) as "microcrystalline albite tuffs" are sediments and ash that were hydrothermally albitized during emplacement of the Mystery Island Intrusive Suite.

Particularly in the lower parts of the formation, calcareous laminae are abundant and there are at least two 1 m thick carbonate beds near Mile Lake; a similar bed was observed underground in the current mine workings on LaBine Peninsula.

In general, the formation coarsens upwards with granules of aphanitic to porphyritic andesite becoming abundant in the upper 100 m. In some locations a polymictic conglomerate near the top of the formation fills channels cut into the finer grained sedimentary rocks.

Echo Bay Formation

The Echo Bay Formation consists of a thick pile of andesite flows and breccias, sparse rhyodacite flows and breccias, intercalated epiclastic rocks and minor beds of reworked tuff that conformably overlie the Port Radium Formation. It is best exposed in a section from Dowdell Point to Echo Bay, where it is nearly 3000 m thick (Fig. 8.3, 8.4).

The formation is divided into 3 informal members – Mile Lake, Surprise Lake, and Sparkplug Lake members. The stratigraphically lowest member (Mile Lake) comprises 400 m of intercalated epiclastic rocks and lava flows while the overlying Surprise Lake member contains only minor beds of epiclastic rocks between lava flows. Andesite flows and breccias with abundant plagioclase phenocrysts that overlie Mackenzie Tuff are collectively termed the Sparkplug Lake member. This member includes a small, composite andesite cone and vent complex approximately 1 km in diameter located south of Lindsley Bay but stratigraphically higher in the section than most rocks of the Sparkplug Lake member.

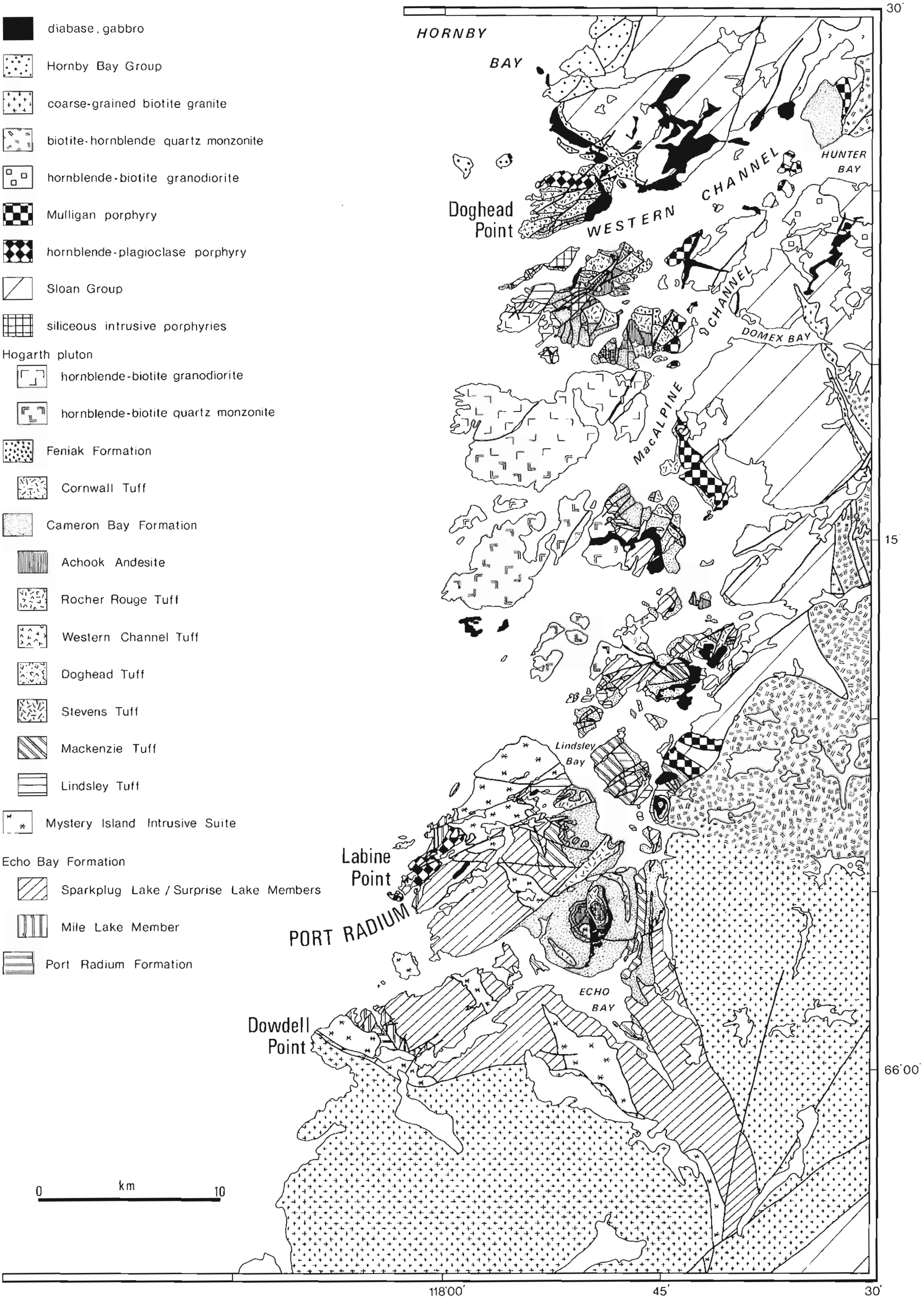


Figure 8.4. *Generalized geological map of the study area*

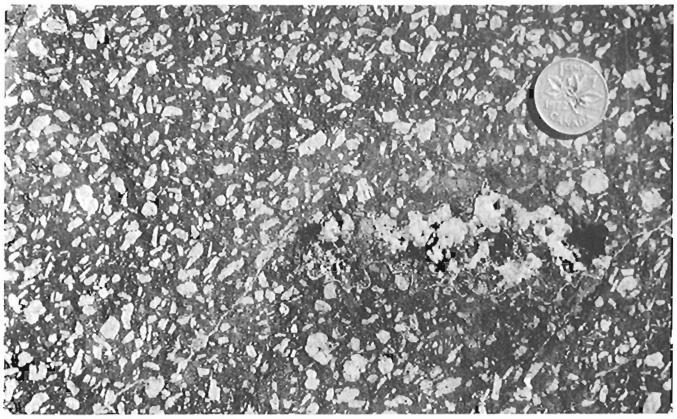


Figure 8.5. *Andesite of the Echo Bay Formation*

In general, lavas of the Sparkplug Lake member are distinguished from those of the Surprise Lake member only by their stratigraphic position.

The lava flows of the Echo Bay Formation are plagioclase porphyritic and the abundance of phenocrysts may range within individual flows from 5 to 40 per cent. Many of the flows show trachytic texture defined by platy plagioclase crystals up to 1 cm across (Fig. 8.5). Before alteration, most flows contained small pyroxene or hornblende phenocrysts, commonly rimmed with opaque Fe-oxides. Relict olivine euhedra occur in some of the lower flows.

All Echo Bay Formation lavas are altered to some degree: ferromagnesian minerals are often replaced by chlorite, opaque Fe-Ti oxides, epidote, and clay minerals. The matrix is commonly a fine grained mosaic of sphene, albite, quartz, chlorites, zeolites, and clay minerals.

Amygdules, commonly sparse but locally up to 20 per cent of the rock, are predominantly silica but in some flows located within alteration haloes around granitoid intrusions, contain mixtures of quartz, epidote, chlorites, pyrite and calcite.

In outcrop, flows are generally massive or columnar jointed but some have platy-jointed or brecciated bases. Brecciated bases grade upward into massive lava, and are often partially oxidized to a brick-red colour. Many individual flows have reddened flow breccias at their tops and margins.

Eruption centres for the lavas are not exposed and lay outside the map area. Flows that intertongue with sediments of the Cameron Bay Formation become thinner and sparser eastward, suggesting that one centre lay west of the map area. Similarly, flows that pinch out to the south and west, and interfinger with conglomerate in the Vance Peninsula region, were presumably erupted from centres to the east or north of the map area.

Epiclastic rocks of the Mile Lake member consist almost entirely of volcanic detritus and are coarse arkosic to lithic sandstone and clast-supported conglomerate, occasionally polymictic but normally dominated by clasts of subrounded to rounded andesite. Some massive conglomerate beds, up to several metres thick, are probably laharic as they contain blocks of a wide variety of size, shape, and rock type floating in a muddy or silty matrix. Many of the sandstones display normal grading and comprise imbricated blocky plagioclase crystals with rounded corners and semispherical augite grains. These beds are probably reworked crystal tuff. Ripple lamination and crossbedding are commonly preserved in the sandstones but exposures are insufficient for paleocurrent analysis.

Cameron Bay Formation

The Cameron Bay Formation is a varied assemblage of volcanic and sedimentary rocks that overlies the Echo Bay Formation, except in the southern part of the area, where both formations interfinger. As used in this paper, the Cameron Bay Formation comprises clastic rocks, rhyolite flows, 6 major units of ash-flow tuff, and several andesite complexes. Major ash-flow tuff units are assigned informal member status within the Cameron Bay Formation. Ash-flow terminology is that of Smith (1960). For convenience, epiclastic rocks stratigraphically near ash-flow tuff members are discussed with those members.

From Vance Peninsula south to Echo Bay massive to poorly-bedded andesite bouldery conglomerates and breccias interfinger with the Echo Bay Formation. These clastics are interpreted as fluvial and debris-flow deposits that formed parts of the volcaniclastic aprons on the vent flanks from which Echo Bay lavas were erupted. An unnamed, flow-banded and flow-folded rhyolite (Fig. 8.6) extrusive with abundant silica-lined cavities to 8 cm occurs on top of gritty sandstone on Mackenzie Island. The eruption centre for this flow is unknown.

Lindsley Tuff Member

The Lindsley Tuff is discontinuously exposed over much of the belt. It is absent in the southwest, overlies Echo Bay Formation in the Lindsley Bay region, and directly overlies both Echo Bay andesite or conglomerate, breccia and rhyolite of the Cameron Bay Formation on Vance Peninsula, Stevens Island, and Achook Island. The Lindsley is a maximum of 1000 m thick and is a single, compositionally zone cooling unit composed of many flow units. Intensely fractured and broken quartz phenocrysts (up to 20%) dominate the lower ash-flows. Altered plagioclase and mafic phenocrysts become progressively more abundant upward in overlying ash-flows. Lithic fragments constitute at most only a few per cent of the unit and are dominantly andesite.

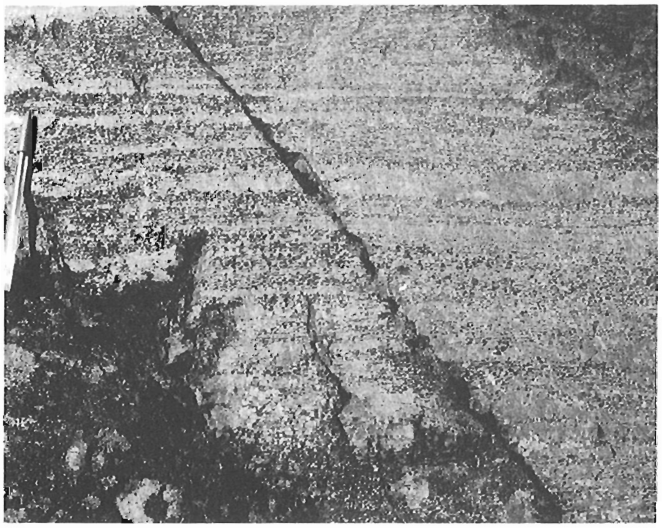


Figure 8.6. *Flow-banded rhyolite flow of the Cameron Bay Formation.*

The tuff is densely welded throughout and pumice is generally not recognizable, probably due to postdepositional recrystallization and alteration. However, on Achook Island, there is a zone with well-developed eutaxitic structure about 35 m above the contact with the underlying Echo Bay Formation.

On eastern Vance Peninsula and northeast of Echo Bay the tuff is always less than 30 m thick and is often absent. Channels, now filled with sandstone and conglomerate, were incised in the tuff after deposition. On Mackenzie Island, 2 km away from Vance Peninsula in pre-transcurrent fault reconstructions, 500 m of Lindsley Tuff is overlain by Mackenzie Tuff with no evidence of erosion. One kilometre to the west of Mackenzie Island, the Lindsley Tuff is overlain by epiclastic rocks and a rhyolitic flow or dome at least 1 km in diameter. Elsewhere, as on Stevens Island and Achook Island, the tuff is overlain by coarse sedimentary rocks, but with no apparent extensive channelling.

Massive to poorly-bedded wedges of breccia containing blocks up to 3 m across are interbedded with the Lindsley Tuff in the Lindsley Bay region. These are interpreted as talus breccias and indicate considerable topographic relief during deposition. The inferred high relief may account for the rapid lateral thickness variations of the tuff.

The abrupt pinchout of thick sections of tuff, coupled with the presence of talus breccias suggests that the Lindsley Tuff ponded against a topographic barrier. As there is no indication of a topographic barrier in the underlying sedimentary rocks, it must have developed during eruption of the tuff.

One possible mechanism to explain these relations is cauldron collapse concurrent with the ash-flow eruptions. The thick sections could represent intracauldron deposition while the thin sections may be remnants of the outflow sheet. The talus breccias would have been shed from the high-standing wall of the cauldron. Lipman (1976) described similar breccias intercalated with intracauldron tuffs in several cauldrons in the San Juan volcanic field of southwestern Colorado. He attributed the breccias in these cauldrons to landslides that resulted from the caving of the steep cauldron margins.

Mackenzie Tuff Member

The Mackenzie Tuff is a composite ash-flow tuff sheet that contains abundant foreign rock fragments and less than 10 per cent phenocrysts of quartz, altered potassium feldspar, plagioclase and sparse ferromagnesian minerals. Eutaxitic texture is commonly well developed, especially on Mackenzie Island (Fig. 8.7). On Mackenzie Island the upper cooling unit contains abundant accretionary lapilli (Fig. 8.8).

The Mackenzie Tuff is exposed only in the southern half of the belt. It is at least 1 km thick on Mackenzie Island and consists of 3 cooling units, with no sediment interbeds. South of Mackenzie Island the cooling units are separated by lenses of ripple-laminated siltstone and mudstone or by flows of the Sparkplug Lake Member of the Echo Bay Formation. The cooling units fill broad paleovalleys and pinch out over local paleo high areas. On Vance Peninsula there are at least 6 cooling units, each less than 20 m thick, interbedded with arkosic sandstone and pebbly gritstone.

From Vance Peninsula north to Doghead Peninsula the Mackenzie Tuff is absent and its stratigraphic position is occupied by a thick pile (1 km) of sedimentary rocks. This sequence is interpreted as a series of braided stream, alluvial fan and lacustrine complexes. Hematitic, polymictic and polymodal conglomerate of mainly volcano-plutonic provenance dominate this interval. Locally, however, there are beds of conglomerate containing 90 per cent well-rounded orthoquartzite clasts. Volcanic lithic and feldspathic sandstones are planar or trough crossbedded and commonly contain mudchips. Ripple-laminated siltstone, sometimes with mudstone drapes, occur throughout the sequence. Local mudstones, occasionally with mudcracks may mark the sites of lacustrine sedimentation or in some cases, the distal ends of debris flows.

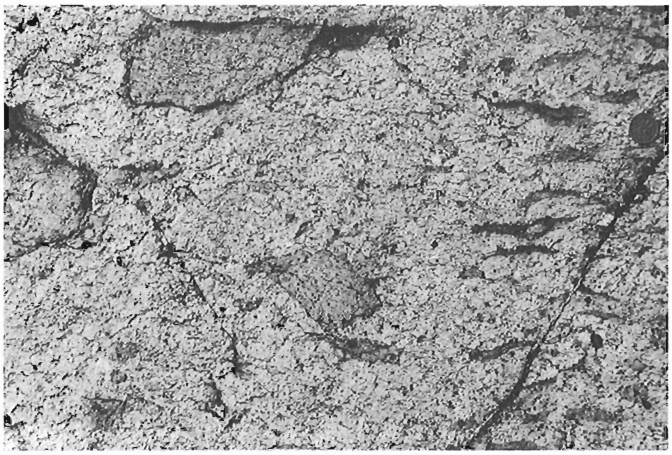


Figure 8.7. *Moderately-welded Mackenzie Tuff*

Figure 8.8. *Accretionary lapilli in the Mackenzie Tuff*

Devitrified ashstone beds (to 2 m) are common, especially on Vance Peninsula. Remnants of the beds, occurring as freestanding pinnacles of ashstone surrounded by sandstone, indicate they were channelled and eroded before deposition of the sandstones. The ashstone beds may represent coignimbrite ashes of the type described by Sparks and Walker (1977).

Directly overlying the Mackenzie Tuff west of Lindsley Bay is a 50 m coarsening-upward succession consisting of basal mudstone to crossbedded sandstones and pebbly sandstone, overlain by bouldery polymictic conglomerate. The entire sequence is capped by andesite lavas of the Sparkplug Lake Member of the Echo Bay Formation. The finely laminated mudstones, with lenses of fine grained sandstone, are locally mudcracked, and are likely lacustrine in origin. Sandstones overlying the mudrocks are medium- to coarse-grained and commonly gritty. Some beds are graded and contain imbricated pebbles that become smaller upsection. These beds are typically about 2 m thick and are overlain by trough and planar crossbedded sandstones. This part of the succession is interpreted as braided stream deposits. Clast-supported, crudely bedded bouldery polymictic conglomerates, that overlie the sandstones, probably represent alluvial fan deposits. The entire sequence is interpreted as a prograding alluvial fan complex, related to renewed volcanism from volcanic centres of the Echo Bay Formation, over lacustrine sediments.

Stevens Tuff Member

Stevens Tuff member consists of moderately to densely welded ash-flows that form one cooling unit characterized by large partly resorbed phenocrysts of quartz. Lithic fragments are ubiquitous near the base of the unit on Cornwall Island, Achook Island and Doghead Point, and commonly comprise up to 30 per cent of the rock. Pumice fragments are common, but in thick sections are obscured by alteration and welding. On Achook Island at least 75 thin ash-fall tuff beds underlie the Stevens Tuff (Fig. 8.9).

The tuff occurs throughout the belt from Echo Bay to Doghead Point. It is 20 m thick on Doghead Point and thickens gradually to the east-southeast. On Achook Island, the tuff is 400 m thick and farther east, on Cornwall Island, an incomplete section of several hundred metres outcrops in individual fault blocks. The Stevens Tuff thins drastically against a northeast-trending transcurrent fault on the eastern end of Cornwall Island. South and east of this fault, the tuff, which is less than 100 m thick, commonly appears to fill paleovalleys or is extensively channelled and eroded. North and west of the fault there is no evidence of a structural break or great topographic relief in the underlying sedimentary rocks, so it is unlikely that the fault was active before ash-flow tuff eruptions. The fault is interpreted to have been active during eruption of the ash-flows and later reactivated during strike-slip faulting.

It appears that the Stevens Tuff was deposited in a low-lying area that deepened to the southeast and was bounded by a major structural break, against which ash-flows ponded to a considerable thickness. The most probable explanation for these relations is that the Stevens Tuff was erupted concurrent with cauldron collapse.

As the central cauldron block appears to have been fault-bounded on only one side, it is reasonable to conclude that the central block subsided in a "trap-door" fashion; that is, bounded on one side by a large fault and on the other by a monoclinal flexure. Trap-door subsidence of central blocks in Cenozoic calderas has been documented by several workers (Seager, 1973; Seager and Clemons, 1975; Lambert, 1974; Steven and Lipman, 1976; Elston et al., 1976).

Achook Andesite Member

Amygdaloidal, sparsely-porphyritic lava flows, tuff, and breccia of andesitic composition that overlie Mackenzie Tuff from Echo Bay to Doghead Point are collectively termed the Achook Andesite. These flows and breccias contain abundant amygdules and sparse phenocrysts of altered plagioclase and amphibole. In these respects the Achook Andesite is different from the Echo Bay Formation which is sparsely amygdaloidal phenocryst-rich andesite. Lava flows in the Achook are generally less than 10 m thick, and commonly flow-banded. Amygdules are commonly mixtures of blood-red and snow-white chalcedony and in some flows are so common that they outline flow folds.

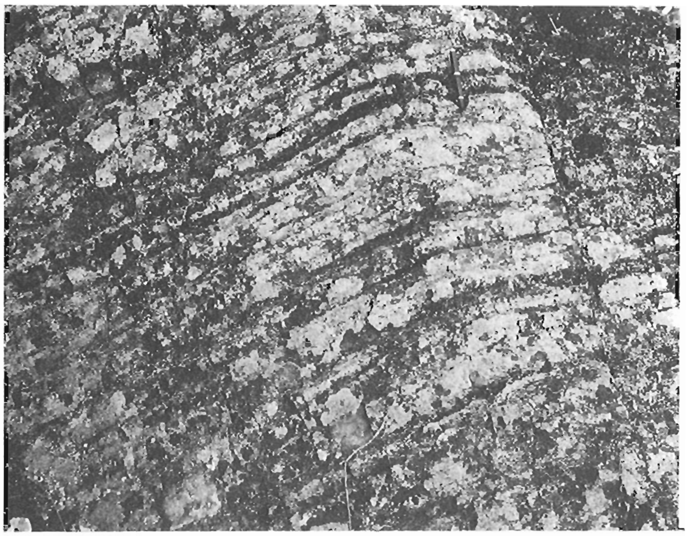


Figure 8.9. *Ash-fall beds at the base of Stevens Tuff. Note pen for scale.*

The Achook Andesite is intercalated with several ash-flow sheets (Fig. 8.3) which indicates that andesitic eruptions occurred sporadically over a time span sufficient to deposit several major ash-flow sheets, and that different volcanoes in the area were active concurrently.

The thickest sections (up to 1 km) of the Achook Andesite occur on Cornwall Island and the unit thins rapidly away from this area. Intensely altered andesite breccia several hundred metres thick on Achook Island may mark the site of an eruptive centre. The breccias, consisting of andesite blocks of varying sizes in a fine, comminuted matrix of andesite microbreccia and broken crystals, are interpreted as explosion breccias.

On Doghead Point lapilli tuff and ashstone contain normal and reverse graded beds. Contacts between these beds are commonly gradational. The beds are probably products of Strombolian-type eruptions. The lack of sharp contacts, coupled with the graded bedding, indicates that eruption and deposition went on more or less continuously. The graded bedding developed as eruptive strength waxed and waned and/or as wind velocity and direction fluctuated (Fisher, unpublished manuscript), or possibly as conduit size changed during eruption (Wilson et al., 1980).

Doghead Tuff

The Doghead Tuff is a single crystal-rich cooling unit that contains up to 30 per cent broken phenocrysts of highly altered hornblende, biotite, and plagioclase (Fig. 8.10). It is exposed only on Achook Island and Doghead Point. Most of the unit is very densely welded, with highly flattened pumice fragments near the base and top. Flattened blocks (to 50 cm) on Doghead Point contain inclined tension fractures, which indicate post-emplacement flowage of the tuff (Schmincke and Swanson, 1967). The orientation of the fractures indicates that movement was toward the northeast.

On Achook Island the tuff overlies an eastward-thickening wedge of bouldery, polymictic conglomerate and breccia, thin ash-flows with similar mineralogy to the Doghead Tuff, and andesite flows. Interbeds of thin ash-flows in the conglomerates indicate that eruptions began during conglomerate deposition.

Figure 8.10. *Densely-welded, crystal rich zone of Doghead Tuff.*

The Doghead Tuff is a minimum of 700 m thick on Achook Island. Both the tuff and underlying conglomerate pinch out eastward in less than 400 m against a series of closely-spaced syndepositional (?) faults which were subsequently reactivated during postvolcanic deformation (Fig. 8.11). The tuff thickens toward the west and northwest and is over 1 km thick on Doghead Point.

The above relations suggest that the Doghead Tuff was deposited within a depression and ponded against a topographic barrier. The exact nature and cause of the depression is unknown but a caldera origin is suspected.

Western Channel Tuff Member

The Western Channel Tuff consists of a single brick-red weathering cooling unit that varies from moderately to densely welded and displays spectacular eutaxitic texture (Fig. 8.12) and prominent columnar joints. The lower 2 m is everywhere drift or talus covered, which suggests the unit has an unwelded base.

Western Channel Tuff is exposed only on Doghead Point and Achook Island. It is everywhere porphyritic, but crystals of altered plagioclase, potassium feldspar, biotite and quartz together generally do not exceed 10 per cent of the rock, except near the top where they occupy nearly 20 per cent by volume. Plagioclase is absent in the lower third of the tuff, but becomes abundant in the upper two thirds.

The tuff is over 500 m thick on Doghead Point and thins to about 75 m on Achook Island. On Achook Island the Western Channel overlies a westward-thickening wedge of gritstone, sandstone, and bouldery polymictic conglomerate.

Rocher Range Tuff

Rocher Range Tuff is exposed only on Doghead Point where it is at least 300 m thick. The top of the member is not exposed. It is not present on Achook Island, only 3 km away. The tuff is densely welded, locally flow banded, and pumice seldom occurs, perhaps due to postdepositional recrystallization. The lower 30 m are clogged with andesite fragments of unknown provenance up to 1 m across. Phenocrysts form 15 to 30 per cent of the rock and are altered plagioclase, hornblende, and biotite.

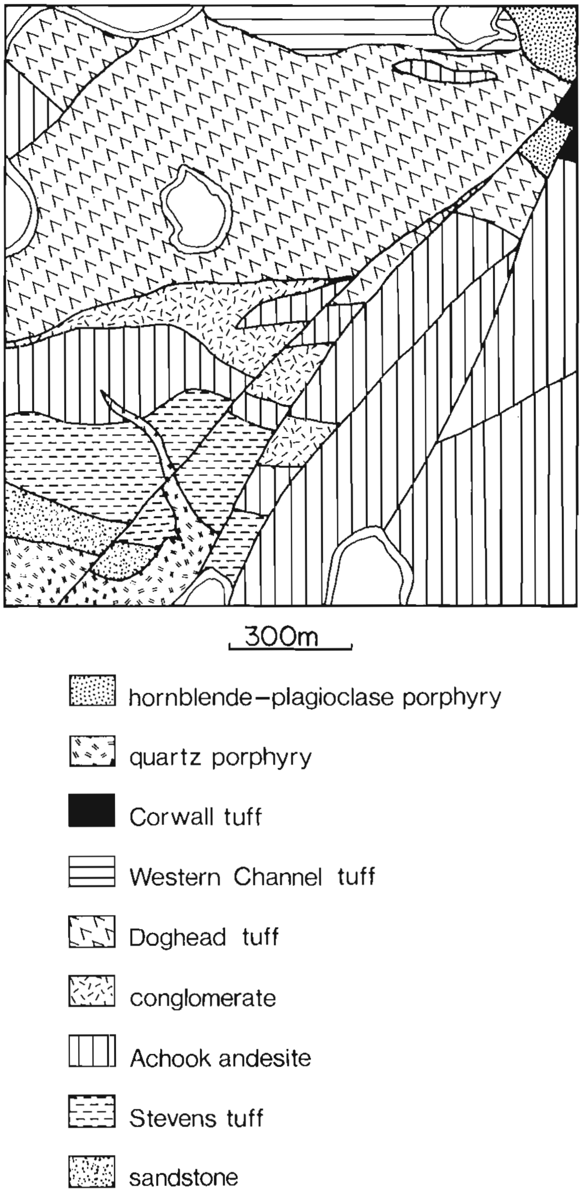


Figure 8.11. *Geological sketch map of north-central Achook Island illustrating relations at postulated cauldron margin. All units are dipping steeply to the north-northeast. Note thickness variations in the Doghead Tuff, believed to have been erupted concurrently with subsidence.*

Feniak Formation

The Feniak Formation occurs throughout the region from Vance Peninsula to Doghead Point. It is best exposed north of Stevens Island. As defined here, the unit includes all extrusive and sedimentary rocks between the base of the Cornwall Tuff and the Sloan Group. Contained within this interval are one major ash-flow tuff sheet (Cornwall Tuff), a thick, stubby dacite flow, and a diverse assemblage of waterlaid crystal tuff, devitrified ashstone, thin ash-flow sheets, fine grained epiclastic rocks, and rare beds of stromatolitic dolomite. It differs from the Cameron Bay Formation in that pyroclastic rocks (i.e. ashstone, crystal tuff and ash-flow tuff) make up the majority of the interval and coarse epiclastic rocks are relatively minor.

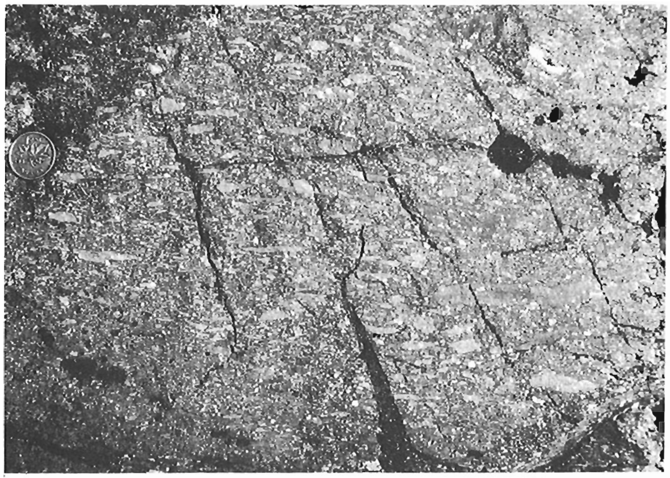


Figure 8.12. *Eutaxitic texture typical of the Western Channel Tuff.*

Cornwall Tuff Member

The lowermost unit of the Feniak Formation is the Cornwall Tuff Member, a non- to densely-welded composite ash-flow sheet containing 5 to 15 per cent altered plagioclase, hornblende, quartz, and potassium feldspar phenocrysts. The unit is well exposed on Achook Island and on Cornwall Island where it is over 1 km thick and highly propylitized. North of Stevens Island the tuff could be as thick as 1.8 km, but the central portion of the unit is not exposed and continuity cannot be demonstrated.

On Achook Island the Cornwall Tuff contains a 4 m thick, probably lacustrine, stromatolitic dolomite bed which suggests that the tuff at this locality is composed of at least 2 cooling units. On Doghead Point the Cornwall Tuff consists of three, or possibly four, thin cooling units intercalated with a thin sequence of diverse epiclastic rocks and a stromatolitic dolomite bed similar to that on Achook Island.

This epiclastic sequence consists, in its lower parts, of finely laminated sandstone and mudstone interbedded with thin beds of cryptalgal limestones, commonly with well-developed tepee structures. Numerous beds of devitrified ashstone also occur in this section. These weather various shades of pink, white and green and range from 30-70 cm thick. They are interpreted to be of airfall origin and may represent co-ignimbrite airfall deposits.

Fine grained arkosic sandstones and wedges of volcano-polymictic conglomerate, which pinch out or thicken at syndepositional faults are also present in this lower part of the section.

Clastic rocks in the middle part of the section contain numerous slump folds and lenses of water-laid crystal tuff. Overlying these beds are 100-150 m of interbedded crystal tuff and ashstone.

A 1 km thick dacite flow overlies the Cornwall Tuff on both Achook and Cornwall islands. It is generally highly altered but where alteration is low, it is plagioclase porphyritic and contains a flow-banded base.

The thickness of much of the Cornwall Tuff suggests that it is intra-cauldron facies tuff that ponded within a topographic depression created by the subsidence of a central block during the ash-flow eruptions. The thin, multiple cooling units preserved on Doghead Point are likely the remains of the outflow sheet.

Intrusive Rocks

A wide variety of intrusive rocks are exposed in the belt, but only the largest, geologically most significant, bodies are discussed here. For information regarding the intrusives not discussed in this paper, the interested reader is referred to Geological Survey of Canada Open File 709 (Hildebrand, 1980). Modal mineral proportions were estimated in the field and the nomenclature follows that recommended by Streckeisen (1973). The noun porphyry, as used in this paper, refers only to intrusive rocks which consist of phenocrysts in an aphanitic groundmass.

Cobalt Porphyry

Podiform to irregular-shaped intrusions of hornblende-plagioclase porphyry and microdiorite, collectively termed the Cobalt porphyry, are abundant at LaBine Point and cut only the Port Radium and lower Echo Bay formations. The Cobalt porphyries are of unknown age but as they are lithologically similar to the host andesite lavas of the lower Echo Bay Formation, they are interpreted as subvolcanic magma chambers from which some of the stratigraphically higher lava flows were erupted.

Brecciated zones up to 2 m wide occur within individual Cobalt porphyry bodies adjacent to the wall rocks. The matrix between the blocks consists of unbrecciated porphyry or finely brecciated and comminuted porphyry. Wall rocks near the contacts are also brecciated.

The brecciation of both the porphyries and the wall rock may have commenced before the porphyries had completely crystallized and could reflect the inflation and deflation that would result if these bodies were subvolcanic magma chambers that vented at the surface. As the chambers were emptied during eruption, solidified magma at their outer margins would be fractured and broken as the magma chambers collapsed. Alternatively, the breccias could be the product of steam explosions if the sediments were wet when the porphyries were intruded. There appears to be less brecciation where the porphyries intrude lava flows. Thus, the steam explosion concept may be of local importance, but the occurrence of breccias in both areas suggests that both mechanisms did occur.

Mystery Island Intrusive Suite

The Mystery Island Intrusive Suite comprises several (Fig. 8.4) semi-concordant sheets of medium grained diorite, quartz syenite, and granodiorite. They are widely distributed throughout the southern half of the map area and intrude both the Port Radium and Echo Bay formations. Characteristic of these intrusions are alteration haloes up to 2 km wide, comprising an inner bleached and albitized zone, a central zone of apatite-actinolite-magnetite pods, breccias, veins, and replacement, and an outer zone of chalcopyrite and pyrite gossan.

One member of this suite, the Tut pluton (Fig. 8.1), is likely contemporaneous with LaBine volcanism because it intrudes the Echo Bay Formation and conglomerate of the Cameron Bay Formation southwest of Lindsley Bay contains abundant clasts up to 1 m in diameter, of diorite and quartz monzonite identical in grain size, texture and lithology to phases of the Tut pluton. Furthermore, paleocurrents in associated sandstones show transport directions to the east, away from the pluton (Fig. 8.13). The conglomerate is overlain by the Stevens Tuff, indicating that the pluton was unroofed before the tuff was deposited.

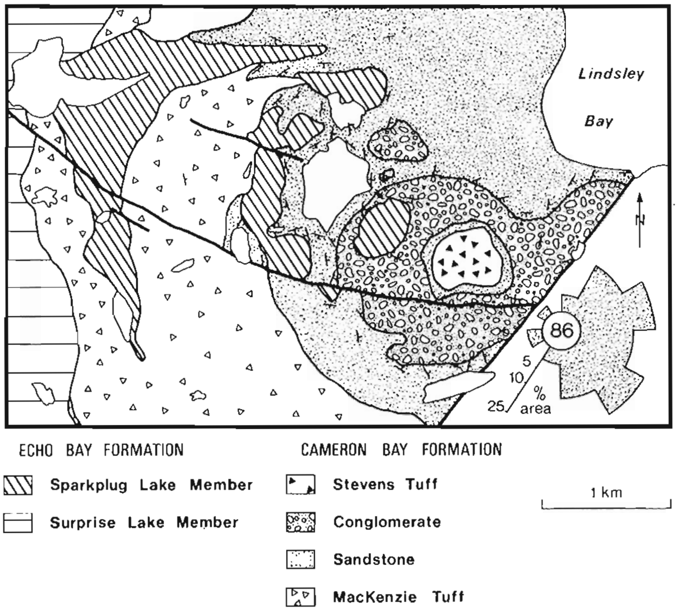


Figure 8.13. *Geological sketch map of an area west of Lindsley Bay showing intercalation of the Echo Bay Formation with the Mackenzie Tuff. Note westward pinchout of sandstone and conglomerate. The Tut pluton is located 1 km west of this figure. A paleocurrent rose diagram for the sandstone underlying the conglomerate rich in Tut pluton clasts is also shown.*

Subvolcanic Porphyries

Porphyries of varied compositions outcrop in the region. Most common are biotite-quartz and hornblende-plagioclase porphyries of unknown age which often intrude a thick conglomeratic horizon above the Lindsley Tuff. Another important group of porphyries, which Hoffman (1978) termed the Mulligan Porphyries, intrudes the Sloan-LaBine contact in several parts of the belt (Fig. 8.4). They are sill-like bodies of plagioclase-quartz porphyry believed by Hoffman (personal communication) to be partly coeval with Sloan volcanism.

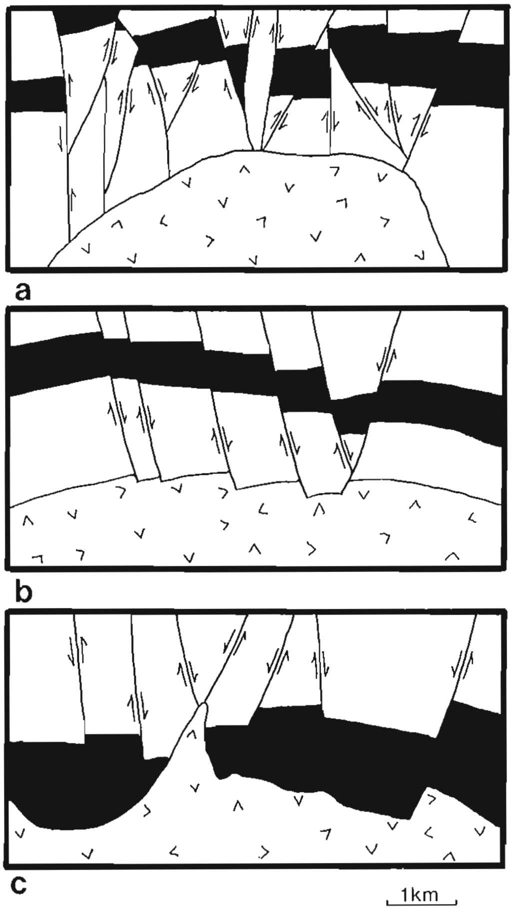
A distinctive plagioclase-potassium feldspar-quartz porphyry outcrops on Cornwall Island, where it intrudes and intensely alters conglomerate of the Cameron Bay Formation. The overlying Stevens Tuff locally contains up to 30 per cent lithic fragments identical to this porphyry. If they were derived from the porphyry, then the porphyry was intruded to within 1.5 km of the surface – the stratigraphic separation between it and the Stevens Tuff.

Hogarth Pluton

The Hogarth pluton intrudes volcanic and sedimentary rocks of the LaBine Group and is exposed from Vance Peninsula northward to Achook Island (Fig. 8.4). It consists of medium grained hornblende-biotite (chlorite-epidote), granodiorite and monzogranite. The granodiorite generally occurs in the upper portions of the pluton, while monzogranite dominates the lower part. Contacts with the wall rock are invariably razor-sharp and alteration is minimal. No miarolytic cavities were found. Xenoliths, of partly digested country rock up to 0.5 m across, are sparse.

A group of block faults that cuts rocks of the LaBine Group occurs above the roof of the Hogarth pluton on Achook Island, Cornwall Island and on Stevens Island. These faults typically have different trends than postvolcanic transcurrent faults or their splays and do not cut the Sloan Group except where reactivated by the younger transcurrent faults. The early faults must predate the Sloan Group because one is left laterally separated on Doghead Peninsula by another fault, probably dip-slip with west side down, which is overstepped by ash-flow tuff of the Sloan Group.

The faults are truncated by the Hogarth pluton near its apex but at deeper structural levels they penetrate the outer shell of the pluton. These relations are interpreted to indicate that the faults were active synchronously with emplacement of the Hogarth pluton and that magma near the margins of the lower part of the intrusion had already crystallized when the uppermost portions of the pluton were emplaced. If this interpretation is correct then the Hogarth pluton must predate the Sloan Group, and barring significant pre-Sloan Group erosion of the LaBine Group the maximum depth of emplacement of the pluton is approximately the stratigraphic thickness between its roof and the LaBine-Sloan contact – about 2.5 km.



a) a map view of the Cornwall cauldron (Hogarth pluton)

b) interpretive cross section of Valles caldera (Smith et al., 1970)

c) interpretive cross section of Timber Mountain caldera (Byers et al., 1976)

Figure 8.14. *Comparison of cross-sections through the central portions of 3 large resurgent calderas. A stratigraphic unit in each area has been blackened to show the doming and development of the central grabens. Note the occurrence of both normal and reverse faults above the Hogarth pluton in a.*

Interestingly, the faults are topographically coincident with the thickest and most altered parts of the Cornwall Tuff, suggesting that the tuff, faults and pluton are genetically related. Structural relations above the Hogarth pluton display striking similarities to resurgent domes of large collapse calderas in other volcanic fields. Blocks above its roof are jostled and lifted and there is a graben located in the central part of this uplift (Fig. 8.14). Similar relations are present in resurgent domes of Creede Caldera (Steven and Ratte, 1973; Steven and Lipman, 1973), Valles Caldera (Smith et al., 1970), Long Valley Caldera (Bailey, 1976; Bailey and Koeppen, 1977), the Timber Mountain Caldera (Byers et al., 1976), and many others. The lack of miarolytic cavities or associated pegmatites in the Hogarth pluton which intruded within a few kilometres of the surface, suggests that the pluton had already lost most of its volatiles before final emplacement. A possible mechanism for their loss appears to be voluminous ash-flow tuff eruptions which resulted when volatile pressure exceeded the containment capability of the roof.

Later Granitoids

Several granitoid plutons that postdate the Sloan volcanics occur within the region covered by this paper. They are biotite-hornblende (chlorite-epidote) monzogranite and granite, typically with narrow alteration haloes and sharp contacts. These plutons were emplaced by block stoping and wall-rock assimilation. Xenoliths of partly digested country rock are common in some outcrops, most notably at LaBine Point, as are miarolytic cavities lined with pegmatite (Fig. 8.15). Contacts with the wall rock are invariably razor-sharp and leucocratic border phases are common (Fig. 8.16).

Figure 8.15. *Outcrop near contact of granitic pluton showing abundant partly digested xenoliths and pegmatite-lined miarolitic cavities.*

VOLCANIC EVOLUTION

The fine grained, locally mudcracked, volcaniclastics of the Port Radium Formation were deposited in a lacustrine environment and are interpreted as material from growing, but distant volcanoes. With further growth and development of the volcanoes, alluvial fan complexes, preserved as the Mile Lake Member of the Echo Bay Formation, prograded across the lacustrine beds and in turn were succeeded by thousands of metres of monotonous andesite flows erupted from at least two volcanic centres (Fig. 8.17). Similar facies relations related to growth and progradation of volcanoes have been described by Williams and McBirney (1979), Clemons (1979), Lipman (1975), and Smedes and Prostka (1972).

Andesitic volcanism had waned, but had not ended, when the first of several major ash-flow sheets was erupted and ponded within a steep-walled topographic depression. Of seven major ash-flow sheets, at least three can be related to cauldrons (Fig. 8.18). Cauldron subsidence coeval with ash-flow eruptions is demonstrated by the order of magnitude thickness variations of the ash-flow tuff sheets. However, the size and shape of the cauldrons are indeterminable because of post-eruptive tectonic complications, including two episodes of granitic plutonism, folding about northwest-trending axes, and separation by a multitude of northeast-trending transcurrent faults.

Figure 8.16. *Contact of granitic pluton. Note leucocratic border phase.*

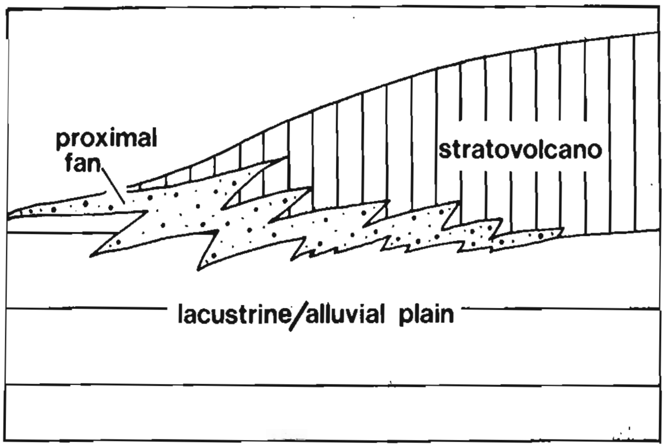
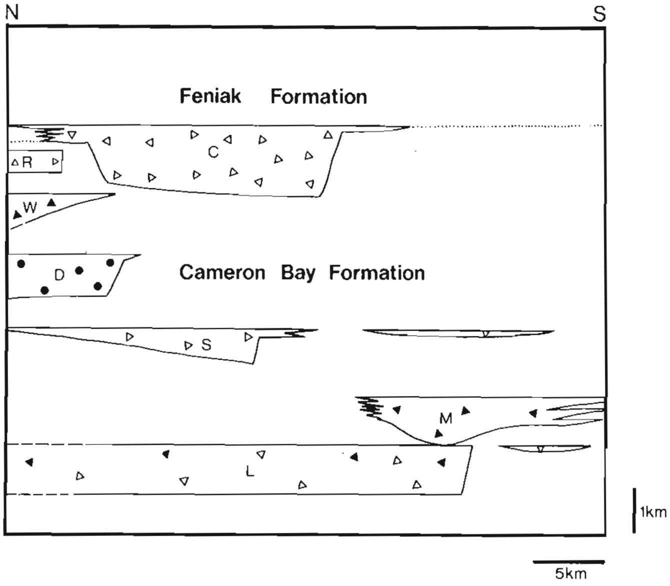


Figure 8.17. *Stratigraphic model for growth and progradation of stratovolcano.*



L = Lindsley Tuff
M = Mackenzie Tuff
S = Stevens Tuff
D = Doghead Tuff
W = Western Channel Tuff
R = Rocher Rouge Tuff
C = Cornwall Tuff

Figure 8.18. *Restored thickness variations of major ash-flow sheets (diagramatic and scale only approximate).*

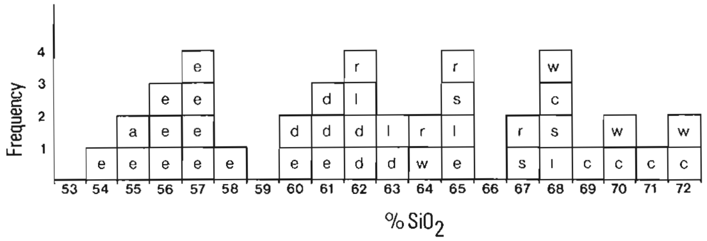
Several, and probably all, the ash-flow sheets are compositionally zoned. Typically, the earliest flows in each sheet are more siliceous, while later flows are more intermediate. This compositional zoning indicates that the source magma chambers for the ash-flows were compositionally zoned, with more differentiated upper portions (see Smith, 1979). These magma chambers are interpreted to have been individual plutons, such as the Hogarth, which probably coalesced at depth forming a batholith of regional dimensions.

The change from gas-poor andesitic volcanism to more highly gas-charged ash-flow eruptions could represent progressive differentiation of the batholith as it rose toward the surface, or perhaps it temporally reflects a higher degree of crustal input in the zone of magma genesis. Alternatively, the magmas may have scavenged volatiles during their rise to the surface with the andesites being erupted earlier, and from a deeper level.

CHEMISTRY

Volcanic rock petrochemistry is highly complicated by post-eruptive processes which modify the original magmatic composition. These processes include devitrification, deuteric processes, vapour phase transport and crystallization, fumarolic alteration, and hydration through interaction with ground water (Smith, 1960; Keith and Muffler, 1978; Lipman, 1965). Contact metamorphism and hydrothermal systems, related to contemporaneous or later events, may further modify earlier alteration making it difficult to determine the original magmatic composition.

As one might expect, the entire LaBine Group of the Echo Bay-Hornby Bay region is altered to some degree. Chemical analyses[1] were performed on the least altered rocks to ascertain their broad chemical affinities and for alteration studies in progress.



e = Echo Bay Formation
d = Doghead Tuff
l = Lindsley Tuff
r = Rocher Rouge Tuff
w = Western Channel Tuff
c = Cornwall Tuff
s = Stevens Tuff

Figure 8.19. *Histogram showing silica variation (recalculated H_2O free) in major stratigraphic units of the LaBine Group.*

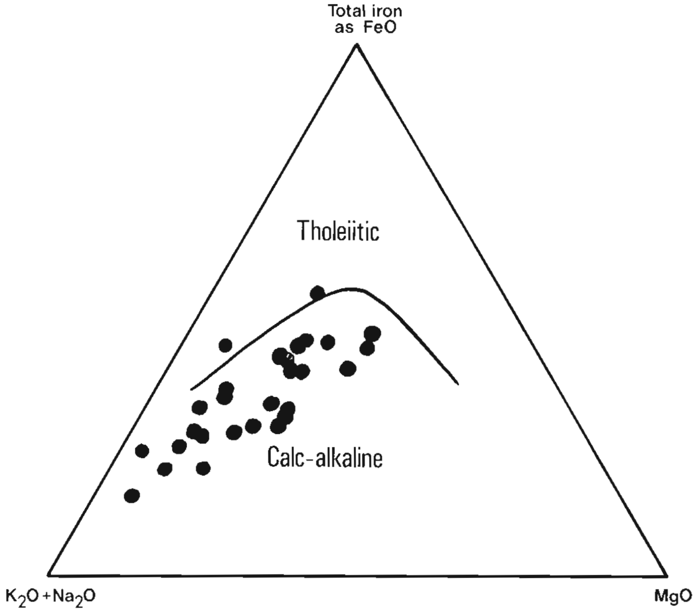


Figure 8.20. *AMF diagram for rocks of the LaBine Group. Tholeiitic-Calc-alkaline dividing line from Irvine and Baragar, 1971.*

In general, the LaBine Group is of intermediate composition with most SiO_2 values clustering between 55 and 68 per cent (Fig. 8.19), a chemical characteristic of calc-alkaline volcanic rocks (Green, 1980). Alkali and alkaline-earth variations indicate that these elements were extremely mobile during alteration and cannot be used for classifications although the suite shows no Fe enrichment trend on an AMF diagram (Fig. 8.20).

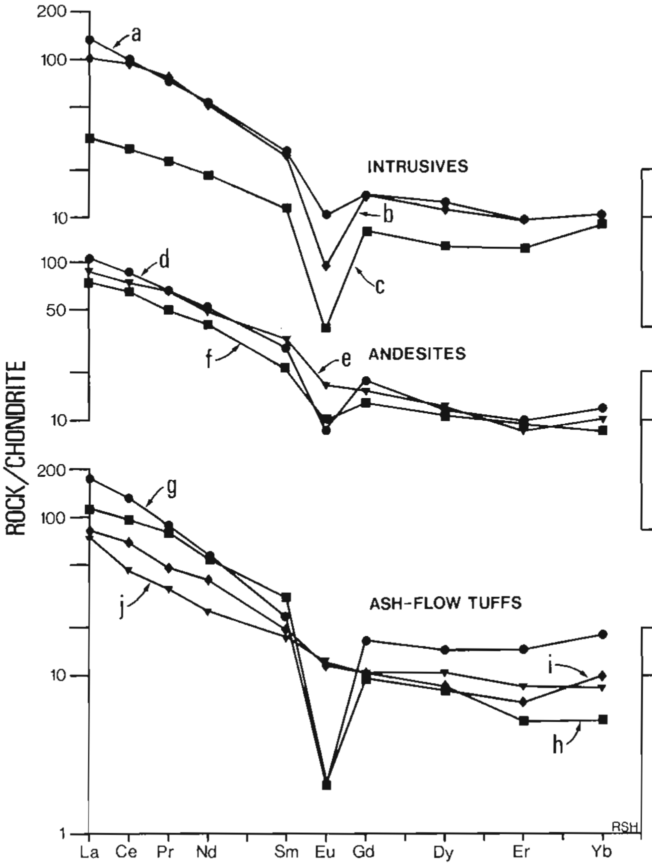
Titanium, while certainly mobile to some degree under appropriate conditions, may be less mobile than most other elements (Pearce and Cann, 1973). TiO_2 values for all rocks analyzed are less than 1.0 per cent. Intermediate rocks with low TiO_2 (<1.75%) dominate Tertiary-Recent volcanic provinces classified as orogenic (i.e. volcanic arcs) by Ewart and LeMaitre (1980). Green (1980) believed that typical TiO_2 values for island arc and continental arc rock series are less than 1.2 per cent. Furthermore, calc-alkaline extrusive rocks of continental arcs such as the Taupo Zone of New Zealand (Ewart et al., 1977; Cole, 1978, 1979), the Andes (for example: Kussmaul et al., 1977; Deruelle, 1978), Papua (MacKenzie, 1976) and the Pontid arc (Egin et al., 1979) nearly always have TiO_2 less than 1.0 per cent.

[1] Complete chemical analyses are available from the author on request.

Rare earth element (REE) analyses of rocks from the LaBine Group (Fig. 8.21) exhibit light REE enrichment patterns and the high overall abundances typical of high-K continental volcanic arcs such as the Chilean Andes (Thorpe et al., 1976, 1979), the Taupo Zone (Ewart et al., 1977; Cole, 1979), and Sardinia (Dupuy et al., 1979).

ALTERATION

The most prevalent alteration in the LaBine Group is pervasive potassium metasomatism in which potassium is enriched and soda depleted. While not yet studied in detail, many rocks contain greater than 6.0 per cent K_2O and less than 0.5 per cent Na_2O. In these rocks the plagioclase feldspars are completely replaced by an unidentified K-rich mineral(s) which is yellow-reactive to sodium cobaltinitrate, as is the groundmass. Trace elements such as Rb and Sr are also affected by this alteration and altered rocks with high K_2O/Na_2O ratios have Rb/Sr ratios greater than 10. Obviously, Rb-Sr ages obtained from the LaBine Group (Robinson and Morton, 1972) may not represent cooling ages but rather are related to hydrothermal events.



a) Hogarth pluton
b,c) Mystery Island Intrusive Suite
d,e,f) Echo Bay Formation
g) Rocher Rouge Tuff
h) Lindsley Tuff
i) Western Channel Tuff
j) Mackenzie Tuff

Figure 8.21. *Representative rare earth element abundances of rocks occurring in the study area.*

Alteration of the type described above has been documented in many volcano-plutonic terranes by numerous workers (for example: Ratte and Steven, 1967; Kisversanyi, 1972; Chapin et al., 1978; Wodzicki and Bowen, 1979). Fenner (1936) and Keith et al. (1978) described similar alteration of Recent age from shallow boreholes in the Yellowstone geothermal field and it has been widely recognized that hot spring waters are commonly depleted in potash relative to soda (Allen, 1935; Orville, 1963; Grindley, 1965; Nathan, 1976; Taylor, 1976; Sorey et al., 1978; Stauffer et al., 1980; Sammel, 1980; Parry et al., 1980; Rinehart, 1980). Thus, the LaBine Group is interpreted to have been affected by a fossil geothermal field in which hot springs were abundant.

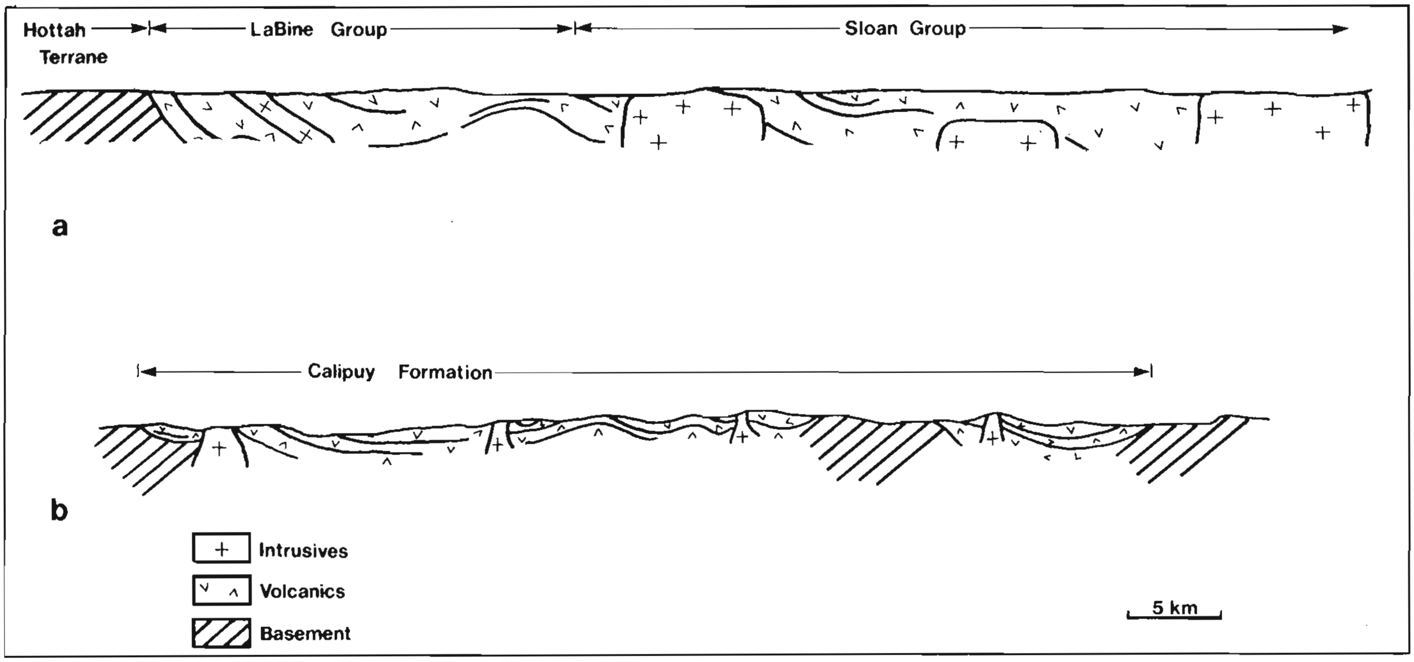
INTERPRETATION AND TECTONIC SETTING

Although alkali and alkaline earth metals were mobile during hydrothermal alteration, the original phenocryst mineralogy (quartz, potassium feldspar, biotite, hornblende, and plagioclase) coupled with SiO_2, TiO_2, and REE values indicate that the LaBine volcanic field is a high-K, calc-alkaline belt of mainly intermediate composition rocks that fall within the broad class of orogenic volcanic rocks (Ewart and LeMaitre, 1980). In detail, they are chemically similar to continental arcs related to subduction such as the Andes. In overall stratigraphy, mode of eruption, and mineralogy the LaBine Group resembles Cenozoic volcanic fields of the western United States such as the San Juan volcanic field (Steven and Lipman, 1976), the Datil-Mogollon volcanic field (Elston et al., 1976), and the Elkhorn Mountain volcanic field (Klepper et al., 1971). Cogent arguments have been made by several authors that the calc-alkaline volcanic rocks in those fields were related to oblique, low-angle subduction of the Farallon plate beneath the North American continent during the Eocene-Oligocene (Lipman et al., 1971, 1972; Elston, 1976; Coney and Reynolds, 1977; Lipman, 1980).

Although genetic details of volcanic arc magmatism are still controversial, there seems little doubt that arc magmatism is a multistage product of lithospheric subduction (Marsh, 1979). I see no compelling reason to invoke an ad hoc model to explain the origin of LaBine Group volcanic rocks as they have readily identifiable Cenozoic analogs. Therefore, I conclude that the LaBine Group represents a remnant of an early Proterozoic continental volcanic arc and that subduction, which may be the principal driving mechanism of plate tectonics (Forsyth and Uyeda, 1975; Richter, 1977; Chapple and Tullis, 1977), was occurring at least by about 1.9 Ga ago.

Stratovolcanoes are not likely to be preserved in the geologic record because they are topographically high-standing features, yet clearly there are tremendous thicknesses of andesite preserved in the LaBine Group. A probable explanation is that the LaBine Group developed in a basin which subsided concurrent with eruptions. The hypothesis that the Great Bear Volcano-Plutonic Belt was a region of subsidence during volcanism was first put forth by Hoffman and McGlynn (1977) who argued that the belt subsided in response to bending of a strike-slip fault.

Volcanic arcs often contain basins of various kinds. For example, grabens presently being filled with volcanics and related sediments are well-developed in the Cascades (Fyfe and McBirney, 1975), Nicaragua (McBirney, 1969), Ecuador (Williams and McBirney, 1979), and New Zealand (Ewart et al., 1977; Cole, 1979; Reyners, 1980). The Central American arc contains other types of basins besides grabens.



a. modified from Hoffman and McGlynn (1977)
b. modified after Hollister and Sirvas B (1978)

Figure 8.22. *Schematic cross-sections illustrating the similarity between the Great Bear Volcano-Plutonic Belt (a) and the Calipuy Formation (b).*

In Honduras, "intermontane tectonic troughs" developed during and after eruption of andesitic to basaltic lavas and breccias of the early Tertiary Matagalpa Formation, and many Miocene ash-flow sheets filled those, as well as other, broad, shallow basins (Williams and McBirney, 1969). Williams and McBirney (1969) also described a series of north-south trending, en echelon basins such as the Sula basin and the huge Comayagua Valley of Honduras. Furthermore, many individual Central American volcanoes, such as those found in Guatemala (Williams et al., 1964), are located within sags or depressions.

Yet another type of basin developed in arc terranes is found in northern Peru (Hollister and Sirvas B, 1978). There basaltic and andesitic volcanoes of the Calipuy Formation were erupted in a linear basin concomitant with folding of the volcanic and sedimentary basin-fill. Structurally the basin is strikingly similar to the Great Bear Volcano-Plutonic Belt (Fig. 8.22). Folds in both regions are en echelon with axes that are oblique to regional trends and to their respective outcrop areas.

Wilcox et al. (1973) showed how en echelon structures are related to wrench zones generated by horizontal shear couples. A given area in a wrench zone can undergo alternating periods of extension and compression because the stress regime at any particular place in the system depends on factors which change with time, such as bends and gaps in the braided fault system (Crowell, 1974a, b) or whether the system is one of parallel, divergent, or convergent wrenching (Wilcox et al., 1973). Crowell (1979), using the broad San Andreas transform system as an example, pointed out that wrench zones must be considered as complex moving systems in which local tectonic patterns, such as pull-apart basins, strike-slip faults, stretching, squeezing, dismemberment and rotation of individual fault-bounded blocks, are rapidly transformed as plate movement continues.

Wrench systems are not confined to transform margins but are also common in regions of oblique convergence (transpression) where the wrench system may appear in the magmatic arc (Fitch, 1972; Nakamura and Uyeda, 1980). There are numerous places where this situation exists with some of the more spectacular examples found in: New Zealand where the Taupo Zone and the Alpine and Hope Faults appear to be related to transpression (Sporli, 1980; Cole and Lewis, 1981); the northern Andes (Campbell, 1974), where the Dolores-Guayaquil Fault system formed in response to oblique motion of the northern portion of the Nasca Plate with respect to South America; Guatemala where Williams et al. (1964) interpreted conjugate sets of oblique faults to have been produced by strike-slip movements parallel to the long axis of the Central American Trench; and in Sumatra where the Sunda Arc is being folded and splintered by the Semangko fault system in response to oblique convergence of the Indian-Australian Plate with the Eurasian Plate.

The Semangko, or Barisan, Fault System of Sumatra exhibits most of the features typically found in wrench zones and is particularly interesting because it slices through the magmatic front (Fig. 8.23). Areas in the arc have undergone several periods of extension and compression leading to en echelon folding of basins filled with volcanic and sedimentary rocks (van Bemmelen, 1949; Westerveld, 1953). Topographic depressions have developed in extensional regimes along the fault itself (Page et al., 1979), especially near junctions of, and gaps between, en echelon fault segments (Tjia, 1978; Posavec et al., 1973). It is this type of environment (wrenched arc) that I envision for the LaBine Group because it satisfies all known constraints (i.e. arc volcanism, deposition in a basin, and en echelon folding). Although not touted in the literature, perhaps many exposures of ancient arc rocks represent the fill of wrench-generated basins related to oblique convergence, for this

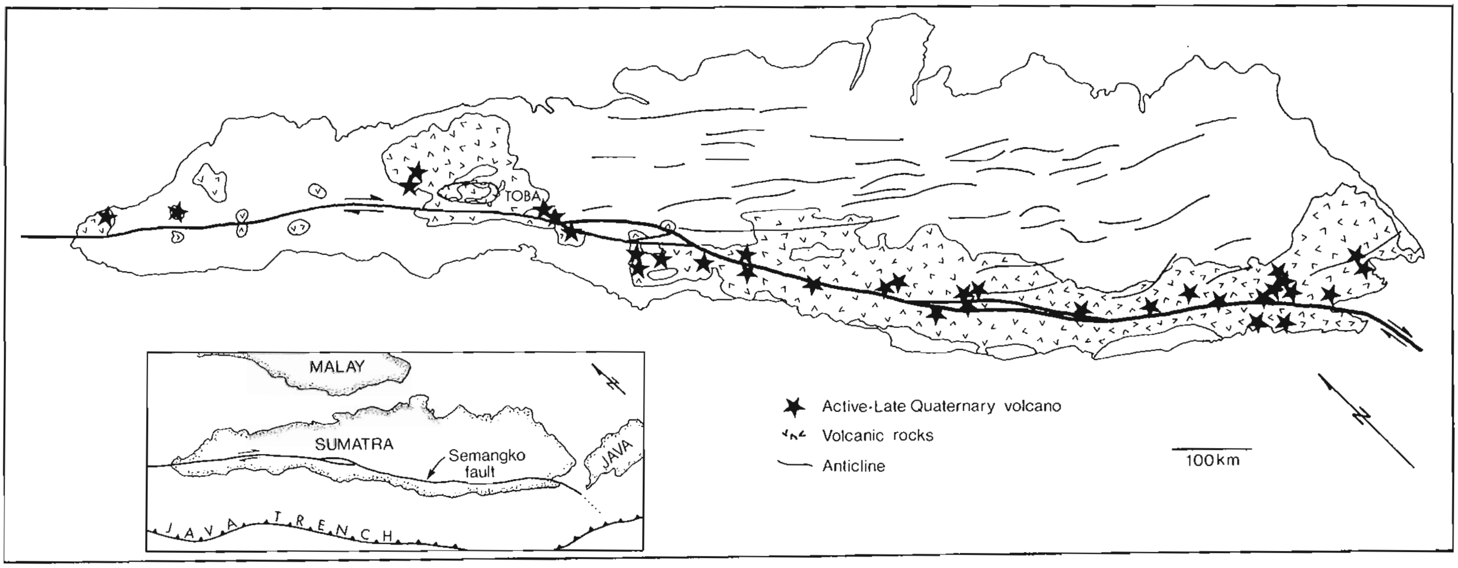


Figure 8.23. *Generalized and schematic geological map of Sumatra showing relationship between arc volcanics, en echelon folds, and the Semangko Fault zone.*

concept provides a simple and logical explanation for the preservation of high-level arc volcanoes which otherwise might be eroded to their roots.

TECTONIC MODEL

The tectonic model presented here is similar to that presented by Hoffman (1980a) but some refinements and modifications have been made in light of new geochronological and field data. The model is shown schematically in Figure 8.24.

In this model the Hottah Terrane is assumed to be allochthonous with respect to the Slave Craton and to be the remnant of a microcontinent or arc which collided with the Slave Craton over a westward-dipping Benioff zone (Fig. 8.24a). The collision resulted in accretion and deformation of the microcontinent and deformation of the western edge of the Slave Craton with its westward-facing passive margin sequence (Fig. 8.24b).

Continent-microcontinent or continent-arc collisions are by no means rare in the geologic record. Excellent examples of more recent continent-small plate collisions are present along the northwestern edge of the Australian continent where the edge of the Australian-New Guinea shelf is presently colliding with the Banda arc (Von der Borch, 1979). During the Miocene, an early Tertiary arc was accreted to the continent at New Guinea (Hamilton, 1979). Other examples of continent-microcontinent collision occur in the Eastern European Alpine System (Burchfiel, 1980) where several collisions are believed to have occurred from mid-Cretaceous to the Recent. In the northern Canadian Cordillera Tempelman-Kluit (1979) interpreted geologic relations in terms of a Late Jurassic-early Cretaceous continent-microcontinent collision.

In Wopmay Orogen the age of the collision is interpreted to have occurred between about 1.92 and 1.89 Ga. Metamorphic isograds, which postdate the major pulse of thrusting in the deformed passive margin sequence (Hoffman et al., 1980), are related to mesozonal S-type plutons (St-Onge and Carmichael, 1979) whose mean age is 1.89 ± 0.01 Ga[1] (Van Schmus and Bowring, 1980). Deformation of the Hottah Terrane must postdate a deformed pluton found at Hottah Lake dated at 1.92 ± 0.01 Ga (Van Schmus and Bowring, 1980). If deformation in both belts was related to the same event, as postulated here, then the age of deformation is bracketed between 1.92 ± 0.01 Ga and 1.89 ± 0.01 Ga.

The LaBine Group, which rests unconformably on the Hottah Terrane and lacks its penetrative fabric, must be younger than the microcontinent-continent collision. If the LaBine Group is a volcanic arc related to subduction, then it must have developed over an eastwardly-dipping subduction zone, as the ocean east of the microcontinent had already closed. This interpretation requires that following collision subduction changed from westward-dipping on the east side of the microcontinent to eastward-dipping on the west side (Fig. 8.24c).

Many examples of continent-arc or microcontinent collisions appear to have involved a reversal of subduction direction following collision. Hamilton (1979) presented evidence for incipient subduction reversal north of the island of Alor, as a result of collision between the Banda Arc and the Australian Continent. He also suggested that reversal of subduction direction occurred after arc-continent collision at New Guinea. The Miocene collision of the Apulian fragment with Euro-Russian continental crust was along a southward-dipping subduction zone while present day subduction under the Hellenic Arc is northward (Burchfiel, 1980). In the northern Canadian Cordilleran example of continent-microcontinent collision subduction is also believed to have stepped outboard of the accreted terrane and reversed direction (Tempelman-Kluit, 1979).

Independent support for an eastward-dipping subduction zone following collision in Wopmay orogen occurs in Athapuscow Aulacogen, located 300 km southeast of Port Radium (Fig. 8.1). There a group of calc-alkaline laccoliths, strikingly similar in composition, alteration and metalliferous deposits to the Mystery Island Intrusive Suite, are distributed axially over the length of the aulacogen, which trends normal to the Wopmay continental margin. The laccoliths exhibit compositional changes ranging from diorite in the west to quartz monzonite in the east (Hoffman et al., 1977).

[1] Age determinations by Van Schmus and Bowring are U-Pb zircon ages.

Badham (1978) considered this to be an oversimplification but stated that both potassium feldspar and biotite content in the laccoliths increased eastward.

The compositional trend in these laccoliths is similar to those of magmatic arcs (Moore, 1959, 1961; Ninkovitch and Hays, 1972; Kistler, 1974; Dickinson, 1975) – a similarity first pointed out by Hoffman et al. (1977) who suggested that the intrusions might be a result of subduction.

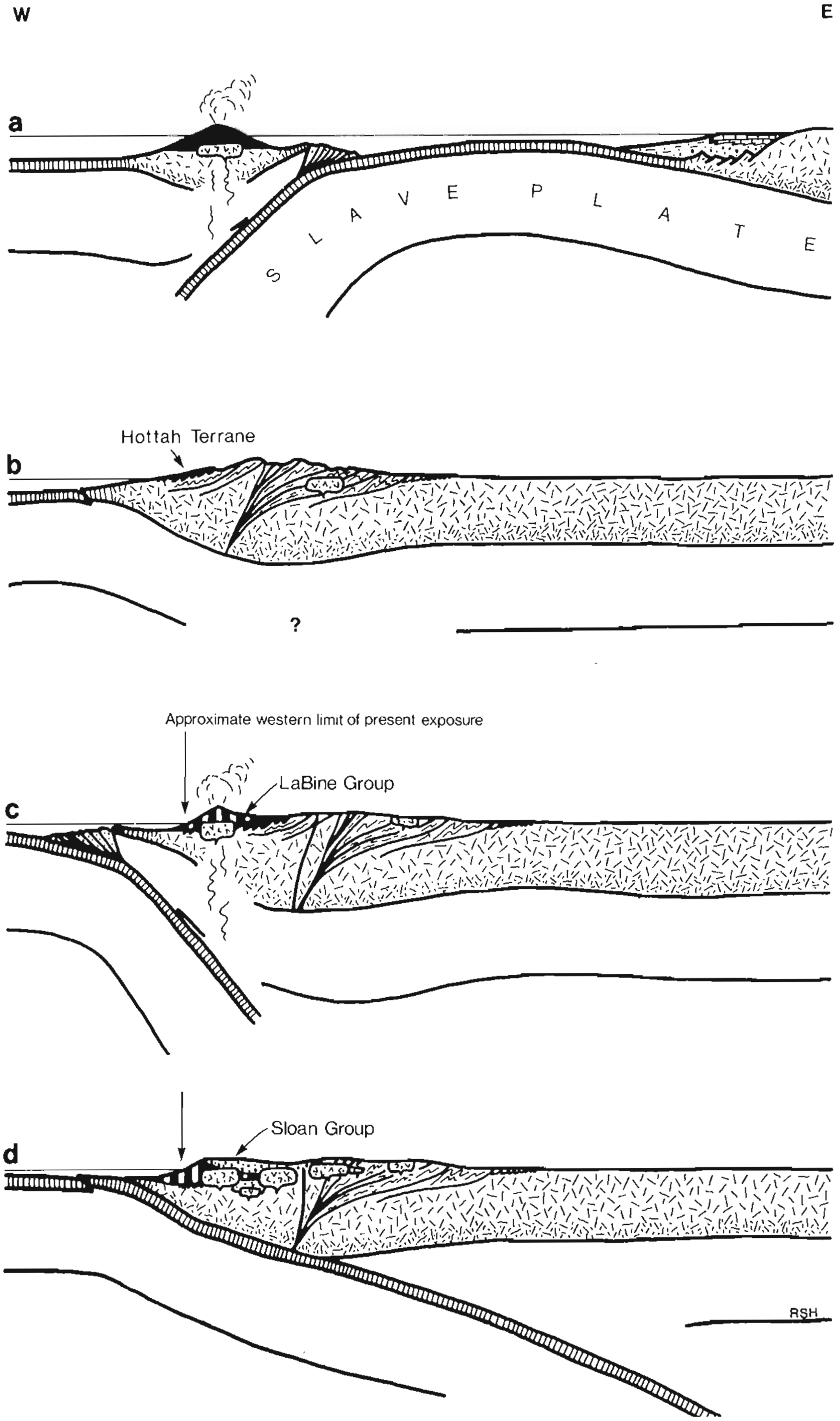


Figure 8.24. Proposed tectonic model for the origin of the LaBine Group and related rocks. See text for explanation.

The laccoliths postdate westerly-derived orogenic molasse presumably produced during collision and have an apparent age of 1.86 Ga ± .02 Ga (Van Schmus and Bowring, personal communication) – the same age or slightly younger than the LaBine Group. Thus, they support the concept of an eastward-dipping subduction zone that postdated the microcontinent-continent collision.

At the present time magmatism occurs above Benioff zones where they are about 100-200 km below the surface (see for example: Isacks and Barazangi, 1977). If this was also the case during the early Proterozoic then the Benioff zone postulated to have generated the laccoliths must have been fairly shallow, for they occur up to 250 km from the trench believed to have existed west of the accreted microcontinent.

A shallow Benioff zone might explain the conspicuous absence of similar magmatism in the Slave craton which should have resulted if a lithospheric slab was being subducted in an eastward direction. Perhaps the dip of the slab was so shallow that there was no asthenospheric wedge above the Benioff zone except under the aulacogen, where it presumably had upwelled during the initial rifting which created the Wopmay continental margin. The possibility that the presence of asthenospheric mantle above a Benioff zone is necessary for arc magmatism to occur has been proposed by Lipman (1980) and Dewey (1980). They both believed that extinction of magmatic activity in the Peruvian Andes is related to extreme flattening of the Benioff zone such that there is no asthenospheric mantle wedge present above it.

If this hypothesis is correct then why was there magmatism of the LaBine Group? I suggest that it may have been for one of three reasons: 1) possibly the subducting lithospheric slab was segmented, in much the same manner as modern slabs (Carr et al., 1979; Isacks and Barazangi, 1977) so that the segment dipping under the aulacogen was dipping at a shallower angle than the segment descending beneath the LaBine region, or; 2) if LaBine volcanism is slightly older than the laccoliths in the aulacogen, the dip of the downgoing slab could have decreased with time or; 3) the presence of thin lithosphere in the suture zone, which the LaBine Group likely buries.

It is the region of thickest and oldest lithosphere where, after collision, subduction would likely initiate because old lithosphere would tend to sink into the asthenosphere faster than young, hot lithosphere (Molnar and Atwater, 1978). With time, progressively younger lithosphere would be subducted resulting in a Benioff zone that becomes shallower with time. Assuming this, I speculate that the voluminous volcanism of the Sloan Group, located east of and stratigraphically above the LaBine Group, reflects progressive shallowing of the downgoing slab as younger, hotter and thinner lithosphere was subducted.

Burke et al. (1976) argued that during the Precambrian, convergence rates would have been greater, and Benioff zones more numerous, than in the Phanerozoic due to the greater thermal output of the earth at that time. If true, then Proterozoic Benioff zones would have been generally flatter than those of the Phanerozoic because the lithosphere would have been thinner and hotter during the Proterozoic.

It is likely that shallower subduction would lead to stronger coupling of the convergent plates (Dewey, 1980). When stronger coupling and oblique convergence occur together, wrench zones in arcs should be more common. Was this the case during the Proterozoic?

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9.

TECTONISM AND DEPOSITIONAL HISTORY OF THE HELIKIAN HORNBY BAY AND DISMAL LAKES GROUPS, DISTRICT OF MACKENZIE

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Abstract

The Hornby Bay and Dismal Lakes groups, the basal components of the Coppermine Homocline, are a middle Proterozoic succession of terrestrial and shallow marine sedimentary rocks deposited over the northern margin of the early Proterozoic Wopmay Orogen. Lithologic sequences and facies patterns within both groups reflect the response of sedimentation to changes in the loci of uplift and/or subsidence, and syndepositional normal fault activity. The two Groups represent two terrestrial siliciclastic to marine carbonate cycles of deposition, each in excess of 1.5 km thick.

Hornby Bay Group sedimentation began with infill of crustal depressions by two separate, in part coeval, fluvial systems (Bigbear system, Fault River system). A basement peneplain (Teshierpi-Bigtree Peneplain) formed in the source area as a consequence of the extended erosion that occurred during deposition of these systems. Cratonic tilting, manifest in Bigbear and Fault River systems as syndepositional normal faults, resulted in a northwest-dipping paleoslope and caused incision of the peneplain. With subsidence, deposition of the regionally extensive westward-flowing Lady Nye fluvial system buried the earlier fluvial deposits and infilled the dissected peneplain. Continued subsidence and/or eustatic sea level rise led to marine transgression and deposition of siliciclastic deltaic-marine and subtidal-intertidal stromatolitic dolostone facies. Cratonic uplift and the onset of a second siliciclastic-carbonate cycle resulted in drowning of regressive supratidal algal marshes and carbonate mudflats by deltaic-floodplain siliciclastic facies of upper Hornby Bay Group. Synchronous normal movement along a major fault (Teshierpi Fault) resulted in erosion of Hornby Bay Group on the upthrown block and controlled fluvial drainage patterns of upper Hornby Bay and lower Dismal Lakes groups.

The basal Dismal Lakes Group siliciclastic unit records the transition from southwest crustal tilting, to regional subsidence and formation of the Dismal Lakes platform with a terrestrial source area to the southeast. With decreased fault activity, continued erosion and gradual marine inundation, the lower Dismal Lakes Group siliciclastic fining-upward sequence was deposited on a stable, northwest-sloping, low-relief platform. Peritidal to open shelf carbonates, which form the bulk of the Dismal Lakes Group, accumulated on a broad platform.

Localized uplift in the September Lakes area exposed portions of this platform to subaerial erosion and karst development, while shallow marine deposition continued uninterrupted west of Dismal Lakes. A final transgression preceded extrusion of the Coppermine River flood basalts and submerged all previously exposed parts of the platform under an epeiric sea. Dismal Lakes Group deposition was terminated by an east-west extensional event marked by extrusion of the Coppermine River Group plateau basalts and intrusion of the Muskox Complex and Mackenzie-age diabase dykes.

Comparison of Dismal Lakes Group with the Ellice, Parry Bay, and Kanuyak formations of the Elu Basin reveals a close lithologic and tectono-depositional correlation indicating that both areas represent remnants of a more extensive Helikian shelf.

Résumé

Les groupes de Hornby Bay et de Dismal Lakes, qui constituent la base de l'homoclinal de Coppermine, représentent une succession d'âge protérozoïque moyen, constituée de roches sédimentaires qui s'étaient déposées en milieu terrestre et en milieu marin peu profond sur la marge nord de l'orogène de Wopmay, formé au début du Proterozoïque. Les successions lithologiques et les distributions de faciès observées dans les deux groupes reflètent la correspondance entre les modes de sédimentation et les déplacements des centres de soulèvement ou de subsidence, et aussi l'activité tectonique (failles normales) accompagnant la sédimentation. Les deux groupes représentent deux cycles de sédimentation, l'un de caractère continental (silicoclastique), l'autre de caractère marin (carbonaté); ils dépassent chacun 1,5 km d'épaisseur.

La sédimentation qui a créé le groupe de Hornby Bay a débuté par le comblement de dépressions crustales par deux systèmes fluviatiles distincts, en partie contemporains (système de Bigbear, système de Fault River). Une pénéplaine basale (pénéplaine de Teshierpi-Bigtree) a été formée dans la zone source, par l'érosion prolongée qui a accompagné la formation de ces systèmes. Le basculement du craton, qui s'est traduit dans les systèmes de Bigbear et de Fault River par la formation de failles normales pendant l'épisode de sédimentation, a créé un paléogradient incliné vers le nord-ouest; de ce fait, la pénéplaine a été entaillée par l'érosion. Avec la subsidence, s'est formé

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le vaste système sédimentaire fluviatile de Lady Nye, orienté vers l'ouest, qui a recouvert les dépôts fluviatiles plus anciens et comblé la pénéplaine disséquée. La subsidence prolongée, peut-être accompagnée d'une montée eustatique du niveau de la mer, a donné lieu à une transgression marine et à la formation d'un faciès sédimentaire silicoclastique, de type deltaïque-marin, et d'un faciès de dolostone, de type stromatolitique subtidal à intertidal. En raison du soulèvement cratonique et du déclenchement d'un second cycle silicoclastique à carbonate, les marais supratidaux régressifs occupés par des formations algales et les marais littoraux carbonatés ont été recouverts par les faciès silicoclastiques deltaïques et de plaine inondable caractéristiques de la partie supérieure du groupe de Hornby Bay. Un mouvement normal synchrone, apparu le long d'une importante faille (faille de Teshierpi), a entrainé l'érosion du groupe de Hornby Bay sur le bloc soulevé, et déterminé les caractères du réseau hydrographique, dans la partie supérieure du groupe de Hornby Bay et la partie inférieure du groupe de Dismal Lakes.

L'unité basale silicoclastique du groupe de Dismal Lakes montre la transition entre le basculement crustal vers le sud-ouest, et la subsidence régionale et la formation de la plate-forme de Dismal Lakes; il existait une source de sédimentation terrestre au sud-est. Avec la diminution de l'activité tectonique, la continuation de l'érosion et une transgression marine graduelle, la succession silicoclastique de la partie inférieure du groupe de Dismal Lakes, qui s'affine vers le haut, s'est déposée sur une plate-forme de faible relief inclinée vers le nord-ouest. Les carbonates du milieu péritidal passant à un milieu de plate-forme ouverte, qui forment la majeure partie du groupe de Dismal Lakes, se sont accumulés sur une vaste plate-forme.

Dans la région des lacs September, un soulèvement localisé a exposé des parties de cette plate-forme à l'érosion subaérienne et à la karstification, tandis que la sédimentation marine peu profonde s'est poursuivie sans interruption à l'ouest des lacs Dismal. Une transgression finale a précédé d'épanchement des basaltes des plateaux de la région de la rivière Coppermine, et toutes les parties anciennement exposées de la plate-forme ont été submergées par une mer épicontinentale. La sédimentation du groupe de Dismal Lakes s'est terminée par une phase d'extension est-ouest, marquée par l'épanchement des basaltes de plateaux du groupe de Coppermine River, et l'intrusion des dykes de diabase du complexe de Muskox et de ceux de l'étage du Mackenzie.

En comparant le groupe de Dismal Lakes aux formations d'Ellice et Parry Bay, et à celle de Kanuyak dans le bassin Elu, on découvre une étroite corrélation lithologique et tectono-sédimentaire, qui indique que ces deux régions représentent les vestiges d'une plate-forme plus vaste d'âge hélikien.

REGIONAL GEOLOGY

The Hornby Bay and overlying Dismal Lakes Group are the lowermost groups in the undeformed, north-dipping Coppermine Homocline (Fig. 9.1) of the Amundsen Embayment (see Young, 1981) (which also contains the Coppermine River and Rae groups in this area). The Hornby Bay Group rests unconformably on Aphebian basement ranging in age from 1.93-1.85 Ga (Van Schmus and Bowring, 1980) and is overlain with local unconformity by the Dismal Lakes Group. The Dismal Lakes Group is conformably overlain by basalt flows of the Coppermine River Group, dated at 1.2 Ga (Baragar, 1972). Both groups are cut by the Muskox Intrusion and by Mackenzie-age diabase dykes (Fig. 9.2a).

The Hornby Bay and Dismal Lakes groups represent two cycles in excess of 2.5 km thick of terrestrial siliciclastic to marine-carbonate deposition that took place over older stable cratonic crust (Wopmay Orogen). Deposition occurred on a generally quiescent post-orogenic basement although periodic movement along reactivated basement faults caused locally important facies changes.

The basement of the southern part of the Amundsen Embayment includes the Aphebian Hepburn metamorphic-plutonic complex, the Great Bear Batholith (and associated volcanic rocks) and sedimentary rocks of the Coronation Supergroup (Hoffman, 1981a). The Hepburn complex is composed of foliated granitoid rocks of the Hepburn and Wentzel batholiths and metasedimentary rocks of the Akaitcho and Epworth groups. Granitic and volcanic rocks of the Great Bear Batholith and McTavish Supergroup (see Hildebrand, 1981) probably underlie most of the Hornby Bay and Dismal Lakes groups and contributed significant clastics to the basal Hornby Bay Group.

Northeast- and northwest-trending faults significantly influenced the depositional history of both the Hornby Bay and Dismal Lakes groups. According to Hoffman (1981b) and Hoffman and St-Onge (1981) these are conjugate wrench faults that formed as a result of Aphebian plate collision west of the present exposure of the Wopmay Orogen and were characterized by strike-slip and reverse motion. During and following deposition of the Hornby Bay and Dismal Lakes groups the sense of movement on these faults was dip-slip

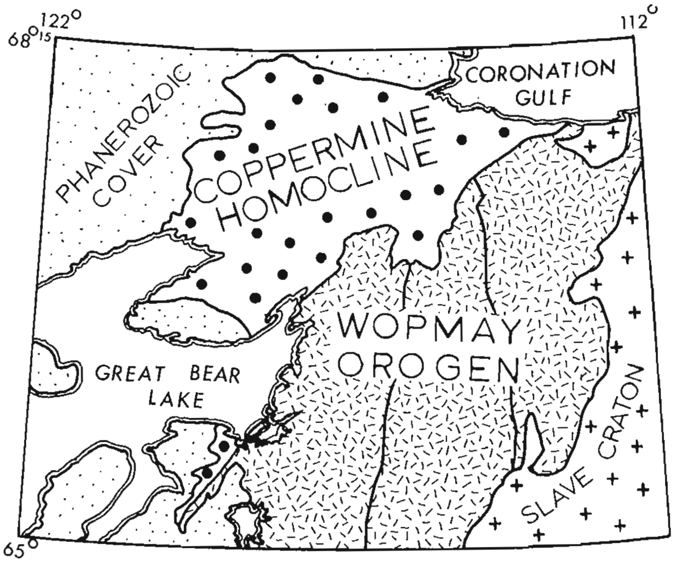


Figure 9.1. *Principal lithologic components of the north-western Canadian Shield.*

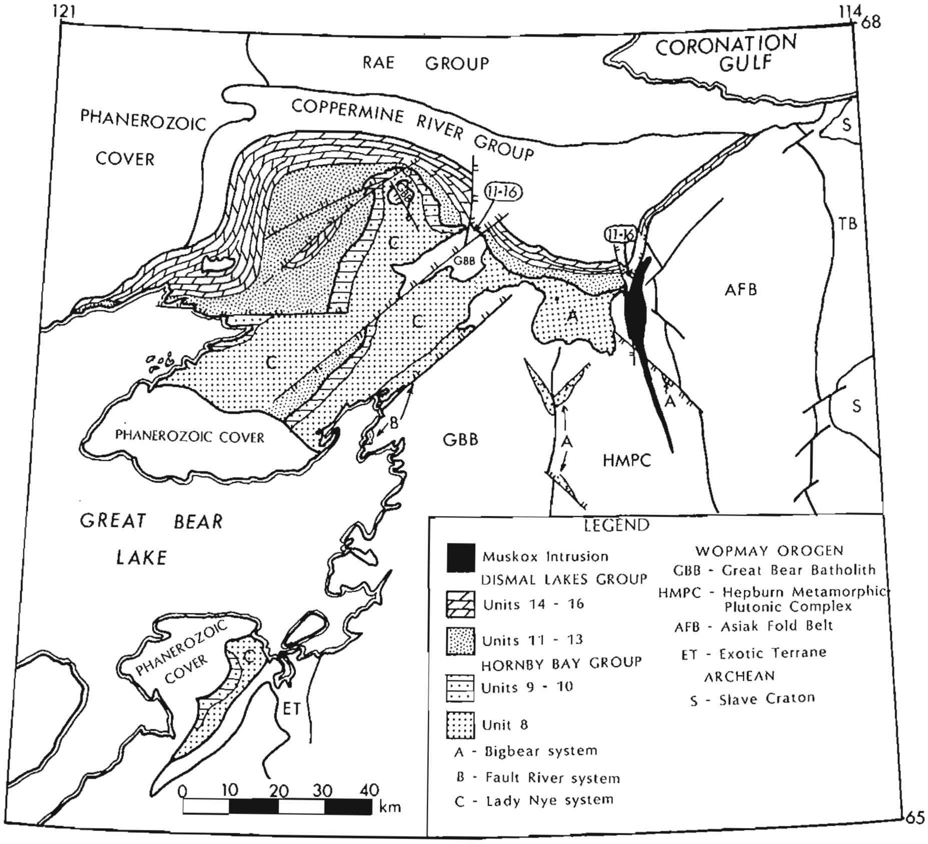


Figure 9.2a. *Geologic map and generalized map units of the Proterozoic rocks of the Coppermine River – Great Bear Lake area.*

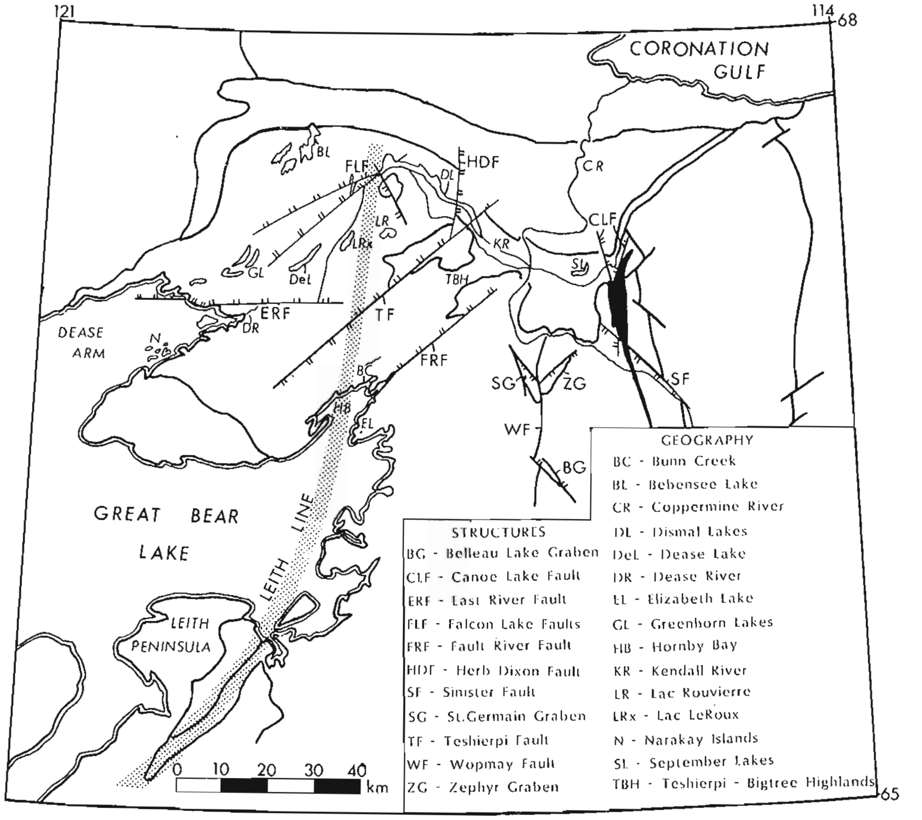


Figure 9.2b. *Map of geographic locations and major structural components referred to in the text.*

and oblique dip-slip. These faults were reactivated in response to tensional stresses developed during periods of regional uplift.

North-trending faults such as the Canoe Lake Fault, Herb Dixon Fault, and unnamed faults along the Leith Line (see below) and farther west, are also an important structural component of the Hornby Bay and Dismal Lakes groups. These faults are characterized by dip-slip movement that was for the most part restricted to late- to post-Dismal Lakes Group deposition (except for the Leith Line). The faults parallel the trend of Mackenzie-age diabase dykes, Coppermine River Group basalt feeders, and the Muskox Intrusion, which suggests that these features formed under the same stress system.

HORNBY BAY GROUP

Introduction

The Hornby Bay Group is an approximately 1500 m thick succession of shallow water sediments dominated by fluviatile and marine siliciclastics (Fig. 9.3). The basal unit of sandstone and conglomerate (Unit 8 – terminology of Baragar and Donaldson, 1973) was deposited in three separate but interrelated depositional systems: the Bigbear, Fault River, and Lady Nye fluvial systems (Fig. 9.2a). Facies patterns and paleocurrents do not reflect control by fault activity related to the final stages of compression in the Wopmay Orogen, therefore indicating a cessation of orogenic activity prior to deposition of Hornby Bay Group sediments. The Bigbear and Fault River systems were in part coeval, and both are conformably overlain by the Lady Nye system. The fluvial sediments of the Lady Nye system grade laterally and vertically into deltaic-marine siliciclastics (undifferentiated component of Unit 8) and a distinctive unit of stromatolitic dolostones (Unit 9). Carbonates were deposited along a northeast-trending shoreline that migrated to the east during marine transgression. Basement uplift in the northeast resulted in progradation of deltaic-marine siliciclastics of Unit 10 over carbonate algal marshes of upper Unit 9. Reactivation of a major northeast-trending basement fault (Teshierpi Fault) during regional uplift resulted in an unconformable relationship between the Hornby Bay Group and Dismal Lakes Group on the upthrown fault block.

Westward thickening of deltaic-marine facies of upper Unit 8 across Leith Line records the first activity of

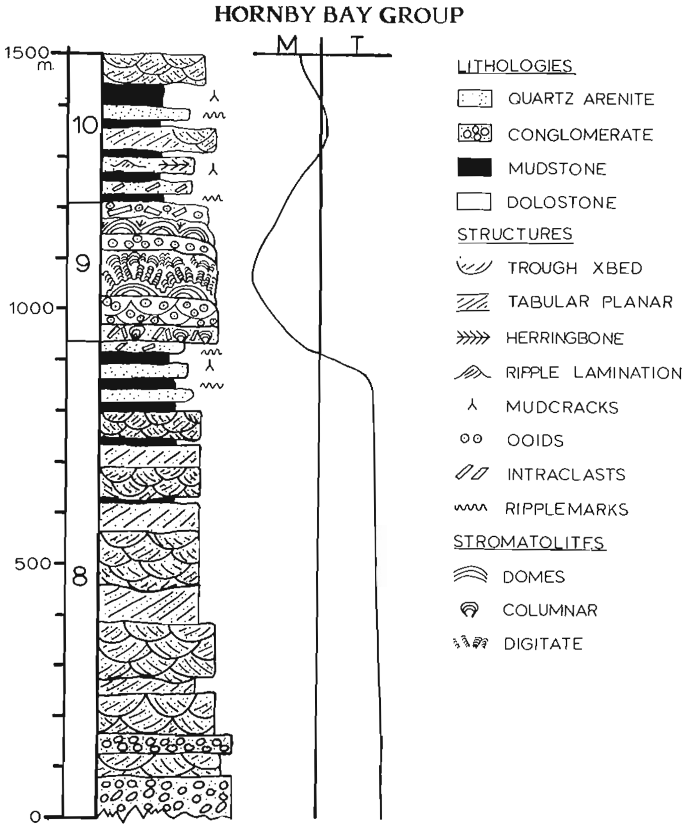


Figure 9.3. Generalized stratigraphic column for the Hornby Bay Group. This column applies only to Lady Nye system and overlying units and does not include subdivisions in lower Unit 8.

this inferred hinge zone. Leith Line extends from the informally named "Leith Ridge" (Balkwill, 1971) on the south shore of Great Bear Lake, north through the eastern (and deepest) part of Great Bear Lake, and continues northward through the Hornby Bay and Dismal Lakes groups (Fig. 9.2b). Indirect evidence of this feature is suggested by the concentration of north-trending (presumably Mackenzie-age) dykes in areas west of Leith Line, a marked change in fault patterns across the Line (northeast and northwest trends east of the Line; north and east trends west of the Line), lack of basement exposures west of the Line, and a sharp change in the strike of map units across the Line (eastward in the east; northward in the west). Consistent westward thickening of map units and interpreted deepening of depositional environments of western facies of both the Hornby Bay and Dismal Lakes groups indicates greater subsidence west of the Leith Line. Steeply-dipping to vertical and locally overturned beds of the Hornby Bay Group along this line suggest significant post-depositional movement.

Bigbear Fluvial System

Sedimentary rocks of the Bigbear system are exposed east and north of Coppermine River and as scattered fault-bounded erosional outliers along Wopmay Fault (St. Germain Graben, Zephyr Graben, Belleau Graben), Sinister Fault, and Canoe Lake Fault (Fig. 9.2b). These outliers occur up to 60 km away from the main exposures and suggest that the present deposits of the Bigbear system represent remnants of a once continuous cover. The Bigbear system overlies fresh to slightly weathered rocks of the Great Bear Batholith west of Wopmay Fault and rocks of the Hepburn metamorphic-plutonic complex east of the fault. The system is in part conformably overlain by basal arenites of the Lady Nye fluvial system and in part unconformably overlain by basal siliciclastics of the Dismal Lakes Group. The sedimentary sequence is subdivided into three stages (Fig. 9.4) that record infill of a rugged basement topography (stage I), stable braidplain deposition (stage II), and fault reactivation (stage III) (see Fig. 9.5).

Stage I

The three map units ($8a_{1-3}$) which comprise the deposits of this stage of sedimentation onlap basement rocks. The basal unit ($8a_1$) ranges from calcareous volcanic paraconglomerate to micaceous arkosic siltstone. It rests on weathered basement with a hackly blocky fracture and grades into overlying conglomerate and pebbly arenite ($8a_2$). It ranges in thickness from 0 to 40 m. The siltstones pinch out to the northwest concomitant with thickening of the paraconglomerates. The paraconglomerates are composed of locally derived angular volcanic rock fragments in a calcareous (locally stromatolitic) matrix and are interlayered with lenses of flat laminated to crossbedded feldspathic quartz arenite. These lithologies occur as erosional remnants along the flanks of present-day basement hills. Feldspathic siltstone with mud rip-ups, granitic and volcanic granules, and parting plane lineation occurs in valleys surrounded by

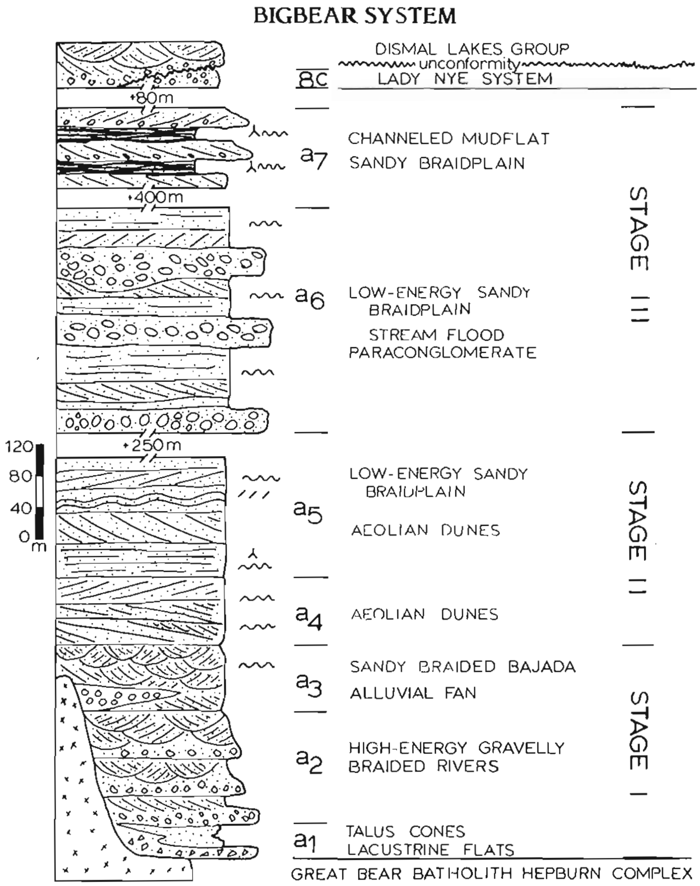


Figure 9.4. Lithofacies succession and depositional stages for the Bigbear system (Unit 8A).

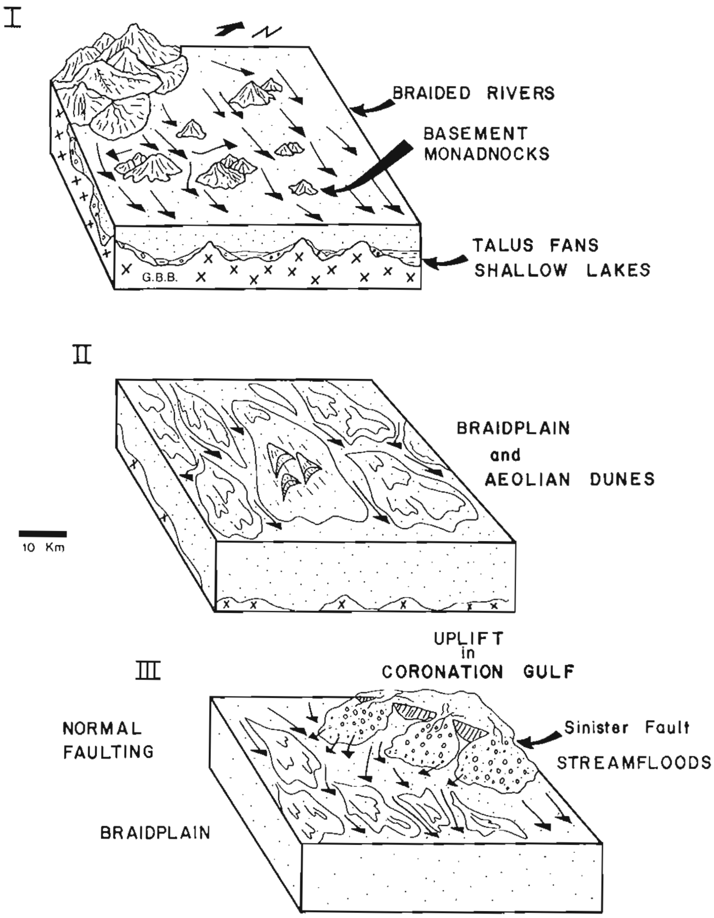


Figure 9.5. Diagrammatic reconstruction of depositional environments recorded by the Bigbear system.

basement hills. Paleocurrent indicators are sparse but generally suggest a southeast-dipping paleoslope with transport affected locally by basement topography.

Unit $8a_2$ is composed of interlayered arenaceous lithic pebble paraconglomerate and pebbly to granular lithic to feldspathic hematitic quartz arenite. It has a maximum thickness of 125 m and pinches out to the west-northwest. The unit has gradational contacts with overlying and underlying rocks. The paraconglomerates are massive and contain locally derived volcanic and granitic pebbles. The moderately sorted and submature arenites contain trough and tabular planar crossbeds up to 2 m. The unit becomes finer and less lithic upward. Paleocurrents are unimodal to the southeast with local deflections about basement highs.

The overlying unit ($8a_3$) is a variegated red and white feldspathic to subfeldspathic, medium grained, moderately sorted quartz arenite up to 120 m thick. In the northwest it is interlayered with volcanic boulder to pebble paraconglomerates which are absent in the southeast. Unit $8a_3$ is the basal segment in contact with basement in all outliers investigated. It is ubiquitously trough crossbedded with less common tabular-planar crossbeds. Current, wave, and run-off (ladderback) ripple marks are common. The amount of feldspar and the average crossbed thickness (10-50 cm) decreases upward. Paleocurrents are unimodal to the southeast with no indication of deflections around basement highs.

Interpretation. Unit 8a deposits formed local alluvial fan/talus cones by both mass movement (paraconglomerates) and traction run-off (arenites) (Fig. 9.5). The feldspathic siltstones are interpreted as fluviolacustrine deposits that accumulated as distal equivalents in topographic lows. The occurrence of this unit at the base of many present-day basement hills and the stratigraphic onlapping observed suggests that the present topography is an exhumed paleotopography. The overlying paraconglomerates and pebbly arenites ($8a_2$) are high-energy braided river deposits similar to the Donjek River facies of Miall (1978). Their deposition and dispersal was controlled to some extent by paleotopography. Feldspathic arenites ($8a_3$) are moderate to low energy braided stream deposits. Their occurrence as the basal unit in the outliers and consistent southeast paleocurrents suggest that by the time of their deposition the rugged paleotopography had been infilled and a regional, essentially planar, depositional surface existed.

Stage II

Deposition during this stage is recorded by two units ($8a_4$, $8a_5$), both of which are composed almost completely of medium grained, well sorted hematitic quartz arenite. Unit $8a_4$ is a maximum of 200 m thick in the southeast, but pinches out 45 km to the northwest. It is in gradational contact with overlying and underlying units. The unit has a friable, slabby-weathering appearance as a result of heavy hematite staining. Tabular- and wedge-planar crossbeds (maximum 1.5 m) are the dominant structure. These crossbeds typically are composed of sets of crosslaminae that have low angle discordance between sets (2nd order surfaces of Brookfield, 1977) and a maximum foreset dip of 26°. Current ripple lamination occurs with directions of transport both parallel and perpendicular to the down-dip direction of the foreset lamination. Climbing ripples are also present. Paleocurrents are unimodal to the southeast but commonly have a strong northeast mode.

The overlying unit ($8a_5$) has a maximum thickness of 380 m and is in sharp contact with the overlying conglomerate-bearing unit. It is composed completely of medium grained, variably hematitic quartz arenite although thin (<1 m) siliceous volcanic granulestones occur in the northwest. Plane bed to undulose current lamination is typical of the unit and produces a distinctive "table-top" weathering profile. Bedding surfaces commonly have wave ripples and thin red mudstone veneers with sand-filled mud cracks. Shallow (<20 cm) mudchip-filled scours occur within the plane bedded unit. Interlayered with plane bedded arenites are large (up to 3 m) tabular planar crossbeds with foreset dips of 10 to 26° that do not scour the underlying arenite. Very thick lenses (10-12 m) of low angle trough crossbeds with festoon widths up to 30 m are also present. These large bed forms are composed of smaller bundles of laminae, 0.5-1.5 m thick, with discordant surfaces between bundles. Paleocurrents show moderate dispersion but trend south-southeast.

Interpretation. Unit $8a_4$ is interpreted as an aeolian deposit based on the similarity of the style of crosslamination to recent and ancient aeolianites from other localities (Brookfield, 1977; Hunter, 1977). The overlying unit was deposited in a mixed aeolian and extremely low energy fluvial setting. The plane to undulose bedded arenites formed by migration of low amplitude sand waves or sand sheets on a very shallow braidplain. Analogous structures in the Platte River (Smith, 1971) form in water depths of several centimetres. The interlayered crossbedded arenites are also interpreted as aeolian, based mainly on the lack of scour at their bases, an expected feature if they formed during fluvial flood stages. The thick megatrough crossbeds were deposited by barchan dune migration. The occurrence of these units suggests that sedimentation took place in a stable and slowly subsiding basin.

Stage III

Unit $8a_6$ is composed of mature medium grained quartz arenite interlayered with massive beds of intraformational cobble-boulder paraconglomerate. The base of the unit is marked by the abrupt appearance of intraformational boulder paraconglomerate. The unit grades into the overlying unit $8a_7$. Thickness estimates of unit $8a_6$ are impossible due to the lack of marker beds and rubbly exposure.

The arenites commonly display flat- to undulose lamination with tabular-planar and trough crossbeds comprising about 20 per cent of the unit. Intraformational paraconglomerates are composed of subrounded to subangular clasts of red quartz arenite in a hematitic sandy matrix. The paraconglomerates rest on surfaces scoured into the underlying arenite. Where present, pebble imbrication and rare cross stratification in the paraconglomerates indicate transport to the southwest, in contrast to the interlayered arenites which have southeast-trending paleocurrents.

Unit $8a_7$ is composed of interlayered flat laminated quartz arenite, mudcracked, very hematitic, muddy granulestone, and clay-rich granulestone and coarse quartz arenite. The base of the unit is distinguished by the first occurrence of vein quartz pebbles in unit 8. Unit $8a_7$ is overlain conformably by basal Lady Nye system conglomerate (Bunn Creek Facies) and unconformably by basal Dismal Lakes Group clastics. The coarse quartz arenites occur in shallow lenses at the base of the unit but become tabular bodies of trough crossbedded pebble-granule arenite towards the top. Vein quartz pebbles are the most common clasts, but red quartz arenite pebbles also occur. Paleocurrents are unimodal to the west-southwest.

Interpretation. The interlayered arenites and paraconglomerates ($8a_6$) are interpreted as products of deposition on a braidplain (arenites) that was interrupted by influxes of locally-derived stream flood paraconglomerate. The variation in crossbed orientation between the arenites and paraconglomerates, and intrabasinal cobble origin, reflect periods of southwest-side-down normal movement along the Sinister Fault (Fig. 9.2b). This resulted in local paleoslope change and erosion of earlier Bigbear deposits.

Unit $8a_7$ records a gradual change from low-energy braidplain deposition to higher-energy braided river deposition, as shown by the upward increase in both grain size and crossbed thickness. The gradual southeast to southwest change in paleocurrents reflects a change from the preceding paleoslope to that of the overlying Lady Nye conglomerate.

Reactivation of the Sinister Fault during Bigbear deposition is interpreted as a manifestation of large-scale regional tilting, which gradually changed the paleoslope from southeast-dipping during Bigbear deposition to southwest-dipping during Lady Nye deposition.

Fault River Fluvial System

Fault River system strata are exposed in a series of steep cuestas along the northwestern side of the Fault River valley, in fault-bounded wedges southeast of the valley, and on islands in Great Bear Lake (Fig. 9.2b). At least 150 m of postdepositional dip-slip movement (northwest side down) has

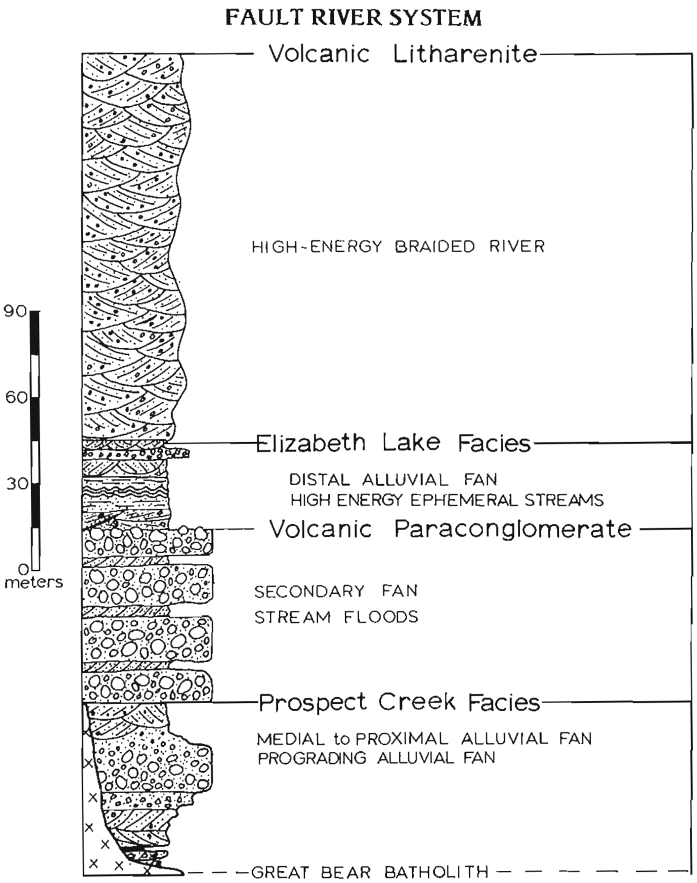


Figure 9.6. *Lithofacies succession and proposed depositional environments for the Fault River system.*

occurred on the Fault River Fault. Facies in ascending stratigraphic order onlap weathered, hematite-rich volcanic and granitic rocks of the Great Bear Batholith to the northeast. In addition, all units display a significant coarsening of grain size in this same direction. The sequence is conformably overlain by the basal conglomerate of the Lady Nye fluvial system. The Fault River system, which has a thickness of 330 m, is subdivided into four facies (Fig. 9.6), inferred to have been deposited in distinct environmental settings.

Prospect Creek Facies

This facies rests on weathered volcanic rocks of the McTavish Supergroup and coarsens upward from siltstone-arenite at the base to boulder paraconglomerate near the top as the facies grades into the overlying Volcanic Paraconglomerate Facies.

The Prospect Creek Facies ranges erratically in thickness from 0-40 m. Where absent, the overlying Volcanic Paraconglomerate Facies rests directly on basement. The lower part of the facies is composed of hematite-rich siltstones, mudstones, and poorly sorted arkosic arenites. The arenites occur in trough crossbeds that fill scours in underlying beds, and in tabular-planar and massive beds interlayered with flat- to ripple-laminated siltstones and mudstones. Angular felsite and granite pebbles become increasingly abundant higher in the facies. These pebbles appear first as thin lags in arenite beds and then in massive mud-rich paraconglomerates.

The middle part of the facies is composed of superposed massive paraconglomerate beds with lenticles of flat-laminated, sparsely mudcracked argillaceous siltstone, and lenses of trough crossbedded arkosic arenite. Towards the top of the facies, boulder beds of orthoconglomerate and paraconglomerate contain abundant round to subround felsitic and granitoid clasts. Boulders of red quartz arenite and vein quartz are conspicuous components of the uppermost conglomerate beds. Interlayered thin specularite-rich litharenites are flat-laminated and have scoured upper contacts. The Prospect Creek Facies is capped by flat-laminated to trough crossbedded pebbly subarkose. Paleocurrents from the lower beds of the facies show transport to the northwest (away from the present bounding fault) but change to southwest-directed in the upper part of the facies.

Interpretation. Based on textural and compositional immaturity and prevalence of probable debris-flow paraconglomerates, the Prospect Creek Facies is interpreted as alluvial fan deposits (cf. Nilsen, 1969). The composition of the lower beds, the northwest directed paleocurrents, and irregular lateral thinning and thickening indicate deposition of locally derived detritus, perhaps shed off of the Fault River Fault scarp, into an area with a maximum of 40 m paleotopographic relief. The upward increase in grain size and abundance of debris-flow beds high in the facies suggest fan progradation. The increased maturity of these units suggests a secondary (fan apex) rather than primary (basement) source area. Clast composition and paleocurrents suggest a source area to the northeast.

Volcanic Paraconglomerate Facies

This facies is composed of coarse quartz arenite with paraconglomerate interbeds up to 5 m thick. It is in contact with both the underlying Prospect Creek Facies and basement and is overlain by the fine grained Elizabeth Lake Facies. The paraconglomerate beds are massive and composed of well rounded to subrounded boulders of felsite, red quartz arenite, and vein quartz in a matrix of coarse quartz arenite. The basal conglomerate has a 2:1 ratio of volcanic to red quartz arenite boulders, whereas the uppermost bed is composed almost completely of boulders of red quartz arenite. The paraconglomerate beds are separated by coarse quartz arenites with trough and tabular-planar crossbeds with low-angle lamination. The amount of interbedded arenite decreases to less than 10 per cent to the northeast. Paleocurrent measurements from the crossbedded arenite layers are strongly unimodal to the west-southwest.

Interpretation. This facies was deposited as a result of stream flood rather than stable fluvial processes (cf. Miall, 1977). The red quartz arenite boulders, lithologically identical to rocks in the Bigbear system, are inferred to have been eroded as the Fault River system drainage incised outlying Bigbear system deposits. The compositional maturity and relatively well-sorted texture of the associated arenites suggests that this facies contains secondary reworked fan deposits. Spasmodic fault-related basin floor subsidence was the probable mechanism for secondary fan formation, and accounts for the relatively abrupt introduction and termination of this facies.

Elizabeth Lake Facies

This facies extends from the type area (Elizabeth Lake, north of Doghead Point, Great Bear Lake) to the easternmost exposures examined, in the Fault River valley. In the east the base of the facies rests directly on the boulder surface of the underlying Volcanic Paraconglomerate Facies. The facies is composed mainly of well-sorted, fine-grained, mudchip-bearing red quartz arenite. Towards the top of the facies, thin lags and massive beds of hematite-rich volcanic debris occur. The dominant sedimentary structures are thin (<25 cm) trough crossbeds with low-angle foreset lamination transitional into current lineated upper flow regime plane beds. In-phase aggradational ripples and mudcracks are abundant. Volcanic detritus first appears as isolated granule lags which gradually decrease in abundance, giving way to several massive, pebbly to granular, matrix-supported conglomerate beds in the uppermost part of the facies.

In easternmost exposures, both average bed thickness and grain size increase. Volcanic granulestones and red quartz arenite cobbles occur throughout the facies in this same area. Paleocurrents are unimodal, with transport directions ranging from north-northwest in easternmost exposures to south-southwest in the type area.

Interpretation. The characteristic upper flow regime stratification, thin bedding, unimodal paleocurrents and periodic desiccation, suggest this facies accumulated in a shallow-water, high-energy fluvial environment, possibly an ephemeral stream complex characterized by sheet flooding. Variation from recent analogues (Picard and High, 1973; McKee et al., 1967), such as lack of scour features, is attributed to the paucity of channelized flow in a nonvegetated fluvial system (cf. Schumm, 1968). The mature, well-sorted character of this facies indicates extensive reworking and sorting, and suggests a distal, secondary, fan setting.

Volcanic Litharenite Facies

This is the thickest facies in the Fault River system, with a minimum thickness of 125 m. It has a gradational lower contact (5-10 m) with the underlying Elizabeth Lake Facies and is conformably overlain by the Bunn Creek conglomerates of the Lady Nye fluvial system. The facies is composed of medium- to coarse-grained lithic and quartz arenites with abundant granules and small pebbles of volcanic detritus. Trough crossbeds, the only sedimentary structures within the facies, range from an average 1.5 m to a maximum of 15 m thick. The transition from the Elizabeth Lake Facies into the Volcanic Litharenite Facies is marked by a gradual upward increase in bed thickness and grain size, an increase in abundance of volcanic detritus, and a loss of hematitic pigmentation. Average grain size and the abundance of volcanic debris increases to the northeast. Paleocurrents are strongly unimodal to the west, and are consistent throughout the facies. Concentrations of conglomeratic material are absent.

Interpretation. The large scale cross stratification in this facies and distinct unimodal paleocurrent pattern, indicate a high-energy fluvial environment. The lack of smaller crossbeds indicative of emergent falling-water stage sedimentation precludes identification of depositional cyclicity. The low paleocurrent dispersion indicates a paleoslope of regional extent. The scale of cross stratification in this facies may reflect an increase in discharge volume as a result of source-area incision (to the east) and piracy of a Bigbear(?) drainage net (cf. Coleman, 1969). The contained lithologies and structures suggest this facies formed during a period of ongoing uplift in the east that caused a regional (basinwide) change in paleoslope to the west.

The four facies of the Fault River system display an overall upwards increase in textural and compositional maturity, and also an increase in fluvial, rather than debris flow, style of deposition. Although this is atypical of terrestrial clastic wedges affected by repeated movement on basin-bounding faults (cf. Heward, 1978; Steel, 1976), it is suggested that the present bounding fault of the Fault River system was active during deposition. The distribution and character of facies within the system suggest that this and related faults (Hoffman, 1978) were important in restricting deposition to a northeast-trending, wedge-shaped depression. Most facies record derivation from the east via second-cycle reworking of earlier deposits that were in part Fault River alluvial fan facies, and in part Bigbear system facies. Regional uplift in the east was punctuated by pulses of (dip-slip?) movement along faults and subsidence of the depression (Fig. 9.7).

Lady Nye Fluvial System

This is the thickest and areally most extensive fluvial system within Unit 8 (Fig. 9.2a). Facies of ascending stratigraphic order rest unconformably on fresh granitoid and volcanic rocks of the system, except where basal conglomerate conformably overlies the Volcanic Litharenite Facies of the Fault River system, and west of the Muskox Intrusion where basal pebbly arenite conformably overlies the Bigbear system. The Lady Nye system is conformably overlain by, and is gradational into, Unit 9 dolostone, with the exception of areas adjacent to and east of the Teshierpi Fault, where basal siliclastics of the Dismal Lakes Group rest unconformably on Unit 9 (Fig. 9.2a).

Lady Nye system lithologies record a completely gradational fining-up sequence from fluvial into deltaic-marine deposits, and are subdivided into three facies: Bunn Creek Facies, Cape MacDonnel Facies, and an unnamed deltaic facies (Fig. 9.8).

Basement-Cover Relationships

A basement regolith is conspicuous in its absence beneath the Lady Nye system. Basement lithologies typically display minor hematitic enrichment of potassium feldspars, mafic phases, and volcanic glass, but are otherwise remarkably fresh. Inliers and large bodies of basement (Fig. 9.2a) are topographically rugged, positive features, that have embayed or re-entrant contacts and apophyses that extend into the surrounding arenite cover. Arenite commonly outcrops in these embayments and also occurs as erosional remnants high on the flanks of hills of basement. Clasts in these arenites are rarely local, and paleocurrents are directed parallel to and/or into basement regions, but not away from these areas. In addition, there are areas, in particular southeast of the Teshierpi Fault, where basement hills are capped by a horizontal to westerly-dipping planar surface that extends across an intervening arenite-filled valley to adjacent basement outliers (Fig. 9.2a).

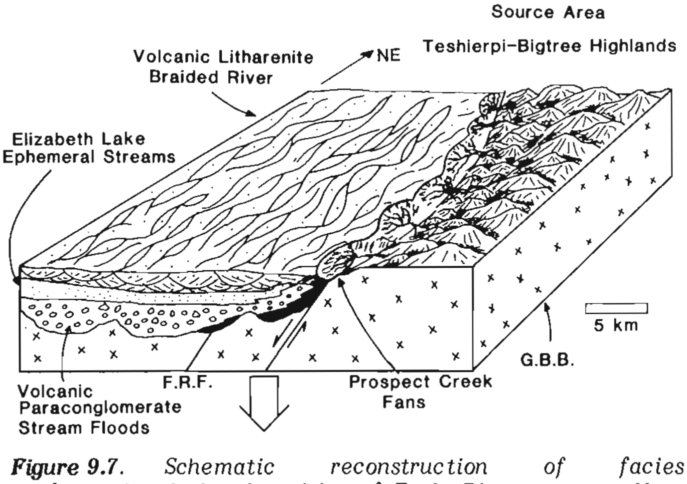


Figure 9.7. *Schematic reconstruction of facies configuration during deposition of Fault River system. Note that the bounding fault is not the Fault River Fault.*

The planar surface is interpreted as a peneplain that formed when this area was the source region for the Bigbear system. The present rugged topography of the basement and the occurrence of outcrops of arenite on the flanks of outliers suggests that the present topography is an exhumed paleotopography. Basement inliers formed as the result of peneplain dissection during uplift (early during Lady Nye deposition) and were subsequently buried by Lady Nye system rivers as the peneplain subsided.

Bunn Creek Facies

This facies is named for exposures along Bunn Creek (north shore of Hornby Bay, Great Bear Lake; Fig. 9.2b) and can be mapped as far north as Teshierpi Highlands. The total thickness of the facies is unknown, as it is gradational into the overlying Cape MacDonnel Facies. In the Hornby Bay area the basal conglomerate overlies Fault River system rocks; to the north and northeast basal conglomerate and pebbly arenite overlie Great Bear Batholith rocks. The basal

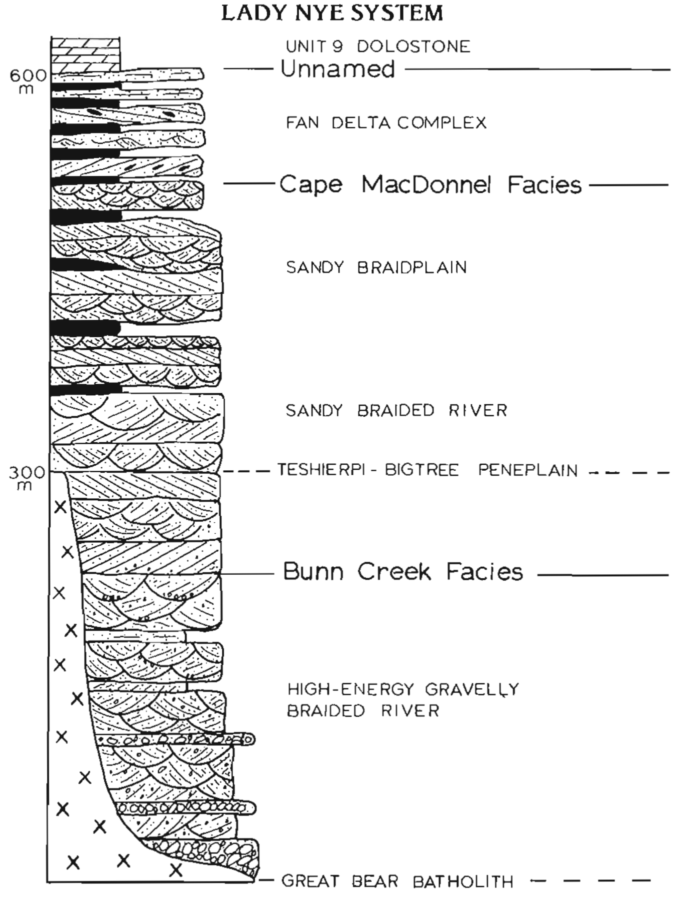


Figure 9.8. *Lithofacies succession and depositional environments for the Lady Nye system. The top of the Great Bear Batholith represents the surface of the Teshierpi-Bigtree Peneplain.*

part of this facies is composed of polymictic orthoconglomerates interlayered with coarse pebbly quartz arenites. The conglomerates are massive, but show well-developed pebble imbrication. Clasts include red quartz arenite, peloidal ironstone, vein quartz, algal-laminated chert, and low-grade metasedimentary rocks. Interlayered pebbly to granular quartz arenites display trough crossbeds up to 2 m. Conglomerate beds and pebble-sized clasts decrease in abundance upward, as do overall bed thickness and grain sizes. Flat-laminated to tabular-planar crossbedded micaceous quartz arenites, with scoured upper contacts, are common in the middle part of the facies. Paleocurrents are unimodal to the southwest and west.

Interpretation. The Bunn Creek Facies represents the proximal to medial part of a gravelly, braided river complex similar to the Donjek River (Williams and Rust, 1969) or the Scott outwash fan (Boothroyd, 1972). It marks the beginning of deposition and subsidence following peneplain incision. Pebbles of only very resistant lithologies, absence of locally-derived material in the conglomerates, and lack of regolith are the result of prolonged reworking during peneplain incision. The distinctive clasts of peloidal ironstone probably were derived from ironstones of the Tree River (Hoffman, 1981a) or Mara formations (Campbell and Cecile, 1981). The nearest present exposure of this lithology is the Tree Basin (Hoffman, 1980a,b), approximately 200 km east of Dismal Lakes. Red quartz arenite clasts in the Bunn Creek Facies conglomerates indicate that Lady Nye fluvial incision of the basement peneplain involved erosion of parts of the Bigbear system. Paleocurrents, in addition to clast provenance, indicate regional westward tilting with uplift in the east-northeast, in the Coronation Gulf-Tree Basin area (Fig. 9.1). The basinwide westward tilting reflects continuation of a process first recorded in the upper parts of both the Bigbear and Fault River systems.

Cape MacDonnel Facies

This facies outcrops in rubbly low-relief bluffs in the central part of the Lady Nye system (Fig. 9.2b) and as steep cliffs capped by diabase sills on the north shore of Hornby Bay. The facies is transitional into underlying Bunn Creek Facies, the overlying deltaic-marine clastics, and Unit 9 dolostones. The true thickness is unknown, but is estimated in excess of 600 m.

The facies is composed almost completely of medium grained quartz arenite. In the Teshierpi Highlands and Lac Rouvière area, the arenites are well-sorted, hematitic, and occur in trough crossbed sets up to 1.5 m. Rippled surfaces and thin quartz pebble lags are common. Towards the north shore of Hornby Bay (Cape MacDonnel area), the arenites are nonhematitic and occur in trough and tabular-planar 1-2 m crossbedded cosets, separated by thin scoured micaceous siltstones and/or green to red mudstone. Paleocurrents are consistently unimodal and range from west to northwest.

Interpretation. The Cape MacDonnel Facies was deposited by westward-flowing sandy braided rivers (braidplain) similar to the Lower Platte River (cf. Smith, 1970) and Tana River (cf. Collinson, 1970). Relatively high-energy proximal fluvial environments in the Teshierpi Highlands/Lac Rouvière area are suggested by the predominance of large trough crossbeds and a lack of mudstone intercalations. Because the present Teshierpi Highlands are the result of dip-slip movement (northwest side down) on Teshierpi Fault, the through-going westward-directed paleocurrents of the Lady Nye system predate reactivation of this fault.

Fan Delta/Marginal Marine Facies

This facies, exposed on the south shore and islands of Dease Arm and from the west end of Dismal Lakes south to East River, marks the transition from fluvial Lady Nye arenites into marine dolostones of Unit 9. Its maximum thickness is 100 m in the westernmost exposures (Narakay Islands, Dease Arm) and it thins eastward to zero at the south shores of Dismal Lakes. The thinning is most marked across a north-trending lineament referred to here as Leith Line (Fig. 9.2b).

Black to red and green lenticular-bedded shales and mature quartz arenites are characteristic of this facies. The arenites range from tabular sheets to scour-based lenses, both with flat- to ripple-lamination and commonly rippled top surfaces. Siliceous and locally calcareous angular mudchips are common in the arenites and the latter are usually associated with oolites. The interlayered shales are commonly mudcracked and infilled by the overlying arenites.

Interpretation. This facies, bounded below by fluvial Lady Nye arenites and above by basal Unit 9 dolostone, is interpreted as a fan delta complex (cf. McGowen, 1971). A typical depositional cycle involved deposition by traction currents during fluvial flood run-off (arenites) followed by suspension deposition (mudstones) and subaerial exposure. Oolites and calcareous intraclasts were deposited by onshore currents during marine inundation. The eastward thinning and eventual pinchout of this facies is attributed to shallowing to the east of the Leith Line.

Unit 9

This unit gradationally underlies and in part interfingers with underlying Lady Nye system siliciclastics, and is transitional into the overlying mudstones and arenites of Unit 10 (Fig. 9.3). It is exposed from the Narakay Islands (Dease Arm) eastwards to areas adjacent to, but not across, the Teshierpi Fault (Fig. 9.2b). A maximum thickness of at least 450 m occurs in Dease Arm area, diminishing eastwards to less than 60 m. Thinning is most pronounced across the Leith Line. The lithologic sequence is divided into three facies (Fig. 9.9): a lower clastic facies (9A); a domal stromatolite facies (9B); and an upper clastic facies (9C).

Lower Clastic Facies (9A)

This facies ranges from 20 m thick near Teshierpi Fault to more than 90 m thick along the north shore of Dease Arm. Its base is defined by the first appearance of dolostone. The gradational upper boundary is arbitrarily designated to be the last appearance of microdigitate domal stromatolites. This facies records the transition from predominantly siliciclastic to predominantly carbonate deposition.

The lower part of the facies consists of interbedded quartz arenite, siliceous mudstone, intraclastic dolostone, quartzose dolarenite, and dololutite characterized by domal stromatolites. The stromatolites are symmetric domes which display laterally continuous smooth laminations (LLH), intraclastic channel fill (SH-C) (Logan et al., 1964) and small digitate columnar protuberances. Arenaceous beds are massive or show tabular-planar cross stratification and commonly display herringbone crossbeds. Overlying trough crossbedded quartz arenite and silicified oolitic dolarenite commonly enclose flat-based bioherms of digitate micro-domal stromatolites, precursors to the more extensive bioherms/biostromes at the top of the facies. Paleocurrents from the crossbedded units show large dispersion, but commonly have a strong maximum to the northwest or northeast.

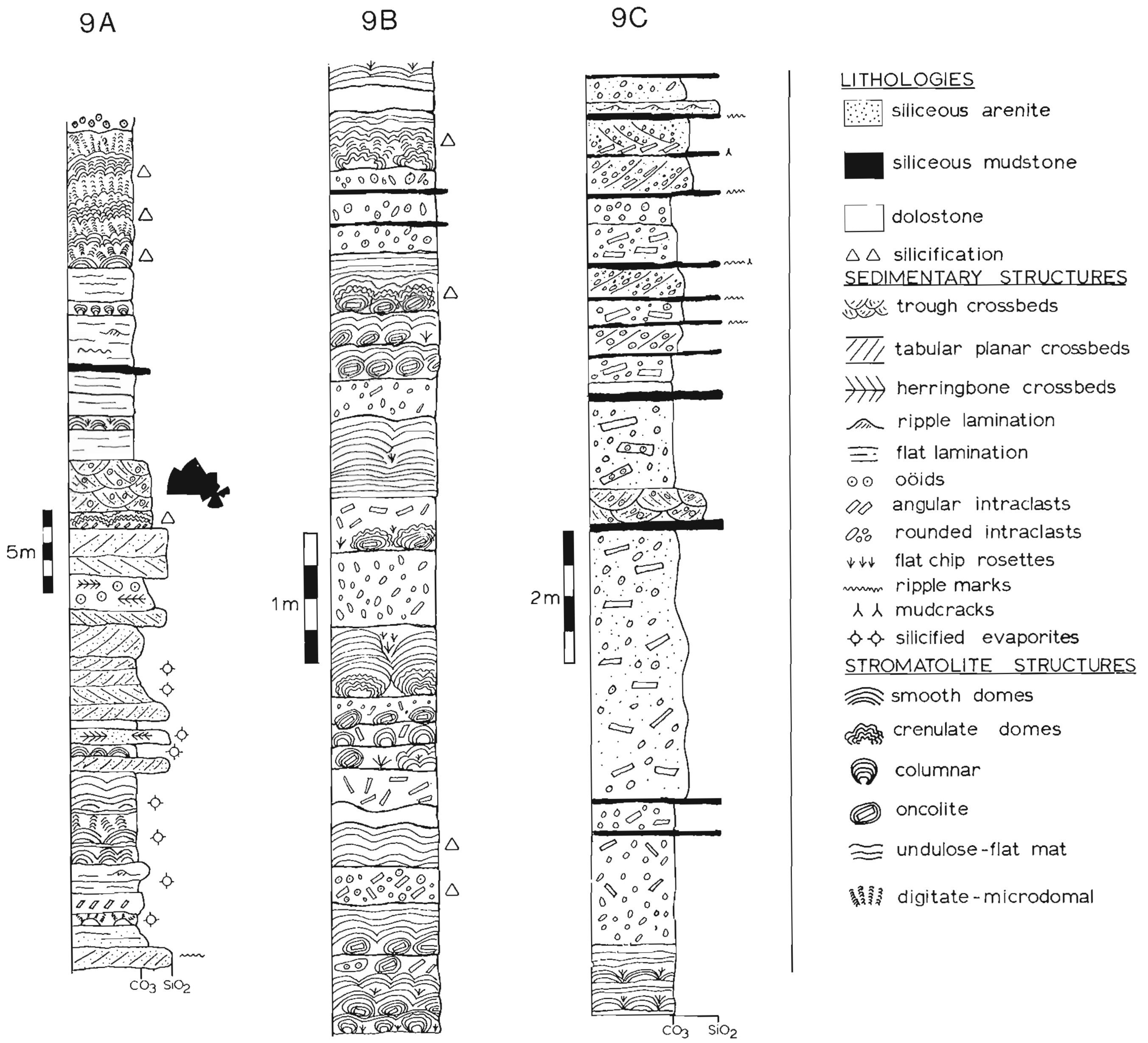


Figure 9.9. *Lithofacies subdivision for Unit 9. Note different scales at right of columns.*

The upper part of the facies is composed of crossbedded oolites, low-relief, laterally linked, domal stromatolites (LLH), and coarsely crystalline dolostone with rare ripple marks and current laminations. The top of the facies is marked by several prominent basinwide marker beds of silicified bioherms and biostromes of compound domal stromatolites. These stromatolites exhibit a smooth domal morphology that grades into crenulate domes, and finally into radiating digitate microdomes. South of Dismal Lakes the domes are relatively small and occur only in one bed, whereas up to five distinct domal beds occur in the Dease Arm area. These domes are overlain by, and in places display channel fills of, massive well-sorted oolites.

Silicified nodular rosettes with bladed pseudomorphs that resemble the early diagenetic "cauliflower evaporites" described by Chown and Elkins (1974), Walker et al. (1977), and Milliken (1979) occur sporadically throughout this facies. They occur in cross-laminated quartzose dolarenite, scattered along laminae within stromatolites, and as coarse pods in interdomal channels. They are most abundant in a northeast-trending zone within the facies.

Domal Stromatolite Facies (9B)

The upper and lower boundaries of this facies are gradational into underlying (9A) and overlying (9C) facies. It is composed of smoothly laminated to slightly crenulate domal stromatolites, commonly with oncolitic cores, and interbeds of mixed oolitic and intraclastic debris (Fig. 9.9). Thin beds of flat to ripple-laminated red to green siliceous mudstone and dololutite also occur. The oncolites and domal stromatolites are commonly transitional upward into flat algal mats. Small (<1 cm) arborescent stromatolites form layers of palisade-like structures (cf. Hardie, 1977) within some of the smooth domes and flat mats. The domes are symmetric to slightly elongate and commonly have flat-chip intraclasts in interdomal depressions. The interbedded clastics, composed of oolites and round to subround intraclasts with rare coarse dololutite fragments are usually massive and only rarely show cross stratification. Although desiccation cracks are absent, tepee structures (cf. Assereto and Kendall, 1977) and especially silicified rosettes (cf. 9A) are locally common.

Upper Clastic Facies (9C)

This facies, and the transition from Unit 9 into Unit 10, is poorly exposed. It represents the transition zone from stromatolitic dolostone of Unit 9 into the siliciclastics of Unit 10. It is composed of massive to crossbedded units of intraclastic dolarenite with some oolites, intercalated with thin siliceous argillite layers (Fig. 9.9). The intraclastic beds are well sorted, but commonly contain slabs up to 10 cm of massive to finely laminated tan dololutite (cryptalgal laminite), suspended in the sand-size matrix. Red and green argillite layers which drape rippled surfaces are commonly mudcracked and in rare cases display runzel marks. Conglomerate beds with clasts of silicified Unit 9 lithologies, some exhibiting weathering rinds, are prominent in exposures of this facies close to Teshierpi Fault. The amount and size of this material decreases substantially to the west. The boundary with Unit 10 is arbitrarily chosen as that point at which siliciclastic layers form greater than 50 per cent of the lithology.

Interpretation of Unit 9

Unit 9 dolostones represent the deposits of a marine transgression. Interfingering of siliciclastics and carbonates in lithofacies 9A is most pronounced in the west but absent in the east, reflecting relatively rapid eastward shoreline progradation over the deltaic/marine and distal fluvial facies of the Lady Nye system. The northeast trend of maximum abundance of silicified sulphate rosettes is inferred to reflect an approximate shoreline orientation. This interpretation is supported by paleocurrents both perpendicular and parallel to this trend. Lithofacies 9A, recording deposition in a nearshore intertidal to subtidal setting, is remarkably similar to the late Proterozoic succession of deltaic and intertidal siliciclastics and biostromal clastic dolostones described by Tucker (1977). The silicified stromatolites in upper 9A have a morphology similar to the subtidal mound and channel belt stromatolites of the middle Proterozoic Pethei Formation (Hoffman, 1974), and probably formed shallow subtidal buildups. They are the only recognized subtidal facies in Unit 9, and their basinwide occurrence represents the maximum extent of marine transgression during Hornby Bay Group deposition.

Gradual shoaling of the short-lived "Hornby Bay Platform" is recorded by progradation of the 9B lithofacies assemblage, interpreted as a supratidal to high intertidal algal marsh and tidal complex (cf. Hardie, 1977; Hoffman, 1975). Continued regression led to deposition of the 9C lithofacies assemblage in a transitional marine tidal flat to outer deltaic plain setting (lower Unit 10). Weathered and silicified clasts of Unit 9A in the 9C lithofacies close to Teshierpi Fault are indicative of the first stage of reactivation of this northwest-side-down fault.

Lithified 9A and 9B Facies exposed on the upthrown, southeast block were subaerially silicified(?), eroded and deposited as an apron on the downthrown block. Teshierpi Fault reactivation is interpreted as a local manifestation of regional cratonic uplift which resulted in deposition of Units 10 and 11.

Narakay Islands Volcanic Complex

The Narakay Islands Volcanic Complex, on islands of the same name in Dease Arm of Great Bear Lake (Fig. 9.2b), records a period of mafic volcanism that was coeval with late Unit 8 to Unit 9 deposition. Four eruptive events are recognized, spanning more than 400 m of stratigraphic section. The volcanic rocks are almost completely pyroclastic and were deposited as agglomerates, airfall and pyroclastic surge, lapilli and ash tuffs. Carbonate alteration of the volcanics is ubiquitous except for a single massive porphyry composed of fine grained olivine-plagioclase basalt, interpreted to record the original composition of the pyroclastics.

The tectonic significance of the volcanic complex is not completely understood, but it may represent mantle-derived magmas released during crustal fracture produced during north-south extension and formation of the Dismal Lakes Platform.

Unit 10

Exposures of Unit 10 occur discontinuously in a belt that extends from the south shore of Dismal Lakes to the mouth of Dease River (Great Bear Lake); it is absent east of the Teshierpi Fault (Fig. 9.2a,b). It has a gradational to sharp (nonconformable) contact with underlying Unit 9 and a gradational contact with Unit 11, except adjacent to the Teshierpi Fault where it has been partially to completely eroded prior to deposition of Unit 11. The unit thickens from 120 m along the south shore of Dismal Lakes to more than 500 m at East River (Fig. 9.2b). It thins to less than 80 m to the northwest where it nonconformably overlies Unit 9. Only two nearly complete sections are exposed and are presented below.

On the south shore of Dismal Lakes, Unit 10 is composed mostly of interbedded red to green mudstone and quartz siltstone and arenite with granular arkoses and thin dolostones lower in the unit (Fig. 9.10). The lower third of the sequence displays interrupted fining-up cycles composed of mudchip-bearing feldspathic to arkosic arenite, with granules of Units 8 and 9, that fine-up to siltstone, mudstone, calcareous mudstone and finally, dolostone. The arenites display small scale trough to ripple crossbedding, plane to massive bedding, rest on weakly scoured surfaces, and have wave-rippled tops. The siltstones commonly display wavy to lenticular bedding and a variety of current and wave ripple marks. Mudstones are commonly mudcracked and grade into thin flat laminated to intraclastic dolostone.

The upper two thirds of the unit is composed of mudchip-rich, fine grained, quartz arenite and siltstone and pervasively mudcracked red mudstone. These lithologies occur in thin (<10 cm) arenite-mudstone couplets that comprise larger scale (4-8 m) poorly organized fining- or coarsening-upward cycles. The thicker arenites (20-40 cm) display a tabular geometry and are commonly massive to plane bedded with thin mud flasers. In situ desiccation breccia and compaction obscure sedimentary structures in thinner bedded parts of the unit. The amount of arenaceous material increases as the unit grades in to Unit 11.

A much thicker and coarser grained succession is exposed at East River. There, the base of Unit 10 is composed of lenticular arenites and desiccated red mudstones, similar to the south Dismal Lakes section. This is overlain by three facies assemblages which define two coarsening-upward/fining-upward cycles. Flaser, wavy, and lenticularly bedded green argillite, siltstone, and arenite (facies a, Fig. 9.10), commonly with soft sediment deformation and injection structures coarsens up into trough crossbedded pebbly to well sorted quartz arenite (facies b). The

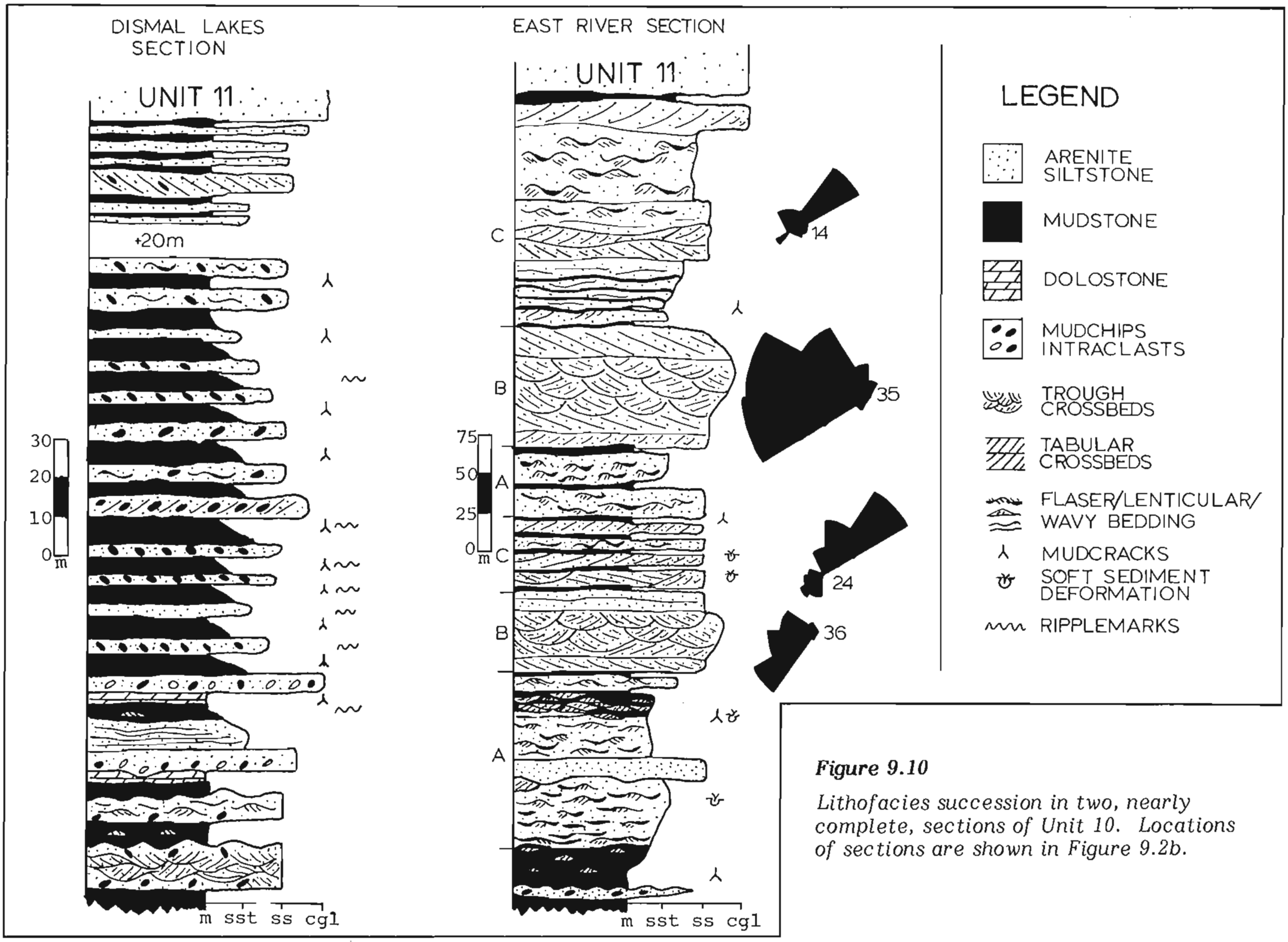


Figure 9.10

Lithofacies succession in two, nearly complete, sections of Unit 10. Locations of sections are shown in Figure 9.2b.

arenites have strong unimodal southwest paleocurrents, and contain pebbles of vein quartz and Unit 8 arenite. Flat laminated intervals less than 2 m thick occur near the top and bottom of this facies. The arenite facies fines upward into interlayered fine grained arenite bodies (20-40 cm thick) separated by thin mud films (facies c). These arenites display small scale (<20 cm) tabular planar and trough crossbedding and undulose planar bedding. Paleocurrents are weakly bimodal-bipolar with a strong northeast mode. The a-b-c facies sequence is repeated twice before Unit 10 grades into overlying Unit 11 arenites.

Interpretation of Unit 10

The East River succession is interpreted to represent two cycles of constructive/destructive delta building in a marginal marine setting (cf. Vos, 1977). Floodplain facies at the base of the section grade into facies a, a tidally influenced delta plain to tidal flat (Vos, 1977; Klein, 1977). The arenites of facies b record progradation of a braided fluvial distributary (fan delta lobe?) facies with flat laminated intervals representing marine swash bar or beach accretion (cf. Vos and Ericksson, 1977) or fluvial, emergent bar plane beds. Facies c formed during periods of fluvial quiescence when deposition was dominated by destructive phase shallow marine reworking via longshore and onshore currents (cf. Hayes, 1975).

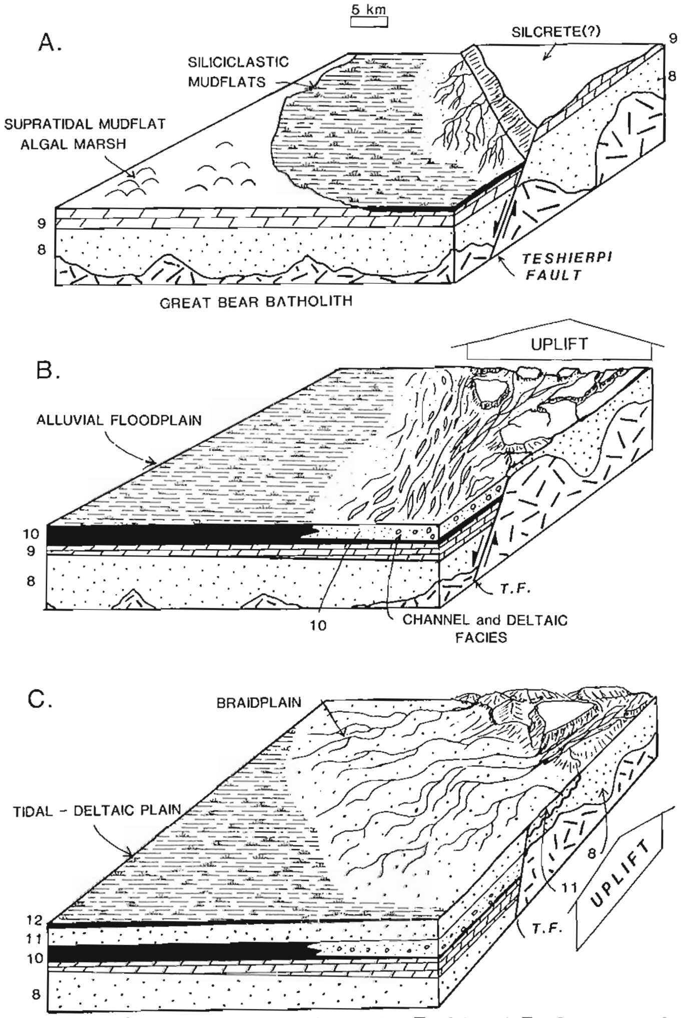
The Dismal Lakes section represents coeval floodplain/overbank deposition (cf. Allen, 1965) with each fining-up cycle recording thin sheet flood events. The lack of evaporites and scarcity of traction current structures, in addition to the terrestrial provenance of the lower arenites, favour a floodplain rather than high intertidal flat environment (Klein, 1970).

Uplift of lower Hornby Bay Group and Aphebian basement to the east-northeast of, and including, the Teshierpi-Bigtree Highlands, shed clastics to the west-southwest and overwhelmed the carbonate flats of facies 9C (Fig. 9.11). Syndepositional movement on the Teshierpi Fault may have locally affected subsidence rates and restricted coarse clastic deposition to the south (East River) with a laterally fining and thinning floodplain facies to the north (Fig. 9.11). Siliciclastic deposits of Unit 10 signal the beginning of the second siliciclastic to carbonate cycle.

Summary of Hornby Bay Group

Deposition of the Hornby Bay Group commenced after the cessation of orogenic activity in the underlying Wopmay Orogen. Uplift and peneplanation of the Teshierpi-Bigtree Highlands shed a fining-upward sequence of fluvial siliciclastics (Bigbear system) to the southeast. Late stage deposition in this system was punctuated by intrabasinal normal faulting along the Sinister Fault. This was coeval with normal faulting in the Fault River area and formation of the trough-shaped Fault River system depocentre. These events reflect a period of crustal tilting and a change from a southeast to southwest paleoslope, during which the Teshierpi-Bigtree Highlands were incised.

Burial of the Bigbear and Fault River systems and infill of the incised peneplain occurred as the westward-flowing Lady Nye system was deposited on the regionally subsiding basement. Thickening of upper Lady Nye system arenites and distal deltaic facies to the west are the first indication of increased subsidence across the Leith Line. Continuing basinal subsidence resulted in marine transgression and formation of a short-lived carbonate platform (Unit 9) with local mafic volcanism. Carbonate lithofacies record a gradual shoaling of the platform from a subtidal stromatolitic barrier complex (9A) into intertidal and supratidal algal marshes (9B). Shoaling coincident with the beginning of basement uplift and rejuvenation culminated in drowning of the carbonate marshes and mudflats by the fluvial-deltaic siliciclastics of Unit 10. Normal faulting, in response to regional crustal warping, occurred along Teshierpi Fault and locally enhanced source area relief and restricted the trend of Unit 10 channel facies to a west-southwest pattern, subparallel to the fault. On the upthrown eastern block of the Teshierpi Fault, Unit 9 and Lady Nye system lithologies were eroded, resulting in an unconformity between basal Dismal Lakes Group siliciclastics and Hornby Bay Group. Erosion of Hornby Bay Group arenites (Lady Nye system) provided abundant arenaceous debris that was deposited as the supermature second-cycle fluvio-marine arenites of Unit 11.



A) Normal fault movement along the Teshierpi Fault exposed Unit 9 dolostone to subaerial silicification with erosion and transportation to the southwest.

B) Regional uplift caused erosion of Units 9, 8 and basement. Drainage was in part controlled by oblique dip-slip movement towards the southwest along the Teshierpi Fault.

C) The culmination of uplift resulted in further erosion of Hornby Bay Group and basement rocks and deposition on a north-northwest facing fluvial (Unit 11) – Deltaic (Unit 12) complex.

Figure 9.11. *Schematic representation of tectonic events and depositional environments during upper Unit 9 (9C) of the Hornby Bay Group through Unit 12 of the Dismal Lakes Group. View is from East River looking northeast along the Teshierpi Fault (see Fig. 9.2b).*

DISMAL LAKES GROUP

Introduction

Basal Dismal Lakes Group sandstones (Unit 11) rest with local unconformity on Hornby Bay Group sediments and Aphebian basement (Fig. 9.2a). This locally unconformable relationship is inferred to be related to syndepositional west-side-down movement along Teshierpi Fault. Fault movement began during late Hornby Bay Group deposition (see above), and culminated prior to deposition of Unit 11 arenites.

Fault-related tectonic activity initiated a second siliciclastic (Units 10 to 13) to carbonate (Units 14 to 16) depositional sequence (Fig. 9.12) of similar character to the "grand cycle" of Aitken (1966). Basal Dismal Lakes Group

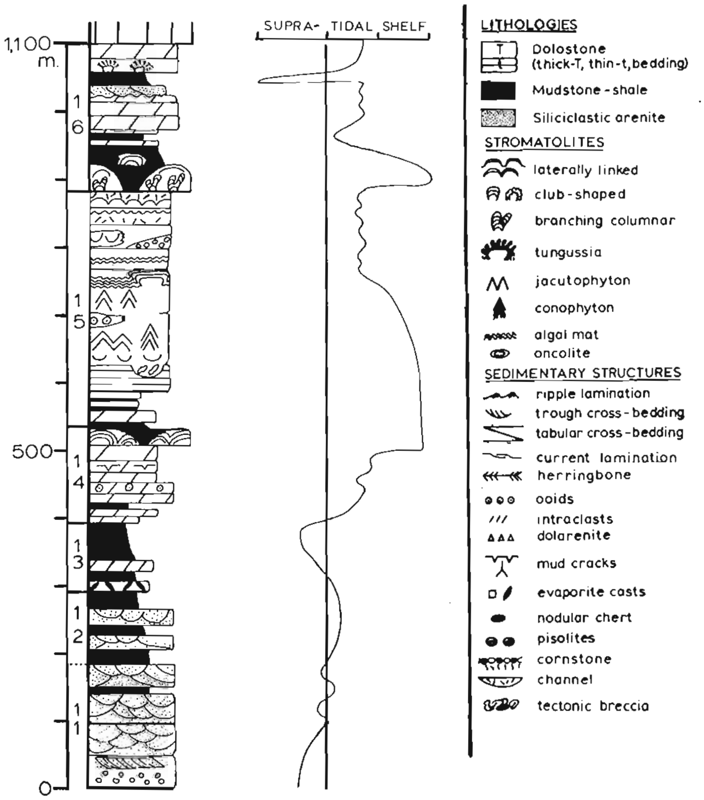


Figure 9.12. *Stratigraphic column for Dismal Lakes Group showing environments of deposition.*

siliciclastics fine upwards from quartz arenites (Unit 11) to mixed arenites and shales (Unit 12), and finally into mixed mudstones and dolostones of Unit 13 (Fig. 9.10). This fining-upward siliciclastic sequence reflects gradual diminution of topographic relief and development of a broad stable platform.

Upper Dismal Lakes Group dolostones were deposited during eastward transgression of this northward and westward-sloping platform. Extreme lateral continuity of contained dolostone units reflects a period of regional tectonic stability. Interruption of stable platform deposition occurred near the end of Dismal Lakes deposition (Unit 16) as marked by a regionally extensive disconformity which extended as far east as correlative rocks in Bathurst Inlet (Campbell, 1978).

Unit 11

Outcrops of Unit 11 occur in a belt that extends from northeast of the Muskox Intrusion to the north of Dease River (Fig. 9.2a,b), although a complete section is not exposed. West of the Teshierpi Fault, this unit is gradational and conformable with underlying Unit 10; in areas east of the fault it rests unconformably on Unit 8 and locally basement. In the region immediately northwest of the fault where the Herb Dixon and Teshierpi faults join (Fig. 9.2a,b), Unit 11 rests unconformably on fault-bounded blocks of Units 10, 9, 8 and basement. The base of Unit 11 across the Teshierpi Fault is not considered to be a time equivalent surface.

Although the contact is poorly exposed, Unit 11 grades into the overlying sandstone-shale of Unit 12. The thickness of this unit is unknown; a maximum of 300 m has been measured and the unit thins towards the east to an estimated 70 m.

Unit 11 is composed of mature to supermature medium grained siliceous quartz arenite, rarely with scattered vein quartz granules, granular coarse grained quartz arenites with clay-silica cement, and vein quartz pebble-bearing quartz arenite. A thin paraconglomerate with cobbles of Unit 8, vein quartz, and locally Unit 9, is found at the unconformity. Three representative partial sections of Unit 11 are shown in Figure 9.13.

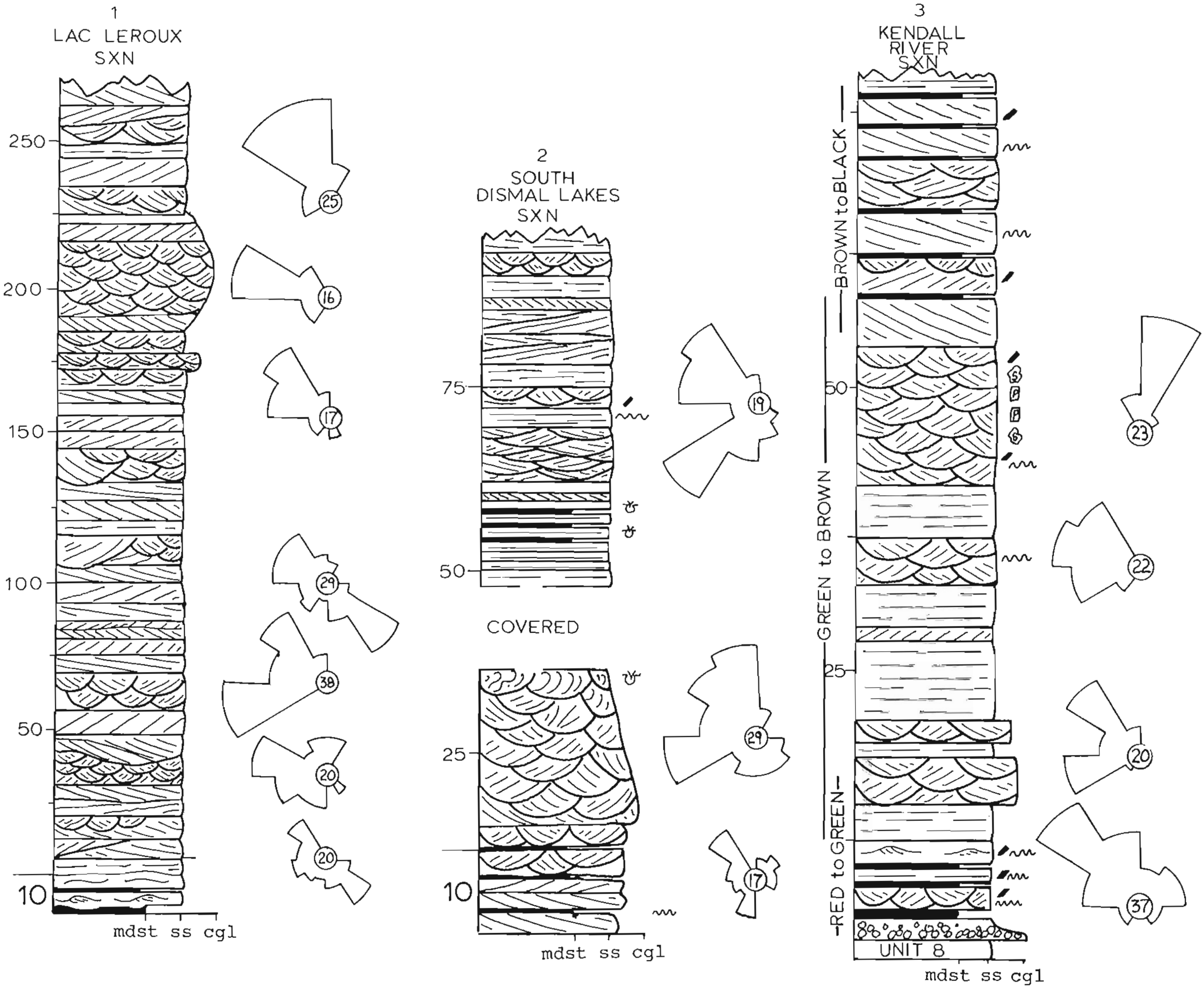


Figure 9.13. Partial stratigraphic columns for Unit 11. Section locations on Figure 9.2b. Scale is in metres and differs between sections. None of the sections have an exposed top (Unit 12). Colours to the left of column three refer to colour of mudstone and argillite interlayers and reflect the change from oxidized (red) mudstones in lower Unit 11 to reduced (black-brown) mudstones in upper Unit 11 and Unit 12. S is symbol for nodular sulphate and P for pyrite.

Two lithofacies assemblages are mappable on a regional scale and irregularly interlayered throughout the unit with several assemblages of only local extent. Submature to supermature pebbly to granular quartz arenites typically display trough crossbeds with strong unimodal paleocurrents. Transport directions are west-southwest in the lower part of the unit, and north-northwest in the upper part.

The second facies assemblage is composed of well-sorted quartz arenite that has a resistant to friable weathering character and displays a variety of sedimentary structures. Thin (<20 cm) trough and tabular-planar crossbeds and plane bedding are typical. Current and wave ripple marks, megaripples, soft sediment deformation, thin mud films, rare reactivation surfaces, and herringbone tabular planar crossbeds also occur. Paleocurrents from this facies are commonly bimodal-bipolar to the northwest and southeast, polymodal, and rarely unimodal to the southeast.

Interpretation

Unit 11 is interpreted to have been deposited in a mixed marginal marine-fluvial environment. On the basis of unimodal paleocurrents, commonly submature character of the lithology, and the lack of sedimentary structures and/or cycles of tidal creek (Barwis, 1978) or tidal inlet origin (Hayes, 1975; Kumar and Sanders, 1974) the submature lithofacies represents periods of deposition in a braided fluvial channel. The thin-bedded facies was deposited in a tidally-influenced shallow marine environment, as suggested mostly by the paleocurrent patterns. Inferred depositional processes include wash-over fans (Andrews, 1970), beachface and swash bar accretion (Vos and Ericksson, 1977) for flat-bedded intervals, subtidal to intertidal sand and flats (Klein, 1970, 1977) or estuarine tidal deltas (Boothroyd and Hubbard, 1975) for units with bimodal to polymodal paleocurrents.

Uplift east and northeast of Dismal Lakes provided source area relief and led to deposition by southwest-flowing fluvial systems of mature to submature second-cycle quartz arenite (Fig. 9.11). Dip-slip movement on the Teshierpi Fault during this time is interpreted to be a local response to stresses generated by regional uplift. Terrestrially-derived arenites were reworked by marine currents during periods of fluvial inactivity. Regional subsidence was initiated during deposition of Unit 11. The paleoslope changed from southwest to north-northwest and led to formation of the north- and west-facing Dismal Lakes Platform. Subsidence/sea level rise also allowed onlap of Unit 11 onto the upthrown eastern block of the Teshierpi Fault (Fig. 9.11).

Unit 12

Mixed sandstone-shale lithologies of Unit 12 conformably overlie and interfinger with shale-poor quartz arenites of Unit 11. The contact is arbitrarily defined as the first appearance of metre-thick shale beds in the section. Shales range in colour from greenish brown to black. The minimum thickness for Unit 12 is estimated at 100 m in the Dismal Lakes area, where the unit is best exposed. Lateral thickness variations cannot be determined because of recessive weathering. Essential lithologic components of Unit 12 are 1-20 m sequences of lenticular or flaser-bedded mixed sandstone-shale, and 2-25 m cross-bedded quartz arenite bodies.

Sedimentary structures in the shaly intervals include symmetric, asymmetric and interference ripple marks, injection-modified mudcracks, small channel scours and rill marks. Both lenticular and flaser bedded units, the dominant bedding types, show polymodal paleocurrent distributions. Rare isolated oolitic units less than 20 cm thick are interspersed in this otherwise entirely siliciclastic section.

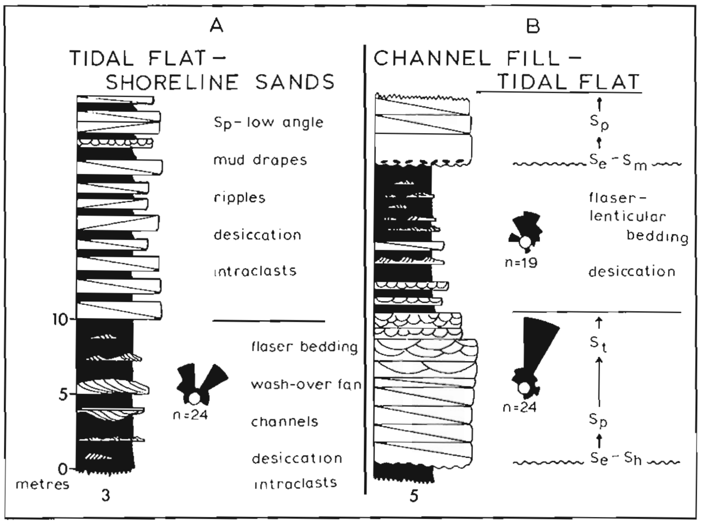


Figure 9.14. *Generalized measured sections showing common sandstone-shale sequences of Unit 12 and interpreted depositional settings. See text for discussion. Section numbers (3, 5) correspond to locations of Figure 9.19.*

Sandstone sequences are grouped into two categories (A and B) based on sandstone body geometry and bedforms. Sequence A (Fig. 9.14) displays tabular- and wedge-planar crossbed sets which gradationally coarsen upward from the underlying lenticular-bedded shaly facies described above. Individual tabular-planar crossbed sets (average 1 m) are intercalated with green sandy argillite beds up to 20 cm. Individual sandstone beds form compound units up to 6 m thick.

Sequence B consists of rhythmically-repeated, scour-based, sandstone- to mixed sandstone-shale sequences (Fig. 9.14). Sandstone bodies average 15 m and pinch out laterally over 750-1000 m. Massive and tabular-planar bedsets (20-120 cm) at the base of a sandstone body pass upward into thinner trough crossbedded arenites and finally into mixed sandstone-shale beds. These intervening shaly units range from 5-10 m thick. Paleocurrents in lower portions of the sandstone bodies are unimodal to the north becoming polymodal or bimodal-bipolar in the capping trough crossbedded arenites and in shaly lenticular bedded units.

Siliciclastic deposits of upper Unit 12 interfinger with Unit 13 sandy stromatolitic marine dolostones in the west (Fig. 9.14) whereas in the east, they are overlain abruptly by terrestrial wadi-fan deposits of Unit 13.

Interpretation

Lenticular- and flaser-bedded mixed sandstone-shale intervals are interpreted as tidal flat deposits, possibly deposited in part on marine-marginal portions of a delta plain (Fig. 9.15). Sequence A sandstones gradationally overlie and interfinger with these tidal flat deposits producing an overall coarsening-upward sequence. Tabular-planar crossbed sets are inferred to have formed during sandwave migration, and may represent longshore deposits similar to those described by Coleman (1966) from the Mississippi River delta, or by Meckel (1975) from the Colorado River delta. Wedge-shaped sets resemble beach shoreface deposits and isolated oolitic units may indicate storm washover (cf. Hayes, 1967) from a laterally associated marine carbonate facies (see Unit 13A).

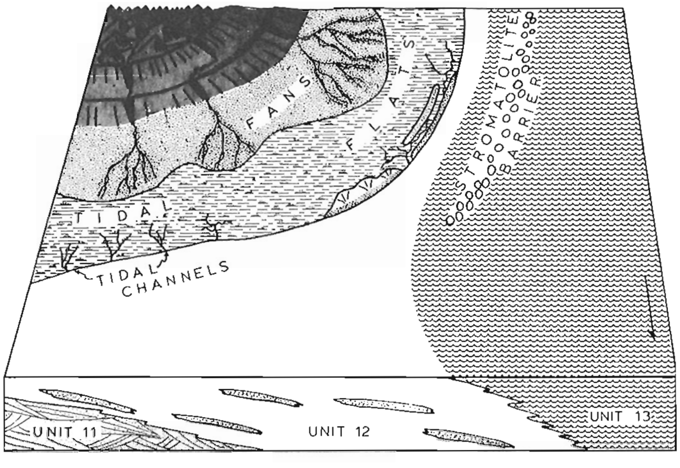


Figure 9.15. *Paleogeographic reconstruction of Unit 12 and lower Unit 13 looking south. Source area in southeast corner corresponds to Teshierpi Highlands. Cuestas represent uplifted Hornby Bay Group sediments.*

Sequence B fining-upward cycles are interpreted as channel-fill deposits. The unimodal transport, shown by tabular-planar crossbeds in the basal fill, probably represents migration of transverse bars in distal deltaic distributary channels (cf. Oomkens, 1974; Eriksson, 1977; Van de Graaf, 1972) or in ebb-dominated tidal channels (cf. Barwis, 1978). Greater paleocurrent dispersion in upper portions of Sequence B sand bodies indicates tidal current reworking of the channel fill which eventually was succeeded by deposition of mixed sandstone-shale on tidal flats. Cyclic repetition in sequence B probably resulted from repeated channel migration.

A westward transition into sediments deposited in more open marine environments, and north- and northwesterly-directed paleocurrents suggest a northeasterly depositional strike approximately parallel to Teshierpi Fault and adjacent highlands (Fig. 9.15).

Unit 13

The transition from black shales and sandstones of Unit 12 into red mudstones and dolostones of Unit 13 is sharp and conformable in the Dismal Lakes area. To the west though, shales in the upper part of Unit 12 interfinger with quartzose stromatolitic dolostones of basal Unit 13 (Fig. 9.16). Unit 13 thickens westward from approximately 100 m in the Dismal Lakes area, to 150 m in the Dease Lake area, 30 km to the west.

In the Dismal Lakes area, three poorly defined lithologic zones are recognized (Fig. 9.16): a) basal massive red mudstone containing thin scour-based sheets and channel-fill deposits of lithic arenite; b) red mudstones with intraclastic rip-up zones, rare sand-sized siliciclastic detritus, and c) uppermost, red shaly mudstones containing halite hoppers and abundant mudcracks interbedded with current-laminated red-tan siltstones. Silty dolostone beds appear near the top of this section and grade upward into clastic-stromatolitic dolostones of Unit 14. Overall trends in the section are an upward fining and an increase in the carbonate/siliciclastic ratio.

In the Dease Lake area three distinct lithofacies (equivalent to informal members) are recognized in Unit 13 (Fig. 9.16). Lithofacies 13A consists of quartzose-oolitic-intraclastic dolarenite and stromatolitic dolostone. Tabular-planar and herringbone crossbed sets are common.

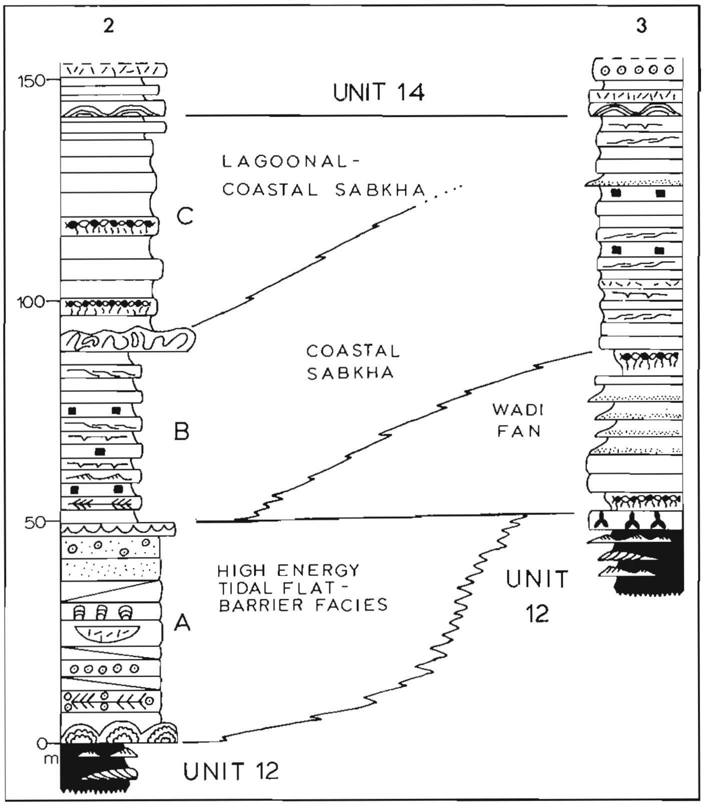


Figure 9.16. *Stratigraphic sections of Unit 13 from Dease Lake (2) and Dismal Lakes (3) showing facies changes and interpreted depositional environments. Symbols as in Figure 9.10. Section locations (2, 3) shown in Figure 9.19.*

Associated structures such as small channel fills, flat-chip conglomerates, mudcracks, symmetrical and interference ripples and ball-and-pillow structures are typical. This lithofacies has been recognized only west of the Leith Line, where it interfingers with underlying Unit 12 black shales. Lithofacies 13A passes conformably upwards into 13B.

Lithofacies 13B consists of 20-60 cm interbeds of red mudstone and tan silty dolostone. Red mudstones contain halite hoppers and mudcracks, whereas the tan dolostones typically contain small-scale ripple cross strata and parallel current laminae. The contact between lithofacies 13B and 13C is best exposed near Dease Lake where red mudstones and tan dolostones of 13B pass gradationally upward into pink mudstones and cream dolostones of 13C.

Lithofacies 13C consists of half-metre thick interbedded pink mudstone and pink-grey dololutite. Parallel current lamination is the only primary sedimentary structure observed in this lithofacies.

Extensive soft sediment deformation typifies lithofacies 13C in the area between Dease Lake and Dease Arm, Great Bear Lake. Sediments become progressively more deformed westward. A basal 2 m dololutite breccia in the east thickens westward to a maximum of 6 m. This breccia is overlain west of Dease Lake by a zone of recumbently-folded dololutites which pass upwards into upright isoclines. Fold axes are subhorizontal throughout and parallel to a major east-trending fault – the East River Fault (Fig. 9.2b). The basal breccia parallels the lithofacies 13B-13C contact and the upper contact shows strongly brecciated lithofacies 13C draped by gently-dipping Unit 14 dolostones. Unit 14 dolostones show minor folding and brecciation for a maximum of 1 m above the contact but are otherwise unaffected.

Interpretation

Deposition of Unit 13 represents the final stage of siliciclastic source-area stabilization and the formation of the broad, low-relief Dismal Lakes Platform. Lateral facies changes illustrated in Figure 9.16 suggest an east-to-west terrestrial-to-marine transition marked by a westward decrease in terrigenous detritus and concomitant increase in marine carbonates. Gradual eastward transgression produced a vertical succession of wadi-fan and tidal flat deposits overlain by coastal sabkha deposits which were onlapped by marine dolostones of Unit 14. An arid to semi-arid climate is implied for Unit 13, based upon the abundance of halite hoppers and gypsum pseudomorphs within this unit.

Soft sediment deformation, restricted to lithofacies 13C, was probably produced by high pore-fluid concentration within clay-rich portions of this lithofacies which preferentially weakened the rocks. This unit is bounded by tectonic contacts, indicating a post-depositional deformation origin. The tectonic nature of brecciation and correspondence of fold axes to the East River Fault trend suggest post-Dismal Lakes Group fault movement, as opposed to solution collapse.

Unit 14

Shaly to clean dolostones of Unit 14 conformably overlie mudstones and dolostones of Unit 13 (Fig. 9.17). The contact between these two units is gradational and is arbitrarily defined as the first appearance of stromatolitic carbonates. The upper contact of Unit 14 is marked by a thin stromatolite biostrome which can be mapped throughout the area. Unit 14 is an estimated 120 m thick in the Dismal Lakes area, and thickens westward to 270 m north of Dease Lake.

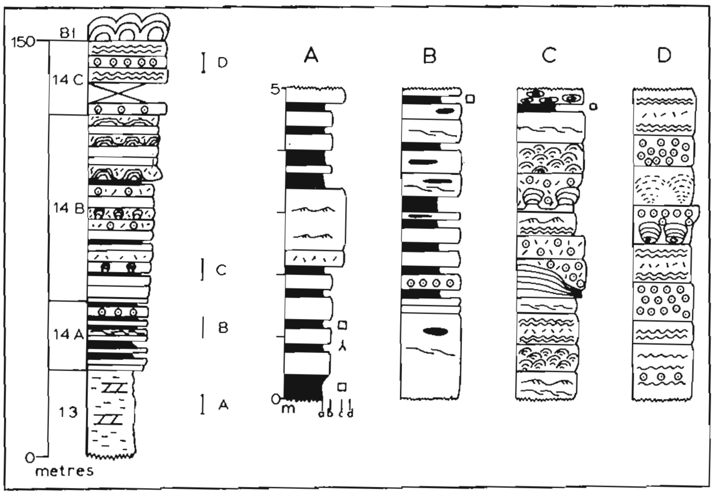


Figure 9.17. *Lithostratigraphy of Unit 14. General column on left is subdivided into gradational lithofacies as follows:*

- **A)** *silty dololutite and red mudstone (upper Unit 13);*
- **B)** *silty dololutite with black shale partings (lithofacies 14A);*
- **C)** *clastic, stromatolitic dolostone (lithofacies 14B);*
- **D)** *oolitic dolarenite and cryptalgal laminite (lithofacies 14C).*

Distinctive biostrome which caps unit provides principal marker bed for correlation (a-lutite; b-siltite; c-arenite; d-rudite).

Unit 14 is divided into three lithofacies, on the basis of contained lithologies and associated sedimentary structures (Fig. 9.17). The biostrome which caps Unit 14 is discussed separately.

Lithofacies 14A

Lithofacies 14A is a uniform sequence of current-laminated dololutite with thin shale partings (Fig. 9.17, section B). Broad low domal stromatolites (LLH-C) occur sparsely on bedding planes; some are draped by 2-4 cm beds of oolitic-intraclastic doloarenite. No scour, rip-up or desiccation structures occur within the lithofacies. The transition from lithofacies 14A to 14B is marked by an increased percentage of intraclastic-oolitic dolarenite (up to 30%) relative to interbedded cryptalgal and current-laminated dololutite (Fig. 9.17, section C).

Lithofacies 14B

Low domal and club-shaped stromatolites (SH-C and SH-V), oncolites and flat algal mats are the dominant cryptalgal structures of lithofacies 14B. The SH-C and SH-V stromatolites are laterally bounded by intercolumn scour channels. Broad shallow channel structures (1.0 x 0.2 m) dissect the dololutite beds. Current and oscillation ripples are common bedforms; desiccation cracks and halite hoppers are rare.

Lithofacies 14C

Lithofacies 14C contains variable amounts of oolitic dolarenite intercalated with densely laminated cryptalgal laminite with abundant synaeresis features. Oolitic units comprise no more than 40 per cent of this lithofacies. Individual oolite beds (average 40 cm) are separated by cryptalgal beds (Fig. 9.17, section D).

A stromatolite biostrome forms a continuous 1-6 m marker which rests sharply on lithofacies 14C. The succession of stromatolite forms in this biostrome consists of flat mats, low domes, and branching columnar forms, with the uppermost branching stage developed only in the Dismal Lakes area. Shaly dololutites of Unit 15 abruptly overlie the biostrome.

Interpretation of Unit 14

The vertical succession grading from siliciclastic-dominated mudstones of Unit 13 into clean clastic carbonates of upper Unit 14 is interpreted to have been produced by a steady marine transgression across the broad stable siliciclastic coastal plain deposits. Unit 14 lithofacies (Fig. 9.17) represent laterally equivalent environmental zones, illustrated in Figure 9.18. Shaly dololutites of lithofacies 14A probably accumulated in a quiet subtidal lagoon sparsely populated by stromatolites. Thin beds of oolitic and intraclastic detritus were brought in by storms from tidal flats (lithofacies 14B) seaward of the lagoon. Mixed clastic-stromatolitic deposits of lithofacies 14B are interpreted to have formed in a complex of intertidal and shallow subtidal environments including tidal flat, tidal channel and beach environments similar to the Three Creeks region of Andros Island (Hardie, 1977; Shinn et al., 1969). Interfingering oolitic and algal mat sediments (lithofacies 14C) represent relatively high-energy intertidal and supratidal deposits.

The extreme lateral continuity of the uppermost Unit 14 biostrome reflects uniformity of this low-relief platform; the absence of intercolumnar clastic detritus within the biostrome suggests a relatively low-energy depositional environment.

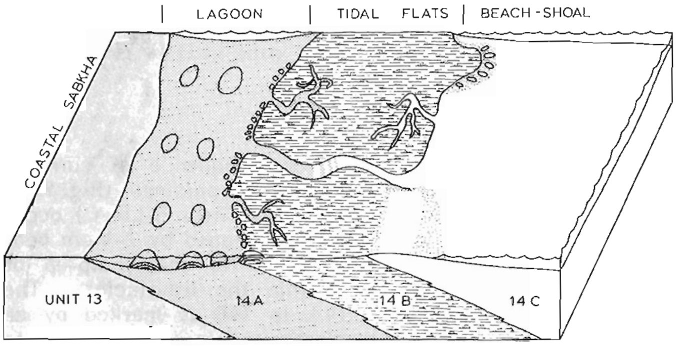


Figure 9.18. *Paleoenvironmental reconstruction of Units 13 and 14.*

Unit 15

Shaly dololutites of basal Unit 15 rest abruptly on the uppermost Unit 14 biostrome in all areas. The upper contact of Unit 15 is defined by a second biostromal marker which is continuous from Great Bear Lake to Coppermine River. Where this biostrome pinches out to the east, a stratiform chert-clast breccia defines the Unit 15-16 contact (Fig. 9.19). Unit 15 is 300 m thick in the Dismal Lakes area, thins to 90 m eastward in the September Lake area, and thickens to 360 m westward, north of Dease Arm, Great Bear Lake.

Unit 15 is divided into three distinct lithofacies, each of which has conformable contacts with successive units (Fig. 9.19). September Lake High (Fig. 9.19) is introduced here as a distinct tectonic element of the Dismal Lakes Platform which subsided at a slower rate relative to the remainder of the platform. The following description and interpretation of Unit 15 contrasts depositional histories of the Dismal Lakes Platform and the September Lake High for each of the three lithofacies.

Lithofacies 15A

Lithofacies 15A consists of shaly, flaggy-bedded, current-laminated to massive dololutite. These dololutites sharply overlie the stromatolites of the Unit 14 biostrome. Shallow-water clastic textures are common in lowermost portions of this lithofacies, but these pass rapidly upwards into structureless or parallel current-laminated dololutites. On the September Lake High, oxidized siliciclastics give this lithofacies a red colour, whereas green to black siliciclastics and grey to black dololutite rhythmites typify the deposits on the Dismal Lakes Platform to the west. Graded bedding, small-scale stratiform breccias, soft sediment folding, sedimentary boudinage and low-angle truncation surfaces (cf. Wilson, 1969; Cook and Taylor, 1977) occur in the rhythmites.

Lithofacies 15B

Lithofacies 15B is composed entirely of massive stromatolitic dolostones which form prominent beige-coloured bluffs along the entire outcrop belt of the Dismal Lakes Group. It is a maximum of 100 m thick in the Great Bear Lake area. On the September Lake High, the vertical stromatolite succession of branching conical-columnar (Jacutophyton-type), conical-columnar (Conophyton-type) and broad domal stromatolites (Fig. 9.19) produced an algal build-up that averages 40 m. Stromatolite inter-areas are filled with massive dololutite or fibrous botryoidal cements.

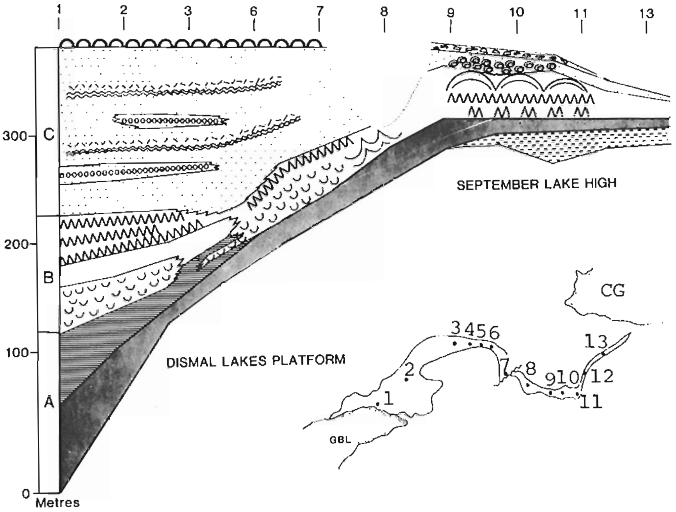


Figure 9.19. *Stratigraphic cross-section of Unit 15 showing three-fold lithofacies subdivision. Lithofacies A subdivided into siliciclastic-rich dololutite (dashed), platy dololutite (grey tone) and grey-black carbonate rhythmites (horizontal ruling). Note westward thickening and facies changes which distinguish September Lake High and the western portion of the Dismal Lakes Platform. Lower contact is Unit 14 biostrome (not shown). Upper contact marked by Unit 16 biostrome, and in the east, a silicic breccia. Location of measured sections used shown in lower right.*

Clastic intercolumnar debris is absent. North of Dease Arm, Great Bear Lake, several 1-2 m growth cycles of flat mat- to conical-columnar- to branching conical-columnar stromatolites occur. A more variable association of curly beds (Kerans and Donaldson, 1979) and conical-columnar stromatolites occurs on the central platform.

Lithofacies 15C

On the September Lake High, lithofacies 15C consists of 20 m of pisolitic and fenestral dolostone, dense cryptalgal laminite and stratiform silicified evaporite breccias. This lithofacies is significantly thicker from Dismal Lakes west to Great Bear Lake (120 m), consisting predominantly of stromatolitic, oolitic and intraclastic dolostones (Fig. 9.19). Tepee structures and edgewise conglomerate are common.

Interpretation of Unit 15

Rapid transgression, initiating growth of the Unit 14 biostrome (Fig. 9.20A), continued during deposition of lithofacies 15A. Low-energy current-laminated, dololutites accumulated on the September Lake High, while on deeper portions of the Dismal Lakes Platform, carbonate rhythmites similar to subtidal shelf facies (Cook and Taylor, 1977; Wilson, 1969; Pfeil and Read, 1980) formed below storm wave base (Fig. 9.20B). The paucity of intraformational truncation surfaces (cf. Wilson, 1969) and slump structures suggests the absence of a significant shelf-slope break in the outcrop area. Onset of sub-wave base conditions was followed by the development of conical and associated stromatolites which formed a unique build-up on the September Lake High, and smaller scale cyclic growth sequences on more actively subsiding portions of the Dismal Lakes Platform (Fig. 9.20C). The absence of wave-generated structures and clastic-filled channels within the September Lake build-up suggests that it

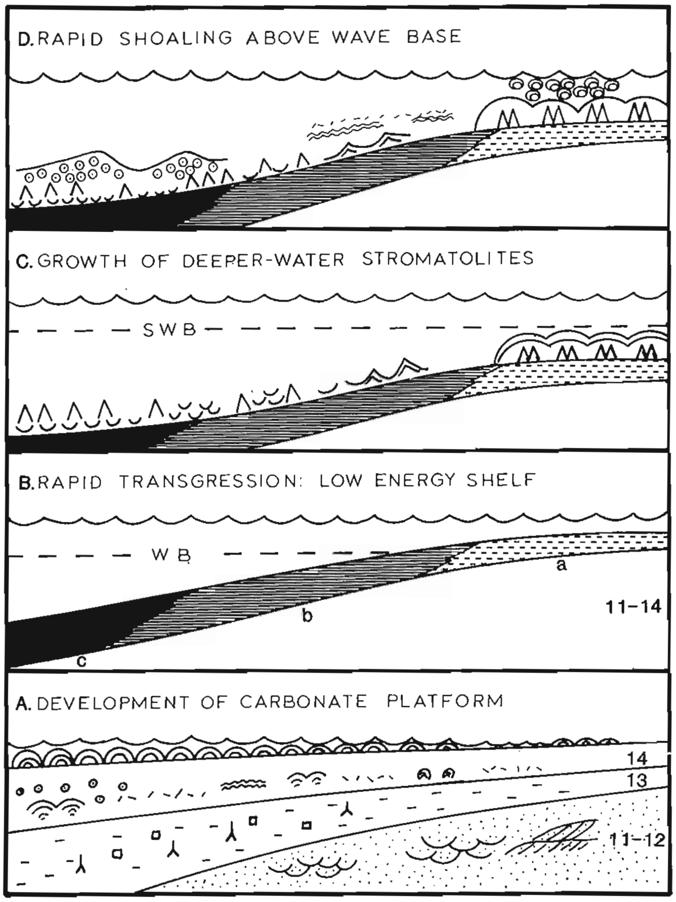


Figure 9.20. *Stages of deposition inferred for Unit 15. See text for discussion. Subdivisions a, b and c of lithofacies 15A as in Figure 9.19.*

did not act as a wave baffle, but was located on a relatively deep portion of the shelf. Lithofacies 15B appears to onlap underlying facies eastward, and records the maximum transgression of the Dismal Lakes sea. Assuming that a similar conical-columnar stromatolite unit in the Parry Bay Formation of the Elu Basin (Campbell, 1979) is time-correlative (see Fig. 9.22), a continuous north-sloping platform probably joined the Elu Basin and the Dismal Lakes Platform at this time.

Shoaling of this submerged shelf occurred rapidly on the September Lake High, where subtidal stromatolites (15B) pass directly upwards into pisolitic and evaporitic deposits (15C). While intertidal and supratidal deposits accumulated on the High, shallow subtidal oolitic shoals and intertidal algal flats predominated on the western portions of the Dismal Lakes Platform (Fig. 9.20D).

Unit 16

Unit 16, the uppermost unit of the Dismal Lakes Group, is well exposed in cliff sections capped by Coppermine River Group basalts and in north-dipping cuestas along Greenhorn and Great Bear lakes. The basal contact is conformable and is defined by a stromatolite biostrome complex in the west and a stratiform breccia in the east (Fig. 9.19). Unit 16 is 190 m thick in the Dismal Lakes area, thins to 90 m on the September Lake High and thickens westward to 370 m north-east of Great Bear Lake (Fig. 9.21).

The upper contact of Unit 16 with basalt flows of the Coppermine River Group is sharp and conformable. The basal basalt flow is characteristically crumbly (punky) and highly altered, commonly containing calcite, epidote and agate amygdules (Baragar, personal communication, 1980). Pipe vesicles also occur in the basal flow, in addition to plastically-deformed rip-up clasts of underlying dolostone. These data support the interpretation that the initial lavas were extruded onto wet sediments of probable shallow marine (peritidal) origin. Rare pillows just above the basal flow (Baragar, personal communication, 1980) suggest that local shallow flow-top lakes developed following the initial extrusion.

Lower and upper dolostone subdivisions of Unit 16 are separated by a complex of collapse breccias and mixed siliciclastic-carbonate sediments formed during uplift and erosion of the eastern Dismal Lakes Platform (Fig. 9.21).

Lower Dolostone

On the western portion of the Dismal Lakes Platform, the lower dolostone subdivision consists of a basal cyclically-repeated stromatolite biostrome complex which rests on peritidal dolostones of upper Unit 15, and is in turn overlain by 40-80 m of current-laminated red and green shales. These shales, lacking structures indicative of subaerial exposure, become increasingly more carbonate-rich upwards, and eventually are succeeded by intraclastic and oolitic dolarenite. The dolarenites attain a thickness of approximately 60 m. On the September Lake High, the equivalent of the biostrome-shale-dolostone sequence is a monotonous sequence of flat cryptalgal laminites which are in continuous conformable contact with peritidal dolostones of upper Unit 15.

Disconformity Complex

Carbonate deposition was interrupted abruptly on both the September Lake High and the eastern Dismal Lakes Platform by a period of regional uplift which created an extensive disconformity. However, coeval dolomitic and stromatolitic shales were deposited without apparent break

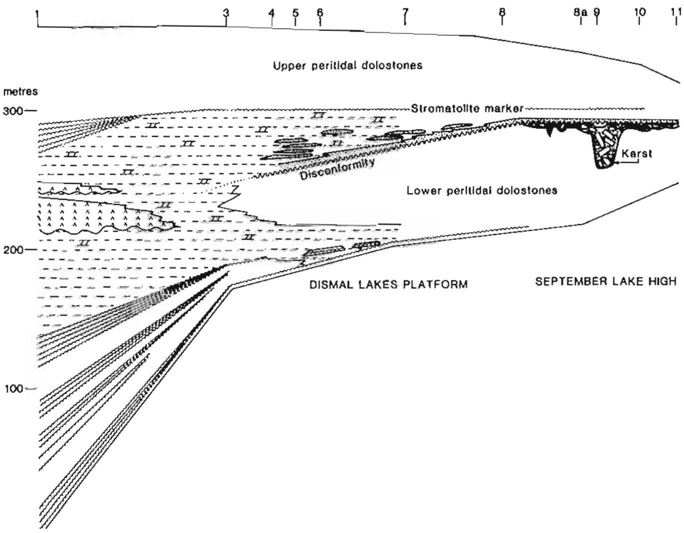


Figure 9.21. *Stratigraphic cross-section of Unit 16. Lower and upper dolostone subdivisions of Unit 16 are shown separated by a disconformity marked by paleokarst in the east and less evident surfaces of erosion westward. Section locations as in Figure 9.19.*

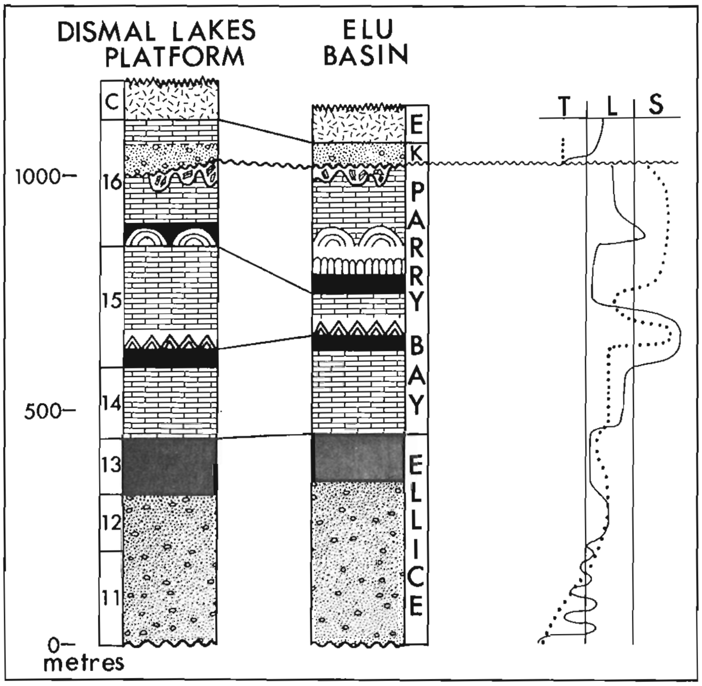


Figure 9.22. *Lithostratigraphic correlation of Dismal Lakes Group and Ellice, Parry Bay and Kanuyak formations of Bathurst Inlet. Sea-level curve on right compares depositional histories of Dismal Lakes (lined) and Elu (dotted) basins. See text for discussion. Inferred depositional environments are: terrestrial (T), littoral (L) and shelf (S).*

west of the Leith Line (Fig. 9.21). Disconformity-related sediments north of Dismal Lakes consist of a thin (5-20 cm) regolith developed on underlying dolostones, a metre-thick conglomeratic sublitharenite sand sheet, and a mixed siliciclastic-carbonate wedge which thickens to the west and grades upwards into clean upper Unit 16 dolostones (Fig. 9.21). On the September Lake High, an irregular erosion surface developed on underlying dolostones, with a local relief up to 40 m. Syndepositional normal faults resulted in local relief up to 35 m, and provided a source for localized talus breccias. Spectacular karst features, including large-scale collapse breccias, cave-floor fluvial sediments and laminar-fibrous carbonate speleothem deposits are preserved. Upper portions of the talus and collapse breccias are reworked and incorporated into the basal sand sheet, and solution joints and pockets in the weathered surface are downfilled by the sand. As in the Dismal Lakes area, mixed siliciclastic-carbonate sediments grade upwards into clean dolostones of upper Unit 16. A stromatolitic marker bed composed of Tungussia-like forms marks the upper limit of mixed siliciclastic-carbonate lithologies on the western Dismal Lakes Platform and the September Lake High, and is useful in regional correlation (Fig. 9.21).

Upper Dolostone

The upper dolostone subdivision above the stromatolite marker is lithologically uniform from the Dismal Lakes Platform onto the September Lake High, and consists of intraclastic and oolitic dolarenite with abundant club-shaped (SH-V) stromatolites. The assemblage is very similar to lower Unit 16 dolostones. The upper dolostone thins slightly from 80 m to 50 m between Great Bear Lake and September Lake. Basalts of the Coppermine River Group rest conformably upon these dolostones, but do not interfinger with them.

Interpretation of Unit 16

A short-lived local transgression on the western portion of the Dismal Lakes Platform is recorded by the subtidal to intertidal biostrome-shale-clastic carbonate sequence which overlies supratidal dolostones of upper Unit 15. On the September Lake High, subtidal equivalents of the biostrome and shale units are missing from basal Unit 16, as peritidal conditions prevailed through the Unit 15-16 transition. This period of local transgression and eventual return to intertidal conditions was followed by major regression. An abrupt disconformity was produced, leaving no record of gradual offlap in pre-disconformity sediments. This abrupt change reflects a rapid change in conditions, most likely caused by renewed basement uplift rather than a eustatic fluctuation, which is characteristically more gradual. Syndepositional block-faulting is a further indication of tectonically controlled deposition. Uplift and erosion was most intense on the September Lake High, where karst processes eroded the underlying platform carbonates to a depth of 40 m (Fig. 9.21). Less intense uplift and erosion occurred as far north as the limit of exposure of Unit 16, just south of Coronation Gulf. Similar breccias in the Bathurst Inlet area (Kanuyak Formation of Campbell, 1979) appear to be at the same stratigraphic level (Fig. 9.22), suggesting that this erosional event was regional in extent. In Bathurst Inlet, as in the Dismal Lakes area discussed above, erosion and karstification is restricted to depositional highs (Hiukitak Platform), with the unconformity shaling-out basinward (Campbell, personal communication, 1980).

A final transgressive phase led to deposition of an onlap sequence of basal sandstones and conglomerates which gradually became more carbonate-rich until the area once again was covered by a uniform peritidal platform. Eastward thinning of Unit 16 suggests a north-trending depositional strike for the Dismal Lakes Platform at the close of deposition. Extrusion of continental flood basalts onto the platform occurred when the carbonate sediments were still wet, as indicated by pipe vesicles and minor soft-sediment folds at the contact.

Summary of Dismal Lakes Group

The basal Dismal Lakes Group fining-upwards siliciclastic sequence (Units 11-13) records gradual stabilization of the source terrane, and development of a low-relief, north- and west-sloping platform (Fig. 9.15, 9.23). Peritidal

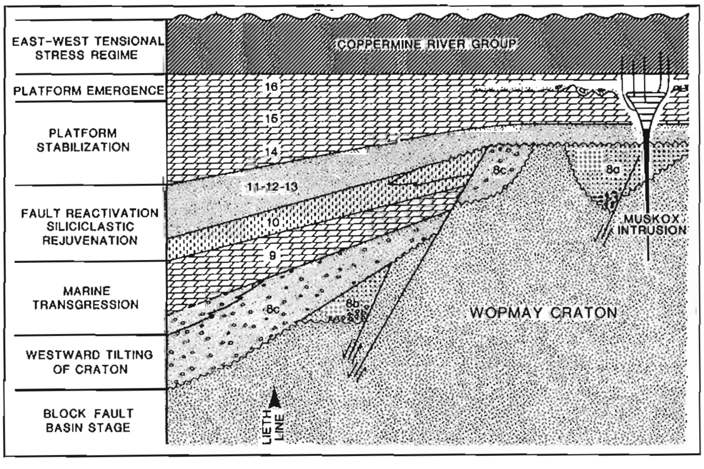


Figure 9.23. *Schematic representation of depositional history of Hornby Bay, Dismal Lakes and lower Coppermine River groups.*

dolostones of Unit 14 deposited over these coastal-plain siliciclastic sediments record eastward transgression and initiation of carbonate platform deposition (Fig. 9.18, 9.23).

A period of rapid subsidence and/or eustatic rise of sea level ensued, resulting in platform submergence and deposition of deeper-water laminated carbonates (lithofacies 15A) and stromatolite-rich carbonates (lithofacies 15B). During deposition of lithofacies 15B, the western part of the Dismal Lakes Platform subsided at a greater rate than the September Lake High, where a massive subtidal stromatolite build-up developed. Platform shoaling led to renewed accumulation of shallow-water dolostones (lithofacies 15C and lower Unit 16). As a result of subsequent uplift and subaerial exposure of the platform, deposition was interrupted, and a period of erosion and karstification ensued (Fig. 9.23). Marine incursion flooded the karsted platform surface, depositing a final 100 m thick siliciclastic-to-carbonate cycle (upper Unit 16).

Syndepositional doming of the September Lake High during deposition of units 15 and 16 may have occurred in response to Mackenzie-event basaltic magmatism (Muskox Intrusion and Coppermine River Group).

REGIONAL CORRELATION

New stratigraphic data presented in this paper justify a re-evaluation of past correlations of the Hornby Bay and Dismal Lakes groups with other Helikian sequences of northwestern Canada. In particular, excellent correlation with strata in nearby Bathurst Inlet can now be established through comparison of lithostratigraphy, regional unconformities, distinctive stromatolite units and tectono-sedimentary histories.

Table 9.1 summarizes previous attempts at correlation of the Helikian strata of the Bathurst Inlet and Coppermine areas. Initial work by Fraser and co-workers (Fraser and Tremblay, 1969; Fraser et al., 1970), based on early helicopter mapping (Fraser, 1960), allowed a gross correlation between a lower sandstone and an upper dolostone unit bracketed below by Aphebian sediments (Epworth Group and Goulburn Group) and above by continental flood basalts (Coppermine River Group). Based on revised stratigraphy of the Coppermine region by Baragar and Donaldson (1973) and the Bathurst Inlet region (Campbell, 1978, 1979), Campbell (1978) suggested that the Ellice and Parry Bay formations were correlative to the Hornby Bay Group and considered the Kanuyak Formation to represent a greatly thinned equivalent of the Dismal Lakes Group.

Further revision of Coppermine region Helikian stratigraphy presented herein adds two critical elements pertinent to the problem of correlation. First, a previously unrecognized disconformity has been identified just below the Coppermine River Group basalts, within Unit 16 of the Dismal Lakes Group (Fig. 9.21). This disconformity is essentially identical in nature and stratigraphic position to the Parry Bay-Kanuyak disconformity described by Campbell (1978). Secondly, detailed mapping along the Hornby Bay-Dismal Lakes unconformity has demonstrated its relation to local fault activity, which has rendered the unconformity unreliable for regional correlation because of lateral variability as a stratigraphic element.

It is suggested here that the Unit 16 disconformity is equivalent to the Parry Bay-Kanuyak disconformity, and that the conformable Ellice-Parry Bay clastic-to-carbonate sequence is correlative with the Dismal Lakes Group clastic (Units 11 to 13) – to carbonate (Units 14 to 16) sequence (Fig. 9.22).

Comparison of these two sequences (Fig. 9.22), compiled in collaboration with Campbell from data presented in this paper and that of Campbell (1978, 1979), illustrates the close lithostratigraphic similarity. Furthermore, northward-directed paleocurrents from both basal siliciclastic sequences (Fig. 9.24), and comparable interpreted sea level curves and siliciclastic/carbonate ratios throughout both sequences (Fig. 9.22) support a close correspondence of their respective tectono-depositional histories. Of particular importance in this respect are the distinctive conical-columnar stromatolite units which represent maximum transgression of both platform areas, as well as the karstic disconformities between the Parry Bay and Kanuyak Formations and within the upper part of the Dismal Lakes Group (Unit 16).

No single piece of evidence presented above unequivocally demonstrates that the Dismal Lakes Group is correlative with the Ellice-Parry Bay-Kanuyak succession,

Table 9.1. Summary of proposed correlations for Helikian Strata of Bathurst Inlet and Coppermine Homocline

BATHURST INLET	COPPERMINE HOMOCLINE		
Campbell(1978)	Fraser et al.(1970)	Campbell(1978)	This Paper
ALGAK FM.	LOWER COPPERMINE RIVER GROUP	ALGAK FM.	HUSKY CREEK FM.
EKALULIA FM.		COPPER CREEK FM.	COPPER CREEK FM.
KANUYAK FM.	HORNBY BAY GROUP DOLOSTONE	DISMAL LAKES GROUP	DISMAL LAKES GROUP (upper Unit 16)
PARRY BAY FM.		HORNBY BAY GROUP	DISMAL LAKES GROUP (Units 11–16)
ELLICE FM.	HORNBY BAY GROUP SANDSTONE		
TINNEY COVE FM.			HORNBY BAY GROUP

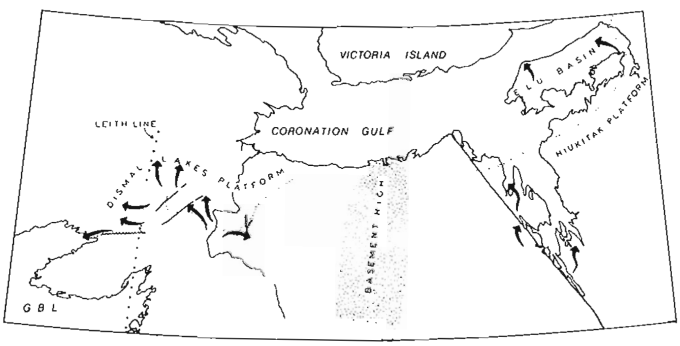


Figure 9.24. Tectonic elements of the Dismal Lakes Platform and Elu Basin. Northward tilting of both areas is indicated by paleoflow in basal siliciclastic strata of both areas. Paleocurrent data for Bathurst Inlet area are from Campbell (1979).

but taken together, the near identical position of unconformities and stromatolitic marker units as well as the similar interpreted tectono-sedimentary histories, strongly suggest that these two areas are inliers of a continuous northward-sloping platform during Helikian time.

The Tinney Cove Formation of the Bathurst Inlet area (Campbell, 1978) and the Hornby Bay Group both rest unconformably on Aphebian basement and are unconformably overlain by basal siliciclastics of the Ellice Formation and basal Dismal Lakes Group respectively. On this basis, these two units may be considered grossly time-equivalent. Importantly, however, the Hornby Bay and Tinney Cove sandstones differ in (terms of) their tectonic relationships to Wopmay Orogen.

The Hornby Bay Group clearly postdates all fault activity related to the terminal collision phase of Wopmay Orogen, whereas the Tinney Cove accumulated during this late Wopmay event along the syndepositionally active Bathurst Fault (Campbell, 1978). This relationship implies that the Tinney Cove is more closely related to the Et Then fanglomerates of East Arm, Great Slave Lake. In addition, minimum pre-Hornby Bay Group erosion is estimated to be at least 1 km (Hoffman and St-Onge, 1981). This relationship and the relatively late post-orogenic initiation of Hornby Bay deposition suggest that it is significantly younger than either the Tinney Cove or Et Then formations.

Predominantly shallow-water siliciclastics and dolostones of the Wernecke Supergroup of the northern Cordillera (Delaney, 1978, 1981) also are approximately time equivalent to Hornby Bay and Dismal Lakes groups ("Sequence A" of Young et al., 1979). Tectono-sedimentary relationships between this 14 km thick sequence and the thinner, cratonward, Hornby Bay-Dismal Lakes sequence are uncertain. In particular, westward deepening facies in both the Hornby Bay Group and Dismal Lakes Group are difficult to reconcile with shallow-water Wernecke Supergroup sediments if these two areas were part of a continuous westward-sloping shelf.

HISTORICAL SUMMARY

In view of the lack of resemblance of Unit 8 to sequences deposited in active strike-slip regimes (e.g. Cordilleran molasse basins, Eisbacher, 1974; Hornelen Basin, Steel et al., 1977; Ridge Basin, Crowell, 1974) we conclude that Hornby Bay Group deposition began well after cessation of compression-related faulting and orogenic activity in the basement (Wopmay Orogen). Deposition began as the Teshierpi-Bigtree Highlands (Fig. 9.2b) shed clastics to the southeast, resulting in peneplain formation during deposition of the Bigbear system. Prolonged crustal stability is suggested by the high maturity and low-energy depositional environments evident in large parts of this system. The appearance of fault-related intraformational paraconglomerates in the Bigbear system signalled the beginning of west-southwest cratonic tilting and uplift. Normal faulting in the Fault River area at this time led to the formation of a trough-like depocentre (Fault River system), and erosion of western portions of the Bigbear system. Incision of the Teshierpi-Bigtree Peneplain by westerly-flowing rivers occurred during this period of uplift.

Subsequent basinwide subsidence was accompanied by deposition of the Lady Nye system, which was derived from a source area in the Coronation Gulf area, and burial of the Bigbear system, Fault River system, and incised Teshierpi-Bigtree Peneplain. Marine transgression and formation of a carbonate platform (Unit 9) resulted from continued regional subsidence. Carbonate deposition on the short-lived Hornby Bay Platform was terminated by progradation of deltaic-marine siliciclastics of Unit 10, which reflects uplift in the northeast and the beginning of a second siliciclastic-carbonate cycle.

In response to regional uplift, reactivation of the Teshierpi Fault commenced during late Unit 9 time. Erosion of lower Hornby Bay Group (Units 8 and 9) and Aphebian basement occurred on the upthrown southeastern block. Southwest paleocurrents in Unit 10 and lower Unit 11, in addition to facies changes in Unit 10, suggest that deposition was controlled by uplift in the northeast and/or hinged dip-slip movement on the Teshierpi Fault. Regional subsidence during deposition of Unit 11 was accompanied by marine transgression and formation of the Dismal Lakes Platform coincident with a shift in the source area from the northeast to southeast. In contrast to the localized tectonic activity and changes in cratonic paleoslope during deposition of the Hornby Bay Group, stable conditions prevailed during deposition of the Dismal Lakes Group sediments on a north- and west-sloping platform. This platform likely extended eastward from the Dismal Lakes area to the Bathurst Inlet area (Elu Basin and Hiukitak Platform of Campbell, 1979), a distance of more than 400 km.

The lower Dismal Lakes Group siliciclastic to carbonate transition (upper Unit 11-14) records a period of stable epeiric sea conditions (Fig. 9.23). The succeeding deeper-water carbonates of Unit 15 represent the period of maximum transgression of the entire Dismal Lakes-Elu Basin shelf. Differential subsidence during transgression isolated the September Lake High as a positive feature. Steady progradation of the carbonate platform proceeded (upper Unit 15, lower Unit 16) until interrupted by a rapid sea level fall which resulted in subaerial weathering and erosion of craton-marginal areas of the platform (middle Unit 16). Syndepositional doming and block-faulting, again localized in the September Lake High area, accentuated erosion of the platform as shown by the karstic paleotopography with up to 40 m relief.

A causal relationship between doming during deposition of the upper Dismal Lakes Group and succeeding Mackenzie-age magmatism (see below) is strongly suggested by their close spatial and temporal relationship. The increased cyclicity and marked thickening of Unit 16 west of the Leith Line, relative to underlying units, indicate greater activity of this tectonic feature as a basin hingeline during deposition of the upper Dismal Lakes Group.

Deposition of the Dismal Lakes Group was terminated by an east-west extensional event (Fig. 9.23), as suggested by Hoffman (1980b). This event was accompanied by extrusion of Coppermine River Group basalts, and intrusion of the Muskox Complex and Mackenzie diabase dykes.

The September Lakes High area is a major locus of this Mackenzie-age magmatic activity, as shown by the coincident occurrence of the Muskox Complex, a high concentration of Mackenzie-age dykes and the depocentre of the Copper Creek – Ekalulia formations continental flood basalt plateau.

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10. HORIZONTAL MOTIONS AND ROTATIONS IN THE CANADIAN SHIELD DURING THE EARLY PROTEROZOIC

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McGlynn, J.C. and Irving, E., Horizontal motions and rotations in the Canadian Shield during the Early Proterozoic; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 183-190, 1981.

Abstract

Paleomagnetic evidence from the Coronation Geosyncline suggests that large relative rotations of small blocks about local vertical axes have commonly occurred in the middle Proterozoic, as they have in Phanerozoic orogenic belts. Paleomagnetic results from early Proterozoic intrusive rocks from the Canadian Shield as a whole, including new data from the 2.2 Ga old Easter Island Dyke, are used to construct separate APW paths for blocks that include the Archean Slave and Superior structural provinces. If we have drawn and calibrated these paths correctly then large scale relative motions in a horizontal sense have occurred between the Slave and Superior provinces during the middle Proterozoic. There is some paleomagnetic evidence to indicte that horizontal motions in the middle Proterozoic were more rapid than motions in the later Phanerozoic.

Résumé

Les indices paléomagnétiques provenant du géosynclinal du Couronnement suggèrent que vers le milieu du Protérozoïque, ont fréquemment eu lieu d'importantes rotations de petits blocs par rapport à des axes verticaux locaux, comme dans les zones orogéniques datant du Phanérozoïque. Les données magnétiques provenant des roches intrusives du début du Protérozoïque, obtenues sur l'ensemble du Bouclier canadien et incluant de nouvelles données relatives au dyke d'Easter Island, dont l'âge est de 2,2 Ga, ont servi à reconstruire les trajets séparés APW des blocs englobant les provinces structurales archéennes des Esclaves et du lac Supérieur. En tracant et étalonnant correctement ces trajets, on voit que des mouvements relatifs de grande envergure ont eu lieu dans le sens horizontal entre les provinces des Esclaves et du lac Supérieur pendant le Protérozoïque moyen. Certains indices paléomagnétiques indiquent qu'à cette époque-là, les mouvements horizontaux ont été plus rapides qu'à la fin du Phanérozoïque.

INTRODUCTION

Directions of magnetizations in rocks (paleodirections) are specified relative to the paleohorizontal – the bedding plane in sediments, for instance. The total vector is resolved into horizontal and vertical components, the former being in the paleohorizontal plane and the latter perpendicular to it. The angle which the horizontal component makes with the present geographic or true north is called the declination. The declination is the sum of all rotations about local vertical axes that the sampling location has undergone relative to present north since the paleodirection was acquired. Hence, comparisons of declinations from rocks of comparable age from different localities yield information about relative rotations (RR) among localities. Paleomagnetically determined RR's are common in Phanerozoic orogenic belts, and have been reported, although less commonly, from Precambrian terranes, for example by Baragar and Robertson (1973). We shall describe further examples of RR's from the northwest Canadian Shield.

Usually a paleodirection has a substantial vertical component so the total vector makes an angle (the inclination) with the paleohorizontal. The inclination is a measure of the paleolatitude of the locality when the paleodirection was acquired. Comparisons of inclinations from different localities yield estimates of relative paleolatitudinal displacements (RPD) among localities. In principle, this provides a means of measuring the RPD's of the Slave and Superior structural provinces. We shall now describe our current view of this important problem.

Because paleodirections are observed from different places account must be taken of the spatial variations of the geomagnetic field and a model for the field must be assumed. Except in special circumstances where the localities are very close to one another comparisons of inclination and declination are not made directly but by reference to the corresponding paleopole. Sequences of paleopoles may be connected to form a path of apparent polar wander (APW) relative to the region of observation. It is assumed that the field on average was that of a geocentric dipole. For the geologically recent past this model is accurate to 3-5° (Hospers, 1954; McElhinny and Merrill, 1975). Paleoclimatic evidence indicates that this model is also approximately correct for the Phanerozoic (Irving, 1956; Drewry et al., 1974).

POSSIBLE INSTANCES OF RELATIVE ROTATIONS IN THE NORTHWEST SHIELD

The APW for the interval 1900 to 1650 Ma compiled from many different parts of the Canadian Shield is shown in Figure 10.1 (documentation in Irving and McGlynn, 1980). Notable is the fact that paleopoles derived from overprints (secondary magnetizations) outnumber those from primary magnetizations. Paleopoles from the Great Slave Supergroup and their equivalents in the Takijuq Basin (Hoffman, in press) map out a crude loop (SA, TA, DP, PPS, PA, ET) which can be extended southeast to include results from Flin-Flon (FFB) and the Richmond Gulf and Belcher Islands (RG, RGM). Paleopoles FFB, RGM and RG are included as part of the

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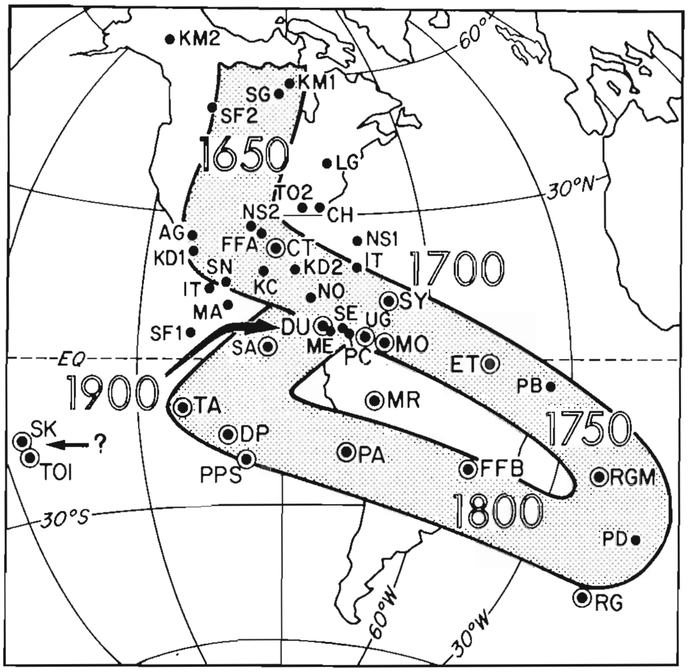

Figure 10.1. Middle Proterozoic, the Coronation Loop. Modified from McGlynn and Irving (1978) and Irving and McGlynn (1980). The rock unit from which each paleopole was obtained is identified below, and the documentation is given in Irving and McGlynn (1980). Paleopoles with double circles are considered to be primary, the others being overprints.

AG	*Amitsoq gneiss, Greenland*
CH	*Charlton Bay Formation, N.W.T.*
CT	*Castignon Complex, Labrador*
DP	*Douglas Peninsula Fm., N.W.T.*
DU	*Dubawnt Group, N.W.T.*
ET	*Et-then Group, N.W.T.*
FFA	*Flin-Flon greenstones (younger paleopole), Manitoba*
FFB	*Flin-Flon Greenstones (older), Manitoba*
IT	*Itivdleq Dykes and gneisses, Greenland*
KC	*Kahochella overprint, N.W.T.*
KD1	*Kangamuit Dykes, Greenland*
KD2	*Kangamuit Dykes, Greenland*
KM1	*Ketilidian metavolcanics, Greenland*
KM2	*Ketilidian metavolcanics, Greenland*
LG	*Gibraltar Formation (lower), N.W.T.*
MA	*Melville-Daly Bay metamorphics, N.W.T.*
ME	*Menihek Formation, Labrador*
MO	*McLeod Bay Formation, N.W.T.*
MR	*Martin Formation, N.W.T.*
NO	*Nonacho Group, N.W.T.*
NS1	*Nagsug gneiss, Kaellingehaetten, Greenland*
NS2	*Nagsug gneiss, Kaellingehaetten, Greenland*
PA	*Pearson Formation A, N.W.T.*
PB	*Pearson Formation B, N.W.T.*
PC	*Pearson Formation C, N.W.T.*
PD	*Pearson Formation D, N.W.T.*
PPS	*Peninsula Sill, N.W.T.*
RG	*Richmond Gulf andesites and redbeds, Quebec*
RGM	*Richmond Gulf-Manotouk Is. basalts, Quebec*
SA	*Seton Formation A magnetization, N.W.T.*
SE	*Slave Province post-Hudsonian overprint, N.W.T.*
SF1	*Seward Formation (B), Labrador*
SF2	*Seaward Formation (C), Labrador*
SG	*Sagdlerssuag Dykes and gneisses, Greenland*
SK	*Stark Formation, N.W.T.*
SN	*Sokoman Iron Formation, Labrador*
SY	*Sparrow Dykes, N.W.T.*
TA	*Takiyuak Formation, N.W.T.*
TO1	*Tochatwi Formation, N.W.T.*
TO2	*Tochatwi Formation overprint, N.W.T.*
UG	*Gibraltar Formation, upper, N.W.T.*

Coronation Loop, although it is possible that they represent different positions for that part of the Superior Structural Province relative to the northwest Canadian Shield. Overprints from the Pearson Formation (PB, PD) provide some support for this construction but they, like the Richmond Gulf rocks, are poorly dated. For the moment we use the deep loop of Figure 10.1 as given in Irving and McGlynn (1979), but we stress the uncertainty about its southeasterly extension; the paleopoles that have a degree of geochronological control (and surely these ages must come under critical review shortly) from the southeast part of the Coronation loop including that from the Eskimo Volcanics (1700 Ma near paleopole PD of Fig. 10.1) recently reported by Schmidt (1980) are from localities remote from the Coronation Geosyncline. Given the age uncertainties, these paleopoles therefore could reflect relative motion between these units and the Coronation Geosyncline. Results presented by Evans and Hoye (1981) at this symposium add further important evidence from the Kilohigok Basin. Certain paleopoles from the upper part of the Great Slave Supergroup (SK, TOI) fall to the west of this path. The question of why this should be so is now considered.

The horizontal component of the paleodirections for rock units in the Christie Bay and Pethei groups and their equivalents in the Coronation Geosyncline are shown in Figures 10.2 and 10.4. The corresponding paleopoles are plotted in Figure 10.3. Four paleodirections, two from the Athapuscow Aulacogen and two from the Takijuq Basin, are south-southeast. In the Athapuscow Aulacogen the trend with time is from DP to PA defined by superposition, and in the Takijuq Basin it is from TA to PPS defined by intrusive relationships. Two units from the Athapuscow Aulacogen (SK, TOI), that stratigraphically immediately underlie PA and which are the probable equivalents of TA in the Takijuq Basin, have southwest paleodirections. Tables showing the stratigraphic relationship of these rock units are given by Hoffman (1981) and Campbell and Cecile (1981).

The Takiyuak Formation from the Takijuq Basin lies stratigraphically between the Douglas Peninsula Formation and the Pearson Formation of the Athapuscow Aulacogen, but its paleopole (TA) falls to the west of both their paleopoles. The Peninsular Sill, which could be co-magmatic with the Pearson Formation, also has a paleopole to the west. It is therefore possible that the localities in the Takijuq Basin have been rotated 10-20° clockwise relative to the localities from which the paleopoles were obtained in the Athapuscow Aulacogen. Rotations could have been caused by the emplacement of nappes that occur within the Great Slave Basin (Hoffman et al., 1977), or by movement along strike-slip faults that cut the Great Slave Supergroup. It seems more likely that the source of the rotation is in the Athapuscow Aulacogen rather than in the Takijuq Basin, because the Takiyuak Formation and the Peninsular Sill are from that part of the Epworth Group that is autochthonous and rests unconformably on Archean basement.

The Stark (SK) and Tochatwi (TOI) paleopoles lie far to the west. It has been suggested that this might have been caused by a local 60° clockwise rotation about a vertical axis of the sampling area, which is about a kilometre in extent (Bingham and Evans, 1976; Irving and McGlynn, 1979).

Gough et al. (1977) have argued against such a rotation from studies of the anisotropy of susceptibility of the rocks in question. Irving and McGlynn (1979) have criticized their argument, firstly on the grounds that uniformity of directions of maximum susceptibility is observed at localities in which the remnant magnetizations are demonstrably secondary, and secondly, the evidence from remanent magnetization for rotation occurs in different sequences from those in which the uniformity in directions of anisotropy of susceptibility is observed.

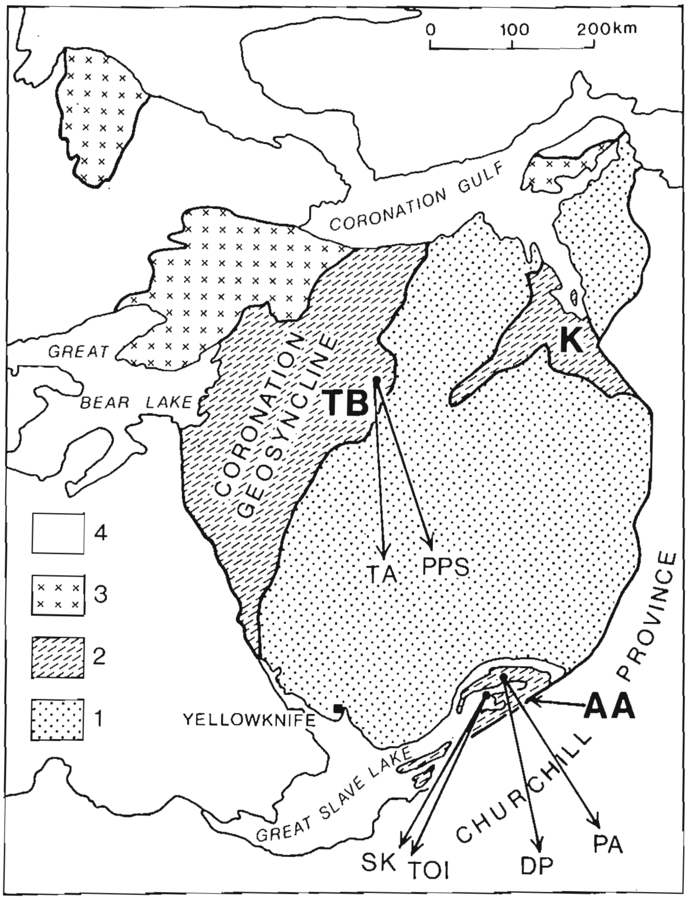

Figure 10.2. Regional geological map of Coronation Geosyncline and its associated tectonic elements.

1. *Archean of Slave Structural Province*
2. *Early Proterozoic rocks of the Coronation Geosyncline, the Athapuscow Aulacogen (AA), and the Kilohigok Basin (K); TB is the Takijuk Basin within the Coronation Geosyncline (Hoffman, 1980)*
3. *Middle and late Proterozoic cover rocks*
4. *Phanerozoic rocks to the west. Precambrian Churchill Structural Province to the east is labelled*

Arrows give the directions of the horizontal component of the mean direction of magnetization observed from the Douglas Peninsula Formation (DP, Irving and McGlynn, 1979), the Pearson Formation (PA, McGlynn and Irving, 1978), the Peninsular Sill (PPS, Irving and McGlynn, 1979), the Stark Formation (SK, Bingham and Evans, 1976), the Takiyuak Formation (TA, Irving and McGlynn, 1979), and the Tochatwi Formation (TOI, Evans and Binghamn, 1976).

Such rotations are commonly found close to Mesozoic and Cenozoic plate-margins, for example the transported thrust sheets of western Sicily that were observed (paleomagnetically) to be rotated dextrally by large and variable amounts relative to the stable (Iblean) platform of southeast Sicily (Channell et al., 1980). The presence of rotated blocks in the mid-Proterozoic orogenic belt of the northwest Shield is consistent with the occurrence of plate tectonics at that time.

It could, of course, be argued that the differences among paleopoles SK, TOI, TA, DP, PPS and PA (Fig. 10.4) arises from short-term departures of the field from an average dipole configuration. We regard this as unlikely because it is only declinations that are affected whereas inclinations are in excellent accord. It would be a coincidence beyond what may be reasonably assumed if an aberration that had its source in the earth's core should systematically affect one component of the field but not the other.

PALEOLATITUDE VARIATIONS 1900 TO 1650 Ma

Paleolatitudes (plotted in Fig. 10.5) depend only on the inclination and do not reflect local rotations recorded by differences in declination. It is notable than results from earlier Proterozoic localities ranging from Greenland, the Labrador Trough, Richmond Gulf to the central and western Shield are consistent with a single paleolatitude cycle from high to low and then back to high paleolatitudes (documentation in Irving and McGlynn, 1980). This indicates that local rotations constitute a sort of "tectonic noise" that otherwise tends to disguise the underlying simplicity of the motions of Laurentia relative to the pole. Laurentia, although it appears to have been generally coherent, may not as we have seen been perfectly rigid.

Interaction with neighbouring plates may have caused deformation extending far into the block, perhaps in a manner comparable to the deformation of central Asia north of its boundary with the Indian plate (Molnar and Tapponier, 1975; Hoffman, 1980). It is important to note that the "general coherence" of Laurentia referred to above applied only to the interval of time after the Hudsonian Orogeny, that is after about 1850 Ma.

There are two items of further interest in Figure 10.5. The turning-point at about 1800 or 1750 Ma signifies a marked change in paleolatitude corresponding to the hairpin of the Coronation Loop. This is a period of time when fault-controlled sedimentary basins, such as the "Et-Then" and

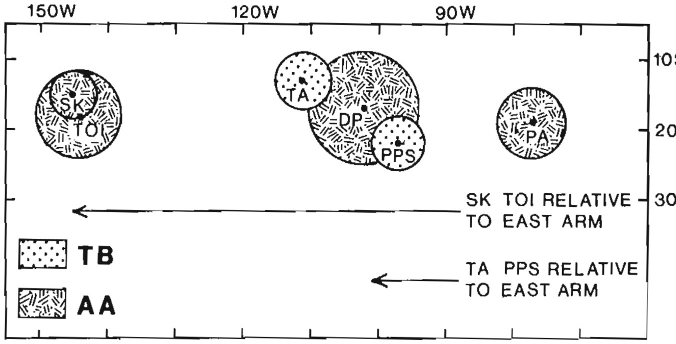

Figure 10.3. Paleopoles for the six rock units compared. Labelled as in caption to Figure 10.2. The standard error circles (P = 0.63) are given. The arrows show mean relative rotation of SK and TOI, and of TA and PPS relative to DP and PA of the East Arm of the Great Slave Lake, the latter assumed fixed. TB stipple are paleopoles from Takijuq Basin AA stipple from the Athapuscow Aulacogen.

"Tinney Cove" basins, were formed in terrain that was deformed, and in places metamorphosed, during the Hudsonian Orogeny. This hairpin represents a large change in direction of drift of Laurentia, and presumably reflects a period of major plate reorganization following the Hudsonian Orogeny. Secondly, the average rate of change of paleolatitude for the 300 Ma interval is 5 to 6 cm per year. If the paleolatitudinal changes are entirely caused by continental drift (and we do not, of course, know the contribution from true polar wander) then this is the minimum drift rate because movements in paleolongitudinal sense are not recorded paleomagnetically. These minimum drift rates persisting for long time intervals are much higher than the average minimum rates of motion (about 2 cm per year) recorded by APW during the past 300 Ma (Gordon et al., 1979).

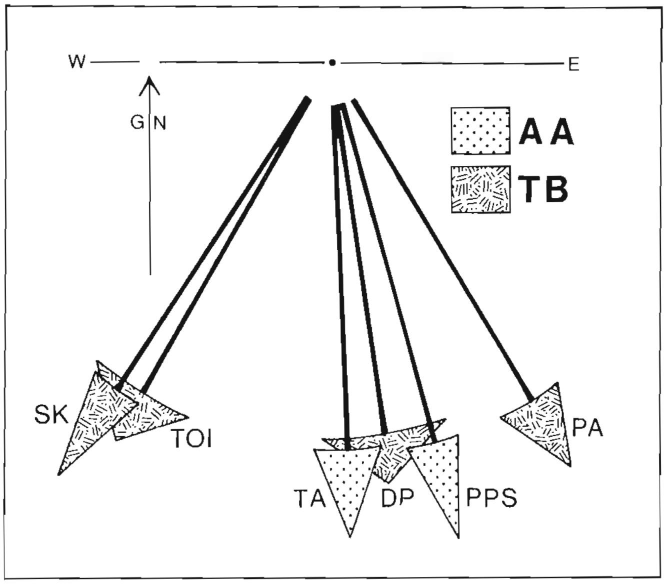

Figure 10.4. *Directions of horizontal component of mean directions of magnetization for six rock units compared. Labelled as in caption to Figure 10.2. GN is present geographic north.*

We conclude that the paleomagnetic results provide evidence of tectonism occurring on two scales. Firstly, the general change in paleolatitude, calculated from the variation in inclination, records continental drift over distances in excess of 10^4 km that appear to have occurred at about twice the typical later Phanerozoic rate. Secondly, the detailed structure of the declination variations record the relative rotation of local blocks.

SLAVE-SUPERIOR COMPARISON

If the Churchill Structural Province is a product of plate tectonic processes then the Archean Slave and Superior structural provinces should have moved relative to one another, and their polar tracks for the interval prior to the Hudsonian Orogeny should be separate. Although numerous results are now available from the Slave and Superior provinces they have not provided an unique answer to this problem; different authors have devised different interpretations. A review of earlier discussions has been given by Irving and McGlynn (1980). Here we explain our current opinion incorporating recently obtained data and ideas of other authors.

Figure 10.6 shows the paleopoles from sampling localities in Slave and Superior provinces noted in the inset. The paleopoles from the Otto Stock and Abitibi dykes of the Superior Province differ by 35° even though their ages do not differ significantly (Rb-Sr isochron of 2114 ± 80 Ma (Bell and Blenkinsop, 1976) for the stock, and Rb-Sr isochron of 2100 ± 68 Ma (Gates and Hurley, 1973) and $^{40}Ar/^{39}Ar$ of 2150 ± 25 Ma (Hanes and York, 1979) for the dykes). Both paleopoles are supported by contact tests and both are from the same general area in which there is no geological evidence of local rotations. The stock and the dykes have vertical contacts indicating that subsequent tilting has been small or negligible. The contact tests for both stock and dykes are good particularly so for the latter because the baked material (in part for the stock, entirely for the dykes) is Matachewan diabase whose magnetization away from the contact is itself known to be primary from contact tests (Irving and Naldrett, 1977). Hence there is good evidence for believing that the paleopole was close to the Superior Province segment of North America about 2150 Ma ago.

Numerous studies of the apparently contemporaneous Nipissing diabase (Rb-Sr isochrons of 2100 ± 50 Ma Van Schmus, 1965) and have yielded a wide distribution of

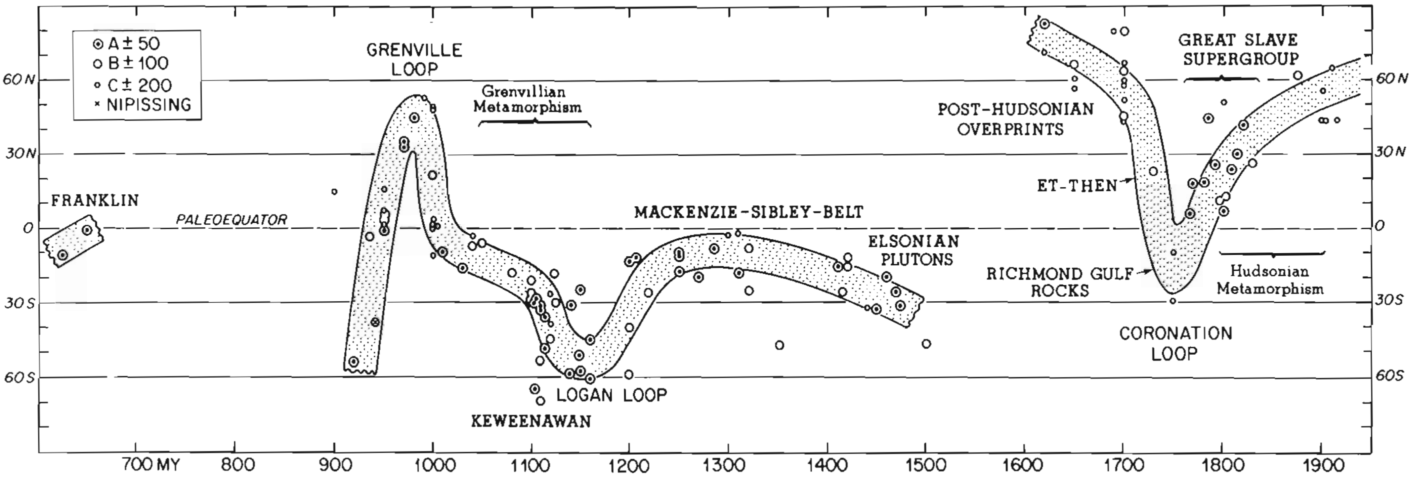

Figure 10.5. *Paleolatitude variations for the Laurentian Shield. Calculated for Winnipeg (50° N, 97° W). From Irving and McGlynn (1980).*

poles (ND1 to N11, Fig. 10.6). The metamorphic grade of the Nipissing diabase varies from subgreenschist to amphibolite. It would appear that generally low-grade rocks have been studied, although this is not always clear.

The Nipissing paleopoles have been considered to fall into a northerly (N) and a southerly (S) group (Fig. 10.6). Considerable uncertainty exists about the relative and absolute ages of the corresponding magnetizations. Symons (1967) believed that the southerly paleopoles were derived from the original magnetization and described a contact test in support. Later, Roy and Lapointe (1976) regarded S as an overprint and N as the original magnetization, basing their argument mainly on a contact test for paleopole ND7. Although this contact test is not conclusive, we believe it to be very good because the Firstbrook Formation, which is reset in the direction of the Nipissing diabase near contacts, has, as Roy and Lapointe (1976) showed from detailed studies, a very stable magnetization which is almost certainly of pre-Nipissing age. Subsequently Morris (1979) has described a contact test applicable to the S palaeopoles, which could support Symons' original argument. A further uncertainty is that the geological relationships between the rocks studied magnetically and those studied radiometrically is not always clear. Thus, the Nipissing diabase palaeopoles of Figure 10.6 could have been derived from rocks of several different intrusive events. For example, ND11 is indistinguishable from paleopoles for the Franklin intrusions (approximately 650 Ma) and from paleopoles from west-northwest trending dykes of the Grenville swarm that are common in this area (Murthy, 1972; Park, 1974). Hence, ND11, ostensibly derived from Nipissing diabase, could have been derived from an intrusion very much younger than the Nipissing, but difficult to distinguish from it in the field because of their petrological similarity. Such situations however are undoubtedly rare – the great majority of observations ascribed to the Nipissing have been obtained from true Nipissing diabase.

The interpretation of results from the Nipissing diabase is further complicated by uncertainty in application of tilt corrections. The diabases are concordant or transgressive sills intruded into Huronian sedimentary rocks. According to Card and Pattison (1973) the Nipissing intrusions "were emplaced after early faulting during the later stage of an early major folding of the Huronian sequence, but before secondary deformational events and regional metamorphism". In earlier studies Symons (1970) showed, in some instances, that the directions of magnetization were somewhat better grouped before rather than after corrections were applied for tilting observed in country rocks. In later studies paleopoles have been calculated on the assumption that the various magnetizations were acquired when the rocks were in their present relative positions. This assumption would be justified if all Nipissing paleopoles were in close proximity, but they are not. The effect of omitting possible tilt correction is not clear, but because of the general east-west strike of the Huronian rocks it could account for some of the meridional spread of Nipissing paleopoles.

Several paleopoles from sedimentary rocks of the Huronian Supergroup have been obtained (GG1 to GG5 and TS). Of these GG1, 3, and 5 are derived from what are probably original magnetization, whereas GG2 and 4 are postfolding overprints. The TS magnetization may or may not be original. The upper broad arrow in Figure 10.6 indicates the polar trend suggested by Roy and Lapointe (1976). Irving and McGlynn (1980) have argued that all evidence, some of it obtained after the work of Roy and Lapointe (1976), is compatible with their suggested polar track if the possibility of local RR's is admitted. The track of Roy and Lapointe may then be connected with the

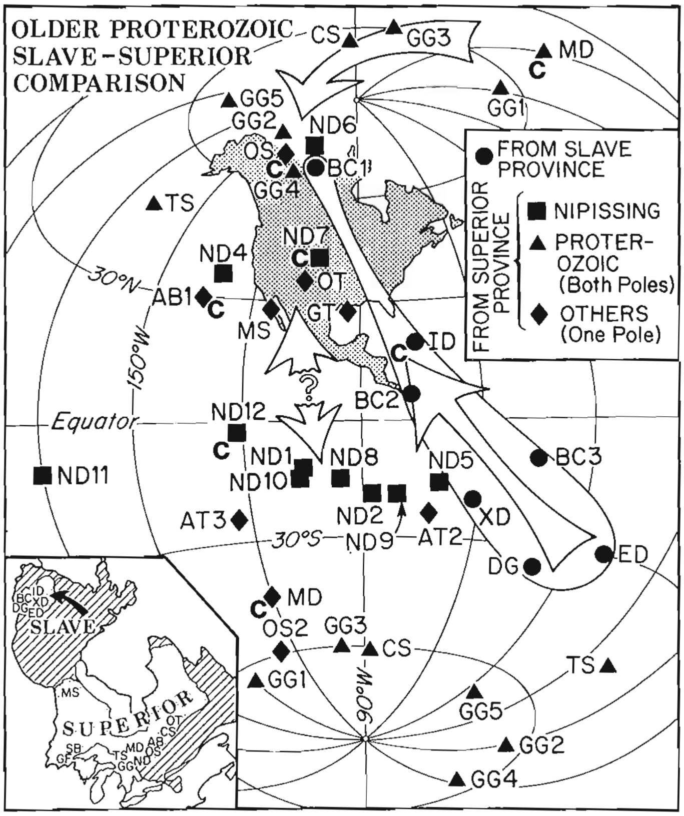

Figure 10.6. *Older Proterozoic, Slave-Superior comparison. The broad arrow passing over the present pole indicates the probable trend of the pole obtained mainly from studies of Huronian sedimentary rocks (Roy and Lapoint, 1976). The SE-NW arrow gives the probable time trend of poles from the Slave Structural Province (McGlynn and Irving, 1975). The double arrow with question mark indicates the uncertainty in the trend of paleopoles from the Nipissing diabase. The main sampling localities are shown in the inset. The paleopoles are labelled as follows, and the documentation is given in Irving and McGlynn (1980).*

- *AB1* Abitibi Dykes, primary
- *AT2* Abitibi Dykes, north directions
- *AT3* Abitibi Dykes, southwest directions
- *BC1 to 3* three magnetizations of Big Spruce Complex
- *CS* Chibougamau sills
- *DG* Dogrib Dykes
- *ED* East Island Dike (McGlynn and Irving, in preparation)
- *GG1 to 5* Gowganda Formation
- *GT* Gunflint Formation
- *ID* Indin Dykes
- *MD* Matachewan Dykes
- *MS* Molson Dykes
- *ND1-12* Nipissing diabese
- *ND12* from Morris, 1979
- *OS1* Otto Stock
- *OS2* Otto Stock overprint
- *OT* Otish gabbro
- *TS* Thessalon Volcanic Series
- *XD* "X" Dykes

Paleopoles with "C" besides them are supported by contact tests. Antipoles of CS, GG, MD and TS are plotted to allow study of alternative northerly directed polar path to be made.

southerly trending "Nipissing polar track" and Track 5 of Irving and McGlynn (1976). Alternatively, as in Figure 10.6, Huronian paleopoles may be plotted in the southern hemisphere and the polar track brought from the south to connect with a northerly trending Nipissing polar track. This northerly trending option of Huronian and Nipissing paleopoles does not very adequately explain the age relationships and we favour the southerly trending option for reasons now given.

We regard the contact tests of OS, AB1 and ND7 as satisfactory, so that these paleopoles truly reflect the direction of the field at the time the rocks were intruded about 2100 Ma ago. The inclinations, like those of the roughly contemporaneous units MS and OT, are steep, indicating high paleolatitudes and a paleopole near to the Superior Province. The contact tests (ND12) for the southerly group of Nipissing poles (ND1, 2, 5, 8, 10, and 12) are less convincing. That given by Symons (1967) is based on very scattered paleodirections. To make the second test, Morris (1979) studied five sites, two in diabase and three in country rock. Two sites in baked sedimentary rock and diabase of the chilled contact have the same paleodirections (yielding paleopole ND12). The sampling site at the centre of the dyke has complex magnetizations with shallow inclinations, but "a few specimens. . . show a consistent directional shift to steeper inclinations", that is towards the direction corresponding to ND12 (Fig. 10.6). The site in the unbaked Huronian argillite has another direction of magnetization corresponding to paleopole GG2, and this magnetization is postfolding (Morris, 1977). The data allow several interpretations. It is possible, as Morris contends, that the baked sediment and chilled diabase contact rock have indeed preserved the original magnetization acquired at the time of intrusion and cooling. It seems to us probable that the magnetization (yielding ND12 paleopole) postdates the magnetization of the country rock, but this is by no means certain. Since, the magnetization of the country rock is itself secondary (Morris, 1979) and of uncertain age, the age of the ND12 magnetization is not well constrained.

The evidence is in an unsatisfactory state, but we would argue that it is permissible to retain the hypothesis of Roy and Lapointe (1976) that the southerly Nipissing paleopoles reflect possible secondary overprints. The age of the overprinting is unknown. The wide longitudinal spread of the southerly Nipissing paleopoles could have been caused by late stage RR's of small blocks related to the longitudinal east-west faults that are common in the area.

The paleopoles from the Slave Structural Province are crudely aligned along a southeast to northwest trend (Fig. 10.6). DG, from the Dogrib Dykes, is probably the oldest because it is derived from the oldest body studied and although contact tests cannot be made (the country rock is magnetically unstable) the uniformity and stability of their distinctive magnetization directions suggest that the magnetization could be original. Estimates of the age of the Dogrib Dykes by K-Ar and Rb-Sr methods fall in the range 2200 to 2600 Ma. The Easter Island Dyke (paleopole ED) is an alkaline gabbro body which has yielded K-Ar ages of about 2200 Ma. It intrudes Archean granitic basement rocks. It is overlain by the Great Slave Supergroup and is thought to be older than the Union Island Group (Hoffman et al., 1977). The Easter Island dyke paleopole falls near that for the Dogrib Dykes (McGlynn and Irving, in preparation). XD is a paleopole form an undated unnamed small group of dykes that are probably of middle Proterozoic age. Paleopole ID from the Indin Dykes (2049 ± 86 Ma, Rb-Sr isochron) is supported by a contact test. The Big Spruce Complex (2066 ± 40 Ma, Rb-Sr isochron) has yielded paleopole BC1, BC2 and BC3

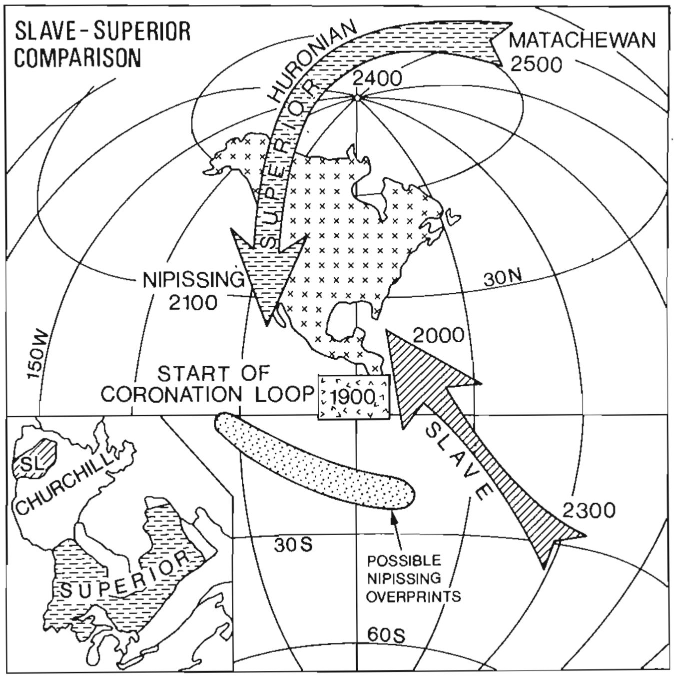

Figure 10.7. Possible polar paths for the Slave and Superior structural provinces. The approximate ages given in millions of years. SL in the inset is the Slave Structural Province.

attributed to three distinct magnetizations. A fourth magnetization. BC4, is probably very much younger. At first we considered BC1 to be the primary magnetization and BC2 and BC3 to be overprints related to the younger Coronation Loop (Irving and McGlynn, 1976). BC1 is statistically indistinguishable from OS1 (the Otto Stock) and we originally considered this agreement to be good evidence against the idea of large relative motions between the Slave and Superior structural provinces since about 2000 Ma. We are now less certain that this is a sound argument (Irving and McGlynn, 1980). The magnetization corresponding to BC1 is directed along the present field and could have been caused by incipient weathering or by some as yet unstudied process related to the effects of ice-loading or permafrost.

Our preferred interpretation of the polar tracks for the Slave and Superior structural provinces are shown in Figure 10.7. the Superior path extends from the Matachewan to Nipissing paleopoles and is similar to the path given by Roy and Lapointe (1976). The Slave path extends from the paleopole for Dogrib and Easter Island dykes to that for the Indin Dykes paleopoles and is like that originally given by McGlynn and Irving (1975). The former begins earlier and apparently ends a little earlier. The Coronation Loop (Fig. 10.1) begins at about 1900 Ma. There are no paleopoles that can be assigned with confidence to the interval from about 2000 to 1900 Ma so that the manner in which the Slave and Superior polar tracks come together and became linked with the Coronation Loop is unknown. The possibilities are too numerous at present for fruitful speculation.

The differing paths for the Slave and Superior blocks imply that they were moving separately in the interval 2300 to 2100 Ma. The absence of reliable information in the interval 2000 to 1900 Ma means that relative motion could have continued up to about 1850 Ma after which it appears to have ceased.

In recent years paths of APW relative to Laurentia as compiled by different workers have tended to become increasingly complicated. These complications could be real, but it is at least equally possible that the segments of the APW path relative to each crustal block within Laurentia are comparatively simple, and that the apparent complexity, exhibited by the data for Laurentia when considered as a whole, reflects relative motions (rotations or latitudinal motions) of the different crustal blocks. The main purpose of this short review has been to argue the latter case.

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11.

PALEOMAGNETIC RESULTS FROM THE LOWER PROTEROZOIC ROCKS OF GREAT SLAVE LAKE AND BATHURST INLET AREAS, NORTHWEST TERRITORIES

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Abstract

Paleomagnetic results are reported from some 300 redbed horizons in the Athapuscow aulacogen and Kilohigok Basin involving the Sosan, Kahochella, and Christie Bay groups in the former and the Western River, Mara, and Peacock Hills formations in the latter. Magnetizations are commonly multiphased, requiring the recognition and removal of recent and ancient overprints before original primary remanences are revealed. For the most part the poles deduced from these formations form a group off the west coast of South America with a mean at 92°W, 14°S (N=8, K=23, A_{95}=12°); no significant polar wander is detectable within the interval represented. This pole (referred to here as the Lower Coronation Pole) indicates that for much of its sedimentation history the Coronation Geosyncline occupied tropical latitudes (about 10°) which is compatible with the abundance of stromatolites and the occurrence of evaporites. In addition to this Lower Coronation Pole, there is a widespread overprint, thought to be associated with uplift and cooling in the Coronation Geosyncline, corresponding to a pole at 091°W, 21°N (K=14, A_{95}=5°).

The pole obtained from the Western River Formation, which is the oldest unit studied, lies off the west coast of Africa, some 80° from the Lower Coronation Pole. One speculative interpretation of this discrepancy involves relative motions between subplates of the Canadian Shield, but it is more likely that it simply represents a time gap of perhaps as much as 200 Ma.

The results from the Stark and Tochatwi formations are aberrant, possibly because of a local tectonic rotation. This suggestion has been tested by collecting new material from the Stark Formation in an area thought to be undisturbed. Unfortunately, the strata there are brecciated and the specimens yield highly scattered directions which do not resolve the rotation problem.

Since the samples were collected in ordered stratigraphic sequences it has proved possible to delineate geomagnetic polarity zones. The magnetostratigraphy of these ancient sedimentary basins is thus beginning to emerge, enabling certain tentative correlations between the various elements of the Coronation Geosyncline to be made.

Résumé

On présente les résultats de levés paléomagnétiques effectués sur environ 300 horizons à couches rouges (red beds), situés dans l'aulacogène d'Athapuscow et le bassin de Kilohigok, englobant les groupes de Sosan, Kahochella et Christie Bay dans l'aulacogène, et les formations de Western River, Mara et Peacock Hills dans le bassin. Généralement, les magnétisations sont multiphasées et exigent l'identification et l'élimination de surimpressions récentes et anciennes, pour laisser apparaître les rémanences primaires originales. En majorité, les pôles déduits de ces formations forment un groupe au large de la côte ouest de l'Amérique du Sud, de coordonnées moyennes 92°W, 14°S (N=8, K=23, A_{95}=12°), et l'on ne peut déceler aucun mouvement notable des pôles correspondant à l'intervalle de temps représenté. Ce pôle (appelé ici pôle de la partie inférieure de la formation de Coronation), indique que pendant une grande partie de son histoire sédimentaire, le géosynclinal de Coronation a occupé des latitudes tropicales (à environ 10°), ce qui concorde avec l'abondance des stromatolites et la présence d'évaporites. Outre ce pôle, on observe une vaste surimpression, que l'on croit associté aux périodes de soulèvement et de refroidissement subies par le géosynclinal de Coronation, et qui est représentée par un pôle de coordonnées 091°W, 21° (K=14, A_{95}=5).

Le pôle obtenu pour la formation de Western River, la plus ancienne unité étudiée, se situe au large de la côte ouest de l'Afrique, à environ 80° du pôle de la partie inférieure de la formation de Coronation. Pour tenter d'expliquer cette discordance, on a considéré les mouvements relatifs entre les sous-plaques du Bouclier canadien, mais il est plus vraisemblable qu'elle correspond à une lacune stratigraphique de peut-être 200 Ma.

Les résultats obtenus sur les formations de Stark et de Tochatwi sont aberrants, peut-être en raison d'une rotation tectonique locale. On a mis à l'épreuve cette suggestion, en recueillant de nouveaux détails sur la formation de Stark, dans une zone que l'on jugeait non perturbée. Malheureusement, les strates de cette formation contiennent des brèches, et les échantillons examinés montrent une importante dispersion de leurs directions, ce qui ne permet pas de résoudre le problème de rotation.

Étant donné que l'on a recueilli les échantillons en suivant le profil stratigraphique, on a pu délimiter les zones de polarité géomagnétique. On commence ainsi à reconnaître la magnétostratigraphie de ces anciens bassins sédimentaires, et donc à établir certaines corrélations entre les divers éléments du géosynclinal de Coronation.

INTRODUCTION

The Athapuscow aulacogen and Kilohigok Basin (Fig. 11.1) both contain several thousand metres of sediments which are for the most part well-exposed and relatively unmetamorphosed. Of particular interest to the present work are the strata of the Great Slave Supergroup of the Athapuscow aulacogen and the Goulbourn Group of the Kilohigok Basin. The formations involved (Table 11.1) are genetically related to the development of the Coronation Geosyncline, which developed along the northwest border of the Slave Province in late Aphebian time, i.e. 2000 to 1750 Ma (Hoffman, 1973; 1980; 1981). Paleomagnetic investigations of several formations have been undertaken to aid in understanding the tectonic evolution of this part of the Canadian Shield and to provide a possible means of correlation between the various elements of the Coronation Geosyncline.

In this paper geological details are kept to a minimum since several excellent accounts already exist (e.g. Hoffman, 1968; Hoffman et al, 1970; Campbell and Cecile, 1976), and others appear in this volume. Procedural details of paleomagnetic field sampling and laboratory work are kept as brief as posible, except where they depart from well-established techniques (see textbooks by Irving (1964) and McElhinny (1973)). The data base for each formation, some of which have already been published, is described briefly before turning to the interpretation and significance of the results.

KILOHIGOK BASIN

New results are reported from three sedimentary formations – the Western River, Mara, and Peacock Hills formations. Data from two sites in a gabbro sill intruding the Western River formation are also described.

Western River Formation

This formation rests unconformably on Archean Yellowknife-like sediments, locally with a regolith and thin conglomerate at the base. These are succeeded by a Lower member, a red siltstone member, a quartzite member, and finally an Upper Argillite member (Campbell and Cecile, 1981). The Red Siltstone member, which ranges in thickness from 100-200 m, was sampled at three sections

Table 11.1. Stratigraphy of the Athapuscow aulacogen and Kilohigok Basin.

			Athapuscow	Kilohigok
	Et-Then		Preble	
			Murky	Tinney Cove
GREAT SLAVE SUPERGROUP	Christie Bay		Unnamed (Hoffman, 1981)	Amagok
			Pearson	
			Portage Inlet	
			Tochatwi*	Brown Sound
			Stark*	Omingmaktook Mbr.
	Pethei		Hearne	
			Wildbread	
			Pekanatuï	
			Blanchet	Kuuvik
			McLean	
			Utsingi	
			Taltheilei	
			Douglas Peninsula	
	Kahochella	Seton	Charlton Bay*	
			McLeod Bay*	Peacock Hills*
			Gibraltar*	
			Unnamed (Hoffman, 1981)	Quadyuk
	Sosan		Akaitcho River*	Mara*
			Kluziai	Burnside River
			Duhamel	Western River*
			Hornby Channel	

Note: Asterisks indicate formations for which paleomagnetic data are reported in this paper

(Fig. 11.2); section A (140 m; 10 sites = horizons; 34 drill cores), B (53 m; 10 sites; 31 cores), and C (23 m; 5 sites; 15 cores). Thermal demagnetization studies revealed a stable remanence well removed from the present field at the sampling locality at almost all sites (Fig. 11.3, Table 11.2). Most have magnetic vectors pointing eastwards with shallow positive dip, but 5 sites are oppositely polarized. Application of a bedding correction slightly increases the precision parameter (from 11 to 12) but this improvement is not statistically significant, due to the fact that dips are small and fairly constant throughout the region studied.

Mara Formation

This formation generally consists of red siltstones and fine subarkose, and varies in thickness from 100 to 500 m. It was sampled in the uppermost portion immediately underlying the Quadyuk Formation.

The bulk of the samples were collected from a locality on the shores of Bathurst Inlet (107°53'W 67°08'N, Fig. 11.1), but a few samples were also obtained from an area about 100 km to the south (D in Fig. 11.2) that was visited briefly. At this southern locality 5 stratigraphic horizons (15 drill cores) were obtained spread over a thickness of about 500 m. At the Bathurst Inlet locality 43 horizons (155 samples, mostly drill cores) were sampled over about 300 m.

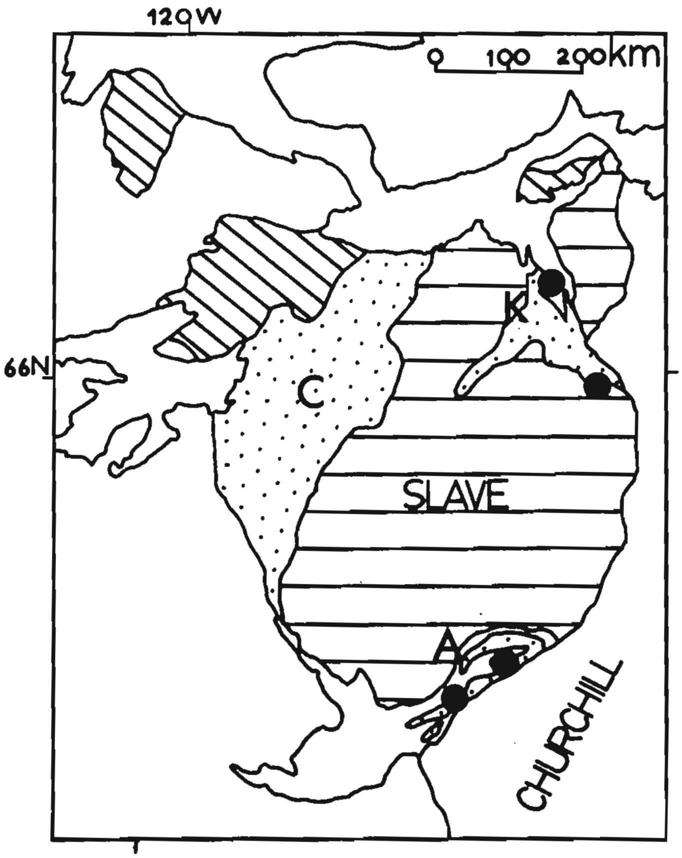

Figure 11.1. Sketch map showing the Slave Province (horizontal lines) and the Coronation Geosyncline with associated aulacogens (dots). Oblique lines represent Helikian and Hadrynian cover rocks. Sampling areas are indicated by large dots.

C = Coronation Geosyncline
K = Kilohigok Basin
A = Athapuscow Aulacogen

Thermal demagnetization was carried out on all samples and most sites yield statistically significant mean magnetic vectors well removed from the present field at the sampling locality (Fig. 11.4, Table 11.3). Two opposed polarity groups emerge but there are also a number of transitional directions, particularly in the upper part of the Bathurst Inlet section, in the 60 m underlying the Quadyuk Formation. These transitional sites are eliminated before averaging the remaining data to obtain a mean direction. In addition rocks at several sites are pale in colour and are very weakly magnetized – these yielded scattered remanence vectors that were highly unstable on demagnetization. It seems that these sites either never had, or have not retained, a remanence measurable with currently available equipment, and they are therefore excluded from further discussion. This leaves 28 site means, of which 9 have mean remanence vectors directed to the south and 19 to the north.

The results from locality D serve to confirm the two polarity groups observed at Bathurst Inlet but the site means are more scattered and the two polarities are not antiparallel. It is possible that some of these sites record

Table 11.2. Site mean results for Western River Formation, after application of bedding correction

Site	N	T	k	D	I	
WAA†	-	-	-	-	-	
WAB	3	600	400	058	+21	
WAC	3	600	22	128	+17	
WAD	3	600	26	079	+15	
WAE	3	600	32	090	+40	
WAF	3	600	220	257	-16	
WAG	3	600	64	111	+27	
WAH	3	600	152	094	+18	
WAI	3	600	52	109	+37	
WAJ	3	600	19	072	+18	
WBA	1	600	-	280	-05	
WBB	2	600	135	266	-25	
WBC	3	600	14	300	-07	
WBD	3	600	6	274	-28	
WBE	2	660	75	064	+35	
WBF†	-	-	-	-	-	
WBG	3	600	28	076	+24	
WBH	3	600	29	086	+17	
WBI	3	600	168	082	+23	
WBJ	3	600	8	126	+22	
WCA	3	600	113	096	+42	
WCB	3	500	27	065	+39	
WCC	3	600	43	050	+37	
WCD	3	500	9	089	+08	
WCE	3	500	61	040	+30	
MEAN	22 sites		12	086	+25	α_{95}=9°

N = no. of cores
T = temperature (°C)
k = Fisher's precision parameter
D = declination (°E of N)
I = inclination (+ downwards)

†These two sites yielded specimen results spread between the two polarity groups indicated by the remaining sites. They appear to contain both polarity vectors that have not been resolved by thermal demagnetization.

N.B. Site WBA is not included in the mean since N is only 1; although its result is in good agreement with the bulk of the other sites.

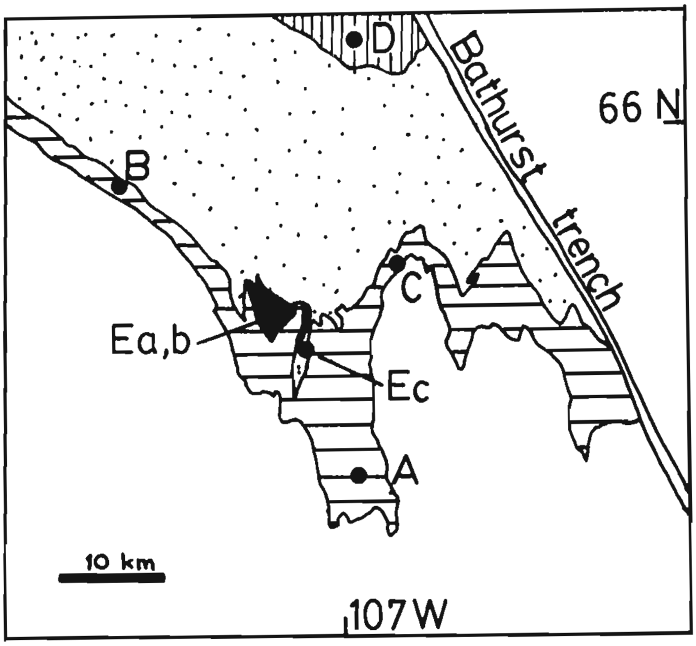

Figure 11.2. Sketch map showing sections sampled in the Western River Formation (A,B,C), gabbro sill (Ea,Eb), and baked contact (Ec).

Horizontal lines = Western River Formation
Dots = Burnside River Formation
Vertical lines = Mara Formation

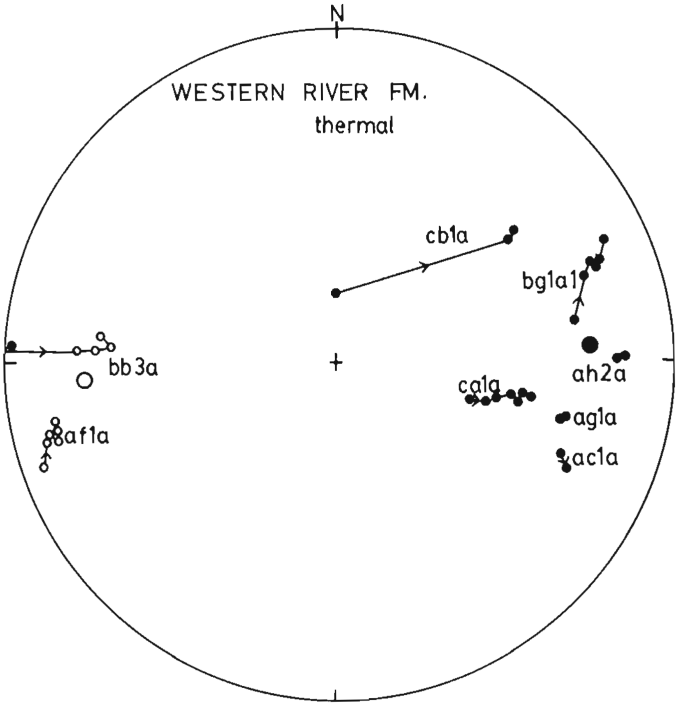

Figure 11.3. Typical thermal demagnetization results for the Western River Formation. Equal area net, solid (open) symbols on the lower (upper) hemisphere. Large symbols represent the overall mean (see Table 11.2).

transitional directions – indeed there is a definite trend in the sequence DA-DB-DC (Table 11.2). With only 5 sites however it is impossible to demonstrate this convincingly, and the D section data are therefore used only as supporting evidence, and are not included in the overall mean for the Mara Formation. This is not a critical decision – inclusion of these 5 sites changes the overall mean given in Table 11.2 by only 3°.

Declination logs (Fig. 11.5) indicate that the lower horizons in both sections are normally polarized (magnetic poles near South America are designated normal for the purposes of this discussion). Between 150 and 200 m below the Quadyuk Formation in the Bathurst Inlet area a reversal takes place, but no sustained northerly group of directions occurs above the transition. Instead there is a complex alternation of normal and reversed horizons. In some cases (e.g. Fig. 11.6) the transition from one polarity state to the other has been recorded. At the D section a reversal also takes place and declination changes from southerly to northerly, but 5 sites are insufficient to reveal finer details. The correlation suggested by Figure 11.5 is possible, but without firmer marker horizons at both localities it must be regarded as promising but tentative.

Peacock Hills Formation

Sixteen sites were samples in a 70 m thick section starting immediately above the Quadyuk Formation in the same locality as the main Mara Formation section. An additional two sites were sampled on a small island 2.5 km to the south. Most of the collection consists of oriented hand samples (48 out of 68) due to difficulties encountered in field drilling. Twelve drill cores were also collected from the Quadyuk Formation, but this stromatolitic carbonate material was very weakly magnetized and yielded unstable, scattered remanence vectors, which are not discussed hereafter.

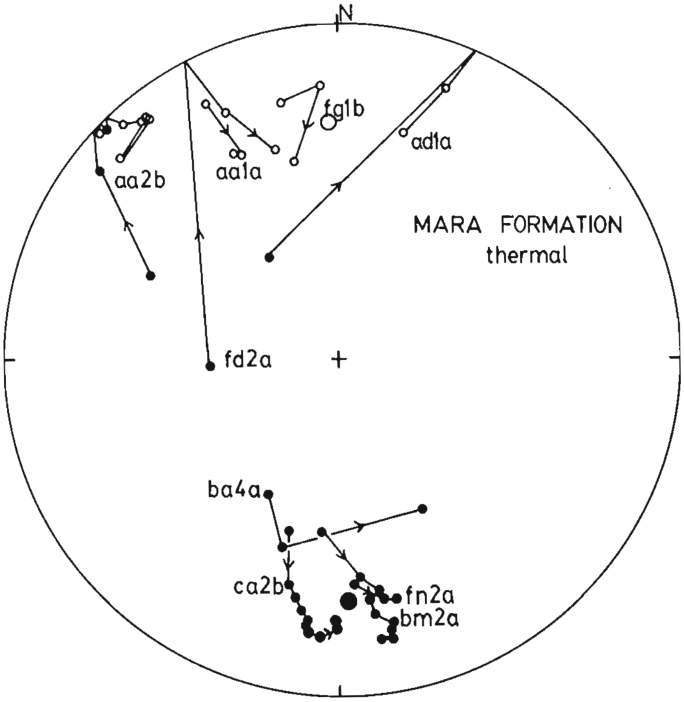

Figure 11.4. Typical thermal demagnetization results for the Mara Formation. Symbols as in Figure 11.3.

During thermal demagnetiztion many Peacock Hills specimens move first to southerly declinations and then abruptly switch to northerly declinations at temperatures between 500 and 700°C (Fig. 11.7); other specimens exhibit northerly directed NRM's which move very little on demagnetization; a few specimens possess stable southerly directions. This type of behaviour suggests the presence of two opposed polarities as described by Roy and Park (1972) and others. It is usually interpreted in terms of a primary remanence with an overprint acquired after one or more field reversals; thus the overprint may not greatly postdate the primary remanence. Roy and Park (1972) have demonstrated that chemical leaching can be an effective technique for resolving the various magnetic components, and we plan to undertake such work with a high-pressure device currently under construction. Thus the Peacock Hills Formation data reported here are of a preliminary nature, with only one sample per site studied in most cases.

Table 11.3. Site mean results for the Mara Formation, after bedding correction

Site	N	T	k	D	I	
BAA	3	500	31	332	-23	
BAC	3	600	188	352	-60	
BAD	4	500	44	018	-03	
BAE	3	600	184	231	+15	
BAF	3	500	11	199	+30	
BAG	3	500	5	163	+01	
BBA	4	500	58	184	+38	
BBB	3	500	46	192	+36	
BBC	3	600	17	164	+39	
BBD	3	500	130	182	+24	
BBE	3	500	32	185	+33	
BBF	3	500	61	183	+32	
BBL	4	500	43	185	+52	
BBM	5	500	46	172	+41	
BBN	4	500	31	179	+55	
BCA	4	500	289	188	+21	
BCB	4	600	153	184	+30	
BCD	4	660	10	003	+15	
8BFC	4	500	9	209	+21	
8BFD	5	600	14	340	-40	
8BFE	4	500	260	166	+28	
8BFF	4	500	122	164	+21	
8BFG	4	600	807	357	-23	
8BFH	4	600	401	338	-09	
8BFI	4	600	12	142	-13	
8BFK	4	600	76	348	-41	
8BFN	4	500	324	176	+26	
8BEA	4	600	867	007	-27	
MEAN	28 sites		12	179	+29	α_{95}=8°
DE	3	630	49	325	-05	
DD	3	630	29	338	+17	
DC	3	600	22	146	+38	
DB	3	600	9	182	+55	
DA	3	630	288	154	+71	

- This table excludes 'transitional' sites and sites too weakly magnetized to be reliably measured.
- See Table 11.2 for explanation of symbols.
- Sites starting with D refer to locality D of Figure 11.2.

At this stage we regard the high-temperature results as the best indicator of the "original" magnetization although endpoints are commonly not well-established. The appropriate data are illustrated in Figure 11.8, and summarized in Table 11.4.

Declination logs indicate that the lowermost parts of the sections are normally magnetized, with a reversal taking place within the first few metres above the Quadyuk. With one exception (8BGJ) this polarity is maintained throughout the rest of the section (Fig. 11.9). Further work may remove this exception since the two specimens demagnetized to date agree up to 500°C, (the result given in Table 11.4), but at higher temperatures one swings to a northerly direction as do specimens from the other sites indicating that this site also possesses an underlying reversed magnetization.

Gabbro Sill

Three cores were drilled at each of two sites in a gabbro sill intruding the Western River Formation (Fig. 11.2, Ea, Eb) at a point where these sediments dip at about 5° to the northeast. One site was centrally located and one was near the margin of the mapped outcrop. Site Ec is situated in the Red Siltstone member of the Western River Formation about 20 m from an outcrop of gabbro.

Igneous material was subjected to alternating-field demagnetization and sediments to thermal cleaning. The results are summarized in Table 11.5, and typical demagnetizations are illustrated in Figure 11.10. The remanence vectors move away from the present field towards a shallow northerly direction reminiscent of results from the Mara and Peacock Hills formations (Fig. 11.4, 11.8).

It is important to note that the data from site Ec agree with those of the intrusion but diverge markedly from the other Western River Formation results. This constitutes a convincing baked contact test which strongly favours the view that the magnetization of the intrusion dates from its time of emplacement. Tremblay (1971) reported a K-Ar

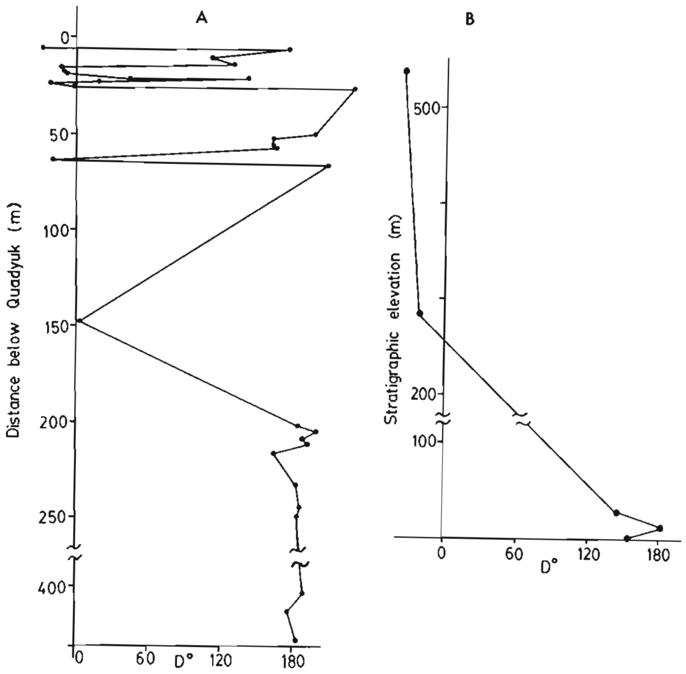

Figure 11.5. *Declination magnetograms for the Mara Formation.*

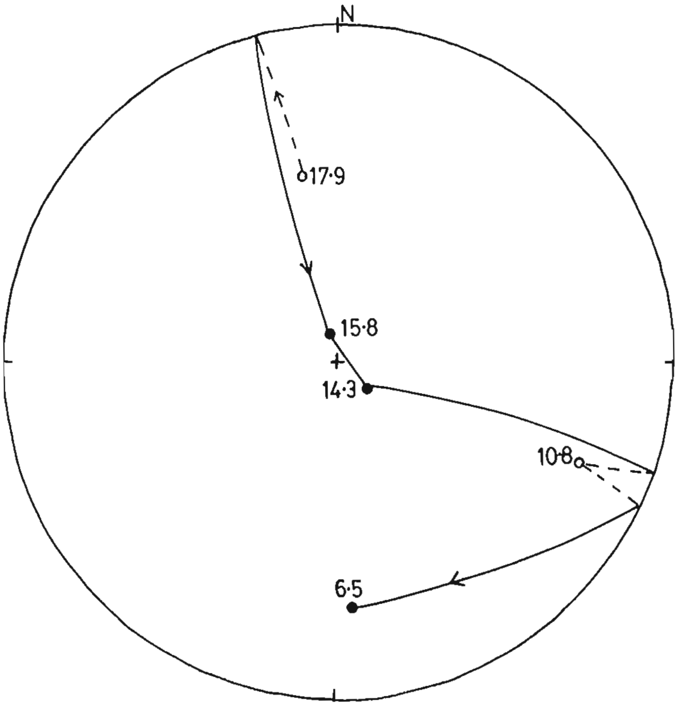

Figure 11.6. The geomagnetic polarity transition recorded between 17.9 m and 6.5 m below the Quadyuk Formation. Symbols as in Figure 11.3.

whole rock age of 1215 Ma for this intrusion which suggests a correlation with the widespread Mackenzie Igneous Episode, but this conflicts with the paleomagnetic evidence. The Mackenzie intrusions have been studied paleomagnetically in areas extending across much of the Shield and yield a tight cluster of poles with a mean at 171°W, 01°N (Irving et al., 1972). The data presented here indicate that the intrusion is magnetized in a direction similar to the Mara and Peacock Hills formations, which yield poles at 107°W, 07°S and 090°W, 15°S respectively (see below). It therefore seems more probable that the intrusion was emplaced about 1800 Ma, and does not, in fact, belong to the Mackenzie Igneous Episode.

ATHAPUSCOW AULACOGEN

Investigations of Athapuscow aulacogen rocks carried out by the University of Alberta paleomagnetic group have appeared in the literature (McMurry et al., 1973; Evans and Bingham, 1976; Bingham and Evans, 1976; Evans et al., 1980; Reid et al., 1981) and discussion is therefore restricted to the points essential for this review.

Akaitcho River Formation

Evans et al. (1980) reported results from 3 sections each about 200 m thick; 40 horizons were sampled, the total number of oriented cores being 120. Thermal demagnetization reveals 2 polarity groups, but these are not exactly antiparallel. This is attributed to a previously recognized widespread magnetic overprint thought to have been acquired during uplift and slow cooling in the Coronation Geosyncline (Irving and McGlynn, 1979). A novel statistical technique was employed to remove this overprint and obtain a best estimate of the underlying original magnetization, which has a mean direction of D=160°, I=+36° (N=35, k=13, α_{95}-7°).

Table 11.4. Site mean results for the Peacock Hills Formation, after dip correction

Site	N	T	k	D	I	
8BGE	1	630	-	298	-10	
8BGG	1	600	-	171	+11	
8BGH	1	500	-	173	+21	
8BGI	1	660	-	342	+02	
8BGJ	2	500	16	188	+25	
8BGK	1	660	-	323	-19	
8BGM	1	660	-	355	+07	
8BGN	1	600	-	324	+09	
8BGO	2	660	12	322	-05	
8BGP	1	660	-	334	+07	
8BGQ	1	600	-	343	+02	
8BGR	1	600	-	353	-11	
8BGS	2	600	22	352	-35	
8BIA	4	600	106	207	+37	
8BIB	4	600	33	342	-27	
MEAN	15		11	342	-12	α_{95}=12°

Symbols are explained in Table 11.2

Kahochella Group

Reid (1972) investigated a thick stratigraphic sequence located on the north coast of Keith Island centred at 111°57'W, 62°03'N. Forty-nine sites were sampled with a typical stratigraphic spacing of 25 m. Some of this work was reported by McMurry et al. (1973) and a full discussion can be found in Reid et al. (1981). After alternating-field and thermal demagnetization a dominant magnetic component in the southeast quadrant with moderate inclinations is observed, with a few sites indicating reversed directions, although never with well-established end-points. As discussed more fully in the next section, Reid (1972) recognized a postfolding overprint which appears to have totally overprinted the original remanence at some sites, but 18 sites retain, at least in part, an earlier remanence from which Reid et al. (1981) deduced a mean direction of D=129°, I=+21° (N=18, k=12, α_{95}=10°). Both polarities occur and are stratigraphically intermingled, with at least 7 reversals implied.

Reid et al. (1981) also reported on five sites in the biogenic, primarily calcareous rocks of the Pethei Group, but poor results from these sites did not provide new information about the ancient geomagnetic field.

Christie Bay Group

Bingham and Evans (1975, 1976) have reported in detail on results from the Stark Formation. They sampled a total of 55 sites, 45 of them located in a single 600 m thick section. The two most significant findings were (a) a profile through an early Precambrian geomagnetic field reversal, and (b) an apparently reliable but highly aberrant paleomagnetic pole. This latter point has been the subject of some debate (Gough et al., 1977; Irving and McGlynn, 1979), the question being whether or not the small area in which the collection was made has undergone rotation about a local vertical axis. There is some structural evidence for this since the bedding strikes in this area diverge 50-60° from the aulacogen as a whole (Bingham and Evans, 1976). Allowance for a rotation of this kind brings the pole into excellent agreement with results from correlative units. Use of such rotations requires extreme caution if circular arguments are to be avoided. We have therefore attempted to resolve this problem by collecting Stark Formation material from another area.

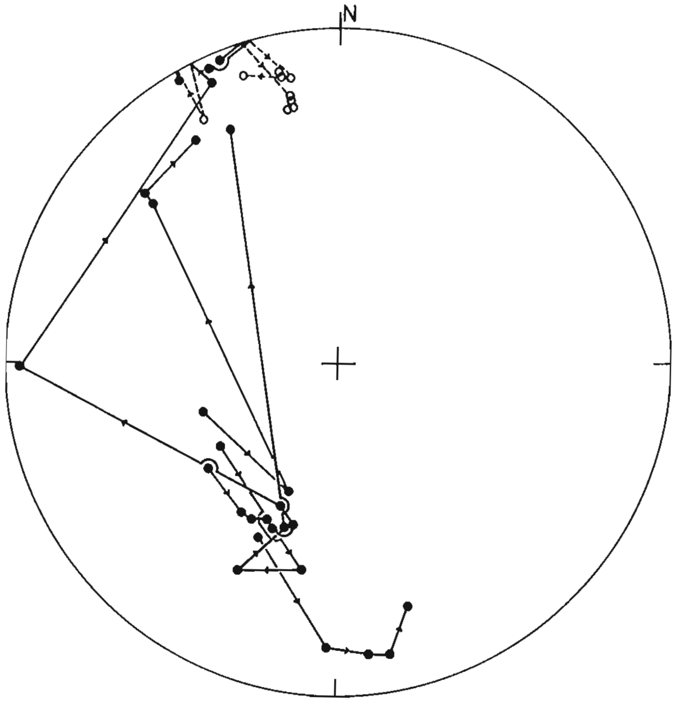

Figure 11.7. *Typical thermal demagnetization results for the Peacock Hills Formation. Symbols as in Figure 11.3.*

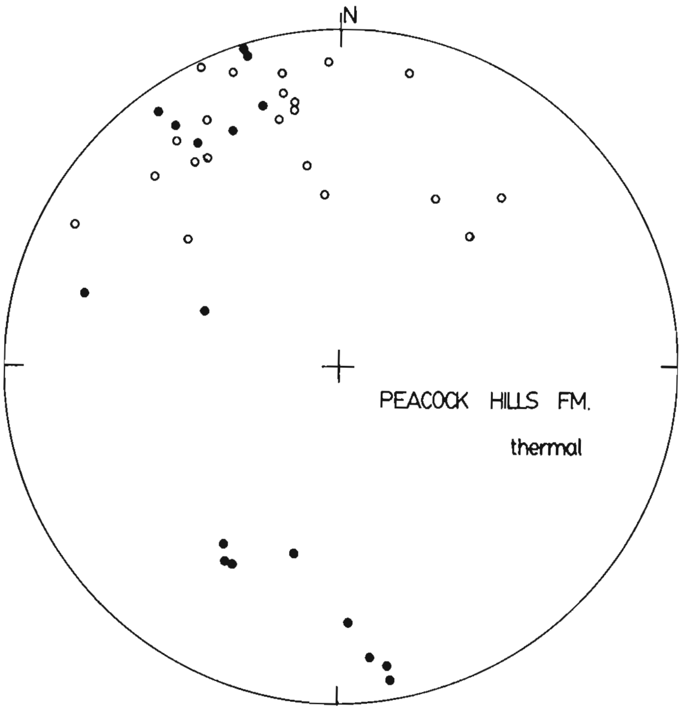

Figure 11.8. *Summary of high temperature remanence (600°C and above) in Peacock Hills Formation samples. Symbols as in Figure 11.3.*

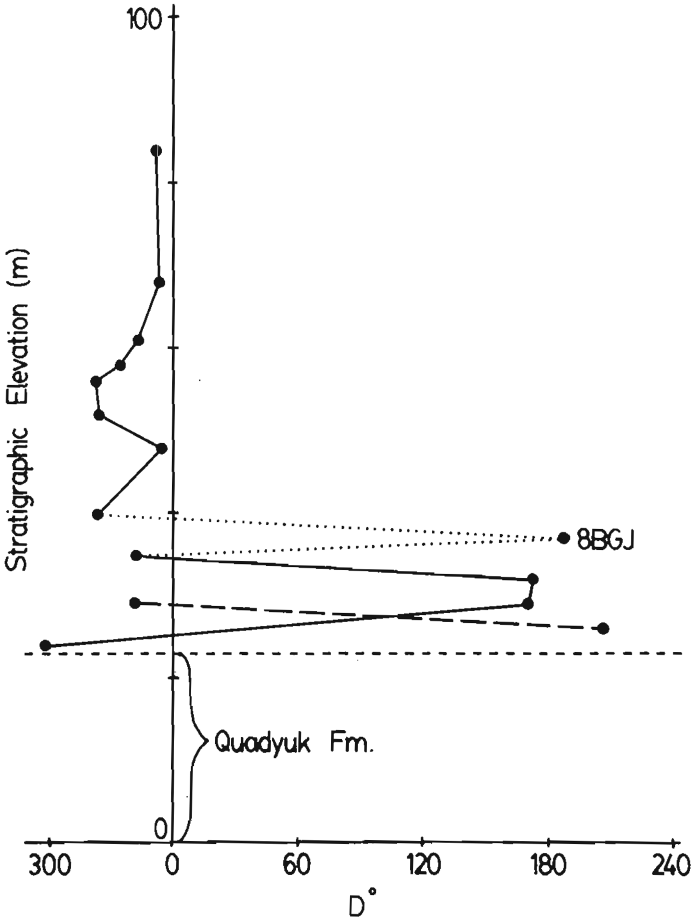

Figure 11.9. *Declination magnetograms for the Peacock Hills Formation. The dashed line connects the two sites collected at a separate locality (see text), and the lines to site 8BGJ are dotted as the results from this site are complex (see text).*

Eighteen sites were samples on the north shore of Keith Island (111°48'W, 62°06'N) for this purpose. Pilot thermal demagnetization was carried out on 16 specimens, and in most cases acceptable endpoints were reached. The results are very scattered, however, and this remains true when the remaining specimens (after heating to 600°C) are included (Fig. 11.11). This scatter is not only between sites, but also between samples, and even between specimens from a single oriented sample. It almost certainly results from the fact that the Stark Formation here is brecciated, as it is in most outcrops (Hoffman et al, 1977). In fact, one of the few localities where it is not brecciated is the Snowdrift area where the original sampling was done and which yielded the aberrant pole. These new data provide what is essentially a conglomerate test (Graham, 1949), and since the brecciation followed very shortly after deposition (Hoffman et al., 1977) provides tight control on the age of the magnetization of the Stark Formation. However, they obviously do not help to resolve the rotation problem.

Evans and Bingham (1976) have studied the Tochatwi Formation at a total of 29 sites in the Snowdrift area. Most of these possess some kind of magnetic overprint (see below), but 8 sites retain a pre-folding magnetization in close agreement with the Stark Formation and thus yield a similarly aberrant pole.

Table 11.5. Site mean results for the gabbro intrusion and contact sediments

Site	N	T/M	k	D	I
Ea	3	40 mT	46	338	-01
Eb	3	30 mT	226	347	+17
Ec	3	600°C	12	354	+04

Symbols are explained in Table 11.2

M = peak alternating field

Table 11.6. Summary of recent overprints obtained by difference vectors.

Formation	N	R	D	I	k	α_{95}
Tochatwi[1]	16	0-200°C	266	+85	8	14°
Stark[2]	34	0-40mT	347	+87	18	6°
Akaitcho River[3]	33	0-500°C	075	+83	31	5°
Seton[4]	17	0-400°C	073	+86	34	6°

[1]Evans and Bingham, 1976
[2]Bingham and Evans, 1976
[3]Evans et al., 1980
[4]Irving and McGlynn, 1979

N = number of sites
R = temperature or AF range
Other symbols as in Table 11.2

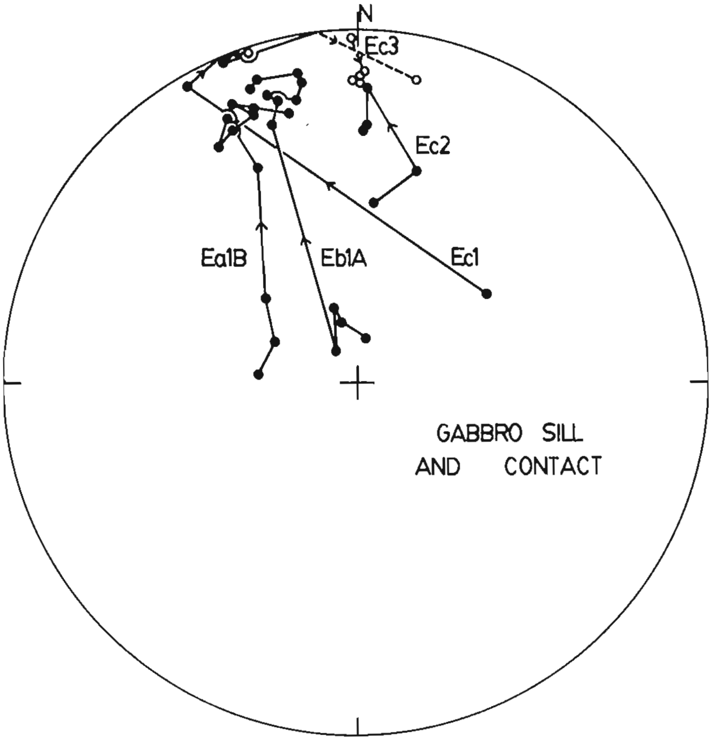

Figure 11.10. Typical thermal (Ec) and magnetic (Ea, Eb) cleaning results from the gabbro sill and baked contact. Symbols as in Figure 11.3.

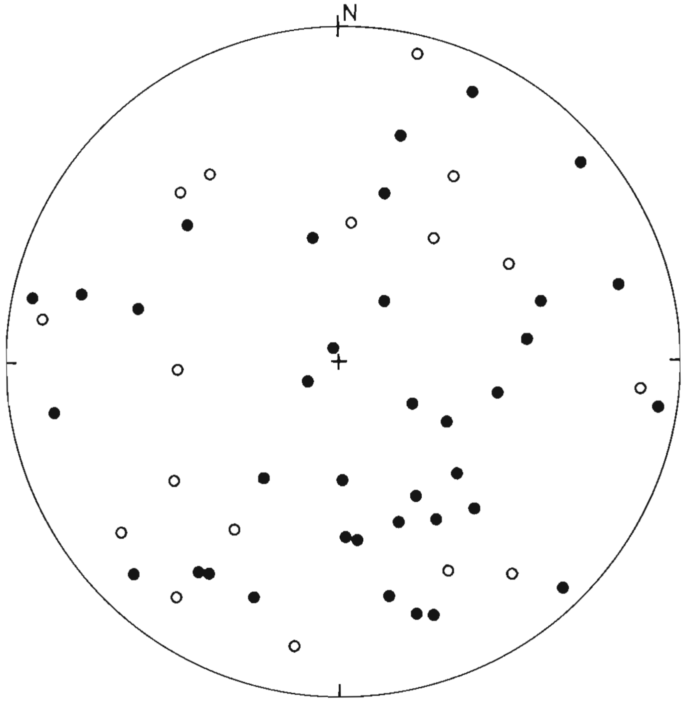

Figure 11.11. Thermal end-point results from brecciated Stark Formation collected on Keith Island (see text). Symbols as in Figure 11.3.

DISCUSSION

Magnetic Overprints

Polyphase magnetization is the rule rather than the exception in these early Proterozoic formations. Detailed demagnetization studies are sometimes effective in isolating the various magnetic components but it is quite probable that many apparently aberrant data are caused by unresolved composite magnetizations.

One effective way of investigating polyphase remanence is to consider the magnetic vectors removed during each step in incremental demagnetization runs. This technique was developed in the early 1970s by E.W. McMurry and A.B. Reid at the University of Alberta and has become a widely accepted procedure in the last decade. It commonly enables one to recognize a recent (probably viscous remanent magnetization, VRM) magnetization acquired in the present ambient field and preferentially carried by grains of low blocking temperature. The Stark, Tochatwi, Akaitcho River, and Seton formations clearly show such an overprint (Fig. 11.12a, Table 11.6). Where geological dips allow a fold test (Stark and Akaitcho River formations) the standard statistical test (Graham, 1949; McElhinny, 1964) demonstrates that this vector was acquired after folding, as expected.

Ancient overprints can also be isolated by vector subtraction, as in the example shown in Figure 11.12b. Here the overprint can be convincingly attributed to heating by a nearby Mackenzie-age intrusion or unexposed equivalent (Evans and Bingham, 1976).

Even where apparently stable remanences occur two (or possibly more) phases may remain. This can sometimes be recognized if reversals occur. A formation which carries a secondary magnetization superimposed on a primary remanence with both normal and reversed polarity generally will yield two polarity groups that are not exactly antiparallel. This is very important to a proper understanding of the Coronation Geosyncline paleomagnetic data. In many of the formations where two polarities are observed these are significantly non-antiparallel. For example the means of the

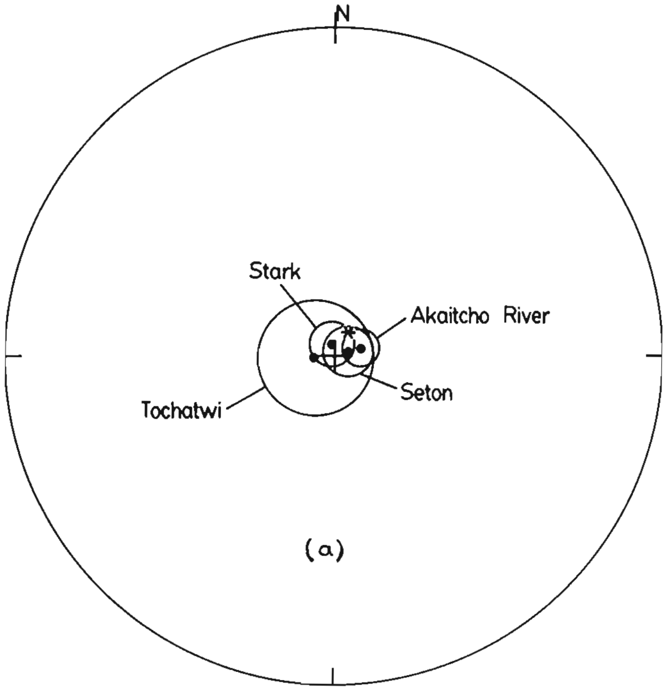

Figure 11.12a. *Recent magnetic overprints in Great Slave Supergroup formations. The asterisk represents the present field direction.*

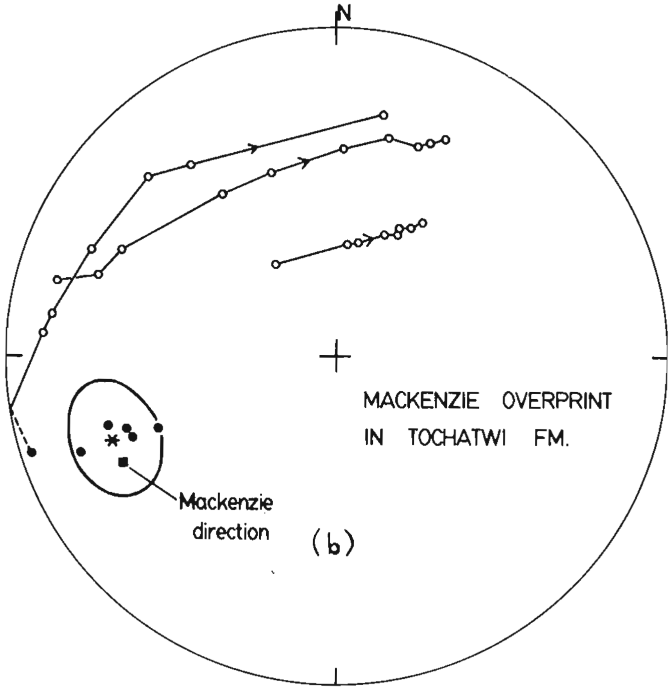

Figure 11.12b. *Mackenzie Igneous overprint carried by Tochatwi Formation samples. The curves illustrate typical behaviour during thermal demagnetization. As the Mackenzie overprint is removed the magnetization vectors move gradually towards the primary Tochatwi magnetization (030,-11). The dots represent the magnetic vectors removed between 20 and 500°C during thermal demagnetization, and the asterisk their mean. Symbols as in Figure 11.3.*

normal and reversed groups of the Akaitcho River Formation are separated by 142° ± 14° (Evans et al., 1980). Similar results are obtained from the Seton Formation (Irving and McGlynn, 1979) and formations of the Kahochella Group (Reid et al., 1981). The key to this discrepancy is the recognition of an ancient overprint first described by Reid (1972; see also McMurry et al., 1973) and subsequently found by other workers (Evans and Bingham, 1976; Irving and McGlynn, 1979). Difference vectors for moderate to high temperature intervals (mostly between 400° and 600°) form a tight cluster directed steeply downwards in the southeast quadrant. This direction is also recorded by a few sites that seem to have been totally overprinted and can be recognizd without resorting to difference vectors. The relevant data have been compiled by Evans et al. (1980) and their mean lies at D=152°, I=+64° with N=75 sites, k=30, and α_{95}=3°. This overprint is postfolding (Reid, 1972) and is attributed by Irving and McGlynn (1979) to uplift and cooling in the Coronation Geosyncline. Thus we refer to it as the Coronation overprint, although it may, in fact, be much more widespread, extending as far as Greenland (e.g. Morgan, 1976). This overprint was acquired about 1700 Ma, and provides a useful paleomagnetic pole which will be discussed below in the context of apparent polar wander.

Given that the Coronation overprint direction is precisely determined, it is possible to mathematically subtract a suitable vector from those formations exhibiting non-antiparallel reversals to reveal the underlying original magnetization. The details of the methods used are described in Reid et al. (1981) and Evans et al. (1980). The former authors deduced that the thermally cleaned Kahochella samples carry an overprint whose intensity is, on average, 30 per cent that of the underlying 'primary' remanence: the latter deduced a corresponding value of 45 per cent for the Akaitcho River Formation. The 'primary' poles deduced from these 'decontamination' procedures are discussed more fully below.

Magnetostratigraphy

Bingham and Evans (1975, 1976) have reported a detailed study of a geomagnetic polarity reversal in the Stark Formation, and a further example is reported here from the Mara Formation (Fig. 11.6). Such features are of interest to geomagnetists as records of field behaviour from very ancient times. They also offer the possibility of correlation between widely separated, undated, unfossiliferous strata; we have attempted to assess this possibility by collecting from a number of stratigraphically ordered sequences.

Polarity reversals found in the Mara and Peacock Hills formations are summarized in Figures 11.5 and 11.9 in the form of declination magnetograms. They indicate the following simplified sequence – at the base some 200 m of normally magnetized strata (southerly declinations), followed by about 150 m of alternating polarities (at least 9 reversals) up to the Quadyuk Formation, above which the first 60 m of the Peacock Hills Formation is mostly reversely magnetized.

The Akaitcho River Formation of the Great Slave Lake area exhibits a very similar pattern (Evans et al., 1980; Fig. 11.4). It seems likely, therefore, that the reversal which occurs about 200 m below the Quadyuk Formation in the Bathurst Inlet area is equivalent to the one occurring about 100 m above the base of the Akaitcho River Formation in the Great Slave Lake area (Fig. 11.13). This is consistent with the broad correlation suggested by Hoffman et al. (1970), but is by no means compelling. Above this reversal in the Bathurst Inlet area a 150 m gap contains only one sampled horizon. Two possibilities arise from this unfortunate circumstance. One is that the 100 m of mixed polarities plus the 200 m of reversed Akaitcho River Formation strata

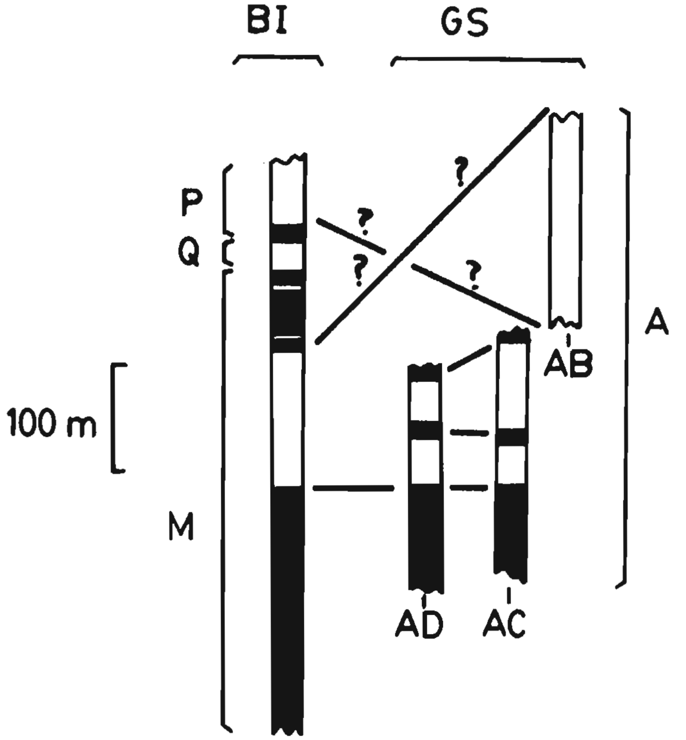

Figure 11.13. *Schematic summary of geomagnetic reversals in the Bathurst Inlet (BI) and Great Slave Lake (GS) areas. GS sections are labelled AB, AC, and AD as in Evans et al. (1980)*

Black = normal polarity
P = Peacock Hills Formation
Q = Quadyuk Formation
M = Mara Formation
A = Akaitcho River Formation

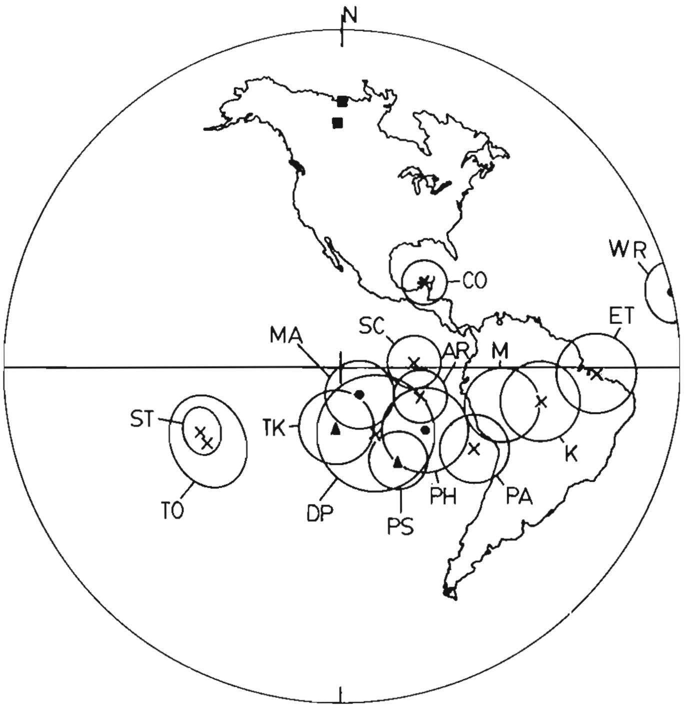

Figure 11.14. *Paleomagnetic poles. The asterisk labelled CO represents the Coronation overprint pole (see text). Squares represent the sampling localities.*

crosses	=	*Athapuscow Aulacogen*
dots	=	*Kilohigok Basin*
triangles	=	*Coronation Geosyncline*
AR	=	*Akaitcho River Formation*
MA	=	*Mara Formation*
DP	=	*Douglas Peninsula Formation*
ET	=	*Et-Then Group*
K	=	*Kahochella Group*
M	=	*Martin Formation*
PA	=	*Pearson Formation "A"*
PH	=	*Peacock Hills Formation*
PS	=	*Peninsula Sill*
SC	=	*Seton Formation "C" (wrongly labelled SA in Irving and McGlynn, 1979, Fig. 11)*
ST	=	*Stark Formation*
TK	=	*Takiyuak Formation*
TO	=	*Tochatwi Formation*
WR	=	*Western River Formation*

(section AB, Fig. 11.13) fit into this gap. The other is that the AB section correlates with the Peacock Hills Formation. More detailed sampling is necessary before paleomagnetic data can provide a firm choice between these alternatives.

Paleomagnetic Poles

Poles deduced from the studies reported here are plotted in Figure 11.14 with those from related formations (see Irving and McGlynn, 1979). Most of them form a group with overlapping 95 per cent uncertainty limits, which supports the general synchroneity between developments in the various elements of the Coronation Geosyncline and the tectono-stratigraphic model of Hoffman et al. (1970). Irving and McGlynn (1979) have summarized apparent polar wander (APW) for North America for the interval 2200 to 1700 Ma, using many of these poles. They argue for an APW curve starting in Alaska, sweeping down the west coast, turning east across Chile, then north to Brazil, and finally west to cross itself in Central America. This latter part they refer to as the Coronation Loop. It represents the best summary currently available, although the loop is not as pronounced as they suggest due to minor plotting errors in their Figure 11.

The results from the Great Slave Supergroup, Goulbourn Group, and Epworth Group provide no compelling evidence for coherent polar movement and we therefore represent them collectively as a single "lower Coronation pole" (Fig. 11.15; 092°W, 14°S, N=8, K=23, A_{95}=12°), composed of the data marked with an asterisk in Table 11.7. The Seton "C" magnetization is not included as it consists of two polarity groups which are not antiparallel. Evans et al. (1980) argued that this reflects the presence of a "Coronation overprint" component, which causes a slight northerly shift in the pole. The effect is not large as can be judged from the proximity of the Seton pole to that of the Akaitcho River Formation, which is laterally equivalent (Fig. 11.14; see also discussion above). The result from the Martin Formation of northern Saskatchewan is included in Figure 11.14 to illustrate that this formation is probably correlative to the Great Slave Supergroup, but it is not used in calculating the summary pole shown in Figure 11.15. Poles from the Stark and Tochatwi formations are also excluded because of the still-unresolved structural complication discussed above.

The APW curve can be connected from this lower Coronation pole, through the Et-Then result (Irving et al., 1972), to the Coronation overprint pole also shown in Figure 11.15. The paleo-equator corresponding to the lower Coronation pole is included in Figure 11.15 to emphasize the fact that the Coronation Geosyncline occupied a tropical position when it formed.

The Western River Formation yields a pole which diverges markedly from the other data (Fig. 11.14). At face value it implies intra-cratonic drift, as has been widely suggested for the much younger Grenville Province. It is important to realize that the Western River Formation is older than any paleomagnetically studied unit in the Athapuscow Aulacogen, and that the early part (pre Seton/Akaitcho River times) of the North American APW curve advocated by Irving and McGlynn (1979) includes no data from Coronation Geosyncline rocks. A merging of separate APW curves, and corresponding parts of the Shield, at about 1800 Ma is therefore not out of the question and does not conflict with any currently available data. Much further work, however, is necessary before this can be regarded as anything but highly speculative. An alternative, and perhaps more attractive, interpretation is simply that a significant time gap exists between deposition of the Western River Formation and those formations which yield the Lower Coronation Pole. If typical Phanerozoic rates of polar motion apply, the 77° between the Western River Formation pole and the Lower Coronation Pole could represent some 200 Ma.

ACKNOWLEDGMENTS

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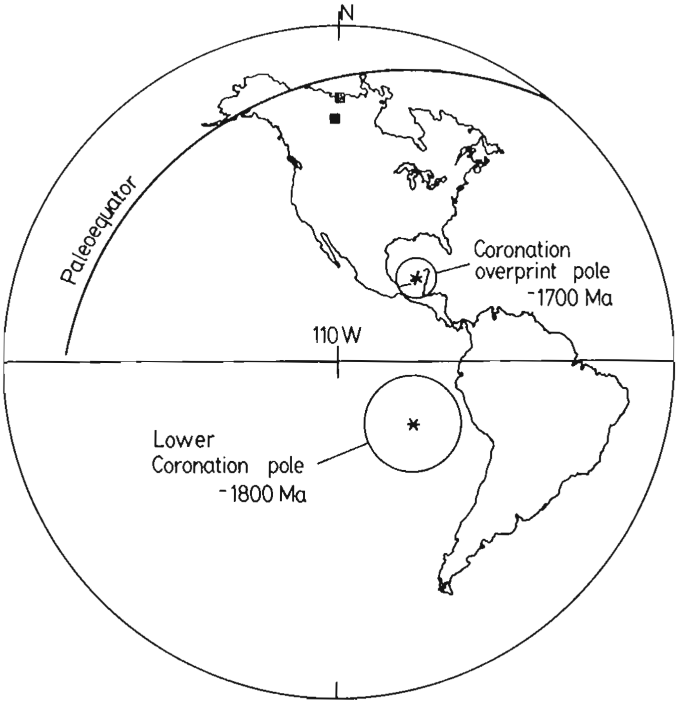

Figure 11.15

The lower Coronation pole and its associated paleoequator, and the Coronation overprint pole.

Table 11.7. Summary of Coronation Geosyncline and related paleomagnetic results

		Directions			Poles		
Formation	N	D,I	k	α_{95}	Long,Lat	K	A_{95}
*Akaitcho River	35	160,+36	13	7°	092W,04S	13	7°
Seton "C"	19	159,+46	24	7°	093W,02N	29	6°
*Kahochella "K"	18	129,+21	12	10°	062W,07S	15	9°
*Douglas Peninsula	6	172,+18	14	19°	102W,17S	19	16°
Stark	39	214,+15	13	7°	148W,15S	9	8°
Tochatwi	8	030,-11	14	15°	143W,18S	24	12°
*Pearson "A"	12	329,-08	21	10°	077W,19S	26	9°
Et-Then	14	294,-21	17	10°	048W,01S	26	8°
Western River	22	086,+25	12	9°	019W,14N	12	9°
*Mara	28	179,+29	12	8°	107W,07S	16	7°
*Peacock Hills	15	342,-12	11	12°	090W,15S	13	11°
*Takiyuak	17	178,+19	15	10°	111W,13S	20	8°
*Peninsula sill	7	345,-02	63	8°	097W,22S	73	7°
Martin	15	333,-29	15	10°	073W,09S	20	9°
Lower Coronation	8	-	-	-	092W,14S	23	12°
Coronation overprint	75	152,+64	30	3°	091W,21N	14	5°

Note: Asterisks indicate formations for which paleomagnetic data are reported in this paper

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12. **THE AMUNDSEN EMBAYMENT, NORTHWEST TERRITORIES; RELEVANCE TO THE UPPER PROTEROZOIC EVOLUTION OF NORTH AMERICA**

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Young, Grant M., The Amundsen Embayment, Northwest Territories; Relevance to the upper Proterozoic evolution of North America; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 203-218, 1981.

Abstract

The Amundsen Embayment comprises late Proterozoic (~1200-~700 Ma) sedimentary and volcanic rocks of the Shaler Group on Victoria and Banks islands, and correlatives in the Brock Inlier and Coppermine area. On Victoria Island the Shaler Group is disconformably overlain by mafic volcanic rocks of the Natkusiak Formation. The sedimentary rocks include limestones and dolostones, some of which are stromatolitic, shales, mudstones, sandstones, and gypsiferous evaporites. The Shaler Group records an initial phase of marine transgression, followed by a succession of marine-to-fluvio-deltaic cycles. A thick carbonate succession was then deposited, under shallow marine conditions. Preservation of two major evaporitic units in the upper part of the Shaler Group indicates basin restriction. An easterly thinning wedge of terrigenous clastics terminates deposition of the Shaler Group. Evaporites, oolites, redbeds, and stromatolites suggest a warm climatic regime. Paleocurrents in fluvio-deltaic units indicate transport to the northwest, but most formations show a random paleocurrent pattern, reflecting the complexity of marine current systems.

Close lithostratigraphic similarity with the Mackenzie Mountains supergroup in the northern part of the Canadian Cordillera suggests that the two regions were formerly parts of one extensive sea. Concentration of evaporites at two stratigraphic levels in both areas indicates deposition in low paleolatitudes, under conditions of restricted circulation. These evaporites suggest that the depositional basin was two-sided. The "other side" of the late Proterozoic sea is not known but might possibly be located in the western Pacific continents. Some of the finer grained terrigenous debris in the Cordilleran trough and the Amundsen Embayment may have been derived from the rising Grenville orogen on the east side of the North American continent.

Résumé

La baie Amundsen comprend des roches sédimentaires du Protérozoïque supérieur (~1200-~700 Ma) et des roches volcaniques du groupe de Shaler sur les îles Victoria et Banks, ainsi que des roches correspondantes dans l'enclave de Brock et la région de Coppermine. Dans l'île Victoria le groupe de Shaler est recouvert en discordance par des roches volcaniques mafiques de la formation de Natkusiak. Les roches sédimentaires sont constituées de calcaires et de dolostones dont certaines sont stromatolitiques, de schistes argileux, de mudstones, de grès et d'évaporites gypsifères. Le groupe de Shaler comprend une phase initiale de transgression marine suivie d'une succession de cycles marins à fluvio-deltaïques. Une succession épaisse carbonatée s'est ensuite déposée dans des conditions marines peu profondes. Une préservation de deux unités évaporitiques dans la partie supérieure du groupe de Shaler indique des étranglements du bassin. Un biseau de roches clastiques terrigènes s'amincissant vers l'est termine la sédimentation du groupe de Shaler. Les évaporites, les oolites, les lits rouges et les stromatolites laissent supposer un climat chaud. Des paléocourants dans les unités fluvio-deltaïques indiquent qu'il y a eu transport vers le nord-ouest, mais la plupart des formations présentent une configuration aléatoire des paléocourants, reflétant la complexité des systèmes des courants marins.

Une similarité lithostratigraphique étroite avec le supergroupe des monts Mackenzie, dans la partie nord de la Cordillère canadienne, laisse supposer que ces deux régions faisaient partie d'une seule mer étendue. La concentration d'évaporites à deux niveaux stratigraphiques de ces deux régions indique que la sédimentation s'est effectuée à des paléolatitudes peu élevées dans des conditions de circulation limitée. Les évaporites laissent supposer que le bassin de sédimentation avait deux côtés. "L'autre côté" de la mer protérozoïque n'est pas connu, mais il serait probablement situé dans les zones continentales du Pacifique ouest. Certains des débris terrigènes, à grains plus fins, du géosynclinal de la Cordillère et de la baie Amundsen, pourraient avoir eu comme origine l'orogenèse du Grenvillien sur le côté est du continent nord-américain.

INTRODUCTION

This paper begins with a brief review of previous work on the formations that comprise the lower part of the Shaler Group. Since there is little published work on the succeeding, less well exposed rocks, a more detailed account is given of the upper three formations. These data are used to provide an overview of the stratigraphy and sedimentary history of the region. Available paleocurrent data and results of radiometric age determinations suggest that deposition of these rocks and correlatives in the northern part of the Canadian Cordillera was contemporaneous with uplift in the region now occupied by the Grenville Province.

PREVIOUS WORK

The foundations of the stratigraphy and structure of the Proterozoic rocks of the Shaler Mountains were laid by Thorsteinsson and Tozer (1962). Subsequent studies of individual units were made by Young (1974), Young and Long (1977a, b), Young and Jefferson (1975), and Young (1977). Miall (1976) reported on late Proterozoic rocks at the south end of Banks Island. Mapping and stratigraphic studies of the Brock Inlier were carried out by Cook and Aitken (1969) and Balkwill and Yorath (1970). Baragar and Donaldson (1973) mapped the Coppermine area and Dixon (1979) proposed some revisions of the Proterozoic stratigraphy of that region. Baragar (1976) gave a preliminary report on studies of the volcanic and intrusive rocks on Victoria Island. A generalized stratigraphic column for the Shaler Group on Victoria Island is shown in Figure 12.1.

Regional correlation with Proterozoic rocks of the Mackenzie Mountains region of the Cordillera (Fig. 12.2) has been suggested by Aitken et al. (1973), Young (1977) and Young et al. (1979).

PROTEROZOIC INLIERS OF THE AMUNDSEN EMBAYMENT

In central Victoria Island Proterozoic rocks occupy the northeast-trending spine known as the Shaler Mountains. This highland area is known as the Minto Arch. Elevations are up to about 700 m. Structurally the Minto Arch is relatively simple, comprising an open syncline (Holman Island syncline) and a smaller anticline to the west (Walker Bay anticline). The precise age of folding is not known but Christie et al. (1972) suggested that it might have taken place in latest Proterozoic time. Northeast- and northwest-trending faults are common in many parts of the Shaler Mountains.

In the southern part of Victoria Island Proterozoic rocks are also present in two smaller areas known as the Wellington and Duke of York inliers (Fig. 12.2). The late Proterozoic rocks of the Richmond Inlier (Dixon, 1979) may be traced southward and westward via the Jameson Islands (Campbell, 1978) and Couper Islands (Baragar and Donaldson, 1973) to the Coppermine area, where equivalent rocks are known as the Rae Group. As discussed by Dixon (1979) there are problems concerning the age of the uppermost units assigned by Baragar and Donaldson (1973) to the Rae Group. The western part of the Amundsen Embayment also includes the Brock Inlier and the southern tip of Banks Island.

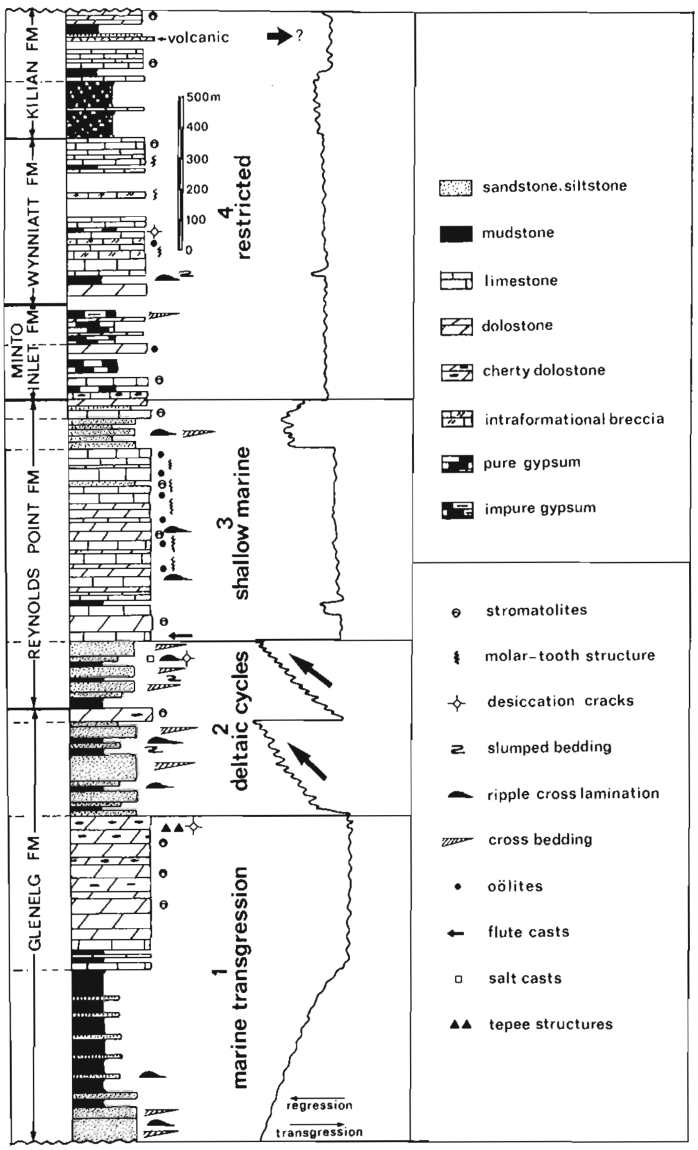


Figure 12.1. Generalized stratigraphic succession of the Shaler Group on Victoria Island. The formational names of Thorsteinsson and Tozer (1962) are retained, but several other mappable subdivisions are also indicated, by dashed lines at left side of diagram. Wavy line to right of stratigraphic column is an interpretation of depositional history in terms of interplay between clastic influx and marine transgression. The depositional history has been divided into four phases as discussed in the text. Arrows at phase 2 (deltaic cycles) indicate northwesterly transport of terrigenous clastic material. Small arrow at the top of phase 4 (restricted conditions) indicates possible westerly provenance of clastics at the top of the Kilian Formation.

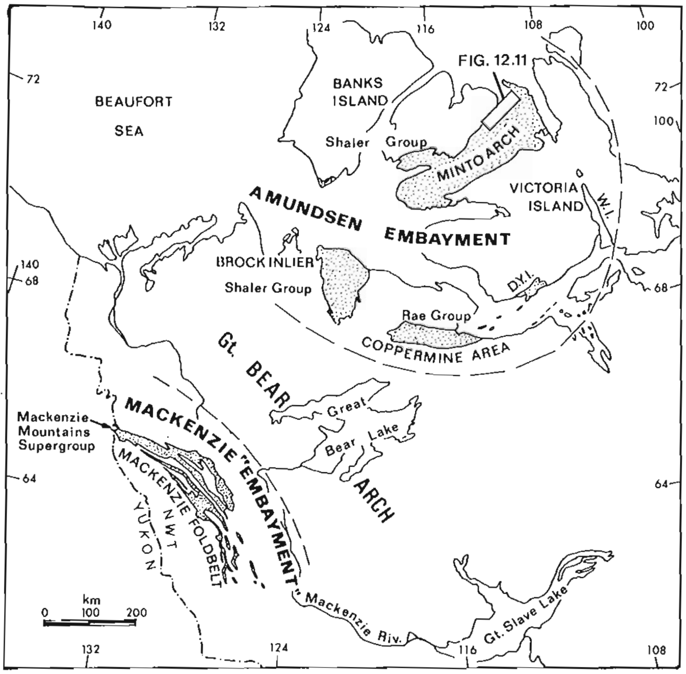


Figure 12.2. Sketch map to show the location of the Amundsen Embayment and Mackenzie Foldbelt in northwest Canada. Major areas of Shaler Group and Mackenzie Mountains Supergroup are indicated by the dotted ornament. D.Y.I. is the Duke of York Inlier; W.I. is the Wellington Inlier. Note the location of Figure 12.11.

AGE OF THE PROTEROZOIC ROCKS OF THE AMUNDSEN EMBAYMENT

At the northeastern end of the Minto Arch in Victoria Island there is an unconformity between quartzites, apparently intruded by granites, and the basal unit of the Shaler Group. An Archean age was reported by Thorsteinsson and Tozer (1962) from the granite, thus providing a lower limit for the age of the Shaler Group of that region. More recently an age of about 1.67 Ga (K-Ar) was obtained (W.A. Gibbins in Campbell and Cecile, 1979) from a granite in the Wellington Inlier. The youngest dates obtained from rocks unconformably below the Shaler Group and equivalents come from lavas of the Coppermine River Group which have yielded Rb-Sr isochron ages of about 1.2 Ga (Wanless and Loveridge, 1972). In the Brock Inlier and on Banks Island the basement to the Shaler Group is not seen.

On Victoria Island and elsewhere in the Amundsen Embayment an upper limit on the age of the Shaler Group is provided by the presence of ubiquitous sills and dykes that have provided K-Ar dates as old as 700 Ma. Lavas of the Natkusiak Formation disconformably overlie the Shaler Group in the central part of the Minto Arch. These volcanic rocks are considered to be co-magmatic with the sills and dykes; all these igneous rocks form part of the Franklin mafic igneous episode.

STRATIGRAPHY AND SEDIMENTATION OF THE SHALER GROUP

The thickest and most representative succession of the Shaler Group is present on Victoria Island (see Table 12.1, Fig. 12.1). The maximum thickness of the Shaler Group is estimated to be about 4000 m but many formations show marked thinning towards the east and northeast. Thorsteinsson and Tozer (1962) divided the rocks of the Minto Arch into five formations. Their scheme is retained in this paper but recent work has shown that some of these formations can be further subdivided (see Fig. 12.1) and that the subdivisions are of regional extent.

Glenelg Formation

The Glenelg Formation is about 1400 m thick and can be divided into four units (Table 12.1). The lowest unit is the least well known, but a recently published section (Dixon, 1979) shows it to be shale, siltstone and sandstone. The predominant rock type is grey and green shale; spherical calcareous concretions are present in some layers. This basal clastic unit is overlain by a thick grey dolomitic member which is cherty in its upper half.

On the south tip of Banks Island a shale and siltstone unit is present above the cherty dolostone (Young and Jefferson, 1975). Neither of the two lowest members of the Glenelg Formation has been the subject of a detailed sedimentological study but Young and Jefferson (1975) reported tepee structures, desiccation cracks and other sedimentary features typical of a supratidal environment from the upper part of the grey cherty dolostone member. Some of the chert occurs as laminae and may be primary. In other places chert formed by an early diagenetic (pre-compaction) process. Detailed descriptions of stromatolites from this unit are given in Jefferson (1977). Excellent preservation of microfossils (Jefferson, 1977) also suggests an early origin for the chert.

On the south tip of Banks Island a shale and siltstone unit occurs above the cherty dolostone (Young and Jefferson, 1975). These rocks may be a deeper water equivalent of the sandstones that occupy a similar stratigraphic position in the eastern part of the Minto Arch. Together these units represent a fluvio-deltaic complex that prograded in a northwesterly direction. A much thicker sandy fluvial facies occurs in this member on Banks Island and in the Brock Inlier (Young, 1977; Miall, 1976) than elsewhere, and may indicate proximity to a major Proterozoic river distributary system. In some areas the shales, siltstones and sandstones of this member are arranged in coarsening-upward sequences that might represent prograding deltas or possibly shallow marine tidal bars (Young, 1974). These sequences commonly contain slump folded shales and sandstones indicative of slope instability and/or rapid sedimentation followed by dewatering. Paleocurrents from Brock Inlier, Banks Island, and Victoria Island suggest northwesterly transport of these clastic sediments (Fig. 12.1).

The topmost unit of the Glenelg Formation is a widespread buff, grey and purple stromatolitic biostrome that characteristically weathers orange. A detailed description of these stromatolites was given by Young and Long (1976) and Jefferson (1977). Many elongate large scale domes and individual stromatolitic columns have a preferred northeast-southwest orientation, possibly normal to a migrating shoreline. It has been suggested (Young, 1979) that this stromatolitic unit extends as far as the Mackenzie Mountains region (Fig. 12.2) where it may be represented by the middle K6 unit (Aitken et al., 1978a) of the Mackenzie Mountains Supergroup.

Summary of Depositional History of the Glenelg Formation

The basal clastic unit of the Glenelg Formation records an influx of clastic debris, at first sandy, then later dominantly fine grained. The coarse to fine upward transition suggests a transgressive depositional setting. Subsequent deposition of the shallow water-to-emergent cherty dolostones indicates a virtual cessation of detrital influx and carbonate build-up in a shallow sea.

The overlying clastic unit shows some lateral variability. In Banks Island and Brock Inlier it consists mainly of fluvial sandstones above a thin deltaic sequence. In northeastern Victoria Island it is mainly a series of coarsening-upward deltaic(?) cycles. These relationships could indicate the presence of a particularly active fluvial regime in the western part of the Amundsen Embayment at that time.

Table 12.1. Table of formations for the Shaler Group on Victoria Island with generalized description and interpretation. Note that the Natkusiak Formation is excluded from the Shaler Group.

Formation	Subdivisions	Lithology	Colours	Special Features	Paleocurrent directions	Environments
Natkusiak (800 m)		mafic lavas agglomerate at base	grey, green	vesicular tops and bases to flows	--	mostly subaerial; subaqueous at base
			disconformable contact			
Kilian (400-500 m)	Upper – non-evaporitic	dolostone, mudstone, limestone, siltstone, basaltic lava near base; to SW may include sandstones	buff, purple, grey	stromatolites desiccation cracks cross lamination	random	shallow marine
	Lower – evaporitic	gypsiferous mudstone, limestone, mudstone, gypsum	purple, red, green, grey	chicken wire texture desiccation cracks	random	shallow marine to emergent
Wynniatt (450-900 m)		limestone, shale, dolostone, gypsum	grey, black, buff	grainstones, oncolites stromatolites, flat chip conglomerate desiccation cracks wavy bedding molar-tooth	random	marine, shallow to moderate water depth
Minto Inlet (300-400 m)	Upper-finely bedded	impure gypsum gypsum dolostone	grey, white, purple	chicken wire texture cross lamination desiccation cracks chert nodules	random	shallow marine – emergent restricted circulation sabkhas
	Lower-coarsely bedded	dolostone, impure gypsum, gypsum	grey, white, purple	flat chip conglomerate grainstones, chert nodules, stromatolites	random	shallow marine – restricted circulation
Reynolds Point (1000 m)	Upper-carbonate	limestone, sandstone	buff, grey, purple	stromatolites silicification	random	shallow marine
	Upper-clastic	sandstone, sandy limestone	grey, buff, white	cross bedding, cross-lamination, desiccation cracks	random	intertidal
	Lower-carbonate	limestone, sandstone, shale	grey, buff, purple	grainstones stromatolites molar-tooth	random	shallow marine, tidal
	Lower-clastic	sandstone, siltstone, mudstone	buff, pink, purple green	cross bedding, ripple marks, salt casts desiccation cracks	NW	marine deltaic complex
Glenelg (1500 m)	Stromatolitic unit	dolostone	grey, buff, orange	stromatolites, domal structures, elongate stromatolite, domes and columns	NE-SW	shallow marine – tidal influence
	Upper clastic	sandstone, siltstone, mudstone	pink, buff, purple, grey, green	cross bedding dewatering pipes ripple marks, current lineation, slumps	NW	fluvial, deltaic, shallow marine
	Cherty dolostone	dolostone, limestone	grey, buff	Tepee structures stromatolites flat chip conglomerate chert	--	tidal, supratidal
	Lower clastic	shale, siltstone, sandstone	grey, green, black	laminations, concretions, cross lamination	–	shallow marine?
			unconformity			
Youngest known basement rocks are granites in the Wellington Inlier which have given a K-Ar age of 1673 ± 42 Ma (W. Gibbons, in Campbell and Cecile, 1979)						

At the southern tip of Banks Island a siltstone-mudstone unit that underlies the main fluvial clastic unit is interpreted as a deeper water equivalent of the fluvial-deltaic complex to the east. A much thicker equivalent of this unit may be present in the Mackenzie Mountains Supergroup – the Tsezotene Formation of Gabrielse et al. (1973) and Aitken et al. (1973).

Deposition of the Glenelg Formation concluded with a transgressive episode during which a widespread stromatolite biostrome (largely composed of the form **Inzeria**) was deposited.

Reynolds Point Formation

Deposition of the Reynolds Point Formation (about 1000 m thick) began with progradation of a deltaic complex (lower clastic member, Table 12.1) over the widespread biostrome of **Inzeria.** Distal mudstones, which buried the stromatolites, were succeeded by siltstone, mudstones, and sandstones interpreted by Young and Long (1977a) as a marine deltaic complex. Paleocurrents in sandstones of this unit indicate a westerly progradation. In the Amundsen Embayment this unit is known only from Victoria Island and the Brock Inlier.

The uppermost sandstones of this deltaic complex are succeeded by a thick succession of carbonates. There are two carbonate-dominated members in the Reynolds Point Formation (Table 12.1). The lower member was divided by Young and Long (1977b) into three units. The lowest of these contains, at its base, evidence of basinal subsidence in the form of slumped beds and flute casts (Young and Long, 1977b). It includes some thick (up to 35 m) stromatolitic biostromes. Unit 2 consists of shaly nodular dolostones and limestones with a thin unit of purple mudstone at its base. Unit 3 is the thickest and consists of oolitic, stromatolitic and shaly carbonates. The stromatolite **Baicalia** is present in this unit (Jefferson, 1977; Aitken et al., 1978a; Young and Jefferson, 1975). Molar-tooth structure (Smith, 1968) is common in the shaly carbonates.

The upper clastic member consists mainly of fine grained quartzarenites with abundant crossbedding. The member shows a facies change from fine sandstones in the northeastern part of the Minto Arch to sandy limestones in the Minto Inlet area (Young and Jefferson, 1975). Individual sandstone units show bimodal-bipolar paleocurrent trends but measurement of about 900 crossbeds from this unit in the southwestern part of the Minto Arch (Young and Jefferson, 1975, p. 1740) showed extremely variable transport directions interpreted as the result of complex interaction of multiple current systems in a shallow shelf sea.

The overlying upper carbonate member includes dolomitic and siliceous carbonates. A characteristic and widespread form of stromatolite, **Acaciella** (Jefferson, 1977) is present. These rocks are transitional upward into the evaporites and purple mudstones of the Minto Inlet Formation.

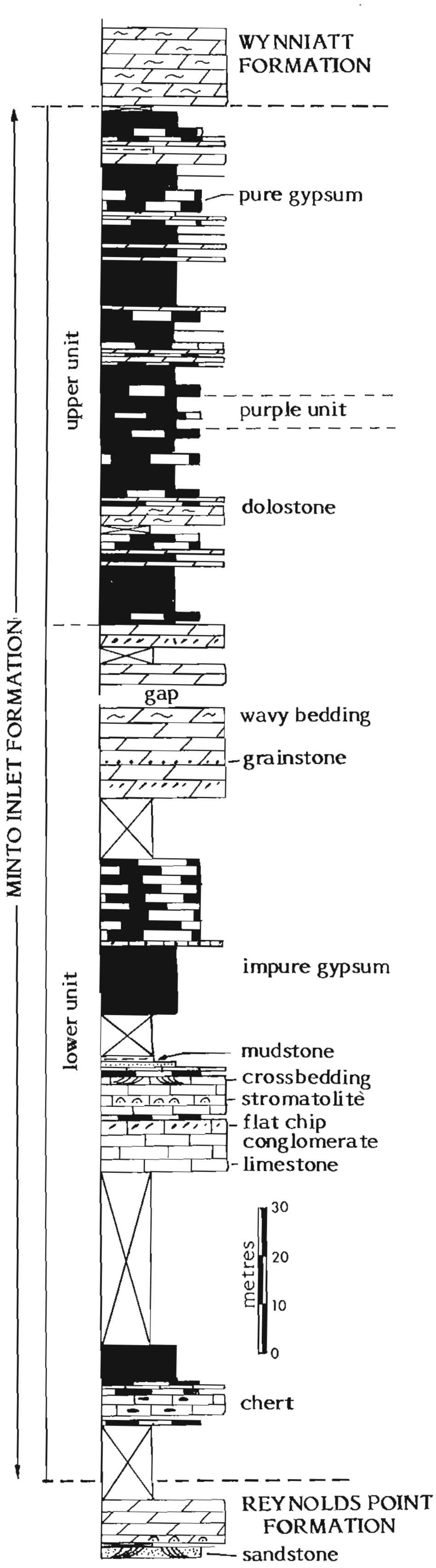


Figure 12.3

Generalized stratigraphic section of the Minto Inlet Formation in the northeastern part of the Minto Arch. This section was compiled from measured sections at F, E, D, C, and G (Fig. 12.11). For more details of the upper part of the Minto Inlet Formation see Figure 12.9.

Summary of Depositional History of the Reynolds Point Formation

Deposition of the Reynolds Point Formation began with a regressional clastic marine deltaic sequence. Basin subsidence and a decrease in supply of terrigenous clastic material ushered in a period of carbonate deposition in a dominantly shallow and agitated marine environment. The upper, fine grained, clastic member formed by progradation of a clastic tidal environment over a carbonate shelf sea. This was followed by a short-lived transgression during which more shallow water dolostones and limestones (some stromatolitic) were deposited. Soon after this transgression the region was subjected to restricted, evaporitic conditions, giving rise to deposition of gypsiferous deposits of the Minto Inlet Formation.

Minto Inlet Formation

The Minto Inlet Formation is the first unit of the Shaler Group to contain substantial amounts of evaporitic minerals. This unit is treated in more detail than the previous ones because there are no published detailed descriptions. The following description is based on sections measured in the Wynniatt Bay region at the north end of the Minto Arch.

The measured thickness of the Minto Inlet Formation is just over 300 m. The map of Thorsteinsson and Tozer (1962) clearly shows a southwesterly thickening of the formation. A thickness of about 400 m was estimated by Thorsteinsson and Tozer (1962) at the southwestern end of the Minto Arch. The basal contact was not seen in the section shown in Figure 12.3 where crossbedded sandstones and cherty stromatolitic dolostones are separated by about 15 m of cover from gypsiferous rocks assigned to the Minto Inlet Formation. These evaporites are overlain by a grey limestone with a minimum thickness of about 20 m. The overlying gypsum-bearing units (up to 13 m thick) are the thickest in the Minto Inlet Formation of this area. A thick dolostone unit (Fig. 12.3) is succeded by a succession composed of interbedded dolostone, impure silty and shaly gypsum beds and clean white gypsum layers. There is marked contrast in the scale of bedding in the lower and upper halves of the formation; the various rock types are interlayered on a fine scale in the upper portion, whereas similar lithologies occur in much thicker units in the lower half.

The Minto Inlet carbonates are shallow water deposits characterized by flat chip conglomerates, some crossbedding, grainstones and rare flat, bun-shaped stromatolites. Some of the shaly beds are red or purple; some display desiccation cracks. The evaporitic units are white, grey, and pink gypsum with minor anhydrite. Many beds have dark grey gypsum rosettes in a finely crystalline matrix of the same material. The evaporites display a variety of textures and structures. Those in the lower part of the formation are mainly fine- to medium-bedded gypsum with minor thin grey

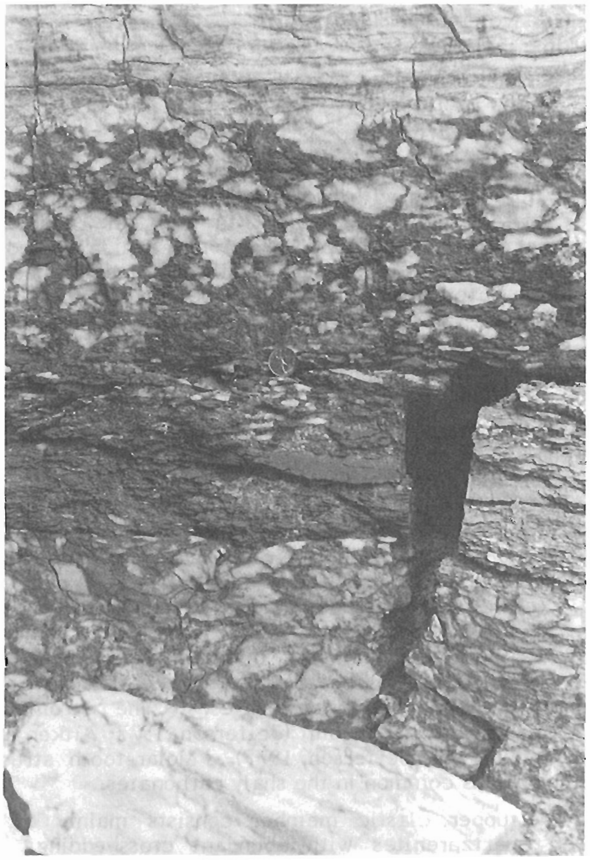


Figure 12.5. *"Chicken wire" texture in evaporites of the upper Minto Inlet Formation. This texture has been attributed to diagenetic growth of gypsum in mud. The "chicken wire" appearance is due to the presence of a thin film of mudstone between gypsum nodules.*

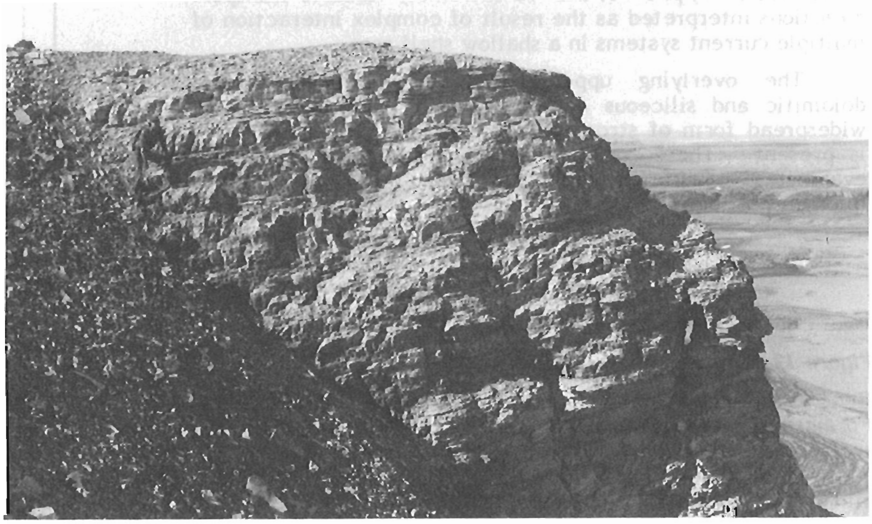


Figure 12.4

Thick evaporitic unit in the lower part of the Minto Inlet Formation, northeast Minto Arch.

clay-rich laminae (Fig. 12.4). The gypsiferous units are thick and are interbedded with equally thick units of dolostone and impure gypsum (Fig. 12.3). These evaporite beds probably formed by precipitation from a standing body of water.

The thinner evaporite beds of the upper half of the formation commonly display "chicken wire" texture (Fig. 12.5), perhaps suggesting an origin by intrastratal early diagenetic deposition of gypsum in a supratidal environment (Kerr and Thompson, 1963; Butler, 1969). Some enterolithic folding is also present (Fig. 12.6). Some evaporite beds include dark grey rounded dolostone bodies a few centimetres across. These are probably intraclasts ripped up from underlying dolostone layers. Crossbedding (Fig. 12.7), possible oolitic gypsum, and intraclastic gypsiferous beds are also present. All these features point to a shallow water, current-dominated depositional environment. Shallow-water to emergent conditions are indicated by the presence of red siltstones with desiccation cracks.

Dark grey to black chert bodies of similar dimensions may have formed in the gypsum layers by diagenetic replacement. Chert is commonly associated with Precambrian evaporites (Wood, 1973; Young and Jefferson, 1975), possibly reflecting greater concentrations of silica in marine waters due to the absence of silica-secreting organisms.

In some gypsum beds, small vertical dyke-like bodies of both dolostone and chert are present. The origin of these structures is not understood but they may represent fillings (chemical or mechanical in the case of dolostone or chemical in the case of the chert) of desiccation cracks or solution channels in the gypsum beds.

A facies analysis of the type described by Cant and Walker (1976) was carried out to examine the upper part of the formation for possible cyclicity. By analysis of 179 transitions from one rock type to another, a cycle involving upward transition from dolostone through impure gypsum to pure gypsum to impure gypsum and finally to dolostone was detected (Fig. 12.8). This cycle indicates fluctuating salinity changes possibly related to fluctuating sea level. Many individual beds or groups of beds can be traced over distances of up to 5 km as shown in Figure 12.9.

Figure 12.6

Enterolithic folds in gypsum layers. Such folds may be caused by volume increases related to the change from anhydrite to gypsum.

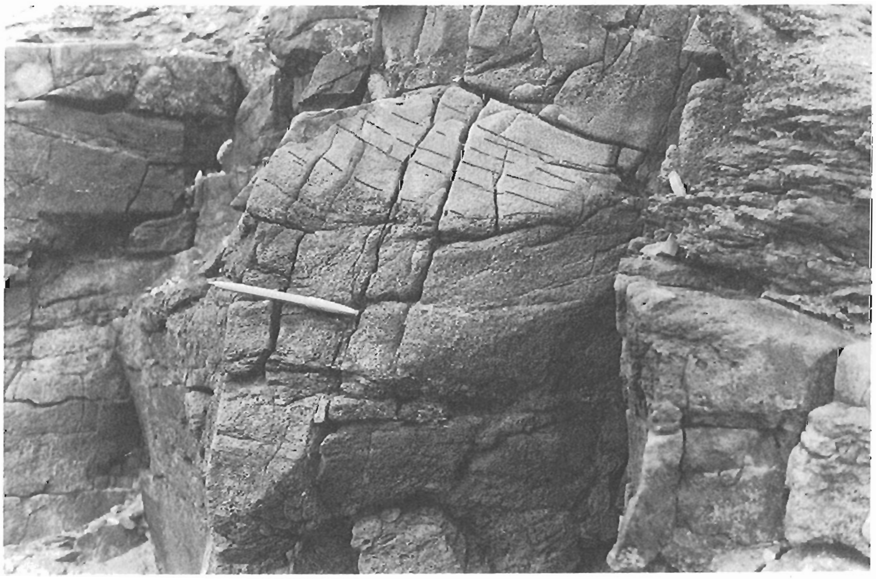
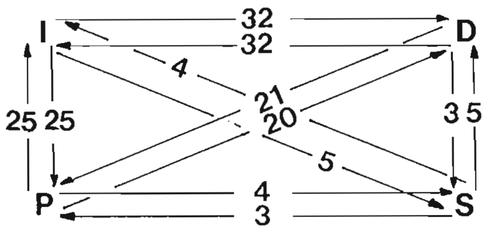


Figure 12.7

Crossbedding (enhanced by use of a marker pen on the outcrop) in evaporites of the upper Minto Inlet Formation, northeast Minto Arch.



	D	P	I	S	
D		5·7 (21)	12·9 (32)	-0·8 (3)	56
P	4·4 (20)		8·3 (25)	0·7 (4)	49
I	12·3 (32)	8·0 (25)		0·8 (5)	62
S	1·2 (5)	-0·3 (3)	-0·1 (4)		12
	57	49	61	12	

D → I → P → I // D → I → P etc.

Summary of Depositional History of the Minto Inlet Formation

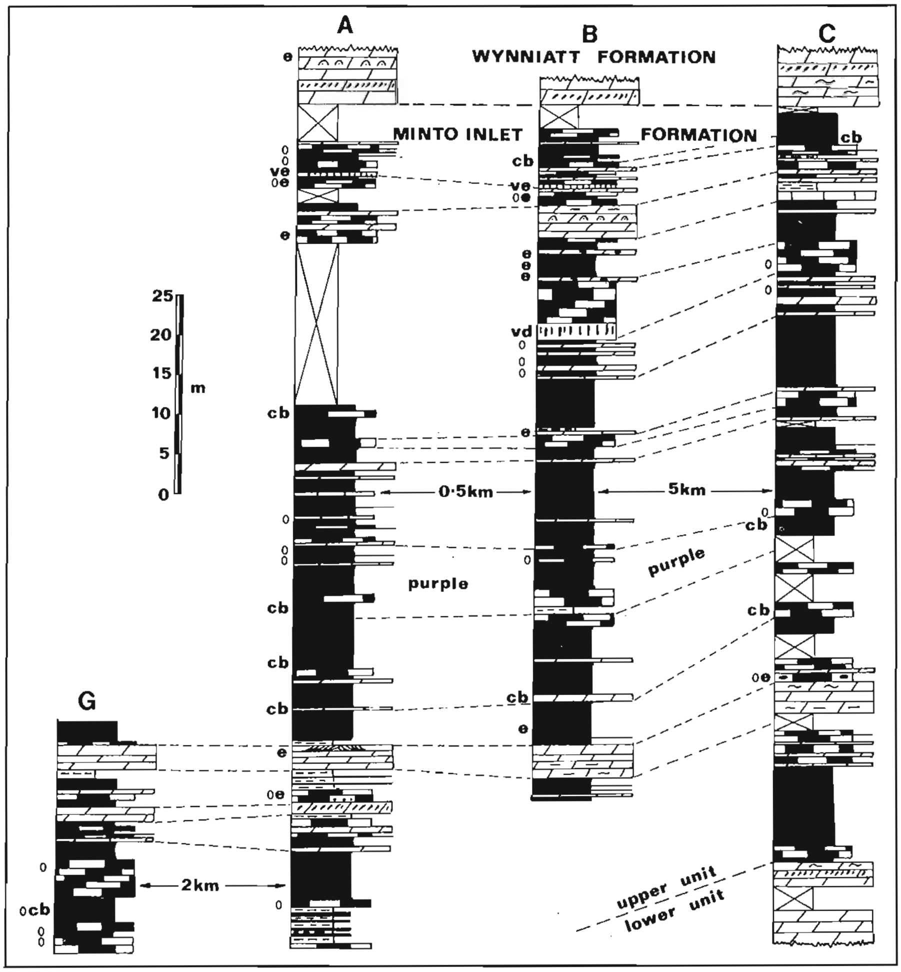
The Minto Inlet Formation consists of beds of gypsum, impure gypsum, red, commonly gypsiferous siltstones, and grey dolostones. In the lower half of the formation bedding of different rock types is on a scale of metres to tens of metres, whereas the upper part displays a cyclical repetition on a much finer scale. These cycles may be due to variations in salinity and clastic influx into the basin, but the exact mechanism is not known. Episodic relative sea level changes could have been caused by sporadic basin subsidence or eustatic sea level changes or possibly related to a feedback mechanism such as that proposed by Bosellini and Hardie (1973). Measurement of small scale crossbeds from grainstones in the upper part of the Minto Inlet Formation showed no preferred orientation.

Wynniatt Formation

No previous detailed descriptions of this unit have been published. A composite measured section about 550 m thick is given in Figure 12.10. The basal contact was not seen; in section C, Figure 12.9, grey carbonates of the lower Wynniatt Formation outcrop about 1 m stratigraphically above the topmost evaporitic unit of the Minto Inlet Formation. Most of the formation is composed of grey micritic shaly limestone but a much more dolomitic section (section G, Fig. 12.11) was noted about 50 km to the northeast of that illustrated in Figure 12.10.

◄ ***Figure 12.8***

Results of numerical analysis of bed transitions in the upper part of the Minto Inlet Formation. Top part of the diagram shows the number of observed upward transitions among P (pure gypsum beds), I (impure, muddy gypsum beds), S (silty mudstones), and D (grey dolostones). Small figures at bottom right of squares are total numbers of transitions from rock type in row to those listed in columns. Numbers at right side of diagram are row totals; those at base are column totals. Large numbers in centre of squares are the difference between observed and expected frequencies of a given transition. Those underlined are of greater magnitude and therefore considered the most significant. Inferred cycle is shown at the base of the diagram.



◄ ***Figure 12.9***

Schematic representation of measured sections of the upper part of the Minto Inlet Formation at the northeast end of the Minto Arch. Lithostratigraphic correlation of thin beds over distances of about 6 km is possible. Symbols are the same as those used in Figure 12.1 with the following additions:

O	–	*"chicken wire" texture*
cb	–	*contorted bedding*
e	–	*chert*
ve	–	*vertical bodies of chert*
vd	–	*vertical bodies of dolostone*

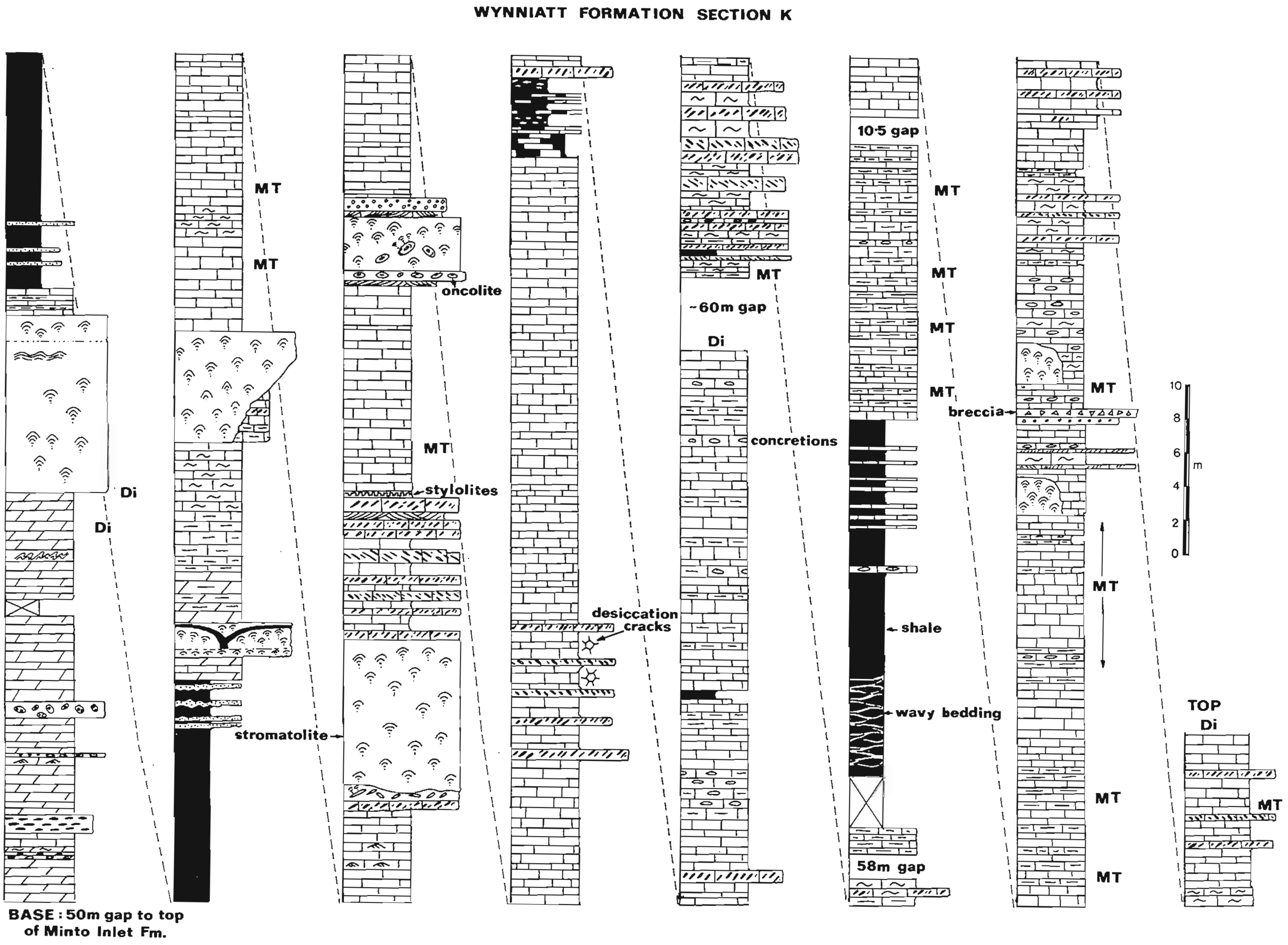


Figure 12.10. Composite simplified stratigraphic section through the Wynniatt Formation in the northeastern part of the Minto Arch. Symbols are as for Figure 12.1 with additions as follows: Di – diabase sill; MT – "molar-tooth" structure. Location of section K is shown in Figure 12.11.

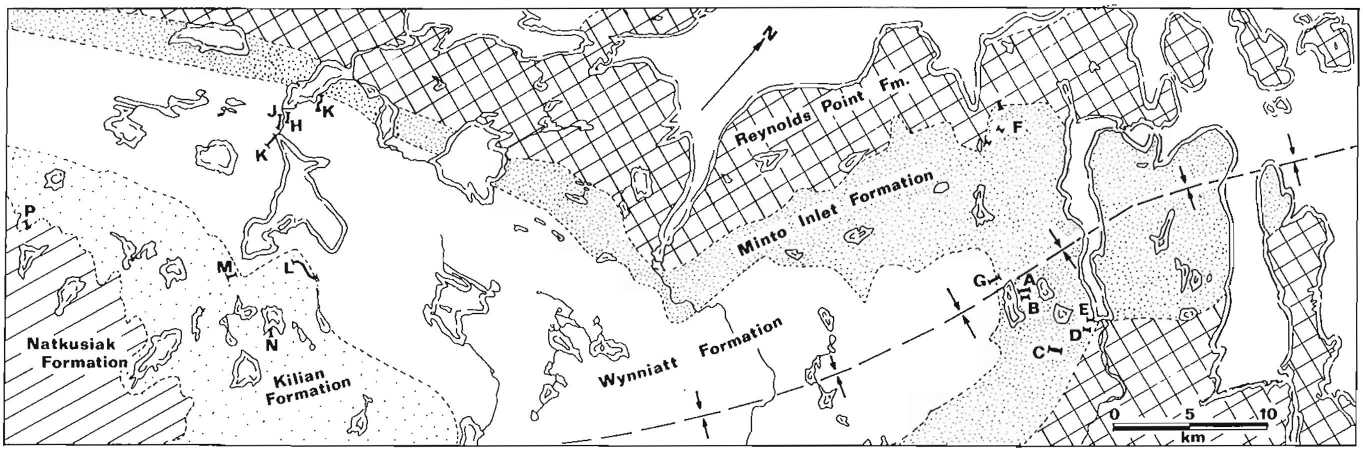


Figure 12.11. *Sketch map of the geology of the northern part of the Minto Arch (see Fig. 12.2 for location). Geology after Thorsteinsson and Tozer (1962). Letters and bars show locations of sections discussed in text.*

Near the base of the formation a thick (10 m) stromatolitic biostrome is overlain by about 28 m of rusty-weathering shale with some thin crossbedded siltstones and sandstones. The coarser clastics are best developed at the top and bottom of the shaly interval. These sandstones are crossbedded and wavy bedded; many beds have slumped upper and lower surfaces. In some cases small sandstone dykes have been injected both upward and downward from sandstone beds. About 50 km to the northeast of section K, the shale-sandstone unit is only about 6 m thick and contains a much higher proportion of sandstone.

Above the shale-sandstone unit there are several stromatolitic biostromes up to 7 m thick. Some of these are tentatively correlated for about 50 km in the northern part of the Minto Arch. They are interbedded with wavy bedded to nodular grey fine grained limestones that contain desiccation cracks and molar-tooth structures (Smith, 1968). The topmost stromatolitic biostrome is locally underlain by about 1 m of coarsely bedded grainstone. In one area of section J (Fig. 12.11) rectilinear shrinkage cracks (Fig. 12.12) are present beneath the grainstone. The biostrome includes a spectacular development of large (ca. 20 cm across) oncolites that are oval in vertical section (Fig. 12.13). In one exposure, columnar stromatolites are nucleated on a large oncolite.

The overlying shales and dolostones contain abundant desiccation cracks and molar-tooth structures, although the two structures appear to be mutually exclusive. There is a cyclic repetition of grey dolomicrite with wavy bedding, overlain by similar rocks with abundant molar-tooth structure and capped by a unit with desiccation cracks. These cycles are from about 9 to 12 m thick and are interpreted as shoaling-upward cycles. They are particularly well developed in section J (Fig. 12.11).

The upper half of the formation contains few stromatolites, but a few small bioherms are present near the top. Nodular to wavy bedded grey fine grained, commonly shaly, limestones are the most common lithology. Some shales have well developed ellipsoidal calcite concretions up to about 20 cm across. Compaction of the surrounding shale indicates that these concretions formed by early diagenetic processes. Many of the carbonate beds in the upper part of the formation contain flat chip conglomerates formed by penecontemporaneous erosion of the thin bedded muddy carbonates. Some fine examples of molar-tooth structures are also present in these beds (Fig. 12.14); an early diagenetic origin is clearly indicated by strong compactional effects on the enclosing layers.

Near the top of the measured section of the Wynniatt Formation (Section K, Fig. 12.11) there is a black to dark grey carbonaceous shale and shaly limestone with a thickness of over 15 m. This unit has extremely fine wavy lamination and has some wavy to nodular calcite-rich layers.

The remainder of the formation is wavy bedded grey limestones with some flat chip conglomerates and molar-tooth structures. Two thin ($\sim$1 m) gypsiferous layers were also noted in the formation (Fig. 12.10).

Summary of Depositional History of the Wynniatt Formation

The Wynniatt Formation records a change from the restricted evaporitic environment of the Minto Inlet Formation to shallow marine conditions of generally normal salinity. Most of the formation is composed of wavy bedded to planar-laminated grey micrite and dolostone. Influx of fine grained terrigenous material resulted in deposition of the dark shaly unit near the base. A second shale unit near the top is carbon-rich, possibly reflecting preservation of organic matter under locally reducing conditions. Measurement of crossbeds in sandstones of the lower (rusty) shale units gave a very scattered distribution with modes in the northeast, southwest and southeast quadrants. Such highly varied paleocurrent directions are typical of shallow marine depositional settings.

Stromatolites are abundant in the lower part of the formation. These rocks also show evidence of vigorous current activity in the form of oolites, flat chip conglomerates and large scale crossbeds. Shallow water conditions are indicated by abundant desiccation cracks. Occasional high salinity is recorded in rare gypsiferous layers. Wavy bedding, concretions, and molar-tooth structures point to early diagenetic migration of $CaCO_3$.

Kilian Formation

The Kilian Formation is about 400 m thick. In the northeastern part of the Minto Arch it can be divided into two units. The lower unit consists mainly of purple, red, green and grey gypsiferous mudstones with subordinate amount of limestone, mudstone and pure gypsum beds. All these rocks are fine grained. A few evaporitic units display "chicken wire" texture. No cyclic development comparable to the upper Minto Inlet Formation was noted. A generalized stratigraphic section is shown in Figure 12.15.

Table 12.2. Chemical analyses of samples from the Natkusiak Formation (GY 77-214, 215, 218) and lavas in the middle part of the upper Kilian Formation. See Figures 12.1, 12.11, 12.15 for stratigraphic position and geographic location of these volcanic rocks. These samples from the two units are chemically very similar. Total iron is reported as FeO. The contribution of total iron to the sume is calculated as Fe_2O_3. Analyses by X-Ray Assay Laboratories, Toronto.

Sample No.	SiO_2	Al_2O_3	CaO	MgO	Na_2O	K_2O	FeO	MnO	TiO_2	P_2O_5	L.O.I.	Total
GY 77 214	49.9	13.5	9.96	7.11	2.12	0.33	12.9	0.18	1.49	0.13	2.12	101.2
GY 77 215	47.4	13.3	7.12	8.93	3.54	0.40	13.4	0.17	1.52	0.12	4.22	101.6
GY 77 218	59.5	13.4	11.0	7.96	2.09	0.40	11.2	0.16	1.25	0.11	1.78	100.0
GY 78 400	49.6	14.1	11.1	7.63	1.83	0.41	11.1	0.20	1.16	0.09	1.58	100.1
GY 78 401	48.5	13.5	10.4	7.12	1.83	0.39	13.3	0.16	1.16	0.09	1.93	99.9
GY 78 402	49.3	14.2	10.9	7.73	1.75	0.40	10.4	0.17	1.17	0.09	1.83	99.0

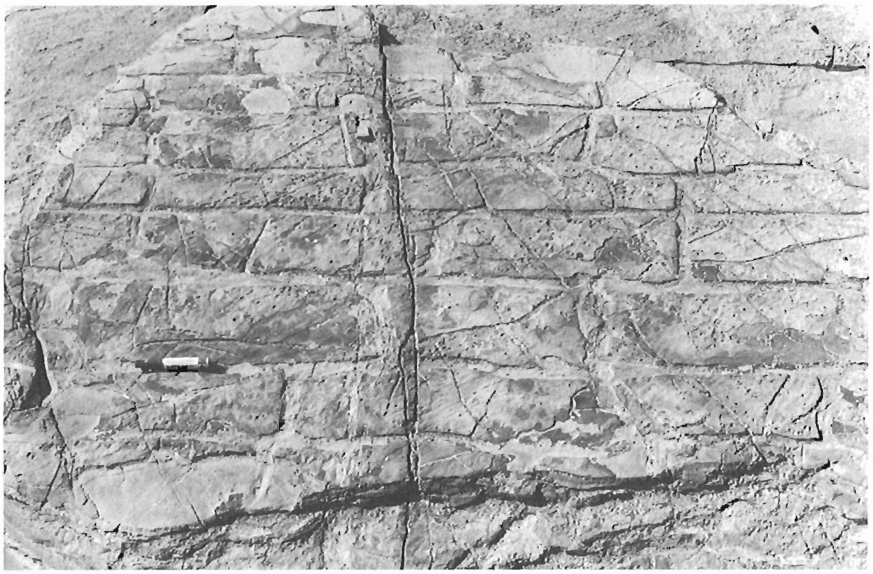


Figure 12.12

Rectilinear desiccation(?) cracks in the lower part of the Wynniatt Formation of section J, Figure 12.11. These structures may reflect down-slope movement of sediment. Note presence of smaller irregular cracks between the straight ones.

The upper part of the Kilian Formation consists mostly of dolostone, mudstone (in part red), and limestone. Some stromatolites are present. The upper portion of the formation is well exposed in a section measured at the north end of the Minto Arch (Section P, Fig. 12.11). The section begins with about 3 m of amygdaloidal basalt, overlain by about 2 m of green volcanic breccia with fragments up to about 3 cm across. These volcanic rocks also occur a few kilometres to the northeast.

The volcanic rocks are overlain by about 80 m of sedimentary rocks. At the base there are cherty dolostones, purple and grey shales and fine grained dolostones. Abundant desiccation cracks indicate shallow-water to emergent depositional conditions. Flat chip conglomerates (some silicified) are also present. The upper part of the section is composed of dolostone and grey to purple shale and mudstone. One thin stromatolitic unit was noted near the top.

Samples of the volcanic rocks were analyzed for comparison with those of the overlying Natkusiak Formation (see Table 12.2). These analyses indicate close chemical similarity between the volcanic rocks in the middle part of the upper Kilian Formation and those of the overlying Natkusiak Formation. If these similarities indicate that the two sets of volcanic rocks were co-magmatic then the disconformity between the Kilian and Natkusiak formations may not have been of long duration and deposition of the youngest rocks of the Shaler Group may have taken place about 700 Ma ago. Paleomagnetic studies currently being carried out by H.C. Palmer and W. Morris should help to clarify these relationships.

Summary of Depositional History of the Kilian Formation

The lower part of the formation indicates a return to restricted evaporitic conditions similar to those under which the Minto Inlet Formation was deposited. There was, however, considerable influx of fine grained muddy material during lower Kilian deposition so that few beds of pure gypsum are present. Grey limestones formed during periods of more normal salinity. Dome-shaped structures in limestones of both the lower and upper Kilian Formation are probably stromatolitic in origin. Following extrusion of tholeiitic basaltic volcanics during deposition of the middle part of the upper Kilian Formation (Fig. 12.15) there was considerable influx of fine grained terrigenous muds, some of which are preserved as purple and red mudstones.

Figure 12.13

Large oncolites in the lower part of the Wynniatt Formation. Stratigraphic position of these organo-sedimentary structures is shown in Figure 12.10.

At the southwestern end of the Minto Arch, Thorsteinsson and Tozer (1962) noted a unit of sandstones and granule and pebble conglomerates in the upper Kilian Formation. This unit has not been studied in detail but is about 130 m thick in the southwest and thins rapidly to the northeast. It has not been observed in the northeastern part of the Minto Arch. It includes pebbles and granules of vein quartz which indicate an extrabasinal origin. Although the nature and origin of this unit must be more closely defined by future studies, it is here suggested that it may represent a westerly derived clastic wedge that spread over the evaporitic basin. It is in many ways analogous to the fluvial sandstones and conglomerates in the upper part of the Redstone River Formation in the Mackenzie Mountains region (Jefferson, 1978a) and may occupy the same stratigraphic position (Young, 1979).

In the Mackenzie Mountains area the sandstones were shed from a shoreline to the east of the main depositional basin (Jefferson, 1978a) to blanket and intertongue with evaporites that form the lower part of the formation. Incorporation of clasts from underlying formations of the Mackenzie Mountains supergroup indicates penecontemporaneous uplift of these units, probably by block faulting. These movements were considered by Young et al. (1979) to signal the beginning of the Hayhook Orogeny (formerly Racklan), marking the change from dominantly platform deposits of the Mackenzie Mountains supergroup to a more active tectonic regime in which the glaciomarine and resedimented rocks of the Rapitan Group were deposited. It is here suggested that the easterly-thinning unit of sandstones and conglomerates in the upper part of the Kilian Formation could be a distal facies of terrigenous clastics shed from a positive area to the west (Great Bear Arch of Fig. 12.2).

Figure 12.14. *"Molar-tooth" structures in the upper part of the Wynniatt Formation of section K, Figure 12.11. These are thought to be an early diagenetic phenomenon involving carbonate precipitation in early-formed tension gashes in the sediment.*

NATKUSIAK FORMATION

The volcanic rocks of the Natkusiak Formation were excluded by Thorsteinsson and Tozer (1962) from the Shaler Group because of the presence of a disconformable contact with the underlying Kilian Formation. The most detailed descriptions of these rocks are those of Christie (1964) and Baragar (1976). On top of the measured section of the Kilian

Formation (Fig. 12.15) there is about 13 m of crudely stratified purple pebbly, cobbly, and bouldery conglomerate or volcanic breccia. The breccia is composed mainly of volcanic fragments but also includes clasts of carbonate rocks and sandstone comparable to those of the underlying Shaler Group. The contact between the Kilian and Natkusiak formations is irregular on a scale of metres. The breccia is overlain by thick, commonly amygdaloidal, basalt flows. The lavas were considered by Christie (1964) to be consanguineous with the ubiquitous sills that intrude the Shaler Group and equivalents in other parts of the Amundsen Embayment.

SUMMARY OF DEPOSITIONAL HISTORY OF THE SHALER GROUP

The Shaler Group is composed of shallow water sedimentary rocks that accumulated in a large embayment of a late Proterozoic sea that extended into the Cordilleran region. Initial transgression resulted in deposition of a fining upward terrigenous clastic unit followed by a thick cherty stromatolitic dolostone. The upper part of the Glenelg Formation and lower part of the Reynolds Point Formation make up two major fluvio-deltaic cycles, interrupted by a transgressive episode during which a widespread stromatolitic biostrome was deposited. A prolonged period of shallow marine agitated conditions existed during deposition of the thick carbonates of the Reynolds Point Formation. Deposition of these carbonates was brought to a close by another influx of detrital clastic material. The upper part of the Shaler Group reflects basin restriction. Evaporites are concentrated at two stratigraphic levels; in the Minto Inlet Formation and in the lower part of the Kilian Formation. The gypsiferous rocks are interbedded with shallow marine carbonates and shales. The youngest sedimentary unit of the Shaler Group is an easterly-thinning wedge of terrigenous clastics. These distinct episodes in the depositional history of the Shaler Group (Fig. 12.1) can be matched in the stratigraphic succession of the Mackenzie Mountains supergroup about 500 km to the southwest. These similarities reflect a common, if not identical, sedimentological and tectonic history for the two regions.

Paleocurrent data from terrigenous clastic units that were not unduly reworked in the marine environment, indicate northwesterly transport. Similar transport directions have been obtained from fluvial sandstones on Banks Island and from the Brock Inlier (Young and Jefferson, 1975; Young, 1977). Highly variable current directions obtained from most other units reflect deposition in a shallow marine environment.

REGIONAL EXTENT OF FORMATIONS AND MEMBERS

A detailed correlation of the late Proterozoic rocks of various parts of the Amundsen basin was proposed by Young (1977). Some modifications were suggested by Dixon (1979). The widespread nature of many stratigraphic units, particularly in the lower part of the Shaler Group suggests that a unified stratigraphic scheme is needed for the entire region. New formational names are needed to accommodate the many subdivisions of the original formations proposed by Thorsteinsson and Tozer (1962).

Following the suggestion of Aitken et al. (1973), Young (1977) proposed a detailed lithostratigraphic correlation between the late Proterozoic rocks of the Mackenzie Mountains supergroup and those of the Amundsen Embayment. Since the documentation of the presence of two evaporitic units in the Mackenzie Mountains supergroup, the

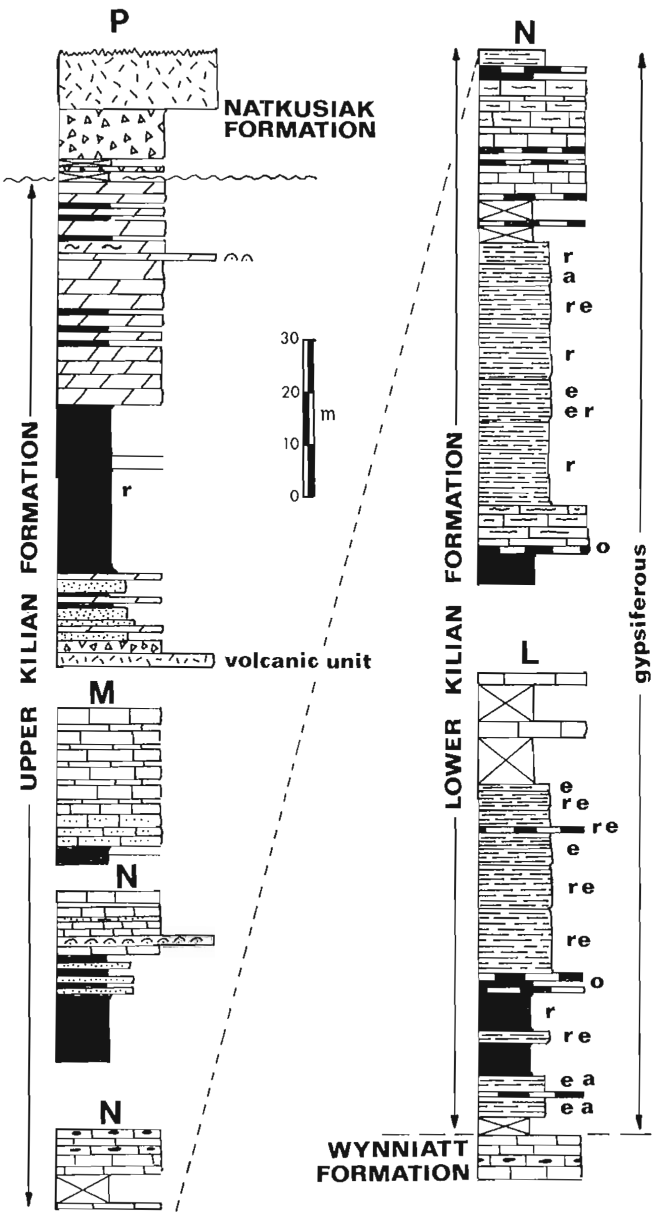


Figure 12.15. Composite generalized stratigraphic section of the Kilian Formation in the northern part of the Minto Arch. See Figure 12.11 for location of measured sections. Symbols as for Figure 12.3 with addition of the following:

- *dashed ornament represents gypsiferous shales and siltstone;*
- *The shale are varicoloured as follows:*
  r – red and purple
  e – green
  a – grey

Note that the upper Kilian lacks evaporites whereas the lower part is characterized by their presence. Note also the volcanic unit in the middle of the upper Kilian Formation (also shown in Figure 12.1). For chemical analyses of this unit see Table 12.2.

correlation has been further refined (Aitken et al., 1978b; Young, 1979). Although arguments have been presented against the former continuity of these two depositional basins (Eisbacher, 1977), the remarkable similarities in stratigraphic succession (Young, 1979) and stromatolites (Aitken et al., 1978a) leave little doubt that the basins were linked.

SEDIMENTATION AND REGIONAL TECTONICS

As pointed out by Donaldson et al. (1973) and Young (1980), terrigenous debris in various upper Proterozoic (~1200 – ~800 Ma) basins along the length of the Cordillera and in the northwestern part of the Canadian Shield was transported mainly to the west. It was during this period that the Grenville Province was undergoing metamorphism, intrusion and uplift (Stockwell, 1973; Baer, 1974). By analogy with present day orogens, such as the South American Cordillera, Young et al. (1979) and Young (1979) suggested that fine erosional debris from the rising Grenville orogen was transported across the continent by rivers to be deposited in basins on the western and northwestern margins of present day North America.

During deposition of the Redstone River Formation (=Kilian Formation of the Amundsen Embayment) a local shoreline was established in the Mackenzie Mountains region (Jefferson, 1978a). Eisbacher (1977) suggested that local faulting may also have contributed to deposition of coarse clastics near the top of the Mackenzie Mountains supergroup. These findings in no way negate the striking lithologic and biostratigraphic similarities between the Shaler Group and the Mackenzie Mountains supergroup nor do they preclude former continuity of the two basins, perhaps around the northern end of a north-northwesterly trending linear positive area (Great Bear Arch of Fig. 12.2), that separated the Amundsen and Mackenzie Embayments of a single extensive late Proterozoic seaway (Young, 1977).

The nature of the late Proterozoic sea is of extreme importance in interpretation of the tectonic setting of the Shaler Group and Mackenzie Mountains Supergroup. The main possibilities are an intracratonic sea with another continental land mass to the west, or that the Cordilleran region, in late Proterozoic times, was the trailing, Atlantic-type, edge of the North American continent. The presence of extensive evaporites in both regions might be of particular significance. Such evaporites require a peculiar set of circumstances, apparently unknown on Earth at present. These include restriction of circulation in a large body of sea water, limited access (by a small seaway or by underground seepage) to an ocean, and climatic conditions suitable for evaporation of large quantities of water. Such conditions existed in the initial phases of rifting that preceded opening of the Atlantic Ocean (Burke, 1975; Evans, 1978). The analogy with the Mackenzie Mountains-Amundsen Embayment is, however, imperfect in that the evaporites of the late Proterozoic successions overlie marine, rather than continental deposits like those of the Permo-Triassic of the Atlantic margins (Evans, 1978). The difference might be explained by the existence, in the late Proterozoic, of an earlier seaway such as an aborted continental rift system (Young et al., 1979) which became restricted by faulting(?) at least twice before final breakup in latest Proterozoic-Cambrian times (Stewart, 1972). Thus the late Proterozoic sea in which the Mackenzie Mountains Supergroup and Shaler Group were deposited may have been two sided. Possible candidates for the "other side" include Siberia (Sears and Price, 1978) and Australia (Jefferson, 1978b). Resolution of these problems must await further studies of Proterozoic rocks on both sides of the Pacific Ocean.

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13.

HUDSONIAN AND HELIKIAN BASINS OF THE ATHABASCA REGION, NORTHERN SASKATCHEWAN

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Abstract

Major late Hudsonian northeast-trending faults divide the Athabasca region into a series of subparallel basins. Textural and sedimentological evidence suggest that the initial basins were about the same size as at present at least in the eastern half of the area.

The pattern and types of faulting, the presence of intrusives, and the nature of the clastics suggest that the basins of the Athabasca region (including those of the Martin Formation) have formed as a series of pull-apart structures in an episode of transcurrent faulting during and following a Hudsonian plate suture in eastern Saskatchewan. The deepness of the Martin basins and the coarseness of their fill indicate that these late Hudsonian or early Helikian basins formed along strongly active faults. The Athabasca basins, several hundred Ma younger, were the result of a less active rejuvenation of the same fault systems. These basins coalesced at an early stage when infilled by sediments of the Athabasca Group, a molasse wedge shed from the region of the Hudsonian suture zone. The deposits are fluvial in the east, but predominantly lacustrine or marine in the western half of the area.

Chemical and petrographic work support the results from this stratigraphic and sedimentologic analysis and suggest that the sediments of the Athabasca Group, like those of the Martin Formation, were immature at the time of deposition and are first-cycle deposits derived from a quartz-rich metamorphic source. Differing diagenetic history has resulted in the present petrographic distinctiveness of these units. The apparent petrographic maturity and much of the evidence cited in the literature for a multicycle origin of the Athabasca Group sediments are due to a strong eolian influence in an environment devoid of sediment binding vegetation at the time of deposition, and the effects of intense diagenetic alterations that continue to the present.

Relations to other Helikian basins will be discussed.

Résumé

De grandes failles de direction nord-est, formées pendant l'Hudsonien supérieur, divisent la région de l'Athabasca en une série de bassins subparallèles. Les détails structuraux et sédimentologiques suggèrent que les bassins initiaux avaient à peu près les mêmes dimensions qu'actuellement, au moins dans la moitié est du secteur.

La configuration des terrains faillés et les types de failles, la présence de roches intrusives, ainsi que la nature des roches clastiques, semblent indiquer que la région des bassins de l'Athabasca (y compris ceux de la formation de Martin) sont initialement apparus sous forme de couches étirées lors d'un épisode de faillage à décrochement horizontal, pendant et après la formation d'une suture de plaques (à l'Hudsonien) dans l'est de la Saskatchewan. La profondeur des bassins de Martin et les matériaux grossiers qui les ont comblés montrent que ces bassins datant de l'Hudsonien supérieur ou de l'Hélikien inférieur se sont formés le long de failles très actives. Les bassins de l'Athabasca, qui les précèdent de plusieurs millions d'années, résultent d'un rajeunissement modéré du même système de failles. Ces bassins ont fusionné tôt dans leur histoire, pendant leur remplissage par des sédiments du groupe de l'Athabasca, qui consistaient en couches de molasses se terminant en coin et issues de la zone de suture de plaques d'âge hudsonien. Les dépôts sont fluviatiles à l'est, mais principalement lacustres ou marins dans la moitié ouest du secteur.

Les recherches chimiques et pétrographiques confirment les résultats de cette analyse stratigraphique et sédimentologique et suggèrent que les sédiments du groupe de l'Athabasca, comme ceux de la formation de Martin, étaient immatures au moment de leur dépôt et provenaient d'un premier cycle de sédimentation à partir d'une source constituée de roche métamorphique riche en quartz. La diagénèse particulière explique le caractère pétrographique distinctif actuel de ces unités. Leur maturité pétrographique apparente et les nombreux détails cités dans la documentation scientifique attribuant plusieurs cycles de sédimentation au groupe de l'Athabasca, résultent d'une importante influence éolienne qui s'est exercée dans un milieu dépourvu de végétation capable de lier les sédiments au moment de leur formation, et traduisent les effets d'intenses altérations diagénétiques qui se poursuivent actuellement.

On discutera des rapports avec les autres bassins de l'Hélikien.

INTRODUCTION

In Northern Saskatchewan three types of basins whose origin may be related to the Hudsonian Orogeny are recognized. The earliest type is filled by sediments of the Thluicho Lake Group and the Ellis Bay Formation. Together, these comprise an early to late Aphebian terrigenous clastic sequence metamorphosed to greenschist grade. These deposits are overlain by late Hudsonian or early Helikian unmetamorphosed volcanics and redbeds confined to deep, narrow fault-bounded basins of which the Martin Formation Basin is the best example. The youngest basin-fill sequence consists of unmetamorphosed quartz-rich clastics, dolomites, and basic intrusives of the Paleohelikian Athabasca Group.

Tectono-stratigraphic relations between these various basins are difficult to decipher due to the lack of mutual contacts and insufficient radiometric dates.

The following discussion of the Thluicho and Martin sequences is based largely on work by Tremblay (1972), Scott (1978) and the author. Discussion of the Athabasca Basin is based on work by the author unless otherwise indicated.

EARLY OR LATE APHEBIAN BASINS: THLUICHO LAKE GROUP AND ASSOCIATED SEDIMENTS

Thluicho Lake Group (Scott, 1978) rests unconformably on the Archean (?) metasediments of the Tazin Group from which it is derived and is preserved in at least two complexly folded synclinal basins (Fig. 13.1). Similar sediments are known from boreholes along the northwest side of Lake Athabasca, where they underlie the Athabasca Group (Harper, 1979) and from Slate Island in Lake Athabasca ('Martin or Tazin' of Donaldson, 1968).

In the type area the Thluicho Lake Group consists of a basal metaconglomerate up to 1000 m thick conformably overlain in turn by meta-arkose and argillite. In all three units chlorite, muscovite, biotite and albite are common. In the meta-arkose and argillite units graded bedding, crossbedding, ripple marks and load structures are locally abundant.

The unconformably overlying Ellis Bay Formation contains some reworked Thluicho Lake Group clasts, but consists largely of monomictic, massive felsic breccia containing rounded fragments of aphanitic albite in a less resistant matrix. It may represent a volcanic rock subjected to brecciation and retrograde metamorphism (Scott, 1978).

Interpretation

No sedimentological analysis of the Thluicho Lake Group has been made, but Scott's (1978) observations are consistent with a turbidite origin for at least part of the Thluicho Lake Group and Ellis Bay Formation.

The depositional age of these units is unknown. An early Aphebian age is possible and might be related to rifting between the Slave and Superior cratons, in which case a correlation with the Wilson Island Group of the Great Slave Lake area (Hoffman, 1979) is possible. A Hudsonian origin is also possible. Hudsonian tectonism decreased markedly west of the Virgin River-Black Lake Shear Zone (Lewry et al., 1978) and consequently marine deposits may be expected in this area as a sedimentary record of the major uplifts to the east. The Thluicho Lake Group and the Ellis Bay Formation may be analogous to the Cretaceous marine units of the Plains in their relation to the Nevadan-Laramide uplifts in the Cordillera or analogous to the flysch units of the northern Alpine foredeep. Detailed sedimentological analysis involving paleocurrent studies in the Thluicho Lake Group sediments should help to distinguish between these hypotheses.

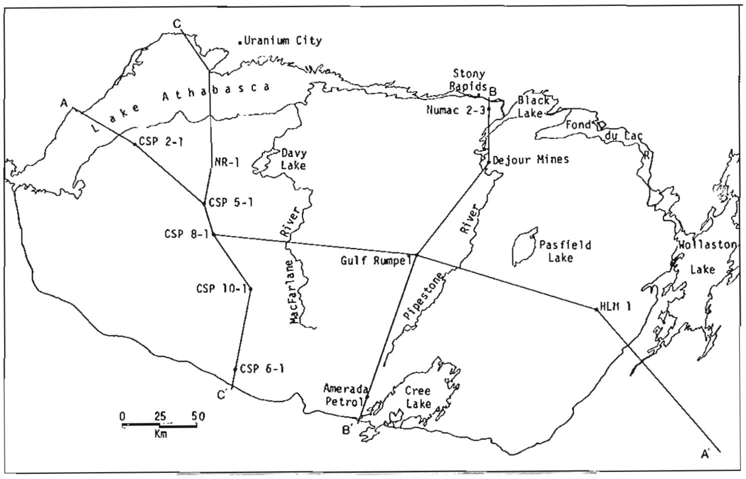
LATE APHEBIAN WRENCH FAULT BASINS: MARTIN FORMATION AND ASSOCIATED DEPOSITS

Associated deposits in northern Saskatchewan of small, late Aphebian unmetamorphosed redbed units occur over a wide area. They unconformably overlie the metamorphosed basement, are themselves folded and faulted, and lie in fault-bounded basins. To date, only part of the Martin Formation has been studied in detail in the Uranium City area (Tremblay, 1972).

Basins similar to the one containing the Martin also occur around Tazin Lake and between Tazin Lake and Lake Athabasca (Scott, 1978). An isolated series of redbeds occur along the northern Lake Athabasca shoreline (Fig. 13.1). Locally in northeastern Saskatchewan, boulders identical to the Martin are common, and suggest the presence of small redbed outliers in this area.

In the Virgin River-Black Lake Shear Zone and the Cree Lake Zone late Aphebian redbeds are missing. As this is an area of exposed Hudsonian high-grade metamorphic rocks their absence may be attributed to a greater degree of post-Hudsonian uplift.

Beds of 'uncertain stratigraphic position' in the Maurice Bay area (Harper, 1979) may be remnants of another basin equivalent to Martin Lake Basin rocks as it contains the usual unmetamorphosed polymict conglomerates. The matrix of these sediments contains largely quartz grains and amygdular or vesicular grains with abundant microliths and spicular structures. The other framework grains have been largely replaced by calcite, making identification difficult, but included structures suggest a volcanic or perhaps a pedogenic origin. Siltstone clasts are present in these conglomerates as are clasts of Thluicho Lake Group sediments (Harper, 1979).



1 Thluicho Lake Group
2 Ellis Bay Formation

Redbed occurrences:

3 Martin Formation
4 Charlot Point Formation
5 Tazin Lake redbeds
6 Slate Island
7 Stewart Island
8 Johnstone Island
9 Gruchy Island
10 Maurice Bay
11 Junction Granite
12 Chipman Sills

Figure 13.1. *Location of sedimentary basins, intrusives and of sections through the Athabasca Basin shown in figures.*

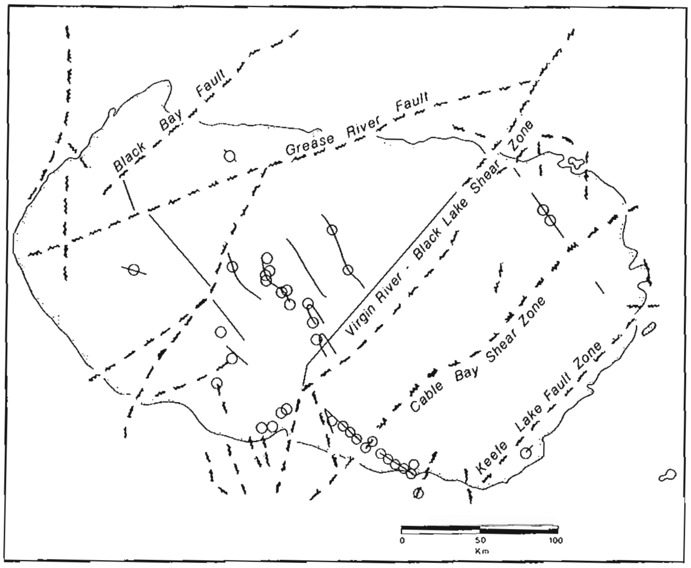


Figure 13.2. *Major faults and shear zones (∿∿∿), dyke outcrops (o), and magnetic lineaments indicating the presence of dykes (——) in the Athabasca Basin.*

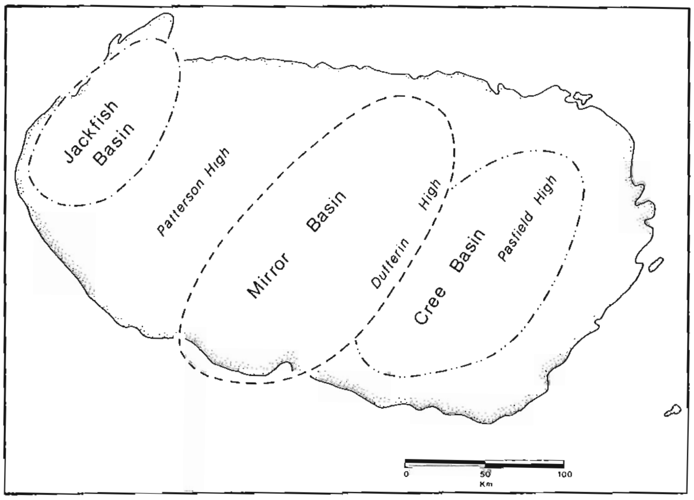


Figure 13.3. *Subbasins in the Athabasca Basin.*

Martin Formation:

The Martin Formation, which fills the Martin Lake Basin, consists of about 5000 m of breccias, conglomerates, arkoses, siltstones, basic volcanics and is intruded by gabbroic dykes and sills. Volcanics of the Martin Formation have been dated by K-Ar at 1630 ± 180 Ma (Fraser et al., 1970). Evans and Bingham (1973), incorporating their paleomagnetic data with the published radiometric ages, suggested an age of 1730 ± 1830 Ma for the Martin Formation. They also established that the hematitization preceded the folding, and thus must have taken place almost immediately after deposition.

Paleocurrent measurements (Tremblay, 1972, Fig. 5) show that sediments in the Martin Lake Basin were derived from both sides of the basin and suggest that the basin was initially not significantly larger than at present. The locally-derived basal conglomerates reinforce that conclusion (Fraser et al., 1970). The depositional environments of the Martin Formation redbeds range from talus slopes, alluvial fans and braided streams to lacustrine.

Discussion:

Scott (1978) showed that details of sedimentation, paleocurrent directions and diagenesis differ significantly in the various different redbed successions, suggesting the existence of separate basins. For example, the Charlot Point Formation (Langford, 1980), formerly a subdivision of the Martin Formation (Scott, 1978) or the Athabasca Formation (Fahrig, 1961) is distinguished from the type Martin Formation by the presence of red siltstone clasts. The large siltstone clasts, relative to quartzite cobbles in the Charlot Point Formation, indicate the siltstones were well indurated prior to deposition. These clasts were interpreted by Langford (1980) to have been derived from the lower Martin Formation and hence he suggested the Charlot Point Formation is younger than the Martin. Charlot Point Formation clasts, however, appear richer in chlorite than the type Martin Formation (compare Tremblay, 1972; Macey, 1973; Scott, 1978).

Alternatively, the siltstone clasts may have been derived from the underlying Thluicho Lake which contains abundant argillites, as the redbeds all show evidence of very local derivation (Tremblay, 1972; Scott, 1978).

At Slate Island, redbeds which unconformably overlie Thluicho Lake Group sediments contain siltstone clasts, suggesting similar derivation. These conglomerates, assigned by Donaldson (1968) to the Athabasca Formation, probably record another redbed basin similar to the Martin Lake, as their lithologic composition and contained sedimentary structures differ sharply from the overlying Fair Point Formation (Ramaekers, 1979a), the lowermost unit of the Athabasca Group in the area.

Erosion and reworking in an oxydizing environment, followed by 1600 Ma of diagenetic alteration easily could have altered some of the green Thluicho Lake argillites to the red and tan clasts of the Charlot Point Formation. This process has been documented in the type Martin Formation where most mafic minerals have been altered, liberating the red iron oxides (Macey, 1973).

Thus, the Charlot Point Formation is here considered to be approximately equivalent to the Martin Formation.

Identification of separate redbed basins means that the name 'Martin Formation' should be restricted to the heterogeneous sediments of the Martin Lake Basin until this complex succession (Tremblay, 1972), deposited in widely differing depositional environments, is better understood. The designation of the Charlot Point area redbeds of the Charlot Point Formation (Langford, 1980) is a first step in the unravelling of the area's Late Hudsonian history.

Interpretation:

The dimensions of the known and inferred redbed basins, contained sediments and igneous rocks, locations relative to orogenically uplifted areas, formation during closing phases of the Hudsonian Orogeny, close association with major faults, and structural complexity all suggest an origin as pull-apart basins associated with wrench faulting similar to that described by Crowell (1974) in southern California, or in the Devonian of Norway by Steel (1976).

In a wrench fault environment the basinal stratigraphy in more or less contemporaneous adjacent basins is commonly dramatically different (e.g. Steel, 1976). Similarities and differences in the basin fill reflect details of tectonic history

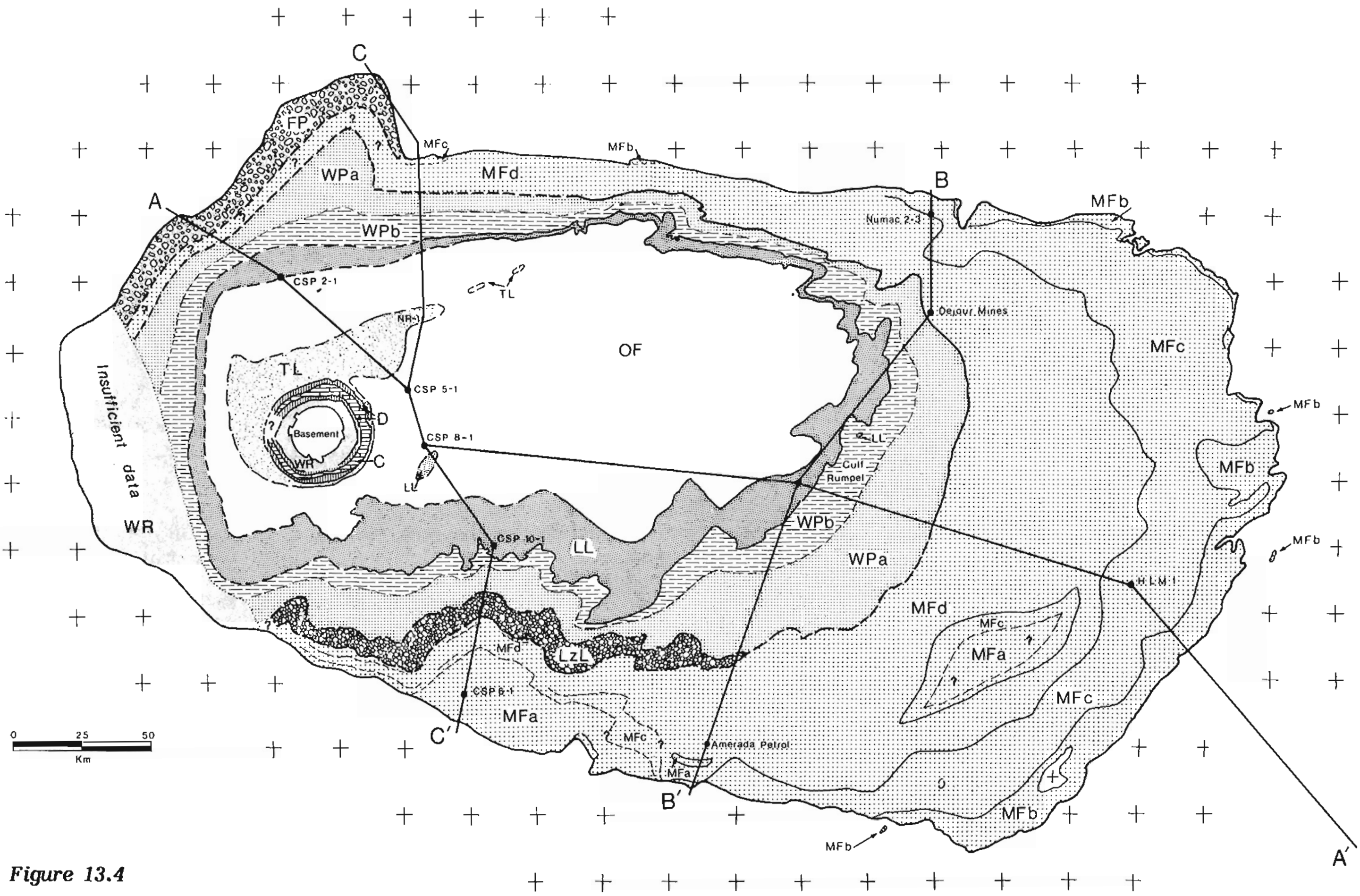


Figure 13.4

and of source areas, and cannot provide a basis for age relationships between the various basins. This seems the case in northern Saskatchewan where some redbed sequences begin with siltstones and arkoses rather than conglomerates (Scott, 1978).

In the Virgin River-Black Lake Shear Zone a number of un- or weakly metamorphosed intrusives of Late Hudsonian age occur. The Junction Granite, age 1745 Ma and the Chipman Sills (Gilboy, 1980) may represent the intrusive roots of wrenchfault basins. This type of basin may reasonably be expected along this long-active shear zone – now exposed at a deeper crustal level than the tectonically "quieter" areas of the redbed basins, typified by the Martin Lake Basin.

Major wrench faults may be expected along a continental suture to accommodate the lateral component of movement at the plate junction. This interpretation seems more plausible than a short-lived phase of rifting which would require dramatic, unexplained changes in mantle circulation to explain these basins.

Modern analogues to the Martin Lake and associated basins are in southwestern British Columbia (Price, 1965) where the Oligocene Kishenehn Formation displays the same diversity in lithologic types and sedimentary environments in the same structural setting as does the Martin Formation.

Summary

The Thluicho Lake-Martin succession has some similarities to successor basin sequences in that coarse fluviatile deposits overlie what appears to be a turbidite unit, and the whole sequence contains limited volcanism (Ramaekers and Hartling, 1979; Langford, 1980). The underlying Tazin Group, however, is not the typical eugeosynclinal sequence that definitions of 'successor basin' require ('epieugeosyncline' of Kay, 1951; King, 1966). While much work on the stratigraphy and depositional environment of the Tazin Group remains to be done, its quartzites, calc-silicates, and few volcanic units suggest a more stable shelf environment (Sibbald and Lewry, 1980). The Martin Formation is not a typical molasse sequence; as discussed above, it more closely resembles the sediments of a wrenchfault basin.

The post-Hudsonian Athabasca Group cover of the Thluicho Lake-Martin sequence has no analogues among Phanerozoic successor basins, perhaps because Phanerozoic uplifts are too young for such a sequence to have developed and because there are few old intracratonic metastable zones on the modern continents.

Much work on the geochronology, stratigraphy and structure of the erosional remnants of these basins needs to be done before they can be incorporated into a plate-tectonic model.

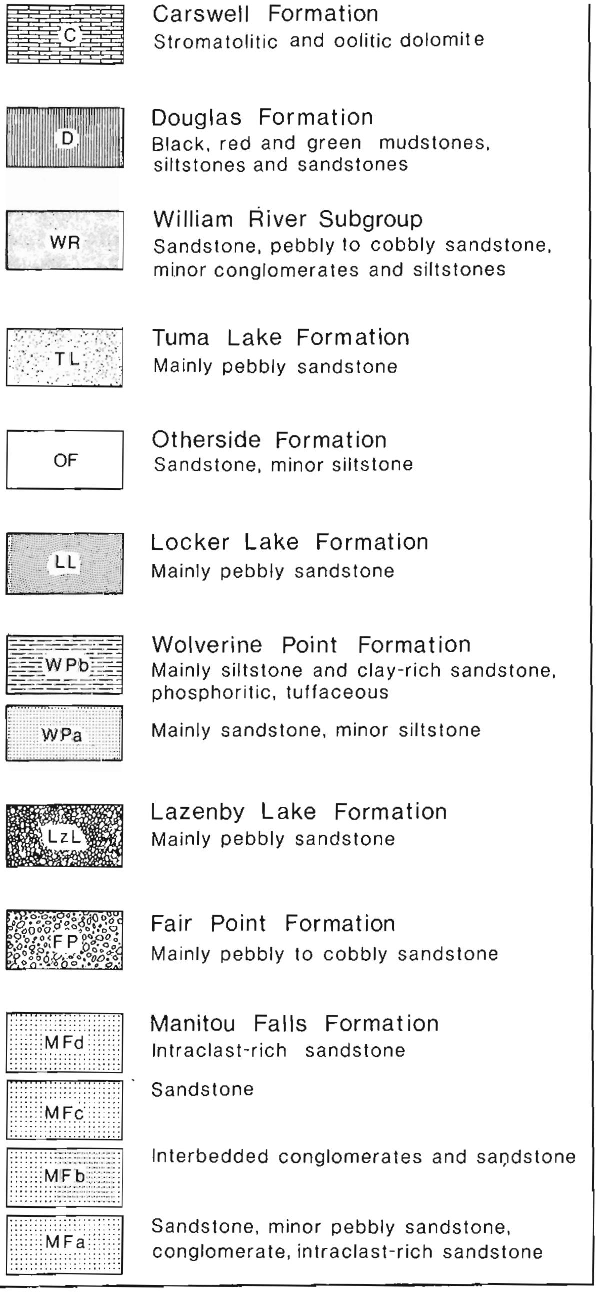


Figure 13.4. *Geological Map of the Athabasca Basin. The locations of the sections are of those shown on Figure 13.5.*

PALEOHELIKIAN CRATONIC BASINS: ATHABASCA GROUP

A period of instability followed the phases of the Hudsonian Orogeny and a series of epeirogenetic uplifts and subsidence zones developed in northern Saskatchewan. The seismic study by Hobson and MacAuley (1969) subdivided the Athabasca Basin into three northeast-trending subbasins (Fig. 13.3), and this has been confirmed by stratigraphic analysis (Ramaekers, 1979a, 1980a).

Sub-Athabasca Group Regolith

The basement underlying the Athabasca deposits was intensively weathered. The paleoweathering profile, up to 50 m of which is preserved, has characteristics typical of lower levels of present day laterites (Hoeve, 1977; Hoeve and Sibbald, 1978). A mineralogical zoning with kaolinite in the upper levels and illite and some chlorite in the lower levels is common. Red hematite staining is pervasive, except at the base of the profile. The profile is variably affected by superimposed reducing alteration which removed hematite and emplaced chlorite in the upper levels of the profile and also affected the overlying sandstones and conglomerates of the Athabasca Group (Ramaekers and Dunn, 1977).

Basal Conglomerates

At the lower contact of the Athabasca Group at least five types of conglomerate may occur, all of which have been termed 'basal Athabasca conglomerates' although they have widely differing origins and ages.

1. As discussed above, a number of the 'basal Athabasca Formation conglomerates' of various authors are Late Hudsonian units deposited in a dramatically different tectonic setting and which are perhaps up to 200 Ma older. Included in these are the conglomerates on Slate Island, Charlot Point and on the islands bordering Crackingstone Peninsula.
2. Some of the sub-Athabasca Group regolith contains conglomeratic lags, and these reflect winnowing and minor reworking of the upper layer of the regolith. Their lithology directly reflects that of the underlying rocks, and they are typically well developed (1-2 m thick) over fine grained pelitic rocks with large quartz segregations (e.g., at Newnham Lake). More commonly though, these lags are thin and discontinuous. By comparison with modern analogues, they may be considerably older than the Athabasca sediments.
3. Locally, as in the Maurice Bay area, the regolith mixed with fresher bedrock is reworked into typically thin, locally-derived debris flows with a quartz sand matrix. These debris flows may be interbedded with lowermost Athabasca Group sediments. As debris flows also occur in the older underlying Late Hudsonian basins, differentiation of Athabasca Group debris flows may be difficult unless angular unconformities are present. These debris flows may be related to initial tectonics which formed the first Athabasca subbasins, before deposition of the major deposits had obliterated bedrock itself.
4. Less frequently regolith material is sorted, transported and incorporated into the overlying Athabasca Sandstone units.
5. A number of the major marine and basal fluvial units of the Athabasca Group are conglomeratic or consist of interbedded sandstone and conglomerate. Clasts are generally well rounded and relatively well sorted.

Minor Units and Formation Contacts within the Athabasca Group

At present, the Athabasca Basin is known only at a reconnaissance level. Major lithostratigraphic units have been mapped and a regional paleographic reconstruction is possible. Detailed mapping in conjunction with drilling programs will doubtless refine the stratigraphy and delineate additional minor units.

Although the basement-Athabasca Group contact is generally sedimentary, fault contacts are common. Debris flows in the vicinity of the faults suggest that at least some faulting was syndepositional. Postdepositional faulting seems more common, and its extent is likely underestimated due to poor exposure of fault zones. Formational contacts within the Athabasca Group are virtually never exposed. Their nature is generally inferred from the lithofacies sequence and the pebble content of the overlying units.

Rarely, at the basement contact and at unconformities within the Athabasca Group, thin local units define sedimentary environments different from those of the surrounding formations. They typically represent environments dominated by local erosional relief. One example is the clay-rich unit with rare desiccation cracks that overlies the basal debris flows of the Athabasca Group in the Maurice Bay area (Harper, 1979; Mellinger, 1980). Similarly, in fluvial units of the eastern part of the Athabasca Basin, paleocurrent patterns at the basal contact may vary dramatically from the uniform trends in the overlying sediments. This also suggests local topographic control during the initial phases of sedimentation. At or near the contacts of the fluviatile units, thin clay-rich zones occur in some borehole cores suggest the presence of short-lived coastal lagoons.

Major Stratigraphic Units of the Athabasca Group

The Athabasca Group (Ramaekers, 1979a) is composed of four marine transgressive sequences and one thick fluvial regressive wedge. Locally pronounced unconformities and disconformities changing basinwards to apparently conformable contacts separate the transgressive sequences. Location, thickness, maximum grain size, and generalized paleocurrent directions in these and other formations described below are illustrated in Figures 13.3-13.6 and 13.9.

The oldest major units are the Fair Point Formation and the lower member of the Manitou Falls Formation (Manitou Falls A) deposited in the Jackfish and Mirror basins respectively.

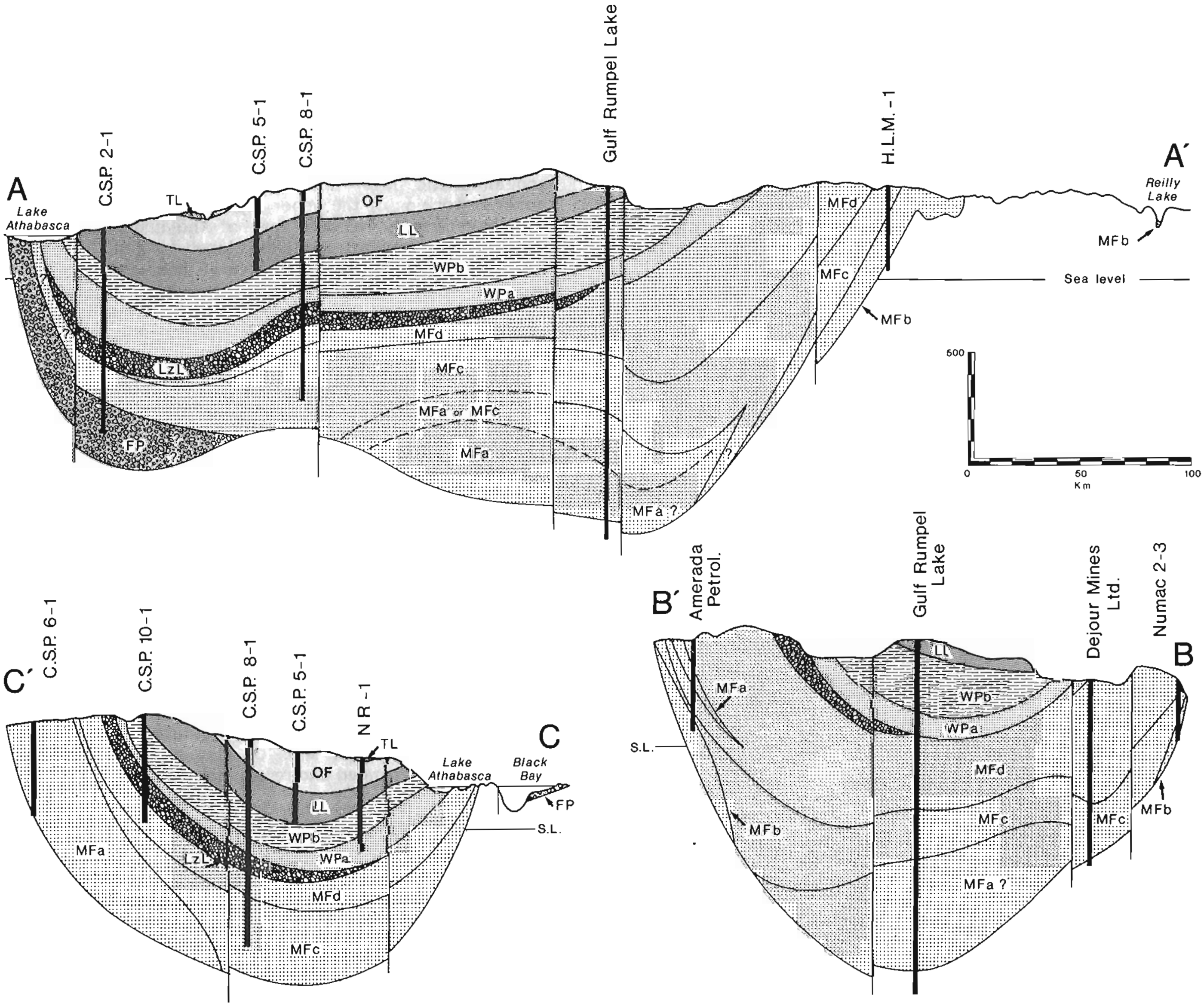


__Figure 13.5.__ Cross-sections through the Athabasca Basin, drawn through the locations as indicated on Figure 13.4. The legend is the same as for Figure 13.4.

Fair Point Formation

The Fair Point Formation (Ramaekers, 1979a, 1980a) is a 300 m thick, cobbly to pebbly, clay-rich marine sandstone, probably of local derivation. Horizontal bedding, low angle and hummocky crossbedding are common. Planar trough and herringbone crossbedding are also present. Paleocurrent directions indicate that the sediment transport was predominantly to the northwest, but their distribution is commonly multimodal and their variances are high. The sedimentary structures and lithologic evidence indicate a marine, high energy, lower shoreface depositional environment dominated by wave, and to a lesser extent, tidal effects.

Manitou Falls Formation

Manitou Falls unit A: This unit described by Ramaekers (1980a) is a 300 m thick sequence of probably fluvial and marine sandstones, pebbly sandstones and minor conglomerates. It underlies the main fluvial wedge of the Manitou Falls Formation (Manitou Falls B, C and D). For lack of detailed mapping, the heterogeneous Manitou Falls A sequence is appended to the base of the Manitou Falls Formation at present. The Manitou Falls A sediments probably were derived largely from the south. However, sedimentary structures and the variability of paleocurrent directions reflect the inhomogeneity of this unit.

Manitou Falls units B, C and D: Overlying the Fair Point Formation and the Manitou Falls A sequence are Manitou Falls members B, C and D, which form an eastward-thickening (to 1000 m) wedge of fluviatile sandstones that comprise the regressive phase of the Athabasca Group (Ramaekers, 1979a, 1980a).

At the base is the conglomeratic member (Manitou Falls B), consisting of eastward-thickening and coarsening interbedded sandstones and clast-supported conglomerates of which 300 m is preserved. Planar and trough crossbedding are predominant in the sandy units, whereas gravels show massive or crude planar bedding and with pebbles imbricate normal to the transport direction. Ripples and clay intraclasts are rare. Both fining- and coarsening-upward sequences up to 15 m thick are present. Paleocurrent directions are quite uniform at outcrop scale, with variances typically less than 1000. The paleocurrent directions define a number of radiating alluvial fans or bajadas divisible into two major deposystems, one derived from the northeast, the other from the eastern side of the basin. Minor detritus supplied from the south is also indicated. In contrast to the Late Hudsonian basins no debris flows occur in these fans.

Distally the conglomeratic fans grade into sandy braided stream deposits (Manitou Falls C) with abundant planar and trough crossbedding, crosslamination, some climbing ripple lamination, and common linguoid current ripples. Clay intraclasts are generally present but not common. Paleocurrent directions show low variances and are directed uniformly to the west.

The sandy braided stream deposits grade westward into similar but more uniformly finer grained deposits (Manitou Falls D). These sands contain abundant clay intraclasts, occasional thin (7.5 cm) lenses of bedded silt and mudstone, abundant current ripples and a few oscillation ripples. Paleocurrent directions show increased variances, but are still directed unimodally to the west. The evidence suggests that the depositional environment was a rapidly aggrading alluvial plain of braided streams with short-lived intermittent lakes similar in many respects to modern wadis.

The Manitou Falls D unit thickens dramatically east of the Virgin River-Black Lake Shear Zone, suggesting that the Dufferin High formed in this shear zone at this time, creating the Cree Basin to the East (Fig. 13.2).

Lazenby Lake Formation

The Lazenby Lake Formation (Ramaekers, 1980a) disconformably overlies the Manitou Falls D sequence in the central part of the Athabasca Basin and marks the base of the second transgressive sequence. It is a pebbly quartz sandstone coarsening and thickening to the south to a maximum thickness of about 130 m, suggesting a southern source area. Sedimentary structures are similar to those of the Fair Point Formation except that convolute bedding and sand volcanoes are common. The unit probably accumulated in a marine, high-energy, lower shoreface depositional environment.

Wolverine Point Formation

The poorly exposed Wolverine Point Formation (Ramaekers, 1979a, 1980a) conformably and gradationally overlies the Lazenby Lake Formation or the Manitou Falls Formation wherever it overlaps the Lazenby Lake Formation. The lower member (Wolverine Point A) consists of 100-200 m of sandstones interbedded with a few thin siltstones. It thickens to the north and west at the expense of the Wolverine Point B member. Planar, trough and possibly hummocky crossbedding are common as are oscillation rippled surfaces. Paleocurrent variances are high at the few good outcrops of this unit (>4000). The evidence suggests a marine lower shoreface to shallow inner shelf depositional environment.

The Wolverine Point A grades upwards into the 100-300 m thick sandstones, siltstones and mudstones of the upper Wolverine Point member (Wolverine Point B). Horizontal bedding, large planar crossbeds, trough and hummocky crossbeds are common. Current lineation and current crescents indicate strong currents at high angles to current directions in underlying bedforms. Paleocurrent directions are polymodal. The member was probably deposited in a relatively shallow marine inner to open shelf environment subject to strong tidal or storm driven currents. Thin phosphorites are widespread, which rarely contain preserved tuffs. The tuffs are preserved exclusively in the quietest depositional environment of the Athabasca Basin and only where they are replaced by an early diagenetic stable mineral. This suggests that an indeterminate amount of Athabasca Group clastics may be tuffaceous, but that the intense diagenetic effects have elsewhere altered the tuffs beyond recognition.

Locker Lake Formation

The third marine transgressive sequence begins with the pebbly sandstones of the 75-200 m thick Locker Lake Formation (Ramaekers, 1979a, 1980a) – another marine shoreface unit which also thickens southward. Coarse clastic input is concentrated in the south, east, and northeast, suggesting fluvial input from these directions. Sedimentary structures show the same range as those of the Fair Point and Lazenby Lake formations. Paleocurrent directions have high variances and are polymodal in the south and east but unimodal northwest in the northern part of the Athabasca Basin. This suggests the dominance of unimodal wind-and-wave-driven currents in the northwest. To the south the high paleocurrent variances, the coarsening and thickening of the

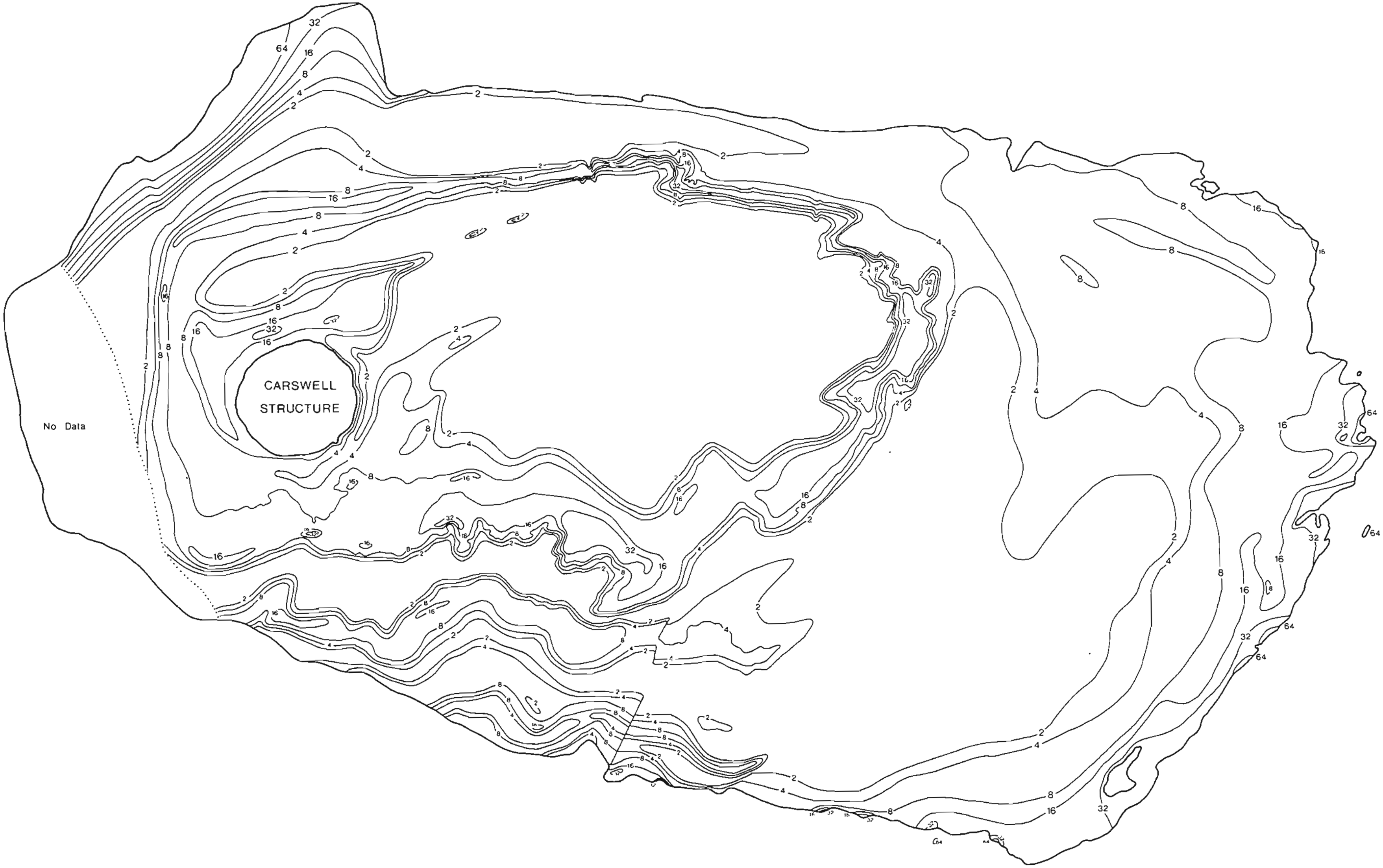


Figure 13.6. Maximum grain sizes in the sandy formations of the Athabasca Basin. The distribution of the basin and the unit designations are as on Figure 13.4.

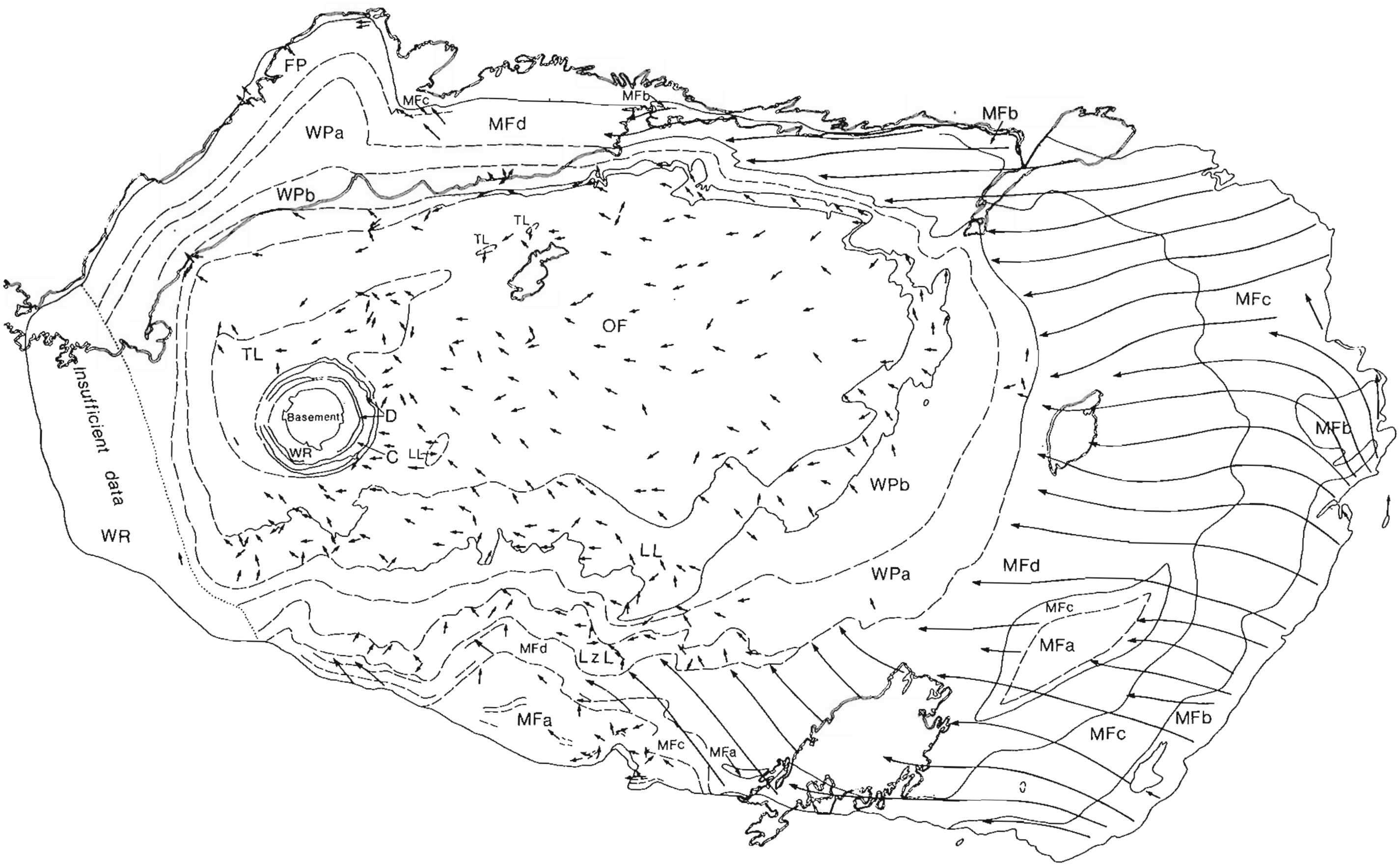


Figure 13.7. Generalized paleocurrent directions of the sandy formations of the Athabasca Group. Arrows reflect average directions at outcrops and are based on 10 023 measurements from 1179 outcrops. The unit designations and distributions are as shown on Figure 13.4.

Locker Lake Formation (as the Lazenby Lake and Wolverine Point formations) suggests that a large sandy delta entered the Athabasca Basin from the south, of which only the marine beds have been preserved (except for the fluvial units of the Manitou Falls A).

Otherside Formation

The 200 m thick, largely sandy Otherside Formation (Ramaekers, 1979a, 1980a) conformably overlies the Locker Lake Formation. It is interpreted as a shallow marine, lower shoreface to inner shelf deposit which accumulated under tidal influence. Sedimentary structures are similar to those of the Wolverine Point Formation and the paleocurrent pattern is again polymodal, occasionally bipolar, and shows high variance at outcrop scale.

Tuma Lake Formation

The fourth marine transgression is poorly documented because of poor exposure, lack of boreholes in the area, and structural complications. It is conceivable that its basal unit, the 100 m pebbly sandstones of the Tuma Lake Formation (Ramaekers, 1979a, 1980a), represents progradation and a retreat of a shoreface unit rather than a new transgressive sequence. However, critical areas needed to test this hypothesis are not preserved, and at present even the nature of the basal contact is uncertain. The Tuma Lake Formation is similar to both the Lazenby Lake and Locker Lake formations in its range of sedimentary structures and in showing a complex paleocurrent pattern.

Douglas and Carswell Formations

Overlying the Tuma Lake Formation is the Douglas Formation (Amok, 1974) interpreted as a quiet water, shallow marine shelf sequence of sandstones, carbonaceous siltstones and mudstones with abundant graded beds that contain compacted sandfilled syneresis cracks. The Douglas overlies the uppermost unit of the Athabasca Group, the oolitic and stromatolitic dolomites of the Carswell Formation (Blake, 1956; Fahrig, 1961; Currie, 1969; Ramaekers, 1979a). Channels in the Carswell are filled with intraclasts of micrite, micrite spar or algal laminates, stratified debris, stromatolites, algal laminates and both quiet water and agitated water oolites. Together, these indicate a warm, shallow water environment of deposition containing both low and high energy locales. These uppermost two formations are preserved only in the Carswell structure (Fig. 13.3-13.5) where intense tectonic complications make it difficult at this stage to determine thicknesses and contact relationships.

Intrusives

Diabase dykes were emplaced in the Athabasca Basin from about 1360 Ma to 1000 Ma (Armstrong and Ramaekers, in press). They reoccupied Hudsonian north-south faults, formed or occupied extensive northwesterly fracture zones or faults (Ramaekers and Hartling, 1979), and were most extensive and penetrated highest in the Athabasca Group sediments of the Mirror Basin (Fig. 13.2, 13.3).

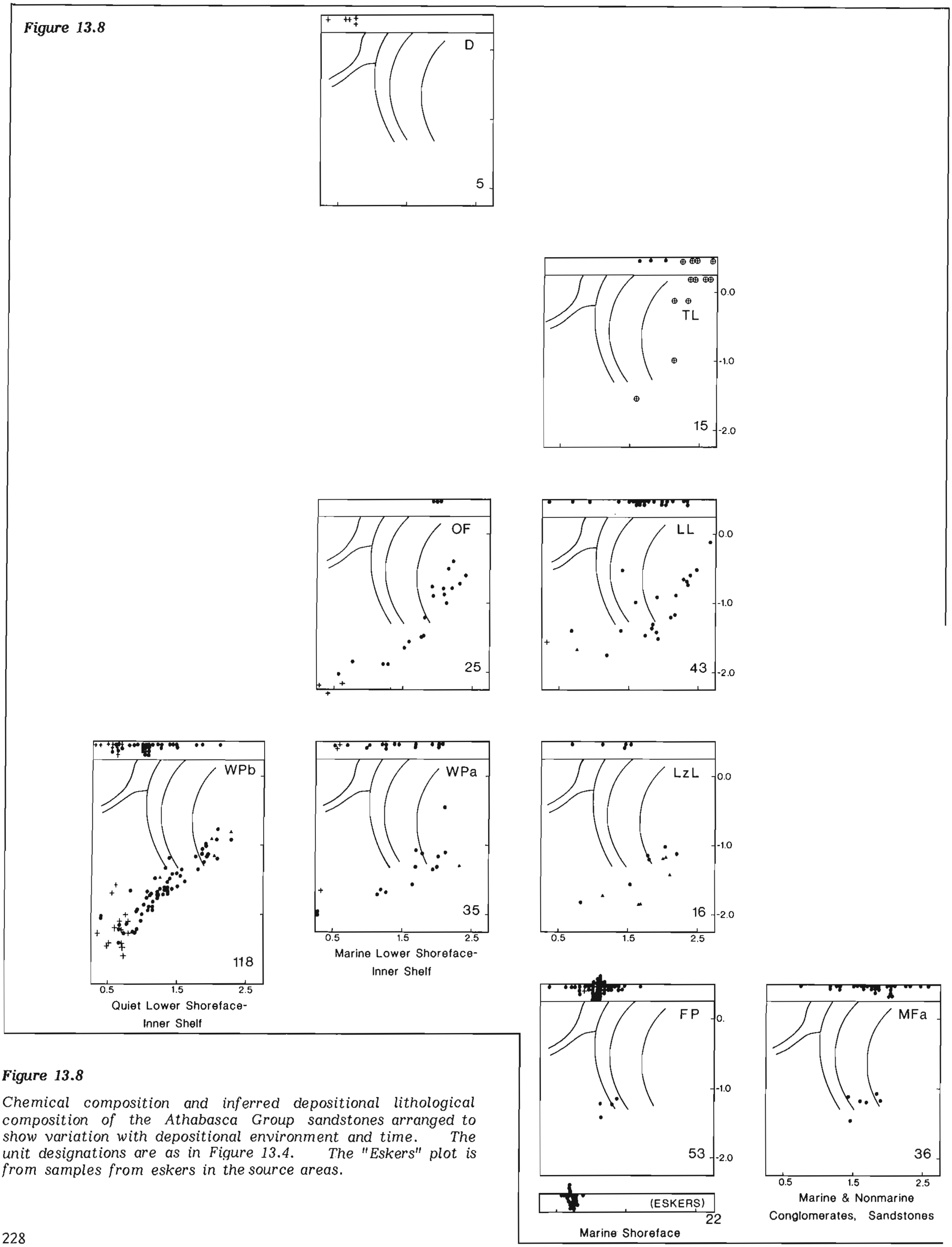


Figure 13.8

Chemical composition and inferred depositional lithological composition of the Athabasca Group sandstones arranged to show variation with depositional environment and time. The unit designations are as in Figure 13.4. The "Eskers" plot is from samples from eskers in the source areas.

PALEOGEOGRAPHY OF THE ATHABASCA GROUP

Source Areas

The Athabasca Basin sediments were derived from four main source areas. Local source areas dominated sediment supply in the Jackfish Basin during deposition of the basal Fair Point Formation. The remaining three source areas were northeast, east and south of the Athabasca Basin, and continuously through intermittently contributed detritus throughout the depositional history of the basin. The relative amount of detritus derived from these three source areas varied with time, and this gives an indication of the sequence and location of the uplifts (Table 13.1). Criteria used to delineate source areas for marine units were formation thickness and maximum grain size. In adjacent nonmarine units paleocurrent directions support this interpretation convincingly (Ramaekers, 1978).

The maximum grain size distributions of the Lazenby Lake, Locker Lake, and Manitou Falls A (Fig. 13.6) units suggest that clastics from the southern source region probably entered the Athabasca Basin via a major river system debouching in the Virgin River Shear Zone area, suggesting that large Proterozoic rivers had the same tendency to occupy major fault zones as do modern streams (Potter, 1978). The fluvial and deltaic units of this postulated delta complex are either not preserved or have not been recognized to date above the Manitou Falls A sequence, perhaps because the unstable sandy shore deposits were reworked shortly after deposition or the shoreline lay farther to the south.

Paleoslope

Paleocurrents in fluvial units trend predominantly to the west (Fig. 13.7), indicating a west-dipping paleoslope during Manitou Falls deposition. The changes in source areas with time (Table 13.1) probably record concomittant changes in the regional paleoslope.

Original Distribution of the Athabasca Group

Initially, Athabasca Basin may have occupied an area similar to the present basin at its southern and eastern margins, where there is evidence of fluvial clastics. Thickening and coarsening of sediments in the marine part of the Athabasca Group to the south and east suggests that the margins of the basin remained more or less in the positions throughout the deposition of the Athabasca Group.

There is little evidence to suggest clastic input from the north. One of the rare exposures of the northern basal contact shows flat-lying fluviatile Manitou Falls sediments in depositional contact with steeply inclined basement. Paleocurrents in the Manitou Falls parallel the contact, suggesting that the formation was deposited against a steep pre-existing slope. Thus, in its early stages, the northern boundary of the Athabasca Basin locally may have been a fault scarp. The drainage pattern on adjacent uplands may have been predominantly to the north and away from the basin centre. Fining of some marine units to the north and west suggests that the Athabasca Group may have extended farther to the north than at present in its later stages.

The western part of the Athabasca Basin was apparently continuously a shallow marine environment during sedimentation in the basin. Phosphorites of the Wolverine Point B member suggest a broad connection to the open ocean.

The absence of Helikian sediments west of the Athabasca Basin suggests the sediments were stripped from the uplifted Slave Arch (Fraser et al., 1970) in post-Helikian to pre-Devonian time.

Paleoclimate

To date no evaporites or pseudomorphs have been confirmed in the Athabasca Group. Oolites and dolomites of the Carswell Formation suggest the seas were warm, but apparently circulation was sufficiently open to prevent

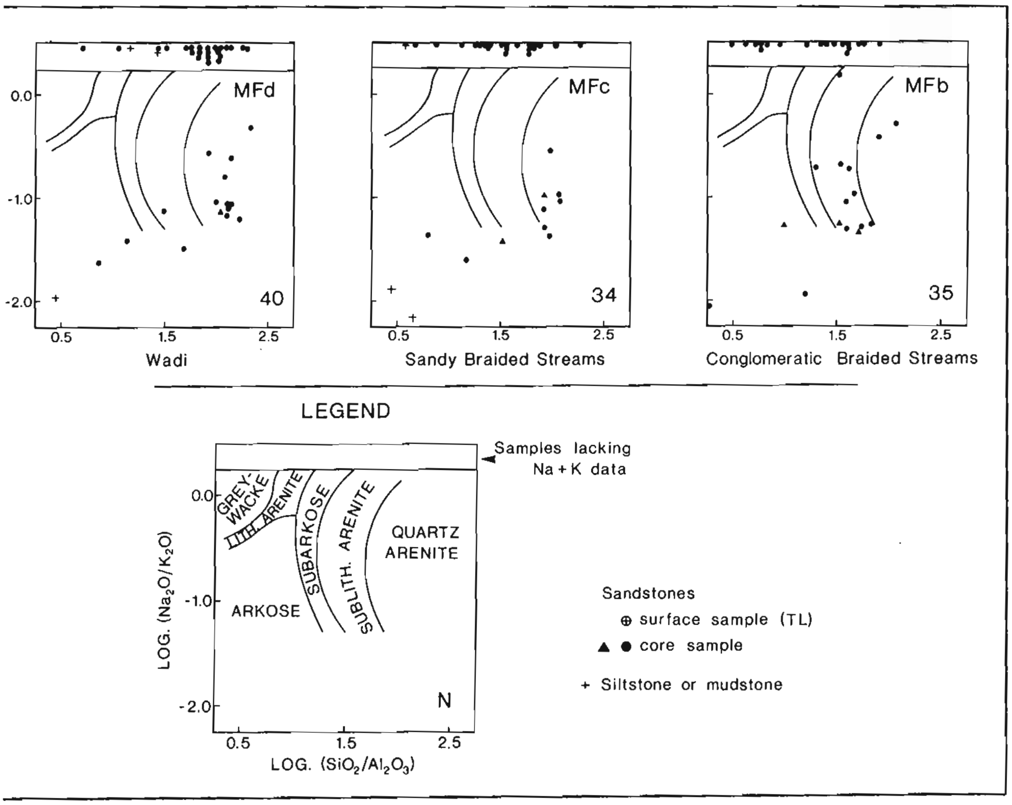


Figure 13.8

Table 13.1. Formations of the Athabasca Group showing depositional environments, geographic distributions, and source areas

West	Jackfish Basin	Mirror Basin	Cree Basin East	Source Area
marine	Carswell Fm.			
marine	Douglas Fm.			
marine	Tuma Lake Fm.			NE ?
marine	Otherside Fm.	Otherside Fm.		
marine	Locker Lake Fm.	Locker Lake Fm.		S, E, NE
marine	Wolverine Point B	Wolverine Point B		S, E
marine	Wolverine Point A	Wolverine Point A	Wolverine Point A	S
marine		Lazenby Lake Fm.		S
nonmarine	Manitou Falls D	Manitou Falls D	Manitou Falls D	E, NE
nonmarine	Manitou Falls C	Manitou Falls C	Manitou Falls C	E, NE, S ?
nonmarine			Manitou Falls B	E, NE, S (minor)
Marine	Fair Point Fm.	Manitou Falls A (marine in part)		local; S, E ?

formation of supratidal evaporites. Similarly, the lack of debris flows in the alluvial fans suggest at least a moderately humid climate. The very well rounded sands and the local well-developed ventifacts in the conglomerates shows that there was strong eolian influence on the vegetationless fluvial plains.

Age and Duration of the Athabasca Group Deposition

The tuffaceous unit of the Wolverine Point B member was dated at 1428 ± 30 Ma (Rb-Sr whole rock; Armstrong and Ramaekers, in press) in two widely separated boreholes. A maximum age 1513 ± 24 Ma (Bell and Blenkinsop, 1980; Ramaekers, 1980a) were obtained on the same unit, suggesting a depositional age of approximately 1500 Ma.

Similar Phanerozoic sequences such as the Northern and Southern Basins in western Egypt (Van Houten, 1980) suggest the Athabasca Group may have been deposited in a 100-200 Ma period.

TECTONIC ENVIRONMENT OF THE ATHABASCA GROUP SEDIMENTS

The Athabasca Group with its thick, stacked transgressive sequences and alluvial fans indicates it was deposited in at least an intermittently active tectonic environment. The transgressive sequences are similar to the Cape Sebastian Sandstone (Upper Cretaceous) of the Oregon Coast (Bourgeois, 1980). Such units typically occur at tectonically active continental margins and are absent in more stable regions such as the modern Atlantic coasts (Bourgeois, 1980).

The increase in silica content of equivalent nearshore environments in successive transgressive sequences (Fair Point, Lazenby Lake, Locker Lake, Tuma Lake) suggests an upward-increase in the amount of reworking. This may be a consequence of slower rates of deposition caused by a temporal decrease in tectonic activity.

Diagenesis

Athabasca Group sandstones, like a number of the thick sandstones filling Proterozoic and later basins are now composed of orthoquartzites with a clay-rich matrix and a significant but variable hematite content. Hematite is ubiquitous in the Athabasca Group sandstones which commonly contain evidence of intense leaching and redeposition of this mineral. The surface rock is typically leached white, commonly to a depth of about 100 m, and other leached zones follow at depth. A limonitic leached zone (indicating relatively recent activity) is erratically present just above the basement contact.

Previous workers (Fahrig, 1961; Fraser et al., 1970) have interpreted the sandstones as supermature, multicycle sediments and suggested a broad, stable marine shelf covered much of northern Canada. However, sedimentological and stratigraphic data presented above suggest Athabasca Group sandstones accumulated rapidly in a tectonically active area. Petrographic evidence also suggests a complex and long diagenetic evolution has produced the first-cycle Athabasca Group orthoquartzites.

Clay pseudomorphs after detrital framework grains, the same size as adjacent quartz grains, are ubiquitous in the Athabasca Group sandstones. Pseudomorphs range from rare in the Manitou Falls D, to abundant in the Fair Point Formation. Where biotite has been altered to clay the warping of the contained diagenetic hematite layers indicates the flakes were distorted by varying amounts of compaction, ranging from zero to 100 per cent. Clay pseudomorphs after feldspars lose their identity much more rapidly. Where early cements or detrital clays impeded diagenesis (Douglas Formation, Wolverine Point B) fresh feldspar is preserved. Thus, at least part of the present interstitial clay in Athabasca sandstones has been produced by intrastratal solution followed by compaction, rather than infiltration. Hence, the sandstones were more polymict at the time of deposition than at present.

XRD studies of the clays and whole rock chemical analyses of the Athabasca sandstones (Fig. 13.8) support the petrographic data. In the Athabasca Group, Na concentrations are low and relatively constant (around 0.01 per cent). Thus the Na_2O/K_2O ratio reflects the amount of illite relative to kaolinite and is largely determined by diagenesis (Fig. 13.8). The Al_2O_3/SiO_2 ratio of the sandstones indicates their total clay content.

The Na/K ratio of the interstitial clays in the fluvial sandstones shows no marked correlation with total clay content and indicates illite/kaolinite mixtures of varying proportions with no obvious distribution patterns. However, in the marine sandstones (in particular the offshore marine sandstones) a strong correlation between interstitial clay content and high K values indicates that illite is the dominant clay.

The Al_2O_3/SiO_2 ratios of the sandstone samples of the various formations are interesting in that they are less affected by diagenesis than the Na/K ratios. They show an increase in silica content in the distal fluvial deposits, concomitant with a decrease in clay content downstream. The sandy facies of the neighbouring marine environments show an increase in clay content. In a modern fluvial or multicycle marine environment one might expect the opposite. However, the data are consistent with first-cycle deposition during the Proterozoic. At that time, fluvial abrasion and intense eolian activity on an unvegetated fluvial plain would preferentially remove minerals with good cleavage such as micas, feldspars, amphiboles and pyroxenes by decreasing their grain size relative to quartz. Differential abrasion during passage through high energy environments has been shown to effectively concentrate feldspars in low energy marine environments (Odom, 1975), thus accounting for the higher Al content in the marine facies.

Infiltration, while a possible mechanism to introduce detrital clays into the coarse fluvial deposits, is much less likely in a marine environment not subject to the constantly fluctuating hydraulic conditions present in fluvial systems.

If the interstitial clays of the Athabasca Group sandstones are considered as largely diagenetic, as suggested by the above data, then estimates of the original compositions of the sandstones may be made (see Pettijohn et al., 1972, p. 62). The original compositions of the sandstone facies of the various formations varied from arkose to orthoquartzite (Fig. 13.8).

Diagenesis of the Athabasca Group continues to the present (Ramaekers, 1979b, 1980b). Identifying the time at which some of these changes took place is difficult, as the chemical reactions involved are not restricted to a near-surface environment. For example, hematite coatings antedating quartz overgrowths, intense stratabound silicification and zones of quartz dissolution containing diaspore and illite may mark original paleosols, particularly in the fluviatile units. Postcompaction silicification with hematite precipitation in pores between quartz overgrowths on detrital grains mark changes after deep burial. Siderite deposition common near some fault zones, followed by decarbonatization, has resulted in a secondary porosity and the formation of still later hematite. Overgrowths on tourmaline are common, and their presence on crystals fractured during compaction indicates that they formed after deep burial.

Paleomagnetic data on sandstone samples (Fahrig et al., 1978; Ramaekers 1979b, 1980b), intense alteration of the intrusives in the Athabasca Basin, and young (1000 Ma; Rb-Sr) ages on clays in the Manitou Falls Formation (Armstrong and Ramaekers, in press) indicate intense diagenetic alterations in the Athabasca Group sediments for hundreds of millions of years after deposition.

ORIGIN OF ORTHOQUARTZITE FILLED BASINS

The Athabasca Group sandstones are first-cycle orthoquartzites. A quartz-rich source area, eolian activity in a vegetationless environment, and long and intense diagenesis have variably contributed to their formation. The Athabasca Group sandstones are similar in many respects to other thick orthoquartzites such as the Thelon, Sioux and Lower Hornby Bay sandstones, although details of stratigraphic sequence differ. Factors in common are the clay-rich quartz arenites, hematite content, moderately deep basins and an intracratonic setting. A structural link not previously noted is their spatial proximity to Hudsonian, Penokean, Wopmay orogenic zones (Fig. 13.9), and their time of formation, several hundred million years after the main orogenic events. This suggests that these 'orthoquartzite basins' developed in well-defined structural and temporal settings – areas of moderate instability between ancient cratons in a vegetationless environment. Modern analogues in character of fill, general stratigraphic sequence, vegetation-poor source areas and structural relations may be the Nubian basins of North Africa described by Van Houten (1980) which occupy a similar unstable zone between two, more stable, parts of the African craton.

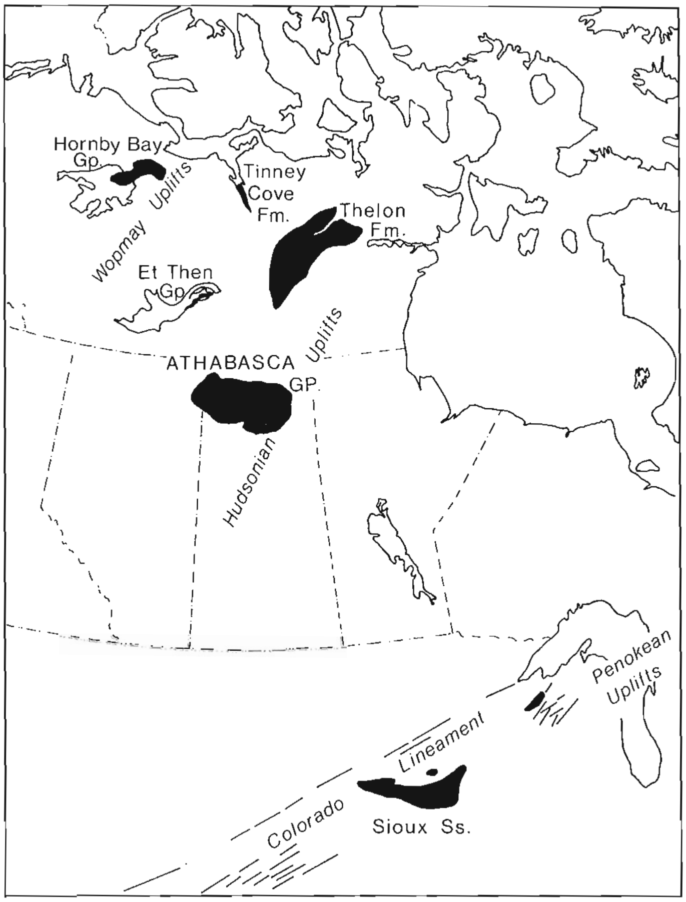


Figure 13.9. Distribution of Paleohelikian sandstones relative to early Proterozoic tectonically active belts.

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14. SEDIMENTARY HISTORY OF THE BELCHER GROUP OF HUDSON BAY

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Ricketts, B.D. and Donaldson, J.A., Sedimentary history of the Belcher Group of Hudson Bay; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 235-254, 1981.

Abstract

The Belcher Group provides a record of the evolution of the Ungava Craton margin in eastern Hudson Bay during middle Precambrian (Aphebian) times. Growth of an initial miogeoclinal stage, represented by a shallow, southwest-dipping marine platform containing evaporitic and stromatolitic dolostones (Kasegalik Formation), was terminated by extrusion of plateau-type basalts of the Eskimo Formation. The Eskimo volcanics probably are correlative with tholeiitic basalts of the Persillon Formation (Richmond Gulf) and may represent the products of incipient rifting in a zone that was transverse to the Ungava Craton; this zone is called the Richmond-Belcher Rift, and originated in earliest Proterozoic times.

A second miogeoclinal stage is composed of: (a) a transgressive carbonate platform-to-basin depositional phase (Fairweather-Laddie Formations) formed under conditions of gradual subsidence, and possessing clastic and stromatolitic dolostone including some spectacular platform-margin stromatolite buildups, and rhythmic, hemipelagic slope carbonates, succeeded by; (b) a prograding carbonate buildup (Rowatt Formation) and associated restricted basin depositional phase (banded ironstone of the Kipalu Formation). The ironstones can be correlated several hundred kilometres around the eastern periphery of Hudson Bay and represent deposition within an extensive arcuate basin. The latter phase indicates that deposition was no longer conformed to the Richmond-Belcher Rift, but embraced the eastern margin of Hudson Bay; a result of regional uplift prior to extensive submarine volcanism (Flaherty Formation). Building of the thick volcanic edifice, here interpreted as a volcanic arc, resulted in reversal of the paleoslope to east-dipping (possible volcanic equivalents extend from Ottawa Islands, south to the Sutton Inlier). A south- to southeast-trending exogeosynclinal basin formed between the volcanic arc and craton margin, and subsequently was filled by turbidites of the Omarolluk Formation (submarine fan facies) and arkosic red beds of the Loaf Formation (distal-molasse facies); the final depositional phase of the Belcher Group. Isoclinal folding occurred during the Hudsonian Orogeny, an event that is considered in terms of ocean-closing and continental collision tectonics.

Résumé

Le groupe de Belcher représente un vestige de l'évolution de la marge du craton d'Ungava dans l'est de la baie d'Hudson pendant le Précambrien moyen (Aphébien). Le développement d'une phase miogéosynclinale initiale, représentée par une plate-forme marine peu profonde plongeant vers le sud-ouest et contenant des dolomies évaporitiques et stromatolitiques (formation de Kasegalik), s'est terminé par l'extrusion des basaltes de plateau de la formation d'Eskimo. On peut probablement corréler les roches volcaniques d'Eskimo avec les basaltes tholéiitiques de la formation de Persillon (golfe de Richmond); elles représentent peut-être les produits de la formation initiale d'un rift dans une zone transversale par rapport au craton d'Ungava; cette zone est appelée rift de Richmond-Belcher, et date des temps protérozoïques les plus reculés.

Une seconde phase miogéosynclinale se subdivise en: (a) une phase de sédimentation carbonatée, transgressive, marquant le passage d'un milieu de plate-forme à un milieu de bassin (formations de Fairweather et de Laddie), indiquant par conséquent une subsidence graduelle, et contenant des dolomies clastiques et stromatolitiques, en particulier des édifices stromatolitiques spectaculaires en bordure de la plate-forme, et des couches carbonatées rythmiques de caractère hémipélagique, inclinées, suivies; (b) de sédiments carbonatés progrades (formation de Rowatt) et d'une phase associée de sédimentation en bassin fermé (niveaux ferrifères stratifiés de la formation de Kipalu). Les niveaux ferrifères stratifiés peuvent être suivis sur plusieurs centaines de kilomètres, dans l'est du périmètre de la baie d'Hudson; ils se sont déposés à l'intérieur d'un vaste bassin en arc de cercle. La seconde phase montre que la sédimentation ne se limitait plus au rift de Richmond et Belcher, mais englobait le rebord est de la baie d'Hudson, par suite d'un soulèvement régional antérieur à un volcanisme sous-marin important (formation de Flaherty). L'édification du puissant ensemble volcanique, considéré dans cet article comme un arc insulaire, a inversé la pente initiale qui plonge maintenant vers l'est (les équivalents volcaniques possibles prolongent les îles Ottawa, au sud de l'enclave de Sutton). Un bassin exogéosynclinal orienté sud à sud-est s'est formé entre l'arc insulaire et le craton, et a ensuite été comblé par des turbidites de la formation d'Omarolluk (faciès de cône alluvial sous-marin) et les red beds arkosiques de la formation de Loaf (faciès distal de molasses); ceci correspond à la phase sédimentaire finale du groupe de Belcher. Pendant l'orogenèse de l'Hudsonien ont eu lieu des plissements isoclinaux, évènement qui entre dans le cadre de la tectonique globale (fermeture des océans et collisions continentales).

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INTRODUCTION

The Belcher Islands was one of the last groups of islands in Hudson Bay to be systematically explored, although various sightings of the islands had been reported well before the end of the last century. The first geological expedition was undertaken by Flaherty (1918), who soon after was followed by several other workers (Moore, 1918; Woodbridge, 1922; Young, 1922). Moore (1918), working mostly in the Laddie Harbour area, proposed the name Belcher Series for strata exposed in the Belcher Islands. An important discovery at this time was the presence of possible algal structures (stromatolites). Moore is also credited with the first report of microfossils from Precambrian rocks.

The first detailed mapping of the Belcher Islands was undertaken by Jackson (1960); his map has provided a sound stratigraphic framework for subsequent studies. Jackson afforded group status to the sequence and subdivided it into 16 units. These units subsequently were given formation status by Dimroth et al. (1970). K-Ar age dating of volcanic rocks collected by Jackson confirmed the Aphebian age of the Belcher Group, and an age of about 1.76 Ga (Rb-Sr) was subsequently obtained by Fryer (1972) from volcanic rocks and shales of the Flaherty and Omarolluk formations respectively. Precambrian microfossils associated with stromatolites in the Kasegalik and McLeary formations were studied by Hofmann and Jackson (1969). A diverse assemblage of filamentous and coccoid cyanophytes, morphologically similar to some Gunflint Formation and Bitter Springs Formation forms, have been reported by Hofmann (detailed systematics in Hofmann, 1976, and several other papers). Stromatolites from the McLeary and Mavor formations have also received attention (Donaldson, 1976).

Paleocurrent determinations in pre-Flaherty strata by Jackson (1960) and Dimroth et al. (1970) indicated that the paleoslope probably dipped west to southwest. This was confirmed in a later study by Barrett (1975). A geochemical and petrographic study of the Eskimo Formation volcanics was undertaken by Stirbys (1975); volcanic and volcaniclastic rocks of the Flaherty Formation also have been examined by Schenk (1959), Leggett (1974) and Ware (1978).

In the first general synthesis of the Belcher Group, Hofmann and Jackson (1969) recognized three major episodes of sedimentation and volcanism that took place before the Hudsonian Orogeny, and a fourth episode defined by intrusion

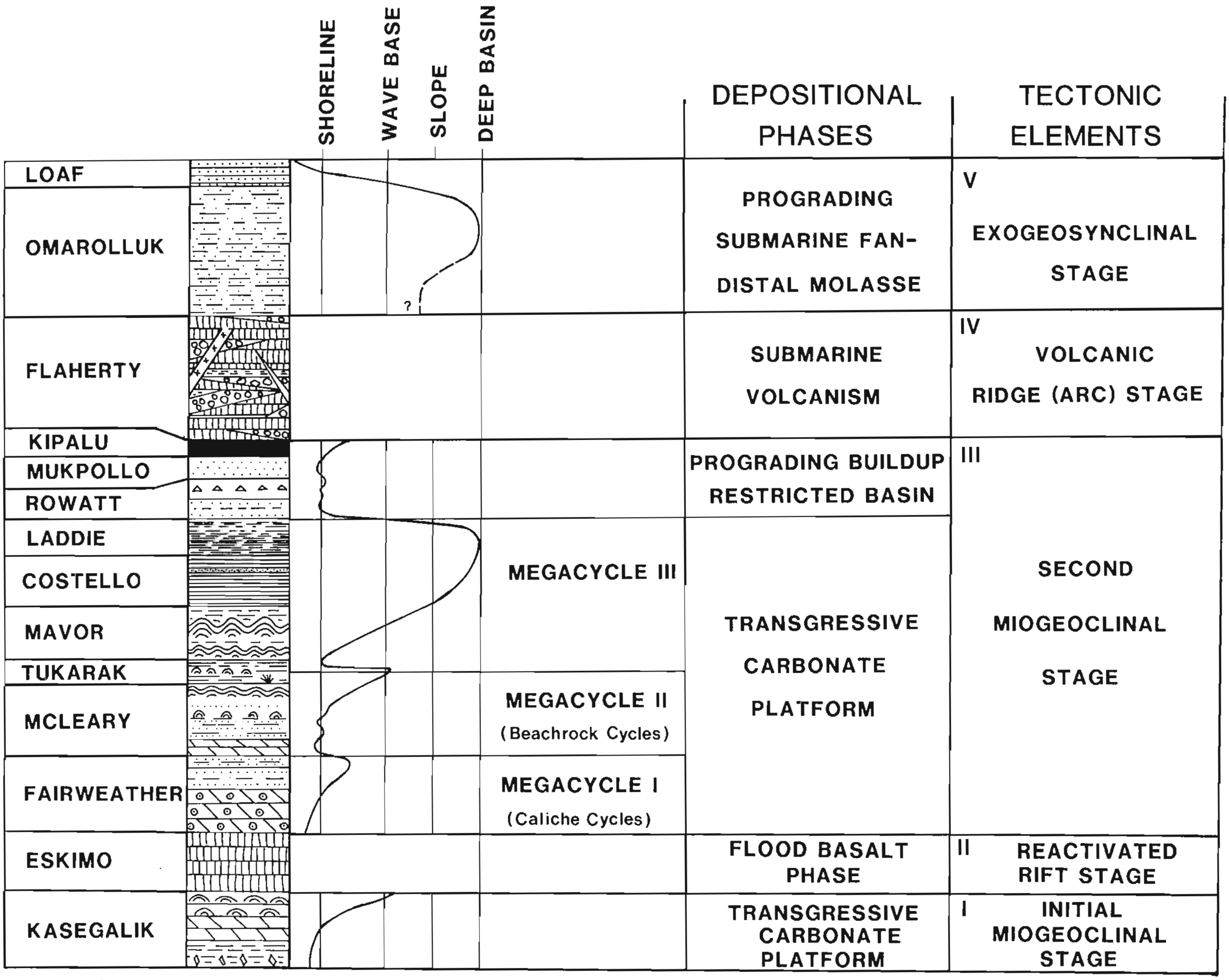


Figure 14.1. Summary of depositional phases and tectonic elements represented by the Belcher Group.

of a few gabbro dykes. This last episode was related to a mild thermal or tectonic event during the Grenvillian Orogeny. Recent investigations (Ricketts, 1979) have permitted refinement of this subdivision and in the present paper, a number of depositional phases are defined (Fig. 14.1).

I. THE FIRST TRANSGRESSIVE PLATFORM PHASE

Kasegalik Formation

This, the lowest stratigraphic unit in the Belcher Islands, is discontinuously exposed along the Kasegalik anticline and Churchill Sound, where it attains a thickness of more than 1200 m. Previously unmapped outcrops are also exposed in a glacially carved valley in Central Tukarak Island. In the northern Churchill Sound area, Bell and Jackson (1974) subdivided this formation into five zones: (a) a lowermost zone consisting of grey dolostone and 20 per cent red mudstone; (b) 60 per cent red mudstone; (c) stromatolitic dolostones; (d) cherty stromatolitic dolostone; (e) argillaceous and tuffaceous dolostones near the Eskimo Formation contact. The above subdivision is extended to Kasegalik Lake sections in central Flaherty Island. The lower two zones (a,b) contain halite casts and sulphate moulds, desiccation cracks and tepee structures, all indicative of prolonged subaerial exposure, possibly in a sabkha-like environment. Cryptalgal structures in zones c and d exhibit a general upwards progression from small low mounds containing flat and digitate mats, to large domal structures (2-5 m across) containing some furcate and digitate branching. Interbedded clastic dolostones contain flat-pebble conglomerates, including vertically stacked varieties resembling stone rosette structures that are considered to have formed in a wave-washed environment (Ricketts and Donaldson, 1979).

The overall lithologic sequence in the Kasegalik Formation indicates deposition on a shallow marine platform with water depths gradually increasing from supratidal conditions (sabkha) at the base of the succession, to shallow subtidal conditions near the top. Sedimentation was interrupted by extrusion of thick basalt flows that make up the Eskimo Formation.

II. PLATEAU-TYPE VOLCANISM PHASE

Regional trends of the Eskimo Formation have been described by Jackson (1960) and in Dimroth et al. (1970). The Eskimo Formation ranges in thickness from a maximum of 900 m in the Eskimo Harbour region to about 600 m in central Tukarak Island, thinning southward and westward to only a few metres. The succession consists of a stack of columnar-jointed basalt flows that attain thicknesses of 20 m. A few pillow lavas occur, but unlike many in the Flaherty volcanics, these are of limited lateral extent. The tholeiitic basalts are fine grained, consisting predominantly of ophitic plagioclase and augite (Stirbys, 1975). A few prophyritic and variolitic basalt flows occur.

Relatively minor interflow sediments include thin red and green argillites, cherts, and thin beds of poorly sorted lapilli tuff that overlie flows, infilling shallow fractures formed by cooling of the flow surface. The rims of angular volcanic fragments in these deposits display reddish brown staining suggestive of subaerial weathering. Authigenic sphene, a common accessory mineral in the matrix, provides additional evidence of subaerial weathering of the volcanics. In these respects, the Eskimo volcanics are different from those in the Flaherty Formation, higher in the Belcher succession. On Tukarak Island, the top of the Eskimo Formation is characterized by massive agglomerates intercalated with thin tuffs and basalt flows. The agglomerates contain a mixture of angular and poorly sorted basalt and argillite blocks, and are possibly flow-front breccias.

On the basis of chemistry, a preponderance of jointed flows, and evidence of subaerial weathering, Stirbys (1975) concluded that the Eskimo volcanics were extruded as flood-type basalts. The overall thinning trends of the formation, together with the asymmetry of some columnar joints, suggest that eruption fissures were east and northeast of the Belcher Islands. The absence of pyroclastics precludes the likelihood of phreatomagmatic-type eruptions.

Metamorphism in the Eskimo Formation attained a grade of prehnite-pumpellyite to subgreenschist facies. This grade is slightly higher than that observed in the Flaherty volcanics, probably as a consequence of greater depths of burial.

III. THE SECOND TRANSGRESSIVE PLATFORM PHASE

The episode of effusive volcanism of the Eskimo Formation appears to have ended almost as abruptly as it started; tuffs are rare in all but the lower few metres of the overlying Fairweather Formation. Nevertheless, epiclastic volcanic fragments are ubiquitous in sandstones of the Fairweather, indicating that the volcanics were exposed during the early stages of this phase.

The second platform phase consists of a succession of almost 2000 m of carbonates and siliciclastics that has been subdivided into six formations, each documenting the evolution of distinctive sedimentologic and biologic facies.

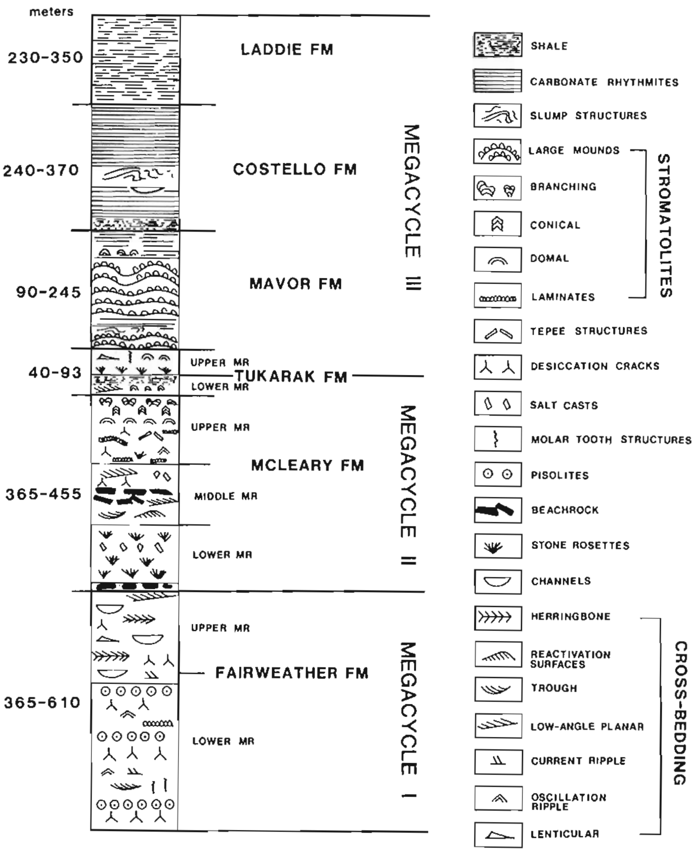


Figure 14.2. Schematic representation of formations composing the second platform phase of the Belcher Group.

Detailed analysis of these facies has also demonstrated that this succession is composed of three successive transgressive 'megacycles' (Fig. 14.2).

Megacycle I

Fairweather Formation

In the original type section, Fairweather strata are poorly exposed (Dimroth et al. 1970); a reference section is proposed at the southeast corner of Tukarak Island, 10 km southwest of McLeary Point, where at least 338 m of almost continuous exposure of the Fairweather Formation occur.

The Fairweather Formation is informally subdivided into two new members (Fig. 14.2): a Lower member consisting predominantly of interbedded pisolitic dolostones, sandstones and red mudstones; and an Upper member consisting of sandstones, siltstones and thin grit beds. The Eskimo-Fairweather contact is a sharp but slightly irregular disconformity, and is characterized by lenticular beds of banded ironstone that occur in depressions of the Eskimo surface up to 2 m deep. Jackson (1960) also reported an occurrence of granule-bearing jasper at a similar stratigraphic level at the northern end of Churchill Sound. The ironstones are interpreted to have formed in small lakes or lagoons near the seaward limit of the exposed Eskimo volcanics, isolated from the open sea by a broad intertidal apron. Iron-rich waters were supplied by runoff from the adjacent volcanic plateau.

Lower member. The dominant feature of this member is a series of tabular, buff pisolitic dolostone beds, each forming the upper part of shoaling-upwards cycles that are 1-30 m thick (Fig. 14.2). Each cycle is divided into:

1. tabular and lensoidal arenites (mainly subarkoses) and siltstone beds that contain a variety of sedimentary structures indicative of a tidal flat environment (herringbone crossbeds, reactivation surfaces, interference ripples and flaser bedding). Crossbed azimuths generally indicate bipolar-bimodal current directions. Interbedded red mudstones commonly display desiccation cracks superimposed on sets of interference ripples.
2. overlying pisolitic beds which, in each cycle, display features similar to more recent examples of vadose pisolites (the pisoliths display composite growth forms, polygonal fitting, geotropic elongation and some inverse grading; see discussion in Esteban, 1976). Lower contacts of these beds are diffuse, grading into the subjacent mudstones, whereas the upper contacts typically show channeling beneath the overlying sandstones. There is no evidence of bedload transport or reworking of pisoliths into overlying sediments; petrographic evidence demonstrates that multiple stages of dissolution-precipitation occurred in individual beds, a feature characteristic of processes occurring in the vadose zone.

The vadose pisolites formed during periods of prolonged exposure in the upper limits of a broad supratidal apron and bears some resemblance to the coniatolite belt (supratidal tuffas) forming at present in Persian Gulf (Purser and Loreau, 1973). Here, the associated mudstones and sandstones in each cycle represent laterally-equivalent tidal flat deposits.

Upper member. The Upper member consists of red and green siltstones and sandstones cut by quartz arenite-filled channels up to 80 m across. Paleocurrent directions are again bimodal; some interbedded mudstones contain desiccation cracks. Upper member rocks probably formed in lower intertidal, sand and mud flats, probably inundated by slightly deeper water than sediments of the Lower member.

The top 15 to 20 m of this member consist of tabular-bedded quartz arenites that mark a brief return to shallower water conditions. This uppermost segment is capped by a distinctive beachrock-bearing unit at the base of the overlying McLeary Formation.

Megacycle II

McLeary Formation

Strata of the McLeary Formation in general record the evolution from a clastic-dominated coastal apron (Lower and Middle members), to a coast that was characterized by prolific growth of algal buildups (Upper member). The Upper member of the McLeary records a gradual transition from a supratidal to shallow subtidal environment (above wave-base).

The McLeary Formation is best known for its diverse stromatolite flora, first illustrated by Moore (1918), and later by Donaldson (1976) and Golubic and Hofmann (1976). Hofmann (1974, 1975, 1976) and Hofmann and Jackson (1969) also have identified a diverse and well-preserved microflora assemblage from cryptalgal structures in the Upper member of this formation.

Lower member. This lithologically distinct unit is well exposed on Tukarak and Mavor islands, but has not been observed along the Kasegalik anticline. On Flaherty Island, basal strata of the McLeary Formation are similar in most respects to Middle member rocks, and hence the Lower member may pinch out west of Tukarak Island.

Contact with the underlying Fairweather Formation is conformable. On Tukarak Island, this boundary is marked by a distinctive grainstone bed that contains good exposures of beachrock (Donaldson and Ricketts, 1979), and will henceforth be referred to as the Beachrock Marker Bed. This bed is the culmination of a brief period of shoreface progradation that commenced during deposition of the upper part of the Fairweather Formation. The remainder of the Lower member (up to 140 m thick) consists of thinly bedded dololutites intercalated with lenses of edgewise-stacked, intraformational conglomerates displaying all the features characteristic of stone rosette structures. On bedding surface exposures stone rosette pavements can be traced for several tens of metres. A few beds containing gypsum casts provide additional evidence for periodic exposure, evaporation and desiccation.

Middle member. The return to a clastic-dominated high-energy shoreline represented by the Middle member is marked by the appearance of a number of sandstone- to dolograinstone cycles terminated by beds of beachrock. Each of these beachrock cycles represent a brief period of shoreface progradation. However, the upper few metres of this member exhibit a progressive change in lithology to interbedded dololutites and dolarenites. The transition is accompanied by the appearance of desiccation structures (mudcracks and tepee structures), pisolitic beds, edgewise conglomerates, vague cryptalgal laminates and sparse gypsum casts.

Upper member. For convenience, the Upper member of the McLeary Formation is divided into three broad zones. This subdivision is not intended to be formal.

Table 14.1. Comparison of cryptalgal laminate types and sedimentary structures in the Upper Member, McLeary Formation.

		DESICCATION CRACKS	TEPEE STRUCTURES	GYPSUM CASTS	RIP-UPS	EDGEWISE CONGLOMERATES	PULL-APARTS (JOINTS)
UPPER ZONE	BRANCHED STROMATOLITES	R			R		
	CONOPHYTON	R			R		
MIDDLE ZONE	ELONGATE DOMES (NON-BRANCHING)	R			C	C	R
LOWER ZONE	STRATIFORM DIGITATE LAMINATES	R			R		R
	PARALLEL LAMINATES	C	C	R	C	C	C
	WAVY LAMINATES	C	C		C	C	C
	TUFTED LAMINATES	C		R	C	C	C
	ONCOLITES				C	C	

INCREASING EXPOSURE ↓

Lower Zone: The transition from the Middle to the Upper member is gradational over several metres. In addition to the sedimentary structures noted in underlying members, the tabular dolarenites and dololutites also contain dololutite pull-aparts (probably originating as penecontemporaneous joints in preferentially lithified crusts), stone rosettes and a few quartz arenite-filled channels, all indicative of shallow water conditions with extended periods of exposure. Towards the top of this zone a variety of cryptalgal laminates appear. Detailed analysis reveals a specific association of inorganic sedimentary structures with each type of algal mat, which in turn are correlated with the degree of desiccation and water depth (Table 14.1).

Cryptalgal laminates in the McLeary Formation compare remarkably well with a variety of forms described from some modern tidal flats, for example Shark Bay and Andros Island (Logan et al. 1974; Hardie and Ginsburg, 1977), wherein a similar zonation of cryptalgal morphotypes has been recognized. The overall succession of morphotypes observed in the McLeary, together with the vertical zonation of associated inorganic sedimentary structures, demonstrates a gradual increase in water depth within an intertidal environment.

Middle Zone: The most important feature in this zone is the appearance of low synoptic relief domes composed of non-branching stromatolites. The domes become more abundant towards the top of the sequence. Dome elongation (L:W) ratios are commonly as high as 4:1. Hoffman (1967) has demonstrated that stromatolite mound elongations generally parallel the direction of wave propagation and hence provide valuable criteria for determining the orientation of ancient shorelines. In the McLeary Formation, the general trends of stromatolite elongation closely parallel the principal current directions determined from crossbeds, indicating that regional strike of the paleoshoreline was approximately northwest.

Upper Zone: The acme of morphological development and abundance of stromatolites in the McLeary Formation is achieved in this zone. The diverse stromatolite assemblage consists of bulbous, furcate, digitate, dendroid and conical forms. The minor interbedded sediments display rare desiccation cracks, interference ripples and rip-up clasts.

In the uppermost 12 m of this zone, a broad morphological zonation of stromatolites is recognized wherein all four supergroups, as defined by Raaben (1969), are present (Fig. 14.3). All the major groups recognized, except the conophytonids, were previously interpreted as temporally restricted to the Riphean or Vendian. Within the group **Conophyton**, some Belcher forms closely resemble the form **C. garganicus**, thus extending its time-range into the Aphebian.

The top 2 m of the Upper Zone is a conspicuous buff dolostone consisting entirely of closely packed columnar stromatolites (cf. **Inzeria**) which form parts of large, interconnected elongate mounds with amplitudes up to 1 m, and intermound spacings up to 10 m. This algal buildup (reef) can be traced throughout most of the Belcher Islands, extending over a known area of about 5000 km^2. The reef exhibits few changes in internal structure and content throughout its extent.

Tukarak Formation – Lower member

The upper limits of the second megacycle are preserved in strata of the Lower member of the Tukarak Formation. Typical lithologies include fine grained immature sandstones, dolomitic sandstones, grey shaly mudstones and minor stromatolitic dolostones. Stromatolite mounds occur sporadically near the base of this member, and mound elongation trends are consistent with those observed in the McLeary Formation (i.e. southwest). Crossbed azimuths are commonly bimodal but have a dominant southwest trend. There is no evidence of subaerial exposure.

The Lower member thus represents the maximum water depths attained during this transgressive cycle; here, shales near the top of the member are probably indicative of deposition at or below wave-base.

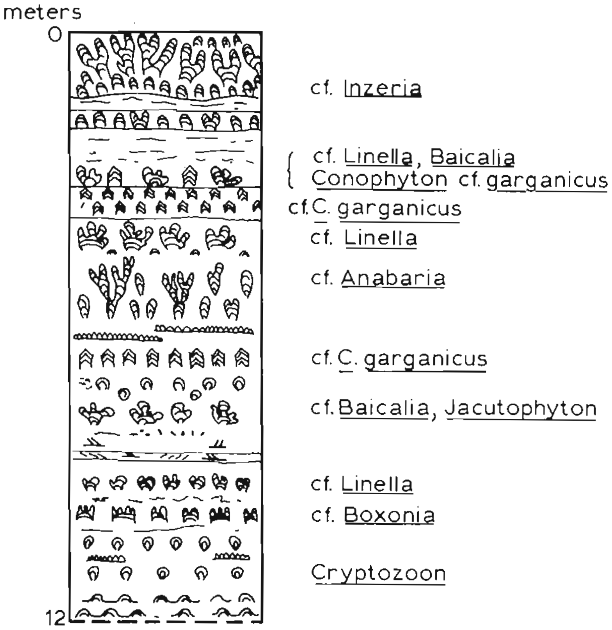


Figure 14.3. Stromatolite zonation in the upper 12 metres of the McLeary Formation.

Megacycle III

Tukarak Formation – Upper member

A brief return to shallow-water conditions is marked by the appearance of a particularly distinctive sequence of thinly bedded, brick-red mudstones and dolostones containing lenticular and flaser bedding, and abundant stone-rosette pavements exposed on bedding planes. The latter structures suggest deposition in a shallow water, swash-backwash zone. A few thin stromatolitic dolostones appear near the top of the member. One particularly distinctive bed has been mapped over central and eastern Belcher Islands and provides an excellent stratigraphic marker. Small stromatolite bioherms exhibit morphology and elongation trends similar to those in the Lower member of the Tukarak Formation.

Mavor Formation

The Mavor Formation is characterized by the appearance of thick, laterally extensive stromatolitic dolostones and thin shaly lutites. Most sections on Tukarak Island display three distinct major stromatolite buildups, with thinner stromatolite-bearing units occurring higher in the sequence.

However, an exceptionally thick unit (146 m) west of Laddie Harbour is almost continuously exposed over the entire length of eastern Tukarak Island. It is in these sections that the nature of what must have been remarkably extensive stromatolite buildups becomes apparent. The individual mounds composing each buildup range in synoptic relief from 30 cm at the base, increasing to 200 cm at the top of the unit. This trend is accompanied by a concomitant increase in lateral mound spacing from 1-2 m at the base, to 20 m near the top. Thin laminated and shaly dololutites commonly drape over the mounds and thicken in adjacent troughs. The mounds commonly coalesce 'up section' into single larger mounds, with inheritance of individual mounds continuing through 100 m. Consistent east-northeast elongation directions are present in these sections – a more easterly component than elongations in the underlying transgressive cycle.

Internally, the Mavor mounds consist of digitate stromatolites associated with forms identical to the Riphean taxon **Tungussia**; superposed 'second order' mound structures have elongation directions identical to the larger mounds.

No evidence of shallow-water deposition has been observed in the Mavor stromatolite facies, and hence growth of these extensive buildups probably took place in deeper water. Mound elongation is interpreted to have developed in response to offshore tidal currents.

Shallow-water sediments, laterally associated with the stromatolite facies, at the northern end of Tukarak Island and the southern part of Eskimo Harbour, are informally referred to as the Salty Bill Hill member and Sanikiluaq member respectively. The mainly crossbedded grainstones of this facies display stone rosettes, thrombolitic structures, and beachrock at one locality in the Sanikiluaq member.

Costello Formation

Shaly dololutites at the top of the Mavor Formation pass conformably into a grey argillaceous shale, 10-20 m thick, the base of which defines the lower boundary of the Costello Formation. The remainder of this formation consists predominantly of rhythmically bedded dololutites and calcarenites with shale partings. The lowermost beds, up to 15 centimetres thick, are laminated, and some are graded. Lunate and straight-crested ripples indicate west-flowing currents. Other lithologies include thin graded calcarenites commonly overlain by crossbedded calcilutites, intraformational conglomerates and slump folds. These all occur in a well-defined zone that can be mapped throughout most of eastern Belcher Islands. Structures in the thin graded calcarenites resemble the A, B and C divisions of Bouma cycles, although complete Bouma cycles are rarely present in the Costello. The intraformational conglomerates occur in small channels, and are particularly interesting because they contain flat, crudely imbricated calcilutite slabs up to 25 cm in length. These may have been derived by erosion of local submarine carbonate hardgrounds, similar to a number of modern, relatively deep-water hardgrounds noted recently (e.g. Schlager and James, 1978).

The array of lithologies and sedimentary structures in the Costello Formation is similar to carbonate rhythmite sequences that are considered by J.L. Wilson (1969) and McIlreath and James (1979) to be characteristic of foreslopes of carbonate platforms. Most of the detrital carbonate and minor siliciclastic silt (hemipelagic deposits) were probably derived from the adjacent marine platform. Dolomitized ooids that are a common constituent of the calcarenites almost certainly were derived from shallow water, and transported into a deeper water environment by turbidity currents.

Laddie Formation

A change in the definition of the Laddie-Rowatt Formation boundary leads to a reduction in Laddie Formation thickness from 330 m to 230 m at its type section (see discussion of the Rowatt Formation below). As a result, there is little need for subdivision of the Laddie Formation into members, as the redefined formation consists predominantly of red, and intercalated red-green, argillites and shales. The contact with the underlying Costello Formation is gradational over several metres.

Summary of the Third Phase

The third depositional phase (Fig. 14.1) of the Belcher Group consists of three megacycles: the first is characterized by supratidal pisolite and intertidal to shallow subtidal siliciclastic deposits. The second consists of both supratidal and intertidal clastic carbonates, together with abundant shallow-water stromatolitic dolostones. The third megacycle is dominated by extensive subtidal algal buildups (reefs), and carbonate rhythmites similar to Phanerozoic slope deposits.

Each megacycle represents deposition on a shallow-marine platform under conditions of increasing water depths. However, the third megacycle records the transition from a marine platform to foreslope and relatively deep basin (Costello-Laddie formations). Here the extensive algal bioherms of the Mavor Formation represent platform margin buildups. As facies transitions between the Mavor and Costello formations are gradational, the transition from platform to adjacent foreslope was most likely gradational. Thus, the platform margin probably corresponds to what McIlreath and James (1979) have called a depositional margin. Transitions from shoal-water platform to deeper slope environments have been documented in Cambro-Ordovician carbonate sequences in Nevada (Cook and Taylor, 1977), the Central Appalachians (Reinhardt, 1977), and also in the Lower Proterozoic Pethei Group, Great Slave Lake (Hoffman, 1974). However, in all these cases the successions are regressive. The thick transgressive sequence represented by Fairweather-to-Laddie strata is therefore unusual in this respect. A closer analog is the transgressive

Silurian to Devonian Helderberg Group of New York State, which records the transition from shallow water intertidal to deeper outer shelf environments (Laporte, 1969).

IV. PROGRADING CARBONATE BUILDUP – RESTRICTED BASIN PHASE

This phase contains three formations: the Rowatt, Mukpollo and Kipalu formations. The transition from the preceding transgressive platform regime marks important changes in tectonism, depositional environment and paleo-coastline configuration.

In Jackson's original definition (Jackson, 1960, restated in Dimroth et al., 1970), the base of the Rowatt Formation was placed at the contact between well-bedded sandstones and a distinctive buff dolostone unit on Tukarak Island (Fig. 14.4). However, the dolostones pinch out to the west, and in central Flaherty Island they are replaced by a completely conformable sequence of sandstones; the above criterion thus becomes untenable in western Belcher Islands. A new definition is proposed: the base of the Rowatt Formation is placed at the first sequence of tabular, cross-bedded sandstones that overlie red and green argillites typical of the Laddie Formation. This proposal requires changes to the internal subdivision of the Rowatt Formation (summarized in Fig. 14.4).

Rowatt Formation

Lower member

The overall vertical and lateral facies relationships in this member are illustrated in Figure 14.5. The Lower member is characterized by terrigenous clastics, although clastic carbonates become increasingly important from the middle of the sequence upward towards the dolostones of the Upper member. Lower strata consist of numerous sandstone-shale cycles, up to 150 cm thick, that contain a variety of sedimentary structures indicative of asymmetric tidal currents (herringbone crossbeds, reactivation surfaces, large-scale ripples with smaller current ripples superimposed at right angles, interference ripples and double crested ripples). Superposition of ripple trains, a relationship commonly produced as a result of late-stage runof during ebb tides, is abundant. Crossbed azimuths are commonly bimodal or quadrimodal. Interbedded shales and siltstones contain lenticular bedding, interference ripples and desiccation cracks, and are consistent with deposition on an upper tidal flat.

A notable feature exhibited by many large dune structures (up to 100 cm thick) is their predominantly northwest direction of transport. This is almost at right angles to the principal bimodal directions of tidal bed-load transport indicated by other sedimentary structures (northeast and southwest). These dune structures probably formed in response to longshore currents that flowed approximately northwest, parallel to the ancient shoreline.

The spectrum of sedimentary structures in the Lower member is analogous to that observed in many siliciclastic tidal flat and sand-bar environments (e.g. Klein, 1970). Based on Walther's Law of Correlation of Facies, the vertical succession of structures and lithologies in each cycle is identical to that which is predicted for deposition under conditions of shoreline progradation.

The increase in dolarenites, thin dololutites, channel-fill conglomerates, beachrock and laterally associated beachrock conglomerates towards the top of this member is accompanied by an increase in sedimentary structures

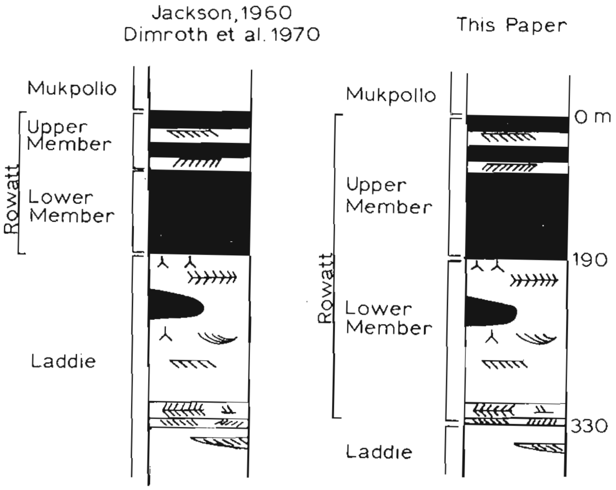


Figure 14.4. Proposed changes in the Rowatt Formation-Laddie Formation boundary.

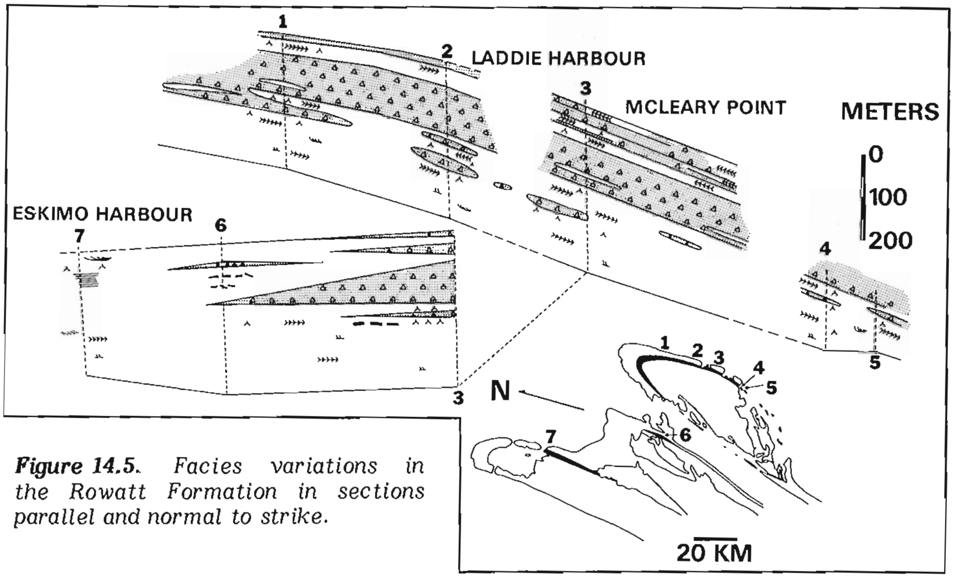


Figure 14.5. Facies variations in the Rowatt Formation in sections parallel and normal to strike.

indicative of subaerial exposure, such as desiccated mudstones, runsel marks and rill structures. Most sedimentary structures in this member also persist throughout the clastic dolostones, and an intertidal environment can be inferred. A number of beachrock conglomerates near the top of the member likely formed as beach ridges during storm surges. By comparison, many channel-fill conglomerates contain clasts whose lithologies are identical to dolostones in the overlying Upper member – a feature of prime importance for the overall interpretation of lateral facies relationships.

A number of isolated dolostone bodies in the Lower member (Fig. 14.5), up to 400 m in length, display most of the features attributed to Upper member dolostones, and are included in the discussion of that member to avoid repetition.

Upper member

Distinctive buff dolostones of the Upper member form prominent ridges on Tukarak Island. However, as shown in Figure 14.5, these dolostones pinch out west of Tukarak Island and are transitional into terrigenous and carbonate clastics similar to those observed in the Lower member on Tukarak Island.

The dolostones are composed of what probably were well-bedded sandy grainstones interbedded with banded dololutites, but which now display various degrees of brecciation and folding. Most of the grainstone beds are less than a metre thick and display subhorizontal laminae and planar crossbeds. These beds can be traced more than 100 m along strike before pinching out, or before bedding is destroyed by brecciation. Relict meniscus cements suggest that cementation of the grainstones was in part vadose.

Numerous intervals (each up to a metre thick) of thinly bedded dololutites are interbedded with the grainstones in a crude cyclical fashion. Characteristically, the lutites have been plastically deformed into small isoclinal folds (amplitudes up to 30 cm). The synsedimentary nature of this deformation is clearly demonstrated where folds are truncated by the overlying crossbedded grainstones.

The grainstones and folded lutites exhibit varying degrees of brecciation ranging from relatively mild, with little relative movement of displaced blocks (fractures parallel and normal to bedding), to zones of chaotic angular blocks. Much of the brecciation clearly occurred in situ, at several stages following deposition and lithification of the carbonate sediments In particular, the style of brecciation in the grainstones demonstrates that a considerable degree of lithification must have existed in these beds at the time of deformation. Fractures that developed as a result of brecciation were filled with carbonate mud; refracturing along old planes of weakness took place, and fracture walls were lined with several generations of fibrous cements, now replaced by radiaxial fibrous dolomite.

Zones of chaotic breccia are commonly overlain by, or grade laterally into, dolostones that exhibit very little deformation, other than folding in the dololutites, indicating that brecciation was essentially penecontemporaneous. This deformation was confined entirely to the Rowatt dolostones and did not affect adjacent rocks.

One isolated dolostone body of the Lower member on eastern Tukarak Island was interpreted by Barrett et al. (1976) as an allochthonous debris flow. However, detailed examination of the Rowatt Formation in the present study has revealed a number of criteria that suggest an alternative interpretation: (i) contact between the dolostones and Lower member clastics is either sharp (but conformable) or gradational – where gradational a continuous sedimentary sequence is implied; (ii) dolostones in the Upper and Lower members overlie mudrocks and beach deposits that form the upper limits of a shoaling-upward, clastic intertidal apron; (iii) deformation is confined entirely to the dolostones; (iv) relict cements in the grainstones indicate that the initial stages of cementation took place above the water table.

It is suggested here that the Upper member dolostones formed as positive physiographic elements relative to mean sea level, and that they represent an ancient carbonate buildup complex that was bordered on its seaward margin by an intertidal apron. In this model, sandy grainstones in the Upper member may have accumulated as spillover lobes, thus providing a mechanism for vertical accretion of the buildups. A mechanism of penecontemporaneous solution collapse, or possibly karstification, is proposed to account for brecciation of the carbonates.

The Upper member dolostones are cut by channels, some more than 200 m wide, that are filled with well-sorted, trough crossbedded grainstones. Paleocurrents in these channels are strongly unimodal, indicating westward flow (i.e. downslope). Channel margins are locally rimmed by beachrock. Unlike the other dolostones in this member, the channel deposits show only minor brecciation. This contrast is ascribed to the absence of early, rapidly-formed cements in the bulk of the channel sands.

Mukpollo Formation

Although the Mukpollo Formation is characterized by quartz arenites, it is poorly exposed. On Tukarak and Innetalling islands, contacts with the subjacent Rowatt Formation and overlying Kipalu Formation are sharp but conformable. On Flaherty Island the Rowatt-Mukpollo contact is gradational over several metres.

The lower half of the Mukpollo consists of well-bedded quartz arenites and purple siltstones, with a few dolomitic sandstones near the base. Tabular-planar crossbedding is abundant in the coarser lithologies, but herringbone crossbeds and a variety of ripples also occur. Although crossbed measurements indicate predominantly northeast-trending currents, the distribution is weakly bimodal. Thin interbedded siltstones commonly display evidence of desiccation.

Small quartz arenite-filled channels containing basal lag conglomerates cut siltstones in several places. Two notable constituents of the lags are chert granules and ironstone pebb' both of which are remarkably similar to ironstones in the overlying Kipalu Formation. This important occurrence provides evidence for lateral association of these two units.

The upper 50 to 60 m of the Mukpollo consist of massive quartz arenites with rare shale partings. Festooned planar crossbeds and large ripples are common.

In general, Mukpollo strata are interpreted as deposited under shallow-water to intertidal conditions, possibly on sand flats, sand bars or shallow subtidal shoals.

Kipalu Formation

The type area for this unit, originally named the Keepalloo Formation by Moore (1918), flanks the east side of Kipalu Inlet. It is from these rocks that Moore compiled the first report of possible microfossils in Precambrian rocks, although Hofmann (1972) regarded them as dubiofossils. LaBerge (1967) also has illustrated a number of spheroidal structures of possible algal origin from samples of Kipalu chert.

Three major rock types occur in the Kipalu Formation: laminated ferruginous argillites, laminated micrites, and granule-bearing jasper. Minor tuffaceous argillites occur at the top of the Kipalu Formation. This lithological subdivision gives rise to three principal depositional subfacies:

1. Banded Argillite Subfacies: consists mainly of highly fissile ferruginous argillites. The grain size in individual beds is in the silt-clay range, and indications of bed-load transport are rare. Cryptocrystalline hematite is the dominant Fe-bearing mineral, but magnetite and pyrite occur locally, especially near contacts with diabase sills.
2. Micrite Subfacies: thin well-bedded micrites (siderite, dolomite and serpentine) are commonly interbedded with the argillites and represent periods of low clastic influx and high carbonate production. Apart from small, isolated pockets of ooids, there are no sedimentary structures indicative of high-energy deposition or reworking.
3. Jasper Subfacies: lenticular and wavy-bedded jasper units up to 30 cm thick contain chert granules, oncolites, argillite rip-ups and scattered quartz grains that form a clast-supported framework. Cementation occurred in two stages with an initial chert rim cement, and later void-filling chalcedony. No evidence of any precursor carbonate fabric is evident in the jasper, and the only carbonate grains in the cherts (dolomite and siderite) are late diagenetic phases that replace hematite and chert.

The sharp grain boundaries and the cementation patterns suggest a primary origin for the chert. Furthermore, thin argillite and micrite beds enclosing the jasper units show no evidence of silica replacement, the chert being confined entirely to the granule-bearing beds. In contrast to other subfacies in the Kipalu Formation, a relatively high-energy environment comparable to oolite or peloid banks and shoals is inferred. Argillite rip-ups also provide evidence of storm activity, although indications of sub-aerial exposure are lacking.

Comparison with other Major Iron Formations

The principal features of five major iron formations from northwest Australia, South Africa and the Canadian Shield are summarized in Table 14.2. Perhaps one of the most striking aspects of these and other Proterozoic iron formations is their great areal extent. For example, in the Brockman Iron Formation, individual beds can be traced over many tens of kilometres. Although somewhat less impressive, zones of strata a few metres thick can be traced over most of the Belcher Islands, over a palinspastically reconstructed area of about 8500 km^2. Crossbedding is rare in most iron formations, with principal exceptions being the Gunflint and Sokoman formations. The Sokoman also differs from other successions in the variety of carbonate lithologies, or in many cases chert-after-carbonate. Whereas micrite-type lithologies compose only about 30 per cent of the Sokoman Formation (Dimroth, 1977), the Brockman and Penge formations are almost exclusively composed of this

Table 14.2. Comparison of iron formations from five major Proterozoic sedimentary basins.

	(TRENDALL,1968)	(GOODWIN,1956; HOFMANN,1969)	(BEUKES,1973, BUTTON,1976)	(DIMROTH ET AL. 1970; CHAUVEL AND DIMROTH,1974).	(YOUNG,1922; WOODBRIDGE,1922 JACKSON,1960; PRESENT STUDY)
THICKNESS	915M IN THE HAMERSLEY GROUP. ABOUT 600M CONSTITUTES THE BROCKMAN IRON FM.	GUNFLINT FM. 140-170M OTHER IRON FORMATIONS RANGE FROM 0-600$^+$M.	PENGE FM. ABOUT 460M MAXIMUM. KURUMAN FM. ABOUT 750M.	SOKOMAN FM. 120-210M.	KIPALU FM. 107-125M.
STRATIGRAPHIC CONTINUITY	CAN CORRELATE MESOBANDS ABOUT ONE INCH THICK OVER 20,000 SQ. MILES.	DIFFICULT TO DEMONSTRATE STRATIGRAPHIC CONTINUITY	CAN TRACE SOME UNITS 40KM AND MORE.	THE SOUTHERN LABRADOR TROUGH EQUIVALENT, WABUSH IRON FM. 305M.	THICKNESS RELATIVELY CONSTANT OVER THE BELCHER ISLANDS (8500$^+$SQ. KM.). PROBABLE CORRELATIVES ALONG HUDSON BAY, ABOUT 600 KM.
SEDIMENTARY STRUCTURES	CROSS-BEDS. RARE	GRADED BEDDING AND SOME CROSS-BEDDING.	CROSS-BEDDING RARE INTRAFORMATIONAL BRECCIAS.	CROSS-BEDDING, LENTICULAR BEDDING, SCOUR CHANNELS, LOAD CASTS.	RARE CURRENT AND INTERFERENCE RIPPLES. GRADED BEDDING IN BANDED ARGILLITES.
MICROBANDING	CONSPICUOUS	RARE	CONSPICUOUS	PRESENT	CONSPICUOUS.
STROMATOLITES		VARIETY OF STROMATOLITES IN GUNFLINT CHERT.		PISOLITES AND ONCOLITES.	POSSIBLE ONCOLITES IN JASPERS
EVAPORITES			NO EVIDENCE FOR EVAPORITES.	LENGTH-SLOW CHALCEDONY IN INTRACLASTS.	NONE OBSERVED.
CLASTICS	NOT REPORTED	INTERBEDDED TUFFS. LOCALLY QUARTZ GRAINS.	OOLITIC JASPERS IN KURUMAN FM.	TERRIGENOUS CLASTICS UNCOMMON. OOLITIC AND PELOIDAL CHERTS COMMON.	BANDED ARGILLITES AND SILTSTONES ABUNDANT. RARE FINE GRAINED SANDSTONES. OOLITIC AND GRANULAR CHERTS COMMON.
RIEBECKITE	MASSIVE AND FIBROUS.	MASSIVE AND FIBROUS.	LESS WELL KNOWN.		NOT COMMON
AGE	ABOUT 2.0 GA.	GUNFLINT FM. ABOUT 2.0 GA.	ABOUT 2.3 GA.	1.9 GA.	ABOUT 1.8 GA.
BASIN SIZE	OVOID BASIN WITH MAJOR AXIS ABOUT 480 KM.				ARCUATE BASIN WITH MAJOR AXIS ABOUT 600 KM$^+$.
VOLCANIC ASSOCIATION.	UNDERLAIN BY FORTESCUE GROUP VOLCANICS AND INTERBEDDED WITH ACID VOLCANIC OF WOONGARRA FM. SHARDS IN THE BROCKMAN FM.	OBVIOUS INTERSTRATIFIC-ATION OF TUFFS AND FLOWS.	LOWER GRIQUATOWN IRON FORMATION OVER-LAIN BY 1100M OF ONGELUK VOLCANICS. PENGE FM. APPARENTLY NOT ASSOCIATED WITH VOLCANICS.	OVERLAIN BY MENIHEK SLATES AND VOLCANICS SLIGHTLY HIGHER IN STRATIGRAPHIC SEQUENCE. UNDERLAIN BY SHALES AND SOME TUFFS OF THE RUTH FM.	OVERLAIN BY FLAHERTY FM. VOLCANICS. FEW THIN TUFFS IN THE UPPERMOST KIPALU FM.

rock type. The Kipalu Formation contains upwards of 60 per cent banded argillites and micrites, and together with the sedimentological features noted above, appears to most closely resemble the Brockman and Penge formations.

The association of iron formations with volcanic rocks differs considerably from basin to basin, from those containing interstratified flows and volcaniclastics (for example the Gunflint), to sequences having no obvious volcanic affinities, as in the Penge Formation. In the case of the Kipalu ironstones, evidence of volcanism appears only at the top of the sequence.

Summary of the Fourth Phase

The Rowatt and Mukpollo formations are interpreted as an upwards-shoaling succession of terrigenous and carbonate clastics representing a prograding intertidal-subtidal apron, giving way to a carbonate buildup that accumulated in an upper intertidal to supratidal environment. The intertidal apron was characterized by actively migrating sand bars and small, isolated carbonae buildups. The Mukpollo Formation also records a sequence of tidal sand flats and bars. However, the Kipalu ironstones are indicative of more restricted basin conditions with local development of shoals. Here, the ironstones are considered to represent a barred or restricted basin which was protected from the open ocean by a system of carbonate buildups and offshore bars (Ricketts, 1978b).

V. SUBMARINE VOLCANISM PHASE

The Flaherty Formation, the "backbone" of the Belcher Islands, underlies about 60 per cent of the area exposed on the islands and forms many of the narrow arcuate peninsulas and prominent ridges characteristic of Belcher terrane. Flaherty strata are well exposed along some 2000 km^2 of coastline.

Moore (1918) first assigned the names Tookarak Diabase and Basalt, to what are now called the Flaherty and Eskimo formations, and Haig Intrusion. The volcanics were subsequently examined by Young (1922), Schenk (1959) and Jackson (1960). Although Schenk (1959) concluded that the volcanics had spilitic affinities, chemical analyses reported by Leggett (1974), from a section on the east side of Eskimo Harbour, indicate typical tholeiitic basalt compositions.

Because of considerable lithological variation, Dimroth et al. (1970) refrained from defining a type section. Nevertheless, it is useful here to define two reference sections:

1. a continuously exposed 1000 m section 7 km east of Sanikiluaq at Katuk entrance to Eskimo Harbour;
2. a less complete, but well-exposed 200 m section along the south shore of McLeary Point, east Tukarak Island.

Whole rock K-Ar dating of pillow-rim and core samples has revealed two broad age groups (Hofmann and Jackson, 1969): (1) an older 1620-1693 Ma age group that has been correlated with metamorphic events during the Hudsonian Orogeny, and (2) an 830-1054 Ma age group tentatively correlated with a mild thermal or tectonic event during the Grenville Orogeny. Whole rock Rb-Sr isochron determinations on a number of volcanic and shale samples from the Flaherty and Omarolluk formations respectively, gave ages of 1717 ± 69 Ma and 1752 ± 89 Ma, with a composite isochron defining an age of 1760 ± 38 Ma (Fryer, 1972).

General Features

The Flaherty volcanics rest disconformably upon the Kipalu Formation. They consist of a sequence of flows and pillows, and also include a variety of volcaniclastic lithologies. Pronounced vertical and lateral variations in the volcanic and sedimentary facies are accompanied by considerable changes in thickness; from 1950 m in western Belcher Islands, to 290+ m on east Tukarak Island.

Volcanic Components

Massive and columnar jointed basalt flows compose more than 50 per cent of the formation throughout the area. The thickest flows occur in western sections, where individual flow units attain thicknesses of 30 m, and composite flows are up to 145 m thick. Two types of flow-top structures are noted:

1. monomict flow-top breccias up to 5 m which grade into the underlying massive flows. These breccias resemble the rough, clinkery surface of aa-type flows.
2. festooned ropy lavas typical of more fluid, more rapidly extruded pahoehoe-type lava flows; these provide valuable indications of flow direction.

Flow-base structures such as inclined pipe vesicles provide additional information for determining flow directions. Both flow-top and flow-base structures consistently demonstrate that flow was predominantly to the east.

The second major constituent of the volcanics consists of pillow lavas. Pillow-piles, ranging in thickness from 5 to 230 m, can be traced laterally for many tens of metres. Some thick piles are composite, with individual units draped by wedges of pillow talus. Lava tubes occur in several of the piles. Whereas pillows make up about 40 per cent of most sections on west Flaherty Island, there is a sharp decrease to about 18 per cent on Tukarak Island.

Volcaniclastic Rocks

The decrease in the proportion of pillow lavas towards the eastern part of the Belcher Islands is accompanied by a concomitant increase in volcaniclastics, from less than 10 per cent on Flaherty Island, to 27 per cent on Tukarak Island. Several types of volcaniclastics are recognized on the basis of the style of bedding, sedimentary structures, sediment composition and texture, and association of broad lithological groups with volcanic flows and pillows. These include: thinly-bedded pyroclastics and water-laid tuffs and lapillistones, some of which contain accretionary lapilli, representing shallow-water or subaerial phreatic and phreatomagmatic eruptions; turbidites consisting entirely of fine grained volcanic debris (ash); thick, laterally extensive, pyroclastic flows (one flow that attains a maximum thickness of 85 m covers a palinspastically reconstructed area of at least 3000 km^2); and pillow talus deposits formed by rapid quenching of hot lavas.

Pyroclastic and thinly-bedded water-laid tuffs predominate in sections west of Flaherty Island and are interbedded with massive and pillowed flows. To the east, however, particularly on Tukarak Island, volcaniclastic turbidites become the important sedimentary component and these too are intercalated with massive and pillowed flows. Crossbed azimuths from both groups of sediments demonstrate pronounced unimodal current directions, with transport of reworked pyroclastic and epiclastic volcanic debris towards the east, in accord with that inferred for the lava flows.

Summary of the Fifth Phase

Volcanism during deposition of the Flaherty was of two main types: effusive, producing a thick sequence of lava flows and pillows; and explosive activity that gave rise to pyroclastics and water laid tuffs (including turbidites) composed of pumice fragments and shards with bubble-wall structures. Tephra of this type is most commonly produced during phreatic or phreatomagmatic eruptions, and these are usually restricted to subaerial and shallow-water environments. Cross-stratification in several tuffs also provides evidence of subaqueous conditions, although the presence of accretionary lapilli in some beds also indicates that some eruption columns were subaerial. Most of the massive and jointed lava flows are also considered to have formed under subaqueous conditions, especially flows that are "sandwiched" between pillow piles or turbidite deposits.

Flaherty volcanism followed a period of regional uplift, shoreline progradation, and formation of a carbonate buildup-restricted basin system (fourth phase). On the basis of regional thinning trends, Jackson (1960) and in Dimroth et al. (1970) presumed that eruption centres and fissures lay to the west of Belcher Islands. This has been substantiated by Leggett (1974), Ware (1978) and in the present study by establishing paleocurrent and paleoflow directions. Subsequent reversal of the paleoslope from west- to east-dipping can be attributed to construction of this thick volcanic edifice.

Volcaniclastic sediments are composed entirely of locally derived volcanic debris. Craton-derived material apparently was not introduced into the Flaherty depositional system, even though the entire volcanic succession is superposed on an older platform sequence.

VI. SUBMARINE FAN – DISTAL MOLASSE PHASE

Omarolluk Formation

Regularly bedded argillites, shales and graded wackes of the Omarolluk Formation are exposed on Gilmour Peninsula and Young Point, eastern Flaherty Island, and on a number of outlying islands, including Baker's Dozen and King George islands. Contact with the underlying Flaherty Formation is a sharp, non-erosional disconformity exposed along a narrow strip of coastline between Gilmour Peninsula and Young Point. The lowermost 15 m of the Omarolluk consists of black pyritic shales, fine grained volcanic sandstones, and rare interbeds of thin graded tuff. The transition from the shale- to graded wacke-dominated lithologies is mostly covered.

Originally, the contact between the Omarolluk with the overlying Loaf Formation was based primarily on the change in colour from grey arkosic wackes, to red arkoses (Dimroth et al., 1970). Within this definition, the Omarolluk contains strata typical of both turbidite and shallow-water fluviatile facies. Mapping during the present study indicates that a more logical basis for subdivision is the presence or absence of graded beds. The Omarolluk Formation is thus restricted to lithologies characteristic of turbidite facies. Hence, although the contact with the Loaf Formation also is gradational on the basis of this criterion, it is more easily established in the field, and has more significance in terms of changes in depositional environments.

The Omarolluk is characterized by extremely regular and tabular-bedded greywackes, some arkosic wackes, intraformational conglomerates, and shales. The sequence of sedimentary structures in most beds is diagnostic of the classic Bouma cycle. The ubiquity and repetitive nature of Bouma cycles leaves little doubt that the Omarolluk Formation can be regarded as a thick turbidite sequence. Four principal facies are present: an arenaceous facies represented by individual turbidite beds (divisions A to D), a pelitic or non-turbidite facies representing a return to normal, low-energy deep-water sedimentation, a conglomerate facies, and a transitional facies.

Arenaceous and Pelitic Facies

Beds can be traced more than 1000 m along strike with little change in thickness or lithology, although a few beds grade laterally into distinct composite beds. The average bed thickness is 40 cm, but can be as thick as 160 cm. The sandstone/shale thickness ratio is consistently about 10:1 over most of Gilmour Peninsula and Young Point. Approximately 1000 m of greywacke and shale outcrop on Walton and Camp islands, 25 km south of Gilmour Peninsula. Here, bed thickness averages 20-30 cm and the sandstone/shale ratio is as low as 3:1.

Flute casts, parting lineations and groove casts are common on exposed bedding soles and show the direction of sediment transport of individual flows. Sole structures indicate that sediment dispersal was predominantly to the southeast. Current ripple measurements, mostly from the pelitic division overlying the turbidites, demonstrate that the flow of bottom currents was east to southeast and, as expected, these show a wider dispersion than do paleocurrent indicators for the turbidites.

Conglomerate Facies

Shallow pebble-filled channels occur at several intervals in the lower 600 m of the section on Gilmour Peninsula. Channels are rarely more than 120 cm deep, but range from 80 m wide to less than a metre, and in many instances are associated with sequences of thick greywacke beds. Channel-fill consists of flat rounded pebbles of siltstone and calcilutite and subspherical calcite concretions derived from nearby greywacke beds. Some of the flat lutite pebbles contain ripple crossbeds and parallel laminae not present in the surrounding matrix. In many cases the flat lutite pebbles resemble elongate concretions from the C and D divisions of greywacke beds. It is evident that concretion growth in the turbidites was a very early post-depositional phenomenon.

Although no large-scale crossbedding was observed in the channel deposits, a crude imbrication of flat pebbles suggests transport to the east and southeast.

Transitional Facies

Omarolluk Formation rocks in the Baker's Dozen Islands define a transitional zone between turbidite facies and shallow-water facies of the overlying Loaf Formation. On Twin Cairns Island both graded and massive arkosic wackes are present, intercalated with minor siltstone or shale beds. Intraformational conglomerates are common, but large-scale festoon crossbeds and red mudstones typical of the Loaf Formation are absent at this stratigraphic level.

Interpretation of the Omarolluk Formation

Compared with many other turbidite successions, the Omarolluk exhibits a relatively limited range of facies, the most prominent ones being defined by parallel and laterally persistent bedding, and alternating arenaceous-pelitic facies. Bouma cycles are ubiquitous.

This association corresponds to the 'C' Facies (outer fan facies) described by Mutti and Ricci Lucchi (1978). Some intervals of thick and composite-bedded greywackes and channel deposits may correspond to middle to inner fan facies. Extensive deep-basin facies have not yet been recognized, although the southward thinning trend and decreasing sandstone/shale ratio at Walton and Camp islands may indicate that lateral deep basin equivalents formed south of the Belcher Islands.

Loaf Formation

The Loaf Formation contains tabular and lenticular beds of arkose which display abundant trough crossbedding, and lesser amounts of red and grey siltstone and mudstone. The lowermost strata on Loaf Island also contain massive arkose beds with a few calcite concretions. Festoon crossbedding is particularly common in the arkoses where trough axes are up to 5 m. Some scour-and-fill structures are lined with basal pebble lags, mostly of intraformational conglomerates, but also containing rare pebbles of white chert, jasper, quartzite, granite and dacite. Exhumed scour troughs are commonly veneered by current-rippled shales. Other forms of crossbedding include tabular-planar sets, low-angle planar sets, herringbone crossbeds and interference ripples. Several mudstone beds display large desiccation cracks.

Trough crossbeds indicate that sediment transport was predominantly to the south and southeast. Large planar crossbeds show similar trends, whereas herringbone structures suggest minor transport to the east and west.

The abundance of trough crossbedding and associated synsedimentary structures (ball-and-pillow, detached load balls and recumbent folding of crossbeds) evidence of subaerial exposure, and strongly unimodal paleocurrent directions substantiate interpretation of a fluvial environment of deposition.

Based on lithology and sedimentary structures, four major fluvial lithofacies are recognized and correspond to lithofacies identified by Miall (1977) as forming in a braided-stream environment:

1. a trough crossbedded sand subfacies formed during stages of river flood;
2. a planar crossbeded subfacies, which may have developed in response to lateral migration of point bars during flooding;
3. a horizontally-bedded sandstone subfacies, usually with parting lineations;
4. and a desiccated mudstone subfacies, representing periodic exposure in an overbank environment.

Shoaling Upwards Cycles

Crude upwards-shoaling cycles, 15-20 m thick, have been identified in the Loaf Island sections. The lowermost strata in each cycle consist of medium grained arkoses containing parallel laminated, low-angle planar and a few herringbone crossbeds. These pass upwards into laminated siltstones and finally trough crossbedded arkoses and desiccated mudrocks. Structures lower in the cycle such as herringbone crossbeds, lenticular and flaser bedding all are suggestive of a weak tidal influence, possibly in a marginal marine environment. The Loaf cycles are somewhat similar to alternating marine and nonmarine regressive cycles in the Devonian Catskill Formation (termed 'motifs' by Walker, 1971). The existence of prograding, marginal marine deposits in the Loaf is reasonable because the transition from relatively deep-water turbidites to fluvial deposits requires passage through a strandline intermediate between these two facies.

Petrographic Trends in the Omarolluk and Loaf Formations

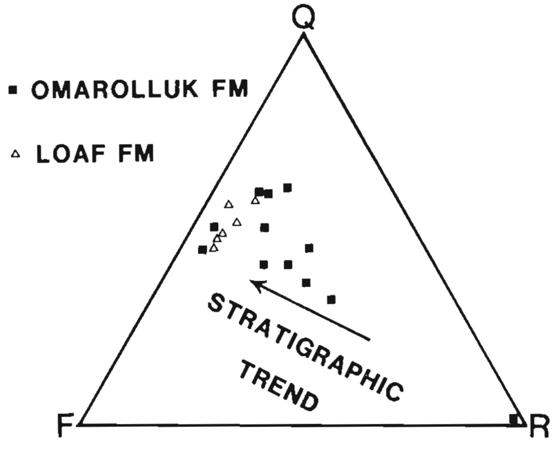
Petrographic analysis of the Omarolluk and Loaf formations has delineated two distinct source terranes; a volcanic terrane (Flaherty Formation), and a cratonic terrane. Based on paleocurrent data, the cratonic source appears to have been northwest of the Belcher Islands.

Broad stratigraphic trends in the composition of Omarolluk and Loaf rocks (Fig. 14.6) show a gradual upward decrease in the amount of plagioclase and volcanic fragments towards the Loaf arkoses, a concomitant increase in potash feldspar, and a tendency towards matrix depletion. Samples from the Loaf Formation form a tighter cluster compared to Omarolluk sands, reflecting the two fundamentally different depositional environments, and also the differences in sources of clasts. Omarolluk sediments were derived from volcanic and cratonic sources, indicating that the Flaherty volcanics were exposed at this time, in addition to uplifted cratonic rocks to the north of Belcher Islands. Loaf sediments on the other hand, were derived solely from a cratonic source.

Summary of the Sixth Phase

A 2100 m thick Omarolluk turbidite sequence, consisting of both proximal and distal submarine facies, disconformably overlies relatively deep-water Flaherty volcanics along eastern Flaherty Island. However, in western parts of the Belcher Islands Omarolluk strata overlie shallower water representatives of the Flaherty Formation. The possibility that the Omarolluk succession thins towards western Belcher Islands, against the eastern flank of the Flaherty volcanic ridge, should therefore be considered.

The fluvial and marginal marine Loaf Formation appears comparable to some examples of distal molasse described in the European Alps (e.g. Van Houten, 1974). Fanglomerates typically associated with Alpine molasse are not present in the Belcher Islands. Although the Omarolluk-Loaf Formation contact is nowhere exposed on the Belcher Islands, the transitional nature of both the sedimentary facies and petrographic trends suggests that the contact is both conformable and gradational. Thus, Omarolluk and Loaf formations are considered lateral equivalents.



Q = quartz
F = total feldspar
R = lithic fragments

Figure 14.6. *Petrographic trends in the Omarolluk and Loaf Formations.*

The trend of the turbidite basin axis was approximately south to southeast, as indicated by paleocurrents, lateral differences in bed thickness, and sandstone/shale ratios and hence was parallel to the trend of the (Flaherty) volcanic ridge. The dominant direction of sediment transport in the Loaf arkoses was parallel to this trend.

In searching for a modern analogue for the Omarolluk-Loaf System, the close relationship between deltas and base-of-slope deposits is noted, a relationship that Stanley and Unrug (1972) have noted may be easily overlooked in studies of anciant turbidite sequences. A number of modern deep basins are in fact fed by delta-fronted river systems, some particularly prominent examples being the Bengal Fan (Curray and Moore, 1971), and the Nile Cone (Maldonado and Stanley, 1978). A similar depositional regime is envisaged for the Omarolluk-Loaf sequence, where fluvial, marginal marine and prograding submarine fan facies are laterally associated.

RÉSUMÉ OF THE BELCHER GROUP

The Belcher Group consists of a thick (7000-9000 m) but remarkably regular succession of distinctive rock units. All major formation boundaries within the succession are either conformable or disconformable; no angular unconformity has yet been recognized in this area. Collectively, formations of the Belcher Group present a complex record of basin evolution whereby long-term changes in depositional environments have taken place in response to fundamental changes in the tectonic regime.

The succession can be subdivided into a number of depositional phases, and their corresponding tectonic elements can be reconstructed (Fig. 14.1). Although crystalline basement is nowhere exposed in the Belcher Islands, the succession is interpreted as ensialic.

The first phase of carbonate platform development, about 2.0 Ga ago, was abruptly interrupted by extrusion of plateau-type basalts (Eskimo Formation). A second miogeoclinal platform stage immediately followed, suggesting that the regional tectonic pattern was little altered by the volcanic event. The second miogeoclinal stage actually consists of two phases of platform development (Fig. 14.1): the first phase involved evolution of platform, slope and deep-basin carbonate facies that formed under conditions of gradual subsidence; the second phase, dominated by a prograding carbonate buildup-restricted basin complex, accumulated as a result of gradual uplift heralding the onset of Flaherty volcanism.

Growth of a thick volcanic pile (Flaherty Formation), mostly under submarine conditions, resulted in reversal of the paleoslope to east-dipping, accompanied by down-sagging of the underlying platform rocks. Continued subsidence and rapid infilling of the basin by a southward-prograding clastic wedge consisting of submarine fan and distal molasse facies, signaled the final depositional phase of the Belcher Group.

Paleocurrent data suggest that most of the Omarolluk Formation turbidites and Loaf Formation molasse was provided by uplifted cratonic terrane northwest of the Belcher Islands. Deposition in the turbidite-molasse basin (exogeosyncline) can logically be inferred to have been syn-orogenic, although actual deformation of the strata did not take place until later.

The highly sinuous outline of the Belcher Islands is a manifestation of open, predominantly isoclinal, doubly plunging anticlines and synclines. Fold axes trend approximately north-south, but as Jackson (1960) has noted, the fold axes become curved west of Flaherty Island, opening towards the west. Crustal shortening ranges from 25 per cent on Tukarak Island to more than 40 per cent in the Kasegalik anticline. The structural style of the Belcher Islands can be explained by a single period of deformation: initial buckling led to formation of extension and conjugate shear joint patterns, followed by predominantly flexural slip folding and concomitant development of major faults. Folding of the entire Belcher succession probably took place during the Hudsonian Orogeny.

CORRELATIONS

The pioneering efforts of Bell (1879), Low (1902) and Leith (1910) established the presence of relatively undeformed Precambrian rocks overlying crystalline basement along eastern Hudson Bay, and also provided evidence for regional correlations based on the presence of distinctive rock units such as banded ironstones, 'concretionary' (cryptalgal) dolostones and basaltic cap rocks. These correlations were extended to the Belcher Islands, first by Flaherty (1918), and later by Moore (1918), Young (1922) and Woodbridge (1922), and also to the Sutton Lake area (Dowling, 1905; Hawley, 1926).

More recently, correlations have been made with Precambrian fold belts ouside the Hudson Bay region, including Cape Smith-Wakham Bay, the Labrador Trough and the Mistassini area (Wahl, 1953; Bergeron et al. 1962). Correlation of these fold belts around the ancient Ungava craton was suggested by Bergeron (1957), and the name Circum-Ungava Geosyncline was subsequently proposed to include the Cape Smith Fold Belt, Labrador Trough and Belcher Fold Belt (Dimroth et al. 1970).

The structure of the Belcher Fold Belt from Ottawa Islands to James Bay has been summarized by Jackson (in Dimroth et al. 1970). Folding in the Belcher Islands diminishes north and south of the islands, and the intensity of deformation also decreases eastwards towards Richmond Gulf. The interpretation that volcanic strata in the Ottawa Islands are a continuation of the Flaherty Formation and similar lithologies in James Bay, is supported by bathymetric and magnetic data (Hobson, 1969; Hood et al., 1969). Baragar and Lamontagne (1980) have further demonstrated correlation between the volcanics on Sleeper Islands and the Flaherty Formation, and that komatiitic rocks of Ottawa Islands are probably correlative with komatiites of the Cape Smith Belt; in both the Sleeper and Ottawa islands localities these authors consider the volcanics to have been extruded into deeper water environments than the Flaherty equivalents.

Correlations in Richmond Gulf

Lithologic units in Richmond Gulf (forming part of the Manitounuk Supergroup) have been correlated with specific formations in the Belcher Islands (e.g. volcanic rocks of the Persillon Formation, formerly Pachi volcanics; the stratigraphy in Richmond Gulf has recently been redefined by Chandler and Schwarz, 1980) and Nastapoka Group in Richmond Gulf have been correlated with the Eskimo Formation and Flaherty Formation respectively, and banded ironstones of the Nastapoka Group with the Kipalu Formation. However, the remaining units show few lithological similarities and at best, Richmond Gulf strata can only be considered as continental and shallow marine equivalents of the Belcher Group. An additional problem is the presence of an angular unconformity between the Richmond Gulf Group and Nastapoka Group which developed during a period of block faulting. According to Chandler (1978), most of this deformation occurred before deposition of Nastapoka sediments and volcanics. An equivalent episode of faulting affecting the Belcher Group is not known. With this in mind, the Qingaaluk (formerly Richmond Gulf Formation) and Persillon formations possibly are time equivalents of the Eskimo Formation. The contention that block faulting predates the Fairweather Formation is further supported by the predominance of immature subarkoses and volcanic litharenites in the Fairweather, that were derived from the uplifted terrane in Richmond Gulf.

The Pachi Formation, predominantly of continental origin, has been correlated with the Kasegalik Formation in Belcher Islands (Dimroth et al, 1970). However, the exact stratigraphic relationship between these two fundamentally different rock units is difficult to demonstrate. For example, the 1220+ m of marine platform carbonates making up the Kasegalik Formation are representative of a transgressive regime; the much thinner Pachi arkoses (interpreted as braided stream deposits by Chandler and Schwarz, 1980) show no evidence of encroaching seas. Furthermore, there is no indication that any of this terrigenous sediment was deposited on the carbonate platform, as might be expected given the dynamic fluvial environment envisaged for the Pachi rocks. Therefore, the Pachi Formation may represent an earlier phase of deposition than the Kasegalik Formation.

North-south thinning of the sedimentary and volcanic units in Richmond Gulf supports the hypothesis of graben-controlled deposition (Kranck, 1951; Woodcock, 1960; Chandler, 1978). Extension of this graben system to the Belcher Islands was first suggested by Kranck (1951).

SUMMARY OF TECTONIC ELEMENTS IN THE BELCHER FOLD BELT

The various tectonic elements composing the Belcher Fold Belt are illustrated in Figure 14.7.

Initial Miogeoclinal-Graben Stage

Continental deposits of the Pachi Formation and platform carbonates of the Kasegalik Formation, accumulated in an east-trending graben within the Ungava craton, here called the Richmond-Belcher Rift. At present, the age of this rift can only be bracketed between pre-Pachi

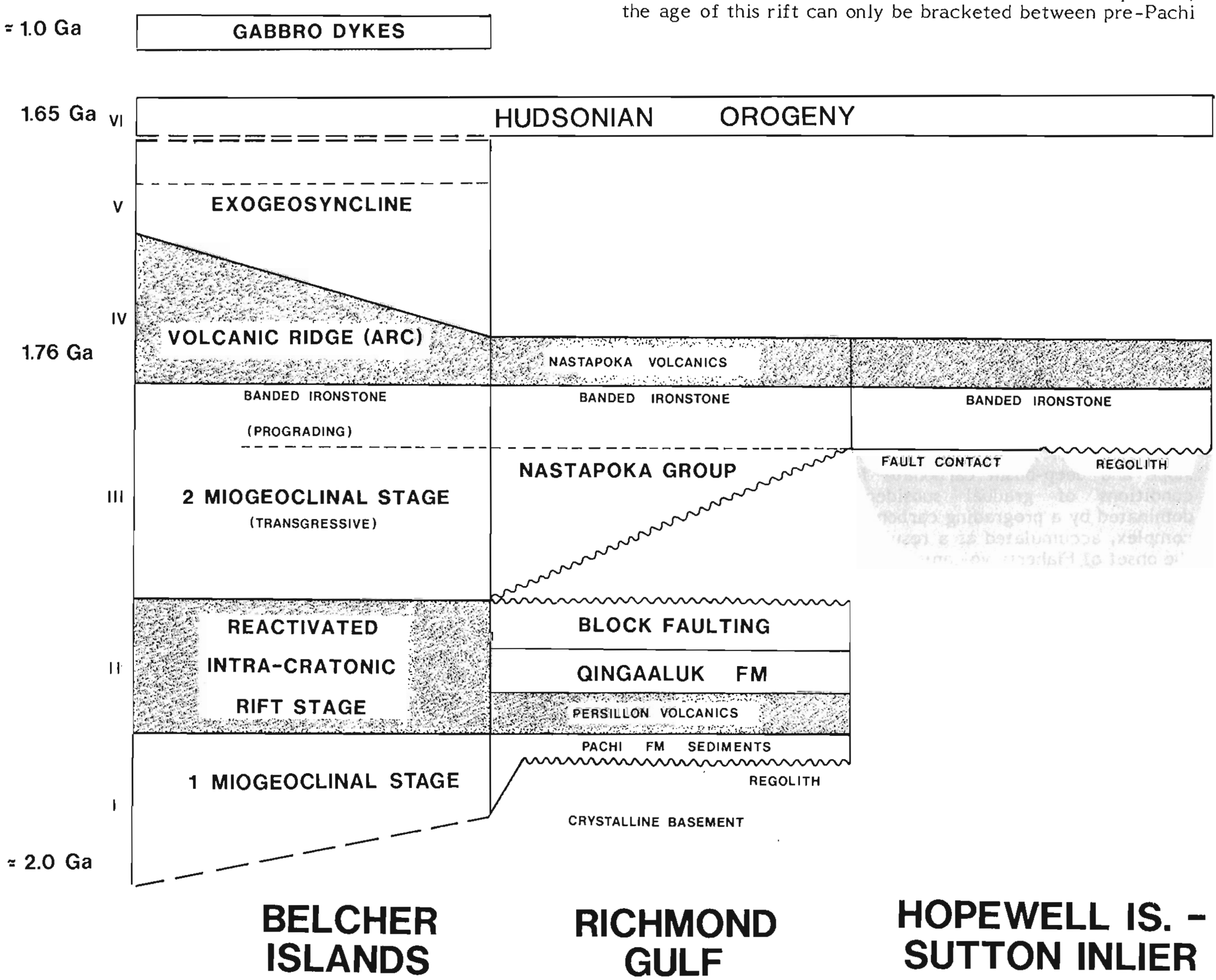


Figure 14.7. *Correlation of tectonic elements throughout the Belcher Fold Belt.*

and post Archean. Igneous rocks with alkalic affinities, commonly associated with continental rifting, are unknown in the immediate Richmond Gulf area, although Dimroth (1967) has reported "post-tectonic" carbonatite dykes, diatremes and pyroclastics due east of Richmond Gulf (actual age unknown).

Rift Reactivation Stage

Deposition in the early miogeocline was abruptly interrupted by outpouring of flood-type basaltic lavas within the Richmond-Belcher Rift: tholeiitic basalts of the Eskimo Formation, and high-Mg tholeiites of the Persillon Formation (see Hews, 1976, Table 11). Continental redbeds of the Qingaaluk Formation disconformably overlie the Persillon volcanics and are considered here to be similar in age to later stages of Eskimo Formation volcanism. Following redbed deposition, renewal of an extensional tectonic regime in the Richmond Gulf area produced a horst and graben topography upon which Nastapoka Group sediments were deposited.

Second Miogeoclinal Stage

Following volcanism and block faulting, deposition of a thick sequence of carbonates and clastics re-established a broad, shallow marine platform in the Belcher region (Fairweather and Laddie formations) with continental equivalents probably represented by basal strata of the Nastapoka Group. Once again deposition was largely confined to the Richmond-Belcher Rift. Conditions of gradual subsidence prevailed.

In the Belcher Islands, a major change in the pattern of sedimentation is recognized at the Laddie Formation-Rowatt Formation boundary, with development of prograding carbonate buildup-restricted basin phase (Rowatt to Kipalu Formation); probable facies equivalents occur in Hopewell and Nastapoka islands, Manitounuk Sound and Sutton Inlier. At these localities banded ironstones are interbedded with, or underlain by, quartz-rich sandstones (Woodbridge, 1922; Lee, 1965; Eade, 1966; Sanford et al., 1968). The Rowatt-Kipalu model is thus extended to include these areas. On this basis, the buildup and barrier bar complex and associated restricted basin are thought to have extended around the entire perimeter of eastern Hudson Bay to Sutton Inlier, and is envisaged as a large arcuate basin having an arc of about 800 km.

At this stage in the development of the Belcher Fold Belt the Richmond-Belcher Rift was no longer the locus of deposition. Furthermore, in the Richmond Gulf area, reactivation of faulting during and after deposition of Nastapoka sediments was only of minor importance.

Volcanic Ridge Stage

The preceding stage of sedimentation was abruptly terminated and the paleoslope reversed, by the accumulation of a thick pile of basaltic and pillowed lava flows, and abundant volcaniclastic sediments (the Flaherty Formation, and correlative strata on Ottawa Islands and Sleeper Islands). Equivalent strata farther east in the Nastapoka Island chain and south to Sutton Inlier, consist predominantly of massive and jointed lava flows and sills. The volcanics are considered to have originated from fissures and eruptive centres west of Belcher Islands, representing segments of an extensive volcanic ridge (? arc) that evolved along the margin of (and which subsequently buried) the former miogeoclinal prism.

Exogeosynclinal Stage

Representatives of this stage occur only in the Belcher and outlying islands. Early synorogenic turbidites and laterally associated distal molasse sediments were deposited in a submarine fan-delta system that prograded southwards over relatively undeformed Flaherty volcanics and the older miogeoclinal succession. The axis of the turbidite basin appears to have been parallel to the volcanic ridge. Sediments were mostly derived from a cratonic source area to the northwest of Belcher Islands, and also from exposed portions of the adjacent volcanic ridge.

Compression Stage

The main period of deformation of the Belcher Fold Belt took place during the Hudsonian Orogeny, and resulted in a single phase of isoclinal folding in Belcher Islands, Sleeper Islands and possibly Ottawa Islands. Metamorphism only attained grades of prehnite-pumpellyite and lower greenschist facies. There is some indication in Belcher and Ottawa islands that the metamorphic grade increases from east to west, a feature that could reflect either proximity to a 'tectonic front', or deeper levels of erosion in the western exposures. Whole-rock K-Ar ages of 1620-1693 Ma determined for Flaherty volcanic samples, probably represent re-setting by this low-grade metamorphism (Hofmann and Jackson, 1969).

DISCUSSION

A comparison of the fold belts composing the Circum-Ungava Geosyncline with Alpine-type tectonic sutures was made by Wilson (1968). On the basis of preliminary gravity data, Innes et al. (1967) had earlier postulated that a northward extension of the Kapuskasing "high", representing a zone of rifting probably passes through the Belcher Islands, terminating about 100 km west of Ottawa Islands. Deformation in the Belcher Fold Belt was thus explained as folding marginal to the rift zone and was coeval with volcanism, but this proposal is here considered to be unlikely, in view of known geological relationships. Following Wilson's (1968) ideas, Gibb and Walcott (1971) described these fold belts as being the products of ocean-closing and continental collision during the Hudsonian Orogeny, a hypothesis reiterated by Burke and Dewey (1973). The latter authors also noted the possibility of a triple junction where a northward extension of the Kapuskasing magnetic high intersects the Belcher suture, although Watson (1980) has questioned the rift-hypothesis of the Kapuskasing structure, citing evidence suggesting an origin related to deep shearing, early in the history of the Superior Structural Province.

In this paper the Belcher Fold Belt is examined further in terms of ocean closing hypothesis, and an attempt will now be made to explain the striking differences in content and style of deformation between this fold belt and others in the Circum-Ungava Geosyncline.

Ocean Opening (Early Proterozoic)

Identification of the fundamental tectonic elements composing the Belcher Group and Richmond Gulf succession (Manitounuk Supergroup) is based on the evolution of these platform and basin-fill sequences as interpreted in this paper, and their striking similarity to Phanerozoic successions.

Following this, criteria suggesting that the Richmond-Belcher Rift originated as the failed arm of a triple junction are listed below (Fig. 14.8a):

a. the trend of rifting is transverse to the margin of the Ungava Craton. Rifting predates sedimentation and volcanism and therefore the term aulacogen is considered to be valid (rather than the term impactogen, the development of which is related to a collision event).

b. In Richmond Gulf, block faulting that predates the Hudsonian Orogeny parallels the axis of the rift zone and is associated with reactivation of rifting, probably coeval with extrusion of basaltic magmas of the Persillon and Eskimo formations. Faulting in the Belcher Islands is related to isoclinal folding during the Hudsonian.

c. The sedimentary fill is broadly similar to that described for several modern and ancient aulacogens (e.g. Benue Trough, Ethiopian Rift, Athapuscow Aulacogen; Hoffman, 1973; Hoffman, 1980; Burke, 1980). The oldest sediments in Richmond Gulf are interpreted to represent early development of a westward-prograding delta, whereas the oldest strata in Belcher Islands represent a sabkha-like environment. Burke (1980) has noted that evaporites commonly are stratigraphically higher than clastic sediments during the early stages of aulacogen development. However, the exact age relationship between the arkosic and evaporitic deposits in the Richmond-Belcher Rift have yet to be established;

d. The characteristic 'arc' of eastern Hudson Bay is similar in shape and magnitude to more recent examples of rifting at triple junctions (e.g. Gulf of Guinea, Gulf of Aden).

The triple junction was probably situated west of Belcher Islands. At present the timing of this event can only be bracketed between post-Archean and pre-Pachi time, although initial rifting in the Belcher area possibly was coeval with an event postulated by Burke and Dewey (1973) at the junction between Cape Smith Fold Belt and Labrador Trough about 2.1 Ga. A similar date for the initial rifting stage of the Wopmay Orogen also has been postulated by Hoffman (1980).

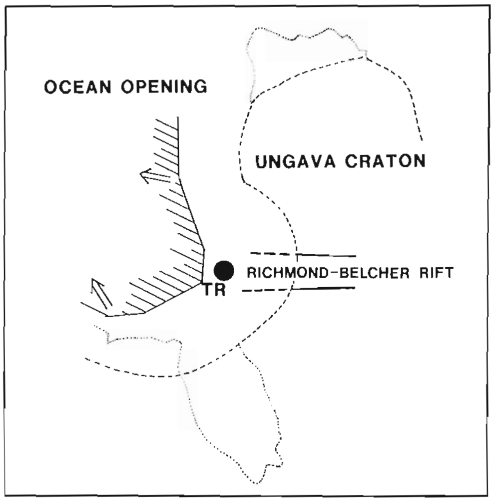


Figure 14.8a. *Initial stage of rifting early in the Proterozoic and generation of the Richmond-Belcher Rift, transverse to the Ungava Craton (TR = triple junction). The dashed line represents the approximate depositional edge along the trailing edge of Ungava craton.*

Deposition along the trailing edge of the Ungava craton was largely confined to the Richmond-Belcher Rift and occurred in two different stages, with carbonate platforms established in the distal portions of the rift zone (Belcher Islands), and associated continental redbeds (deltaic) within Richmond Gulf. These two stages were separated by reactivation of rifting, represented by an outpouring of flood basalts (Eskimo and Persillon volcanics), and later by a brief period of block faulting within Richmond Gulf.

Whereas the early phase of carbonate platform sedimentation was mostly confined to the Richmond-Belcher Rift, later deposition of carbonate buildups, bars and associated restricted basin facies (Rowatt-Kipalu formations and their correlatives) embraced the entire eastern coast of Hudson Bay. This suggests that the rift zone no longer served as the principal repository for sediments, a situation that may have heralded the approaching collision event.

Ocean Closing

Growth of an extensive volcanic ridge (? arc) in the Belcher Fold Belt (Flaherty Formation and correlatives) is considered to have been a product of ocean closing (subduction) prior to continental collision. Probable time-equivalents of these volcanics occur in the Cape Smith Fold Belt which is characterized by a bimodal volcanic suite having island arc affinities (Baragar, 1974). The scenario presented in Figure 14.8b shows synorogenic turbidite and molasse sediments derived from an uplifted cratonic source to the north of the Belcher Islands, and also the exposed volcanic ridge to the west, and shed over the older miogeoclinal succession during the collision event. Collision was first initiated at the "Cape Smith Salient", progressing

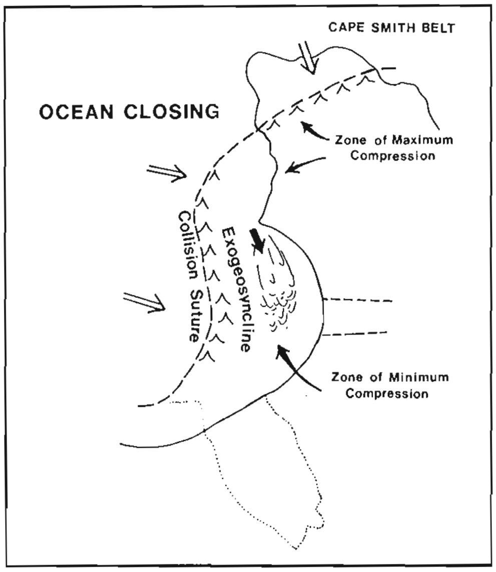


Figure 14.8b. *Ocean Closing: This scenario depicts a sinuous volcanic arc from Cape Smith to Belcher Islands, and probably continuing west to the Nelson River, Flin Flon area. The subsequent exogeosynclinal stage was situated between the volcanic arc and Ungava craton margin, trending parallel to the arc. Continental-type collision and deformation occurred during the Hudsonian Orogeny with minimum deformation occurred during the Hudsonian Orogeny with minimum deformation occurred during the Hudsonian Orogeny with minimum deformation taking place in the recessed craton margin of eastern Hudson Bay (heavy dashed lines mark the approximate position of the collision stuture).*

southwards towards the eastern Hudson Bay re-entrant. Continued convergence of cratonic blocks resulted in isoclinal folding of the Belcher Group during the Hudsonian Orogeny (although penetrative deformation is scarce), and only gentle tilting of strata bordering eastern Hudson Bay. The transition from volcanic-rich lithologies to stratigraphically higher craton-derived lithologies observed in the turbidite-molasse succession, also reflects the diachronous nature of the tectonism where southwards progradation of the delta-fan system (i.e. longitudinal dispersal) probably kept pace with progressive uplift of the craton during collision.

The relatively mild deformation imposed on the Belcher Fold Belt can be explained in terms of the concave geometry (opening westwards) of the trailing edge of the Ungava craton; here, continental convergence was minimal and only mild compression ensued, decreasing in intensity from west to east across the fold belt (Fig. 14.8b). Such mild deformation is in marked contrast to that observed in the Cape Smith Fold Belt where remnants of volcanic arcs were thrust over the Archean foreland, and high-grade metamorphic equivalents occur in the adjacent hinterlands.

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15.

PALEOMAGNETISM OF THE CIRCUM-UNGAVA FOLD BELT II: PROTEROZOIC ROCKS OF RICHMOND GULF AND MANITOUNUK ISLANDS

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Schwarz, E.J. and Fujiwara, Y., Paleomagnetism of the Circum-Ungava Fold Belt II: Proterozoic Rocks of Richmond Gulf and Manitounuk Islands; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 255-267, 1981.

Abstract

Three hundred and eighty-eight oriented drill cores were collected from Proterozoic volcanics and redbeds at 68 sites along the east coast of Hudson Bay. Detailed AF demagnetization in fields up to 60 mT and thermal demagnetization resulted in 35 accepted sites most of which come from Richmond Gulf. The Nastapoka basalt caprock (remanence carrier: magnetite) yielded five such sites which show no indication of the presence of more than one high coercivity and blocking temperature component. The pole position is 9°W, 39°S; 6°, 4°. South of Richmond Gulf, the same layer (130 m thick) yielded 6 sites with low negative inclinations which are thought to be contaminated by a secondary component and 2 sites with a stable magnetization parallel to a postfolding direction in the nearby Belcher Islands. Three sites from the underlying Qingaaluk Formation sediments yielded a comparable pole position at 13°W, 40°S; 11°, 6° (remanence carriers: hematite and magnetite). The older Persillon volcanics collected on the south shore of Richmond Gulf yielded a pole position of 12°E, 46°S; 12°, 9° based on 5 sites with magnetite and hematite as remanence carriers. These three pole positions are not significantly different. However, the Persillon sites from the northern part of Richmond Gulf show a different direction based on 11 sites resulting in a pole at 46°W, 49°S; 4°, 2°. This group of sites shows a small improvement in grouping statistics after correction for the small difference in bedding attitudes in the different fault blocks suggesting that at least most of the magnetization is pre-faulting. The faulting is pre-Nastapoka so that the general coincidence of the Nastapoka and Persillon directions would suggest that these are primary or essentially so. Furthermore, the pole positions correspond to the pole based on pre-folding magnetization in the Eskimo volcanics from the nearby Belcher Islands suggesting an approximate time-stratigraphic correlation. Preliminary results from La Grande 4 redbeds and Sutton Lake basalt suggest that these may be younger and that they are correlatable with the Flaherty volcanics and Haig intrusives from the Belcher Islands.

Résumé

En soixante-huit sites, situés le long de la côte est de la baie d'Hudson, on a recueilli trois-cent quatre-vingt-huit carottes de forage orientées, dans des terrains volcaniques et à couches rouges (redbeds) d'âge protérozoïque. Les études détaillées de démagnétisation AF dans des domaines atteignant 60 mT et de la démagnétisation thermique, ont permis de trouver 35 sites acceptables, dont la plupart se situent dans le golfe de Richmond. La couverture basaltique de Nastapoka (élément rémanent: magnétite) avait cinq sites de ce type ne contenant apparemment pas plus d'un élément, de coercivité et température de blocage élevées. La position du pôle est 9°W,39°S; 6°,4°. Au sud du golfe de Richmond, la même couche (de 130 m d'épaisseur) a donné 6 sites caractérisés par de faibles inclinaisons négatives, et sans doute contaminés par une composante secondaire, et deux sites de magnétisation stable et parallèle à une direction ultérieure à une période de plissement, dans les îles Belcher avoisinantes. Trois sites centrés sur les sédiments sous-jacents de la formation de Qingaaluk étaient caractérisés par une position comparable de leur pôle, soit 13°W,40°S; 11°,6° (éléments rémanents: hématite et magnétite). Les roches volcaniques plus anciennes de Persillon, recueillies sur la rive sud du golfe de Richmond, ont donné un pôle situé à 12°E,46°S; 12°,9°, d'après l'étude de cinq sites où la magnétite et l'hématite étaient les éléments rémanents. Les positions de ces trois pôles ne diffèrent pas fortement l'une de l'autre. Cependant, dans le partie nord du golfe de Richmond, les sites de Persillon présentaient une direction différente, d'après l'examen de 11 sites, dont le pôle se situait à 46°W,49°S; 4°,2°. Ce groupe de sites se laisse légèrement mieux regrouper statistiquement, si l'on corrige les petites différences de l'attitude des divers blocs faillés; ceci suggère qu'en très grande partie, la magnétisation date d'avant le période de faillage. Cette dernière est antérieure à l'épisode de Nastapoka, de sorte que la coincidence générale que manifestent les directions deNastapoka et Persillon indique peut-être que ces directions sont essentiellement primaires. D'autre part, les positions des pôles correspondent à celle du pôle déterminé d'après la magnétisation qu'ont subie, avant la période de plissements, les roches volcaniques d'Eskimo, près des îles Belcher; il existerait donc une corrélation chronologique approximative. Les résultats préliminaires dérivés de l'étude des quatre couches rouges (redbeds) de La Grande et des basaltes de Sutton Lake indiqueraient que ces terrains sont plus récents, et qu'on peut les corréler avec les terrains volcaniques de Flaherty et les roches intrusives de Haig, sur les îles Belcher.

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INTRODUCTION

The Circum-Ungava belt is a large sinuous structure of Proterozoic rocks in northeastern Canada ranging from the Labrador Trough to Wakeham Bay-Cape Smith in Northern Quebec and extending to the Belcher and Manitounuk islands in Hudson Bay. It is a sedimentary-igneous sequence overlying the Archean rocks of the Superior Province.

A large variety of rock types of different degrees of deformation and metamorphism occurs within the belt. The first part of the study centred on Cape Smith Island at the western edge of the almost east-west trending Cape Smith Range (Fujiwara and Schwarz, 1975). The thick succession of mafic and ultramafic flows and sills yielded good paleomagnetic results but the age of magnetization could not be ascertained, the bulk of the evidence suggesting a Hudsonian direction. This result is of limited value in attempting to correlate various parts of the belt. Therefore, the slightly deformed and weakly metamorphosed igneous and/or hematitic sedimentary rocks of Richmond Gulf and the Manitounuk Islands were investigated (see also Schwarz, 1976).

Geology and Sampling

The strip of Proterozoic rocks along the East coast of Hudson Bay (Fig. 15.1) between Inoucdjouac and James Bay consists mainly of basaltic and clastic sedimentary rocks (Stevenson, 1968; Eade, 1966). The basalts form a monocline which dips gently towards the Hudson Bay coast near Richmond Gulf and on islands to the north and south. In the Richmond Gulf area, basalts and sediments show subgreenschist facies metamorphism and are affected by weak east-west folding and faulting. Similar Proterozoic basalts form the Hopewell Islands but the metamorphic grade is greenschist. In the Richmond Gulf area, the basaltic rocks are overlain conformably by clastic sediments which form the Nastapoka Islands. This sequence of basalts and sediments is

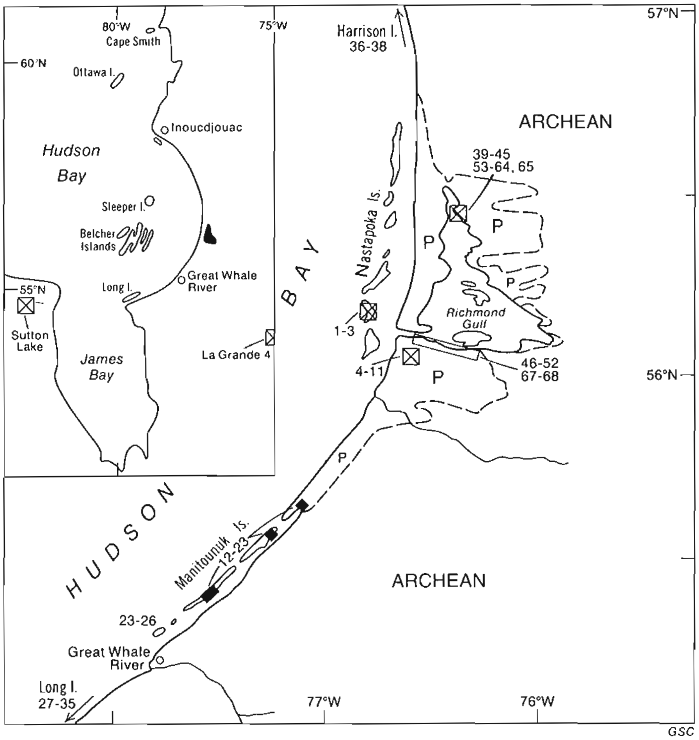


Figure 15.1. Map indicating the distribution of Proterozoic rocks (P) in the vicinity of Richmond Gulf on the east coast of Hudson Bay. Distribution of the sampling sites is indicated by their numbers. Sites collected from Harrison Island near Inoucdjouac and from Long Island are off the main map but can be located on the inset.

Table 15.1. Stratigraphic column showing approximate levels of sampling

Group	Formation	Lithology	Thickness	Sampling sites
Nastapoka		Clastic sediments	>60 m	1-3
		Basaltic flows	130 m	4-38
		Dolomite	150 m	
		Unconformity		
Richmond Gulf	Qingaaluk	Pink sandstone	500 m	39-44
	Persillon	Basaltic flows	30 m	48-64
	Pachi	(Red) Arkose	150 m	65-68
Archean rocks				

NOTE: The Richmond Gulf Group and Pachi Formation occur only in the Richmond Gulf area; elsewhere the Nastapoka Group rests on Archean basement (Chandler and Schwarz, 1980).

called the Nastapoka Group, which unconformably overlies the Richmond Gulf Group pinkish arkosic sandstone in the Richmond Gulf area, and elsewhere Archean basement rocks. This sandstone overlies unconformably the Persillon Formation andesitic volcanics and Pachi Formation redbeds which also occur exclusively in the Richmond Gulf area. Table 15.1 shows a stratigraphic column based on Chandler and Schwarz (1980).

A total of 388 oriented drill cores were taken from 68 sites. Orientation was done by sun compass whenever feasible and otherwise by magnetic compass. In the latter case, checks were performed to avoid substantial errors due to strong local magnetization by sighting prominent topographic features. The cores (diameter: 2.5 cm) each yielded at least two cylindrical specimens of nearly 2.5 cm in length.

RESULTS OF THE MEASUREMENTS

The direction and intensity of the remanent magnetization of the specimens from the volcanics were measured with an astatic magnetometer and those of the sediment samples with a spinner magnetometer with higher sensitivity. Alternating field and thermal demagnetization was performed in up to 27 steps on at least two specimens per site in order to evaluate in detail coercivity and blocking temperature ranges. If the direction remained constant in all test specimens and showed good coherency the rest of the specimens of the same site were treated at a single temperature or field strength. In all other cases, the rest of the specimens of the site were treated in a range of field strengths (up to 60 mT) or heated to a series of increasing temperatures and the remanence remeasured after each step. The site data selection was based on whether endpoints were reached during treatment and, if so, on data yielding the best grouping. For 35 of the 68 sites, 2 specimens of at least three cores showed well-defined endpoints and within-core and between-core coherency and these results were accepted. Only the directions of the Natural Remanent Magnetization (NRM) for the remaining sites are given in Table 15.2 because no stable direction of magnetization could be isolated. Only a few specimens from a few of these sites yielded endpoints, and these results are used as evidence for rough stratigraphic correlation but not for the calculation of a paleomagnetic pole.

NASTAPOKA GROUP

The experimental results are discussed by main sampling area: Richmond Gulf, Manitounuk Islands, Long Island and Harrison Island (see Fig. 15.1).

Richmond Gulf Area

The youngest rocks sampled are the clastic sediments which form the Nastapoka Islands (Fig. 15.1). The NRM directions for the three sites are given in Table 15.2. Detailed AF and thermal testing yielded highly variable directions in most cases without within-core coherency (angle between specimen directions >25°) and they are therefore unfit for further consideration.

Eight sites were collected from the stratigraphically lower basalts on the mainland. The NRM directions are towards the southeast with either up (sites 7 to 11, Fig. 15.1) or down average inclinations (Table 15.2). Representative single specimen results obtained after AF demagnetization are shown in Figure 15.2. They show that sites 7 to 11 generally contain one remanence component occupying essentially the whole coercivity spectrum with, in most

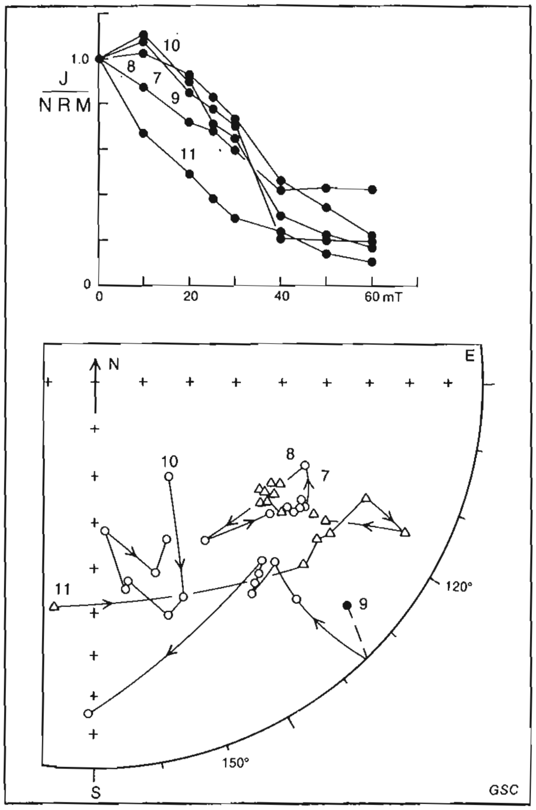


Figure 15.2. Effect of AF demagnetization on Nastapoka basalt from Richmond Gulf: top shows intensity change relative to original NRM and bottom part shows concomitant changes in direction. Samples are indicated by their site number and closed and open symbols indicate respectively downward and upward direction.

cases, a superimposed low-coercivity viscous component. The best within-site grouping was observed after the 25 mT treatment yielding the results in Table 15.2. Thermal demagnetization yielded comparable results with essentially one component occupying almost the full blocking temperature range above 400°C (Fig. 15.3). The average direction for this group of sites is listed in Table 15.2.

Sites 4, 5, and 6, collected from the middle and upper levels of the same basalt yielded different results (Fig. 15.4). Firstly, good endpoints are generally not indicated. Secondly, poor within-site coherency of the directions is due to some core directions displaying negative inclinations (upward directions) while others have positive inclinations. Both generally show southeastern declinations. Stable magnetization could not be isolated in the majority of the specimens from these sites. One component of the magnetization may well be the southeast-up component carried by the majority of the specimens from sites 7 to 11. The decay of the intensity is more rapid than that generally observed for specimens from sites 7 to 11 (Fig. 15.2, 15.4).

The 24 accepted core directions (out of the total of 49) show best grouping after the 25 mT treatment (Fig. 15.5). The average direction of this group is given in Table 15.2 and the corresponding pole position is given in Table 15.4 both before and after rotation of the flow planes to horizontal. These results will be discussed in a later section together with results obtained from the other sampled areas.

Table 15.2. Data for sites from Nastapoka sediments and basalts

Site	Formation	NRM					After demagnetization						
		n	D	I	k	α_{95}	AF	T	n	D	I	k	α_{95}
1	Nastap. Sed. NI	5	326	+71	65	9							
2		5	281	+82	31	7							
3		6	72	+83	7	16							
4	Nastap. Bas. RG	7	146	+71	14	14							
5		6	107	+34	2	44							
6		6	153	+29	6	24							
7		6	133	-31	11	9	25		3	135	-38	15	18
8		6	128	-32	14	18	25		5	135	-38	46	9
9		6	128	-29	53	9	25		4	127	-41	331	3
10		6	130	-41	103	6	25		6	129	-41	119	6
11		6	123	-39	8	19	25		6	121	-41	9	19
Average sites 7 to 11									24	127	-40	32	5
12	Nastap. Bas MI	4	296	+85	169	7							
13		6	149	+75	4	28							
14		3	216	+88	113	8							
15		6	139	+68	17	14							
16		6	205	+70	1500	2	60		6	197	71	36	11
17		6	189	+66	138	5	60		6	210	70	61	9
18		6	57	+18	2	47	20	560	5	113	-15	339	4
19		6	115	+54	2	42	25		3	118	-16	25	25
20		6	91	+34	4	20	50	560	6	92	- 8	85	7
21		6	117	+35	5	26	25		6	112	-12	90	8
22		6	95	-26	8	20			4	100	-17	114	9
23		7	104	+36	5	25	25		4	100	-16	164	7
24		6	118	+69	7	23							
25		6	184	+86	9	19							
26		6	120	+72	4	32							
Average sites 18 to 23									28	107	-13	50	4
27	Nastap. Bas LI	6	277	+ 3	3	33							
28		6	122	+70	9	20							
29		6	294	+65	10	16							
30		6	69	+82	10	18							
31		7	256	+ 8	38	9							
32		7	304	+25	4	28							
33		6	286	+58	12	16							
34		6	346	+76	18	13							
35		7	260	+78	17	13							
36	Nastap. Bas HI	6	98	+79	5	29							
37		6	99	+63	15	18							
38		6	111	+68	8	23							

NI - Nastapoka Islands
RG - Richmond Gulf
MI - Manitounuk Islands
LI - Long Island
HI - Harrison Island

Site average direction based on n cores is given by declination (D)

Inclination (I; positive and negative represent respectively downward and upward) and grouping statistics are indicated by the normally used parameters k and α_{95}

Demagnetization treatment is indicated as AF (alternating field with peak intensity in mT) and T(thermal, in °C), yielding best grouping of endpoints.

Table 15.3. Data for sites from Qingaaluk sediments, Poersillon volcanics and Pachi sediments all from the Richmond Gulf area.*

		NRM					After demagnetization						
Site	Formation	n	D	I	k	α_{95}	AF	T	n	D	I	k	α_{95}
39	Qingaaluk N	6	228	+86	23	14		550	3	149	-36	9	24
40		5	93	+67	48	11		550	5	129	-48	34	9
41		6	280	+81	31	17		550	4	123	-24	38	20
42		6	343	+73	75	11							
43		5	309	+76	24	26							
44		5	48	+87	46	11							
45		6	315	+79	47	11							
Average sites 39 to 41									12	135	-41	14	10
46	Qingaaluk S	6	161	+11	2	20							
47		6	267	+70	5	19							
48	Persillon S	6	156	-21	21	13	30	560	6	148	-35	42	9
49		7	165	-62	2	39	30	520	6	137	73	7	20
50		6	115	-47	13	16	30	540	6	112	55	21	12
51		6	77	+17	2	41	25	620	4	98	-56	18	19
Average sites 48 to 52									25	125	-54	12	9
53	Persillon N	5	232	+50	2	66							
54		5	77		1	138			3	174	-25	80	9
55		5	199	+47	4	42			4	171	-13	90	10
56		6	185	-13	2	65		620	4	157	-51	155	5
57		5	176	-53	14	21		620	4	157	-49	261	3
58		4	135	-48	13	26		640	3	149	-43	123	6
59		5	178	-13	16	17		640	4	171	-25	91	5
60		6	190	+25	5	38		640	4	173	-25	31	12
61		5	169	-43	22	15		620	6	160	-50	80	5
62		5	216	+63	2	74		640	4	161	-34	144	8
63		5	159	-18	2	67		550	4	141	-46	21	14
64		6	160	- 7	98	6	30		6	157	-28	484	3
Average sites 58 to 67									46	161	-43	23	4
65	Pachi	5	224	+75	35	13							
66		6	315	+48	106	4	60		6	313	46	90	6
67								670	5	313	44	41	10
68		6	322	+55	65	7	30		6	301	55	65	7

*Symbols as in Table 15.2

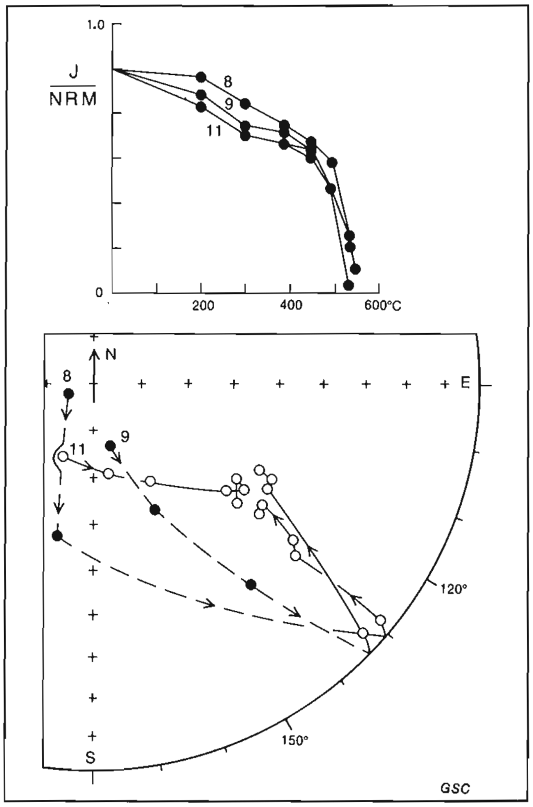


Figure 15.3. *Effect of heating and subsequent zero-field cooling on Nastapoka basalt from Richmond Gulf. Top part shows relative intensity change. Symbols and labelling as in Figure 15.2.*

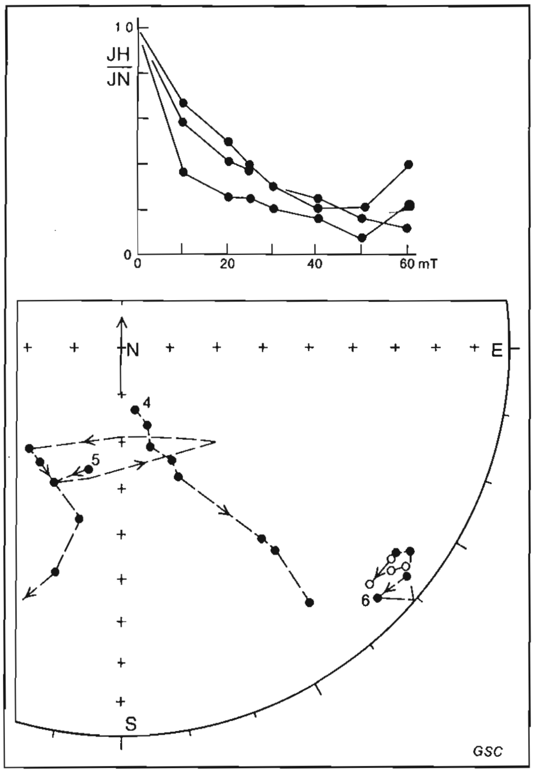


Figure 15.4. *AF demagnetization results for representative specimens from sites 4, 5 and 6, Nastapoka basalt, Richmond Gulf.*

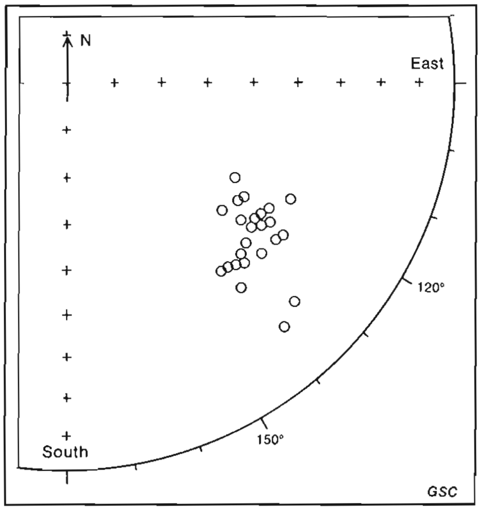


Figure 15.5. *Core magnetic directions accepted for Nastapoka basalt from Richmond Gulf as determined by averaging 25 mT AF results of two specimens per core.*

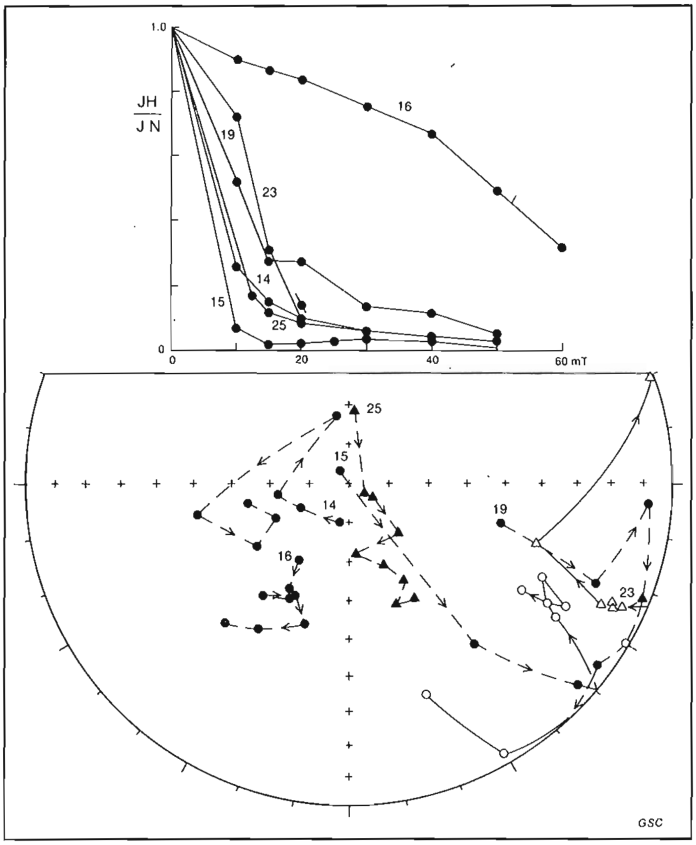


Figure 15.6. *Representative AF results for specimens of Nastapoka basalt from the Manitounuk Islands. Conventions as in Figure 15.2. NRM of most specimens contains large low-coercivity component compared to the data given in Figure 15.2.*

Table 15.4. Average direction before and after tilt correction and their corresponding pole positions for Nastapoka basalt.

Formation		Average before tilt				Pole position				Average after tilt				Pole position			
		D	I	k	α_{95}	Long.	Lat.	dm	dp	D	I	k	α_{95}	Long.	Lat.	dm	dp
Nast. Bas RG	24	127	-40	32	5	5W	39S	6	4	130	-37	32	5	9W	39S	6	4
Nast. Bas MI	35	108	-13	50	4	2E	15S	4	2								
Nast. Bas HI	12	203	+70	74	4	91W	19N	6	5								
Qi Sed.	12	135	-41	14	10	12W	43S	12	7	133	-37	14	9	13W	40S	11	6
Pe. Volc. N.	46	161	-43	23	4	45W	55S	5	3	160	-33	35	3	46W	49S	4	2
Pe. Volc. S.	25	125	-54	12	9	7E	47S	12	9	122	-55	13	8	12E	46S	12	9
Haig Int. BI		20	-45	4	26					40	-46	23	6	123W	1N		
Fla Volc. BI		2	-45	4	26					40	-44	46	7	126W	0		
Esk. Volc. BI		114	-64	3	49					119	-46	36	10	2E	40S		

RG - Richmond Gulf
MI - Manitounuk Islands
Qi - Qingaaluk sediments
Pe - Persillon volcanics from the northern (N) and southern (S) parts of the Richmond Gulf area

Number of cores is indicated by n. Average directions and pole positions for the Haig intrusives, Flaherty volcanics and Eskimo volcanics all from the Belcher Islands are also listed from Schmidt (1980).

Manitounuk Islands (MI)

These islands are formed by Nastapoka Group basaltic rocks similar to those that form the ridge between Richmond Gulf and Hudson Bay (sites 4 to 11).

All sites (12 to 26, Table 15.2) except 22 display downward NRM directions which are generally poorly grouped as was also observed for sites 4, 5, and 6. The declination is highly variable. AF and thermal demagnetization indicate few well-defined endpoints, the remanence direction changing either continuously along a circular path or erratically when the intensity has reached relatively low levels as shown in Figures 15.6 and 15.7. Specimens which show endpoints such as those from sites 18, 20, and 12 in Figure 15.7 indicate the presence of two basic directions: (1) southwest-steeply-down, and (2) southeast – up. The southwest-down direction is consistently displayed by cores from sites 16 and 17 as well as by individual specimens from many other sites. Sites 16 and 17 show high coercivities but in all other cases intensity decay is relatively rapid (compare Fig. 15.2 and 15.6). This direction is similar to that of the Cape Smith komatiitic basalts which is likely to be secondary (Fujiwara and Schwarz, 1975). A similar direction reported by Schmidt (1980) from the Belcher Islands, is demonstrably postfolding (Hudsonian).

The other basic direction in the Manitounuk Islands is southeast-up. This direction is somewhat comparable to that for sites 7 to 11 from Richmond Gulf. In fact, the behaviour of some specimens (e.g. one from site 15 on Fig. 15.6) closely resembles that of specimens from sites 7 to 11 (see 8 and 9 in Fig. 15.2). Sites 18 to 23 rather consistently show this direction (Table 15.2). The remainder of the sites do not show a reasonably well-defined average direction because (1) many specimens do not show endpoints, (2) core directions would be based on widely divergent specimen directions where endpoints exist and (3) less than 3 cores define a site direction.

The accepted core directions from sites 18 and 23 are shown in Figure 15.8. The declinations exhibit a large spread and their average is 20° less than that of sites 7 to 11. The inclination shows little spread compared to that of sites 7 to 11, but its average is much lower than that of the latter group. If the magnetization of sites 18 to 23 is a single component magnetization the difference between the MI and Richmond Gulf results must be due to:

1. A difference in age. Although precise time-stratigraphic correlation between MI and RG has not been established, it seems unlikely that the deposition of about 130 m of basalt without substantial intercalations of sediment would have taken a long period. In both the RG and MI data, however, secular variation has probably not been averaged out as sampling was confined to basal and intermediate layers. It could have produced as much as half the difference of 32° between the average RG and MI directions.
2. Difference in bedding attitude is confined to strike which is essentially south in Richmond Gulf and roughly southwest on the Manitounuk Islands. Rotation to horizontal about strike hardly affects the difference in the average magnetic directions.

On the other hand, if the MI magnetization contains a small component of the MI southwest-down overprint, it should lie on a great circle between the overprint direction and the average RG direction. This is not the case (Fig. 15.8). Furthermore, the MI average is not in the plane which includes the RG average and the present geomagnetic field (GMF). Thus, the difference between the average MI and RG directions is not resolved by considering two remanence components only.

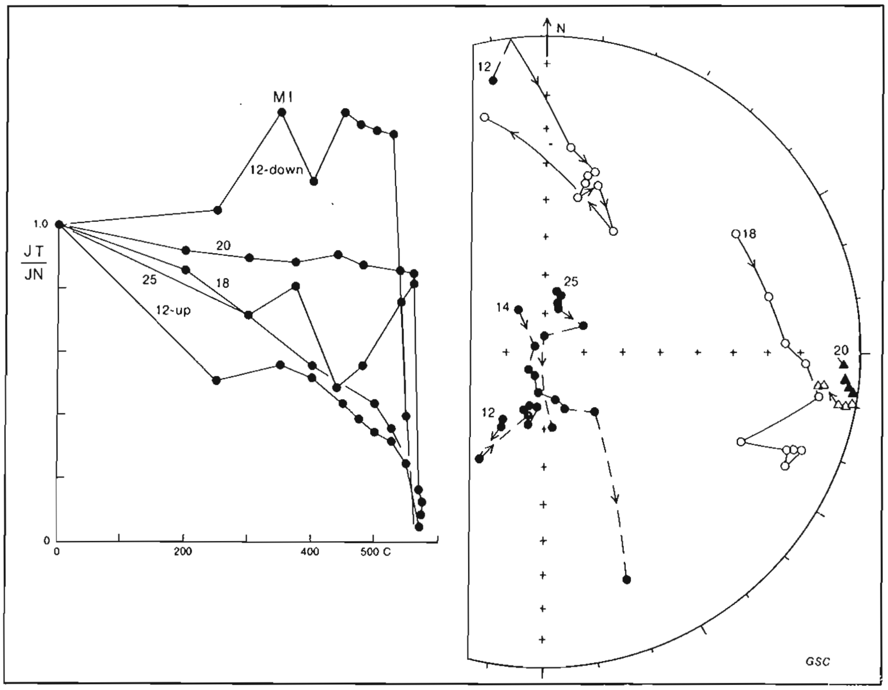


Figure 15.7

Thermal demagnetization of Nastapoka basalt from the Manitounuk Islands. Magnetization is carried by magnetite. See Figure 15.2 for symbols.

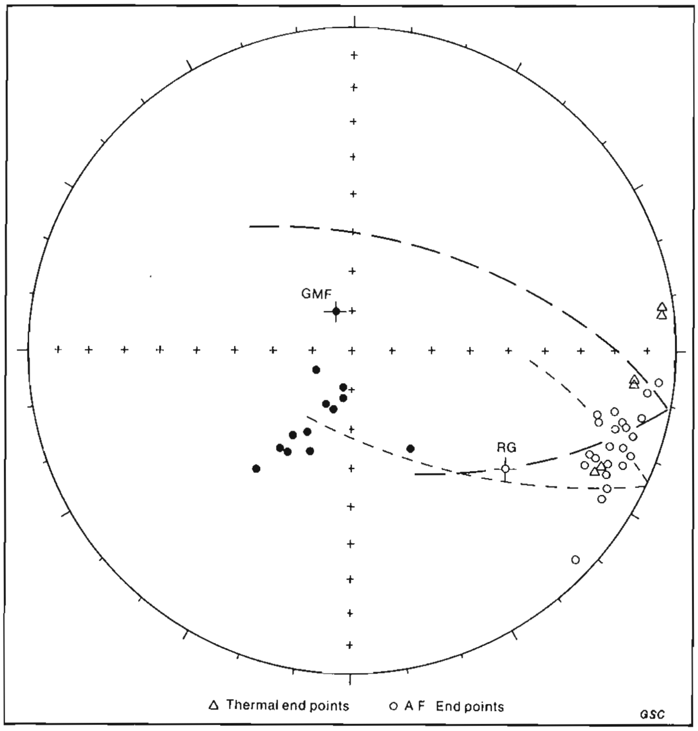


Figure 15.8. *Stable core directions after AF (circles) or thermal (triangles) treatment for basalts from the Manitounuk Islands. RG gives the average direction for the basalt cap at Richmond Gulf. Dashed great circles show that the Manitounuk direction cannot be regarded as a simple hybrid between two components i.e. RG and either the present field direction (GMF) or RG and the probably secondary direction to the southwest-down detected in two sites of Manitounuk basalt.*

Long Island (LI) and Harrison Island (HI)

The large number of cores collected on Long Island near James Bay yielded only a few individual specimen results that showed some stability in direction during AF or thermal treatment. One of these is shown in Figure 15.9 (site 32, thermal). The three sites from Harrison Island near Inoucdjouac yielded two such individual specimens (site 37, AF). Intensity decay is rapid. These results suggest that an easterly-up direction is present along the coast of Hudson Bay between Inoucdjouac and Long Island.

RICHMOND GULF GROUP

The Richmond Gulf Group (RGG) occurs in the Richmond Gulf area between the Nästapoka Formation and the Archean basement. Chandler and Schwarz (1980) showed that extensive faulting affected Richmond Gulf Group before deposition of the Nastapoka Group sediments and basalt and subdivided the group into three formations (Table 15.1).

Qingaaluk Formation

Samples of reddish sediments were collected from nine sites in the Qingaaluk Formation. The NRM directions for these sites vary strongly with, in many cases, poor within-site grouping (Table 15.3). In all except three, thermal and AF treatment did not result in three or more consistent core directions with good individual specimen endpoints per site. However, although within-site grouping of the directions of these sites (39, 40, 41) is not good, their mean corresponds to that of the Nastapoka basalts (Table 15.2, 15.3). Detailed thermal demagnetization results for some individual specimens from these sites are shown in Figure 15.10. The following observations can be made:

1. The northwest-down direction displayed by some specimens generally contains both low and high blocking temperature components. In most specimens, the direction is rather stable throughout the blocking temperature range but in a few (e.g. 41 in Fig. 15.10) the highest

blocking temperature component is towards the southeast and therefore may correspond to the stable direction isolated for the Nastapoka basalt. Ascribing the low blocking temperature component to viscosity in the present Earth's field, the high blocking temperature northwest-down component may be a chemical remanence acquired recently rather than an original component representing a field reversal with respect to the southeast-up component.

2. Many specimens display a rotation of the magnetization vector in a subvertical plane from down to up and subsequent stabilization in a southeast-up direction. Clearly, thermal treatment results in elimination of a low blocking temperature viscous component along the Earth's field direction. However, a few specimens display at least a tendency to yield a high blocking temperature component oriented (north) east-up, carried either by magnetite (40 in Fig. 15.10) or by hematite (39).

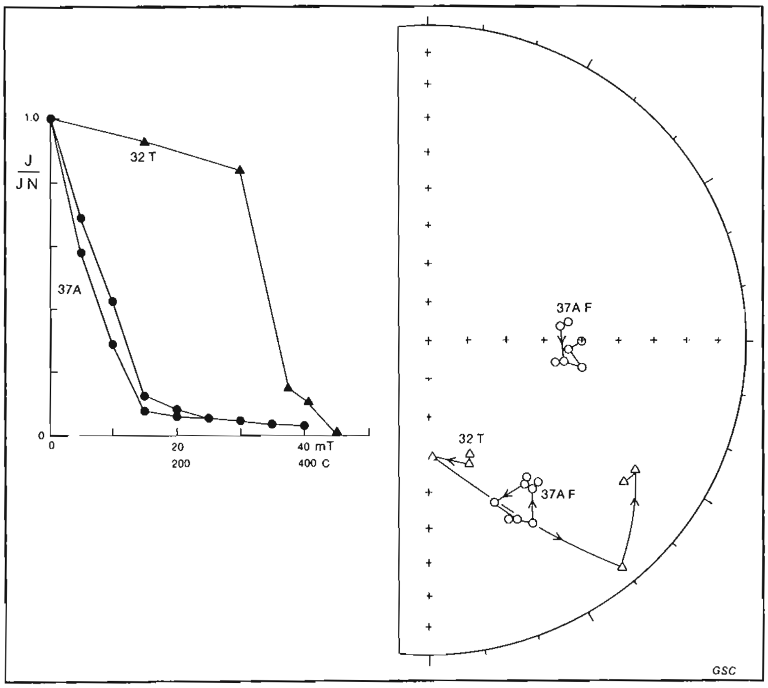


Figure 15.9. *Alternating field (AF) data for two cores from Harrison Island, and thermal data for one (32) core from Long Island.*

Persillon Formation

The large majority of the sites collected from the Persillon volcanics carry a large stable component, their NRM directions being towards the southeast and upward (Table 15.3). The sites have been divided into two groups because of a difference in declination which persists after detailed AF and thermal cleaning. The Persillon South (S) group is formed by 5 sites (48 to 52) collected in the southwestern corner of Richmond Gulf. The other group of 12 sites (53 to 64) was taken in the northern part of Richmond Gulf (Fig. 15.1).

Detailed AF and thermal results are shown in Figures 15.11 and 15.12 respectively. Figure 15.11 shows that the AF demagnetization to 60 mT does not break down the high coercivity portion of the NRM. Directional stability is high and there are no indications that the high coercivity portion of the NRM is multi-component. Typical thermal results show that magnetite carries a large part of the NRM in a few specimens but that hematite is generally an important carrier. There are no indications that the components carried by the two minerals differ in direction. The directions usually show good endpoints comparable to those obtained by AF demagnetization. The main NRM component is oriented to the southeast-up as observed earlier for Qingaaluk Formation sediments and the Nastapoka Group basalt with the exception of higher declination values for the Persillon volcanics collected in the northern part of Richmond Gulf.

As is the case for a few specimens from Qingaaluk Formation, a few specimens from the Persillon also show a high blocking temperature (660°C or higher) component with different declination (specimen 57 in Fig. 15.12). The great majority of the individual specimens, however, do not, show a systematic change in direction for a weak high blocking temperature component (e.g. specimens 50, 51, 60). The southeast-up component, carried by both magnetite and hematite represents the stable magnetization of these volcanics and the weak high blocking temperature components indicated by a few specimens probably represent CRM overprints due to recrystallization of hematite. Whatever its origins, there is very little if any coherency in the direction of this component.

For most of the sites, up to three cores out of six showed a northwest-down direction carried by magnetite, a southwest-down direction with blocking temperatures between 600 and 670°C (see also 57 and 58 in Fig. 15.12), and in a few cases an indication of a weak component towards the southeast (specimen 62 in Fig. 15.13). The low blocking temperature hematite southwest-down component represents about one third of the original NRM intensity. A reasonably accurate isolation of this component was not achieved because of the large scatter in direction in its blocking temperature range and also in vectors removed. However, the general direction of this component corresponds to the predominant low stability direction of the Nastapoka basalts from the Manitounuk Islands, and also to the direction interpreted by Schmidt (1980) as secondary on the basis of a fold test for the Belcher Islands.

Figure 15.14 shows all core directions accepted on the basis of coherent specimen endpoints. The population is streaked due to a comparatively large variation in declination, but is roughly divisible into two more or less isometric parts when the directions are plotted according to sampled area. This subdivision is arbitrary in the sense that there were two main sampling areas of the Persillon volcanics – one on the southern and one on the northeastern shore of Richmond Gulf. However, the populations from the south and north are significantly different as is borne out by the average direction for the two groups of data and their statistics (Table 15.3). The question is clearly why do they not coincide.

1. It is unlikely that there is a substantial difference in age between the volcanics sampled on the south shore and those on the north shore as the volcanics are only about 30 m thick.
2. It is not clear that one of the groups is a hybrid direction due to the concurrence of at least two directions which are inseparable by blocking temperature and coercivity. A possibility would be that the north shore data are contaminated by a relatively weak[1] southwest steeply-down magnetization; a component indicated by several specimen results as mentioned previously. Although a

[1]The angles between the mean Persillon vectors and their similar average intensities indicate that this secondary component could amount to about 50% of the mean vector of the Persillon south group after thermal and AF treatment as in Figure 15.14.

reasonably precise estimate of the direction of this component has not been possible, a great circle through the centres of the north and south shore data groups contains the general direction of this southwest component (Fig. 15.13, 15.14). The north and south shore data are therefore treated separately for the calculation of paleopoles.

3. Differential rotation between north and a south blocks around a roughly vertical axis of, say, 20° is not supported by geological data given by Chandler (1979). However, since detailed geological mapping has not extended to the Archean rocks immediately east of Richmond Gulf, this possibility cannot be discarded. As discussed in a later section, correcting for tectonic tilt results in a slight improvement of the grouping of the data but the average directions for both groups of samples do not change appreciably.

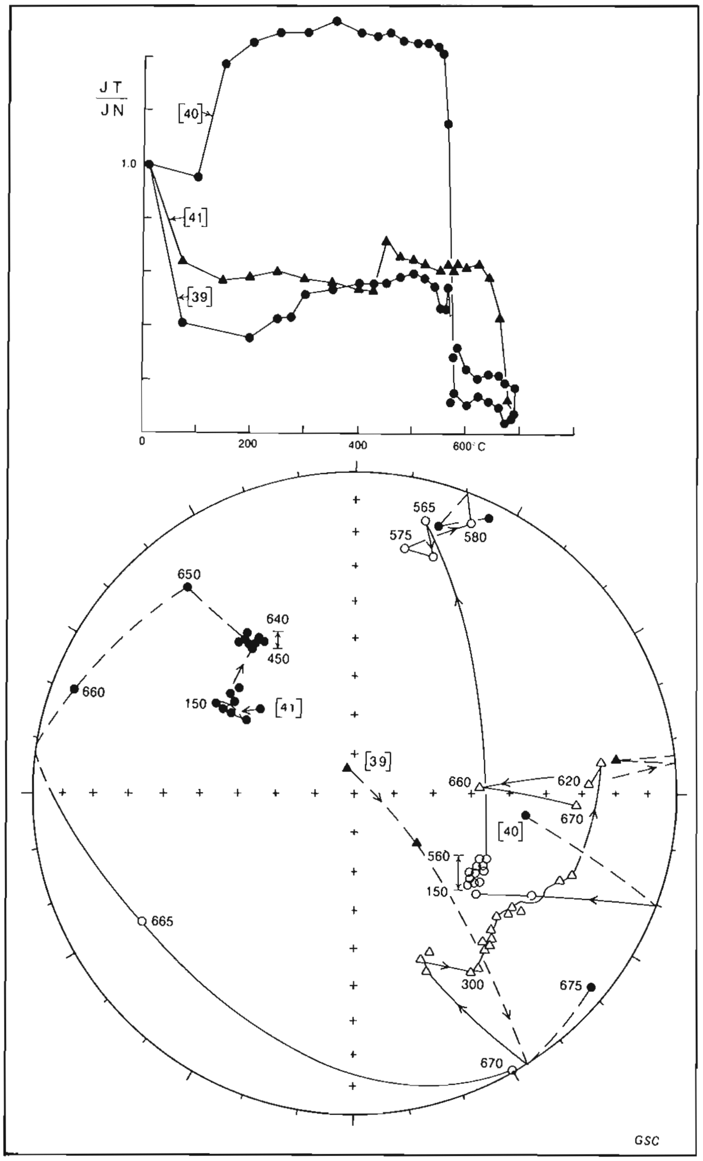


Figure 15.10. *Thermal demagnetization of Qingaaluk Formation redbeds from Richmond Gulf. Numbers in brackets are the site number and those without brackets indicate a temperature of heating. Magnetite and hematite carry the NRM.*

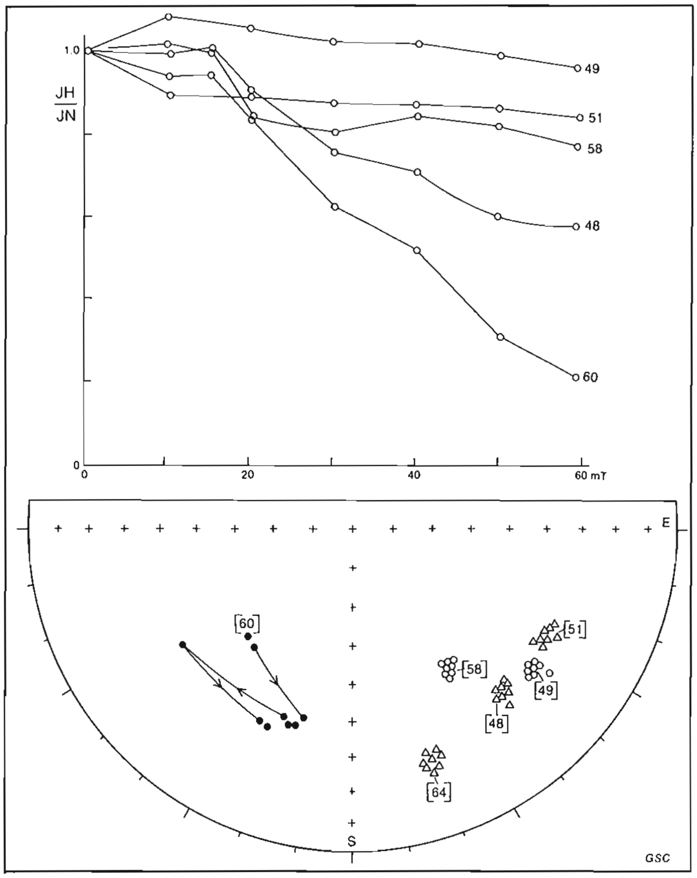


Figure 15.11. *AF demagnetization of Persillon Formation andesite specimens. Most specimens reveal a large high-coercivity NRM component.*

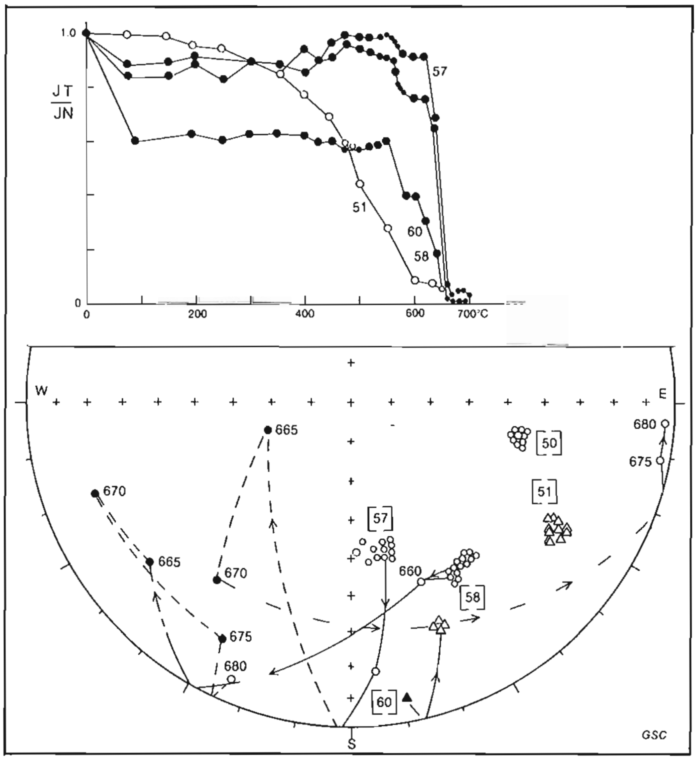


Figure 15.12. *Thermal demagnetization of Persillon andesites. NRM is carried by hematite and magnetite.*

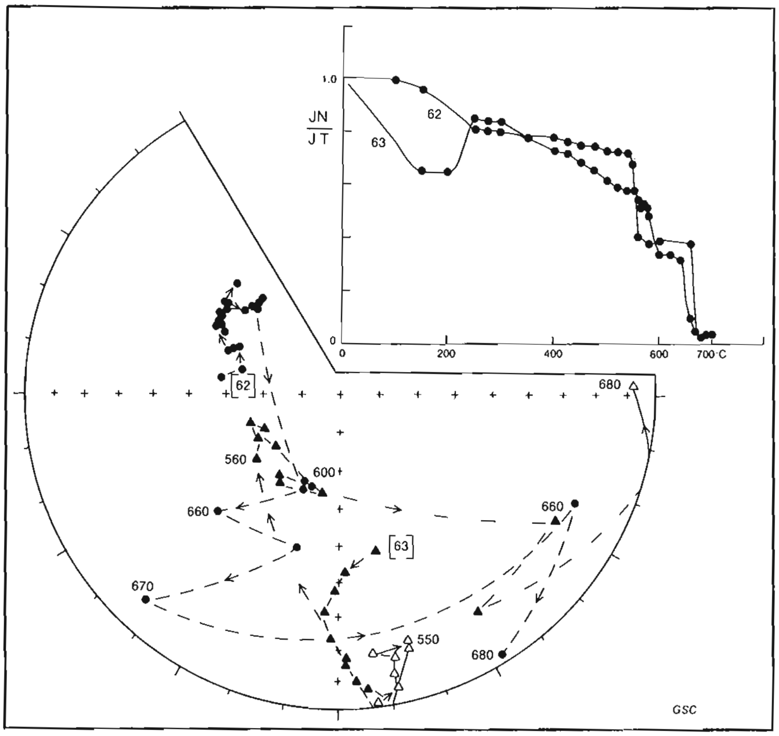


Figure 15.13. *Evidence for remanence components in different blocking temperature ranges in Persillon andesite.*

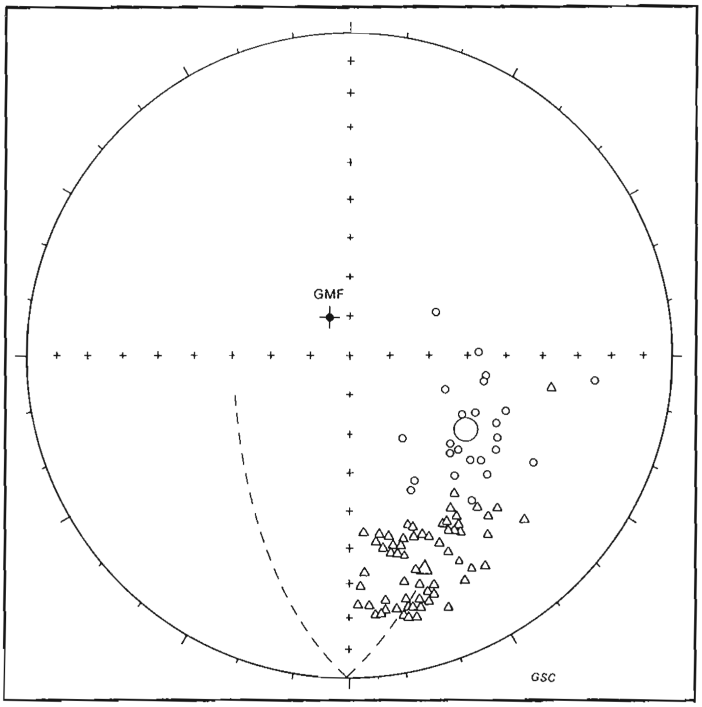


Figure 15.14. *All accepted core directions for Persillon andesite along the south shore (circles) and the north shore (triangles) of Richmond Gulf. The large symbols indicate the mean directions as given in Table 15.4.*

SIGNIFANCE OF THE RESULTS

Table 15.4 shows average directions and pole positions before and after correction for bedding plane tilt for the various groups of samples. These are based on stable and coherent core directions pointing southeast-up, and, for sites 16 and 17 (MI), southwest-down. The latter direction probably is a secondary direction and the tilt correction which involves a common rotation of 6°, has not been applied. The direction for Nastapoka Group basalts from the Manitounuk Islands shows a much lower inclination and median coercivity than for the same group near Richmond Gulf. As discussed earlier the MI direction may be a hybrid although it is not clear what its other component would be if there are only two. The Nastapoka direction for Richmond Gulf, on the other hand, is essentially similar to the Persillon (S) direction, carried by both magnetite and hematite. Therefore, the Nastapoka direction determined for the Richmond Gulf sites must be regarded as the more reliable. The Manitounuk Islands direction is therefore not further considered.

The Richmond Gulf Group was block faulted; the overlying Nastapoka Group in the Richmond Gulf area forms a monocline with a 7° westerly dip. Consequently, only the Richmond Gulf Group sites can be expected to show a change in grouping by rotating the bedding planes to horizontal assuming zero primary dip. The bedding dip is up to about 20° and strike values vary by up to about 30°. The results of this rotation to horizontal are given in Table 15.4 and the test is not significant McElhinny (1964). However, all groups of samples, particularly the Persillon (N) samples, show a small improvement in grouping. Thus, the data suggest that the magnetization of the Richmond Gulf Group is pre-faulting. Also, the essentially similarly directed magnetization of the Nastapoka basalt of Richmond Gulf would have been acquired soon after faulting and therefore would be a primary thermoremanent magnetization. The problem here is that the Persillon North direction is significantly different from the Persillon South, Qingaaluk, and Nastapoka directions and that the effect of tilting the beds to horizontal would be decreased if a secondary component has indeed been acquired after the faulting.

Another result, with a bearing on the age of magnetization in Richmond Gulf Proterozoic rocks is the positive fold test obtained by Schmidt (1980) for the Belcher Island volcanics. This test shows that the Belcher magnetization is pre-folding or pre-Hudsonian. Comparing the pole positions after tilt correction of the Belcher Islands and Richmond Gulf Proterozoic rocks (Fig. 15.16) shows that the Nastapoka-Richmond Gulf sequence may be of similar age as the Eskimo volcanics of Belcher Islands.

Preliminary results for the lower and middle sections of the Sakami Formation sediments (Schwarz and Freda, 1980) and the Sutton Lake diabase (see Fig. 15.1) suggest that these may be correlative with the Flaherty and Haig formations of the Belcher Islands. Thus, the paleomagnetic results available to date suggest a time correlation between various parts of the Circum-Ungava Belt in the manner depicted schematically in Figure 15.15.

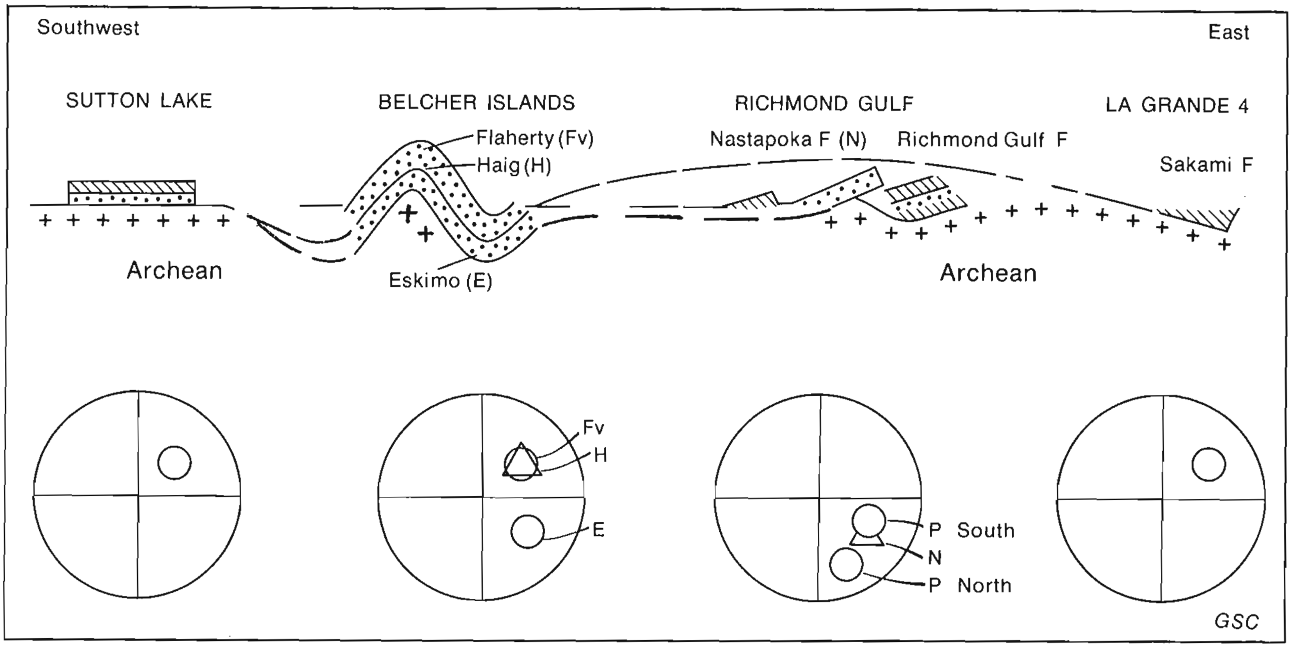
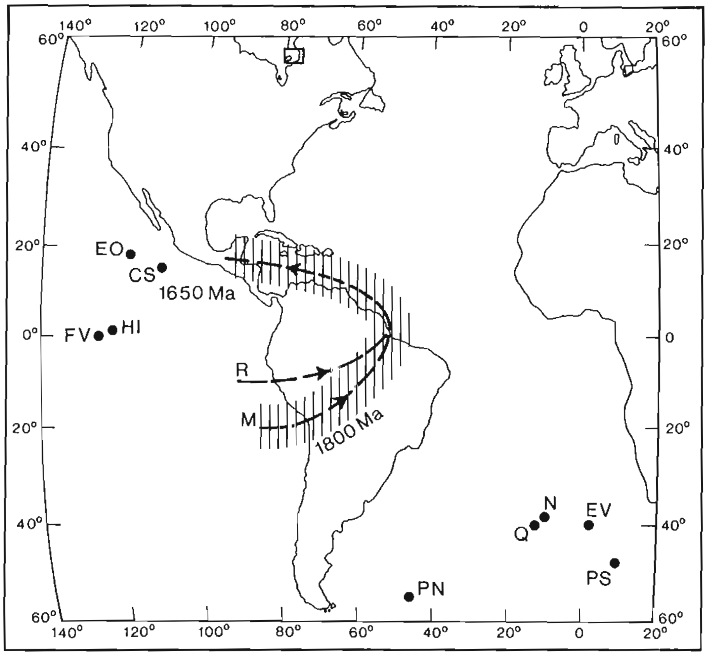


Figure 15.15. *Schematic representation showing possible stratigraphic correlation on the basis of remanent magnetization of Proterozoic rocks in the various regions sampled. Sutton Lake and La Grande 4 data are preliminary.*

Pole Position

The pole positions derived for the Circum-Ungava Belt are shown in Figure 15.16. EO and CS are probably complete overprints and PN may well be based on a multicomponent magnetization. The Richmond Gulf area poles are closely spaced with overlapping areas of 95 per cent confidence (Table 15.4). EV also falls in the group suggesting a similar age of formation for the Eskimo volcanics of the Belcher Islands. Similarity in age between the Eskimo volcanics and the Persillon to the Nastapoka sequence is also suggested by their radiometric ages. Proterozoic ages available for the Richmond Gulf area, which are listed by Chandler and Schwarz (1980), shows a spread from 1600 to 1800 Ma. A reasonable interval seems to be 1700 to 1750 Ma. Schmidt (1980) has listed ages for the Belcher Island rocks and selected a probable age of 1750 Ma for the Eskimo volcanics and 1640 Ma for the Flaherty volcanics and Haig intrusives.

The Circum-Ungava poles with suggested age of about 1750 Ma form a cluster about 50° southeast of the apparent polar wander path (APWP) drawn by McGlynn and Irving (1978). The 1750 Ma segment of APWP is based on poles from the Coronation Geosyncline (Reid et al., in press) and elsewhere in the Shield (McGlynn and Irving, 1978). Assuming the 1750 Ma age is correct, the large deviation of the Circum-Ungava poles from the APWP (Fig. 15.16) may indicate either large-scale pre-Hudsonian relative movement between the Circum-Ungava Belt and the rest of the Churchill Province or that APWP must be redrawn to include these poles rather in the manner proposed by Irving and McGlynn (1979).



EO - Eskimo overprint, all from Schmidt (1980).
PN and PS - Persillon north and south groups respectively,
Q - Qingaaluk sediments,
N - Nastapoka basalt,
CS - Cape Smith volcanics.

Figure 15.16. *Pole positions listed in Table 15.4 after tilting the bedding plane to horizontal. Study area outlined by square at the top of the figure. Band (M) indicates 1800-1700 Ma segment of APWP according to McGlynn and Irving (1978) and line (R) according to Reid et al. (in press).*

ACKNOWLEDGMENTS

K.R. Clark collected the Persillon North volcanics and did all experimental work on this group of samples (1977). K.W. Christie participated in the field work in the period 1974-1976.

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16. **RIFT-RELATED CYCLIC SEDIMENTATION IN THE NEOHELIKIAN BORDEN BASIN, NORTHERN BAFFIN ISLAND**

G.D. Jackson[1] and T.R. Iannelli[2]

Jackson, G.D. and Iannelli, T.R., Rift-related cyclic sedimentation in the Neohelikian Borden Basin, northern Baffin Island; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 269-302, 1981.

Abstract

Mid-Proterozoic sedimentation in the Borden Basin began as 1000 m of braided fluvial to marginal marine quartz arenites and plateau basalts accumulated over a local regolith developed on a peneplaned gneiss complex. Deposition was initially restricted to a narrow fault-controlled channel that merged northwestward into an alluvial braidplain. The overlying 1100 m of strata were deposited during major faulting in large sandstone-shale, marine-influenced delta fan complexes that grade laterally northward into subtidal shales.

Subordinate faulting continued during deposition of the succeeding 1700 m of supratidal to shallow subtidal stromatolitic shelf carbonates that include a subtidal shale zone, and contain economic lead-zinc deposits and many gypsiferous coastal sabkha cycles.

Major faulting accompanied by local erosion and karsting occurred as about 1000 m of interbedded sandstones, shales, carbonates and boulder conglomerates accumulated in alluvial fan to subtidal environments. The uppermost 1200 m of strata were deposited during a relatively stable tectonic interval. Fluvial to intertidal sandstones grade upward into shallow subtidal shelf quartz arenites, and are overlain disconformably by lower Paleozoic strata.

Borden Basin was initiated as a Neohelikian aulacogen in the North Baffin Rift Zone. The Borden is one of several similar, penecontemporaneous, temporarily interconnected, basins which developed by rifting along the northwest edge of the Canadian-Greenlandic Shield. The basins are probably related to the 1200-1250 Ma old opening of the Proto-Arctic Ocean – the Poseidon Ocean.

Résumé

Dans le bassin de Borden, la sédimentation du Protérozoïque moyen a débuté par l'accumulation d'arénites quartzeuses d'origine fluviatile (cours d'eau anastomosés) à marine (milieu marginal) et l'accumulation de basaltes de plateaux, au-dessus d'un régolite local formé sur un complexe gneissique pénéplané. La sédimentation était initialement restreinte à un étroit chenal limité par des failles, qui fusionnait au nord-ouest avec une plaine alluviale réticulée. Les 1100 m de couches sus-jacentes se sont déposés pendant la formation de grandes failles dans les vastes complexes deltaiques qui montrent des influences marines, sont constitués de grès et de schistes argileux, et passent latéralement à des schistes argileux de type subtidal vers le nord.

Ensuite s'est manifestée une phase moins forte de fracturation pendant le dépôt de 1700 m de sédiments stromatolitiques carbonatés sur une plate-forme supratidale à subtidale peu profonde, contenant un niveau subtidal de schistes argileux et des gîtes minéraux exploitables de plomb et zinc, ainsi que de nombreux cycles de sebkha littorale à couches gypsifères.

Une phase de forte fracturation, suivie de processus locaux d'érosion et karstification a accompagné l'accumulation d'environ 1000 m de couches interstratifiées de grès, schistes argileux, carbonates et conglomérats de blocs, dans des milieux de cône alluvial ou de type subtidal.

Les 1200 m supérieurs de strates se sont formés pendant un intervalle de stabilité tectonique relative. Les grès fluviatiles à intertidaux passent progressivement vers le haut à des arénites quartzeuses de plate-forme subtidale peu profonde, et sont recouverts en discordance d'érosion par des strates datant de la fin du Paléozoïque.

Le bassin de Borden a tout d'abord été au Néohélikien un aulacogène dans la zone de rift du nord de l'île Baffin. Le bassin de Borden est l'un de plusieurs bassins similaires, presque contemporains, temporairement reliés entre eux, qui résultent de la fracturation du rebord nord-ouest du bouclier canadien – Groenlandais. La formation de ces bassins résulte probablement de l'ouverture, il y a 1200 à 1250 Ma, de l'océan Proto-arctique – ou océan de Poséidon.

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INTRODUCTION

As much as 6100 m of late Proterozoic (Neohelikian) strata are spectacularly exposed in towering castellated cliffs along inlets and fiord-like sounds of rugged, mountainous northern Baffin and Bylot islands (Fig. 16.1). These strata nonconformably overlie an Archean-Aphebian gneiss complex, are intruded by Hadrynian Franklin diabase dykes, and are overlain unconformably by Paleozoic and Cretaceous-Eocene strata. Lemon and Blackadar (1963) and Blackadar (1970) documented early exploratory work, established the stratigraphic succession, and defined the formations. Other recent studies include those by Blackadar (1968a-d), Galley (1978), Geldsetzer (1973a,b), Graf (1974), Iannelli (1979), Jackson and Davidson (1975), Jackson et al. (1975, 1978a,b, 1980), Olson (1977), and Sangster (1981).

This paper presents primarily results obtained during Operation Borden (1977, 1978 and 1979 field seasons) and results from some previous studies. All Rb-Sr ages quoted have been calculated, or recalculated, using $\lambda^{87}Rb = 1.42 \times 10^{-11} a^{-1}$.

The basic formational nomenclature erected by Lemon and Blackadar (1963) and by Blackadar (1970) for the Neohelikian strata has been retained in this paper, although individual formations have been divided into members (e.g. see Iannelli, 1979; Jackson and Davidson, 1975; Jackson et al., 1975, 1978a,b, 1980). Some formations need to be better defined and reference sections selected to augment the type sections. It is also considered more logical to classify the formations into 3 groups rather than the 2 groups of Lemon and Blackadar (1963).

The three groups – lower clastic (Eqalulik), middle carbonate platform (Uluksan), and upper clastic (Nunatsiaq[1]) are commonly separated by intra-basinal disconformities, and are here referred to collectively as the Bylot Supergroup (Table 16.1, Fig. 16.2).

The Neohelikian strata of the Bylot Supergroup were assigned to the Borden Basin by Christie et al. (1972). Deposition in the eastern part of the basin occurred predominantly in three grabens of the North Baffin Rift Zone (Jackson et al., 1975) separated by basement horsts that both plunge and die out toward the northwest (Fig. 16.2, 16.26).

The Bylot Supergroup has been considered to be of Helikian and/or Hadrynian age by recent workers (Blackadar, 1970; Geldsetzer, 1973b; Jackson, 1969; Jackson et al., 1975, 1978a,b, 1980; Olson, 1977). Ages noted immediately below have been determined by the Geochronological Laboratories of the Geological Survey of Canada.

Seventeen whole rock K-Ar ages determined for Nauyat volcanics range from 762 ± 26 to 1221 ± 31 Ma and have a mean age of 946 Ma (Fig. 16.3, 16.34). In addition, 6 of 14 sample analyses yield a 1129 Ma, 6-point Rb-Sr age with an initial intercept of 0.7120 and a MSWD of 5.5.

Fahrig et al. (1981) obtained paleomagnetic poles for the Bylot Supergroup which they interpreted as Mackenzie. They also interpreted the Nauyat-Adams Sound pole to be about 1220 Ma, and the Strathcona Sound pole to be about 16.5 Ma younger. Both Fahrig et al. (1981) and Chandler and Stevens (1981) considered the mid-late Proterozoic volcanism to be a Mackenzie igneous event, with which we concur.

Sphalerite-galena deposits lie chiefly within Society Cliffs dolostone in the Nanisivik region of Milne Inlet Trough (Fig. 16.26) and are associated with pyrite and hematite-goethite deposits. The sulphide bodies are secondary cavity fillings and replacement bodies emplaced some time after Victor Bay sedimentation and prior to Franklin dyke intrusion (see also: Geldsetzer, 1973a,b; Graf, 1974; Olson, 1977; Sangster, 1981).

Northwest-trending tholeiitic Franklin (Fahrig et al., 1971; Fahrig and Schwarz, 1973) diabase dykes up to 200 m thick cut all Precambrian rocks of the map area (Fig. 16.4), postdate some folding and faulting of the Bylot Supergroup, and are overlain nonconformably by lower Paleozoic strata. Franklin dykes appear less altered than Nauyat Formation volcanics, which have probably undergone subgreenschist facies metamorphism (Jackson and Morgan, 1978). Chemically, the Franklin diabases are more highly differentiated than the Nauyat volcanics. A similar relationship between Neohelikian volcanics and Hadrynian dykes has been observed in the Thule Basin (Fig. 16.34; Frisch and Christie, in press).

Many of the Franklin dykes were emplaced along fault zones and dominant northwest-trending dykes are concentrated along the Tikerakdjuak Fault Zone (Fig. 16.4, 16.25, 16.26). Some subordinate, thinner, northerly trending dykes cut northwest-trending dykes, while others seem to be co-intrusive and to simply be splays off the main northwest-trending dykes. The two trends probably developed pene-contemporaneously in response to brittle fracturing of a block related to tension and torsion caused by movement along bounding fault zones. Dykes of both orientations yield similar whole rock K-Ar ages (Fig. 16.3, 16.34; Blackadar, 1970; Jackson, 1974). To date, all paleomagnetic measurements indicate a single paleomagnetic pole for Franklin dykes and Fahrig (personal communication, 1981) estimates that the Franklin Igneous Event is probably 700-750 Ma old.

BASEMENT COMPLEX

The Archean-Aphebian basement is a variable complex of high-grade gneisses and igneous rocks. The gneisses are chiefly irregularly banded intermediate to acidic migmatites, in part fluidal, that are commonly intruded by at least two

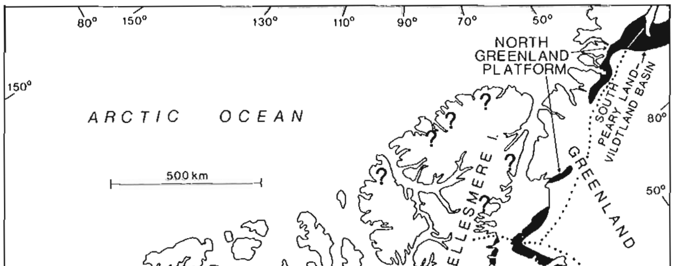


Figure 16.1. Location of strata within Borden Basin and in other Neohelikian basins, northwest edge of Canadian-Greenlandic Shield.

[1]Nunatsiaq (from Nunatsiaq Point, 81°09'W Long., 73°25'N Lat.) is Inuit for "good land".

Table of Formations

Era	Supergroup	GP	Formation	Units
HADRYNIAN			Franklin Intrusions: Diabase	
			Intrusive Contact	
NEOHELIKIAN	BYLOT SUPERGROUP	NUNATSIAQ	Elwin Fm. (470-1220 m):	EL_1: Quartzarenite, siltstone EL_2: Sandstone, siltstone, dolostone
			Gradational	
		NUNATSIAQ	Strathcona Sound Fm: (430-910 m+)	SS_6: SS_{1-5} lithologies interbedded SS_5: Polymictic conglomerate SS_4: Siltstone, greywacke SS_3: Arkose-greywacke, shale SS_2: Dolostone, dolostone conglomerate SS_1: Shale, siltstone
			Gradational	
		NUNATSIAQ	Athole Point Fm: (0-585 m) Limestone, sandstone, shale	
			Gradational	Gradational to Unconformable
		ULUKSAN	Victor Bay Fm: (156-735 m)	VB_2: Limestone, dolostone, flat pebble conglomerate VB_1: Shale, siltstone, sandstone, limestone
			Conformable, Abrupt to Gradational	
		ULUKSAN	Society Cliffs Fm: (263-856 m)	SC_2: Stromatolitic & massive dolostones SC_1: Stromatolitic dolostone, shale, sandstone, gypsum
			Chiefly Unconformable?	Gradational to Unconformable
		EQALULIK	Fabricius Fiord Fm: (400-2000 m+) FF_4: Arkose, conglomerate, dolostone FF_3: Subarkose, conglomerate FF_2: Shale, quartzarenite FF_1: Quartzarenite, shale	Arctic Bay Fm: (180-770 m) AB_4: Shale, dolostone AB_3: Shale, siltstone AB_2: Shale, quartzarenite AB_1: Siltstone, quartzarenite
			Conformable, Abrupt to Gradational	
		EQALULIK	Adams Sound Fm: (0-610 m) AS_3: Quartzarenite, conglomerate, shale AS_2: Quartzarenite AS_1: Quartzarenite, conglomerate	AS_U: Quartzarenite AS_L: Quartzarenite, conglomerate
			Conformable	
		EQALULIK	Nauyat Fm: (0-430 m)	NA_2: Plateau basalt NA_1: Quartzarenite, subarkose, basalt
			Nonconformity	
ARCHEAN-APHEBIAN			Granitic gneiss basement complex: Migmatite, foliated granitic rocks, granite, charnockite, supracrustal relics	

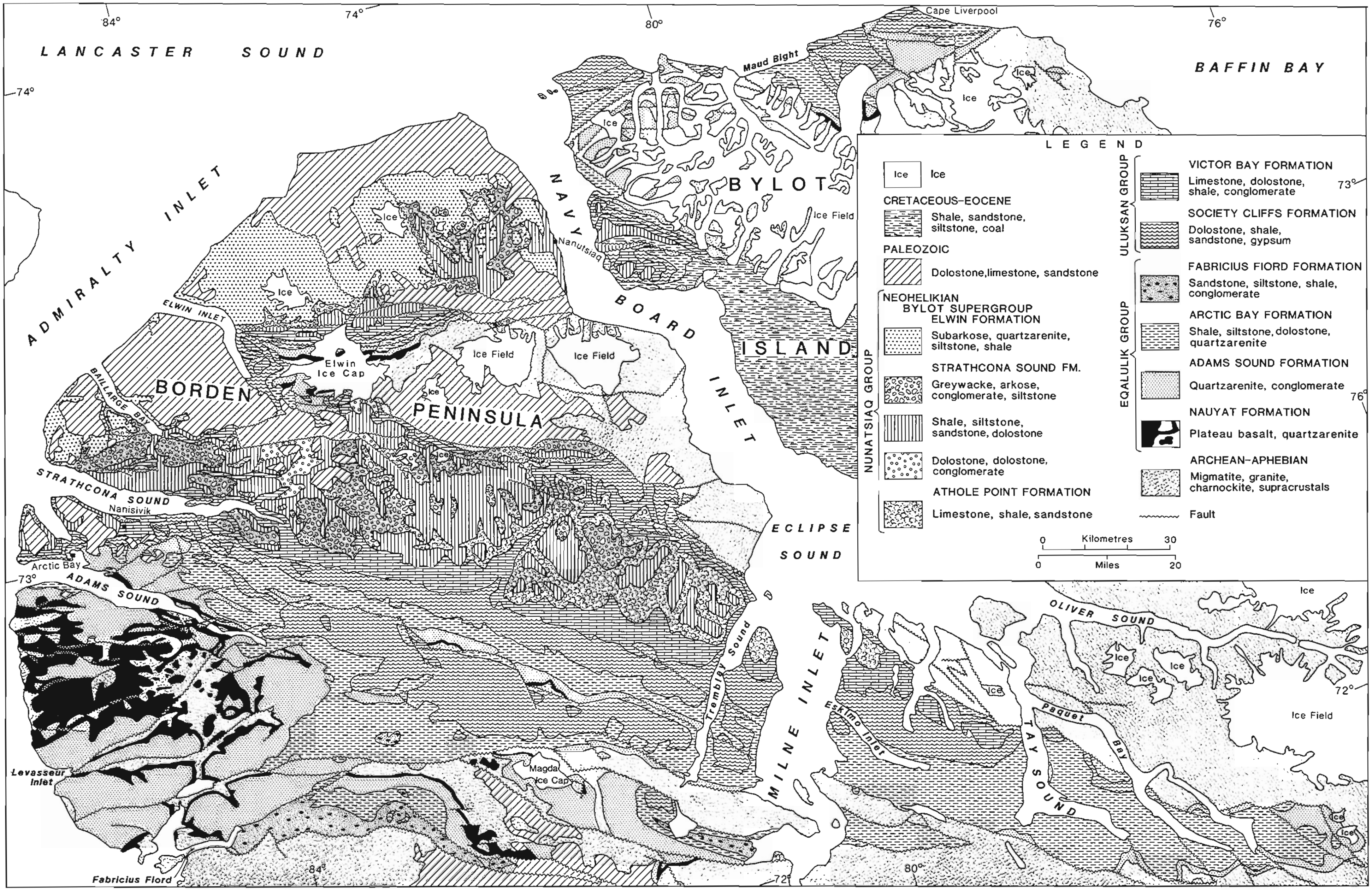
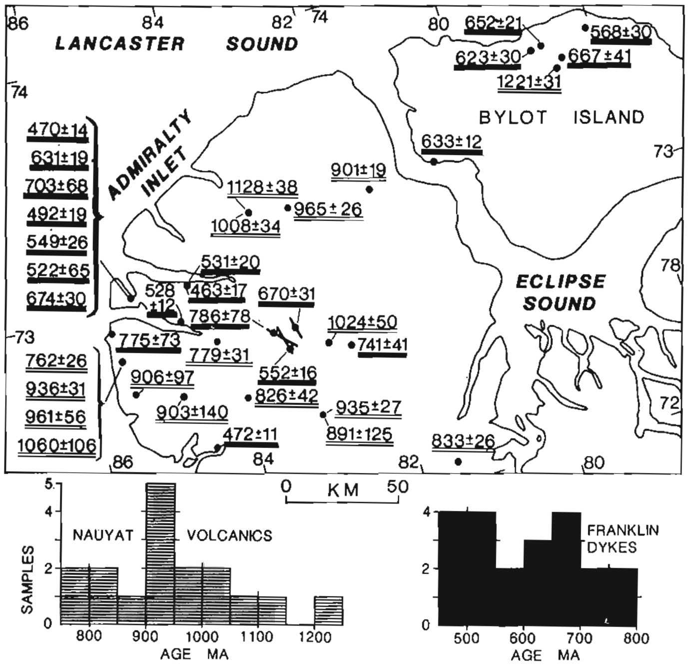


Figure 16.2. *Geology of Borden Basin.*



Double underline = Nauyat volcanics
Solid underline = Franklin diabases

Figure 16.3. *Whole rock K-Ar ages in Borden Basin.*

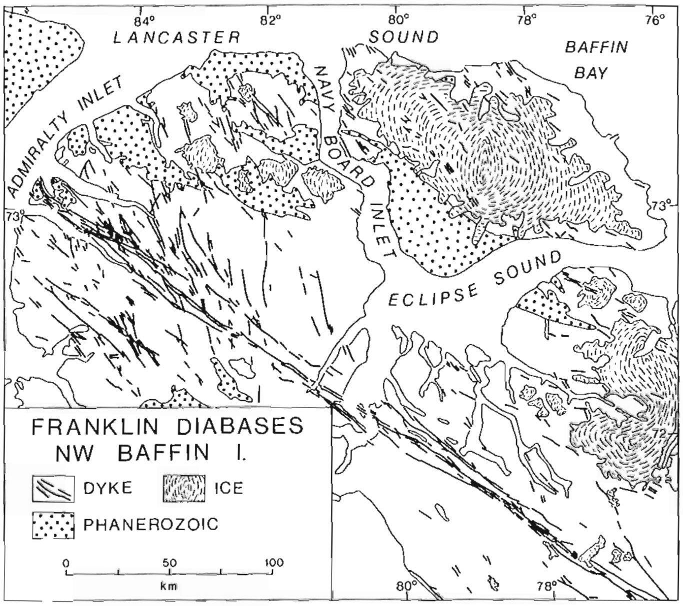


Figure 16.4. *Distribution of Franklin diabase dykes on northern Baffin and Bylot islands. Note main concentration along Tikerakdjuak Fault Zone (see Fig. 16.26).*

generations of granites. Upper amphibolite regional metamorphism predominates in the complex except on Bylot Island where granulite grade predominates. On northwestern Baffin Island granulite facies rocks are most abundant adjacent to the Borden Basin and occur in Navy Board and Byam Martin highs (Jackson and Morgan, 1978).

A thin regolith is locally present at the contact between Bylot Supergroup rocks and the gneissic basement. The gneisses are commonly stained red for several metres below the unconformity. Up to 6 m of regolithic material is preserved south and east of the head of Tremblay Sound, in the Paquet Bay area and on Bylot Island (Fig. 16.2). In these areas basement gneisses pass gradually upward into friable, varicoloured, massive to poorly stratified rocks that consist essentially of recessive masses of kaolinized feldspar and more resistant granular layers and lenses of quartz, all in a finely crystalline chloritic matrix. The contact between the regolith and basal Bylot Supergroup strata is unconformable and planar to slightly undulatory. Where the regolith is absent, the contact between the gneisses and overlying Neohelikian strata is also planar to undulatory, with a maximum observed relief of 2 m. Local draping and pinching-out of sedimentary beds occurs over the small basement topographic highs.

EQALULIK GROUP

In the revised nomenclature presented here, the Fabricius Fiord and Arctic Bay formations have been reassigned to the Eqalulik Group along with the Nauyat and Adams Sound formations (Fig. 16.2).

Nauyat Formation

The Nauyat, lowermost formation of the Eqalulik Group (Table 16.1), outcrops primarily south of Adams Sound and extends to south of Tremblay Sound (Fig. 16.2). Nauyat strata average almost 240 m in thickness in the western part of the area, and may be 430 m thick south of Adams Sound.

The Nauyat Formation is divided into two members: a lower NA_1 member composed chiefly of quartz arenite, and an upper NA_2 member consisting almost entirely of basalt flows. Nauyat quartz arenites are similar to those of the overlying Adams Sound Formation, and are included in the latter formation where NA_2 basalt flows and sills are absent. Nauyat volcanics are thickest along a line trending north-northeast through Elwin Icecap, thin southeastward, and do not occur east of Tremblay Sound (Fig. 16.2, 16.5, 16.6).

NA_1 Member

The NA_1 member consists predominantly of multi-coloured (chiefly buff to pink) quartz arenite interbedded with minor subarkose, lesser amounts of quartz granule-pebble conglomerate and siltstone, and rare shale. The subarkose and conglomerate beds occur mainly in the basal part of the member; a thin conglomeratic unit commonly rests on basement gneisses (Fig. 16.5, 16.6).

The NA_1 member is entirely quartz arenite south of Tremblay Sound and on northern Bylot Island. Elsewhere, a conformable volcanic subunit containing one or two amygdaloidal plateau basalt flows, similar to those in the NA_2 member, occurs sporadically in the middle part of the member. The lower contact is sharp and the underlying sediments are baked, while the upper contact is sharp, but possibly erosional. Locally, rare volcanic fragments occur in the overlying strata.

This member varies in thickness from 0 to more than 160 m in the western part of the area and increases in thickness to the northwest. The intra-member volcanics, from 27 m thick southeast of Elwin Inlet to 60 m north of Levasseur Inlet, are under- and overlain by quartz arenites.

South of Adams Sound, the member is characterized by conglomerate-based fining-upward cycles 2-8 m thick (Fig. 16.7). These cycles consist of a thin basal conglomerate overlain by a thick, graded quartz arenite, and capped by thin

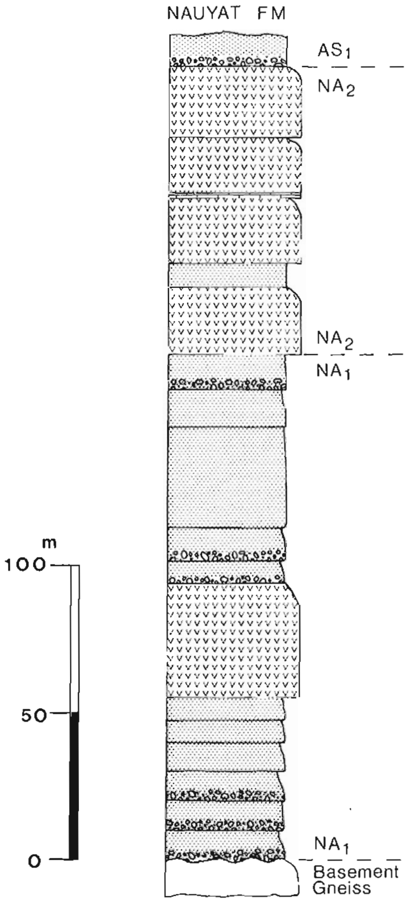


Figure 16.5. Representative stratigraphic section of Nauyat Formation (NA) south of Adams Sound.

siltstone, or rarely, siltstone-shale. These cycles are less well defined upward and die out to the northwest where coarser units are lacking.

Small to medium trough and planar crossbeds are the most common sedimentary structures. Minor structures include scour channels, current ripples, and small load casts. Large-scale current ripples and crossbeds are rare. Crossbeds indicate trimodal paleocurrent trends of low to moderate dispersion to the southwest and west to northwest (Fig. 16.8). The dispersion of the modes is considered a consequence of local pre-depositional topography.

The NA_1 member is sharply and conformably overlain by NA_2 volcanics and the uppermost NA_1 strata have been baked and highly indurated.

NA_2 Member

The NA_2 member consists of a variable sequence of up to 7 basalt flows and minor interflow sediments (Fig. 16.5, 16.6). The NA_2 may be 200 m thick south of Adams Sound, but is commonly 100-130 m thick both there and in the Elwin Icecap region. The member thins southeastward, and is 8 m thick south of Tremblay Sound and 30 m thick on northern Bylot Island. Southeastward, the number of flows decreases to 1 or 2. NA_2 rocks on northern Bylot Island may include massive mafic to ultramafic sills. Where NA_1 strata are absent, the NA_2 volcanics rest directly on basement.

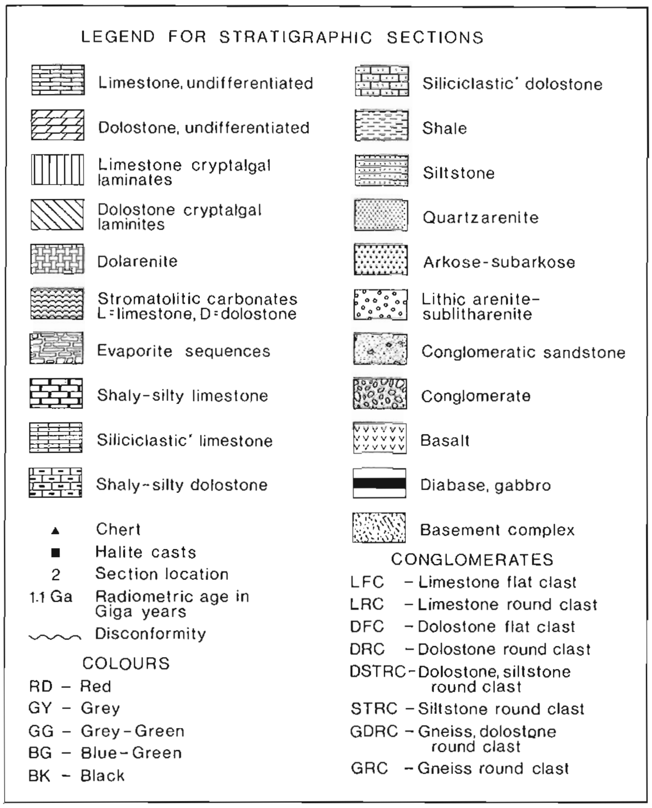


Figure 16.6. Legend for stratigraphic sections and cycle diagrams. "Siliciclastic*" refers to clastic silicate grains, chiefly quartz and feldspar, disseminated in the carbonate.

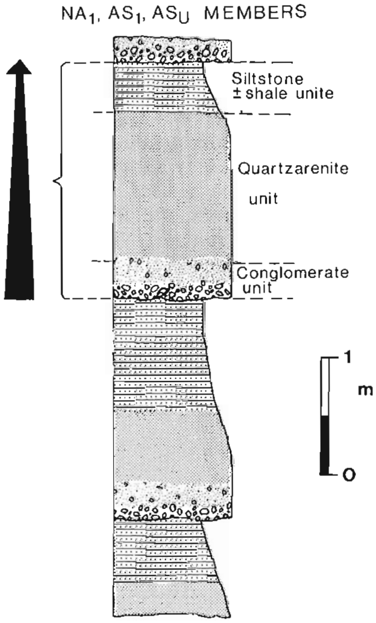


Figure 16.7. Representative conglomerate-based fining-upward cycles in NA_1, AS_1, and AS_U members, southern Borden Basin.

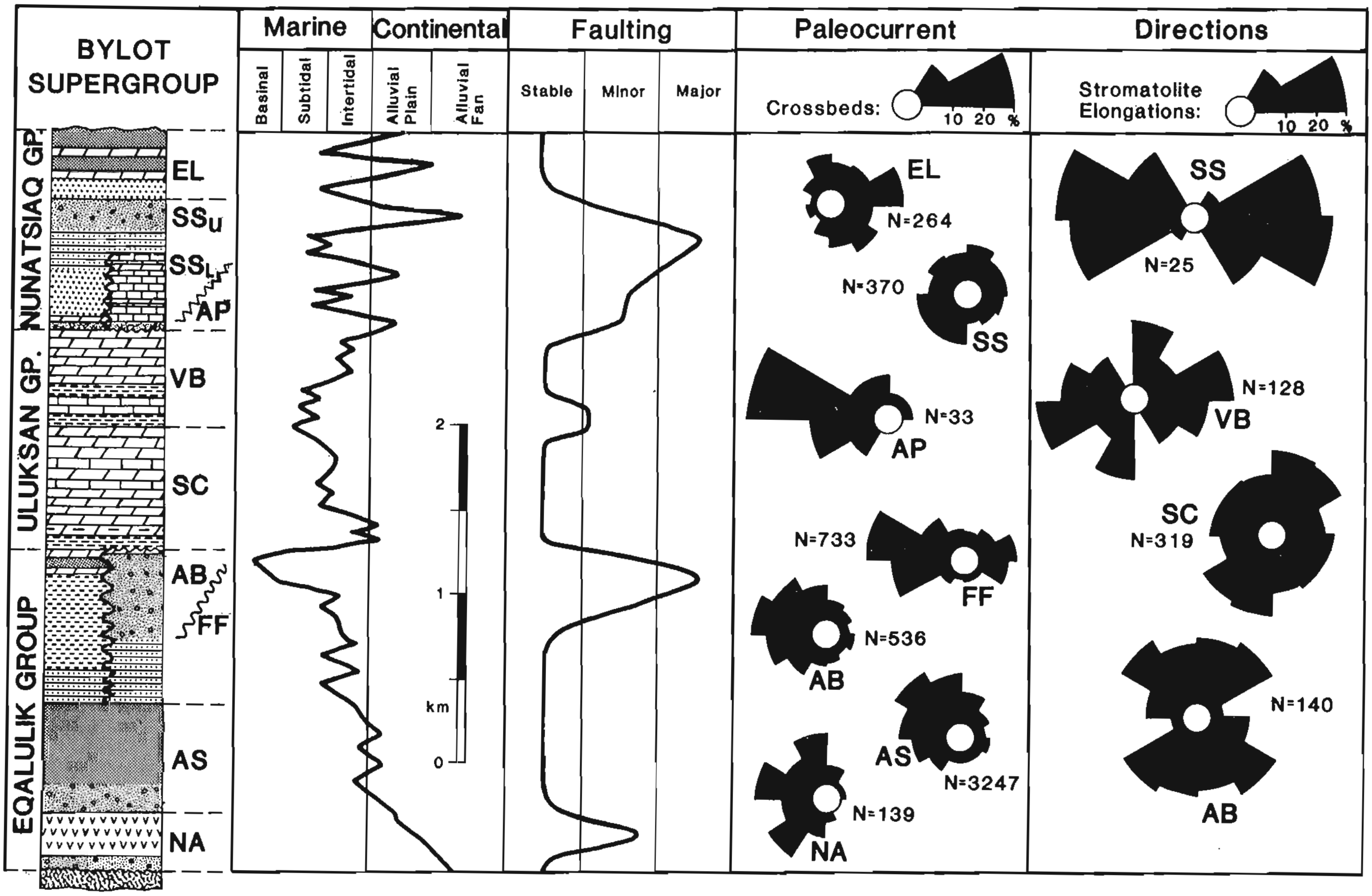


Figure 16.8. *Generalized section of Bylot Supergroup as well as depositional environments, relative intensity of faulting, and cumulative crossbed and stromatolite elongations for each formation. Formation abbreviations as in text and for formation sections.*

The contact with the overlying Adams Sound Formation is sharp and apparently conformable, but local volcanic clasts in the suprajacent quartz arenites suggest minor local erosion.

Individual flows are commonly 2 to 20 m thick but may be as much as 60 m. All flows are fine grained and dark green to green-grey. Most flows are increasingly amygdaloidal upward but the uppermost flow is massive, resistant and amygdule-free. Chemical analyses indicate that both NA_1 and NA_2 flows are tholeiitic basalts which become more alkalic toward the top of the sequence (Galley, 1978; Fig. 16.9, 16.10).

Columnar joints are commonly well-developed. Rare pillows, chiefly 25 cm or less in diameter with indistinct rims, occur south of Adams and Tremblay sounds. Flow-banding occurs locally and volcanic breccia lenses occur north of Levasseur Inlet. Upper parts of some flows contain partially resorbed quartz grains, quartz arenite clasts up to cobble size, and quartz arenite-filled cracks. This suggests that some flow surfaces were probably semi-fluid when the clastics were deposited on them (Galley, 1978).

Thin, baked, interflow sediments up to several metres thick occur locally in NA_2. These include vuggy quartz arenite, thinly laminated stromatolitic(?) limestone, iron-rich carbonate, minor chert and siltstone, and rare volcanic breccia lenses. Small-scale quartz arenite-dominated fining-upward cycles occur within some of these sequences.

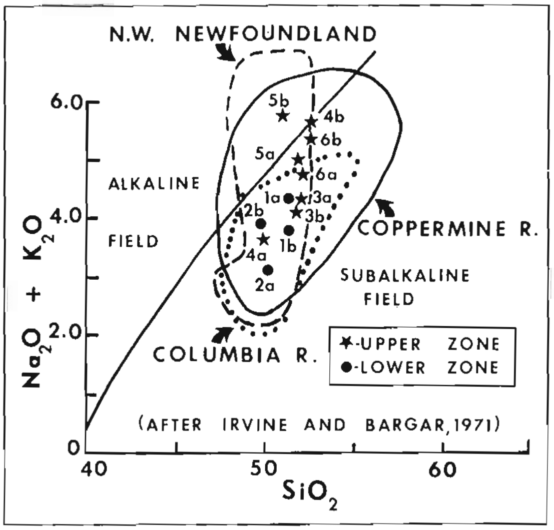


Figure 16.9. *Alkali-silica diagram for Nauyat basalts. From Galley (1978).*

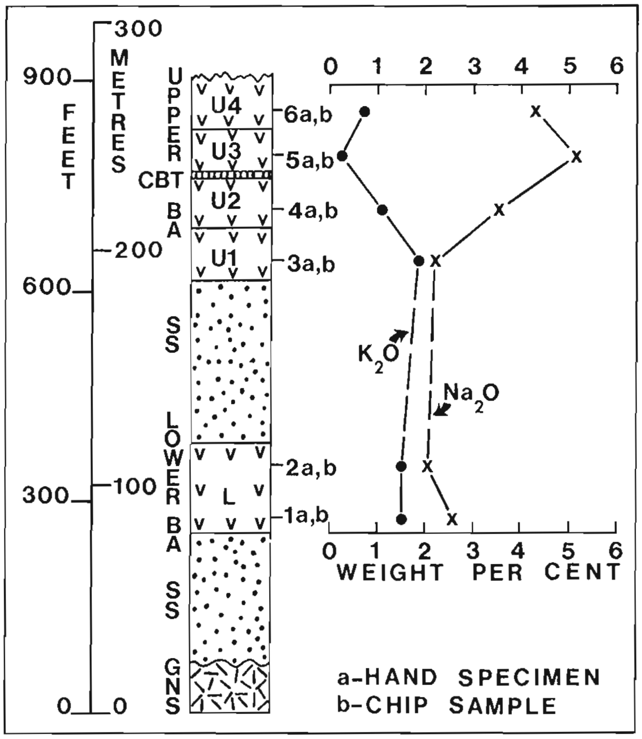


Figure 16.10. Stratigraphic positions of chemically analyzed Nauyat basalt samples and variations in K_2O and Na_2O. From Galley (1978).

The volcanic pile is spatially related to the present Central Borden, Tikerakdjuak, White Bay, Hartz Mountain and Cape Hay fault zones (Fig. 16.2, 16.26).

Interpretation

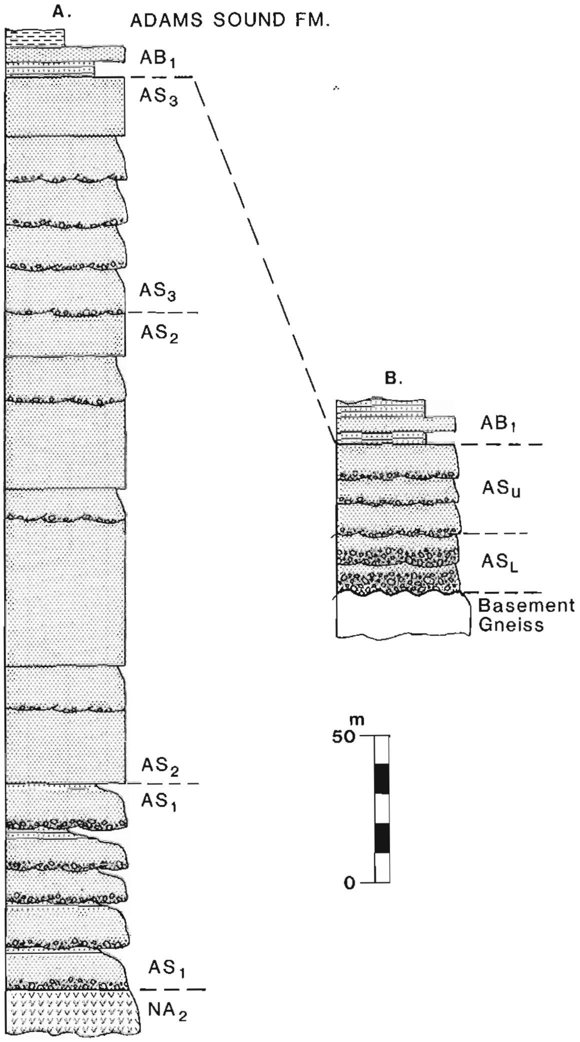
Unimodal paleocurrent trends (crossbeds are unimodal locally but cumulative rose is trimodal – Fig. 16.8) and conglomerate-based to quartz arenite-dominated fining-upward cycles indicate a fluvial depositional environment for most of the Nauyat sediments.

The composition (Fig. 16.9, 16.10) and physical properties of the Nauyat volcanics indicate that they are predominantly subaerial tholeiitic plateau basalts that were extruded quietly along fissures or linearly-controlled eruptive centres (Galley, 1978). They were covered by sediments before significant erosion could occur.

Adams Sound Formation

The Adams Sound Formation is the middle formation of the Eqalulik Group, and outcrops extensively in southern Borden Peninsula (Fig. 16.2). The Adams Sound conformably overlies NA_2 volcanics wherever the latter are present, although a thin (8 cm) layer of volcanic arenite locally overlies the volcanics (Geldsetzer, 1973b). Elsewhere the Adams Sound rests upon basement gneisses. The formation ranges from 610 m thick on western Borden Peninsula, to 45 to 65 m east of Tremblay Sound, and at least 415 m on northwestern Bylot Island.

The Adams Sound Formation consists predominantly of thin- to thick-bedded, rarely massive, quartz arenites, with minor interbedded subarkose, quartz-pebble conglomerate and rare siltstone and shale (Fig. 16.11). Sedimentary structures are common and include current ripples, trough and planar crossbeds (Fig. 16.8), channels, scours, load casts, soft sediment deformation structures, clastic dykes and microfaults.



A – Adams Sound area
B – West of Paquet Bay

Figure 16.11. Representative stratigraphic sections of Adams Sound Formation (AS).

Two broad regions of Adams Sound strata have been delineated on the basis of subtle changes in lithology, the nature of depositional cycles, and contained sedimentary structures. These regions are: Borden Peninsula, and the Eastern Region (southeast of Milne Inlet, and Bylot Island, Fig. 16.2).

Borden Peninsula

The Adams Sound in this region has been divided into three intergradational members.

AS_1 Member: Most of this member is composed of buff to red, planar to crossbedded quartz arenite that contains minor interbedded and interlensed granular to pebbly quartz

arenite, oligomictic to polymictic pebble to cobble orthoconglomerate, subarkose, siltstone and shale. Conglomeratic rocks occur mostly in the basal part of the member; clasts are predominantly quartz, but locally include quartz arenite, chert and gneiss. Some sequences contain well-developed conglomerate- to pebbly quartz arenite-based, quartz arenite dominated, fining-upward cycles (1 to 3 m) rarely capped by thin units of siltstone or siltstone-shale (Fig. 16.7). AS_1 thicknesses range from 130 m at Elwin Inlet to 8-16 m near Tremblay Sound.

Sedimentary structures, especially planar and trough crossbeds are ubiquitous. Paleocurrents are primarily unimodal northwest to northeast, with subsidiary southwest to southeast trends.

AS_2 Member: The entire member is dominated by buff to light pink quartz arenite. Poorly defined 3-5 m thinning- and fining-upward cycles occur locally. These consist of lower units of medium- to coarse-grained, medium-bedded quartz arenite, with large trough crossbeds and scours, grading upward into fine grained, thin bedded to laminated, quartz arenite. This member has a maximum thickness of 128 m at Adams Sound but thins to 33 m at Tremblay Sound.

The relatively few sedimentary structures present include crossbeds, current ripple marks, load casts, and synaeresis or desiccation cracks. Paleocurrent patterns show high dispersion, unimodal northwest to northeast transport with subsidiary east and southeast trends.

AS_3 Member: This member is chiefly white, light to dark grey and pink quartz arenite with interbeds and lenses of granular to pebbly quartz arenite, and quartz-pebble conglomerate. Thin shale and siltstone interbeds occur in the uppermost part of the member. Rare channels filled by quartz arenite-clast breccia occur in the Tremblay Sound area, where some sections also contain 4-8 m thick fining- and thinning-upward cycles. Fining-upward cycles in the Adams Sound area are 1.5-4.5 m thick and have large basal scour channels filled with quartz granule to quartz-pebble conglomerate. AS_3 strata are commonly 68-100+ m and are somewhat thicker locally at Tremblay Sound.

Megaripple marks with wavelengths up to 2 m, wave ripple marks and graded and overturned crossbeds also occur. Paleocurrent trends are polymodal, northwest-southeast bimodal, and northwest and northeast to southeast unimodal.

Eastern Region

Adams Sound strata on Bylot Island and southeast of Milne Inlet are similar to those on Borden Peninsula, but are here informally divided into only a lower (AS_L) and upper (AS_U) member that are approximately equivalent to the Borden Region strata. Measured thicknesses are in excess of 374 m on Bylot Island where one partial section was estimated to be 415 m (Jackson and Davidson, 1975). The Adams Sound Formation southeast of Milne Inlet has a much higher proportion of conglomerate than elsewhere, but is much thinner (45-65 m).

AS_L Member: Buff to pink, orange and purple strata dominate this member. On Bylot Island quartz arenite predominates and contains some interbeds of granular to pebbly quartz arenite and quartz-pebble conglomerate. The upper part of this member contains a distinctive dark purplish-red hematite-stained quartz arenite sequence with abundant siltstone and shale partings.

Southeast of Milne Inlet most of the AS_L is thick-bedded to massive, oligomictic to polymictic, quartz-pebble to quartz-cobble orthoconglomerate in which quartz arenite, feldspar, and gneiss clasts are locally abundant. The gneiss clasts decrease in abundance upward in the member. Subarkose, pebbly subarkose and quartz arenite are interbedded with the conglomerate.

Fining-upward cycles occur locally throughout the member. On Bylot Island the beds in the cycles also thin upward, and cycles are 2-5 m thick. These cycles consist of a lower unit of medium-bedded, medium-grained to granular quartz arenite with large trough crossbeds, channels and load casts. This unit grades upward into medium-bedded to thick-laminated fine grained quartz arenite with smaller trough crossbeds, current ripple marks and synaeresis cracks. Fining-upward cycles southeast of Milne Inlet are conglomerate based and conglomerate dominated (Fig. 16.7). The member is typically 110-196 m thick on Bylot Island and 1-15 m southeast of Milne Inlet.

Few sedimentary structures are present southeast of Milne Inlet, but include large scattered planar crossbeds up to 2.5 m thick that contain current aligned quartz pebbles along foreset and bottomset beds. Paleocurrent patterns show north-northwest, north, and southeast unimodal transport. Paleocurrent patterns on Bylot Island vary from northwest-southeast bimodal-bipolar to unimodal northwest to southwest.

AS_U Member: White to buff and brown-grey quartz arenite predominates in this member. Minor interbeds of granular to pebbly quartz arenite and quartz-pebble conglomerate thin and decrease in abundance upward. Siltstone partings are rare. Some 1-4 m fining-upward cycles are present, and are similar to those present in AS_L member from the respective areas (Fig. 16.7). These cycles also thin-upward on Bylot Island. The member ranges from 140-200 m thick on Bylot Island to 18-35 m southeast of Milne Inlet.

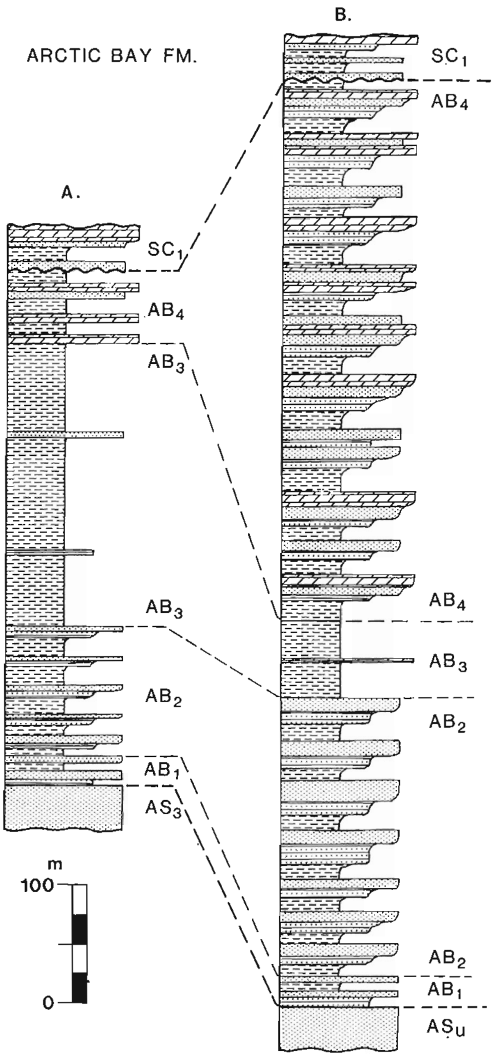
Large-scale planar crossbeds with graded foresets megaripples, and scour channels are common locally southeast of Milne Inlet. The crossbeds show unimodal northwest to north transport, whereas transport in the Bylot Island succession was southwest to northwest unimodal and bimodal.

Interpretation: The conglomerate-dominated, conglomerate-based and quartz arenite-dominated fining-upward cycles, predominance of strongly unimodal paleocurrents (Fig. 16.8), and the abundance of festoon crossbeds and other structures, indicate that AS_1 member of the Borden Peninsula region and the AS_L and AS_U members southeast of Milne Inlet were deposited under proximal to distal braided fluvial environments.

The AS_2 and AS_3 members on Borden Peninsula and AS_L and AS_U on Bylot Island were deposited in mixed fluvial-marine environments, as suggested by thinning-upward clastic shoreline cycles, large tidal or fluvial bar-like structures, and strong unimodal to bimodal-bipolar paleocurrent trends.

Arctic Bay Formation

The Arctic Bay Formation outcrops in a broad belt across most of the southern basin (Fig. 16.2). Locally, pyritiferous shale is the predominant lithology, with siltstone and quartz arenite interbedded with shale in the lower part of the formation, and siltstone, dolostone and quartz arenite interbedded with shale in the upper part.



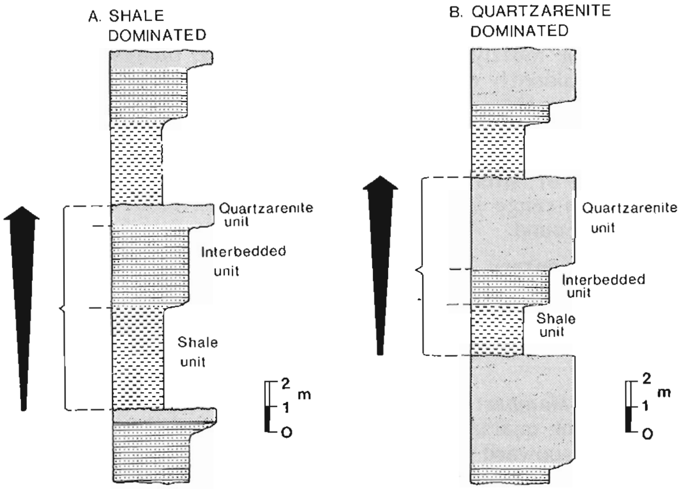
A – Upper Tremblay Sound area
B – Tay Sound-Paquet Bay area

Figure 16.12. *Representative stratigraphic sections of Arctic Bay Formation (AB).*

The formation is 180 m thick at Arctic Bay, and ranges from 500 to 770 m throughout most of the rest of the area. The lower contact with the Adams Sound Formation is conformable and gradational.

Sedimentary structures are most common in the siltstone, quartz arenite, and stromatolitic dolostone beds. They include wave and current ripples, planar and trough crossbeds, synaeresis and desiccation cracks, molar tooth and tepee structures, load casts, rip-ups, scours, rill marks, convolute or soft sediment-folded beds, and dewatering structures. Some dolostones contain small vugs lined with calcite, dolomite, quartz, and rarely celestite, siderite, and black bituminous material. Some shale beds contain concretions and cone-in-cone structures. White gypsum efflorescence and calcareous coatings are common on the shales and some strata emit a strong petroliferous odour.

Four intergradational members are distinguished in the formation throughout the area (Table 16.1, Fig. 16.12). However, the entire formation is distinctly different southeast of Milne Inlet.



A – West of Milne Inlet
B – East of Milne Inlet

Figure 16.13. *Generalized coarsening-upward cycles of the AB_2 member.*

AB_1 Member

This member comprises grey-green to white interbedded siltstone and quartz arenite and minor thin interlayers and partings of black shale that increase upward. Buff-white to grey quartz arenite with minor thin grey siltstone and shale interlayers and partings predominate east of Milne Inlet. Thicknesses range from 12-48 m in the Milne Inlet Trough and are similar elsewhere.

Characteristic sedimentary structures include herringbone crossbeds, megaripples, small clastic dykes, and lensoid bedding. Paleocurrents everywhere are bimodal-bipolar northwest-southeast, with polymodal transport trends also present in the Borden-Bylot area.

AB_2 Member

The entire AB_2 member is composed of 12 to 15 coarsening-upward shale-dominated cycles that consist of three intergradational units (Fig. 16.13). Individual cycles are 4.5-21 m thick in the Borden-Bylot region. However, there are only 4 to 6 cycles in the east Milne Inlet area, and these are quartz arenite-dominated, and 6.5-17 m thick. The member is typically 48-115 m thick, but is 150-260 m around Tremblay Sound.

The lowermost unit of the cycles, is chiefly black shale with minor lensoid to thin-bedded silty shale and quartz arenite. It is typically 2-5 m thick east of Milne Inlet and 3-9 m thick elsewhere. The middle unit of the cycles consists of lenticular- to planar-interbedded grey-white siltstone-quartz arenite and shale. The shale component decreases upward from more than 85 per cent at the base to less than 50 per cent at the top. The middle unit is 1.5-4.5 m thick east of Milne Inlet and 1.5-6 m thick elsewhere.

The upper unit of the cycles is composed of grey-brown to grey and white thin- to thick-bedded quartz arenite. The unit contains some interlayers and partings of grey-black siltstone and shale and decreases upward in thickness throughout the member. Thickness ranges from 3-7.5 m east of Milne Inlet, and 0.3-1.5 west of the inlet.

Characteristic sedimentary structures include herringbone-crossbeds, megaripples and channels. Crossbeds indicate unimodal southwest to northwest transport throughout the area. Bimodal-bipolar southeast-northwest trends occur east of Milne Inlet, and polymodal trends locally occur elsewhere.

AB_3 Member

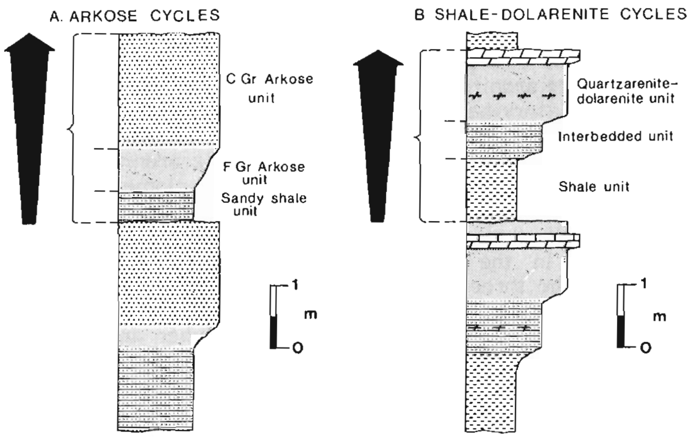
Planar-laminated micaceous black shale predominates in the AB_3 member. In addition, it contains minor (less than 10%) thin grey-black silty shale, siltstone, dololutite and dolosiltite interlayers throughout the Borden-Bylot area. However, these constitute up to 25 per cent of the member east of Milne Inlet, where a 1 m concretionary limestone bed in the middle of the member is a distinct marker. There are few sedimentary structures and the member is recessive. This member is up to 400 m thick east of Elwin Inlet and west of Tremblay Sound, but only 64-92 m east of Milne Inlet.

AB_4 Member

Grey-black shale is the chief lithology of this member, except southeast of Milne Inlet where it is merely a prominent lithology and the member makes up most of the Arctic Bay Formation. AB_4 sequences and lithologies are more variable than in other Arctic Bay members, and a cyclical lithological sequence occurs at most localities. All cycle types contain the same basic three-fold intergradational subdivision as in AB_2 member cycles. Member thicknesses range from 370-615 m southeast of Milne Inlet to 17-212 m elsewhere.

Small- to large-scale shallowing-upward cycles occur in the AB_4 member throughout most of the Borden Basin. At the head of Adams Sound the cycles are 8-40 m thick and consist of a lower black shale unit, a thin middle interbedded shale-calcisiltite unit with minor siltstone and quartz arenite, and an upper unit of thinly interbedded calcisiltite-limestone layers.

East of Milne Inlet shallowing-upward shale-dolarenite cycles constitute most of the AB_4 member and as many as fifty cycles occur in a single section (Fig. 16.14). Individual cycles are 5-25 m thick, and contain a lower unit (1.5-8 m) of shale with minor siltstone. Interlayered siltstone, quartz arenite, and dolarenite occur with minor shale and subarkose in the middle unit (0.6-6 m). The upper unit (3-15 m) consists chiefly of calcareous siltstone, quartz arenite, dolarenite, stromatolitic dolostone, and flat-pebble conglomerate.



A – arkose cycles
B – shale-dolarenite cycles
(Paquet Bay area)

Figure 16.14. *Coarsening-upward cycles in the AB_4 member east of Milne Inlet.*

Coarsening-upward cycles occur in the AB_4 member adjacent to the White Bay Fault Zone east of Milne Inlet and southeast of Elwin Inlet (Fig. 16.2). Partial sequences east of Milne Inlet contain between five and ten 1-5 m thick arkose cycles adjacent to the fault and these pass southward (basinward) into shallowing- and coarsening-upward, clastic-carbonate cycles (Fig. 16.14). Typically, the arkose cycles consist of a thin lower sandy-shale unit that grades upward through silty fine grained arkose into thick coarse grained conglomeratic arkose.

Southeast of Elwin Inlet the AB_4 member consists of about 25, 2.5-22 m thick, coarsening upward, shale-siltstone cycles. Cycles in the middle part of the sequence contain minor pisolitic, concretionary to brecciated carbonate beds. Those in the upper part contain interbedded siltstone-dolosiltite to stromatolitic dolostone.

Adjacent to Tremblay Sound and on northern Bylot Island the AB_4 member is mostly shale with thin interbeds of siltstone, dolosiltite, stromatolitic dolostone and limestone. The carbonates occur in light-grey, orange-brown weathering, planar-based lenses and undulatory beds up to 3.5 m thick which constitute 20-30 per cent of the member.

Locally, the upper 10 m of the AB_4 member is entirely stromatolitic dolostone. The carbonates occur both in situ and as local slump or debris flow blocks in the uppermost part of the AB_3 member. Some stromatolitic dolostone contains biohermal mounds in which basal tabular stromatolites pass upwards through low domal types into expanding, branching columnar (**Baicaila**) types. The columns are 1-2 cm high and upright to inclined or overturned. The mounds contain local dolostone breccia lenses, and are elongate approximately northwest-southeast to northeast-southwest (Fig. 16.8).

Interpretation

The herringbone crossbeds, bimodal-bipolar to polymodal paleocurrent patterns and the gross stratigraphy suggest that the AB_1 member was deposited in an environment that ranged from intertidal east of Milne Inlet to mixed intertidal-shallow subtidal in the rest of the basin (Fig. 16.8).

The types of coarsening-upward cycles, the associated structures and paleocurrent data indicate that AB_2 depositional environments ranged from marine-influenced deltaic and clastic shoreline regimes in the southeastern part of the basin to clastic shoreline and muddy shelf regimes in the north. The monotonous black shales of the AB_3 member accumulated in shallow to deep subtidal to basinal environments.

The AB_4 member west of Milne Inlet was probably deposited in mixed clastic-shoreline and shallow carbonate shelf environments. This is suggested by the fining- and shallowing-up cycles, biohermal mounds and their associated stromatolites and sedimentary structures. Similar environments are indicated by the presence of shale-dolarenite cycles for most of the area east of Milne Inlet. The fault-fringing arkose cycles in this area probably formed within marine-influenced delta to alluvial fan complexes, that interfingered basinward with the cyclic shale-dolarenite sediments.

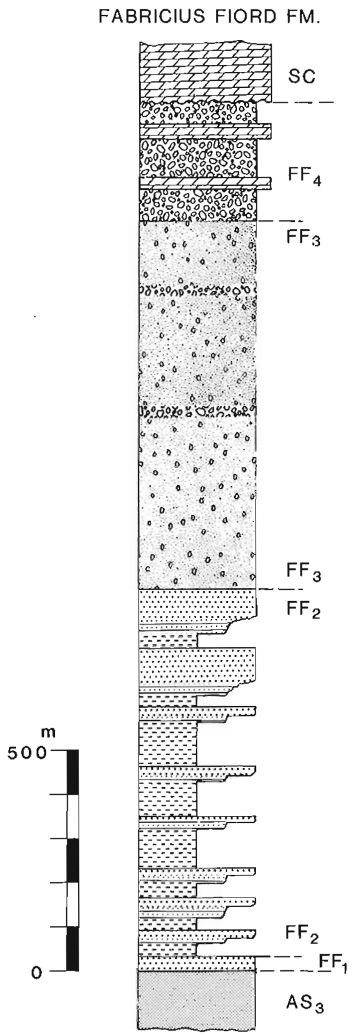


Figure 16.15. *Generalized stratigraphic section of Fabricius Fiord Formation (FF) east of Fabricius Fiord.*

Fabricius Fiord Formation

The Fabricius Fiord Formation outcrops from Fabricius Fiord eastward to south of the Magda Icecap (Fig. 16.2). A few small isolated areas of Fabricius Fiord-like strata, which will not be discussed further, occur adjacent to major fault zones at several localities throughout the basin (Fig. 16.2).

In general, the Fabricius Fiord Formation contains a lower sequence composed chiefly of black shale, grey to grey-white siltstone, and brown-grey to white quartz arenite in which the alternation of recessive and resistant beds gives a distinctive ribbed appearance to the outcrop. The remainder of the formation is predominantly subarkose and conglomerate, and several minor lithologies (Fig. 16.15). Locally the strata outcrop in large fan-shaped structures. Sedimentary structures are most abundant in the lower quartz arenites and siltstones and include: current and wave ripples, trough, planar and herringbone-crossbeds, channels, small-scale load casts, convolute beds, clastic dykes, rills and synaeresis cracks. The crossbeds yield well-defined unimodal to bimodal-bipolar paleocurrent patterns that indicate major distributing currents flowed west and northwest (Fig. 16.8).

The Fabricius Fiord Formation ranges from 400 to more than 2000 m thick. Four intergradational members have been defined within the formation. In addition, three interlensing lithologic associations have been differentiated in the uppermost FF_4 member.

The contact with the underlying Adams Sound Formation is conformable and sharp to gradational. Fabricius Fiord strata are invariably downfaulted along, and/or marginal to major fault zones as along the Central Borden Fault Zone. The lower FF_1 and FF_2 members grade laterally northward (basinward) into facies-equivalent Arctic Bay strata. The FF_3 and FF_4 members are overlain disconformably by massive dolostone that closely resembles the Society Cliffs Formation in both lithology and contained stromatolites. Lenses (possibly biohermal) and thin beds of similar dolostone also occur within the FF_4 member. Therefore, although contact relations between upper Fabricius Fiord strata and lower Society Cliffs (SC_1) strata have not been seen, the two are considered tentatively to be facies equivalents.

FF_1 Member

In the Fabricius Fiord area the lower part of this member is mostly white to rust-brown, hematite-stained quartz arenite which grades upward into shale, and interbedded siltstone and quartz arenite. Eastward, in south-central Borden Peninsula, the FF_1 member contains a lower sequence consisting largely of grey-green quartz arenite and subarkose interbedded with lesser amounts of pebbly subarkose, quartz- and feldspar-pebble conglomerate, siltstone and shale. Upper beds in this area are massive quartz arenite containing graded planar- and trough-crossbeds and load casts. The FF_1 member is an average 18 m thick in the west and ranges from 15-23 m thick in the east.

FF_2 Member

The thick, distinctive, FF_2 member consists entirely of coarsening-upward cycles that range from 9 to 40 +m and up to 55 in number. The member ranges from 370 to 892 m in thickness.

Each coarsening-upward cycle contains three distinct intergradational units (Fig. 16.16). The lower unit, 3-15 m thick, contains shale, silty shale, and minor siltstone, and grades upward into the middle unit (1.5-6 m thick), which consists of interbedded quartz arenite, siltstone and shale. The quartz arenites and siltstones grade upward from lensoid and undulatory-bedded into planar-bedded as their proportion increases from less than 10 to more than 60 per cent of the unit. The upper unit (0.5-15 m) is quartz arenite with thin siltstone or shale partings, and with conglomerate in the upper part of the member. These quartz arenites contain nearly all the herringbone-crossbeds, ripple marks and channels in the formation. Lower FF_2 member cycles contain the three units in equal proportions. Cycles in the middle of the member are dominated by shale (70 per cent), and those in the upper part of the member are sandstone dominated.

FF_3 Member

The massive, resistant FF_3 member is mostly medium bedded to massive light grey to buff subarkose and pebbly subarkose interbedded with thin- to thick-bedded quartz- and quartz-feldspar-pebble conglomerate. Quartz arenite is abundant locally. The FF_3 member is 736 m thick at Fabricius Fiord and 840 m thick in south-central Borden Peninsula.

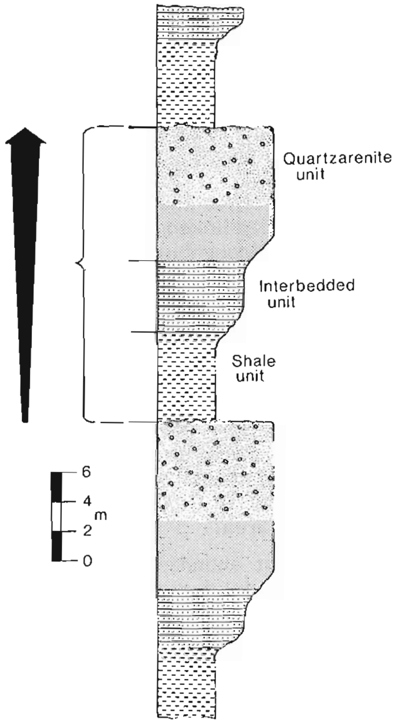


Figure 16.16. *Representative coarsening-upward cycles in FF_2 member east of Fabricius Fiord (same location as for Fig. 16.15).*

Poorly defined 4-30 m thick coarsening-upward cycles are best developed southwest of Magda Icecap. These consist of a lower unit of medium- to thick-bedded, coarse grained to pebbly subarkose that grades upward into massive pebbly subarkose and conglomerate. Sedimentary structures are rare, but a few large planar crossbeds occur locally. They have a maximum thickness of 1.5 m and current-aligned quartz pebbles occur along foreset and bottomset beds.

FF_4 Member

Thick alternating beds and lenses of grey to yellow-brown and red-orange subarkose and arkose, light to medium-grey gritty to stromatolitic dolostone, and pink to brown-grey breccia-conglomerate constitute the FF_4 member. This member occurs adjacent to the Central Borden Fault Zone, and individual beds locally increase in thickness towards the fault zone. The member is 150-245 m thick and has been divided into three laterally and vertically intergradational lithological associations. The FF_4 member comprises more than 65 per cent of the FF_{4A} association, the rest consisting of scattered lenticular zones of FF_{4B} association and unstratified FF_{4C} wedges adjacent to the Central Borden Fault Zone. FF_{4B} and FF_{4C} associations occur at various stratigraphic positions within FF_{4A}. Both the FF_{4A} and FF_{4B} associations weather a diagnostic chocolate brown colour due to breakdown of contained iron-rich carbonate.

FF_{4A} Association. This association is mostly chocolate-brown subarkose and pebbly subarkose with minor interbeds of arkose and pebbly arkose. The strata commonly have a calcareous matrix and the arkoses typically occur as thin wedges which increase in thickness toward the fault zone. Sedimentary structures are rare and consist of a few scattered medium- to large-scale planar crossbeds, poorly-developed scours, and shallow channels.

FF_{4B} Association. Carbonate-rich strata predominate in the FF_{4B} and consist of stromatolitic dolostone and relatively massive gritty dolostone. The latter forms one of the two lithologic end members, the other being calcareous subarkose. The grit in the dolostone is chiefly subrounded to subangular sand to pebble-sized clasts of quartz and feldspar which make up 5 to more than 25 per cent of the rock. The stromatolitic dolostone is commonly interlensed with flat pebble conglomerate and calcareous, pebbly subarkose.

Stromatolites of the FF_{4B} association occur in isolated lenses, in small biohermal mounds in the arkosic rocks, and as clasts in the flat pebble and boulder conglomerate. They include planar, or low domal forms 5-10 cm high, and unbranching expanding-upward columnar forms up to 20 cm high. The stromatolites commonly occur in a gritty or arkosic matrix and some are overturned and partially eroded.

FF_{4C} Association. The entire association comprises structureless massive breccia-conglomerate wedges up to 5 m thick adjacent to the Central Borden Fault Zone, but nearly all wedge out within 0.5 km north of the fault zone. Granule- to cobble-size quartz and feldspar clasts and pebble to boulder granite and gneiss clasts predominate. In addition, cobble- to boulder-sized clasts are common adjacent to the fault zone but clast size decreases northward until pebble-sized clasts predominate 0.5 km away. The clasts are supported in a dolostone matrix that constitutes 10-40 per cent of the rock and contains scattered biotite flakes and quartz and feldspar sand grains.

Interpretation

The large scale coarsening-upward trend within the formation, cyclic deposition, and contained sedimentary structures, suggest the Fabricius Fiord strata were deposited in large marine-influenced delta fan complexes. These complexes extend more than 10 km basinward (north), are 1-2 km thick, and dominated the southern margin of the Borden Basin and Milne Inlet Trough throughout Fabricius Fiord deposition.

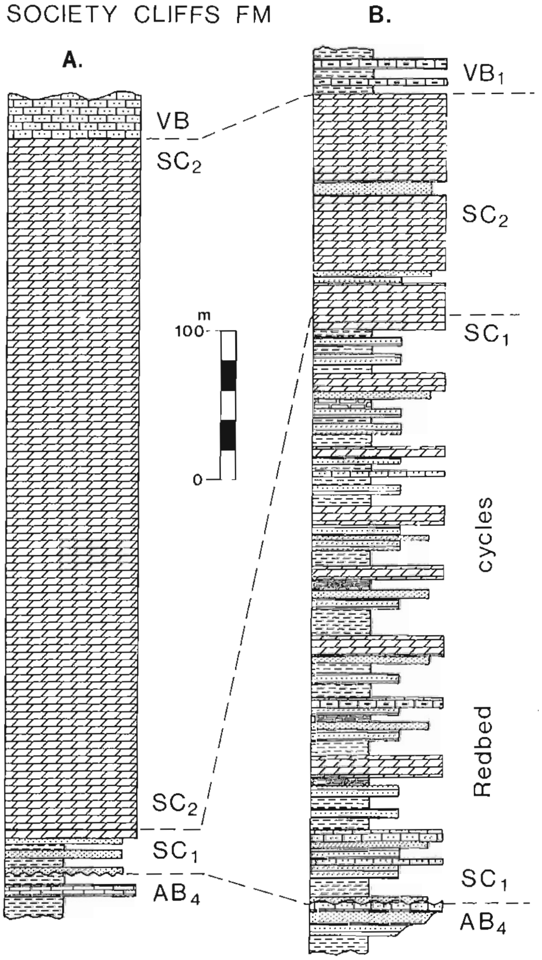
FF_1 beds are thin coastal to shallow marine shelf blanket sandstones and siltstones that were buried by prograding FF_2 coarsening-upward cycles of the lower and mid-fan complexes. These cycles in turn grade upward and toward the fault zone into thick FF_3 and FF_{4A} sheet sandstones and conglomerates of the upper fan platform. The FF_{4B} rocks originated within interdistributary basins on the emergent platform while FF_{4C} breccia-conglomerates were sporadically deposited as wedges along the fault zones during periodic tectonic activity. With cessation of major faulting along the basin margins, and subsequent expansion of the Society Cliffs carbonate platform Fabricius Fiord deposition ended.

ULUKSAN GROUP

The term Uluksan Group is restricted here to strata of the Society Cliffs and Victor Bay formations. These distinctive strata, together with the Nauyat, are the strata within the Bylot Supergroup that can probably best be correlated with strata in other regions.

Society Cliffs Formation

The Society Cliffs Formation outcrops chiefly in a broad lenticular belt that extends from Arctic Bay east-southeast to Paquet Bay (Fig. 16.2).



A – Elwin Ice Cap area
B – west side of Tay Sound

Figure 16.17. *Representative stratigraphic sections of the Society Cliffs Formation (SC).*

Most of the formation consists of several types of laterally and vertically interfingering grey to brownish grey, light buff weathering dolostones that commonly emit a petroliferous odour and contain fine laminae and/or sparsely disseminated small blebs of bituminous material (Fig. 16.17).

Interbedded with the dolostones, mostly in the lower part of the formation, are lesser amounts of quartz arenite, arkose, varicoloured shale, and gypsum. Limestone lenses and beds and black to grey thinly bedded to nodular chert occur scattered through the formation. Varicoloured chert, primarily replacing stromatolitic dolostone, is locally abundant in the east, closely associated with redbeds and sabkha sequences.

The formation thickens from 260 m at Arctic Bay to 856 m at the mouth of Tremblay Sound, and thins to the east and south. Thicknesses range from 370 m to 610+ m in the northern part of the basin.

The basal contact with the underlying Arctic Bay Formation ranges from conformable and sharp as at Arctic Bay, to unconformable as at the head of Adams Sound or Elwin Inlet where a dolostone-clast or limestone-clast conglomerate, respectively, occurs at the base. At Milne Inlet and Tay Sound the basal beds are commonly quartz arenite, subarkose, or sublitharenite with clasts of Arctic Bay shale. Polymictic conglomerates in the lower part of the formation, and fault juxtaposition of Society Cliffs and basement gneisses, suggest that the Society Cliffs Formation overlapped the gneisses. The contact with the overlying Victor Bay Formation is conformable and sharp to gradational, and the shale content in the Society Cliffs locally increases upward in the uppermost few metres.

Abundant stromatolites include planar laminites, and individual to laterally-linked undulose, domal, and hemispheroidal types throughout most of the formation. Most individuals are less than 25 cm in diameter. Bun-shaped and bulbous types are common, and some small domes are attached to the periphery of larger domes up to 3 m in diameter. Individual to laterally-linked columnar types are 2-75 cm high. They, together with tubular types about 5 cm in diameter, types resembling **Baicalia**, laterally linked box or rectangular types, and low domes with radially attached small columnar types occur in the vicinity and east of Tremblay Sound. Conical stromatolites (**Conophyton**) 5-10 cm in diameter and 10-30 cm high were noted only in the western Elwin Inlet area.

Bioherms are most abundant in the upper part of the formation in the Milne Inlet region. The larger ones are commonly 30-60 m in diameter. One bioherm southeast of Nanisivik is 1500 m in diameter.

Sedimentary structures are abundant and include tepee and other dewatering structures, molar tooth, synaeresis, desiccation cracks, convoluted to disrupted bedding, stylolites, load casts, salt casts, scours, wave ripples, micro-faults and local unconformities. Planar crossbeds are rare. Vugs up to 5 cm, and lined with various minerals, are abundant.

Rare crossbeds around and southeast of Tremblay Sound indicate northwest-moving currents. Abundant stromatolite and bioherm elongation measurements from the same region (Fig. 16.8) indicate northeast or southwest currents are more common than northwest or southeast currents. A few elongation measurements in the west indicate northwest or southeast currents.

An abrupt vertical change in the terrigenous clastic content of the Society Cliffs Formation permits its division into two members: a lower, SC_1 member containing abundant terrigenous clastics, and an upper, SC_2 member with minor terrigenous detritus (Fig. 16.15). Major carbonate lithologies are similar in both members and are described below, prior to describing the two members.

Major Dolostone Lithologies

The major dolostone types are classified into 4 categories:

1. Thick bedded to massive dolostone.
2. Regularly laminated algal dolostone (cryptalgal laminite) to thin bedded dolostone.
3. Nodular to ellipsoidal, irregularly laminated dolostone.
4. Dolostone conglomerate and breccia.

The dolostones include dololutite, dolosiltite, dolarenite and dolorudite. Dololutite and dolosiltstone appear relatively more abundant in SC_1 member, whereas cryptalgal laminites (2) are more abundant in SC_2.

Categories 1, 2. The first two categories are the most abundant and occur in units up to 40 m thick throughout the formation. Thick bedded to massive dolostones are commonly internally faintly laminated to very thin bedded, or mottled buff and light brown. Disseminated quartz and feldspar grains occur in some beds and silty and clayey terrigenous material in others. Regularly laminated dolostones (cryptalgal laminites) are mostly planar laminated and are in general most abundant in the upper part of the formation. Minor wavy laminated varieties occur chiefly in the lower part of the formation (SC_1 and lower SC_2). Cryptalgal laminites are composed of thick, light-coloured laminae (2-5 mm) that alternate with thinner, darker-coloured to black laminae, mostly less than 2 mm thick. Very thin- to thin-bedded dolostones are commonly interbedded with the cryptalgal laminites.

Category 3. Nodular, irregularly-laminated dolostones make up much of the lower part (SC_1 and lower SC_2) of the formation in the Arctic Bay area, but are only a minor constituent elsewhere. The nodular dolostone contains nodules and lenticular beds of massive vuggy dolostone in, or separated by, irregularly-laminated dolostone that contains relatively abundant closely-spaced black carbonaceous laminae. Nodules are commonly outlined by a gypsiferous coating and some nodule cores contain gypsum crystal casts (Geldsetzer, 1973b; Olson, 1977).

Category 4. Several types of dolostone conglomerates and breccias occur in the Society Cliffs Formation. In most, both the clast and matrix are dolomite. Flat pebble conglomerate beds are a common, although minor, lithology throughout the formation east of Milne Inlet, and in the lower half to the west.

Round clast conglomerates occur locally throughout the Borden Basin, are less abundant than flat pebble conglomerates, and occur chiefly at the base or near the top of the formation.

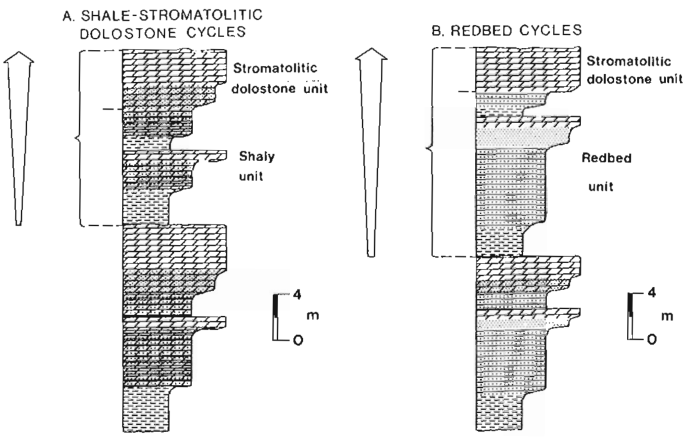
Dolostone breccias are abundant in the Strathcona-Adams Sound region and locally make up almost the entire formation. They decrease in abundance eastward from this area, and are rare elsewhere. Most breccias are related to karstification and/or solution collapse. They and associated dolostones are commonly hematite stained. Some upper breccia contacts are marked by reworked breccia redeposited as sediments. Chutes, channels and lenses of breccia, and carbonate and quartz veins are common in adjacent unbrecciated dolostone.

SC_1 Member

Most of this member comprises various combinations of dolostones described above in variously interbedded units. Terrigenous clastics, interbedded with the dolostones, contain gypsum beds in the eastern part of the basin.

Throughout most of the Borden Basin the terrigenous clastics are chiefly white to light grey quartz arenite, sublitharenite, arkose, and dolomitic quartz-feldspar granule conglomerate, with minor local grey to black shale. Locally, adjacent to the White Bay Fault Zone in the eastern part of the basin, the terrigenous clastics are varicoloured, and polymictic conglomerate clasts include granitic and dolomite granules and cobbles.

Evaporitic redbed sequences in the SC_1 consist of green, red, and black shale with interbedded grey-green and pink dolostone and white gypsum. Salt casts occur sporadically. Individual gypsum beds, commonly less than 1.5 m thick, range up to 3 m thick on northern Bylot Island. Elsewhere, gypsiferous coatings, gypsum casts, and crystal aggregates ("desert roses") occur locally in the dolostones.



A – shale-stromatolitic dolostone cycles near Tremblay Sound
B – redbed cycles on west side of Tay Sound

Figure 16.18. *Generalized shallowing-upward cycles of the SC_1 member east of Tremblay Sound.*

The SC_1 and SC_2 members are intergradational and the contact is arbitrarily defined as the horizon above which terrigenous clastics abruptly cease to be abundant. Thus the SC_1 member is 10-15 m thick in the westernmost part of Borden Basin, thickens gradually eastward to about 45 m on eastern Borden Peninsula, then thickens abruptly to 460 m near Tay Sound, and comprises the whole formation (480 m) locally on western Bylot Island.

Most of the SC_1 member consists of shallowing-upward cycles (Fig. 16.18) composed of two intergradational units. The lower, thicker, unit is composed of variously arranged terrigenous and/or dolostone clastics. The upper unit is composed chiefly of stromatolitic dolostone with minor clastic dolostone and, locally, shale. Gypsum occurs in the upper unit of redbed evaporitic cycles (Fig. 16.18B).

Only a few simple 2-3 m cycles occur in the western part of the Borden Basin. Cycles are commonly up to 30 m thick in the eastern part of the basin and locally are stacked in large complex shallowing-upward cycles as much as 135 m thick.

SC_2 Member

The SC_2 member consists predominantly of the four major buff to brownish grey dolostone lithologies already described. Rare terrigenous clastics are almost entirely shale. Poorly defined fining-, coarsening-, shallowing-, thickening-, and thinning-upward cycles up to 15 m thick are locally common. The SC_2 member ranges in thickness from 30 m to more than 600 m.

Interpretation

The widespread shale-dolostone shallowing-upward cycles, dewatering structures, variation in stromatolite types, and internal unconformities, suggest that the SC_1 member was deposited in shallow subtidal to intertidal environments west of Milne Inlet (Fig. 16.30). Redbed cycles and coastal gypsiferous sabkha sequences, in addition to features listed above, are interpreted as a variety of environments ranging from alluvial plain to supratidal and intertidal east of Milne Inlet and on Bylot Island.

The extensive planar cryptalgal laminites, stromatolite varieties, common biohermal zones, abundant local unconformities and dewatering structures, indicate that the SC_2 member formed in shallow subtidal to intertidal environments. Redbeds in both members east of Milne and Navy Board inlets and an abundance of bioherms in Milne Inlet region indicate clastic source areas were close in these regions and that there may have been syndepositional north-south faulting or an active hinge line in the vicinity of Navy Board and Milne inlets. Modest uplift of the Navy Board High is also indicated.

Some karsting and solution of Society Cliffs strata occurred prior to deposition of basal Victor Bay strata. Most, however, occurred subsequent to Victor Bay deposition. Deformation of Society Cliffs strata on northeast Bylot Island may be related to solution and removal of salt from the formation as well as to fault movement along the north Baffin and/or Cape Hay fault zones (see Kerr, 1980).

Victor Bay Formation

Strata of this formation outcrop primarily in two irregular bands rimming the axial zone of an open asymmetrical syncline on central Borden Peninsula (Fig. 16.2). The Victor Bay Formation consists primarily of variously interbedded dark shales, siltstones, dolostones and coarse carbonate clastics (Fig. 16.19). A petroliferous odour is common in the formation in eastern and northern Borden Basin. The major lithologies are commonly interbedded in sequences up to 55 m thick that differ from one another in thickness of beds (up to 8 m in VB_1), the major lithology, the relative proportions of contained lithologies, and the cyclical nature. Victor Bay strata range from 160 m thick at Arctic Bay to 730 m east of Milne Inlet.

The contact between the Victor Bay Formation and the overlying Strathcona Sound or Athole Point formations is variable. A dolostone flat boulder (clasts to 30 cm) conglomerate up to 6 m thick occurs locally in the Victor Bay Formation at or near the contact with both overlying formations. A carbonate round boulder (clasts to 1 m) conglomerate bed up to 10 m thick is common at or a short distance above this same contact in both overlying formations. The Victor Bay-Athole Point contact appears conformable and gradational to abrupt. The Victor Bay-Strathcona Sound contact ranges from gradational and sharply conformable to unconformable.

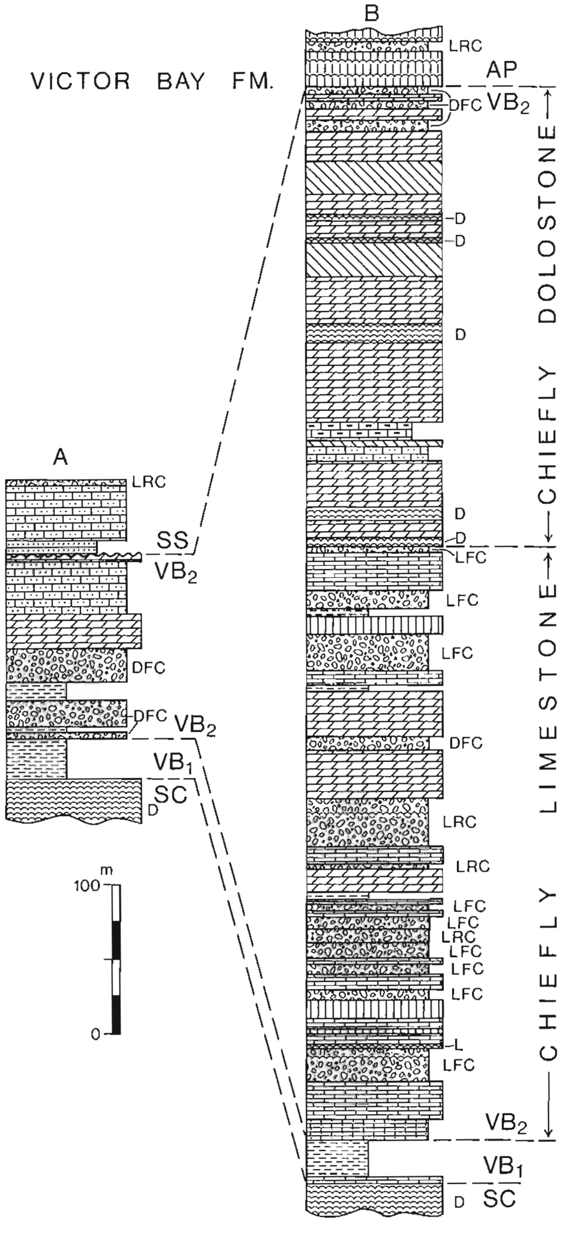
Most of the Victor Bay Formation south of the White Bay Fault Zone (Fig. 16.19) is divisible into two locally intergradational members: a lower shale-dominated (VB_1), and an upper carbonate (VB_2) member.

VB_1 Member

The VB_1 member occupies a broad fan-shaped area that is limited almost entirely to the Milne Inlet Trough (Fig. 16.26) and widens abruptly southward from the White Bay Fault Zone. This member is 30-50 m thick at Arctic Bay and east of Milne Inlet, but is 100-210 m between Nanisivik and Tremblay Sound.

Predominant lithologies (Fig. 16.19) are grey, brownish-grey or black laminated to thin-bedded locally pyritiferous shale, siltstone, calcilutite-calcisiltite, and dololutite-dolosiltite. Minor lithologies include calcarenite, dolarenite, siliciclastic carbonates, graphitic shale, quartz arenite, quartzwacke, and subarkose. Rare limestone and dolostone flat clast conglomerate interbeds occur in the upper part. Individual carbonate beds and carbonate-dominated units commonly increase in number and thickness upward in the member, as does the proportion of dolostone to limestone. Some sequences show shallowing-upward features and consist of shale through interbedded carbonate and shale, to carbonate or carbonate clast conglomerate. Others are thickening-upward, thinning-upward, and a few are deepening-upward cycles. The shale content of the VB_1 member is markedly less in the eastern and western parts of Milne Inlet Trough where the member is almost entirely dark limestone- and dolostone-cryptalgal laminites.

Rare sedimentary structures in the VB_1 include scour channels, crossbeds, ripples, wrinkle marks, synaeresis cracks, grooves, load structures, ball and pillow structures, flame structures, molar tooth, slump structures, and soft sediment folds.



A – near type section west of Arctic Bay
B – east of Milne Inlet

Figure 16.19. *Generalized stratigraphic sections of the Victor Bay Formation (VB).*

VB_2 Member

The VB_2 member is present everywhere in the formation and constitutes the entire formation north of the Milne Inlet Trough. It is composed of several major carbonate lithologies, several minor terrigenous clastics and shaly, silty and quartzitic to arkosic carbonates (Fig. 16.19), and locally abundant varicoloured chert. The VB_2 member ranges from 130 m thick at Arctic Bay to 702 m east of Milne Inlet. The contact with the underlying VB_1 member varies from gradational to locally disconformable. Where disconformable, carbonate, conglomeratic calcareous shale, or siliciclastic limestone beds lie on the VB_1.

Grey, brownish grey to black and white clastic carbonates constitute most of the VB_2 member. All are similar to those of the Society Cliffs Formation, but commonly contain more disseminated terrigenous clay, silt and sand. Laminated to thin bedded carbonates are the dominant lithologies, but medium- to thick-bedded, very thick-bedded, and lumpy-bedded (nodular) carbonates, as well as cryptalgal laminites, are also common. Carbonate flat pebble-boulder conglomerates are abundant in the south, and rare in the northwest. Round clast conglomerate and chaotic breccia occur locally, while edgewise conglomerate and angular-rectangular clast breccias are rare. Conglomerate clasts are predominantly intraformational, but a few may have been derived from the Society Cliffs Formation. Most clasts are less than 20 cm but range up to 70 cm; rare, deformed slump blocks up to 5 m across occur in the west.

Carbonates within the Milne Inlet Trough are chiefly limestone in the lower part of the member and dolostone in the upper part. Dolostone predominates throughout the member to the north.

Grey to black, locally pyritiferous, shales are the dominant terrigenous clastics, and occur as lamellar partings or thin interbeds, and in interbedded units up to 5 m thick. Shale is most abundant in the lower part of the VB_2 where the underlying VB_1 member is thickest. Minor siltstone, siliciclastic dolostone, locally conglomeratic arkose, quartz arenite, and quartz-pebble quartz arenite occur near the upper contact or adjacent to the White Bay Fault Zone.

Shallowing-up sequences (to 30 m) in the VB_2 consist of a lower unit of laminated to medium bedded carbonates, shale, or interbedded shale and carbonate, and an upper unit of massive carbonate or carbonate flat-clast conglomerate. Minor deepening-up cycles begin with carbonate flat-clast conglomerate which grades upward into massive vuggy carbonate, or into interbedded conglomerate and shale which in turn is overlain by interbedded carbonate and shale.

Stromatolites are most abundant in the VB_2 member in the eastern part of Milne Inlet Trough, and are uncommon in the Arctic Bay-Nanisivik area. Stromatolites occur as planar, and isolated to laterally-linked undulose, low domal (to 30 cm diameter), hemispheroidal (to 70 cm diameter), and columnar to digitate columnar (2-70 cm high) types. Together, these types occur in bioherms, which are 20-1500 m long and occur mainly in the upper part of the member.

Common sedimentary structures in the VB_2 include: molar tooth, tepee and other dewatering structures, synaeresis, desiccation cracks, scour channels, load casts, rip-ups, convolute bedding, soft-sediment folds, slump structures, graded bedding, boudinaged beds, birds eye structure, planar crossbeds, ripples, and microfaults. Rare structures include ball and pillow, stylolites, rain prints and quartz arenite dykes.

Elongated domal stromatolites in the eastern part of the Milne Inlet Trough indicate chiefly east-northeast or west-southwest currents with subordinate secondary trends (Fig. 16.8). A few crossbed measurements from two locations show southeast and east transport.

Interpretation

The abundant grey and interbedded black pyritiferous shale suggests that the lensoid VB_1 was deposited in a euxinic starved subtidal environment. The distribution of the VB_1, the absence of a basal unconformity, and the abrupt lithologic change from the underlying Society Cliffs Formation, indicates that the depositional area rapidly subsided and was tilted southward, possibly by movement along the Central Borden Fault Zone. This down-drop was accompanied by regional basinal sagging initially centred in east-central Borden Peninsula and later in the Milne Inlet area. The relative abundance of sandstone and the presence of quartz pebble-cobble conglomerate adjacent to the White Bay Fault Zone east of Milne Inlet suggests that there may have been some syndepositional movement along this zone as well. Carbonate-capped shallowing-up cycles, together with an upward-increasing carbonate content, suggest gradual shoaling punctuated by sudden syndepositional, fault-related deepening.

The regional abundance of carbonate flat clast conglomerate, the interbedded lithologies, and extensive variety of sedimentary structures indicate that most of the VB_2 member was deposited in intertidal to supratidal environments. Round clast conglomerates in the lower and uppermost VB_2 member are therefore probably beach deposits. Isolated large bioherms in the upper part of the member, however, indicate shallow subtidal environments were continuously maintained in some areas. The siltstone and arkose at the top of the VB_2 herald uplift of the Byam Martin High and Navy Board High source areas.

NUNATSIAQ GROUP

Strata overlying the Victor Bay Formation are assigned to the newly named Nunatsiaq Group, which includes the Athole Point, Strathcona Sound and Elwin formations (Table 16.1). Except for the Athole Point Formation, the group is dominated by terrigenous clastics.

Athole Point Formation

The Athole Point Formation outcrops southeastward from central Borden Peninsula in the eastern part of the Milne Inlet Trough (Fig. 16.2).

The Athole Point is composed chiefly of various dark-coloured limestones interbedded with subordinate siliciclastic limestones, calcareous siltstones, and sandstones. As in the Victor Bay Formation, these lithologies occur in variously interbedded sequences commonly 2-40 m thick. The limestones typically have a petroliferous odour. The Athole Point ranges from 525-585 m in thickness. It consists of three intergradational members east of Milne Inlet (Fig. 16.20): the lower, AP_1 member is predominantly limestone; the middle, AP_2 member is composed of limestone cryptalgal laminite; and the upper, AP_3 member is composed of interbedded limestones, cryptalgal laminites, and siliciclastic limestones and sandstones. The AP_2 member thins abruptly west of Tremblay Sound where only the AP_1 and AP_3 are presently differentiated.

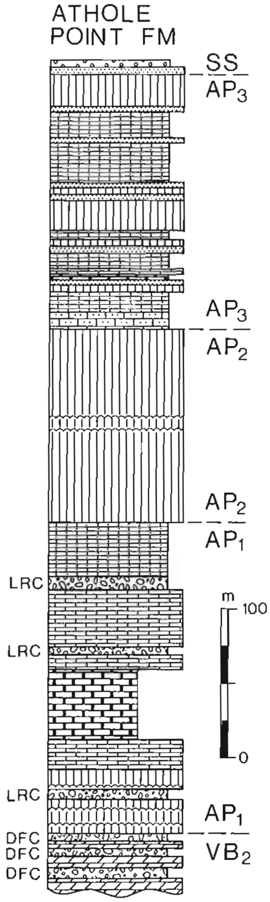


Figure 16.20

Generalized stratigraphic section of the Athole Point Formation (AP) east of Milne Inlet.

Athole Point strata are gradational into and interfinger with Strathcona Sound sandstones and shales, both vertically and laterally, and are overlain by SS_3 strata from the Tremblay Sound area eastward (Fig. 16.2, 16.8). Athole Point strata are characterized by a westward-increasing amount of terrigenous material, and grade into Strathcona Sound strata in central Borden Peninsula.

AP_1 Member

The lower AP_1 member consists predominantly of interbedded, thinly-laminated to medium-bedded calcilutites and calcisiltites, in part stromatolitic. The limestones commonly contain shaly or silty terrigenous material and some are siliciclastic. Limestone, cryptalgal laminites and lumpy bedded limestones are common. Minor calcarenite, limestoney, flat- and round-clast conglomerate, calcareous sandstone, orange-weathering siliceous limestone units (1-10 m), and rare dolomite, also occur.

Fining-up sequences include (but are not limited to) complete and incomplete Bouma cycles, and are particularly common in the upper part of AP_1 member. Calcareous quartz arenite- or siliciclastic limestone-based, calcisiltite-calcilutite cycles are typical. Rare coarsening-upward conglomeratic cycles occur near the base.

The AP_1 member is 127-210 m thick in the Milne Inlet-Tremblay Sound area.

AP_2 Member

The AP_2 member consists almost entirely of limestone cryptalgal laminites interbedded with units of laminated to thin-bedded limestone. The limestones are calcilutites and calcisiltites, and bedding is planar to undulatory. The lower part of the member coarsens slightly upward and is dark grey to brown and black. A few thin siliciclastic limestone and orange-weathering limestone beds separate the lower and upper parts of the member. The AP_2 member is 90-124 m thick.

AP_3 Member

Most of the AP_3 member consists of interbedded units of planar to undulose, thinly-laminated to very thick-bedded, grey to brown and black limestone, limestone cryptalgal laminite, and calcareous sandstone. Sporadic orange-weathering limestone beds are locally excellent market horizons. Minor limestone lithologies include lumpy-bedded strata, flat-clast breccias, and chip to boulder breccia. Limestones are dominantly calcilutites and calcisiltites with subordinate calcarenite and many contain terrigenous shaly or silty material. Dolostone is rare. The AP_3 member ranges from 190 to 370 m in thickness.

Terrigenous clastics in the AP_3 include fine- to coarse-grained, thin- to medium-bedded grey calcite-cemented quartzarenite to subarkose, sublitharenite, minor locally reddish shale and siltstone, and rare arkose and litharenite. These strata grade vertically into siliciclastic limestones and non-calcareous sandstones.

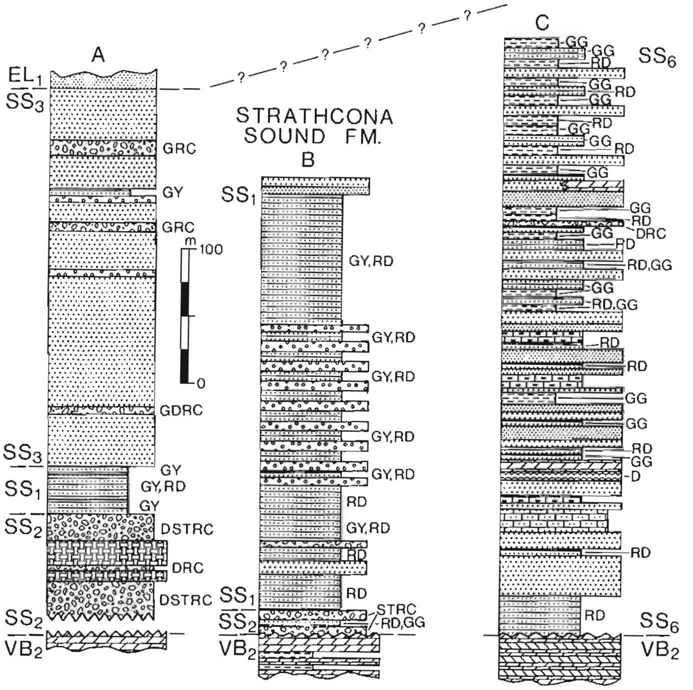
The AP_3 member contains numerous fining-up cycles, and a few coarsening-up cycles. At some localities there is an overall coarsening upward, with a sympathetic increase in terrigenous clastics. Abundant turbidite sequences (1-1.5 m) occur interbedded with limestones in zones to 100 m, or make up entire sections to 30 m thick. Basal beds are generally calcareous sandstone or sandstone and the uppermost beds are limestone or shale.

Sedimentary structures are most common in the AP_3 member and least in AP_2 member. Those common throughout the AP_1 and AP_3 members include: graded beds and Bouma cycles, flutes and load casts, flame structures, small-scale crossbeds, scour channels, convoluted bedding and soft sediment folds, slumped blocks and boudinaged beds, rip-up clasts, birds eye structure, small asymmetrical ripples, and microfaults. Minor structures include concretions, desiccation cracks, and clastic sandstone dykes. Stromatolites, although common, are predominantly planar to undulose lamellar types with poor lateral linkage. Low domal and hemispheroidal varieties to 70 cm across, and columnar types up to 10 cm high are common. Small bioherms and rare large ones (up to 1.5 km long) occur locally. Elongations of individual and biohermal stromatolites is minimal and apparently random. Crossbeds from isolated localities indicate unimodal west-northwest transport.

Interpretation

The restriction of the Athole Formation to the eastern Milne Inlet Trough, and abrupt vertical change in lithologies between the Victor Bay and Athole Point indicate the eastern Milne Inlet Trough probably rapidly deepened by normal faulting which terminated Victor Bay deposition.

Limestone round-clast and flat-clast conglomerates locally associated with desiccated beds in the basal Athole Point probably include some debris flows. The cryptalgal laminites associated with these conglomerates suggest the basal Athole Point accumulated in an intertidal-supratidal environment. The widespread cryptalgal laminites and turbidites throughout the remainder of the formation suggest intertidal to subtidal deposition. The turbidites may have been generated by storms or by syndepositional fault movement. Presence in AP_3 of a few coarsening-up



A – east of head of Baillarge Bay
B – north of head of Strathcona Sound
C – western Bylot Island

Figure 16.21. *Generalized stratigraphic sections of the Strathcona Sound Formation (SS).*

sequences and an upward increase in both grain size and amount of terrigenous clastic material suggests a progressive shallowing and infilling of the basin.

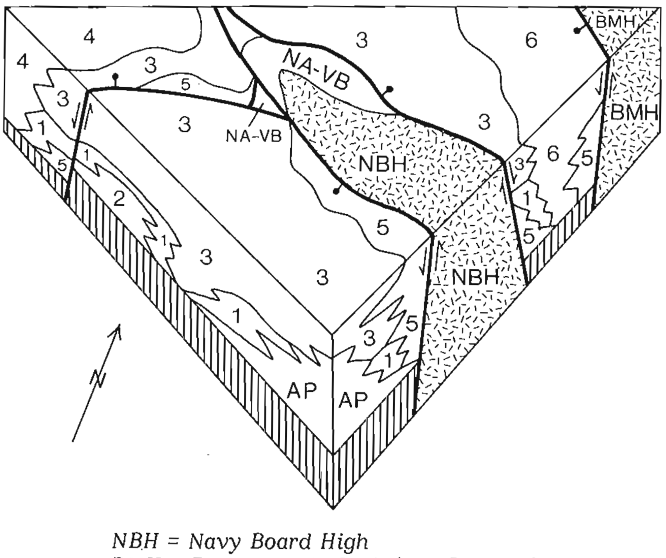
Strathcona Sound Formation

The Strathcona Sound Formation is most abundant in the axial portion of a major syncline from Baillarge Bay to Tremblay Sound (Fig. 16.2). In this paper, the formation includes most of the strata mapped previously as undivided Victor Bay-Society Cliffs on western Bylot Island (Jackson and Davidson, 1975).

The formation comprises a wide variety of complexly interfingering lithologies. These are mainly laminated to thin-bedded shales and siltstones, thin- to very thick-bedded sandstones, stromatolitic dolostones, dolostone conglomerates, and polymictic conglomerates (Fig. 16.21). These various lithologies occur singly, or interbedded with others in units to 30 m, and rarely, 110 m.

Faults, and the varied nature of the formation, make thickness determinations difficult. Partial sections up to 813 m are present and the formation is estimated to exceed 910 m. The contact between the Strathcona Sound and Victor Bay formations varies from conformable (abrupt to gradational) to disconformable. Contacts of the Strathcona Sound with the laterally equivalent Athole Point and the overlying Elwin Formations are conformable and gradational.

The Strathcona Sound Formation has been subdivided into six intergradational members (Table 16.1; Fig. 16.22). Only two, however, are typically present: a lower shale-siltstone member (SS_1), and an upper arkose-greywacke member (SS_3).



NBH = Navy Board High
BMH = Byam Martin High (see Fig. 16.26)

Figure 16.22. *Schematic diagram, looking north-northwest, showing relationships of Strathcona Sound members (1 = SS_1, 3 = SS_3, etc.). Left section is along Milne Inlet Trough. Right section extends from the Trough northward to western Bylot Island.*

The greatest variety of sedimentary structures occurs in SS_1 and SS_3 members, and include: ripples (symmetrical, asymmetrical, interference, climbing), planar trough and herringbone crossbeds, flutes, flames, load casts, ball and pillow, graded beds (normal and reverse), rip-ups, convolute bedding and slumps, microfaults, scour channels, tool marks, desiccation cracks (SS_1), and clastic dykes. Relatively few of these structures are common in SS_5 member and they are rare in both the SS_2 and SS_4.

Crossbeds from the SS_1 and SS_3 members indicate predominantly southwest to west transport south of the White Bay Fault Zone (Fig. 16.8, 16.25, 16.36), and westerly and northerly to easterly transport north of the Hartz Mountain Fault Zone.

Lower Shale-Siltstone Member (SS_1)

Red to red-brown or locally grey to green interbedded calcareous shales and siltstones predominate in this member (Fig. 16.21B). Both commonly contain disseminated quartz and feldspar sand grains. Very fine grained arkose and subordinate litharenite are locally abundant. Carbonate and medium to coarse grained arkose, litharenite, greywacke and rare quartz arenite beds and lenses occur sporadically throughout the member. The carbonate is in part stromatolitic, commonly contains terrigenous material, is chiefly dolostone in central Borden Peninsula and limestone to the northwest and southeast.

Fining- and thinning-upward cycles (to 2 m) occur at several localities. Fining-upward cycles consist chiefly of arkose, litharenite or siltstone grading upward into siltstone, shale, or calcisiltite.

Carbonate-clast orthoconglomerates commonly occur in the lower part of the member, particularly at or a short distance above the base of the member. Most conglomerate beds are less than 5 m thick and contain clasts of: grey to reddish massive or laminated or stromatolitic carbonate to

30 cm (rarely 100 cm); minor shale or siltstone; and rarely, gneisses. Clast rounding increases westerly and easterly away from the centre of Borden Peninsula. In general, degree of rounding and proportion of shale, silt and gneiss clasts increase upward, and gneiss clasts commonly predominate in conglomerates in the upper part of the member.

The SS_1 member is best developed in south- and north-central Borden Peninsula and constitutes most of the formation in the latter. It is laterally intergradational with the SS_2, SS_4, SS_5, SS_6 members, and with the Athole Point Formation (Fig. 16.22). SS_1 red siltstone overlies karsted Victor Bay dolostone south of Elwin Ice Cap. Relationships with the overlying SS_3 member vary from interfingering and intergradational to unconformable. Measured partial sections indicate a maximum thickness of more than 348 m.

Dolostone Member (SS_2)

This member is composed almost entirely of dolostones. Light grey, stromatolitic dolostone predominates. Locally, stromatolites are rare and interbeds of breccia, round-clast and flat-pebble conglomerate, and shaly dolostone with local synaeresis cracks are present.

The SS_2 member extends from the head of Strathcona Sound to just north of Elwin Ice Cap (Fig. 16.2). The biohermal part of the member is conformable and gradational with the underlying Victor Bay Formation. SS_2 strata laterally interfinger with, and are locally overlain by, SS_1 red shale and siltstone. Maximum thickness may be at least 300 m.

Laterally the central part of the member is composed mostly of elliptical, narrow, elongate bioherms up to 1 km long and 120 m thick. Bioherms have a variety of orientations, but easterly and northeasterly elongations predominate. They are composed largely of vertically stacked and laterally-linked hemispheroids with some undulose planar or branching columnar stromatolites, and rare, more complex types. Each bioherm is developed from a number of coalescing growth centres 2-5 m high.

The marginal areas of the SS_2 member are predominantly oligomictic to polymictic boulder orthoconglomerates and breccias which occur in irregular lenses (2-140 m thick) interbedded with dolosiltite, dolarenite, siltstone and shale (Fig. 16.21A). Most clasts are dolostones derived from the Victor Bay, Society Cliffs, and from upper Victor Bay-lower Strathcona Sound biohermal carbonates, but granitic gneiss boulders occur locally. Most clasts are 30 cm or less across, but south of Baillarge Bay (Fig. 16.2) they range up to 2.5 m. In this same area olistoliths up to 1 km long, and underlain by olistostromes, were probably derived from the Society Cliffs Formation to the south.

Arkose-Greywacke Member (SS_3)

This member is composed mostly of grey to green, locally red-brown, sandstones interbedded with conglomerates, siltstones, shales, and locally with minor dolostones and limestones. The sandstones are fine to very coarse grained and are commonly conglomeratic, with scattered granules and pebbles of granitic rocks, quartz and feldspar, and locally

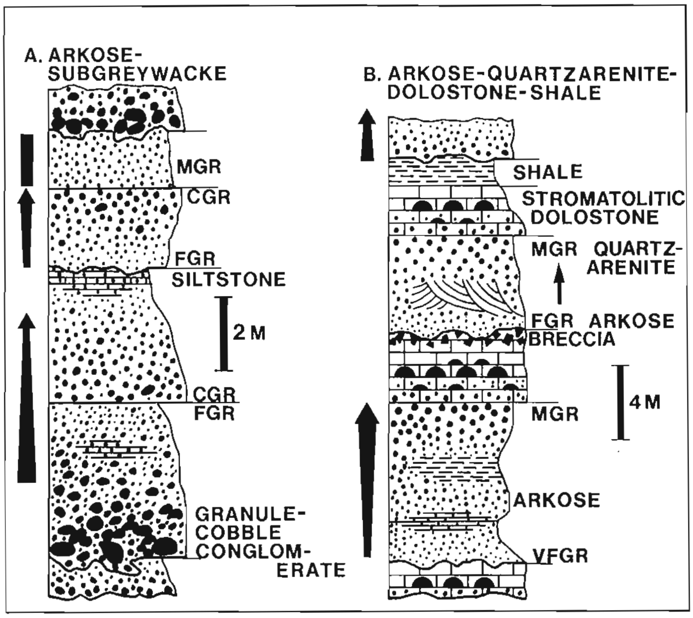


Figure 16.23. *Representative cycles in Strathcona Sound and Elwin members. A – fining- and coarsening-upward cycles in SS_3 and SS_5 members in central Borden Peninsula; B – coarsening- and shallowing-upward cycles in SS_6 and EL_1 members along northern Navy Board Inlet. VFGR = very fine grained, etc.*

of shale, siltstone, and carbonate. The sandstones are invariably calcite cemented. Lithic arenites and arkoses predominate, although greywackes, sublitharenites and subarkoses are abundant; quartz arenite and quartz wacke are minor.

The SS_3 member conglomerates are polymictic orthoconglomerates with moderately to highly rounded and spherical clasts. Most are pebble-granule size, but a few range to more than a metre across. Granitic gneiss, quartz, and feldspar are the predominant clasts but sandstone, siltstone, shale and carbonate clasts are common. Carbonate clasts are predominant locally in the basal part of the member, and conglomerate or chert clasts occur locally.

SS_3 strata are interbedded cyclic packets of fining-upward units 1-90 m thick. In addition to the individual fining-upward beds and cycles, the packets also typically fine upward. Basal beds commonly rest on scoured surfaces and local conglomerates fill channels up to 15 m deep. A large variety of fining-up cycles are present (Fig. 16.23A). In addition, thinning-up cycles are common, and some coarsen upward. Locally, a disconformity within the SS_3 may divide lower fine grained varied lithologies and upper coarse calcareous feldspathic wacke.

The SS_3 member lies disconformably on Victor Bay strata in the vicinity of Strathcona Sound, grades laterally into and interfingers with SS_4, SS_5 and SS_6 members, and interfingers laterally with and conformably overlies Athole Point Formation (Fig. 16.2, 16.22). The contact with the overlying Elwin Formation is gradational. Partial sections indicate the member is more than 410 m thick. Locally, it may constitute the entire formation, as in the vicinity of Strathcona Sound (Fig. 16.22). The member is generally much thinner on northern Borden Peninsula.

Siltstone-Greywacke Member (SS_4)

This member consists predominantly of a monotonous sequence of laminated grey siltstone interbedded with minor very thin to medium-bedded grey subarkose and feldspathic wacke. It occurs between Strathcona Sound and Baillarge Bay. The strata are commonly calcareous and a few thin beds of limestone and dolostone are present. Sedimentary structures are rare, but increase in abundance immediately below the Elwin Formation, and include ripple marks, cross-lamination, small low angle planar crossbeds, load structures, flutes, and scoured bases.

About 270 m of SS_4 strata were measured in a partial section but the member may be nearly 600 m. It is transitional eastward into the SS_1 and SS_3 members, and is gradational into the overlying Elwin Formation.

Polymictic Conglomerate Member (SS_5)

This member is most abundant along the White Bay Fault Zone (Fig. 16.22, 16.30, 16.31), and lenses-out abruptly a short distance away from it. The SS_5 member consists chiefly of red to pink and grey-green, fine- to coarse-grained arkoses, polymictic pebble-boulder orthoconglomerates interbedded with subordinate lithic arenites, siltstones, shales and calcisiltites, and minor quartz arenites. Most strata are calcareous; calcisiltites and stromatolitic limestones (to 20 m) occur locally in the lower part of the member. Slumped fault blocks of massive, bedded (VB_2?) and stromatolitic dolostones up to 70 m thick and 2 km long occur southeast of Elwin Ice Cap.

The member contains ortho-, para-, oligomictic, and polymictic conglomerates which range from a few centimetres to a lens more than 186 m thick containing rounded carbonate clasts up to 10 m. Most clasts are less than 1 m across.

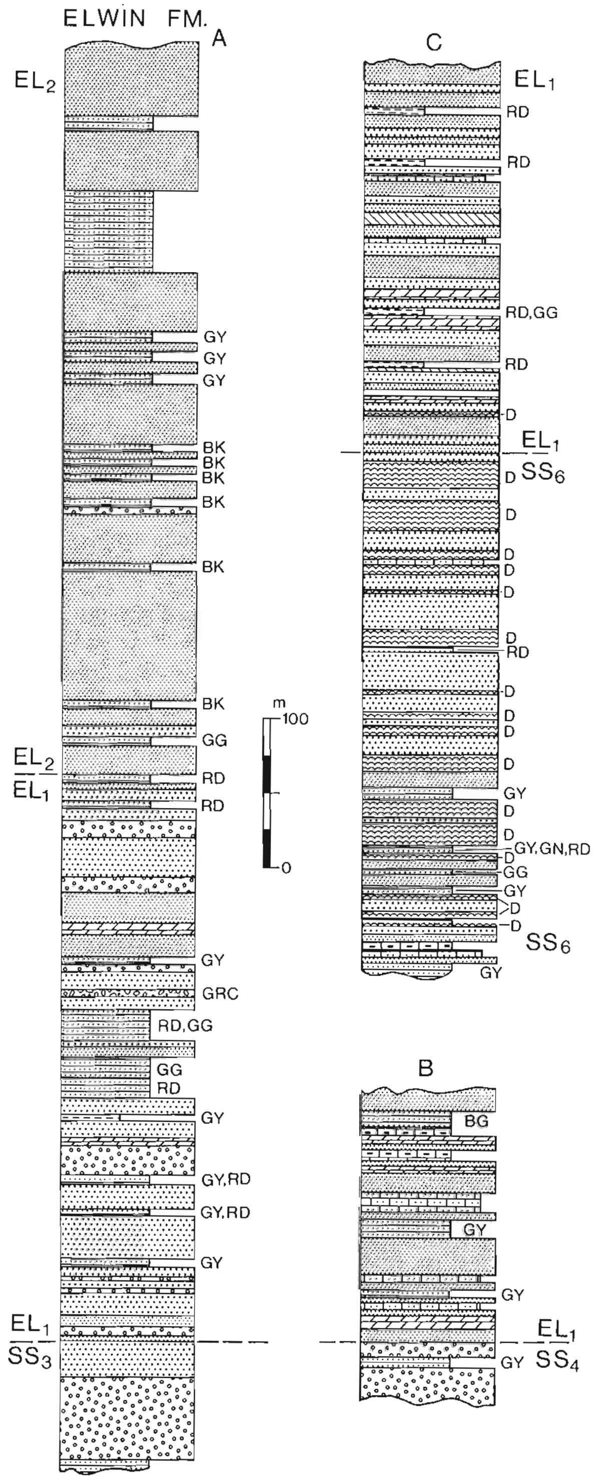
The polymictic conglomerates contain clasts mostly of carbonate and basement gneisses, minor siltstone, shale, and sandstone, as well as local conglomerates and quartz arenite. The conglomerates are concentrated in fans which grade along and away from the White Bay Fault Zone into typical SS_1, SS_3 and Athole Point strata.

SS_5 strata units, up to 50 m thick in the fans, are commonly less than 5 m elsewhere. Sedimentation in the SS_5 is dominated by fining-upward cycles (Fig. 16.23) but a few shoaling-upward and coarsening-upward cycles are also present. At least 812 m of SS_5 strata occur along the White Bay Fault Zone.

Sedimentary structures in the member include: ripple marks, trough crossbeds, convolute beds, load casts, rip-up clasts, microfaults, and stromatolites. Crossbeds and imbricate clasts indicate dominantly southerly to westerly transport.

Interbedded Member (SS_6)

Whereas only a few lithologies predominate in the other Strathcona Sound members described above, several lithologies are thinly interbedded throughout all but the lowermost 60 m of Strathcona Sound strata on western Bylot Island and along the west side of Navy Board Inlet (Fig. 16.21C, 16.24C). This thinly interbedded nature is characteristic of the lower EL_1 member of the Elwin Formation, which, however, commonly has more quartz arenite and less carbonate strata than does the Strathcona Sound Formation (Fig. 16.21, 16.22). Therefore strata bordering northern Navy Board Inlet (Fig. 16.2) are assigned to the Interbedded member (SS_6).



A – composite section, Elwin Inlet
B – south side of Baillarge Bay
C – west side of northern Navy Board Inlet

Figure 16.24. *Generalized stratigraphic sections of the Elwin (EL) and Strathcona (SS) formations.*

This member is composed of interbedded arkoses, lithic arenites, quartz arenites, siltstones, shales, dolostones, and siliciclastic and stromatolitic dolostones that are red, green, grey, buff, white, laminated to medium bedded, planar to wavy bedded, and locally lumpy bedded. Oolites, pisolites and frosted sand grains are common in the upper part, and some beds contain disseminated pyrite.

Although each lithology predominates in units commonly 10 m or less thick, much if not most of the member consists of red-coloured lithologies in 5-130 m sequences separated by grey-green lithologies in 8-50 m sequences. Fining-, shallowing-, and thinning-upward cycles 2-40 m thick are characteristic (Fig. 16.23B), and include sandstone-siltstone-dolostone or shale cycles. Coarsening- and thickening-upward cycles are less common.

The SS_6 member may be divided into a lower sequence characterized by relatively thick lithological units, a middle sequence containing abundant interbedded quartz arenite, and an upper sequence of arkose interbedded with either siltstone and shale or stromatolitic dolostone (Fig. 16.21C, 16.24C). The contact with the underlying Victor Bay and overlying Elwin formations is conformable and gradational. Tentative correlation between partial sections of 450 m and 350 m (Fig. 16.21C, 16.24C) indicates the SS_6 member is at least 550 m thick and probably constitutes the entire formation in the Navy Board Inlet area.

Sedimentary structures typical of the SS_1 and SS_3 members are common throughout the SS_6 (excluding flutes and tool marks). Synaeresis, birds eye, tepee and other dewatering structures, and molar tooth occur in the carbonates, which also contain planar, low domal and laterally-linked hemispheroidal stromatolites. Crossbeds, locally to 2 m, and stromatolite elongations indicate west-northwest transport was dominant.

Interpretation

Lithologies, cycles and contained structures of the Strathcona Sound Formation, as well as facies relations between members, indicate that the SS_5 member is composed of partially coalescing alluvial fan complexes that accumulated rapidly along the active White Bay Fault Zone adjacent to the rising Navy Board High (Fig. 16.22) that was stripped of a cover of previously-deposited Bylot Supergroup strata. The few shoaling-up sequences suggest that local parts of the fan complexes may be lacustrine or shallow marine.

The SS_5 member interfingers laterally southward with the lower shale-siltstone (SS_1) and upper arkose-greywacke (SS_3) members (Fig. 16.22). The lower SS_1 member probably accumulated in overbank alluvial and intertidal environments throughout the Borden Basin. The carbonate clast conglomerates immediately above the Victor Bay Formation in the Milne Inlet Trough are interpreted as debris flows, although some may be breccias related to the solution of interbedded evaporites. Channel deposits, conglomerate-based fining-upward cycles, and terrigenous clastic-carbonate cycles, indicate the SS_3 was deposited in mixed alluvial and intertidal environments.

Carbonate beds in the Milne Inlet Trough are best developed near the White Bay Fault Zone, which suggests that as the Trough subsided, it was also tilted downward toward the north, and carbonate deposition occurred in a tongue of the sea or in shallow ephemeral lakes. The few northeasterly to easterly directed crossbeds in SS_3 lithic arenites that disconformably overlie Victor Bay and SS_1 strata southeast of Strathcona Sound (Fig. 16.2) suggest that some SS_3 strata may have been derived from a southerly or westerly source related to renewed activity along the Central Borden Fault Zone or a fault along Admiralty Inlet.

The SS_2 member accumulated as a small biohermal carbonate platform in a shallow subtidal to intertidal environment. Westward- and southward-prograding SS_1 clastics buried the locally-emergent carbonate platform. West of this platform, the monotonous grey siltstone-greywacke turbidites of the SS_4 member accumulated in a downfaulted wedge-shaped basin between Strathcona Sound and Elwin Inlet (Fig. 16.22).

The large scale lateral variations, rapid lateral changes within individual beds, the oolites, pisolites and frosted quartz and feldspar grains, and the abundance of fining- and shallowing-upward cycles, all indicate that the SS_6 member was deposited in closely-spaced alluvial plain to supratidal and intertidal environments.

The distribution of SS_2, SS_4 and SS_6 members suggests that north- to northeast- as well as northwest-trending faults continuously influenced sedimentation. The vertically and laterally variable nature of the Strathcona Sound and Athole Point formations, the presence of large olistoliths underlain by olistostromes along the western side of the SS_2 member, and large slumped carbonate fault blocks in the SS_5 member attest to continuous tectonic instability during Strathcona Sound deposition.

Elwin Formation

The Elwin Formation is the uppermost formation of the Nunatsiaq Group, and outcrops only on northern Borden Peninsula (Fig. 16.2). It is composed of varicoloured quartz-rich sandstones, with minor siltstones, dolostones and shales (Fig. 16.24). The strata are very thin- to thick-bedded, and the sandstones are equigranular and generally contain even less matrix than those of the Strathcona Sound.

The Elwin Formation is conformable and gradational with the underlying Strathcona Sound Formation. This contact is difficult to define, but is commonly marked by the abrupt appearance of quartz arenite, and a slight but distinct diminution in sandstone grain size. The Elwin is overlain unconformably by lower Paleozoic strata, ranges from 870 to 1220 m thick, and is divided into two intergradational members.

Planar crossbeds, small scour channels, synaeresis, symmetrical and asymmetrical ripples, and frosted sand grains are common throughout the formation. Larger clasts of rounded quartz, feldspar, and dolostone, and angular rip-up clasts of siltstone, sandstone and shale, are locally present. Desiccation cracks are sparse.

Lower Member (EL_1)

The EL_1 member is composed of interbedded red, grey-green, white and buff subarkoses, quartz arenites, lithic arenites, siltstones, and dolostones (dololutite, dolosiltite, intraformational dolorudite). Commonly, the sandstones contain kaolinized feldspar and are very fine to medium grained, although coarse grained sandstones are locally common (Fig. 16.24). Rare stromatolitic dolostone contains planar and wavy laminated, low domal, and laterally-linked hemispheroids.

The contact between the lower and upper Elwin members is gradational, and is marked by an abrupt increase in the amount of quartz arenite and a corresponding decrease in the number of other lithologies. The EL_1 member ranges from more than 265 m to 375 m in thickness.

Strata occur variously interbedded in units up to 30 m thick but most are 10 m or less. Depositional cycles are common but are not as apparent as in the underlying Strathcona Sound. Cycles range from simple to complex and some include two of the following types in compound cycles: thinning-up (bedding), thickening-up, coarsening-up (grain), fining-up, and shallowing-up (Fig. 16.23B). Fining- and shallowing-up cycles include: arkose to quartz arenite to dolostone, with red shale below and/or above the dolostone; arkose to shale to locally brecciated dolostone; coarsening-up (and deepening?) cycles include dololutite to shale to arkose.

Trough crossbeds, climbing ripples, concretions, slump structures, microfaults, load casts, and sole marks are common in the member in addition to structures already noted for the formation as a whole. Tepee structures occur in some dolostones and oolites and pisolites are common locally, as are halite casts 80-100 m above the base of the member.

Paleocurrent patterns are polymodal and have high dispersion (Fig. 16.8).

Upper Member (EL_2)

The EL_2 member is predominantly very fine to medium grained, white to light grey, buff to orange, red and light-green quartz arenites interbedded with minor siltstones (Fig. 16.24A). Although well to poorly sorted, they are finer grained, better sorted and more mature than sandstones in the lower member. Rare subarkose and sublitharenite occur in the lower part of the member. Most siltstones are black in the lower part of the member and grey in the upper part. The EL_2 member is at least 495 m thick at Elwin Inlet and may be as much as 850 m thick elsewhere on Borden Peninsula.

Strata of the upper member are thicker bedded than those in the lower member and occur interbedded in units up to at least 90 m thick, although most are less than 17 m. Typically the member comprises alternating quartz arenite units (65 per cent), and quartz arenite-shale units (35 per cent) that are 70 per cent quartz arenite. Where present, cyclic deposition is vague, ill-defined, and constitutes a few graded beds and coarsening-up cycles.

In addition to the sedimentary structures characteristic of the whole formation, rare current lineations occur in the middle of the member. Paleocurrent patterns from crossbeds show high dispersion, and indicate polymodal, east-northeast transport (Fig. 16.8). Directions most commonly indicated in both members are: north-northwest, northeast, and southeast (Fig. 16.8).

Interpretation

The large variety of lithologies, depositional cycles and sedimentary structures, indicates the EL_1 member accumulated in intratidal to supratidal and alluvial braidplain environments. The local evaporitic strata probably accumulated in ephemeral pools on tidal flats or in lagoonal areas.

The abundant trough and planar crossbeds, asymmetrical and symmetrical and climbing ripples, small channels, desiccation cracks, and coarse subarkose lenses suggest a braided fluvial environment for the upper 100 m of the EL_1 at Elwin Inlet. East-northeast to southeast trends predominate in the paleocurrent pattern for this unit, as well as for the rest of the EL_1 member of western Borden Peninsula, and suggest that some detritus may have been supplied from sources in the northern Brodeur Pininsula and/or Devon Island areas.

Predominant west-southwest paleocurrent trends adjacent to Navy Board Inlet suggest the Byam Martin High may have been the major EL_1 source area near Navy Board. However, northerly paleocurrent trends on north-central Borden Peninsula indicate that here the Navy Board High may have been the major source area. In these regions, as in the others, subsidiary northeast and east-southeast paleocurrents suggest that tidal currents were active.

The kaolinized feldspar, abundant quartz arenite, and better sorted nature of the lower Elwin as compared with Strathcona Sound strata, suggest an abatement of tectonic activity and EL_1 strata are more reworked, possibly during prolonged transport. The thin nature of the EL_1 member, and relative paucity of contained carbonate strata on central Borden Peninusla compared with areas to the west and east, suggest that environments more typical of EL_2 deposition prevailed slightly earlier in central Borden Peninsula, or that this area was farther from source areas during EL_1 deposition compared with areas to the west and east.

The lithologies, structures, and polymodal paleocurrent patterns indicate the EL_2 accumulated chiefly in a subtidal sandstone shelf environment. The major north-northwest and south-southeast paleocurrent trends are interpreted as a result of tidal activity and the east-northeast trends as being related to longshore currents (Fig. 16.8). The predominance of equigranular quartz arenites suggests that previously-deposited sands were reworked and redeposited during a period of prolonged relative stability.

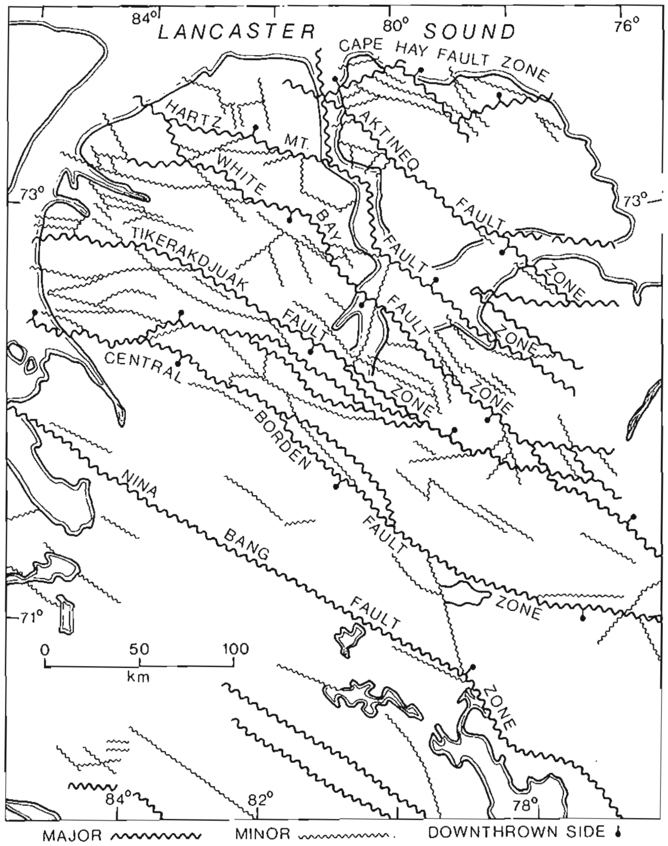


Figure 16.25. *Major fault zones and associated minor faults in northwestern Baffin Island east of Admiralty Inlet.*

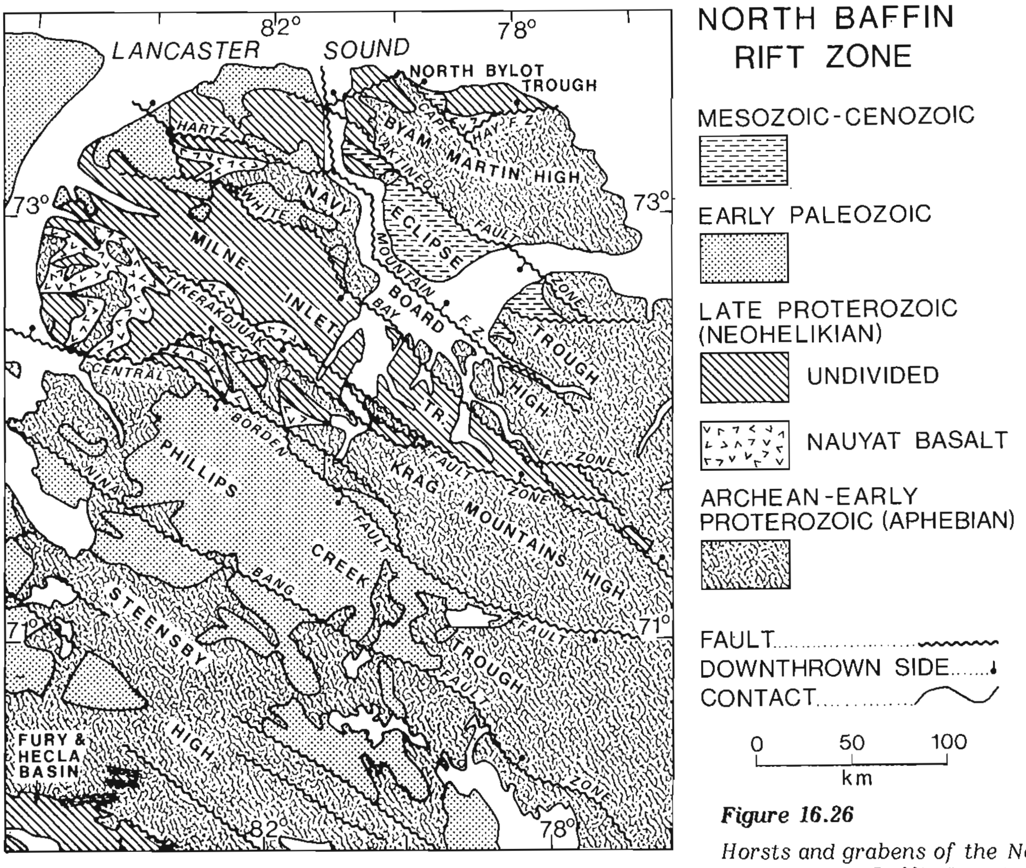


Figure 16.26

Horsts and grabens of the North Baffin Rift Zone, northwestern Baffin Island.

The cumulative paleocurrent diagram indicates predominantly easterly transport for the Elwin Formation compared with predominantly westerly transport for all of the other formations (Fig. 16.8). This change in direction may signify increased tectonic activity to the west, perhaps in the vicinity of the Boothia Arch or farther out in the basin.

FAULTING

Faulting, predominantly along northwest-trending faults (Fig. 16.25, 16.26) has taken place in northwest Baffin Island from the end of Aphebian to Recent time (Jackson et al., 1975, 1978a; Jackson and Morgan, 1978). The gentle folds and local vertical dips in Bylot Supergroup strata were caused by this faulting. Several small northwest-dipping, low-angle thrusts may be mainly syndepositional but may also be related to postdepositional compression from the northwest.

Most of the faulting during deposition of the Bylot Supergroup occurred as vertical movements along steep northwest-trending faults. Movements were episodic and culminated during deposition of the Arctic Bay-Fabricius Fiord formations and later during Strathcona Sound-Athole Point Formation deposition.

Northerly to northeasterly trending faults were much less widespread and of relatively local extent during sedimentation. Parts of the Milne-Navy Board Inlets zone acted as a hinge zone throughout sedimentation, whereas faulting occurred along other parts. Alignment of this zone with the east margins of Devon and Ellesmere islands and Nares Strait (Fig. 16.1) suggests (vertical) movement may have occurred along this linear in Neohelikian time. Penecontemporaneous basic volcanics in basal strata of the aligned Fury and Hecla, Borden and Thule basins, apparently absent at Somerset Island to the west, suggest most volcanism probably occurred along northerly to northeasterly trending fissures.

Although much faulting postdates Bylot Supergroup deposition, the present fault distribution and nature of the North Baffin Rift Zone (Fig. 16.25, 16.26) probably approximates the nature of horst and graben development at the close of Elwin deposition. A low negative Bouguer anomaly over southwest Bylot Island suggests that although components of the Eclipse Trough (Fig. 16.26) were in existence in the Neohelikian, they were modified subsequently. Lancaster Sound may have been a graben area within the North Baffin Rift Zone, although the present graben (Kerr, 1980) may have a more easterly trend than the Neohelikian one.

EVOLUTION OF BORDEN BASIN

In this section the deposition of the Neohelikian Bylot Supergroup is summarized, brief comparisons and correlations are made with Neohelikian strata in adjacent basins, and the origin of Borden Basin is discussed.

Filling of the Basin

Deposition of Bylot Supergroup commenced with faulting, terrigenous clastic sedimentation and basic volcanism (Eqalulik Group) during initial marine transgression of a northwesterly dipping peneplaned Archean-Aphebian grani-toid basement complex. Later broad downwarping resulted in regional shallow marine carbonate sedimentation (Uluksan Group). Continued downwarping resulted in renewed

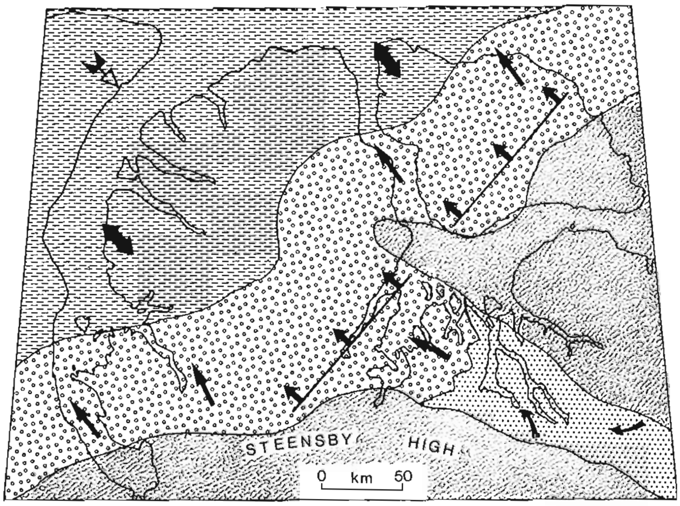


Figure 16.27. *Paleoenvironmental reconstruction for Nauyat and lower to middle Adams Sound deposition.*

faulting, collapse, and mild compression, and finally, basin stability (Nunatsiaq Group). Presence of a marine basin to the northwest is indicated during most of Bylot Supergroup deposition.

The dominantly braided fluvial sandstones of the lower Nauyat and Adams Sound accumulated in a narrow channel southeast of Milne Inlet that may have originated by faulting. This channel is transitional into an alluvial plain delineated by an anastomosing network of upward-coalescing braided stream deposits (Fig. 16.27, 16.28). Lower Nauyat strata in the northern part of the basin were deposited in mixed fluvial to marginal-marine environments.

Regionally, sedimentation was interrupted by quiescent extrusion of southeast-thinning Nauyat terrestrial tholeiitic plateau basalts. These volcanics were probably not deposited in the southeastern part of the basin. The present distribution of the flows is determined by major northwesterly trending faults. However, northwestward thickening of the volcanics suggests the erupting fissures were probably northeasterly trending. Renewed sedimentation rapidly buried the volcanics beneath prograding braided fluvial clastics before appreciable erosion or incision.

Throughout deposition of the Adams Sound, fluvial environments predominated in the southeast part of the basin, whereas mixed fluvial and marginal marine environments predominated in the northeast. In the western (north and south) part of the basin fluvial sedimentation occurred throughout AS_1 deposition and mixed fluvial and marginal marine sedimentation during AS_{2-3} deposition.

Periodic regional subsidence was accompanied by marine transgression until marine sedimentation prevailed during deposition of the interfingering and intergradational Fabricius Fiord and Arctic Bay formations. Lower Fabricius Fiord (FF_1) coastal and shallow shelf sediments accumulated along the southern margin of the basin. Coeval lower Arctic Bay (AB_1) intertidal sediments were deposited in the southeast and mixed intertidal to shallow subtidal clastics in the rest of the basin.

Major intrabasinal faulting and extrabasinal uplift began late during lower Arctic Bay-Fabricius Fiord deposition, and continued sporadically throughout deposition of both formations. The most pronounced effect of this faulting and uplift was in the southern part of the basin, where large coalescing marine-influenced delta fan complexes (FF_{2-4}) were deposited along the Central Borden Fault Zone (Fig. 16.26, 16.28, 16.29) and extend upward to interfinger with lower Society Cliffs strata.

While these Fabricius Fiord delta-fan complexes formed, the coeval Arctic Bay member (AB_2) was deposited in marine-influenced deltas and on clastic shorelines in the southeast, and in a shelf environment elsewhere. The AB_3 member accumulated in a subtidal, locally starved, basin environment. AB_4 strata, including deltaic-alluvial fan complexes, were deposited in shallower mixed clastic shoreline and shelf environments.

The upward-increasing carbonates in the Arctic Bay are precursors of the thick platformal carbonates of the Society Cliffs and Victor Bay formations. However, the disconformity common at the base of the Society Cliffs suggests that faulting or regional warping again may have played a major role in changing the nature of sedimentation.

Carbonate shelf environments were maintained throughout most of the basin during deposition of the Society Cliffs, with subtidal to intertidal sedimentation predominating. Syndepositional faulting produced local karst topography and elevated the proto-Navy Board and Byam Martin highs. Terrigenous clastics shed from these highs accumulated in alluvial plain and intertidal to supratidal or sabkha environments throughout most of the eastern Society Cliffs (Fig. 16.28, 16.30).

Downfaulting of the Milne Inlet Trough, accompanied by a sagging of the south-central part of the basin at the beginning of Victor Bay sedimentation, resulted in deposition of a northward-thinning tongue of subtidal shale (Fig. 16.28, 16.31). Elsewhere, intertidal to supratidal, with rare subtidal, carbonate deposition predominated.

Major syndepositional faulting during accumulation of the Strathcona Sound and Athole Point formations produced a large variety of local and regional depositional environments and interfingering lithologies (Fig. 16.22, 16.28, 16.32). With uplift of the Navy Board and Byam Martin highs, previously deposited sandstones and carbonates were stripped off the

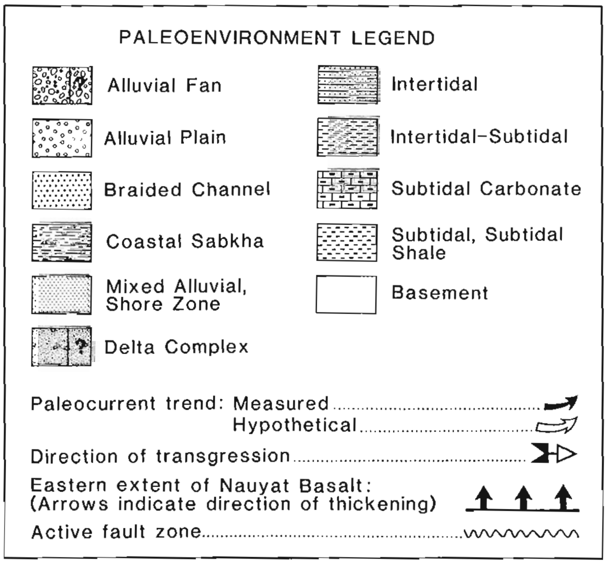


Figure 16.28. *Legend for paleoenvironmental reconstructions in Figures 16.27 and 16.29-16.32.*

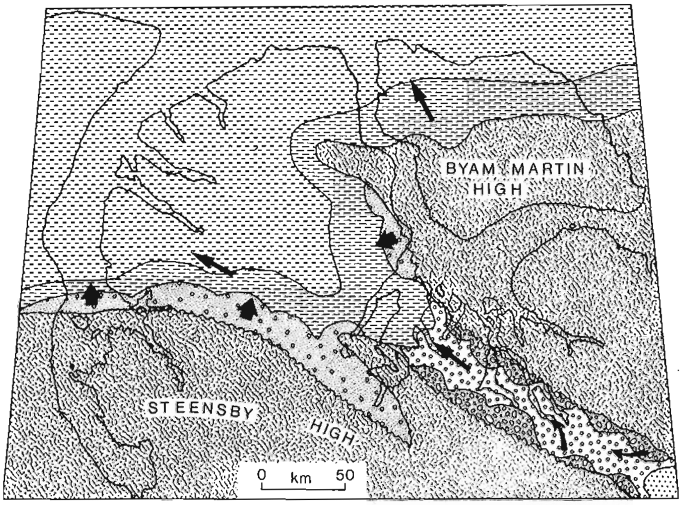


Figure 16.29. *Paleoenvironmental reconstruction for Arctic Bay and laterally equivalent Fabricius Fiord deposition.*

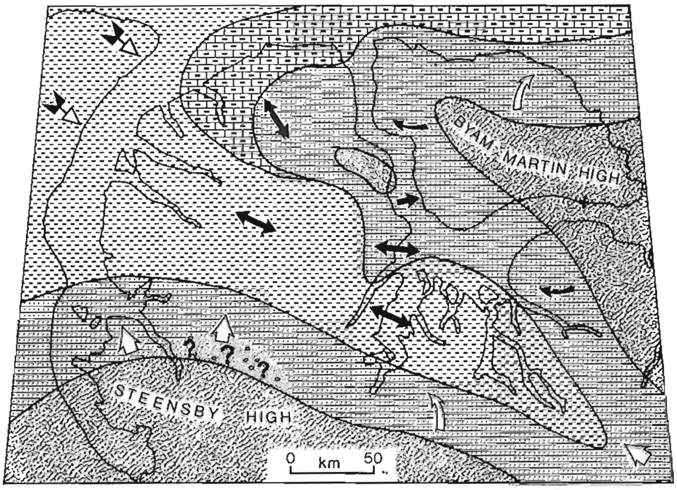


Figure 16.31. *Paleoenvironmental reconstruction for lower Victor Bay deposition.*

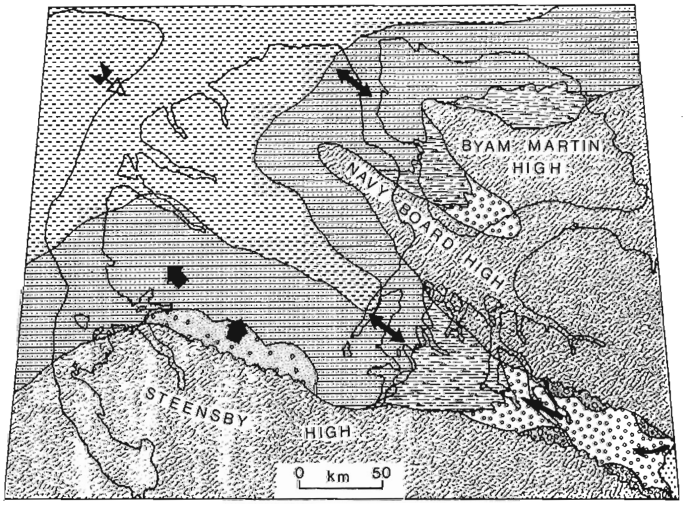


Figure 16.30. *Paleoenvironmental reconstruction for Society Cliffs and uppermost Fabricius Fiord deposition.*

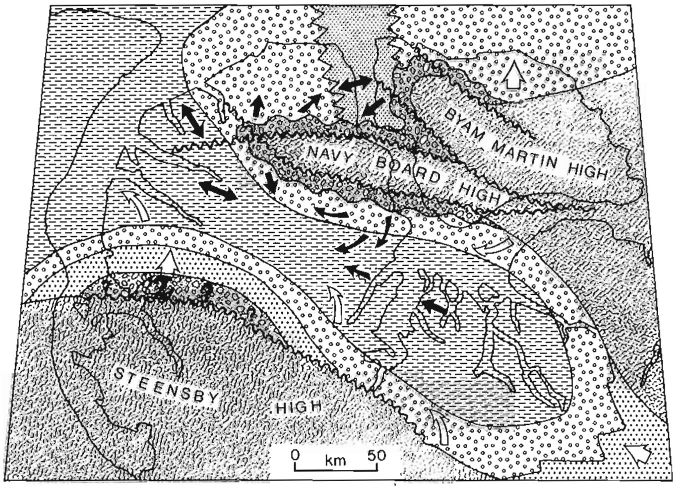


Figure 16.32. *Paleoenvironmental reconstruction for Athole Point and lower Strathcona Sound deposition.*

Navy Board High and a thick sequence of coarsely conglomeratic coalescing alluvial fan complexes (SS_5) were deposited south of the White Bay Fault Zone (Fig. 16.26, 16.28, 16.32) during virtually the entire Strathcona Sound depositional period. At the same time, finer grained, varied sediments (SS_6) were deposited in the north-central part of the basin, and a monotonous sequence of turbidite clastics (SS_4) accumulated in a down-dropped area in the central western part of the basin.

Faulting during lower Strathcona Sound deposition involved northward down-tilting of the Milne Inlet Trough and scissor-type movement between the Trough and Navy Board High. This resulted in deposition of lacustrine or shallow marine limestones in the lower SS_5 member adjacent to the Navy Board High. It also resulted in deposition of the predominantly intertidal-subtidal Athole Point limestones with turbidites and minor interbedded debris flow conglomerates in the southeastern part of the basin. When the deepened portion of the Milne Inlet Trough was eventually filled, the Athole Point strata were buried by prograding upper Strathcona (SS_3) fluvial and shallow marine sands.

In the northern and south-central parts of the Borden Basin, lower Strathcona Sound (SS_1) intertidal and overbank alluvial strata include scattered channel deposits and debris flow breccia and conglomerate lenses. An isolated small subtidal-intertidal biohermal carbonate platform (SS_2) developed rapidly in the west-central part of the basin, east of the SS_4 member, during deposition of lower Strathcona Sound sediments. Syndepositional faulting shed prograding alluvial SS_1 clastics westward, terminating carbonate platform development.

SS_1, SS_2 and Athole Point strata are overlain by upper Strathcona Sound (SS_3) mixed channel and overbank deposits. The coarseness of SS_3 compared with SS_1 strata suggests rejuvenation of the source areas. Uplift of the area south of

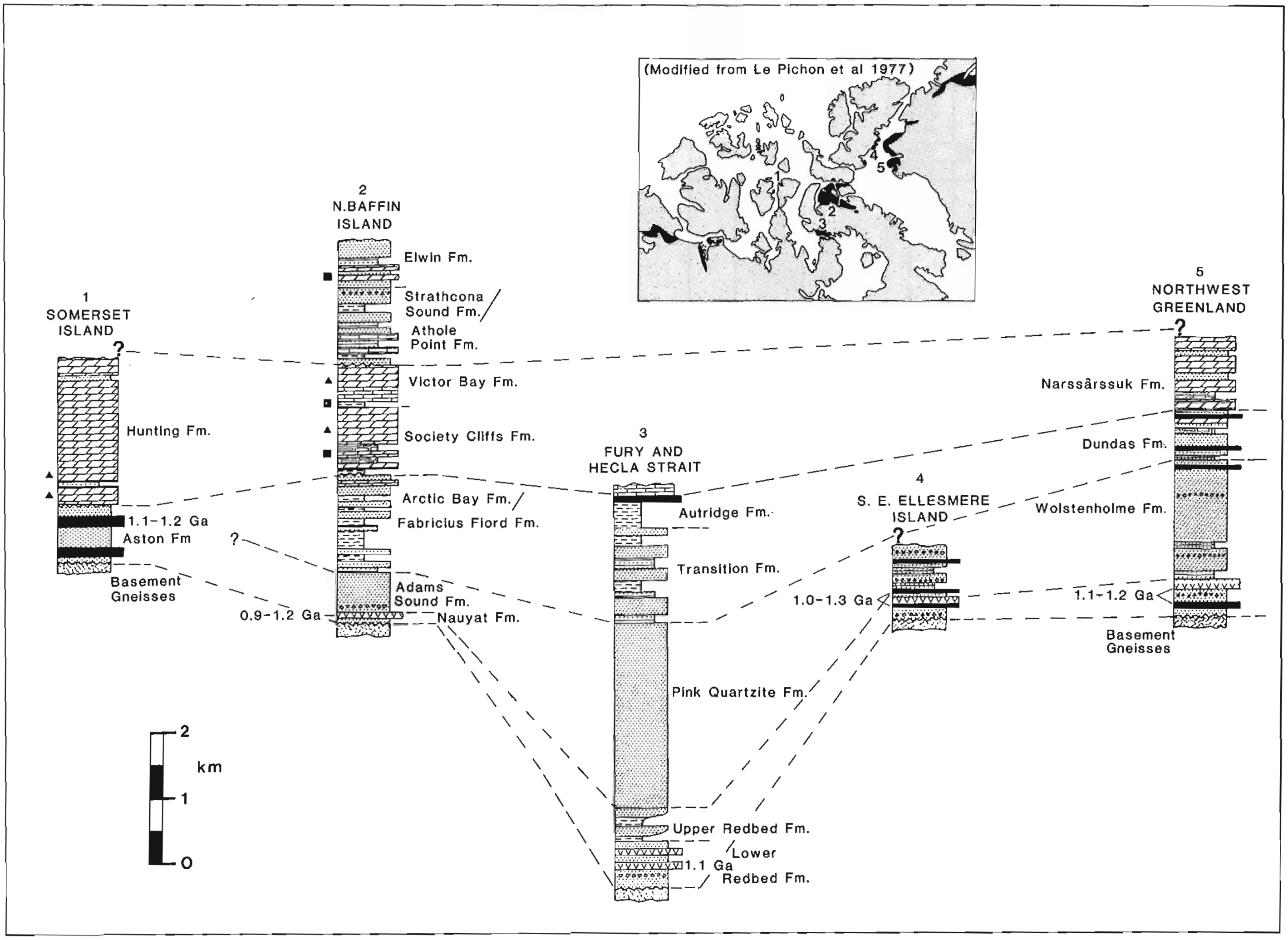


Figure 16.33. Correlation of Neohelikian strata in Somerset Island-Prince of Wales Island area (1), Borden Basin (2), Fury and Hecla Basin (3), and Thule Basin (4,5).

the Central Borden Fault Zone (Steensby High) may have shed fluvial and shallow marine sands north and northeastward across the Milne Inlet Trough. While the eastern Milne Inlet Trough was probably connected to a westerly sea during most of Strathcona Sound deposition, this connection may have been temporarily severed during deposition of upper SS_3.

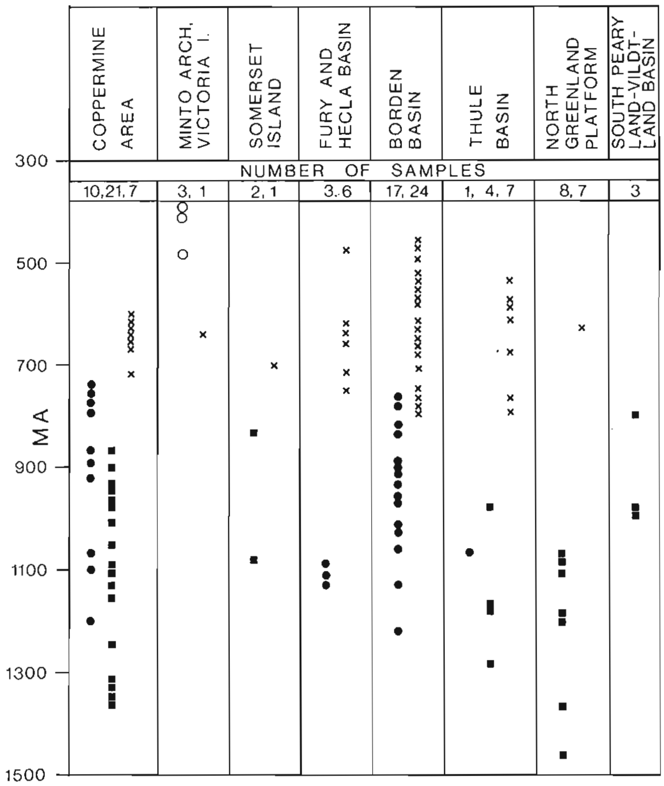
Strathcona Sound strata on southeastern Devon Island (Fig. 16.1) rest unconformably on basement gneisses (G.M. Narbonne, personal communication, 1979). This suggests that this area may have been a source area during deposition of much of the Bylot Supergroup although fault-initiated uplift along Lancaster Sound may also have stripped earlier Bylot strata and redeposited them elsewhere.

Waning source area uplift and minimal subsidence during lower Elwin Formation deposition in intertidal to supratidal environments was superseded by deposition of relatively clean and well-sorted deeper water sands of the upper Elwin.

Correlations

Neohelikian sedimentary sequences are preserved in several basins and areas along the northwest edge of the Canadian-Greenland Shield (Fig. 16.1, 16.33-16.35). Sequences in the Thule (Canada and Greenland) and Fury and Hecla basins and in the vicinity of Somerset Island most closely resemble the Bylot Supergroup (Fig. 16.33).

These sequences lie nonconformably on gneissic basement, contain tholeiitic basalts and/or basic sills about 1.2 Ga in age, yield Mackenzie paleomagnetic poles, and contain similar lithologies arranged in similar stratigraphic sequences and deposited in similar environments; some syn-depositional faulting is common (Fig. 16.1, 16.33-16.35). Regional correlations involving these Neohelikian strata along the northwest edge of the Canadian-Greenlandian Shield have been made by Blackadar (1970), Blackadar and Fraser (1960), Christie et al. (1972), Lemon and Blackadar (1963), Dawes (1976), Fahrig et al. (1981), Kerr (1979), and Young (1979).



Solid circles = Mackenzie (Neohelikian) volcanics
Solid squares = Mackenzie intrusions
Open circles = Franklin (Hadrynian) volcanics
Crosses = Franklin intrusions

Figrue 16.34. *K-Ar ages for some basic igneous rocks along northwest edge of Canadian-Greenlandian Shield. References in text, except for Minto Arch, from Wanless (1970).*

Thule Basin

As much as 4.5 km of Thule Group strata are preserved in the fault-bounded Thule Basin 300 km northeast of the Borden Basin (Davies et al., 1963; Dawes, 1976, 1979; Dawes et al., 1973; Frisch et al., 1978; Frisch and Christie, in press; Vidal and Dawes, 1980). The basal Wolstenholme Formation comprises two to four members, the lowermost of which consists of varicoloured sandstone, basic tholeiitic plateau basalt and basic sills. The igneous rocks yield whole rock K-Ar ages (Fig. 16.34) similar to those obtained for the Mackenzie Nauyat volcanics, and the lowermost Wolstenholme member is correlated with the Nauyat Formation (Fig. 16.33). The overlying Wolstenholme members are composed of locally conglomeratic sandstones and are similar to Adams Sound members. The Wolstenholme Formation thins to the north, south, and east, where the lowermost member may be absent.

Shales and fine sandstones predominate in the Dundas Formation which is gradational with the underlying Wolstenholme Formation. Coarsening-upward cycles occur in the lower part. Gypsum beds are present and dolostones in the upper Dundas thin northward (marginward?). Hadrynian Franklin sills and dykes intrude Dundas strata (Fig. 16.34) and the Dundas is correlated with the Arctic Bay-Fabricius Fiord formations (Fig. 16.33).

The Narssârssuk Formation contains three members, of which the lower two contain gypsum. The contact with the underlying Dundas Formation is not exposed. Lower and upper red members are composed of varicoloured carbonates, shales, siltstones and sandstones that are arranged in cycles (Davies et al., 1963), similar to those in the Society Cliffs Formation. The middle Aorfêrneq dolomite member contains oolites, minor limestone, and breccia. The Narssârssuk is considered to be equivalent to the Society Cliffs and possibly to the Victor Bay formations.

Thule Group strata were deposited in environments, in part semi-restricted, similar to those in which the equivalent Eqalulik and Uluksan Group strata were deposited. Crossbed measurements indicate westward transport (Frisch and Christie, in press), and thinning of strata indicates a northern to eastern and southern shoreline.

Fury and Hecla Basin

About 6 km of sedimentary strata occur in the vicinity of Fury and Hecla Strait 200 km south of Borden Basin. Blackadar (1958, 1963, 1970) differentiated a thick lower Fury and Hecla Formation and an upper Autridge Formation. Chandler (in Chandler et al., 1980) divided the Fury and Hecla Formation into four informal formations: lower redbed formation, upper redbed formation, pink quartzite formation, and transition formation. These Neohelikian strata are intruded by Hadrynian Franklin dykes (Fig. 16.34; Blackadar, 1970; Chandler and Stevens, 1981), although Chandler and Stevens (1981) noted that a slightly older Hadrynian event might also be represented.

The two redbed formations are chiefly red sandstone, siltstone and shale, with minor light-coloured sandstone and stromatolitic dolostone. An orthoconglomerate in the lower redbed formation thins and fines westward. Whole rock K-Ar ages of about 1100 Ma have been obtained for two thin plateau basalt flows in the upper part of the same member (Fig. 16.34; Chandler and Stevens, 1981). Coarsening-upward cycles occur in the upper redbed formation. The two redbed formations are correlated with the Nauyat Formation, and laterally equivalent sandstones in Borden Basin.

The pink quartzite formation thins westward and is correlated with the Adams Sound Formation. The transition formation, composed of varicoloured shales, siltstones and sandstones, and the overlying Autridge Formation, chiefly black shale, are correlated with the Arctic Bay-Fabricius Fiord formations. About 80 m of grey limestone locally overlies Autridge shale (Blackadar, 1958, 1963, 1970; Heywood, personal communication, 1980) and is correlated with the Society Cliffs Formation.

Fury and Hecla Basin strata, like the Thule Group, were deposited in environments similar to those in which correlative Bylot Supergroup strata were deposited. Crossbed measurements indicate northwest transport for the lower redbed formation, western transport for the upper redbed formation, and southerly transport for the pink quartzite and transition formations (Chandler et al., 1980). The strata are displaced by east-west and northwest-southeast faults. However, it is not yet known whether or not syndepositional faulting was important or whether the present distribution of strata reflects the site of a depositional basin (Chandler, personal communication, 1980). It is inferred here from the presence of basal polymictic, possibly fault-related breccia at two localities (Chandler et al., 1980), the great thickness of the succession and the similarity of its setting and depositional history as compared with the fault-controlled Bylot Supergroup, that syndepositional faulting was prominent.

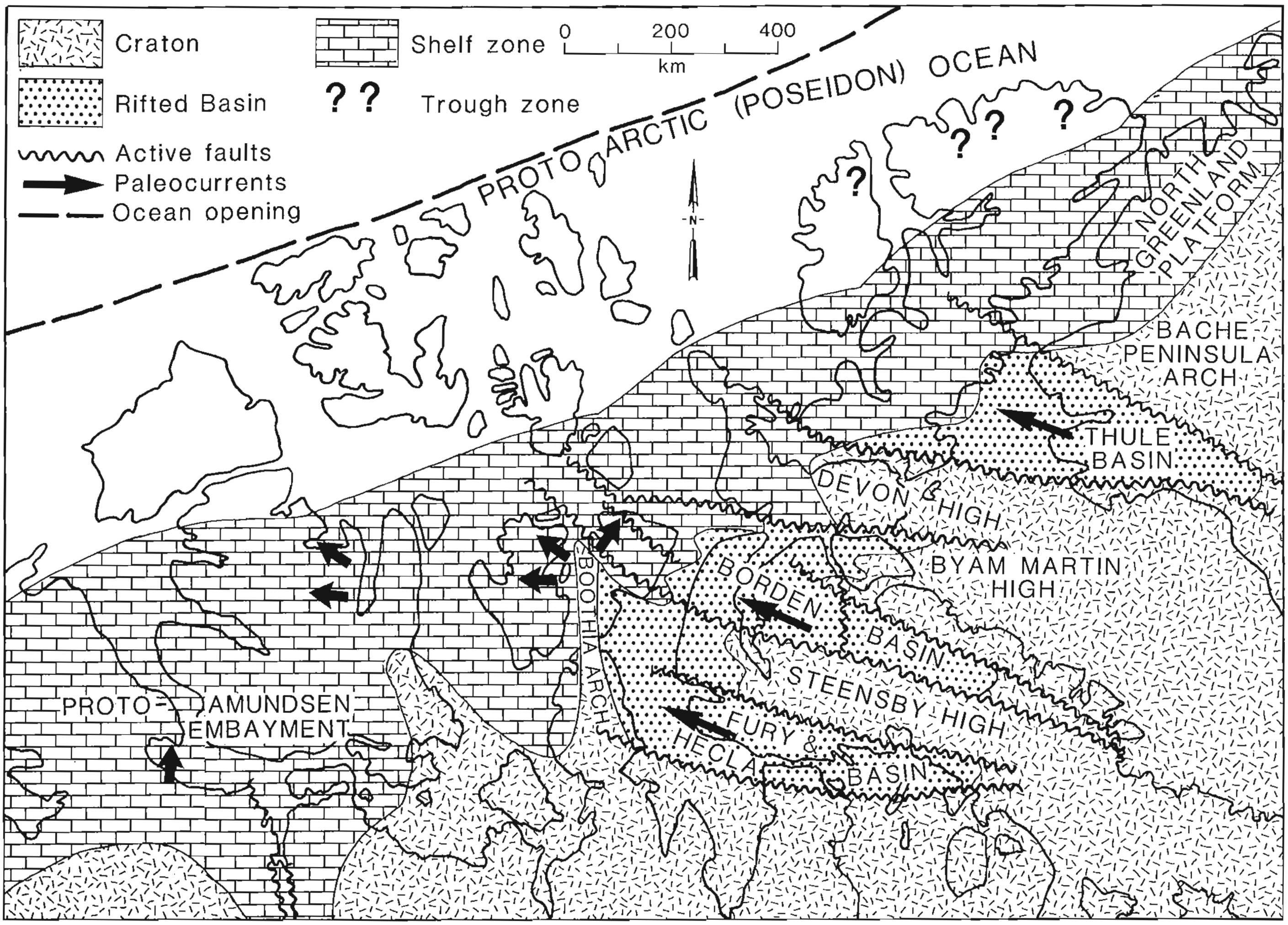


Figure 16.35. *Schematic reconstruction for 1200-1250 Ma old opening of the early Proto-Arctic (Poseidon) Ocean.*

Somerset Island Area

At least 3 km of Neohelikian strata outcrop in the vicinity of western Somerset and eastern Prince of Wales islands (Fig. 16.1, 16.33; Blackadar, 1967; Brown et al., 1969; Dixon et al., 1971; Dixon, 1974; Kerr, 1977a; Kerr and deVries, 1976; Miall, 1969; Reinson et al., 1976; Tuke et al., 1966). The lower Aston Formation is composed of sandstones and minor siltstones and stromatolitic dolo stones. Locally, an angular gneiss-clast cobble-boulder breccia lies directly on basement gneisses. Immature sandstones, with minor shale layers, reappear locally in the upper part of the formation. The Aston Formation both thickens and dips away from the Boothia Arch (Horst) of basement gneisses, and is thickest to the west in the vicinity of eastern Prince of Wales Island.

The Hunting Formation overlies the Aston on Somerset Island and is composed chiefly of several commonly stromatolitic, dolostone lithologies. At some localities a basal conglomerate grades upward through finer clastics into sandy dolostone. Quartz sand content increases upward in the upper part of the formation and interbeds of red sandstone and siltstone occur near the base and top of the formation, while chert occurs chiefly in the middle.

The contact between the Hunting and Aston formations apparently ranges from conformable and transitional to unconformable (Dixon, 1974; Kerr, 1977a; Kerr and deVries, 1976; Tuke et al., 1966). Kerr and Kerr and deVries (op. cit.) concluded that the Aston was eroded from westernmost Somerset Island prior to westward encroachment of the Hunting across the Aston, and its contained intrusions, onto basement gneisses. Aston and Hunting formations were deformed into north-trending folds prior to intrusion of northeast-trending dykes.

Neohelikian Mackenzie paleomagnetic poles have been interpreted for the Aston Formation and for a truncated basic sill in the Aston (Jones and Fahrig, 1978; Fahrig et al., 1981). Hadrynian Franklin paleomagnetic poles have been interpreted for several northeast-trending dykes that intrude both formations and a possible Tertiary pole was obtained for one dyke. K-Ar ages (Fig. 16.34; Dixon, 1974; Jones and Fahrig, 1978) also indicate at least two ages of basic intrusions.

Crossbeds consistently indicate northeast transport on Somerset Island (Tuke et al., 1966) and westward transport in the vicinity of Prince of Wales Island to the west. Some Aston lithologies in the latter area may reflect changes in position of the Boothia Arch (Horst) source area and a shoreward east and southeast transition toward the Arch (Dixon et al., 1971). Emergence of the Boothia Arch during Aston deposition probably involved some faulting.

The Aston and Hunting formations are very similar to, and are correlated with, the Nauyat-Adams Sound Formation and with the Society Cliffs Formation respectively. Strata equivalent to the Arctic Bay and Fabricius Fiord formations, and possibly the upper Adams Sound Formation, were either not deposited or were eroded prior to Hunting deposition. Descriptions of the Aston-Hunting contact suggest the unconformity is locally well developed, but may die out away from the Boothia Horst and so represents a small time interval. Therefore, we presently consider the Hunting to be older than the lithologically similar strata of the Shaler Group on Victoria Island that is commonly considered to be Hadrynian (Thorsteinsson and Tozer, 1962; Young, 1974, 1979). Upper Hunting strata may in part be equivalent to lower Victor Bay strata, although Dixon (1974) has identified an unconformity between the Hunting and overlying lower Paleozoic strata. Dixon (1974) considered black shales that Tuke et al. (1966) placed in the uppermost Hunting to be lower Paleozoic.

The Aston and Hunting formations were deposited in a tectonically active semi-restricted basin in which the Boothia Arch (Horst) was probably a source for Aston sediments. Depositional environments were very similar to those of the correlative formations in Borden Basin.

Origin, Regional Relations

Borden Basin evolved within the North Baffin Rift Zone, in which movement along dominantly vertical northwest-trending faults, probably from the close of Aphebian to Recent time, resulted in formation of several horsts and grabens (Fig. 16.26). Post-Aphebian northwest-trending faults extend throughout Baffin Island (Jackson and Morgan, 1978), and the structures shown in Figure 16.26 probably extend much farther than shown. Alignment of faults, graben structures, Franklin dykes, mineral deposits and topographic features (Blackadar, 1967; Jackson and Morgan, 1978; Jones and Dixon, 1977; Kerr, 1977a, b; Kerr and deVries, 1977; Miall and Kerr, 1977; Reinson et al., 1976) indicates that the Milne Inlet Trough structure probably extends 1200 km from Somerset-Cornwallis islands southeast to the northwest corner of Home Bay on the east coast of Baffin Island (Fig. 16.35). Information presently available, however, indicates that significant Phanerozoic faulting has not affected the Brodeur Homocline west of Admiralty Inlet (Blackadar, 1968c, d; Trettin, 1969).

Borden Basin probably developed along a failed arm or aulacogen, as first suggested by Olson (1977), during a 1.2 Ga ocean opening to the northwest (Fig. 16.35). Doming and crustal extension accompanied by early faulting initiated the aulacogen and basin. Some of the features within the Borden Basin also found in aulacogens (e.g. Hoffman et al., 1974; Freund and Merzer, 1976), but not necessarily restricted to them, include:

1) Faults have developed a zig-zag pattern (Fig. 16.25); individual faults are irregular, steep, and most of the movement has been dip-slip.
2) Faults locally follow pre-existing structures, but grabens, such as the Milne Inlet Trough, do not seem to follow a specific structure, although it is subparallel to the regional Archean-Aphebian gneissosity.
3) Once developed, fault movement has occurred over a long period of time.
4) Central horsts such as Navy Board High, have been elevated but not as high as marginal horsts (Byam Martin) or bordering areas.
5) Relative abundance of granulite facies rocks adjacent to Borden Basin (Jackson and Morgan, 1978) may be a result of early updoming.
6) Paleocurrents are chiefly longitudinal. Northwest transport directions predominated except during Elwin deposition when a reversal resulted in easterly transport predominating.
7) Metamorphism is weak to absent.

Figure 16.35 shows schematically what the relationships may have been between several of the Neohelikian basins along the northwest edge of the Canadian-Greenland Shield during early sedimentation in these basins. The positions of Greenland and Ellesmere Island are from Le Pichon et al. (1977). Similarities in lithology, stratigraphy, inferred ages and depositional environments, and the general indications of a marine basin to the northwest suggest that sedimentation in Fury and Hecla and Thule basins and the Somerset Island area was penecontemporaneous with sedimentation in Borden Basin, and that the other three basins were interconnected with the Borden Basin at least temporarily (Fig. 16.35). The most likely period for interconnection was relatively late in the sedimentary history, during deposition of the platform carbonates (Uluksan Group, Hunting Formation, Narssârssuk Formation, etc.). At this time the basins may have been modified to merely broad embayments along the southeastern edge of an ocean.

Its on-strike nature with the Bylot Supergroup suggests that the Aston-Hunting succession was probably deposited in the western part of the Borden Basin, while the Bylot Supergroup was deposited in the eastern part of the same basin. The southeastern edge of a major ocean may have been only a short distance northwest of Somerset Island area at this time, and Borden Basin may have been at least partially divided by development of a north-south horst or upwarp in the area of Brodeur Peninsula (west of Admiralty Inlet, Fig. 16.2) near the close of Borden Basin sedimentation.

As suggested in Figure 16.35, Fury and Hecla Basin may have been connected with the Borden Basin to the west. During early development of these basins, however, the Steensby High may have extended westward to the west side of Somerset Island, at least temporarily, thus including part of the Boothia Arch. During these times the Fury and Hecla Basin may have been landlocked or connected to an ocean across the southern part of the Boothia Arch. During deposition of the platform carbonates the western part of the Devon High may have been an island to the east of which the Borden and Thule basins were interconnected.

A Rb-Sr isochron age of 1257 ± 45 Ma (Baragar, 1972) has been obtained for plateau basalts of the Copper Creek Formation in the Coppermine region (Fig. 16.1). Mackenzie paleomagnetic poles have been obtained for these volcanics and for coeval intrusions (Fahrig and Jones, 1969; Robertson, 1969). In northeastern Greenland a Rb-Sr isochron age of 1230 ± 25 Ma has been obtained for red granophyric sheets and dykes in mid-Proterozoic sandstones. The granophyres have been interpreted as being genetically related to dolerite sills and basaltic to rhyolitic flows (Jepsen and Kalsbeek, 1979). K-Ar ages (Fig. 16.34) from the Coppermine region (Fahrig and Jones, 1969; Baragar and Robertson, 1973; Robertson and Baragar, 1972), and from eastern Ellesmere Island and northwestern Greenland (Dawes, 1976; Dawes et al., 1973; Frisch and Christie, in press; Frisch, personal communication, 1981; Henrikson and Jepsen, 1970) support the Rb-Sr and paleomagnetic data, and a mid-Proterozoic (Neohelikian) age for some of the sedimentary and volcanic rocks in these regions.

We conclude that several basins developed about 1200-1250 Ma ago and extended for about 3200 km, between the Cordilleran Geosyncline in western North America, and the Carolinidian Geosyncline in eastern Greenland, along what is now the northwest edge of the Canadian-Greenlandic

Shield (Fig. 16.35). Most of these basins were initiated and/or evolved as grabens. The evolution of these basins is consistent with the interpretation that they are related to a 1250 Ma ocean opening to the northwest that formed the Proto-Arctic Ocean of Trettin et al. (1972).

Harland and Gayer (1972) proposed the term Pelagus (Greek god who ruled over part of the ancient ocean) for an ocean that was probably present north of Greenland and Ellesmere Island, at least in early Paleozoic time, in order to differentiate it from the present Arctic Ocean. Trettin and Balkwill (1979) interpreted the eastern Arctic Ocean as being closed between 1000 Ma old and earliest Cambrian. If so, the 1250 Ma ocean is distinctly older. Rather than call it the Proto-Pelagus Ocean, which would be confused with the initial Pelagus opening, the name Poseidon Ocean is proposed here for the Neohelikian ocean which was ancestral to the early paleozoic Arctic ocean called Pelagus. Poseidon, in Greek mythology, was the brother of Zeus, and was the god of water and the father of Pelagus.

This Neohelikian ocean opening and the related strata and structures may be analogous to the early Mesozoic opening of the Atlantic Ocean and the related strata and structures along the Northern American Atlantic coast (Austin et al., 1980; Burke, 1976; Falvey, 1974; Jansa and Wade, 1975; Haworth and Keen, 1979). Structures related to the early Arctic, Poseidon, ocean opening (e.g. Milne Inlet Trough, Boothia Arch) are commonly subperpendicular to one another and oriented at 40-70 degrees to the opening axis, whereas they are chiefly subparallel or subvertical to the axis of the Atlantic opening. This apparent difference may be an indication that a different reconstruction than that of Le Pichon et al. (1977) is required for the Neohelikian, which is not surprising. The Boothia Arch (Horst) and Milne Inlet Trough are known to have been active over a much longer period of time and may be larger than the Mesozoic Atlantic structures, although the late Triassic graben extending from Delaware to Vermont (Sanders, 1963) is about the same size as the Boothia Arch.

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17.

PALEOMAGNETISM OF THE BYLOT BASINS: EVIDENCE FOR MACKENZIE CONTINENTAL TENSIONAL TECTONICS

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Abstract

A number of apparently related Helikian sedimentary and volcanic basins in northern Canada are here referred to collectively as the Bylot basins. These include the Borden and the Fury and Hecla basins of northern Baffin Island, and the Thule Basin of Greenland. The basins developed in response to widespread rifting activity 1220 Ma ago which resulted from tensional tectonics of continental dimensions. Mackenzie magmatism, which took place throughout the western Canadian Shield, and basaltic magmatism in southern Greenland and Fennoscandia, appear to be related to the same tensional event. The precise time correlation of early Borden Basin development with Helikian basaltic magmatism throughout the Canadian Shield and throughout a reconstructed supercontinent is based chiefly on paleomagnetic evidence.

The lowest units of the Borden sequence, Nauyat volcanics and Adams Sound Formation, are normally magnetized and yield poles at 167.5°W, 4.5°N, $\delta m = 13.0°$, $\delta p = 6.5°$ and 166.0°W, 1.5S, $\delta m = 6.0°$, $\delta p = 3.0°$. Their combined pole for 12 sites is at 166.5°W, 1.5°N, $\delta m = 6.0°$, $\delta p = 3.0°$. This is indistinguishable from Mackenzie igneous poles. The Strathcona Sound Formation, which occurs near the top of the Borden sequence, exhibits six reversals in the sequence sampled and has a pole at 155.5°W, 8.0°N, $\delta m = 4.0°$, $\delta p = 2.0°$ which is distinguishable from the Mackenzie pole.

The Aston Formation, which occurs on Somerset and Prince of Wales islands, has a pole position at 178.5°W, 2.0°S, $\delta m = 13.4°$, $\delta p = 6.7°$, which suggests that it is the time equivalent of the lower Borden sequence.

The fill that makes up the Borden Basin was accumulated over a period of 18 Ma, commencing about 1220 million years ago.

Résumé

Dans cette étude, on désigne collectivement par le terme de bassins de Bylot un certain nombre de bassins sédimentaires et volcaniques datant de l'Hélikien, et sans doute apparentés les uns aux autres. Ils comprennent le bassin de Borden et celui de Fury et Hecla dans le nord de la terre de Baffin, et le bassin de Thulé au Groenland. Ces bassins se sont formées il y a 1220 Ma sous l'influence d'une vaste activité tectonique, les fractures résultant de tensions s'exerçant à une échelle continentale. L'épisode magmatique de Mackenzie, qui s'est exercé sur tout l'ouest du Bouclier canadien, et le magmatisme basaltique du sud du Groenland et de la Fennoscandie, ont probablement été engendrés par le même épisode de tensions. Pour établir une corrélation temporelle précise entre le développement initial du bassin de Borden et le magmatisme basaltique de l'Hélikien, sur l'ensemble du Bouclier canadien et le supercontinent original, on s'est principalement basé sur des indices paléomagnétiques.

Les unités les plus anciennes de la succession de Borden, c'est-à-dire les couches volcaniques de Nauyat et la formation d'Adams Sound, sont magnétisées de façon normale et comportent des pôles de coordonnées 167,5°W, 4,5°N, $\delta m = 13,0°$, $\delta p = 6,5°$ et 166,0°W, 1,5S, $\delta m = 6,0°$, $\delta p = 3,0°$. En combinant les pôles de 12 sites, on obtient les coordonnées suivantes: 166,5°W, 1,5°N, $\delta m = 6,0°$, $\delta p = 3,0°$. Ces pôles se confondent parfaitement avec les pôles d'activité volcanique de Mackenzie. La formation de Strathcona Sound, que l'on rencontre près du sommet de la succession de Borden, présente 6 inversions de polarité dans la succession échantillonnée, et possède un pôle de coordonnées 155,5°W, 8,0°N, $\delta m = 4,0°$, $\delta p = 2,0°$, qui se confond parfaitement avec le pôle de Mackenzie.

La formation d'Aston, que l'on rencontre sur les îles de Somerset et du Prince de Galles, se caractérise par un pôle de coordonnées 178,5°W, 2,0°S, $\delta m = 13,4°$, $\delta p = 6,7°$ ce qui suggère qu'elle est l'équivalent stratigraphique de la partie inférieure de la succession de Borden.

Le matériel sédimentaire qui constitue le bassin de Borden s'est accumulé pendant une période de 18 Ma qui a débuté il y a environ 1220 Ma.

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INTRODUCTION

The stratigraphy and development of the Borden sequence is dealt with in detail elsewhere in this volume so only the briefest description of Borden stratigraphy is given here to provide a setting for the paleomagnetic data and for the synthesis of these data. The Borden sequence was divided into two groups (Blackadar 1963, 1970; Lemon and Blackadar, 1963). The lower group, the Eqalulik, comprises two units, the Nauyat, largely volcanic, which is conformably overlain by the Adams Sound, consisting of buff to purple quartzite. The Adams Sound is thought to be largely fluvial and grades upward (Blackadar, 1970) into the lowest unit of the Uluksan Group. The Uluksan consists of several formations and is largely shallow marine. Jackson et al. (1980) have indicated unconformities at two levels within the Uluksan, but these are unlikely to represent major lapses in time.

Most of the material collected for paleomagnetic study was obtained from three formations: the Nauyat and Adams Sound (Eqalulik Group) and the Strathcona Sound (Uluksan Group). The Strathcona Sound is the second highest unit of the Uluksan Group. In the lower Strathcona sampled section it consists chiefly of medium to dark chocolate-red siltstone and fine sandstone. The poles for the three formations may be used to indicate the direction and amount of polar wandering that took place during almost the entire time involved in filling the Borden Basin.

NAUYAT FORMATION

The Nauyat Formation consists largely of thick basaltic flows, which in the major area of exposure south of Adams Sound, extend over an area of at least 300 km^2 (Fig. 17.1). Six flows were sampled in the section indicated. These consist of four flows above and two below a thin pink quartzite unit. The flows are all at least 10 m thick and some are several tens of metres thick. A seventh flow probably occurs below the sampled section. The weathered surfaces of the flows vary considerably in colour, being rust, brick red, dark red, purple or grey-green. Individual flows have a number of distinguishing features which were described by Galley (1978) from a section a few kilometres to the southwest. The flows vary considerably in composition even within the same flow, but are chiefly subalkaline tholeiites. Alkalis total 5 per cent, and silica about 52 per cent (Galley, 1978).

Sixty cores were drilled from seven Nauyat sites (31 to 37), with site 31 representing the highest exposed flow. The collection includes 55 cores from the 6 flows mentioned above and 5 cores from a pink quartzite layer (site 35).

Two specimens were cut from each core and the 110 specimens from the six volcanic sites were A.F. demagnetized in five or six steps. With the exception of site 33, specimen directions shifted systematically to a fairly flat westerly direction during the step cleaning. At fields of 500 or 600 oe the directions within the 5 sites are highly

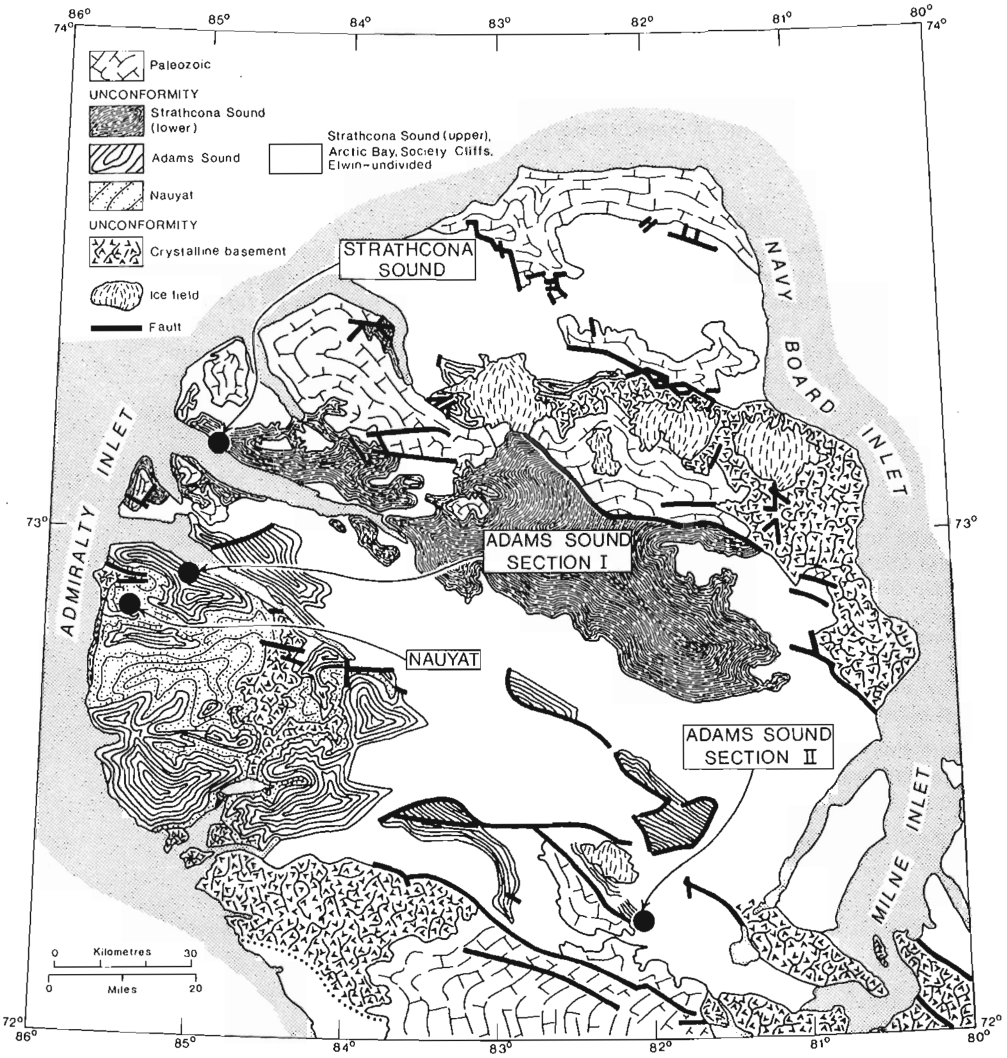

Figure 17.1

Geological map (simplified after Blackadar et al., 1968) showing location of sections sampled for paleomagnetic work. Only the basement gneisses, Paleozoic cover and the three formations sampled for paleomagnetic work, have been patterned.

dispersed. Specimen directions for site 33 showed the same westerly shift during step cleaning but were streaked towards the present earth's field direction. For this reason, data from this site were not included in group statistics for the formation (Table 17.1[1]). Figure 17.2 shows characteristic A.F. demagnetization curves for the six volcanic sites.

The pink quartzite (site 35) samples were thermally demagnetized in steps to 680°C. Sample directions were highly dispersed above 600°C and below this temperature retained a steeply down direction not far from that of the present earth's field. Figure 17.3 shows the demagnetization curves for one specimen of each core. These specimens were very weakly magnetic and it is unlikely that a primary magnetization was isolated during the thermal cleaning.

ADAMS SOUND FORMATION

The Adams Sound Formation was sampled at two localities (Fig. 17.1). At Section I, 36 cores were drilled at 7 sites through a 54 m very gently north-dipping sequence.

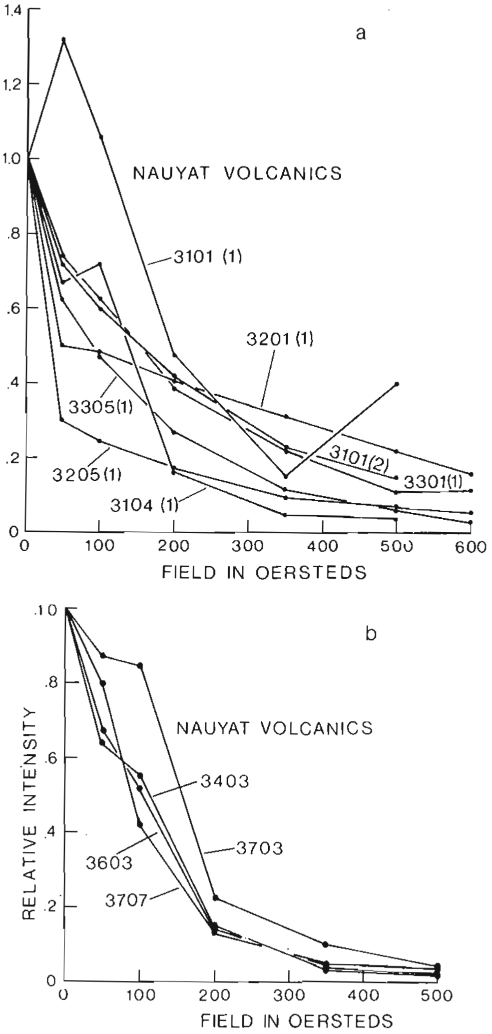

Figure 17.2. Alternating field demagnetization curves for Nauyat volcanics.

Site 38 is the highest site and site 44 the lowest. The sandstones here vary from buff to dark purple and drilling was concentrated in purple layers because it seemed likely that these would yield measurable cores.

One specimen from each core was thermally step cleaned and Figure 17.4 shows the response of these specimens to cleaning. The N.R.M. directions of the cores were well away from the earth's field direction (Table 17.2) and thermal demagnetization at 645°C caused a shift to a flatter westerly direction. The angles between the magnetization directions of specimens cut from the three cores of site 44 were all greater than 25° both before and after cleaning. Results for this site are not included in the formation mean even though their mean cleaned direction is very close to that of the remainder of the group.

A second set of cores (site 11) was collected at Adams Sound Section II (Fig. 17.1) about 130 km southeast of Section I. This collection consists of 17 cores drilled through a 15 m section. These cores were collected west of a 43 m thick diabase dyke. The core nearest the dyke was 76 m distant and the farthest core was about 300 m.

All of the specimens of this site were thermally cleaned in great detail. Their N.R.M. directions were steeply down and it was only as a result of cleaning at high temperatures that the majority of cores yielded the flat westerly directions characteristic of the cores from Section I (Fig. 17.5, Table 17.2). The difference in behaviour between these cores and those of Section I is ascribed to the proximity of Section II samples to the diabase dyke, which has a steeply down magnetization.

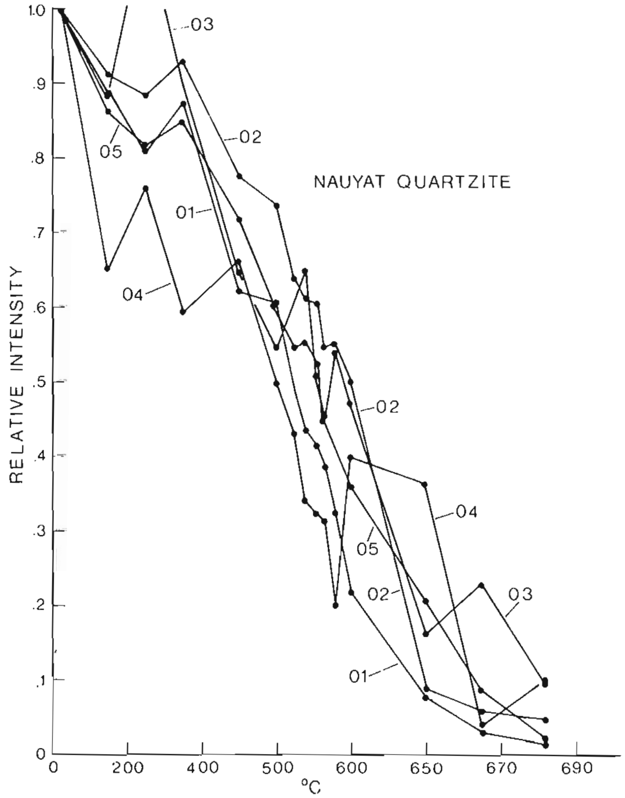

Figure 17.3. Thermal demagnetization curves for Nauyat quartzites.

[1]Tables for this paper appear after "Discussion"

The combined pole position for the Nauyat and Adams Sound formations (12 sites) is not statistically different from that for Mackenzie igneous rocks (Fahrig and Jones, 1969). The Adams Sound samples are all normally magnetized.

STRATHCONA SOUND FORMATION

Sixty-six samples were collected from a 245 m section in the Strathcona Sound Formation (Fig. 17.1). About 60 m of the section are repeated by faulting so the sampled stratigraphic interval is about 180 m. Site 5 was stratigraphically the highest and site 28 the lowest (Fig. 17.6). Sixty of the samples are from the lower Strathcona (Blackadar 1970) and the remaining six are from the upper Strathcona. The upper Strathcona at this locality is grey-green silty sandstone with some thin dark maroon layers. Site 5 consisted of this material and the three samples from this site proved unstable. Site 15, on the other hand, though similar in appearance, gave stable results. The lower Strathcona at this locality consists of rather friable red shales with layers of more massive dirty-red to chocolate-red silty sandstone. These more massive layers were sampled for paleomagnetic study.

Six clear polarity reversals were encountered in descending the section (Table 17.3, Fig. 17.6). The highest occurs several metres below the top of the lower Strathcona. This reversal is repeated by faulting (Fig. 17.6) and is referred to as the 'Strathcona Reversal'. Samples from the lower Strathcona were subjected to both alternating field and thermal demagnetization. The N.R.M. directions were well away from the present field direction and during cleaning their directions shifted only slightly. This shift took place in fields up to 400 oe and temperatures up to 400°C. Above these fields and temperatures no discernible changes in direction were detected (Fig. 17.7, 17.8). The site mean directions given in Table 17.3 are those obtained after cleaning at 600°C and 700 oersteds.

The magnetization direction of the Strathcona Sound is shown with that of the combined Adams Sound and Nauyat formations on Figure 17.10.

ASTON FORMATION

The Aston Formation was first mapped during "Operation Franklin" (Blackadar in Fortier et al., 1963). This is an unfossiliferous Proterozoic assemblage outcropping south of Aston Bay, Somerset Island. It consists predominantly of sandstone with minor shale and is overlain by the Hunting Formation, also Proterozoic, which is predominantly dolomite with minor shale and conglomerate. More detailed stratigraphic study of these formations has been carried out by Tuke et al. (1966), Dixon (1974) and Kerr (1977). Kerr indicated that the Hunting lies unconformably on the Aston.

Christie et al. (1971) correlated the basal sedimentary unit on the east coast of Prince of Wales Island with the Aston (Fig. 17.10, 17.11). In this area the unit consists predominantly of sandstone with conglomerate, dolomite and stromatolitic dolomite. It has been suggested that the depositional environment of the Aston ranged from supratidal to shallow subtidal. Paleomagnetic and K-Ar work on diabase from Savage Point suggested that the Aston is at least 1220 Ma old (Jones and Fahrig, 1978).

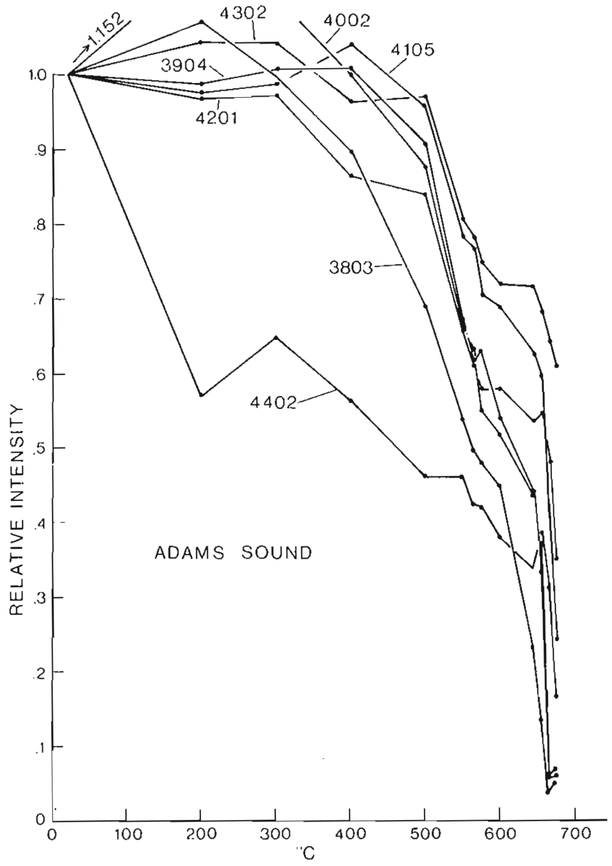

Figure 17.4. Thermal demagnetization curves for specimens of Adams Sound Formation from Section I (Fig. 17.1).

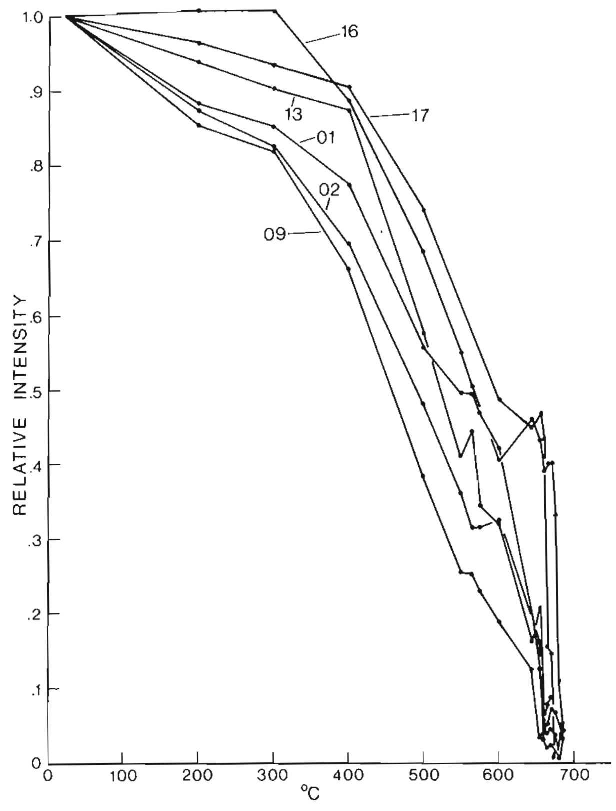

Figure 17.5. Thermal demagnetization curves for specimens of Adams Sound Formation from Section II (Fig. 17.1).

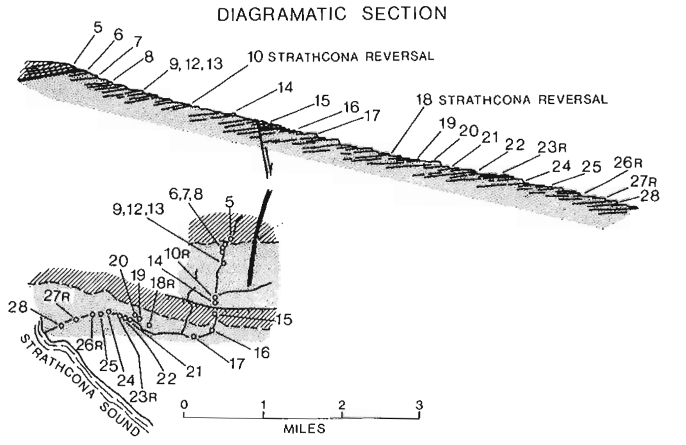

Figure 17.6. Below – plan view showing geology of lower Strathcona Sound Formation collecting site. Lower Strathcona stippled, upper Strathcona cross-hatched. "R" following site number indicates reversed polarity. Above – diagrammatic section indicating repetition of the Strathcona reversal by normal faulting.

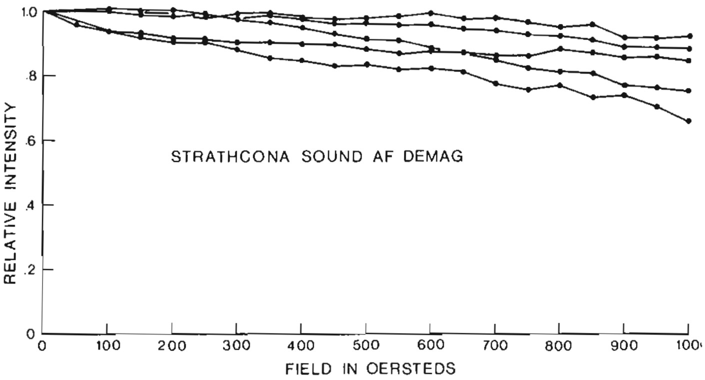

Figure 17.7. Alternating field demagnetization curves for lower Strathcona Sound Formation samples.

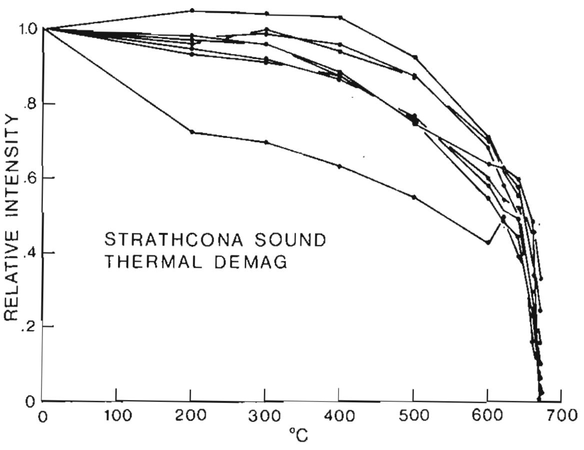

Figure 17.8. Thermal demagnetization curves for lower Strathcona Sound Formation samples.

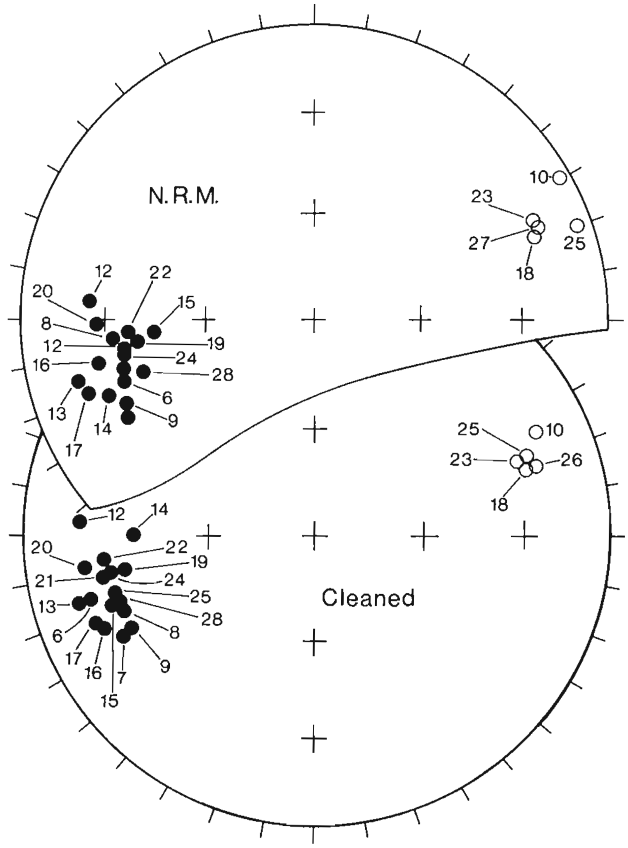

Figure 17.9. Equal area plot of site mean directions of Strathcona Sound before (upper) and after (lower) cleaning at 700 oersteds or 600°C.

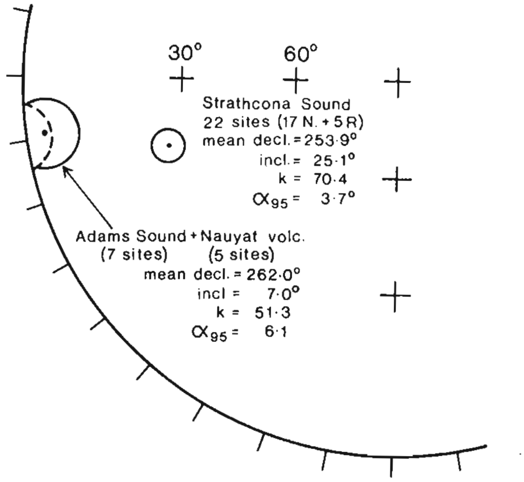

Figure 17.10. Stereographic (Wulff) plot of mean magnetization direction of the lower Borden sequence (Nauyat volcanics and Adams Sound Formation combined) and of the lower Strathcona Sound Formation.

Material for paleomagnetic work was collected from twelve sites near Aston Bay (Somerset Island) and Savage and Whitehead points (Prince of Wales Island). Results from only the five Savage Point sites are reported here (Table 17.4) because work was not completed on the remaining seven sites. Enough has been done, however, to indicate that at least some of the uncompleted sites will not give reliable paleomagnetic results.

Sites 3 and 5 consisted of well layered maroon silty sandstone, site 6 of buff to maroon feldspathic sandstone, and sites 12 and 13 of purple or maroon feldspathic sandstone.

DISCUSSION

The continent-wide nature of late Helikian Mackenzie igneous events was first established by Fahrig and Jones (1979) and was based chiefly on paleomagnetic evidence for the correlation of basic igneous rocks, particularly of diabase dykes, which suggested a remarkable period of North American crustal tension. The contemporaneity of the development of the Bylot rift basins with the Mackenzie events, as established in the present paper, extends evidence for this tension into the Arctic Islands (Fig. 17.13). In addition, Patchett et al. (1978) have pointed out that basic intrusive rocks in southern Greenland and Fennoscandia are very similar in age to the Mackenzie igneous rocks and have used the paleomagnetism of these rocks to reconstruct a North Atlantic supercontinent. Their reconstructed continent, to which the Bylot basins have been added, is shown in Figure 17.14.

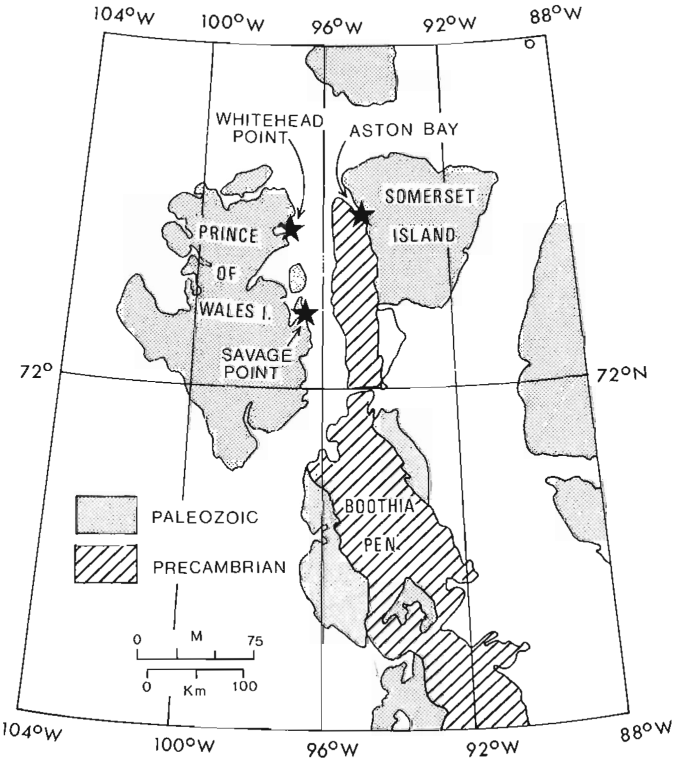

Figure 17.11. Index map showing the location of Aston Formation.

The correlation of Bylot basins with Mackenzie rocks has interesting implications regarding the Mackenzie Magnetostratigraphic Interval, which is a chronostratigraphic unit of the Helikian defined by Fahrig et al. (1971). At that time the upper and lower limits of the Interval could not be defined as virtually all the supporting paleomagnetic data were derived from intrusive rocks. The upper boundary of the Mackenzie Interval is now defined as the first reversal within the Borden sequence. At present the earliest known reversal is that in the lower Strathcona Sound Formation, so this is defined as the upper limit of the Mackenzie Magnetic Interval unless or until a reversal is discovered lower in the sequence.

As our knowledge of the importance of the Mackenzie tensional episode expands, support is provided for the suggestion by Douglas (1980) that the Mackenzie Magnetostratigraphic Interval be used as a fundamental marker in subdividing Helikian time.

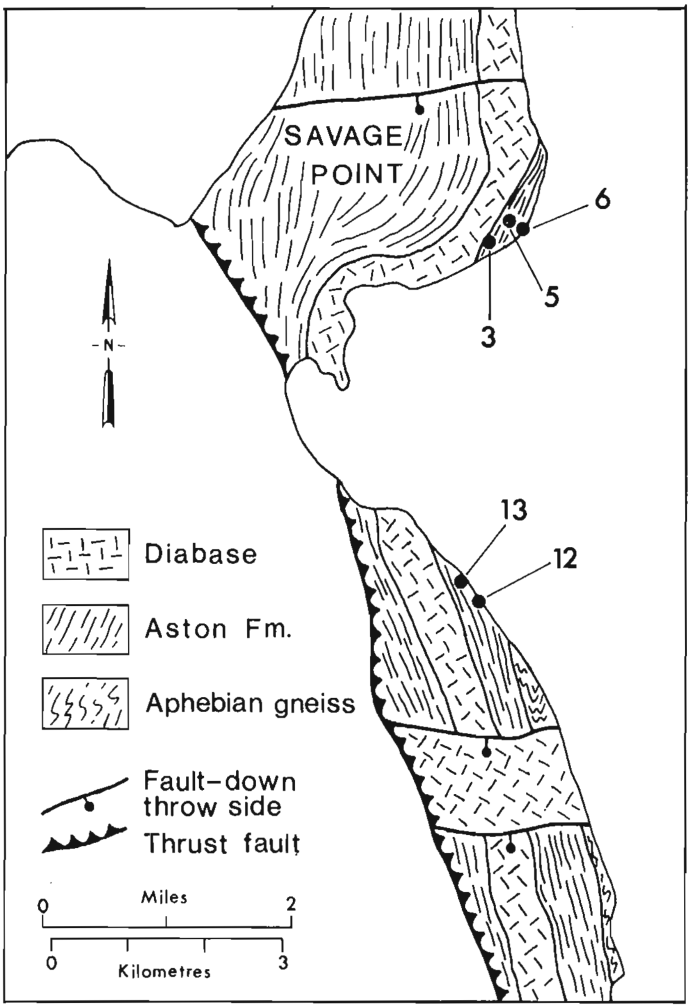

Figure 17.12. Local geology of the Savage Point area (after Christie et al., 1971) showing collecting sites of Aston Formation.

It should be possible from paleomagnetic data to estimate the time involved in filling the Borden Basin. The Mackenzie pole is the best dated of all North American Precambrian poles and the Logan dyke pole which forms a second anchor on the Logan Loop (Robertson and Fahrig, 1971) is fairly well dated at 1100 Ma. The intervening Logan Loop is the best defined part of the Precambrian North American polar wandering curve (Fig. 17.15). If it is assumed that the polar wandering rate between Nauyat volcanism and lower Strathcona sedimentation was the same as the average rate between 1220 and 1100 Ma, Nauyat to Strathcona time was 16.5 Ma. If 1.5 Ma are ascribed to upper Strathcona and Elwin sedimentation then the filling of the Borden Basin took 18 Ma.

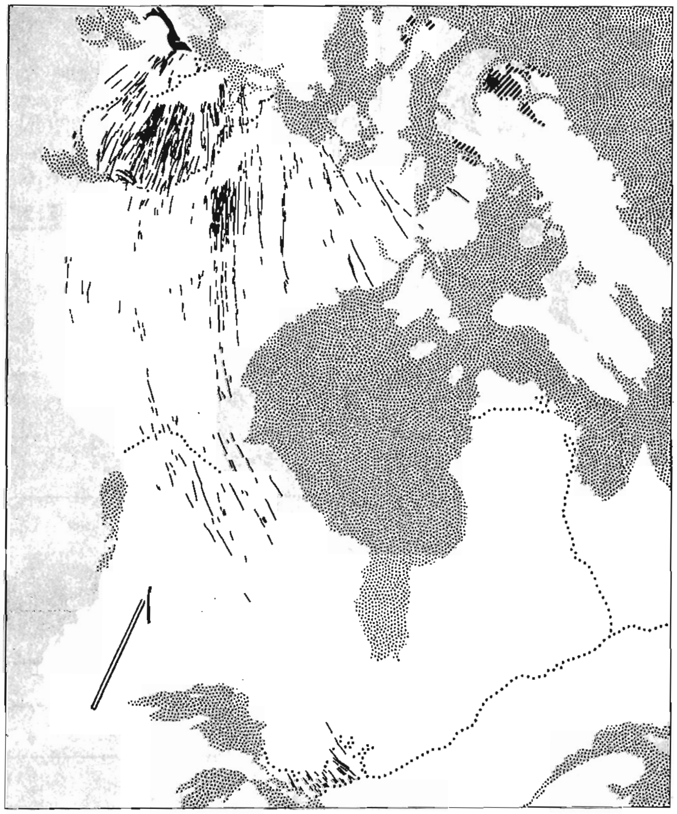

Figure 17.13. *Mackenzie igneous rocks; Coppermine volcanics, Muskox Complex, East Arm sills, Mackenzie dykes, Sudbury dykes and the Bylot basins (shown cross-hatched with the Nauyat volcanics in solid black; fine stippled areas are Paleozoic).*

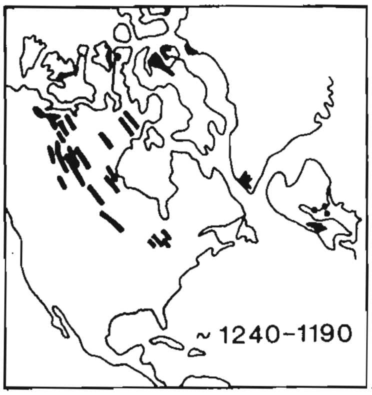

Figure 17.14. *Mackenzie supercontinent (Patchett et al., 1978) with the Bylot basins added to the Mackenzie igneous rocks.*

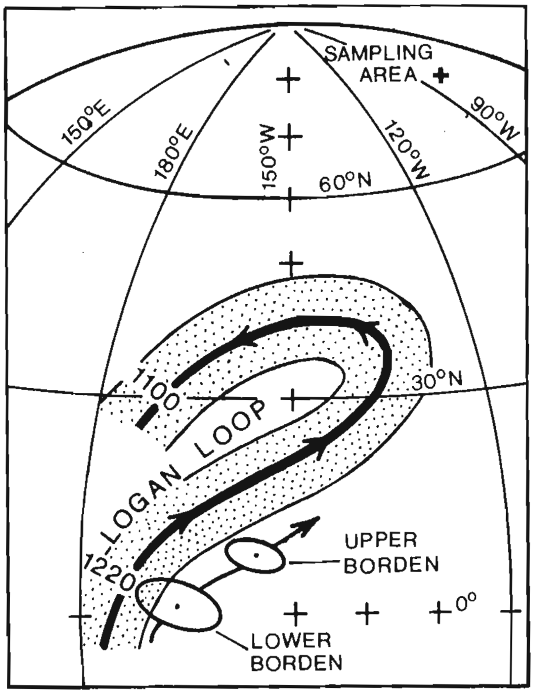

Figure 17.15. *Pole positions with ellipses of confidence for lower and upper Borden sequence.*

Table 17.1. Paleomagnetic data for Nauyat Volcanics

	NRM						CLEANED				POLE FROM MEAN DIRECTION			
SITE	N	n	D	I	k	FIELD	D	I	k	α	LONG.	LAT.	δm	δp
31	10	6	247.4	42.6	31.1	200 oe	261.0	22.4	45.3	10.1				
32	10	9	263.3	55.0	14.5	500 oe	273.4	0.4	50.5	7.3				
33*	9	9	264.6	66.1	99.4	200 oe	264.7	42.7	40.6	8.2				
34	9	9	240.3	50.6	26.1	200 oe	258.9	27.6	67.5	6.3				
36	10	10	254.7	27.6	32.4	200 oe	263.4	12.3	63.9	6.1				
37	7	7	256.8	17.7	93.8	350 oe	262.1	1.9	94.0	6.3				
SITES	6	5	252.7	39.0	23.1		263.9	13.0	37.2	12.7	167.7W	4.5N	13.0	6.6
SAMPLES	55	41	252.9	39.2	14.6		264.2	12.8	27.3	4.3	168.0W	4.5N	4.4	2.3

*Site 33 omitted from means and pole positions

N = number of samples collected at site
n = number of samples per site after rejecting unstable samples
D = declination (degrees)
I = inclination (degrees), negative numbers indicate north-seeking directions above the horizontal
k = Fisher's (1953) best estimate of precision
α = half angle (degrees) of the cone of confidence at P = 0.95
δm, δp = semi-axes of ellipse of confidence at P = 0.95

Table 17.2. Paleomagnetic data for Adams Sound Formation

	NRM						CLEANED				POLE FROM MEAN DIRECTION			
SITE	N	n	D	I	k	TEMP.	D	I	k	α	LONG.	LAT.	δm	δp
38	6	4	250.9	11.8	25.7	645°C	247.5	2.6	19.1	21.6	SECTION I			
39	5	5	255.0	9.1	51.6	645°C	252.9	-0.6	24.2	15.9				
40	5	5	264.2	17.6	47.5	645°C	262.2	6.8	91.3	8.1				
41	5	6	262.3	8.6	72.7	645°C	262.3	1.3	50.8	9.5				
42	6	4	261.7	15.7	222.9	645°C	262.8	5.2	181.2	6.8				
43	5	5	269.6	7.4	14.3	645°C	267.7	0.5	29.7	6.7				
44	3	0	-	-	-		-	-	-	-				
11	17	12	265.4	42.7	20.7	*645°C 675°C	264.2	5.3	25.6	8.7	SECTION II			
SITES	8	7	261.1	16.1	35.1		260.7	2.7	105.2	5.9	165.8W	1.4S	5.9	3.0
SAMPLES	53	41	262.0	20.4	15.4		261.7	3.1	32.7	4.0	166.7W	1.0S	4.9	2.0

* Optimum temperatures used for this site varied with sample:
1 at 645°, 3 at 655°, 3 at 660°, 4 at 665°, 1 at 675°.

Symbols as in Table 17.1.

Table 17.3. Paleomagnetic data for Strathcona Sound Formation

	NRM				CLEANED					POLE FROM MEAN DIRECTION			
SITE	N	D	I	k	D	I	k	α	POLARITY	LONG.	LAT.	δm	δp
6	1	251.9	32.8		254.5	20.2			N				
7	3	241.9	30.0	69.1	242.3	27.7	64.0	15.5	N				
8	4	264.2	32.5	67.5	248.8	29.8	29.1	17.3	N				
9	2	244.7	30.9		242.6	29.1			N				
10	2	59.5	-3.2		64.6	-18.5			R				
12	3	274.3	24.4	933.8	272.5	20.2	1546.6	3.1	N				
13	3	254.2	19.0	35.5	254.2	16.1	31.4	22.4	N				
14	1	248.5	26.7		269.8	37.5			N				
15	3	264.8	44.3	12.4	251.0	17.6	23.9	25.8	N				
16	3	257.3	26.3	10.2	246.5	22.1	17.3	30.6	N				
17	3	251.1	10.9	230.7	248.5	10.1	265.5	7.6	N				
18	4	69.2	-20.2	687.1	71.4	-26.4	464.0	4.3	R				
19	3	263.2	39.3	249.1	260.0	33.3	215.9	8.4	N				
20	2	268.9	28.1		262.3	20.8			N				
21	4	260.4	34.8	191.5	259.1	27.5	143.7	7.7	N				
22	4	264.9	38.0	171.7	263.5	27.6	326.7	5.1	N				
23	4	66.3	-18.9	75.1	69.7	-28.0	102.6	9.1	R				
24	3	258.8	34.8	212.6	259.6	28.3	134.8	10.7	N				
25	3	254.6	34.0	238.7	254.0	28.7	136.5	10.6	N				
26	3	70.2	-4.3	850.1	72.3	-22.9	240.7	8.0	R				
27	3	66.9	-18.3	3660.1	69.0	-25.1	915.8	4.1	R				
28	2	252.2	39.7		250.9	29.3			N				
N SITES	17	257.3	31.3	53.9	255.3	25.3	59.8	4.7		156.9W	8.5N	5.0	2.7
N SAMPLES	47	258.4	31.5	32.8	255.1	25.1	39.1	3.3		156.8W	8.4N	3.5	1.9
R SITES	5	66.4	-13.0	74.0	69.3	-24.2	311.8	4.3		151.6W	6.3N	4.6	2.5
R SAMPLES	16	67.0	-14.4	75.1	69.8	-24.9	172.2	2.8		151.8W	6.9N	3.0	1.6
ALL SITES	22	254.5	27.2	34.9	253.9	25.1	70.4	3.7		155.6W	8.0N	4.0	2.2
ALL SAMPLES	63	255.2	27.2	26.8	253.7	25.1	47.5	2.6		155.5W	8.0N	2.8	1.5

Polarity N = normal
R = reversed

Other symbols as in Table 17.1

Table 17.4. Paleomagnetic data for Aston Formation

	NRM						CLEANED				POLE FROM MEAN DIRECTION			
SITE	N	n	D	I	k	TEMP.	D	I	k	α	LONG.	LAT.	δm	δp
03	5	5	262.4	5.5	145.2	500°C	261.6	0.9	170.8	5.9				
05	10	9	261.4	6.4	132.1	500°C	259.2	1.2	122.9	4.7				
06	8	8	278.3	6.4	14.0	550°C	280.5	-4.5	18.6	13.2				
12	6	5	244.9	18.8	66.0	550°C	242.4	5.5	68.0	9.3				
13	8	8	268.9	15.0	20.0	550°C	264.1	1.3	11.6	18.5				
SITES	5	5	263.3	10.6	37.0		261.6	0.9	33.8	13.4	178.3°W	2.1°S	13.4	6.7
SAMPLES	37	35	264.7	10.2	20.6		263.0	0.5	18.6	5.9	179.7°W	1.9°S	5.9	2.9

Symbols as in Table 17.1

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18.

THE APHEBIAN RAMAH GROUP, NORTHERN LABRADOR

I. Knight[1] and W.C. Morgan[2]

Knight, I. and Morgan, W.C., The Aphebian Ramah Group, Northern Labrador; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 313-330, 1981.

Abstract

The Ramah Group is a 1700 m thick outlier of lower Aphebian sedimentary strata resting unconformably on peneplained Nain Province Archean basement rocks at the Nain-Churchill Structural Province boundary. The group consists of two contrasting sequences reflecting a change from shallow siliciclastic shelf to deep basinal deposition. Shelf sedimentation followed transgression of the basement and well washed to muddy clastics were deposited in sand-dominated shallow shelf and shoreline environments interrupted by some muddy shelf deposition and minor subaerial volcanism. The shelf sands were covered by a northward advancing deltaic complex, capped by shallow carbonate deposits reflecting a hiatus. Rapid subsidence followed and black, pyritic muds, sulphide iron formation, chert, volcanogenic muds and thin silty turbidites were deposited in an initially fetid, deep quiet basin. A submarine fan of thinly stratified argillaceous carbonate muds, megabreccias and sands prograded west into the basin from a shelf located outside the area. Renewed black mud deposition gradually succeeded the carbonates and culminated with sandy turbidites rich in fresh basic volcanic detritus. Diabase sills, locally layered with ultramafic cores, intruded the strata prior to the Hudsonian Orogeny, which deformed the group into a doubly plunging, north-trending synclinorium in which deformation and metamorphism, of greenschist to amphibolite rank, increase west and south. Amphibolite and granulite facies Archean basement west of the group was reworked during the Hudsonian and thrust east over the Nain Archean craton. An extensive cataclastic zone west of the craton is a continuation of the Nagssugtoqidian boundary in Greenland.

Résumé

Le groupe de Ramah est un témoin de 1 700 m d'épaisseur de roches sédimentaires de l'Aphébien inférieur reposant en discordance sur le soubassement archéen pénéplané de la province de Nain, à la limite entre les provinces structurales de Nain et de Churchill. Le groupe est constitué de deux successions contrastées, indiquant un passage d'une sédimentation de plate-forme peu profonde, avec dépôt de roches silico-clastiques, à une sédimentation profonde de bassin. La sédimentation de plate-forme a suivi la transgression marine sur le soubassement; des sédiments clastiques bien délavés à boueux se sont alors déposés sur une plate-forme peu profonde généralement sablonneuse et sur le littoral; ces faciès ont été interrompus par un épisode de sédimentation boueuse, et un bref épisode de volcanisme subaérien. Les sables de plate-forme ont été recouverts par un complexe deltaïque progressant vers le nord, lequel a été recouvert par des dépôts carbonatés peu profonds, qui indiquent une lacune stratigraphique. Il y a ensuite eu subsidence rapide, et des boues noires pyriteuses, des dépôts de sulfure de fer, des cherts, des boues volcanogéniques et de minces turbidites silteuses se sont déposés dans un profond bassin aux eaux tranquilles, initialement fétides. Un cône alluvial sous-marin composé de minces stratifications boueuses à matériaux carbonatés et argileux, de mégabrèches et de sables a progressé vers l'ouest à l'intérieur de bassin, à partir d'une plate-forme extérieure à la zone. De nouveau, la sédimentation de boues noires a succédé aux dépôts de carbonate, et a culminé avec le dépôt de turbidites sableuses, riches en débris volcaniques basiques frais. Des sills de diabase, contenant localement en leur centre des stratifications ultramafiques, ont pénétré à l'intérieur des strates avant l'orogénèse de l'Hudsonien; celle-ci a déformé le groupe, qu'elle a transformé en un synclinorium à plongement double, orienté vers le nord, dans lequel les effets de la déformation et du métamorphisme, allant du faciès roches vertes au faciès amphibolite, augmentent vers l'ouest et le sud. A l'ouest du groupe, le soubassement archéen, métamorphisé dans le faciès amphibolite et granulite, a été remanié pendant l'Hudsonien, et rejeté vers l'est par-dessus le craton archéen de Nain. Une vaste zone cataclastique, située à l'ouest du craton, est le prolongement de la limite de Nagssugtoqid au Groenland.

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INTRODUCTION

The Ramah Group, consisting chiefly of sedimentary rocks, forms a linear north-trending fold belt in northern Labrador (Fig. 18.1) that separates Archean basement rocks of Nain Structural Province from reworked Archean basement of Churchill Province to the west (Morgan, 1975). Although the group extends discontinuously about 112 km north from the Hebron Fiord area, it has been mapped in detail only between Saglek and Nachvak fiords (Morgan, 1979; in press). In this area, the group is preserved in a warped doubly-plunging synclinorium (Fig. 18.2) commonly about 8 km wide, with a maximum width of 16 km south of Little Ramah Bay. The eastern margin of the Ramah fold belt is a gently inclined unconformity on the Archean basement. This contrasts with the strongly folded western margin, bounded by west dipping thrusts and by faults, which is in contact with a mobile zone of reworked intensely mylonitized, amphibolite and granulite facies Archean basement rocks.

The group is considered Aphebian on the basis of a Rb-Sr whole-rock age of 1892 Ma (Morgan, 1978). This metamorphic age, derived from a tholeiitic basalt near the base of the Ramah, is interpreted to be related to the Hudsonian Orogeny. Despite metamorphism ranging from greenschist to amphibolite facies, and local intense deformation, sedimentary features are commonly well preserved in the Ramah Group.

The group consists of approximately 1700 m of chiefly sedimentary rocks (Morgan, 1972; Knight, 1973) which have been formally subdivided into six formations (Knight and Morgan, 1977). Volcanic rocks form only a minor component

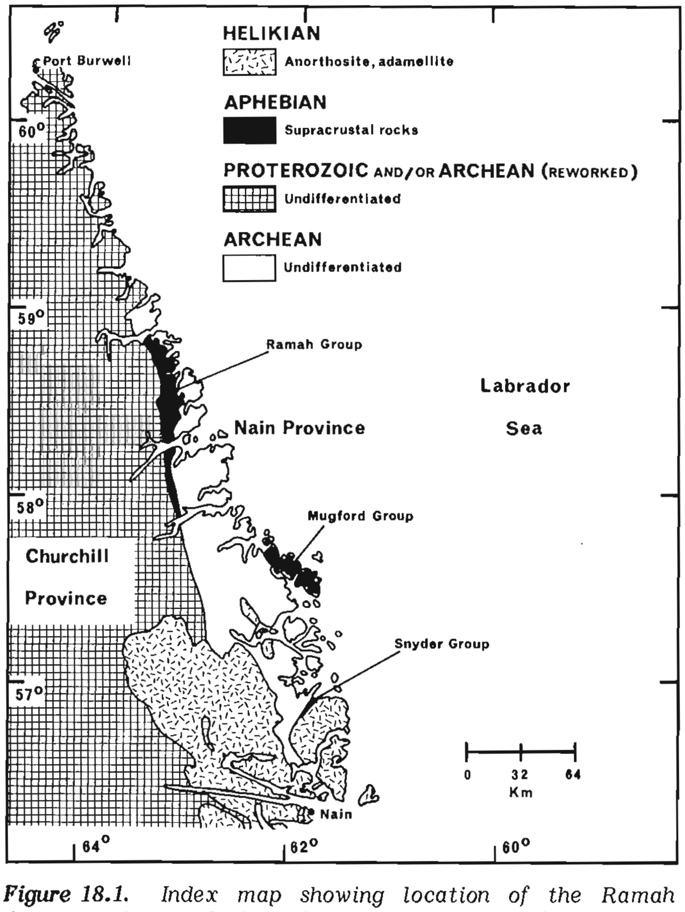


Figure 18.1. Index map showing location of the Ramah Group and correlative Aphebian supracrustal rocks in northern Labrador.

in the lower sequence, although numerous transgressive diabase sills, locally with ultramafic cores, intrude the entire group.

The formations (Fig. 18.3 and Table of Formations) form two contrasting sequences: a lower sequence of shelf sediments, comprising the Rowsell Harbour and Reddick Bight formations, predominantly quartzite and sandstone capped by an important dolomite marker horizon, and; an upper sequence comprising the remaining four formations of basinal fine grained sediments, predominantly argillaceous and carbonate. Thickness and facies of some of the shelf sediments change across Reddick Arch, an intrabasinal tectonic high, located near Reddick Bight, that separated a northern and southern basin. The upper sequence is dominated by the thick Nullataktok Formation composed of thinly stratified shale, mudstone and fine carbonate with a few coarser units of breccia and sandstone. This passes upward into the Warspite Formation; a mixed assemblage of carbonate breccia, minor sandstone and thinly stratified carbonate and siliciclastic mudstone. The Warspite is in turn overlain by black shale of the Typhoon Peak Formation. The sequence is capped by the turbidite sandstones and shales of the Cameron Brook Formation.

THE RAMAH UNCONFORMITY AND REGOLITH

The unconformity at the base of the Ramah Group, first described by Coleman (1921), is magnificently exposed in the mountains along the eastern margin of the group, particularly where long narrow ridges, overlain by gently west dipping Ramah strata project east for several kilometres from the main Ramah outcrop, as in the area between Little Ramah Bay and Bear's Gut (Fig. 18.2). The unconformity sharply truncates the underlying subvertical Archean banded gneisses and granitoid rocks (Fig. 18.4). It forms an extensive peneplain, the exhumed surface of which is probably represented by the present westward-dipping mountain slopes marginal to the eastern contact of the group (Fig. 18.5). Small outliers of basal Ramah quartzite cap the summits of several mountains east of the main fold belt (Morgan, 1979).

Corrections for dip from a number of localities between Little Ramah Bay and Rowsell Harbour indicate that the unconformity dipped at about 4 or 5 degrees to the northwest. Variations in strike, however, suggest the existence of a slightly rolling topography, with only local deviation from a planar geometry elsewhere along the contact.

Irregularities in the erosion surface are, however, present at several localities, particularly south of Bear's Gut. There, quartzite filled, southwest-trending channels, up to 13 m deep and 6 m wide, cut through a thick regolith and into unaltered basement gneiss; these can be traced for up to 70 m. The erosion surface nearby, defined by rounded to blocky pillars of gneiss, has an undulating relief of approximately 3 m. Vase-shaped hollows are incised into these pillars, which in the depressions, are onlapped and draped by crossbedded pebbly quartzite. Beneath the unconformity in the area south of Bear's Gut, linear fractures up to 16 cm wide parallel and crosscut relict gneissic banding in a thick regolith. Some fractures expand downward into enlarged cavities up to 40 cm high and 218 cm long filled by coarse- to medium-grained pink sandstone with faint red mud drapes, suggestive of geopetal fill. Large neptunean dykes filled by weathered and brecciated gneiss near Saglek Head at the mouth of Saglek Fiord (K.D. Collerson, personal communication, 1977) are interpreted as pre-Ramah.

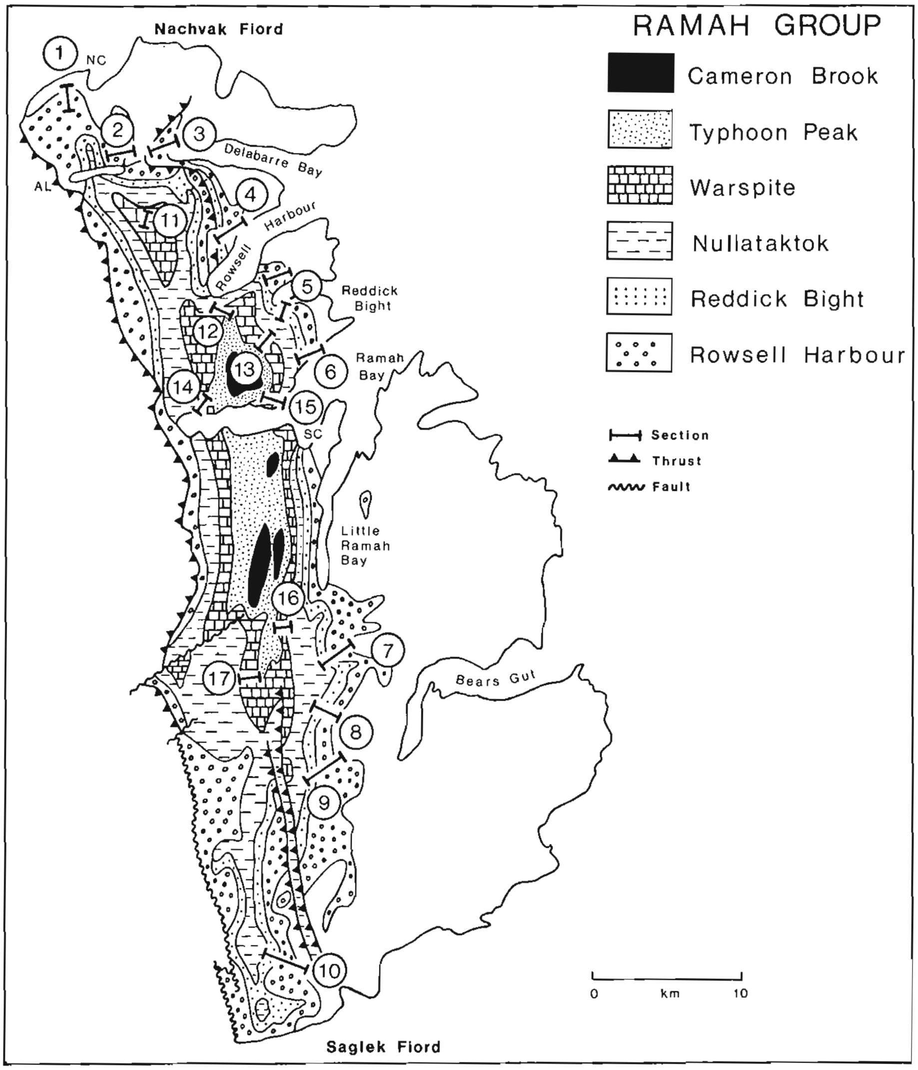


Figure 18.2

Distribution of formations within the Ramah Group in the Saglek Fiord-Nachvak Fiord area. Locations of measured sections of Rowsell Harbour and Reddick Bight formations (1 to 10; see Fig. 18.8) and of Warspite Formation (11 to 17; see Fig. 18.11) are indicated.

AL - Adams Lake
NC - Naksaluk Cove
SC - Schooner Cove

The preserved thickness of the pre-Ramah regolith at the basal contact of the group varies considerably. It averages 1-8 m, with a maximum of 12 m south of Bear's Gut, and is generally more poorly developed north of Reddick Bight. As the degree of alteration decreases downwards, it is difficult in places to define where the regolith passes into unweathered gneiss (Fig. 18.6). Bleaching, reddening and green coloration characterize the friable altered gneiss. Quartz is reddened and feldspars are altered to creamy green or grey clays. Both magnetite and hematite are locally concentrated as irregular spherical bands in the gneiss. Red jasper concretions, associated with mafic gneiss, occur within either the top 10 cm of the regolith or at the contact, and are particularly common in the Ramah and Little Ramah Bay areas. Dolomite zones produced by carbonatization of mafic and ultramafic rocks are common.

Pre-Ramah diabase dykes of an east trending swarm are deeply weathered within the regolith, and their original crystalline texture is obliterated. Spheroidal weathering of the basement gneiss is sporadically preserved in the regolith with the best examples at the south end of Little Ramah Bay and on the ridge north of Reddick Bight (Fig. 18.6, 18.7). Spheroids of less weathered gneiss 10-20 cm in diameter are surrounded by rusty weathered, decomposed gneiss 2-3 m below the unconformity. A weathered margin to the spheroids at depth gives way to onion shell peeling near the unconformity surface.

Interpretation and Depositional Environment of the Regolith

Preservation of different thicknesses of regolith, characterized by total decomposition of basement gneiss to clay minerals with local spheroidal weathering, suggests deep weathering under humid, probably subtropical or tropical climatic conditions. Mobilization and removal of iron was incomplete and was locally fixed as nodules, Liesegang-like rings and crystals of magnetite in the gneiss and diabase beneath the unconformity. This suggests the presence of some free oxygen in the atmosphere at that time.

The deep weathering probably resulted in the relatively easy erosion of the regolith to enhance an extensive peneplain during initial Ramah transgression. Where the regolith is almost completely removed by erosion, resistant granitic plugs apparently formed local rounded highs above the gently undulose peneplain surface. South of Bear's Gut abundant channels, depressions, pillars and sand-filled fractures suggest a rugged rocky shoreline. Neptunean dykes near Saglek Head, 30 km southeast of the main Ramah Group outcrop, suggest that the regolith and unconformity extended southeast close to the present-day land surface.

SHELF SEDIMENTATION

The regolith is overlain by sand-dominated clastics of the Rowsell Harbour and Reddick Bight formations. Stratigraphic sections were measured along the well-exposed eastern margin of the synclinorium (Fig. 18.2, 18.8), but not in the more highly deformed western limb, although the units are present. A ubiquitous dolomite caps this shelf succession.

Rowsell Harbour Formation

Rowsell Harbour Formation consists of up to 470 m of varicoloured sedimentary quartzite, phyllite and shale, with a single thin volcanic flow. Five members are present in the formation (Fig. 18.8 see Table of Formations) and all but the volcanic unit can be traced throughout the synclinorium.

Lower White Quartzite Member

This member consists chiefly of quartzite, overlying a granite wash, and local conglomerate. The member is capped by rusty weathering, dolomitic, pebbly sandstone and conglomerate. The basal and uppermost strata are subarkose

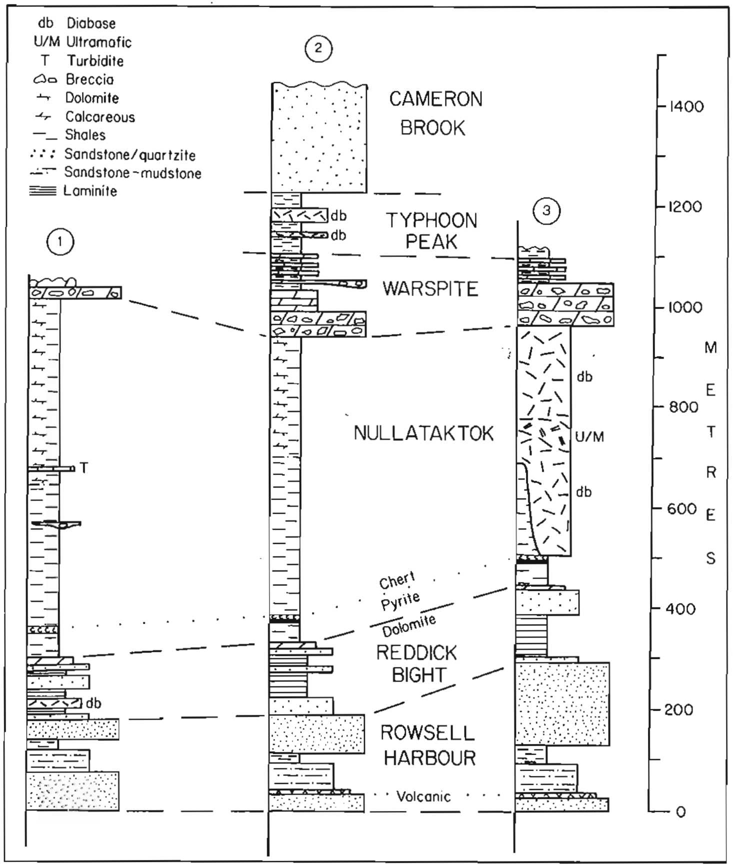


Figure 18.3. Stratigraphic sections of the Ramah Group at 1 – Delabarre Bay; 2 – Ramah Bay; 3 – Little Ramah Bay.

and arkose rich in argillaceous matrix, fresh microcline and plagioclase that is decomposed to clay minerals in some of the granite wash. Argillaceous quartz arenite occurs above these basal strata, with from 1-20 per cent matrix, and carbonate cement increasing with decreasing matrix.

The granite wash and conglomerate unit above the regolith is rarely greater than 6 m thick. It is crudely bedded to locally crossbedded and consists of poorly sorted, argillaceous arkose. Pebbles are 1-4 cm, dominantly quartz, with some granite, pink feldspar, jasper and banded ironstone. This basal unit is overlain by a continuous 10-20 m thick sequence of moderately to poorly sorted, coarse, white quartzite with abundant heavy mineral laminae (Fig. 18.9). This sequence rests directly on basement north of Reddick Bight. The heavy mineral laminae are 90 per cent magnetite with some amphiboles. They delineate small scale (4 cm thick) festoon trough cross stratification, with the opaques concentrated along the basal scour and lower half of the foresets. Granular and pebbly layers are common near the base, but the sequence gradually fines upwards and becomes better sorted. The bulk of the member is composed of 19-96 m of yellow weathering, white quartzite which grades from coarse to fine upward. Pebbles and opaque grains are rare. The quartzite is dominantly trough crossbedded with some parallel lamination. Large areas of straight to sinuous, symmetrical and asymmetrical ripple marks are common, and are locally crosscut by runoff channels. Locally mudcracked mud drapes fill the troughs.

South of Nachvak Fiord, the quartzite is striped and dolomitic. The dolomitic cement occurs near the top of the quartzite in 9-45 cm layers separated by 8-75 cm dolomite-free quartzite. Gradational boundaries to the dolomite layers and their irregular distribution in the lower parts of the crossbedded quartzite suggest the dolomite is secondary.

Capping the member is a rusty weathering marker bed of dolomitic pebbly sandstone and conglomerate. It is chiefly less than 60 cm thick and consists of poorly sorted, very coarse to medium sandstone overlain by 4-6 cm of conglomerate or pebbly sandstone. Pebbles are similar in size

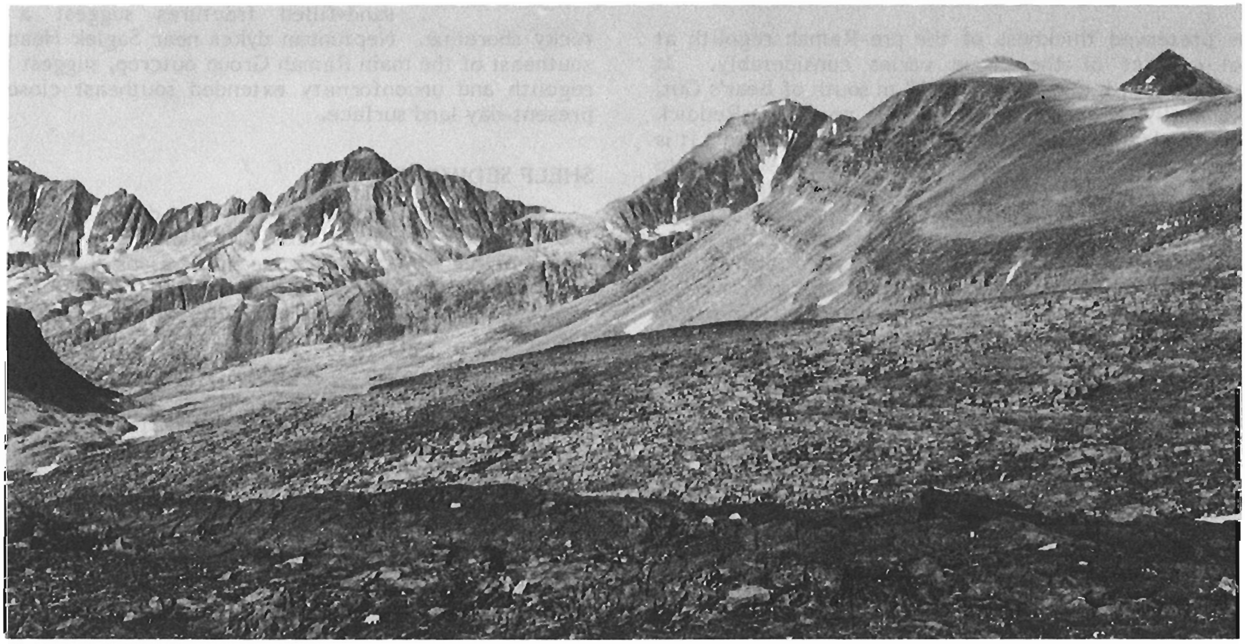


Figure 18.4. *View northeast from south of Little Ramah Bay showing high angle angular unconformity separating rocks of the Ramah Group from underlying banded Archean gneisses. GSC 203678-A*

Figure 18.5

View north across Ramah Bay showing inclined mountain tops, underlain by Archean rocks, that are located east of the Ramah Group. The inclined surface represents the exhumed pre-Ramah peneplain. GSC 203678-B

Figure 18.6

Unconformity (upper arrow) at base of the Ramah Group on the north side of Reddick Bight, showing bedded quartzite overlying a cleaved regolith. The cleavage is replaced by a coarse fracturing as the regolith passes down into less strongly weathered granitoid gneiss. A spheroid of unweathered gneiss (lower arrow) is preserved in regolithic gneiss.

and composition to those in the basal conglomerates. The facies is trough crossbedded and has opaque minerals concentrated along scours.

The only variation in this member is a 2.24 m sequence of laminated sandstone and mudstone that is restricted to the west side of the valley south of Naksaluk Cove on Nachvak Fiord. This unit consists of 13-23 cm of thinly stratified fine grained sandstone and mudcracked mudstone alternating with 15-88 cm of laminated fine, grey sandstone. The top of the sequence is cut out by a 50 cm deep channel filled by a structureless sandstone which upwards encloses large (up to 25 cm) and small clasts of thinly stratified sandstone and mudstone.

Volcanic Member

A 6-9 m, highly altered grey tholeiitic basalt flow (Morgan, 1978) rests upon the rusty weathering, pebbly sandstones of the lower white quartzite member, and forms the only volcanic unit in the group (Fig. 18.8). It is present from near the south shore of Rowsell Harbour to west of Bear's Gut. Locally, north of Reddick Bight, a brown to black, iron-rich crust on the pebbly sandstone is the only evidence of the flow. At Bear's Gut the volcanic unit is intermittently developed.

Thin hematitic, altered dykes and minor intrusions of similar basaltic volcanic rock cut basement gneisses east of the main outcrops of the Ramah Group northwest of Bear's Gut and north of Saglek Fiord (Morgan, in press). Two thin dykes also cut Rowsell Harbour Formation quartzites southwest of Bear's Gut, but appear to be at a higher horizon than the stratigraphic position of the volcanic flow.

The altered fine grained flow is composed of carbonate, sericitized plagioclase, tremolite, chlorite, muscovite, quartz, and opaques, with few microscopic textures preserved. Iron staining is common in the reddened flow top.

TABLE OF FORMATIONS

EON	ERA	GROUP	FORMATION		THICKNESS	LITHOLOGY
PROTEROZOIC	APHEBIAN					Diabase, sills; transgress bedding; locally differentiated with ultramafic core.
		Intrusive Contact				
		RAMAH GROUP Maximum measured thickness 1702 m	Cameron Brook Formation		200 m +	Greywacke – sandstones and mudstones.
			Typhoon Peak Formation		85-130 m	Slates; some sandstone and limestone.
			Warspite Formation		165 m	Dolomitic breccias and sandstones; dololutites; some limestone and calcareous mudstone; argillite, mudstone.
			Nullataktok Formation		595 m	Varicoloured mudstones and shales; graphitic pyritiferous; chert; pyrrhotite-pyrite unit; calcareous and dolomitic mudstones.
			Reddick Bight Formation		53-143 m	Black quartzite; grey muddy sandstone; sandstone-siltstone laminites; yellow weathering dolomite unit.
			Rowsell Harbour Formation 251-470 m	Upper White Quartzite Member	46-267 m	White quartzite; some conglomerate; interbedded shale and mudstone.
				Phyllite Member	15-44 m	Laminated purple mudstone and very fine grained sandstone.
				Purple Quartzite and Mudstone Member	75-157 m	Pink quartzite; alternating units of purple quartzite and purple mudstone; grey sandstone and shale.
				Volcanic Member	0-9 m	Altered tholeiitic basalt flow.
				Lower White Quartzite Member	31-97 m	Granitic wash; pebble conglomerate; white quartzite with heavy mineral laminae; white quartzite; pebbly, coarse grained sandstone.
		Unconformity				
					0-12 m	Regolith
						Diabase dykes; eastward trending swarm.
		Intrusive Contact				
ARCHEAN						Granitic gneiss; migmatite; granulite; mafic gneiss; metasediments.

Figure 18.7

Detail of Figure 18.6 (lower arrow) showing spheroid in regolithic gneiss.

In the thickest sections the flow is massive, with scattered chlorite-filled amygdules. Tear to oval shaped vesicle pipes (Waters, 1960) are common at Ramah Bay and are locally bent or inclined towards the west. At Little Ramah Bay, the flow contains flattened, spherical structures, similar to pahoehoe toes, but the southernmost outcrops are generally highly fractured and rubbly, with carbonate and other secondary minerals such as chalcedony and jasper filling the voids. Secondary weathering of the flow has produced spherical patterns that resemble, and have been interpreted as, pillows (Douglas, 1953). The flow, however, represents a subaerial extrusion, and volcanic clasts occur locally in the overlying ripple marked, mudcracked quartzite.

Purple Quartzite and Mudstone Member

This 75-157 m thick member (Fig. 18.8) has the most diverse assemblage of lithologies in the formation, but contains intercalated units of typical pink to purple quartzite and grey to purple grey mudstone in both northern and southern basins. North of Reddick Arch, however, the typical purple quartzite and mudstone overlie a succession of coarse, phyllitic sandstones with some units of ferruginous, laminated sandstone at the base. These basal sediments form a feather edge near Rowsell Harbour but thicken to 98 m north of the arch (Fig. 18.8). South of the arch pink quartzite occurs beneath the typical purple quartzite and mudstone.

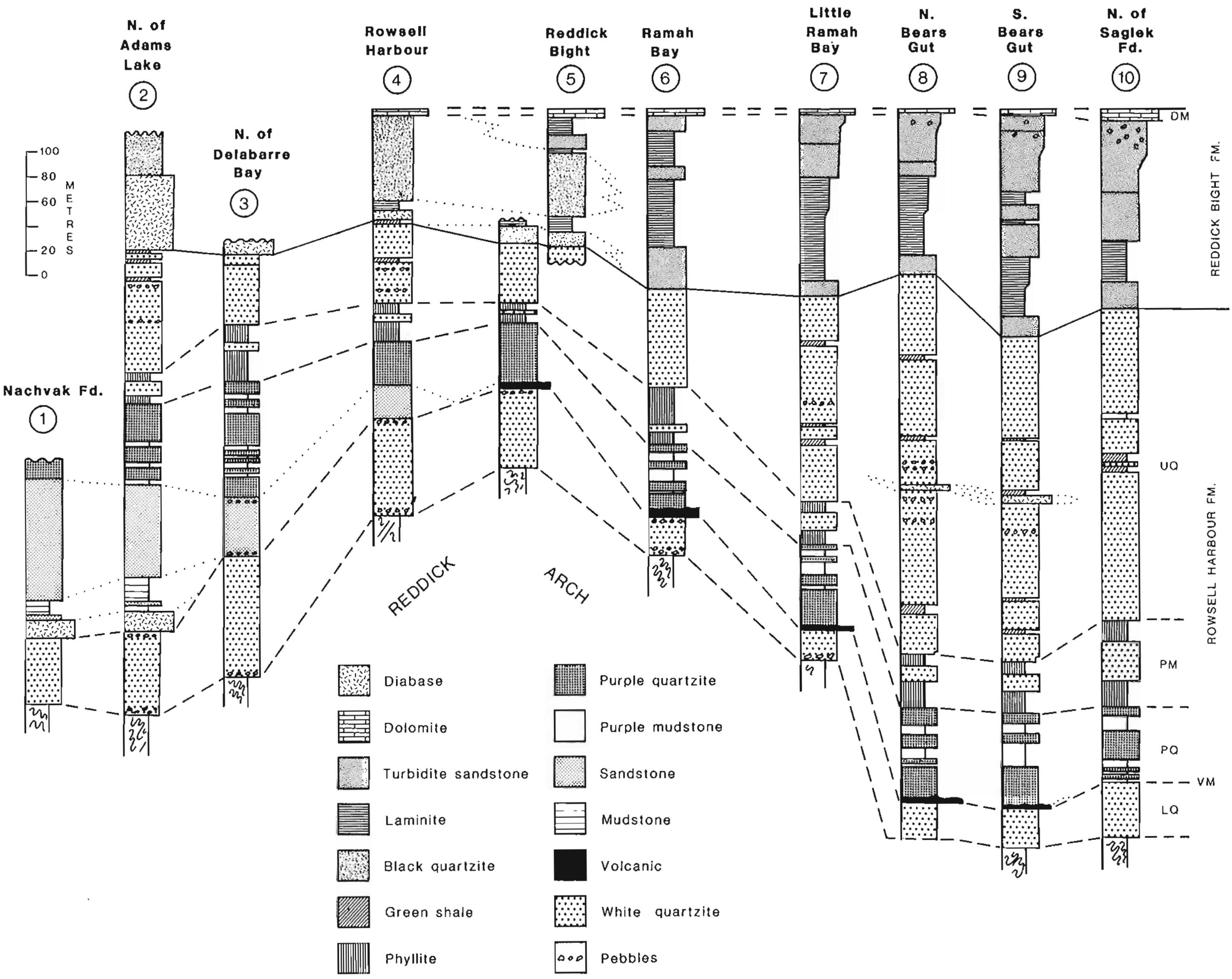
Basal ferruginous sandstones of the northern basin occur only north of Adams Lake, are 16-20 m thick, and consist of thinly stratified, yellow weathering, very fine sandstone alternating with white to red phyllites. Lamination and minor crosslamination are the dominant structures. The yellow and red coloration is produced by pyrite and iron carbonates.

The coarse phyllitic sandstone overlying the ferruginous sediments in the northern basin consists of white, purplish white, and grey, very coarse to medium grained sandstone, with minor fine sandstone, grit and conglomerate. Black to grey phyllitic laminae delineate stratification in the sandstones, and thin shale drapes separate 30-140 cm thick beds. Pebbly horizons are composed of 1-1.5 cm quartz pebbles and shale intraclasts. Large and small scale trough crossbedding and some planar cross stratification are the dominant sedimentary structures. Several large composite structures are composed of tabular cross sets that pass laterally into flat lamination and straight, sinuous or locally linguoid ripple marks. Many ripple crests are smoothed, and rippled surfaces are covered by mudcracked mud drapes. Fining upward sequences several metres thick are common near the base of this unit, but upwards the typical coarse phyllitic sandstones are intercalated with very fine sandstones to form coarsening-upward sequences. These very fine sandstones are interbedded with grey shales and are 72-168 cm thick. The sandstone beds have sharp flat bases, and are internally laminated with only minor crosslamination and ripple marks.

In the southern basin, the base of the member consists of 14-28 m of well sorted, medium- to fine-grained, locally herringboned, trough and planar crossbedded pink quartzite that may be partly equivalent to the coarse phyllitic sandstones of the northern basin.

The uppermost unit of the member in both basins consists of purple quartzites and mudstones, intercalated on a 10-70 cm scale, or as major 12 m thick quartzite units and 20 m thick mudstone sequences. These quartzites are chiefly medium- to fine-grained and, unlike the white quartzites, are generally well sorted quartz arenites with well rounded grains, quartz overgrowths, hematite coatings and minor carbonate cement. The mudstone-quartzites are dominantly coarsening-upward sequences, but some quartzites show grain size uniformity or fining-upward sequences.

The quartzite units locally contain mud clasts and lie with sharp planar, erosive to locally channelled bases on the mudstones. Trough and planar cross stratification indicating current reversals, and cross strata climbing up the backs of earlier sets, are common. Shale beds 1-7 cm thick occur between the quartzite beds and load casts are common at the contacts. The geometry of the quartzite units appears to be broad sheets, many of which clearly have concave upward upper surfaces.



LQ - lower white quartzite member
VM - volcanic member
PQ - purple quartzite and mudstone member
PM - phyllite member
UQ - upper white quartzite member
DM - dolomite member

Figure 18.8. *Stratigraphic sections of Rowsell Harbour and Reddick Bight formations. Locations of sections (1 to 10) are indicated on Figure 18.2.*

Ripple marked thin sandstone and shales in 20 cm and locally 200 cm thick units cap many quartzites. B and C type ripple drift (Jopling and Walker, 1968) is abundant and the strata in many areas are deformed by convolution and cut by sandstone dykes.

The mudstone sequences are characterized by 5-15 cm thick laminated, cleaved mudstones rich in laminae and thin beds of sandstone. In some mudstones there is an upward increase in grain size and number of sandstone beds towards the overlying quartzite sequence. These sandstones typically occur as thin flat sheets, lenticular bodies, or isolated ripples within the mudstone. Purple quartzite lenses are common and can locally be traced into thin beds. Whereas some quartzites are capped by very large ripple marks of 20 cm amplitude and 112 cm wavelength, others have smooth curved upper surfaces and planar bases.

Phyllite Member

This member is of variable thickness (15-44 m), and is thinnest over Reddick Arch and near Nachvak Fiord (Fig. 18.8). Elsewhere it consists of two varicoloured phyllite units, 14-28 m and 3-15 m thick, separated by a 3-14 m thick quartzite unit that occurs at different levels in the member. A thick purple quartzite unit caps the member at Little Ramah Bay, but is absent to the south and north.

The phyllites occur in alternating 0.5-6.9 m thick colour sequences of yellowish pink phyllite and purplish grey phyllite. Both types consist of 5-30 cm graded beds that commence with sharp, irregular to planar erosional bases. The scours are overlain by a few centimetres of laminated, silty phyllite which grades into structureless phyllite. In the purplish grey facies and in some beds of yellow-pink phyllite,

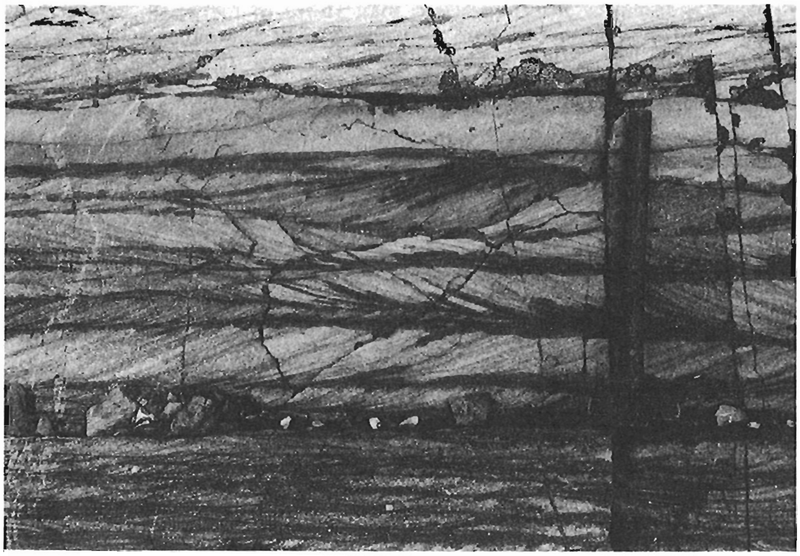


Figure 18.9

White quartzite of Rowsell Harbour Formation with heavy mineral lamination, chiefly formed by magnetite, outlining small scale crossbeds. 8 km NW of the head of Rowsell Harbour. GSC 162673

rusty weathering, very fine laminated and crosslaminated sandstones overlie the scours. Channel bound sandstone beds rich in mud clasts also occur.

The quartzite units, comparable to those of the underlying member, are white in the north but become purple to the south. They commonly show erosive bases and large scale crossbedding, and contain phyllite intraclasts.

Upper White Quartzite Member

The upper white quartzite member lies sharply and conformably upon the phyllite member. It is thinnest (46-68 m) over Reddick Arch but thickens southwards to 267 m at Bear's Gut and northwards to 192 m at Naksaluk Cove (Fig. 18.8). The member is generally composed of monotonous white quartzites, with some green beds in the south. At the base of the member the sandstones are petrographically similar to the lower white quartzites, but at higher levels they include up to 10 per cent fresh plagioclase and microcline-perthite. Conglomerate is scattered through the sequence but the major other lithology consists of thin beds of green phyllite or units of thinly stratified, rippled sandstone and phyllite.

The white quartzites are chiefly fine- to medium-grained with some coarser grades. Coarsening-upward sequences occur, formed of sandstone capped by pebbly sandstone or conglomerate, or of thinly stratified sandstone and shale overlain by sandstone capped by pebble beds above which mudcracked units may occur. These cycles vary in thickness from a few metres to 30 m, with a maximum of 60 m recorded at Bear's Gut.

The upper white quartzite member commences with 3-5 m of laminated, planar-stratified, very fine quartzite with shale partings. This sequence is overlain by predominantly trough crossbedded white quartzites. Planar cross stratification, flat lamination with parting lineations, and large and small ripple marks are also present in the quartzites. Green phyllite drapes and partings are common. Mud-filled ripple troughs are mudcracked and rain prints, convoluted foresets, and sandstone dykes are also present. Many mudcracks, however, are incomplete, and are similar to the subaqueous syneresis cracks of van Straaten (1954) and Burst (1965).

Conglomerates in the quartzites are 10-50 cm thick and consist of 1 cm pebbles of quartz and rare feldspar, jasper and phyllite. They overlie erosive bases, and are either crossbedded or structureless. Thinly stratified sandstones and shale form sequences a few centimetres to 3 m thick. Both shale and sandstone-dominated sequences occur, with the latter generally typified by flaser bedding. In the shale dominated types, 1-9 cm thick, very fine sandstones are either lenticular or thin flat bedded sheets which lens out, and are internally laminated or structureless. Rippled sandstone sheets also occur.

Paleocurrents

Paleocurrents collected from the Rowsell Harbour Formation (Fig. 18.10) give polymodal distributions (Morgan, 1975). However, the distribution varies somewhat from member to member, and also within facies of members. The main points are summarized below:

1. Crossbeds with polymodal, bimodal, and unimodal distributions dominated by southwest vectors occur in the quartzites of the lower white quartzite member. However, a westerly paleoflow with only minor reversal dominates the crossbedded, heavy mineral-rich quartzites of the lower part of this member. Crossbeds in the rusty weathering, pebbly sandstone give southwesterly directed currents.
2. Bimodal-bipolar distributions of crossbeds are dominant in the purple quartzite and mudstone member. The north-south directions are dominated by the northerly vector in the purple quartzites. This direction is closely paralleled by data from the coarse phyllitic sandstones of the base of the member north of the Reddick Arch.
3. Polymodal and bipolar-bimodal crossbed distributions occur in the upper white quartzite member, but are again dominated by a northwest and west vector.

Reddick Bight Formation

The Reddick Bight Formation gradationally overlies the Rowsell Harbour Formation and is 53-143 m thick (Fig. 18.8). It consists of a lower member formed by a complex of grey sandstones, shales, brown laminites, black quartzites, a mudflow, and an upper dolomite member. The top of the formation is placed at the top of the ubiquitous yellow-weathering dolomite member. The formation thins northward and westward, decreasing from greater than 100 m in the south to less than 60 m over Reddick Arch, and to less than 20 m near Nachvak Fiord and along the western contact of the group northwest of Bear's Gut.

The lithology of the lower member changes across Reddick Arch. Black quartzites dominate the sequence in the northern basin, but terminate abruptly near the south margin of the arch (Fig. 18.8). The quartzites are associated with laminites and turbidite sandstones.

In the southern basin, the lower member is typically composed of lower and upper sandstone units separated by a thick laminite unit. Whereas some sandstones show turbidite-like structures, others are chiefly crossbedded. The upper sandstone unit coarsens upwards, becoming pebbly and granular near the top. Both lower and upper sandstone sequences, dominated by crossbedded sandstone with thick

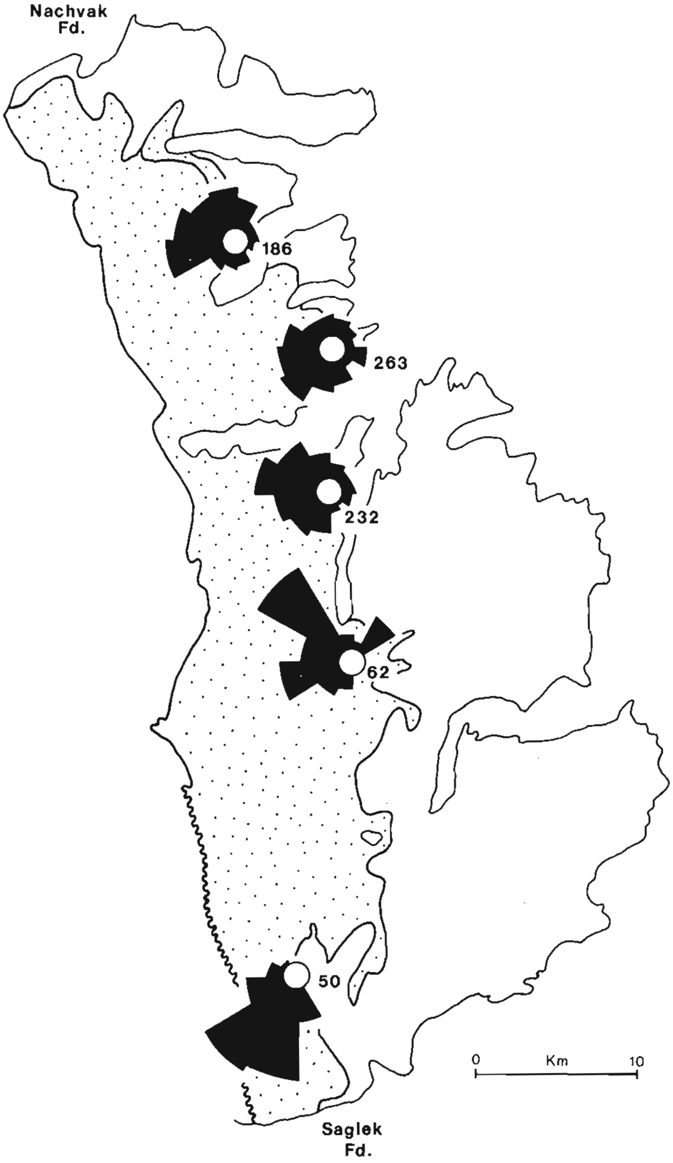


Figure 18.10. *Paleocurrent data derived from crossbedding in the Rowsell Harbour Formation. Diameter of centre circle in histograms is 10 per cent.*

laminite interbeds (less than 10 m) in the south, are transitional into turbidite sandstones with a sympathetic decrease in bed thickness and grain size north of Little Ramah Bay. For example, pebbly and granular sandstones south of Little Ramah Bay and Bear's Gut pass northward into medium to coarse sands and eventually into very fine sands and coarse siltstones.

A northward decrease in grain size also occurs in the laminite facies, reflected by a change from a sandstone:shale ratio of 70:30 in the south to 30:70 or more in the north. Laminae of fine sandstone in the southern laminites are siltstone in 50 per cent of the laminites in the north.

Sandstones

The turbidite sandstones, interbedded with shales, are mostly 5-70 cm thick and have erosive bases with local channels overlain by predominantly structureless, commonly intraclast-rich argillaceous sandstones. Most beds are graded and some display lamination and rare crosslamination. Internal deformation is common and tubular water escape pipes and sheet-like structures, similar to pillar structures (Lowe, 1975), are present in the massive sandstones. Sandstone dykes, load casts, flame structures and local post-depositional thickening also occur.

Crossbedded sandstones are intercalated with massive sandstones and are developed as thick 145 cm to 13 m units interbedded with shale or laminite. Small scale trough crossbedding occurs in the lowest sandstones, and the set size increases upwards. Planar cosets become important accompanying this upward coarsening grain size. The sets overlie planar scours and are continuous for several metres down current.

Clasts in the pebbly sandstones are chiefly of grey and opalescent quartz, fresh plagioclase, amphiboles, opaque minerals and mud chips. The sandstones are argillaceous subarkose with minor arkose, and have a matrix that varies up to 40 per cent. Sand grains are composed of fresh microcline, perthite and plagioclase detritus mixed with quartz, some of which is polycrystalline. Minor tourmaline, amphibole and zircon also occur. In the uppermost sandstones, rare grains of dusty cryptocrystalline quartz intergrowths, and embayed clear quartz with fine felsic texture matrices in the embayments, suggest a minor felsic volcanic source for some detritus.

Laminite

Throughout the unit, shaly sandstone or siltstone laminae alternate with silty shale laminae on a millimetre scale. Although the laminae are generally smooth, small scale discontinuity surfaces, lensing out, local scour and fill, and small ripple lenses are common. Rusty, very fine to fine sandstones up to 16 cm thick are spaced in the laminites. They are massive, laminated or crosslaminated, with irregular scour bases and some ripple marked tops.

Structures on bedding planes resembling foam impressions (Reineck and Singh, 1975) and longitudinal ripple marks (van Straaten, 1951) were previously misinterpreted as possible trace or fossil structures (Knight, 1973; Knight and Morgan, 1977).

Quartzites

Black quartzites form 12-35 m thick sequences from Reddick Bight to Delabarre Bay. They are very fine to medium grained and occur in 0.5-5.0 m beds separated by 10-55 cm beds of shale with thin sandstones. Grits and pebble beds and lenses also occur in the quartzites. Large scale trough crossbedding is present, but is not easily distinguished. There is also some minor flat lamination and crosslamination associated with straight, sinuous and linguoid ripple marks. Ripples have maximum amplitudes of several centimetres and wavelengths of up to 40 cm. Shale drapes in some ripples are mudcracked, and mudcracks also cut the shale beds.

Mudflow

The mudflow facies, seen nowhere else in the formation, occurs at the base of the upper turbidite sandstone sequence on the south side of Reddick Arch at the Reddick Bight type section (Knight and Morgan, 1977). It is 2.8 m thick, and is composed of deformed, rolled and attenuated laminite, shale and sandstone lithoclasts, set in a sand-mud matrix. Sharp planar contacts occur. The flow is at the base of a 12 m thick turbidite sandstone sequence, and lies only 45 cm above quartzite with mudcracked shale.

Dolomite Member

This 4-17 m member caps the formation throughout the greater part of the synclinorium. It consists of a lower dolomitic sandstone with an irregular gradational base, and an upper 3-4 m thick, pure dolomite. The basal sandstone preserves some original bedding structures, but these are commonly obliterated. Grain boundary solution is visible in hand specimen, and pods of pure dolomite are common. The upper, pure, dolomite consists of a brecciated and folded, thinly stratified, fine dolomite, cemented by white sparry dolomite. The latter is prismatic and radial, and developed on and perpendicular to the breccia fragments. The brecciation appears to have occurred during the formation of 50-100 cm high domes which have a wavelength of 0.5-2 m. Dolomite spar and chert fill pores and late fractures.

Remnants of thinly stratified, fine grey to pale grey dolomitic calcilutite are preserved close to the base of the pure dolomite near Adams Lake. This is the only evidence of an original bedded carbonate precursor to the dolomite.

Paleocurrents

Paleocurrent data from Reddick Bight Formation are generally scarce, except in the black quartzites north of Reddick Bight where crossbeds give a bimodal NNW-SSE distribution, with the northwest sector dominant. Ripple marks yield a polymodal spread, again dominated by the north-northwest trend. In laminites from the same area, a northeast trend for linear, very thin groove casts occurs and this is generally paralleled by current ripple marks, although symmetrical ripple crests trend at right angles. A northeast direction for current ripple marks occurs on the south side of Rowsell Harbour.

In the Bear's Gut area, ripple marks in the basal crossbedded sandstones have a polymodal distribution. Crossbeds, however, indicate a westerly to northwesterly paleoflow, which appears consistent throughout the formation in this area. The planar cross stratified sandstones at the top of the formation give southwest and northerly vectors. Each vector is restricted within its own crossbedded unit and this suggests that each sandstone deposit is essentially unidirectional with an overall northwestward progradation. A few crossbeds at Little Ramah Bay give the same vector. Linear water escape structures and sandstone dykes in the Little Ramah Bay area have a southwest trend suggesting a northwesterly inclined paleoslope.

Interpretation and Depositional Environment of the Shelf Sequence

Shallow shelf sedimentation influenced by a number of transgressive-regressive events prevailed throughout the deposition of the Rowsell Harbour and Reddick Bight formations. Whereas the Rowsell Harbour was deposited in somewhat fluctuating, shallow shelf conditions with marine and possibly some fluvial processes dominant, Reddick Bight strata record a prograding, shallowing upward, deltaic complex which ultimately produced regression and blanketed the shallow shelf. The Reddick Arch was an important intrabasinal tectonic influence during deposition of the two formations. Stratigraphic sections hung from the dolomite member of Reddick Bight Formation (Fig. 18.8) reveal a positive high influencing thickness near Reddick Bight and separating two basins. Reddick Arch was active during the deposition of part of the Rowsell Harbour Formation as only a number of units are restricted to north of the arch. Facies in the Reddick Bight Formation also change close to the arch. The arch probably trended just south of east, but no supporting evidence is available.

Quartz-rich sands devoid of fines are commonly considered to be deposits of shallow shoreline and shelf environments, formed under generally high energy conditions. Peculiar to the Rowsell Harbour quartzites is a high argillaceous content which reaches 20-40 per cent in some strata. Nevertheless, bimodal and polymodal paleocurrents, herringbone cross stratification, large ripple fields, runoff channels, lenticular and flaser bedding, mudcracks and rain prints suggest intertidal, sand dominated environments such as beach and barrier sands, tidal deltas and intertidal sand flats (Boothroyd, 1978; Davis, 1978; Reinson, 1979).

The general upward fining of the lower white quartzite member reflects the response of a barrier sand complex migrating across the shelf, as transgression advanced across the basement. Fluvial sedimentation may have been important in the lower part of this member where arkosic granite wash and poorly sorted, argillaceous, heavy mineral-rich quartzites occur. Although transgression probably advanced rapidly, the sandy shelf remained generally shallow as indicated by the runoff structures and mudcracks on ripple fields near Delabarre Bay. Consequently, when the transgression ceased, a thin blanket of pebbly sandstones, possibly deposited by shallow braided streams, blanketed the shelf. Current directions for this deposit remain southwest, indicating a source area to the northeast and east, so that transgression must have advanced eastward, and possibly southeastwards, since the Rowsell Harbour Formation wedges out north of the Mugford Group (Smyth and Knight, 1978).

The volcanic member, characterized by subaerial structures, blanketed the braided sandy alluvial plain during this regressive stage.

Although shallow marine and shoreline deposition is envisaged for the lower white quartzite member, during formation of the purple quartzite and mudstone member, conditions changed to deltas with high sediment supply influenced by tidal effects similar to type II deltas (Wright, 1978), and possibly offshore barrier sands. The depositional setting was then a mud dominated shelf that formed after a second transgression which deposited shallow, tidal sand on flats and mudflats of the lower part of this member. The deltas built northwards across the mud dominated shelf, producing coarsening upward sequences with bipolar crossbedding suggesting an active tidal influence. Starved sand waves and ripples were common in the muds.

The shelf may have deepened somewhat during deposition of the overlying phyllite member, although this deposit might also represent a tidal mudflat crossed by delta channels in which associated thicker quartzite units were deposited. The former interpretation is preferred, however, since no mudcracks were seen and there is no mud chip conglomerate at the base of the upper white quartzite member suggesting that transgressive reworking occurred. Instead, flat stratified, laminated, very fine quartzites are present, and these probably represent lower foreshore deposits (Harms, 1975), as would be expected if the shoreline migrated seaward.

The upper white quartzite member with its thick, upward coarsening cycles, suggests progradation of barrier island and sandy shoreline complexes across the shelf (Elliot, 1978). The shelf may have been fairly deep at this time. Thinly stratified planar and rippled sandstones and shales at the base of some of the sequences lack evidence of tidal flat deposition, but are comparable to units described by Anderton (1976) and de Raaf, et al. (1977), which are interpreted as deeper subtidal deposits marginal to migrating sand bodies.

The depositional environment of the Reddick Bight Formation is interpreted as a deltaic complex in the south contrasting sharply with a dominantly shallow intertidal sand and mudflat north of Reddick Arch. The northern sand flat represented by the black quartzites, with polymodal ripple marks, bimodal crossbeds, clean sands and common mudcracks in mud drapes on rippled surfaces, suggests intertidal deposition reflecting a depositional control by the Reddick Arch. Laminites associated with the quartzites display foam impressions and longitudinal ripple marks similar to modern tidal flats (Reineck and Singh, 1975).

The other facies of the formation, however, reflect a deltaic complex prograding northwestwards across a somewhat deeper shelf and becoming gradually dominant over shelf-tidal and current processes. The predominance of crossbedded and turbidite sandstones suggests a fluvial dominated delta heavily charged with coarse sediment. The crossbedded sandstones were probably deposited as an extensive lobate sand sheet (Miall, 1979) that formed in response to shifting distributary channels on the delta front and top. Instability of the delta front, or high density sand and mud discharge along channels, produced the turbidite sandstones at the foot of the delta front, possibly as subaqueous fans (Walker, 1978). The abundance of intraformational clasts indicates that the depositing currents cannibalized earlier deposits. The mudflow and numerous sandstone dykes, water escape structures, and penecontemporaneous deformation of the sandstones suggest that the delta front environment was unstable, probably because of rapid sediment accumulation and oversteepening of the delta front. The laminites lacking dessication features were subaqueously deposited as a quiet, deep water, prodelta facies. The thin sand-shale alternation was interrupted by thicker sandy beds showing structures again reminiscent of turbidity deposition. The growth of the complex appears to have become less active during deposition of the middle part of the formation, as the lower turbidite sandstones are succeeded by a laminite blanket over the entire area. Reactivation of the source area, possibly due to increasing tectonism, initiated deposition of the upper sandstone sequence which coarsens upwards reflecting active progradation of the fluvially-dominated delta into the basin at this stage. Foredelta turbidite sandstone then invaded the laminite basin, building a fan over which the distributary channels and delta top fluvial channels advanced to ultimately form a wide, lenticular sand blanket centred on the southern end of Little Ramah Bay. The unidirectional, planar-trough cross stratification of the upper part of the formation suggests straight channels, or a braided system, switching direction frequently from southwest to north across the delta top. The braided system interpretation is supported by the southward increasing pebble content of the sandstones and it is probable that the delta passed south into a pebbly alluvial plain. Detritus for this river system was supplied from basement and a minor felsic volcanic source. However, no felsic volcanics are known in eastern Labrador.

Delta advance appears to have ceased quite rapidly and was followed by formation of secondary dolomite which possibly represents a depositional hiatus. This dolomite clearly replaces the uppermost beds of the delta, regardless of whether they are laminite or sandstone. Pure dolomite replacing original thinly stratified dolomitic calcilutite was observed at one locality. This suggests that shallow carbonate deposition over the delta plain, possibly in response to gentle flooding of the abandoned delta, preceded dolomitization. Diagenesis of the calcilutite to fine dolomite, which was subsequently buckled and cemented by sparry dolomite, is interpreted as supratidal diagenesis of the thin carbonate mud flat. Volume increase due to crystallization pressures and temperature fluctuations during diagenesis led to lateral stresses which buckled and fractured the bedding and formed tepee structures similar to those described by Assereto and Kendall (1971, 1977). Semi-arid or arid conditions caused evaporative "pumping" of carbonate saturated groundwaters into this fractured, hummocky terrain. This led to deposition of the fringing sparry cement on fragments and cavity linings and probably caused precipitation of dolomite, solution of silicic grains, and destruction of sedimentary textures in the underlying siliciclastics.

BASINAL FINE-GRAINED SEDIMENTATION

Contrasting markedly with the basal siliciclastics is a 1000 m sequence of predominantly mud and silt grade succession. Four formations, which are chiefly argillaceous with some carbonate strata in the middle, are recognized.

The basal black shales sharply overlie the dolomite member at the top of Reddick Bight Formation, suggesting that basinal conditions were established rapidly.

Nullataktok Formation

The Nullataktok Formation (Fig. 18.3; see Table of Formations) is a 595 m thick sequence of shales, varicoloured mudstones, calcareous and dolomitic mudstones, thin dolomitic sandstone, some intraformational breccias, limestone, sedimentary chert, pyrite iron formation, siliceous dolomite and argillite. Fine grain size, structural complexity and a well developed cleavage produce generally poor exposure, so that only a few localities, the best of which is the north shore of Ramah Bay, provide a good section.

A fairly consistent vertical sequence, with slight local variations, occurs in the formation. It commences with black shale and varicoloured slaty mudstones which occur below a sedimentary pyrite and chert association that is overlain by thin siliceous dolomite and argillite. The varicoloured slaty mudstones, with localized channel-bound dolomitic breccias and sandstones, then resume and continue for more than half the thickness of the formation. Above this point, however, the succession gradually becomes calcareous and dolomitic and the dark mudstones form units interlayered with light grey, calcareous mudstones, green shales with white limestone, yellow-weathering dolomitic mudstones, and some limy turbidites. Intraformational breccias, local thick slump beds and thin dolomitic fine sandstones break the monotony of the fine carbonate succession. Bedding throughout is generally thin with stratification showing persistent lateral continuity except for some local erosion.

Black Pyritiferous Shale

Pyritiferous, graphitic, sulphurous black shales form the basal 12-20 m of the formation. The jet black shales are generally structureless except for some single or grouped, very fine, light coloured lamination and pyrite laminae which locally may outline flat lying, recumbent slump folds. Spherical or irregular 21 by 5 cm pyrite nodules also occur. A 2-3 m black chert unit occurs at the base of this sequence, near Saglek Fiord, directly upon the dolomite of the Reddick Bight Formation.

Varicoloured Mudstones

This, the dominant facies in the formation, is approximately 350 m thick and is divided into two segments by the pyrite-chert association that occurs about 60 m above the base of the formation. Green-grey and blue-black alternating colour banding on a 1-80 cm scale typifies the facies. Structureless, planar green-grey mudstone beds are transitionally overlain by finer grained, finely laminated darker mudstone. The ratio of one band to the other varies throughout the sequence. Local brown weathering, structureless and/or finely laminated thin siltstone beds occur at the base of some colour couplets and may form multiple beds up to 15 cm thick. Petrographically, the siltstones are composed of quartz and plagioclase detritus set in a carbonate-sericite-chlorite matrix.

Sedimentary Pyrite-Chert-Dolomite-Argillite

This 8-20 m thick sequence interrupts the colour banded mudstones and was traced, varying somewhat in thickness and succession, from north of Adams Lake to south of Bear's Gut. The most prominent change is the absence of pyrite but thickening of the chert in the northernmost outcrops.

A 40-57 cm thick pyrite bed consists of pinkish, granular pyrrhotite, probably resulting from contact metamorphism by an overlying ultramafic cored diabase sill in the south, and silvery grey pyrite in the north. In the south, from Bear's Gut to Little Ramah Bay, the pyrrhotite bed is pebbly, at first massive and then bedded. The pebbles are less than 7 cm in diameter, appear to decrease in size northward, and are composed of white quartz, chalcedonic quartz and green shale. They are not present north of Little Ramah Bay, where finely laminated, though locally brecciated and recemented, pyrite occurs.

The rusty brown weathering Ramah chert is about 4.5 m thick where it overlies the pyrite bed, but thickens to 17 m north of Adams Lake where the pyrite is absent. The basal contact is sharp though nodular, and the white, grey, and black vitreous to rarely granular chert is well bedded on a 6-45 cm scale. Green pyritic shales and mudstones separate some beds, particularly near Adams Lake. Dolomite nodules are common in the chert and fine lamination is preserved locally.

Above the chert, a 220-236 cm thickness of thinly bedded, rusty weathering, siliceous dolomite and green and white striped argillites is consistently present. Bedding is smooth though undulose, and both lithologies are finely laminated.

Intraformational Breccias

Channel-bound dolomitic intraformational breccias and sandstones occur in the colour banded mudstones about 200 m above the chert. The breccias are only 10 cm thick at Ramah Bay, but occupy two deeper channels, separated by 290 cm of grey mudstone, southeast of Adams Lake. Whereas the lower channel is 59 cm wide and 26 cm deep, the upper channel is at least 18 m wide and 290 cm deep. The breccias are composed of pebble- to cobble-sized angular clasts of dolomitic mudstone, sandy or massive dolomite, and quartz sandstone, supported by a quartz dolomite sand and mud matrix. The matrix-supported breccias in the thicker channel occur at three horizons interbedded transitionally with massive grits and sandstones, and with laminated sandstone at the top.

Calcareous and Dolomitic Mudstones

The upper 220-250 m of the Nullataktok Formation comprises calcareous and dolomitic mudstones interbedded with less calcareous mudstones. Carbonate content rarely exceeds 40 per cent (Jefferson, 1973) and although weathering delineates sedimentary structures, the fine uniform crystallinity associated with recrystallized quartz, silt and fine clays defy petrographic description even in sandy beds. The limy layers produce rusty grey to silvery grey coloured lutites whereas dolomitic units weather bright yellow. Contacts are gradational.

Thin, flat, continuous stratification of alternating structureless and laminated mudstone is ubiquitous in all lithologies. Laminated and crosslaminated dolomitic siltstones or very fine sandstones, up to 3 cm thick, are common near the top of the formation. The monotony of the sequence is broken only by small to quite large scale sedimentary deformation structures including convolution, slump folds, faults, sedimentary dykes, boudinage and slump breccias. The thickest slump deposits are 1-2 m thick, and appear to have moved substantial distances. Some slumps carry slabs of relatively undeformed strata, since layers locally peel downwards into the apparently liquified and brecciated sole of the slump. Other slumps are overlain conformably by flat-lying strata which infill fractures in their upper surfaces.

Units 50-200 cm thick of green shales with nodular to planar thin beds of white to grey calcitic limestone, which are invariably boudinaged or reworked in slumps, are characteristic of the upper 70 m of the formation.

Several limestone turbidites, 20-56 cm thick, interbedded with grey shales, lie in the lower part of the succession near Delabarre Bay. They have erosive bases, massive intervals, and convoluted lamination that contains lensoid crosslamination structures.

Warspite Formation

Warspite Formation comprises a lower sequence of stratified fine dolomites, dolomitic breccias, dolomitic sandstones, and dolomitic mudstones; and, an upper sequence of argillites and mudstones with some dolomite and sandstone. Distribution and vertical sequences for the formation are shown in Figures 18.2 and 18.11. The formation is 110-165 m thick, and is separated from the calcareous mudstones of the underlying Nullataktok Formation by the basal Warspite breccia (Fig. 18.3). Between Ramah Bay and Rowsell Harbour, the basal breccias are composed exclusively of clasts of the Nullataktok Formation, but these are mixed with sandy dolomite and dolomitic quartz sandstone clasts at Delabarre Bay, Little Ramah Bay and along the western limb of the synclinorium.

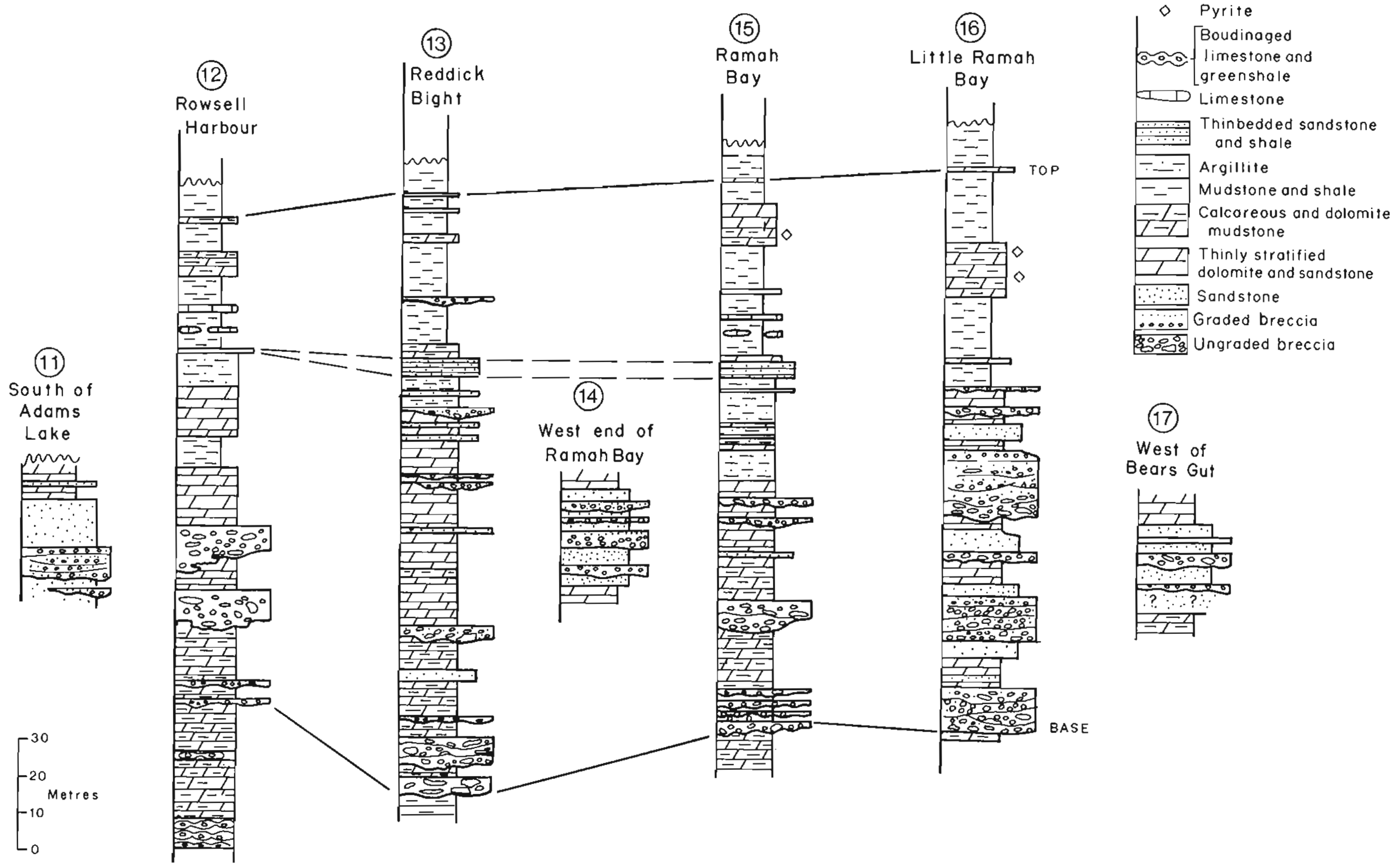


Figure 18.11. Stratigraphic sections of Warspite Formation. Locations of sections (11 to 17) are shown on Figure 18.2.

Although original grain textures are mostly obliterated by the fine crystallinity of the dolomite and recrystallization of quartz in the sandstones and dolomites, some albitic plagioclase grains can be discerned.

Breccias and Associated Sandstones

Breccias, ranging from matrix to clast supported, occur with or without thick bedded sandstones in a number of 1-12 m thick sequences in the basal 115 m of the formation. The breccias along the eastern margin of the synclinorium are channel bound. The smaller channels have gently inclined to steep, ragged, margins with local breccia injection and abundant evidence of plucking. A possible canyon, 75 m deep, crosscuts Nullataktok strata 4 km southwest of Little Ramah Bay, but this feature has not been examined in detail. Blocky, angular, commonly cobble-sized Nullataktok clasts, supported by a structureless sandy mud matrix, typify the basal deposits at Reddick Bight. The deposits near Little Ramah Bay include chiefly ungraded, clast-supported breccias, composed of small to large pebble-sized sandy clasts.

Nongraded and normally graded beds up to 3 m, as individual strata and associated with thick bedded grits and sandstones, typify the breccias along the western limb of the synclinorium. The sandstones and grits are either nongraded and structureless, or grade upwards from a basal massive division into coarse laminated sands with thin crosslaminae. Isolated and multiple sets of planar and trough crossbedding locally occur in the sandstones.

Thinly Stratified Dolomites with Sandstones

Thick beds of thinly stratified, yellow- to orange-weathering dolomites with thin dolomitic sandstones and white quartzitic sandstones are important lithologies associated with the breccias. They are developed only with the upper breccia sequences in the east, but occur with the lower breccias in the west. Laminated 5-15 cm dololutite beds are interlayered with 1-3 cm structureless dolosiltite beds. Planar, inclined, irregular scours and channel scours cross the generally planar stratification, which may be deformed by decollement structures, slump folds and convolution. Ripple drift and laminated, very fine, 1-3 cm sandstones also interrupt the dolomites. These thicken to 15 cm thick sequences near the breccias at the western margin of the formation. Scouring and convolution of the ripple drift structures are common.

Lenses and sheets of apparently planar crossbedded, and locally laminated, white quartz sandstones are interspersed in the dolomites at Reddick Bight and Delabarre Bay. These 10-15 cm beds have erosive-deformed bases and are locally up to 130 cm thick.

Dolomitic and Calcareous Mudstones

Mudstones similar to those in Nullataktok Formation occur in the northeast of the area within the lower 40 m of Warspite Formation.

Argillites, Thinly Stratified Sandstones and Shales

Argillites overlain by thinly stratified sandstones and shales occur 40-70 m from the top of the formation. They thin both north and south away from Ramah Bay and Reddick Bight. At Reddick Bight the argillites are crosscut by, and also overlie, a channel bound breccia. The argillites are green grey and black, are characterized by locally deformed thin, flat lamination, and contain interspersed 1-3 cm laminated, fine white sandstone beds.

White sandstones and black shales gradually replace the argillite in the upper 4-6 m of the sequence. These 5-20 cm sandstones scour into the interbedded shales, and are typically laminated or crosslaminated with flat or rippled tops. Millimetre-sized silt laminae and thin sandstone beds occur in the shales, which are locally cut by sand filled fractures.

Grey Mudstones with Minor Carbonates

Thinly stratified, laminated, rusty weathering light and dark grey slaty mudstones dominate the upper 35-55 m of the formation. They contain rusty weathering, pyrite speckled, laminated dolomitic mudstones and black, faintly laminated crystalline limestones. The black limestones are either nodular or form irregularly bedded 1-1.5 m units.

Paleocurrents

Paleocurrent data for Warspite Formation derived from crossbeds, ripple marks, ripple drift crosslamination, channel orientations and lateral variations of lithology indicate a westward transportation of sediment. The cross strata fan from southwest to north, and deformation suggests north and southwest paleoslopes. Channels, however, are orientated at about 230° at Reddick Bight and Rowsell Harbour, and a westward change from matrix-supported breccias to graded and ungraded breccias with massive and crossbedded sandstones, suggests a regional east to west depositional slope.

Typhoon Peak Formation

Typhoon Peak Formation (Fig. 18.2, 18.3; see Table of Formations) is the most poorly exposed formation in the group. It is intruded by several diabase sills and is highly folded and cleaved. The 85-130 m thick formation was measured in two sections, north and south of Ramah Bay. The northern section is composed almost exclusively of rusty-weathering grey slates. South of Ramah Bay, however, white laminated sandstones and siltstones, grey quartzite concretions, and grey limestones are interbedded with the slates.

Alternating 1-3 cm thick massive and laminated layers, can be observed on cleavage surfaces of the slates. Pyrite cubes speckle the slates and only rarely do 1-12 cm thick sandstone beds break the monotony of the succession. They are internally massive, bounded by sharp bedding planes, and are commonly boudinaged.

The white sandstones and siltstones occur south of Ramah Bay to Little Ramah Bay, and form 1-18 m units that thicken southwards. They are thinly bedded and laminated, with sharp planar bedding planes. Large (100 by 75 cm) quartzitic concretions are associated with one of the units south of Schooner Cove, Ramah Bay. They have an outer spheroidal shell composed of crystalline grey quartz and an inner massive calcareous core.

Minor strata of black to grey, thinly stratified limestone are associated with the sandstones and concretions. These beds are fine grained and structureless with rare faint laminations.

Cameron Brook Formation

The uppermost unit of the group (Fig. 18.2, 18.3; see Table of Formations), the poorly exposed 200 m thick Cameron Brook Formation, gradationally overlies the dark slates of Typhoon Peak Formation. It is composed of cleaved grey to black, argillaceous turbidite sandstones alternating with black to grey shales. At some points in the sequence, either the shale or sandstone thickens at the expense of the other lithology, and some distinct beds of laminated shaly siltstone also occur. The sandstones are subarkoses to lithic subarkoses composed of white and opalescent blue quartz, fresh plagioclase, shale, siltstone and albitized basic volcanic grains, set in a fine matrix of recrystallized micas, epidote and carbonate (Morgan, 1975). Grain size in the sequence is variable, though dominated by fine and medium sizes. Coarser sandstones form intervals up to 750 cm thick in the Ramah Bay area, and grits and pebbly sandstones occur west of Little Ramah Bay.

The sandstones contain Bouma cycles (Bouma, 1962) dominated by A→B→E and A→E sequences (Walker, 1967), with some A→B→C→E and a few A intervals. Beds are commonly 10-70 cm thick, with 10 cm shale interbeds, but are locally up to 125 cm in the coarsest A turbidites.

The turbidites have sharp planar to locally undulose or irregular bases, which are commonly deformed by load casts and flame structures. In some beds a thin, inversely graded horizon is overlain by a normally graded massive A division. Mud content is high and black shale clasts, although commonly only centimetre-sized reach up to 30 cm. Faint, flat stratification is present locally.

Laminated B and crosslaminated C divisions are commonly thin, but locally are up to 30 cm. They are overlain by laminated siltstone and shale of division D. Dish and pillar structures (Stauffer, 1967; Chipping, 1972; Lowe, 1975) and asymmetrical convolutions are common.

The 70 cm thick laminated shaly siltstone beds are bounded by sharp planar bedding planes and are ungraded.

Interpretation and Depositional Environment of the Basinal Sequence

Rapid subsidence of the shallow shelf, apparently without uplift of adjacent land areas, led to the abrupt establishment of deep water basinal conditions in the Ramah area.

Stagnant, quiet bottom conditions prevailed during initial basin formation and euxinic, pyrite and carbon-rich, black muds were deposited at the base of the Nullataktok Formation. Such conditions could form either in a barred basin or by density stratification (Byers, 1977). The presence of black pyritic slates at a comparable stratigraphic level in several other correlative Aphebian sequences in Labrador (Smyth and Knight, 1978) suggests that the environment was regional. Despite minor temporal variations, the pyrite-chert-dolomite-argillite association was probably deposited under similar basinal conditions. However, the pebbly massive pyrite which passes northward to bedded and then to laminated pyrite, as the bed dies out, suggests that the pyrite may have been in part resedimented, possibly as a mudflow or turbidite.

Quiet conditions suitable for the deposition of the thick colour-banded mudstones were prolonged. The planar, undeformed and continuous rhythmic layers were most probably deposited as pelagic muds. Variation of Eh and pH conditions in the basin, or compositional variation of the detritus, may explain the colour variation. As similar mudstones are interbedded with rubbly and fine grained pyroclastic rocks in the Mugford Group (Smyth and Knight, 1978) it is possible that volcanic dust may be a prime source of the detritus.

Thin beds of rusty, carbonate-rich sandstone and siltstone in the colour-banded mudstones are similar to thin distal turbidites (Walker, 1975, 1978) and suggest that a shelf environment was located outside the present basin. This interpretation is supported by the occurrence of channel-bound dolomitic breccias and massive sandstones in mudstones well below the first calcareous beds of the formation. These coarse sediments are similar to facies developed in deep water fans (Walker, 1975, 1978) and indicate that a sand and carbonate shelf environment existed adjacent to the basin.

Encroachment of thin planar, stratified, calcareous, dolomitic and noncalcareous mudstones over the deep basin deposits then occurred. These strata resemble shelf slope deposits (Cook and Taylor, 1977, Fig. 24) and deep water limestones of Utah (Bissell and Barker, 1977), and fulfill the deep water criteria of Wilson (1969). The nodular bedding and local in situ deformation as well as major slumps are also common in deep water slope deposits (Garrison and Fischer, 1969; Corbett, 1973; Cook and Taylor, 1977; Bissell and Barker, 1977; Walker, 1978) and indicate that slope deposits prograded over the basinal deposits. The thin stratification and ripple crosslamination suggest deposition of thin fine grained turbidites with sediment derived from a mixed silicic and carbonate shelf.

The breccias and sandstones at the base of Warspite Formation are comparable to the resedimented breccias and massive and pebbly sandstone facies of the deep water fan model of Walker (1975, 1978, 1979). Channels are envisaged to have cut across the shelf slope deposits and to have funnelled sandy dolomitic detritus into the basin to produce the deep sea fans. Steep-sided channels may have developed major rotational slumps that produced the matrix-supported breccias of Nullataktok lithoclasts at Reddick Bight. Alternatively, channels may have formed by enlarging earlier slump scars. Local breccia-filled channels as at Rowsell Harbour and near Ramah Bay, suggest that a fan-shaped network of small channels fed the main arteries (Fig. 18.12). The coarse detritus moved along the channels as mass flow (Hendry, 1972) or debris flows (Middleton and Hampton, 1973; Walker, 1975) choking the channels with debris before emerging into the basin where the deep water fan formed. The irregular distribution of the breccias and the possible canyon inland from Little Ramah Bay suggest that the shelf margin had an irregular, perhaps fault controlled, form. Westerly-directed paleocurrents and the change from ungraded channelled breccias in the east to graded pebbly sandstones and breccias in the west, suggest that the main channels pass westward into braided suprafan deposits (Walker, 1978; and Fig. 18.12). The associated finer grained sediments are similar to fine levee and interchannel fan deposits (Ricci-Lucchi, 1975; Walker, 1978).

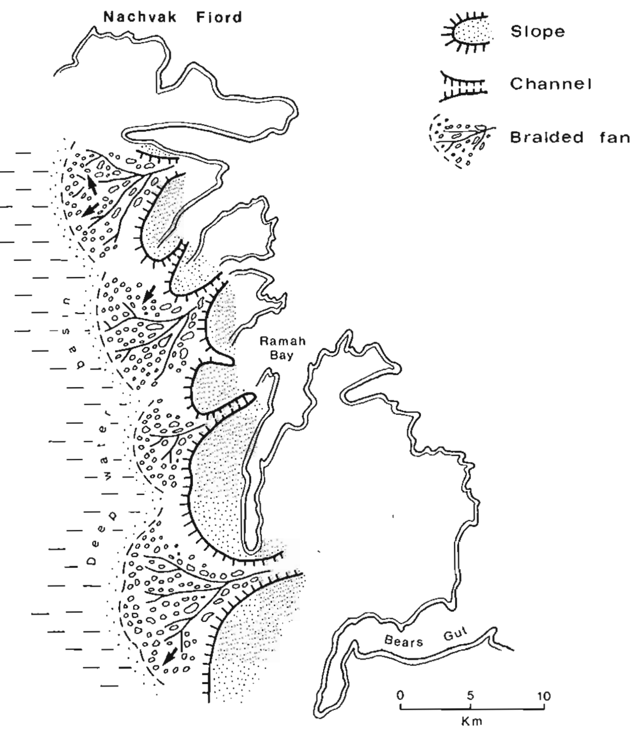


Figure 18.12. *Reconstruction of the depositional environment of the Warspite Formation. Arrows indicate paleocurrent directions.*

The influence of carbonate sedimentation, an adjacent shelf environment and tectonic instability diminished rapidly as quiet water muds were again deposited in the upper part of the Warspite Formation and in the Typhoon Peak Formation. This quiet phase of deposition was, however, gradually interrupted by a sequence of turbidites of the Cameron Brook Formation. The detritus in these sediments indicates a mixed basement and basic volcanic provenance. The freshness of the detritus plus the predominance of A→B→E and A→E turbidites, typical of a proximal environment (Walker, 1967), suggest a close source. The grits and pebbly beds near Little Ramah Bay suggest that the turbidites coarsen southeast, and indicate that the basic volcanic rocks of the Mugford Group may have been the source of the flysch.

SUMMARY

The 1700 m thick Ramah Group forms a linear north-trending outlier of chiefly sedimentary lower Aphebian strata preserved at the contact between Nain and Churchill Structural provinces in northern Labrador. The base of the group is marked by a spectacular unconformity which abruptly truncates basement rocks comprising steeply inclined, north trending, banded Archean gneiss crosscut by an intense west trending pre-Ramah diabase dyke swarm. A 1-12 m thick pre-Ramah regolith is characterized by decomposed basement gneiss and diabase dykes, spheroidal weathering, magnetite, hematite, jasper concretions and dolomitic zones.

The six formations of the group form two contrasting sequences: a basal siliciclastic sequence and an upper fine grained argillaceous and carbonate sequence. Deposition of the Ramah Group was preceded by regolith formation due to deep weathering in a humid subtropical to tropical paleoenvironment. The unconformity surface originally formed a gently northwest-inclined peneplain with a slightly undulose surface. Prior to the transgression of the basal Ramah, erosion had removed the regolith in the north, near Adams Lake, and a rocky shoreline had developed south of Bear's Gut.

The basal siliciclastic sequence represents shallow shelf sedimentation with several transgressive-regressive events. Rowsell Harbour Formation was deposited by dominantly marine and fluvial processes. Sedimentation commenced with the formation of a beach and barrier sand complex that migrated across the shelf as transgression advanced rapidly over the basement. The shelf remained shallow, and when transgression ceased it was covered by a thin blanket of

pebbly sandstone deposited by braided streams. The source area lay to the northeast and east, in the Greenland portion of the North Atlantic Archean craton. During this regressive stage a thin subaerial basalt flow was extruded over the braided sandy alluvial plain. During a second transgression a mud-dominated shelf, covered by shallow tidal sand flats and mud flats, developed. Deltas advanced northwards across this shelf, which then became deeper and again mud-dominated. Sedimentation of Rowsell Harbour Formation terminated with progradation of barrier island and sandy shoreline complexes across a fairly deep shelf.

The depositional environment of Reddick Bight Formation was influenced considerably by Reddick Arch, a topographic high that separated the basin into two parts. North of the arch, shallow intertidal black sand and mud flats evolved. South of the arch, however, a fluvial-dominated deltaic complex prograded northwestward across the shelf. Subaqueous turbidite fans formed at the foot of the delta front, and thin, sand-shale laminites were deposited as a deeper water pro-delta facies. Regression and blanketing of the shelf resulted. Shallow carbonate mud flat deposition then occurred in response to flooding of the deserted delta plain. Supratidal semi-arid or arid conditions prevailed at this time.

The 1000 m thick upper sequence of mud and silt argillaceous and carbonate strata record sedimentation under deep basinal conditions. As the basal black shales of Nullataktok Formation directly overlie the dolomite at the top of Reddick Bight Formation, these conditions must have been established rapidly, without coeval uplift of adjacent land areas. Pyrite and carbon-rich black muds, a pyrite-chert-dolomite-argillite association, and rhythmic layered pelagic muds were deposited in a stagnant, quiet, deep-water basinal environment. A silicic and carbonate shelf environment then developed in a distant area and supplied thin distal carbonate turbidites to the basin. Deep-water slope deposits followed and prograded over the basin. Dolomite breccias and sandstones in the lower part of Warspite Formation represent deep water fan deposits. Channels were cut westward across the shelf and slope deposits and these funnelled detritus into the basin as deep sea fans and braided suprafans. Quiet deep-water conditions returned to the area, and muds were deposited, forming the upper part of Warspite Formation and the Typhoon Peak Formation. Deposition of Cameron Brook Formation proximal turbidites, apparently terminated deposition of the Ramah Group.

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19. EARLY PROTEROZOIC SEQUENCES IN LABRADOR

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Richard J. Wardle and David G. Bailey, Early Proterozoic sequences in Labrador; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 331-359, 1981.

Abstract

Early Proterozoic (Aphebian) sedimentary and volcanic rocks in Labrador are described from two tectonic belts: the Churchill Province of western Labrador, and the Makkovik Subprovince of eastern Labrador.

The Aphebian in the Churchill Province is represented mainly by the Labrador Trough (Kaniapiskau Supergroup) but also in the Laporte Group, Lake Harbour Formation and Petscapiskau Group. The Labrador Trough is the most completely exposed of these sequences and displays a transition from shelf sedimentation in the west, to deeper water basinal conditions in the east. Rifting along the eastern margin of the Labrador Trough may have produced a narrow proto-oceanic rift during this stage.

Aphebian sequences in the Makkovik Subprovince are represented by the Moran Lake and Aillik groups. The Moran Lake and lower Aillik groups were deposited in environments broadly similar to those of the Churchill Province and in approximately the same time. The upper Aillik Group (1750-1670 Ma) represents a younger assemblage dominated by felsic volcanics intricately intruded by granites and may indicate the onset of radically different tectonic conditions towards the end of Aphebian time.

Résumé

La présente étude décrit les roches sédimentaires et volcaniques du Protérozoique récent (Aphébien), provenant de deux zones orogéniques au Labrador, soit la province de Churchill dans le Labrador occidental et la sous-province de Makkovik dans le Labrador oriental.

Dans la province de Churchill, l'Aphébien est surtout représenté par la fosse du Labrador (supergroupe de Kaniapiskau) mais aussi par le groupe de Laporte, la formation de Lake Harbour et le groupe de Petscapiskau. La fosse du Labrador est la mieux exposée de ces séquences; les sédiments représentent une transition d'une plate-forme dans l'ouest à un bassin plus profond dans l'est. Durant cette période, la formation de fissures le long de la marge orientale de la fosse aurait produit un fossé proto-océanique étroit.

Dans la sous-province de Makkovik, les groupes de Moran Lake et d'Aillik représentent les séquences aphébiennes. les sédiments des groupes de Moran Lake et d'Aillik inférieur ont été déposés à peu près en même temps que ceux de la province de Churchill et dans des environnements plus ou moins semblables. Le groupe d'Aillik supérieur (1750-1670 Ma) représente un assemblage plus récent, dominé par des sédiments volcaniques felsiques pénétrés par des granites; il pourrait indiquer la création, vers la fin de l'Aphébien, de conditions tectoniques très différentes.

GENERAL STATEMENT

Early Proterozoic (Aphebian) supracrustal sequences occur in a variety of metamorphic states in all tectonic provinces of Labrador and adjacent New Quebec. The most completely preserved sequences occur in the Churchill and Nain provinces, and in the Makkovik Subprovince (Fig. 19.1). This paper is concerned predominantly with those sequences of the Churchill Province, in particular the Labrador Trough, and the Makkovik Subprovince. Those of the Nain Province have been treated by Knight and Morgan (1981) and by Smyth and Knight (1978). Extensive areas of Aphebian supracrustal rocks are also present in the Grenville Province, but in general their complexly deformed and metamorphosed state precludes detailed stratigraphic analysis. The Churchill Province (R. Wardle) and Makkovik Subprovince (D. Bailey) are discussed separately in Parts 1 and 2 of this paper. In the Churchill Province, emphasis has been placed on the Labrador Trough, which provides the best preserved Aphebian sequence in Labrador and provides the most insight into Early Proterozoic basinal processes in this area.

At present, insufficient information is available to allow development of a comprehensive tectonic model unifying the Churchill Province and Makkovik Subprovince. Models for each region, therefore, are discussed individually.

PART I – EARLY PROTEROZOIC OF THE CHURCHILL PROVINCE: THE LABRADOR TROUGH

Introduction

The Churchill Province of Labrador-New Quebec is a mobile belt of Hudsonian age (1750-1800 Ma; Stockwell, 1972), forming a southern splay off the main Churchill Province of western Canada and the Arctic Islands. The belt is bounded to the east and west by the stable Archean cratons of the Nain and Superior provinces. Aphebian sequences of the Churchill Province are shown in Figure 19.1. The principal area of sedimentary and volcanic rocks occurs in the Circum-Ungava Foldbelt or Geosyncline (Dimroth et al., 1970), which forms the western margin of the Churchill Province and lies unconformably on the rim of the Superior Province. The southeastern part of this belt is termed the

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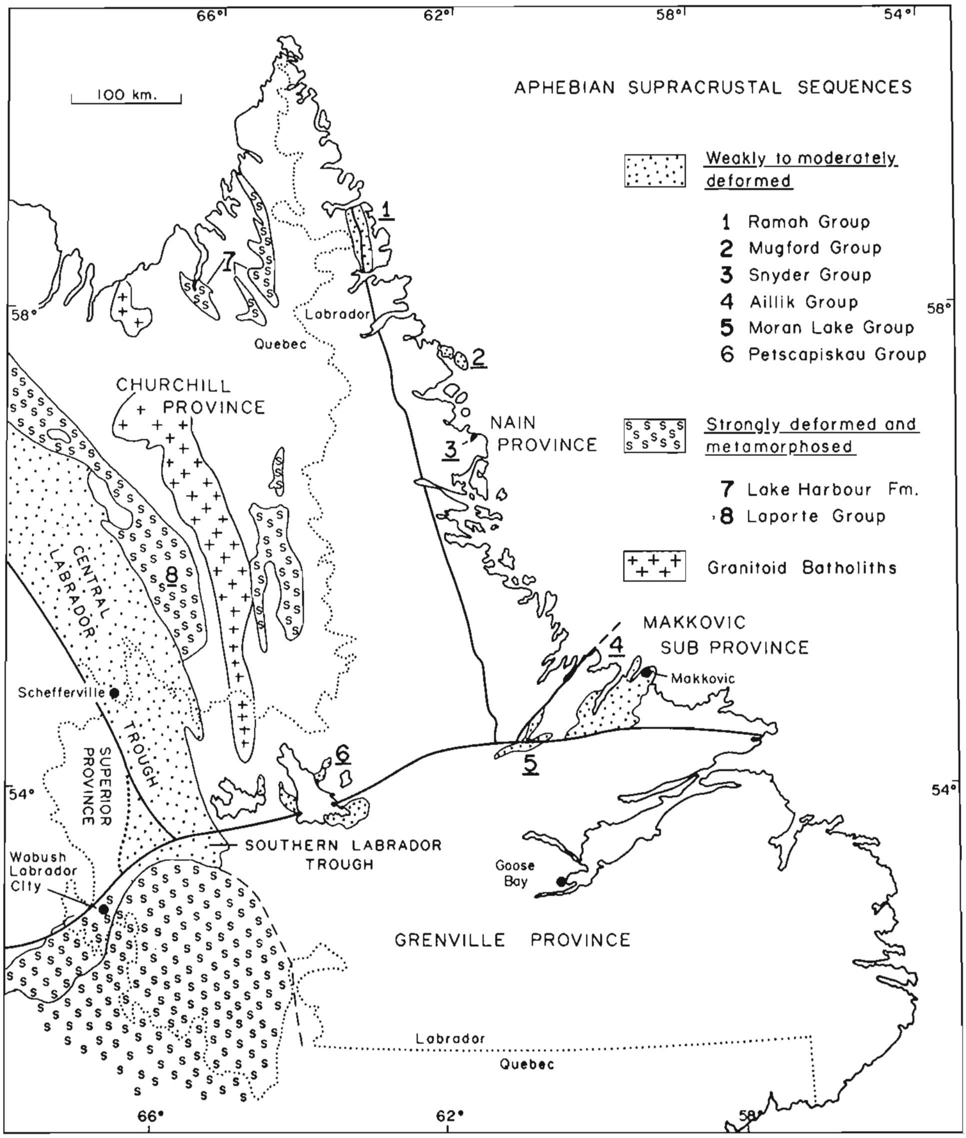


Figure 19.1. *Early Proterozoic (Aphebian) sequences of Labrador and adjacent New Quebec.*

structures is present in the Ramah Group on the eastern margin of the Churchill Province where the direction of thrusting and overturning was eastwards, onto the Nain Province craton. The interior of the Churchill is largely underlain by granitic gneisses with diverse structural trends defining basinal and domical structures. The age of these rocks is largely unknown; in large part they probably represent remobilized Archean basement (e.g. Dimroth, 1964; Dimroth et al., 1970; Wardle, 1979a). Undoubtedly though, they also include some highly metamorphosed and migmatized Aphebian granites and supracrustals (e.g. Taylor, 1979). Where exposed along the eastern margin of the Labrador Trough the Archean gneisses are grouped in the Wheeler Complex (Dimroth, 1978) and Eastern Basement Complex (Wardle, 1979a).

A further important feature of the Churchill Province is a belt of late kinematic granitoid batholiths (Taylor, 1979) which form an axial zone 450 km long (Fig. 19.1).

The following discussion is concerned largely with the stratigraphy and basinal development of the Labrador Trough and Laporte Group. The other sequences of the Churchill Province are, in general, too highly deformed and poorly known to permit detailed analysis.

Labrador Trough. An extensive sequence of more highly metamorphosed supracrustal rocks occurs in the Laporte Group adjacent to the eastern margin of the Trough.

Small units of highly metamorphosed supracrustal rocks (marble, metaquartzite, calc-silicate, amphibolite and pelitic gneiss) occur scattered throughout the central Churchill Province, some of which in northeastern New Quebec have been correlated with the Lake Harbour Group of Baffin Island (Jackson and Taylor, 1972) and named the Lake Harbour Formation (Taylor, 1979). A small sequence of pelitic schist at the southern end of the province is known as the Petscapiskau Group (Emslie, 1970).

Clastic and local mafic volcanic rocks also comprise the Ramah Group which lies largely in the Nain Province of eastern Labrador (Knight and Morgan, 1981). The eastern margin of this group, however, has been affected by Hudsonian deformation and therefore lies in the Churchill Province.

Structural trends within the Churchill Province are dominantly northwest-southeast. The western margin is characterized by steep, easterly-dipping thrust faults, and folds overturned to the west. A mirror image of these

Labrador Trough (Including Laporte Group)

The Labrador Trough (2150 Ga – 1800 Ga; Dimroth, 1972; Fryer, 1972) has been divided into three geographic segments by Dimroth et al. (1970): the Northern Trough, north of 57°N; the Central Trough between 57°N and the Grenville Front; and the Southern Trough, south of the Grenville Front.

The stratigraphy of the Trough is most completely developed in the central segment, and it is with this region that the following discussion is concerned. The northern part of this area, referred to as the North-Central Trough (Fig. 19.2) has been extensively described by Dimroth (1968, 1970, 1971a, b, 1972, 1978) and Dimroth et al. (1970). The regional geology of the South-Central segment (that part south of 55°15'N) has been established by Frarey (1961), Baragar (1967), Wynne-Edwards (1960, 1961) and Fahrig (1967) and, more recently, by Wardle (1979a), Evans (1978), and Ware and Wardle (1979).

The Labrador Trough comprises a succession of sedimentary and mafic volcanic rocks, the Kaniapiskau Supergroup, which is intruded by a sequence of gabbro and ultramafic sills, approximately 6000 m thick, of the Montagnais Group (Frarey and Duffell, 1964; Baragar, 1967). The Kaniapiskau Supergroup (Fig. 19.2) is divided into two distinct, lithic assemblages: a western, predominantly sedimentary succession, the Knob Lake Group (6500 + m), and an eastern, predominantly mafic volcanic unit, the Doublet

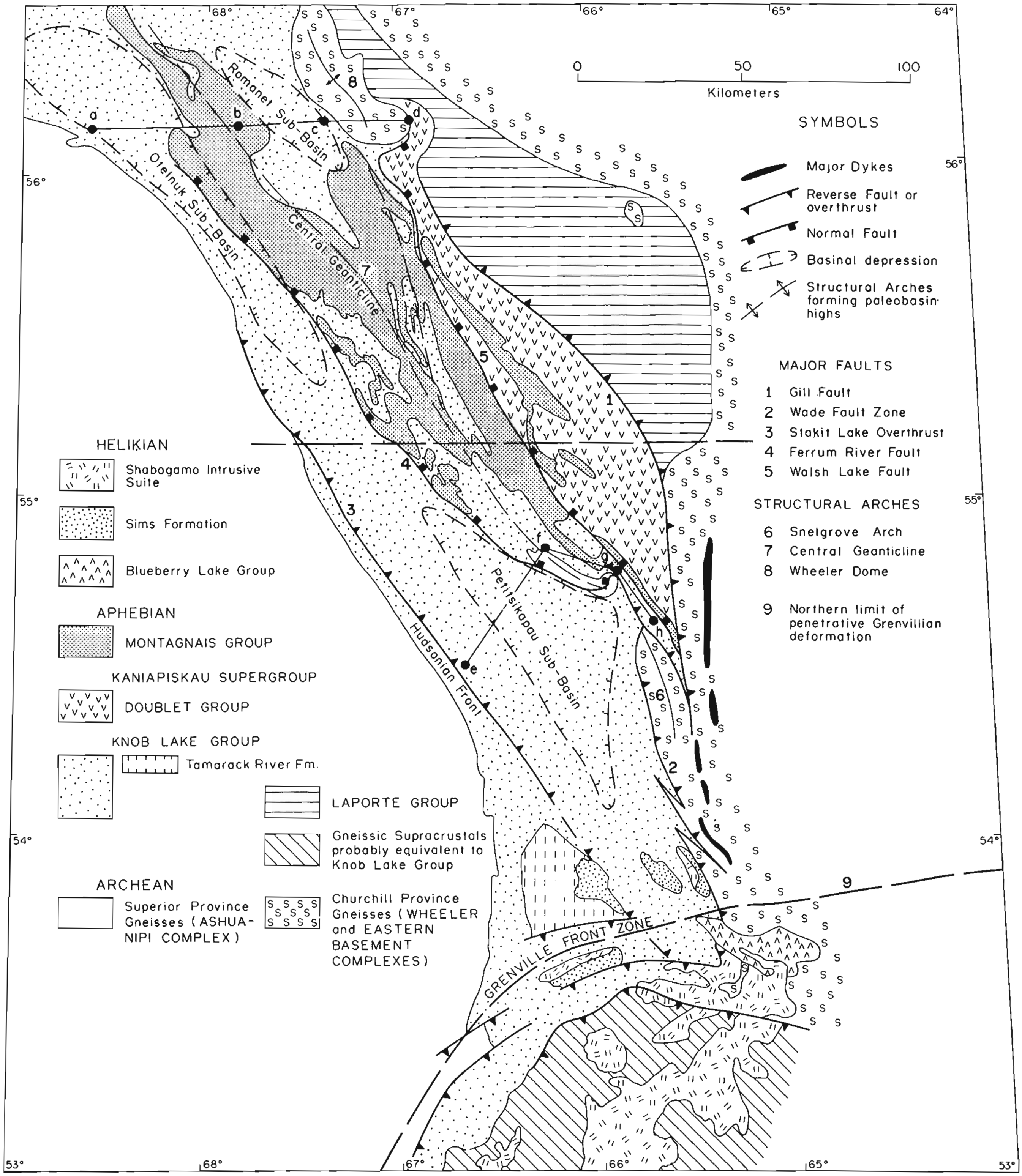


Figure 19.2. *General geology of the Central Labrador Trough. Dashed line divides north-central and south-central parts.*

Group (5000 + m). The Knob Lake Group may in turn be divided into a western zone containing predominantly shallow water sediments, and an eastern zone dominated by deep-water sediments and abundant mafic volcanics. The Doublet Group locally conformably overlies the Knob Lake Group on the eastern margin of the Trough but the two are generally in tectonic contact along the Walsh Lake Fault (Fig. 19.2). A thick succession of pelitic to semi-pelitic schist and amphibolite, termed the Laporte Group, is in overthrust contact with the eastern margin of the Doublet Group. Dimroth et al. (1970), did not consider the group to be part of the Labrador Trough, largely because of its uncertain stratigraphic position. Dimroth (1978), however, has recently demonstrated the group to be laterally equivalent to the major part of the Knob Lake Group, and it is evident that it should be considered as an integral component of the Labrador Trough.

The Labrador Trough has often been considered as a miogeosynclinal-eugeosynclinal couple (e.g. Harrison et al., 1972) with the western part of the Knob Lake Group representing a shallow shelf and the eastern part of the Knob Lake Group and the Doublet Group an offshore basinal equivalent. It is now clear, however, that this concept has to be modified since it is the Laporte rather than the Doublet Group which forms the offshore component. The Doublet Group is a younger sequence developed across the Knob Lake – Laporte Group transition.

In the following sections, the basinal evolution of the Trough is divided into several stages, each of which represents a distinct lithofacies assemblage and tectonic environment. Emphasis throughout is placed on facies patterns rather than formal stratigraphic division. Stages 1-9 deal with the evolution of the shelf (Knob Lake Group) and its transition into deeper water lithologies to the east (Laporte Group). Stage 10 concerns the development of the Doublet Group. Each stage is summarized in facies maps and palinspastic sections.

The Shelf Sequence (Knob Lake Group)

The stratigraphy of the Knob Lake Group is summarized in Figure 19.3. Terminology for the South-Central Trough is that of Frarey and Duffell (1964) modified by Wardle (1979a), Evans (1978), and Ware and Wardle (1979); that for the North-Central Trough is from Dimroth (1978).

As proposed by Dimroth et al. (1970), the Knob Lake Group is informally divided into two segments, here termed the upper and lower Knob Lake group. Both begin with fluviatile or intertidal terrigenous clastics, pass through an interval of shallow marine shales, carbonates, or chemical precipitates and culminate with deep water turbidites.

Deposition of the lower Knob Lake Group was followed by broad regional warping. While sedimentation was continuous in the axial region of the Trough, uplift and erosion occurred at the margins. As a result, the upper Knob Lake Group disconformably oversteps earlier units and locally rests directly on basement at the Trough margins.

Deposition of both segments was strongly influenced by uplift of two major basement structures which acted as intermittent basin arches. The most important of these structures is a linear arch located along the eastern margin of the Knob Lake Group. The basement gneisses underlying this structure are exposed in two culminations of the arch referred to as the Wheeler Dome and Snelgrove Arch (Fig. 19.2). West of these structures, the entire Knob Lake Group is dominated by shallow water clastics, shale, carbonate and chemical precipitates. These structures define the eastward limit of this environment and were themselves the site of shallow water environments. East of the structures, the Knob Lake Group is largely buried beneath thick Doublet Group cover, but where exposed east of the Snelgrove Arch, it consists of a thick shale-siltstone succession apparently deposited in moderately deep water. The arch, therefore, marks the break between shallow water shelf deposition in the west and basinal or basin slope deposition to the east.

The other intrabasinal arch, termed the Central Geanticline (Dimroth, 1968), was more short-lived and aerially restricted to the North-Central Trough. The structure is discordant to the general trend of the Trough and merges with the Snelgrove Arch (Fig. 19.2). The structure was first an active source area during the early stages of the Knob Lake Group and was covered during later transgression.

The basinal evolution of the Knob Lake Group shelf sequence is discussed under the following 9 stages.

Stage 1 – Continental Rifting (Lower Seward Subgroup)

Development of the Knob Lake Group began with deposition of a thick (1500-3000 m) continental clastic sequence throughout the Central Trough. In the north, the sequence is known as the Chakonipau Formation (Dimroth, 1968) and in the south as the "Discovery Lake and Snelgrove Lake formations" (new terms proposed by Wardle in a manuscript in preparation).

In both north and south, the rocks are very similar and consist of red and grey crossbedded arkoses, quartz granule conglomerates, and local pebble-boulder conglomerates. Most clasts are derived from adjacent basement terranes. Locally, andesite clasts in the North-Central Trough indicate a contemporaneous volcanic source (Fig. 19.4a, b). Baragar (1967) has also described trachyandesites and alkaline trachy-basalt flows interbedded with the Chakonipau Formation.

Festoon and planar crossbedding are abundant in sandstones and conglomerates in the south-central area. Unimodal paleocurrents and the lack of typical point bar cycles have been used to infer deposition in a gravelly, braided fluviatile regime (Wardle, 1979a). Scarce interbeds of mudcracked shale and dolomite concretions (caliche?) provide evidence of local flood plain deposition, or filling of abandoned channels. Granule conglomerate beds within the sequence are probably longitudinal gravel bars. A similar braided fluviatile depositional environment has been inferred for the Chakonipau Formation in the north-central area (Dimroth, 1978). In this same area, a thick fanglomerate sequence developed as an alluvial fan on the northern scarp of an inferred east-west fault (Dimroth et al., 1970). The fault defines the southern margin of a structure known as the Castignon Lake graben (Fig. 19.4a), which funneled detritus into the Trough from the west (Dimroth et al., 1970).

Paleocurrent data (Wardle, 1979a) in the south-central area suggest a source area for trough-filling sediments near the present southern end of the Trough (Fig. 19.4a), an interpretation which is strengthened by the presence of cobble conglomerates in this area and the absence of the Seward Subgroup to the south.

The thickness of the clastic sequences, and their apparent restriction to the axial region of the Trough, suggest deposition in a north-northwest trending rift valley, an interpretation strengthened by the presence of synsedimentary alkaline volcanics. The valley was probably fed from both northern and southern ends by a system of braided alluvial fans. The east-west Castignon Lake graben at the north end probably represents a subsidiary rift, similar to those in modern rift systems (e.g. East African Rift). However, there is no evidence to suggest the graben was an aulacogen or failed arm, as suggested by Burke and Dewey (1973).

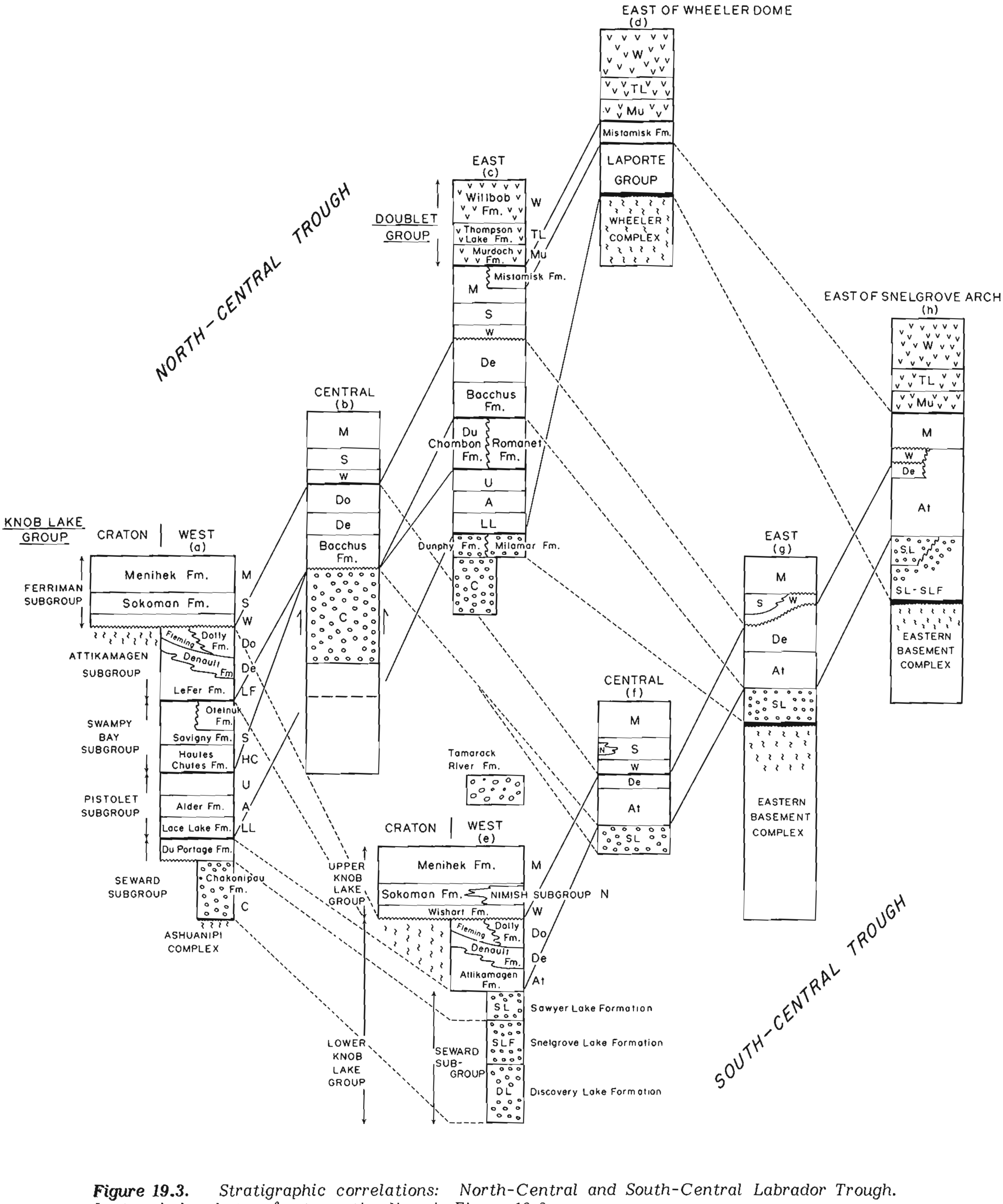
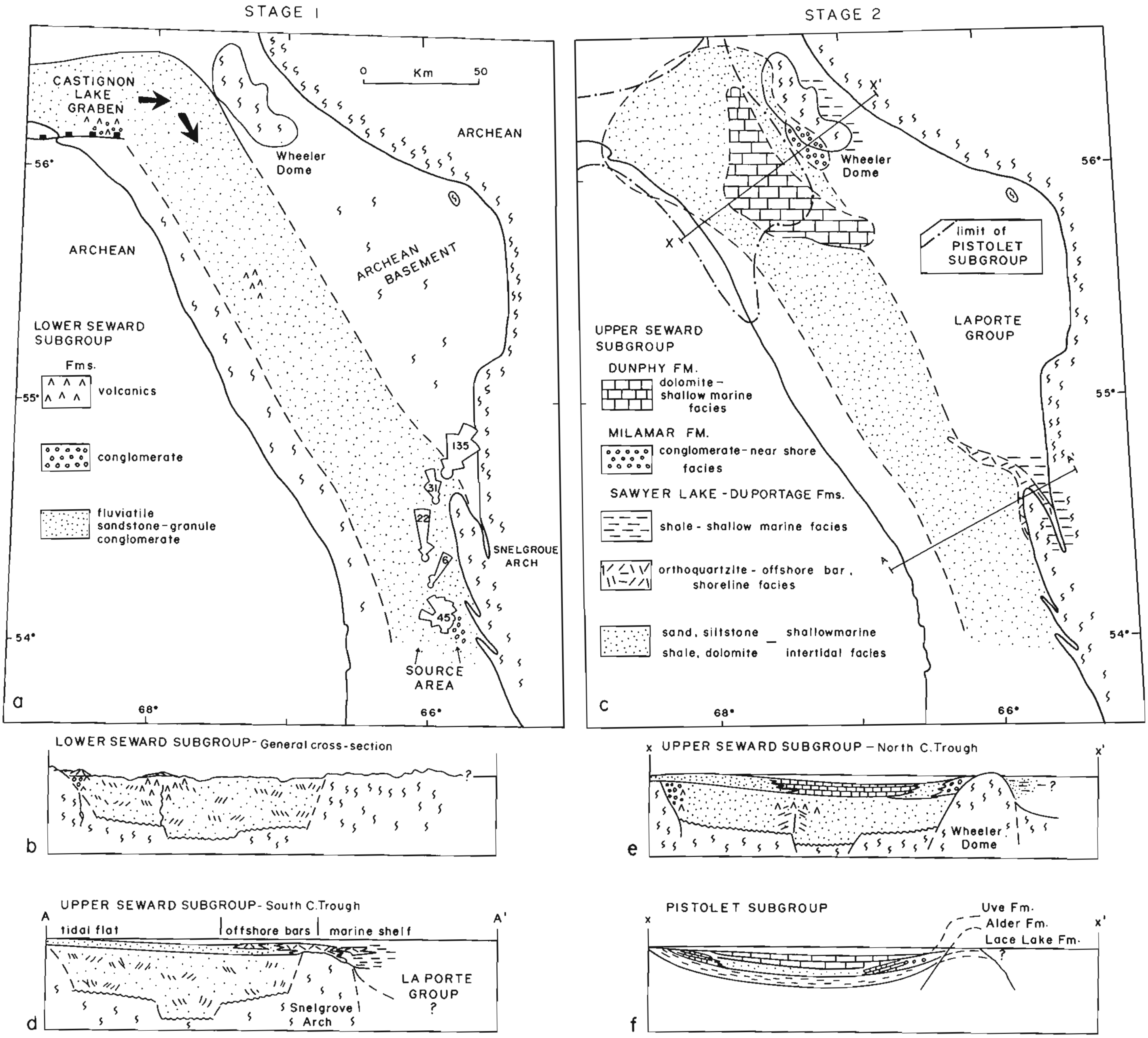


Figure 19.3. *Stratigraphic correlations: North-Central and South-Central Labrador Trough. Letters in brackets refer to section lines in Figure 19.2.*



a,b) Stage 1: *Lower Seward Subgroup. Deposition in braided fluviatile system. Paleocurrent directions measured on crossbedding (Wardle, 1979a).*

c,d,e) Stage 2: *Upper Seward Subgroup (Du Portage, Dunphy and Milamar formations, in North-Central Trough; and Sawyer Lake Formation in South-Central Trough). Note transition from intertidal to shallow marine deposition in south and presence of shallow marine carbonate basin in north.*

c,f) *Pistolet Subgroup (Lace Lake, Alder and Uve formations). Deposition restricted to shallow marine carbonate-clastic basin in North-Central Trough.*

Figure 19.4. *Lithofacies distribution and palinspastic cross-sections, Labrador Trough.*

Stage 2 – Marine Transgression and Development of Shelf Environment (Upper Seward Subgroup and Pistolet Subgroup)

This stage marks the filling of the rift valley and the beginning of a shallow marine transgression across the Trough.

In the South-Central Trough (Wardle, 1979a), this stage is represented by the "Sawyer Lake formation" (new term proposed by Wardle in a manuscript in preparation), a lithologically variable unit consisting of a 300 m thick red and purple shale-siltstone facies in the west; a 300-500 m thick sandstone-orthoquartzite facies around the Snelgrove Arch; and a shale-phyllite-siltstone sequence over 800 m thick east of the arch. The transition is interpreted as a progression from tidal flat deposition in the west, through a system of offshore sandbars or strand line deposits around the Snelgrove Arch, into a basinal environment in the east (Fig. 19.4c, d).

In the North-Central Trough, this stage is represented by a more complex sequence comprising the upper Seward Subgroup and the Pistolet Subgroup, both deposited in a minor basin with a depocentre located west of the Wheeler Dome (Fig. 19.4c, e). The upper Seward Subgroup consists of a western sequence of sandstones and stromatolitic dolomite (Dunphy Formation) which passes eastwards into fine grained arkoses (Du Portage Formation), and then into arkoses, conglomerates and quartzites (Milamar Formation). These sequences accumulated in environments ranging from shoreline in the west and east, to shallow marine in the basin centre. Wind-borne sand in the western shoreline facies (Dimroth, 1968) indicates the probable existence of terrestrial environments west of this area.

Upper Seward Subgroup formations are overlain by the Pistolet Subgroup (700 m, Dimroth, 1978), which is restricted to the North-Central Trough (Fig. 19.4e, f). Deposition commenced with shallow water shales (Lace Lake Formation) followed by accumulation of stromatolitic dolomite, dolomitic sandstone and quartzite (Alder Formation), then by dark, massive dolomite (Uve Formation). Sedimentation in the basin centre was dominated by a shallow marine-intertidal carbonate environment, interfingering towards the basin margins with near shore and shoreline clastics (Fig. 19.4f).

The Wheeler Dome and Snelgrove Arch probably acted as locally emergent highs during this stage and separated shallow water – shoreline environments in the west from deep-water conditions in the east. Deep-water deposits are apparently represented by the thick metashale – siltstone sequence (Sawyer Lake Formation) east of the Snelgrove Arch and by pelitic schists of the equivalent Laporte Group east of the Wheeler Dome (Fig. 19.4c, d, e).

Stage 3 – First Stage of Subbasin Development on Shelf (Swampy Bay Subgroup)

This stage was also restricted to the North-Central Trough. Uplift of a central block or arch, termed the Central Geanticline (Dimroth et al., 1970) was accompanied by subsidence in the marginal Otelnuk and Romanet subbasins (Fig. 19.5a).

Deposition in both subbasins was dominated by rhythmically bedded shale, siltstone and quartz wacke turbidites, but conglomerates are locally present on the eastern side of the Romanet Subbasin. Turbidites of the central zones of the subbasins laterally interfinger with shallow water sandstones, shales and conglomerates towards the margins (Dimroth, 1978; Fig. 19.5b).

The Otelnuk and Romanet subbasins were later, as now, separated by uplifted Chakonipau Formation. The source of the Swampy Bay sediments was therefore probably upper Seward and Pistolet subgroups. We speculate that development of these structures was fault-controlled and related to accelerated rifting on the shelf.

Stage 4 – First Collapse of Shelf (Attikamagen, Bacchus and Le Fer formations)

This stage signifies a temporary halt to shallow-water clastic – carbonate deposition and the prograding incursion of shales and deep-water mafic volcanics, concomitant with shelf collapse.

In the South-Central Trough, this stage is represented by the Attikamagen Formation and is divisible into two distinct depositional lithofacies: a western sequence of grey, yellow and green shales and siltstones; and an eastern sequence of grey shale, siltstone and greywacke interbedded with mafic pyroclastics and thick basaltic lavas (Fig. 19.5a). Limited data suggest that the formation thickens from about 500-1000 m in the west to about 2000-3000 m in the east.

The western sequence is characteristically thinly bedded with abundant ripple cross-lamination and was deposited in shallow to moderate water depths. The eastern facies contains greywacke and siltstone turbidites deposited in a considerably deeper-water environment (Fig. 19.5d).

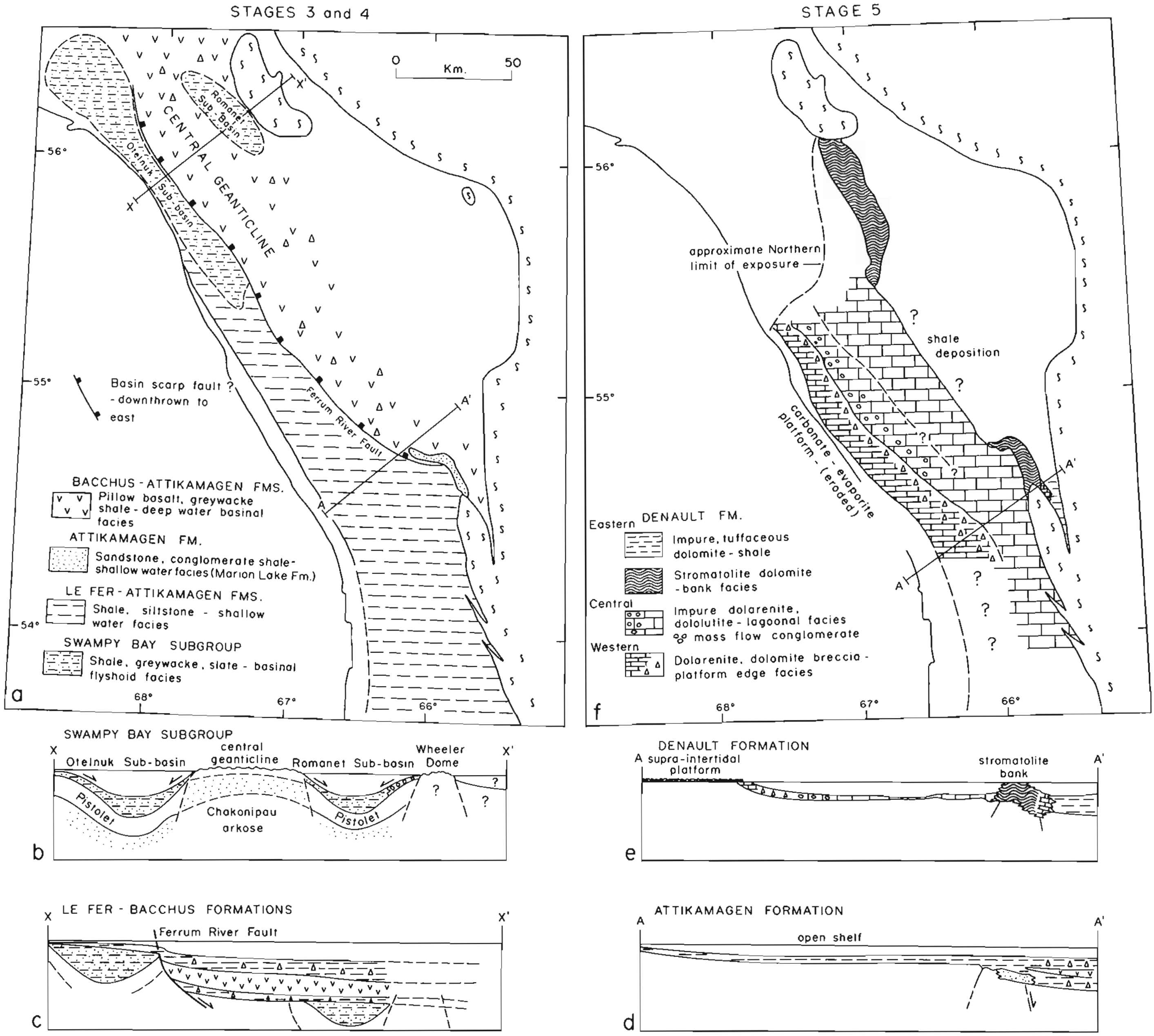
Eastern facies volcanics occur as a single unit of tholeiitic flows about 400 m thick. The flows are predominantly massive but are locally pillowed. Pillows are typically nonvesicular and show little internal structure in the form of radial joints, variolites, cavities, etc. Comparisons with textures seen in modern pillow basalt (e.g. Wells et al., 1979) suggest that they formed in intermediate water depths below 1000 m. The massive nature of many of the flows is not incompatible with deposition in this environment and probably indicates rapid eruption (Dimroth et al., 1978).

The flows and overlying sediments have been intruded by numerous basalt, diabase and fine grained gabbro sills assigned to the Montagnais Group. These are petrographically and chemically similar to the lavas and are generally interpreted as comagmatic (Baragar, 1967; Dimroth; 1971a; R.A. Doherty, personal communication). The general fine grained nature of the sills and the lack, or restricted nature, of contact metamorphism in the host sediments suggest intrusion at very high levels.

A unique sedimentary facies developed around the northern edge of the Snelgrove Arch (Fig. 19.5a) has previously been termed the Marion Lake Formation (Donaldson, 1966). The facies is a thick sequence (approximately 1000 m) which occurs in the middle Attikamagen Formation and comprises shallow water impure sandstones, conglomerates and shales. The sequence is interpreted as a local nearshore deposit developed around the margin(s) of the Snelgrove Arch as it subsided. East of the arch, the Attikamagen Formation consists of a "normal" thick sequence of shale, siltstone (turbidites?) and mafic volcanics (Fig. 19.5a, d).

The division between the western shallow water facies and the eastern basinal facies is abrupt and is defined by the steep, easterly dipping, Ferrum River Fault (Fig. 19.5a), a structure which also marks the westward limit of Montagnais Group gabbro sills. The abrupt facies change across the fault suggests that it may have been active as a syndepositional basin scarp fault, with downthrow to the east (Fig. 19.5d). The fault was subsequently modified to a reverse fault during deformation.

Similar facies relationships occur in the North-Central Trough (Fig. 19.5a, c), where the intermittently-exposed, shallow-water western facies, termed the Le Fer Formation (Dimroth, 1978) is transitional eastward into the thick Bacchus Formation slates, sandstones, tuffs, hyaloclastites



a,b) Stage 3: *Swampy Bay Subgroup. Restricted to two flysch basins in North-Central Trough: the Otelnuk Subbasin (Hautes Chutes, Savigny and Otelnuk formations) in the west and the Romanet Subbasin (Du Chambon and Romanet formations) in the east. Subbasins separated by Central Geanticline source area.*

c,d) Stage 4: *LeFer, Bacchus and Attikamagen formations. Ferrum River Fault interpreted as basin scarp fault separating shallow water shale environment in west from deep water shale – greywacke – mafic volcanic environment to east.*

e,f) Stage 5: *Denault Formation. Shallow basin containing impure, muddy carbonates is bounded to east by intertidal, stromatolite bank; and to west by an inferred supra-intertidal, carbonate-evaporite platform. Edge of platform marked by slump breccias and debris flows.*

Figure 19.5. *Lithofacies distribution and palinspastic cross-sections, Labrador Trough.*

and massive to pillowed basaltic lavas (Dimroth, 1978). The Bacchus lies unconformably across the Central Geanticline but may be conformable with parts of the Swampy Bay Subgroup in the Romanet and Otelnuk subbasins. Dimroth (1978) has shown the Le Fer and Bacchus formations to be separated by a reverse fault. In Figures 19.2 and 19.5a, c, the fault is inferred to be the northern continuation of the Ferrum River Fault.

Both the Bacchus and Attikamagen formations appear to pass eastwards into the pelitic-amphibolitic schists of the Laporte Group. It is inferred, therefore, that the Snelgrove Arch and Wheeler Dome were largely passive features at this time and only locally influenced sedimentation.

Stage 5 – Shelf Re-established as Shallow Carbonate Environment (Denault Formation)

The Denault Formation dolomite marks the re-establishment of shallow water environments across the shelf and provides an excellent picture of the basinal geometry of the Trough at this stage. The Denault is best developed in the South-Central Trough and pinches out between 55° and 56°N (Fig. 19.5f). The formation is divisible into western, central, and eastern depositional facies; a fourth facies is also inferred to have existed west of the present limit of the Trough (Fig. 19.5e, f).

Western Facies. The western facies (500-600 m) is composed of thin-bedded, rarely graded, dolarenites with incomplete Bouma sequences, interbedded with lensoid units of coarse dolomite breccia 1 to 10 m thick. Clasts in the breccias were derived from a variety of dolarenites and vary from angular in the west to subangular and subrounded in the east. The dolomite-breccias are interpreted to have formed by slumping (debris flows) on a steep, easterly dipping slope, adjacent to a supra-intertidal platform associated with evaporites. Evidence for the existence of this platformal facies is derived from the overlying Fleming Formation and is discussed further below.

Central Facies. The central facies (200 m) comprises brown, muddy, commonly crossbedded and slump folded dolomites, interbedded with massive diagenetic dolomite. The thickness of the central facies is highly variable, however, and the formation locally pinches out between underlying and overlying shale. The facies is interpreted as having formed in shallow basinal conditions (Fig. 19.5e).

Eastern Facies. The eastern facies is a thick (1000-3000 m) accumulation of stromatolitic dolomite which formed a discontinuous carbonate bank along the eastern margin of the Trough. Tepee structures, pisolite and oolite horizons, and intraclastic rip-up breccias of algal laminite, indicate deposition in an intertidal-supertidal environment, subject to storm agitation (Donaldson, 1963, 1966; Dimroth, 1971b; Wardle, 1979a). Interbedded units of crossbedded dolarenite probably represent channel sequences within this environment.

The stromatolite banks extend out from the Wheeler Dome and the Snelgrove Arch. These areas may have been partially emergent at this time. Whether the banks were connected is uncertain, since faulting has removed much of the intervening Denault Formation. Small scattered exposures of thin-bedded, non-algal, dolomite, however, indicate that the two banks were probably separated by deeper water, subtidal regions.

East of the Snelgrove Arch the stromatolitic dolomite is transitional into a discontinuous sequence (50-700 m thick) of impure dolomite interbedded with mafic tuff. The Denault Formation is not recognized east of the Wheeler Dome but marble and calc-silicate units in the Laporte Group to the east may be its lateral equivalents. Some of the shales assigned to the Attikamagen Formation east of the Snelgrove Arch may also in fact be equivalents of the Denault (Fig. 19.5e)

The thick stromatolitic facies was probably developed over a subsident zone which may have marked the shelf-basin slope break at this time. It separated a shallow shelf-lagoon environment in the west from a deeper-water basinal environment in the east. Stromatolite growth was apparently able to keep pace with subsidence and continuously maintained an intertidal environment.

Stage 6 – Second stage of Subbasin Development on Shelf; Broad Warping, Uplift and Erosion (Dolly and Fleming formations)

Carbonate deposition was followed by warping of the shelf in the South-Central Trough into a major central subbasin (Petitsikapau Subbasin) in which the Dolly Formation accumulated (Fig. 19.6a, b); and a minor western subbasin in which the Fleming Formation was deposited.

Dolly Formation. The Dolly Formation (Harrison et al., 1972; Dimroth, 1978) has been recognized over a large area (Wardle, 1979a) and defines an area termed the Petitsikapau Subbasin (Fig. 19.6a). At the subbasin margins, predominantly crosslaminated red, green and black shales and minor siltstone accumulated in a shallow-water environment. Toward the basin centre, their equivalents are rhythmically-bedded grey shale and siltstone of turbidite origin (Fig. 19.6b). In many respects, the Dolly Formation is similar to the Swampy Bay Subgroup.

Fleming Formation. The Fleming Formation is a distinctive chert breccia unit restricted to the western margin of the South-Central Trough, and occupies a stratigraphic position equivalent to the Dolly Formation. However, the two formations do not interfinger and we interpret the Fleming to be slightly younger than the Dolly Formation.

In central and western areas, the Fleming consists of massive to thick-bedded chert breccia of pebble- to boulder-sized clasts of colloform chert and drusy quartz set in an impure quartzite-chert matrix. Slabs of colloform chert up to several metres long have also been noted (Wardle, 1979b). In eastern exposures the formation consists of well-defined (1-2 m) breccia beds interbedded with quartzite and siltstone. All chert clasts are clearly void-fill structures derived from a pre-existing terrain.

Previously, the Fleming has been interpreted both as a residual karstic breccia formed by erosion of silicified Denault Formation (in Harrison, 1952) and as a submarine breccia produced by collapse and subsequent silicification of an interbedded sequence of siliciclastics and carbonates (Howell, 1954; Dimroth, 1971b). Dimroth (1978) recognized length-slow chalcedony in the clasts and proposed a replaced evaporite origin for the sequence (Pittman and Folk, 1971). Brecciation was proposed to have resulted from a combination of solution collapse and slumping. The authors agree with an evaporitic origin for at least part of the sequence and propose that the source must have been a western facies of the Denault Formation which was eroded during upwarp of the western margin of the Trough and whose representatives occur only as clasts in the Fleming Formation. However, the chaotic nature of the sequence is believed to be the result of large scale slumping, possibly as an olistostrome sheet, rather than solution collapse. The necessary slope instability required by this mechanism would have been provided by the

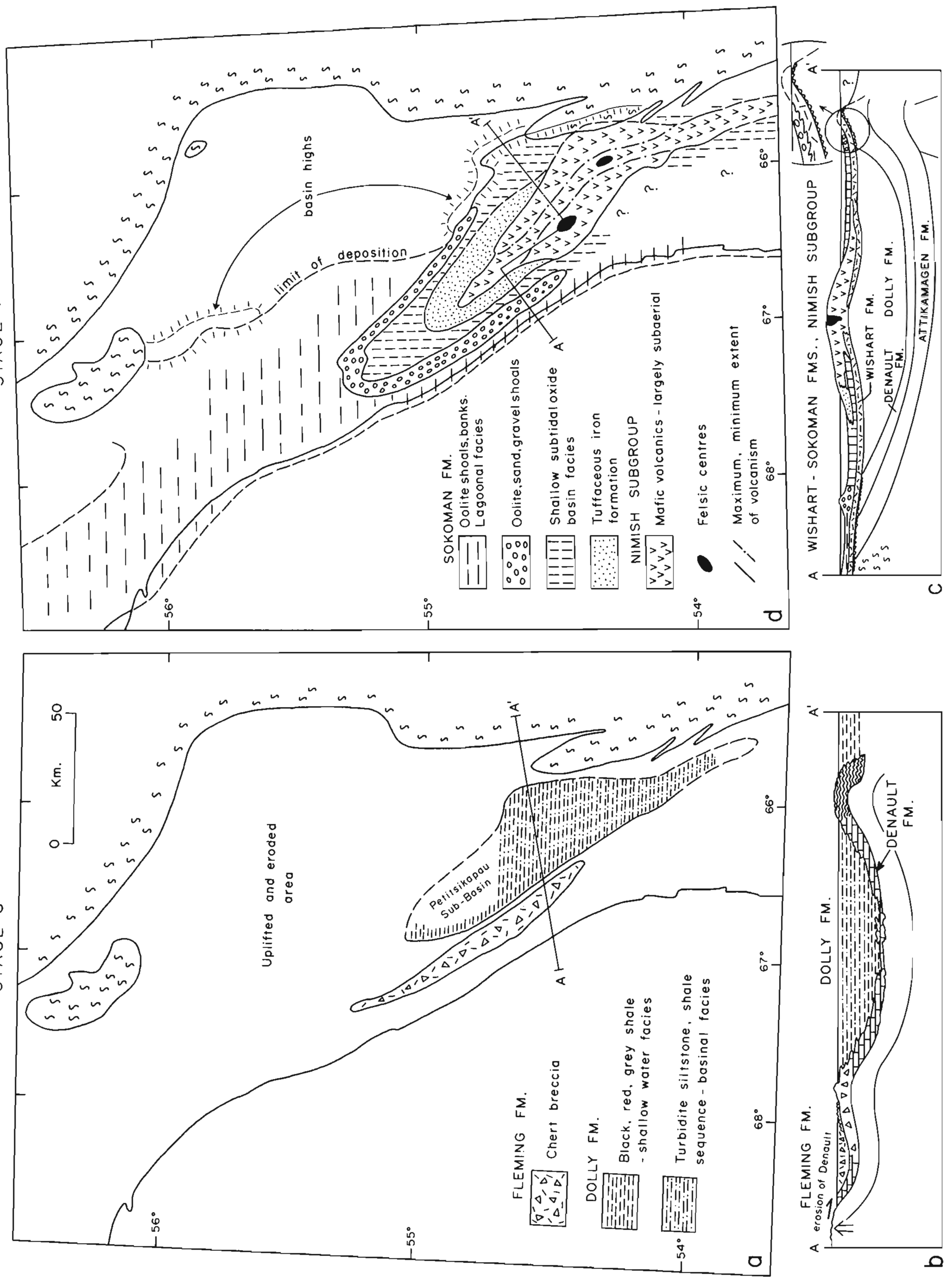

STAGE 6
Uplifted and eroded area
0 Km. 50
Petitsikapau Sub-Basin
FLEMING FM.
Chert breccia
DOLLY FM.
Black, red, grey shale - shallow water facies
Turbidite siltstone, shale sequence - basinal facies
56°
55°
54°
68°
67°
66°
a
A
A'
FLEMING FM.
erosion of Denault
DOLLY FM.
DENAULT FM.
b
STAGE 7
basin highs
limit of deposition
SOKOMAN FM.
Oolite shoals, banks. Lagoonal facies
Oolite, sand, gravel shoals
Shallow subtidal oxide basin facies
Tuffaceous iron formation
NIMISH SUBGROUP
Mafic volcanics - largely subaerial
Felsic centres
Maximum, minimum extent of volcanism
d
WISHART - SOKOMAN FMS., NIMISH SUBGROUP
WISHART FM.
DENAULT FM.
DOLLY FM.
ATTIKAMAGEN FM.
c

Figure 19.6

a,b) Stage 6: *Fleming and Dolly formations. Fleming Formation interpreted to be derived by erosion of silicified Denault carbonate-evaporite platform. Dolly Formation restricted to Petitsikapau Subbasin, locally interdigitates with Denault Formation.*

c,d) Stage 7: *Wishart and Sokoman Formations; Nimish Subgroup. Wishart forms a transgressive sandstone-siltstone blanket of littoral-shallow marine origin, and lies disconformably upon most older units. Sokoman Formation facies show a concentric deepening towards the Nimish volcanics. Volcanics are largely subaerial and developed over the previous Petitsikapau Subbasin. Centre of volcanism moved westward during Sokoman deposition.*

Figure 19.6. *Lithofacies distribution and palinspastic cross-sections, Labrador Trough.*

broad warping which apparently initiated development of the Fleming and Petitsikapau subbasins, and then continued to deform the Trough into broad basins and arches (Fig. 19.6b). Deposition continued in the basinal areas but erosion occurred on the arches and on the margins of the Trough. This period of deformation defines the break between the lower and upper parts of the Knob Lake Group.

The cause of this period of broad warping is not obvious, but may be related to deep-seated, fault-block adjustment on the shelf. Sedimentation on the eastern margin of the Trough was continuous during this period and dominated by shale-siltstone deposition (Laporte Group and shale succession east of Snelgrove Arch, Fig. 19.6b).

Stage 7 – Transgression Across Shelf; Shallow Water Clastic and Chemical Precipitate Deposition (Wishart and Sokoman formations and subaerial volcanism (Nimish Subgroup)

The Wishart and Sokoman formations occur together over most of the Central Trough; the Nimish Subgroup, however, is restricted to the area of the Petitsikapau Subbasin in the South-Central Trough (Fig. 19.6d). The Wishart Formation marks the beginning of a new transgressive cycle which progressively overstepped older units from east to west. On the margins of the Trough it disconformably overlies earlier units (Fig. 19.6c).

Wishart Formation. The Wishart is composed of quartzite, siltstone and shale, and varies in thickness from 50 m in the west to 300 m in the east. The well sorted, mature nature of most of the formation and abundant small scale cross-stratification indicates deposition in a near shore, shallow, shelf environment. Thick, highly mature, orthoquartzite units, up to 70 m thick, occur in eastern parts of the formation near the Snelgrove Arch and may represent offshore sand bar complexes.

East of the Snelgrove Arch, the formation pinches out into shales of the Attikamagen and Menihek formations. The Wishart Formation is not exposed near the Wheeler Dome, therefore, it is impossible to ascertain what effect, if any, this structure had on lithofacies development.

Sokoman Formation. The Sokoman Formation is a cherty iron formation, and for economic reasons is the most widely studied unit of the Labrador Trough. The unit is a typical Superior-type iron formation (Gross, 1970) and is comparable with similar formations found throughout the Circum-Ungava Belt (Dimroth et al., 1970) and Southern Province (Sims, 1976). The formation varies from 120 to 500 m thick and is generally thickest in the area of the Petitsikapau Subbasin, where it is intimately interbedded with the Nimish Subgroup volcanics (Fig. 19.6d, c). The cherty iron formation is usually underlain by Ruth Formation black ferruginous shale-lean chert. Following the suggestion of Zajac (1974), this has been included in the Sokoman Formation. The cherty iron formation is generally divisible into three stratigraphic units: a lower silicate-carbonate member; a middle oxide member; and an upper silicate-carbonate member. However, this is frequently complicated by local facies changes.

The silicate-carbonate iron formation is typically thinly plane-laminated and is inferred to have formed in quiescent, albeit shallow, lagoonal environments. The oxide iron formation, however, shows a variety of features such as oolitic-pisolithic texture, intraclastic breccia and conglomerate, ripple marks and cross-stratification, which indicate deposition in a variety of shallow water, turbulent environments.

Based on these sedimentary textures, Chauvel and Dimroth (1974), Zajac (1974), and Wardle (1979c) define paleogeographic and facies models for separate parts of the Trough. These correlate remarkably well and are summarized in Figure 19.6d.

The north-central part of the Trough formed in a complex environment dominated by oolite shoals and intraclastic (sand) banks, separated by lagoons (thinly laminated silicate-carbonate iron formation) and tidal channels. This was fringed to the southeast by a horseshoe-shaped ring of oolite shoals and coarse, intraclastic sand and gravel bars deposited in turbulent intertidal conditions. To the southeast, this complex was transitional into a basin around the volcanics of the Nimish Subgroup. The periphery of this basin is dominated by fine grained, oxide iron formation formed in shallow, subtidal conditions, and thinly laminated silicate-carbonate iron formation. Around the Nimish volcanic pile these lithologies pass into tuffaceous, silicate- rich iron formation, and interfinger with mafic flows and tuffs.

The eastern limit of the Sokoman Formation is delineated by the Snelgrove Arch and by the stromatolite banks of the Denault Formation. The formation may, however, have extended to the east through gaps between these barrier structures. Iron formation does not occur in the Laporte Group, however, and cannot have extended far beyond the barrier system. The barrier system probably separated a protected and restricted shallow shelf in the west from a deeper water, open sea in the east.

Nimish Subgroup. The Nimish Subgroup (Evans, 1978) is a sequence of mafic volcanics, varying between several metres and 1000 m in thickness, interbedded with various levels of the Wishart and Sokoman formations. The subgroup extends into the Southern Trough where it is interbedded with the Denault Formation in addition to other units (Noel and Rivers, 1980).

The Nimish volcanic complex consists of a central pile of vesicular, largely subaerial mafic lavas, interbedded with thin basaltic conglomerates and breccias. Many of these are interpreted to be laharic; some, however, show evidence of a shallow marine or fluviatile origin. Two volcanic centres have been identified (Fig. 19.6d) and are characterized by rhyolite- and comendite- bearing fanglomerates, trachytes and rhyodacites of alkaline affinity (Evans, J.L., 1978, 1980). The width of the pile varied considerably during its formation, as shown in Figure 19.6d.

The subaerial pile passes laterally into aprons of pillow breccia, hyaloclastite and bedded tuff which interfinger basinward with tuffaceous, cherty iron formation.

Stage 8 – Second Collapse of the Shelf (Menihek – Mistamisk formations)

The Menihek Formation occurs throughout the Central Trough and marks the beginning of a major transgression within the Knob Lake Group and the collapse of the shallow shelf environment characteristic of the underlying formations.

Two major facies have been recognized in the formation. In the western area of the Knob Lake Group the formation consists of grey, rhythmically bedded, turbidite siltstones and shales (Harrison et al., 1972; Wardle, 1979b). In the eastern area of the Knob Lake Group the formation is a black shale-siltstone-greywacke turbidite sequence interbedded with mafic tuffs, breccias and tuffaceous greywackes (Fig. 19.7a). In the extreme eastern part of the area, adjacent to the Doublet Group, the sequence contains massive and rarely pillowed, basaltic flows. Minimum thickness of the formation varies from about 1000 m in the west to 1500 m in the east.

The upper Menihek Formation facies in the vicinity of the Snelgrove Arch is atypical and consists of crossbedded, clean, quartz rich sandstones and siltstones, which accumulated in shallow water environments and reflect the local influence of the Snelgrove Arch as a basin high in upper Menihek time (Fig. 19.7a, b).

The Menihek Formation also occurs around the Wheeler Dome but it appears that this structure exerted little influence on depositional facies. The formation is probably transitional eastward into pelitic schists of the Laporte Group.

The Menihek has been intruded by numerous sills of the Montagnais Group and distinctions between fine grained intrusions and flows are difficult. Thin black graphitic shale sandwiched between the flows commonly contain lenses of syngenetic, massive pyrite-pyrrhotite, apparently formed by submarine, synvolcanic, exhalative activity.

Immediately north of the Snelgrove Arch, the eastern and western facies of the Menihek Formation are separated by the Ferrum River Fault. Further north, however, the two facies are separated by a large area of uplifted older rocks. Development of the two facies in the South-Central Trough may have been controlled by reactivation of the Ferrum River Fault as a basin escarpment (Fig. 19.2, 19.7a, b). This relationship may have extended into the North-Central Trough.

The Mistamisk Formation is a series of basalts disconformably overlying earlier units in the North-Central Trough (Fig. 19.7a). Dimroth (1978) has proposed the formation to be in part equivalent to the upper Menihek Formation. Its stratigraphic position would also suggest a possible correlation with the lower Doublet Group which overlies the Menihek Formation to the east (Fig. 19.7d).

Stage 9 – Local Uplift and Erosion of Shelf; Red Bed Deposition (Tamarack River Formation)

This unit, recently recognized from the South Central Trough, was previously mapped as part of the Menihek Formation (Fahrig, 1967). The unit has not yet been formally incorporated into the Knob Lake Group, but for the purposes of this paper, it is included as such.

The formation generally consists of red, crossbedded arkose, frequently dolomitic; red pisolithic dolomite, green-red argillites and siltstones (Ware and Wardle, 1979; Ware, 1980). The general depositional environment fluctuated between fluviatile and shallow marine or possibly lacustrine. Paleocurrents indicate derivation from a westerly source area (Fig. 19.7c, e).

The lower contact of the formation is not exposed but is presumed to be a disconformity as arkoses within the unit contain clasts typical of the upper Knob Lake Group. The lithological character of the formation indicates derivation by uplift and erosion of at least the western part of the shelf. This uplift may represent an early phase of the Hudsonian Orogeny. As such, it resembles the Loaf Formation of the Belcher Islands (Dimroth et al., 1970; Ricketts and Donaldson, 1981), a molasse facies overlying similar rocks to the Knob Lake Group in the western part of the Circum-Ungava Belt.

This formation completes the shelf sequence in the Central Trough. Another sequence of post-Menihek age occurs in the northern Trough where it consists of shallow shelf dolomites, iron formation and shale (Abner and Larch River formations; Dimroth et al., 1970; Clark, 1978) which conformably overlie Menihek equivalents. These may be broadly equivalent in age to the Tamarack River Formation.

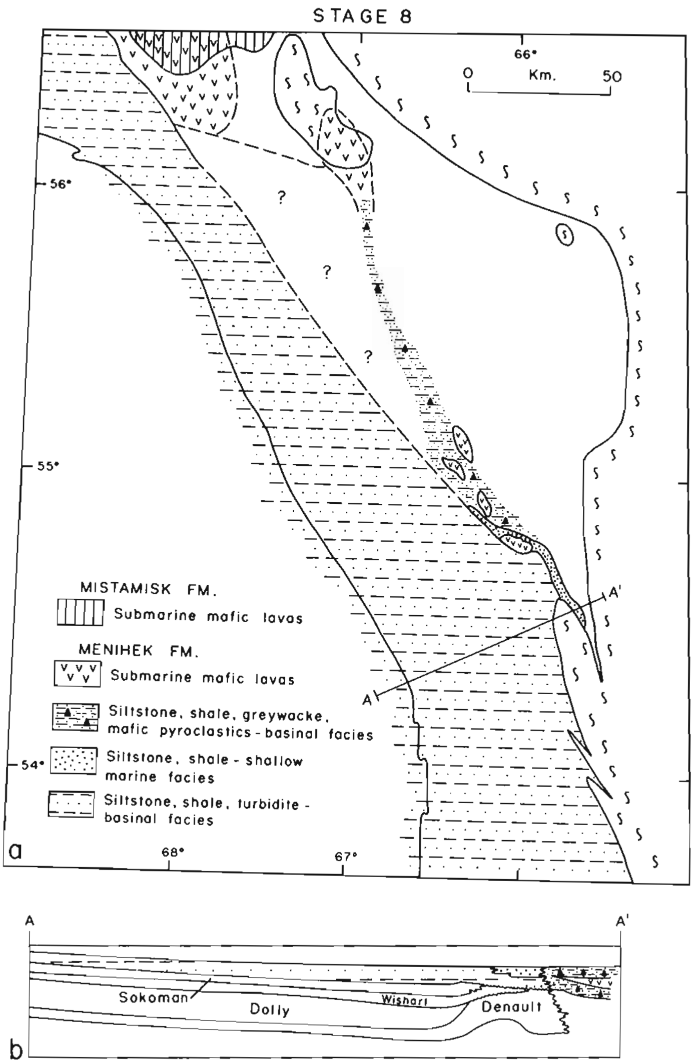
Igneous Activity on the Shelf. Volcanic rocks and comagmatic intrusives (dominantly sills) of the shelf sequence occur in two divisions: a western division comprising subaerial-shallow water volcanics of the Seward and Nimish Subgroups (Stages 1 and 7); and an eastern division of deep-water volcanics associated with shelf collapse and flysch incursion in the Attikamagen and Menihek formations (Stages 4 and 8).

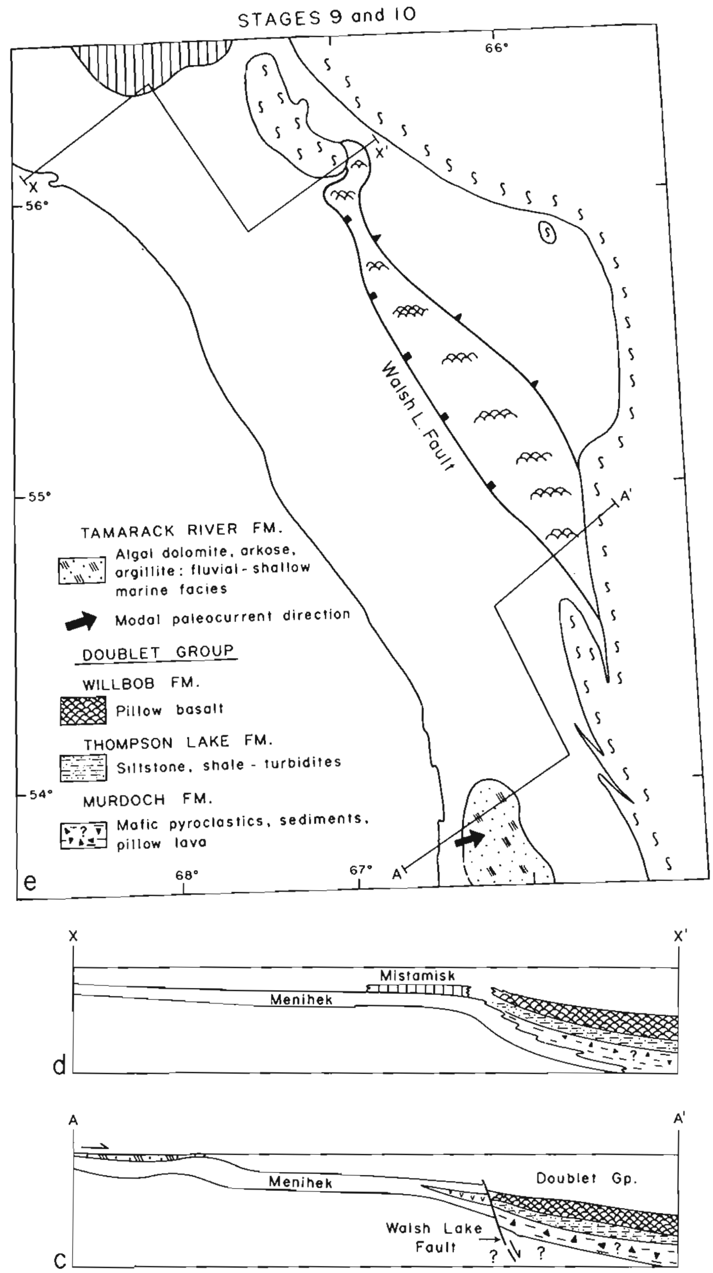
The Seward Subgroup lavas are largely of trachybasalt-trachyandesite composition, show a strong alkaline affinity, and were apparently associated with rifting and graben formation during the initial stage of Trough development. Volcanics in the Castignon Lake Graben were apparently formed along active graben faults (Dimroth et al., 1970).

The Nimish Subgroup is a predominantly basaltic suite of subalkaline-alkaline affinity (Fig. 19.8), (Evans, J.L., 1978, 1980) with local trachyte, trachyandesite, comendite and rhyolite. J.L. Evans (1980) has noted the strong similarity between these rocks and those of the Little Aden Suite (Cox et al., 1970) developed on the margin of the Red Sea Rift. The volcanics are locally affected by strong potash metasomatism which is largely responsible for the pronounced alkaline trend in Figure 19.8.

A suite of kimberlites, carbonatites and melilitite tuffs (Dressler, 1975) has been recognized in the North-Central Trough and may be of approximately the same age as the Nimish Subgroup, indicating a northerly extension of this alkaline environment. Dimroth (1970), however, has proposed a post Kaniapiskau age for these rocks.

Basaltic volcanics in the Attikamagen, Menihek and Mistamisk formations are tholeiitic and chemically identical to the associated sills of the Montagnais Group, and younger volcanics of the Doublet Group. The petrology and chemistry of these rocks in the North-Central Trough have been described by Baragar (1960, 1967), Dimroth (1971b), and Dimroth et al. (1970) and are only summarized here. Recent analyses from the South-Central Trough (Fig. 19.9, from unpublished information by R.A. Doherty and R.J. Wardle) show general similarity with the volcanics to the north. The basalts and associated sills are all low-K tholeiites with





a,b) Stage 8: *Menihek and Mistamisk formations. The Menihek Formation shows a transition from non-volcanic flysch deposition in the west to a volcanic-flysch association in the east. The facies change may be controlled by reactivation of the Ferrum River Fault as a basin scarp fault.*

c,d,e) Stages 9 *and* ***10:*** *Stage 9. Tamarack River Formation arkoses and dolomite represent uplift and erosion of western shelf. Stage 10. Doublet Group (Murdoch, Thompson Lake and Willbob formations). Deposition in South-Central Trough is inferred to have been controlled by a basin scarp fault, the Walsh Lake Fault. This fault dies out to the north where the Doublet Group is in conformable contact with the Menihek Formation. The Doublet Group is inferred to be in part a lateral equivalent of the upper Menihek and the Mistamisk formations. It may possibly also be equivalent to Tamarack River Formation.*

Figure 19.7. *Lithofacies distribution and palinspastic cross-section, Labrador Trough.*

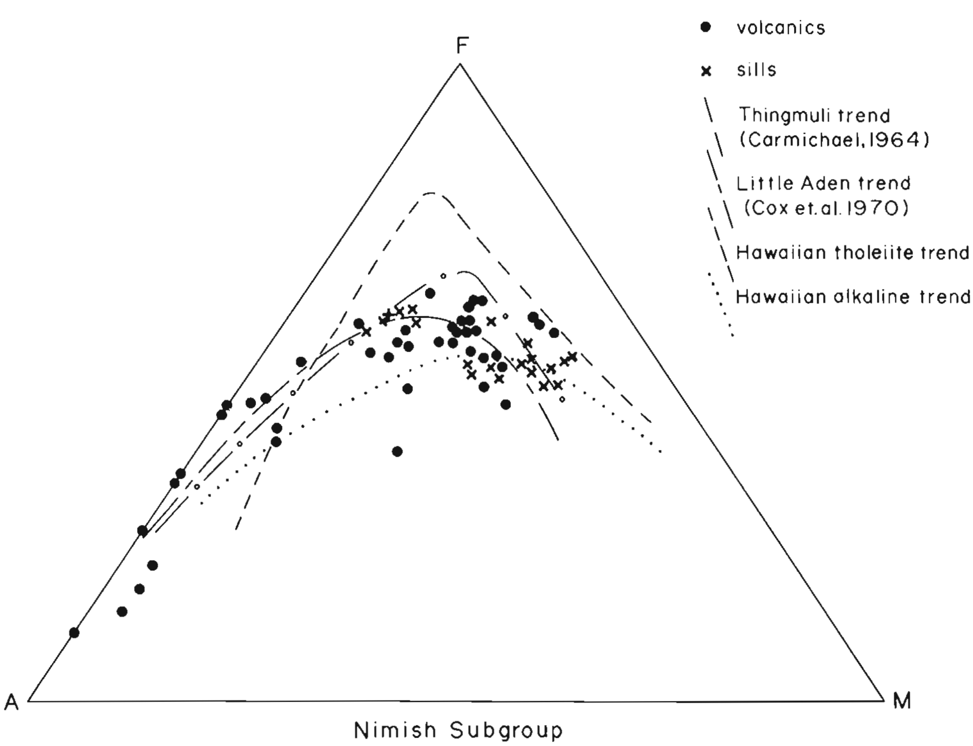


Figure 19.8

AFM plot for Nimish Subgroup volcanics compared to differentiation trends from recent volcanic series.

typical iron enrichment differentiation trends (Fig. 19a, b; Baragar, 1967), and fall largely into the oceanic tholeiite field (Engel et al., 1965). This is also borne out by the K_2O-TiO_2-P_2O_5 diagram (Fig. 19.9c) of Pearce et al. (1975).

The Basinal Sequence

The Laporte Group

The Laporte Group (Harrison, 1952), also known as the Younger Complex (Baragar, 1967), is a poorly defined unit which includes most of the recognizable metasedimentary and metavolcanic rocks that occur along the eastern margin of the Trough (Fig. 19.2).

The contact with the Trough is generally an easterly dipping overthrust. However, in the vicinity of the Wheeler Dome, Dimroth (1978) has recognized the group as a lateral equivalent to that part of the Knob Lake Group above the lower Seward Subgroup. The Laporte Group is also interpreted as a distal, lateral correlative of the shale-siltstone sequence which characterizes the Knob Lake Group east of the Snelgrove Arch. Much of the group has only been mapped on reconnaissance scale; consequently, no stratigraphic subdivisions have been made and only the general nature of the sequence is known.

Over much of its outcrop area, the Laporte Group comprises monotonous biotite schists and gneisses (Fahrig, 1964) of a pelitic - semi-pelitic composition for which Baragar (1967) has recognized a shale-greywacke protolith.

From the considerable extent and generally monotonous pelitic nature of these rocks, it is inferred that they were deposited in a deep water, basin slope-basinal environment. Intercalated within the pelitic metasediments are units of meta-igneous rocks including amphibolites (metavolcanics) and talc-tremolite-fosterite schists (meta-peridotite sills?). Also present are local units of metaquartzite, meta-arkose, dolomite, marble and calc silicate (Taylor, 1979); all apparently representing periods of shallowing within the Laporte basin.

Stage 10 – Rapid Rifting and Submarine Volcanism on Eastern Margin (The Doublet Group)

The Doublet Group is a dominantly mafic volcanic sequence divided into the Murdoch, Thompson Lake, and Willbob formations. The group has a minimum thickness of 5000 m of which 3000 m is formed by the pillow basalts of the Willbob Formation. The Murdoch and Thompson Lake formations have been intruded by numerous gabbro and peridotite sills of the Montagnais Group, which generally appear comagmatic with the volcanics. The Doublet Group occurs in a synclinorium overturned to the west. The eastern margin of this structure is overthrust by Laporte Group and reworked Archean basement rocks. The western limit, where the group is in contact with the Knob Lake Group, is generally the Walsh Lake Fault (Fig. 19.2). However, in the vicinity of the Snelgrove Arch and Wheeler Dome this fault dies out and the Murdoch Formation conformably overlies the Menihek Formation. In diagramatically reconstructed sections across the Trough, Dimroth et al. (1970) suggested the lower Doublet Group interfingers with the Menihek Formation (cf. Fig. 19.7d), suggesting that the Walsh Lake Fault is not, therefore, a particularly profound structure. The normal movement on the fault (downthrow to the east) may indicate that it was an active syndepositional basin scarp fault as proposed for the South-Central Trough (Fig. 19.7c). A 5 km long sliver of rhyolite located along the fault (Frarey, 1961) that may have been the source of similar rhyolite clasts in the Murdoch conglomerates supports this interpretation. The rhyolite may have been intruded during early development of the fault and exposed along the scarp during later movement.

Murdoch Formation

The Murdoch Formation is the most varied unit of the Doublet Group and consists of fine grained, schistose mafic pyroclastics (chlorite phyllites), tuffaceous siltstones, pillow lava and minor conglomerate (Frarey, 1967). Deformation has obscured primary textures and the environment of deposition is unknown. The relatively fine grained nature of most of the formation suggests that it represents the distal products of explosive volcanism, the locus of which presumably lay somewhere to the east.

Baragar (1967) reported a local, discontinuous, iron formation approximately 200 m thick, at the Murdoch-Thompson Lake Formation contact. The unit is thinly bedded and consists of chert-oxide and chert-carbonate-silicate assemblages. The depositional environment of the unit is unknown.

Thompson Lake Formation

The Thompson Lake Formation is a relatively thin unit (500-700 m) composed of rhythmically bedded siltstone, slate and argillite. Although the primary character of the unit is metamorphically obscured, the rhythmic alternation of slate and siltstone suggests a distal turbidite environment.

Willbob Formation

The Willbob Formation is the most prominent unit of the Doublet Group and is composed entirely of tholeiitic flows interbedded with thin units of pillow breccia and hyaloclastite. Willbob flows are predominantly pillowed in contrast to those of the Knob Lake Group. Pillow morphology and texture is similar to that of the Attikamagen Formation.

Volcanic rocks of the Willbob Formation and associated Montagnais sills are also low-K oceanic tholeiites, chemically identical to those of the Attikamagen and Menihek formations with which they are plotted in Figure 19.9a, b, c.

Also shown in Figure 19.9a, b, c are plots for analyses from large amphibolite and metagabbro dykes which occur in the Eastern Basement Complex adjacent to the South-Central Trough (Fig. 19.2). These were once interpreted to be Archean (Wardle, 1979a), but on the basis of their chemical similarity to the Montagnais Group are now interpreted as feeders for the overlying volcanic-sill complexes of the Knob Lake and Doublet groups.

The Doublet Group represents a renewal of submarine rifting activity. This was initiated by a period of explosive volcanism (Murdoch Formation), passed through a period of relative quiescence (Thompson Lake Formation), and culminated in voluminous lava eruption in a deep water environment (Willbob Formation). This is interpreted as the precursor of rapid crustal rifting.

Summary of Basinal Development

The restored stratigraphy of the Labrador Trough, approximately corrected for about 100 km of Hudsonian shortening, is shown in Figure 19.10. Stages in the evolution of this system are illustrated in Figure 19.11 and summarized below.

The Central Trough west of the Wheeler Dome-Snelgrove Arch developed initially in a continental rift environment (Stage 1), then evolved into a shallow marine shelf for the greater part of its existence (Stages 2 to 9). The shelf was subjected to broad warping and possibly block faulting on two occasions: the first (Stage 3) resulted in formation of the Central Geanticline and the flyschoid deposits of the Swampy Bay Subgroup; the second (Stage 6) saw the formation of the Petitsikapau Subbasin and accumulation of the Fleming and Dolly formations. Both periods of broad warping were probably the result of block faulting in the basement in response to rifting activity on the eastern margin.

The shelf was also subjected to two periods of partial or complete inundation during which the shelf-basin slope break moved well to the west. The first period occurred with collapse of the shelf east of the Ferrum River Fault (Stage 4) and incursion of Attikamagen and Bacchus Formation flysch and mafic volcanics. The second inundation occurred during deposition of Menihek and Mistamisk formations (Stage 8) when the entire shelf was drowned and covered by flysch. During both incursions the shelf edge is believed to have been defined by a basin escarpment (listric?) fault. The final stage seen in the evolution of the shelf, at least in the Central Trough, was uplift, local erosion and deposition of Tamarack River Formation redbeds (Stage 9).

Volcanic activity on the shelf was predominantly alkaline, and occurred largely in a terrestrial environment. The eastern margin of the shelf during periods of normal shallow water sedimentation was defined by the Wheeler Dome - Snelgrove Arch line. This line may have consisted of upwarped arches, upfaulted blocks or fault scarps at various stages in its history. Eventually, they were deformed and upthrust into their present position during the Hudsonian Orogeny.

East of the Wheeler Dome-Snelgrove Arch line, the shelf environment was transitional into deeper-water basin slope and basinal environments in which the Laporte and eastern Knob Lake groups were deposited. The chemistry of the volcanics in these units and their association with active synsedimentary faulting indicate that this was an area of intermittent rifting activity through the evolution of the Trough. Incursions of flysch and mafic volcanics from this environment onto the shelf (Stages 4 and 8) were probably the result of periods of increased rifting activity and widespread subsidence east of the Trough.

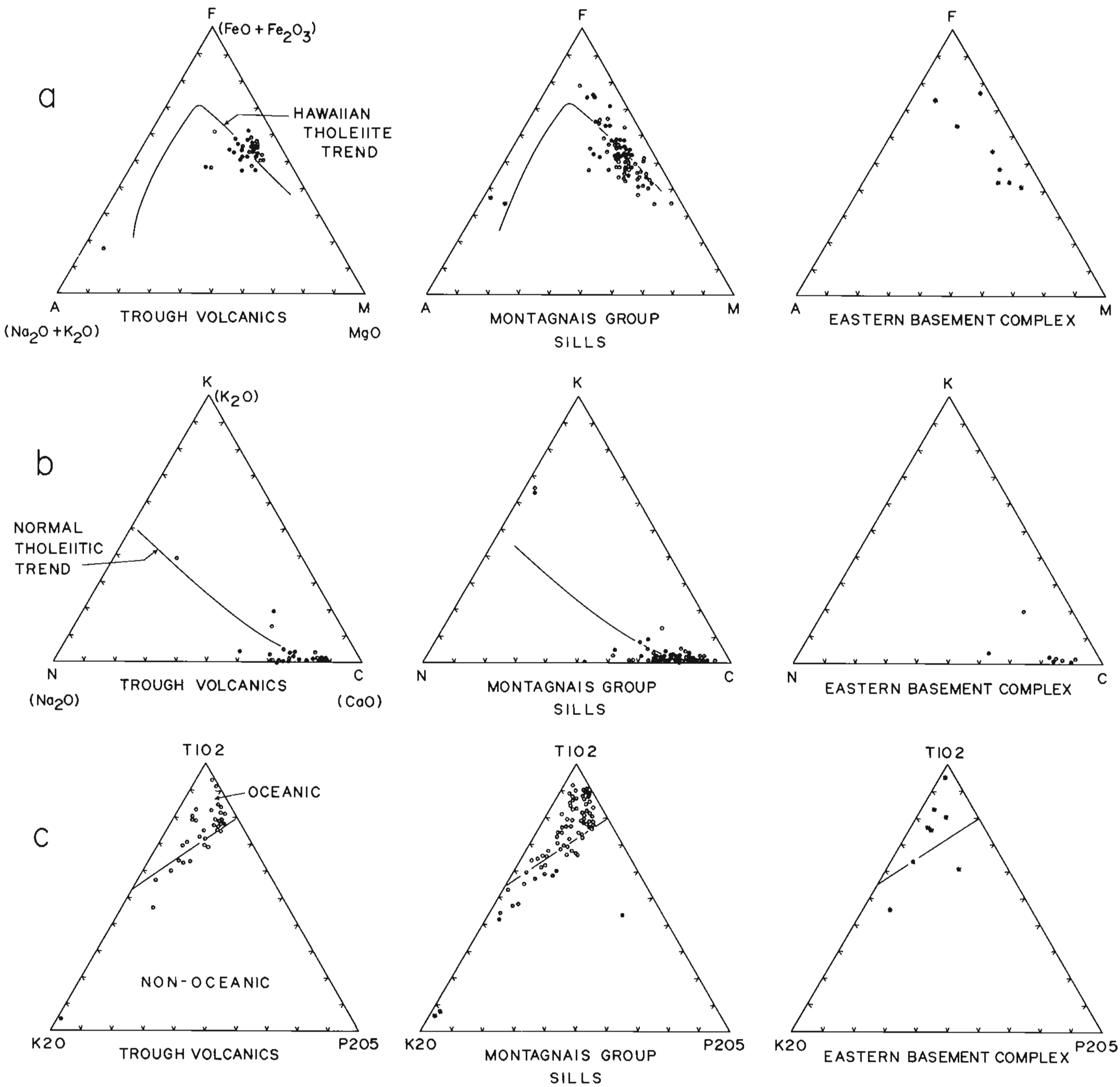
The Doublet Group (Stage 10) represents a renewed period of rifting activity and it spread across the earlier shelf-basin slope break. The massive pillow lava eruption forming the end of this stage (Willbob Formation) is interpreted as the result of rapid rifting at the end of Trough evolution.

The original eastern extent of the Doublet Group is unknown. The western limit is believed to have been a major basin escarpment (Walsh Lake Fault). Shelf equivalents of the Doublet Group west of this structure have not been definitely identified. Conceivably, however, they could be the Tamarack River Formation or the post-Menihek succession in the Northern Trough.

Speculation on Tectonic Setting

In the preceeding sections, the Labrador Trough has been established as a shelf-basin slope-basin system whose evolution was controlled by crustal rifting. The locus of rifting originally lay under what was to become the shelf and then moved to a deep water environment on the eastern margin of the Trough where it was associated with voluminous mafic volcanism. In plate tectonic terms, the most obvious interpretation of this system is that the Trough represents an ancient continental shelf-slope-rise system developed on the passive margin of an oceanic rift system, as suggested by Wilson (1968), Burke and Dewey (1973), and Dewey and Burke (1973).

In this respect, the profiles in Figure 19.10 and 19.11 are similar to those derived for the North Atlantic margin (Drake et al., 1959; Rabinowitz, 1974) where offshore fault horsts (cf the Snelgrove Arch) form prominent subsurface ridges. Burke (1968) and Talwani and Eldholm (1972) have suggested that similar ridges form at the line about which the



a) AFM plots of volcanics, Montagnais Group sills and Eastern Basement Complex amphibolites.

b) NKC plots for above groups.

c) K_2O-TiO_2-P_2O_5 plot (Pearce et al., 1975) for above groups.

Figure 19.9. *Chemistry of Labrador Trough magmatic rocks.*

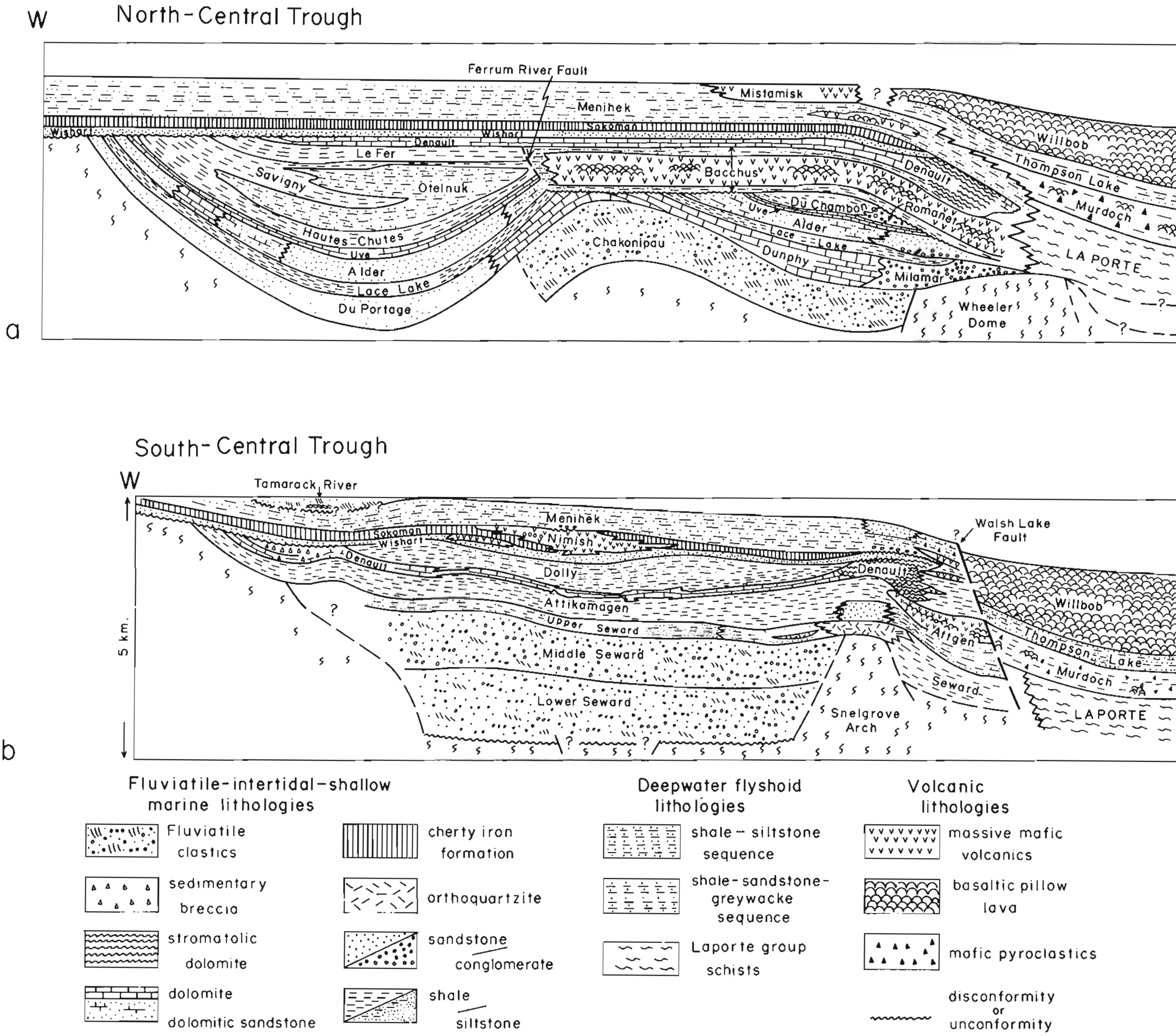


Figure 19.10. *Summary of stratigraphic development of: a) North-Central Trough (modified after Dimroth et al., 1970); and b) South-Central Trough. The Wheeler Dome-Snelgrove Arch line represents the eastern limit of shallow water clastic and chemical precipitate deposition on the shelf. East of this line deposition is dominated by shales, siltstones and mafic volcanics of deep water origin, which correlate with the pelitic schists of the Laporte Group.*

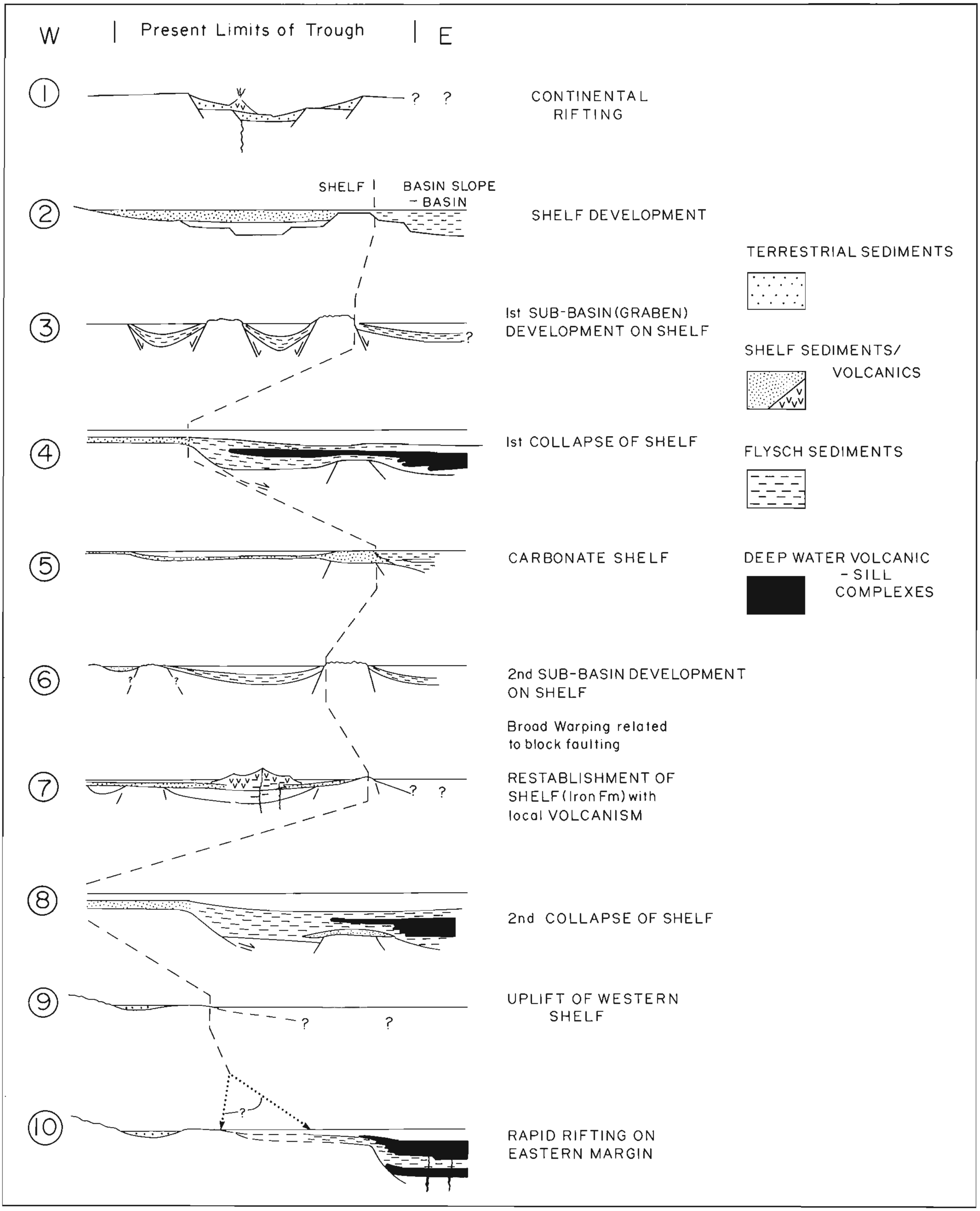


Figure 19.11. *Diagramatic summary of basinal evolution of Labrador Trough. Individual stages are described in text.*

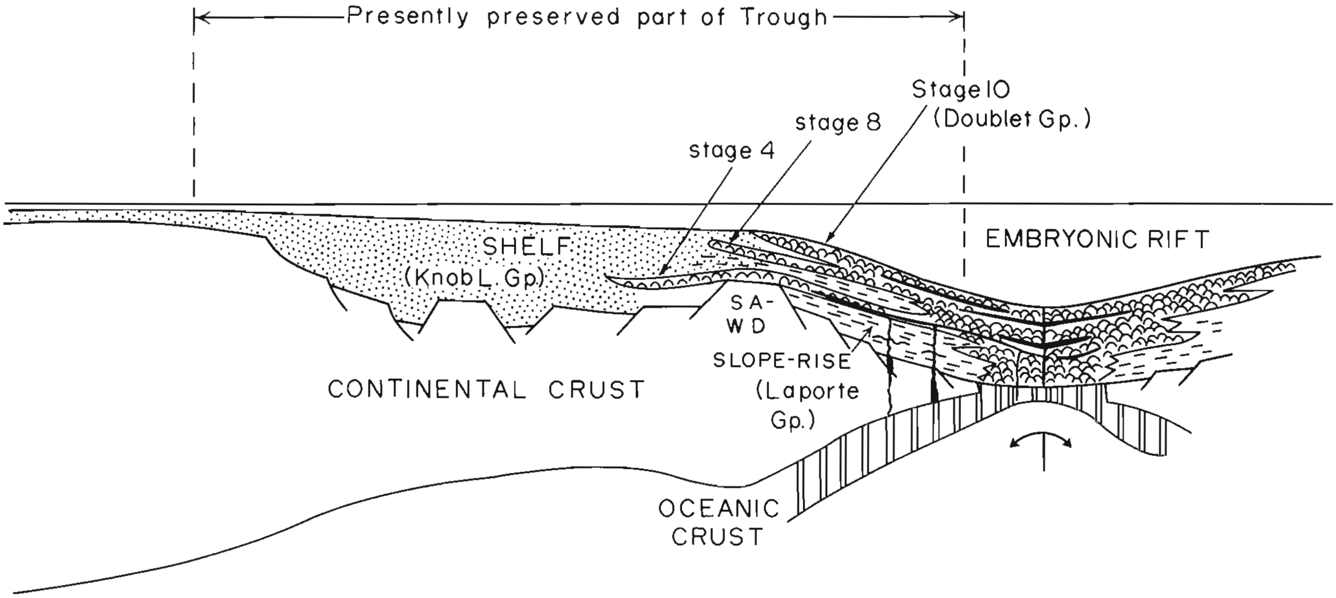


Figure 19.12. *Schematic model for embryonic ocean rift development on eastern margin of Trough.*

continental slope and rise hinge downwards during early oceanic rifting. The ridges subsequently come to mark the shelf-slope break. Lorenz (1981) has also suggested a similar origin for the Sweetgrass Arch marking the offshore margin of the Williston Basin.

It is probable, therefore, that structures such as the Snelgrove Arch originated as horsts during continental rifting and that subsident structures such as the various subbasins originated as grabens.

However, there are no preserved remnants of oceanic crust, either in situ, or obducted, to confirm the original presence of oceanic crust to the east. The flysch- volcanic-sill swarm complexes of the eastern Trough have been described as ophiolitic (Baragar, 1967; Dimroth, 1972) using the term in its classical sense (Steinmann, 1905). However, whilst the chemistry of these rocks is characteristic of ophiolite, they clearly do not display the sequence and structure required in the more modern definition of ophiolite as fragmented oceanic crust (Dietz, 1963; Moores and Vine, 1971). However, recent studies in the Gulf of California – (Einsele et al., 1980) a narrow embryonic oceanic rift – have shown that during the initial stages of opening the turbidite fill of the rift was intruded by numerous sills generated over the developing spreading ridge. Sill formation, accompanied by sea floor eruption of pillow lava, in this case apparently took place at the expense of construction of normally layered crust. In this situation, it is apparent that rapid flysch deposition will keep burying the ridge, particularly if spreading rate is low, and prevent formation of the pillow lava-sheeted dyke system typical of mature ocean basins (e.g. Le Pichon, 1969; Christensen, 1970). The rift, therefore, will develop an interstratified assemblage of continentally-derived flysch, sill swarms and pillow lava, a situation analogous to the eastern margin of the Trough. A schematic evolution for the eastern Trough utilizing this concept is shown in Fig. 19.12.

Basement to this rift would consist of continental crust on the margins and a primitive two-layer (gabbro-peridotite) oceanic crust in the centre. It is evident that the entire eastern margin of the Trough is underlain by Archean continental crust and, therefore, can represent only the eastern half of such a paleo-rift zone.

From Figure 19.11, however, it is apparent that rift-related volcanism was intermittently active on the eastern margin of the Trough throughout practically the whole history of the shelf sequence. This situation would not be expected if rifting had been rapid and resulted in early plate separation and removal of the oceanic ridge axis well to the east. Rather, it is speculated that rifting during evolution of most of the shelf sequence was slow and intermittent and did not result in the formation of anything more than a very narrow oceanic rift.

The complete drowning of the shelf during Menihek Formation deposition, and ensuing voluminous volcanism of the Doublet Group, however, may indicate the onset of rapid rifting and ocean basin formation.

If the remainder of the rift did develop as an ocean basin, the oceanic crust must have been subsequently removed during plate closure – either by obduction to the west, or subduction to the east under an approaching Nain plate. A model for such a system, which accounts for deformation of the Churchill Province by plate closure and subsequent collision, has recently been proposed by Thomas and Kearey (1980). There is at present little evidence (e.g. island arc volcanics, obducted ophiolite) to confirm a plate closing event, but this could be a function of deep erosion levels. It should be noted that Thomas and Kearey interpreted the Labrador Trough as a trench complex – presumably developed at the leading edge of a Nain plate. The Labrador Trough, however, shows none of the characteristics of an accretionary trench prism (e.g. mélange, ophiolite slices, telescoped stratigraphy) and clearly formed on the passive margin of the Superior craton. This does not, however, invalidate the rest of their model which remains a good working model for the evolution of the eastern Churchill Province.

In summary, it is speculated that the Labrador Trough formed as a continental shelf-slope-rise system on the western edge of a proto-oceanic rift system. Whether this system evolved into a large ocean basin or remained as a narrow rift is unknown and must await more detailed work in the interior of the Churchill Province.

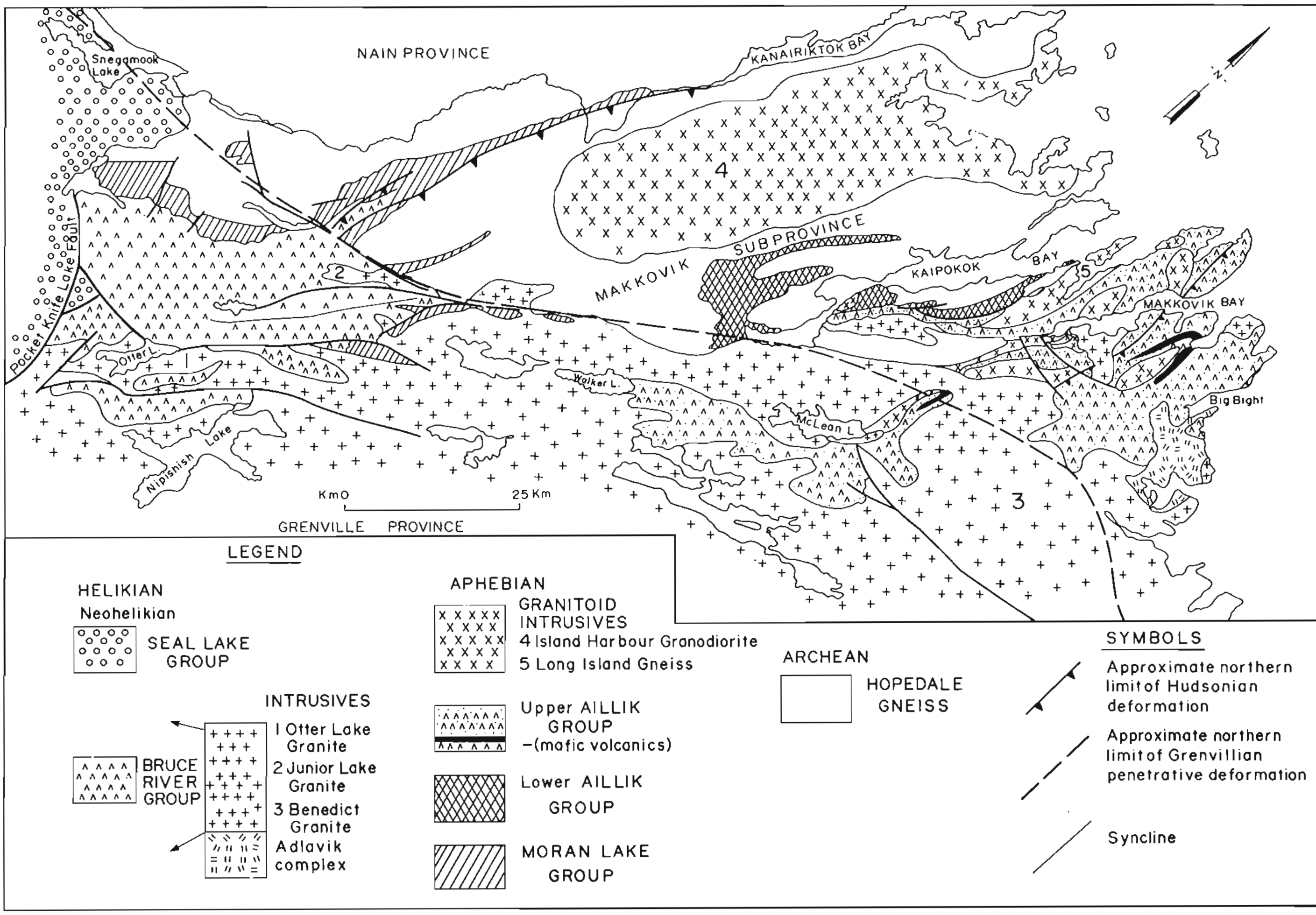


__Figure 19.13.__ General Geology of the Makkovik Sub-Province. Modified after Marten (1977), Smyth et al. (1978) and Bailey (1979).

PART II – EARLY PROTEROZOIC OF THE MAKKOVIK SUBPROVINCE

Introduction

Following Taylor's (1971) proposal, the Makkovik Subprovince is defined as that area of the southern Nain Province which has been affected by deformation of approximately Hudsonian age. The northern and southern limits of the subprovince (Fig. 19.13) are defined as the northern limits of significant penetrative Hudsonian and Grenvillian deformation respectively. The subprovince is separated from the Churchill Province by Helikian rocks, but is presumed to have once been contiguous. It also forms a westerly extension of the Ketilidian Mobile Belt of southern westerly Greenland (Sutton et al., 1972).

The oldest rocks in the subprovince are the Archean gneisses of the Hopedale Complex (Sutton et al., 1971) and are unconformably overlain by two major Aphebian sequences: the Moran Lake Group in the west (Smyth et al., 1978; B. Ryan, personal communication) and the Aillik Group in the east (Ghandi et al., 1969; Marten, 1977; Bailey, 1979; Clark, 1979).

The Moran Lake Group is unconformably overlain by Paleohelikian felsic volcanic rocks of the Bruce River Group which are in turn overlain unconformably by the Neohelikian Seal Lake Group (Smyth et al., 1978; Ryan, 1981).

Within the Makkovik Subprovince, the Moran Lake and Aillik groups have been variably deformed, and metamorphosed during an event, approximately equivalent to the Hudsonian Orogeny of the Churchill Province. The Aillik Group was also intruded by a number of pre-, syn- and post-tectonic plutons. Initial K-Ar dating of rocks of the Aillik Group gave ages of 1728 to 1600 Ma (Gandhi et al., 1969). However, recent Pb-Pb and Rb-Sr dating (Kontak, 1980; Brooks, 1979) indicates that older ages may predominate in the deformed granitoid plutons and Aillik Group rocks and may more closely correspond to Hudsonian dates in the Churchill Province (Stockwell, 1972). Younger ages, which predominate in undeformed plutons intruding the Aillik Group (Gandhi et al., 1969; Archibald and Farrar, 1979), document a Paleohelikian magmatic event which post-dates Hudsonian orogenesis.

Much of the area of the Makkovic Subprovince has been metamorphosed at greenschist facies conditions and displays a relatively simple style of deformation consisting of northeast-southwest trending folds, faults and cleavage in the supracrustal rocks; and refoliation and cataclasis in the basement gneisses. In localized zones, particularly in the area of Kaipokok Bay, the main phase of metamorphism occurred at middle and upper amphibolite facies conditions and was accompanied by intense polyphase deformation.

The Moran Lake Group

This group originally comprised the lower part of the Croteau Group (Fahrig, 1959; Williams, 1970; Roy and Fahrig, 1973). However, Smyth et al. (1975, 1978) recognized the presence of a major internal unconformity within the Croteau Group and proposed a new subdivision into the Moran Lake Group (Aphebian) and the Bruce River Group (Helikian).

The group is estimated to have a minimum thickness of 1600 m and has been subdivided into a number of informally defined units (Smyth et al., 1978). This stratigraphy is summarized in Figure 19.14.

In the area of the southwestern part of the group, basement granites are unconformably overlain by moderately sorted, grey-white sandstones and quartzites which become interbedded with purple and grey laminites towards the top of the unit. These are locally overlain by a thinly bedded chert-oxide iron formation less than 10 m thick.

To the northeast, the basal sandstones and quartzites are inferred to be overlain by a middle unit comprising a succession of black shale, dolomite and greywacke. The greywackes are in turn overlain by a thick (800-1000 m) sequence of massive and pillowed mafic volcanics (Fig. 19.14). The greywackes and shales onlap over the older units of the group and indicate northerly transgression onto the Nain craton.

The environment in which the group was deposited appears to have evolved from one of shallow water, near-shore, deposition at the base of the group, through a shallow marine, shelf environment, into deep-water basinal conditions associated with greywacke turbidite deposition and submarine basaltic volcanism.

Nonvesicular pillows in the submarine flows suggest that the basalts were extruded into relatively deep water, perhaps in a similar environment to that of the Labrador Trough basalts. The maximum basin depth is indicated by chert-carbonate interbeds which suggest extrusion at depths less than the carbonate compensation depth of approximately 5000 m.

The stratigraphic sequence of the Moran Lake Group is strongly reminiscent of the upper Knob Lake Group. In particular, the sequence of sandstone-quartzite-iron formation-black shale is practically identical to that of the Wishart, Sokoman and Menihek formations sequence (Fig. 19.14). The upper part of the Moran Lake Group, comprising greywacke and pillow lava, could also be favourably compared to the upper Menihek Formation on the eastern margin of the Trough. The intermediate unit of the Moran Lake Group, comprising shallow water shales and dolomite, may represent a shallow water facies equivalent to the lower Menihek Formation.

The Moran Lake Group iron formation appears to be of particular significance with respect to comparisons between the two groups. From its mineralogy and its association with shallow water clastics and shales, it would appear to be of Superior type (Gross, 1970). Evidence accumulated to date suggests that most Early Proterozoic iron formations of this type in the Canadian Shield (e.g. Biwabik and Gunflint iron formations of the Lake Superior region; the Temiscamie Formation of the Mistassini Group; the Sokoman Formation and Kipula Formation of the Circum-Ungava Belt) are all of approximately the same age (Goldrich, 1973), i.e. 1900-2000 Ma. Thus, the occurrence of a unit of Superior type iron formation, albeit thin and poorly exposed, may be highly significant in long-range comparisons between the Labrador Trough and Makkovik Subprovince. It should be noted, however, that if the Nain and Superior cratons were once separated by an ocean, then the Knob Lake and Moran Lake Group cannot be directly correlated.

Regardless of whether direct correlations are possible it is evident that both groups show similar patterns of stratigraphic development and formed in similar basinal settings.

Aillik Group

The Aillik Group (Stevenson, 1970) initially known as the Aillik Series (Kranck, 1939; King, 1963; Gandhi et al., 1969) is a heterogeneous sequence comprising a lower unit of mafic volcanics and metasediments, and an upper unit of felsic volcanic and volcaniclastic rocks.

Total thickness of the group has been estimated at between 7620 m (Gandhi et al., 1969) and 8500 m (Clark, 1974). In view of the locally highly deformed state of parts of the Aillik Group, these figures must be interpreted as speculative.

Rb-Sr whole rock determinations on the felsic volcanics have yielded ages of 1676 ± 8 Ma, (Watson-White, 1976) and 1767 ± 4 Ma (Kontak, 1980). These dates are younger than the 1812 Ma Rb-Sr age proposed by Stockwell (1972) for the Aphebian-Helikian boundary. The Aillik Group, however, has been deformed in an event correlated with the Hudsonian Orogeny and must therefore be regarded as Aphebian. The dates on the felsic volcanics, however, indicate that volcanism either immediately preceeded, or was synchronous with orogeny.

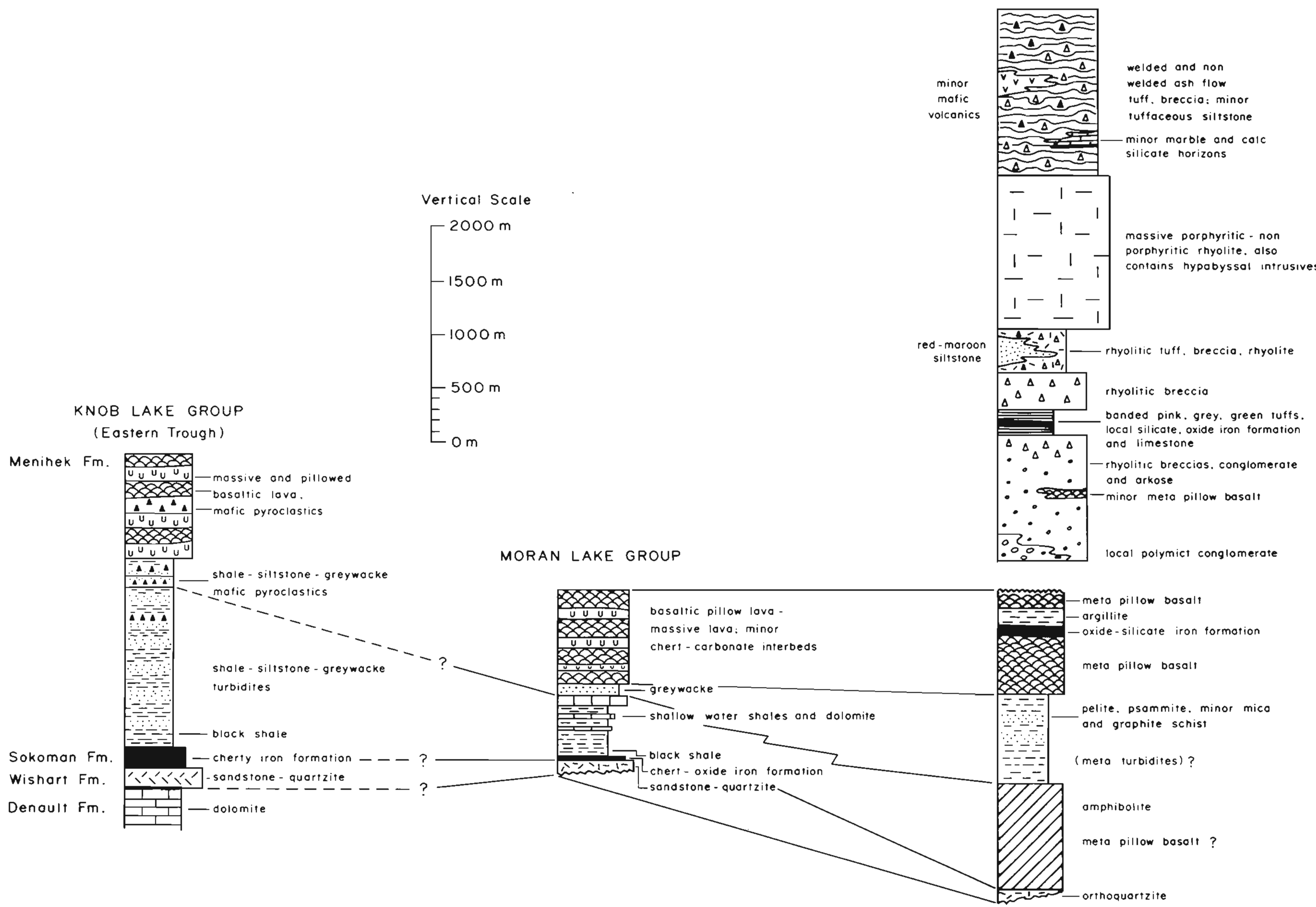
The upper Aillik Group exhibits a relatively simple structural style and low grade of metamorphism in comparison to the intense polyphase deformation and high grade metamorphism characteristic of the lower Aillik Group, suggesting the presence of an angular unconformity. The contact, however, is a zone of strong cataclasis and the depositional contact has nowhere been found, despite detailed mapping (Marten, 1977). Sutton et al. (1971) and Marten (1977), have attributed the deformation in the Aillik Group to a single orogenic event, and suggested that the marked structural contrast between upper and lower parts of the group is due to inhomogeneous deformation and metamorphism concentrated within the lower Aillik Group. Marten (1977), however, has recognized the local development of a polymict conglomerate unit at the lower- upper Aillik contact, which he believes may indicate a disconformity.

Lower Aillik Group

The lower Aillik Group is restricted in lateral extent to the area around the western and eastern shores of Kaipokok Bay (Fig. 19.13). West of the bay, the lower Aillik consists of intensely deformed basaltic pillow lavas, largely converted to actinolite phyllite, with minor pelitic units (Sutton, 1972). Towards the southern end of Kaipokok Bay the volcanics are structurally underlain by a white orthoquartzite (Ryan, 1979). Intense deformation has produced tectonic interleaving of the lower Aillik rocks with refoliated Hopedale gneisses. However, on the basis of structural evidence, Sutton (1972) has inferred the contact to be a tectonized unconformity.

On the eastern side of Kaipokok Bay, Marten (1977) recognized a sequence of amphibolite, (consisting largely of metamorphosed mafic volcanics) metasiltstones and sandstones, and metabasaltic pillows lavas associated with silicate-oxide iron formation and tuffaceous metasandstone approximately 2700 m thick (Fig. 19.14). Marten (1977) has

Figure 19.14. Stratigraphy of the Aillik and Moran Lake Groups and comparison with the Knob Lake Group. Note that upper Aillik Group stratigraphy is generalized and that thicknesses are very approximate.



recognized occasional relict pillow textures in the amphibolite but the origin of the remainder of the unit is uncertain. These units have been severely deformed and interleaved with refoliated basement gneiss in a similar manner to that seen west of Kaipokok Bay.

A depositional environment for the lower Aillik Group can only be generally inferred. The transition from orthoquartzite to thick sequences of pillow lava is similar to that seen in the Moran Lake Group and would likewise suggest a progression from shallow water, nearshore to deeper water, basinal conditions.

A correlation has been suggested by Sutton et al. (1972) and Marten (1977) between the lower Aillik and Moran Lake groups and has been reinforced by the recent mapping of Ryan (1979; personal communication). The meta-pillow basalts occupying the upper part of the lower Aillik Group may be correlative with the pillow lavas of the upper Moran Lake Group (Fig. 19.14). The pelites and psammites, which form the middle unit of the lower Aillik, have been inferred by Marten (1977) to be metaturbidites and are probably equivalent to the greywacke turbidites occurring below the pillow lavas of the Moran Lake Group. The lower amphibolite unit of the Aillik Group, which may represent meta-pillow basalt does not have a correlative in the Moran Lake Group. The white orthoquartzite found on the west side of Kaipokok Bay, however, may be correlative with the basal Moran Lake sandstones. An overall comparison between the two units suggests that the lower Aillik Group represents a deep water, more basinal facies equivalent to the Moran Lake Group. This is consistent with its more southeasterly location.

Upper Aillik Group

The upper Aillik Group comprises a thick sequence of felsic volcanics and associated volcaniclastics with minor intervals of mafic volcanics. Figure 19.14 illustrates a generalized stratigraphic scheme for the unit which may be recognized in both the northeastern belt near Makkovik, and in the inland belt between Walker and MacLean lakes (Bailey, 1979). This stratigraphy cannot, however, be uniformly applied throughout the sections, largely because of the laterally discontinuous nature of the volcanic units. However, a distinctive banded tuff unit is traceable from the southwestern to the northeastern part of the belt (Bailey, in press).

The apparent base of the upper Aillik is seen on the eastern side of Kaipokok Bay where mafic volcanics of the lower Aillik are overlain by grey-white and pink arkoses interbedded with rhyolitic tuff and tuffaceous sandstone. Marten (1977) has also described a localized polymictic conglomerate, containing clasts of granite, felsic volcanics and hyperabyssal felsic intrusives which possibly marks a disconformity surface with the lower Aillik Group (Fig. 19.14).

The arkoses and conglomerates pass up into a thick sequence of conglomerate, arkose and rhyolitic breccia, then into a thin sequence of banded tuffs containing minor units of silicate-oxide iron formation and limestone. The iron formation is poorly described but would appear to be of exhalative origin and related to volcanic-hydrothermal processes. The tuffs pass up into a sequence of rhyolitic breccias, tuffs and minor rhyolite flows; then into a thick unit of massive rhyolite flows and associated hypabyssal intrusives. The uppermost unit of the group is a thick succession of welded and nonwelded ash flow tuffs interstratified with various rhyolitic breccias of monolithologic and heterolithologic nature. The monolithologic breccias are composed predominantly of rhyolite clasts and are interpreted to have formed by explosive volcanism. The heterolithologic breccias consist of rhyolite clasts of varying texture and origin, and are interpreted as laharic breccias similar to those described by Fisher (1960). Small amounts of mafic tuff and lava occur in the sequence.

In the southwestern belt of the upper Aillik Group, which occurs between Walker and MacLean lakes (Fig. 19.13), the ash flow tuffs are underlain by red sandstones interbedded with ash fall tuffs. In the northeast, however, the redbeds are absent and the tuffs contain minor intercalations of limestone and calc-silicate.

The limestone, iron formation and crossbedded sandstone and conglomerate in the lower, volcaniclastic part of the upper Aillik Group indicate deposition in a shallow marine environment. The upper part of the sequence, containing the massive rhyolite and ash flow tuffs, was evidently formed largely under subaerial conditions. The thin limestone and calc-silicate horizons in the ash flow tuffs of the northeastern part of the area indicate that this area was subject to shallow marine or lacustrine incursions.

Plutonic Rocks of the Makkovik Subprovince

The Aillik Group has been intruded by numerous granitoid plutons of varying composition. In general, they can be divided into late Aphebian plutons, intruded during the Hudsonian Orogeny; and Paleohelikian plutons mainly intruded in the period 1600-1400 Ma. The Aphebian granites vary from strongly foliated to gneissic and are generally small, with the exception of the Island Harbour granodiorite (Fig. 19.13) located north of Kaipokok Bay. Plutons of Helikian age (e.g. Benedict and Otter Lake granites) are unaffected by Hudsonian deformation and form a large batholith extending between the coast and Nipishish Lake (Fig. 19.13). Locally, some of these plutons show synvolcanic relationships to the upper Aillik volcanics, suggesting that batholith development began in very late Aphebian time, during or immediately following the Hudsonian Orogeny, and continued into the Paleohelikian.

Chemistry of the Upper Aillik Group

In terms of overall composition, the upper Aillik Group is a bimodal suite composed predominantly of rhyolite with minor basalt. The limited information available on the chemical composition of the volcanic rocks suggests that the felsic volcanics are dominantly of rhyolitic composition (Watson-White, 1976; Bailey, 1979; Evans, D.F., 1980). Compared to modern rhyolites (e.g. Ewart and Stipp, 1968; Lowder and Carmichael, 1970) from island arc environments, rhyolite of the Aillik Group is slightly more potassic but is comparable to rhyolite of comenditic composition (e.g. Nicholls and Carmichael, 1969). However, the typical anorthoclase of comendite is lacking from the Aillik Group rhyolites which also show trace element contents similar to those of calc-alkaline suites (Bailey, 1979).

Plots of total alkalies against silica (Fig. 19.15a) show that most of the upper Aillik volcanic rocks fall into the sub-alkaline field as defined by Irvine and Baragar (1971). Similarly, on Na-K-Ca diagram (Fig. 19.15b), the few analyses available suggest a calc-alkaline trend (Irvine and Baragar, 1971), a suggestion also made by Evans, D.F. (1980). White and Marten (1980), however, have proposed a marginally sub-alkaline character for the rhyolites and note that they have locally been subjected to a strong peralkaline autometasomatism which has completely altered the original character of the rocks. They suggested that the apparent calc-alkaline character of the rhyolites may be due to

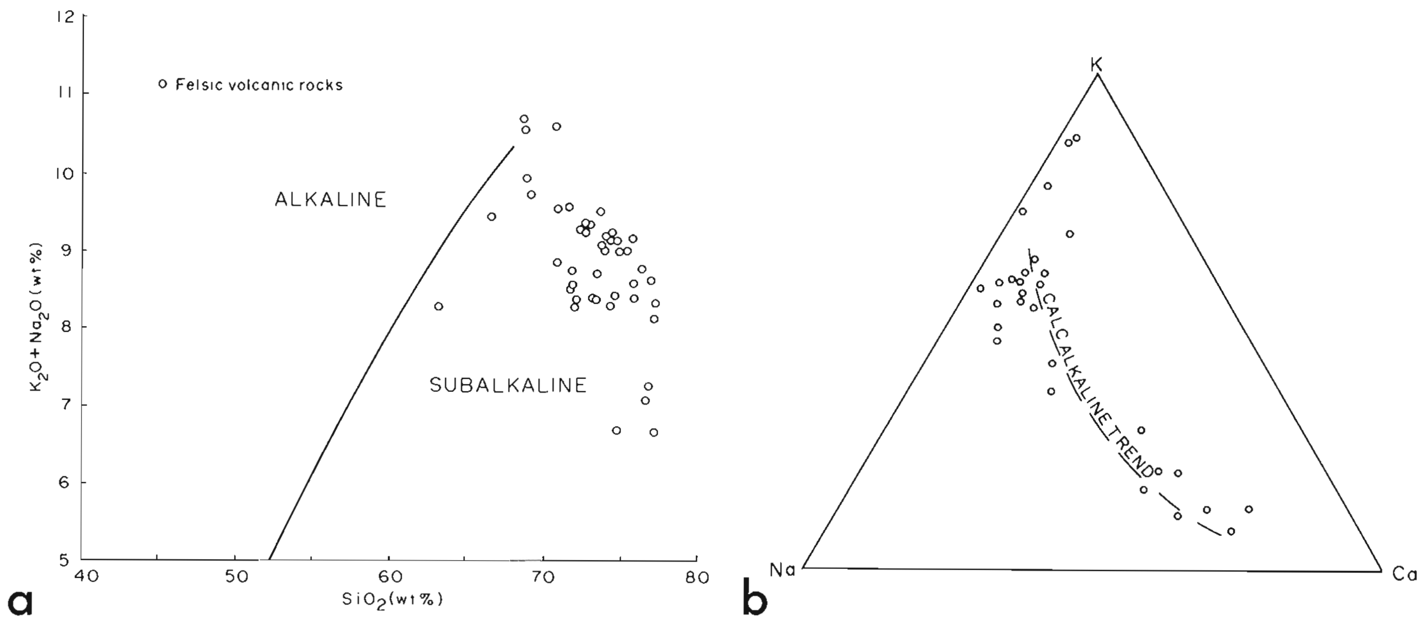


Figure 19.15. *a) Alkali-silica plot and b) NKC plot of upper Aillik Group Volcanics. Dividing lines are those suggested by Irvine and Baragar (1971).*

metasomatic alteration. At this stage, however, insufficient data exist to allow firm conclusions as to classification of the volcanic rocks. Certainly, the greenschist mineralogies of the Aillik volcanics suggest considerable chemical exchange may have taken place during metamorphism and thus present chemical compositions may not accurately reflect those of the unaltered rocks. The work of Kontak (1980) and White and Martin (1980) also indicates that autometasomatism may locally have strongly modified the original chemistry of these rocks.

Strontium isotope analyses have given slightly different initial Sr^{87}/Sr^{86} ratios of 0.704 (Watson-White, 1976) and 0.7022 (Kontak, 1980). Whereas Watson-White has proposed a mantle origin for the Aillik Group on these grounds, Kontak favoured a crustal origin. Both origins are possible. Low initial ratios may result from mantle partial melting and basaltic liquid differentiation, although current models of strontium evolution (Faure and Powel, 1972) suggest that initial ratios of rhyolites formed by this method should be somewhat lower than the results obtained (0.700-.702). On the other hand, the low initial strontium ratios could have resulted by melting of very old (3.5 Ga) sialic crust with low initial Sr ratios such as is exposed in the northern Nain Province (Collerson and Fryer, 1978), and which could also be present in the Hopedale Complex basement. This latter possibility, however, requires single-stage strontium evolution, a requirement which is unlikely at best (Carmichael et al., 1974; Faure and Powell, 1972), and which is almost certainly not the case in an open system such as a sialic crustal environment subjected to several periods of metamorphism and attendant metasomatism.

Speculation On Tectonic Setting

The Moran Lake and lower Aillik groups developed in a similar basinal setting to that of the Labrador Trough. A shallow-water shelf around the southern end of the Nain craton was transitional southwards into a deeper-water environment dominated by submarine rifting and tholeiitic volcanism.

The change to felsic volcanism in the upper Aillik Group suggests a fundamental change in the tectonic setting of the basin for which four possible models may be proposed:

1. Ensialic arc: the upper Aillik volcanics developed in response to northward subduction under a Nain "plate". If this is the case, however, the volcanics should be largely pre-orogenic, whereas in fact they appear to be practically synorogenic. The lack of andesites also detracts from the model, since these form important components of Phanerozoic ensialic arcs (e.g. Andean chain of Peru, Taupo Zone of New Zealand).
2. Collisional products: the volcanics were developed on the overriding plate in a continent-continent collision, such as has been proposed for the dacite-andesite volcanics of the Tibetan Plateau (Dewey and Burke, 1973) and the early Proterozoic Bear Province (Hoffman, 1980). This model is compatible with the synorogenic nature of the volcanics but again suffers from the lack of andesites, generally held to be volumetrically important in collisional orogeny (e.g. Burke and Kidd, 1980).
3. Strike-slip fault: a variation on the collisional model has been proposed by Clark (1979) who suggested that the upper Aillik volcanics were produced over a strike-slip fault zone following collision of two crustal blocks. However, most of the major faults in the Aillik Group appear to have originated as normal faults active during sedimentation and volcanism (Bailey, 1979), and were later modified into reverse faults during the Hudsonian Orogeny.
4. Rift environment: the volcanics represent a continuation of the rifting environment evident in the preceding shelf-mafic volcanic sequence, but with added partial melting of sialic crust. This model is compatible with the bulk composition of the upper Aillik volcanics (Evans, D.F., 1980; White and Martin, 1980), in particular their slight bimodal nature, and analogies could be drawn with the late Cenozoic Rio Grande Rift (Chapin and Seager, 1975) and the late Proterozoic volcanics of eastern Newfoundland (Strong, 1979). The apparent synorogenic timing of volcanism is a problem in this respect, however.

On the basis of present evidence, it is difficult to adequately discriminate between these models. On the basis of bulk volcanic composition alone, a rift origin is favoured (e.g. White and Martin, 1980) and a sequence could be invoked which began with development of a mantle plume and melting of mantle to produce lower Aillik volcanics, then evolved to melt sialic crust and produce upper Aillik felsic volcanics. With time, this plume may have migrated southwards producing increasingly greater amounts of crustal melt which eventually crystallized as the late Aphebian-Paleohelikian batholith terrane.

If radiometric dates on the Aillik Group are taken at face value, the apparent synorogenic timing of volcanism presents a problem. A possible explanation may be that rifting in the Aillik area occurred contemporaneously with deformation and metamorphism south of the Makkovik Subprovince. Deformation and metamorphism may then have advanced north and overprinted the Aillik Group immediately following the cessation of rifting. The precision of radiometric dating techniques is probably not sufficient to separate the rifting and deformational events. Crystallization of the granite batholiths presumably outlasted deformation, hence their posttectonic nature.

This hypothesis is clearly tentative, however, and open to revision pending work in progress south of the Makkovik Subprovince.

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20.

VOLCANISM, SEDIMENTATION, PLUTONISM AND GRENVILLIAN DEFORMATION IN THE HELIKIAN BASINS OF CENTRAL LABRADOR

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Ryan, Bruce, Volcanism, sedimentation, plutonism and Grenvillian deformation in the Helikian basins of central Labrador; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 361-378, 1981.

Abstract

The Helikian geological history of central Labrador includes several periods of plutonism, sedimentation and volcanism and culminated with the Grenvillian Orogeny. The earliest extensive granitic plutonism (1600-1650 Ma) was followed by deposition of the Bruce River Group circa 1530 Ma. The Bruce River Group, the main focus of this paper, comprises the Heggart Lake Formation alluvial fan conglomerate-sandstone, the Brown Lake Formation volcaniclastic blanket sandstone, and the Sylvia Lake Formation potassic, mafic to felsic volcanic rocks. Peralkaline rhyolite porphyries and associated sediments of the Letitia Lake Group may be an equivalent of the Bruce River Group. Post Bruce River plutonism is represented by the Bruce River Group. Post Bruce River plutonism is represented by the Otter Lake granite (circa 1496 Ma), and the Elsonian Harp Lake anorthosite – adamellite suite (circa 1450 Ma).

The Seal Lake Group (circa 1323 Ma) unconformably overlies the Bruce River and Letitia Lake groups, the Otter Lake granite and Harp Lake Complex. It comprises seven formations of terrestrial sediments and associated volcanic and intrusive rocks.

Helikian depositional basins in Labrador appear related to two distinct phases of regional extension. The first (Bruce River – Letitia Lake groups) may be related to events accompanying the emplacement of the early Helikian granitoids, whereas the second (Seal Lake Group) may be a consequence of pre-Grenvillian rifting of the eastern Canadian Shield.

Résumé

L'histoire géologique de l'Hélikien du Labrador central comprend plusieurs périodes de plutonisme, de sédimentation et de volcanisme et s'est terminée par l'orogénèse du Grenvillien. Le plutonisme granitique étendu le plus ancien (1 600 à 1 650 Ma) a été suivi par l'accumulation du groupe de Bruce River, il y a environ 1 530 Ma. Le groupe de Bruce River, objet de la présente étude, comprend le conglomérat-grès de cône de déjection de la formation de Heggart Lake, le grès volcanoclastique étendu, peu épais, de la formation de Brown Lake et les roches volcaniques potassiques, mafiques à felsiques de la formation de Sylvia Lake. Les rhyolites porphyriques hyper-alcalines et les sédiments associés du groupe de Letitia Lake pourraient être équivalents au groupe de Bruce River. Le plutonisme post-Bruce River est représenté par le granite d'Otter Lake (environ 1 496 Ma) et la série elsonienne d'anorthosite-adamellite de Harp Lake (environ 1 450 Ma).

Le groupe de Seal Lake (environ 1 323 Ma) repose en discordance sur les groupes de Bruce River et de Letitia Lake, sur le granite d'Otter Lake et sur le complexe de Harp Lake. Il comprend sept formations de sédiments terrestres et de roches volcaniques et intrusives associées.

Les bassins de sédimentation hélikiens du Labrador pourraient être reliés à deux phases distinctes d'extension régionale. La première (groupes de Bruce River et de Letitia Lake) se rapporterait aux évènements accompagnant la mise en place d'anciens granitoides hélikiens, tandis que la deuxième (groupe de Seal Lake) résulterait de la fissuration pré-grenvillienne du Bouclier canadien oriental.

INTRODUCTION

Folded and metamorphosed middle Proterozoic supracrustal rocks in central Labrador outcrop in a cuspate area along the northern foreland zone of the Grenvillian Orogen. They comprise the Paleohelikian Bruce River Group (formerly the upper Croteau Group of Fahrig (1959) and the middle and upper Croteau Group of Williams (1970)), the Paleohelikian Letitia Lake Group and the Neohelikian Seal Lake Group (Fig. 20.1). All three are part of the Central Mineral Belt (Beavan, 1958; Smyth et al., 1978). The Proterozoic geological evolution of this belt involves several distinct periods of volcanism, sedimentation, plutonism and deformation (Smyth et al., 1978) of which only the Helikian events are described here. These events include intrusion of early Helikian granitic plutons, deposition of the Bruce River and Letitia Lake groups, intrusion of middle Helikian granitoid and anorthositic rocks, deposition of the Neohelikian Seal Lake Group, and deformation and metamorphism during the Grenvillian Orogeny.

The Bruce River Group (Smyth et al., 1975, 1978) comprises a lower formation dominated by conglomerates, a middle formation of volcaniclastic sandstones, and an upper formation of mafic to felsic extrusive rocks. The volcanic rocks yield Rb-Sr whole rock ages in the range of 1526 ± 44

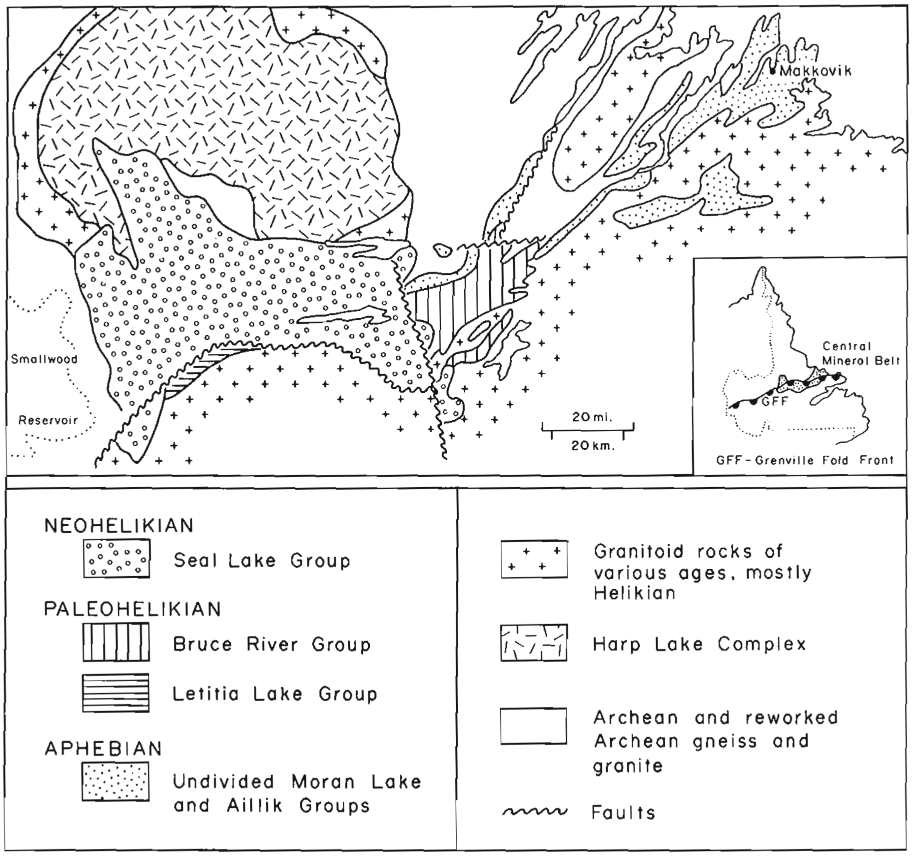


Figure 20.1. *Regional setting of Helikian rocks in the Central Mineral Belt.*

to 1538 ± 25 Ma (Wanless and Loveridge, 1972; Kontak, 1979)[1]. The Letitia Lake Group (Brummer and Mann, 1961) consists of felsic volcanic rocks which differ chemically from similar rocks of the Bruce River Group (W.R.A. Baragar, personal communication, 1979; Thomas, 1980) but are probably time equivalent. The Bruce River Group was intruded by a granitic suite circa 1496 Ma (Kontak, 1979), and subsequently underwent erosion prior to deposition of the overlying Seal Lake Group. The Seal Lake Group, which consists of seven formations of sandstones, siltstones and basaltic flows, with numerous gabbroic sills (Brummer and Mann, 1961; Baragar, 1969, 1974), has recently yielded a Rb-Sr whole rock age of 1323 ± 92 Ma (Baragar, 1978).

Although this paper focuses mainly on the Bruce River Group, the general geology of the Letitia Lake Group, the middle Helikian granitoid and anorthositic rocks, and the Seal Lake Group are also described. The development of the Helikian supracrustal basins in central Labrador is outlined in terms of their time-space tectonic relationships.

PRE-HELIKIAN ROCKS

Pre-Helikian supracrustal deposition in the Central Mineral Belt is recorded by the Aphebian Moran Lake and Aillik groups (Gandhi et al., 1969; Smyth et al., 1978, Wardle and Bailey, 1981). Both were deposited on a basement of Archean gneiss, granite and metavolcanic rocks (Smyth et al., 1978; Ryan, 1977; Ermanovics and Raudsepp, 1979). The Moran Lake and lower Aillik groups are predominantly black shale, carbonate, pillow lava and their metamorphosed equivalents (Smyth et al., 1978; Marten, 1977). The upper Aillik Group comprises felsic and mafic extrusives and volcaniclastic sediments (Bailey, 1978; Bailey et al., 1979; Clark, 1980). The supracrustal rocks and underlying basement were deformed during a period of regional tectonism equated with the Hudsonian Orogeny (Gandhi et al., 1969; Wardle and Bailey, 1981).

EARLY PALEOHELIKIAN GRANITOID INTRUSIONS

South central Labrador was the locus of intrusion for voluminous post-Hudsonian granites. These rocks have been recognized from the Makkovik – Cape Harrison area (Gower, 1980), west to Smallwood Reservoir (Fig. 20.1), and constitute a major plutonic belt at least 400 km long and up to 75 km wide. The rocks range from granite (s.s.), to granodiorite, monzonite, monzodiorite, alkali granite and syenite (Gower, 1980; Thomas and Hibbs, 1980) and yield Rb-Sr and ^{40}Ar-^{39}Ar ages in the range of 1600 – 1650 Ma (Brooks, 1979; Archibald and Farrar, 1979).

PALEOHELIKIAN SUPRACRUSTAL SEQUENCES

Paleohelikian deposition in central Labrador is recorded by the sedimentary and volcanic rocks of the Bruce River and Letitia Lake groups. Both are characterized by the presence of felsic porphyries, but recent investigations indicate that direct lithological correlation is untenable, although the groups may be time equivalents.

Bruce River Group

Introduction

The Bruce River Group occupies an open syncline and comprises three major formations (Smyth et al., 1978; Fig. 20.2). From oldest to youngest, these are: the Heggart Lake Formation, predominantly coarse conglomerates and sandstones; the Brown Lake Formation, composed of volcaniclastic sandstones; and the Sylvia Lake Formation, chiefly andesitic to rhyolitic flows and volcaniclastic rocks. The group unconformably overlies polydeformed Aphebian supracrustals and early Helikian granitoids in the north, and is intruded by middle Helikian granitoids in the south. The major attributes of the group have been described by Smyth et al. (1978) and are expanded here as a result of subsequent work.

Heggart Lake Formation

Distribution. The Heggart Lake Formation (Fig. 20.2) comprises a monotonous sequence of conglomerates and sandstones with minor mafic flows and sills. It rests unconformably on the Aphebian Moran Lake Group and the early Helikian Junior Lake granite, and is overlain by the Brown Lake Formation. The lack of sufficient sedimentary criteria for stratigraphic control and the probable repetition of parts of the sequence by folding and faulting, precludes definition of a type section. Chaulk (1979) mapped the formation in the

[1] All Rb-Sr ages reported here were calculated using the ^{87}Rb decay constant $\lambda = 1.42 \times 10^{-11} a^{-1}$ as recommended by the IUGS Subcommision on Geochronology (Steiger and Jager, 1977)

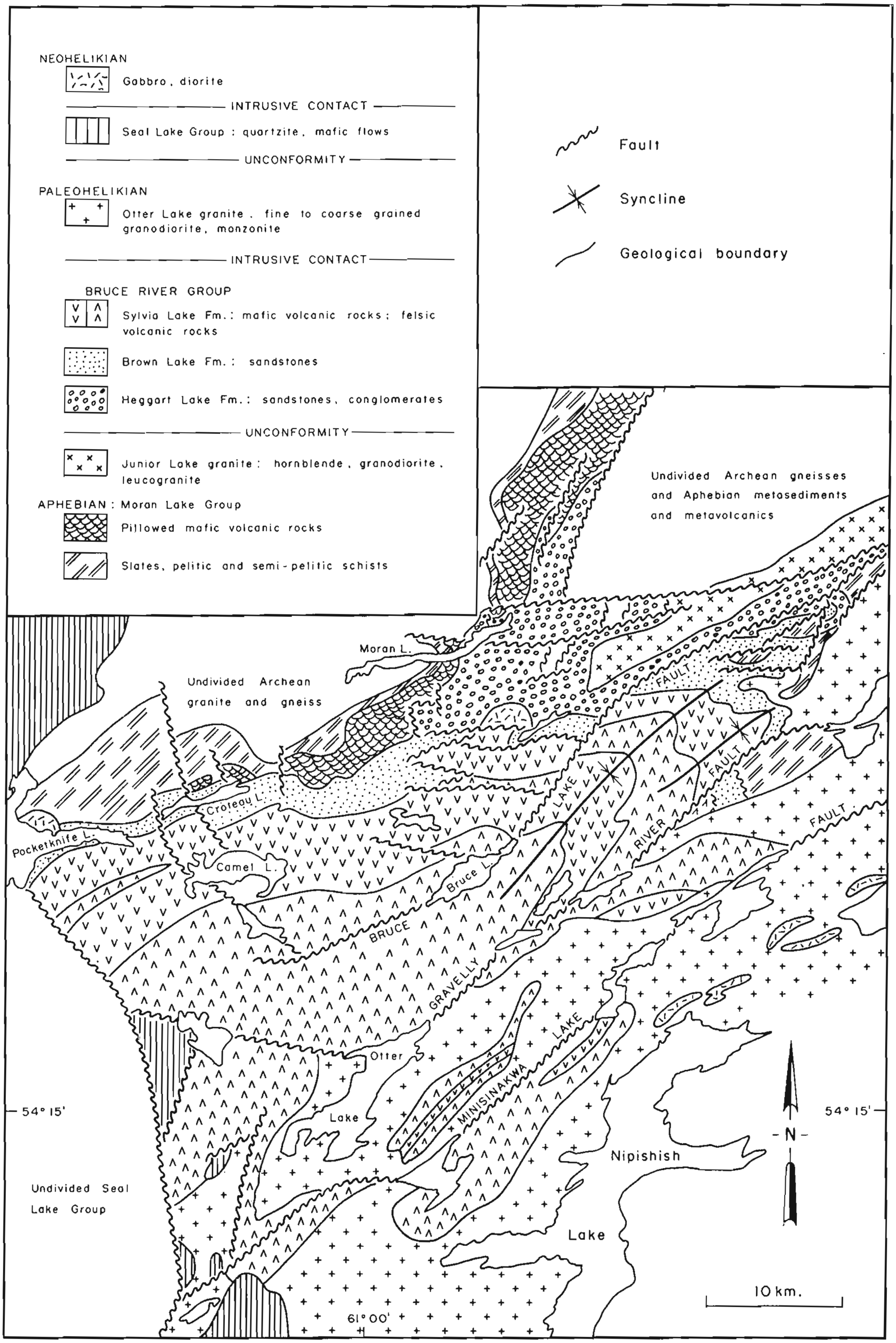
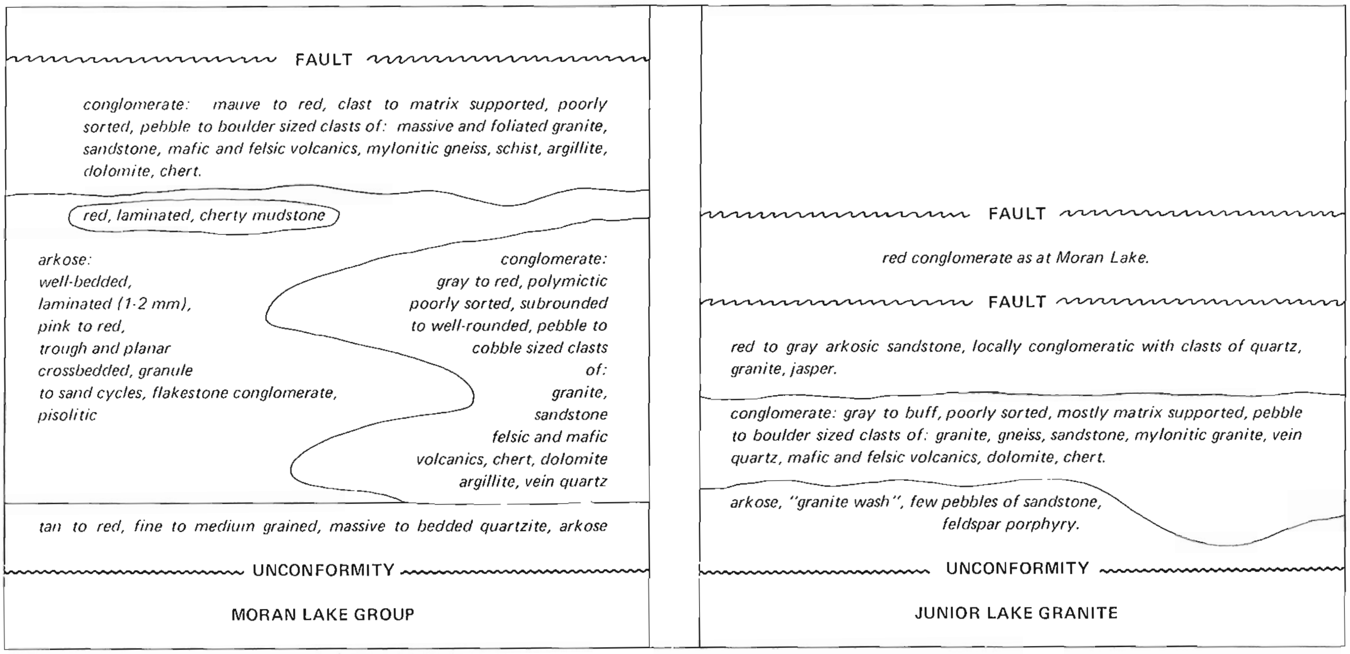


Figure 20.2. *Generalized geological map of the Bruce River Group and surrounding rocks (modified from Smyth et al., 1978).*

Table 20.1. Summary of sedimentary lithologies in Heggart Lake Formation and schematic interpretation of their relationships



vicinity of Moran Lake and suggested that it could be divided into an interdigitating conglomerate-sandstone sequence (Table 20.1) which has been folded into a fault-disrupted asymmetric syncline whose southeastern limb is overturned (Fig. 20.3, 20.4). Slight differences in the stratigraphic sequence on each limb probably reflect lateral facies variations.

Sedimentary Rocks – Lithology and Petrography. At Moran Lake the Heggart Lake Formation rests unconformably on pillowed lava of the Aphebian Moran Lake Group. It comprises a 2000 m thick basal unit of quartz arenite, arkose and interbedded conglomerate, overlain by 2250 m of massive polymictic conglomerate.

The lowermost arenites are commonly structureless, fine- to medium-grained tan coloured rocks. These grade upwards into well bedded, laminated (1-2 mm scale) pink to red arkoses which display small scale trough and planar crossbedding, and some cyclic repetitions of granule conglomerate grading to fine sand in individual beds. Some arkoses display bimodal grain size and rounding (e.g. larger rounded quartz grains in a finer angular feldspathic matrix). Ellingwood (1958) noted ripple marked beds and carbonate-rich, concentrically layered, spherical, "concretions" in this sequence. Some of the thin conglomerate beds in the sandstones contain clasts which exhibit pre-depositional weathering rinds; others are dominated by flaky mudstone clasts.

Two large tongues of conglomerate and two small lentils of dark red and massive to thin bedded mudstone (the latter not shown in Fig. 20.3) occur within the quartzite-sandstone unit. The conglomerate is a massive, grey to red, clast supported, poorly sorted, polymictic, pebble- to cobble-bearing type, with subrounded to well rounded clasts. The most common clasts are granite, sandstone, quartzite, mafic and felsic volcanics, and vein quartz, with lesser amounts of jasper, argillite, red mudstone, quartz-feldspar porphyry, dolomite and chert. The matrix varies from a red, mauve, pink, grey to green sand, with the colour differences appearing to be a result of varying degrees of oxidation.

Overlying this basal quartzite-sandstone-conglomerate unit is a mauve to red, clast- to matrix-supported, poorly sorted, polymictic pebble to boulder conglomerate of which approximately 2250 m is exposed. The contact with the underlying sandstones is generally gradational, but locally appears to be erosional. Clasts in this conglomerate are generally larger than those in the underlying conglomerate and range up to 60 cm. However, the lithologies are similar, with massive and foliated granite (some aplite), quartzites, sandstones, mafic and felsic volcanics, quartz, mylonitic gneiss, and schist clasts most abundant, with lesser amounts of jasper, argillite, red mudstone, quartz-feldspar porphyry, dolomite and chert. The conglomerate matrix is commonly reddish coloured sand.

On the southeast limb of the syncline adjacent to the Junior Lake granite (Fig. 20.3), the lowest unit of the Heggart Lake Formation is a steeply dipping to overturned, discontinuous, 200 m thick medium- to fine-grained arkosic sandstone and granular "granite wash". The "granite wash" is composed of quartz, feldspar and mafic minerals derived from the Junior Lake granite, but scattered throughout the rock are granules and pebble-size clasts of quartz, sandstone, feldspar porphyry and jasper. In places, the "wash" shows an interlocking texture like that of the parental granite, and without the presence of granules and pebbles of "foreign" compositions it is difficult to distinguish the granite from its weathered product.

Overlying and overstepping the granite wash is a grey to buff, poorly sorted, mostly matrix-supported, pebble to boulder conglomerate. Clasts in the conglomerate range up to 80 cm in diameter and are usually subangular to subrounded. Where the conglom-erate occurs adjacent to a pink alaskite along the southern margin of the Junior Lake granite, alaskite clasts constitute 60-70 per cent of the rock, and form the largest boulders. Other lithologies represented by clasts in the conglomerate are sandstone, foliated and gneissose granite, mylonitic granite, quartz, and mudstone, with lesser amounts of feldspar porphyry and rare clasts of jasper, mafic and felsic volcanic rocks and diorite. The matrix is usually a coarse arkosic sand or grit but a black mud is locally predominant. This grey conglomerate and the predominantly red conglomerate north of the Junior Lake granite are in fault contact, but Chaulk (1979) has suggested that the red conglomerate is a younger blanketing deposit that covered all the earlier sedimentary units.

A grey sandstone is interbedded with the conglomerate south of Conglomerate Lake. It is massive and fine- to medium-grained, but locally becomes gritty or conglomeratic and contains a few thin black mudstone lenses. Scattered clasts in the sandstone are chiefly jasper, quartz, granite, and mudstone.

Heggart Lake Formation sandstones are predominantly immature with the degree of rounding and sorting varying on thin section scale. The majority are arkose, and contain detritus from older sedimentary, volcanic, metamorphic, and plutonic rocks of the area, as well as glass shards from contemporaneous volcanism. Some contain sufficient fine grained detritus to be termed wackes according to the classification of Pettijohn et al. (1972).

Volcanic Rocks – Lithology and Petrography. Dark green to grey nonporphyritic and porphyritic mafic and intermediate rocks occur as thin units scattered throughout the conglomerate between Moran and Alvin lakes (Fig. 20.3). These mafic rocks may be autobrecciated and/or amygdaloidal, and locally have behaved in a less competent manner and exhibit more intense deformation than the surrounding sediments.

The volcanic rocks retain no primary mineralogy and range from fine- to medium-grained, displaying relict intersertal, trachytic and microporphyritic textures. Outlines of feldspar laths are preserved but other mineral habits are masked by abundant dusty and granular hematite, chlorite, and a ramifying network of carbonate veins. Plagioclase, where present, is albite.

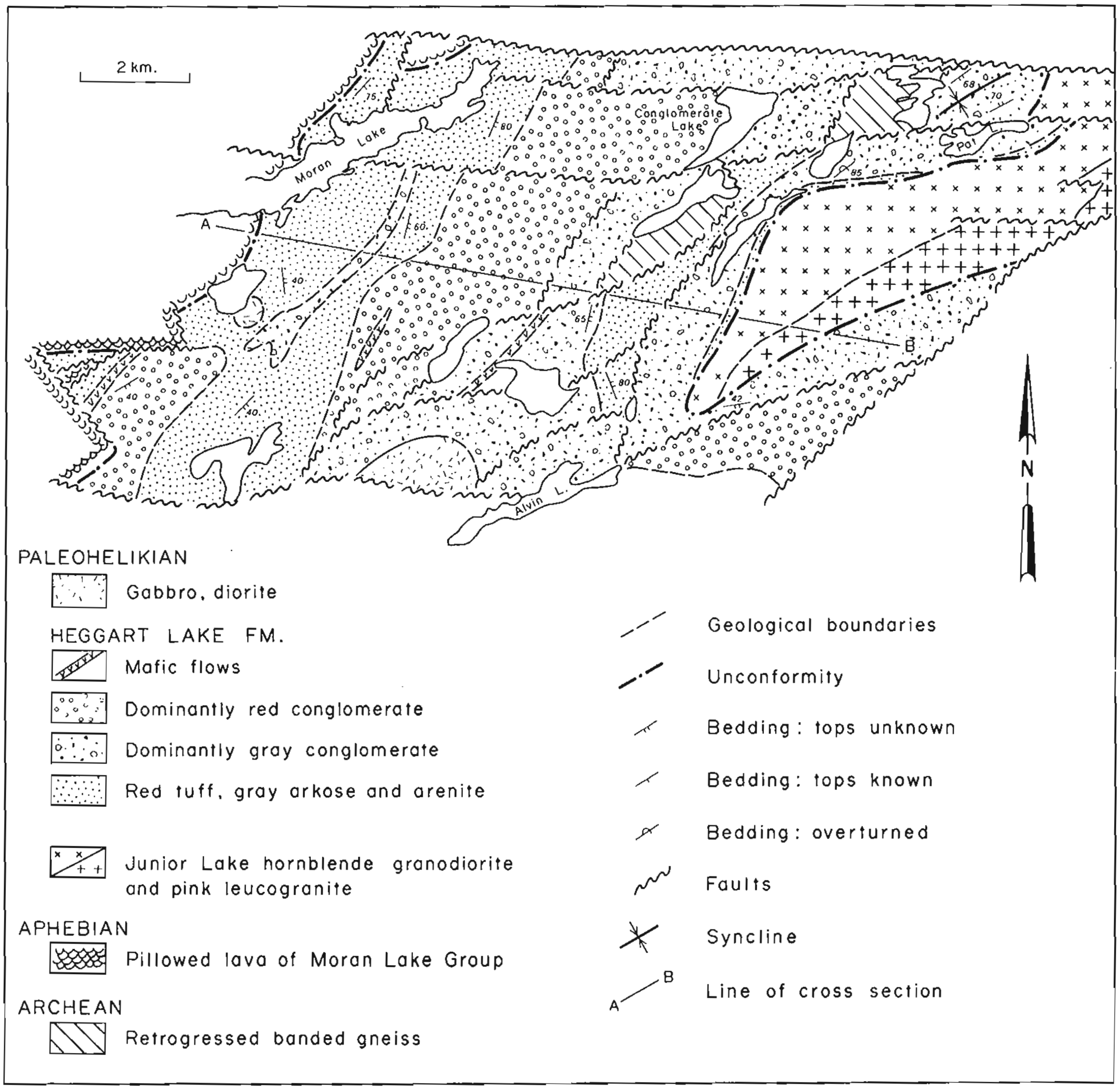


Figure 20.3. Distribution of the Heggart Lake Formation of the Bruce River Group in the Moran Lake area.

Interpretation. The Heggart Lake Formation as a whole is a coarsening upward sequence and by analogy with similar sandstone-conglomerate associations (Nilsen, 1968; Bull, 1972; Steel, 1976; Miall, 1979), is interpreted as an alluvial fan-floodplain assemblage. Judging from the frequency of felsic and mafic volcanic clasts in the conglomerates, volcanic rocks were widespread outside the depositional basin.

A summary and possible facies interpretation for the sedimentary components of the Heggart Lake Formation are provided below and shown in Figure 20.5.

Massive rocks in the Moran Lake area, may represent shoreline deposits developed at the margins of a lake or lakes on a floodplain, and the flat stratified and crossbedded units may be distal fan facies deposits. Sand dunes on the distal portions of the fan, may have contributed wind-blown or stream eroded, well rounded material to the less mature alluvial outwash.

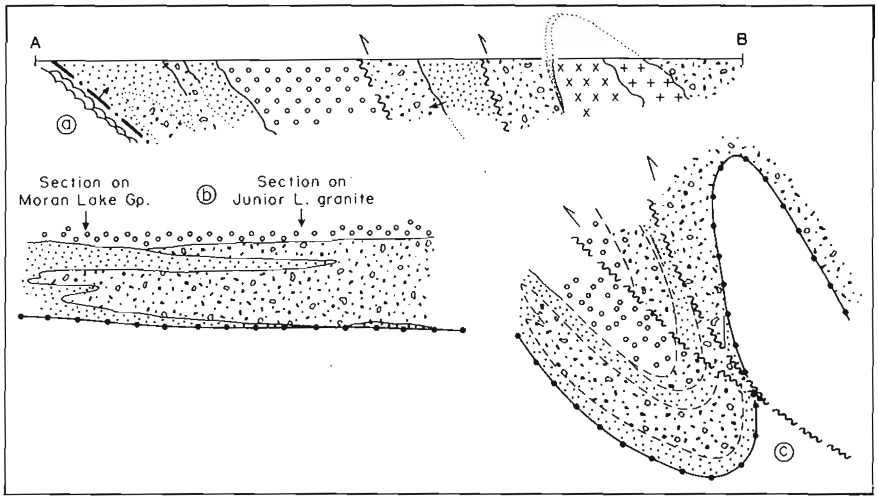


Figure 20.4. *(a) Cross-section along line A-B of Fig. 20.3, (b) assumed original facies relationships and (c) present fold/faults structure. Solid line with dots represents unconformity surface, and arrows in (a) represent facing directions.*

The thin conglomerates, some with clasts showing well developed weathering rinds, and local intraformational flakestone or mud-flake conglomerates reflect flash flooding and resedimentation of earlier deposits following a period of non-deposition on the distal portions of the fan. The sandstones with zoned carbonate "concretions" (pisolites?) appear similar to cornstones and may represent the restricted development of caliche zones in this environment (cf. Steel, 1974; Esteban, 1976). Local coarser crossbedded units within this sandstone sequence could record channel lag or point bar deposits formed by braided streams on the floodplain.

Erosional based, extensive, coarse, poorly sorted conglomerates southeast of Moran Lake reflect a radical change in the sedimentary processes on the floodplain, and record a progradation of coarser fan deposits over the distal plain. Sandstones overlying these coarser deposits indicate an upward return to the previous distal fan depositional environment. Red laminated mudstones in this upper sandstone sequence may be either overbank flood deposits on the plain, or playa lake accumulations.

Extensive, monotonous reddish conglomerates above the upper sandstone sequence are interpreted as a fan fringe prograding over the floodplain from a fault scarp. The characteristic polymodal clast distribution and crude stratification suggest that these conglomerates were transported and deposited by viscous debris flows (Walker, 1975; Larsen and Steel, 1978) rather than by normal bed-load stream activity. This environment was dominant during deposition of the remainder of the formation, and indicates significant uplift and rejuvenation of the source area.

The immature, "granite wash" exposed on the Junior Lake granite apparently developed "in situ", and represents a soil or regolith. The "wash" is only locally developed, and is rapidly overstepped by a coarse grey conglomerate. The local provenance of many of the clasts in the conglomerate suggests that this grey conglomerate originated near the base of fault scarps through rock falls and avalanches. Arenaceous sediments in the conglomerate formed in a fluviatile regime and probably define a channelway on the fan.

Heggart Lake Formation sediments may therefore be interpreted as fanglomerate deposits, largely of debris-flow origin, and fluvial deposits of an alluvial fan, developed in a tectonically active, block faulted, area. The uplifted metamorphic and plutonic basement rocks and older volcanic rocks were exposed on basin margin fault scarps. Rapid sedimentation, probably accompanied by syndepositional faulting, led to the formation of alluvial fans radiating from fault scarps (Fig. 20.5). The basin was quickly filled with poorly sorted, immature detritus which was reworked by fluvial processes at the distal portion of the fans and on the floodplain. Minor amounts of lava were erupted during sedimentation, possibly from localized fissures.

Lack of paleocurrent data and postdepositional faulting precludes detailed depositional analyses of basin and basement elements. However, if the facies interpretation is correct, then the main horst structures were probably along the southeast and northern portions of the area presently occupied by the Heggart Lake Formation.

Brown Lake Formation

Distribution. The Brown Lake Formation (Smyth et al., 1978; Fig. 20.2) corresponds to the lower part of the Upper Croteau Group of Williams (1970), and consists of a discontinuous basal conglomerate (30-70 m) overlain by 1000 m of volcaniclastic sandstones with minor intraformational conglomerate. In some areas, the basal conglomerate member disconformably overlies the Heggart Lake Formation, but in others there is a gradational contact between the two (Collins, 1958). Between Pocketknife Lake and Moran Lake, the Brown Lake Formation oversteps the Heggart Lake Formation and rests unconformably upon the Moran Lake Group.

Lithology and Petrography. The basal conglomerate of the Brown Lake Formation is distinguished from Heggart conglomerates by its bright red weathering and by the preponderance of red sandstone clasts (Smyth et al., 1975). At Croteau Lake (Fig. 20.2) a conglomerate composed predominantly of white quartzite clasts derived from nearby Moran Lake Group sediments forms the base of the sequence, and is vertically separated from the overlying red conglomerate by a few metres of arkose.

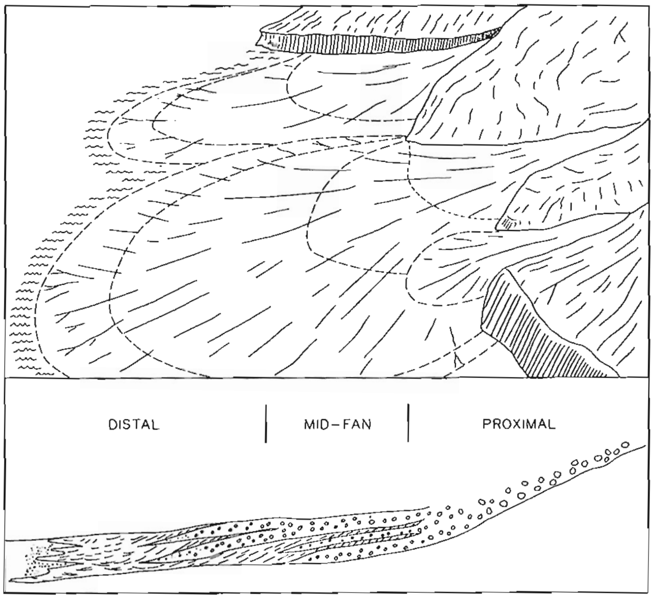


Figure 20.5. *Diagram illustrating postulated depositional environment of the Heggart Lake Formation. Based on Collinson (1978).*

The red conglomerate consists predominantly of moderately rounded, but poorly sorted, pebble- to boulder-sized, matrix-supported clasts of maroon quartzite, red chert, vein quartz, grey quartzite, grey-blue mudstone, phyllite, felsic porphyry, and welded ash flow tuff. Larger clasts are generally better rounded. The matrix varies from red arkose to buff, red, and black mud.

Brown Lake Formation sandstones are dominantly fine- to medium-grained varicoloured volcaniclastites. Thin cherty dust tuffs, and red and grey siltstone are interbedded with the arenaceous rocks. The sandstones are red, mauve, pink, green, grey and buff. Bedding varies from centimetre to metre scale; beds may be massive or finely laminated. Trough crossbedding defined by purple-black muddy laminae or heavy mineral concentrations is common locally. In some instances in the coarser varieties, granule units define the bases of crossbeds. Rare isolated pebbles of chert, granite and felsic porphyry also occur in some beds.

Field and petrographic evidence suggest that the Brown Lake sediments become "less volcaniclastic" to the northeast. Those at Croteau Lake are dominated by red, feldspathic and shard-rich varieties, while those in the northeast are chiefly buff and grey, more mature varieties with less feldspar and little or no glassy component.

Sandstones of the Brown Lake Formation are generally arkosic, with a significant feldspar component; winnowing or size sorting is not pronounced, and the degree of compaction varies. The variation in glass-crystal-rock fragment ratios indicates that most are lithic-vitric and crystal-vitric arenites and lithic-crystal wackes. In the latter, a high degree of sericitization makes it difficult to distinguish feldspar grains from matrix.

Two textural/compositional varieties dominate the sandstones, namely (i) one in which glass shards are abundant (usually these sandstones are red), and (ii) one in which the granular quartz, feldspar and lithic constituents are surrounded by a drusy matrix of cryptocrystalline quartz and feldspars. One main difference between Brown Lake and underlying Heggart Lake sandstones is that the vast majority of Brown Lake quartz grains are of volcanic origin, whereas predominantly metamorphic and mildly undulose "common" (plutonic?) quartz grains characterize the Heggart Lake.

Lithic fragments in the sandstones are predominantly felsic in composition. The felsic components, chiefly rhyolite, are a diffuse, intergrown, fine, quartz-feldspar aggregate and, with the exception of quartz and feldspar phenocrysts, display few primary features. Many exhibit a fine banding suggestive of flow-banded rhyolite or welded tuff.

Type (i) sandstones are composed of abundant glass shards intermixed with splintery to subangular grains of mildly undulose quartz, variably sericitized plagioclase, and felsic and mafic volcanic fragments. Relative sizes of constituents vary; generally crystals are less than 0.5 mm, whereas flattened pumice and rhyolite clasts reach several millimetres.

Type (ii) sandstones show no evidence of glass-shard forms. Otherwise their mineralogical and lithic constituents are the same as type (i) sandstones.

Some sandstones are gradational between types (i) and (ii). These exhibit subrounded to well rounded coarse grains of quartz and rock fragments in a matrix of fine quartz + feldspars + rock fragments + glass shards.

The pink to pale green, cherty, dust tuffs or porcellanites, interbedded with the sandstones, are generally 10-20 cm thick, and are commonly spotted due to spherulites developed by devitrification. The spherulites commonly have a carbonate core, surrounded by a white or grey siliceous rind and rarely exceed 5 mm in diameter. In some porcellanites the spherulites are more resistant to weathering than the surrounding rock and stand out in refief as ovoid protrusions.

Interpretation. The Brown Lake Formation blankets the earlier alluvial fan sediments of the Heggart Lake Formation. It oversteps the Heggart Lake to rest directly on the Moran Lake Group (Fig. 20.2), suggesting "capping" of the original graben and possibly the bounding highland area. The Brown Lake accumulated in a tectonically quiescent basin with the disconformity at the base of the sequence recording the terminal disturbances at the close of Heggart Lake deposition. Minor intraformational conglomerates record local unstable conditions.

The overwhelmingly volcaniclastic sediments of the Brown Lake are interpreted to be products of wind-blown ash clouds introduced to the basin from a distal volcanically active area. The paucity of lapillisized ashes suggests that the area was somewhat removed from the tephra source although, as Walker (1971) and Lajoie (1979) pointed out, fine grained deposits do not necessarily indicate distance from source vents.

Delicate glass-shard shapes in the shard-rich sandstones suggest they are primary air-fall sand-tuffs. Postdepositional abrasion has been minimal. Sandstones with the ultrafine groundmass may be mixtures of pyroclastic and epiclastic debris.

Despite local variation, the whole Brown Lake succession appears to have been deposited either under totally aeolian conditions or in a very shallow water lacustrine environment, as small scale planar and trough crossbedding are common throughout the sequence.

Crossbedding, present even in the shard-rich tuffs seems to be a result of short-lived phenomena, for the shards are not significantly abraded. This may be a function of rapid burial and/or cementation. However, texturally more mature units in the northeast portion of the area indicate that some reworking of the volcaniclastic material took place.

The co-existence of rounded clastic components and delicate vitroclastic material in some of the sandstones suggests that these rocks are dual-source deposits, formed either by pyroclastic glass-shard ash settling on an area in which sediments present were already subjected to prolonged abrasion, or by transportation of the vitric components from their site of initial deposition and their admixture with more mature sediments. Alternatively, the rounded grains represent products from a high energy environment (shoreline? dune?) swept slightly offshore by storms or wind, but deposited in water depths close to wave base.

Sylvia Lake Formation

Distribution. The Sylvia Lake Formation (Smyth et al., 1978) comprises predominantly mafic and felsic flows and sills, coarse pyroclastic breccias, agglomerates, welded and nonwelded ash flow tuffs, with lesser volcaniclastic sediments. The various elements of the formation are excellently exposed in the type area, located along the axial trace of the Bruce syncline (Fig. 20.2; Ryan, 1978). The bulk of the formation outcrops northwest of the Gravelly River Fault, but several roof pendants occur in the granite southeast of the fault in the Otter Lake – Nipishish Lake area (Fig. 20.2). The formation is at least 8000 m thick where a continuous sequence is preserved, but numerous faults in its upper portions considerably disrupt the succession.

Lithology, Petrography, and Geochemistry. Mafic and intermediate volcanic rocks predominate in the lower parts of the formation. Felsic volcanic rocks increase in volume upwards in the sequence, with the major portion of the formation in the Bruce Lake – Otter Lake sector comprising approximately 3000 m of predominantly welded and nonwelded ash flow deposits. The major characteristics of the formation are summarized below.

The least altered rocks occur northwest of the Gravelly River Fault (Fig. 20.2) where, although greenschist facies metamorphic assemblages are present, primary mineralogy and textures are widely preserved. South of the fault, however, the rocks are more metamorphosed and sheared and although commonly retaining gross primary features, are considerably recrystallized with few primary igneous textures (Ryan and Harris, 1978). The descriptions which follow refer to that portion of the formation northwest of the Gravelly River Fault.

The volcanic rocks of the Sylvia Lake Formation are rarely quartz porphyritic, but characteristically contain feldspar phenocrysts. In most cases the groundmass is so fine grained that an accurate determination of modal mineral content is not possible. The porphyritic rocks are andesites, trachyandesites (latites) and trachytes (Williams et al., 1954). Although the above terminology is used, the rocks may not correspond chemically to their petrographic name (cf. Creasey and Krieger, 1978; Streckeisen, 1979). For example, many rocks termed trachytes in the field, are chemically rhyolites.

The lowermost units of the volcanic pile are olivine basalt, andesite and trachyandesite (latite). Interlayered with the flows are mafic to intermediate volcanic breccias and agglomerates and crystal-rich volcaniclastic sediments. Several thin lensoid flow banded and autobrecciated rhyolite flows and oval rhyolite plugs are present in the lower sequence in the vicinity of Camel Lake (Smyth and Ryan, 1979; Fig. 20.2). Northeast of Bruce Lake, the intermediate-mafic flow sequence is split by a 2500 m unit of welded felsic tuff and massive felsic porphyry.

The volcanic rocks vary from grey to green to mauve, and from massive aphanitic to feldspar and pyroxene phyric; pyroxene phenocrysts are less common than feldspars. The flows locally show a crude flow lamination defined by tabular feldspar alignment and subtle compositional and grain size variations on the scale of several centimetres. Locally they are amygdaloidal, with vesicles filled by one or more of quartz, calcite, chlorite, epidote and green biotite. Flow tops are commonly scoriaceous, and infilled with volcaniclastic sediment where the latter occurs between flows. Several bifurcating sills of andesitic feldspar porphyry intrude the lower part of the Sylvia Lake Formation. Locally they reach the surface as porphyritic flows (Smyth and Ryan, 1979).

Angular autoclastic breccias, pyroclastic breccias, agglomerates and volcanic rudaceous rocks of epiclastic origin are common components of the mafic succession. Individual blocks in these fragmentals are generally less than 0.5 m in maximum dimension. They are generally poorly sorted and may be either mono- or polylithologic. The matrix of many of these coarse deposits is a crystal-rich volcaniclastic sand.

Sedimentary screens which separate flow and fragmental units are generally less than 20 m thick. They vary from red through grey to green, and in composition from coarse volcaniclastic sandstones to siliceous mudstones. Bedding is well defined, and the finer grained varieties are commonly laminated. The sediments infill blocky flow surfaces, and are locally incorporated as clasts and large contorted rafts at the base of the overlying flow.

The felsic pile is dominated by a monotonous succession of porphyritic pyroclastic rocks, most of which are ignimbrites or welded ash flow tuffs. Interbedded with these are stratified felsic airfall (?) crystal tuffs and lapillistones, coarse pyroclastic breccias, massive felsic porphyries (laccolithic intrusions?), and thin mafic-intermediate flows and pyroclastics, the latter similar to those described above.

The welded tuffs are pink, grey, red, purple and black, and display well developed eutaxitic structure expressed by elongate rock fragments and hairlike, lenticular, collapsed pumice fragments moulded around crystal and lithic constituents (cf. Ross and Smith, 1961). The crystal to lithic fragment ratio varies, but in general fragments constitute less than 30 per cent of the deposits. The degree of welding and flattening of pumice fragments also varies. A diffuse zonation from a weakly or nonwelded fragmental, upwards through a dense vitrophyric portion, to a less welded portion is locally present. Hiatuses in deposition of the ash flows are indicated by intercalated units of sediments and narrow mafic flows.

In the upper part of the Sylvia Lake Formation volcaniclastic sedimentary rocks are common as screens between the flows. They are mostly lapilli tuffs, comprising fragments of light-coloured pumice(?), andesite, and rhyolite porphyry set in a mixture of ash size rock and mineral fragments. The tuffs are buff to green, and commonly bedded on a millimetre to metre scale; syndepositional deformation is common. One such sedimentary unit northeast of Bruce Lake contains high angle aeolian (?) crossbeds up to 3 m thick.

Massive porphyry units within the sequence are mesoscopically similar to the welded crystal tuffs. However, there is no indication of a glass shard component, so these may represent lava flows or synvolcanic laccoliths in the volcanic pile.

Geochemical studies of the rocks of the Sylvia Lake Formation are in progress. The suite exhibits a calc-alkaline trend on an AFM diagram, and plots in the subalkaline field on the Ol'-Ne'-Q' diagram of Irvine and Baragar (1971). On the $Na_2O + K_2O$ versus SiO_2 diagram using the terminology of Middlemost (1980) the suite ranges from subalkalic and alkalic basalt and basaltic andesite through trachyandesite and trachyte to high alkali rhyolite. The rocks are relatively high in K_2O compared to similar classes of the calc-alkaline and tholeiite series, and appear to be best compared with the "shoshonite association" of Joplin (1968).

Interpretation. The Sylvia Lake Formation represents a major volcanic accumulation in an area which had previously been dominantly a sedimentary depocentre. This volcanism was the most important contributing factor through the remainder of Bruce River Group deposition.

The earliest volcanism is dominated by alkalic basalt, basaltic andesite and trachyandesite. The minor autobrecciated and flow banded rhyolite domes of the Camel Lake area are interpreted to be products of viscous felsic lava eruptions from flank vents coeval with ash flow tuffs and rheoignimbrites northeast of Bruce Lake. Porphyritic sills in the Camel Lake area are synvolcanic intrusions which locally breached the surface on the flank of the shield volcano.

The upper pile, which is dominated by ignimbrites, records repeated explosive eruptions. These rocks range from trachyte to high-K rhyolite, and are the stratigraphically highest rocks of the volcanic pile. Hiatuses in ash flow eruption are commonly marked by coarse polylithologic breccias which appear to represent volcanic "belching" or vent cleaning phenomena prior to onset of the next massive ash flow or lava eruption. Some, however, appear to be laharic breccias (D.G. Bailey, personal communication, 1978). Lapillistone and crystal tuff deposits, some of which are characterized by syndepositonal isoclinal folding and large scale crossbedding, may in part be due to base surge processes (cf. Moore, 1967; Fisher and Waters, 1970; Crowe and Fisher, 1973).

By analogy with ignimbrite plains and shield volcanoes described from other parts of the world (Macdonald, 1972) massive ash flow eruptions such as those present in the upper Sylvia Lake Formation are followed by caldera formation over the roof of an evacuated magma chamber (cf. Williams, 1941; Smith et al., 1961; Steven and Lipman, 1976; Lipman et al., 1978; Myers, 1975; Christiansen, 1979). The thicknesses of ignimbrites present in the Bruce River Group are far in excess of what would be expected from one simple fissure or crater. Instead, the whole pile would seem to be the product of multiple vents, a phenomenon found in such areas as the Toba region of Sumatra (van Bemmelen, 1929), the western United States (Hamilton and Myers, 1966; Christiansen and Lipman, 1972), all of which are major volcano-tectonic depressions. In all these areas silicic volcanism in the form of extensive ignimbrite eruptions accompanied normal faulting and block subsidence. The Bruce River volcanics, with their generally cyclic nature, in comparison to sequences represented in these more recent fields, are therefore probably the product of several episodes of overlapping stratavolcano construction and caldera formation.

Summary of Depositional History of Bruce River Group

The Heggart Lake Formation conglomerates and sandstones are products of rapid erosion and alluvial fan – floodplain deposition in a restricted graben (Fig. 20.6a). Flanking syndepositional faults exposed earlier Helikian (?) volcanic rocks and granitoids, Aphebian supracrustals, and Archean gneisses reworked during the Hudsonian Orogeny. Minor tephra eruptions and fissure flows accompanied sedimentation.

During deposition of the Brown Lake Formation the basin stabilized and topographic relief was significantly reduced. Brown Lake Formation sands blanketed the earlier graben and overstepped its flanks to rest on older rocks on both sides. The sandstones were derived through fall out from ash clouds blown in from active volcanoes southeast (?) of the basin, and were deposited in a low energy aeolian or shallow lacustrine environment (Fig. 20.6b).

The Sylvia Lake Formation prograding lavas and other volcanic products dominate the remainder of the depositional history of the Bruce River Group. The earliest basaltic to andesitic lavas represent initial "quiet" eruptions at the base of a growing stratovolcano. The upper Sylvia Lake Formation is characterized by extensive sheets of trachytic and rhyolitic ignimbrites indicative of repeated explosive volcanism, interrupted periodically by minor intermediate to mafic lavas and pyroclastic eruptions (Fig. 20.6c). The cyclic nature of the volcanic pile suggets that it records a repetitive history of stratovolcano development and caldera formation.

Letitia Lake Group

Introduction

The Letitia Lake Group (Fig. 20.7) outcrops as a narrow belt south of the Seal Lake Group (Fig. 20.2; Mann, 1959; Brummer and Mann, 1961; Thomas, 1979). It is dominated by felsic porphyries which have been traditionally considered to be equivalent to those of the Bruce River Group (cf. Brummer and Mann, 1961). However, the presence of phenocrysts of quartz, and the peralkaline character of the rocks (Thomas, 1980) rule out direct lithological correlation, although a time equivalence is possible. The Letitia Lake Group is intruded by a series of alkaline rocks, referred to as the Red Wine intrusive suite and Arc Lake intrusive suite, which have yielded a Rb-Sr age of 1392 ± 75 Ma (Curtis and Currie, 1981). It is unconformably overlain by the Seal Lake Group, and has been polydeformed during the Grenvillian Orogeny (cf. Marten, 1975; Thomas, 1979; Calon and Hibbs, 1980). No formal subdivision of the group has yet been attempted.

Lithology and Geochemistry

The Letitia Lake Group comprises subvolcanic quartz-feldspar porphyry, its extrusive equivalents, and porphyry derived sediment.

Felsic quartz-feldspar porphyry, with buff to white alkali feldspar and quartz phenocrysts in a white weathering, fine grained siliceous groundmass occurs at the base of the group. Quartz phenocrysts increase and feldspar phenocrysts decrease in proportion upwards in the unit.

The lower intrusive unit grades vertically and horizontally into rhyolite flows and ignimbrites which consist of euhedral and broken feldspar phenocrysts (av. 4 mm) in a pink aphanitic groundmass interpreted as extrusive equivalents of the lower unit.

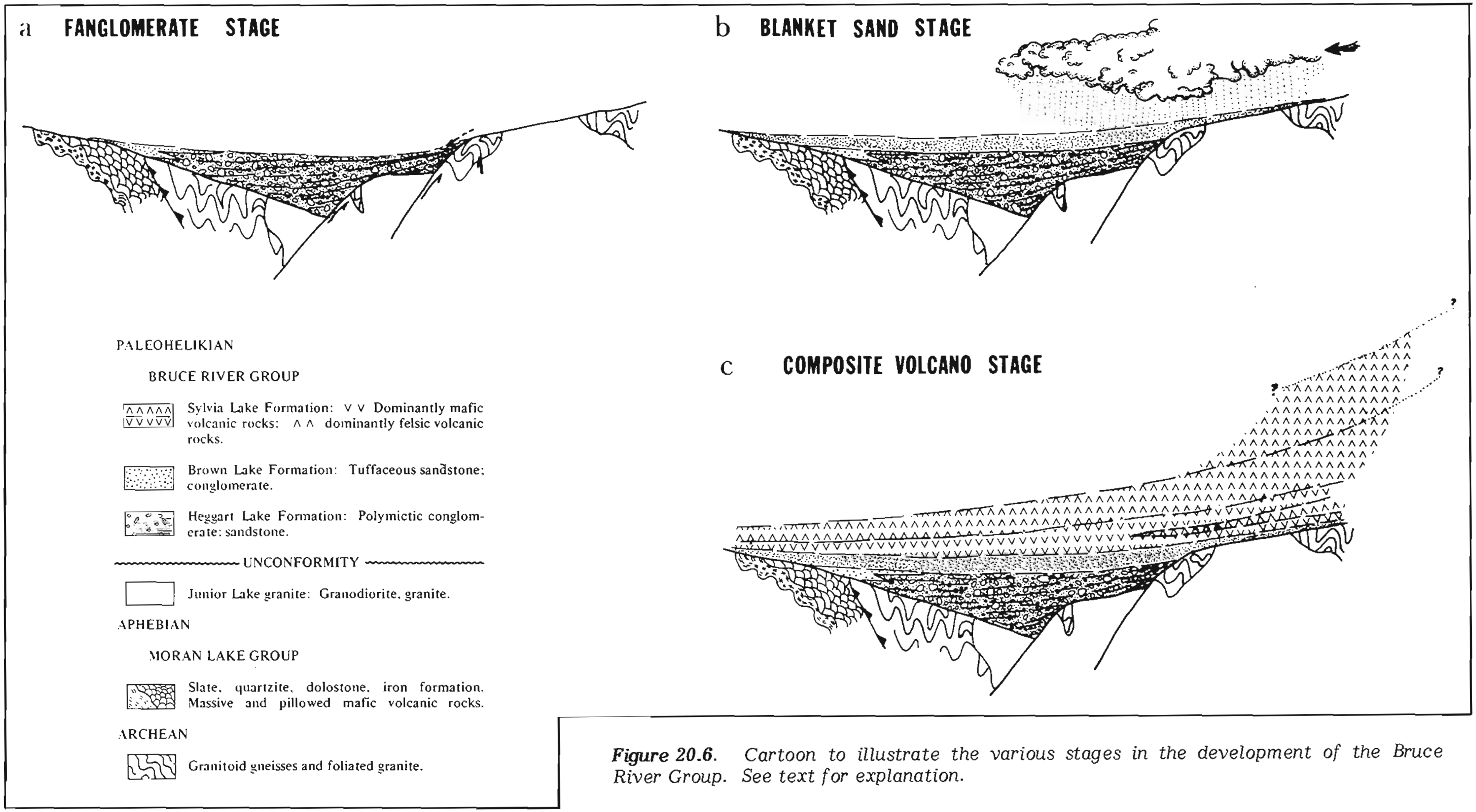


Figure 20.6. *Cartoon to illustrate the various stages in the development of the Bruce River Group. See text for explanation.*

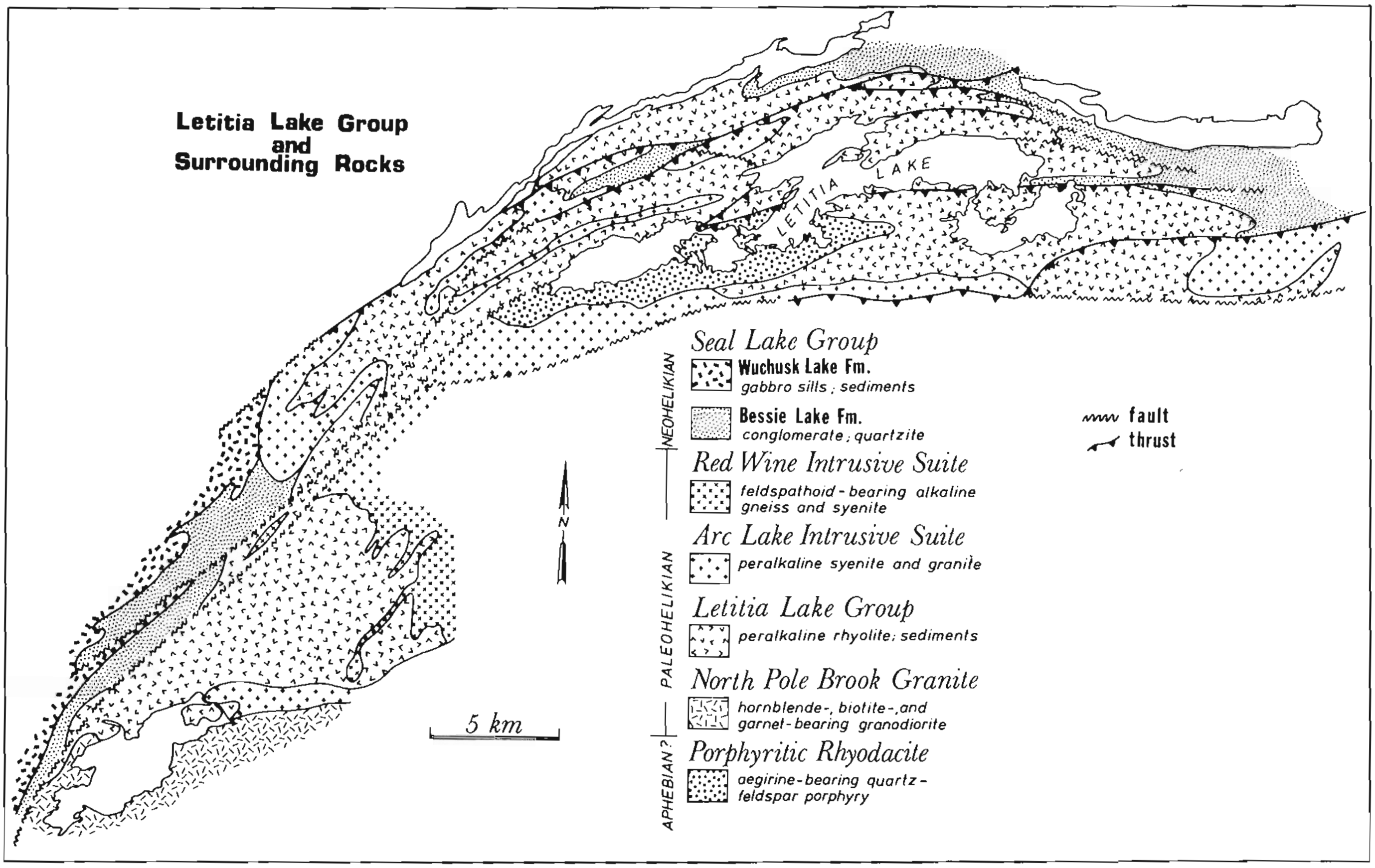


Figure 20.7. Generalized geology of the Letitia Lake Group and surrounding rocks (from Thomas, 1980).

Both porphyritic units described above were weathered during the late Paleohelikian or early Neohelikian and a regolith consequently developed. The overlying porphyry derived sediments have been considered basal Seal Lake Group by some workers, but Calon and Hibbs (1980) considered them to be an integral part of the Letitia Lake Group, and place the unconformity with the Seal Lake Group at the base of the overlying quartzite-conglomerate units.

Both major and minor element chemistry indicate a strong peralkaline affinity for Letitia Lake Group volcanics (Thomas, 1980). Fresh, unweathered, acid porphyries have agpaitic indices greater than or equal to 1, and exhibit major element oxide distributions typical of crystalline comendites. In addition, all rocks of the group contain high contents of the elements Zr, Y, La, and Ce, also characteristic of comendites.

MIDDLE HELIKIAN INTRUSIVE ROCKS

Middle Helikian plutonic activity in the Central Mineral Belt is recorded by the Otter Lake granite and the Harp Lake Complex. The Otter Lake granite intrudes the Bruce River Group, and has been dated at 1496 ± 37 Ma (Kontak, 1979). The Harp Lake Complex comprises anorthositic, gabbroic and granitic rocks emplaced 1450 ± 25 Ma ago (Emslie, 1980) as part of the Elsonian event (Emslie, 1978a, b).

Otter Lake Granite

Medium- to coarse-grained biotite ± hornblende granodiorite and monzonite are the two most widespread phases of the Otter Lake pluton (Fig. 20.2). However, in the vicinity of Nipishish Lake these rocks are intruded by fine grained muscovite-biotite granite and aplite sheets up to 8 km in width.

The coarse granodiorite and monzonite are locally characterized by numerous irregular xenoliths, generally a few centimetres to a metre in maximum size, of massive and porphyritic diorite. However, discreet diorite bodies up to several kilometres in size occur in part of the granite.

The similarity between the composition of the intrusive rocks of the Otter Lake granite and the extrusive rocks of the Sylvia Lake Formation of the Bruce River Group, coupled with their similar ages suggest that the plutonic rocks were the parent magma of the volcanic rocks. The magma probably rose into the volcanic pile in an intimate relationship with its extrusive products (cf. Myers, 1975; Hoffman and McGlynn, 1977) because of cauldron collapse.

Kontak (1979) noted that the Sr_O of 0.7051 ± .0013 obtained from this granite suggests a lower crust or upper mantle source, but pointed out that the association with vast volumes of high potassium felsic ignimbrites is more compatible with a crustal origin.

Harp Lake Complex

The Harp Lake Complex (Emslie, 1980) which outcrops north of the Seal Lake Group (Fig. 20.1), is typical of Elsonian magmatism. It comprises at least 20 major igneous subunits ranging from gabbro, troctolite, leucotroctolite, leucogabbro, leuconorite, anorthosite and diorite to granodiorite, adamellite and granite, reflecting several pulses of magmatic activity.

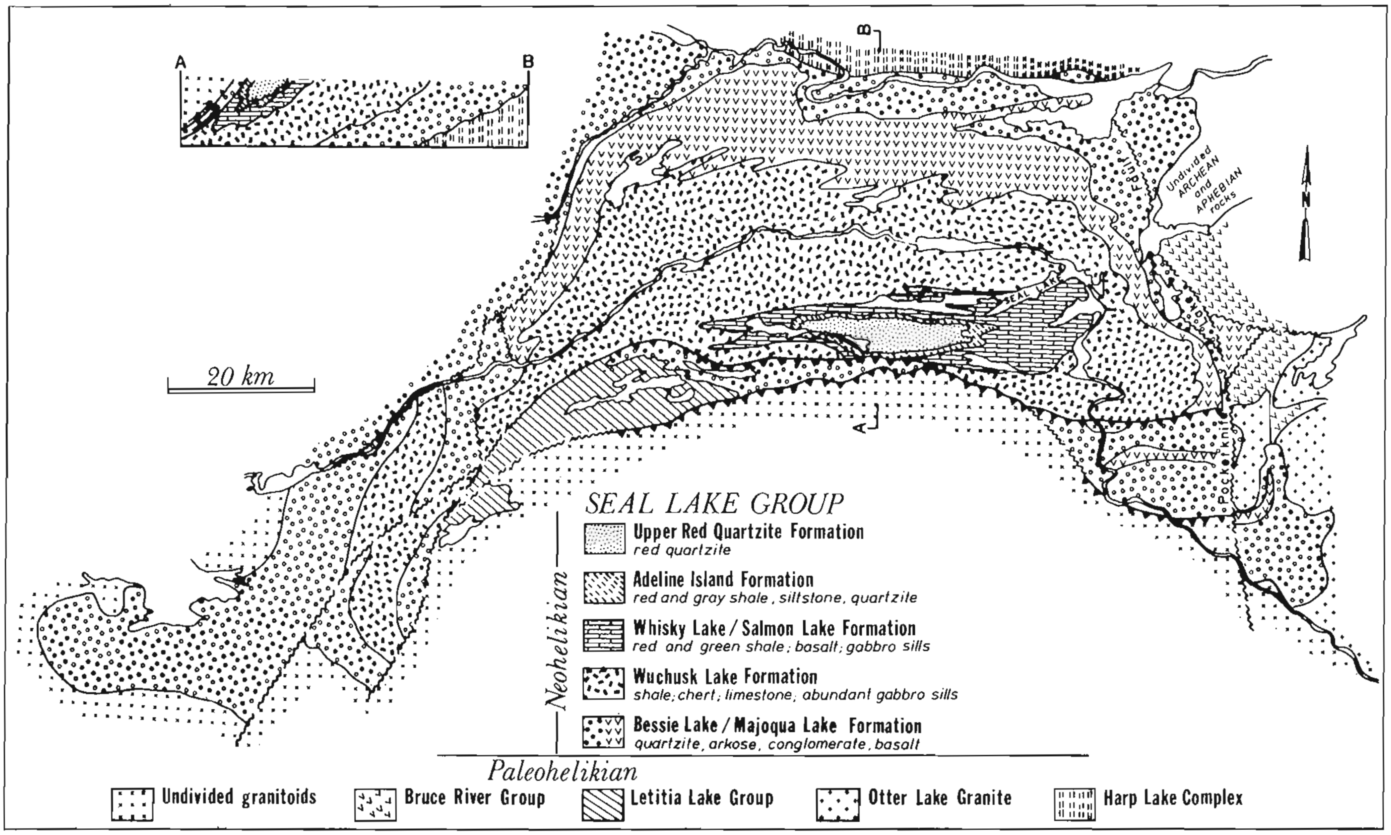


Figure 20.8. *Generalized geological map of the Seal Lake Group and surrounding rocks.*

Emslie (1978b) has argued that the bimodal plutonic assemblage of the Elsonian event was the result of the emplacement of olivine tholeiite magmas along the crust-mantle interface. These magmas fractionated at the base of the crust to produce high-Al gabbro magmas which migrated upwards to crystallize plagioclase in intracrustal chambers producing the gabbro-anorthosite suite. Crystallization of the olivine tholeiite magmas at the base of the crust also led to partial fusion of the lower crust, and these granitoid melts also coalesced and rose upward into the crust forming the adamellite suite.

NEOHELIKIAN SUPRACRUSTALS

Deposition during the Neohelikian subera in central Labrador is represented by the Seal Lake Group. This is the youngest supracrustal sequence in the Central Mineral Belt, yielding a Rb-Sr age of 1323 ± 92 Ma (Barager, 1978). It is in unconformable contact with all rocks described above.

Seal Lake Group

Introduction

The Seal Lake Group (Fahrig, 1959; Brummer and Mann, 1961; Roscoe and Emslie, 1973) is a sequence of deformed sedimentary, volcanic and intrusive rocks exposed in an east-west trending, doubly-plunging, synclinorium west of the Bruce River Group (Fig. 20.1, 20.8). The southern limb of this fold is overturned and overridden by older granites which have been thrust northward over the supracrustals. Along the northern limb of the synclinorium the group rests unconformably on Archean granite and gneiss, and the Helikian Harp Lake Complex (Emslie, 1980) of anorthosite, gabbro and granitoids. The major portion of the group outcrops west of the Pocketknife Lake Fault zone, but the basal formation is locally preserved east of the fault where it is in unconformable contact with the Sylvia Lake Formation of the Bruce River Group, and the Otter Lake granite (Marten and Smyth, 1975).

The stratigraphic terminology employed here is chiefly that of Brummer and Mann (1961) but subsequent modifications are noted. The thickness estimated by Brummer and Mann (1961) for the group was 12 600 m, but this was reduced to 5820 m by Knight (1972). The latter estimate takes into account repetition of parts of the sequence by folding and thrusting (see also Gandhi and Brown, 1975, p. 147).

Bessie Lake Formation/Majoqua Lake Formation

The Bessie Lake Formation outcrops along the southern margin of the Seal Lake Group. In the east it rests unconformably on the Bruce River Group and Otter Lake granite; in the west it rests unconformably on the Letitia Lake Group. It comprises clean, massive, pink and white quartz arenites, arkoses, grey and black shales, and intercalated basic volcanic rocks. Granule and pebble conglomerate beds are present locally.

Along the northern limb of the Seal Lake synclinorium, the basal sequence is referred to as the Majoqua Lake Formation, and comprises arkoses, conglomerate and thick amygdaloidal mafic flows, which rest unconformably on Archean granite and gneiss and on the Harp Lake Complex. Knight (1972) suggested that "Majoqua Lake Formation" should refer only to the volcanic rocks; the basal sedimentary rocks being given formal status as the Arkose Lake Formation. In this account the name Bessie Lake Formation is used to refer to the basal association of sediments and volcanic rocks on both limbs of the syncline, although it is recognized that such subdivision is warranted. Thickness estimates of the formation given by Brummer and Mann (1961) range from 1260 m in the south to 3750 m in the north; Knight (1972) considered that 1500 m is a maximum for the northern succession.

Bessie Lake Formation sediments along the southern limb of the synclinorium are at least 90 per cent quartz arenites. These are commonly white, or blue-grey, but pink, buff, and pale green varieties are also present. They are massive, composed predominantly of equigranular quartz with minor altered feldspar; magnetite grains outline bedding. Strongly foliated overturned, bedded quartzite and quartz-sericite schists are common along the southern margin of the Seal where it abuts foliated granite overthrust from the south. Along the northern margin of the group, the sediments are less mature, and comprise chiefly arkoses and conglomerates.

Basalts become common towards the top of the succession, and constitute a significant portion (900 m) of the formation in the north. In the south the volcanics are metamorphosed and deformed to dark grey and green chlorite schists with quartz stringers, epidote and specularite. In the north they are fine grained, locally columnar jointed, subaerial, basalts with pyroxene (augite) and plagioclase phenocrysts. Quartz, calcite, and zeolites occur as vesicle fillings. Laterites are locally developed on surfaces of the flows, and rare pillows have been documented (Knight, 1972).

Wuchusk Lake Formation

This formation overlies the Bessie Lake Formation and comprises thin bands of buff to red quartzite, red shale, cream to black chert, and grey to pink stromatolitic and oolitic limestone. The sedimentary strata have been injected by numerous diabase and gabbro sills that vary from 30-600 m in thickness, and are locally very coarse grained and pegmatitic. Although Brummer and Mann (1961) included the sills in their definition of the formation, recent amendments to the Stratigraphic Code (Sohl, 1977) dictate that they should be redefined as a separate intrusive suite within the Seal Lake Group.

Whisky Lake Formation

This formation consists of 900 m of red shale, grey and green slates, red siltstones and red quartzites (Brummer and Mann, 1961). Baragar (1969) omitted this formation from his stratigraphic compilation of the group. Kidd (1968) and Hale (1968) suggested that the Whisky Lake Formation is equivalent to the upper part of the Wuchusk Lake Formation and the lower part of the overlying Salmon Lake Formation; Gandhi and Brown (1975) included the Whisky Lake with the Salmon Lake.

Salmon Lake and Adeline Island Formations

These formations, with a combined thickness of 1350 m (Baragar, 1969) are similar although numerous basalt flows occur in the Salmon Lake. The sedimentary rocks are chiefly red, maroon, green-grey, and black shales, and pink and white locally crossbedded quartz arenites. Bimodal arenites (Kidd, 1968) are more common in the Adeline Island Formation where they display small scale current bedding features and asymmetrical ripple marks.

Amygdaloidal basalts of the Salmon Lake Formation are chiefly grey to green, locally flow banded and columnar jointed flows. They are transected by native copper- and chalcocite-bearing quartz-calcite veins. The majority of the more than 250 copper mineral showings known in the Seal Lake Group occur in these basalts, but the uppermost grey and green shales of the Adeline Island Formation are also cupriferous (cf. Gandhi and Brown, 1975).

Upper Red Quartzite Formation

This is the highest stratigraphic unit of the Seal Lake Group. It comprises massive to well bedded red quartzites, estimated by Brummer and Mann (1961) to be 780 m thick, but by Baragar (1969) to be only 450 m thick.

Geochemistry of Seal Lake Group Basalts

The Seal Lake volcanic rocks are typical plateau basalts ranging from 3-50 m in thickness, and grading from massive bases to amygdaloidal, oxidized flow tops (Baragar, 1969). Chemically they are transitional between alkali and tholeiitic magma types (Knight, 1972; Baragar, 1974, 1977).

Environment of Deposition of the Seal Lake Group

Deposition of the Seal Lake Group was initiated on a peneplaned surface of pre-Neohelikian granite, gneiss, anorthosite and volcanic rocks. The gross character of the red fluvial clastics and oxidized columnar jointed basalts with rare pillow lava are indicative of a continental environment. Deposition probably occurred in an elongate east-west (?) basin in which conditions changed from wholly terrestrial in the north to shallow marine (?) in the south. Climatic conditions were probably arid to semi-arid (Brummer and Mann, 1961; Knight, 1972). Paleomagnetic data suggest that this portion of Labrador was located near 10°S latitude at this time (Roy and Fahrig, 1973; Irving, 1979). The following evolutionary history of deposition is taken largely from Mann (1959) and Brummer and Mann (1961).

The Bessie Lake Formation along the south limb of the Seal synclinorium appears to have been deposited in a shallow marine (?) environment. Contemporaneous coarse arkoses and conglomerates in the north (the Arkose Lake Formation of Knight, 1972) indicate more terrestrial conditions dominated by alluvial fan and braided stream deposition from a northwest receding source area undergoing active block faulting (Knight, 1972).

Volcanism associated with the early deposition of the Seal appears to have been almost totally subaerial. There was a sufficient interval of exposure between each basaltic eruption for the top of the earlier flow to undergo lateritic weathering and erosion prior to burial by the next flow (Knight, 1972). Pillows may be the result of lava flowing into ephemeral lakes on older flow surfaces.

Brummer and Mann (1961) suggested that the finely banded shales and cherts of the Wuchusk Lake were deposited during transition to quieter, possibly lacustrine conditions. Ripple marked and crossbedded quartzites and oolitic and algal limestones were deposited on shallow, high energy (?) shoals.

Red shales and quartzites of the Whisky Lake Formation accumulated in a shallow water (playa lake?) environment which persisted through Salmon Lake time. Sedimentation was accompanied by extrusion of basaltic flows, and concomitant intrusion of diabase sills in the Wuchusk Lake Formation.

During Adeline Island deposition the environment of sedimentation changed from quiet, shallow water to high energy(?) strandline which continued through deposition of the Upper Red Quartzite Formation.

The stratigraphic and geochemical evidence from the Seal Lake basin suggests the sequence formed in an intracontinental rift zone (cf. Baragar, quoted by Baer, 1974; Gandhi and Brown, 1975; Baragar, 1977).

GRENVILLIAN DEFORMATION

The present disposition of the supracrustal rocks discussed in this paper suggests northward compression during Grenvillian deformation. The foreland zone rocks exhibit increasing metamorphic grade and structural complexity towards the mobile belt, and the south margin of the Seal Lake Group is characterized by northward directed lobate thrust and high angle reverse faults.

A north-to-south increase in metamorphic grade from prehnite-pumpellyite to chlorite-epidote is well displayed by the Seal Lake Group (Mann, 1959; Baragar, 1974). Similarly the Bruce River Group shows a transition from chlorite grade to biotite grade and the Letitia Lake Group shows a transition from middle to upper greenschist grade (Thomas, 1980) from north to south.

The Bruce River Group contains one regional penetrative slaty cleavage or mineral foliation attributed to Grenvillian deformation. However, along the southern margin of the Seal Lake Group and within the Letitia Lake Group there is evidence of a polyphase structural history (Marten, 1975; Thomas, 1979, 1980; Calon and Hibbs, 1980). In these areas, first generation isoclinal reclined non-cylindrical folds are refolded by open to tight, second generation structures associated with reverse faults.

The Seal Lake Group was overridden along its southern margin by older granitoid rocks thrust northward during the late stages of deformation. This thrust fault has a lobate form, characteristic of the whole of the Grenville Front Zone in southern Labrador (Greene, 1974; Gower et al., 1980). The attitude of this structure changes from a reverse fault in the zone of maximum curvature, to a strike-slip fault where the trend is subparallel to the transport direction (Fig. 20.8). Although there is some evidence of local thrusting in the Bruce River Group (Fig. 20.2), the major faults (Bruce Lake, Gravelly River, and Minisinakwa Lake) appear to be for the most part strike slip structures.

Post-Grenvillian scissor movement along the Pocketknife Lake Fault (Fig. 20.8) is postulated to be a rupturing of the extension of the junction between the Nain and Churchill structural provinces which outcrops along the northern boundary of the Central Mineral Belt (Smyth et al., 1975). The upthrown block, on which the Bruce River Group occurs, seems to have undergone "intra block" normal faulting along the Bruce Lake, Gravelly River and Minisinakwa Lake faults. This is suggested by the thicker and more extensive sequences of lower Seal Lake Group rocks south of the Gravelly River and Minisinakwa Lake faults, indicative of increasing downthrow from northwest to southeast.

DISCUSSION

The earliest Helikian rocks in central Labrador are the voluminous felsic plutonic associations along the southern border of the Central Mineral Belt and the northern margin of the Grenville Province. Radiometric ages indicate that this magmatism commenced shortly after the close of the Hudsonian Orogeny. Emplacement of such an extensive belt of plutonic rocks into the crust could have been accompanied by regional crustal doming and extension and the formation of graben in which contemporaneous supracrustal rocks were deposited (cf. Hills, 1959; Bridgwater et al., 1974; Bridwell, 1978; Ramberg, 1971). The Bruce River – Otter Lake volcano-plutonic complex (circa 1540-1495 Ma) may be the product of the late stages of this post-Hudsonian magmatic event, the supracrustal rocks being fortuitously preserved because they lie north of the main belt of crustal uplift associated with the Grenvillian Orogeny. Any supracrustal rocks which formed in the southern area have subsequently been eroded from this uplifted zone.

As noted earlier, the volcanic rocks of the Bruce River Group are calc-alkaline, high-K assemblage comparable to the shoshonite association. It is now commonly accepted that rocks of the "calc-alkaline series" are diagnostic members of volcanic provinces developed above subduction zones (cf. Garcia, 1978), and that where such rocks occur in abundance in the geologic record, they reflect sites of former lithospheric plate convergence.

In subduction zones the high-K calc-alkaline and shoshonite subseries are erupted on crust above the deepest portions of the subducted slab or in the late stages of arc development (Dickenson, 1968; Jakes and White, 1972; Johnson et al., 1978). However, high-K calc-alkaline and shoshonitic suites are also found in volcanic provinces which are related to epeirogenic movements during crustal stablization in orogenic zones (Joplin, 1968; Mackenzie and Chappell, 1972; Carmichael et al., 1974) following either cessation of subduction (Girolamo, 1978) or complete continental suturing (Dewey and Burke, 1973; Burke et al., 1974; Kidd, 1975; Sengor and Kidd, 1979). Therefore, without considering other aspects (such as overall geological setting) of high-K volcanic rocks, it becomes nearly impossible to distinguish between collisional and arc-related magmatic products by their chemical character alone (Burke and Kidd, 1980). The Bruce River Group overlies a Hudsonian deformed terrane of Aphebian supracrustals and reworked Archean gneisses, contains fanglomerates (the Heggart Lake Formation) which are interpreted to have formed under a block-faulting regime, and includes a volcanic pile (the Sylvia Lake Formation) which appears to be the surface expression of the late stages of a long lived (1650-1500 Ma) episode of post-Hudsonian magmatism. These features suggest that the Bruce River Group accumulated in a depositional basin formed by vertical crustal movements during post-Hudsonian continental stablization.

The peralkaline affinity of the Letitia Lake Group volcanic rocks is also evidence for an extensional regime (Macdonald, 1975) in the Paleohelikian, but this volcanism is obviously not directly related to that of the Bruce River Group.

If the Hudsonian Orogeny in the Makkovik subprovince to the north of the Paleohelikian supracrustals is a product of lithospheric plate convergence (cf. Wardle and Bailey, 1981) then the formation of the Paleohelikian granites and depositional basins in central Labrador may be related to postsubduction/postsuturing processes like those in the areas referred to above.

The next major event in the development of middle Proterozoic basins of central Labrador was the emplacement of the anorogenic Harp Lake Complex circa 1450 Ma ago (Emslie, 1980). This is just one member of a number of such plutons emplaced during the Elsonian plutonic event (Emslie, 1978a).

Intrusion of the Elsonian plutons was followed by rapid uplift and a new period of crustal attenuation. Erosion exposed the Harp Lake Complex in a new basin system in which the terrestrial to shallow water sediments and subaerial plateau basalts of Seal Lake Group accumulated circa 1325 Ma ago. Baragar (1977) has suggested that the Seal Lake Group eruptions took place in a graben system related to a major Helikian rift zone in the Canadian Shield which paralleled the trend of, and lay within, the present Grenville Province.

All Helikian rocks have been affected by the Grenvillian Orogeny. Generally there is evidence of only one period of deformation, but along the southern periphery of the supracrustal belt a polyphase history is apparent, and basement is thrust over cover.

In conclusion the Helikian basins in Labrador reflect two distinct episodes of crustal extension. The Paleohelikian strata indicate linear crustal attenuation probably associated with a long lived episode of post-Hudsonian granitic magmatism. The Neohelikian strata may be remnants of a more extensive rift zone assemblage possibly developed in a zone of significant pre-Grenvillian crustal separation. The only feature they have in common with respect to the Grenvillian Orogeny is that both are deformed by it.

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21.

EVOLUTION OF EARLY PROTEROZOIC BASINS OF THE GREAT LAKES REGION

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Sims, P.K., Card, K.D., and Lumbers, S.B., Evolution of early Proterozoic basins of the Great Lakes region; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 379-397, 1981.

Abstract

The two early Proterozoic supracrustal sequences of the Southern Province – one in the Lake Superior region, and the other extending east from Lake Huron into the Grenville Province – are composed of clastic sedimentary and volcanic rocks with subordinate chemical sedimentary rocks, chiefly iron formation. Although of somewhat different age, the two sequences have similar asymmetric patterns of stratigraphy and structure. Both sequences thicken southward; their thicker parts are intensely deformed, metamorphosed, and intruded locally by granitoid plutons. Except in the Grenville Province, the two sequences have a common tectonic setting and analogous ancestral sedimentary basins.

The Proterozoic basins of the Southern Province are aligned along a zone of weakness that represents the boundary between late Archean greenstone-granite complexes (2750-2600 Ma) of the Superior Province on the north and an early Archean (3500-3000 Ma) gneiss terrane on the south. These two crustal segments were welded together at the end of the Archean. Though apparently tightly juxtaposed thereafter, the two Archean basement terranes differed vastly in tectonic stability. Their boundary was a hinge during deposition of the supracrustal rocks, and a tectonic front during two main stages in the development of the early Proterozoic orogen.

During the first stage, sedimentary basins evolved in an intracratonic setting, probably far from a continental margin, but possibly crossing eastward to such a margin in the Grenville Province. Initial rift-faulting, concentrated in a zone along and adjacent to the boundary between the two basement segments, probably migrated westward with time and provided sites for deposition of detritus shed from the inner part of the craton to the north. During subsidence of the more mobile gneissic Archean crust (3500-3000 Ma) thick coalescing turbidite sheets spread southward into the resulting basins. The basin in the Lake Superior region also received tholeiitic and calc-alkalic volcanic rocks intimately intercalated with the turbidites.

The terminal, compressional stage ended deposition. The supracrustal rocks and the gneiss basement were deformed together and metamorphosed under low to moderate temperature-pressure conditions, as a result of widespread diapiric doming of the gneiss crust and lateral transport of this crustal segment and its deformed cover against the more rigid greenstone-granite crust.

Thick, early Proterozoic supracrustal sequences of the northwestern Grenville Province were highly deformed, metamorphosed, and invaded by felsic plutonic rocks during early and late Proterozoic events. Here, deposition and subsequent orogenesis probably occurred in a continental margin environment.

Résumé

Les deux successions supracrustales de la province du Sud, qui sont du début du Protérozoïque – l'une située dans la région du lac Supérieur, l'autre allant du lac Huron à l'est jusqu'à l'intérieur de la province de Grenville – consistent en roches clastiques d'origine sédimentaire et volcanique, et contiennent accessoirement des roches formées par sédimentation chimique, en particulier des formations ferrifères. Bien qu'elles soient d'âge légèrement différent, les deux successions présentent une stratigraphie et une structure similaires, de caractère asymétrique. Elles montrent un épaississement vers le sud; les couches épaisses sont intensément déformées, métamorphisées, et localement pénétrées par des plutons granitoïdes. Excepté dans la province de Grenville, les deux successions ont un cadre tectonique commun, et pour origine des bassins sédimentaires analogues.

Les bassins protérozoïques de la province du Sud sont alignés le long d'une zone tectoniquement fragile, qui représente la limite entre les complexes granitiques et à roches vertes de la fin de l'Archéen (2750-2600 Ma) situés dans la province du lac Supérieur au nord, et un terrain gneissique du début de l'Archéen (3500-3000 Ma) situé au sud. Ces deux segments crustaux se sont soudés l'un à l'autre à la fin de l'Archéen. Bien qu'apparemment, ils soient ensuite restés étroitement juxtaposés, les deux soubassements archéens ont présenté de grandes différences de stabilité tectonique. Leur ligne de séparation constituait une charnière pendant le dépôt des roches supracrustales, et un front tectonique pendant les deux principales phases d'orogénèse du début du Protérozoïque.

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Pendant la première phase, les bassins sédimentaires ont évolué dans un cadre intracratonique, probablement loin d'une marge continentale, sauf dans la province de Grenville, où ils rejoignaient peut-être cette marge continentale à l'est. Les effondrements initiaux, qui avaient principalement touché une zone longeant et bordant la limite entre les deux parties du soubassement, se sont probablement déplacés vers l'ouest au cours des temps, et ont ainsi créé des bassins de sédimentation, où se déposaient les débris issus de la portion interne du craton, au nord. Pendant la subsidence de la croûte archéenne gneissique plus mobile (3500-3000 Ma), d'épaisses nappes de turbidites ont fusionné et se sont étalées vers le sud dans les bassins ainsi formés. Dans le bassin de la région du lac Supérieur, se sont déversées des roches volcaniques tholéiitiques et calco-alcalines, interstratifiées avec les turbidites.

La phase finale de compression a mis fin à la sédimentation. Les roches supracrustales et le soubassement gneissique ont été simultanément déformés et métamorphisés en présence de températures et pressions faibles à moyennes, sous l'effet d'un vaste bombement diapirique de la croûte gneissique, et du transport latéral de ce segment crustal et de sa couverture déformée contre la croûte granitique et à roches vertes, qui était plus rigide.

Les épaisses successions supracrustales datant du début du Protérozoïque du nord-ouest de la province de Grenville ont été fortement déformées et métamorphisées, puis envahies par des roches plutoniques felsiques au début et à la fin du Protérozoïque. Dans cette région, il y a probablement eu sédimentation, puis orogénèse dans un milieu de marge continentale.

INTRODUCTION

Major sedimentary-volcanic sequences of early Proterozoic age in the Great Lakes region, the Marquette Range Supergroup and equivalent successions in the Lake Superior region, the Huronian Supergroup in the Lake Huron region, and equivalent rocks in the Grenville Province, are aligned along strike but physically separated by the late Proterozoic (Keweenawan) Midcontinent Rift System and the Grenville Front. The Southern Province sequences have many similar internal features, including asymmetrical patterns of stratigraphy and structure, and share a common tectonic setting. Both unconformably overlie Archean basement rocks and are interpreted as having formed in analogous ancestral intracontinental basins. A unifying structure that can account for many of the shared features is the boundary in the basement between the two Archean crustal segments that has been delineated in the Lake Superior region (Morey and Sims, 1976; Sims, 1980) and is proposed herein to exist in the Lake Huron region.

Early Proterozoic sequences of the northwest Grenville Province suffered more prolonged, intense metamorphism, deformation, and plutonic activity than did their Southern Province counterparts. Archean basement rocks have not been recognized in the Ontario Grenville except locally near the northwest margin of the Province. Accordingly, a different tectonic regime, probably a continental margin environment, is postulated for the early Proterozoic sequences of the northwest Grenville Province.

The main purpose of this paper is to document the relationships between the early Proterozoic basins and the contrasting types of basement rocks. We contend that the boundary zone between the two basement crustal segments not only localized initial rifting of the crust during the early Proterozoic but played an active part in subsequent tectonism in the Southern Province. The discontinuity was established in the late Archean, when the two crustal segments were imperfectly welded together, and was repeatedly rejuvenated in the Proterozoic, being the locus of both regional extension and compression. It controlled basin evolution in the Southern Province in an intracratonic setting and extended eastward to a continental margin regime in the Grenville Province. Because of its geologic significance and extent, this boundary zone has been named the Great Lakes tectonic zone (Sims et al., 1980).

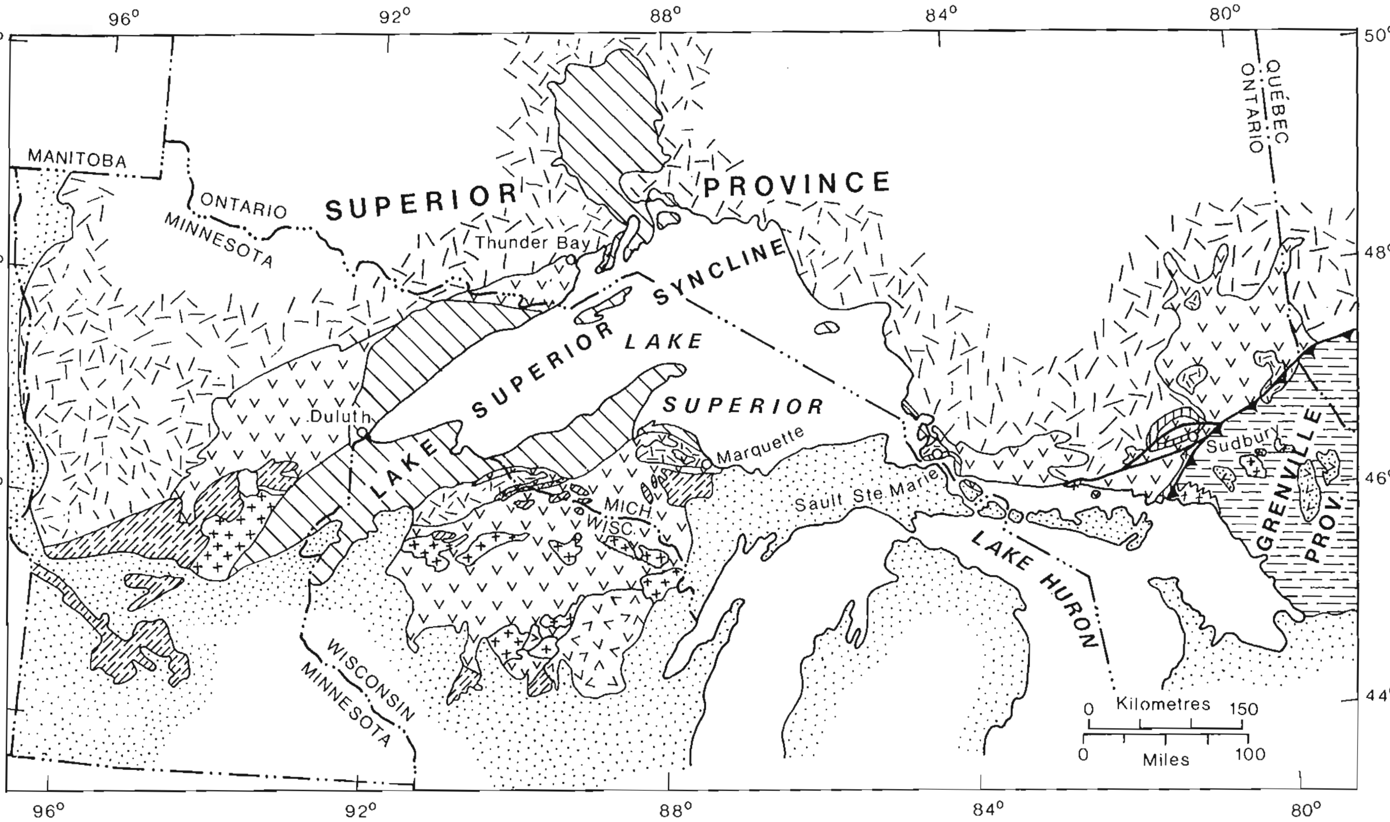
For convenience, in this report late Archean and early Archean are used as informal terms to distinguish rocks that are respectively younger and older than 3000 Ma. Early Proterozoic refers to rocks that are 2500 to 1600 Ma old and late Proterozoic to rocks 1600 to 570 Ma.

GENERAL GEOLOGY

The early (1600-2500 Ma) Proterozoic supracrustal sequences in the Lake Superior and Lake Huron regions constitute a discontinuous linear fold belt some 1300 km long extending from Minnesota into eastern Ontario along the southern margin of the Archean Superior Province (Fig. 21.1). They compose the major part of the Southern Province (Stockwell et al., 1970) of the Hudsonian foldbelt along the United States-Canada border on the Tectonic Map of North America (King, 1969) and also extend into the northwestern Grenville Province in Ontario. The sequences are transected at both ends of Lake Superior by the late Proterozoic (Keweenawan) Midcontinent Rift System, and in northeastern Ontario by the late Proterozoic Grenville Front tectonic zone, which largely obliterates the primary features of the early Proterozoic rocks in the Grenville Province. The sequences are overlain to the south by flat-lying Paleozoic strata.

From west to east, the sequences consist of the Animikie and Mille Lacs groups in Minnesota (Morey, 1978); the Marquette Range Supergroup in northern Michigan and Wisconsin; and the Huronian Supergroup in northeastern Ontario and rocks of comparable age in the Grenville Province. These strata were deposited between 2500 and 1900 Ma ago, and were subsequently deformed, metamorphosed, and intruded by plutonic rocks. Except for those early Proterozoic rocks in the Grenville Province, the major tectonic event which affected these successions was the Penokean Orogeny (1900-1800 Ma). The last major tectonic event affecting the early and late Proterozoic rocks in the Grenville Province was the Grenvillian Orogeny (1200-1000 Ma).

The early Proterozoic sequences are composed mainly of clastic sedimentary and volcanic rocks with subordinate chemical sediments, chiefly iron formation. They thicken southward from a thin erosional edge to a maximum of more than 10 km. Although in general they bear some similarities to Phanerozoic geosynclinal sequences, they lack rocks characteristic of the ophiolite, synorogenic flysch, and post-orogenic molasse suites of Phanerozoic orthogeosynclines.



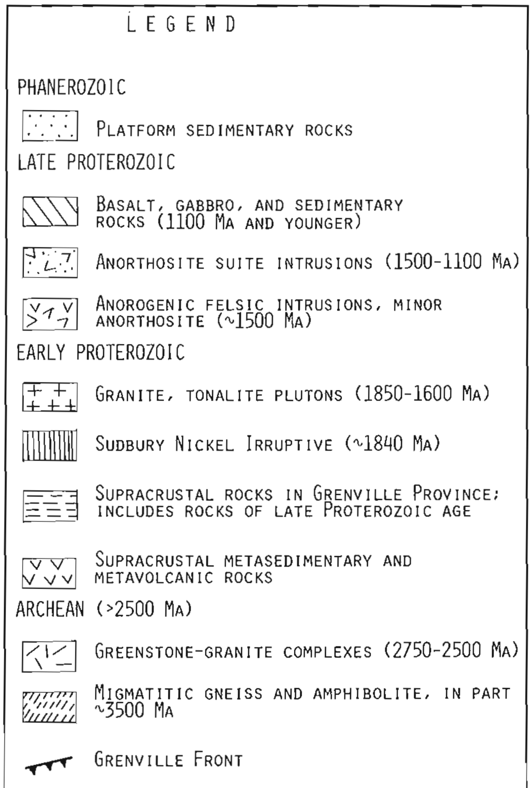


Figure 21.1. *Generalized geologic map of the Great Lakes region. Modified from King (1969), with additions from Card (1978a), Sims (1976), and Sims et al. (1978).*

The successions begin with shallow water sediments and end with deep water sediments and volcanics, the opposite of many younger geosynclinal sequences. The sediments were derived mainly from the adjacent Archean craton to the north and consequently seem to lack the distinctive bipolarity characteristic of many younger geosynclines.

The Archean rocks that compose the basement for the Southern Province supracrustal rocks are of two contrasting types, which differ in age, rock assemblages, metamorphic grade, and structural style (Morey and Sims, 1976). Greenstone-granite complexes of late Archean age (2750-2600 Ma), typical of most of the southern part of the Superior Province (Peterman, 1979), underlie the northern parts of the basins. Migmatitic gneiss and amphibolite, in part about 3500 Ma old, underlie the southern part of the basin in the Lake Superior region (Van Schmus and Anderson, 1977), and are inferred also to be present in the Lake Huron region. Although gneisses of early Archean age are not exposed in the Lake Huron region, felsic gneiss and amphibolite beneath the Paleozoic cover of Manitoulin Island (see Fig. 21.7; Van Schmus et al., 1975a) are possible correlatives. The type area for the gneiss terrane is the Minnesota River valley in southern Minnesota (Goldich et al., 1961, 1970). The location of the boundary between the two basement terranes is shown in Figure 21.3.

LAKE SUPERIOR REGION

The early Proterozoic sedimentary and volcanic rocks of the Animikie and Mille Lacs groups and Marquette Range Supergroup were deposited in a basin that formed some 2200 to 1900 Ma ago (Goldich, 1973). The sequence was deformed, metamorphosed, and locally intruded by granitic plutons during the Penokean Orogeny, 1900-1800 Ma ago (Van Schmus, 1980).

Table 21.1. Correlation chart of early Proterozoic bedded rocks in the Lake Superior region.

AGE	Subdivisions of Morey (1973;1978)	Gunflint Range, Minnesota and Ontario Goldich et al. (1961)	Mesabi Range White (1954)	Cuyuna Range Morey (1978)	Subdivisions of Cannon and Gair (1970)		Gogebic Range Leith et al. (1935)	Western Marquette Range Cannon and Klasner (1972)	Eastern Marquette Range Gair (1975)	Menominee Range James (1958)
PROTEROZOIC		(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	Paint River Group	Marquette Range Supergroup	(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	Fortune Lakes Slate Stambaugh Formation Hiawatha Graywacke Riverton Iron-formation Dunn Creek Slate
PROTEROZOIC	Animikie Group	Rove Formation ↕	Virginia Formation ↕	Rabbit Lake Formation ↕	Baraga Group	Marquette Range Supergroup	Tyler Slate ↕	Michigamme Formation: Upper slate member Bijiki Iron-fm. Mbr. Lower slate member Clarksburg Volcanics Member Greenwood Iron-fm. Member Goodrich Quartzite Unconformity	Goodrich Quartzite Unconformity	Badwater Greenstone Michigamme Slate Fence River Formation Hemlock Formation Goodrich Quartzite Unconformity
PROTEROZOIC	Animikie Group	Gunflint Iron-formation Kakabeka Quartzite	Biwabik Iron-formation Pokegama Quartzite	Trommald Formation Mahnomen Formation Unconformity	Menominee Group	Marquette Range Supergroup	Ironwood Iron-formation Palms Quartzite Unconformity	Negaunee Iron-formation Siamo Slate Ajibik Quartzite	Negaunee Iron-formation Siamo Slate Ajibik Quartzite Unconformity	Vulcan Iron-formation Felch Formation Unconformity
PROTEROZOIC	Mille Lacs Group	(No equivalent rocks) Unconformity	(No equivalent rocks) Unconformity	Trout Lake Formation Little Falls Formation Glen Township Formation Denham Formation Unconformity	Chocolay Group	Marquette Range Supergroup	Bad River Dolomite Sunday Quartzite Unconformity	(No equivalent rocks) Unconformity	Wewe Slate Kona Dolomite Mesnard Quartzite Enchantment Formation Unconformity	Randville Dolomite Sturgeon Quartzite Fern Creek Formation Unconformity
ARCHEAN		Granite; greenstone	Granite; greenstone	Gneiss; granite; greenstone			Granite; greenstone	Gneiss	Gneiss; Mona Schist	Gneiss

Younger, post-Penokean rhyolite, coeval epizonal granite (Smith, 1978; Van Schmus, 1978), and platform quartzite were deposited south of the erosional edge of the early Proterozoic basinal sequence between 1765 Ma and 1500 Ma. A large anorogenic granitic intrusion and anorthosite, the Wolf River batholith, was emplaced in central Wisconsin about 1500 Ma ago (Van Schmus et al., 1975b).

The early Proterozoic basin is split by the Midcontinent Rift System (1100 Ma), effectively separating it into two distinct segments. The northwest segment contains the sequences of the Gunflint, Mesabi, and Cuyuna iron ranges of Ontario and Minnesota, and the southeast segment contains those of the Gogebic, Marquette, and Menominee ranges of northern Wisconsin and Michigan. Because of this physical separation, the strata in the northwest segment have been assigned to the Animikie and Mille Lacs groups (Morey, 1973, 1978) and the rocks of the southeast segment have been assigned to the Marquette Range Supergroup (Cannon and Gair, 1970). With the possible exception of basal formations north of and in the Marquette district (Puffett, 1969; Table 21.1), these early Proterozoic sequences are considered to be younger than the Huronian Supergroup.

Early Proterozoic Supracrustal Rocks

The early Proterozoic rocks were deposited in an east-trending basin on the Archean craton and form a generally southward-thickening wedge of clastic and chemical sedimentary rocks, including the vast iron formations of the region, and mafic to felsic volcanic rocks. Infilling of the basin tended to be cyclic and was terminated by the Penokean Orogeny.

Stratigraphic successions in each of the iron mining districts (Fig. 21.4) have been correlated throughout much of the basin (James, 1958; Bayley and James, 1973), although physical continuity between districts generally has not been firmly established. Bedded rocks in the southernmost part of the basin are predominantly mafic and felsic volcanics, but because they are poorly exposed it is impossible at this time to integrate them adequately with the better known succession to the north. However, it is probable that they are broadly correlative with the strata to the north.

Marked facies and thickness differences across the basin divide it longitudinally into two zones: (1) a thin succession of predominantly sedimentary rocks in the north, and (2) a much thicker succession of interbedded sedimentary and volcanic rocks in the south. Because of the virtual coincidence of these facies zones with early Proterozoic tectonic zones, it is possible to delineate these two longitudinal zones on a tectonic map (Fig. 21.5). Stratigraphic sections representative of the two facies are given in Figure 21.6. In general, the sequences in the north are 1500-2000 m thick, whereas those in the south locally attain a thickness of about 9000 m (Iron River district, Michigan).

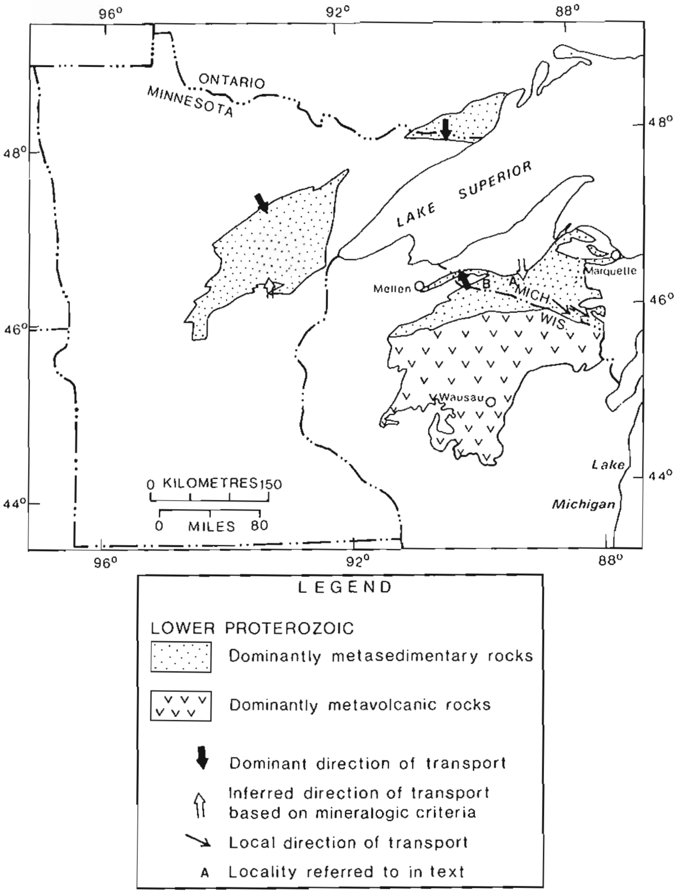


Figure 21.2. *Inferred directions of sediment transport for early Proterozoic clastic rocks of the Lake Superior region.*

Three grossly fining-upward depositional cycles are present in the Marquette Range Supergroup, each of which began with deposition of quartz sands and/or conglomerate. In parts of Michigan, basin filling began with local deposition of conglomerate on the beveled Archean basement surface in diversely oriented fault-controlled troughs (Larue and Sloss, 1980). Deposition continued with accumulation of platform sediments, stromatolitic dolomite, and, locally, slate. These deposits constitute the Chocolay Group (Table 21.1). Because of irregularities on the basement surface and subsequent erosion, thicknesses of the deposits vary considerably over the region, ranging from nil in the Western Marquette Range to about 1400 m in the Menominee district (Table 21.1).

Rocks of this age were apparently not deposited in the Mesabi Range and vicinity in the northern part of the basin, where strata of the younger Animikie Group rest directly on Archean basement. In the Cuyuna Range, submarine tholeiitic basalt was deposited intermittently during accumulation of quartz sandstone and dolomite of the Mille Lacs Group (Table 21.1); deposition was terminated by mild uplift and erosion.

The second cycle began with deposition of quartzose sandstone followed by the major iron formations of the region. These strata constitute the Menominee Group in Michigan and Wisconsin, and the lower part of the Animikie Group in Minnesota. In northern Minnesota and Ontario, and in the Gogebic Range, the iron formations record nearly contemporaneous transgressive sedimentation over the Archean craton (Morey, 1973). The iron formations in this part of the basin are about 200 m thick and possibly were connected prior to Keweenawan rifting (Sims, 1976). In the southern part of the basin, however, iron formations of the Menominee Group differ greatly in thickness, stratigraphic detail, and depositional facies, suggesting that they were deposited in isolated second-order troughs within the larger basin (see, e.g. Gair, 1975). Deposition of the second cycle ended during minor deformation and uplift, now recorded by a local unconformity.

At the beginning of the third cycle, deposition of the Goodrich Quartzite and equivalent units at the base of the Baraga Group (Table 21.1) ushered in a period of pronounced crustal disturbance. This period was characterized by mafic to felsic volcanism and by differential subsidence that culminated with deposition of 3000+ m of greywacke turbidites of the Baraga Group in the southern part of the basin (James et al., 1961). Virtually contemporaneous deposition of greywacke and shale in the northern part of the basin (Animikie Group) indicates some foundering of the shelf here as well. These deposits accumulated directly on iron formation in gradually deepening water (Morey, 1973). The volcanics of the Baraga Group in the southern part of the basin are locally 3000+ m thick, and lenticular. Gabbroic dykes and sills in northern Michigan were emplaced at this time, as subvolcanic equivalents of the basaltic lavas (Cannon, 1973).

The deep water environment initiated during deposition of the Baraga Group continued with additional accumulation of about 2000 m of greywacke and shale and local lenses of iron formation (Paint River Group) in the Iron River-Crystal Falls district of the Menominee Range (Fig. 21.4). These units are of limited geographic extent and apparently were deposited mainly in an oval-shaped depression about 50 km in maximum dimension. Sedimentation was terminated by the major deformational pulse of the Penokean Orogeny.

The depositional patterns of the early Proterozoic sequences reflect increasing crustal instability and southward-prograding differential downwarping. The lateral and vertical facies variations record a complete transition from a stable craton to an unstable, deep water environment in early Proterozoic time (Bayley and James, 1973).

Until the latter part of Baraga deposition, detritus was supplied principally from exposed Archean rocks to the north (Fig. 21.2), as indicated by facies and paleocurrent patterns and sedimentological characteristics of units such as the Pokegama Quartzite, the younger Virginia Formation, and their stratigraphic equivalents in northern Minnesota and Ontario (Morey, 1973). A northerly Archean greenstone-granite provenance is also indicated by low initial $^{87}Sr/^{86}Sr$ ratios (0.705) of rocks of the Virginia, Rove, and Michigamme formations (Peterman, 1966; written communication, 1978).

However, equivalent formations in more southerly parts of the basin were derived in part from local sources. In the Cuyuna Range, there is evidence for derivation of sediments from early Archean gneiss (Morey, 1978) and from a mixed-age terrane of rocks 2500-3500 Ma old (Peterman, 1966). Near Marenisco, east of the Gogebic Range, arkose and conglomerate at the base of the Marquette Range Supergroup (Baraga Group) contain clasts identical to the underlying 2700 Ma granite. Taken together, these features indicate a shelf prograding into a deep water environment, with increasing amounts of clastic debris derived from intrabasinal positive areas and possibly from its southern margin. Some of these positive areas coincide with the present sites of Archean gneiss domes.

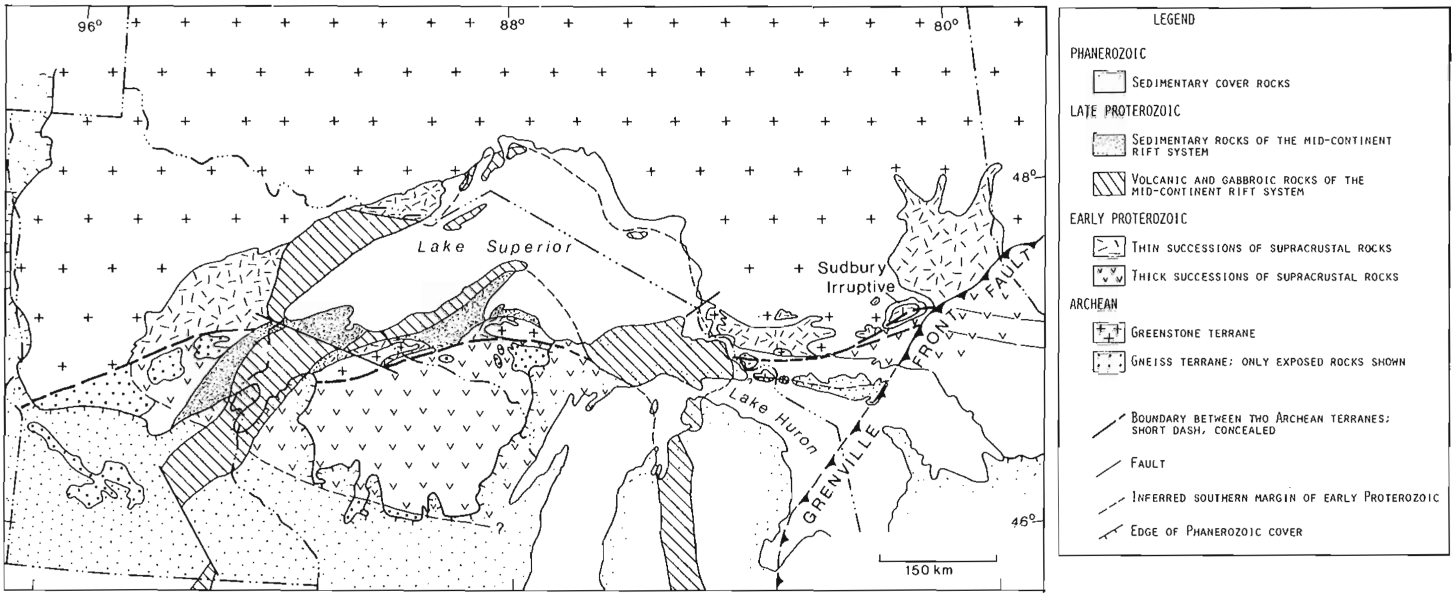


Figure 21.3. *Map of the Great Lakes region showing location of boundary between Archean greenstone-granite and gneiss terranes.*

Deformation, Metamorphism, and Igneous Activity

The early Proterozoic supracrustal rocks of the Lake Superior region can be divided into two broad longitudinal zones on the basis of contrasting styles of deformation and grades of metamorphism: a northern stable cratonic zone and a southern foldbelt (Fig. 21.5). The tectonic front separating the two zones is marked approximately by the northern limit of penetrative cleavage (Fig. 21.8A).

North of the tectonic front, in the stable cratonic zone, the Proterozoic supracrustal rocks unconformably overlie late Archean greenstone-granite basement and remained virtually undeformed and unmetamorphosed during the Penokean Orogeny (Morey, 1973; Floran and Papike, 1975; Hanson and Malhotra, 1971).

South of the tectonic front, early Proterozoic supracrustal rocks within the foldbelt display two contrasting tectonic styles. A broad subzone 60 to 70 km wide immediately south of and subparallel to the tectonic front (Fig. 21.5) is intensely deformed and characterized by a nodal distribution of metamorphic zones (James, 1955). South of this subzone is a belt of comparable intense tectonism, but with linear rather than nodal patterns of deformation and metamorphism and containing mesozonal granitic plutons of early Proterozoic (~1850 Ma) age (Sims and Peterman, 1980; Van Schmus, 1980). All supracrustal rocks within the foldbelt are presumed to lie unconformably on early Archean gneiss, as gneisses are exposed in the cores of several gneiss domes or upraised fault blocks and crop out to the south of the erosional edge of the supracrustal rocks (Fig. 21.5).

The northern part of the foldbelt characterized by nodal metamorphism has become the "type area" for the Penokean Orogeny (Cannon, 1973). Here, early regional folds trending subparallel to the tectonic front were modified by superposed folds related to diapiric gneiss domes and basement uplift. Nodal metamorphism of the low-pressure type was contemporaneous with recrystallization and anatexis of the Archean gneiss in the cover of the uplifts (Sims and Peterman, 1976) and was superimposed on a regional low grade metamorphism (James, 1965) developed during the early folding (Fig. 21.8A).

The southern zone of the foldbelt is beneath extensive Pleistocene cover and details of its deformation and metamorphism are not known. In general, fold trends are linear but discontinuous and diversely oriented (Sims et al., 1978). The deformation pattern is tentatively interpreted as indicating separate regimes within different crustal blocks, suggesting strong control by Archean basement structure.

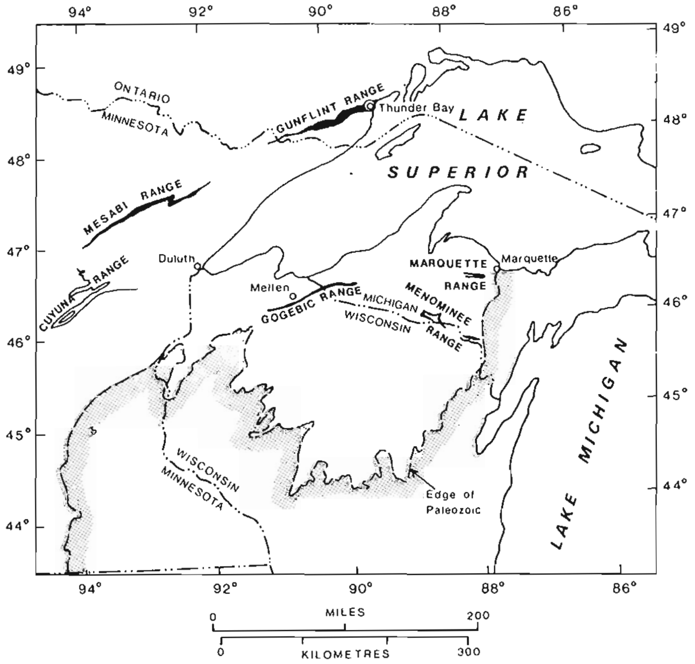


Figure 21.4. *Proterozoic iron ranges of the Lake Superior region.*

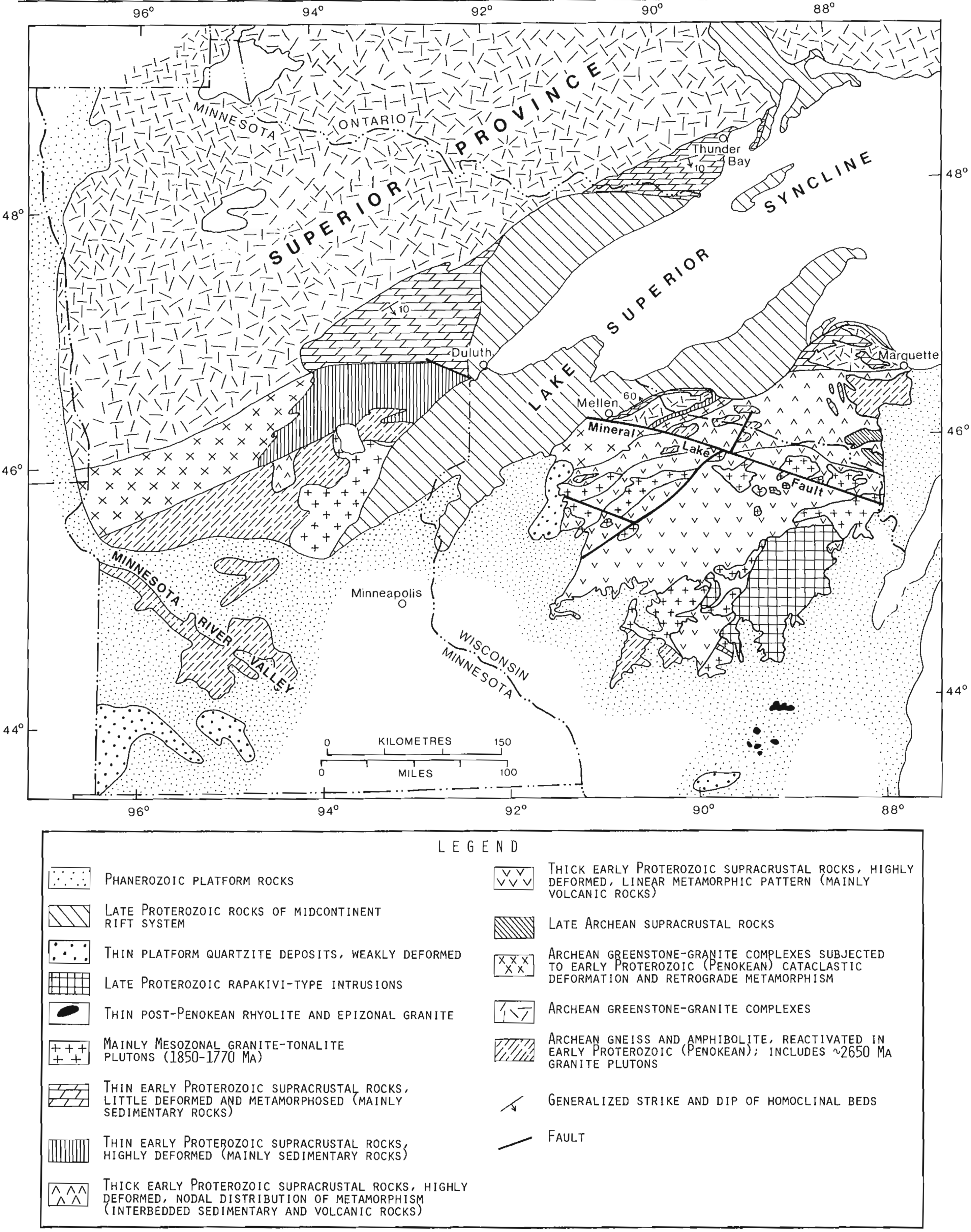


Figure 21.5. *Tectonic units of the Lake Superior region. Modified from King and Beikman (1974).*

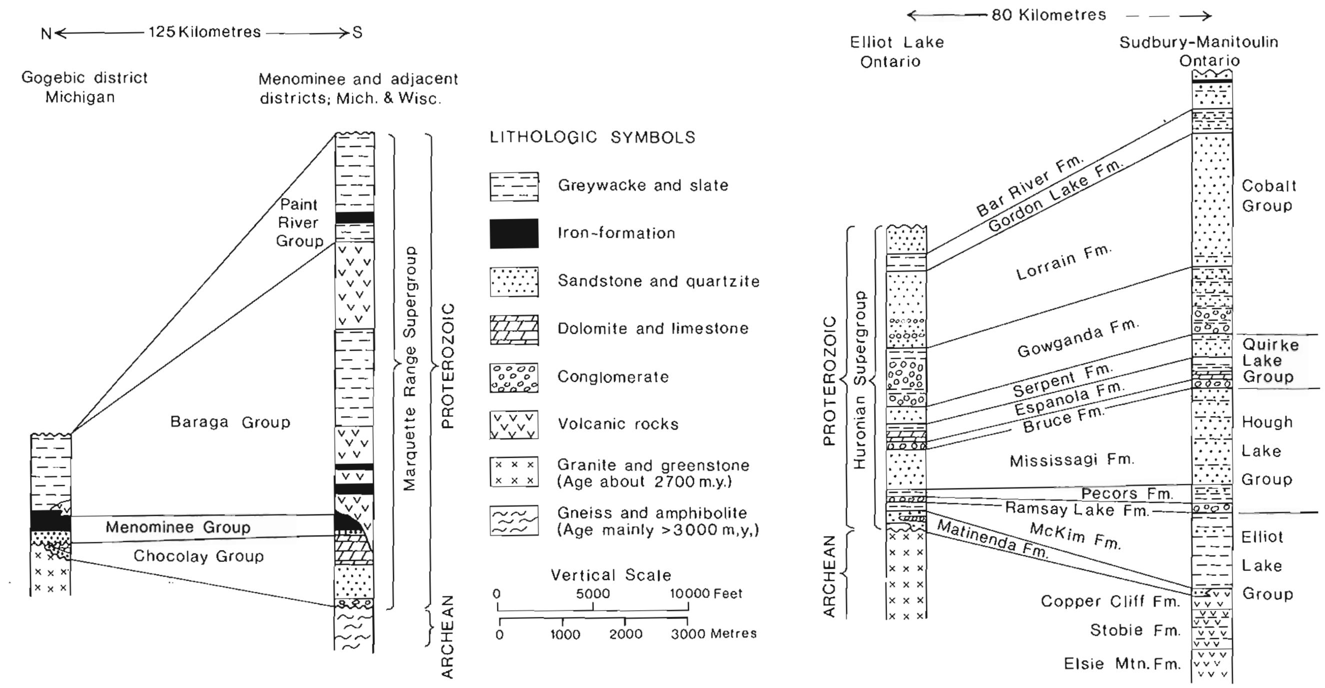


Figure 21.6. *Selected stratigraphic sections of early Proterozoic rocks of the Lake Superior and Lake Huron regions.*

Depositional and Structural Zonation

Patterns of early Proterozoic deposition, deformation, metamorphism, and plutonic-volcanic activity in the Lake Superior region are notably asymmetric. The close spatial mimicry of these patterns indicates that they are part of a tectonic continuum that began with the development of the structural basin and culminated with the major tectonothermal pulse of the Penokean Orogeny.

The epicratonic basin developed initially through foundering of the Archean craton, initiated by rifting, along an easterly axis that coincided with the boundary between the two Archean crustal segments. As the basin evolved, adjustments along the trace of this inherited zone of weakness propagated vertically, leading to the development of the postdepositional tectonic front.

During the early stages of infilling, shallow water shelf facies were deposited across the axis of the basis upon both Archean basement segments. Contemporaneous faulting produced local troughs within the platform that were diversely oriented and largely controlled by older structures in the basement rocks (Fig. 21.4). Submarine basalt flows were extruded locally on the gneiss basement segment during accumulation of clastic and chemical sediments. Platformal conditions continued, at least around the margins of the basin, during succeeding iron formation deposition, but increasing foundering in the southern part developed into major downwarping, concentrated along the pre-existing fault-bounded troughs. Deposition of iron formation in the southern part of the basin ceased, probably at somewhat different times in different troughs, because of slight deformation. In contrast, deposition of iron formation in the northern part of the basin was halted by prograding deposition of greywacke and shale (Morey, 1973), apparently as a result of uplift of the Archean greenstone-granite terrane to the north and minor foundering of the shelf. Foundering intensified in the southern part of the basin after deposition of the major iron formations, with local rapid subsidence and accumulation of thick, deep water greywacke turbidites and basaltic to felsic volcanics. At this time, a southward-thickening sheet of greywacke and shale derived from both northerly and internal sources coalesced across the basin axis, with detritus from both sources crossing it. Extrusion of volcanics was confined, however, to areas adjacent to, and south of, the tectonic front. The wide regional disparity in thicknesses of the volcanic rocks indicates substantial structural relief in the basin and, apparently also, local sources for the volcanics. Existing deep depressions continued to founder in the closing stages of accumulation, especially in a part of the Menominee Range, with deposition of clastic sediments, iron formation, and black pyritic, graphitic shale. Sedimentation then ceased, before or during the main pulse of deformation accompanying the Penokean Orogeny.

Deformation in the foldbelt involved compression accompanied by substantial shortening of the supracrustal sequence (Cannon, 1973) and some transport toward the stable craton. Adjacent to the tectonic front, the main compressive phase was followed by pronounced vertical tectonism with the development of diapiric gneiss domes caused by partial remobilization of the basement rocks.

LAKE HURON REGION

Early Proterozoic rocks of the Lake Huron region, the Huronian Supergroup, consist mainly of clastic sedimentary rocks, with subordinate chemical sediments and local basal accumulations of volcanic rocks. They were deposited unconformably on Archean basement rocks 2500 – 2100 Ma ago, as indicated by the approximate radiometric ages of the Archean granitoid basement rocks and the intrusive Nipissing Diabase, respectively (Card et al., 1972; Van Schmus, 1965). Strata grossly correlative with the Huronian Supergroup extend into the Grenville Province, and are now represented by several types of high grade gneisses (Lumbers, 1975) that have been subdivided at reconnaissance mapping scale.

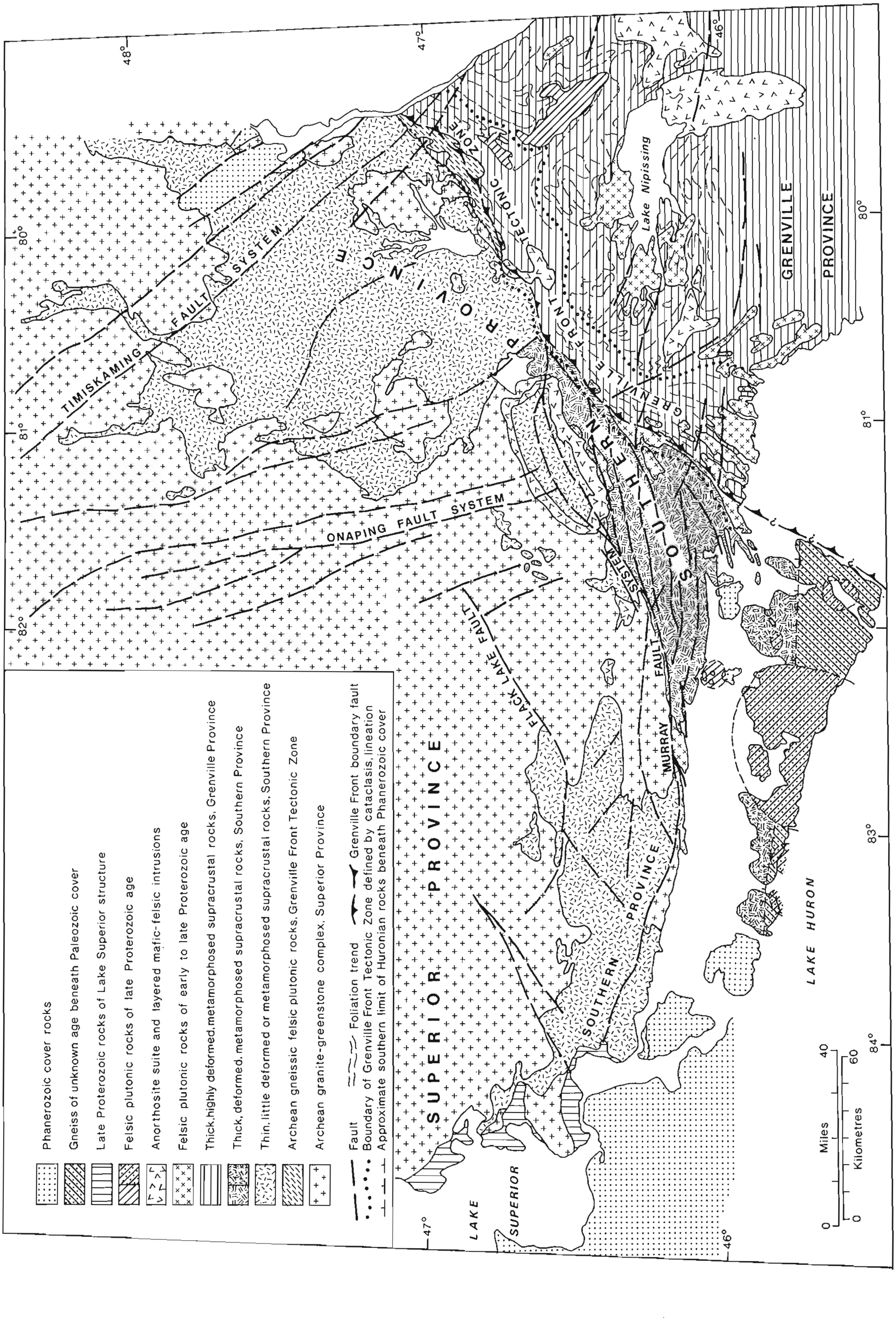


Figure 21.7. Geologic-tectonic map of the Lake Huron region.

The Huronian supracrustal rocks were deformed, metamorphosed, and intruded by mafic and felsic igneous rocks during a protracted series of events extending from 2200 Ma to 1700 Ma. The Sudbury structure (Card and Hutchinson, 1972), including the Sudbury Nickel Irruptive (Souch et al., 1969), was formed during this interval. Later, anorogenic granitic plutons (about 1500 Ma old) and mafic dyke swarms (about 1250 Ma old) were emplaced in the Southern Province.

Huronian Supergroup

The early Proterozoic Huronian Supergroup consists of a southeastward-thickening wedge of sedimentary and volcanic rocks. Southeastward thickening and accompanying facies changes are neither uniform nor simple; instead, they occur abruptly across zones now marked by the Murray fault system (Fig. 21.7). Accordingly, it is convenient to discuss these rocks with respect to two areas: the Sault Ste. Marie-Elliot Lake area in the northwest, and the Sudbury-Cutler area in the southeast. Representative sections of the Huronian Supergroup for the two areas are given in Figure 21.6. A generalized diagram showing facies variations across the Murray fault system and the Grenville Front tectonic zone is shown in Figure 21.9.

The Elliot Lake Group, the basal unit of the Huronian Supergroup, is a heterogeneous assemblage of arkose, conglomerate, siltstone, greywacke, and mafic to felsic volcanic rocks. Lateral facies changes are particularly abrupt near volcanic epicentres. In the northwest, the Elliot Lake Group is about 600 m thick and consists mainly of coarse feldspathic sandstone with lenses of uraniferous quartz-pebble conglomerate (Roscoe, 1969; Frarey, 1977). Thin successions of volcanic rocks, mainly tholeiitic basalt and mildly alkalic hawaiite, mugearite, and ankaramite occur at the base of the group in the Sault Ste. Marie-Elliot Lake area (Bennett and Innes, 1979). In the southeast, near Sudbury, the group consists of about 1500 m of tholeiitic basaltic lavas (Card et al., 1977) and intercalated sedimentary rocks, 2000-3000 m of rhyolite, and 2000-3000 m of greywacke turbidites. Associated with the volcanic accumulations in the Sudbury-Cutler area are several layered, lopolithic gabbro-anorthosite intrusions. The McKim Formation displays a distinct facies variation from a thick, thick-bedded proximal facies greywacke turbidite sequence in the southeast to a thin, thin-bedded distal facies siltstone-argillite sequence in the northwest. Possibly the McKim represents a flyschoid facies derived from a positive element to the south or east.

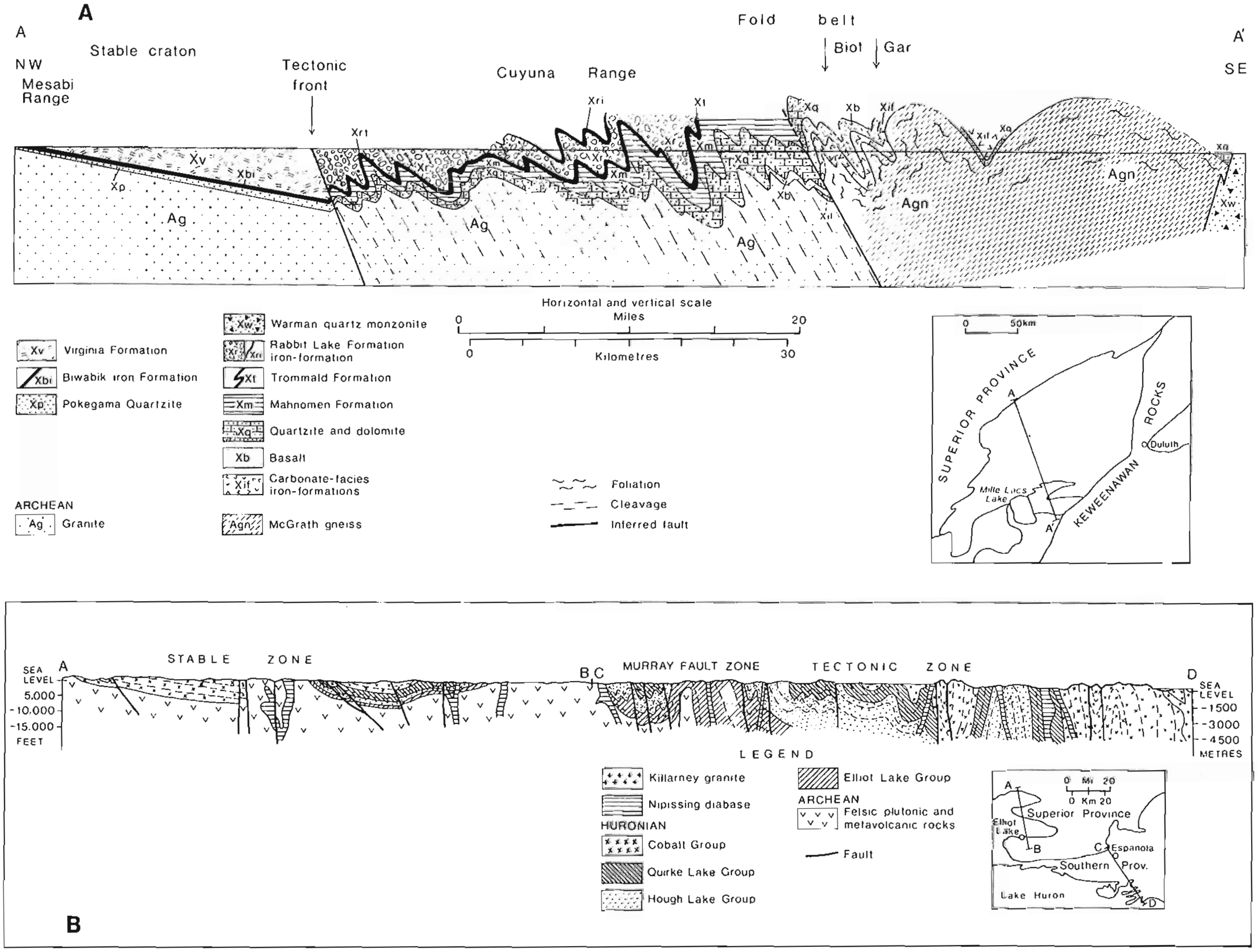


Figure 21.8. Generalized geologic sections across the early Proterozoic basins of east-central Minnesota (A) and Ontario (B).

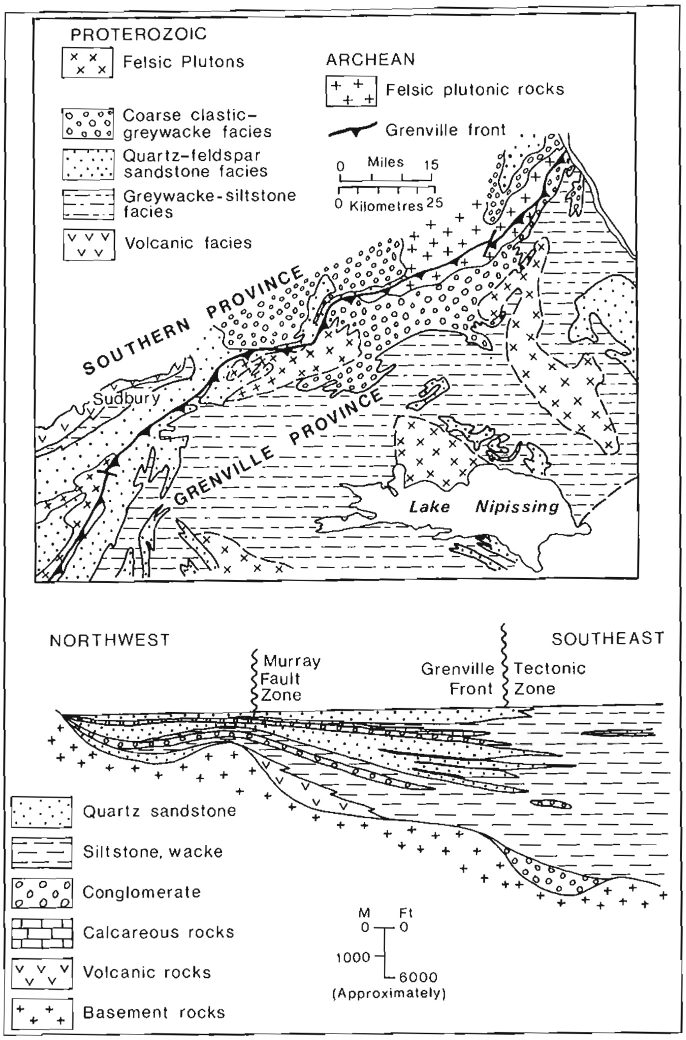


Figure 21.9. *Diagrammatic cross-section and map showing stratigraphic facies variations of early Proterozoic supracrustal sequences across the Murray fault and Grenville Front zones, near Sudbury, Ontario.*

The overlying Hough Lake and Quirke Lake groups (Fig. 21.6) constitute two major sedimentary cycles. Each consists of a lower conglomeratic unit, a middle shaly unit, and an upper sandstone. Typically, the conglomerates are thin, massive, or coarsely interstratified sheets of pebbly mudstone and wacke. Shaly units consist of thin-bedded siltstone and thick-bedded greywacke turbidites. Sandstones are crossbedded, characteristically quartz- and feldspar-rich and form thick wedges. Both Hough Lake and Quirke Lake, and their contained formations, thicken markedly southward. The combined thickness of the two groups is about 1500 m in the northwest and about 6000 m in the southeast. Southeastward thickening of individual sandstone formations is accompanied by a relative increase in shaly material. For example, the Mississagi Formation is approximately 600 m thick in the northwest and consists of coarse grained, cross-bedded feldspathic sandstone. In the southeast it is as much as 3000 m thick and consists of regular alternations of sandstone and siltstone with minor carbonate-rich units.

The Cobalt Group, which comprises the upper part of the Huronian Supergroup, consists of a heterogeneous assemblage of conglomerate, siltstone, greywacke, and silty arkose (Gowganda Formation); a thick succession of arkose and orthoquartzite (Lorrain Formation); a thin unit of vari-coloured siltstone and fine sandstone, locally containing gypsum and anhydrite (Gordon Lake Formation); and an upper unit of orthoquartzite and hematitic siltstone (Bar River Formation). The preserved Cobalt Group has a maximum thickness of about 3500 m in the northwest and about 5000 m in the Sudbury-Cutler area. The contact between the Cobalt Group and underlying strata is commonly a disconformity. Locally, in the more northerly parts of the basin, it is an erosional unconformity, indicating local tectonic activity between deposition of the two parts of the Huronian Supergroup.

Abrupt changes in stratigraphic thickness and depositional facies occur near the Grenville Front tectonic zone (Fig. 21.9). Southeast of Sudbury, the lower half of the Huronian Supergroup thickens toward the Grenville Province. Most of this thickening occurs in the McKim and Mississagi formations. In contrast, other formations, for example the Espanola and Serpent, thin markedly near the Grenville Front (Card et al., 1977). Also, as shown diagrammatically in Figure 21.9, greywacke turbidites thicken markedly across the front at the expense of quartz sandstone. Thus, the Grenville Front tectonic zone coincides with an abrupt transition from mainly shallow water to deep water depositional environments.

Near the Grenville Front northeast of Sudbury, the lower part of the Huronian Supergroup pinches out, suggesting that prior to deposition of the Cobalt Group a positive area existed in this region along the northwestern margin of the Grenville Province. This positive element may have formed concurrently with the emplacement of mafic dyke swarms in the Archean rocks of this area.

Summary of Depositional Environments

The rocks in the lower part of the Huronian Supergroup are interpreted as having been deposited during a series of major regressive marine cycles controlled by syndepositional faulting and differential movements of basement blocks. Each cycle began with normal faulting, subsidence, and a sudden increase in paleoslope and water depth that triggered submarine debris flows, resulting in the deposition of the conglomerate units (Ramsay Lake and Bruce formations). These events were followed by the deposition of wacke and siltstone under relatively deep water conditions, in part by turbidity currents (McKim and Pecors formations). As basin infilling brought about a return to shallow water conditions, crossbedded sandstones were deposited (Mississagi and Serpent formations).

The upper part of the Huronian Supergroup, the Cobalt Group, can be characterized as a clastic wedge consisting of a lower assemblage of coarse clastic debris grading upward into shallow-water sandstone. There is evidence that the rocks of the Gowganda Formation were deposited by glacial and glacial-related processes (Lindsey, 1971).

Petrographic and paleocurrent data for the Huronian Supergroup, mainly from sandstone units, indicate that the sediments were derived primarily from Archean basement to the north and were deposited by generally south-flowing currents (Fig. 21.10). Most formations show a distinctly polymodal pattern, interpreted as an interaction between southward-flowing contributory currents and east-west redistributing currents (Palonen, 1973). Exceptions to this

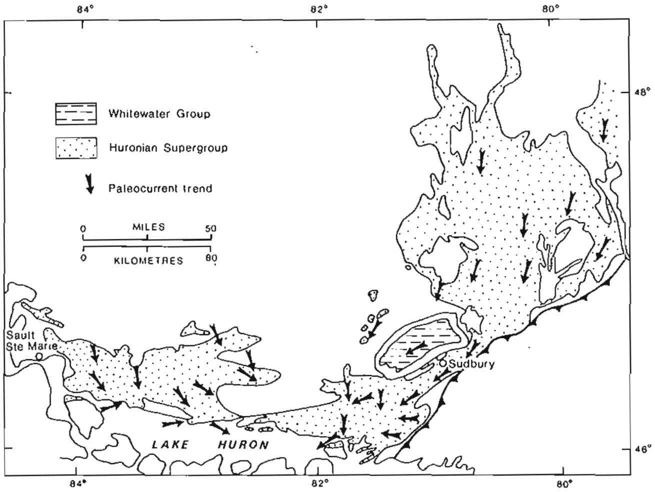


Figure 21.10. Generalized map of the Lake Huron region showing inferred directions of sediment transport for early Proterozoic clastic rocks.

general pattern have been noted near the Grenville Front, where crossbedding in the Mississagi and Serpent formations records westward-flowing paleocurrents (Long, 1976).

Deformation, Metamorphism, and Igneous Activity

Early Proterozoic rocks in the Lake Huron region show systematic differences in intensity and style of deformation, regional metamorphism, and igneous activity that conform generally to the asymmetric depositional pattern. Changes are most abrupt and notable across the Murray fault system, and the contrasting patterns are best discussed with respect to the two areas mentioned earlier: Sault Ste. Marie-Elliot Lake and Sudbury-Cutler.

The Murray fault system marks an abrupt transition from relatively thin, little deformed, little metamorphosed supracrustal rocks in the northwest to relatively thick, highly deformed, highly metamorphosed rocks in the southeast. Murray faults, as shown in Figures 21.8 and 21.10, dip steeply southward and form an anastomosing east-trending system which records several episodes of movement. Early movement, indicated by abrupt thickening of formations of the Huronian Supergroup across the zone, was dominantly dip-slip, with the south side down. Later, after deposition of the Huronian Supergroup and intrusion of the Nipissing Diabase and Sudbury Irruptive, reverse dip-slip and right-lateral strike-slip movements resulted in separations up to 3000-5000 m on some structures. The repeated movements, great displacements, and continuity of the fault system indicate that a major crustal structure is represented.

Early Proterozoic tectonism in the Lake Huron region was protracted and occurred either in pulses or intermittently from about 2200 Ma to 1700 Ma. It was renewed still later during major deformation of the Grenville Province. Folding of the supracrustal rocks began prior to emplacement of the Nipissing Diabase (2160 Ma; Van Schmus, 1965; Fairbairn et al., 1969). Folding was accompanied or succeeded by the emplacement of small granitic plutons, the Creighton and Murray Granites (about 2200 Ma; Gibbins and McNutt, 1975). Following emplacement of the Nipissing Diabase, further deformation, regional metamorphism and igneous intrusive activity occurred during the 1950-1700 Ma interval. Deformation both preceded and followed emplacement of the Sudbury Irruptive (1840 Ma; Krogh and Davis, 1974). The Killarney plutons (1700 Ma and 1500 Ma; Krogh et al., 1971), a group of composite granitic intrusions along the Grenville Front (Fig. 21.7), were emplaced after the main early Proterozoic deformation and metamorphism within the Southern Province but prior to the late Proterozoic deformation and metamorphism in the Grenville Province (Lumbers, 1975; Card, 1978a). The Cutler pluton (1750 ma; Wetherill et al., 1960) cuts previously metamorphosed and deformed Huronian rocks but was affected by still later deformation (Cannon, 1970).

Huronian strata in the Sault Ste. Marie-Elliot Lake region are flat-lying or gently folded (Fig. 21.8). Folds are open, upright concentric structures with gentle but variably-plunging hinges. Regionally, fold trends are variable and minor tectonic structures are weakly developed. Metamorphic grade is low and has been termed subgreenschist facies (Card, 1978b).

In the Sudbury-Cutler area to the southeast, supracrustal rocks are moderately to intensely deformed into open to subisoclinal, flattened "buckle folds" with upright to slightly overturned axial surfaces. Fold hinges plunge moderately and reversals in plunge are common, yielding elongate basins and domes. The folds trend east to northeast, subparallel to the Murray fault system. Penetrative axial plane cleavage is generally well developed and is accompanied by a steeply plunging rodding or mineral lineation. Other planar fabrics, including early bedding-plane foliation and late strain-slip cleavage, are present.

Near the Grenville Province, late Proterozoic cataclasis, mylonitization, and accompanying recrystallization, related to the Grenville Front tectonic zone, are superimposed upon the major folds described above. Near the Grenville Front boundary fault the supracrustal rocks have been refolded about southeast-dipping axial planes, which subparallel the boundary fault (Lumbers, 1978).

Metamorphism of supracrustal rocks in the southeast is moderately intense and ranges from low greenschist to low amphibolite facies (Card, 1964, 1978b). Higher grade metamorphic rocks contain staurolite, andalusite, and kyanite-bearing assemblages and occur in two northeast-trending zones (Fig. 21.11), one immediately south of and parallel to the Murray fault system and the other northwest of the Grenville Front tectonic zone. Within each zone, the highest grade rocks have a nodal distribution pattern. The northern zone of high grade metamorphism coincides with an anticlinorium; the southern zone with a series of major anticlines. Intervening lower grade rocks coincide with major synclinoria in the supracrustal rocks. In detail, however, metamorphic isograds transect fold axes. Higher grade metamorphism either was superimposed on previously folded rocks or continued after the early major folding.

The main regional metamorphism culminated after intrusion of the Nipissing Diabase (2150 Ma) and prior to emplacement of the Sudbury Irruptive (1840 Ma). Radiometric age dating of Huronian metasedimentary rocks indicates that the metamorphic culmination occurred at about 1900 Ma (Fairbairn et al., 1969), at approximately the same time as the Penokean metamorphic culmination in the Lake Superior region (Van Schmus, 1976).

Depositional and Structural Zonation

Asymmetric patterns of deposition, deformation, metamorphism, and felsic plutonic activity in the Lake Huron region reflect persistent tectonic environments throughout the early Proterozoic. In the northwest, a stable cratonic environment existed throughout deposition of the supracrustal rocks. In the southeast, however, relatively unstable conditions persisted from the beginning of deposition through subsequent tectonic events. The demarcation between these two contrasting tectonic zones, the Murray fault system, was a hinge line during deposition and a tectonic front during deformation.

During early stages of basin filling, the Murray fault system was tectonically active. Several thousand metres of mafic and felsic volcanic rocks were deposited along and immediately south of the fault zone. Volcanism was accompanied by emplacement of layered gabbro-anorthosite complexes, which intruded the Archean granitic rocks underlying the northern part of the basin (Card, 1978a). Succeeding sheets of clastic sediments were deposited across the fault zone, each thickening substantially into the southern part of the basin. This pattern of intermittent tectonic activity and wedge-shaped depositional units persisted through the remainder of lower Huronian deposition. During later stages of deposition (Cobalt Group), tectonism across the hinge line abated, and sediments spread across it without conspicuous changes in thickness.

Throughout deposition, detritus continued to be supplied from the Archean rocks to the north, but at times clastic material from a southerly or easterly source was supplied to the basin.

Asymmetric deformation and metamorphic patterns in the Lake Huron region indicate that during the early Proterozoic the basement south of the Murray fault system was more mobile than the basement to the north. By analogy with the Lake Superior region, we suggest that this southern mobile basement is composed in part of early Archean gneiss. The location of the boundary between the gneiss and the late Archean greenstone-granite terrane, exposed north of the Huronian sequence, is somewhat uncertain. We propose that the basement boundary in the Cutler area (Fig. 21.10) coincides with the Murray fault because of the abrupt change in tectonic style on either side of the fault. East of Espanola, the location is less certain but the boundary may coincide with the faults that traverse the Sudbury structure or with another fault trace nearer the north shore of Lake Huron (Fig. 21.7).

GRENVILLE PROVINCE

The thickest accumulation of early Proterozoic rocks in Ontario occurs within the northern part of the Grenville Province, east of the Sudbury Nickel Irruptive (Fig. 21.12). At least some of these rocks were deposited contemporaneously with those in the lower part of the Huronian Supergroup (Lumbers, 1978). A younger sequence of supracrustal rocks, late Proterozoic in age, is present in the southern part of the Grenville Province of Ontario.

Metamorphism and plutonic activity were more prolonged and intense in the Grenville Province than in the Proterozoic basins of the Southern Province, and accordingly it represents a profoundly different orogenic regime than that in the Southern Province. Tectonism culminated between 1300 Ma and 1000 Ma, when nearly all the supracrustal and plutonic rocks were converted into highly recrystallized and deformed gneiss (Grenvillian Orogeny, Stockwell et al., 1970).

In Ontario and for an unknown distance northeastward into Quebec, the northwest boundary of the Grenville Province is defined by the Grenville Front boundary fault,

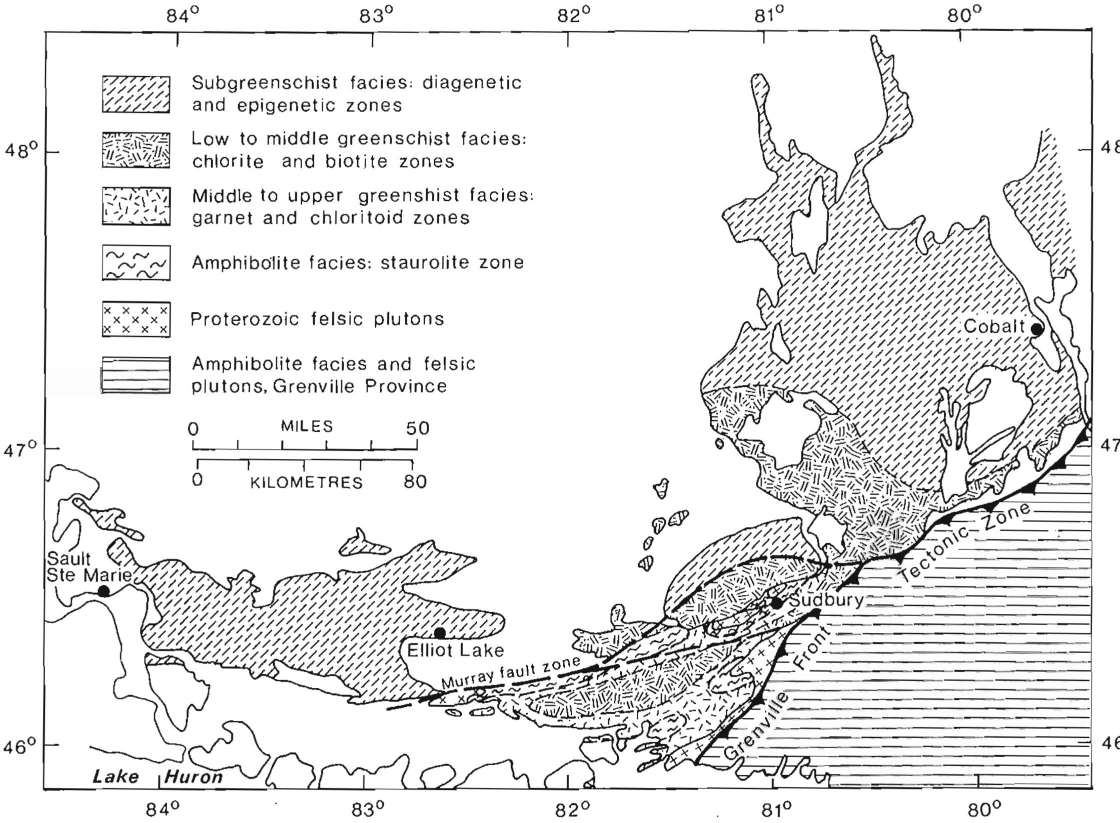
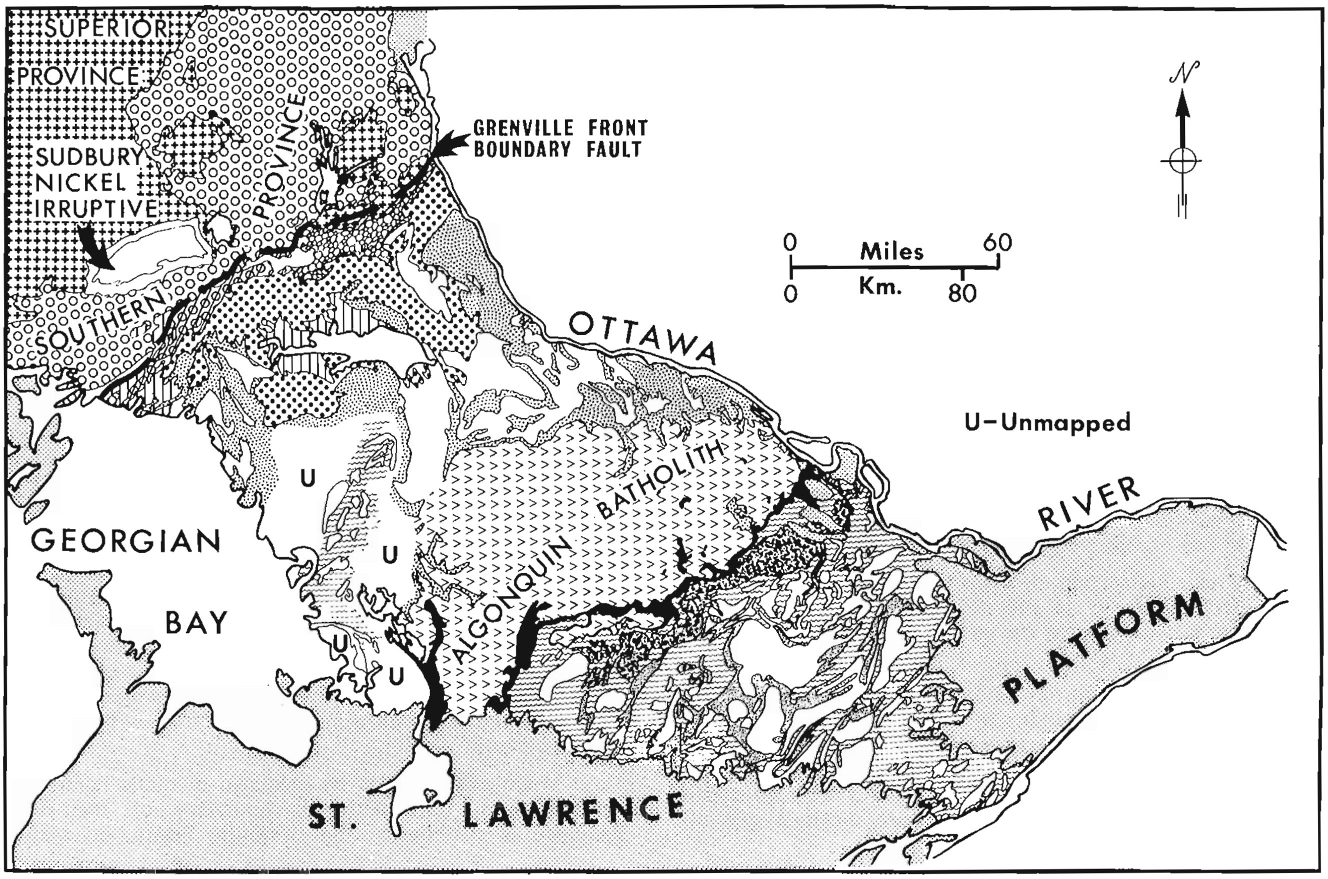


Figure 21.11. *Metamorphic zones in the Lake Huron region.*



LEGEND

PHANEROZOIC

- Unsubdivided Paleozoic sedimentary rocks

PRECAMBRIAN

GRENVILLE PROVINCE

Late Proterozoic

- Alkalic plutonic rocks
- Unsubdivided plutonic rocks
- Metavolcanics
- Carbonate and other metasediments
- Coarse clastic sequence

UNCONFORMITY

- Algonquin batholith (unsubdivided)

Early Proterozoic

- Unsubdivided plutonic rocks
- Shallow to deep water metasandstone and other metasediments
- Deep water turbidite deposits (mainly greywacke-siltstone)
- Gneissic metasediments derived from the Huronian Supergroup
- Coarse clastic sequence

UNCONFORMITY

Archean

- Gneissic Archean plutonic and supracrustal rocks

GRENVILLE AND SUPERIOR PROVINCES

- Unsubdivided early Proterozoic rocks of the Southern Province
- Unsubdivided Archean rocks of the Superior Province

Figure 21.12. *Generalized geologic map of the Grenville Province in Ontario.*

which separates high grade gneisses of the Grenville Province from relatively low grade rocks of the Superior and Southern Provinces (Fig. 21.12). This fault is near the northwest margin of the Grenville Front tectonic zone, a structural zone as much as 30 km wide that is superposed on older structures of the adjacent Southern, Superior, and Grenville provinces. The zone is characterized by a northeast-trending cataclastic foliation and a prominent southeast-plunging rodding. Its boundaries are gradational, and visible cataclasis extends as much as 6 km into the adjacent Southern and Superior provinces. In these two provinces, all rocks near the fault are intensely deformed, metamorphosed, and generally slightly overturned toward the northwest by relatively tight northeast-trending folds. Although the Grenville Front tectonic zone reflects a long history, probably beginning in the late Archean, the last major deformation occurred in the interval 1200 Ma to 900 Ma ago (Lumbers, 1975, 1978).

Early Proterozoic Supracrustal Rocks

Supracrustal sequences in the northern part of the Grenville Province in Ontario were deposited between 2500 Ma and 1800 Ma (Lumbers, 1975, 1978), and consist mainly of clastic siliceous metasedimentary rocks. Near the northwest margin of the province these rocks lie unconformably on Archean granitic rocks that extend into the Grenville Province from the Superior Province. The base of the accumulation is a coarse clastic succession (Fig. 21.9). Southeastward, this coarse facies grades into thinly bedded, medium grained greywacke and argillite of deep water affinity preserved in a linear depression or trough along the northwestern margin of the Grenville Province. Recent mapping (Lumbers, 1975) has shown that these early Proterozoic metasedimentary rocks in the Grenville Province are probably correlative with at least part of the Huronian Supergroup. South of the Sudbury Nickel Irruptive, sandstone of the Mississagi Formation grades laterally southeastward into a succession dominated by greywacke and argillite which appears to be contiguous with the greywacke succession in the Grenville Province. The facies change occurs within the Grenville Front tectonic zone across the Grenville Front boundary fault.

South of the metasedimentary rocks in the trough, thick successions of impure sandstone contain several units of moderately to well-sorted arkose, subarkose, orthoquartzite, aluminous clay-rich material, and iron formation. These strata are intruded by granitic rocks as old as 1800 Ma, indicating that they also are early Proterozoic in age. The age of these metasediments relative to those in the trough is not definitely known. If they are a part of the trough sequence, they mark a significant change in the paleogeography of the developing early Proterozoic basin. Alternatively, they could represent a somewhat younger early Proterozoic sequence deposited over the deep water facies in the trough.

Deformation, Metamorphism, and Igneous Activity

The tectonic-metamorphic history of rocks in the Grenville Province is not known in detail because the late Proterozoic high grade regional metamorphism has masked evidence of earlier events. Nevertheless, the existence of early, gneissic granite pegmatite dykes in metasedimentary rocks in the trough and in the Grenville Front tectonic zone suggests that the early Proterozoic supracrustal rocks near the northwest margin of the province were metamorphosed contemporaneously with the Huronian rocks of the Southern Province, between 2160 Ma and 1800 Ma ago (Lumbers, 1978). The pegmatite dykes are older than the granite plutons that were emplaced between 1800 Ma and 1500 Ma in the early Proterozoic supracrustal rocks. Early Proterozoic metamorphism varied in intensity but probably rarely exceeded the temperature and pressure conditions of the lower almandine-amphibolite facies. Late Proterozoic regional metamorphism produced mineral assemblages in this part of the province indicative of middle to upper almandine amphibolite facies. During late Proterozoic metamorphism, the Algonquin batholith, which cuts early Proterozoic rocks and was emplaced between 1460 Ma and 1400 Ma (Krogh and Davis, 1969a, b), was reactivated and diapirically intruded into the overlying rocks, as were several other granitoid bodies (Schwerdtner and Lumbers, in press). Diapirism caused most of the tectonic deformation within the batholithic rocks and produced recumbent folds in the adjacent supracrustal rocks.

The youngest tectonism recognized in the Grenville Province was movement along the Ottawa-Bonnechere graben (Fig. 21.3), which crosses the northern part of the Grenville Province in Ontario and possibly joins the Murray fault system south of Sudbury. The graben probably was initiated in Archean time, and the part of it east of Lake Nipissing has remained active to the present time.

Early Proterozoic intrusive rocks include gneissic mafic dykes and sills, possibly correlative with the Nipissing Diabase in the area near the Grenville Front boundary fault south and east of Sudbury, and abundant 1800 to 1600 Ma old granitic plutons. Most of the remaining intrusions belong to the anorthosite suite and are composed mainly of quartz monzonite and syenite with local anorthositic and tonalitic rocks. The largest of these, the Algonquin batholith, is probably part of a major suite of 1500 to 1400 Ma old anorogenic intrusions that extends from Labrador to California (Silver et al., 1977).

Discussion of Stratigraphic and Tectonic Features

The evolution of the Grenville Front tectonic zone began in late Archean or early Proterozoic time, when the trough now containing the early Proterozoic supracrustal rocks formed. A substantial thickening of the lower part of the early Proterozoic sequence and a change to deep water facies coincide with the position of the trough, indicating that it was a significant paleogeographic feature in the developing Proterozoic basin. The submarine, predominantly deep water clastic rocks in the trough were derived from the Archean terrane to the north and were deposited approximately synchronously with the lower part of the Huronian Supergroup to the west.

Subsequent to its infilling, the trough influenced the location and development of the Grenville Front tectonic zone and was the locus of Proterozoic regional metamorphism, plutonism, and other tectonic activity, which culminated in the formation of the present tectonic zone between 1200 and 1000 ma ago. Cataclastic deformation within the tectonic zone culminated during high grade regional metamorphism in the Grenville Province and continued on a diminished scale into the waning stages of this metamorphism. The boundary fault, which separates gneissic rocks of the Grenville Province from cataclastic rocks of the Southern and Superior provinces, was a zone of intense deformation with little apparent strike-slip and only minor dip-slip components (Lumbers, 1975). During late Proterozoic tectonism, rocks of the Grenville Province were flattened against the stable craton to the west and northwest and were transported upward with respect to the stable craton. Although rocks in the Grenville Province and the adjacent cratonic zone underwent considerable strain near the boundary fault, vertical slip on the fault was relatively small, at most a few kilometres. Accordingly, the boundary fault also marks a thermal boundary to the heat influx into rocks of the Grenville Province during late Proterozoic regional metamorphism.

Except for the late Archean granite exposed along the Grenville Front tectonic zone northeast of Sudbury (Fig. 21.12), which extends a short distance into the Grenville Province, Archean basement rocks are not exposed in the Grenville Province in Ontario. If the early Proterozoic rocks had been deposited on Archean basement, as we propose for the basins to the west, early Archean gneiss probably would have been brought to the surface, at least locally, during the diapiric uplift of the Algonquin batholith and other granitoid bodies in late Proterozoic time. In the absence of such basement rocks, we suggest that the early Proterozoic rocks possibly were deposited on a continental margin. Paleomagnetic measurements of rocks in the Grenville Province are not useful at this time in reconstructing the paleotectonic environment because of the limited accuracy of available pole determinations (Roy and Robertson, 1979).

TECTONIC ENVIRONMENT OF THE GREAT LAKES REGION

Early Proterozoic supracrustal sequences of the Great Lakes region, although physically separated by major structural elements and Phanerozoic cover rocks, have many common features. They are of the same general age

(2500-1800 Ma), form a linear belt south of the Archean Superior Province, and were variably affected by Proterozoic orogenic events. They also display differences, variations in internal stratigraphy, structural style, and metamorphism, that are probably attributable to differences in tectonic setting and especially to differences in the nature of the basement upon which they were deposited.

The early Proterozoic basins were initiated by crustal extension (rifting) that occurred in a cratonic setting in the Southern Province and a continental margin setting in the northwest Grenville. A cratonic environment for accumulation of early Proterozoic rocks in the Southern Province is indicated by the occurrence of Archean basement surrounding and within the basins in the Lake Superior region (Fig. 21.3). In the Lake Huron region, late Archean rocks form the basement to the north, and an early Archean gneiss basement to the south is inferred from similarities in tectono-stratigraphic patterns to the Lake Superior region. In contrast, Archean basement rocks, or indeed basement of any kind, is essentially absent in the northwest Grenville Province.

We have shown that abrupt changes in stratigraphic and tectonic patterns in the Southern Province approximately coincide with the boundary between two Archean basement terranes delineated in the Lake Superior region and proposed to exist in the Lake Huron region. This basement boundary constituted a zone of weakness that was the primary control for localizing the early Proterozoic basins. Faulting resulting from crustal extension was localized along the boundary, and the faulting, together with subsequent subsidence of the gneissic crustal segment, produced the traps required for accumulation of sediments derived from the craton to the north. Faulting along the boundary and subsidiary fracture zones controlled volcanism during evolution of the basin by tapping subcrustal magma sources.

The basement also had a secondary control on deposition and deformation, at least in the Lake Superior region. In the early stages of infilling, local, generally elongate, fault-bounded troughs developed subparallel to pre-existing basement structures (James et al., 1961; Cannon, 1973). These troughs received substantially thicker deposits of clastic and chemical sedimentary material than did intervening platforms (Larue and Sloss, 1980), and subsequently these rocks were deformed by uplift of the bounding Archean rocks. As a consequence, folds are diversely oriented throughout most of the Lake Superior basin known or inferred to be underlain by Archean gneiss.

The boundary between the two Archean basement terranes played an important role in controlling patterns of deformation, metamorphism, and igneous intrusive activity during subsequent orogenesis. The boundary approximately marks the northern limit of penetrative compressive deformation, of early Proterozoic felsic plutons, and of influx of metamorphic heat. The asymmetric patterns of deformation and metamorphism mimic the previously established depositional patterns.

Van Schmus (1976) interpreted the evolution of the early Proterozoic sequences of the Great Lakes region in terms of deposition on a passive continental margin and in foreland or back-arc basins, followed by subduction of oceanic crust beneath the continental margin along a north-dipping subduction zone. Cambray (1978) interpreted the early Proterozoic of the Lake Superior region as deposition on a rifted continental margin followed by continent-continent collision along a south-dipping subduction zone. Available data are not adequate to discriminate between an intracratonic environment, as we favour, or a continental margin environment for the Southern Province sequences. We consider, however, that a continental margin environment was most likely for the early Proterozoic of the northwest Grenville Province. Furthermore, if Penokean tectogenesis was the result of plate collision, the suture must lie somewhere to the south of the region under discussion.

We propose that the tectonic process operative when the Southern Province basins were formed was unique to the early Proterozoic and probably was restricted to the time interval during which the oldest sedimentary basins were formed on the Canadian Shield. There is no evidence that rifting proceeded to the stage where oceanic crust developed between disrupted crustal segments or that the compressional stage involved subduction. Instead, during evolution of the basins the two crustal segments seem to have been tightly juxtaposed, yet they exhibited vastly different tectonic stabilities. The contrasting mobility of the two basement terranes resulted from a thermal regime that produced diapiric reactivation, uplift and, probably, expansion of the gneiss terrane, accompanied by some lateral transport of this crustal segment against the more rigid greenstone-granite crust. For whatever reason, the gneissic crust retained its heat flux resulting from the decay of radio-elements, possibly augmented by abnormal subcrustal heat flow, during early Proterozoic time, whereas the greenstone-granite crust was depleted in lithophile elements by the end of the Archean.

In addition to visible diapirism and other evidence of deformation, rejuvenation of the Archean gneissic basement, culminating 1900-1800 Ma ago, is indicated by widespread cataclasis in the gneisses, several mylonite zones, strong disturbance of isotopic systems (Peterman et al., 1980; Van Schmus, 1976), and minor anatexis (Sims and Peterman, 1976). At least locally in the orogen, Archean structures in the gneisses were nearly obliterated by the Penokean deformation (Maass et al., 1980). Such transposition of Archean structures apparently was most prevalent in the southernmost part of the basin, which underwent the greatest differential uplift.

The pattern of early Proterozoic tectonism indicated in the Great Lakes region has counterparts in other shield areas, notably in South Africa and Western Australia. In these regions, stable Archean nucleii are surrounded by Proterozoic mobile belts which have been interpreted, respectively by Kröner (1977) and Gee (1979), as representing destruction and mobilization of Archean crust rather than lateral accretion. As noted by Gee (1979, p. 363), "...certain segments of Archean shields, particularly the ancient metamorphic complexes, seem to be predestined to continued mobility in the Proterozoic."

Although the early Proterozoic supracrustal rocks of the Grenville Province in Ontario probably were deposited at the margin of an Archean craton and therefore could represent an addition to the craton, other early Proterozoic basins in eastern Canada, such as the Labrador Trough (Dimroth, 1970) and the Cape Smith foldbelt (Thomas and Gibb, 1977), probably formed in the same tectonic environment as the intracratonic basins of the Great Lakes region. Thus, the early Proterozoic may have been a time of both lateral addition to the crust in North America, as suggested by Engel (1963), and of destruction and reworking of Archean crustal material, the remnants of which are preserved in and adjacent to linear early Proterozoic mobile belts.

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22. KEWEENAWAN GEOLOGY OF THE LAKE SUPERIOR BASIN

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Wallace, H., Keweenawan geology of the Lake Superior Basin; in Proterozoic Basins of Canada, F.H.A. Campbell, editor; Geological Survey of Canada, Paper 81-10, p. 399-417, 1981.

Abstract

Keweenawan rocks of the Lake Superior region were deposited during the formation of a continental rift system 1.0 to 1.3 Ga ago. Although rocks of that age are now exposed only around the lake, dense mafic igneous rocks associated with the rift are traceable in subsurface southwest as far as Kansas and southeast beneath the Michigan Basin, altogether forming an arc over 2000 km long.

A thin sequence of shallow marine sandstones 1.3 Ga ago unconformably covered Early and Middle Precambrian terrane in the western part of this region. Between 1.12 and 1.14 Ga, when actual rifting took place, these sediments were buried by enormous volumes of subaerially erupted tholeiitic basalt which accumulated rapidly in several distinct lava plateaus up to 12 000 m thick. Most flows appear to be the products of fissure-type eruptions but, as in the Tertiary of Iceland, central-vent cones have been identified with the flood basalt plateaus.

Rifting and volcanism ceased after only 60-90 km of plate separation, but isostatic subsidence continued under the crustal load of volcanic rock forming a major sedimentary basin roughly congruent with the present Lake Superior. While a succession of fluvial, lacustrine and shallow marine sediments were being deposited, the regional tectonic regime altered to one of moderate compression which steepened the sides of the basin forming the Lake Superior Syncline, and which produced a series of reverse faults parallel to the synclinal axis. Sedimentation into the Lake Superior Basin continued from all sides until early Cambrian time.

Résumé

Dans la région du lac Supérieur, les roches du Keweenawan se sont déposées pendant la formation d'une zone continentale de fracture, il y a 1,0 à 1,3 Ga. Bien que les roches de cet âge n'affleurent qu'autour du lac, on peut retracer au sud-ouest jusqu'au Kansas et au sud-est au-dessous du bassin du Michigan une dense formation de roches mafiques associée au rift et formant un arc de plus de 2000 km de long.

Il y a 1,3 Ga, une mince succession de grès marins peu profonds a recouvert en discordance des terrains d'âge précambrien inférieur et moyen dans la portion ouest de cette région. Il y a 1,2 à 1,14 Ga, alors que se formait le rift, ces sédiments on été recouverts par d'énormes volumes de basaltes tholéiitiques qui se sont écoulés lors d'éruptions sous-marines, et rapidement accumulés en formant plusieurs plateaux, dont certains atteignaient 12 000 m d'épaisseur. La plupart des coulées semblent s'être produites pendant des éruptions issues de fissures volcaniques béantes, mais, comme dans les terrains tertiaires d'Islande, des cônes à cheminée axiale ont été identifiés et associés aux plateaux formés par les épanchements basaltiques.

Les phases de fracturation et de volcanisme n'ont cessé qu'après que les plaques se soient séparées de 60 à 90 km, mais la subsidence isostatique s'est poursuivie en raison du poids exercé par la roche volcanique sur la croûte, et il s'est ainsi formé un vaste bassin sédimentaire qui coincide approximativement avec l'emplacement actuel du lac Supérieur. Pendant que se déposait une succession de sédiments fluviatiles, lacustres et de milieu marin peu profond, le régime tectonique s'est modifié; et une phase de compression modérée a approfondi les versants du bassin, donnant ainsi naissance au synclinal du lac Supérieur et produisant une série de failles inverses parallèles à l'axe synclinal. La sédimentation s'est poursuivie dans le bassin du lac Supérieur, de toutes parts, jusqu'au début du Cambrien.

INTRODUCTION

The name Keweenawan is applied to all the rocks in the Lake Superior region considered coeval with the native copper-producing volcanic and sedimentary rocks which form the Keweenaw Peninsula of the Upper Peninsula of Michigan (Fig. 22.1). Prior to the 1960s, it was generally believed that all of these rocks had been deposited over a relatively short period of time in a single constantly subsiding basin, roughly coincident with the present lake. In the last two decades, however, with the advent of radiometric dating and paleomagnetic techniques, with continued detailed geological and geophysical mapping on both sides of the international border and across the lake itself, and with the development of plate tectonic concepts, our understanding of the Keweenawan and the Lake Superior Basin has radically altered. This paper gives a brief overview of recent developments in Keweenawan geology with special emphasis on studies on the Canadian side of Lake Superior.

DISTRIBUTION OF KEWEENAWAN ROCKS

As shown in Figure 22.2, Keweenawan supracrustal rocks outcrop only in the vicinity of Lake Superior. However, detailed gravity and magnetic mapping combined with bore hole data clearly indicates that the Keweenawan strata continue under Paleozoic cover in a narrow continuous arc at least 2000 km long (Smith et al., 1966; King and Zietz, 1971). Southwest of Lake Superior, the Midcontinent Gravity High, which is generally attributed to dense mafic Keweenawan flows and intrusive rocks (Craddock et al., 1963; King and Zietz, 1971) extends as far as central Kansas in a series of short en echelon segments (see Fig. 22.3). Southeast of the lake, a similar though much weaker geophysical feature known as the Mid-Michigan Gravity Anomaly, extends at least as far as the Grenville Front in southern Michigan (Hinze et al., 1975; Oray et al., 1973). Recently the connection between Keweenawan igneous rocks and this gravity high has been confirmed by deep drilling in the Michigan Basin (Hinze et al., 1975; Ells and Champion, 1976). An extension of Keweenawan rocks even further to the southeast has been suggested by Halls (1978) based on weak gravity anomalies traceable as far as Tennesee.

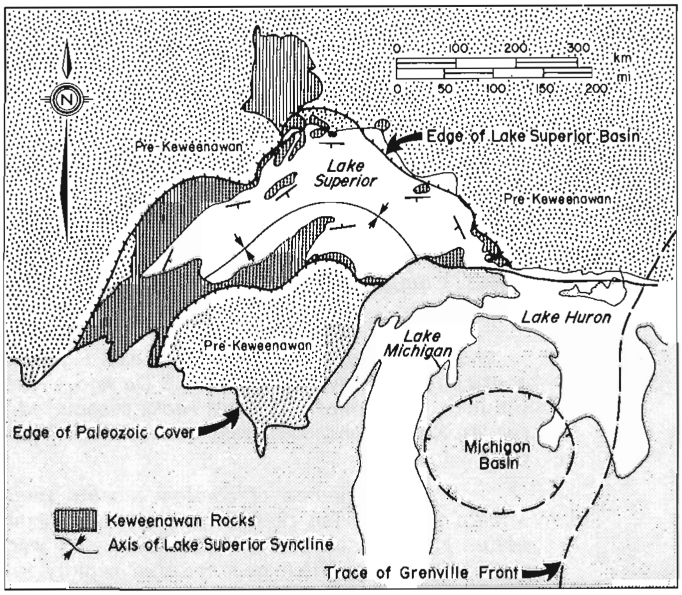


Figure 22.2. An outline of the Lake Superior Basin showing the distribution of exposed Keweenawan strata.

STRATIGRAPHIC CLASSIFICATION, PALEOMAGNETISM AND AGE

Since early work in the Upper Peninsula of Michigan, the Keweenawan Series has generally been subdivided into three divisions (Van Hise and Leith, 1911). Historically, all the shallow dipping volcanic rocks of low metamorphic rank found around Lake Superior were assigned to the Middle Keweenawan. Sandstones around the western part of the lake which these volcanics conformably overlie were referred to as Lower Keweenawan; the sandstones which conformably overlie the volcanics were included in the Upper Keweenawan.

Several paleomagnetic studies of the Lake Superior region have shown that there were only two magnetic pole reversals during the time of extrusion of Keweenawan volcanic rocks (Dubois, 1962; Palmer, 1970; Books and Green, 1972). The oldest rocks are normally magnetized and to date have been found only in a few localities in the Upper Peninsula and Minnesota. Reversely magnetized flows, which overlie these, have been identified in most of the major areas of Keweenawan exposure (Fig. 22.4). In turn these rocks are covered by a second normally magnetized sequence which is by far the most extensive and voluminous of the three. The second change in magnetic orientation is now widely accepted as a time marker useful for correlation purposes throughout the Keweenawan. In several Keweenawan sections investigated in detail the point of magnetic reversal has been marked by a distinct hiatus in volcanic activity. In the Osler Group (McIlwaine and Wallace, 1976) and Mamainse Point (Annells, 1973) area of Ontario, for example, thick conglomerate

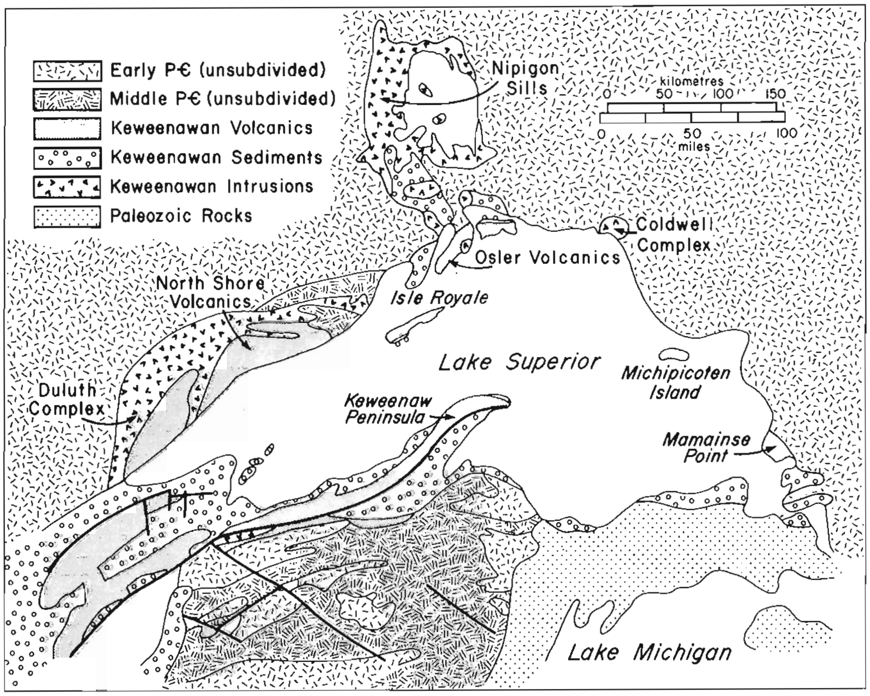
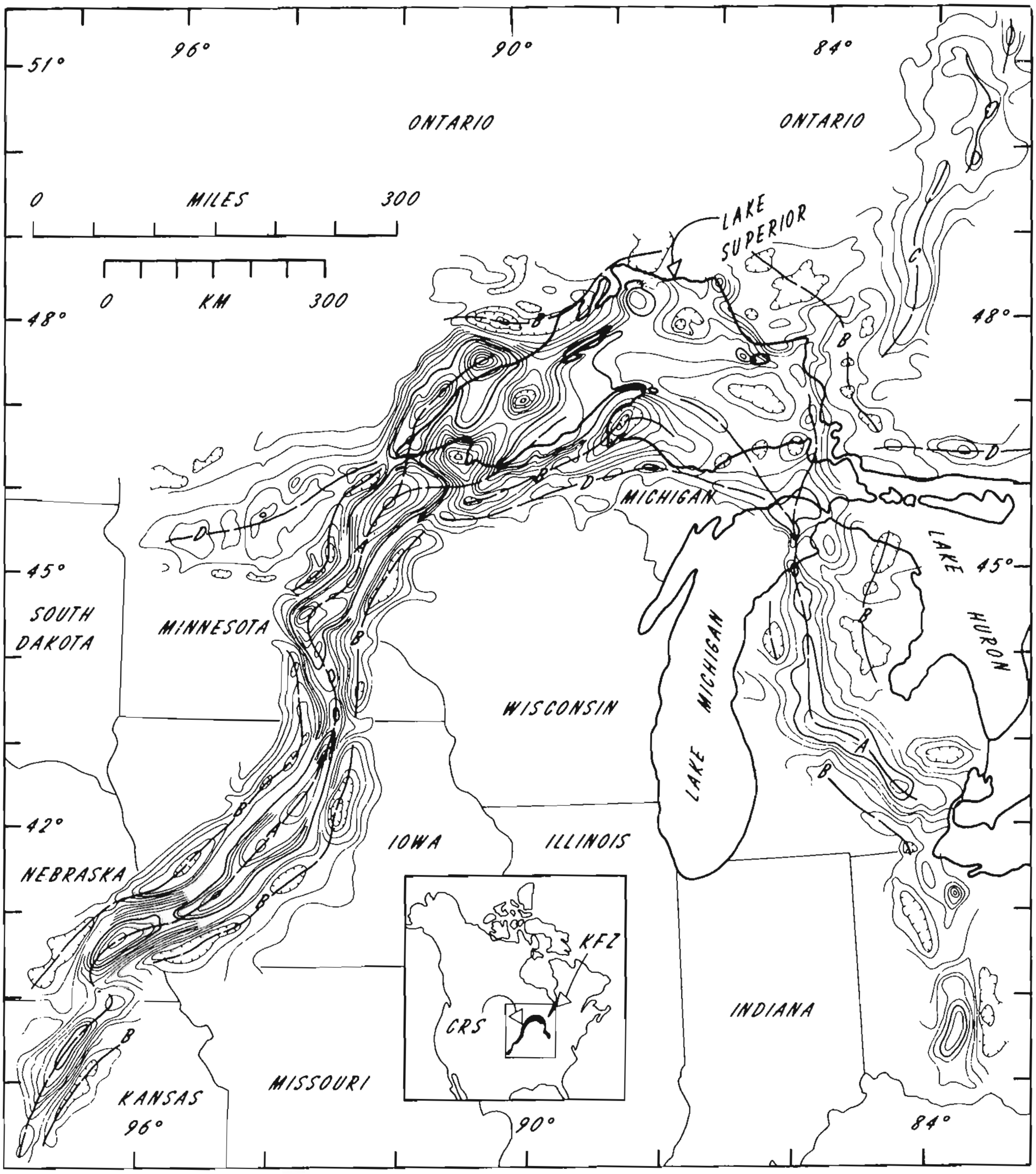


Figure 22.1

General geology of the Lake Superior region showing the distribution of Keweenawan volcanic and sedimentary strata.



heavy contour line = 0 mgals
contour interval = 10 mgals

Figure 22.3. *Bouger gravity map of Central North America. The Midcontinent Gravity High is interpreted to be due to dense Keweenawan igneous rocks. Gravity lows flanking these features are due to thick wedges of Keweenawan clastic sedimentary rocks in sharp fault contact with the volcanics (after Halls, 1978).*

wedges separate the normally and reversely magnetized flows. A thick nonmagnetic sequence separates normally and reversely magnetized flows in western Michigan. Hubbard (1975b) has interpreted these rocks as clastic sediments but they are not exposed and are known only from a few boreholes. In Minnesota, stratiform intrusions such as parts of the Duluth Complex have obliterated the supracrustal stratigraphy in the vicinity of the magnetic reversal, and Green (1972) believed these bodies may have been intruded along sedimentary horizons.

The apparent universality of this paleomagnetic reversal throughout the Keweenawan, and the evidence that the reversal coincided with a substantial gap in volcanic activity have prompted many investigators (e.g. Books, 1968; Green, 1972) to propose a major revision of the original stratigraphic scheme, moving the boundary between Lower and Middle Keweenawan to the point of the magnetic reversal. This change, however, has yet to be generally accepted, and in this paper the old usage of Lower and Middle Keweenawan has been retained. In discussions of Keweenawan volcanic rocks the magnetic orientations will be explicitly referred to where such distinction is necessary.

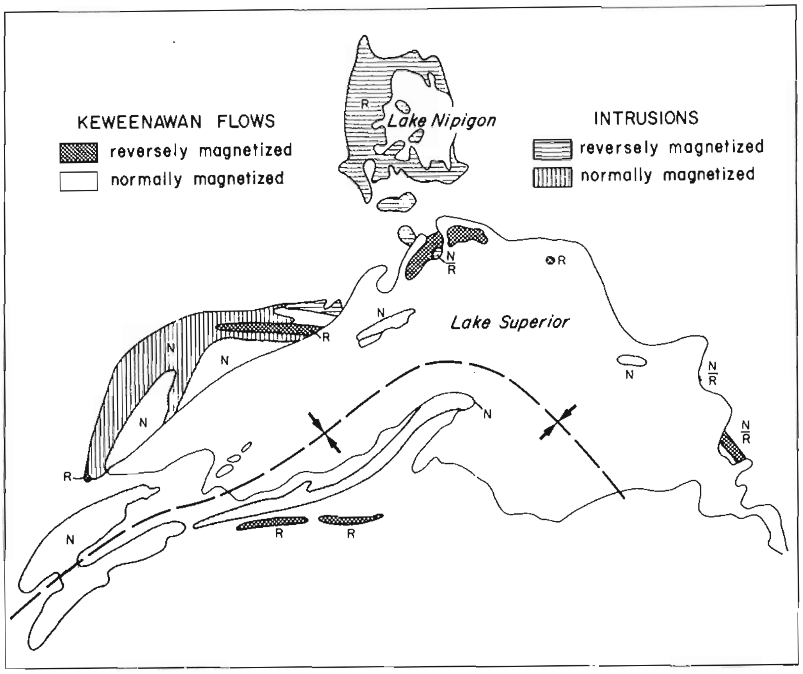


Figure 22.4

Distribution of reversely (older) and normally (younger) magnetized Keweenawan igneous rocks. Reversely magnetized flows are marginal to the Lake Superior Basin.

Radiometric age determinations are now available for Keweenawan volcanic rocks from most parts of Lake Superior (Goldich, 1968; Faure and Chaudhuri, 1967; Silver and Green, 1972). Nearly all the dates determined by Rb-Sr and U-Pb methods fall between 1000 and 1150 Ma, but this range is misleading in two ways. Firstly, the vast majority of dates have been obtained from the relatively young, normally magnetized, flows and intrusions. Sequences from which older ages may be expected, such as the lower, normally magnetized lavas of the Ironwood area in Michigan and the reversely magnetized Osler Group and the Slate Island flows, have not been dated. Secondly, results obtained by individual investigators have shown a much narrower range of ages. Silver and Green (1972), using U-Pb zircon methods, have shown that the North Shore Volcanics (both reversely and normally magnetized), the Duluth Complex and Endion Sill in Minnesota, as well as the Portage Lake lavas and the Mellen Sill of the Upper Peninsula were all emplaced between 1120 and 1140 Ma ago. As periods between magnetic pole reversals have averaged only 0.4 Ma over the last 80 Ma (Strangway, 1970) this relatively narrow range of about 20 Ma appears more reasonable for the span of Keweenawan time than the 100 to 150 Ma estimates previously given. The U-Pb zircon results of Silver and Green also agree well with the highest Rb-Sr ages obtained from several Keweenawan areas (e.g. Faure et al., 1969; Faure and Chaudhuri, 1967).

The age of Lower Keweenawan sedimentary rocks is less well established. Franklin et al. (1980) have reported a whole rock Rb-Sr isochron date of 1339 ± 33 Ma on the Sibley sandstone, roughly 200 Ma older than the igneous rocks dated by Silver and Green (1972). The Sibley date is comparable however, to the oldest ages found for the enormous diabase sills in the Lake Nipigon area which are approximately 1240 ± 34 Ma old (Franklin et al., 1980). Hence the forma-tion of the Sibley Group which is known to contain units of both normal and reverse magnetic polarity (Robertson, 1973), was roughly contemporaneous with the earliest Keweenawan igneous activity.

Upper Keweenawan clastic sediments, once collectively termed the Lake Superior Sandstones, range in age from 1000 Ma for the Oronto Group, including the Freda Formation which is believed to underlie most of Lake Superior, to 600 Ma for the upper part of the Bayfield Group which is unconformably overlain by Cambrian sedimentary rocks (Craddock, 1972a).

REGIONAL STRATIGRAPHY

Stratigraphic relationships between Lower, Middle and Upper Keweenawan rocks and the underlying Middle and Early Precambrian are summarized in Figure 22.5.

Pre-Keweenawan

In the western part of Lake Superior, Lower Keweenawan sediments rest unconformably upon Middle Precambrian metasediments and Early Precambrian granite-greenstone terrane. At the eastern end of the lake, where Lower Keweenawan and Middle Precambrian rocks are absent, Keweenawan flows and intercalated sedimentary units lie unconformably on Early Precambrian metavolcanic, meta-sedimentary and granitic rocks.

Lower Keweenawan

The Lower Keweenawan is a thin (50 to 240 m) sequence consisting predominantly of marine sediments. The Sibley Group and nearby Puckwunge Formation are considered correlative (Mattis, 1972), but their relationship with the Bessemer Formation which appears to occupy the same stratigraphic position south of Lake Superior on the Wisconsin-Michigan border, has not been clearly resolved. In all three of these sequences, discontinuous basal conglomerate units up to 3 m thick, consisting of clasts derived from local Middle and Early Precambrian rocks are overlain by mature quartz-rich arenites. Crossbedding and ripple marks indicate a north and northwest provenance. The well-sorted, well-rounded quartz grains in these arenites are

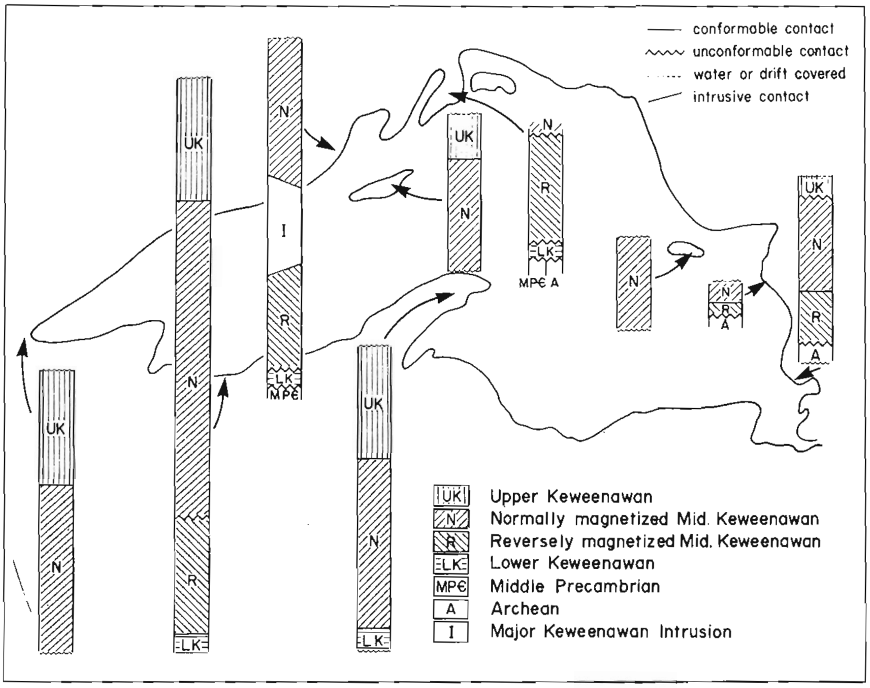


Figure 22.5

Columnar sections through major areas of Keweenawan exposure, comparing stratigraphic thicknesses and contact relationships.

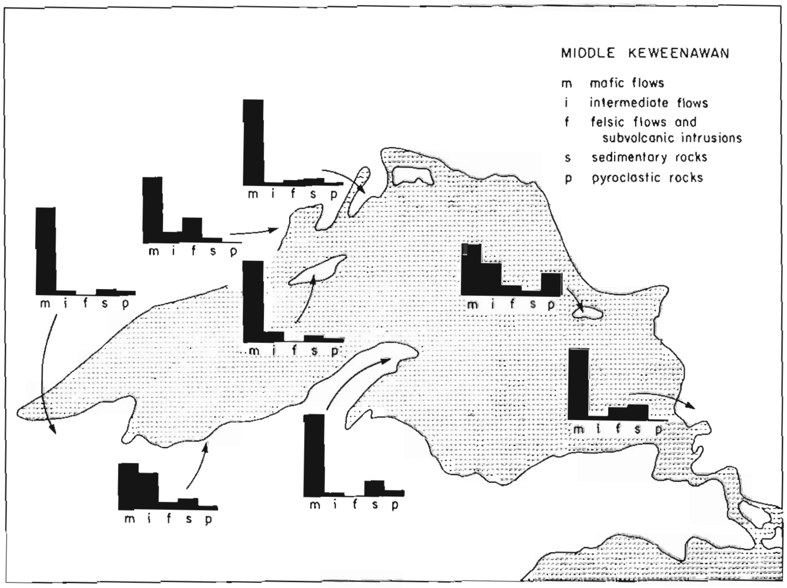


Figure 22.6

Relative proportions of lithologies present in major areas of Keweenawan exposure. Mafic flows predominate but in the western Michigan and Michipicoten Island intermediate volcanic rocks are volumetrically significant.

indicative of deposition in a shallow water environment. Studies of the Sibley Group by Franklin et al. (1980) suggested that sedimentation occurred during a series of northward transgressions and southward regressions of shallow seas in an intracratonic basin of relatively constant depth. From the present distribution of the Sibley strata, Franklin et al. (1980) concluded that deposition for the most part occurred in a north-south elongate, fault-bounded trough. The implications of the orientation of this trough will be discussed later in the text.

Middle Keweenawan

Contact relationships between Lower Keweenawan sedi-ments and overlying Middle Keweenawan volcanics vary across Lake Superior. On the northern side of the lake there is some evidence that the Sibley sediments have been lithified and mildly deformed prior to the extrusion of the first Osler Group flow. This contact, for the most part, is obscured by diabase and felsic porphyry intrusions of Middle Keweenawan age, but where exposed it has been interpreted as an unconformity or an erosional disconformity (Giguère, 1975). In Michigan and Wisconsin, however, several units of the Bessemer Formation occur well above the lowest Middle Keweenawan flow and the contact between the formations is apparently gradational (Hubbard, 1975b).

Although relative proportions of the Middle Keweenawan lithologies present vary considerably across the region, flood basalt flows are predominant in every major area of Middle Keweenawan exposure (see Fig. 22.6). Flows of intermediate and felsic composition are volumetrically far less important, but locally, as on Michipicoten Island

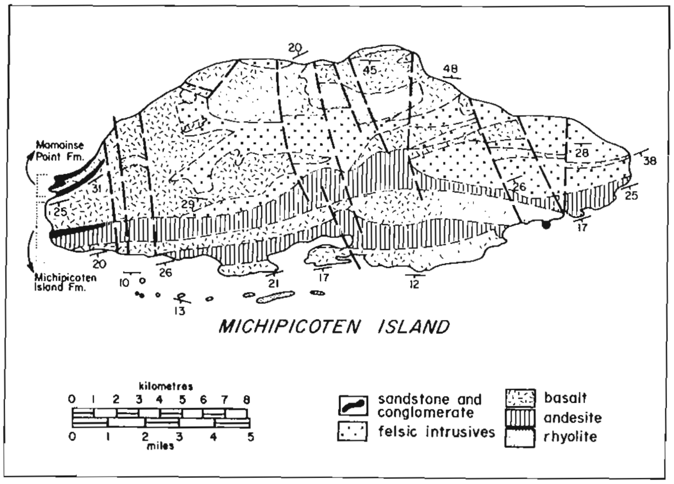


Figure 22.7. *Sketch map of the geology of Michipicoten Island, northeast Lake Superior (after Annells, 1974).*

(Fig. 22.6, 22.7), they form a significant fraction of the pile. The Michipicoten Island sequence is also unusual in that thick pyroclastic units occur at several stratigraphic levels. Elsewhere in the region, pyroclastic rocks are rare (as in most flood basalt provinces). Interflow sedimentary rocks form up to 15 per cent of the Keweenawan stratigraphic column in the Mamainse Point area, but are far less abundant in the west, constituting less than 5 per cent of the Osler Group, and only 1 per cent in the North Shore Volcanics.

Major intrusions of similar age to the Middle Keweenawan volcanics are prominent constituents in most areas. These range in composition from gabbro and anorthosite to granite, and include major alkalic and carbonatite complexes. In northeast Minnesota, the Duluth Complex volumetrically far exceeds the normally and reversely magnetized North Shore Volcanic flows which it intrudes (Green, 1972).

Mafic Volcanic Rocks

Estimates of the total thickness of Middle Keweenawan volcanics range as high as 12 000 m in Upper Michigan (White et al., 1971). In the North Shore Volcanics a section of at least 6400 m has been measured by Green (1972), and on Isle Royale (Huber, 1972), in the Osler Group (Wallace, 1972) and on Michipicoten Island (Annells, 1974) total thicknesses of volcanics are each in the order of 3000 m. The number of flows in these sections is not reliably known because of discontinuous exposure, but in relatively small areas such as Isle Royale over 100 have been documented (Huber, 1972) and on Mamainse Point over 300 flows were recorded by Annells (1973). White (1960) calculated the total volume of mafic Keweenawan volcanic rocks to be in excess of 400 000 km^3 making this one of the world's largest flood basalt provinces.

Keweenawan volcanic activity was almost entirely subaerial. Pillows have been recognized in a few basalt flows at the base of the North Shore Volcanics (Green, 1972) and hyaloclastic units have been reported from the lower portion of the Mamainse Point Formation (Annells, 1973), but elsewhere evidence for subaqueous deposition is virtually absent. Interflow sedimentary units have invariably been interpreted as fluvial, having been deposited by temporal streams crossing relatively flat volcanic surfaces (Green, 1972; Annells, 1973).

The flows are characterized by their remarkable areal extent, with continuity of thickness, morphology, lithologic characteristics and petrochemical composition over many tens of kilometres. Both White (1960) and Annells (1973) report tracing individual flows over 90 km along strike. Flow thicknesses range from a few centimetres to well over 100 m, and average between 7 and 12 m. Flows in excess of 30 m thick are common and several over 100 m thick are known from the Keweenaw Peninsula (Broderick, 1935) and the North Shore Volcanics (Green, 1972). Composite flows consisting of a number of nearly contemporaneous flow units each 10 to 30 cm thick are also common (Green, 1972).

The general morphology of most flood basalts (Fig. 22.8) includes a basal amygdaloidal layer, a central massive layer which generally constitutes two thirds of the total flow thickness, and a thick amygdaloidal flow top. The basal layer is typically very thin, consisting of chilled basalt with numerous ovoid or pipe-shaped amygdules filled with calcite, quartz, epidote, prehnite, chlorite or a variety of zeolites. The massive central layer typically exhibits well developed columnar jointing. Toward the top of each flow, amygdules increase in size and abundance until they are predominant in the upper 1 to 2 m. The degree of autobrecciation of flow tops varies considerably. Intact pahoehoe surfaces have been reported on some flows from most areas of Keweenawan exposure, but more typically the flow tops consist of angular blocks of amygdaloidal basalt in a matrix of aphanitic basalt grading upward into finer fragmental material. With the exception of the flows in eastern Lake Superior where Annells (1973, 1974) found that the major mineral phases were strongly altered throughout most flows, the effects of deuteric alteration and very low (zeolite facies) to low-grade (lower greenschist facies) burial metamorphism occur mostly in the highly permeable upper amygdaloidal zones.

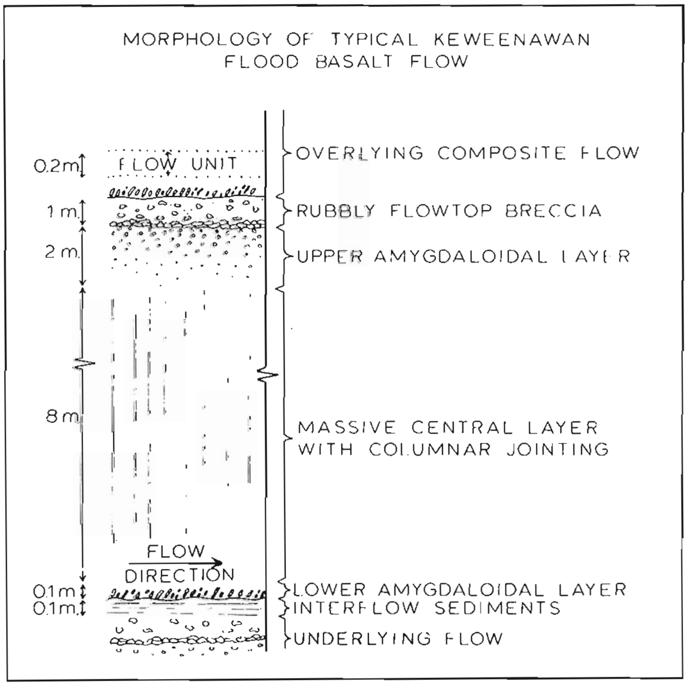


Figure 22.8. *Schematic cross-section through one Keweenawan basalt flow.*

In the massive zone, original igneous textures (e.g. intersertal, intergranular, ophitic, pilotaxitic), and mineralogy (plagioclase An_{45-60}, augite, pigeonite, Fe-Ti oxides, olivine and quartz) are commonly well preserved. Mild decalcification of plagioclase and serpentinization of olivine are the only changes seen in the massive parts of most flows. Green (1972) has reported undevitrified glass in the mesostasis of some of the North Shore Volcanics and glassy rhyolitic flows are known to occur in the Osler Group (Wallace, 1972).

In the amygdaloidal zones, igneous textures have generally been destroyed, and the original aphanitic basalt has been replaced by hydrous calcic minerals such as laumontite, pumpellyite, prehnite and epidote which commonly form extensive, essentially monomineralic domains (Jolly, 1971).

Studies in the Keweenaw Peninsula (Stoiber and Davidson, 1959; Jolly et al., 1972) and in the Osler Group (Wallace, 1972) have outlined distinct metamorphic zonation patterns which roughly parallel the volcanic stratigraphy. These patterns have been attributed to the maximum depth of burial attained by the flows.

Directions of lava flows have been determined in statistically significant numbers in several areas from such features as bent pipe amygdules and pahoehoe tops. Some studies have found no preferred flow direction (Green, 1972; Giblin, 1974), but in other areas distinct flow patterns have been cited (White, 1960; Giguère, 1975; McIlwaine and Wallace, 1976). On the Keweenaw Peninsula, normally magnetized flows moved southward from feeders near the axis of Lake Superior. On the Black Bay Peninsula and islands to the east, feeders for the reversely magnetized flows of the Osler Group were located to the north and west of the present areas of Keweenawan exposure. Similarly, in the Ironwood area of Michigan and Wisconsin flow directions were northward, from the margin of the Lake Superior Basin. The overall pattern appears to be that whereas the younger normally magnetized flows originated from feeders in the axial portion of Lake Superior, reversely magnetized sequences tend to show either random flow directions or were fed from areas outside the present area of Keweenawan exposure. This arrangement has significant implications for the interpretation of the Lake Superior Basin.

Intermediate and Felsic Volcanic Rocks

Stratiform felsic bodies, rhyolite to quartz latite in composition, form 5 to 25 per cent of Middle Keweenawan sequences, second in total volume only to the flood basalt flows. Most of these felsic rocks can be identified either as flows or as subvolcanic domes on the basis of morphology and mesoscopic contact features, but in many cases the mode of emplacement has not been unequivocally determined. Compared to the basalt flows, the felsic bodies tend to be much thicker (20 to 30 m on average), and far less extensive.

Andesitic and dacitic volcanic rocks are relatively rare in the Keweenawan and are all but absent in the Keweenaw Peninsula, in the Osler Group and in the Mamainse Point Formation. In the Michipicoten Island Formation, however, aphyric and glomeroporphyritic andesite flows are interlayered with andesitic pyroclastic units. Annells (1974) interpreted the volcanic activity which formed most of the Michipicoten Island Formation to be of the explosive Plinian-type and identified the southern part of the island as a volcanic centre.

In the Porcupine Mountains of the Upper Peninsula of Michigan andesitic flows with a few dacitic and rhyolitic flows form a pile up to 3000 m thick. This sequence is known almost entirely from subsurface data, but from dip and thickness trends White et al. (1971) interpreted it as a shield-type central vent accumulation surrounded and buried by flood basalts.

Roughly 10 per cent of the North Shore Volcanics of Minnesota consist of andesite, trachyandesite and dacite flows, and a further 20 per cent is made up of rhyolitic to trachytic flows and minor intrusion (Green, 1972).

Chemistry of Keweenawan Volcanics

The major element petrochemistry of Keweenawan volcanics is summarized in Figure 22.9 which is based on analyses from Keweenaw Peninsula (Broderick, 1935), the Ironwood area of western Michigan (Hubbard, 1975b), the North Shore Volcanics of Minnesota (Green, 1972), the Osler Group of the Black Bay area, Ontario (Wallace, 1972), the Slate Island (Patterson Island) flows (Sage, in press), the Mamainse Point Formation (Annells, 1973) and the Michipicoten Island Formation (Annells, 1974). From this data and other recent studies (Phinney, 1970; Brannon et al., 1979) a number of general observations and conclusions have been made regarding the petrogenesis of Keweenawan volcanics.

A. Flows in the reversely magnetized sequences are relatively primitive in composition (i.e. olivine tholeiites), and appear to have been derived from a single homogeneous source, presumably the upper mantle (Annells, 1973; Green, 1972).

   Only limited trace element data have been published for Keweenawan rocks, but the information available when applied to discrimination such as the Ti-Zr-Y and Ti-Zr-Sr plots of Pearce and Cann (1973) suggests that these volcanics are chemically equivalent to modern "within-plate" basalts.

B. Despite their apparent lithologic monotony, distinct and rather unusual petrochemical trends have been reported in the basalts from several areas (Wallace, 1972; Phinney, 1970; Annells, 1973). These trends differ considerably between adjacent areas, such as the North Shore Volcanics and the Osler Group. The Osler flows exhibit a distinct tholeiitic trend of iron enrichment and are uniformly potassium-poor, whereas the North Shore Volcanics are neither clearly tholeiitic nor calc-alkaline and are 5 to 10 times more potassic (Green, 1972).

C. Differentiation trends in at least two areas, the North Shore Volcanics and the Osler Group, have been attributed to fractional separation of plagioclase-rich cumulates such as gabbroic anorthosite (Phinney, 1970; Wallace, 1972). Recent work by Green (1979) however has suggested that the Keweenawan basalts could not have been the products of simple fractional crystallization and has cast doubt on the theory that intrusions such as the Duluth Complex are directly related to the flows they intrude.

D. Simple fractionation by either crystal settling or partial melting cannot explain (1) trace element trends that have been reported from the North Shore Volcanics (Brannon et al., 1979) and Michipicoten Island (Annells, 1974), notably the Ni, Co, Cr variation with stratigraphic position; (2) the high proportion of rhyolitic rocks relative to basalts in the older volcanic sequences; (3) the highly potassic nature of the intermediate and

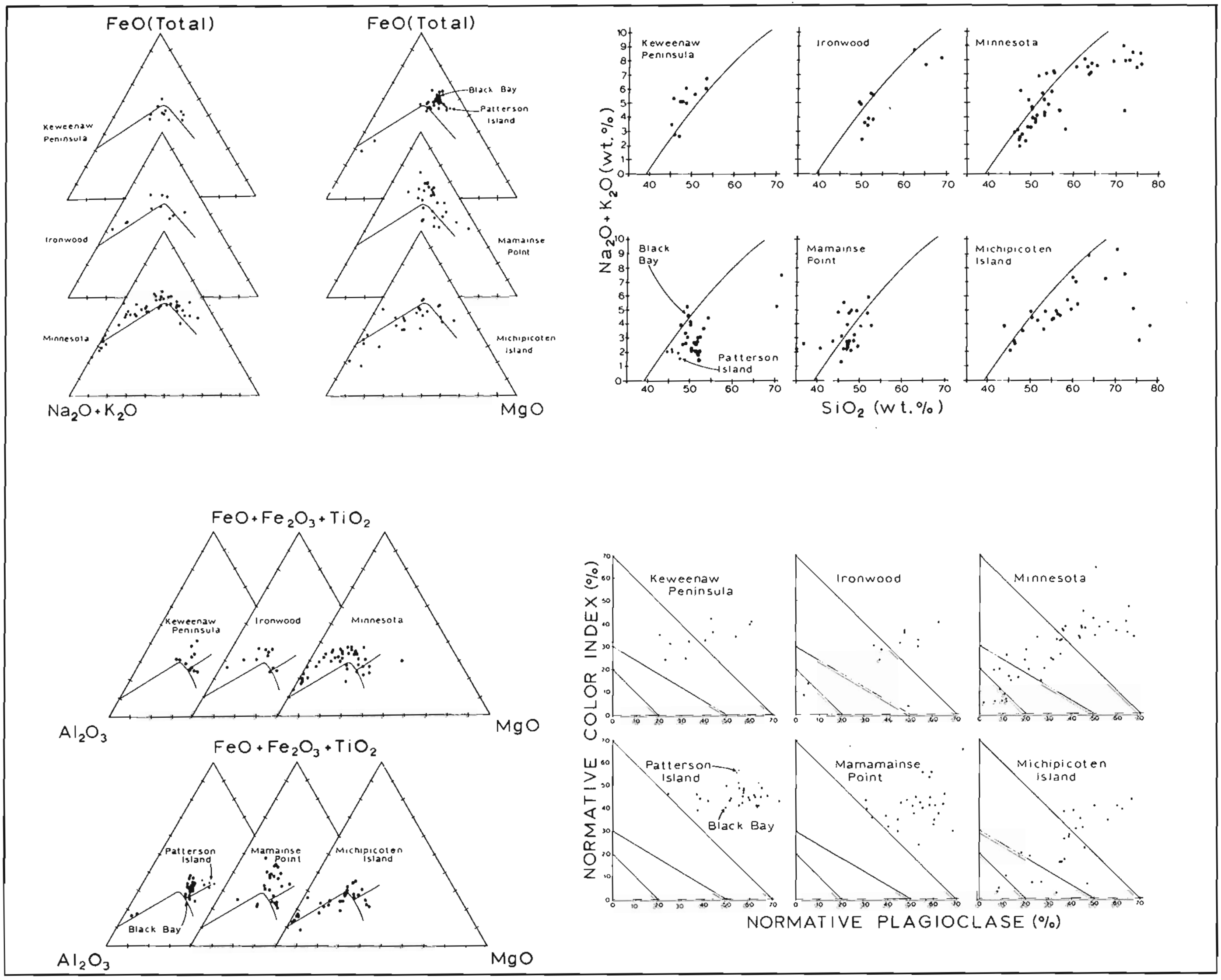


Figure 22.9. *Standard AFM, SiO_2 vs. Na_2O + K_2O, normative plagioclase vs. colour index and MgO-FeO-Al_2O_3 plots (Irvine and Baragar, 1971 and Jensen, 1976) summarizing petrochemical variations within and between six major areas of Keweenawan exposure.*

felsic flows of the North Shore Volcanics; and (4) the petrochemical bimodality evident in most of the Keweenawan sections. These problems can be at least partially resolved by invoking the simultaneous existence of two magmas, one basaltic and one rhyolitic, and allowing for magma mixing in some areas to produce intermediate rocks (Green, 1972; Annells, 1974; Phinney, 1970; Massey, 1980). The upward movement of large volumes of basaltic magma into the granitic crust must have initially produced melting and assimilation on a large scale, hence the preponderance of rhyolitic volcanic rocks in the older sequences.

Middle Keweenawan Sedimentary Rocks

Interflow sedimentary units constitute only a small fraction of the Osler and North Shore Volcanic groups. These are either lithic sandstones consisting of angular volcanic detritus and quartz, or polymictic conglomerates which vary greatly in clast composition depending upon proximity to the margin of the basin. Some interflow units have recently been reinterpreted as felsic pyroclastic rocks such as ash-fall tuffs (Aaquist, 1977).

Both the sandstone beds, which are generally less than 0.5 m thick, and conglomerates, which vary up to 100 m and more thick, occur as lenticular units traceable along strike for only short distances. In some cases the conglomerates are graded and distinctly crossbedded, and where they occur in sequence they contain channel-fill type features. Typically there is little evidence of prolonged erosion of the volcanic unit beneath the sedimentary bed. The volcaniclastic sandstones in many places coarsen downward into the underlying brecciated flow top. Altogether the evidence indicates that these sediments were deposited by ephemeral streams (Green, 1972; Giguère, 1975; McIlwaine and Wallace, 1976).

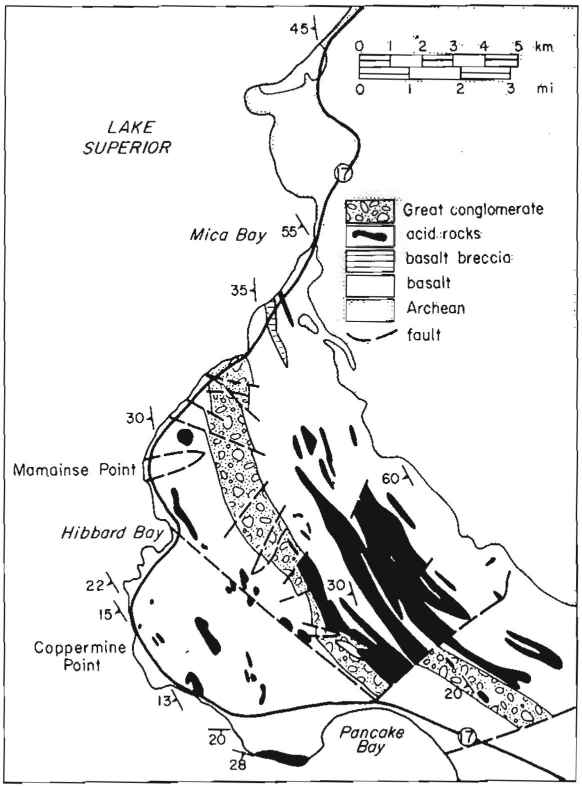


Figure 22.10. *Sketch map of the geology of the Mamainse Point area (after Giblin, 1969).*

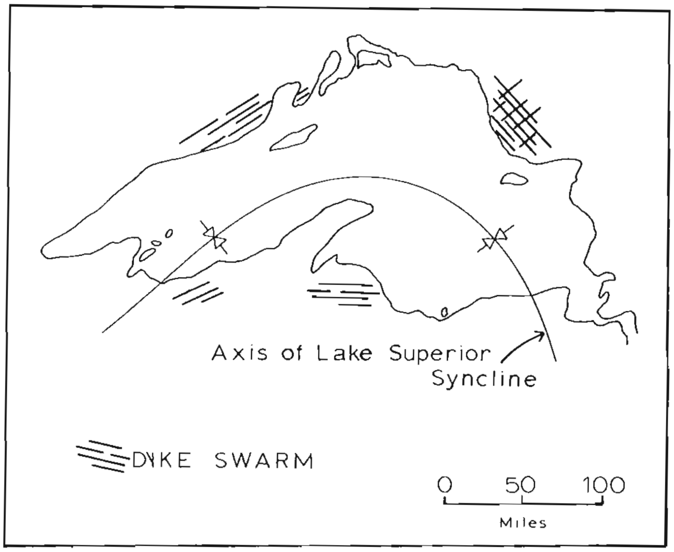


Figure 22.11. *Major Keweenawan dyke swarms around Lake Superior occur parallel and normal to the axis of the Lake Superior Syncline.*

The Mamainse Point Formation is unusual in the Keweenawan in that massive polymictic conglomerate units comprise roughly one quarter of the total section of 4300 m (Annells, 1973). This may be a function of a relatively slow accumulation of the flood basalt in this area. Clasts in the conglomeratic units are predominantly Archean rock types derived from the east and southeast. The conglomerates are commonly over 30 m thick. However, the unit reported by Palmer (1970) to separate reversely and normally magnetized flows, is over 540 m thick (see Fig. 22.10).

On the Keweenaw Peninsula interflow conglomerate beds consist predominantly of rhyolite clasts derived from Keweenawan flows exposed to the south around the margin of the basin (Hubbard, 1972).

Keweenawan Intrusive Rocks

Large volumes of intrusive rocks, mostly gabbroic in composition, were injected into Lower and Middle Keweenawan strata and underlying Middle and Late Precambrian rocks between roughly 1300 Ma and 1000 Ma ago. The oldest of these intrusions appears to be some of the Logan sills of the Lake Nipigon and Thunder Bay area. Many of these tabular sheets of diabase exceed 150 m in thickness and, as can be seen in Figure 22.1, have enormous areal extent. Dykes of similar age occur in swarms which are especially prominent along the northeastern and northwestern shores of Lake Superior (Fig. 22.11). In the vicinity of the lake, the dykes trend roughly parallel to the axis of the lake itself, but further from the shore most dykes of this age occur in swarms radiating from the centre of the lake. The wide-spread distribution of these dykes well outside areas now underlain by Keweenawan supracrustal rocks suggests that the flood basalts may initially have had a far greater areal distribution.

By far the largest of the Keweenawan intrusions is the Duluth Complex of Minnesota, a composite body consisting mostly of gabbroic anorthosite and troctolite with lesser volumes of dunite, norite and ferrogabbro (Phinney, 1972). The complex is roughly 6500 km^2 in area and occupies a subconcordant position between paleomagnetically reversed and normal flows in the North Shore Volcanic Group (Green, 1972) with most phases of the intrusion itself being normally magnetized. Many investigators interpreted the Duluth Complex as the crystal cumulate parts of a magma chamber where the material which formed the overlying volcanic rocks underwent differentiation. A similar relationship has been postulated between the Moss Lake Intrusion, an anorthositic gabbro lopolith (Keeler, 1971) and the Osler Group (Fig. 22.12) (McIlwaine and Wallace, 1976). Many smaller intrusions such as the Beaver Bay Complex and Endion Sill of Minnesota (Green, 1972), and the Mellen Complex of Michigan and Wisconsin (Hubbard, 1975b) are of similar gabbroic to anorthositic compositions, and may also have formed as cumulate concentrations in differentiating magma chambers.

The Coldwell Complex at the northeastern corner of Lake Superior is by far the largest of a series of alkalic and carbonatite intrusions in that area. Age determinations by Gittins et al. (1967) indicate that these are roughly the same age as Keweenawan magmatism. The older rocks in the complex are gabbro and larvikite, and the younger, less abundant phases found in the inner part of the body are syenodiorite and nepheline syenite. The relationship between the Coldwell Complex and the other alkalic and carbonatite intrusions nearby and the Keweenawan basaltic rock just to the south remains to be explained.

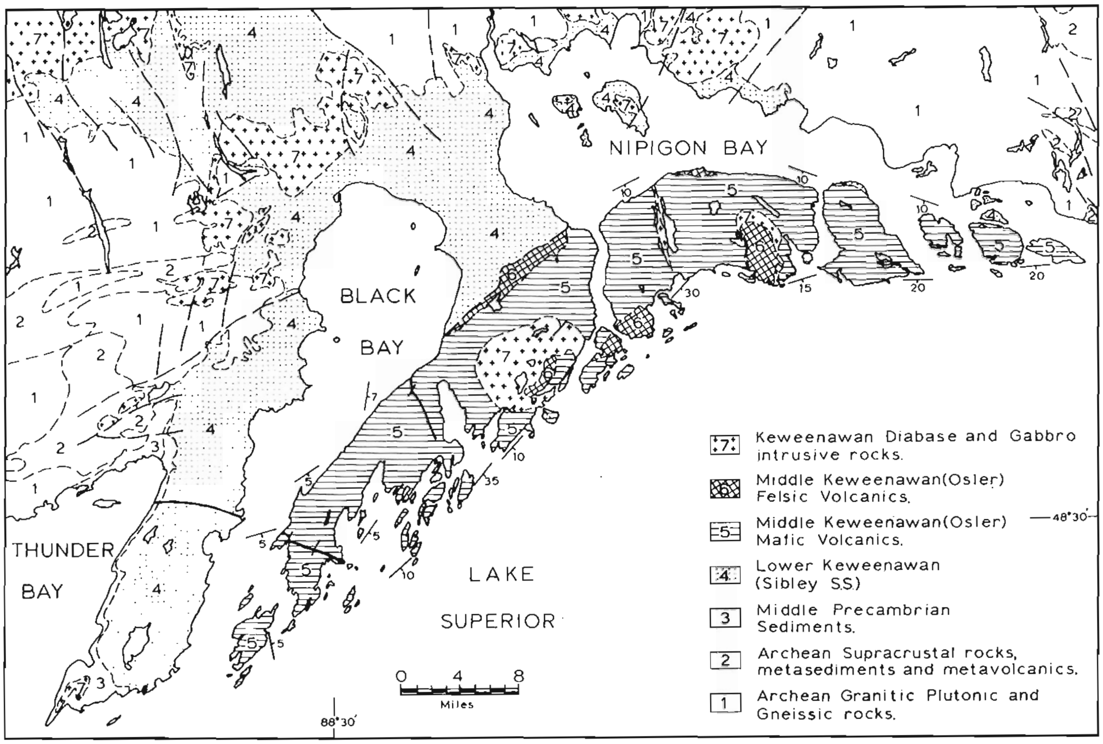


Figure 22.12. *General geology of the Black Bay Peninsula and vicinity (Giguère, 1975; McIlwaine and Wallace, 1976).*

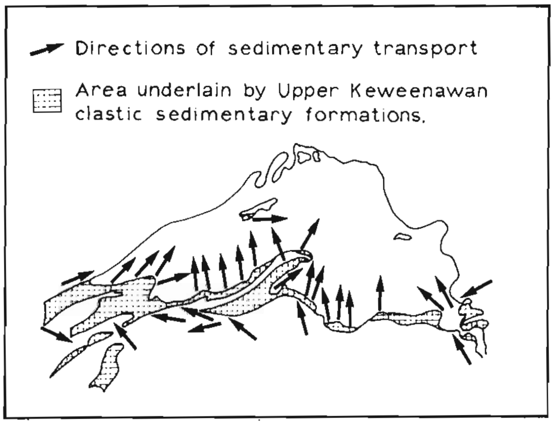


Figure 22.13. *Directions of sedimentary transport based on cross-stratification in Upper Keweenawan sandstones indicate infilling of the Lake Superior Basin with detritus from highlands around its margins (after Hamblin, 1961).*

Upper Keweenawan

After volcanic activity waned in the region, the Lake Superior Basin remained a depository for clastic sediments more or less continuously into Cambrian time. A combined thickness of between 6500 and 10 000 m of fluvial, lacustrine and shallow marine deposits blanketed the area during that time. These rocks are now exposed mostly along the southern shore of Lake Superior and in a few localities along the eastern side. Crossbedding and other paleocurrent indicators show movement of detritus from the margins to the centre of the basin (Fig. 22.13) (Hamblin, 1961; Hamblin and Horner, 1961; Huber, 1972).

The oldest Upper Keweenawan sedimentary rocks are the conglomerate units of the Copper Harbor Formation, exposed in the Keweenaw Peninsula and Isle Royale (see Fig. 22.14). The conglomerate beds, together over 2000 m thick, conformably overlie and are in part intercalated with normally magnetized flows of the Portage Lake lavas. Relative dips and flow directions in the flood basalts and interlayered conglomerate units indicate that basin subsidence occurred continuously during the transitional period between predominantly volcanic accumulation and clastic sedimentation, with slope changes taking place between each eruption and the following period of gravel deposition (White, 1960). These conglomerates have been interpreted as fluvial fanglomerates and flood plain deposits by Huber (1972).

On the Keweenaw Peninsula, the Copper Harbor Conglomerate is overlain disconformably by the Nonesuch Formation which consists of 70 to 200 m of siltstone and coarse arkosic sandstone. The Nonesuch is, in turn, conformably overlain by the Freda Formation (Hubbard, 1975a). Sandstones of the Freda Formation which is in places more than 3600 m thick, are interpreted to be flood plain and lacustrine deposits (Hamblin, 1961) and are considered by many workers to be the youngest Keweenawan strata. A thin sequence of sandstone similar to the Freda occurs in the Alona Bay area just north of Mamainse Point (Giblin, 1974) where sandstone beds unconformably overlie normally magnetized Middle Keweenawan flood basalts.

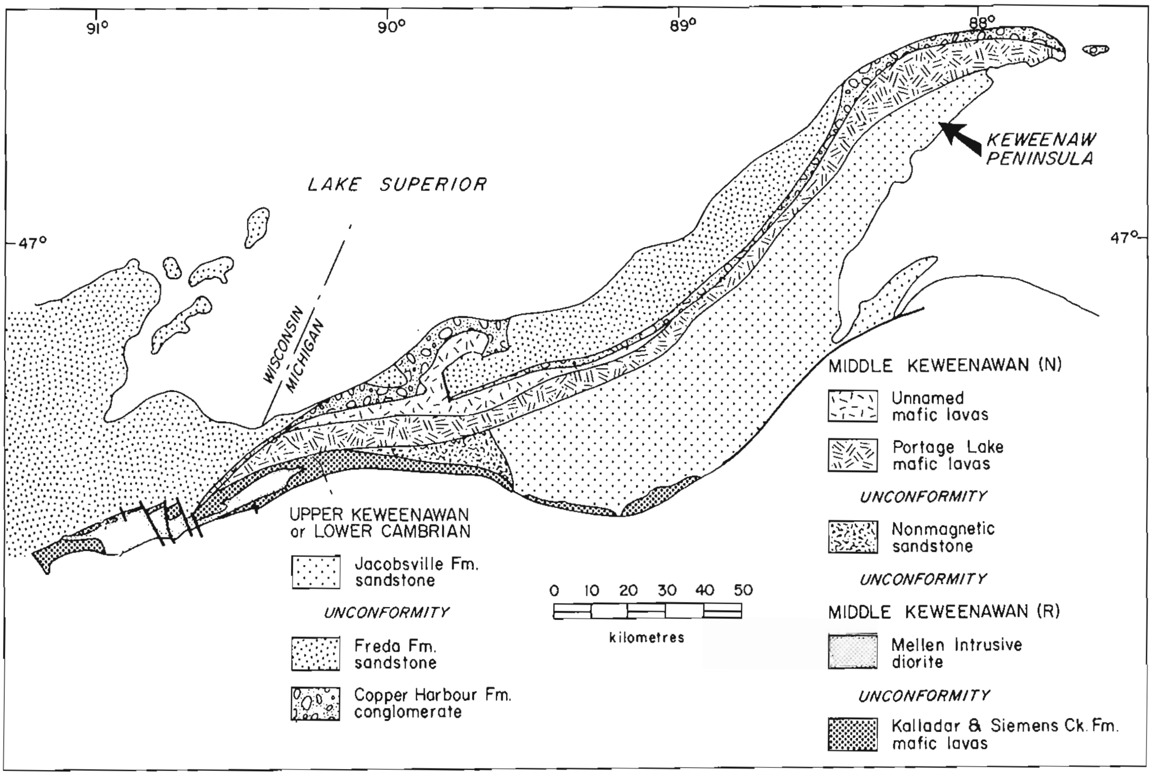


Figure 22.14. *Sketch map of the geology of the Keweenaw Peninsula, Upper Peninsula of Michigan (after Hubbard, 1975a,b).*

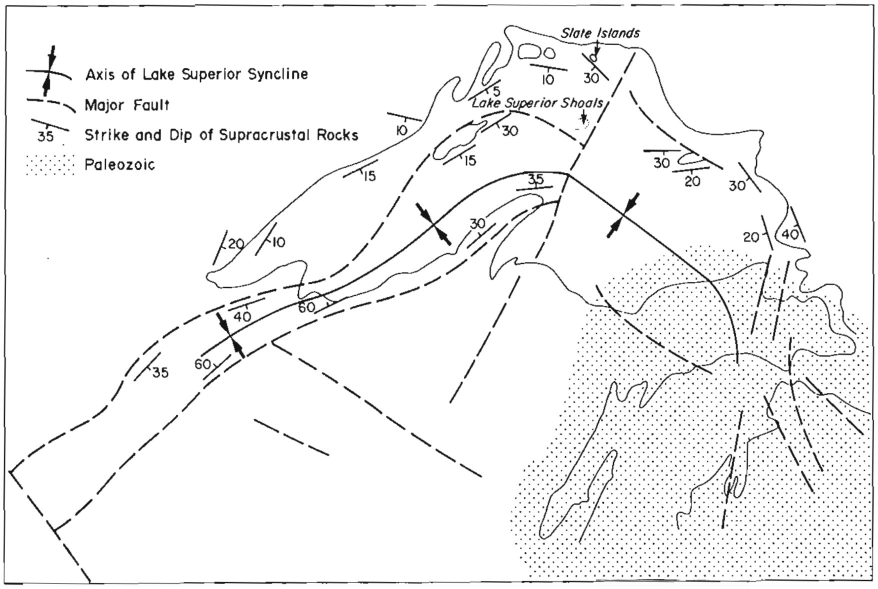


Figure 22.15

Major structural features of the Lake Superior Basin. Fault correlations in western Lake Superior are those of Klasner et al. (1979). Inferred structures under Paleozoic cover interpreted by Oray et al. (1973).

The contact between the Freda Formation and the Jacobsville sandstone (also known as the Bayfield Group) is not exposed and its nature has never been conclusively established. Some evidence suggests that Freda rocks were deformed and eroded prior to the deposition of the Jacobsville sequence (Craddock, 1972a). However, there is still considerable controversy regarding age of the Jacobsville, as some assign it to the Upper Keweenawan (Dubois, 1962; Craddock, 1972a), and others assign it to the Lower Cambrian (Hamblin, 1958). The Jacobsville-Bayfield Group, a maximum of 1400 m thick, is believed to underlie most of Lake Superior (Halls and West, 1971), and the size and shape of the present lake is determined by the distribution of these rocks and their susceptibility to erosion relative to older Keweenawan, Middle Precambrian and Archean rocks (White, 1966a,b; Halls and West, 1971).

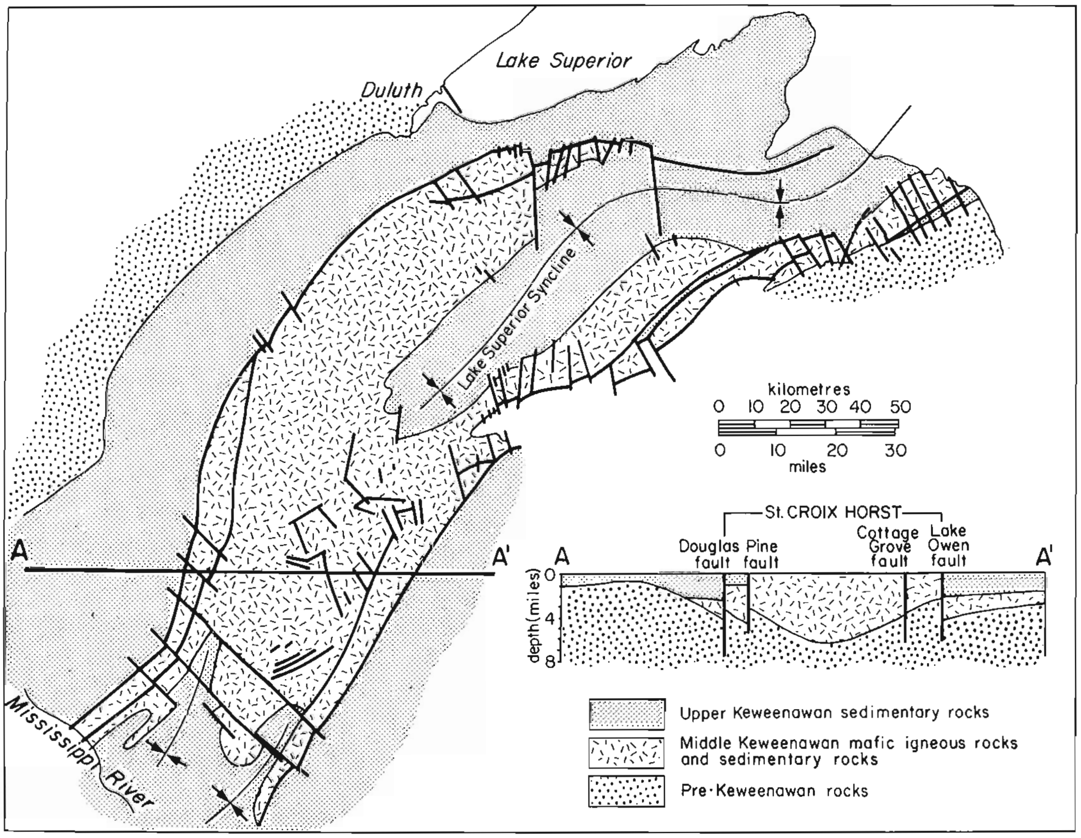


Figure 22.16. *Precambrian geology of east-central Minnesota showing fault-controlled distribution of Keweenawan rocks, and the western termination of the Lake Superior Syncline (after Morey and Mudrey, 1972).*

STRUCTURE OF THE LAKE SUPERIOR BASIN

The Lake Superior Syncline was recognized first by Irving (1883) but it was not named as such until 1923 (Hotchkiss, 1923). As shown schematically in Figure 22.15, Keweenawan volcanic and sedimentary strata are tilted toward the centre of the lake from all sides with considerably steeper dips on the south side than on the northeast and northwest. The second major structural feature of the basin is a set of long reverse faults which parallel the axis of the major syncline. These faults form a horst and graben system which has been documented in the eastern part of Minnesota and adjoining Wisconsin where the main fault block, some 30 to 80 km across, is known as the St. Croix Horst (Craddock et al., 1963). In east-central Minnesota (Fig. 22.16) the central block of the St. Croix Horst is believed to have been elevated 2 to 4 km relative to the surrounding troughs (Morey and Mudrey, 1972). For a time it acted as a topographic high and shed volcanic detritus into the flanking depressions. The movement along the faults brought dense, magnetic Middle Keweenawan volcanic rocks into juxtaposition with Upper Keweenawan clastic sediments, creating high gravity and aeromagnetic contrasts over the structure (Craddock et al., 1963).

Geophysical maps (see Fig. 22.3) clearly show that similar patterns continue northeast of the well-exposed areas (Klasner et al., 1979; King and Zietz, 1971) parallel to the axis of the Lake Superior Syncline, up the Keweenaw Peninsula and across the central part of the lake. The continuation of both the fault system and the Lake Superior Syncline under the southeast corner of Lake Superior and southward under Paleozoic rocks into southern Michigan is somewhat more tenuous (Oray et al., 1973).

The relationship of the major syncline and coaxial horst-graben system has now been well established (Craddock, 1972b). The broad syncline first began to form during the rapid accumulation of dense, normally magnetized Middle Keweenawan flood basalts in the axial portion of the syncline. From the opposing dips of flows and intercalated sediments from all parts of the lake, it has been inferred that the steady downwarping of the crust due to increasing load, and the infilling of the basin were roughly in balance. During deposition of Upper Keweenawan sediments the initially tensional tectonic regime altered. Northwest-southeast compressive stress developed which formed north-trending folds within the Upper Keweenawan sediments and produced a steepening of the southeast limb of the Lake Superior Syncline (Craddock, 1972a). The same compressive stress produced the major reverse faults which bound the axial horst-graben system. Movement along these high angle faults was virtually completed during deposition of the Jacobsville-Bayfield sandstones (Craddock, 1972b) and only slight readjustments have taken place along them since the beginning of the Paleozoic.

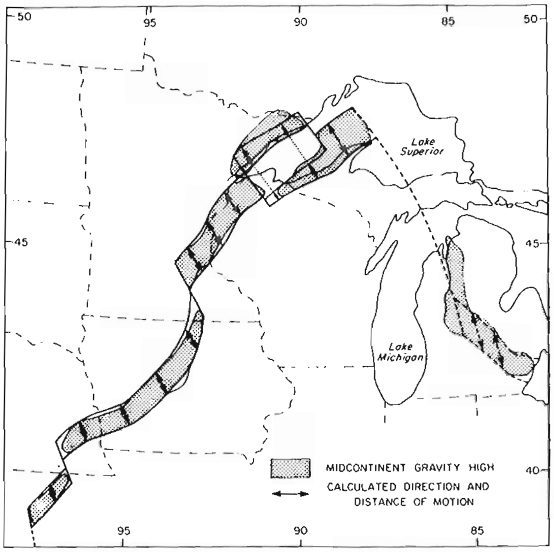


Figure 22.17. *Magnitude of Keweenawan Plate separation. Excellent fit is apparent between the geometry of the Keweenawan Rift System as predicted by the best-fit pole of relative rotation model, and the superimposed Mid-continent Gravity High (Chase and Gilmer, 1973).*

THE KEWEENAWAN AS A PALEO-RIFT

The possibility that the Keweenawan represents an ancient rift system similar to those active today in Iceland or East Africa, was first put forward by King and Zietz (1971) and White (1972), who pointed out that the Keweenawan belt appeared to be a series of en echelon linear segments in a pattern which strongly resembles a spreading ridge-rift system with offsets along transform faults (Morgan, 1968). In a test of this hypothesis Chase and Gilmer (1973) attempted to measure the correlation between the pattern of plate separation predicted by a best-fit pole of relative rotation, and the widths of gravity anomalies over the belt of Keweenawan rocks extending from Lake Superior to Kansas. The agreement between the theoretical pattern of spreading and observed gravity anomalies is remarkably good (Fig. 22.17); however, spreading from the best-fit pole selected by Chase and Gilmer could not explain the apparent extension of the Keweenawan igneous rocks southeast of Lake Superior.

Burke and Dewey (1973) proposed that the two segments of the Keweenawan extending southwestward and southeastward from Lake Superior represented rift arms of a plume-generated triple junction in which a north-trending arm had failed to develop. Again this hypothesis was based solely on the geometry of the Keweenawan igneous belt as defined by geophysics (see Fig. 22.18). This triple junction was believed to have been centred in northeastern Lake Superior, and the Kapuskasing structure, a broad linear fault-bounded slice of amphibole-pyroxene granulite and gneiss, was thought to represent the so-called "failed arm". Carbonatite and alkalic intrusions along the Kapuskasing structure are of two ages. The older are roughly 1.7 Ga, but the younger date at 1.1 Ga (Gittens et al., 1967), the time of peak Keweenawan volcanism. Hence, Burke and Dewey (1973) suggested that the "failed arm" of the Keweenawan triple junction produced reactivation of a pre-existing zone of weakness.

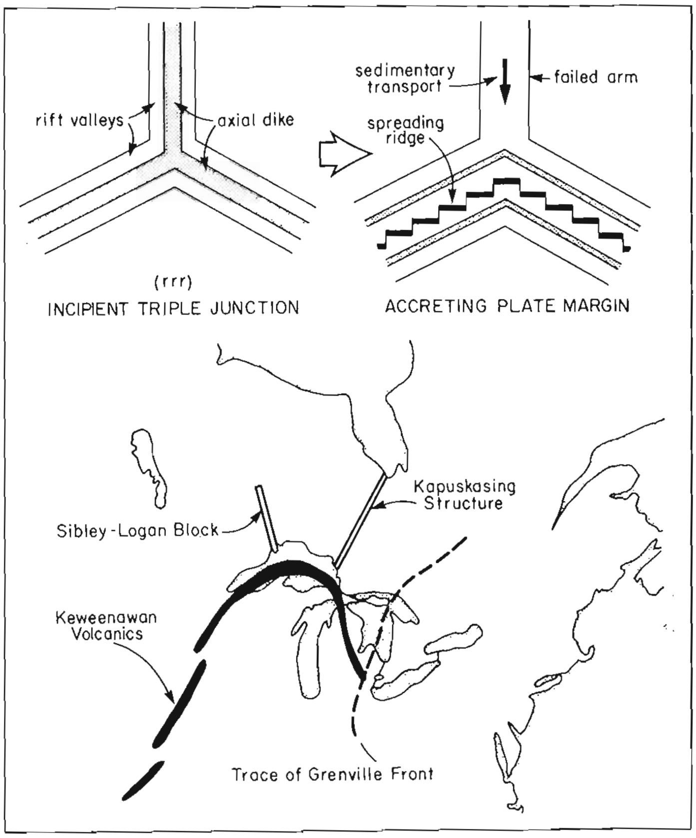


Figure 22.18. *Schematic evolution of a plume-generated triple junction (after Burke and Dewey, 1973), in comparison with the geometry of the Keweenawan Rift System.*

Franklin et al. (1980) proposed a variation of the previous model by suggesting that the "failed arm" might have been a fault-bounded trough extending from Lake Superior to the northern part of Lake Nipigon. This graben they suggested, acted as a channel for the northward transgressive sea in which the Sibley Group sediments were deposited. The early Keweenawan igneous activity which produced some of the Logan Sills in the Lake Nipigon and Thunder Bay areas, is also believed to be related to the formation of this failed arm.

Aside from similarities in geometry between the Keweenawan igneous belt and present ridge-rift systems there are several lines of evidence supporting the paleo-rift theory. Seismic and gravity data (Smith et al., 1966; White, 1966a) strongly indicate that the axial portion of Lake Superior is occupied by a relatively narrow, sharply bounded, vertical zone of very dense, high velocity material extending down to the upper mantle with no intervening granitic crust. This zone may be a single body, or more likely, a massive swarm of sheeted gabbroic dykes which expanded laterally by repeated injection of magma as the continental crust was pulled apart. Such sheeted dyke swarms are a well-documented component of oceanic ridge-rift systems (Dewey and Bird, 1971). Seismic evidence also indicates that both the thickest, and some of the thinnest crust in North America occurs around Lake Superior, with depth to Moho varying from only 20 km just west of the lake to at least 50 km under the central part. The extreme thinness may be attributable to attenuation of the crust during the incipient stages of rifting and plate separation. The thick central portion is due

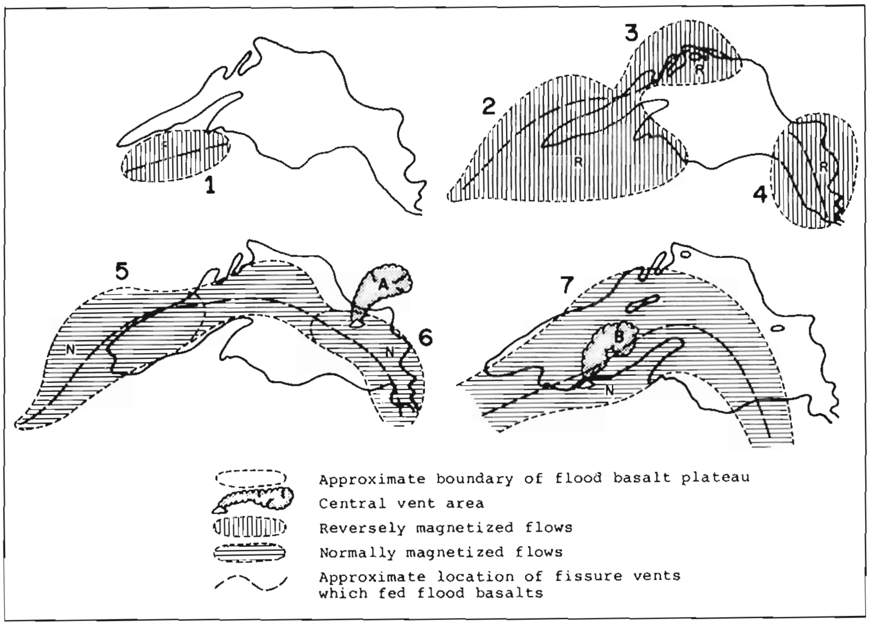


Figure 22.19

Sequential development of Keweenawan flood basalt plateaus (after Green, 1977). A and B represent central vent-type accumulations at Michipicoten Island and in the Porcupine Mountains respectively.

in part to the nature of the sheeted dyke complex and in part to the great accumulation of dense volcanic rock along the axis of the rift, which must have required substantial isostatic readjustment.

Keweenawan trends generally transect Middle Precambrian and Archean structural features, especially around the western end of Lake Superior. Sims (1976) pointed out that a major offset in Archean stratigraphy occurred across the Keweenawan belt in the area south of Duluth, where the contact between supracrustal and granite-gneiss terranes is displaced approximately 160 km. In the same area, Halls (1978) reported a similar offset in gravity features related to Middle Precambrian lithologies.

The areal distribution of reversely (older) and normally (younger) magnetized flows can also be interpreted to support the rift theory of origin for the Keweenawan in that the older flows invariably occur marginal to the younger volcanics (Fig. 22.4). This pattern appears to be at least partly due to differences in the sites of the volcanic activity rather than just the result of later downwarping which produced the Lake Superior Syncline. As cited earlier, flow direction data and the distribution of diabase dyke swarms which probably acted as feeders to the reversely magnetized flood basalts, indicate that those older flows erupted from many subparallel fissure-dykes widely spaced over a large area, at least 150 km wide. In contrast, feeders for the overlying normally magnetized flows were all located in a narrow zone along the axis of the lake. The formation of new crust, in the form of sheeted dykes, along this axis moved the pre-existing reversely magnetized volcanics to their present peripheral position.

The style of volcanism in the Keweenawan has been compared by many workers to that in the active Mid-Atlantic ridge. Flow morphology, stratigraphic relationships and petrochemistry show remarkable similarities to the Tertiary rocks of Iceland (Annells, 1974; Green, 1972; Walker, 1964). The vast majority of flows in the two areas are virtually indistinguishable in terms of thickness, geometry and internal structures; even their present mineralogies modified by deuteric alteration and low grade metamorphism are broadly comparable. Overlapping accumulations of basalt sheets, roughly elliptical in outline, occur in a random pattern along the Mid-Atlantic rift zone, the result of sporadic volcanism which shifts and varies in intensity with time.

A multiple lava basin hypothesis for the Keweenawan volcanics was first proposed by White (1972) and supported by Green (1977). Instead of a single basalt plateau, Green determined that there had been at least seven temporally and/or spatially distinct basins of flood basalt accumulation. The shapes of these basins (see Fig. 22.19) were inferred from outcrop and borehole information which allowed Green to extrapolate thinning of lava piles toward their original margins, and from aeromagnetic and seismic data which were used to define ridges in the Archean basement that could have acted as lava barriers.

Significant lithologic and petrochemical differences between relatively closely spaced areas of Keweenawan exposure such as the North Shore volcanics, the Osler Group and the Isle Royale lavas, are explained by the multiple basin hypothesis; each of these areas form part of a separate plateau (see Fig. 22.19) (Green, 1977).

In Iceland, shield and cone-type volcanic centres have been located, surrounded by and buried within the flood basalt plateaux (Walker, 1964). So far only two such centres are known in the Keweenawan (Annells, 1974; White et al., 1971) but given the relatively small percentage of the Keweenawan presently exposed, many more probably exist.

The petrochemical characteristics of Keweenawan volcanics are similar to the world's major flood basalt provinces, including the Tertiary of Iceland, in that they are predominantly tholeiitic, show pronounced bimodality in most areas, and through most of their section show limited variability. The limited published trace element data for mafic rocks (McIlwaine and Wallace, 1976; Annells, 1974) for example plots of Ti-Zr-Y and Ti-Zr-Sr plots, show the Keweenawan rocks to be similar to modern "within-plate" tholeiitic basalts (Pearce and Cann, 1973). Major element data from several areas (Green, 1972; Annells, 1973, 1974; Wallace, 1972), when plotted on AFM and other variation diagrams, have shown strong similarities to trends derived from the Thingmuli suite from Iceland (Carmichael, 1964).

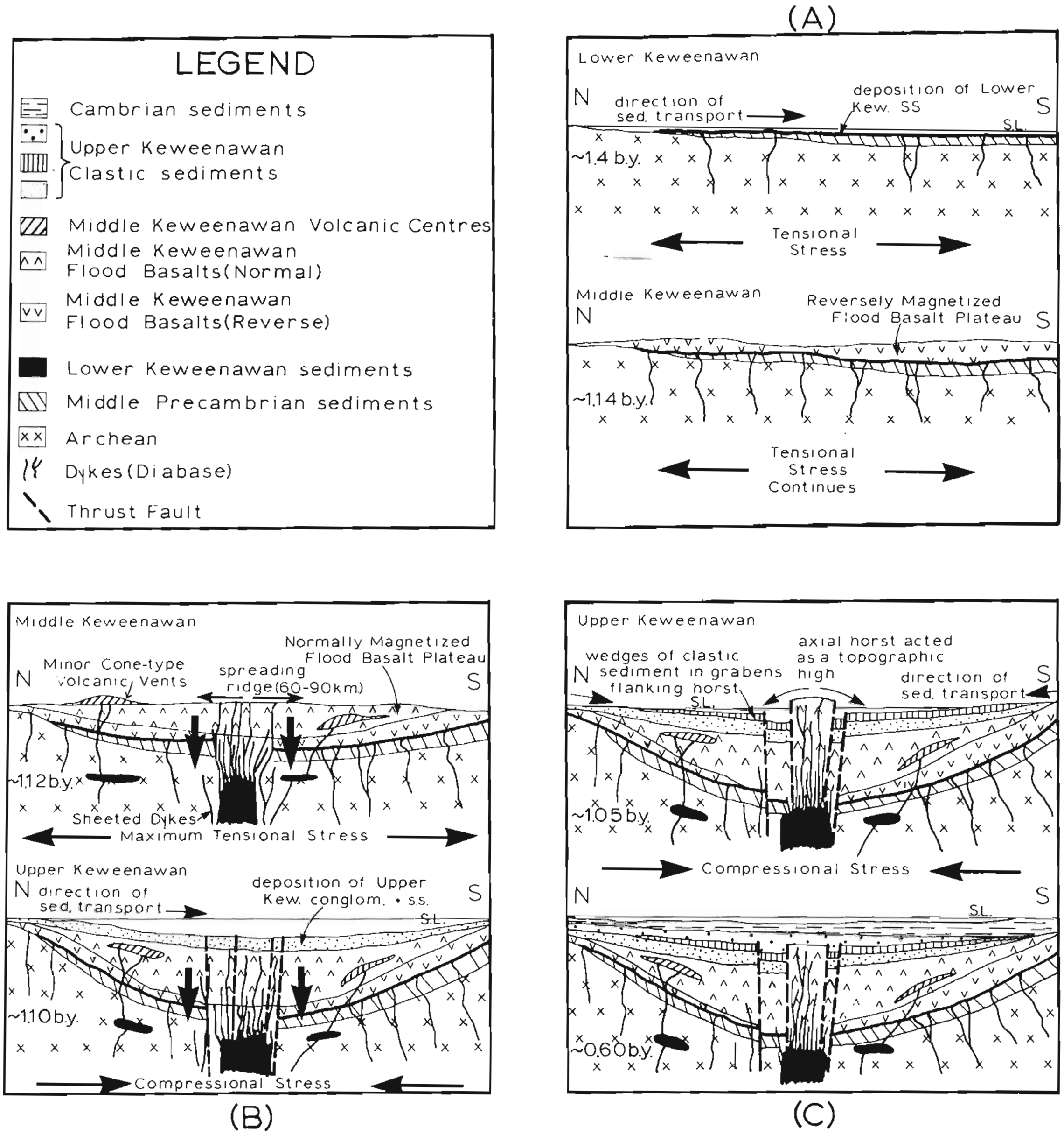


Figure 22.20. *The development of the Keweenawan Rift System and the evolution of the Lake Superior Basin.*

Green (1977) estimated the rate of plate separation along the Keweenawan rift system at roughly 0.5 cm per year. This compares with average rates of from 1.5 to 2.2 cm per year along the Mid-Atlantic Ridge (Strangway, 1970) and is at least one to two orders of magnitude greater than that for the East African Rift System. The wide discrepancies in the rates of distension for these three rift systems have been cited to account for their major differences (White, 1972; Green, 1977). The East Africa Rift is essentially a graben structure, whereas the Lake Superior Basin was formed by a downwarping of the crust. Volcanic activity in the East Africa Rift is very much subordinate to sedimentation, and most of the small stratovolcanoes consist of extremely alkalic rocks, in sharp contrast to the tholeiitic flood basalts of the Keweenawan. White (1972) suggested that relatively fastrates of rifting, such as in the Keweenawan and Mid-Atlantic systems, would produce deep crustal fissuring and allow for rapid movement of magma to the surface. Slower spreading rates would allow strain to be taken up by normal faulting with consequently less dyke formation and volcanic activity. Green (1977), in comparing the Keweenawan plateaux with those of Iceland, attributed the much greater thickness of the former to "abrupt periods of intense eruptive activity, followed by quiescence...with relatively little spreading". In Iceland, the spreading continues at a relatively constant rate which gives rise to thinner plateaus produced from closely adjacent feeder systems. Green (1977) also explained the much thinner crust under the Icelandic rift zone in terms of more rapid spreading.

The development of the Keweenawan can also be compared to and contrasted with the Newark tectonic-volcanic event (de Boer and Snider, 1979) of eastern North America. During the Triassic, crustal rifting with limited spreading (∽100 km) (Rodgers, 1970) took place along the axis of what is now the Appalachian Mountains. This aborted rifting preceded the actual opening of the Atlantic by approximately 50 Ma. The rate of plate separation during the Newark event is believed to have been relatively slow since it resulted in the formation of deep rift valleys bounded by en echelon sets of normal faults. As in the East Africa Rift, redbed fluvial and piedmont fan sedimentation was dominant. Most volcanic activity was alkalic, but major tholeiitic intrusions and relatively minor flood basalt accumulations formed in the northern part of the rift zone from Pennsylvania to Nova Scotia (de Boer and Snider, 1979). As in the Keweenawan, major dyke swarms parallel the graben systems, and extend far beyond the Triassic basins. A major geophysical feature known as the Appalachian Gravity High is roughly congruent with these linear basins. As in Green's (1977) interpretation of Keweenawan volcanism, the Newarkian igneous and tectonic activity shifted in time and space along an active rift arm of a triple junction located in northern Florida (de Boer and Snider, 1977) and formed a number of discrete en echelon elliptical basins.

SUMMARY

The Lake Superior Basin and the Keweenawan rocks within it formed as a result of crustal rifting and plate separation which took place between 1.3 and 0.6 Ga. The rifting cycle began (Fig. 22.20A) during a prolonged period of stable platform, shallow marine sedimentation (Lower Keweenawan) with the formation of extensive dyke swarms parallel to the axis of the incipient rift. With continued crustal distention a series of deep subparallel fissures brought magma rapidly to the surface producing a small number of overlapping plateaus each consisting of hundreds of reversely magnetized subaerially extruded flood basalts. After a short hiatus in volcanic activity, during which deposition of some terrestrial sediments took place in local basins, volcanism recommenced with the eruption of voluminous normally magnetized flood basalts from feeders located along the axis of the rift where crustal spreading of up to 90 km took place (Fig. 22.20B). As the mass of mafic flows rapidly increased along the axial zone, regional subsidence took place which more or less kept pace with the volcanic accumulation. This downwarping continued after volcanism waned, producing the Lake Superior Syncline into which clastic sediments (Upper Keweenawan) were introduced from all sides. At roughly the same time, the tectonic regime in the region changed to one of compression which led to a steepening of the limbs of the syncline, and the development of a system of high angle reverse faults along the axis of the depression. Vertical movement on these faults in the order of 2 to 4 km formed an extensive horst-and-graben system with a prominent central horst which for some time acted as a topographic high providing detritus for the flanking troughs (Fig. 22.20C). Sedimentation into the basin which roughly coincided with the present outline of Lake Superior continued into Cambrian time with only minor reactivations of the older faults.

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23\. PRECAMBRIAN FOSSILS IN CANADA – THE 1970s IN RETROSPECT

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Abstract

At the time of publication of the last inventory of reported Precambrian remains in Canada (Hofmann, 1972), 72 occurrences were known, including 31 of stromatolites, 3 of microfossils, 3 of megafossils, and 19 of dubiofossils, the remainder being demonstrably inorganic (pseudofossils), or younger than Precambrian. Among the important published finds of Precambrian remains made in Canada since 1970 are the following: stromatolites in the Archean Yellowknife Supergroup; stromatolitic microbiotas in the Aphebian Belcher Supergroup of Hudson Bay; carbonaceous megafossils, microfossils, and dubiofossils in the Neohelikian Little Dal Group of the Mackenzie Mountains; microfossils in the Helikian Dismal Lakes Group northeast of Great Bear Lake; Hadrynian trace fossils and microfossils in the Rocky Mountains, and microfossils in 11 Hadrynian formations on the Avalon Peninsula. Stromatolites, as yet inadequately dated but Proterozoic, were reported from 15 new areas and units. In addition, earlier known occurrences received more detailed study and taxonomic treatment, particularly microfossils and stromatolites in Aphebian formations in the Animikie, Great Slave, Epworth, and Belcher basins. Furthermore, some of the 19 remains earlier considered as dubiofossils were reinterpreted as biogenic.

These developments were paralleled by similar ones elsewhere on the globe. A striking increase in interest in Precambrian paleontology has resulted in many publications on all types of remains; several syntheses have appeared, and working hypotheses have been updated to accommodate the new data. The oldest acceptable fossils now known are stromatolites and microfossils from the 3.5 Ga old Warrawoona Group of Western Australia; "microfossils" reported from the 3.8 Ga old Isua Group of southwestern Greenland, are probably nonbiogenic.

Résumé

Lors de la publication du dernier inventaire des restes précambriens signalés au Canada (Hofmann, 1972), on connaissait 72 manifestations, dont 31 de stromatolites, 3 de microfossiles, 3 de mégafossiles et 19 de fossiles douteux, les autres étant des fossiles inorganiques (pseudofossiles) ou des fossiles plus récents que le Précambrien. Parmi les données importantes publiées sur les restes précambriens faites au Canada depuis 1970, on compte des stromatolites dans le supergroupe archéen de Yellowknife; des microbiotes stromatolitiques dans le supergroupe aphébien de Belcher de la baie d'Hudson; des mégafossiles, des microfossiles, et des fossiles douteux charbonneux dans le groupe néohélikien de Little Dal des monts Mackenzie; des microfossiles du groupe hélikien de Dismal Lakes au nord-est du Grand lac de l'Ours; des traces de fossiles et de microfossiles dans 11 formations de l'Hadrynien sur la presqu'île d'Avalon. Des stromatolites non encore datés de façon précise mais qui doivent probablement être du Protérozoïque, ont été signalés dans 15 nouvelles régions et nouvelles unités. En outre, des manifestations déjà connues ont fait l'objet d'études et d'un traitement taxonomique plus détaillés, particulièrement les microfossiles et les stromatolites des formations de l'Aphébien dans les bassins d'Animikie, de Great Slave, d'Epworth et de Belcher. De plus, un certain nombre des 19 restes du Protérozoïque considérés antérieurement comme des fossiles douteux ont été réinterprétés comme étant d'origine biologique.

Parallèlement à ces développements on en a étudié d'autres similaires ailleurs dans le monde. L'intérêt accru porté à la paléontologie du Précambrien a donné lieu à de nombreuses publications sur tous les types de restes; plusieurs synthèses sont apparues et des hypothèses de travail ont été mises à jour pour s'adapter aux nouvelles données. Les fossiles acceptables les plus anciens maintenant connus sont des stromatolites et des macrofossiles du vieux groupe de Warrawoona (3, 5 Ga) de l'ouest de l'Australie; les "microfossiles" signalés dans le vieux groupe d'Isua (3, 8 Ga) du sud-ouest du Groenland ne sont probablement pas d'origine biologique.

INTRODUCTION

During the past decade Precambrian paleontology has continued to grow as a subdiscipline, with many geologists in different parts of the world contributing to this active field, and making discoveries almost anywhere Precambrian sedimentary rocks are accessible. Results of this blossoming activity are reflected in an expanding literature dealing with new finds, as well as with the practical application of the data to Precambrian biostratigraphy and paleogeography. (For general reviews see Schopf, 1975a, 1977). It has become difficult for most geologists to keep up with these developments; this overview presents progress in this field in Canada during the past decade, and brings previous inventories (Hofmann 1971a, 1972) up to date. Altogether 50 new occurrences are registered in the present compilation, and new information is provided for 22 previously known occurrences.

TYPES OF REMAINS CONSIDERED

Precambrian remains can be conveniently classified under 9 headings (Fig. 23.1). Megafossils (body fossils) are the remains of macroscopic organisms, of which the Ediacaran metazoans are the best known examples; included in this category are also macroscopic carbonaceous compressions with uniform size and geometry. Microfossils comprise structurally preserved organisms, chiefly algae and bacteria, visible with the aid of the optical and electron microscopes. Chemofossils are those biogenic remains of ultramicroscopic size identifiable only by their molecular configuration, stable isotope ratios, or composition, that is, by chemical methods (Fig. 23.2); they include the various geochemical alteration products of biogenic substances such as carbon, kerogen, and the porphyrins and isoprenoid alkanes formed by the degradation of chlorophyll (Fig. 23.3), as well as isotopically light $\delta^{13}C$ and $\delta^{34}S$ values indicative of metabolic processes. Ichnofossils are the tracks, trails, and burrows left by mobile animals. Stromatolites are fixed, laminated structures that represent the traces of life-activities of micro-organisms in integrated communities, mainly of algae and bacteria. Oncolites are detached, concentrically laminated stromatolites which can be moved repeatedly on the substrate by currents and waves. Catagraphs represent internally unlaminated detrital grains believed to have formed under the influence of micro-organisms; many of them are microscopic dubiofossils, some are pseudofossil intraclasts. Dubiofossils are structures of undetermined or uncertain origin; they may or may not be biogenic; available evidence is inconclusive for their classification as true fossils or as pseudofossils. Pseudofossils are not evidence of life; they are those remains that have a deceptive resemblance to fossil organisms, but can be shown to be entirely due to nonbiologic processes.

Of the 50 new occurrences (some, in reality, multiple localities), 2 are megafossils, 14 are microfossils, 1 is an ichnofossil, 15 are stromatolites and oncolites, 5 are chemofossils, and 13 are dubiofossils (including catagraphs) and pseudofossils. Each of these is further discussed separately. Summaries of developments in Canada for the period 1970-1980 are given in Figure 23.4 and in the Appendix, where specific references are cited for each occurrence; these are also keyed into the bibliography.

MEGAFOSSILS

The only body fossils known in 1970 were the Hadrynian metazoans in the Mistaken Point Formation of southeastern Newfoundland (Occurrence 49), discovered by P. Thompson and S.B. Misra in 1967. While this fauna has since been found at many different localities on the Avalon Peninsula (Anderson 1978, Fig. 1), no detailed systematic or taxonomic study has so far been published, an unfortunate situation, considering the great importance of this assemblage in Precambrian paleontology. This fauna is dominated by problematic coelenterate genera, including various medusoids as well as leaf-like colonies reminiscent of structures found in the Charnian of England; genera thus far identified are **Charnia, Charniodiscus,** and **Cyclomedusa.**

New occurrences of large medusoids similar to those in the Mistaken Point Formation have also been found in the correlative Hibbs Hole Formation, and the older Briscal and Drook Formations (Occurrence 119) (Hofmann et al., 1979, p. 83, 85). These, too, require systematic study.

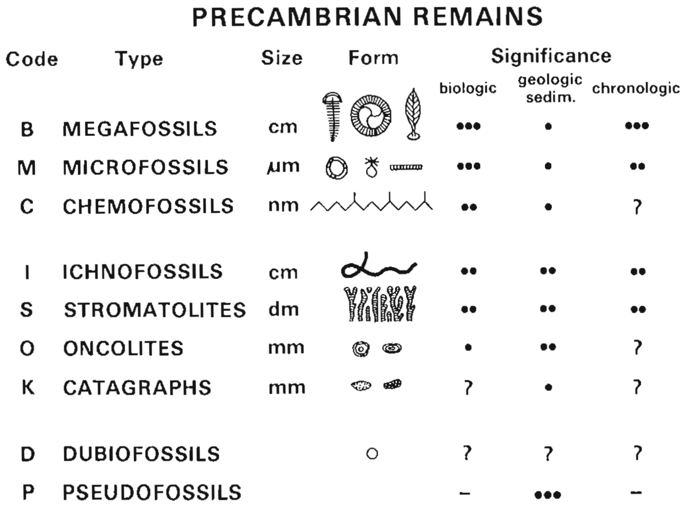


Figure 23.1. Classification and attributes of Precambrian remains. The frequency of dots in the 3 columns on the right indicates the relative quality of information provided by the remains regarding biology, geological conditions and processes, and geochronology.

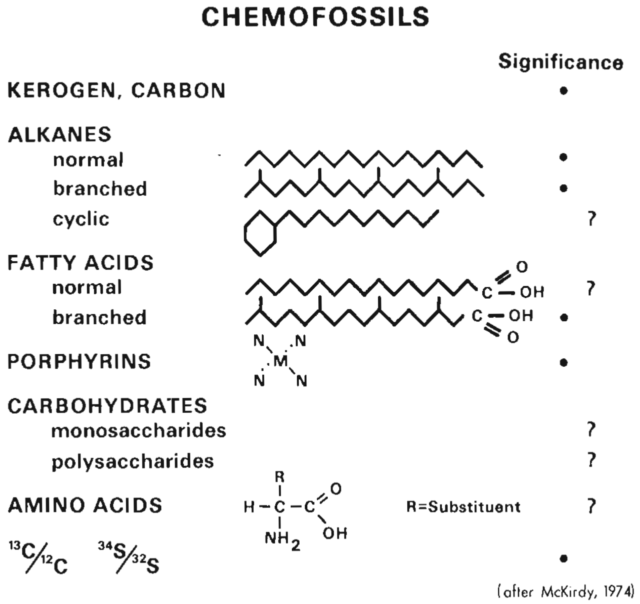


Figure 23.2. Some chemical evidence of life-activity found in Precambrian sediments. The biological significance of the remains is given in the column on the right.

A new biota was discovered by J.D. Aitken in 1976, in the Helikian Little Dal Group of the Mackenzie Mountains (Occurrence 85). This comprises 4 types of carbonaceous compressions: the discoidal **Chuaria circularis**, the ovate **Morania? antiqua**, the ribbon-like **Tawuia dalensis**, and irregular, angulate fragments assigned to **Beltina danai**. This biota is preserved in a finely laminated, basinal dolostone facies, off the foreslope of stromatolite reefs. It is of exceptional interest and evolutionary significance because **Tawuia**, interpreted as an eucaryotic alga, is the largest 1 Ga old organism known, the ribbons attaining as much as 6 mm in width and 150 mm in length. The **Chuaria** found in great abundance in association with **Tawuia** suggests these two taxa may possibly represent an alternation of generations of the same organism. Studies are under way to determine their affinities and possible relations more closely.

The Little Dal fossils, as well as new occurrences of **Chuaria circularis** in the Uinta Mountains, have made it possible to transfer the Hadrynian dubiofossils of Occurrence 24 (Hofmann, 1971a, p. 24, 1972, p. 24) to the body fossils; they support the interpretation of Gussow (1973).

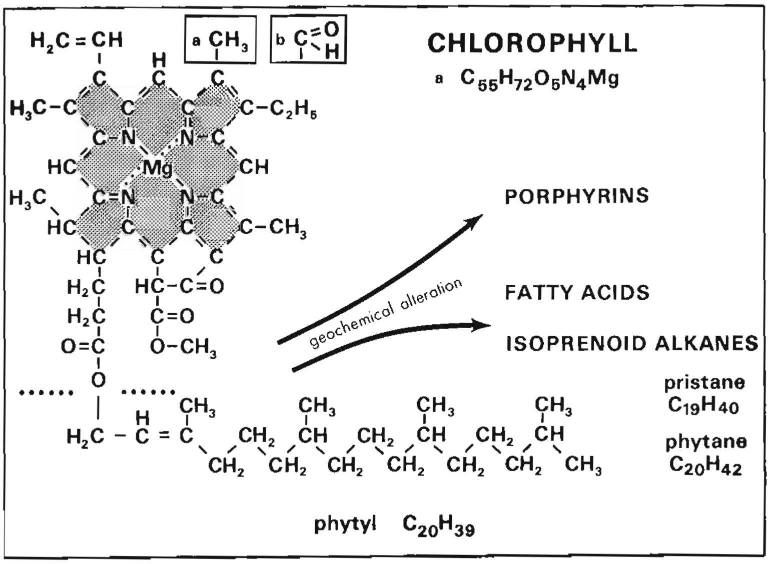


Figure 23.3

Diagram showing the production of some chemofossils. The chlorophyll molecule chosen as an example has bipartite organization; during geochemical alteration the molecule may split along the dotted line, the "head" yielding porphyrin complexes, and the "tail" yielding isoprenoid alkanes or fatty acids. These breakdown products, which may be preserved in the fossil record, are of molecular dimensions and identifiable only by chemical methods.

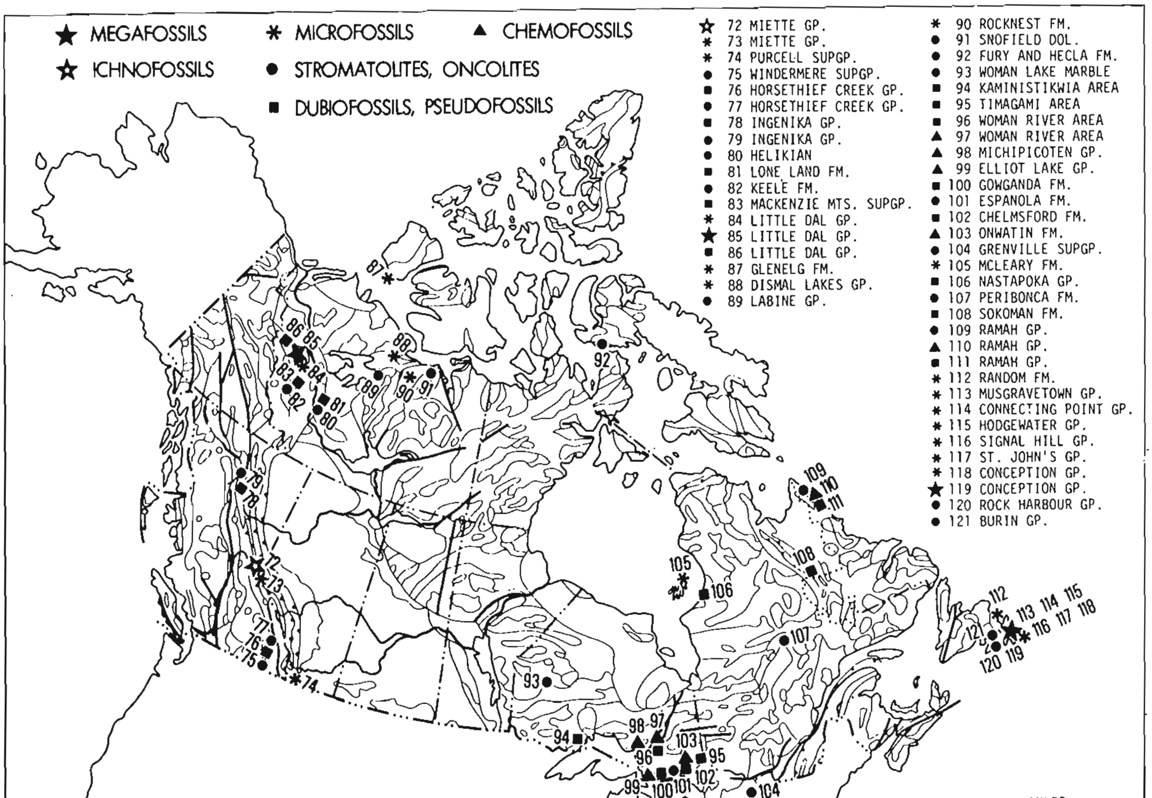


Figure 23.4. *New occurrences of Precambrian remains reported from Canada during the decade 1970-1980.*

Microfossils

Knowledge of Canadian Precambrian microfossils improved considerably in the 1970s. In addition to new data and interpretations of known assemblages (Occurrences 25, 31, 71), 14 new occurrences were reported (73, 74, 84, 87, 88, 90, 105, 112-118), ranging in age from Aphebian to Late Hadrynian, and including both benthic stromatolitic chert-facies and planktonic shale-facies microfossils.

The chert-facies microfossils have proved to be the most diverse taxonomically and the most important with regard to providing paleobiologic and paleoecologic information. Their early diagenetic in situ permineralization by silica has resulted in exceptional fidelity of preservation, permitting the study of life-cycles of individual taxa, and of relationship of elements in ancient microbial communities in their ecologic context. Many microfossils are morphologically similar to modern ones in similar peritidal settings, and the pronounced evolutionary conservatism exhibited by this group of organisms has made it possible to study the biology, ecology, and taphonomy of the modern analogues to be in a better position for evaluating the ancient ones. The most thoroughly studied Canadian occurrences in this respect now are the Aphebian Gunflint Formation (Occurrence 31) and the Belcher microfossils (Occurrences 71, 105), and the Helikian Dismal Lakes Group (Occurrence 88) (for references see Appendix). All these assemblages are dominated by coccoid and filamentous cyanobacteria (cyanophytes). Altogether 21 taxa are recognized in the Gunflint Formation, excluding synonymous ones, another 16 in the Kasegalik Formation, 19 in the McLeary Formation, and 13 in the Dismal Lakes Group (see Appendix).

The Gunflint microbiota includes in addition to the prevalent coccoids and filaments, a number of bizarre forms that are likely planktonic and difficult to interpret for lack of convincing modern analogues. These include the genera **Eosphaera**, considered as a green alga (Kazmierczak, 1976, 1979), or a red alga (Tappan, 1976); **Eoastrion**, a bacterium, or a nonbiologic structure (Barghoorn, 1963, 1971, 1974, 1977), and **Kakabekia** (see references under B.Z. Siegel, S.M. Siegel,); presumed reproductive stages of **Huroniospora** were discussed in several papers (Darby, 1972, 1974), the role of **Huroniospora** and **Gunflintia** in the formation of some Gunflint stromatolites was treated in others (Awramik, 1973a, b, 1976a, b, 1977).

Studies of post-mortem degradation of modern procaryotes were applied in the interpretation of similar microfossils with internal "dark spots" (see references under Awramik, Barghoorn, Golubic, Hofmann), and experimental silicification of modern procaryotes have facilitated the interpretation of Precambrian microfossils generally (Oehler, 1976a, b).

Shale-facies microfossils comprise spheroids and filaments which are associated with the Helikian Little Dal megafossils (Occurrence 84), and which are also found in various Hadrynian units on the Avalon Peninsula in Newfoundland (Occurrences 112-118). Their size is generally larger than that of the chert-facies microfossils by an order of magnitude, and they are probably algae. **Bavlinella**-like colonial aggregates and isolated spheroids in the late Hadrynian Windermere Supergroup (Occurrences 25, 73) have been interpreted, not without question, as different ontogenetic stages in the life-cycle of planktonic algae (Cloud et al., 1975; Moorman, 1972, 1974; Javor and Mountjoy, 1976).

Chemofossils

Relatively few references were published on material considered to be chemical evidence of Archean and Aphebian life-activity. These deal with molecular fossils in the Gunflint Formation (Occurrence 30, 31) and with fractionated stable carbon and sulfur isotopes, the negative $\delta^{13}C$ and $\delta^{34}S$ values being ascribed to the effects of microbial metabolism (Occurrences 30, 31, 97, 98, 99, 103) Barghoorn et al., 1977; Goodwin et al., 1976; T.A. Jackson, 1971a, b; Schidlowski, 1979). Carbonaceous matter in a 5 cm seam was reported from the Aphebian Ramah Group (Occurrence 110).

Ichnofossils

Only one new report of definite trace fossils was made, the **Didymaulichnus miettensis** from the Hadrynian Miette Group (Young, 1972, Occurrence 72). These simple, smooth, gently curving bilobate trails parallel to bedding are about 15-20 mm wide and possibly are molluscan crawling trails.

Trace fossils previously reported from the Random Formation of Newfoundland (Occurrence 45) may in part be Lower Cambrian (Greene and Williams, 1974).

STROMATOLITES (INCLUDING ONCOLITES)

By far the most numerous entries in the bibliography refer to stromatolites. Newly reported occurrences include 2 from the Archean, 4 from the Aphebian, 3 from the Helikian to Early Hadrynian, and 6 from units considered to be Late Hadrynian. The 2 Archean finds (Henderson, 1975b; Thurston and Jackson, 1978: Occurrences 91 and 93) are the most important from a paleontologic point of view, considering that the only other locality known prior to 1970 was at Steeprock Lake (Occurrence 29). Together, these 3 comprise the oldest fossils known in Canada (Fig. 23.5), dating from the Late Archean. No microfossils have yet been found in these structures.

During the past decade geologists engaged in regional mapping and Proterozoic basin analysis made extensive use of stromatolites (see publications by Aitken, Campbell, Cecile, Donaldson, Hoffman, Jefferson, Long, G.M. Young). Results of concurrent work on modern analogues in Shark Bay, the Bahamas, Bermuda, and the Persian Gulf area, aided this analysis to no small extent (see publications by Gebelein, Golubic, Hoffman). The work on modern analogues, however, has, hardly touched the problem of the less accessible environment of subtidal stromatolites, the type most likely represented by the vast majority of Precambrian columnar stromatolites which are used biochronologically.

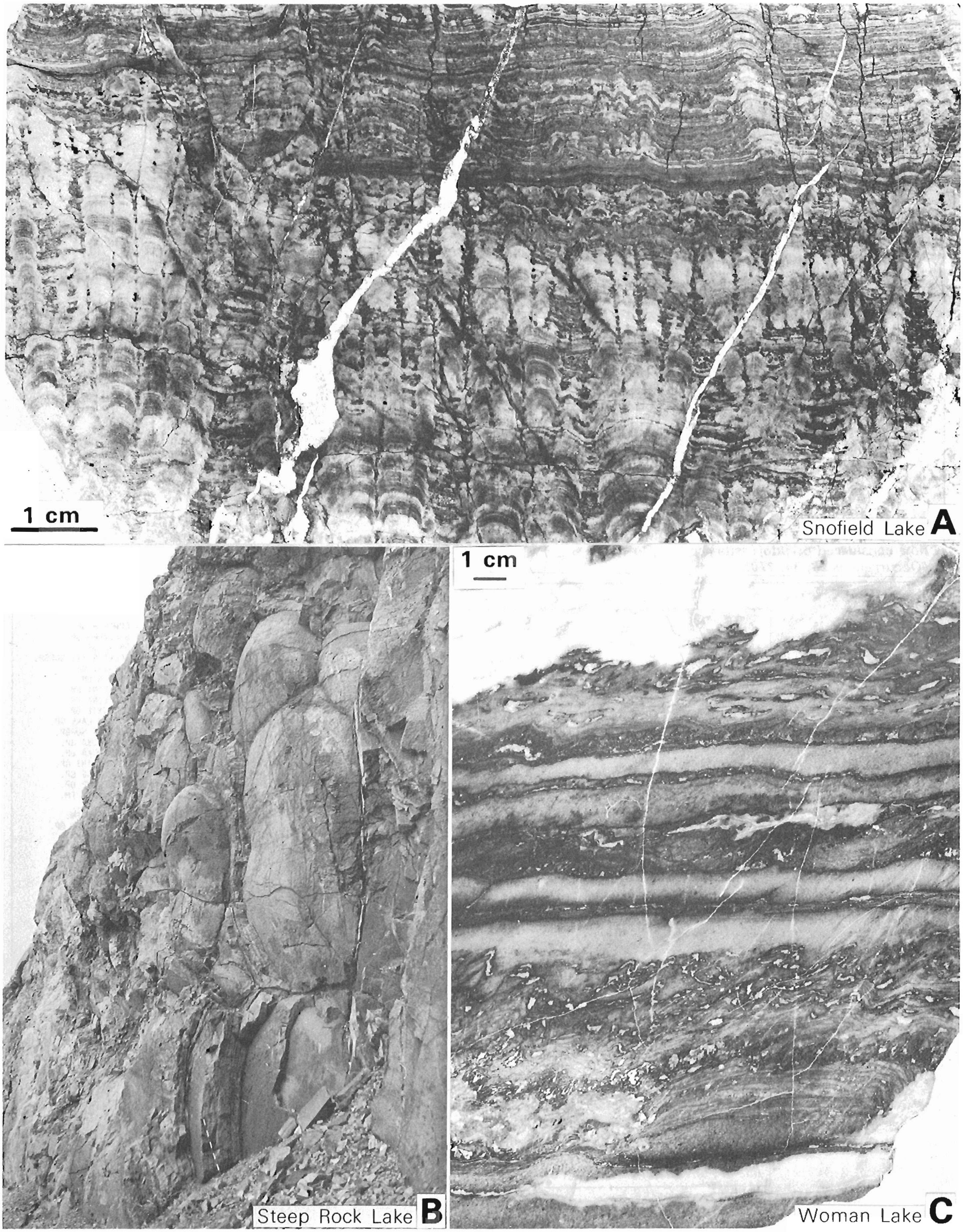
Figure 23.5

Canada's oldest fossils: Archean stromatolites from 3 localities.

A - Yellowknife Supergroup, Snofield Lake, District of Mackenzie (Occurrence 91). ***Pseudogymnosolen*** *and* ***Stratifera****. GSC hypotype 65650.*

B - Steeprock Group, Atikokan, Ontario (Occurrence 29). Metre stick for scale.

C - Unnamed carbonate unit at Woman Lake, northwestern Ontario (Occurrence 93; specimen collected by P.C. Thurston). GSC hypotype 65651. Orientation of specimen not known.


1 cm
Snofield Lake A
Steep Rock Lake B
1 cm
Woman Lake C

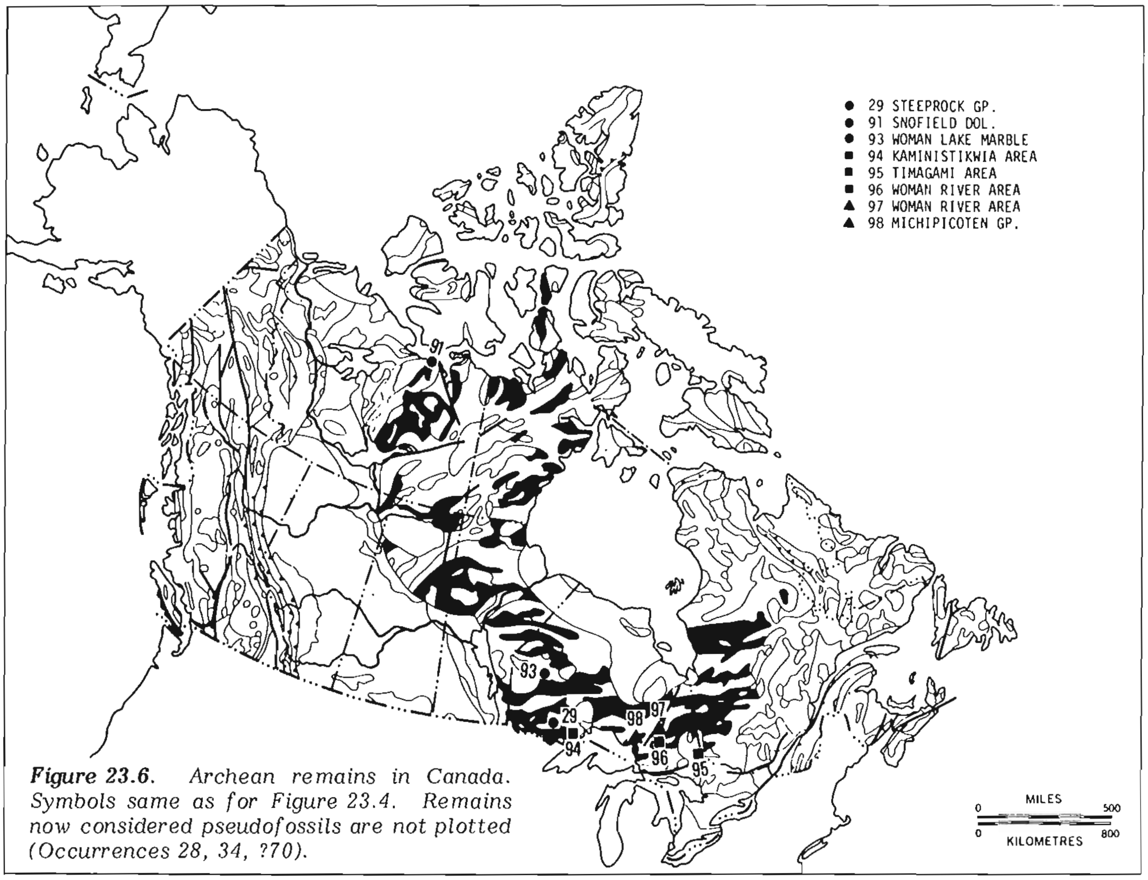


Figure 23.6. *Archean remains in Canada. Symbols same as for Figure 23.4. Remains now considered pseudofossils are not plotted (Occurrences 28, 34, ?70).*

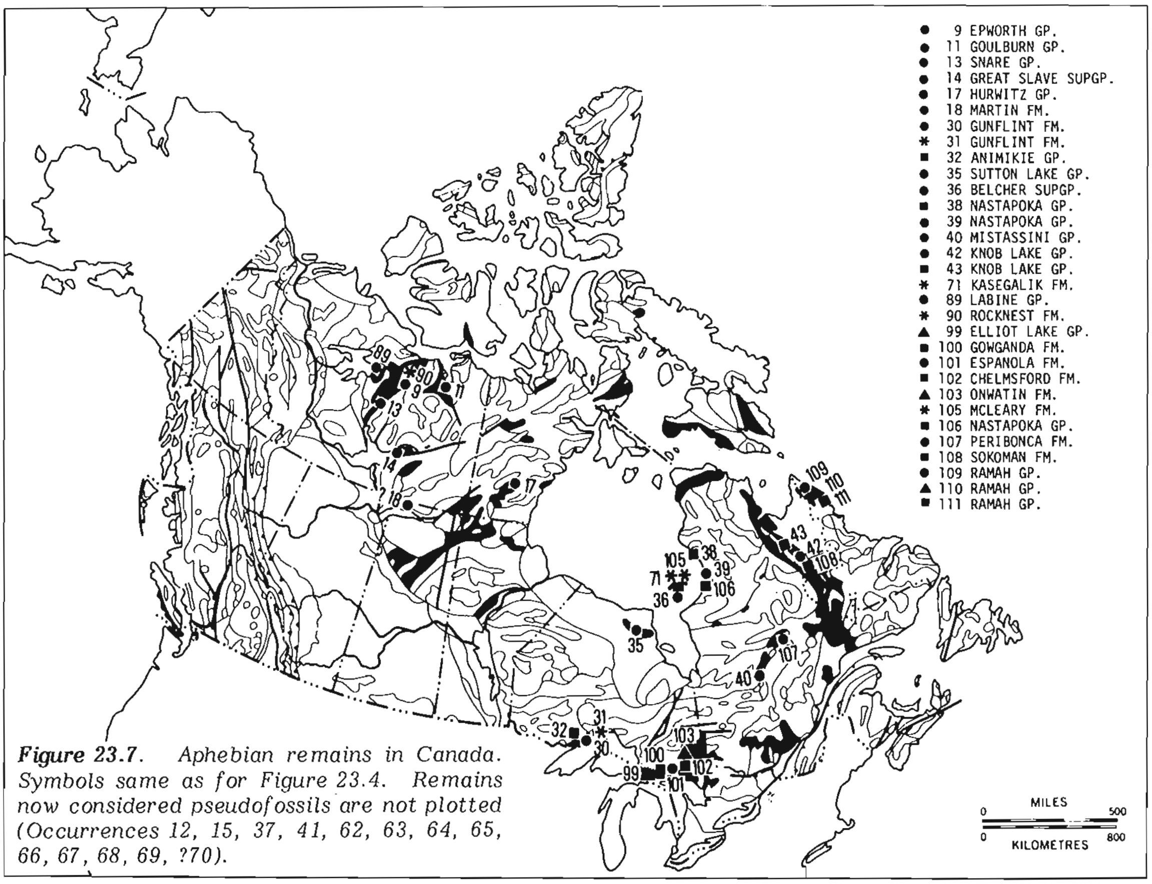


Figure 23.7. *Aphebian remains in Canada. Symbols same as for Figure 23.4. Remains now considered pseudofossils are not plotted (Occurrences 12, 15, 37, 41, 62, 63, 64, 65, 66, 67, 68, 69, ?70).*

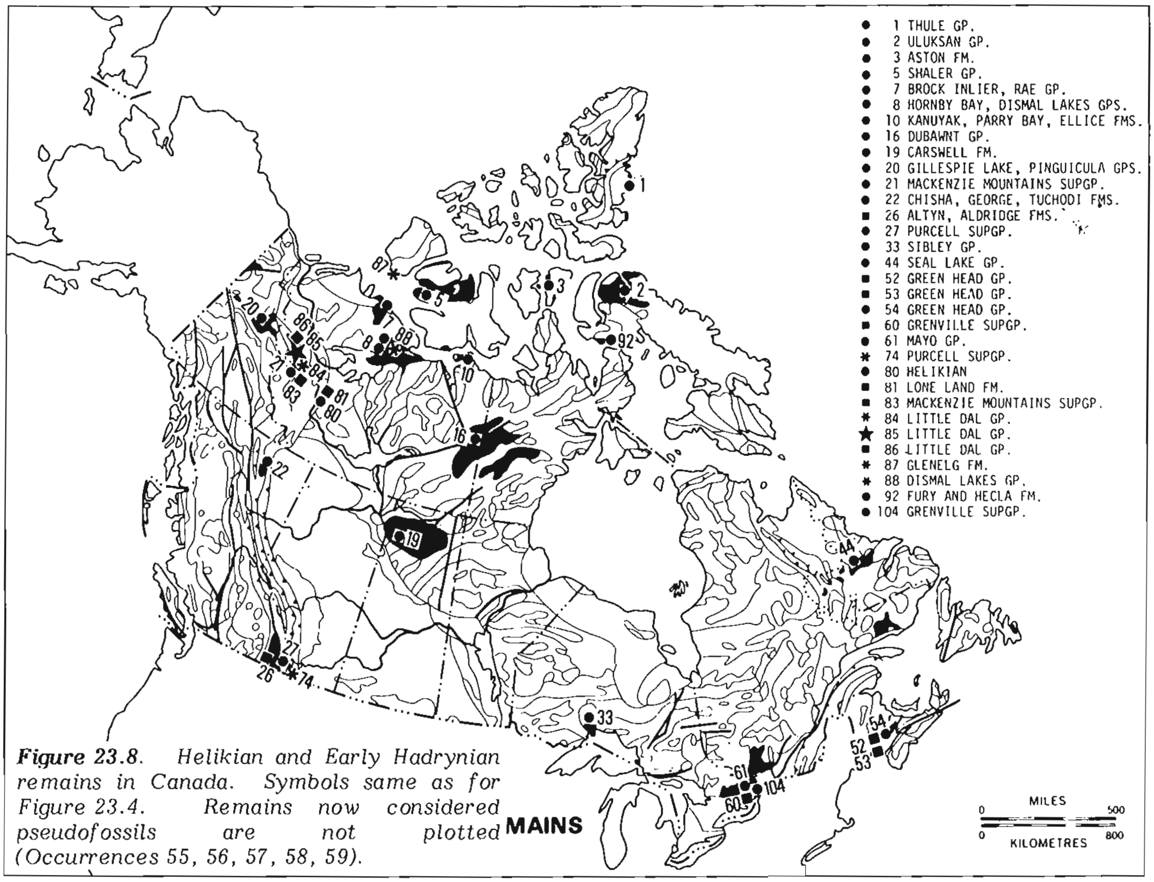


Figure 23.8. Helikian and Early Hadrynian remains in Canada. Symbols same as for Figure 23.4. Remains now considered pseudofossils are not plotted (Occurrences 55, 56, 57, 58, 59).

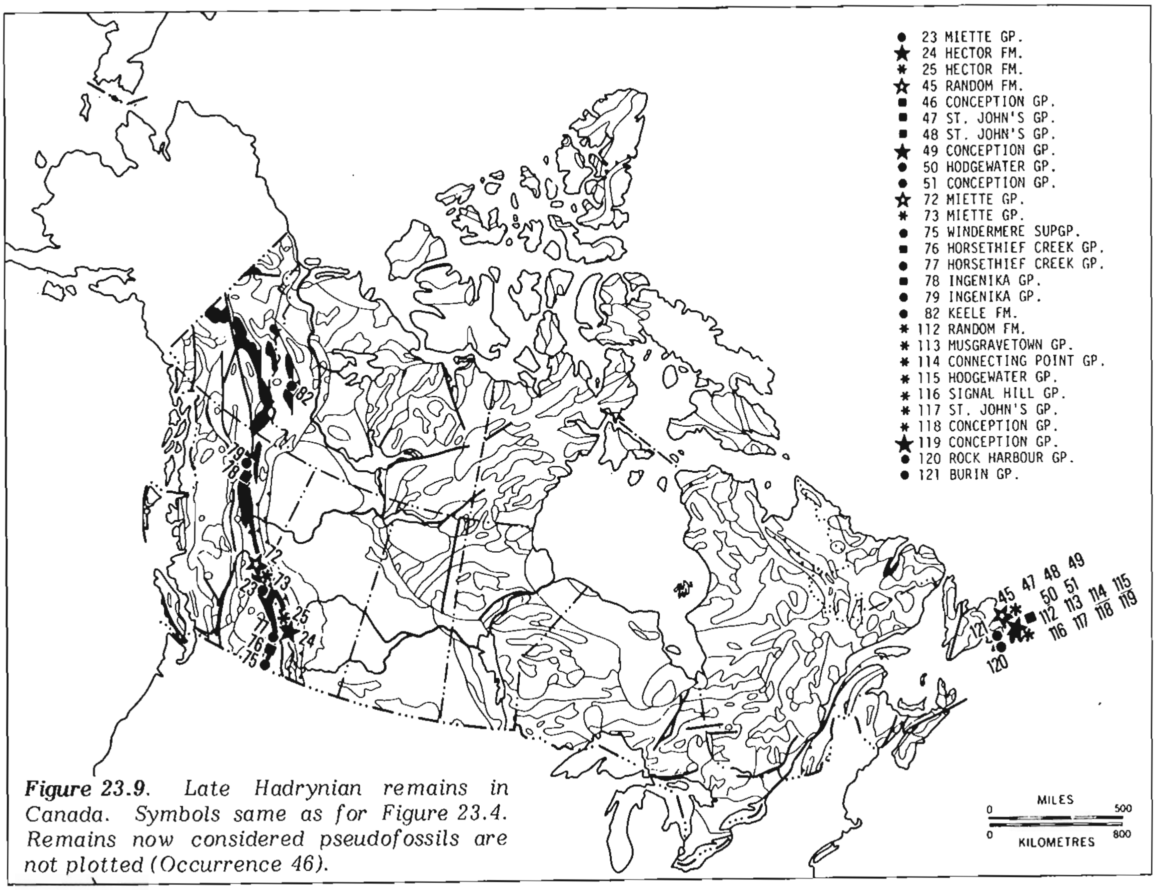


Figure 23.9. Late Hadrynian remains in Canada. Symbols same as for Figure 23.4. Remains now considered pseudofossils are not plotted (Occurrence 46).

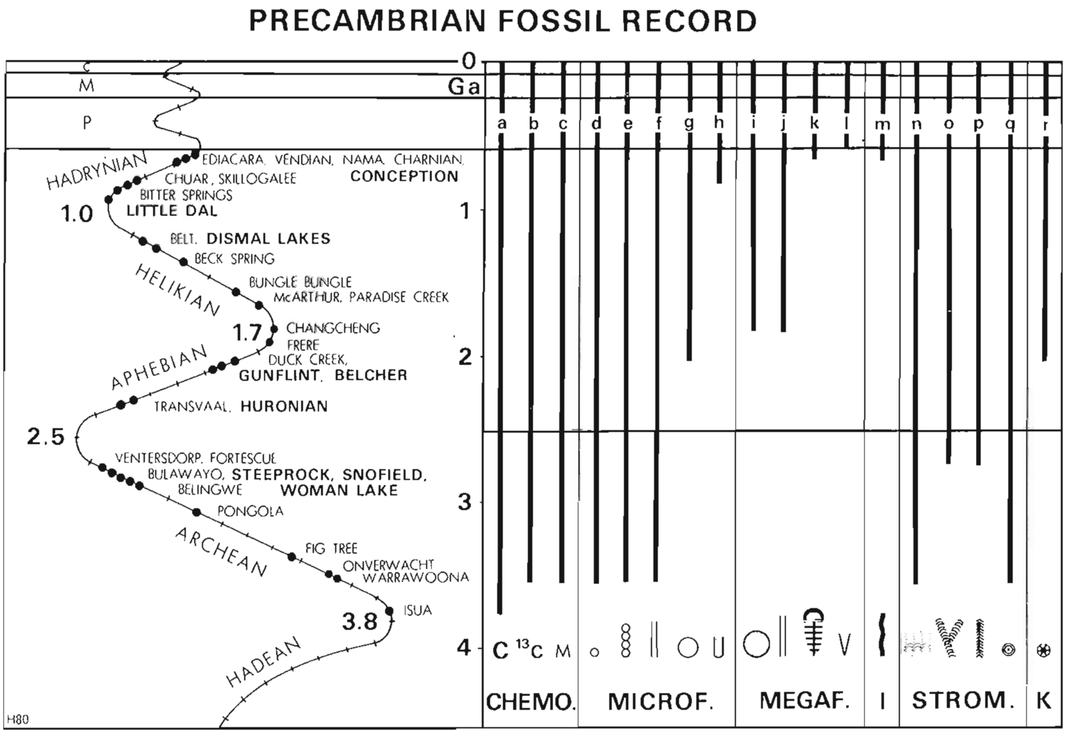


Figure 23.10. *Graphic summary of present status of Precambrian fossil record. Selected important fossiliferous stratigraphic units are placed in their geochronologic context; Canadian stratigraphic units are emphasized in boldface type. The attenuating sinuous curve schematically delineates major chapters in earth history and biological evolution. The stratigraphic range of different types of remains is shown in columns a-p:*

a	*- free carbon, kerogen*	*j*	*- megascopic carbonaceous ribbons*
b	*- fractionated isotopes of C and S*	*k*	*- metazoan body fossils*
c	*- molecular fossils*	*l*	*- shelly fossils*
d	*- coccoid microfossils < 30 µm*	*m*	*- ichnofossils*
e	*- filamentous microfossils*	*n*	*- stratiform stromatolites*
f	*- tubular microfossils, sheaths*	*o*	*- branching columnar stromatolites*
g	*- coccoid microfossils > 30 µm*	*p*	*- conophytonid stromatolites*
h	*- vase-shaped microfossils*	*q*	*- oncolites*
i	*- megascopic carbonaceous discs*	*r*	*- catagraphs*

References giving systematic and taxonomic treatment of stromatolites are few (Awramik, Hofmann, Jefferson, Semikhatov, Walter): 8 new stromatolite groups (one a junior synonym) and 16 new forms were erected using Canadian materials, all from Aphebian units (Occurrences 9, 14, 30, 40). Stromatolites were used for interbasinal correlation, particularly for Helikian and Early Hadrynian sequences in northwestern Canada (see references under Aitken, Jefferson, Semikhatov, G.M. Young).

Certain chert-hematite stromatolites from the Gunflint Formation at Mink Mountain, Ontario (Occurrence 30) formed the basis for studies of rhythmicity of the contained unusually fine and even lamination, to determine possible astronomic cycles in the Precambrian record (Pannella). An obtained minimum value of 448 days per year at the time of Gunflint deposition needs corroboration. The Mink Mountain stromatolites were also compared with laminated geyserite deposits in Yellowstone Park in the context of a nonbiologic origin previously suspected on the basis of the unusually fine and even lamination (Walter, 1972).

A new, quantitative method of studying stromatolites was introduced by H.J. Hofmann, using an image-analysis computer system that allows for more precise specification of attributes of these structures. Finally, many earlier known stromatolite occurrences were restudied, and new paleontologic, paleoenvironmental, and stratigraphic information became available.

DUBIOFOSSILS (INCLUDING CATAGRAPHS) AND PSEUDOFOSSILS

Prior to 1970 structures of problematic origin had occasionally been reported from Precambrian rocks. Many of these are now known to be entirely of inorganic origin and are assigned to the pseudofossil category. The concept of dubiofossils was introduced to accommodate the remaining structures probably or possibly having biological significance (Hofmann, 1972b, p. 27-29). New dubiofossils were reported from 13 places (see Fig. 23.4). About half of these are "possible trace fossils" (Occurrences 76, 78, 83, 86, 102, 111), and 5 are microscopic dubiofossils (Occurrences 94, 95, 96, 106, 108); all require more detailed study.

PRECAMBRIAN FOSSILS IN CANADA – PRESENT STATUS

Our present knowledge of Precambrian remains is summarized in the Appendix, and in a series of maps showing their geographic distribution by type and age, in conjunction with the outcrop belts of rocks of the same age (Fig. 23.6-23.9). Except for the Archean, most of the outcrop belts of unmetamorphosed Proterozoic sequences have yielded remains. The potential for finding additional occurrences in the Archean is good, and excellent for the Proterozoic. If the past is any guide, many new discoveries will be made fortuitously by geologists working in the field on general mapping or stratigraphic problems in Precambrian terranes. It will be useful to have some specialist input in the interpretation of such possible organic remains. What also still remains is a great amount of further work on occurrences already known.

GLOBAL CONTEXT – SUMMARY

The present status of the global Precambrian fossil record, which includes important contributions from Canadian localities, is graphically summarized in Figure 23.10. The oldest remains of free carbon on our planet (excluding meteorites) are in high grade metasedimentary rocks of the Isua Group in southwestern Greenland. Isotopic data and molecular remains from this unit yield no unequivocal information proving biogenic processes, and reported "microfossils" are irregular inclusions at grain boundaries. The oldest compelling evidence of life is found as stromatolites and microfossils in the Warrawoona Group of Western Australia and in the approximately coeval Onverwacht Group of South Africa.

Aside from the 3 Canadian occurrences, Archean stromatolites also are known from the Pongola, Belingwe, Bulawayo, and Ventersdorp units of southern Africa, and the Fortescue Group of Western Australia. The approximately 2.6-2.7 Ga old Canadian stromatolites are thus the only known Archean stromatolites from the Northern Hemisphere; they also include some of the oldest branching columnar forms anywhere.

On the Aphebian leg of earth history, many stromatolitic carbonates make their appearance, although only a few are shown in Figure 23.10, the oldest being in the Huronian and Transvaal supergroups. However, well documented and diverse microfossil assemblages do not occur until about 1.9 Ga ago, with the Gunflint and Belcher microbiotas constituting outstanding markers. Possible contemporaneous microfossiliferous units are the Frere and Duck Creek formations in Western Australia.

The earliest megascopic filaments and compressed spheroids are found in the Changcheng System of North China; assemblages of slightly larger dimensions are found in progressively younger deposits, such as the examples given in Figure 23.10. The Belt and Little Dal units also contain large ribbon-like carbonaceous compressions, veritable giants among Precambrian organisms.

Remains of the Hadrynian interval comprise various outstanding occurrences, including the well studied Bitter Springs microbiota, the Chuar Group of the Grand Canyon with the oldest known chitinozoan-like microfossils, and the Latest Precambrian Ediacaran fauna with its various equivalents, including the Conception Group metazoans of Newfoundland.

It is evident that Precambrian remains from Canada have figured prominently in the Precambrian fossil record, and will continue to do so in the future.

ACKNOWLEDGMENTS

I thank the many colleagues who have contributed photographs, samples, information, and discussion on Precambrian remains in Canada; their names are among those listed in the bibliography. I particularly appreciated receiving reprints of their papers, because this greatly facilitated the onus of compiling and assimilating the data, and made the bibliography more complete. The study was financially supported by research grant A7484 from the Natural Sciences and Engineering Research Council of Canada. W.A. Padgham (Department of Indian and Northern Affairs, Yellowknife) provided generous logistic support for a visit to Locality 91, where new stromatolite material illustrated in Figure 23.5A was obtained.

BIBLIOGRAPHY OF PRECAMBRIAN REMAINS IN CANADA – 2

This bibliography comprises 237 references specifically concerned with the paleobiology, biostratigraphy, paleoecology, or paleobiogeochemistry of Precambrian remains in Canada. It complements the last comprehensive bibliography on the same subjects (Hofmann, 1971a), and brings it up to date by including some 22 pre-1970 articles that were missed during the earlier compilation, and 7 references published in early 1980. Together these 2 bibliographies constitute the most complete collection of references on Canadian Precambrian paleontology yet published.

The references are coded by numbers and letters to facilitate their utilization. The number code refers to the occurrence as listed in the Appendix and as shown on Figures 23.1, 23.4, 23.6-23.9, continuing the numbering system employed in the previous compilations (Hofmann, 1971a, 1972). The letter codes refer to the stratigraphic position, type of remains, and emphasis of the paper with respect to the remains. The letter key is as follows:

Age:
- H - Holocene analogues of Canadian Precambrian remains
- Z - Late Hadrynian
- Y - Helikian and Early Hadrynian
- X - Aphebian
- W - Archean

Type of remains:
- B - Body fossils (megafossils)
- C - Chemofossils
- D - Dubiofossils
- I - Ichnofossils (trace fossils)
- K - Catagraphs
- M - Microfossils
- O - Oncolites
- P - Pseudofossils
- S - Stromatolites

Orientation of paper:
- a - applied to correlation, chronostratigraphy
- e - paleoenvironmental, paleogeographic, lithogenesis
- g - general treatment, synthesis
- r - incidental report of occurrence
- t - taxonomic treatment, systematics, paleobiological

Table 23.1. Composition of Bibliography 2

Year of publication	Number of references
early 1980	7
1979	17
1978	23
1977	25
1976	29
1975	31
1974	21
1973	23
1972	17
1971	14
1970	8
Total 1970-1980	215
pre- 1970	22
Grand total	237

Publisher	Number of post-1969 references	Percentage
Geological Survey of Canada	58	27.0 %
Other Canadian publications	40	18.6
Foreign publications	111	51.6
Theses at Canadian universities	3	1.4
Theses at American universities	3	1.4
Total	215	100.0
Most frequently cited:	Occurrence 31 (Gunflint Fm. microfossils)	

The composition of the bibliography is analyzed in Table 23.1. As can be seen, the publication rate reached a maximum during the mid-1970s. Of the total of 215 references listed for the years 1970 to early 1980, 98 (45.6%) appeared in Canadian publications, while 111-(51.0%) in foreign publications, chiefly American. By number of references, the leading Canadian publisher is the Geological Survey of Canada with 58 (27.0%); 43 (20%) are in Canadian periodicals or are unpublished theses. Most of the references are incidental reports of remains, mainly of stromatolites, in various parts of Canada, or are paleoenvironmentally oriented; relatively few provide detailed systematic and taxonomic treatment of Precambrian biotas; fewer still deal with chronostratigraphy and correlation based on the fossils.

The largest number of references (51) on a single occurrence relate to the Gunflint microfossils (Occurrence 31), indicating the sustained interest of Precambrian paleontologists in those remains, and their relative importance in the global context of Precambrian paleontology.

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APPENDIX

New data on occurrences of Precambrian remains in Canada

Table provides new paleontologic or stratigraphic information for certain remains among occurrences 1-72, previously cited in Hofmann (1971a, reverse of Fig. 1; 1972; Fig. 4), and lists occurrences 73-121, reported in the literature during the past decade (see Fig. 23.4). Symbols for age and type same as in bibliography.

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
1	Thule Gp.	Y?	S	Stromatolites	Christie et al., 1978, p. 380
					Frisch et al., 1978, p. 137
2	Uluksan Gp.	Y			
	Elwin Fm.		S	Stromatolites	Jackson et al., 1980, p. 327
	Strathcona Sound Fm.		S	Stromatolites	Jackson et al., 1978, p. 10, 11
	Athole Point F.		S	Stromatolites	Jackson et al., 1975, p. 9, 24
					Jackson et al., 1980, p. 325
	Victor Bay Fm.		S	Stromatolites	Geldsetzer 1973b, p. 119-121
					Jackson et al., 1975, p. 4, 9
					Jackson et al., 1978, p. 9, 10
					Jackson et al., 1980, p. 325
	Society Cliffs Fm.		S	Stromatolites	Geldsetzer 1973b, p. 111-117
					Jackson et al., 1975, p. 4, 9, 16, 17, 23, 25
					Jackson et al., 1978, p. 7, 8
					Iannelli, 1979 p. 45, 55
					Jackson et al., 1980, p. 324
	Arctic Bay Fm.		S	Stromatolites	Jackson et al., 1975, p. 4
					Jackson et al., 1978, p. 5
					Iannelli 1979, p. 46, 53, 54
					Jackson et al., 1980, p. 321, 324
	Fabricius Fiord Fm.		S	Stromatolites	Jackson et al., 1978, p. 7
					Iannelli 1979, p. 46, 51, 52
5	Shaler Gp.	Y			
	Kilian Fm.		S	Stromatolites	Aitken et al., 1978b, p. 486
	Wynniatt Fm.		S	**Baicalia**	Aitken et al., 1978b, p. 486
			S	**Boxonia**	
			S	cf. **Gymnosolen**	
	Reynolds Point Fm.		S	cf. **Acaciella**	Jefferson 1977, p. 79-84;98-219
			S	**Baicalia**	Jefferson and Young 1977, p. 26
			S	**Boxonia**	Young and Long 1977b, p. 2259
			S	cf. **Gymnosolen**	Aitken et al., 1978b, p. 486
	Glenelg Fm.		S	**Baicalia burra**	Young and Jefferson 1975, p. 1137
			S	**Basisphaera irregularis** ?	Jefferson and Young, 1976, p. 63
					Jefferson 1977, p. 79-84; 98-219
			S	**Conophyton**	Jefferson and Young 1977, p. 26
					Young and Long 1976, p. 2257
			S	**Inzeria**	Aitken et al., 1978b, p. 486
7	Brock Inlier	Y			
	Unit P_4		S	**Baicalia**	Aitken et al., 1973, p. 23
					Young 1977b, p. 1781
					Aitken et al., 1978b, p. 486
	Unit P_2		S	**Baicalia**	Aitken et al., 1973, p. 23
			S	**Colonnella**	Aitken et al., 1978b, p. 486
	Rae Gp.	Y	S	**Conophyton**	Baragar and Donaldson 1973, p. 11
					Donaldson 1976b, p. 532-534
8	Dismal Lakes Gp.	Y	S	**Conophyton**	Baragar and Donaldson 1973, p. 8
			S	cf. **Stratifera**	Donaldson 1976b, p. 526-531
					Schopf 1977, p. 150
					Kerans and Donaldson 1979, p. 456
	Hornby Bay Gp.	Y	S	Stromatolites	Baragar and Donaldson 1974, p. 7
9	Epworth Gp.	X			Hoffman 1976, p. 605

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
	Rocknest Fm.		S	**Asperia aspera** (= **Pseudogymnosolen**)	Semikhatov 1978, p. 120-122, 128-129, 133-138, 140-143
			S	**Colonnella**	
			S	**Conophyton minusculum**	
			S	**Confunda confuta**	
			S	**Discorsia discorsia**	
			S	**Kussoidella** f. indet.	
			S	**Olenia** (= **Pseudogymnosolen**)	
			S	**Omachtenia**	
			S	**Pilbaria perplexa**	
			S	**Stratifera laxa**	
10	Kanuyak Fm.	Y	S	Stromatolites	Campbell 1979, p. 4
	Parry Bay Fm.	Y	S	cf. **Collenia**	Campbell 1978, p. 100
			S	cf. **Conophyton**	Campbell 1979, p. 4, 10-17
			S	cf. **Tungussia**	
	Ellice Fm.	Y	S	Stromatolites	Campbell 1979, p. 5, 9
14	Great Slave Supgp.	X			
	Et-then Gp.		S	**Externia externa**	Semikhatov 1978, p. 118-120, 122-125
	Murky Fm.		S	**Stratifera laxa**	
			S	**Stratifera tenica**	
	Pethei Gp.		S	**Omachtenia**	Hoffman 1976, p. 607
	Hearne Fm.		S	**Stratifera hearnica**	Semikhatov 1968, p. 115-117
	Utsingi Fm.		S	**Conophyton infernum**	Semikhatov 1978, p. 127-133, 143-145
			S	**Jacutophyton infernum**	
			S	**Vertexa termina**	
	Taltheilei Fm.		S	**Conophyton biformatum**	Semikhatov 1978, p. 126-127, 139-140
			S	**Kussoidella limata**	
	Kahochella Gp. Gibralter Fm.		S	**Stratifera kahochella**	Semikhatov 1978, p. 117-118
18	Martin Fm.		S?	Possible stromatolite (= caliche?)	Tremblay 1972, p. 121, 155
20	Pinguicula Gp.	Y	S	Stromatolites	Eisbacher 1978a, p. 54; Young et al., 1979, Fig. 1
	Wernecke Supgp. Gillespie Lake Gp.	Y	S	Stromatolites	Bell and Delaney 1977, p. 34; Goodfellow 1979, p. 333, 336, 337, 347; Young et al., 1979, Fig. 1
21	Mackenzie Mountains Supgp.	Y			
	Coppercap Fm.		S	Stromatolites	Eisbacher 1978, p. 18
	Redstone River Fm.		S	Stromatolites	Eisbacher 1977, p. 230, 232
	Little Dal Gp.		S	**Baicalia**	Gabrielse et al., 1973, p. 13, 17
			S	**Boxonia**	Aitken et al., 1973, p. 6, 15, 16, 17, 57, 58, 66, 67, 87, 89, 93, 95, 110, 111, 112
			S	**Gymnosolen**	
			S	**Jurusania**	Aitken and Cook 1974, p. 8
			S	cf. **Minjaria**	Young 1977, p. 1781
			S	**Parmites**	Aitken et al., 1978a, p. 383, 384
			S	**Stratifera**	Aitken et al., 1978b, p. 485
			O	Oncolites	Aitken and Cook 1974, p. 8
	Katherine Gp.		S	**Baicalia**	Aitken et al., 1973, p. 13, 69, 107
			S	**Inzeria** ?	Aitken and Cook 1974, p. 7; Aitken et al., 1978a, p. 481
	Tsezotene Fm.		S	Stromatolites	Gabrielse et al., 1973, p. 14; Aitken et al., 1973, p. 12; Aitken and Cook 1974, p. 7

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
					Aitken et al., 1978a, p. 481
	Unit 1		S	**Baicalia**	Aitken et al., 1973, p. 10
			S	**Conophyton**	Aitken and Cook 1974, p. 3
			S	**Jacutophyton**	Aitken et al., 1978a, p. 481
			S	**Stratifera**	Young 1977b, p. 1781
			S	**Svetliella**	
24	Windermere Supgp.	Z	B	**Chuaria** sp. cf. **C. circularis**	Hofmann, 1971a, p. 24
	Hector Fm.				Gussow 1973, p. 1108-1112 Cloud et al., 1975, p. 150
25	Windermere Supgp.	Z	M	**Sphaerocongregus variabilis** (= **Bavlinella faveolata**)	Moorman 1974, p. 524-539 Cloud et al., 1975, p. 131-150
30	Animikie Gp.	X	S	**Frutexites microstroma** (= microstromatolites)	Walter and Awramik 1979, p. 23-33 Awramik and Semikhatov 1979, p. 489-492
			S	**Gruneria biwabikia schreiberi** (nom. nud.)	Awramik 1976, p. 314-419 Awramik and Semikhatov 1979, p. 486, 487
			S	**Kussiella superiora** (nom. nud.)	Awramik 1976, p. 314, 319 Awramik and Semikhatov 1979, p. 487
			S	**Stratifera biwabikensis**	Awramik 1976, p. 314, 319 Awramik and Semikhatov 1979, p. 486
31	Animikie Gp.	X	M	**Anabaenidium barghoornii** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 39-40
	Gunflint Fm.		M	**Animikiea septata**	Hofmann 1971a, p. 45 Awramik and Barghoorn 1977, p. 139-140
			M	**Archaeorestis magna**	Awramik and Barghoorn 1977, p. 137-139
			M	**Archaeorestis schreiberensis**	Hofmann 1971a, p. 47
			M	**Chlamydomonopsis primordialis** (= poorly preserved **Huroniospora** sp.)	Edhorn 1973, p. 46, 48 Awramik and Barghoorn 1977, p. 129
			M	**Corymbococcus hodgkissii**	Awramik and Barghoorn 1977, p. 132-134
			P	**Cumulosphaera lamellosa** (= inorganic)	Edhorn 1973, p. 52
			M	**Entosphaeroides amplus**	Hofmann 1971a, p. 46 Awramik 1976, p. 315
			M	**Eoastrion bifurcatum**	Hofmann 1971a, p. 48
			M	**Eoastrion simplex**	Hofmann 1971a, p. 48
			M	**Eomicrhystridium barghoorni**	Hofmann 1971a, p. 50
			M	**Eosphaera tyleri**	Kaźmierczak 1976, p. 40-41 Tappan 1976, p. 636-637 Awramik and Barghoorn 1977, p. 130 Kazamierczak 1979, p. 1-22
			M	**Exochobrachium triangulum**	Awramik and Barghoorn 1977, p. 136
			S	**Frutexites microstroma** (= microstromatolite; see Occurrence 30)	Walter and Awramik 1979, p. 22-33
			M	**Galaxiopsis melanocentra**	Awramik and Barghoorn 1977, p. 136-137

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
			D	**Glenobotrydion aenigmatis** (= inorganic?)	Edhorn 1973, p. 50 Awramik and Barghoorn 1977, p. 129
			M	**Gunflintia grandis**	Hofmann 1971a, p. 47
			M	**Gunflintia minuta**	Hofmann 1971a, p. 46 Awramik 1976, p. 315-319 Awramik and Barghoorn 1977, p. 126-127 Awramik and Semikhatov 1979, p. 484-494
			M	**Huroniospora macroreticulata**	Hofmann 1971a, p. 47 Awramik and Barghoorn 1977, p. 140
			M	**Huroniospora microreticulata**	Hofmann 1971a, p. 47 Darby 1974, p. 1595-1596 Awramik 1976, p. 313-315 Tappan 1976, p. 635 Awramik and Barghoorn 1977, p. 140 Awramik and Semikhatov 1979, p. 488-494
			M	**Huroniospora psilata**	Hofmann 1971a, p. 48 Awramik and Barghoorn 1977, p. 140
			M	**Kakabekia umbellata**	Hofmann 1971a, p. 49
			M	**Leptoteichos golubicii**	Knoll et al., 1978, p. 978-990
			M	**Megalytrum diacenum**	Knoll et al., 1978, p. 987-989
			M	**Menneria levis** (= ? **Trachysphaeridium** sp.)	Lopukhin 1975, p. 169-173 Lopukhin 1976, p. 302-303
			M	**Palaeoanacystis irregularis** (= ?poorly preserved **Corymbococcus hodgkissii**)	Edhorn 1973, p. 40
			D	**Palaeorivularia ontarica** (= displacement of organic matter by growth of fibrous chalcedony)	Hofmann 1971a, p. 49-50 Oehler 1976, p. 123-128
			M	**Palaeoscytonema moorehousei** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 46
			M	**Palaeospiralis canadensis** (= poorly preserved **Gunflintia** or **Animikiea**)	Edhorn 1973, p. 54
			M	**Palaeospirulina arcuata** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 42
			M	**Palaeospirulina minuta** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 42
			M	**Primorivularia thunderbayensis** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 44
			M	**"Schizothrix" atavia** (= poorly preserved **Gunflintia** sp.)	Edhorn 1973, p. 42
			D	**Sphaerophycus gigas** (= ? inorganic; ? poorly preserved **Corymbococcus hodgkissii**)	Edhorn 1973, p. 42
			M	**Thymos halis**	Awramik and Barghoorn 1977, p. 134
			M	**Veryhachium** sp.	Hofmann 1971b, p. 522-524
			M	**Xenothrix inconcreta**	Awramik and Barghoorn 1977, p. 135-136

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
			M	Eucaryotic cells (= **Huroniospora** spp.)	Edhorn 1973, p. 54
			M	Radiolaria-like organism (= cf. **Eoastrion**)	Edhorn 1973, Pl. 9, fig. 1
			M	Desmid-like organism (= ?**Kakabekia**; cf. Barghoorn and Tyler, 1965, Fig. 7, part 10)	Edhorn 1973, Pl. 2, fig. 5, Pl. 9, fig. 5
35	Sutton Lake Gp.	X		Bostock 1971, p. 8-11	
	Nowashe Fm.		S	**Nucleella**	
			S	**Paniscollenia**	
			S	**Stratifera**	
36	Belcher Supgp.	X		Hofmann 1975, p. 1123	
	Mavor Fm.		S	**Colleniella**	Donaldson 1976a, p. 371-380
			S	**Gymnosolen**	Hofmann 1977, p. 175-205
			S	**Kussiella**	Semikhatov 1978a, p. 118-122
			S	**Stratifera**	
			S	**Tungussia**	
	Tukarak Fm.		S	**Gymnosolen**	
	McLeary Fm.		S	**Jacutophyton garganicum**	
			S	**Lenia** (= **Asperia aspera** = **Pseudogymnosolen**)	
			S	**Stratifera**	
			S	**Tungussia**	
			O	**Osagia**	
			K	**Vesicularites**	
	Kasegalik Fm.		S	**Lenia** (= **Asperia aspera** = **Pseudogymnosolen**)	
			S	**Minjaria**	
			S	**Stratifera laxa**	
			O	**Osagia**	
			K	**Vesicularites**	
40	Mistassini Gp.	X			
	Upper Albanel Fm.		S	**Colonnella**	Hofmann 1977, p. 179-180
			S	**Kussiella**	
			S	**Minjaria**	
	Lower Albanel Fm.		S	**Mistassinia wabassinon**	Hofmann 1978, p. 573-579
			S	**Straticonophyton icon**	Hofmann 1978, p. 579-584
			S	**Stratifera icon**	Hofmann 1978, p. 580
42	Knob Lake Gp.	X			
	Denault Fm.		S	**Asperia aspera** (= **Pseudogymnosolen**)	Semikhatov 1978, p. 121-122
49	Conception Gp.				
	Mistaken Point Fm.	Z	B	**Charnia masoni**	Hofmann 1971a, p. 43, 64
				Charniodiscus concentricus	Anderson 1978, p. 147
				cf. **Cyclomedusa davidi**	Williams and King 1979, p. 12-14
				other metazoans	
	Hibbs Hole Fm.	Z	B	Medusoids	Hofmann et al., 1979, p. 83, 85
54	Green Head Gp.				
	Ashburn Fm.	Y	S	**Archaeozoon acadiense**	Hofmann 1974, p. 1098-1115
61	Grenville Supgp.	Y?	S	Stromatolites	Crosby 1978, p. 37-59
	Mayo Gp.			Hofmann, this paper	
71	Belcher Supgp.	X		Hofmann, 1974, p. 87-88	
	Kasegalik Fm.		M	**Archaeotrichion** sp.	Hofmann 1975, p. 1121-1232
			M	**Biocatenoides sphaerula**	Hofmann 1976, p. 1040-1073
			M	**Eoentophysalis belcherensis**	Golubic and Hofmann 1976, p. 1074-1082

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
			M	**Eomycetopsis filiformis**	Golubic and Campbell 1979, p. 201-215
			M	**Eosynechococcus medius**	
			M	**Eosynechococcus moorei**	
			M	**Globophycus** sp.	
			M	**Halythrix** sp.	
			M	**Melasmatosphaera magna**	
			M	**Melasmatosphaera media**	
			M	**Melasmatosphaera parva**	
			M	**Myxococcoides minor**	
			M	**Palaeoanacystis vulgaris**	
			M	**Pleurocapsa**? sp.	
			M	**Rhicnonema antiquum**	
			M	**Sphaerophycus parvum**	
			M	Acritarcha	
72	Miette Gp.	Z	I	**Didymaulichnus miettensis**	Young 1972, p. 10-13 Häntzschel 1975, p. W60-W61
73	Miette Gp.	Z	M	Spheroids and framboids	Javor 1975, p. 1133 Javor and Mountjoy 1976, p. 115-119
74	Purcell Supgp. Altyn Fm. Waterton Fm.	Y	M	Silicified tubes (uniseriate algae?)	McGugan 1973, p. 558-566 Oehler 1976, p. 778-782
75	Windermere Supgp. Monk Fm. Irene Volcanics	Z	S	Stromatolites	Glover and Price 1976, p. 21-22
76	Horsethief Creek Gp.	Z	D	Probable burrows	Poulton 1973, p. 296, 299
77	Horsethief Creek Gp.	Z	S	Stromatolites	Poulton 1973, p. 296, 299
78	Ingenika Gp.	Z			
	Stelkuz Fm.		D	Bifurcating burrows	Mansy and Gabrielse 1978, p. 12
79	Ingenika Gp.	Z			
	Stelkuz Fm.		S	Stromatolites	Mansy and Gabrielse 1978, p. 12, 13
	Espee Fm.		O	Oncolites	Mansy 1978, p. 5
			S	Stromatolites	Mansy and Gabrielse 1978, p. 7, 10, 14
80	Helikian Gp.				Aitken et al., 1973, p. 244
	Map-unit 2	Y	S	Stromatolites	
	Map-unit 1	Y	S	Stromatolites	Aitken et al., 1973, p. 160
81	Lone Land Fm.	Y	D	Ovoid carbonaceous films	Aitken et al., 1973, p. 150-151
82	Rapitan Gp.				Gabrielse et al., 1973, p. 29
	Keele Fm.	Z	S	Stromatolites	Aitken et al., 1973, p. 20 Eisbacher 1978, p. 15, 16
83	Mackenzie Mountains Supgp.	Y			
	Tsezotene Fm.		D	?"burrows"	Aitken et al., 1973, p. 22, 62
	Tigonankweine Fm.		D	Fucoidal markings	Gabrielse 1973, p. 15
	Katherine Gp.		D	**Arthrophycus**-like markings	Aitken et al., 1973, p. 22, 97
	Unit H-5		D	**Arthrophycus**-like markings	Aitken et al., 1973, p. 22-89

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
84	Little Dal Gp.	Y			
	Basinal sequence		M	**Archaeotrichion** sp.	Hofmann and Aitken 1979, p. 155
			M	**Kildinella** sp.	Hofmann and Aitken 1979, p. 157
			M	**Nucellosphaeridium** spp.	Hofmann and Aitken 1979, p. 156
			M	**Siphonophycus** spp.	Hofmann and Aitken 1979, p. 156
			M	**Taeniatum** spp.	Hofmann and Aitken 1979, p. 156
			M	**Trachysphaeridium** spp.	Hofmann and Aitken 1979, p. 156
85	Little Dal Gp.				
	Basinal sequence	Y	B	**Beltina danai**	Hofmann and Aitken 1979, p. 162
			B	**Chuaria circularis**	Aitken et al., 1978, p. 483 Hofmann and Aitken 1979, p. 157
			B	**Morania? antiqua**	Hofmann and Aitken 1979, p. 160
			B	**Tawuia dalensis**	Aitken et al., 1978, p. 483 Hofmann and Aitken 1978, p. 158
86	Little Dal Gp.				
	Basinal sequence	Y	D	**Bergaueria**? sp.	Hofmann and Aitken 1979, p. 163
87	Shaler Gp.				
	Glenelg Fm.	Y	M	cf. **Caryosphaeroides** or **Glenobotrydion**	Jefferson 1977, p. 220-228
			M	**Eomycetopsis**	
			M	cf. **Gloeodiniopsis lamellosa**	
			M	cf. **Siphonophycus kestron**	
88	Dismal Lakes Gp.	Y	M	**Archaeoellipsoides grandis**	Donaldson and Delaney 1975, p. 371-377
			M	**Biocatenoides**? sp.	Schopf 1975, p. 223, 228
			M	**Eoentophysalis dismallakensis**	Horodyski and Donaldson 1980, p. 125-159
			M	**Eomicrocoleus crassus**	
			M	**Filiconstrictosus** sp.	
			M	**Myxococcoides grandis**	
			M	**Myxococcoides**? sp.	
			M	**Oscillatoriopsis curta**	
			M	**Oscillatoriopsis robusta**	
			M	**Oscillatoriopsis** sp.	
			M	**Sphaerophycus medium**	
			M	**Sphaerophycus parvum**	
			M	**Sphaerophycus**? sp.	
89	Labine Gp. (Unit 6; "Elizabeth Lake rhyolite")	X	S	Stromatolites	Hoffman et al., 1975, p. 356 Hoffman and McGlynn 1977, p. 182
90	Epworth Gp.				
	Rocknest Fm.	X	M	Algal filaments (cf. **Palaeoleptophycus**)	Gebelein 1974, p. 591
91	Yellowknife Supgp.				
	Snofield dol.	W	S	**Colleniella**	Henderson 1975 a, p. 325, 327
			S	**Pseudogymnosolen**	Henderson 1975b, p. 1619-1630
			S	**Stratifera**	Henderson 1977, p. 51-53
			O	**Osagia**	Hofmann, this paper
92	Fury and Helca Fm.	Y	S	Stromatolites	Chandler et al., 1980, p. 128
93	Woman Lake marble	W	S	Stromatolites	Thurston and Jackson 1978, marginal notes
	Unit 5d				Thurston 1980, p. 84
94	Kaministikwia area	W	P	Spheroidal microstructures (= nonbiologic crystallites)	LaBerge 1973, p. 1099, 1100

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
95	Timagami area	W	P	Spheroidal microstructures (= nonbiologic crystallites)	LaBerge 1973, p. 1099, 1100
96	Woman River area	W	P	Spheroidal microstructures (= nonbiologic crystallites)	LaBerge 1973, p. 1099, 1100
97	Woman River area	W	C	Sulfur isotope ratios	Goodwin et al., 1976, p. 881 Schidlowski 1979, p. 306
98	Michipicoten Gp.	W	C	Carbon isotope ratios	Goodwin et al., 1976, p. 870, 878-880
				Sulfur isotope ratios	Schidlowski 1979, p. 303, 306, 307
99	Elliot Lake Gp.	X	C	Carbon isotope ratios	Eichmann and Schidlowski 1974, p. 449
100	Gowganda Fm.	X	D	Lenses of carbonaceous matter	Jackson 1971, p. 313-315
			D	Metal sulfides	
			D	Lens-shaped cavities	
101	Espanola Fm.	X	S	Stromatolites	Pearson 1974, p. 15
102	Chelmsford Fm.	X	D	Possible biogenic structures	Rousell 1972, p. 86, 89
103	Onwatin Fm.	X	C	Carbon isotope ratios Sulfur isotope ratios	Goodwin et al., 1976, p. 888 Schidlowski 1979, p. 303, 307
104	Grenville Supgp.	Y?	S	Stromatolites	Crosby 1978, p. 18-36
105	Belcher Supgp.			Hofmann 1974, p. 87-88	
	McLeary Fm.	X	M	**Biocatenoides sphaerula**	Hofmann 1975, p. 1121-1132
			M	**Caryosphaeroides** sp.	Hofmann 1976, p. 1040-1073
			M	**Eoentophysalis belcherensis**	
			M	**Eomycetopsis filiformis**	Golubic and Hofmann 1976, p. 1074-1082
			M	**Eosynechococcus medius**	
			M	**Eosynechococcus grandis**	
			M	**Eozygion minutum**	
			M	**Glenobotrydion majorinum**	
			M	**Globophycus** sp.	
			M	**Kakabekia**? sp.	
			M	**Melasmatosphaera magna**	
			M	**Melasmatosphaera media**	
			M	**Melasmatosphaera parva**	
			M	**Myxococcoides inornata**	
			M	**Myxococcoides minor**	
			M	**Myxococcoides** sp.	
			M	**Palaeoanacystis vulgaris**	
			M	**Sphaerophycus parvum**	
			M	**Zosterosphaera** sp.	
			M	Acritarcha	
106	Nastapoka Gp.	X	K	Asterosphaeroides **(Nelcanella) Radiosus**	Hofmann 1977, p. 202-203
107	Otish Gp.				
	Peribonca Fm.	X	S	Stromatolites	Chown and Caty 1973, p. 58, 70
108	Sokoman Fm.	X	P	Spheroidal microstructures (= nonbiologic crystallites)	LaBerge 1973, p. 1099, 1100 Klein and Fink 1976, p. 471-474 Reeves 1979, p. 73
109	Ramah Gp.	X	S	Stromatolites	Morgan 1972, p. 126

Occurrence	Stratigraphic unit	Age	Type	Taxon	References
110	Ramah Gp. Reddick Bight Fm.	X	C	Carbonaceous matter in 5 cm thick bed	Morgan 1975, p. 24
111	Ramah Gp. Reddick Bight Fm.	X	D	Bioturbation structures	Knight 1973, p. 157 Morgan 1975, p. 24 Knight and Morgan 1977, p. 6, 16
	Rowsell Harbour Fm.	X	D	Possible biogenic structures	Morgan 1975, p. 22
112	Random Fm.	Z?			Nautiyal 1976, p. 609-611 Hofmann et al., 1979, p. 83-98
			M	**Gunflintia** sp. (= ?**Taeniatum** sp.)	
			M	**Heliconema** sp. (= ?**Taeniatum** sp.)	
			M	**Huroniospora** sp. (= ?**Trachysphaeridium** sp.)	
			M	**Leiosphaeridia** sp. (= ?**Trachysphaeridium** sp.)	
			M	**Leiovalia** sp. (= ?**Trachysphaeridium** sp.)	
			M	**Nucellosphaeridium** sp. (= ?**Trachysphaeridium** sp.)	
			M	**Siphonophycus kestron** (= ?**Taeniatum** sp.)	
			M	**Siphonophycus** sp. (= ?**Taeniatum** sp.)	
			M	**Taeniatum** sp.	
113	Musgravetown Gp.	Z	M	**Taeniatum** sp.	Hofmann et al., 1979, p. 83-98
			M	**Trachyspaeridium** sp.	
114	Connecting Point Gp.	Z	M	**Eomicrhystridium**? sp.	Hofmann et al., 1979, p. 83-98
			M	**Taeniatum** sp.	
			M	**Trachysphaeridium** sp.	
			M	**Trematosphaeridium holtedahlii**	
115	Hodgewater Gp.	Z			Hofmann et al., 1979, p. 83-98
	Snows Pond Fm.		M	**Taeniatum** sp.	
			M	**Trachysphaeridium** sp.	
	Halls Town Fm.		M	**Eomicrhystridium**? sp.	
			M	**Taeniatum** sp.	
			M	**Trachysphaeridium** sp.	
116	Signal Hill Gp.	Z			Hofmann et al., 1979, p. 83-98
	Gibbett Hill Fm.		M	**Trachysphaeridium** sp.	
	Cappahayden Fm.		M	**Taeniatum** sp.	
			M	**Trachysphaeridium** sp.	
117	St. John's Gp.	Z			Hofmann et al., 1979, p. 83-98
	Renews Head Fm.		M	**Trachysphaeridium** sp.	
	Fermeuse Fm.		M	**Taeniatum** sp.	
			M	**Trachysphaeridium** sp.	
118	Conception Gp.	Z			Hofmann et al., 1979, p. 83-98
	Drook Fm.		M	**Taeniatum** sp.	
	Gaskiers Fm.		M	**Bavlinella** sp.	Knoll, in Hofmann et al., 1979, p. 86
	Mall Bay Fm.		M	**Taeniatum** sp.	
119	Conception Gp.	Z			
	Briscal Fm.		B	Medusoids	Hofmann et al., 1979, p. 85
	Drook Fm.		B	Medusoids	
120	Rock Harbour Gp.	Z?	S	Stromatolites	Strong et al., 1978, p. 117, 122, 128
121	Burin Gp. Port au Bras Fm.	Z?	S	Stromatolites	Strong et al., 1978, p. 117, 123

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