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Commission géologique du Canada

PAPER
ÉTUDE 88-1C

CURRENT RESEARCH PART C
CANADIAN SHIELD

RECHERCHES EN COURS PARTIE C
BOUCLIER CANADIEN

1988

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GEOLOGICAL SURVEY OF CANADA
PAPER 88-1C
COMMISSION GÉOLOGIQUE DU CANADA
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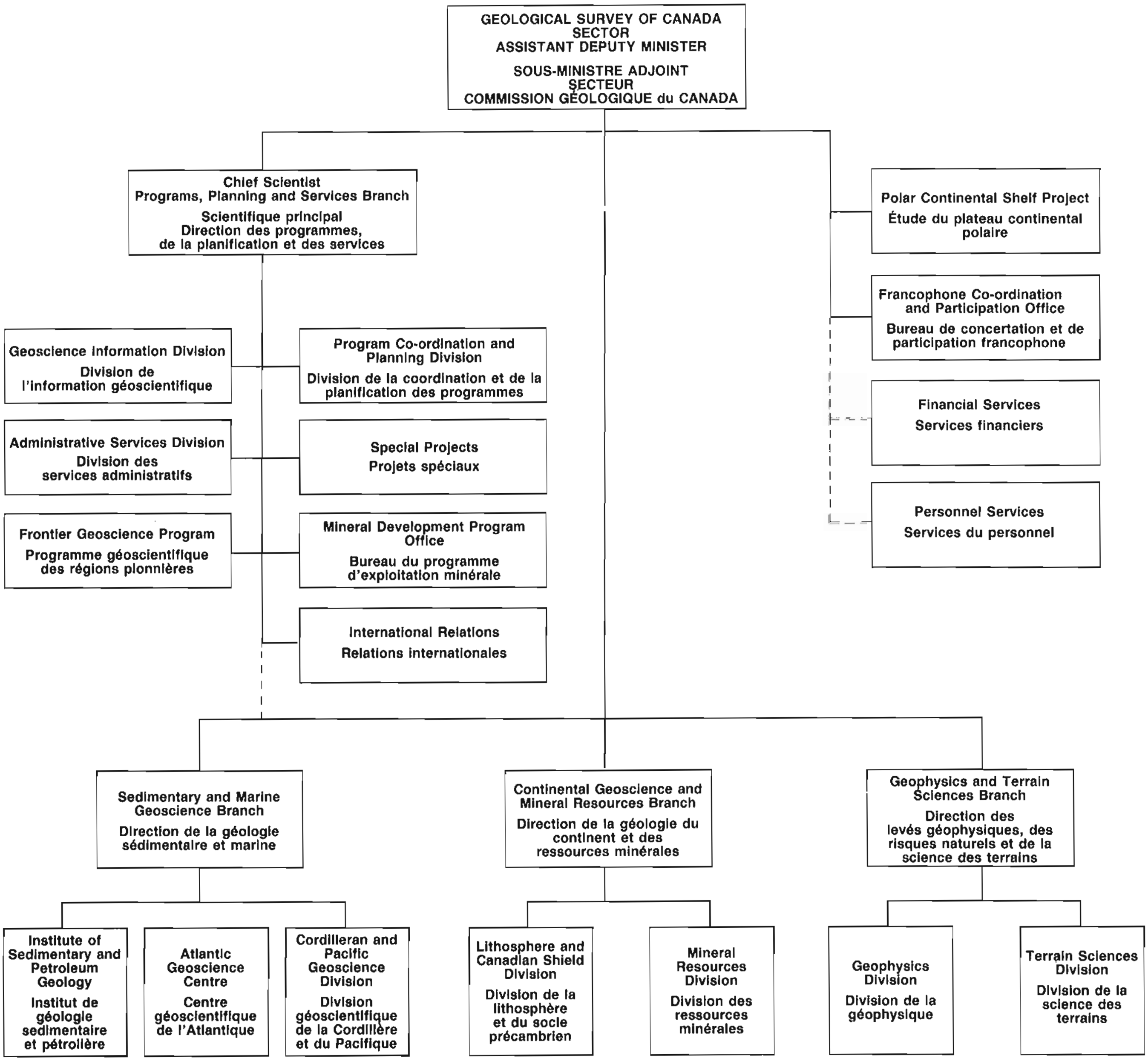
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INTRODUCTION

In 1987 the Geological Survey of Canada became a Sector within the Department of Energy, Mines and Resources, and was re-organized into the four Branches shown on the accompanying organizational chart. The primary role of the Survey, which was founded in 1842, continues to be to provide an overview of all facets of Canadian geology as a basis for national policy, for planning by government and industry, and for public information.

In order to provide interim results of its program a publication titled "Summary of Research" was initiated in 1963. The title was changed to "Current Research" in 1978 and the report was released three times a year (Part A, B and C). After 1982 Part C was no longer issued and Part B was discontinued in 1987 to encourage greater use of journal publication for short contributions.

Current Research, however, is the one series of GSC publications that gives the public a yearly overview of the range of the Geological Survey of Canada Sector activities. From time to time Current Research has been criticized for its size, as it was necessary for the user to buy a large volume to obtain a few pertinent papers. To introduce greater flexibility, this issue of Current Research is therefore available in six parts that can be purchased separately: four regional volumes, one volume of national and general programs, and a volume that contains abstracts of all the reports. The Parts are:

Part A: Abstracts/Résumés
Part B: Eastern and Atlantic Canada
Part C: Canadian Shield
Part D: Interior Plains and Arctic Canada
Part E: Cordillera and Pacific Margin
Part F: National and General Programs

Identification of the Parts by letters is for convenience only and may be subject to change. Each of Parts B to F includes a paginated Table of Contents for the volume: Table of Contents for the other Parts of this series will be found at the back of each volume.

En 1987 la Commission géologique est devenue un secteur à l'intérieur du ministère de l'Énergie, des Mines et des Ressources et a été réorganisée en quatre directions indiquées sur l'organigramme d'accompagnement. Organisme fondé en 1842, la Commission a comme rôle principal de procurer un cadre d'ensemble de toutes les facettes de la géologie du Canada comme base d'une politique nationale pour la planification du gouvernement et de l'industrie et pour informer le public en général.

Afin de fournir les résultats préliminaires de son programme de recherche une publication ayant titre « Summary of Research » est apparue en 1963. Une nouvelle publication, ayant les mêmes buts, est apparue en 1978 sous le titre « Recherches en cours »; cet ouvrage était diffusé trois fois par année (parties A, B et C). Après 1982 la partie C a été abandonnée et ce fut de même pour la partie B en 1987. L'arrêt de ces publications avait pour but d'adopter une nouvelle forme d'édition pour satisfaire davantage l'usager.

La publication « Recherches en cours » appartient à part entière à la série des publications de la CGC et apporte à chaque année une vue d'ensemble des activités de la Commission géologique maintenant au niveau de secteur. De temps à autre la publication « Recherches en cours » a été critiquée pour son fort volume, plusieurs ont constaté qu'il était nécessaire d'acheter un gros volume uniquement pour n'avoir accès qu'à un petit nombre d'articles. Maintenant, cette publication est disponible en six parties en vente séparément, ce qui procure une plus grande flexibilité pour l'usager. La publication est répartie comme suit: quatre volumes régionaux, un volume couvrant les programmes nationaux et généraux et un dernier contenant les résumés de tous les articles. On y trouve les parties suivantes:

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Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust, and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec

M.R. St-Onge, S.B. Lucas[1], D.J. Scott[2], N.J. Bégin[2], H. Helmstaedt[2], and D.M. Carmichael[2]
Lithosphere and Canadian Shield Division

St-Onge, M.R., Lucas, S.B., Scott, D.J., Bégin, N.J., Helmstaedt, H., and Carmichael D.M., Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 1-18, 1988.

Abstract

The 1.9 Ga continental rift sediments and volcanic rocks, transitional-crust and true oceanic crust (Purtuniq ophiolite) are contained in thrust sheets bounded by south-verging thrust faults (D_1) and preserved in the Cape Smith Belt as the result of two post-thrusting episodes (D_2, D_3) of thick-skinned crustal folding. The D_1 thin-skinned thrust-fold belt is characterized by a basal décollement, localized at the basement-cover interface, which displays a frontal ramp that exposes up to 30 m of footwall autochthonous fluvio-deltaic sandstone and ironstone. Imbricates of more proximal rift-basin ironstone and sandstone are structurally overlain by more outboard (rift basin) semipelite and proximal fan arkosic sandstone deposits which pass upwards into LREE-enriched basalts, gabbro sills and minor rhyolite. More internal thrust sheets contain transitional-crust basalts (komatiitic to MORB- like) which are imbricated with distal fan sandstones and semipelites. Deep water laminated graphitic pelites are structurally sandwiched between underlying transitional-crust basalts and hangingwall ophiolitic units. Imbricates of the ophiolite contain either (1) sheeted dykes overlain by pillowed and massive basalt flows intruded by mafic dykes and sills; or (2) mafic and ultramafic cumulates. D_1 out-of-sequence thrusting associated with interleaving of basement and cover units, resulted in re-imbrication of the thrust stack and further southward transport.

Résumé

Des roches métasédimentaires et métavolcaniques datées à 1,9 Ga, à faciès de rift continental, croûte transitionnelle et croûte océanique (ophiolite de Purtuniq), sont incorporées dans une série de nappes de charriage délimitées par des failles chevauchantes (D_1) et préservées dans la zone de Cape Smith en raison de deux périodes, survenues après le chevauchement, de plissement (D_2, D_3) des unités du socle et de la couverture. La zone de chevauchement et plissement des unités de couverture est caractérisée par un décollement de base situé au contact entre le socle archéen et la couverture protérozoïqe. Le décollement de base comprend une rampe frontale mettant en évidence jusqu'à 30 m de grès fluvio-deltaïque et de sable ferrugineux. Les imbrications de roche sédimentaire ferrugineuse et grès de bassin de rift proximal sont sous-jacents à des nappes de charriage renfermant de la semipélite et des grès proximaux de cône de déjection. Reposant stratigraphiquement sur les sédiments sont des basaltes riches en éléments des terres rares légers, des gabbros et une petite quantité de rhyolite. Les nappes chevauchantes d'origine plus lointaine dans l'arrière-pays renferment des basaltes de croûte transitionnelle dont le caractère varie du type komatititiques au type « MORB » (basaltes de dorsale médio-océanique) ainsi que les sédiments distaux de cône de déjection. Des pélites à graphite laminées sont mises en place en contact de faille entre les basaltes sous-jacents de croûte transitionnelle et des unités ophiolitiques sus-jacentes. L'ophiolite est segmentée en nappes de chevauchement contenant: soit 1) des dykes en feuillets sur lesquels reposent des coulées massives ou en coussins de basalte, recoupées par des filons-couches et dykes mafiques, soit 2) des cumulats mafiques et ultramafiques. Des chevauchements désordonnés liés à l'imbrication des unités du socle et de la couverture, ont provoqué une nouvelle imbrication du système de nappes chevauchantes, donnant ainsi lieu à une reprise du transport tectonique en direction sud.

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INTRODUCTION

The 1.9 Ga (R. Parrish, pers. comm., 1986) Cape Smith Belt is a thin-skinned, south-vergent thrust-fold belt (Hynes and Francis, 1982; Lamothe et al., 1984; Hoffman, 1985) which is exposed in a west-plunging oblique section (15 km of structural relief) from low structural levels in the Wakeham Bay area (St-Onge et al., 1986) to high structural levels in the Lac Watts - Lac Cross area (St-Onge et al., 1987). Field work during the summer of 1987 at exposed intermediate structural levels in the Rivière Déception - Lac Vicenza area (Fig. 1) has led to completion of 1 : 100 000 scale mapping of the eastern portion of the belt. As a result, the regional tectono-stratigraphic evolution recorded by the early Proterozoic units and the structural-metamorphic development of the thrust-fold belt can be fully documented. An understanding of both the tectono-stratigraphic record and structural architecture of the belt are essential for future mineral exploration projects in the area. As such, first-order constraints provided by GSC regional field-based research in the eastern Cape Smith Belt are presented in this paper and complement those published by Lamothe et al. (1986) for the western portion of the belt.

SUMMARY OF RESULTS

The principal results of three months of field work during the summer of 1987 in the eastern half of the thrust-fold belt (Fig. 1) can be summarized as follows:

1. In kilometre-scale outliers west of Burgoyne Bay, up to 30 m of early Proterozoic autochthonous cover is preserved in the footwall of the basal décollement. The autochthon is composed principally of trough crossbedded arkosic sandstones (channel deposits) interbedded with sandy ironstone and semipelite. Paleocurrent directions are from south to north.
2. The basal décollement is localized at the basement-cover interface in the Joy Bay area outliers, and in the Cape Smith Belt proper. The transition from up to 30 m of autouchthonous Proterozoic section in the south to none in the north documents the presence of a major footwall ramp associated with propogation of the basal décollement south of the main Cape Smith Belt.
3. Clastic sediments in progressively more internal (northerly) thrust sheets record the sedimentological evolution from a continental-rift setting to a deep water, ocean basin environment. In southern (external) thrust sheets, sediments consist of ironstone, semipelite, proximal fan deposits of arkosic sandstone, and micaceous sandstone. Sediment-dominated imbricates from intermediate structural levels contain distal fan sandstones and semipelite. Finally sediments in the most internal thrust sheets (highest structural levels) are dominated by laminated graphitic pelite.
4. A similar and parallel change in the nature of volcanic and associated mafic intrusive units can be documented from external (southern) to internal (northern) thrust sheets. In the southern, more external imbricates, volcanic rocks consist of LREE-enriched basalt and minor rhyolite. Thrust sheets at intermediate structural levels are dominated by komatiitic to MORB-like tholeiitic basalts, interpreted as part of crust transitional to true oceanic crust. At the highest structural levels, the most internal imbricates consist of "sheeted" mafic to ultramafic dykes, gabbro sills cut by dykes, pillowed basalts cut by mafic dykes and mafic-ultramafic cumulates. These units represent part of an obducted and structurally dismembered ophiolitic suite. The progressive outboard change in the nature of the sediments, volcanics and mafic intrusive rocks document the evolution in tectono-stratigraphic environment from continental rift-basin to transitional-crust, and finally to true oceanic crust.
5. Pre-thermal peak D_1 thrust faults root on the basal décollement. A north to south movement direction for these early D_1 thrusts is documented by the geometry of both footwall and hangingwall ramps and associated fault-bend folds, and by the asymmetry of rare minor, pre-thermal peak D_1 folds. The early D_1 thrust faults show a piggyback stacking sequence, with more external, structurally lower faults being younger than structurally higher faults.
6. Syn-D_1 thermal relaxation following early D_1 crustal thickening resulted in growth of metamorphic mineral assemblages in a hot-side-down thermal geometry, and development of a ductile basal shear zone during continued basal décollement slip.
7. The timing of D_1 out-of-sequence thrust faulting is interpreted to be syn-to-post-thermal peak in age. Kinematic indicators (commonly shear bands) combined with stretching lineations in the high strain zones associated with the out-of-sequence faults document their north to south movement sense. The syn-thermal peak out-of-sequence thrust faults root on an out-of-sequence basal décollement and are responsible for the emplacement of basement imbricates in the thrust belt.
8. The morphology and amplitude of macroscopic D_2 thick-skinned folds of both basement and cover changes from south to north in the early Proterozoic outliers southeast of the Cape Smith Belt proper. In the southern (more external) outliers, the D_2 folds are open, upright structures with rounded profiles and limb dips of 10° to 30°. In the northern (more internal) outliers, the D_2 folds are tight, asymmetric, south-verging structures which preserve the cover rocks in pinched (cuspate) synforms with steep to overturned northern limbs. The foreland to hinterland change in macroscopic D_2 fold character is probably related to the ramp in the basal décollement which exposes thick, undeformed autochthonous sandstone beds, and changes the rheological structure of the basal Proterozoic tectono-stratigraphy.
9. Macroscopic D_3 cross-folds are best developed in areas where there are appropriate rheological contrasts within the tectono-stratigraphy to allow buckling. As a result, these cross-folds are primarily found: (1) at the Archean basement/Proterozoic cover contact; and (2) in areas where Kattiniq intrusive suite peridotite-gabbro sills are emplaced in rift-fill Povungnituk Group clastic sediments.

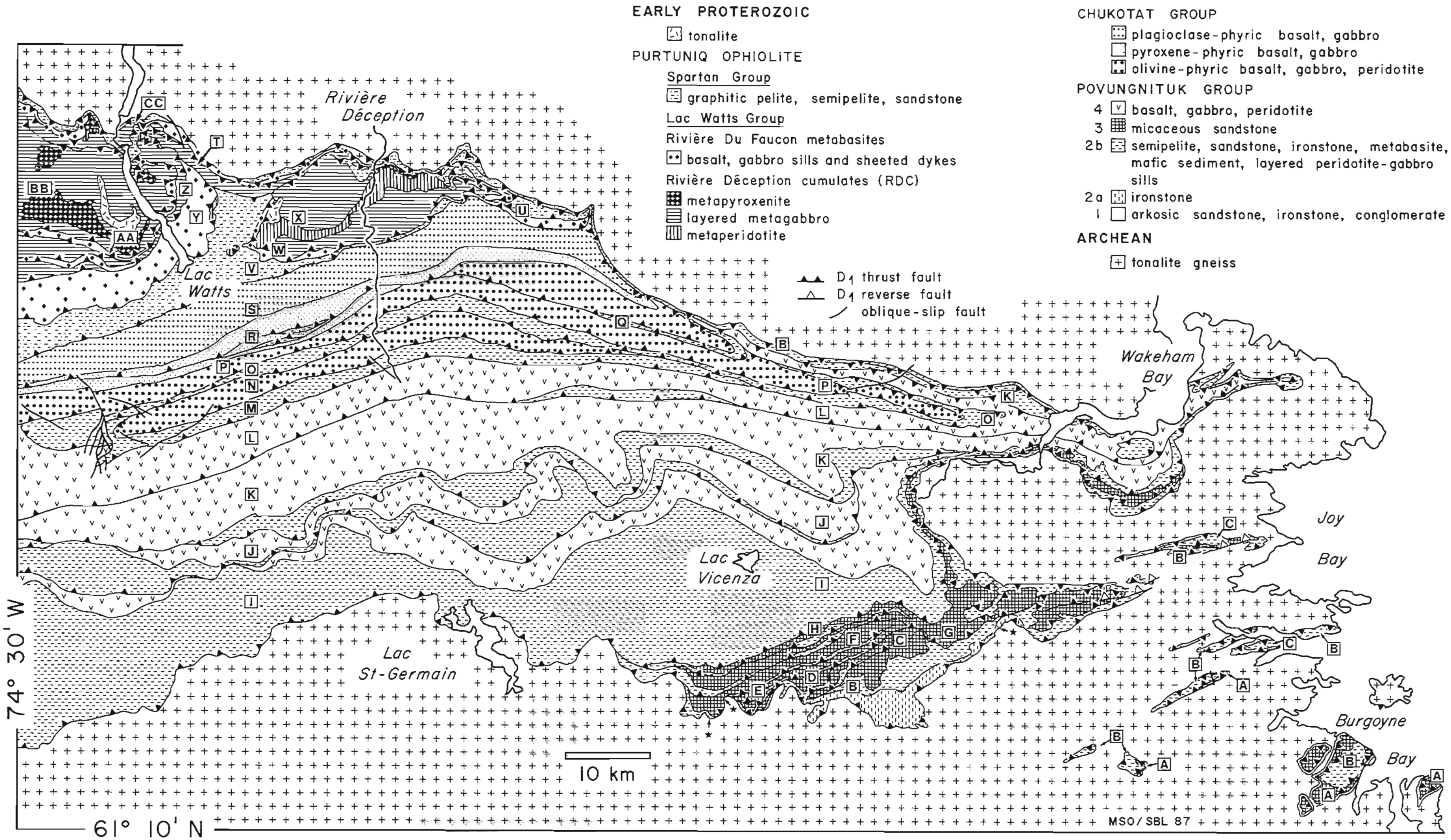

Figure 1. Geological map of the eastern Cape Smith Belt. Descriptions of early Proterozoic units are given in text. Stars denote locations of autochthonous cover rocks. Boxed letters identify thrust sheets referred to in text.

TECTONO-STRATIGRAPHY

General statement

The present distribution of thrust sheets containing sedimentary, volcanic and mafic intrusive rocks in the Cape Smith Thrust-Fold Belt is primarily determined by the age, displacement and stacking order of D_1 thrust faults. A discussion of the tectono-stratigraphic record within the belt must therefore take into account the fundamental architecture of the thrust-fold belt. The following description begins with the autochthon and basal (external) thrust imbricates which preserve the more inboard stratigraphic facies. This is followed by a description of the more internal (intermediate and higher structural level) thrust sheets which expose progressively more outboard units. Intrinsic to the discussion is the notion that tectono-stratigraphic changes from low to high structural levels primarily reflect initial lateral paleogeographic changes from south to north. This is justified by the careful documentation of the stacking order of both regular and out-of-sequence D_1 thrust faults.

Autochthon

Autochthonous early Proterozoic strata occur in three small (single outcrop) occurrences of clastic sediments on the south margin of the main belt, and in more extensive (up to 30 m thick) clastic sequences in nine outliers southeast of Wakeham Bay (Formation 1, Fig. 1). In both the main belt and nine outliers, autochthonous units rest unconformably on more steeply-dipping Archean tonalitic gneisses that show no evidence of a penetrative Proterozoic fabric. With the exception of one outlier, all autochthonous units are capped by a high strain zone corresponding to the position of the basal décollement. In the main belt the autochthon comprises pockets of either garnetiferous polymictic metaconglomerate (St-Onge et al., 1986), quartz pebble-bearing arkosic grit and/or iron-rich sandstone. At all three localities (Fig. 1) the thickness of the autochthon is less than 1 m.

In the Burgoyne Bay area outliers (Fig. 1), the autochthon ranges from 20 m to 30 m thick. Prominent in the sequence is a succession of 2-3 m thick arkosic sandstone beds (Fig. 2) characterized by abundant blue quartz grains. Rounded to subrounded feldspar (dominantly plagioclase) and quartz grains vary from submillimetre to several millimetres in diameter. Graded bedding is well preserved, with the base of individual beds often corresponding to a quartz pebble grit. Some of the thinner (less than 1 m) arkosic sandstone beds display well developed trough crossbedding (Fig. 3) indicating channelized paleo-flow from south to north in a fluviodeltaic environment. The arkosic sandstone is interbedded with 1-2 m thick sandy ironstone beds, showing graded bedding, and with 2-3 m thick intervals of well-laminated semipelite. The accumulation of the autochthonous clastic sediments in both the main belt and the outliers is interpreted to signal the onset of basin subsidence related to epicontinental rifting. Based on lithological correlations, the sediments are assigned to the lower Povungnituk Group as defined by Bergeron (1959).

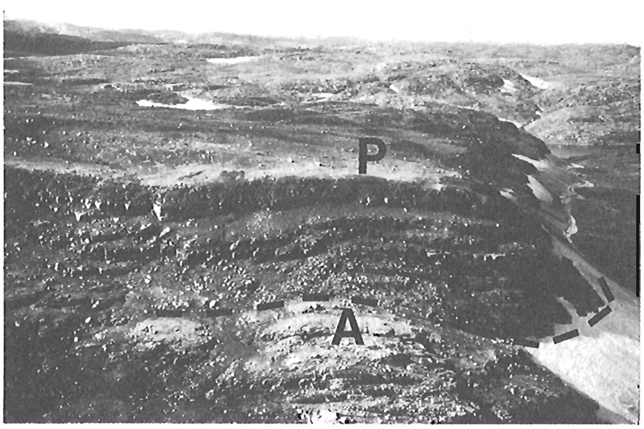

Figure 2. Aerial view of autochthonous Povungnituk Group channel-fill arkosoic sandstones. Individual white sandstone beds are 3 metres thick and are interbedded with semipelite and iron rich sands. Dashed line outlines unconformity between early Proterozoic (P) cover and Archean (A) basement. GSC 204234-F

Basal (external) thrust sheets

Lower Povungnituk Group.

Lower Povungnituk Group sediments preserved in the thrust sheets of the outliers (thrust sheets A,B,C, Fig. 1) southeast of Wakeham Bay and in the basal imbricates along the southern margin of the main belt (thrust sheets B,C,D, Fig. 1), exhibit an internal stratigraphy that is consistent at a regional scale. The lower sediments in the hangingwall of the basal décollement are ironstones which vary in thickness from 50 m to 500 m (Formation 2a, Fig. 1). Different units within the ironstones include ferruginous sandstones, interbedded massive magnetite/grunerite schist, carbonate-rich grunerite schist and garnet-biotite-hornblende-grunerite schist. These units are not very continuous laterally and therefore the ironstones are not subdivided in Figure 1. Overlying the ironstones are 60 m to 340 m of rusty semipelites and proximal fan deposits of arkosic sandstone that grade laterally into mafic sediments interlayered with metabasites (Formation 2b, Fig. 1). The semipelites are thinly bedded (cm-scale), occasionally graded, and are characterized by parallel laminations at the top of beds. The arkosic sandstones are composed of subrounded to rounded mm-scale individual quartz and feldspar grains. Bed thicknesses range from 15 cm to over 2 m (Fig. 4). Tens-of-metre scale channel structures and graded

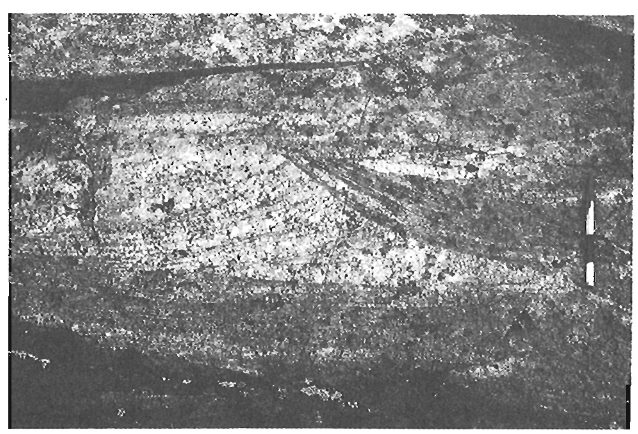

Figure 3. Detail of arkosic sandstone bed showing trough crossbedding. Length of hammer is 34 cm. GSC 204234-G

bedding are commonly preserved in the arkosic sandstones. The mafic sediments are dominated by quartz-biotite beds containing metamorphic actinolite and/or hornblende. The base of the rusty semipelite and arkosic sandstone interval is marked by a narrow 2 m to 5 m discontinuous interval of dolomite and associated calc-silicate units. The rusty semipelites are in turn overlain by massive, homogeneous, ridge-forming micaceous sandstones (Formation 3, Fig. 1), which vary in thickness from 70 m to 270 m (St-Onge et al., 1986; Fig. 1.4). The micaceous sandstone consists of quartz and up to 20 % metamorphic biotite and muscovite. This unit is characterized in the field by the common occurrence of discontinuous, centimetre-scale milky quartz veins and segregations.

At higher structural levels, along the southern margin of the thrust-fold belt, subsequent thrust sheets show progressively more outboard stratigraphic facies in Formation 2 due to the piggyback stacking nature of early D_1 thrust faults (see section on D_1 regular stacking sequence thrust faults). Thus for thrust sheets E to H (Fig. 1), although the sedimentary rock types described above are generally present, the proportions of lithologies within Formation 2 change systematically with each more internal (northerly) thrust sheet. For example thrust sheet G contains volumetrically less ironstone and arkosic sandstone but more semipelite than thrust sheet E. Thrust sheet H has even less ironstone and laterally less continuous beds of arkosic sandstone, but proportionally more semipelite than thrust sheet E or G. The volumetric decrease in ironstone and the proportional increase in semipelite coupled with the transition to more laterally discontinuous and thinner arkosic sandstone beds in structurally higher (more northerly) imbricates of Formation 2 is interpreted as reflecting a progressive increase in rift-basin water depth from south to north. Within the lower imbricates (thrust sheets A to H, Fig. 1), the micaceous sandstones form a laterally extensive blanket overlying Formation 2 deposits. The micaceous sandstones in all thrust sheets show a relatively homogeneous mineral assemblage and uniform grain size both laterally and vertically. Unlike the sediments of Formation 2, the micaceous sandstones do not show outboard (south to north) facies changes, and do not interfinger with volcanic rocks. The accumulation of such a thick, homogenous micaceous sandstone section lacking channel structures is interpreted to mark an outboard shift in local rift-related tectonic activity, providing a relatively more stable environment for clastic sediment deposition.

The relatively large displacement thrust underlying thrust sheet I (Fig. 1) emplaces significantly further outboard lower Povungnituk Group sediments. These consist almost entirely of laminated semipelites interbedded with arkosic sandstones and no overlying micaceous sandstone unit. Stratigraphically above the semipelites, in thrust sheets I, J and K (Fig. 1), pillowed basalts and gabbro sills of the upper Povungnituk Group are found rather than the micaceous sandstone (Formation 3) of the lower imbricates.

This dramatic stratigraphic change indicates that the locus of rift-basin subsidence and active rift volcanism had shifted outboard, permitting inboard micaceous sandstone deposition. The total thickness of the lower Povungnituk Group in thrust sheets I, J and K is greatly inflated by emplacement of numerous gabbro sills interpreted to be part of the feeder system to the overlying upper Povungnituk Group volcanics. The mafic sills range from 10 m to 600 m thick and are essentially conformable with bedding defined by graphitic laminations in the clastic sediments. The gabbro sills have narrow (less than 1 m) chilled margins and rarely show internal layering.

Upper Povungnituk Group.

Ridge-forming metabasalts (Formation 4, Fig. 1) conformably overlie or interfinger with the clastic sediments of the lower Povungnituk Group in thrust sheets I, J, K and L. The volcanic pile varies in thickness from 2100 m to 2500 m, the top of the Povungnituk Group being marked by a tectonic contact. Both thin, pillowed mafic flows and thicker, tabular flows are present in the succession. Rare dolomite and semipelites are preserved between the flows and mafic pyroclastic rocks are most common in the lower part of the volcanic section. Also characteristic of the base of the mafic pile are laterally discontinuous 5 m to 10 m thick volcanic conglomerate units. The latter are typically composed of sub-rounded to angular blocks of basalt and gabbro (Fig. 5) which range in diameter from a few centimetres to several metres. The conglomerate is clast-supported and the infilling matrix consists of millimetre-scale basalt rocks fragments and carbonate.

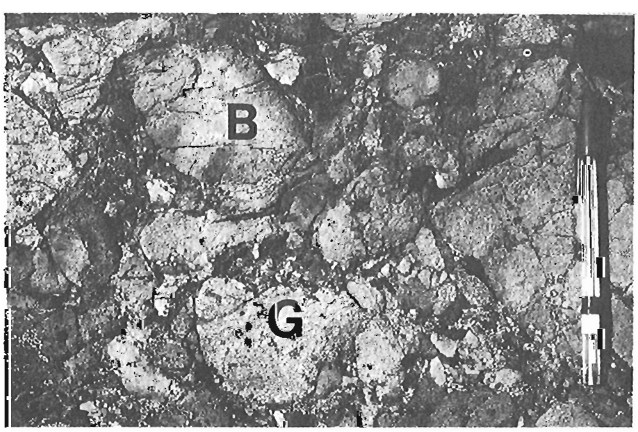

Figure 4. Proximal fan sandstone beds in basal thrust sheet I (Fig. 1) of main thrust-fold belt. Arrow points to 80 cm long hammer. GSC 204234-H

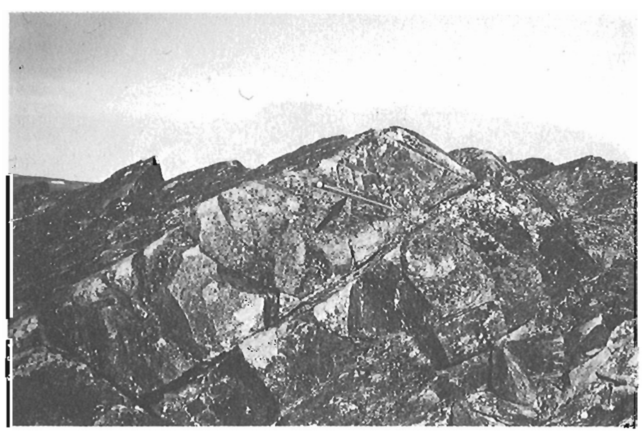

Figure 5. Upper Povungnituk Group fault-scarp conglomerate. Note angularity of blocks and presence of both gabbro (G) and basalt (B) clasts. Length of pen is 15 cm. GSC 204234-I

The little-reworked nature of the clasts, the presence of both basalt and gabbro clasts and the limited lateral extent of the volcanic conglomerate suggest that this unit is possibly a fault scarp breccia related to ongoing (normal?) faulting during accumulation of the Povungnituk Group.

The mafic volcanic rocks are reported by Hynes and Francis (1982) and Francis et al. (1983) to be Fe- and Ti-rich continental tholeiites. Laterally discontinuous and thin (less than 50 m thick) rhyolite bodies occur within the mafic volcanic pile. A rhyolite body 40 km northwest of Lac du Cratère (Fig. 1) has been dated at 1960 Ma (U-Pb zircon) (R. Parrish, pers. comm., 1986).

Intermediate structural level thrust sheets

Lower Povungnituk Group.

Thrust sheets L, M, and O, are found at intermediate structural levels in the thrust-fold belt (Fig. 1). They are further travelled and occur more internally than the lower imbricates described above and therefore bring a more outboard package of lower Povungnituk Group sediments. In all three thrust sheets the dominant sediments are semipelites interbedded with minor distal fan sandstones (Formation 2b, Fig. 1). The semipelites are thin-bedded (centimetre-scale) and interfinger with beds of more pelitic composition. The sandstone beds are tens of centimetres thick and are composed chiefly of rounded submillimetre quartz grains. Individual beds are apparently laterally very continuous and show no evidence of channel structures or crossbedding.

At the intermediate structural levels, the lower Povungnituk Group is either in thrust contact with: (1) footwall upper Povungnituk Group metabasites (regular stacking sequence thrust); or (2) hangingwall and footwall Chukotat Group basalts (both regular and out-of-sequence thrust faults).

Chukotat Group.

Thrust sheets N, P, and S (Fig. 1) are host to the Chukotat Group (Bergeron, 1959), which consists dominantly of pillowed basalt flows but also contains massive flows intruded by thin (5-100 m thick) mafic and layered mafic- ultramafic sills. In contrast to the upper Povungnituk Group, the Chukotat Group is devoid of sediments. The volcanic rocks range from Mg-rich komatiitic basalts to low-Ti, tholeiitic basalts with affinities to modern MORBs (Hynes and Francis, 1982). Primary textures and volcanic structures, such as variolitic pillow margins, ropy lava, pillow tubes (Fig. 6) and flow top breccia are well preserved. Three distinct pillowed basalt types within the Chukotat Group can be recognized based on the presence of a dominant pseudomorphed phenocryst type (Francis and Hynes, 1979; Hynes and Francis, 1982; Francis et al., 1983). As a result, dominantly olivine-phyric, pyroxene-phyric and plagioclase-phyric pillowed basalts were distinguished during mapping (Fig. 1). A complete description of field and petrographic characteristics of the Chukotat Group lavas is given in Hynes and Francis (1982).

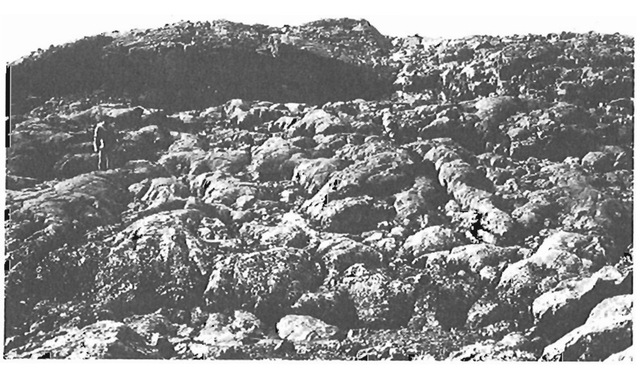

Figure 6. Chukotat pillowed basalt flows. Height of person is 1.8 m. GSC 204234-J

Kattiniq intrusive suite.

The clastic metasediments and metabasalts of the Povungnituk Group in thrust sheets B to Q (Fig. 1) are host to numerous layered peridotite-gabbro sills (Fig. 7) which are tentatively interpreted as consanguineous and are collectively named the Kattiniq intrusive suite (KIS). The layered sills range up to 500 m in thickness and are commonly characterized by a basal pyroxenite chilled border phase overlain by columnar jointed peridotite and capped by gabbro. The thicker sills show a layered gabbroic upper sequence with cyclic plagioclase cumulate layering (St-Onge et al., 1987, Fig. 65.4), ferrogabbro and quartz ferrogabbro zones. Simple peridotite sills and dykes postdate the main phase of layered sill emplacement, and as a result represent the last stage of KIS magmatism. Layered mafic-ultramafic sills emplaced in the Chukotat Group are potentially part of the Kattiniq intrusive suite. However, this correlation requires geochemical and geochronological data.

Higher structural level (internal) thrust sheets

Spartan Group.

At the higher structural levels exposed in the more internal part of the thrust-fold belt, clastic sediments in thrust sheet V (Fig. 1) are distinct from the lower Povungnituk Group and have been assigned to the Spartan Group (Lamothe et al., 1984). The sediments are dominantly laminated dark pelites interbedded with minor semipelite (Fig. 8). Beds in the pelites range in thickness from 2-3 mm to 1-2 cm and are commonly outlined by graphite laminae. Quartz- rich semipelitic beds of similar thicknesses grade upwards into graphite-rich pelite. Towards the structural top of thrust sheet V the sedimentary package is characterized by the occurrence of discrete, white, fine grained sandstone beds generally tens of centimetres thick. The sediments of thrust sheet V are interpreted as deep water pelagic sediments with the monotony of the sequence interrupted near the top by distal fan deposits.

Lac Watts Group.

Rivière du Faucon Metabasites. Thrusted over the Spartan Group sediments in the southern Lac Watts area (Fig. 1) is a 4 km thick sequence of mafic dykes, gabbro sills and pillow basalts (Fig. 9). The metabasites are thrust fault-bound both at the base and top of the mafic sequence. The thrust

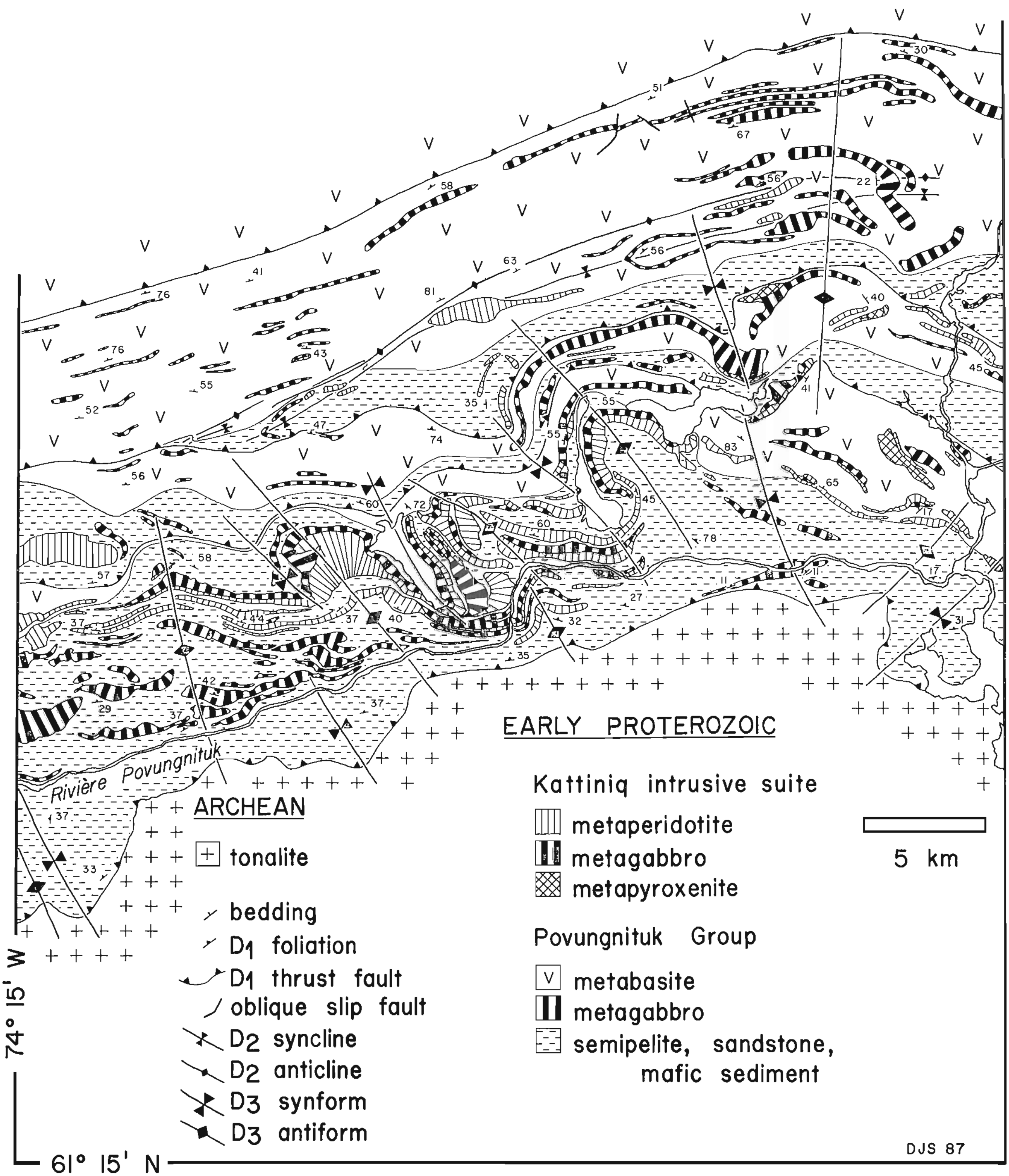

Figure 7. Detailed geological map documenting D_3 folding of lower Puvungnituk Group and Kattiniq intrusive suite sills. Lake in southeast corner is Lac St-Germain.

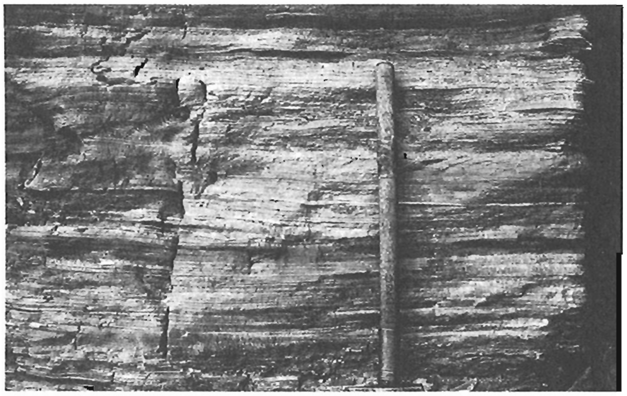

Figure 8. Spartan Group laminated graphitic pelites on southeast side of Lac Watts. Length of hammer is 50 cm. GSC 204234-R

Figure 10. Rivière du Faucon "sheeted dykes". Note that outcrop is 100% mafic half-dykes dipping steeply to the east (right-hand side of photograph). Length of hammer is 34 cm. GSC 204234-O

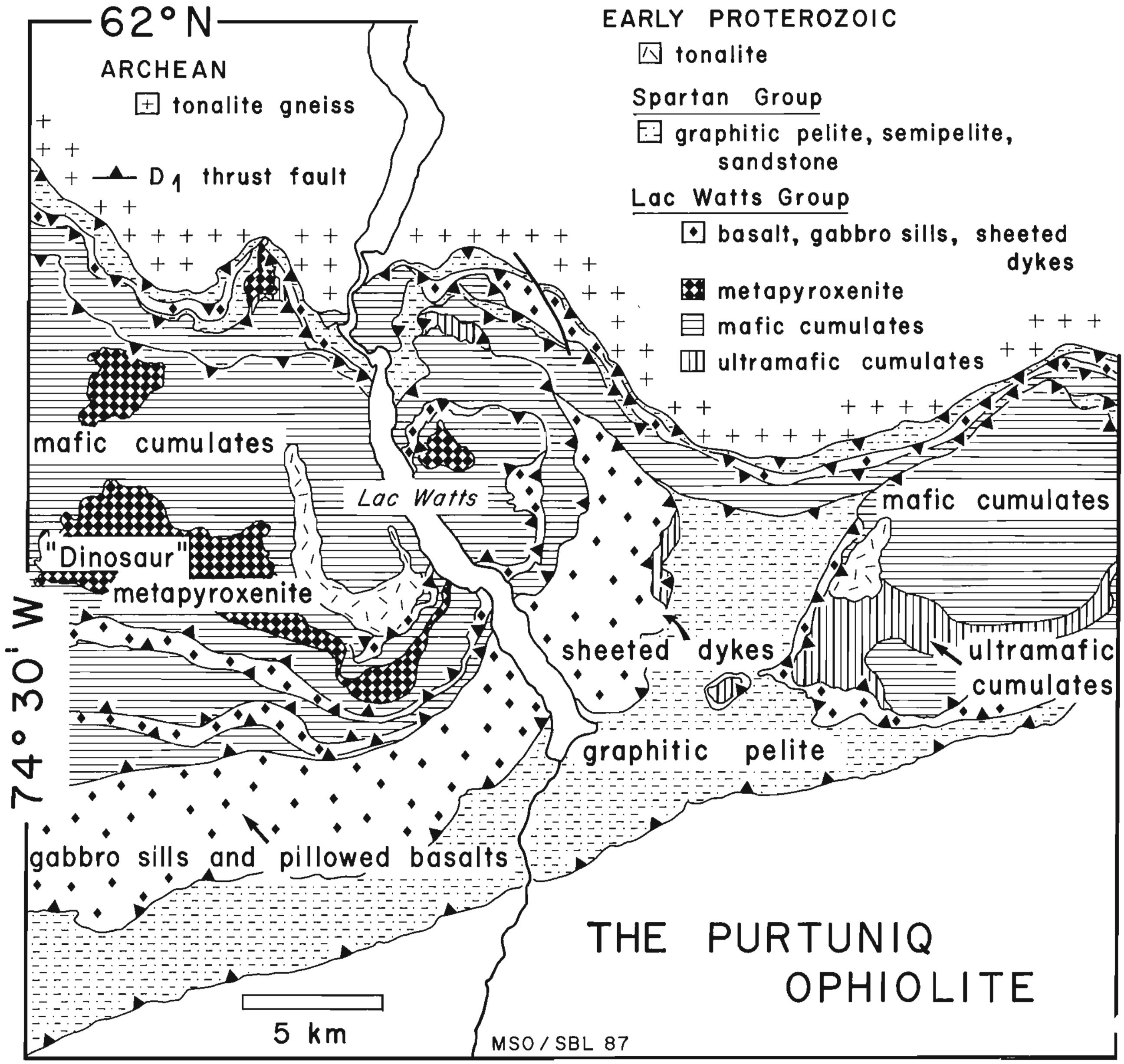

Figure 9. Detailed geological map for Lac Watts area showing the distribution of the tectono-stratigraphic units of the Purtuniq ophiolite.

sheet is upright based on pillow facing directions. On the east side of Lac Watts, the unit near the base of thrust sheet Y is a sheeted dyke complex which is best exposed in a large outcrop 6 km due west of Asbestos Hill (Fig. 9) and on which the following description is based. The composite exposed width of the sheeted dykes is 60 m with a strike length exposure of approximately 500 m. Individual "half-dykes" are 20 cm to 30 cm wide and commonly preserve a 2-3 mm wide one-sided chilled margin. The general orientation of the dykes is 030°, with dips averaging 80° to the southeast. The dykes range in composition from mafic, both with and without plagioclase phenocrysts, to ultramafic with abundant pyroxene. A striking feature in the field is the 100 % intrusive nature of the complex (Fig. 10), requiring multiple dyke emplacement into previously emplaced older dykes to produce the sets of half-dykes with one-sided chilled margins.

Overlying the sheeted dykes and occurring over a structural interval of several hundred metres are 2-3 m thick mafic sills and massive basalt flows striking 030° and dipping 20° to the northwest. The sills are gabbroic with the grain-size ranging from fine to medium. Massive flows and sills are crosscut by a set of discrete single dykes that show the compositional range of the underlying sheeted dykes. The crosscutting dykes are on average 30 cm wide, striking northeast and dipping 80° to the east.

Above the mafic sill-dyke and massive flow sequence, up to 3.5 km of pillowed basalt and mafic sills are exposed in thrust sheet Y. The basalt flows are aphyric with well developed pillow selvages. The flows strike 030° and dip 20° to the northwest on average. Pillow tops indicate a stratigraphically upright sequence. The mafic sills are of similar appearance and apparent composition as the underlying sills. A small number of steeply-dipping mafic dykes crosscuts both pillows and sills.

Rivière Déception Cumulates. A series of extensive, kilometre-scale thick imbricates of metaperidotite, metapyroxenite and layered metagabbro occur in thrust sheets T, X, Z, BB and CC (Fig. 1) at the highest structural levels exposed in the eastern thrust-fold belt. St-Onge et al. (1987) interpreted these units as consanguineous and the collective informal name "Lac Watts intrusive suite" was employed. Detailed field work in 1987 has shown that, with only one exception, the metaperidotites, metapyroxenites and layered gabbros are all mafic or ultramafic cumulates. The only exception is an intrusive metapyroxenite body on the west side of Lac Watts ("Dinosaur Metapyroxenite", Fig. 9) which shows no internal layering and locally contains layered metagabbro xenoliths at its margins. Given the occurrence of identical metaperidotites, metapyroxenites and layered metagabbros east of the Rivière Déception it is here proposed to informally rename this package of rocks the Rivière Déception cumulates (RDC). The Lac Watts Group (Lamothe et al., 1984), consists of the Rivière du Faucon metabasites and the Rivière Déception cumulates.

The RDC metaperidotites show centimetre to metre scale layering defined by colour changes (tan versus reddish brown) and differential weathering (Fig. 11). The changes are interpreted to reflect modal variations in the original igneous mineral assemblage: olivine and pyroxene. Laterally discontinuous layers of chromite, although rare, were documented at several locations in the ultramafic cumulates. The ultramafic cumulates occur in two different settings: as fault-bound lozenges (horses) at the base of thrust sheets of RDC layered metagabbro or within layered gabbro thrust sheets as an integral part of the cumulate sequence (Fig. 9).

The layered metagabbros form the volumetrically most important unit of the RDC. They occur as fault-bound thrust sheets ranging in thickness from 100 m to several kilometres. The centimetre- to metre-scale compositional layering in the metagabbros is defined by the modal abundance of two end-member mineral assemblages: hornblende, which forms black metapyroxenite layers, and plagioclase, which forms striking white meta-anorthosite layers. Most common are layers of gabbroic composition that clearly exhibit metamorphosed primary pyroxene-plagioclase compositional layering (Fig. 12).

An Early Proterozoic Ophiolite. St-Onge et al. (1987) proposed two potential tectono-magmatic models to account for the mafic and ultramafic rocks of thrust sheets W to CC. The first model suggested that the RDC represent an imbricated, transported and metamorphosed large layered intrusion. The second model proposed that the RDC may represent the lower part of an ophiolite suite. After detailed work in the Lac Watts - Rivière Déception area (Fig. 9) there can

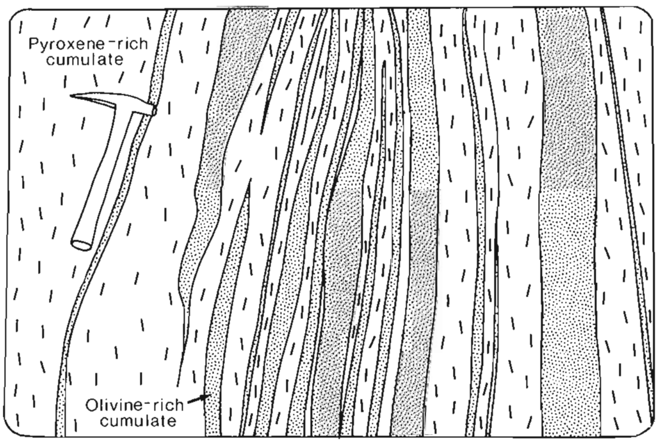

Figure 11. Line drawing of Rivière Déception ultramafic cumulates characterized by alternating metamorphosed olivine-rich and pyroxene-rich beds. Length of hammer is 34 cm.

Figure 12. Rivière Déception mafic cumulates characterized by plagioclase cumulate layering. Length of pen is 15 cm. GSC 204234-Q

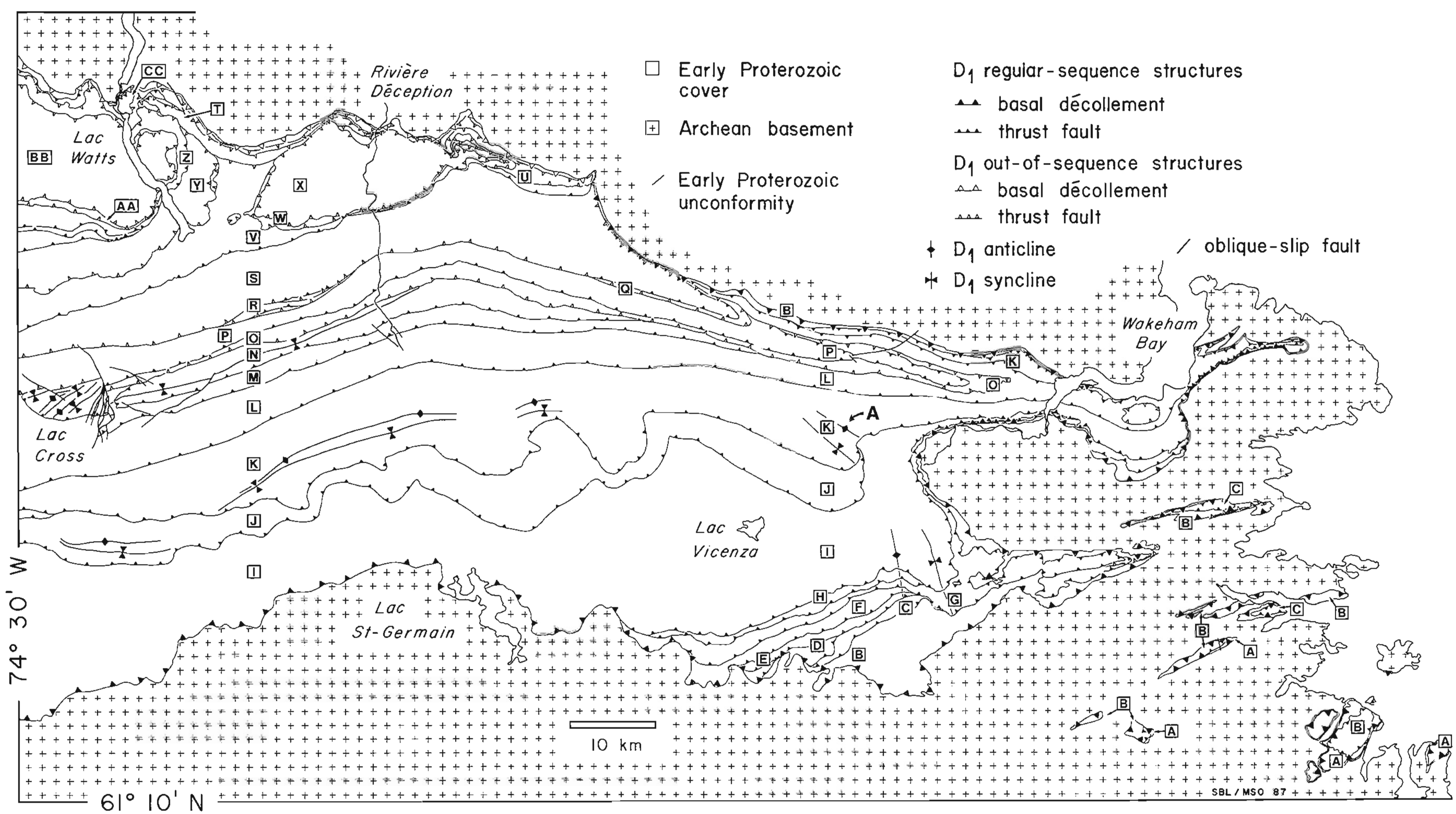

Figure 13. Compilation of D_1 thrust faults and associated folds for the eastern Cape Smith Belt. Note distinction between regular stacking sequence thrust faults and out-of-sequence thrust faults. Boxed letters identify thrust sheets referred to in text.

be no doubt that mafic and ultramafic units in thrust sheets W to CC constitute a dismembered and metamorphosed ophiolite suite. The ophiolite is now in part structurally inverted as a result of regular and out-of-sequence D_1 thrusting. The ophiolite suite in the Lac Watts area comprises, from structural bottom to top, the following units: (1) deep water laminated graphitic pelites; (2) sheeted mafic and ultramafic dykes, gabbroic sills and massive basalt flows cut by dykes, pillowed basalts and sills cut by dykes; (3) mafic/ultramafic cumulates. This ophiolite suite, containing most of layers 1, 2 and 3 of the oceanic crust (Fig. 9), is here informally named the Purtuniq ophiolite.

Tectono-stratigraphic evolution

The tectono-stratigraphic record of the Cape Smith Belt documents the evolution of an Early Proterozoic epicontinental rift which ultimately led to formation of oceanic crust (Hynes and Francis, 1982). The continent-derived sediments of the lower Povungnituk Group document opening and infilling of a basin, at least in part overlying continental crust. Deposition of the clastic sediments in an approximately east-west-trending basin with water depth increasing northward is documented by: (1) the spatial distribution of channel sands, proximal and distal fan deposits, pelagic units; and (2) limited paleocurrent observations in the autochthon and in basal and intermediate structural level thrust sheets.

The epicontinental rift setting for the accumulation of the Povungnituk Group is supported by the similarity of upper Povungnituk Group mafic magmas to modern, within-plate continental tholeiites with respect to overall major element ratios, ranges in TiO_2 content and trace-element ratios (Hynes and Francis, 1982; Francis et al., 1983). The continental-rift nature of the upper Povungnituk Group is consistent with the observation of the contemporaneous eruption of both basalt and rhyolite (St-Onge et al., 1986).

The Chukotat Group, confined to intermediate structural level thrust sheets, records a phase in the magmatic evolution of the Cape Smith Belt which involved formation of transitional oceanic crust. Chemically the Chukotat Group ranges from komatiitic basalt compositions to low-Ti tholeiites with MORB affinities (Francis and Hynes, 1979; Hynes and Francis, 1982; Francis et al., 1983). The tholeiitic lavas are low in incompatible elements and have trace element characteristics very similar to those shown by modern ocean-floor basalts (Hynes and Francis, 1982). Chukotat Group magmatism may therefore be contemporaneous with the complete rifting apart of continental crust producing transitional-crust which evolved to true oceanic crust.

The structurally highest thrust sheets in the most internal (northern) part of the Cape Smith Belt carry the Spartan Group pelagic sediments, the Rivière du Faucon metabasites (including sheeted dykes, gabbroic sills and pillowed basalts) and the Rivière Déception mafic/ultramafic cumulates. There can be no doubt that these units constitute the imbricated and metamorphosed remnants of Early Proterozoic oceanic crust, preserved in the thrust-fold belt as the Purtuniq ophiolite. As such, the ophiolitic suite completes the tectono-stratigraphic record of the northern margin of the Superior craton, which evolved about 1.9 Ga from an epicontinental rift system to a true oceanic domain.

STRUCTURE

Structural overview

The continental-rift, transitional-crust and ophiolitic suites of the Cape Smith Belt are deformed by three temporally and geometrically distinct sets of structures. The cumulative effect of the D_1, D_2 and D_3 deformation events is to preserve the thin-skinned D_1 thrust-fold belt in an east-west-trending D_2 synclinorium, doubly-plunging as the result of D_3 northwest-trending cross-folds. The earliest set of D_1 structures recorded are regular sequence (foreland- propagating or piggyback style) south-verging thrust faults which root on a basal décollement localized at or near the basement-cover interface (Fig. 13). D_1 folds of bedding are developed in the hangingwall of early D_1 thrusts as the result of movement of the thrusts over footwall ramps (fault-bend folds, Suppe, 1983).

Transport of the thrust belt along the basal décollement during thermal relaxation following the early D_1 imbrication of the cover resulted in the growth of a ductile basal shear zone (BSZ). The syn-thermal peak BSZ represents a foreland- tapering zone of continuous ductile strain along which the thrust belt was transported to the south. Syn- to post-thermal peak out-of-sequence thrust faults ramp through the previously assembled thrust stack to achieve late D_1 crustal thickening. Two syn-thermal peak out-of-sequence faults (O and S, Fig. 13) root on the basal décollement, and are associated with the incorporation of laterally discontinuous basement slices into the thrust belt. (Thrust faults are labelled after hangingwall thrust sheets carried by the faults (Fig. 1,13).)

Thick-skinned D_2 and D_3 folding of basement and cover produced a dome and basin (Type 1, Ramsay, 1967) fold interference pattern at all scales (Fig. 14). The northern margin of the Cape Smith Belt is characterized by locally overturned segments of the D_2 northern synformal limb. The lateral terminations of the overturned segments appear to have controlled the position of subsequent D_3 buckle folds initiated at the basement-cover contact. The shape of D_2 folds in the outliers south of Wakeham Bay (Fig. 1,14) changes relatively abruptly from pinched, angular synforms in the north to rounded open synforms in the south at the ramp in décollement which exposes autochthonous section in the footwall to the south. The distribution of D_3 axial traces (Fig. 14) emphasizes the importance of rheologic layering, such as gabbro and peridotite sills in sediments, in producing buckle folds and determining their wavelength.

Structural and thermal evolution of the D_1 thrust belt

Basal décollement.

The position of the pre-thermal peak basal décollement for the regular stacking sequence D_1 thrusts can be traced throughout the eastern Cape Smith Belt map area. Although the ductile basal shear zone overprints the early D_1 basal décollement, the location of the latter can be inferred from several lines of evidence: (1) pre-thermal peak D_1 regular sequence thrusts are shown to root at the basement-cover contact (eg. south margin, Fig. 13); and (2) early D_1 fault-related folds and rare pre-BSZ minor folds are restricted in occurrence to above the basement-cover contact in the main belt.

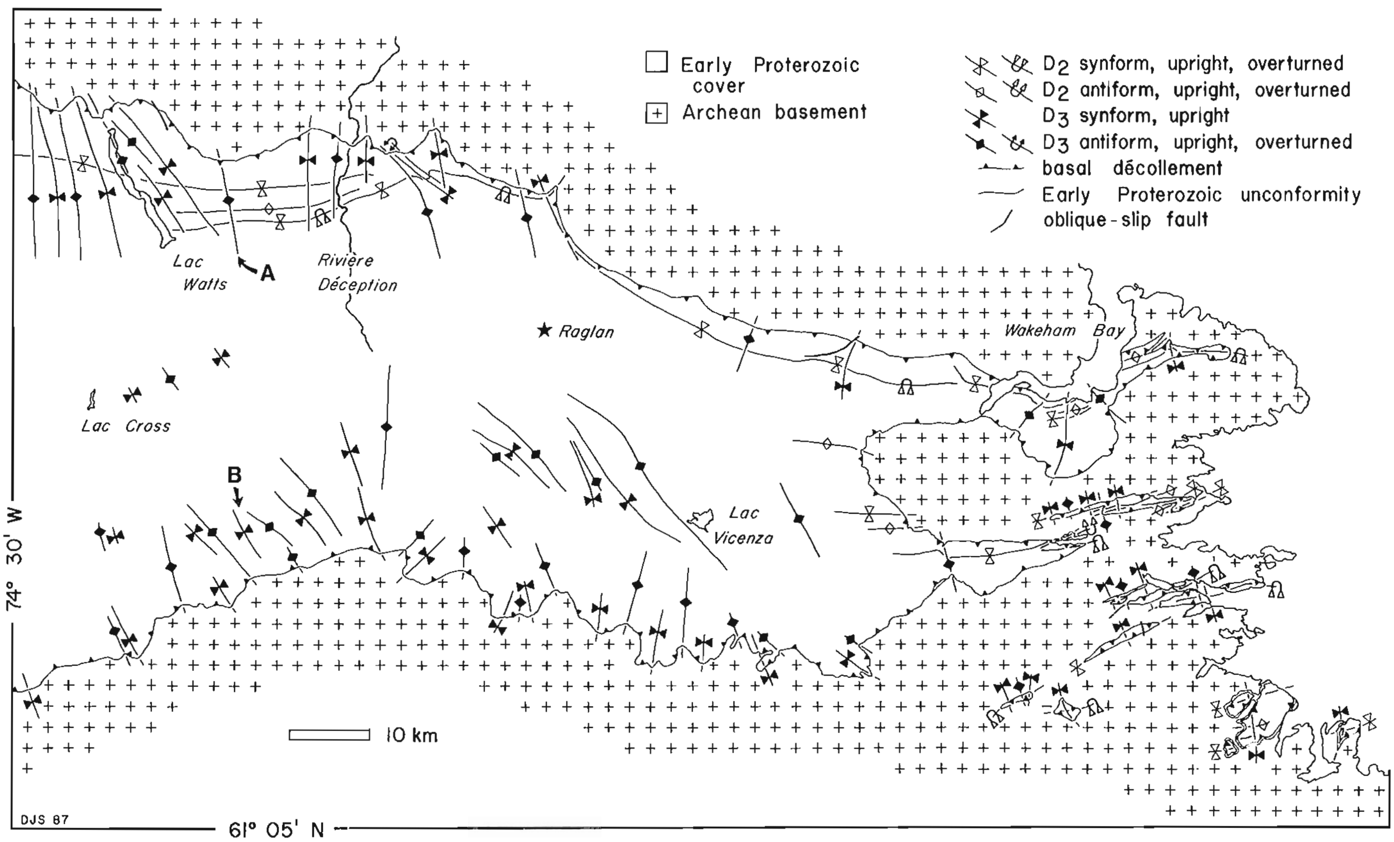

Figure 14. Macroscopic D_2 and D_3 fold axial trace compilation for eastern Cape Smith Belt.

The basal décollement appears to have initially propagated along the basement-cover contact throughout most of the preserved thrust belt, and in general separates footwall Archean tonalite gneisses from the highly strained basal thrust sheets (St-Onge et al., 1986). The extensive ductile deformation and subsequent imbrication of footwall basement gneisses that characterizes the northern margin of the belt in the Rivière Déception - Lac Watts area (St-Onge et al., 1987) is related to syn-thermal peak basal shear zone strain and out-of- sequence faulting; 1-3 m wide strain gradients at the top of the basement slices attest to the existence prior to imbrication of a décollement at the basement- cover contact.

The south margin of the main-thrust-fold belt preserves autochthonous Proterozoic deposits at three locations (Fig. 1). Footwall strain in these autochthonous sections is very low, documented by the undeformed shape of conglomerate and grit clasts and by the absence or poorly developed nature of the D_1 foliation in finer grained sandy beds. These laterally restricted and stratigraphically thin (less than 1 m) autochthonous sections probably represent local paleotopographic depressions above which the décollement propagated, and do not indicate a regional ramping of the basal detachment.

In contrast, the regional step-up of the décollement into the Proterozoic deposits was mapped in the outliers to the south of the Cape Smith Belt proper (Fig. 15). Over 30 m (stratigraphic thickness) of autochthonous sandstone and ironstone are present in the southernmost outliers. The step-up angle for the basal décollement cannot be estimated accurately because of the discontinuous nature of preserved Proterozoic rocks in the outliers (Fig. 1), and because of subsequent D_2 and D_3 folding. The location of the ramp in the décollement varies along strike in the outliers (Fig. 15), possibly in response to lateral changes in the distribution of the thick autochthonous sandstone beds. Footwall strain in the autochthon is limited to a spaced cleavage in the finest-grained semi-pelitic interbeds, in sharp contrast to the highly strained pelites, micaceous sandstone, arkosic sandstone and ironstone above the décollement. Final thermal peak mineral growth of mesoscopically identical garnet-biotite assemblages in semipelites occurred statically in both the autochthon and allochthons of the outliers.

Regular Sequence Thrust Faults and Associated Folds.

Initial imbrication and southward transport of continental-rift (Povungnituk Group), transitional crust (Chukotat Group), deepwater sediment (Spartan Group) and oceanic crust (Lac Watts Group) units occurred along a series of thrust faults which accomplished crustal thickening above the underthrust Superior Province basement. The early D_1 thrust faults, defined by repetition of distinct tectono-stratigraphic packages, in part root on the basal décollement in the preserved belt, and in part are truncated by subsequent out-of-sequence D_1 thrusts (see below).

The southern (external) thrusts imbricating the Povungnituk Group rift sediments are developed in a restricted stratigraphic interval near the top of the basal clastic formation (Formation 2, Fig. 1). At the regional scale, the Povungnituk Group imbricates form two hinterland-dipping duplexes (C-D-E-F, and H, Fig. 1) (Boyer and Elliott, 1982). To the west, the sole thrusts of the duplexes root on the décollement as deeper and more internal structural levels are progressively exposed against the flank of a D_3 antiform (Fig. 13,14; see below).

The roof thrust (I, Fig. 1) to the northernmost of the two duplex structures is a relatively large displacement thrust in comparison to those developed in its footwall (excluding the basal décollement). The hangingwall sequence contains the semipelite and sandstone of Formation 2b, now essentially devoid of ironstone but greatly inflated with gabbro sills and Kattiniq intrusive suite mafic-ultramafic sills. This succession passes stratigraphically upwards into the Povungnituk Group rift basalts (Formation 4, Fig. 1) instead of the shallower water micaceous sandstone formation. Repetition of the sediment and sill (Formation 2b) to basalt (Formation 4) stratigraphic package occurs within four thrust sheets above the roof thrust (Fig. 1,13). Each underlying thrust has a hangingwall glide zone in the semipelite of Formation 2b.

D_1 fault-related folds are found in the hangingwalls of several of the intra-Povungnituk Group thrust faults (J, K, Fig. 13). Restriction of these relatively open, upright folds to the hangingwall of the faults, and their simple geometry as fold pairs (anticline-syncline) and not fold trains suggests that they are fault-bend folds (Suppe, 1983) developed above footwall ramps. The anticline-syncline pair found in the hangingwall of thrust K appears to represent a fold pair developed at the top of the frontal footwall ramp through Povungnituk Group sediments and basalts (A, Fig. 13). In contrast, the syncline-anticline pair documented in the hangingwall of thrust G formed as a result of the passage of thrust sheets over the lateral ramp developed in the footwall at the eastern edge of the southern duplex structure.

The sequence of thrusting during imbrication of the Povungnituk Group rift deposits is clearly piggyback style, with structurally lower faults being younger than structurally higher faults. These early D_1 piggyback style thrusts are termed regular stacking sequence thrusts (St-Onge et al., 1987). Folding of thrusts during passage of the hangingwall

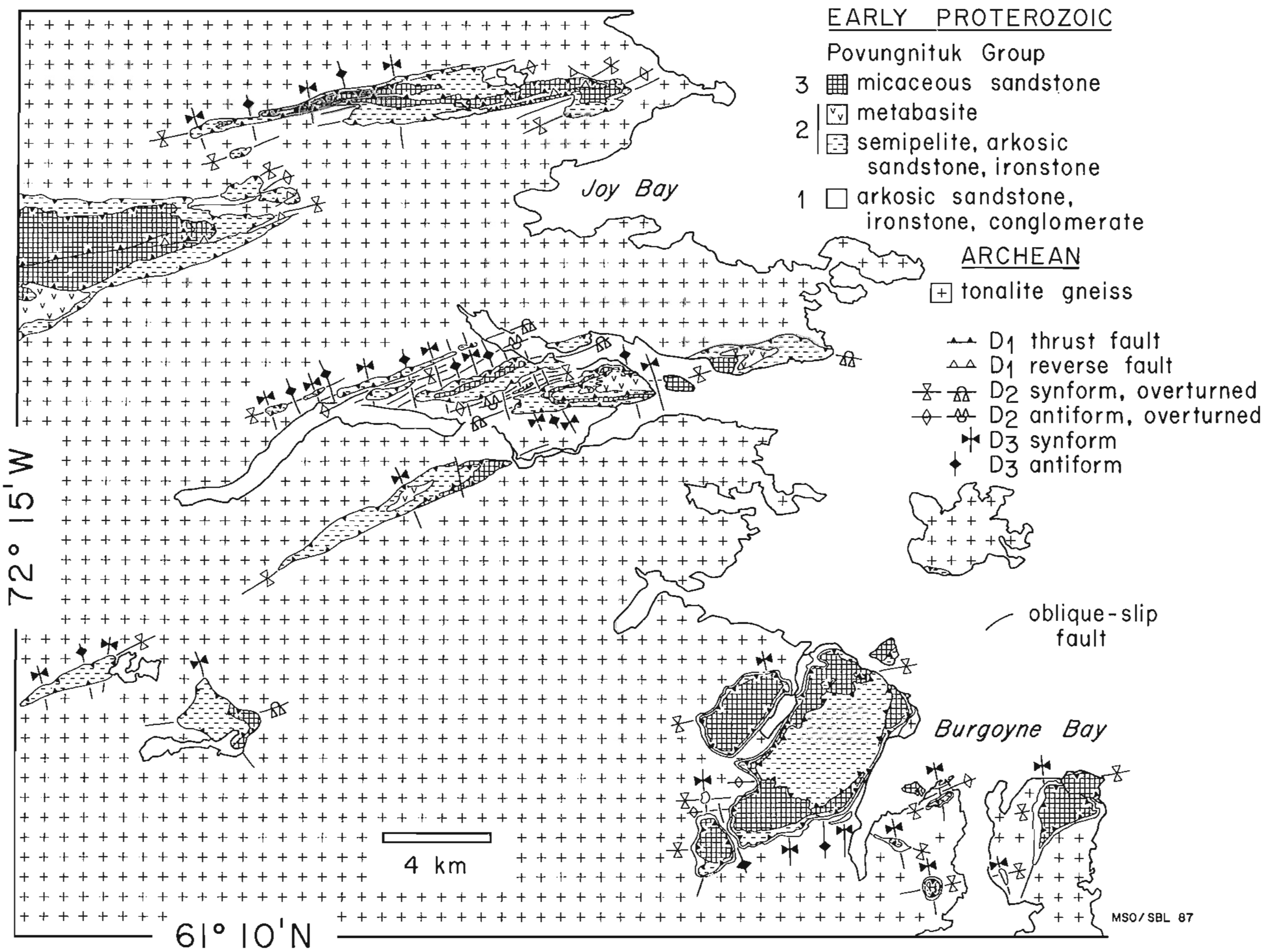

Figure 15. Detailed geological map for Joy Bay - Burgoyne Bay area showing D_2- D_3 fold interference pattern responsible for preservation of the Early Proterozoic outliers. Note ramping-up of basal décollement away from the Archean basement-Proterozoic cover contact in the nine southern outliers.

thrust stack over footwall ramps indicates that structurally higher thrusts became inactive and were deformed during movement on underlying faults. The piggyback-style initial imbrication of the rift deposits is consistent with that documented for the majority of collisional mountain belts (eg. Southern Canadian Rocky Mountain foreland, Price, 1981; Helvetic nappes, Ramsay, 1981; and Idaho-Wyoming belt, Dixon, 1982, Royse et al., 1975).

The pre-thermal peak timing of the intra-Povungnituk Group thrusts is documented by several important observations: (1) thermal peak biotite porphyroblasts overgrowing the post-thrusting S_1 cleavage in the hangingwall sediments of thrusts J and K; (Fig. 13) and (2) deformation of thrusts during syn-thermal peak ductile movement along the basal shear zone. As a consequence, the early D_1 crustal thickening by thrusting is thought to be responsible for the subsequent syn-D_1 thermal relaxation involving heating of the thrust belt and lower plate, and development of the ductile basal shear zone (Lucas and St-Onge, 1986; St-Onge and Lucas, in press).

Initial emplacement of transitional-crust (Chukotat Group) and oceanic crust (Lac Watts Group) thrust sheets occurred along what are interpreted to be early D_1 (pre-thermal peak) regular sequence thrust faults (eg. thrusts N, P, and CC, Fig. 1,13). Their pre-thermal peak age is suggested because (1) these faults are truncated by syn-to-post-thermal peak out-of-sequence thrust faults; and (2) they are deformed at deep structural levels by the syn-thermal peak basal shear zone (see below). As a result of out-of-sequence faulting and the D_2 folding of basement and cover, the relationship of these more internal regular sequence thrusts to the basal décollement cannot be determined. A 0.5 to 1.0 m wide high strain zone along the base of the Chukotat Group olivine-phyric basalts (thrust sheet N, Fig. 1) provides further evidence, in addition to cut-offs of hangingwall and footwall mafic-ultramafic sills (St-Onge et al., 1987), of a fault contact separating the transitional crust (Chukotat Group) basalts from the Povungnituk Group rift sediments and basalts. The relatively narrow zone over which the Chukotat basalts were sheared and consequently foliated is consistent with lower temperature (pre-thermal peak) conditions during thrust movement.

The Lac Watts Group thrust sheets were emplaced on pelagic sediments of the Spartan Group and underlying volcanic rocks. Thrusts T and CC (Fig. 1,13) are apparently the only regular sequence faults remaining that record the initial emplacement of the ophiolite. The persistent occurrence of the RDC mafic-ultramafic cumulates as the lowest stratigraphic unit in the ophiolite suggests that the basal ophiolite thrust fault (CC, Fig. 1) was rooted in the cumulates. The fault zone associated with the regular sequence thrusts T and CC has been reworked at syn-thermal peak amphibolite grade conditions during basal shear zone ductile deformation.

Basal Shear Zone.

A foreland-tapering zone of continuous ductile strain occurs at the base of the thrust belt ranging from 500 m in thickness at the southern (foreland) margin to over 5 km at the northern margin. The basal shear zone (BSZ) incorporates all but the most competent of Proterozoic rocks and from <1 m to >100 m of Archean basement gneisses in penetrative ductile shear (St-Onge et al., 1986; Lucas and St-Onge, 1986). The BSZ was traced from the outliers south of the belt to the northern margin zone of basement imbrication and out-of-sequence faulting (Fig. 13). Along the northern margin of the belt, extensive basement involvement in the BSZ (>100 m structural thickness as documented by St-Onge et al., 1987) terminates at the intersection of the out-of- sequence thrust fault with the basal décollement (thrust S, Fig. 13). To the east, basement involvement is restricted to <3 m of ductile reworking and transposition of Archean fabrics into parallelism with the BSZ mylonitic foliation developed in the cover units. Kinematic indicators in the northern margin BSZ, including C/S fabrics (Berthé et al., 1979) and shear band foliation (Platt and Vissers, 1980; White et al., 1980) consistently indicate a top-side-to-south sense of shear for both cover and basement. The magnitude of the shear strain associated with BSZ deformation is attested to by the occurrence of sheath folds (Cobbold and Quinquis, 1980) in clastic sediments (Fig. 16).

The extent of the BSZ is well defined along the southern margin of the Cape Smith Belt. Abundant gabbro sills and arkosic sandstone beds serve as excellent markers of relative strain state in the thick southern Formation 2b section. The structurally lowest gabbro sills are highly foliated and well lineated, in sharp contrast to essentially undeformed gabbro sills only several hundred metres structurally higher. The arkosic sandstone beds mirror the strain gradient documented in the gabbro sills, with bulk strain increasing down towards the basement-cover contact. In total, about 500 m of structural section above the basal décollement are involved in the BSZ to the west of Lac Vicenza (Fig. 1). To the east of Lac Vicenza, the BSZ includes the basal Formation 2b - Formation 3 imbricates but does not extend above thrust I.

The timing of BSZ deformation is clearly indicated by the internal structure of garnet porphyroblasts in semipelites found near the basal décollement in the outliers south of the belt. Garnet cores show helical inclusion trails related to south-directed BSZ bulk shear, confirming previous observations by St-Onge et al. (1986, 1987) of syn-thermal peak BSZ movement. In addition, these garnet display massive rims, indicating that BSZ strain decayed prior to final growth

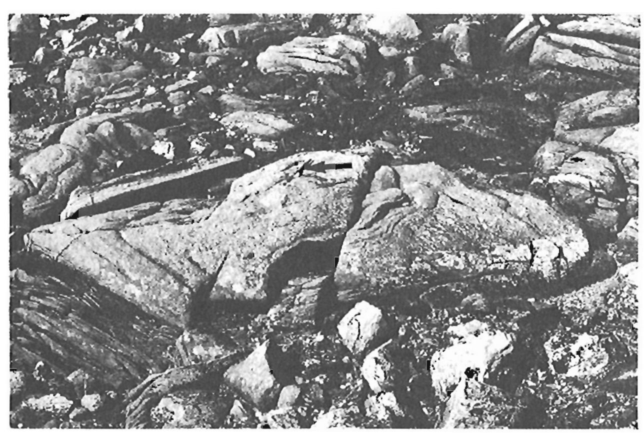

Figure 16. Basal shear zone sheath fold of Spartan Group sandstone, east of Rivière Déception. Arrow points to 15 cm long pen. GSC 204234-M

of thermal peak mineral assemblages. Observations concerning the BSZ amassed during the 1987 field season are consistent with the interpretation of the BSZ as a zone of continuous ductile strain developed at the base of the thrust belt in response to continued slip along the basal décollement during thermal relaxation following initial D_1 crustal thickening (Lucas and St-Onge, 1986; St-Onge and Lucas, in press).

Out-of-sequence Thrust Faults and Basement Imbrication

The thrust belt geometry is further characterized by three dominant systems of out-of-sequence faults. Out-of-sequence faults are simply defined as those which develop in the hangingwall of an earlier formed thrust (Butler, 1982). Out-of-sequence faults in the Cape Smith Belt are recognized by: (1) thrust faults cutting down section in the transport direction in either the hangingwall or footwall; (2) truncation of footwall structures (e.g. thrust faults, folds); and (3) growth of distinct syn- or post-thermal peak metamorphic assemblages in discrete high strain zones associated with individual thrusts (St-Onge and Lucas, 1987).

The three systems of out-of-sequence faults, illustrated in Fig. 13, show a hinterland-younging age progression. The oldest out-of-sequence faults (thrusts O and S, Fig. 13) are syn-thermal peak thrusts which reimbricate previously assembled thrust sheets of the Povungnituk and Chukotat groups. Sitting structurally above these thrusts and forming a klippe in the Lac Watts - Rivière Déception area is post-thermal peak thrust W-X-Y, placing the Rivière du Faucon metabasites or the Rivière Déception cumulates on the Spartan Group sediments. A second post-thermal peak out-of-sequence thrust (BB, Fig. 13), carrying the Rivière Déception cumulates occurs in the hangingwall of thrust W-X-Y and also forms a klippe in the Lac Watts area. Faults W-X-Y and BB are two out-of- sequence faults which belong to a regional set that results in the structural inversion of the Purtuniq ophiolite stratigraphy (Figs. 9,13). Both of the post-thermal peak faults have retrograde (post-thermal peak) actinolite-chlorite high strain tectonic schists (St-Onge and Lucas, 1987) developed over 1 to 3 m in their hangingwalls. It is postulated that the structurally highest thrust (BB, Fig. 13) represents the fault that moved last in the exposed thrust belt. However, a metamorphic study of fault zone assemblages is required to confirm the relative age of these two out-of-sequence thrusts. The age progression of the out-of-sequence faults, younging structurally upwards and towards the hinterland, if correct, is striking in that it is directly opposite to that recorded for the regular stacking sequence faults.

The syn-thermal peak thrusts (O and S, Fig. 13) clearly demonstrate their out-of-sequence character along their southern surface traces. Thrust O truncates the northern limb of the fault-propogation fold syncline found in its footwall, cutting off earlier formed regular sequence thrusts (Lac Cross area, Fig. 1,13). Similarly, thrust S truncates a regular sequence thrust in its hangingwall, and inverts the early stacking sequence by placing thrust sheets of Povungnituk Group arkose and alluvial conglomerate or proximal fan sediments with sills (thrust sheet Q, Fig. 1) on Chukotat Group basalts. Thrusts O and S merge east of Raglan; clarification of the relative ages of these two syn-thermal peak faults must await microscopic examination of mineral assemblages and deformation microstructures associated with their movement. The single thrust north of their merge point (termed thrust O-S, Fig. 13) clearly roots on the basal décollement (Fig. 17). The occurrence of syn-thermal peak (syn-BSZ) basement imbricates to the west rooting on the out-of-sequence décollement segment (Fig. 17) indicates that interleaving of basement and cover in the Cape Smith Belt occurs as a consequence out-of-sequence faulting. Further, it appears that the demise of the regular sequence basal décollement (at the basement-cover contact) and the propagation of out-of-sequence faults occurred following: (1) heating of the thrust belt and the lower plate; (2) hydration and ductile shear of the lower plate beneath the décollement; and (3) initiation of uplift and erosion of the thrust belt. As a result, the basement-involved out-of-sequence faulting may have been in response to a requirement to re-thicken the thrust wedge following hinterland erosion, and to a decrease in the rheological contrast at the early basement-cover contact décollement position.

Syn-thermal peak, late D_1 out-of-sequence faults south of Wakeham Bay (Fig. 15) are reverse faults of limited displacement which place Archean tonalite gneisses on Formation 2b units. These faults are unequivocally D_1 structures since they are deformed by D_2 map-scale folds involving basement and cover.

D_2 fold geometry

The map-scale D_2 fold geometry at deep structural levels in the Wakeham Bay area is characterized by a W-shaped profile involving both basement and cover (Fig. 14). At higher structural levels to the west, the southern D_2 antiform- synform pair decreases in amplitude and eventually become a simple north-dipping monocline. In contrast, the northern synclinal keel appears to step north in an *en échelon* fashion near Raglan (Fig. 14). Its eastern segment (from Wakeham Bay to Raglan) is a tight to isoclinal fold structure with a steeply dipping axial surface which becomes overturned along one small interval west of Wakeham Bay. The western segment of the northern keel (from Lac Watts to east of Rivière Déception) also changes in character along strike, from a relatively open synform with an upright axial surface (Lac Watts area) to an overturned synform with its axial surface dipping at approximately 30° to the north (east of Rivière Déception). The overturned segments of the northern D_2 synformal keel are laterally discontinuous structures reflecting local oversteepening of the basement-cover contact along the northern synformal limb. Local overturning of the basement-cover contact may reflect lateral inhomogeneities in the growth rate of the northern basement-cover antiform, possibly in response to local centres of increased gravitational enhancement of D_2 buckling.

Outliers of early Proterozoic rock south of the belt proper are preserved in a D_2-D_3 dome and basin fold interference pattern (Fig. 14,15). D_2 map-scale folds have a 0.5 to 2.0 km wavelength, and are highlighted because each outlier is preserved in one or more doubly-plunging D_2 synformal structure. The style of D_2 macroscopic folding shows a sharp north to south change coinciding approximately with the position of the footwall ramp in the basal décollement (Fig. 14,15). North of this change in D_2 fold geometry, the map-scale D_2 folds are asymmetric, south- verging structures with the Proterozoic units preserved in the pinched keels. These D_2 synforms have steep to overturned northern limbs, and moderately north- dipping southern limbs. In contrast, the D_2 structures to the south, where up to 30 m of autochthonous sandstone and ironstone are present, are broad, open folds with neither limb dip exceeding 30°. The north to south abrupt change in style of the D_2 folding is thought to reflect primarily the change in the rheological structure of the Proterozoic rocks with the appearance of autochthonous massive sandstone beds. In the northern set of outliers, a penetrative D_1 foliation in the cover allochthons and in the upper few metres of transposed basement was responsible for more cuspate, chevron-style fold profiles during D_2 buckling. The massive sandstone beds of the autochthon and the underlying nontransposed basement in the southern set of outliers appear to have controlled the shape and amplitude of the D_2 folds, producing relatively rounded fold profiles. Fold tightening strain was accommodated in the basement gneisses in the northern pinched D_2 synforms by development of a spaced D_2 cleavage over a 10 m-scale interval below the décollement. No D_2 fabrics were observed in the cover or basement rocks of the southern outlier group.

D_3 fold geometry

D_3 folds were recognized at all scales from 100 km wavelength macroscopic to 1 cm wavelength mesoscopic scale. The map-scale D_3 fold geometry is characterized by thick-skinned, basement-involved folds at two dominant wavelengths: (1) belt-scale 50 km wavelength folds (eg. A, Fig. 14); and (2) tectono-stratigraphically-controlled 5-10 km wavelength folds (eg. B, Fig. 14). The smaller wavelength folds, coaxial with the longer wavelength set, appear to have developed on the limbs of the belt-scale folds during D_3 deformation.

The 5-10 km wavelength D_3 folds are restricted to tectono-stratigraphic levels which contain rheologic layering to buckle during D_3 deformation. The two tectono-stratigraphic levels which buckle during D_3 are: (1) the basement-cover interface (both north and south margins, Fig. 14); and (2) the thrust sheets of Povungnituk Group sediments which contain gabbro, peridotite and layered gabbro-peridotite sills and occur above the south margin to the west of Lac Vicenza (Figs. 7, 14). The rheological contrast across the basement-cover interface, and the rheological stratification of relatively strong, little deformed sills in the dominantly fine-grained sediments of the Povungnituk Group appears to be responsible for the D_3 deformation being accommodated by buckling at these tectono-stratigraphic levels. The only D_3 cleavages observed were found in fine- grained sediments in the cores of tighter D_3 map-scale folds (eg. D_3 antiform- synform pair immediately west of Lac Vicenza, Fig. 14). D_3 shortening at tectono- stratigraphic levels lacking adequate rheologic layering to buckle at the 5-10 km wavelength-scale was most likely accomplished by some form of inhomogeneous grain-scale deformation and/or mesoscopic-scale folding.

IMPLICATIONS FOR RESEARCH AND MINERAL EXPLORATION

The completion of mapping of the eastern Cape Smith Thrust-Fold Belt allows the implications of regional tectono-stratigraphy, structural geometry and tectono-thermal evo-

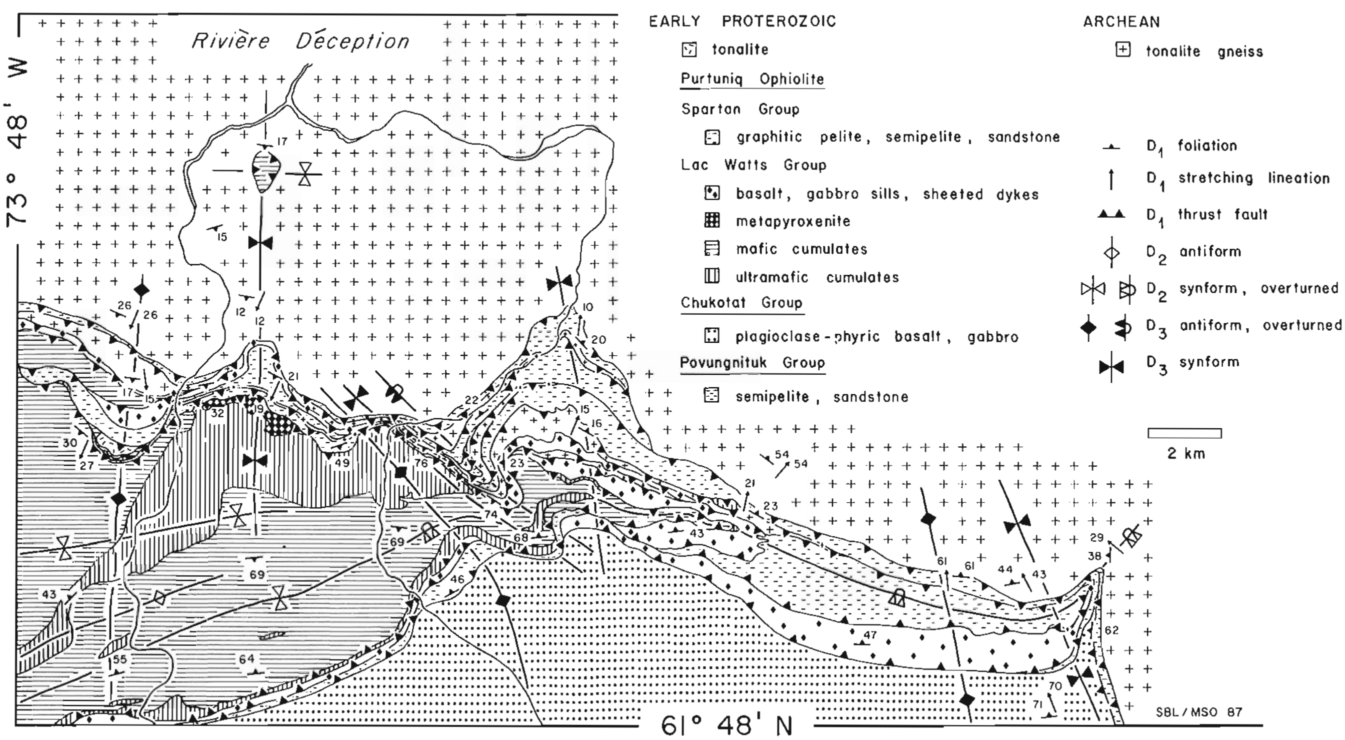

Figure 17. Detailed geological map for Rivière Déception area showing the northern margin basement imbricates.

lution to be evaluated in terms of mineral exploration and further geoscientific research. Identification of 0.5 cm to 3 cm wide chromite seams in the ultramafic cumulates of the Purtuniq ophiolite introduces the possibility of economic potential related to the areally extensive ophiolite thrust sheets. The Cu-Ni sulphide deposits and showings, often with elevated PGE values, are predominantly associated with the Kattiniq intrusive suite layered sills in thrust sheet M (Fig. 1). The along-strike extent of thrust sheet M is structurally controlled by thrust O, a syn-thermal peak out-of-sequence fault which eliminates the thrust sheet to the east of Raglan (Fig. 1,13). Thrust sheet M is folded into an upright D_1 syncline, cored by a Chukotat Group thrust sheet along most of its length, as a consequence of the propagation of the out-of-sequence fault O (Fig. 13). The northern limb of the syncline is cut out by out-of-sequence fault O near Kattiniq (Fig. 1, 13). The obvious structural control on the occurrence of the thrust sheet containing the mineralized sills emphasizes the implications of our study of both the structural architecture and evolution of the Cape Smith Belt for mineral exploration.

The detailed description of the tectono-stratigraphic transition from continental-rift margin to true oceanic crust, and the documentation of the structural geometry and tectono-thermal evolution of the D_1 thrust belt constitute important results from the final summer of field work in the eastern Cape Smith Belt. These results, combined with a series of U-Pb age dates on rift margin, transitional crust, ophiolitic and syn-D_1 plutonic units will allow for the unparalleled study of the birth of an Early Proterozoic ocean basin and its demise in a collisional mountain belt. Any future geoscientific investigation north (hinterlandward) of the Cape Smith Belt in the potential root zone for the thrusts and the ophiolite (Sugluk Inlet area of northern Quebec) will lead to further understanding of not only the geometrical and tectono-thermal evolution of this collisional belt but that of the thrust belts and root zones of collisional mountain belts in general.

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Geology of North River-Nutak map areas, Nain-Churchill provinces, Labrador

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Ermanovics, I.F., Van Kranendonk, M., Corriveau, L., Bridgwater, D., Mengel, F., and Schiøtte, L., Geology of the North River-Nutak map areas, Nain-Churchill provinces, Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 19-26, 1988.

Abstract

Remnants of supracrustal rocks, amphibolite, mafic and ultramafic rocks in feldspathic orthogneiss known to be at least 3.5 Ga old are correlated with established successions of similar age in the Saglek-Hebron fiord areas. Inferred protoliths of the remnants include pelite, psammite, impure quartzite, magnetite-iron-formation, mafic volcanic rocks and associated wacke, gabbro, and layered mafic and ultramafic intrusions including anorthosite.

In central Nain Province late Archean granulite and retrograded granulite facies to amphibolite facies metamorphism coincides with the second (D_2) of three periods of deformation. Upper amphibolite facies metamorphism associated with D_1 and D_3 deformation is largely masked by pervasive high-temperature D_2-activity.

Western Nain craton was restructured along with Proterozoic Ramah Group equivalent metasediments in the Nain-Churchill boundary zone. Eastern Nain craton preserves Proterozoic Mugford Group sediments, sills and lavas in down-faulted blocks of Archean basement rocks with intrusions of Paleohelikian granite.

Résumé

On a établi une corrélation entre des vestiges de roches supracrustales, d'amphibolites, de roches mafiques et ultramafiques présentes dans un orthogneiss feldspathique d'âge connu (au moins 3,5 Ga), et des successions d'âge similaire identifiées dans les régions de Saglek-Hebron Fiord. Parmi les roches originelles dont proviendraient ces vestiges, figurent des pélites, des psammites, des quartzites impurs, des formations ferrifères à magnétite, des roches volcaniques mafiques et d'autre part des grauwackes, des gabbros et des intrusions mafiques et ultramafiques statifiées, qui sont associés aux précédentes, y compris des anorthosites.

Dans la province centrale de Nain, la phase de métamorphisme d'âge archéen supérieur qui a créé le faciès des granulites, puis a causé le passage rétrograde du faciès des granulites au faciès des amphibolites, coïcide avec la seconde (D_2) de trois périodes de déformation. Le métamorphisme qui a créé le faciès supérieur, des amphibilites et qui est associé aux déformations D_1 et D_3, a été largement masqué par l'activité généralisée de D_2 qui résulte d'une température élevée.

Le craton ouest de la provincee de Nain a été restructuré en même temps que les métasédiments équivalents du groupe protérozoïque de Ramah dans la zone limitrophe des provinces de Nain et de Churchill. Le craton de l'est de la province de Nain contient encore des sédiments, filons-couches et laves du groupe protézoïque de Mugford présente dans des blocs du socle archéen, rejetés en profondeur par des failles normales, ainsi que des intrusions de granite paléohélikien.

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INTRODUCTION

Geological mapping during July and August 1987 covered approximately 1800 km^2 at scales suitable for publication at 1:100 000. The project area (Fig. 1) is bounded by latitudes 57°30'-58°00'N and extends from the Labrador Sea (Nutak map sheet NTS 14F) westward into North River map sheet (NTS 14E). The study area adjoins Hebron map sheet (NTS 14L) under investigation by B. Ryan, Newfoundland Department of Mines and Energy, Mineral Development Division, and was mapped previously at 1:250 000 by Taylor (1977).

The project, scheduled for three field seasons, aims to produce a map at 1:100 000 scale that will allow assessment of georesources to be made from a more detailed data base. Fieldwork during 1987 established that, although correlations are possible with lithologies from the adjacent Saglek area, the tectonic history of Nain Province needs to be established, the succession of events along the Nain-Churchill boundary zone require corroboration (c.f. Korstgard et al., 1987), and isotopic age determinations are needed to constrain the timing of complicated kinematics. Although remnants of Ramah Group were discovered in the Nain-Churchill boundary zone, the rocks are probably mainly deep-seated and as such may provide a study area of deep crustal tectonics not available in the central Torngat orogen (Mengel, 1987) to the north.

PREVIOUS WORK

Taylor (1977, 1979) established that Nain Province is underlain by amphibolite and granulite facies Archean felsic plutonic rocks, layered gneiss, metasediments, amphibolite and ultramafic rocks. In Saglek and Hebron Fiord areas (NTS 14L) rocks of Nain Province are a complex of polydeformed supracrustal and plutonic associations formed 3.8 to 2.5 Ga ago (Hurst et al., 1975; Bridgwater et al., 1975; Collerson and Bridgwater, 1979; Collerson et al., 1982; Ryan, 1977; Ryan et al., 1983, 1984). A key to the chronology is the Saglek mafic dykes that cut rocks older than 3.6 Ga. Ancient Archean rocks were first recognized in Nain Province at Lost Channel of Mugford Bay in the present map area, where Hurst (1973) obtained a zircon $^{207}Pb/^{206}Pb$ minimum age of 3460 Ma that subsequently led to extensive isotope age dating in the Saglek and Hebron areas.

Rocks of Nain Province were progressively restructured, metamorphosed and juxtaposed westward against high grade feldspathic sillimanite-garnet mylonite (Tasiuyak gneiss, Wardle, 1983, 1984) during late Archean and early Proterozoic deformation (Taylor, 1979). Mylonite formed during collision of Churchill Province (Rae Terrain, P.F. Hoffman, pers. comm., 1987) with Nain craton along what is now recognized as the Abloviak shear zone, composed of Tasiuyak gneiss and Nain protoliths (Ryan, 1984).

GENERAL GEOLOGY

Mapping proceeded systematically in NTS 14E/9 and in parts of adjoining map sheets. North-south and east-west transects of well exposed bedrock also provided an overview of the area, which showed that Nain Province rocks (ca. 3600 km^2) are composed of a number of lithologically and structurally distinct, commonly fault-bounded, domains. Nevertheless, lithological correlation from Saglek, the reference locale, southward to Hebron (Ryan, 1977), and from there to Okak (latitude 57°30', southern part of present project) is apparently possible.

Nain Province is unconformably overlain by the flat-lying Proterozoic Mugford Group along the Labrador Sea coast (Kaumajet Mountains) in a graben or half-graben. In this domain early Archean prograde amphibolite facies is locally preserved in the basement rocks. Fifty kilometres west of the coast, rocks of Nain Province were progressively restructured during late Archean and early Proterozoic activity that included deposition and deformation of the Proterozoic Ramah Group. The southern part of the area and the faulted

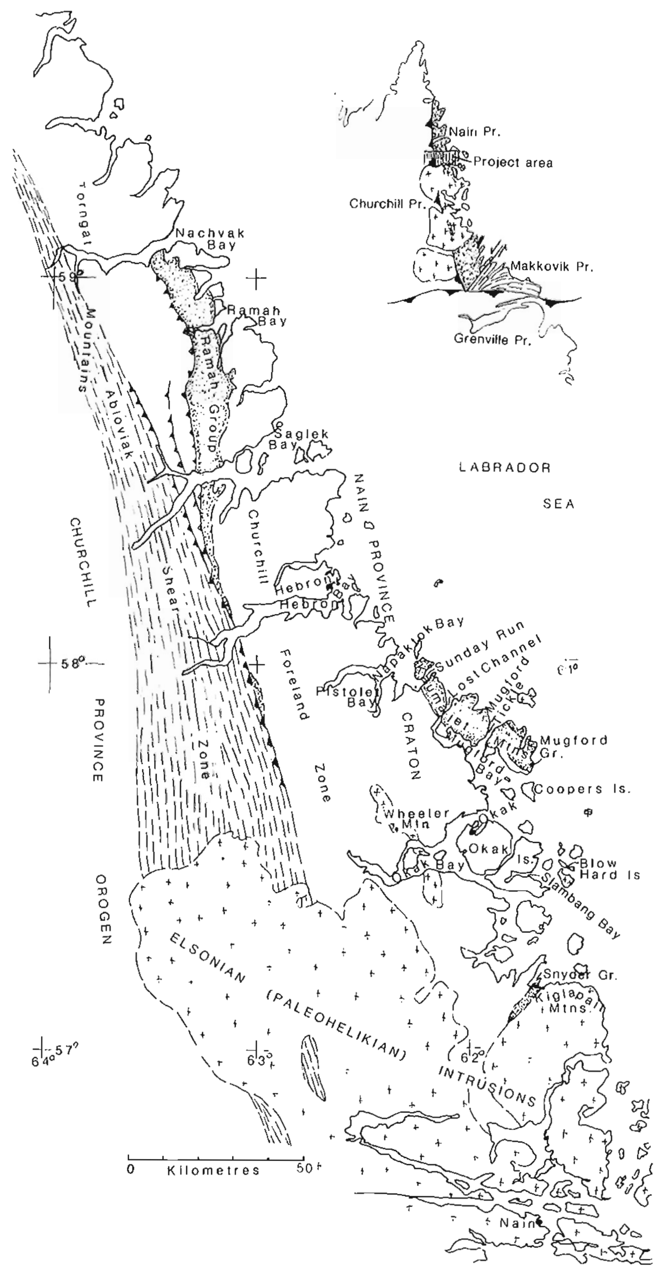

Figure 1. Location of project area and major tectonic units.

terrane around the Kaumajet Mountains are intruded by Paleohelikian igneous rocks. The Wheeler Mountain granites extend from the southern Paleohelikian terrane as an isolated ridge into the central part of Nain Province (Taylor, 1977).

NAIN PROVINCE

Seventy per cent of Nain Province is underlain by homogeneous to heterogeneous, leucocratic to mesocratic, metaplutonic orthogneiss and migmatite of several ages. Metasediments, amphibolite and ultramafic rocks occur as fist-size inclusions and as mappable units (greater than 50 m wide) to make up the remaining 30 %. At least three pervasive Archean deformations are recorded in these rocks and although granulite facies metamorphism of assumed late Archean age masks most of the early events, pre- and post-granulite facies metamorphism was nevertheless detected.

Metaplutonic rocks

Orthogneiss, mainly granodioritic and tonalitic in composition, constitutes a highly variable complex of multi-phase intrusions and deformation. These rocks may be laced with supracrustal or mafic rock inclusions to form layered gneiss, or they may be schlieric, gneissose, less deformed mixtures of several generations and magma types. Early migmatite may form components in younger migmatite. About 60 % of the metaplutonic complex is layered quartzofeldspathic gneiss which is the oldest member of the complex, because it is cut by metamorphosed mafic dykes (Saglek dykes?). The origin of the early gneiss can only be deduced from clean coastal exposures where relationships with supracrustal and mafic units suggest lit-par-lit injection of gneiss modified by subsequent tectonism.

In areas of pre-granulite amphibolite facies metamorphism, the older gneisses are well layered migmatites with a paleosome of medium grained, grey, granoblastic biotite tonalite-granodiorite gneiss veined by leucocratic material on a 10 cm scale. Mafic dykes cut this migmatite. Well preserved examples occur on the southeast coast of Sunday Run and the southwest end of Mugford Tickle structurally beneath the Mugford Group. In areas of granulite facies metamorphism, however, the early structures are largely destroyed. There the rocks range from medium- to coarse-grained dark, hypersthene tonalitic gneiss with mafic xenoliths, to light coloured flecked gneiss showing retrograde recrystallization during late granulite facies migmatization.

Anorthosite, gabbroic anorthosite and associated mafic to ultramafic rocks occur as small and large inclusions up to 3 km^2 in the older gneisses. At Okak Harbour these are cut by the mafic dykes. The anorthositic rocks (colour index of 1 to 20) are coarse grained, up to 10 cm, hornblende-bytownite $\pm$ clinopyroxene $\pm$ orthopyroxene rocks (Hurst, 1973). The widespread distribution of these mafic rocks suggests derivation from larger layered intrusions.

Younger orthogneiss comprises gneissic granodiorite, tonalite and regional pegmatite swarms that intrude and tend to disperse the older gneisses and supracrustals. Some of this material forms relatively inclusion-free homogeneous masses 2 to 10 km^2, but most occurs as lit-par-lit neosome in structurally planar gneissic migmatite. The size, shape and age of the younger orthogneiss intrusions remains largely unknown, although most intrusions seem to predate granulite facies metamorphism because their leucosomes contain orthopyroxene. A late, grey, coarse grained, post-granulite facies, foliated granodiorite underlies most of Coopers Island and parts of northern Okak Island where it forms sheets tens of metres thick.

An estimated 20 % of the area is intruded by swarms of coarse grained biotite granite and pegmatite sills. These occur predominantly in areas of retrograded granulite facies metamorphism. Individual bodies are tens of metres wide, up to hundreds of metres long and are composite with complex internal structures. Schlieren of grey, fine grained, gneiss, late aplitic veins and sheets that cut part of a pegmatitic phase locally grade, along strike, into pegmatite. The structures suggest synkinematic pegmatite emplacement, with early phases deformed before the next phase was emplaced.

Supracrustal rocks and mafic/ultramafic associations

According to Ryan (B. Ryan, pers. comm., 1986), the following main lithological groups form early Archean Nulliak assemblages between the Saglek and Hebron Fiord areas: iron-formation, marble; garnet-sillimanite paragneiss; mafic gneiss and iron-formation. These supracrustal rocks are intruded by quartzofeldspathic gneiss that also contains inclusions of gabbroic anorthosite, ultramafic rocks and clinopyroxene-quartz- feldspar gneiss. At the type locality, these lithologies are cut by Saglek dykes that are generally plagioclase-phyric. A younger, post-Saglek dyke association of supracrustal rocks, the Upernavik assemblage, comprises: paragneiss, marble, and quartzite; mafic gneiss; ultramafic rocks. The similarity of Nulliak and Upernavik assemblages thus requires the distinguishing presence of Saglek dykes. Accordingly, because the absence of Saglek dykes is not a reliable criterion of age, all the supracrustal rocks are described together with the proviso that there may be more than one sequence.

In the present map area supracrustal and associated mafic and ultramafic rocks occur as fist-sized inclusions and lenses in highly dispersed zones of migmatite gneiss and as thin folded belts generally 50-1000 m wide. The widest of these belts may be as thick as 500 m in section, and most extend several kilometres before they are offset by faults or are 'cut-out' by intrusive rocks.

The supracrustal rocks consist of layered (metre to decametre scale) metasedimentary rocks interpreted to have been pelite, semipelite, psammite, quartzite and impure quartzite, magnetite-quartz iron-formation, and minor carbonate and impure calcareous rocks. These are associated with variegated amphibolite, homogeneous amphibolite, and ultramafic rocks probably derived from volcanic and gabbroic rocks that may have also contained ultramafic flows. Ultramafic and mafic sheets, including anorthositic layers, intrude the metasedimentary rocks. These metasedimentary and meta-igneous rocks are commonly adjacent to, or intercalated with, mesocratic layered gneiss that is discontinuously layered on millimetre- to centimetre-scales, and contains up to 10 %

thicker (decametre- to metre-scale) layers and pods of amphibolite. This layered gneiss is also considered to be mainly supracrustal because of the fine scale layering and because of the common occurrence of layers of quartzite, calc-silicate or other clearly metasedimentary components within it.

The largest and best preserved belt of supracrustal rocks outcrops 3 km south-southeast of Pistolet Bay in the form of a 5 km long × 3 km wide re-folded fold (Fig. 1). Supracrustal rocks within the structure consist of ultramafic and mafic meta- volcanic rocks at the base. These are overlain by garnet-sillimanite- biotite ± graphite metapelite and a 2 m thick layer of iron-formation. Subparallel to the layering in the metapelite unit is a sheet (up to 100 m thick) of hornblende- plagioclase ± garnet amphibolite probably derived from gabbro (colour index 40-80), which shows relict igneous layering and locally contains layers of recrystallized plagioclase phenocrysts up to 5 cm long and 2 cm wide. On the western side of the structure, however, metasedimentary rocks are basal. These include graphite-rich schist and graphitic magnetite-quartz iron-formation, possible stromatolite-bearing rocks, minor calc-silicate gneiss, and garnet-biotite-orthopyroxene paragneiss. The basin is surrounded and intruded by the early tonalitic migmatite orthogneiss that contains structurally concordant remnants of the main basin lithologies (Fig. 2).

A 2500 × 500 m supracrustal, mafic and ultramafic structural remnant outcrops on the Kingnektuk Islands. These rocks, which are folded and in the granulite metamorphic facies, are from west to east:

— mafic/ultramafic layered orthopyroxene-clinopyroxene-hornblende rocks, - 50 m.
— pelite-semipelite and thin quartzite layers; circa 90 to 120 m; sillimanite-garnet-biotite (± cordierite?) gneiss grades to psammitic rocks containing biotite-garnet schist that locally contains iron sulphide horizons; the schist is intruded by 50 m mafic layers containing orthopyroxene-clinopyroxene-hornblende.
— the schist grades to a 400 m thick unit of layered pale brown weathering mesocratic gneiss that contains por-

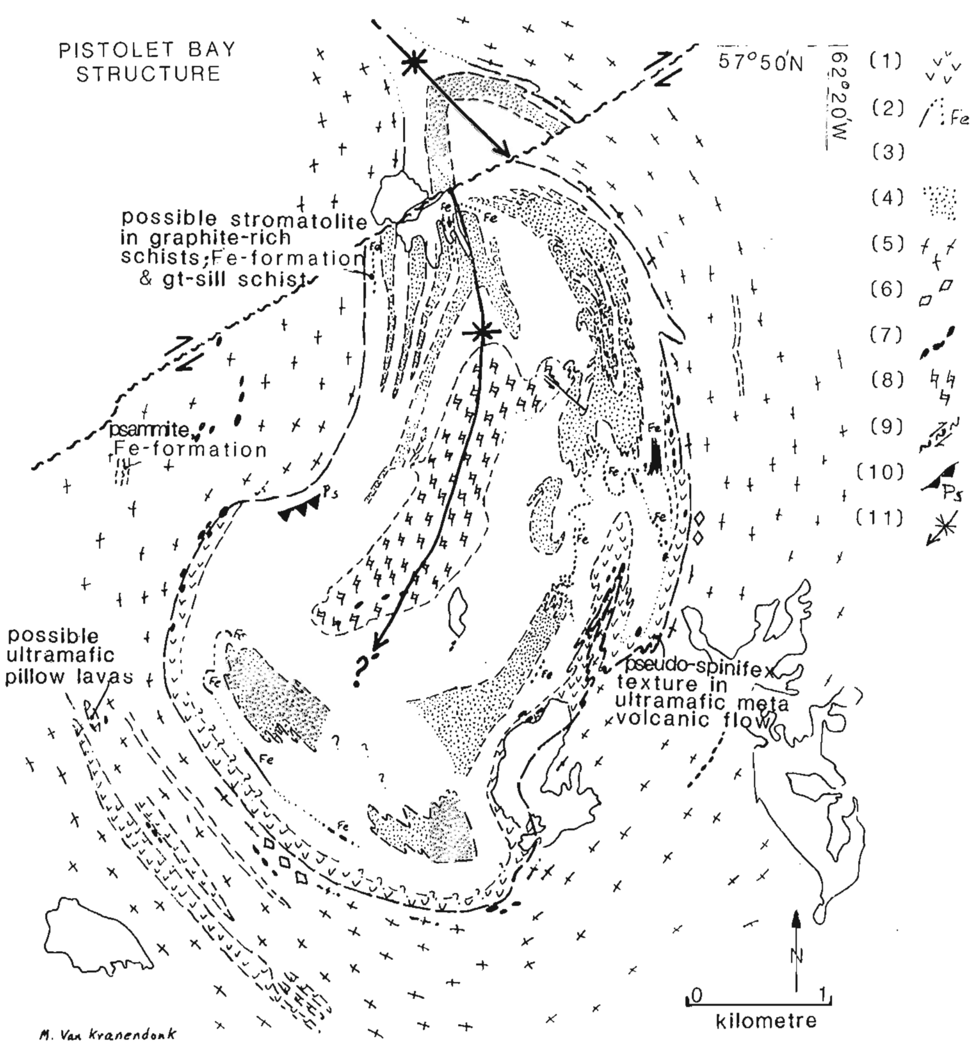

Figure 2. Pistolet Bay structural basin. Legend: 1 - metavolcanics; dominantly layered, variegated amphibolite (orthopyroxene-clinopyroxene-hornblende-garnet), subordinate ultramafic layers and local pillow-lavas. 2 - magnetite iron-formation, (Fe); contains garnet-amphibole, quartz-magnetite ± graphite as contorted poddy trains. 3 - unornamented areas, layered gneiss and metasediments; includes garnet-sillimanite-biotite, garnet-orthopyroxene, garnet-biotite-cordierite and garnet-biotite quartzofeldspathic gneiss; locally contains graphite, calc-silicates and rarely pyrite. 4 - leucogabbro (CI = 40-80); hornblende-plagioclase ± garnet; intrudes metasediments (3). 5 - leucocratic, tonalitic migmatite orthogneiss; long axes of crosses parallel trend of foliation. 6 - inclusion of amphibolite(1) in orthogneiss(5). 7 - ultramafic pods; serpentine ± biotite possibly derived from disaggregated ultramafic dykes. 8 - synkinematic, garnetiferous granite sheet. 9 - fault. 10 - late Archean(?) thrust fault with pseudotachylyte, (Ps). 11 - trend of major D_2 synformal fold axis, and plunge.

phyroblastic crystals of ortho-amphibole and orthopyroxene enclosing plagioclase and minor quartz; ultramafic rocks and green chert occur as scattered elongate pods.

The complete range of association of magnetite-quartz iron-formation within various rocks has not been determined. Iron-formation containing various amounts of magnetite was discovered with psammite, quartzite-pelite and mesocratic layered gneiss. An iron-formation-mafic-ultramafic succession, similar to that seen in the type Nulliak rocks in the Saglek-Hebron Fiord area, outcrops on the southwest coast near Lost Channel. There, thin quartz-magnetite-amphibole $\pm$ hypersthene iron-formation is part of a mafic-ultramafic iron-formation - diopside-rich iron-formation succession. The mafic rocks grade from basaltic through Fe-basaltic compositions in the banded iron-formation repeatedly in 5 m successions.

Metamorphism and structure

Central Nain Province is dominated by D_2-structures and granulite and retrograded granulite metamorphic rocks that yield Rb-Sr whole-rock and U-Pb zircon ages of 2.7 to 2.8 Ga (c.f. Schiøtte et al., 1986; L. Schiøtte, pers. com., 1987). This metamorphism and deformation postdates the Saglek dykes and is a period of major refolding of structures and recrystallization of amphibolite facies assemblages formed 3.6 Ga ago (c.f. Ryan et al., 1983). Amphibolite facies rocks occur beneath the Proterozoic Mugford Group east of a broad 130-140° trending mylonite and brittle shear zone that extends from Shark Gut Island to Blow Hard Island. Structural juxtapositions of granulite and amphibolite facies in the present area may thus be analogous to those in the type Saglek area where the Handy fault also separates these facies.

Preliminary structural and metamorphic analyses suggest three Archean deformations. D_1-deformation developed a planar gneissosity in supracrustal rocks and in the older migmatite orthogneiss (Fig. 3a). In areas of preserved amphibolite facies, mafic dykes cutting these structures are deformed but nevertheless are well preserved. D_2-deformation occurred during granulite facies metamorphism and produced large-scale folds with axial planar fabrics and pronounced mineral lineation parallel to fold axes. In the Pistolet Bay structural basin and in blocks of well preserved granulite facies orthogneiss, D_2-folds are generally steep, tight, isoclinal and overturned on north-trending axes. The plunge of D_2 fold axes is variable, because the folds were refolded during D_3-deformation about open, generally SE trending axial planes. The combination of D_2 and D_3 folding resulted in the basinal form of the Pistolet Bay structure and may account for the refolded pattern of most supracrustal belts in the area.

In a number of domains, D_3-deformation produced pronounced mullion structures in shallowly dipping (15-35°) planes at amphibolite facies (Fig. 3b) and retrograded granulite facies metamorphic grades. Locally stratiform layers of well preserved granulite facies rocks tens of metres thick, and possibly of thrust origin, are bounded by mylonite and amphibolite facies rocks in these shallow-plunging open structures. Thus, some juxtaposed associations of granulite - retrograded granulite - prograde amphibolite facies assemblages may be structural intercalations.

The supracrustals and associated mafic and ultramafic rocks commonly contain the following metamorphic assemblages:

- pelite, semipelite and quartzofeldspathic rocks (psammite): garnet-biotite-sillimanite $\pm$ graphite, garnet-biotite-cordierite, garnet-biotite-orthopyroxene, garnet-biotite
- iron-formation with quartzite and other metasediments: garnet-magnetite, biotite-magnetite, clinopyroxene-magnetite
- iron-formation with mafic and ultramafic rocks: quartz-magnetite-amphibole-orthopyroxene, quartz-magnetite-diopside
- mafic and ultramafic rocks: orthopyroxene-clinopyroxene-hornblende-garnet, orthopyroxene-chrome clinopyroxene-hornblende, serpentine $\pm$ talc $\pm$ biotite $\pm$ orthopyroxene $\pm$ olivine (pseudo-spinifex); cut by orthopyroxene-clinopyroxene-hornblende bearing leucosomes.

Retrograded granulite facies gneisses are characterized by a change from dark grey in granulite to light grey in retrograded rocks in which amphibole, clinopyroxene and biotite rim or overprint and partly or completely replace garnet- orthopyroxene aggregates. Retrograded rocks commonly have a diagnostic flecked or 'blebby' appearance resulting from recrystallization. In several outcrops, however, fresh or variously altered orthopyroxene leucosomes cut layered amphibolite facies gneisses that lack the flecked textures, and thus appear to show no evidence of having been at higher grades of metamorphism. Such gneisses may be interpreted to reflect either early preserved amphibolite facies metamorphism, or dynamic recrystallization following the waning phases of granulite facies metamorphism, or the gneisses may reflect a separate late amphibolite facies metamorphic event.

The most common mode of retrogression in granulite terrane occurs in areas of multiple lit-par-lit pegmatite granite injection (Fig. 2c, 2d). Many of the pegmatites are dark near their margins and locally contain orthopyroxene; small apophyses inject the country rocks which may also contain orthopyroxene. Toward the centre of pegmatite dykes, the feldspars lose their dark colour (except for a few megacrysts) and orthopyroxene is replaced by biotite. Locally, younger pegmatites with or without orthopyroxene cut zoned older pegmatites, suggesting that the presence or absence of hypersthene is controlled by local factors in individual bodies. The overall effect of pegmatite development is retrogression, but locally, the best evidence of granulite facies metamorphism in the retrograded country rocks is in apophyses from the pegmatite or in rootless 'sweats'.

TRANSITION TO THE PROTEROZOIC

In central Nain Province, trends of D_3 mullion structures and refolded D_2 mineral lineations vary from one block of terrane to the next. The cause of these discontinuities may be block rotation brought about by either late granite intrusion, as on Coopers Island or Okak Islands, or by early Proterozoic deformation of the Nain Province margins. Proterozoic activity within Nain Province is represented by intrusion of diabase dykes of several orientations. At the western margin of the craton it is represented by at least two ma-

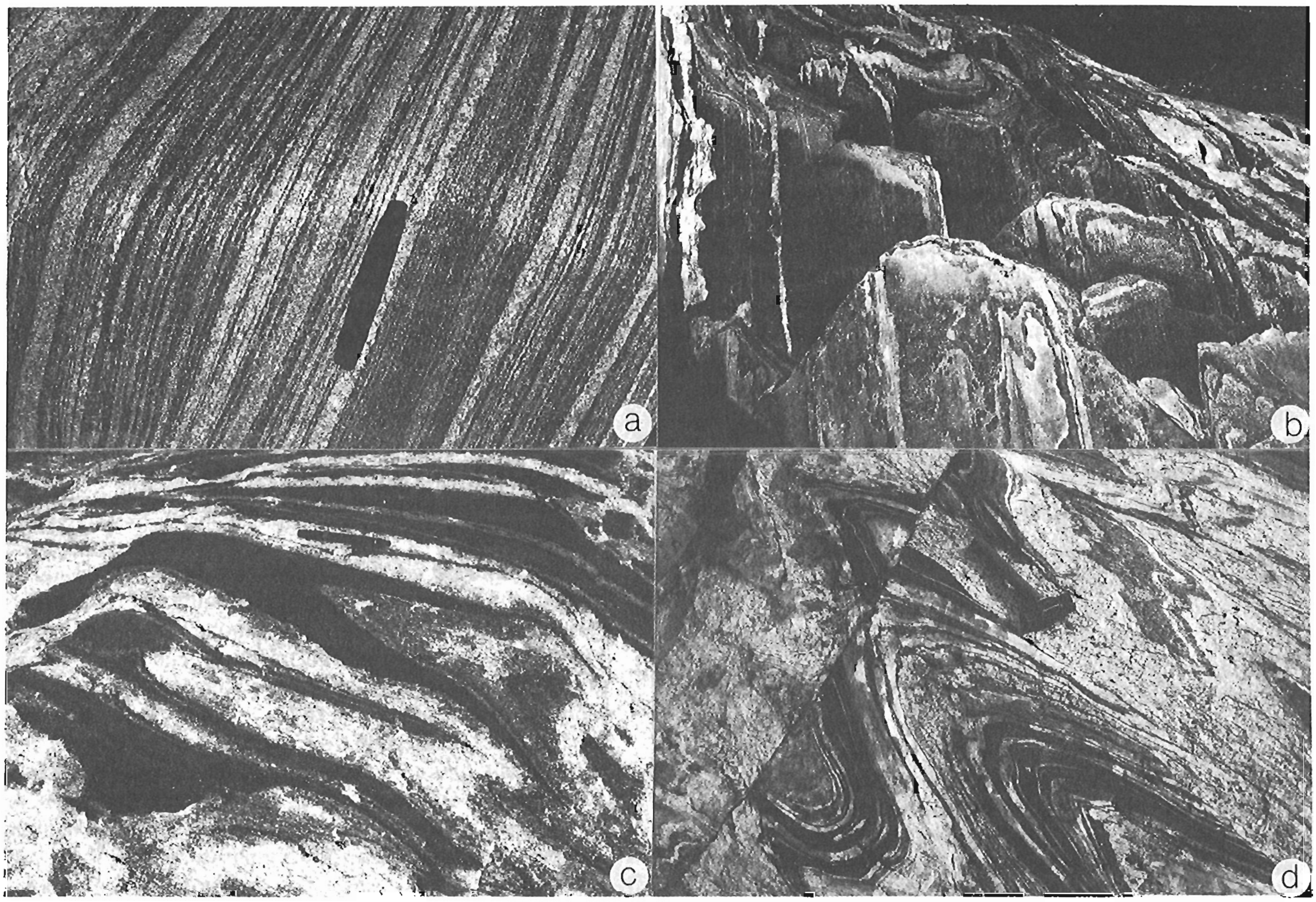

Figure 3. D_3 - deformation structures in amphibolite facies gneisses of Slambang Bay, Okak Islands:
a - tonalitic, migmatite orthogneiss 'straightened' by D_3 extension.
b - L and S fabric formed by extension of D_2-layering.
c - orthogneiss (grey) 'exploded' by and reacting with pegmatitic granite gneiss.
d - brittle deformation following process in (c).

jor deformations, deposition of Ramah Group, and by upper amphibolite facies metamorphism that also alters the diabase dykes. At the coast to the east, Proterozoic activity is represented by deposition of Mugford Group, faulting, and by a number of Paleohelikian bodies of granite.

East

The Kaumajet Mountains, located in the east, consist of up to 1600 m of shale, siltstone chert, volcanic rocks and mafic sills of the Mugford Group (Kranck, 1939; Christie, 1952; Douglas, 1953). This group lies unconformably on regolith on the Archean basement. The Mugford Group is internally faulted but is largely undeformed and unmetamorphosed. However, the southwest margin is in faulted contact with the Archean gneisses along a major northwest-southeast ductile shear system with steeply southeast-plunging lineation that affects the Archean gneisses over a variable width of up to 5 km, or possibly locally more. The group therefore appears to be preserved in a downthrown (west side up) Proterozoic fault block. It is this steep shear deformation which may have caused rotation of domains within Nain Province.

Conglomeratic clastic dykes up to 30 m wide, some of which show graded bedding, occur in the Archean basement as far as 10 km west of the Mugford Group. If these clastic rocks are Mugford Group equivalent sediments, it indicates that Mugford Group sediments were once distributed over a wider area than that in which they are presently exposed. Clastic dykes contain fragments of diabase of presumed Proterozoic age as well as of Archean basement gneiss, suggesting that the diabase dyke material was not of the same source as the Mugford Group lavas, and that the Mugford Group was deposited after intrusion of the diabase dykes. Clastic dykes are locally mylonitized along northwest-trending sinistral shear systems which may be related to the major faulting along the southwest margin of the Kaumajet Mountains.

West

At latitude 57°52′N and longitude 63°W, the Nain-Churchill boundary zone records two separate Proterozoic deformations of unresolved complexity. The older of these deformations has been referred to as the Torngat orogeny (Mengel, 1987), some effects of which are present within the Abloviak shear zone (Korstgard et al., 1987) and the east contact of which marks the Nain-Churchill boundary (Taylor, 1979; Fig. 4). The Abloviak shear zone is characterized by prominent, shallow north-northwest-plunging mineral lineations in all lithological units which include recognizable Nain Province protoliths as well as Tasiuyak gneiss of uncertain origin. The transition from Nain Province protoliths to the Tasiuyak gneiss is gradational over a 5 km wide zone in which these units are interlayered on a 10-300 m scale (Fig. 4). Layering is largely due to tectonic juxtaposition of units across pseudotachylyte breccia zones of the younger deformation but probably originated as an intrusive contact of the Tasiuyak gneiss into Nain Province rocks. All lithologies in the Abloviak shear zone contain granulite facies mineral assemblages: e.g. garnet-biotite ± sillimanite ± graphite in the Tasiuyak gneiss, ubiquitous orthopyroxene in tonalitic gneisses, and in 2-pyroxene-garnet-bearing amphibolites.

The structures of the Abloviak shear zone are characterized by significant north-northwest extension (the principal (x:y:z) axes of stretched quartz, where x > =y > =z, is typically in the range 40-20:5-2:1), and horizontal compression in an east-northeast direction as recorded from intrafolial isoclinal folding on subvertical (± 10°) axial planes with fold axes oriented parallel to the stretching lineation direction. No simple shear direction could be identified within this zone using mesoscopic textural features. It was found that Taylor's (1977) eastern boundary of the Churchill Province (eastern boundary of Abloviak shear zone) coincides with the disappearance of the prominent north-northwest-plunging mineral lineation in the area rather than with a change in lithology.

The younger deformation occurs in Abloviak shear zone and in the Churchill foreland zone (Fig. 4) and is expressed as zones of ductile ultramylonite and/or pseudotachylyte breccia, the largest of which is 50 m thick. These are north-northwest-trending zones on vertical to steeply west-dipping planes (and locally on steeply east-dipping planes interpreted to represent overturned ramps) which transect the ductile fabrics of the Abloviak shear zone. Both the ultramylonite and pseudotachylyte breccia zones show down-dip lineations (mineral stretching lineations and polished slickensides, respectively). Small-scale Z-folds and rotated feldspars in ultramylonite zones indicate west-side-up movement (even on east-dipping planes) suggesting these zones represent east-directed thrusts. Major pseudotachylyte breccia zones (> 10 cm) occur roughly every kilometre across the Churchill-Nain Province boundary zone, with subordinate pseudotachylyte breccia zones (< 10 cm) and veins (0.3-3 cm) common throughout. Pseudotachylyte breccia zones occur also along thrust faults in the centre of Nain craton where they account for the prominent north-trend of valleys in the central portion of the map area.

In the only transect of the Churchill boundary zone, a 1 km wide strip, at least 4 km long, of Ramah Group equivalent metasediments was discovered infolded with amphibolite facies Archean basement. Some 100 m west of the western contact of the metasediments is a 20-50 m thick glassy ultramylonite zone with minor pseudotachylyte that separates Ramah Group and Nain Province protolith in Proterozoic amphibolite facies from Proterozoic granulite facies gneisses to the west (Fig. 4). The eastern contact of the metasediments dips steeply to the west and is not marked by a major deformation zone. The metasediments are composed dominantly of quartzite (locally fuchsite-bearing and conglomeratic), and feldspathic quartzite (biotite psammite), with subordinate semi-pelite (biotite ± garnet ± sillimanite schist). Near the base of this sequence, 5 m of conglomerate contains < 10 cm large cobbles of Archean lithologies. The con-

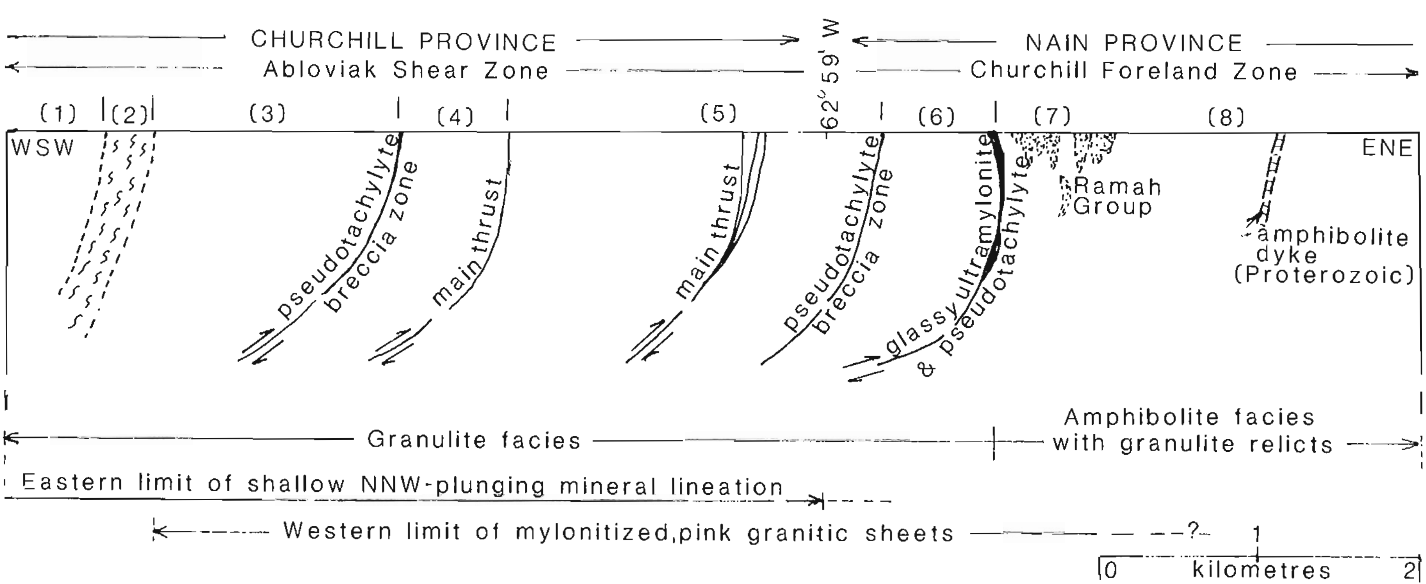

Figure 4. Section of Nain-Churchill boundary zone at latitude 57°52′, longitude 63°. Legend: 1 - homogeneous Tasiuyak gneiss. 2 - granulite facies metapelite. 3 - interlayered zone of Tasiuyak gneiss (1) and orthopyroxene-gneiss (4); pseudotachylyte common. 4 - orthopyroxene-tonalites (Nain protolith). 5 - orthopyroxene-tonalites and minor supracrustals (Nain protolith). 6 - mixed zone of anorthosite, pink granite and amphibolite. 7 - Ramah Group. 8 - amphibolite facies migmatites with granulite facies Nain supracrustals.

tact with the Archean rocks is marked by about 3 m of pink, chaotic, inhomogeneous, coarse grained granitic rock that may be metamorphosed regolith. Bedding and crossbedding are locally preserved in the quartzite despite tight folding about south-southeast-plunging axes, and later shearing, that developed chevron folds on variable north-plunging axes, re-orientation of earlier fold axes, and folding of lineated muscovite. The total thickness of Ramah Group at latitude 57°52′ is about 200 m.

SUMMARY

Lithologies typical of the Saglek and Hebron fiords area have been correlated southward to Okak in the present map area at latitude 57°30′. The centre of Nain Province records mainly late Archean D_2 and D_3 deformations at granulite, retrograded granulite and possibly prograde amphibolite facies metamorphism. An earlier Archean (older than 3.6 Ga) high grade metamorphism may be preserved east of the Proterozoic Mugford Bay fault zone in rocks beneath and peripheral to the base of the Kaumajet Mountains. Conglomeratic sandstone dykes west of the Mugford Bay fault zone attest to once more extensive deposits of Mugford Group sediments on the Nain craton. The presence of Ramah Group equivalent metasediments in the Nain-Churchill boundary zone of the project area extends this important marker horizon of the Torngat orogen to at least 135 km south from Nachvak Fiord.

Interpretations in this report are based on preliminary observations and will serve as working hypotheses during future fieldwork. No assessment is offered for the Fe-oxide and Fe-sulphide facies rocks in supracrustal mafic and ultramafic associations. Isotopic age determinations are required to ascertain the time of ductile deformation in the Nain-Churchill boundary zone and to establish the time of deposition of the Proterozoic Ramah and Mugford groups.

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Umiakovik Lake batholith and other felsic intrusions, Okak Bay area, Labrador

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Abstract

The Elsonian Umiakovik Lake batholith intrudes the northwestern flanks of the Nain anorthosite complex. The batholith contains successively younger rock types from monzodiorite through pyroxene-hornblende quartz monzonite to biotite-hornblende granite, the latter fluorite-bearing. Wallrocks of the batholith along the North River are layered, subvertical, granulitic paragneisses rich in garnet with common graphite-bearing zones. Retrograde reactions involving garnet breakdown in these gneisses may be contact metamorphic effects.

Intrusions within the adjacent Nain anorthosite include syenite (or monzonite) and mafic oxide-rich diorite as well as quartz-bearing rocks similar to those of Umiakovik Lake batholith. Quartz in anorthositic rocks near quartz-bearing intrusions suggests contamination from superjacent granitic magma.

Small, dominantly biotite granite stocks bearing fluorite are exposed on Opingiviksuak and Saddle islands and intrude amphibolitic gneisses of the Archean basement complex. Both are marked by distinctive positive aeromagnetic anomalies which also occur nearby to the north and south offshore.

Résumé

Le batholite d'âge elsonien de Umiakovik Lake est intrusif dans les flancs nord-ouest du complexe anorthositique de Nain. Ce batholite contient divers types de roches progressivement plus récentes, allant des monzodiorites à des monzonites quartziques à pyroxène et hornblende, puis à des granites à biotite et hornblende, ces derniers contenant de la fluorine. Le long de la rivière North, les parois du batholite sont des paragneiss granulitiques stratifiés et subverticaux, riches en grenats, fréquemment accompagnés de zones graphitiques. Les réactions rétrogrades qui ont causé la dégradation des grenats inclus dans ces gneiss résultent peut-être d'un métamorphisme de contact.

Les roches intrusives dans l'anorthosite adjacente de Nain sont des syénites (ou des monzonites) et des diorites mafiques riches en oxydes, ainsi que des roches quartzifères semblables à celles que l'on rencontre dans le batholite de Umiakovik Lake. La présence de quartz dans des roches anorthositiques à proximité des intrusions quartzifères suggère une contamination par un magma granitique injecté juste au-dessus.

De petits stocks principalement composés de granite à biotite, et contenant de la fluorine, affleurent dans les îles Opingiviksuak et Saddle, et sont intrusifs dans des gneiss amphibolitiques du socle archéen. Ces deux types de roches sont caractérisées par des anomalies aéromagnétiques nettement positives qui se manifestent aussi au nord et au sud au large des côtes.

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INTRODUCTION

A considerable variety of felsic intrusive rocks is associated with anorthosite massifs in central and northern Labrador. These form large batholiths (Mistastin, Umiakovik Lake) as well as numerous smaller unnamed intrusions at the margins of, within, and distant from anorthosite massifs. There is no longer much doubt that a close relationship exists in time as well as space for this essentially bimodal anorogenic suite. Although a comagmatic relationship has been rejected by most authors, genetic linking remains viable through crustal melting that occurred in association with the basic magmatism related to development and intrusion of the anorthositic suite.

The source materials for the felsic magmas remain speculative but some of their properties can be broadly characterized, for example:

a) elevated $^{87}Sr/^{86}Sr$ initial ratios in Umiakovik Lake batholith (0.7096 $\pm$ 0.0013, Taylor, 1978) and Mistastin batholith (0.7082 $\pm$ 0.0003, Marchand and Crocket, 1977) imply that older crustal components may have formed a substantial part of the source materials.

b) early, voluminous members of the large felsic suites are relatively K-rich, calcic, quartz-poor and relatively mafic with high $Fe^{2+}/Fe^{2+}+Mg$ (monzonite, quartz monzonite, quartz mangerite).

c) the felsic magmas were fluid-undersaturated through most of their crystallization intervals as indicated by the general lack of pegmatite and vein development and absence of hydrated contact aureoles. Compositions approaching the water-saturated granite minimum were produced only in relatively small volumes at late stages of crystallization.

The fact that similar characteristics are exhibited in intrusions that penetrate both the Archean gneiss terrane of the Nain Province and the compositionally much different Proterozoic terrane of the adjacent Churchill Province to the west in Labrador implies that either; 1) similar crustal source materials were present at depth under both terranes, or 2) the physical and chemical processes involved in crustal melting were capable of yielding similar melt compositions from different crustal source materials.

Some isotopic evidence exists to indicate that a component of reworked early Archean crust is incorporated within the Hudsonian terrane of the Churchill Province (e.g. Ashwal et al., 1986). It is possible that the deep crustal bulk compositions of both terranes are not greatly different. Comparisons of mineral chemistry, isotopic and trace element signatures in Elsonian felsic intrusions from both terranes can provide further tests of the nature of the source materials, even though direct comparison of crustal bulk compositions at depth is not possible.

The rocks investigated during the 1987 field season included the northeastern part of Umiakovik Lake batholith, several intrusions into the Nain anorthosite complex in the vicinity of Umiakoviarusek Lake, and two small granitic stocks intruding Archean gneisses east and south of the mouth of Okak Bay (Fig. 1).

UMIAKOVIK LAKE BATHOLITH

This batholith is one of the two major charnockite-granite batholiths produced by the Elsonian magmatic episode in central and northern Labrador. Many smaller intrusions with similar rock types intrude the anorthositic massifs of this region. A Rb-Sr whole-rock isochron age of 1246 $\pm$ 36 Ma, Sr_i = 0.7096 $\pm$ 0.0013 (^{87}Rb half-life = 1.47 x $10^{-11}yr^{-1}$) on rocks of the batholith has been published (*see* Taylor, 1978).

Two distinct major plutons were identified in the northeastern part of the batholith (Fig. 2); a younger, mainly homogeneous massive biotite-hornblende granite and an older pyroxene-bearing (and probably fayalite-bearing) charnockitic suite which shows considerable textural and some compositional variability, most of which is gradational. Exposures of biotite-hornblende granite are characteristically light grey to pinkish grey whereas the charnockitic suite typically weathers deeply to a buff to deep brown, often crumbly surface.

The charnockitic suite contains rocks that range from fine- to medium-grained monzodiorites to porphyritic and coarse grained equigranular pyroxene quartz monzonite. Fine- to medium-grained equigranular rocks are present at contacts with anorthositic rocks and with the North River gneiss; inward from contacts, at first sparse and then increasingly abundant and larger phenocrysts of alkali feldspar make their appearance. Inhomogeneities in texture are present in

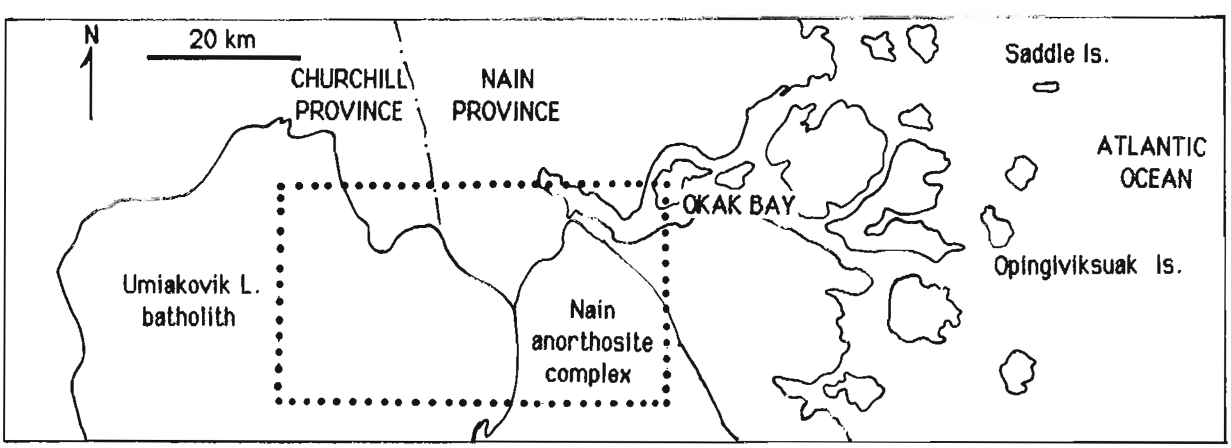

Figure 1. Okak Bay area showing locations of rocks investigated. Heavy dots outline the area shown in Figure 2.

many outcrops principally as variable grain sizes of both alkali feldspar phenocrysts and groundmass. Massive fine- to medium-grained monzodiorite is present in abundance at some localities remote from observed contacts and may reflect proximity to a floor or former roof. Distinctly finer grained xenoliths enclosed within a coarse grained host of similar bulk composition were observed in a number of locations; these seem likely to represent autoliths formed during the intrusion process. Country rock xenoliths are rare and were observed only in proximity to contacts.

At the eastern margin of the batholith, southeast of Umiakovik Lake, the dominant rock is medium grained diorite to monzodiorite with colour index averaging about 30. Small, rounded perthites 0.5 to 1 cm across comprise 5 to 20 % of many rocks with a 2 to 3 mm matrix. These rounded perthites can be highly variable in amount even over a single outcrop. Pyroxene and opaque oxides are the common mafic minerals but traces of biotite and hornblende also may be present. In many places scattered dark plagioclase tablets up to 3 x 6 cm but more commonly 2 to 3 cm long become prominent and perthite ovoids become rare or absent; this transition to diorite appears to be completely gradational and was observed at a number of localities. Similar relations at the margin of the batholith just to the south of the area of Figure 2 have been described by Ranson (1976; 1981).

Coarse grained leuconorite and anorthosite occurs at the contact of the batholith to within about 2 km of Sikkoyavik Brook. This is evidently a narrow wedge or prong that extends northwest from the main part of the Nain anorthositic complex; it intrudes and contains xenoliths of Archean gneiss and is completely enclosed within gneisses at its northern extremity. One large mass of coarse grained anorthosite and leuconorite occurs within the batholith about 1.5 km due south of the eastern end of Umiakovik Lake. It is exposed in a deeply incised streambed and may be a xenolith or the exposed part of a much larger mass underlying the batholith.

Passing westward from the contact, monzodiorite contains increasing amounts of perthite ovoids and quartz and is transitional into coarse grained pyroxene quartz monzonite. This transition is irregular and large patchy areas of fine- to medium-grained monzodiorite are present well into the interior of the batholith.

North and west of Umiakovik Lake, the batholith is readily subdivided into an older pyroxene-hornblende quartz monzonite and a younger pluton of biotite-hornblende granite. Both of these units intrude a distinctive gneiss referred to here as the North River gneiss. The North River gneiss is part of a much larger unit extending far to the north that forms the wallrock around the northern part of the batholith. The larger unit was described by Taylor (1979) as garnet-quartz-feldspar gneiss and considered to be of Aphebian age.

At contacts between pyroxene-hornblende quartz monzonite and North River gneiss, dykes and small bodies of fine- to medium-grained monzodiorite can be found intruding the gneiss at distances of up to 0.5 km from the main contact. Contacts between biotite-hornblende granite and the gneiss are sharply defined and the granite breaks directly

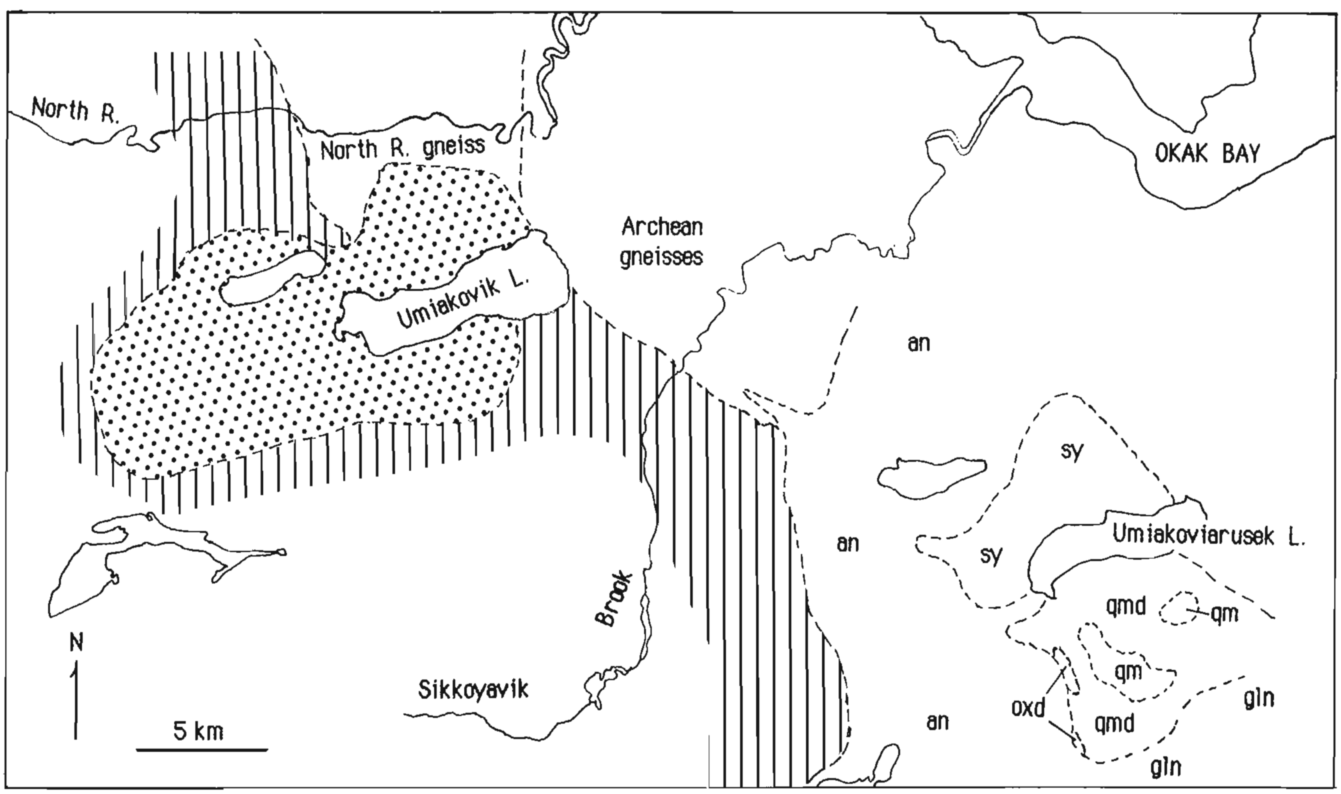

Figure 2. Northeastern part of Umiakovik Lake batholith and adjacent Nain anorthositic complex. Vertical hatching — monzonitic rocks including monzodiorite, monzonite, pyroxene-hornblende quartz monzonite; stipple — biotite-hornblende granite; an anorthositic rocks; sy — syenite; gln — gabbronorite, quartz gabbronorite, leuconorite; qmd — quartz monzodiorite; qm — quartz monzonite; oxd — oxide-rich diorite.

across strike of the subvertical gneisses for a long distance north of Umiakovik Lake; in addition, dykes of biotite-hornblende granite tend to have sharp angular walls, perhaps suggesting that the granite intruded under different physical conditions (cooler? lower pressure?) than the pyroxene quartz monzonite. Large low-dipping sheets of biotite-hornblende granite occur in North River gneiss on the cliffs forming the south valley wall of the North River.

North River gneiss

The major gneiss unit comprising the country rock at the northern boundary of Umiakovik Lake batholith is highly distinctive. It is well exposed on both sides of the North River valley. Because it is part of a very large regionally developed unit (Taylor, 1979) and may be rather variable, the informal name North River gneiss is used here to describe its character and occurrence.

Approximately 7 km of an east-west cross-section through the gneiss is almost continuously exposed along the North River at the eastern margin of Umiakovik Lake batholith. The rocks are distinctively layered mostly on a scale of 10 cm or less to more than 1 m with consistent north-south strikes and subvertical dips. Compositions of individual layers range from felsic to intermediate with numerous interlayers of impure quartzite and quartzite. Slightly darker layers bearing abundant biotite and hypersthene are present in many parts of the section. Truly mafic layers are exceedingly rare; in this respect the unit appears compositionally distinct from the Archean gneiss assemblages exposed immediately to the east. In places variable amounts of injected light grey granitic to dioritic material is present in outcrops and forms irregular dykes, veins and small masses. The injected material, although foliated in varying degrees, is clearly trangressive in most cases and only rarely forms more than 10 % of an outcrop. Two outcrop areas near the northern limit of the batholith were observed to contain 30 to 50 % of similar light grey to whitish medium- to coarse-grained granitic intrusive material.

The presence of abundant garnet in many layers, commonly with grain sizes greater than 1 cm and in places up to 4 cm, is a striking feature of many outcrops. Except for the garnets, medium grain sizes of 1 to 2 mm characterize much of the gneiss. Many garnets show evidence of partial breakdown along fractures to fine grained crystalline products. In some cases the reaction has advanced considerably and only small garnet grains remain, surrounded by a rounded or ellipsoidal mafic clot which has a bluish cast suggesting cordierite.

Graphite is a common accessory mineral and is especially abundant in rusty zones or layers. The rusty zones were probably sulphide-bearing but typical weathered outcrop samples contain little or no visible sulphide minerals.

UMIAKOVIARUSEK LAKE

Felsic intrusions with a range of compositions intrude anorthositic rocks within the Nain complex in this area. It is unclear from field relationships which, if any, members of the felsic suite may be comagmatic.

North of Umiakoviarusek Lake the rocks are dominantly medium grained (1 to 3 mm mostly, but up to 5 mm) syenites or monzonites with variable colour index, typically between 10 and 15 but locally up to 20. These rocks were mapped as syenite by Taylor (1979). Outcrops have a light buff to dark rusty brown weathered veneer. Hornblende and opaque oxide are the chief dark minerals; pyroxene cores may be locally present within hornblende. Most rocks are massive but well-defined, regular, small-scale layered structures occur over intervals up to 1 m thick; several of these sections may be present on a large outcrop. Outcrops with layers dipping about 10o to the northwest and northeast occur in the central part of the mass but much steeper dips were noted near the eastern and western margins. One type of layering has dark concentrations of hornblende and oxide up to 0.5 mm thick separated by feldspar-rich intervals a few centimetres thick in which dark minerals are virtually absent. Above and below the layered section the same minerals are uniformly distributed in homogeneous rock. Locally, large xenoliths of coarse grained anorthosite are present and monzonitic dykes intrude anorthosite wallrocks at the northern and western contacts.

South of Umiakoviarusek Lake a greater compositional variety and more complex relationships characterize the rocks that intrude the anorthosite suite. Anorthositic, leuconoritic, and gabbronoritic rocks are intruded by a suite of rocks that ranges in composition from diorite or ferrodiorite to granite. Medium- to coarse-grained gabbronorite, in part quartz-bearing, forms one of the larger units and carries large and small xenoliths of coarse grained anorthosite in many outcrops; dark tabular plagioclase crystals up to 10 cm, but mostly 2 to 4 cm are also common in this unit. Contacts between this rock and medium grained quartz monzodiorite appear to be gradational with disappearance of tabular plagioclase upward; with further gradual upward increase in quartz and K-feldspar, quartz monzonite composition is attained. Small,pink, fine- to medium-grained granitic dykes intrude all units, and one intrusive mass 30 m across was observed. Highly magnetic oxide-rich diorite or gabbro outcrops at intervals over a distance of more than 5 km along a major south-southeast-trending valley just west of the western end of Umiakoviarusek Lake. At the southern locality noted in Figure 2, along the western valley wall, sulphide- and oxide-bearing blocks are present in mafic dioritic rubble below a cliff face. The oxide-rich dioritic rock is gradational into less oxide-rich diorite locally containing coarse grained anorthositic xenoliths. Weak discontinuous layering and marked subvertical foliation is sporadically present in the diorite but much of it is massive. The larger northern dyke-like mass has a strong subvertical foliation throughout that strikes subparallel to elongation of the body. Colour index of this rock is about 70 with hornblende and opaque oxide the dominant dark minerals. The body may have been intruded along a shear or fracture zone in anorthosite; whitened, recrystallized and altered anorthosite is present along the eastern contact.

Coarse grained quartz monzonite to biotite-hornblende granite is exposed on some of the higher hills and intrudes anorthositic rocks. The rocks are massive and fresh and quartz and K-feldspar contents decrease from higher to lower elevations. The relationship of the main mass of granitic rocks to the dioritic suite is not exposed but sharply-defined fine-to medium-grained granite dykes and small intrusive masses cut dioritic rocks at several places.

OFFSHORE ISLANDS

At least four prominent, subcircular, positive, aeromagnetic anomalies are present in the near offshore area in the vicinity of Okak Bay (*see* map 7452 G, Geological Survey of Canada, Geophysical Series). These have intensities of 1000 gammas or more and diameters of 5 to 10 km. Small stocks of fresh, massive granitic rocks occur on several of the outer islands east of Okak Bay (Taylor, 1979) and two that correlate with well defined positive aeromagnetic anomalies were investigated and sampled. They are of special interest because they intrude the Archean gneiss complex and may be related to middle Proterozoic magmatism in this region.

Barton (1977) reported Rb-Sr whole-rock and mineral isochron ages on a number of granite bodies along the coast in this area, including Opingiviksuak and Saddle Islands. Six samples of the pink facies of Opingiviksuak granite yielded an isochron age of 1376 $\pm$ 38 Ma with initial $^{87}Sr/^{86}Sr$ = 0.7264 $\pm$ 0.0014. A whole-rock-biotite mineral separate isochron age for a single specimen from Saddle Island granite gave 2383 $\pm$ 48 Ma with initial $^{87}Sr/^{86}Sr$ = 0.7149. A similar age was obtained from a whole-rock-biotite pair from White Bear Island to the north which is also distinguished by a pronounced subcircular aeromagnetic anomaly. The ^{87}Rb decay constant used for these calculations was 1.39×10^{-11} yr^{-1}. The whole-rock-biotite ages need to be regarded with caution because biotite combines high Rb levels with low Sr so that small changes in Sr due to secondary alteration or re-equilibration for example, could have large effects on a calculated age.

Opingiviksuak Island

Approximately 12 km^2 of granitic rocks are exposed on the island. Archean gneiss that forms the wallrock, outcrops on the northern edge of the island. On the western margin of the island, a narrow (not more than 100 m wide) strip of dark grey to black quartz-bearing fresh, coarse gabbroic rock is exposed. Both of these units are intruded by granitic dykes, a light grey medium grained variety being common. Much of the central part of the island is grey to pinkish-grey, massive coarse grained biotite granite. The southern part of the island is underlain by large areas of medium grained, pale pink granite. Near the summit of the island purple fluorite coatings are found on fracture surfaces and rare interstitial fluorite grains are present in the granite here as well as in other parts of the island. Also near the summit, some joint surfaces are distinctly reddened, the effect not extending more than a few millimetres into the rock; some of these reddened surfaces are fluorite-coated.

Approaching the Archean gneiss contact to the north, the granite becomes slightly, but distinctly, more mafic and hornblende may accompany biotite in the assemblage. Weak sporadic foliation is present in the granite within about 100 m of the gneiss contact.

The dark subophitic coarse grained gabbro on the west side of the island is a relatively fresh massive rock and seems likely to be associated with the granite intrusion; it is unlike rocks of the Archean basic suite in the area. Sharply-bounded fine, medium, and locally coarse grained to pegmatitic granitic dykes transect the gabbro in abundance. Two dykes were found that have a notably higher colour index and are quartz-poor suggesting they may be products of local hybridization. Several of the pegmatitic dykes carry coarse amazonite crystals in their central zones. The gabbro itself contains quartz grains and small interstitial quartz-rich patches suggesting contamination by a granitic component. Large, dark tabular plagioclase megacrysts from a few centimetres to 20 cm are common in the gabbro. These megacrysts normally comprise less than 10 % of the rock.

Saddle Island

Massive granitic rock underlies the entire island. This is typically medium grained, light grey biotite granite with accessory purple fluorite.. Rare fine grained aplitic dykes up to 1.5 m thick cut the granite.

Dark, amphibolitic Archean gneiss xenoliths are present throughout the granite; most are subangular and they have a restricted size range from a few centimetres up to about 30 cm. The xenoliths have sharply-defined margins and obvious interaction with the enclosing granite appears slight.The xenoliths are fairly uniformly distributed and in places comprise as much as 10 % of the outcrop. The upper size limit of the xenoliths suggests that roof stoping probably characterized at least the final stages of intrusion; buoyancy and viscosity relationships may have dictated that larger xenoliths sank in the granitic magma whereas smaller ones remained suspended.

DISCUSSION

The silicic Elsonian intrusions in Labrador have been confirmed by isotopic dating to be strictly anorogenic, in the sense that their intrusion and crystallization ages are remote in time from known, prior or succeeding, orogenic episodes. In addition, their geological settings are consistent with anorogenic interpretation. The igneous rocks described here are all products of anorogenic magmatism (except perhaps Saddle Island) and intrude adjoining crustal blocks of different age and evolutionary history. If they formed dominantly by partial melting of deeper parts of their respective crustal blocks, then either the bulk composition of the lower crust in both blocks is similar, or extensive crystal fractionation of different primary magmas produced similar end products. Many A-type granites as geochemically defined (e.g. Whalen et al., 1987) occur in settings whose specification as anorogenic is somewhat subjective because their crystallization ages may be less clearly separated from preceding or succeeding orogeny. Because of comparatively well defined chemical and mineral features of A-type granites, there is strong temptation to try to relate them to specific source compositions. This is especially so if the granite forms a single pluton or the overwhelmingly dominant member of a suite. Most recent models of origin for A-type granites favour direct partial melting of specific source materials (e.g. depleted felsic granulites) and relegate subsequent fractional crystallization of the magmas to a subordinate role.

Strong negative Eu anomalies typical of A-type granites imply that feldspar fractionation has played an important role in their origins. An obvious question is whether the Eu anomalies formed during the partial melting process at the source, or are products of subsequent crystal-liquid fractionation of the magmas. Most geochemically-defined examples of A-type granites are high-level intrusions (epizonal), commonly subvolvanic and sometimes associated with lavas and pyroclastics. Their geological setting suggests that they may exhibit only the late products of the A-type granite-forming process. At somewhat deeper levels (mesozonal), charnockitic-rapakivi associations include (either as gradational varieties or as separate plutons) more calcium- and magnesium-rich earlier-crystallized rocks (Wolf River, Wisconsin; Mistastin, Labrador) with higher temperature mineral assemblages (pyroxenes, fayalite, ternary feldspars). Still deeper (catazonal) igneous charnockitic associations are characterized by pyroxene- and olivine-bearing assemblages with dominant ternary feldspars and even more magnesian silicates (Lofoten, Norway; Mealy Mtns, Labrador).

Large Proterozoic felsic suites like the Elsonian in Labrador commonly comprise a range of pluton compositions, many of which individually would probably not be labelled as ''A-types'' by currently used geochemical criteria. These include rapakivi and charnockitic varieties in which mineral and chemical criteria imply comagmatic relations (Wiborg, Wolf River, Mistastin). High degrees of crystal fractionation accompanying passage of anorogenic granitic suites through the crust is an expected consequence of their mode of origin. This would help to compensate for heat losses of the magmas to relatively cool crustal rocks in a stabilized cratonic setting. Heat supplied by fractional crystallization of the parent felsic magma would help sustain diapiric rise by continued subtraction of a crystal fraction which has the additional salutary effects of maintaining a high liquid/crystal ratio in, as well as increasing the relative buoyancy of, the remaining magma.

High temperatures of the initial parent magmas are indicated by pyroxene-, fayalite-, and hypersolvus feldspar-bearing rocks that typically form earlier-crystallized monzonitic and monzodioritic members of the suites. This implies high degrees of crustal melting of the source rocks, probably 50 % or more, requiring major input of heat from the mantle, now widely believed to be due to extensive ponding of mafic magma beneath the crust. Postulated small degrees of melting of specialized sources (e.g. depleted felsic granulites) on the other hand encounter the difficulty of efficient separation of melt from the source materials and the problem of maintaining upward mobility through cratonic crust in the face of continuing large heat losses. Although A-type granites are recognized as products of strong fractionation processes, the site and mechanism for this fractionation remains uncertain; most authors have favoured partial melting at the source as the main control but considerations discussed above provide support for substantial crystal-liquid fractionation during passage through the crust.

One of the clues to the origin of A-type rapakivi-charnockitic suites is the common association of monzonites (monzosyenites, mangerites) that are present in variable volumes but are most abundant in catazonal igneous charnockitic suites. These monzonitic rocks commonly show evidence of having crystallized at high temperatures (ternary feldspars, complexly-exsolved pyroxenes) and have marked positive Eu anomalies indicating that they are crystal-rich fractions. In addition, intrusive relations place them among the earlier crystallized members of individual suites. This relationship strongly suggests that epizonal A-type suites may commonly develop through crystal fractionation from more ''primitive'' magmas with alkali-calcic characteristics that originated by unusually high temperature melting of near ''average'' (andesitic) crust. The liquidus quartz volume for a wide range of ''granitic'' melts is known to expand considerably with increasing pressure under both fluid-saturated and fluid-undersaturated conditions. This implies that ''granitic'' melts formed in the lower crust will have compositions much poorer in quartz than granitic minimum melts at low pressure and be closer to quartz monzonite or quartz mangerite compositions. Transport of such melts to higher crustal levels is accompanied by shrinkage of the quartz liquidus volume so that unless the rise is very slow and steady the liquid must become quartz-undersaturated from time to time. This should result in crystallization intervals in which mangeritic or monzonitic crystal-rich fractions form and may be separated from the remaining melt. The efficiency with which these crystal-rich fractions are able to separate from the melt fraction will determine how much quartz is present in the final products. As noted above, monzonites and quartz-poor monzonites (mangerites and quartz-poor mangerites) are well known associates of catazonal and mesozonal equivalents of these suites.

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Great Slave Lake shear zone meets Thelon Tectonic Zone, District of Mackenzie, NWT.

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Hanmer, S. and Needham, T., Great Slave Lake shear zone meets Thelon Tectonic Zone, District of Mackenzie, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 33-49, 1988.

Abstract

The early Proterozoic Great Slave Lake shear zone and the Thelon Tectonic Zone are co-extensive elements of the same tectonic structure — the Great Slave — Thelon shear zone. The 060° trending elements (Laloche segment) are transcurrent; 030-010° trending elements (Schist-Daisy segment and Thelon Tectonic Zone) are dextral transcurrent on their west side and dip-slip on their east side. The broad Daisy Lakes belt of high grade dip-slip mylonites brings granulites up into the map-plane and postdates the Schist Lakes belt, a broad belt of high grade dextral strike-slip finely homoclastic protomylonites. The trend of all high grade mylonites wraps around the southeast corner of Slave Craton, but is crosscut by an 060° trending narrow belt of greenschist facies strike lineated dextral transcurrent mylonites. The early Proterozoic Great Slave Supergroup unconformably overlies the Great Slave — Thelon shear zone. These observations are compatible with recent modeling of the tectonics of southeastern Slave Craton in terms of early Proterozoic oblique convergence, collision and indentation.

Résumé

La zone de cisaillement du Grand lac des Esclaves, datant du Protérozoïque inférieur, et la zone tectonique de Thelon constituent les deux parties de la zone de cisaillement de Great Slave et Thelon. Les éléments tectoniques à direction 060° (secteur Laloche) ont agi en décrochement. Les éléments à direction 030-010° (secteur Schist et Daisy et zone tectonique de Thelon) ont joué en décrochement dans leur partie ouest, mais en chevauchement dans leur partie est. Une large zone de mylonites chevauchantes à faciès des amphibolites, responsable de la mise en place des mylonites à faciès des granulites, est plus jeune qu'une zone de protomylonites à faciès des amphibolites qui a joué en décrochement. La direction de ces mylonites suit étroitement le coin sud-est du craton de Slave, mais elle est recoupée par une bande étroite de mylonites à faciès des schistes verts à direction 060° qui a joué en décrochement. Les sédiments du supergroupe de Great Slave reposent en discordance stratigraphique sur la zone de cisaillement de Great Slave. Ces données sont compatibles avec un modèle tectonique de convergence oblique, de collision et de penétration entre deux blocs continentaux

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INTRODUCTION

Great Slave Lake shear zone (Hanmer, in press(a)), an early Proterozoic dextral transcurrent structure of crustal proportions, lies along the southeast margin of Slave Craton. Our fieldwork during the summer of 1987 demonstrates that it extends northeastwards from the 25 km-wide Laloche segment previously described (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986), into the Schist Lakes — Daisy Lakes area where it is 13 km wide (Fig. 1). Principal new field results are:

(1) The mylonites of the Schist-Daisy segment form three distinct northeast-trending belts which decrease in metamorphic grade from southeast to northwest — Daisy Lakes, Schist Lakes, and Hornby Channel belts.

(a) The Daisy Lakes belt comprises a 3 km-wide granulite facies southeastern sub-belt and a 2 km-wide upper amphibolite facies northwestern sub-belt. The two sub-belts are apparently contemporaneous. Both sub-belts carry a vertical foliation and a down-dip extension lineation. The kinematic history is complex since both right-slip and southeast-side-up dip-slip shear sense indicators are present.

(b) The Schist Lakes belt is 7 $\pm$ 0.5 km wide and is predominantly composed of upper amphibolite facies finely homoclastic protomylonites. All mylonitic rocks carry a vertical foliation with a sub-horizontal extension lineation and right-slip shear sense indicators.

(c) The Hornby Channel belt is a 0.5-1.5 km-wide belt of chlorite-bearing mylonites and ultramylonites carrying a vertical foliation, a sub-horizontal extension lineation and right-slip shear sense indicators.

(2) The upper amphibolite strike-lineated protomylonites of the Schist Lakes belt are older than the granulite amphibolite facies, dip-lineated mylonites of the Daisy Lakes belt. The former is intruded by the 'Sandwich' granite which in turn is tectonically incorporated into the Daisy Lakes belt.

(3) The strike of the mylonites and ultramylonites is variable, both spatially and temporally. The trend of the upper amphibolite and granulite facies mylonites changes from 060° in the southwest to 030° in the northeast. To the northeast the greenschist facies mylonites trend 060°, thereby cutting across the older tectonites. Reconnaissance indicates that the Hornby Channel belt maintains this trend for at least 65 km

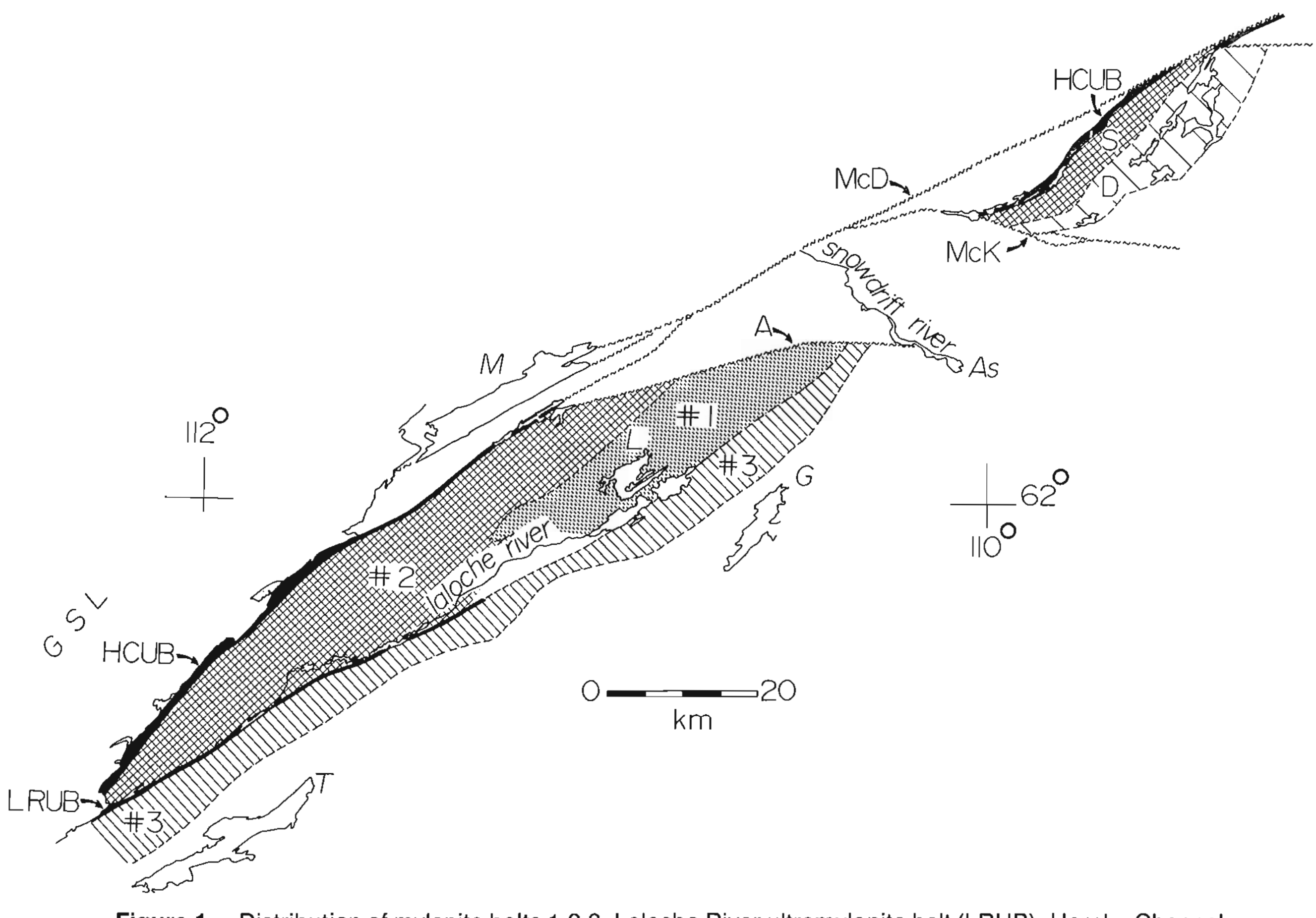


Figure 1. Distribution of mylonite belts 1,2,3, Laloche River ultramylonite belt (LRUB), Hornby Channel ultramylonite belt (HCUB), Schist Lakes (S) and Daisy Lakes (D) within Great Slave Lake shear zone. McDonald-Wilson (McD), Austin (A) and McKee (McK) faults are shown. Abbreviated locations are Austin (As), Gagnon (G), Great Slave (GSL), Laloche (L), McDonald (M), Thubun (T) Lakes. Segments of Great Slave Lake shear zone are Laloche (southwest of Austin fault), Schist-Daisy (northeast of McKee fault) and Snowdrift segment. See text.

beyond the limit of our systematic mapping. Since the Hornby Channel belt truncates the other mylonites and granites of the Schist-Daisy segment, it is the youngest mylonite belt.

(4) Sandstones and conglomerates on the southeast side of Schist Lakes unconformably overly homoclastic protomylonites of the Schist Lakes belt with marked angular discordance. The sediments probably comprise an outlier of the Sosan Group.

(5) Structurally the Schist-Daisy segment is bounded to the south by the McKee fault which corresponds to a well defined linear element in the regional aeromagnetic field at the latitude of McKee Lake (Geological Survey of Canada, 1984). It is a vertical, initially ductile fault which was active during passage through the bulk-rock brittle-ductile transition. The fault cuts across all the rocks of the Schist-Daisy segment with a cumulative dextral offset of 35-40 km.

On the broader scale, our new observations lead us to the following initial conclusions:

(6) The Thelon Tectonic Zone, the Schist-Daisy segment, and by extension the whole length of Great Slave Lake shear zone, are parts of a co-extensive tectonic structure which progressively changes trend towards the south and west from 010 to 060°.

(7) Radiometric dating of tectonic activity in Great Slave Lake shear zone implies that at least the southern part of the Thelon Tectonic Zone is an entirely early Proterozoic tectonic feature (see Thompson et al. 1985, 1986).

(8) The relative ages of the Great Slave Lake shear zone mylonites are compatible with a published model of oblique dextral convergence of Slave Craton and Rae Province, leading to collision and followed by indentation (Hoffman, 1987).

BELTS AND SEGMENTS

Northeast-trending mylonite belts of Great Slave Lake shear zone lie to the southeast of, and are cut by, the McDonald-Wilson fault. They are also cut by the east-west striking Austin and McKee faults, which divide the shear zone into three segments: Laloche, Snowdrift and Schist-Daisy (Fig. 1). The Laloche segment has been described elsewhere (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986). We have mapped the Schist-Daisy segment at 1:25 000 - 50 000 scale and the Snowdrift segment at reconnaissance scale (see also Reinhardt, 1966, 1969). We note here that the principal geological boundaries, both between and within the three segments, are all directly reflected in the regional aeromagnetic field (Geological Survey of Canada, 1984).

The Schist-Daisy segment comprises three mylonite belts: Schist Lakes, Daisy Lakes and Hornby Channel belts (Fig. 1). The mylonites are derived from a protolith of magnetite-bearing biotite — hornblende granitoid (Fig. 2A-C), apparently an extension of the Laloche batholith (Hanmer and Connelly, 1986), with included screens and xenoliths of metasediment, amphibolite and meta-tonalite. The batholith is a leucocratic granitic to tonalitic syn-shear zone intrusion (circa 2.15-1.95 Ga, U/Pb in zircon in the Laloche segment; S.A. Bowring *in* Hanmer and Connelly, 1986). Similar magnetite-bearing granitoid, underlies the Snowdrift segment, but is also present in large volume in the shear zone wallrock southeast of the Schist-Daisy segment (Fig. 3). In contrast to the Laloche segment, the well developed, northeast-trending positive aeromagnetic anomaly underlain by the batholith (Geological Survey of Canada, 1984) is more extensive than the outcrop of Great Slave Lake shear zone itself. The northwestern boundary of the batholith is marked by the Hornby Channel belt (Fig. 1). In the Laloche segment, the batholith is demonstrably intrusive into white sillimanite-cordierite-garnet granite and migmatitic paragneiss along its southeastern contact (Hanmer and Connelly, 1986). A similar garnetiferous granite with migmatitic xenoliths forms the wallrock along the southeastern contact of the Laloche batholith in the present map area (Fig. 3).

SHEAR ZONE AND WALLROCK

In the Schist-Daisy segment (Fig. 3), the shear zone is bounded to the northwest by a medium- to coarse-grained, muscovite-rich, white, two-mica leucogranite. This is intrusive into previously folded micaceous meta-arenite, rhythmically interlayered meta-arenite and metapelite, calc-silicate and amphibolite collectively referred to here as the external paragneisses. Trough cross-stratification is preserved locally (Fig. 4A). The external paragneisses are folded at various scales up to 1 km in wavelength. Folds are moderate to very tight, upright and vertical with east to north-northeast striking axial planes. With the exception of the axial zones of kilometre scale folds located north of Lower Schist Lake, the well developed axial planar schistosity is regionally parallel to well preserved bedding. A mineral-alignment extension lineation is everywhere hinge-parallel and therefore steeply plunging (Fig. 5A). Coarse (< 15 cm) porphyroblastic chiastolite, staurolite, sillimanite and cordierite(?), with rare smaller (< 5 mm) garnet, occur within more pelitic bands (Fig. 4B). Granitic melt(?) pods and stringers occur in pelitic bands close to the McDonald-Wilson fault.

The white leucogranite is itself mildly foliated. Away from the main outcrop of external paragneiss the granite foliation is north-south trending (Fig. 5B). Closer to the paragneisses the granite foliation trend becomes more northeasterly. The paragneisses and the leucogranite are both cut by an isotropic, medium to coarse grained, pink two mica granite which occurs in large and small bodies between the McDonald-Wilson fault and the Hornby Channel belt. Associated pegmatites commonly carry coarse tourmaline. All of the above described lithologies are intruded by vertical sheets of isotropic ophitic metagabbro to metadiabase, one to hundreds of metres thick (Fig. 6A).

The external paragneisses were previously identified along the north shore of McKee Lake as Yellowknife Supergroup (Wright, 1968). However, the presence here of trough cross-stratification and volumetrically important calc-silicate units (Fig. 3) does not support such correlation since they are absent in the type Yellowknife Supergroup as described by Henderson (1975). Rather, we point out the close resemblance of the external paragneisses to outcrops of amphibolite facies Wilson Island Group (circa 1.93 Ga, Bowring et al., 1984) in the Îles du Large, East Arm of Great Slave Lake. However, we also note the resemblance of the white muscovite-bearing leucogranite with its associated mafic

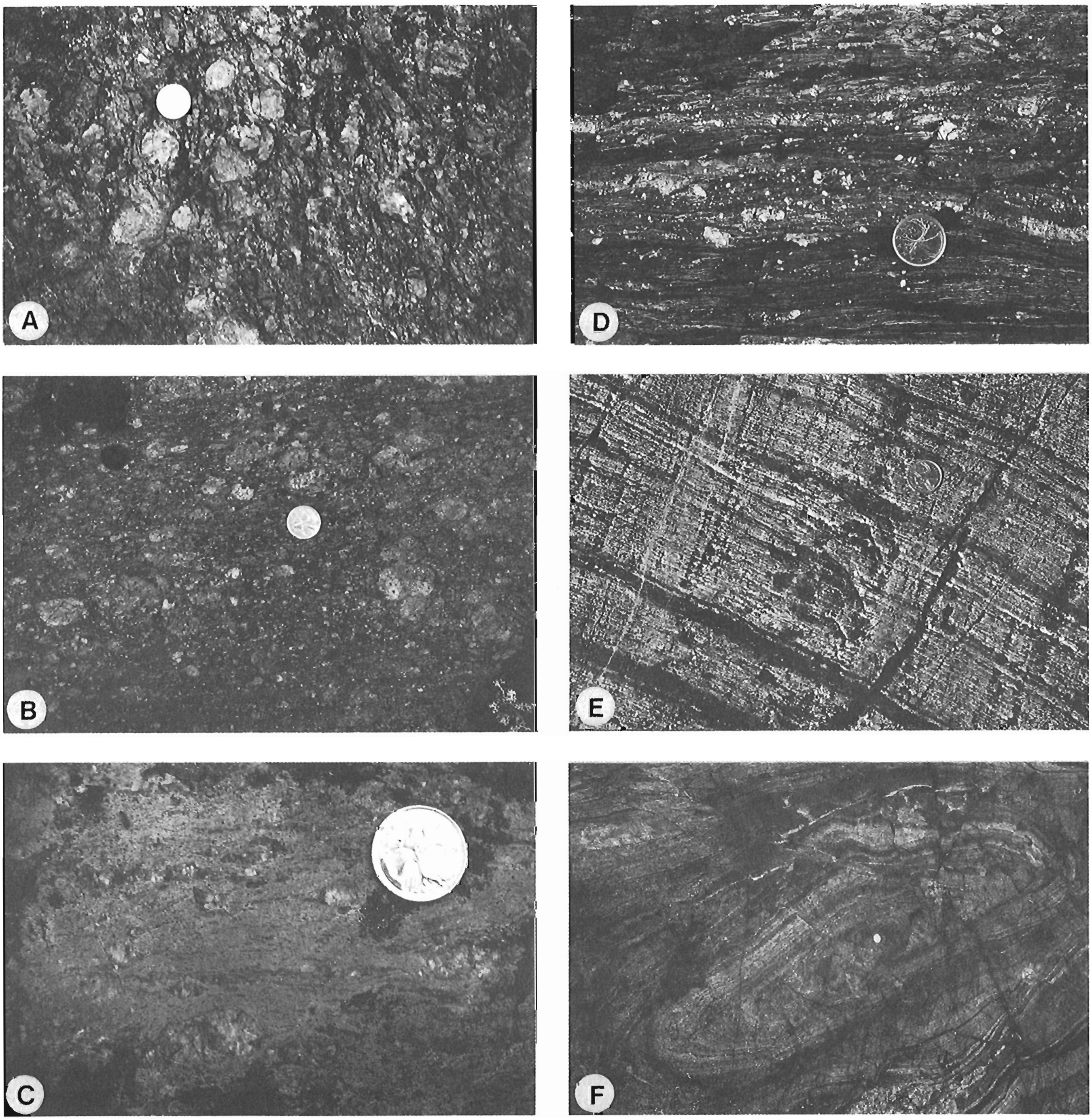


Figure 2. Mylonites of the Schist-Daisy segment of Great Slave Lake shear zone. (A) Foliated, coarse megacrystic granite protolith (GSC-204338-D). (B) Coarsely heteroclastic protomylonite (GSC-204338-K). (C) Coarsely heteroclastic ribbon mylonite. A-C comprise a sequence of progressively developed textures in the northwestern sub-belt of Daisy Lakes belt (GSC-204338-H). (D) Porphyroclastic schist derived by mylonitization of coarse granite sheets intruded into semi-pelitic metasediment host, Hornby Channel belt (GSC-204338-F). (E) Recrystallized (annealed), finely banded homoclastic ribbon mylonites derived from a coarse megacrystic granite protolith, Dion gneiss (GSC-204338-L). (F) Sheath fold in banded recrystallized homoclastic mylonites, observed on horizontal outcrop surface, indicates steeply plunging finite extension direction (GSC-204338-C).

dykes and paragneiss xenoliths to the McDonald Granite (Fig. 3) which occupies an analogous wallrock position with respect to the Laloche segment (Hanmer and Lucas, 1985). U/Pb zircon dating of the McDonald Granite yields ages of circa 2.5 Ga (S.A. Bowring, pers. comm., 1986). At least one of these potential correlations must be invalid.

To the southeast, the Schist-Daisy segment is intruded by a wide swath of magnetite-bearing hornblende-biotite granitoid, the Snowdrift granite (Fig.3). The granitoid is variably foliated to isotropic and carries sporadically developed, very narrow (< 1m) discontinuous dip-lineated mylonite zones. The eastern limit of the Snowdrift granite lies to the east of our mapping. It is marked by a belt of white garnetiferous granite (Fig.4C), with abundant garnet — sillimanite quartzite rafts and xenoliths, which underlies an 030° tending aeromagnetic low to the east of our systematic mapping (Geological Survey of Canada, 1984; our reconnaissance; S. Roscoe, pers. comm., 1987). The Snowdrift and garnetiferous granites also occupy most of the Snowdrift segment (Fig.3), wherein both are locally mylonitized.

SCHIST - DAISY SEGMENT:

Hornby Channel belt

A belt of chlorite-bearing ultramylonites and mylonites, with subordinate protomylonites, up to 2 km wide, runs from McKee Lake, along the length of Schist Lakes (Fig.3) where it was previously identified as Archean Yellowknife Supergroup (Wright 1968)[1]. Along strike to the northeast it is truncated by the McDonald-Wilson fault at the latitude of Upper Daisy Lake. Lithologically, the belt separates magnetite-bearing hornblende-biotite granitoid to the southeast from muscovite-bearing granites with external paragneisses and mafic dykes to the northwest and is visibly derived at the expense of both. However, the northwestern lithologies constitute the predominant protolith. The granites

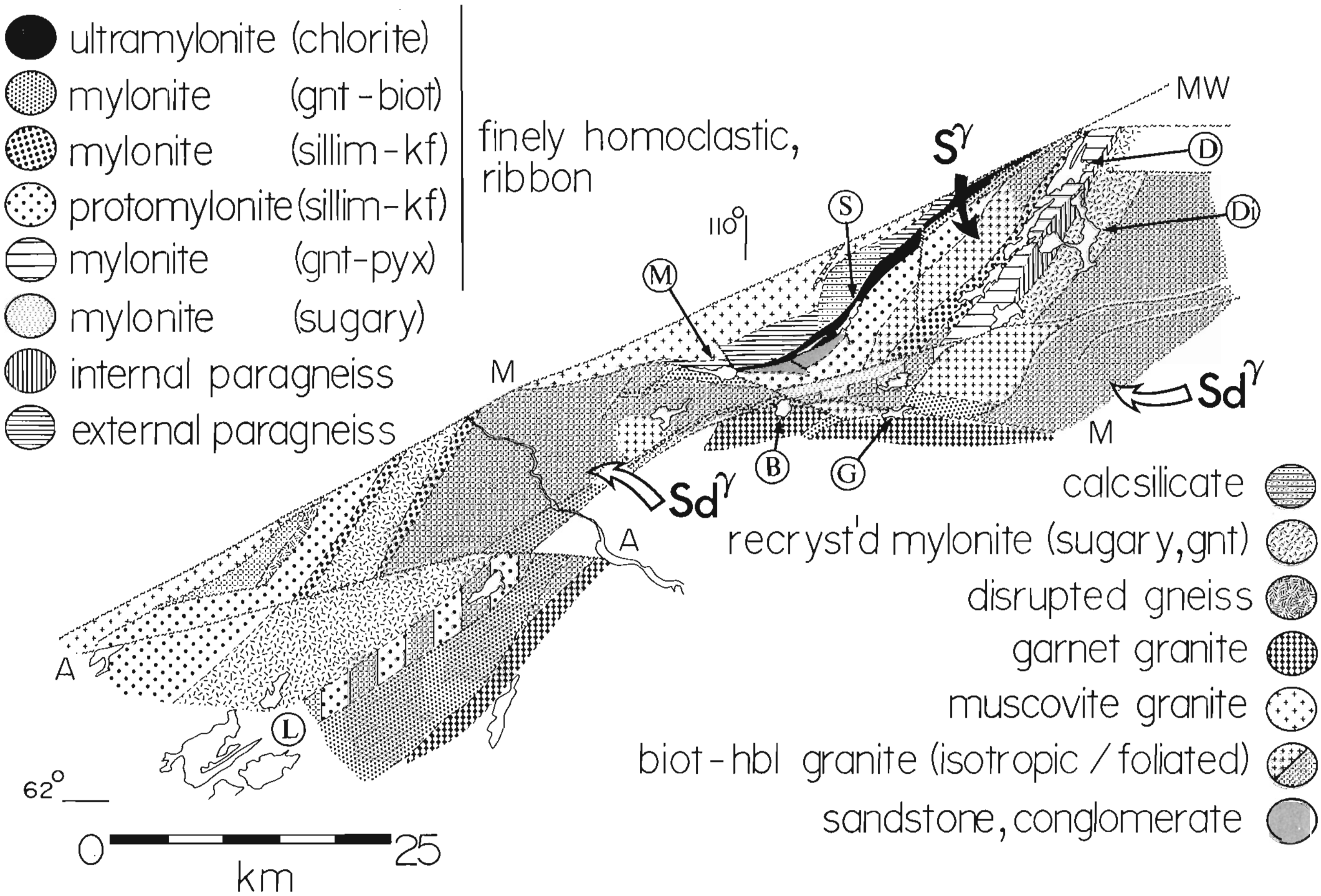


Figure 3. Geology of the Schist-Daisy, Snowdrift and northeast part of the Laloche segments of Great Slave Lake shear zone (see Fig. 1). Sγ is the 'Sandwich' granite; Sdγ is the Snowdrift granite; the muscovite granite northwest of Laloche Lakes is the McDonald granite and the recrystallized sugary garnet mylonites passing through Dion Lake are the Dion gneiss. Note also that garnet - biotite mylonite is too fine grained to determine the presence or absence of sillimanite in the field. Abbreviations are Daisy (D), Dion (Di), Laloche (L), McKee (M), Schist (S) Lakes, Beach (B) and Snowgoose (G) lakes, Austin (A), McDonald-Wilson (MW) and McKee (M) faults. Discussed in text.

[1] Note: Hanmer and Lucas (1985) identified misoriented xenoliths of fine schistose material enclosed within isotropic McDonald granite, just northwest of Hornby Channel ultramylonite belt, as inclusions of recrystallised (annealed) ultramylonite. From our experience this summer and after re-inspection of the pertinent outcrops we find that the inclusions mentioned are schistose metasedimentary inclusions, similar to parts of the external paragneiss of the Schist-Daisy segment. Therefore, the interpretation of Hanmer and Lucas (1985) that the McDonald granite is syntectonic with respect to Hornby Channel ultramylonite belt is erroneous.

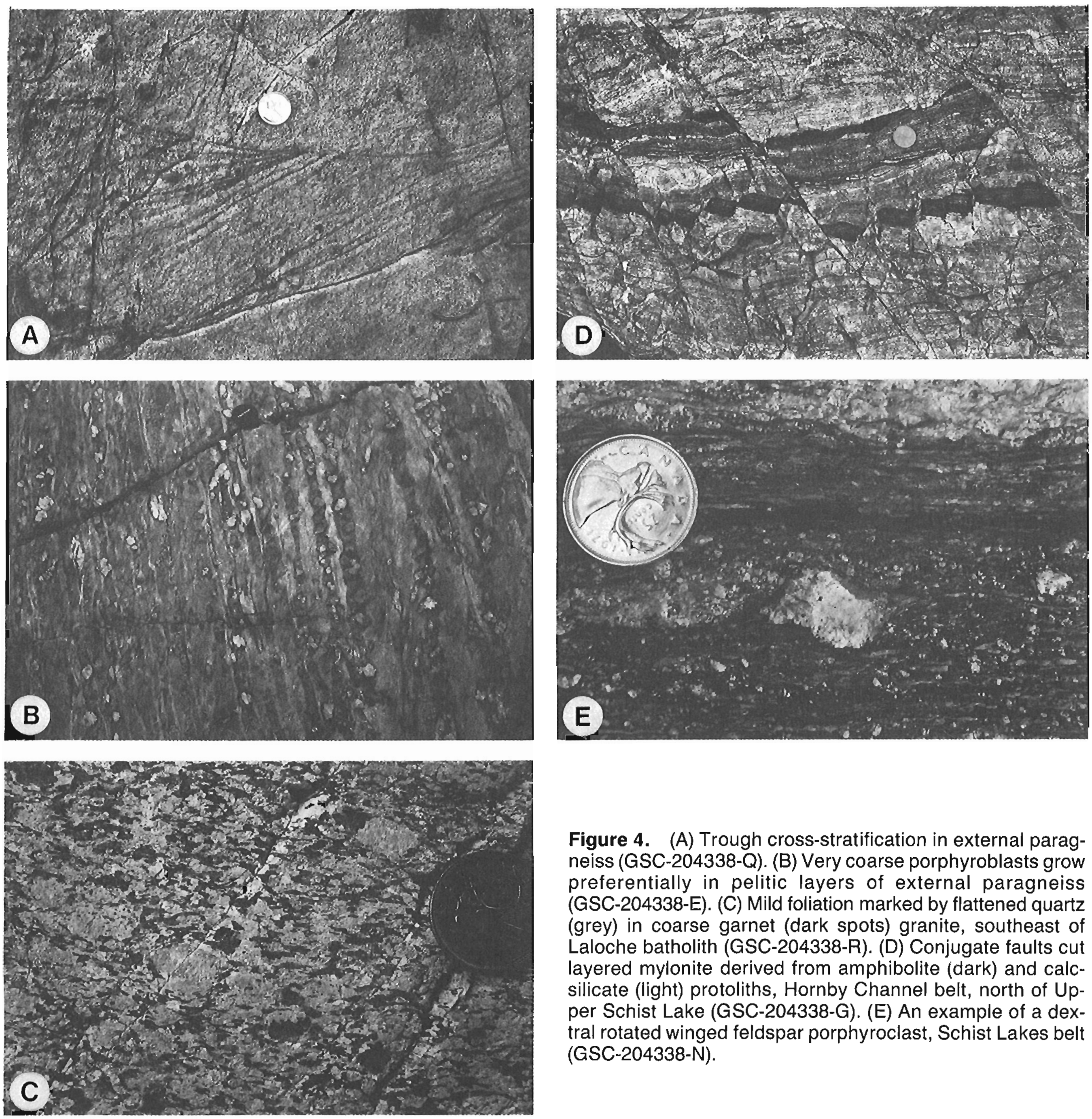
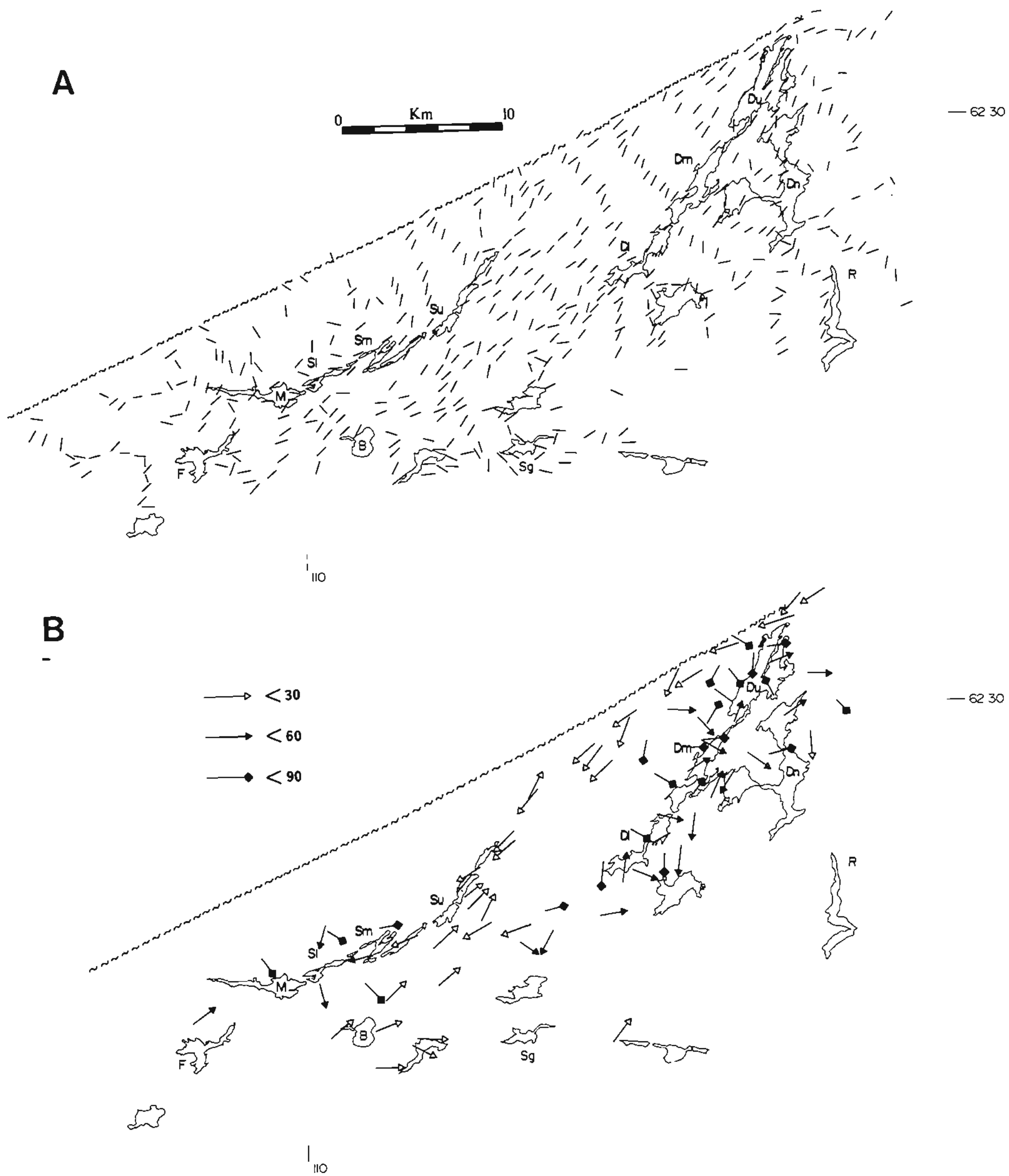


Figure 4. (A) Trough cross-stratification in external paragneiss (GSC-204338-Q). (B) Very coarse porphyroblasts grow preferentially in pelitic layers of external paragneiss (GSC-204338-E). (C) Mild foliation marked by flattened quartz (grey) in coarse garnet (dark spots) granite, southeast of Laloche batholith (GSC-204338-R). (D) Conjugate faults cut layered mylonite derived from amphibolite (dark) and calc-silicate (light) protoliths, Hornby Channel belt, north of Upper Schist Lake (GSC-204338-G). (E) An example of a dextral rotated winged feldspar porphyroclast, Schist Lakes belt (GSC-204338-N).



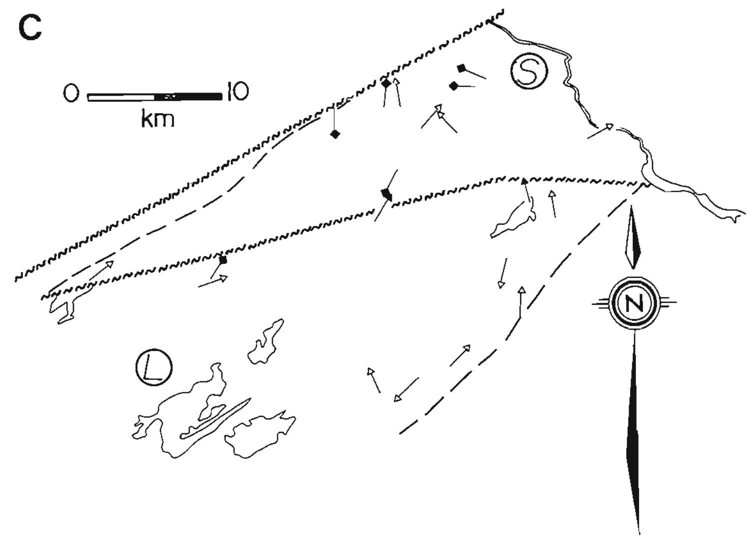


Figure 5. Structural elements of Schist-Daisy and Snowdrift segments of Great Slave Lake shear zone. Abbreviations are Daisy (Du,m,l), Dion (Dn), Face (F), McKee (M), Robert (R) and Schist (Su,m,l) Lakes, Beach (B) and Snowgoose (SG) lakes. Mcdonald-Wilson fault is shown. (A) Foliation trace. All dips are greater than 60 degrees and are omitted for clarity. Note the irregular foliation trends on either side of the corridor formed by the Schist Lakes and Daisy Lakes mylonite belts, especially the folded foliation in heterogeneously to mildly foliated hornblende - biotite granite west of Face Lake, as well as the north - south strike of foliation in muscovite granite northwest of Schist Lakes. (B) Extension lineations, expressed as streaked-out quartz feldspar aggregates, quartz aggregates and mineral lineations, for Schist-Daisy segment, presented in three plunge classes for clarity. Note the shallow plunges except in the Daisy Lakes belt and in the external paragneisses northwest of Schist Lakes. (C) Extension lineations for Snowdrift segment and northeastern end of Laloche segment, presented in three plunge classes as in B. Dashed boundaries correspond to northwestern and southeastern limits of the Laloche batholith.

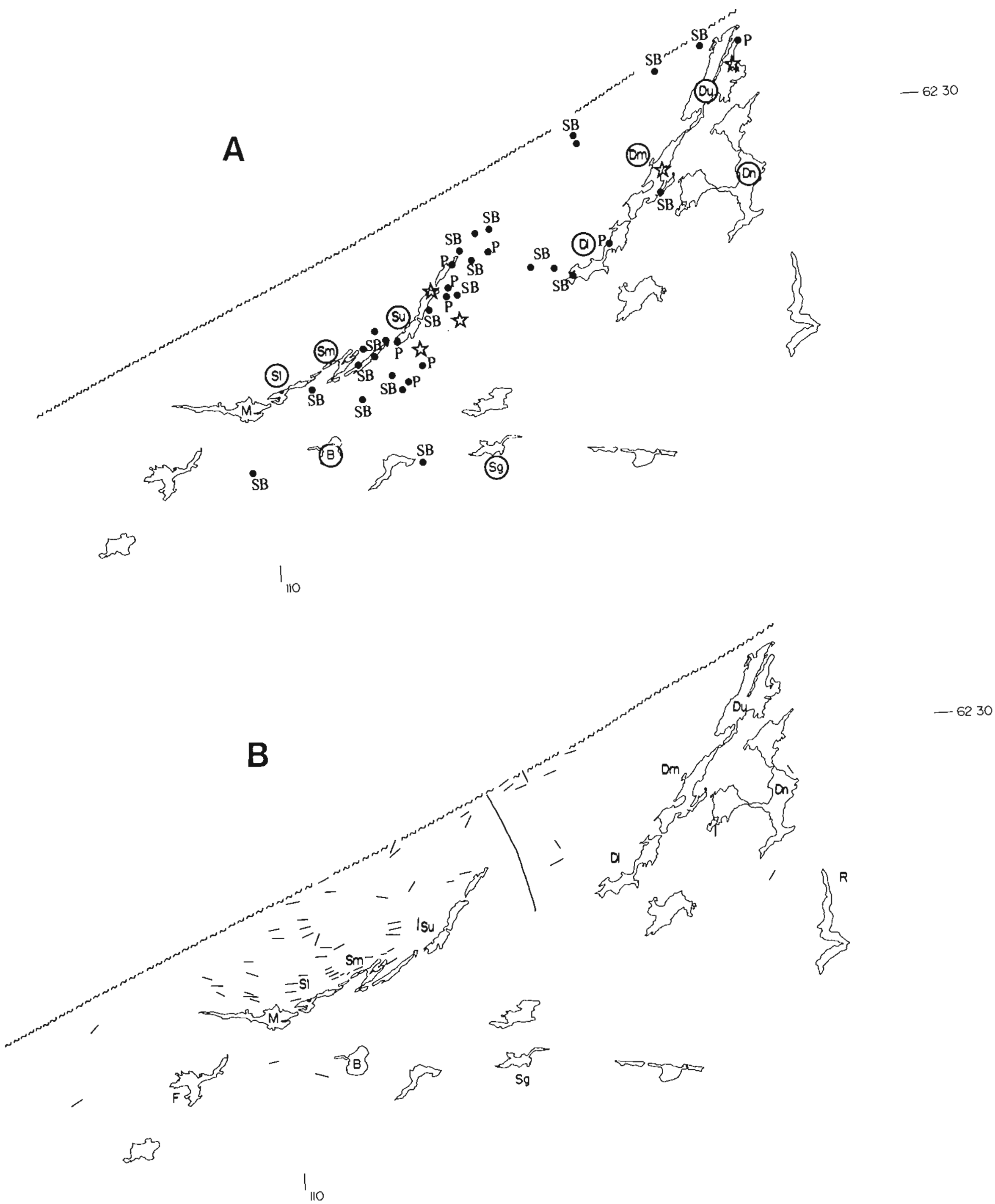


Figure 6. (A) Transcurrent dextral kinematic indicators. Asymmetrical extensional crenulation or shear band foliation (SB; White et al., 1980), rotated winged feldspar porphyroclasts (P; Hanmer, 1984; Passchier and Simpson, 1986). Stars indicate other structures including 'C & S' planes (Berthé et al., 1979), climbing pegmatites (Hanmer, 1984), rotated fold axial planes, asymmetrical pull-aparts (Hanmer, 1986). (B) Amphibolite and diabase dykes. Dykes of 150° trend are diabase. Most others are amphibolite. Most dykes northwest of Schist Lakes are isotropic. Note the east-west dyke trend in the vicinity of McKee and Lower Schist Lakes. We speculate that this east-west 'grain' may have influenced the location of the McKee fault. Abbreviations as in Figure 5.

evolve into white ribbon ultramylonites whereas the metasediments show only a subtle decrease in grain size and increase in fissility. However, where the metasediment protolith contained granite or pegmatite veins, the resulting tectonite is either a banded schist with granitic ribbon ultramylonite layers or a remarkably porphyroclastic schist (Fig.2D). All intermediate stages are observed from coarsely to progressively very finely porphyroclastic schist. Calc-silicate mylonites occur locally between Middle Schist Lake and McDonald-Wilson fault (Fig. 4D).

An extension lineation porpoises gently to moderately in a vertical foliation along the length of the belt (Fig.5). A structural assemblage of rotated winged porphyroclasts, asymmetrical extensional crenulation cleavage and asymmetrical extension fractures in porphyroclasts gives a consistent dextral sense of shear (Fig.6B). With the exception of the presence of calc-silicate mylonites, this belt is identical to the Hornby Channel ultramylonite belt of Hanmer and Lucas (1985).

Schist Lakes belt

A belt of finely homoclastic (Hanmer, in press(b)) ribbon protomylonites borders the southeastern flank of, and is reworked by, the Hornby Channel belt (Fig. 2, 5). The belt changes character both along and across strike. To the southwest, in the vicinity of Schist Lakes, the belt is 4 km wide. The protomylonites contain uniformly round small (1-2 mm) porphyroclasts of plagioclase set in a K-feldspar and quartz ribbon matrix. Locally coarse (2 cm) dispersed porphyroclasts, trains of very coarse (4-5 cm) porphyroclasts after pegmatite and locally preserved volumes of coarse grained (2 cm) granitoid protolith are indicative of the degree of grain size reduction associated with the production of the tectonites. Disruption and pull-apart structures involving competent amphibolite layers in the protomylonite indicate minimum principal extensions of several hundred percent. Inclusions and large mappable screens of paragneiss occur throughout the belt. The paragneiss comprises an assemblage of lithologies strikingly similar to the 'supracrustal assemblage' of the Laloche segment (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986); (a) layered graphitic sillimanite — garnet meta-pelite, often intensely migmatized (metatexite and diatexite), (b) less aluminous biotite-rich, sometimes garnetiferous, metasediment, (c) banded amphibolite, and (d) calc-silicate. Whereas only minor quantities of quartz-rich metasediments were previously reported from the Laloche segment, subfeldspathic meta-arenites are a significant component of the paragneisses of the Schist-Daisy segment. In the southwestern part of the Schist Lakes belt, the stable metamorphic assemblage is sillimanite - garnet - biotite - K feldspar - quartz - graphite. Muscovite is absent. This upper amphibolite facies assemblage is seen in both coarsely feldspar porphyroblastic paragneisses (metatexites) and in their ribbon (proto)-mylonite equivalents. Therefore, mylonitization took place at upper amphibolite facies conditions in the Schist Lakes belt.

To the northeast, the belt narrows progressively to less than a kilometre. There is also a textural transition within the granitic tectonites from finely homoclastic protomylonite to finely homoclastic mylonite. Furthermore, the porphyroblastic metatexites are reduced to garnet-biotite schists, locally with fine feldspar porphyroclasts. A similar transition occurs across strike towards the northwest within the sector bounded by the North-South fault and Upper Daisy Lake (Fig. 7). The across-strike transition is absent between the North-South fault and Schist Lakes where the upper amphibolite assemblages are visibly reworked by the Hornby Channel belt. The significance of these textural and metamorphic transitions are discussed below.

The tectonites of the Schist Lakes belt are characterized by a sub-horizontal extension lineation (Fig. 5A) and a structural assemblage of rotated winged feldspars and other inclusions (Fig. 4E), asymmetrical extensional crenulation cleavage and pull-aparts, rotated fold axial planes, all indicating a dextral sense of shear (Fig. 6B). We are impressed by strong resemblance of the southwestern part of the Schist Lakes belt to belt no. 2 of the Laloche segment (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986).

Daisy Lakes belt

A 5 km-wide belt of mixed tectonites, previously identified as Yellowknife Supergroup (Wright, 1968), forms a corridor through Daisy Lakes and Dion Lake (Fig. 2 and 5). The belt comprises two sub-belts. The northwestern sub-belt occupies the ground to the northwest of and within Daisy Lakes. It is mainly composed of finely homoclastic ribbon mylonites, demonstrably derived from a coarse biotite megacrystic granite, and associated inclusions of mylonitic clinopyroxene metatonalite and migmatized metasediment. The latter contain the stable upper amphibolite facies muscovite-free assemblage sillimanite garnet - biotite - K-feldspar - quartz - graphite. The southeastern sub-belt occupies the ground between Daisy Lakes and part of Dion Lake. It is an intimate mixture of homoclastic garnet - orthopyroxene(?) ribbon mylonite derived from garnet - orthopyroxene - biotite megacrystic granite, very coarse migmatitic sillimanite - garnet biotite - K-feldspar - quartz graphitic metapelite, marble, calc-silicate and coarsely foliated to platy clinopyroxene metatonalite. Clearly the southeastern sub-belt contains granulite facies protoliths and mylonites, the metamorphic grade of which was first recognized in the mid-1960s by E.W. Reinhardt (GSC, unpublished data). The coarse paragneisses have probably recrystallized post to late tectonically judging by the occurrence of randomly oriented sillimanite blades. We cannot determine whether the often coarsely foliated metatonalite resisted/escaped mylonitization or has subsequently recrystallized. Tectonites of the southeastern sub-belt are cut by both concordant and discordant sheets of poorly foliated hornblende - biotite granitoid, hundreds of metres wide, which strongly resemble the Snowdrift granite to the southeast.

All of the tectonites of the Daisy Lakes belt carry a vertical foliation and a rectilinear dip-parallel extension lineation (Fig. 5). Field shear-sense indicators are few and follow-up study of oriented samples is required before drawing kinematic conclusions. This said, field indications (asymmetrical extensional crenulation cleavage and rotated winged porphyroclasts) suggest southeast-side-up movement during the formation of the mylonites. However, dextral 'C & S' fabrics, asymmetrical extensional crenulation cleavage, rotated winged porphyroclasts and obliquely intruded boudined (climbing) pegmatites indicative of dextral transcurrent shear are also observed in the horizontal ('Y/Z') plane. In view of the rectilinear nature of the dip-parallel extension lineation, we suggest that the transcurrent component is relatively weak and relatively late compared to the dip-slip movement.

'Sandwich' granite

The strike lineated protomylonites of the Schist Lakes belt are separated from the dip lineated mylonites of the Daisy Lakes belt by a 2-3 km wide belt of equigranular, medium grained, mildly foliated to isotropic biotite granite (Fig. 7), whose informal name derives from its location between two mylonite belts. The 'Sandwich' granite is intrusive with respect to the Schist Lakes belt protomylonites, large recrystallized (annealed) xenoliths and rafts of which are included within the granite close to the contact. Smaller xenoliths of the same are scattered throughout the granite and are often "assimilated" to the extent that only the biotitic mafic bands of the misoriented and folded granitic xenoliths permit their distinction from the enclosing homogeneous granite. The Schist Lakes belt protomylonites are intruded, disrupted (broken apart, misoriented and folded) and recrystallised (annealed) within a corridor 500 m wide adjacent to the 'Sandwich' granite. From the foregoing, the 'Sandwich' granite closely resembles the similarly located equigranular granite and its vein aureole which cuts the protomylonites of belt no. 2 in the Laloche segment (Hanmer and Connelly, 1986). On its southeastern side, the 'Sandwich' granite is heterogeneously sheared with the local development of narrow (< 100 m) mylonite zones. These mylonites are dip-lineated and pass imperceptibly into the upper amphibolite facies mylonites of the northwestern sub-belt of the Daisy Lakes belt. Clearly the 'Sandwich' granite separates the flanking mylonite belts in time as well as in space.

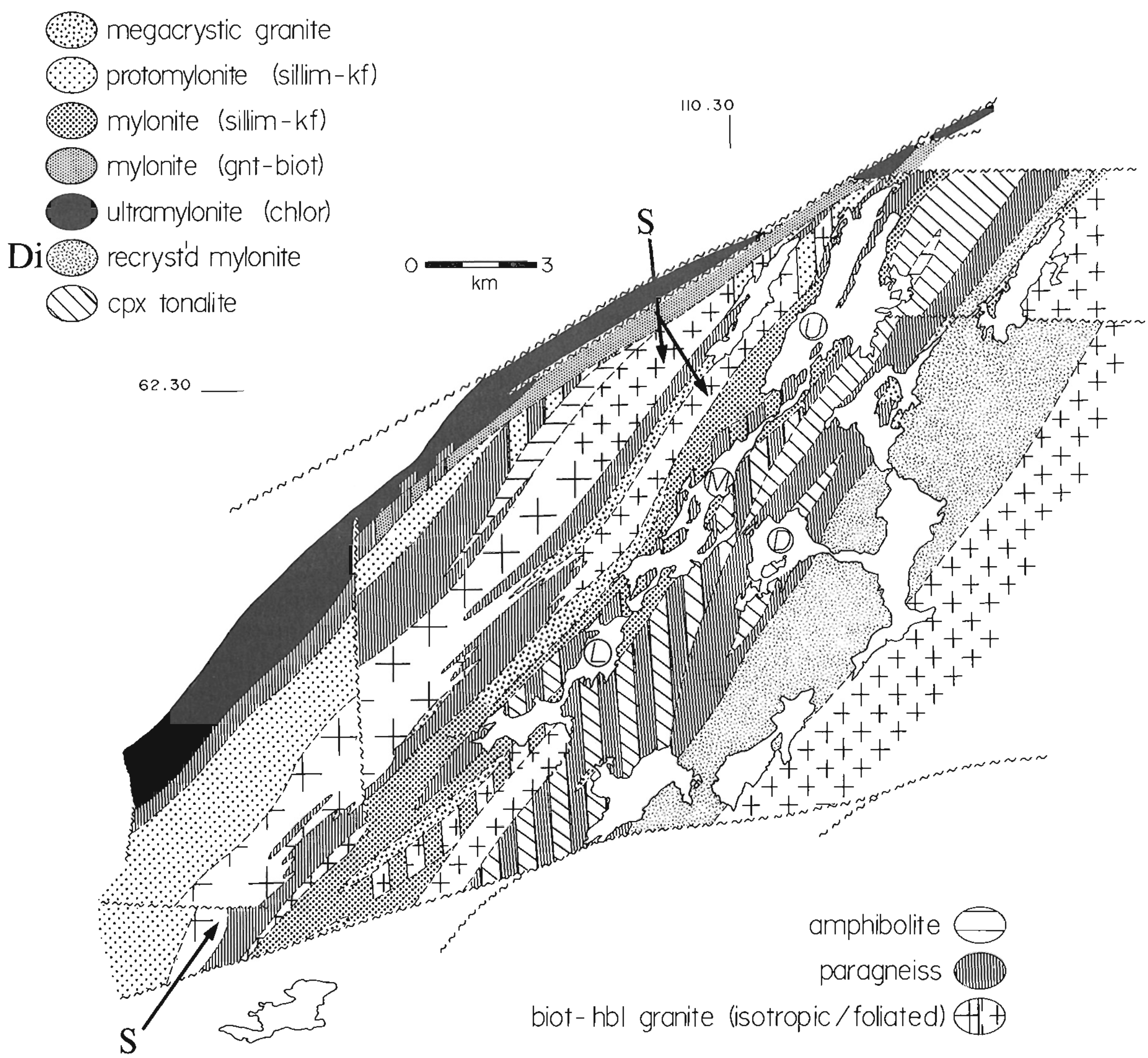


Figure 7. Detail of the geology of the northeast Schist Lakes belt, the isotropic to foliated hornblende - biotite granitoid 'Sandwich' granite and the Daisy Lakes belt. Bar shading indicates map units of mixed lithologies. Note that the fingering-out of the paragneiss is schematically illustrated. The meridional fault to the west is referred to as the North-South fault. Di is Dion gneiss; S is 'Sandwich' granite. Abbreviations are Dion (D) and Schist (U,M,L) Lakes.

While the foregoing is applicable to 90 % of the strike length of the 'Sandwich' granite, time relations at its northeastern end at the latitude of Upper Daisy Lake are somewhat more complex. The garnet-biotite upper greenschist(?) mylonites at the northeastern end of the Schist Lakes belt clearly crosscut the contacts between the Schist lakes belt upper amphibolite facies protomylonites and the 'Sandwich' granite. Furthermore, the northeastern end of the 'Sandwich' granite is a biotite-bearing finely homoclastic protomylonite, demonstrably developed at the expense of the granite. Although we have yet to confirm the metamorphic grade of the fine grained mylonites, it would appear that post-Daisy Lake belt mylonitisation was concentrated on the northwestern side of the Schist-Daisy segment, even before the formation of chlorite-bearing mylonites in Hornby Channel belt. It is possible that the garnet — biotite mylonites represent local preservation of an upper greenschist precursor to the Hornby Channel belt (see Miscellaneous mylonites).

Miscellaneous mylonites

The Daisy Lakes belt is flanked to the southeast by a belt of fine to medium grained (< 1 mm), finely banded hornblende-biotite granitic sugary gneisses with thin continuous amphibolite layers, which underlies much of Dion Lake (Fig. 2 and 5). We refer to these as the Dion gneiss. The upright amphibolite layers are thrown into vertical folds, boudined and disrupted. Coarse, poorly foliated sheets of hornblende-biotite granitoid, hundreds of metres thick, are intruded into the gneisses. Locally the gneisses carry garnet and are demonstrably derived by the deformation of coarse garnet — orthopyroxene(?) megacrystic granitoid protolith. Elsewhere, they locally preserve both quartz and feldspar ribbons (Fig. 2E). Therefore, they are annealed high grade(?) sugary mylonites derived from a once coarse granitic protolith with mafic sheets and/or inclusions. Locally this sugary mylonite - granitoid assemblage is remylonitized in narrow (< 1 m) vertical belts, similar to those described above from the Snowdrift granite to the southeast. The contact between the Dion gneiss and the Snowdrift granite is gradational and intrusive, but is poorly exposed; we emphasise the highly subjective nature of our separation of wallrock with abundant intrusions from pluton with abundant inclusions. The Dion gneiss does not contain the abundant metasediment and clinopyroxene metatonalite components which characterize the adjacent Daisy Lakes mylonite belt. However, in view of the mylonitic nature of the gneiss, its apparent high metamorphic grade and its relationship to the Snowdrift granite, we suggest that it has affinities to, and constitutes the intruded, disrupted and recrystallized equivalent of, the Daisy Lake belt.

A number of discontinuous, geologically disparate, map-scale (500 -1500 m wide) vertical mylonite zones occur at the latitude of the boundary of the Schist-Daisy and Snowdrift segments (Fig. 3). They are all developed from coarse hornblende - biotite granitoid protolith. Although all are biotite-bearing, the finely homoclastic mylonites to ultramylonites do not contain metamorphic mineral assemblages diagnostic of metamorphic grade. While most are ribbon mylonites, an important sugary mylonite zone running east - northeast from Beach lake is now so thoroughly recrystallised that it is isotropic at the grain scale (Fig. 2). By working out progressive grain size reduction and transposition into the zone from the southeast and identifying isoclinal and rare sheath folding of the transposed layering (Fig. 2F), plus relict centimetric feldspar porphyroclasts within the zone, one can recognize that these fine grained gneisses are recrystallised (annealed) mylonites. Despite the destruction of linear fabric elements by annealing, the orientation of isoclinal and rare sheath fold axes indicates that the finite principal extension direction is down-dip. Apart from the annealed fabric, these mylonites resemble, and are probably part of, the dip-lineated mylonites of the Daisy Lakes belt from which they are separated by an east - northeast trending fault.

The ribbon mylonite zones east and west of Snowgoose lake and southwest of Beach lake and McKee Lake show variable lineation plunges (Fig. 2 and 3B). Poor outcrop hindered extensive observation of the lineation. Our impression, however, is that the extension lineation in the ribbon mylonites is predominantly shallow plunging. We emphasise here that we can neither relate these mylonites to the major mylonite belts described above, nor to each other.

Union Island and Sosan Group outliers

Two outliers of mildly to uncleaved sediments occur at Schist Lakes (Fig. 8). The smaller of the two occurs as a small island plus a mainland outcrop at Middle Schist Lakes, on the northwestern boundary of Hornby Channel belt. It comprises a fine grained, finely laminated dolomite with millimetre-thick quartz laminae and centimetre-scale quartz void-fillings. This rock is reminiscent of dolomites of the Union Island Group (Hoffman et al., 1977), with which it may be correlative. It presumably occurs within a fault bound block.

The larger outlier (20 km by 3 km) extends from McKee to Upper Schist Lakes along the southeastern side of Lower and Middle Schist Lakes (Fig. 8; Wright, 1968). It comprises a 250 m thick sequence of quartz arenites, feldspathic arenites and conglomeratic to pebbly granulestones preserved in a doubly plunging syncline. The outlier occurs at the contact between the Schist Lakes and Hornby Channel belts; its contacts with both are generally faulted. The sense of faulting cannot be deduced; the faulted contact is parallel to the foliation in the mylonites and is very sharp, showing no sign of brecciation, fracturing or discolouring. There are two exceptions, both occurring in Upper Schist Lake (Fig. 8). The northwestern contact is very locally mylonitized. Coarse feldspathic sandstones can be traced into a < 10 m wide, upright, strike-lineated quartz mylonite adjacent to extremely flinty, ultrafine reworked ultramylonite demonstrably derived at the expense of ribbon mylonite of the Hornby Channel belt. While no field kinematic analysis was possible in such fine, flinty mylonite, we suspect that it represents a late, low temperature episode of transcurrent shear.

On the southeast side of the lake, an angular unconformity is exposed over several hundreds of metres at the contact between the outlier and finely homoclastic granitic protomylonites and a mylonitized assemblage of quartzites, magnetic iron formation, garnet - sillimanite - andalusite metapelites and pegmatites (Fig. 9A and B). The vertically

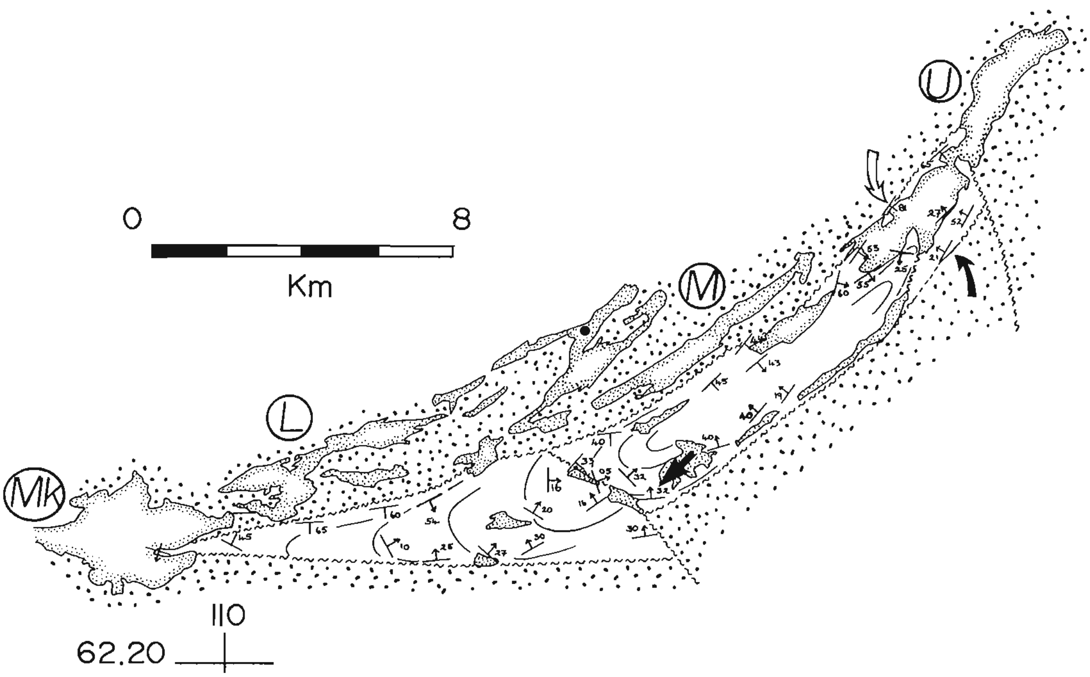
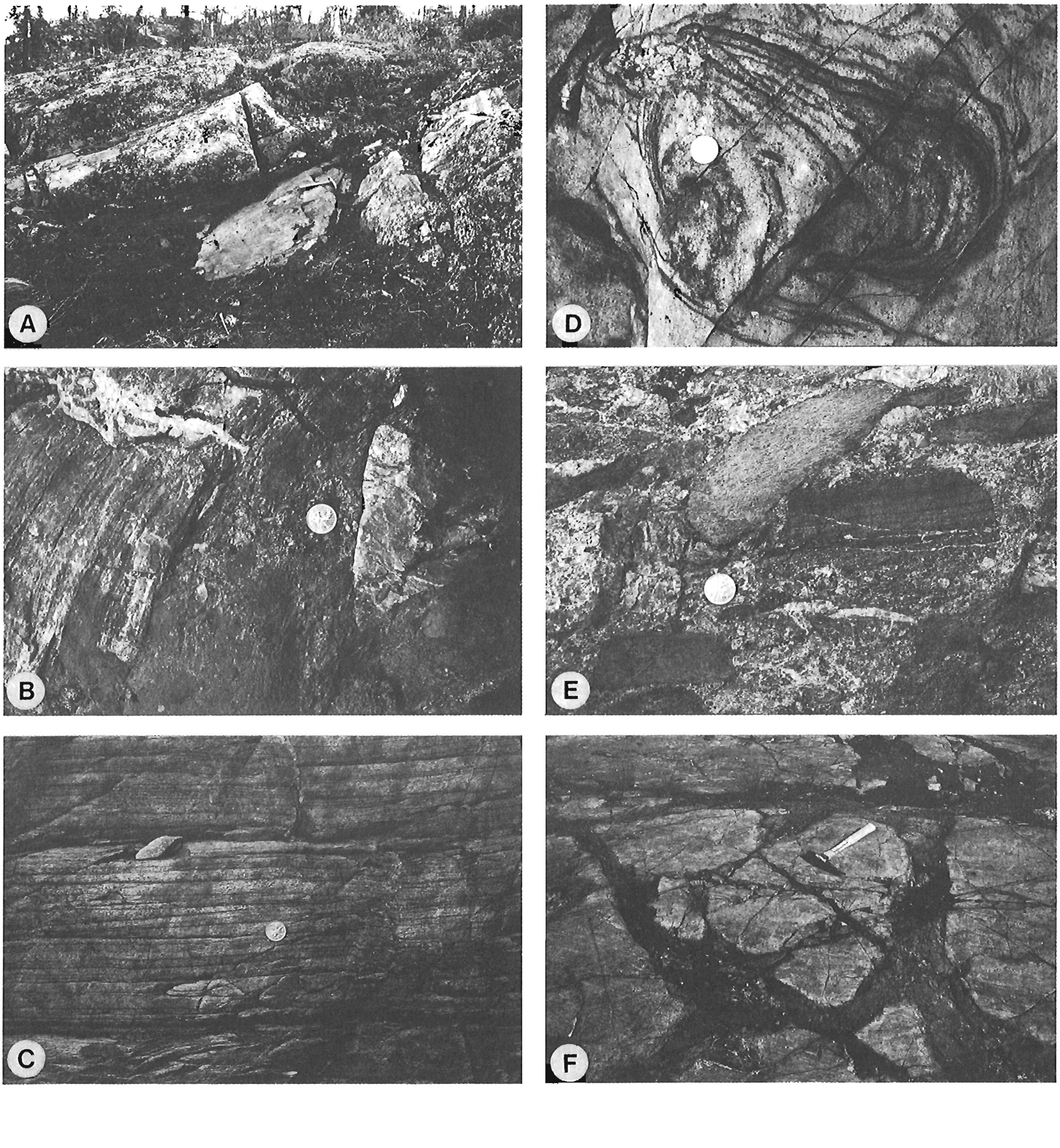


Figure 8. Detail of the Schist Lakes outlier of Sosan Group quartz arenites and conglomeratic pebbly granulestones. The outlier is bounded by the faults shown except at the curved black arrow where it lies unconformably on mylonites and protomylonites of the Schist Lakes belt and at the curved open arrow where a narrow strip of very fine and flinty ultramylonite reworks the contact with the Hornby Channel belt. Bedding orientations, stratigraphic tops and interpolated traces of bedding indicate that the outlier is an open doubly plunging syncline. The black dot locates the outcrop of Union Island laminated dolomite; straight black arrow locates post-cleavage intrusive felsite. Dotted pattern is basement. Abbreviations are McKee (Mk) and Schist (U,M,L) Lakes.

foliated mylonites are overlain by a thin (< 10 cm) basal breccia lag with mylonite clasts, and 10m of thicker (< 1 m) intercalated beds of grey-maroon mudstone and quartz arenite deposited upon an irregular erosion surface of low relief. The unconformity and the bedding dip to the northwest at about 30-40 degrees. Immediately below the unconformity the mylonites are strongly altered and enriched in haematite in a zone several metres thick, presumably representing a weathering horizon. Also confined to the immediate vicinity of the unconformity are vertical sheets, up to 1m wide, of a coarse quartz-feldspar rock composed of large sub-angular grains. The sheets are confined to below the unconformity and are concordant with the mylonite foliation. We tentatively suggest that the quartzo-feldspathic sheets are a texturally and compositionally immature sediment trapped in surface cracks and crevices and preserved below the unconformity. The grey-maroon mudstone is confined to the lower 5 m of section, above which the quartz arenites are overlain by a 5m thick, coarse (<25 cm) polymictic conglomerate containing pebbles of mylonitized quartzite, hematitic weathered horizon material, vein quartz and intraclasts of grey-maroon mudstone. The conglomerate either cuts down to or on-laps the mylonitic basement where it has a strongly erosional irregular base.

The conglomerate is overlain by a compositionally mature quartz arenite, bedded on a 25-100cm scale with well developed trough cross-stratification, convolute lamination, ripples and plane-bed structure outlined by thin (< 1mm) heavy mineral lags (Fig.9C). Thin (< 10cm) pebbly beds occur sporadically as low angle lags within channels. A small number (6) of unambiguous intersections of trough cross-stratification with regional bedding (Fig.9D) consistently indicate that paleocurrents flowed from northeast to southwest. Such quartz arenites are found throughout the section, but are most common in the lower part. The upper half to two thirds of the section is dominated by coarse feldspathic quartz

Figure 9. (A) Exposed unconformity at the base of the Sosan Group, Schist Lakes outlier. The precise contact lies just to the right of the hammer head. The arenites and mudstones above the contact (left field) dip gently to the left (northwest). The mylonites below the contact (right field) are vertical (GSC-204338-. (B) Detail of the unconformity in A. A triangular block of quartzite mylonite (lower field) is detached from the main mylonite mass (upper field) and is supported by Sosan arenite. Mylonite foliations dip steeply into the picture. In the main mylonite mass, foliation is oriented top left - bottom right, while in the fragment it is oriented top right - bottom left (GSC-204338-M). (C) Plane bedded Sosan quartz arenite. Bedding marked by heavy mineral bands (GSC-204338-O). (D) Trough cross-stratification intersection with regional bedding in Sosan quartz arenite, marked by heavy mineral bands. Paleocurrent flowed towards the southwest, away from observer (GSC-204338-I). (E) Heterolithic conglomerate with granulestone matrix. Note bedded sandstone pebble (dark, right of centre) (GSC-204338-S). (F) Felsite emplaced into dilated system of fractures cutting cleaved Sosan quartz arenite, Schist Lakes outlier, Middle Schist Lake (GSC-204338-J).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arenite to granulestone passing up into polymictic pebbly and often conglomeratic granulestone and conglomerate (Fig.9E). On some traverses across the outlier, the quartz arenites pass abruptly into the coarser sands with 20 % pink feldspar grains (< 5mm) at an horizon marked by channels, several metres wide, which coalesce up-section into laterally continuous, internally cross-stratified beds. The abundance of conglomeratic channel fill and beds increases up-section and consequently is best developed at the latitude of Middle Schist Lake. In descending proportion the pebble types include vein quartz, a fine grained siliceous rock, various arenites either exotic or resembling those lower in the section, quartz-phyric rhyolite and rare poorly foliated granitoids. The "siliceous rock" pebbles may include a proportion of rhyolitic clasts. We note the near absence of mylonite clasts throughout the section above the basal conglomerate. We also emphasise that very few of the pebbles were cleaved prior to incorporation, again with the exception of the basal part of the section.

The outlier is folded into a moderate, upright, subhorizontal, doubly plunging syncline. A vertical axial planar cleavage is visible on suitably etched and weathered surfaces in the finer quartz arenites. It is also marked in pebbly to conglomeratic granulestones by cleaved and elongate fine grained siliceous (rhyolitic?) pebbles. No other pebbles are visibly distorted by the folding strain and the matrix granules are only rarely cleavage forming elements. Presumably other pebble lithologies were too stiff to deform and the matrix granules accommodated the folding strain predominantly by particulate flow (Borradaile, 1981). The outlier cleavage and the foliation of the underlying basement mylonites are co-planar. It is improbable that the horizontal shortening associated with deformation of the outlier has modified the orientation of the basement foliation. Furthermore, the shortening of the outlier is too weak to have significantly reoriented the extension lineation in the basement mylonites. The cleavage in the outlier is cut by an undeformed, deep pink, very fine grained, locally porphyritic intrusion emplaced into a series of linked and dilated fractures (Fig.9F) which strongly resemble similar, but less dilated structures in diatremes described from the East Arm of Great Slave Lake (Reinhardt, 1972).

SNOWDRIFT SEGMENT

The geology of this complex segment comprises a number of elements (Fig.3), not all of whose mutual relationships are yet understood. The segment is dominated by a broad swath of Snowdrift granite, identical to that already described above from the Schist-Daisy segment. Here too, the granite is bordered to the southeast by a white garnetiferous granite and is cut by 500-1000 m-wide mylonite zones already described (see Miscellaneous mylonites). We would only add that the garnets of the garnetiferous granite tend to be strongly hornblendized in this segment.

To the west, between the Snowdrift and McDonald granites, is a very complex terrane (Fig. 3). Lithologically it resembles parts of the northeastern end of the Laloche segment, from which it is separated by the Austin fault. We emphasise that we did not directly observe the Austin fault on the ground. Its existence is deduced from the abrupt change of strike across a marked linear break in the aeromagnetic field which also coincides with a fundamental change in the geology on the ground. South of the Austin fault we have extended belts 1, 2 and 3 of the Laloche segment (Hanmer and Lucas, 1985) up to the Austin fault itself (Fig. 3). We can add little to previous descriptions, except to note that the "deceptively bland", sugary hornblende-biotite $\pm$ garnet granitoid mylonites ("granoblastic gneisses"; Hanmer and Lucas, 1985, p13) completely dominate belt 1 northeast of Laloche Lakes. Sense of movement on the Austin fault cannot be directly observed and no correlations across it are intended in what follows.

The 030° striking gneisses lying between the Snowdrift and McDonald granite are dominated by upper amphibolite facies finely homoclastic protomylonites in the west and recrystallized granitoid mylonites in the east. Both closely resemble similar tectonites on the south side of the Austin fault (see Hanmer and Lucas, 1985). The protomylonites are demonstrably derived by transposition and mylonitization of upper amphibolite facies granitic, amphibolitic and pelitic gneisses whose once straight layering has been folded, boudined and intruded by granite sheets ("disrupted straight gneiss" of Hanmer and Lucas, 1985). Such disrupted gneisses are preserved in a narrow strip. While the body of granite to the west of the disrupted gneiss may be the source of the intruded granite sheets, its relationship to the gneisses remains equivocal. The recrystallized granitoid mylonite to the east is flanked on either side by upper amphibolite facies ribbon mylonites. The striking feature of the ribbon mylonites is their very well developed steeply plunging extension lineation (Fig. 5C). On account of their textural preservation, we tentatively suggest that the ribbon mylonites are the youngest tectonites in the western Snowdrift segment.

Allowing for the reconnaissance nature of our mapping in this segment, we suggest that older recrystallized and disrupted mylonites have been reworked initially by a broad zone of protomylonites and subsequently by narrow zones of mylonite. While the relationship of the Snowdrift granite to the mylonites remains indeterminate, we would suggest that it is younger.

FAULT ZONES

A number of important faults mark major geological boundaries within Great Slave Lake shear zone. The McDonald-Wilson fault outcrops in our map area between McDonald and Daisy Lakes (Fig. 3). Easily identified by the topographic scarp it engenders, it is almost undetectable at the outcrop scale. However, this generalization breaks down to the northeast of Snowdrift River as the fault is frequently marked by a 5-10 m thick vertical zone of cataclasite; a fine grained featureless non-foliated matrix containing less than 50 % by volume of rounded porphyroclasts and rock fragments (Sibson, 1977). The McDonald-Wilson fault truncates, and is therefore younger than, all other faults described in this report.

The McKee fault, which marks the boundary between Schist-Daisy and Snowdrift segments, is a zone of anastomosing fault strands, oriented approximately 100°, and running through McKee, Beach and Snowgoose lakes (Fig. 3). Two sets of intersections both indicate a dextral separation of 35-40 km. The first set corresponds to the intersections of the boundary between the Snowdrift granite and the garnetiferous granite with the fault. Its southern intersection occurs at Beach lake (Fig. 3) while its northern intersection is the western side of the marked 030° trending negative aeromagnetic anomaly lying just east of our map area (Geological Survey of Canada, 1984). The second set corresponds to the intersections of Hornby Channel belt with McKee fault, after removal of the effect of the McDonald-Wilson fault. The northern intersection occurs at McKee Lake (Fig. 3) while the constructed southern intersection occurs at a point 35 km to the west. Northeast-trending faults apparently branch off from the eastern part of McKee fault (Fig. 3). The east- and northeast-trending faults are collectively referred to as the McKee fault zone. The following description is a composite picture which applies throughout the fault zone, although elements may be locally absent.

The oldest tectonites associated with the fault zone are biotite-bearing granitoid ribbon mylonites with a vertical foliation and a shallowly plunging extension lineation (Fig. 5). The mylonites form zones up to several hundred metres wide. Mylonites and non-mylonitized wallrock are reworked by penetrative cataclasis resulting in zones, up to 100 m thick, of cataclasite, splendid examples of which form the cliffs of the south shore of McKee Lake. Brittle faulting associated with macroscopic dilation and quartz veining occurs at various scales. In places, a penetrative shattering of the wallrock is high-lighted by centimetre scale quartz veinlets developed throughout the rock volume. Elsewhere, composite quartz veins are tens of metres thick and carry misoriented fragments of wall rock (quartz stockwork). Several examples of thoroughly mylonitized quartz stockwork were identified at McKee Lake and suggest that evolution of the fault zone through the brittle-ductile transition was associated with transient excursions from one behavioural field to the other (Sibson, 1980). Pseudotachylite is only developed at a very local scale. As given, this description is directly applicable to the McKee fault. Furthermore, it adequately describes the geology of the un-named east-west fault located just north of Upper Daisy Lake (Fig. 2 and 5). The development of mylonite is less extensive in the northeast-trending strands of the McKee fault zone, which are dominated by cataclasite and brecciation.

A wide zone of brecciation and cataclasis, 100 m thick, marks the continuation of Hornby Channel belt northeast of the intersection made by the Austin fault (Fig. 3). The zone comprises large (10m) blocks of chlorite-bearing ultramylonite and penetratively shattered granite in a matrix of cataclasite. We consider that the mylonite blocks represent relicts of Hornby Channel belt which has been cut out by brittle faulting. We cannot determine whether this faulting is of pre- or syn-McDonald-Wilson fault age. However, given the extensive development of cataclasis along this fault and the minor cataclasis associated with the McDonald-Wilson fault, we suggest that the Hornby Channel belt was cut out along this section prior to activity on the McDonald-Wilson fault.

The Austin fault comprises an easterly trending part to the east and a northeasterly trending part to the west (Fig.3). The latter intersects, but does not offset, Hornby Channel belt and its cataclastic extension (Fig.1). Either it is earlier than Hornby Channel belt, or it roots into the zone of cataclasis and brittle faulting which affects the ultramylonites (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986). West of McKee Lake, the McKee fault turns to a more northeasterly orientation, geometrically resembling the Austin fault (Fig.3). It strikes into, and is clearly truncated by, the McDonald-Wilson fault. However, the McKee fault truncates Hornby Channel belt at McKee Lake. The presence of early biotite-bearing mylonites in the McKee fault suggests that the fault's mylonitic history commenced prior to the initiation of the chlorite-bearing Hornby Channel belt and that its later cataclastic and fracture associated history continued into post-Hornby Channel belt time. Although the intersection of McKee fault with the southwestern section of Hornby Channel belt is covered by younger sediments northwest of the McDonald-Wilson fault, we can speculate upon its nature. In one scenario the Hornby Channel belt would simply cross-cut the McKee fault mylonites. In a second scenario, the McKee fault mylonites would root into what is now chlorite-bearing Hornby Channel belt and represent preserved early biotite grade mylonites whose equivalents have been obliterated within the chlorite-bearing ultramylonites. The cataclasites would root into the present Hornby Channel belt and would link along strike with the cataclasites which now cut out the chlorite ultramylonites northeast of the Austin fault intersection. In either case, much of the displacement of Hornby Channel belt would be accommodated by brittle faulting and cataclasis. Although we cannot discriminate between these two scenarios, we find the latter more comprehensive.

DISCUSSION

Hornby Channel and Schist Lakes belts on the northwest side of the Schist-Daisy segment of Great Slave Lake shear zone strikingly resemble the similarly located, texturally, metamorphically and lithologically identical Hornby Channel ultramylonite belt and belt 2 of the Laloche segment to the southwest. We propose that they are indeed parts of the same belts and that the northwestern side of the Schist-Daisy segment is a continuation of the Laloche segment. The break in material continuity represented by the Snowdrift segment is due to a combination of (i) vertical(?) movement on the Austin fault bringing a volume of complex older mylonites, analogous to those of belt 1 of the Laloche segment, into the map plane and (ii) the emplacement of post-high grade mylonite Snowdrift granite into the shear zone.

After removal of the effect of 70 km right-slip on the McDonald-Wilson fault, the 010° trending Thelon Tectonic Zone (Thompson and Henderson, 1983) lies due north of our map area (Henderson and Macfie, 1985). Thompson et al. (1985) and Henderson and MacFie (1985) interpreted the regional aeromagnetic field to indicate that the Thelon Tectonic Zone extended south of the MacDonald-Wilson fault, although they were unclear as to its location. The spatial coincidence between eastward passage from amphibolite to granulite facies and from strike-lineated to dip-lineated tectonites is strikingly similar to descriptions of a west-east transect through the western part of the Thelon Tectonic Zone north of Sifton Lake (James, 1986; Henderson et al., 1987). Furthermore, the southeastern sub-belt of the Daisy Lakes belt lithologically resembles James' (1986) descriptions of the dip-lineated granulite facies Darrell River gneiss and pyroxene bearing granitoid 15 km east of the western limit of the Thelon Tectonic Zone. Given the progressive northeastward change of regional strike from 060° to 030° in our map area, we propose that the Schist-Daisy segment of Great Slave Lake shear zone and the southern Thelon Tectonic Zone are coextensive elements of the same tectonic structure. It follows from the previous paragraph that we extend this proposition to include the entire mapped length of Great Slave Lake shear zone. We shall refer to this tectonic structure as the Great Slave — Thelon shear zone. Preliminary radiometric dating of syn-tectonic plutonism in the Laloche segment (U/Pb in zircon, S.A. Bowring in Hanmer and Connelly, 1986, and pers. comm., 1987) yields ages in the range 2.03 — 1.94 Ga. Similar studies within the Thelon Tectonic Zone yielded ages in the range 1.96 - 1.91 Ga (van Breemen et al., 1987a, b). Together, these radiometric data suggest that the Great Slave — Thelon shear zone is early Proterozoic in age (see also Henderson et al., 1987).

In a recently published model of early Proterozoic tectonics of the Great Slave Lake shear zone Thelon Tectonic Zone region, Hoffman (1987) proposed oblique convergence between Slave Craton and the Rae Province (northwesternmost Churchill Province) which generated dextral strike slip shear within the plutonic roots of a continental magmatic arc, leading to collision and initiation of westward thrusting at circa 1.95 Ga, followed by continued indentation by Slave Craton. The relative timing, kinematics and crosscutting relationships of the Schist Lakes, Daisy Lakes and Hornby Channel belts all provide support for such a model. (i) An early, relatively wide belt of upper amphibolite facies, dextral, transcurrent mylonites deforms magnetite-bearing granitoid of the Laloche batholith (oblique convergence?). (ii) Further granitoid is added to the batholith. (iii) The locus of deformation jumps to the southeast and the kinematic framework rotates through 90 degrees such that shearing is dip-slip. Upper amphibolite facies dip-lineated mylonites are flanked to the southeast by granulite facies dip-lineated mylonites (collisional thrusting?). (iv) The high grade mylonites are bent through 50 degrees to wrap around the southeast corner of Slave Craton and minor high grade transcurrent shear is superposed upon the already formed mylonites (initial indentation of Rae Province by Slave Craton?). (v) The locus of deformation jumps to the northwest. The zone of active shearing narrows with decrease in metamorphic temperature (Hanmer, in press(a)), down to lower greenschist facies. The greenschist mylonites maintain the Great Slave Lake shear zone trend, a great circle extending from the Rockies to Artillery Lake (Hoffman, 1987), and cut through the older mylonites (continued indentation, uplift and erosion?). (vi) The Great Slave Supergroup was deposited on the excavated Great Slave - Thelon shear zone.

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Structural character and history of the Ashuanipi complex in the Schefferville area, Quebec-Labrador

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Abstract

In the Schefferville area, the Ashuanipi complex of the Superior Province consists of an older sequence of paragneiss interlayered with orthopyroxene-oikocrystic tonalite and minor pyroxenite-peridotite and gabbro sills; and a younger, voluminous suite of orthopyroxene-bearing diatexite plutons with minor tonalite, granite and syenite. Deformation and migmatization (S_1) of the older sequence preceded the intrusion of diatexites, which are involved in open F_2 folds that determine the large-scale map pattern. Doubly-plunging F_2 synforms form prominent structural basins, characterized by interior zones of relatively massive, homogeneous early tonalite, and flanks of migmatitic rock. Granulite facies metamorphism accompanied D_1, persisted through diatexite crystallization, and continued until crystallization of syn-D_2 orthopyroxene-bearing pegmatites.

Pyroxenite-peridotite sills, up to 80 m thick where well preserved, form variably continuous, boudinaged and attenuated structural markers. Sulphide-, graphite-bearing gossans in pyroxenite are barren, but a sample of basal peridotite with disseminated sulphides yielded a 70 ppb Pt analysis.

Résumé

Dans la région de Schefferville, le complexe d'Ashuanipi de la province du lac Supérieur est composé d'une séquence ancienne de paragneiss, interlité avec de la tonalite à orthopyroxène oikocrystique et des rares filons-couches de pyroxénite et de péridotite et gabbro; et d'une suite volumineuse de plutons plus récents de diatexite à orthopyroxène avec une petite quantité de tonalite, de granite et de syénite. La déformation et la migmatisation (S_1) de la séquence ancienne ont précédé l'intrusion des diatexites, impliquées dans les plis ouverts (F_2) qui déterminent la configuration de la carte. Les synformes (F_2) à double plongement créent des bassins structuraux remarquables qui sont caractérisés par des zones intérieures de tonalite ancienne homogène, relativement massives, et des flancs de roches migmatitiques. Le métamorphisme a accompagné la déformation D_1, a persisté jusqu'à la cristallisation de la diatexite et des pegmatites à orthopyroxène syn-D_2.

Les filons-couches de pyroxénite et péridotite, qui atteignent une épaisseur de 80 m, constituent des marqueurs structuraux, plus ou moins continus, boudinés et atténués. Les chapeaux de fer à sulfures et graphite dans les pyroxénites sont stériles, mais un échantillon de la péridotite basale à sulfures disséminés a donné une teneur de 70 ppm en Pt.

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INTRODUCTION

Reconnaissance mapping of the Archean Ashuanipi granulite complex in the Schefferville area (NTS 23J: Percival, 1987) identified a belt of complexly-deformed paragneiss and orthogneiss that was also found to contain elevated gold levels (Thomas and Butler, 1987) (Fig. 1). As the existing level of detail was inadequate to define either the structural history or gold association, the second field season focussed on resolving detail within the gneissic belt, flanked by mainly homogeneous plutonic rocks. Detailed mapping of similar terrane in the adjacent region to the north (Lapointe, 1986) located gold showings with values up to 18.9 g/tonne, presently under active exploration by Vior-Mazarin Ltd.

In the Schefferville area, the Ashuanipi complex consists dominantly of high- grade gneissic and orthopyroxene-bearing granitoid rocks. An older sequence of paragneiss with interlayered tonalite forms an irregular, 25-75 km-wide northwest-striking belt that is dismembered by homogeneous plutonic rocks of variable composition (Percival, 1987). Dips are generally to the northeast although oval structural basins characterize parts of the gneissic belt. Migmatitic textures and the widespread occurrence of orthopyroxene suggest uniformly high-grade metamorphic conditions.

A pilot U-Pb geochronology study of the Ashuanipi complex in the Lac Clairambault area, 50 km west of the present area, indicated late Archean ages for the igneous and metamorphic events (Mortensen and Percival, in press). Zircon from foliated tonalite gave a date of 2691 $\pm$ 3 Ma and that from a diatexite (homogeneous orthopyroxene granodiorite), probably emplaced near the metamorphic peak, gave 2668 $\pm$ 5 Ma. Monazites, which record regional cooling through a temperature of about 600°C, gave dates of approximately 2650 Ma. Ages of the major igneous and metamorphic events in the Ashuanipi complex are very similar to those from the Quetico (Percival and Sullivan, 1986) and English River (Krogh et al., 1976) metasedimentary gneiss belts of the western Superior Province.

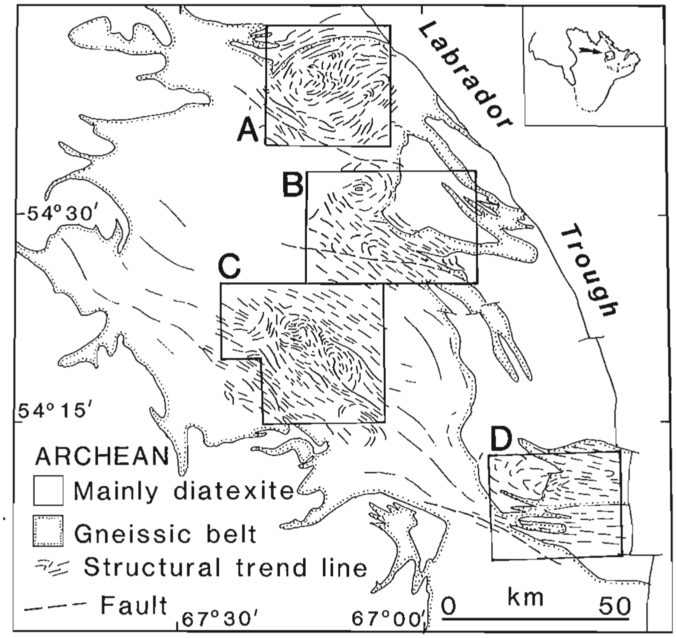


Figure 1. Position of gneissic rocks in the area mapped by Percival (1987). Generalized structural trends are shown for areas mapped in detail. Area A is presented in Figure 3, area C in Figure 5 and area D in Figure 6. Schefferville is located in the Labrador Trough, 7 km north of the map boundary.

LITHOLOGY

The lithological units defined in Percival (1987) were modified and refined as a result of more detailed work. In particular, compositional variation was recognized within paragneiss, early tonalite and pyroxenite. Close examination also revealed that some garnet-orthopyroxene-biotite metatexites (terminology of Brown, 1973), previously considered to be paragneiss, are actually garnet-bearing tonalite. Criteria were established for distinguishing these two units with similar migmatitic structure and identical mineral assemblage.

Paragneiss (garnet-orthopyroxene-biotite metatexite)

This unit generally consists of migmatitic rocks, with 50-95 % foliated mesosome containing biotite, garnet and orthopyroxene, interlayered on a 1 to 5 cm scale with 5-50 % orthopyroxene, biotite or garnet-bearing leucosome of granodioritic to tonalitic composition. Although generally psammitic in bulk composition, one outcrop of sillimanite, cordierite-bearing pelite was recognized and thin (5-25 m) leptynitic layers extend for several kilometres along strike locally. Bedding on the 10-20 cm scale was reported by Percival (1987). Psammitic paragneiss is distinguishable from migmatitic, garnet-bearing tonalite only where characteristic, meta-igneous pyroxene is preserved in the latter (see below).

Early tonalite (Orthopyroxene-biotite metatexite)

This unit varies texturally from homogeneous to gneissic both at outcrop scale and regionally, making description as gneiss or metatexite inappropriate; the present nomenclature accomodates textural varieties developed from a common protolith, at the same time distinguishing early and later tonalites recognized in 1986. Early tonalite occurs as concordant bodies in paragneiss on the metre to 10 km scale and underlies large parts of the gneiss belt. In the least deformed state, early tonalite is a massive green-grey oikocrystic rock with 10-30 % conspicuous, diffuse, inclusion-ridden, spherical to ovoid orthopyroxene crystals up to 1 cm, set in a fine to medium grained matrix of quartz, plagioclase, orthopyroxene, and rare K-feldspar and biotite. Compositional varieties include local diorite, with oikocrysts of both orthopyroxene and clinopyroxene, to rare gabbro, and local leuco- tonalite. Textural varieties include discontinuous 1-10 m layers of plagioclase- oikocrystic tonalite containing up to 25 % euhedral- to subhedral oikocrysts (Fig. 2). Igneous cumulate layering, defined by rhythmic variations in mafic mineral content, are preserved locally. Oikocrysts rarely grow across compositional layer boundaries.

With deformation, orthopyroxene crysts become flattened and gradually destroyed, leaving a foliated, medium grained orthopyroxene-biotite rock. White leucosome of quartz-plagioclase ± garnet in proportions up to 20 % is common in such deformed zones, which vary in scale from discrete zones within single outcrops to kilometre-sized regions. In irregular patches in some areas, up to 20 % garnet occurs as 1-2 cm clusters of anhedral grains or as quartz-filled poikiloblasts, surrounded by leucocratic depletion haloes (Fig. 2). Rare blocky orthopyroxene to 5 mm, surrounded by leucocratic haloes, occurs in relatively massive, garnet-bearing tonalite which also contains relicts of oikocrystic orthopyroxene (Fig. 2). This texture is interpreted to represent partly preserved igneous orthopyroxene, overgrown by metamorphic garnet and orthopyroxene. The orthopyroxene oikocrystic texture, highly resistant to deformation and metamorphism, is recognizable even where present in gneissic inclusions in diatexite.

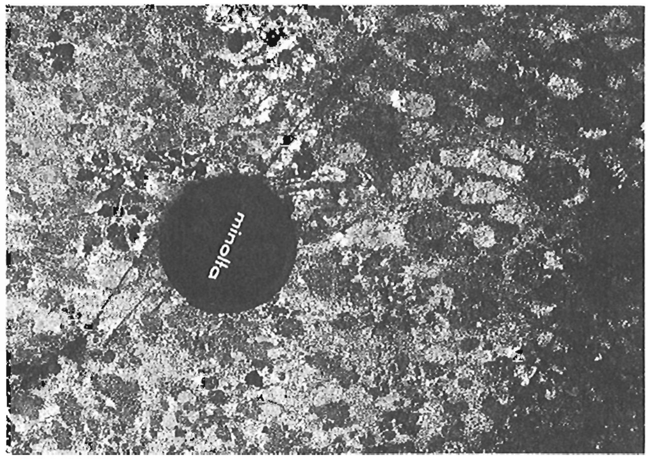


Figure 2. Outcrop photograph showing variable texture of early tonalite. Upper right side of photograph has prominent plagioclase oikocrysts which grade toward the bottom left into diffuse orthopyroxene oikocrysts (medium grey). Blocky metamorphic orthopyroxene (black spots) and garnet clusters (dark grey) are surrounded by mafic depletion haloes.

Pyroxenite-peridotite

Ultramafic rocks occur within the gneissic sequence as boudined lenses and layers ranging in size from sills up to 80 m thick and several kilometres long, to equant 500 m pods, to 3 m-thick, discontinuous layers. Generally pyroxenite in composition, the rocks vary in texture and mineralogy, from weakly foliated, with preserved igneous minerals, textures and structures, to strongly foliated, ultramafic schists.

Ultramafic bodies are most common and best preserved in the ''Pyroxenite'' Lake area (Fig. 1, 3, 4), where three sub-parallel units extend discontinuously over 40 km of strike length. The three bodies have distinct lithological character that is variably obscured by inhomogeneous deformation in and around them. A detailed 80 m section was measured across an exposure of the largest and best preserved sill (Fig. 4). Of bulk pyroxenitic composition, the sill varies from pyroxenite at the structural base of the exposed sequence, through 0.1-3 m interlayered peridotite-pyroxenite, to pyroxenite at the top. Discontinuous, cm-scale rhythmic layering occurs in layers 0.5 m thick. Cumulate textures with intercumulus hornblende are present locally. Minor amounts of plagioclase occur in thin layers, as do sulphides (pyrite, pyrrhotite), graphite and magnetite. Gabbro is present along strike between boudins, where it is generally strongly foliated or gneissic.

A second thick but discontinuous pyroxenite unit in the same region consists of a basal unit of peridotite, containing disseminated pyrrhotite, arsenopyrite and magnetite seams, a main unit of pyroxenite, or thin gabbro horizon and an upper pyroxenite unit. The complete sequence is preserved in the core of a southwest-plunging synform in the northeastern corner of the ''pyroxenite'' lake area, however the limbs are strongly attenuated. It is possible that the first and second sills are the same unit, either deformed (folded, attenuated or thickened) to variable extent, or with compositional/depositional variation along strike.

A third pyroxenitic unit, approximately 3-5 m thick, consists of homogeneous, medium to coarse grained two pyroxene-hornblende-biotite ± plagioclase assemblages. It is commonly foliated and discontinuous in the strike direction. Together the pyroxenites form semi-continuous structural markers in the ''Pyroxenite'' Lake area.

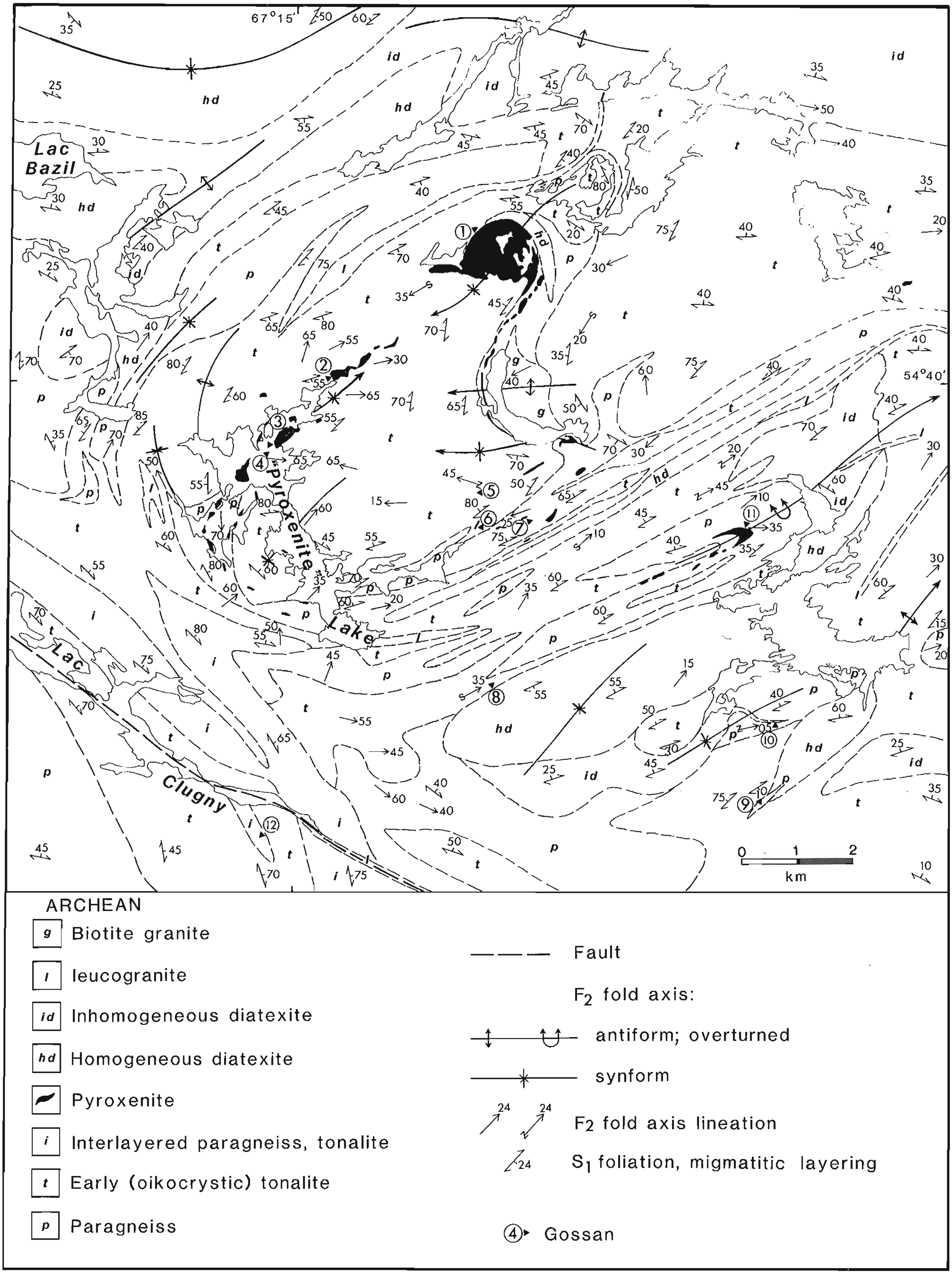
Other varieties of pyroxenite are present in smaller volume elsewhere in the region. In area B (Fig. 1), pods of massive, coarse grained pyroxenite up to 500 m in diameter have cumulate orthopyroxene oikocrysts to 2 cm surrounded by intercumulus hornblende. In the same area, pyroxenite layers up to 10 m thick grade outward to gabbro, diorite, and to the host oikocrystic tonalite. In area D (Fig. 6), minor pyroxenite occurs as layers or pods within a laterally extensive gabbroic unit. Many more small isolated occurrences occur; attempts to trace these were variably successful owing to a combination of incomplete exposure, structural disruption, and intrusive variability.

Gabbroic rocks

Rocks of mafic composition are rare. In addition to the minor amounts associated with pyroxenite and tonalite, several semi-continuous layers are present in the McPhadyen River area (Fig. 6). The most extensive is a 3 to 300 m thick body at least 40 km long, identified as mafic gneiss by Percival (1987). Detailed work showed the gneissic, migmatitic parts to be subordinate to a homogeneous, coarse to medium grained, 2-pyroxene-hornblende-plagioclase metagabbro. Small sulphide-rich zones are ubiquitous and contain elevated gold contents (Thomas and Butler, 1987).

Quartz diorite, diorite, tonalite

Bodies of this composition are common north of Lac Desliens (Fig. 5), interlayered with migmatitic rocks. The rocks are


67°15'
54°40'
Lac Bazil
Pyroxenite Lake
Lac Clugny
0 1 2
km
ARCHEAN
g Biotite granite
l leucogranite
id Inhomogeneous diatexite
hd Homogeneous diatexite
Pyroxenite
i Interlayered paragneiss, tonalite
t Early (oikocrystic) tonalite
p Paragneiss
Fault
F2 fold axis:
antiform; overturned
synform
F2 fold axis lineation
S1 foliation, migmatitic layering
Gossan

generally homogeneous, medium grained and slightly foliated, although massive and migmatitic varieties are present. They consist of assemblages of 2 pyroxenes, biotite and plagioclase, with 0-10 % quartz.

Diatexite

This rock-type forms most of the Ashuanipi complex in the Schefferville area (Percival, 1987); rocks of similar composition and textures are also prominent to the west (Eade, 1966). Diatexite (Brown, 1973; Percival, 1987) is coarse grained to megacrystic granodiorite to granite, named for its association with and similarity to migmatitic paragneiss. Diatexite occurs in homogeneous batholiths as well as in concordant layers within gneiss. It is homogeneous to inhomogeneous (more than 25 % inclusions) with a massive to foliated matrix. Two mineralogical varieties of texturally-similar diatexite were recognized: 1) a more common garnet-orthopyroxene-biotite granodiorite, with large blocky, white, antiperthitic feldspar; and 2) rare orthopyroxene-biotite granite with large, white, perthitic feldspar. Plutons of orthopyroxene-biotite diatexite are generally isolated from surrounding gneiss by an envelope of garnet-bearing diatexite.

Biotite granite

Pink, medium grained, homogeneous biotite granite forms small plutons locally. It varies texturally from massive to foliated.

Leucogranite, pegmatite

Leucogranite is ubiquitous throughout the region, in pegmatite dykes and irregular masses on the cm to km scale. It is generally coarse grained to pegmatitic, and composed of quartz, white feldspar, and minor amounts of garnet, orthopyroxene and biotite. Rare leucocratic syenite occurs as a thin sill in the McPhadyen River area (Fig. 6).

STRUCTURE

Structural elements

Primary structures (S_0) are limited to one example of bedding in paragneiss (Percival, 1987, Fig. 1.3) and locally preserved igneous layering in pyroxenite and tonalite. Structural facing directions could nowhere be determined.

The most prominent structural element is an S_1 foliation, defined by biotite alignment in gneiss, axes of orthopyroxene ovoids in early tonalite and elongate olivine and pyroxene grains in ultramafic rocks. Gneisses contain a parallel S_1 migmatitic layering on the 2-15 mm scale. On the regional scale, S_1 strikes west-northwest and dips moderately to steeply, generally to the northeast.

Figure 3. Geology of the "Pyroxenite" lake area (area A, Fig. 1) including generalized structural data. Analyzed gossans numbered 1 through 12 have the following values in ppb for Au and Pt respectively: (nd = not detected; detection limit = 1 ppb for Au, 10 ppb for Pt); 1:5, 70; 2:1-10, nd; 3:nd, 10; 4:14, nd; 5:17, nd; 6:130, nd; 7:2, 10; 8:2, nd; 9:1-61, nd; 10:nd, nd; 11:13, nd; 12:nd, 10. Measured section through pyroxenite sill in Figure 4 is located south of gossan 2.

D_2 structures are generally mesoscopic to map-scale folds of the S_1 surface, without an associated axial-planar foliation. The orientation and style of large-scale D_2 folds vary regionally and are discussed individually below.

Reversals in plunge direction of D_2 folds may be related to a set of open F_3 folds with no other manifestation, or alternatively, to porpoising of F_2 axes.

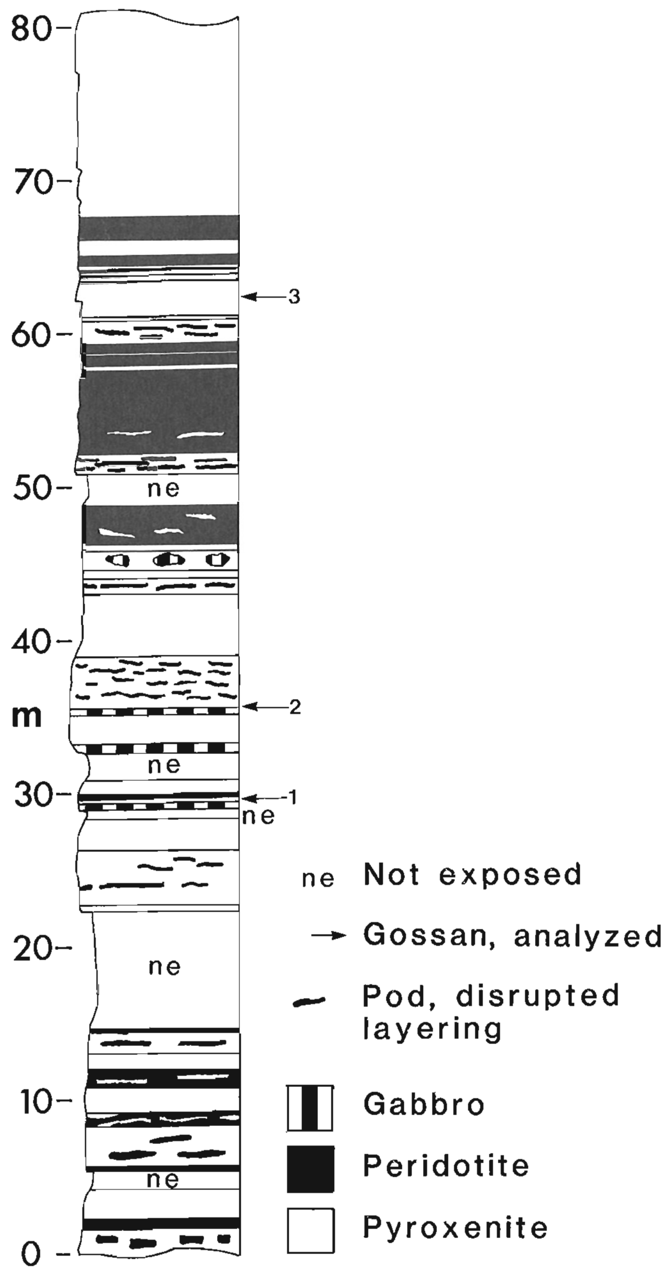


Figure 4. Measured section through layered pyroxenite-peridotite sill at "Pyroxenite" lake, converted to true thickness from layering dipping 30°N. See Figure 3 for location of section. Analyzed gossans labelled 1 through 3 have the following values in ppb for Au, Pt: 1:nd, nd; 2:nd, 10; 3:nd, 10.

A set of late fractures and cataclasite zones present in all rock types is concentrated in prominent west-northwest valleys. Fractures are generally steep with northwest strike and are locally folded, with steeply-plunging axes. The fault-related structures are generally hosted by zones containing abundant leucogranite, which show signs of an earlier, ductile, high-strain event.

Structural character

Large-scale D_2 structures define the map-pattern of units in the central gneissic belt. Four areas are described to illustrate the variation of structural style.

Area A

In the "Pyroxenite" lake area ("A", Fig. 1; Fig. 3), a major open, upright east-plunging F_2 synform makes up the large-scale map pattern. Tight asymmetric folds on the limbs generally have steep to moderate easterly plunges. Reversals in F_2 plunge direction produce dome and basin-style interference structures, the most prominent of which is the 15 km wide basin in the "Pyroxenite" lake area.

Pyroxenite sills form a semi-continuous rim around the basin. One sill can be traced continuously for approximately 8 km on the western edge of the basin. In the northwest corner, the bodies are well-preserved but boudined, with individual

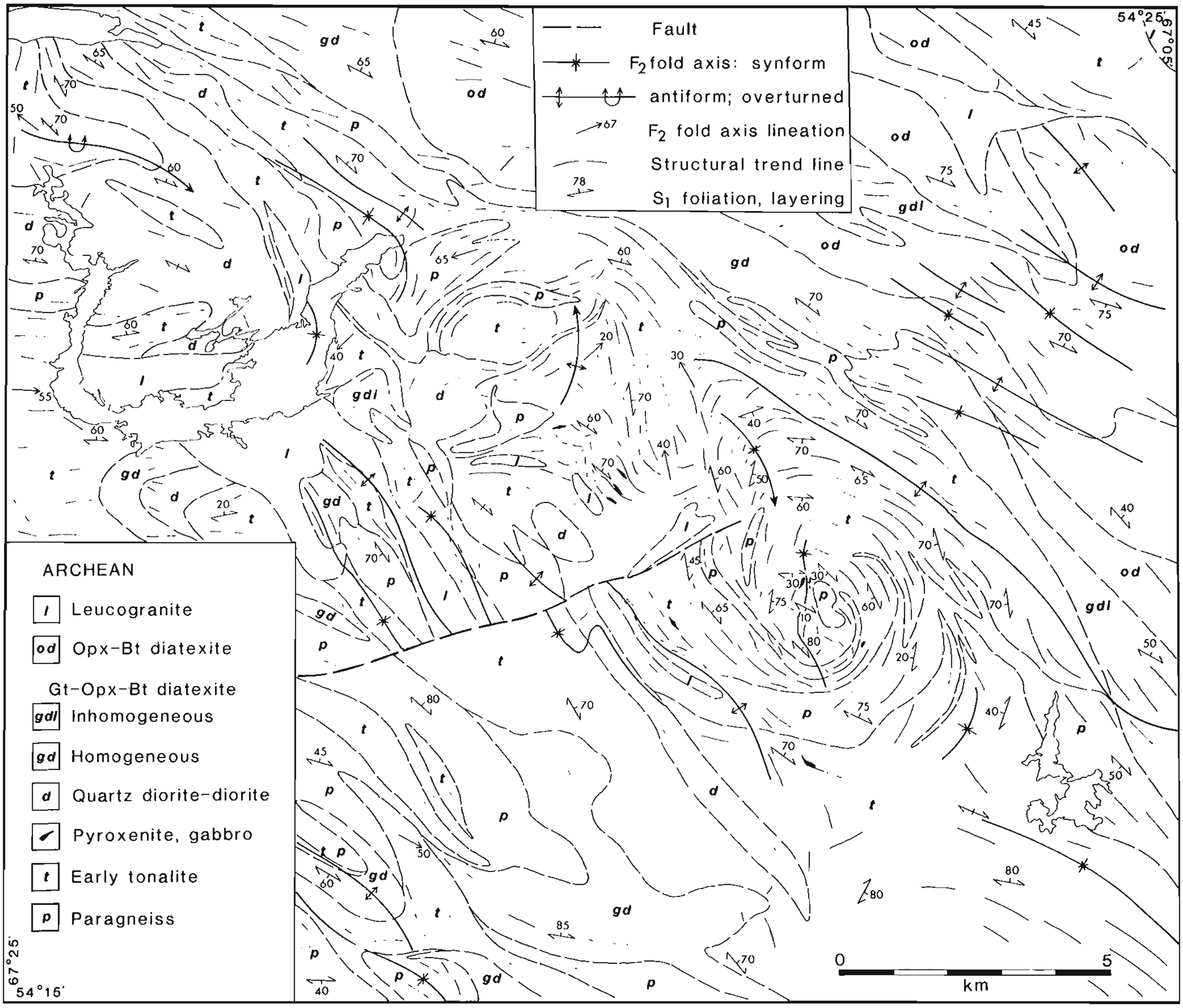


Figure 5. Geology of the area northeast of Lac Desliens (area C, Fig. 1) including generalized structural data. The Lac Desliens basin is located in the southeastern part of the map.

segments rarely more than 500 m long. In the southwest corner, the unit is very thin or absent. Sills are thickest and least internally deformed in the northwest and northeast corners; elsewhere, pyroxenite is schistose and as little as 3 m thick. Thick boudin segments are separated by structurally chaotic zones containing tightly folded country rock gneiss, metagabbro and abundant pegmatite.

Tonalitic rocks vary greatly in their degree of deformation, recrystallization and migmatization. Within the core of the basin, orthopyroxene oikocrysts have little preferred orientation, in contrast to tonalite on the flanks of the structure, where oikocrysts are only locally preserved in layers within strongly foliated, biotite-rich, migmatitic, tonalitic gneiss. Small-scale asymmetric folds of gneissic layering appear to be parasitic with respect to map-scale F_2 folds.

Diatexite occurs as conformable layers within gneiss, as well as in larger bodies to the north. It generally contains aligned gneissic inclusions which define a foliation in an otherwise massive rock. Where involved in tight F_2 folds, a foliation in diatexite matrix is locally defined by aligned biotite. Because diatexite contains gneissic inclusions with S_1 layering and is involved in F_2 structures, its emplacement is bracketed between D_1 and D_2 events.

A small arcuate pod of biotite granite occurs in the nose of an F_2 fold on the eastern side of the basin. It locally contains an east-west foliation, axial planar to the F_2 fold, suggesting syn-tectonic emplacement with respect to D_2.

Ubiquitous, leucogranitic, pegmatite dykes cut S_1 migmatitic layering at a low angle and are commonly folded by small-scale, open D_2 folds. Emplacement is thus pre- to syn-tectonic with respect to D_2. Pegmatite is particularly abundant in zones of strongly foliated, migmatitic, contorted gneiss, suggesting late-tectonic injection along shear zones.

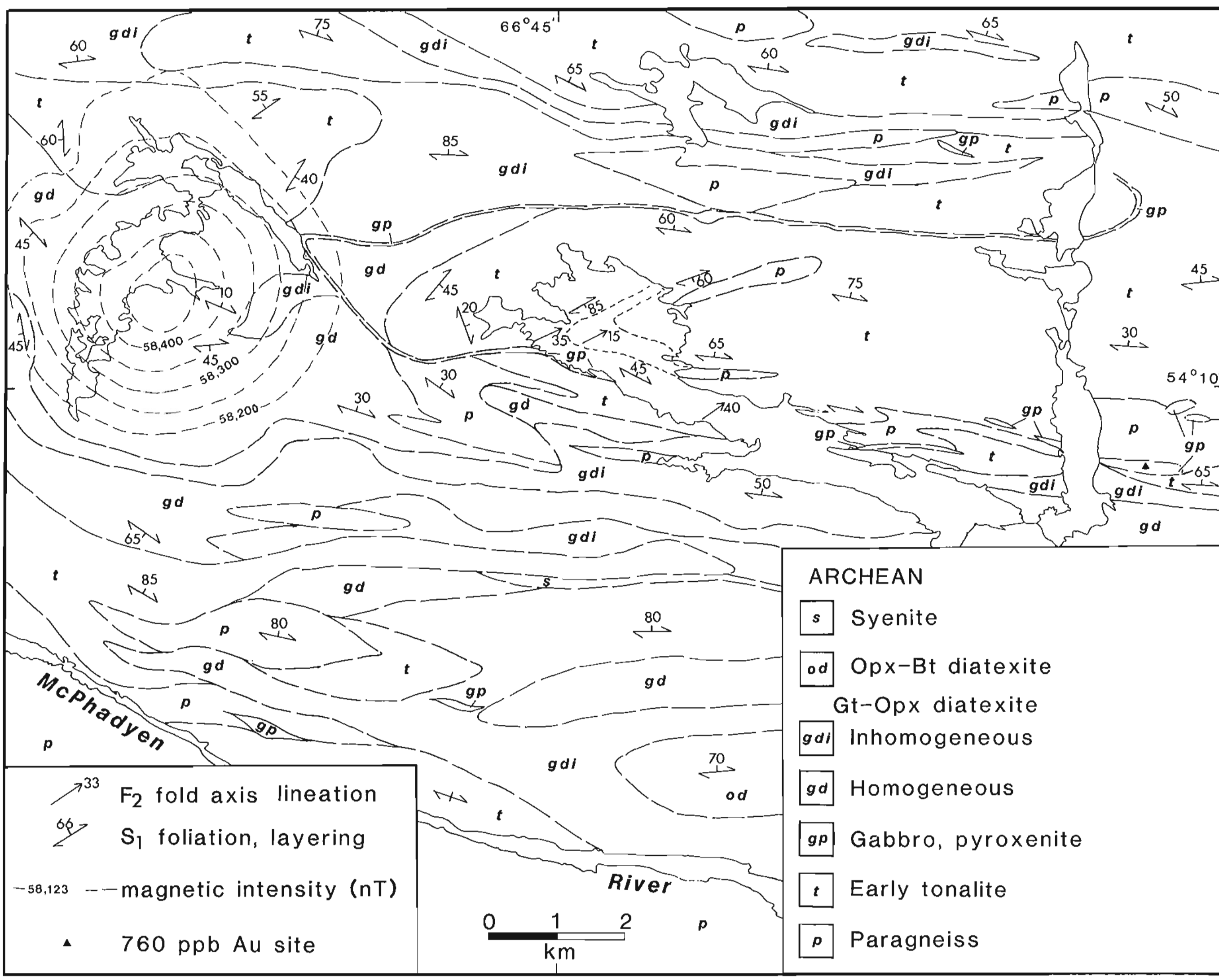


Figure 6. Geology of the area north of McPhadyen River (area D, Fig. 1) including generalized structural data. A thin discontinuous belt of gabbroic and pyroxenitic rocks outlines the limbs of a northeast-plunging fold. Menihek Lake is located 4 km to the east of the map.

Area B

Several map-scale folds are present in the dominantly northeast-facing homocline of interlayered paragneiss and early tonalite of area B (Fig. 1). A major southeast-plunging, isoclinal synform is overturned to the southeast. Smaller scale folds to the east have similar orientation. Complete folds are rarely preserved; rather, isolated closures are separated by long limbs.

S_1 foliation in the northern part of area B defines a small basinal structure which appears to be a canoe-shaped F_2 synform with a northeast-striking long axis. The lithology is asymmetric with respect to the fold axis, suggesting pre-F_2 heterogeneity.

Area B contains lithological variety within both paragneiss and early tonalite units. Quartz-rich biotite-garnet-orthopyroxene leptynite occurs in layers to 10 m thick, locally useful as stratigraphic markers. Units resembling diatexite occur as concordant, 1-10 m layers within paragneiss. These rocks have a fine grained matrix in which biotite defines the foliation and coarse grained garnet, orthopyroxene and feldspar. Most early tonalite has a colour index in the range 20-30. Layers of orthopyroxene-oikocrystic leucotonalite, with colour index of 10-15, are rare, as are dioritic to gabbroic layers, with orthopyroxene and clinopyroxene oikocrysts amounting to 30-50 % of the rock. Pyroxenite layers, up to 10 m thick, are locally associated with mafic and intermediate compositions. Pyroxenite also occurs as isolated pods up to 500 m in diameter, separated by structurally chaotic gneiss.

Late structures are evident in major northwest-trending valleys. Brittle offsets, fractures, cataclasite, fault gouge and breccia indicate proximity to faults. Fractures are locally folded into steeply-plunging, N-S, small-scale asymmetric folds. Common fracture orientations are 290° and 025-045°. Leucogranite sheets, on the 1-5 m scale, are commonly cut by the zones of brittle deformation. Individual lithological units cannot be traced across major valleys; however there appears to be no offset of structural trends.

Area C (Lac Desliens)

The overall northeast-facing homocline of area B continues southeast into area C, where it is modified by several F_2 folds, resulting in a complex map-pattern. Several weakly-deformed, pre-tectonic bodies of tonalite-quartz diorite have possibly concentrated strain by acting as relatively rigid masses.

At the map scale (Fig. 1, 5), the S_1 foliation forms a broad, open "Z" warp, from WNW in the northwest, through NW in the centre, to WNW in the southeast. Similar warps are present at a scale of 1-2 km. Reversals in dip direction of S_1 foliation within the generally homoclinal panel can be related to open, upright, northwest-trending folds with horizontal axes. The folds are generally discontinuous, extending for 2-15 km, and occur in the northwest-trending portion of the broad "Z" warp. In cross-section the folds are open "S" folds, with long, northeast-dipping limbs and short, southwest-dipping limbs.

A large, doubly-plunging F_2 synform forms the NW-SE axis of the oval Lac Desliens basin (Nagerl, 1987). The core of the basin consists mainly of oikocrystic tonalite whereas the margins are interlayered paragneiss and migmatitic tonalite. The corresponding antiform to the northeast terminates at its northwestern end in a structurally complex pattern (Fig. 5). Plunge variation on the F_2 axis from 45°SE to 45°NW defines the basin. Although this could be due to a superposed, northeast- trending F_3 synform, there appear to be no analogous antiform-antiform domal culminations of the same scale. Alternatively, plunge variation could be an inherent D_2 feature, or an effect of interference with an earlier, possibly igneous structure (Nagerl, 1987). An antiformal closure south of the Lac Desliens basin appears to be an isolated nose of a map-scale, ESE-plunging "Z" fold.

Northwest of the basin, a complex structural pattern developed in tonalitic rocks has the appearance of a rootless antiformal fold on a 3 km scale (Fig. 5). The significance of this structure of overall "S" geometry, which contrasts with the broad regional "Z" pattern, is not understood at present.

Structural patterns are complex in the vicinity of bodies of massive to weakly foliated and locally gneissic quartz diorite to diorite. Based on its variably deformed nature, this unit is pre-tectonic, however massive portions suggest a rigid character during deformation. Small bodies may be boudined segments of a larger pluton, separated by gneisses with complex interference patterns. A similar explanation may apply to the Lac Desliens basin, whose relatively homogeneous lithology may have focussed strain in the adjacent rocks.

The location of the discontinuous F_2 folds in the bend of the large-scale "Z" warp suggests a possible genetic relationship. Both could be related to a dextral, northeast-side-up shear component.

Area D (MacPhadyen River north)

Mafic rocks in area D (Fig. 6) have attracted attention as a result of their anomalous gold content (Percival, 1987; Thomas and Butler, 1987). The detailed work in this area was aimed at defining the character and distribution of mafic units within the regional context and resulted in several modifications to the previous version of the map.

Early tonalite underlies much of the northern part of area D. It is generally migmatitic, with locally-preserved orthopyroxene oikocrysts and abundant garnet clusters, with mafic-depletion haloes. Mafic tonalite is common and is locally associated with dioritic and gabbroic layers. Minor paragneiss occurs as thin layers in tonalite. The southern half of area D is underlain by diatexite.

Units of gabbro-pyroxenite occur as widely distributed, variably continuous sills, generally in early tonalite. Mafic bodies are up to 150 m wide whereas pyroxenite is generally less than 20 m in width. Pyroxenite and gabbro are commonly associated and have similar textures, ranging from massive and coarse grained to foliated or layered and medium grained. Heterogeneities such as that previously illustrated (Percival, 1987; Fig. 1.5) appear to be developed from homogeneous rocks in local deformation zones.

Two east-striking, steeply-dipping belts of gabbro-pyroxenite may be related by a fold axis in the western part of area D. However, intrusive bodies disrupt the unit in the axis area. The southern belt is generally thicker, but less continuous than the northern. It is mainly metagabbro, with local pyroxenite. Complex structure in gneisses between gabbro lenses suggests large-scale boudinage. Sulphide-rich pods at the metre scale are particularly common in gabbro on the prominent ridge west of Menihek Lake (Fig. 6). Where best preserved the northern belt consists of 10 m of pyroxenite, flanked by 20-30 m of gabbro. Elsewhere the body thins to 10 m and grades along strike from pyroxenite to gabbro. The unit is associated with a 2-km-wide positive aeromagnetic anomaly.

Foliation and layering strike westerly, swinging to the northwest in the west. Dips are steep to moderate to the north. A large-scale, open, upright antiform characterizes the southern diatexite. A map-scale ''S'' fold between the mafic belts has a strong, gently northeast-plunging, axial rodding lineation.

A concentric, 5 km diameter magnetic anomaly in the western part of area D has no obvious lithological cause. It coincides with a crudely domal structure (Fig. 6).

CHRONOLOGY AND TECTONIC HISTORY

The oldest unit recognized is paragneiss. Early tonalites probably represent pre-metamorphic sheets of variable thickness. Mafic compositions (diorite to gabbro) are present within tonalite and are locally associated with pyroxenite, suggesting a possible cogenetic relationship between tonalite and pyroxenite. Pyroxenite also occurs within paragneiss, indicating the presence of discrete intrusions. The paragneiss-early tonalite-pyroxenite sequence was metamorphosed and deformed (D_1) to produce a foliation of variable intensity and prominent migmatitic layering (S_1), generally of quartz-feldspar $\pm$ garnet composition. Diatexite was emplaced as sheets concordant to layering, prior to formation of F_2 folds. Garnet-, orthopyroxene-bearing pegmatites were emplaced on the regional scale during D_2 deformation. High-grade metamorphism persisted from D_1 migmatization through crystallization of pegmatites late in the D_2 deformation. Reversals of plunge direction of large-scale F_2 synforms, forming prominent structural basins, may be the result of local F_3 warps, or syn-D_2 effects of lithological/mechanical heterogeneity. Basins in both areas A and C are characterized by relatively massive, homogeneous early tonalite which may have behaved competently during deformation with respect to the surrounding migmatitic rocks.

METAMORPHISM

Pre-metamorphic anhydrous minerals and textures are preserved in oikocrystic tonalite and in pyroxenite-peridotite. Assemblages include orthopyroxene-plagioclase-quartz in tonalite, and olivine-orthopyroxene- clinopyroxene-hornblende-spinel in ultramafic compositions. Relatively static metamorphism of tonalite produced clusters of garnet with mafic-depletion haloes and rare blocky metamorphic orthopyroxene, whereas combined with deformation, metamorphism yielded biotite-orthopyroxene $\pm$ garnet, migmatitic, tonalitic gneisses. Ultramafic rocks are recrystallized in zones of deformation and adjacent to granitic intrusions. Metamorphic minerals include hornblende, biotite, magnetite, talc and serpentine.

Metamorphic orthopyroxene occurs in migmatitic paragneiss and early tonalite throughout the region, in association with garnet, quartz, plagioclase, K-feldspar and biotite. Orthopyroxene and garnet in diatexite and pegmatitic leucogranite are probably of igneous origin, based on their coarse subhedral texture in coarse grained massive rocks. Crystallization temperatures for both gneisses and diatexites, based on coexisting orthopyroxene, K-feldspar and quartz, (Bohlen et al., 1983) are estimated at over 750°C. Pressure based on garnet-orthopyroxene- plagioclase-quartz barometry (Newton and Perkins, 1982) in the Lac Desliens area is on the order of 600 MPa (6 kbar) (Nagerl, 1987), similar to regional values of 500-650 MPa (5-6.5 kbar). The mineral assemblages, in conjunction with migmatitic textures, suggest granulite metamorphism and associated intrusion in a water-dominated fluid regime at mid-crustal levels.

ECONOMIC GEOLOGY

Several gossans were sampled for precious metal analysis, based on encouraging results obtained previously for gold and potential for platinum mineralization in ultramafic rocks. Some values are reported in Figures 3 and 4.

In the ''Pyroxenite'' lake area, gossans are widespread, both within and adjacent to pyroxenite and in paragneiss/early tonalite. Discontinuous gossans in pyroxenite contain pyrite, pyrrhotite and graphite and generally low amounts of gold (1-10 ppb) and platinum (below 10 ppb detection limit).

Mineralized zones at pyroxenite contacts also contain sulphides, modest Au anomalies (13-130 ppb) and less than 10 ppb Pt. The basal peridotite unit from the largest, layered, peridotite-pyroxenite-gabbro sill contains disseminated pyrrhotite and arsenopyrite and 70 ppb Pt.

A 30 m wide gossan in interlayered paragneiss and early tonalite contains massive sulphide and some graphite-rich zones. Three grab samples yielded Au assays of 1-60 ppb and detailed work by McConnell et al. (1987) indicated similar levels. Mafic rocks in area D contain up to 760 ppb Au (Thomas and Butler, 1987).

SUMMARY

The Ashuanipi compelx consists mainly of felsic, orthopyroxene-bearing gneissic and intrusive rocks (Eade, 1966; Card and Ciesielski, 1986). In the Schefferville area, a NW-striking, NE-facing belt of paragneiss and early tonalite, flanked by intrusive diatexite, contains semi-continuous pyroxenite-peridotite sills that outline regional structural patterns. Major deformation and granulite facies metamorphism accompanied production of an S_1 foliation and migmatitic layering which, after intrusion of voluminous diatexite, was subsequently warped into open F_2 folds. F_2 structures form the map pattern of lithological units, including several prominent structural basins which are doubly-plunging F_2 synforms. The occurrence in the basins of relatively massive, homogeneous early tonalite suggests that the distribution of large-scale F_2 features may be controlled by lithological and hence mechanical heterogeneities.

Mineral potential is highest in the gneissic belt, in particular in mafic and ultramafic rocks. Metagabbro in the south has gossans with up to 760 ppb Au whereas to the north, layered pyroxenite-peridotite sills contain up to 70 ppb Pt. Gossans in the paragneiss-early tonalite suite contain pyrite, pyrrhotite, arsenopyrite and graphite, with up to 60 ppb Au.

ACKNOWLEDGMENTS

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The Quaternary stratigraphy of the Timmins area, Ontario, as an aid to mineral exploration by drift prospecting[1]

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DiLabio, R.N.W., Miller, R.F., Mott, R.J., and Coker, W.B., The Quaternary stratigraphy of the Timmins area, Ontario, as an aid to mineral exploration by drift prospecting; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 61-65, 1988.

Abstract

From top to bottom, the Quaternary sequence in the Timmins area consists of: glaciolacustrine clays of the Barlow-Ojibway succession, containing the Cochrane Till; the Matheson Till; the fossiliferous nonglacial Owl Creek beds; and at least two older tills and associated stratified sediments. The Owl Creek beds discontinuously underlie an area of at least 2000 km^2 and form a distinctive stratigraphic marker. This unit may represent a complete Sangamon or Middle Wisconsinan nonglacial cycle. Ice-flow directions and provenances for tills under the Owl Creek beds are being investigated because these tills have the largest contact areas with the local bedrock and therefore are prime sampling media for drift prospecting in the areas of thicker drift.

Résumé

Du sommet à la base, la séquence quaternaire de la région de Timmins se compose: des argiles glaciolacustres de la succession de Barlow-Ojibway, qui contient le till de Cochrane; du till de Matheson; des couches fossilifères et non glaciaires d'Owl Creek; et d'au moins deux tills plus anciens et des sédiments stratifiés associés. Les couches d'Owl Creek sont présents de façon discontinue dans le sous-sol d'une région d'au moins 2000 km^2, et constituent un horizon stratigraphique repère facile à distinguer. Cette unité représente peut-être un cycle non glaciaire complet du Sangamon ou du Wisconsinien moyen. On étudie actuellement les directions d'écoulement des glaces et la provenance des tills au-dessous des couches d'Owl Creek, parce que ces tills présentent la plus vaste surface de contact avec le soubassement local, et constituent donc un milieu idéal d'échantillonnage pour les levés des dépôts glaciaires dans les zones où ces dépôts sont relativement épais.

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INTRODUCTION

Since the adoption of the reverse circulation drilling method for deep till sampling in mineral exploration, hundreds of borings have been made through the Quaternary sequence in the Timmins camp (Skinner, 1972; Averill, 1978). This exploration has been aimed at the discovery of gold and base metal deposits by analysis of the geochemistry and lithology of the glacial sediments, mainly till. Theoretically, it should be possible to trace metal-rich till back to its source, but in practice this has not been easy. With the publication of drill logs from an early research project (Skinner, 1972), it became obvious that a thick, complex Quaternary sequence was present in the whole of the Abitibi clay belt, and that it was at its most complex around Timmins.

The main problems related to this sequence are how to correlate discontinuous tills in the subsurface and what ice-flow direction to assign to each till. East of the Timmins area, a great deal of subsurface work has been done in the Black River-Matheson (BRIM) project (Steele et al., 1986). Ice-flow sequences for a large part of the clay belt have been established by Veillette (1986). Closer to Timmins, Bird and Coker (1987) have identified three units of glacial and proglacial sediment below the Cochrane till of Hughes (1965), but the regional correlations and associated ice-flow directions are unclear.

During and after Skinner's project, wood chips and organic-rich sediments were noted, but rarely sampled in many reverse circulation drillholes, particularly north and east of Timmins (Brereton and Elson, 1979; DiLabio, 1982). After a number of such sites had been reported, it became clear that a distinctive organic-rich unit directly underlay the main regional till (the Matheson Till of Hughes, 1965) over a large area. If this easily recognized stratigraphic marker could be fully defined, between tills in the middle of the sequence, it could be most useful to mineral exploration. Determination of its age, paleoecology, and rank as interstadial or interglacial would also be crucial to the problem of possible multiple Wisconsinan deglaciations of the Hudson Bay Lowlands since the last interglaciation (Andrews et al., 1983).

This study began in 1984 with the donation by Cominco Ltd. to the Geological Survey of Canada of two Rotasonic drill cores (SMO-65 and 66, Fig. 1) from Stock Township.

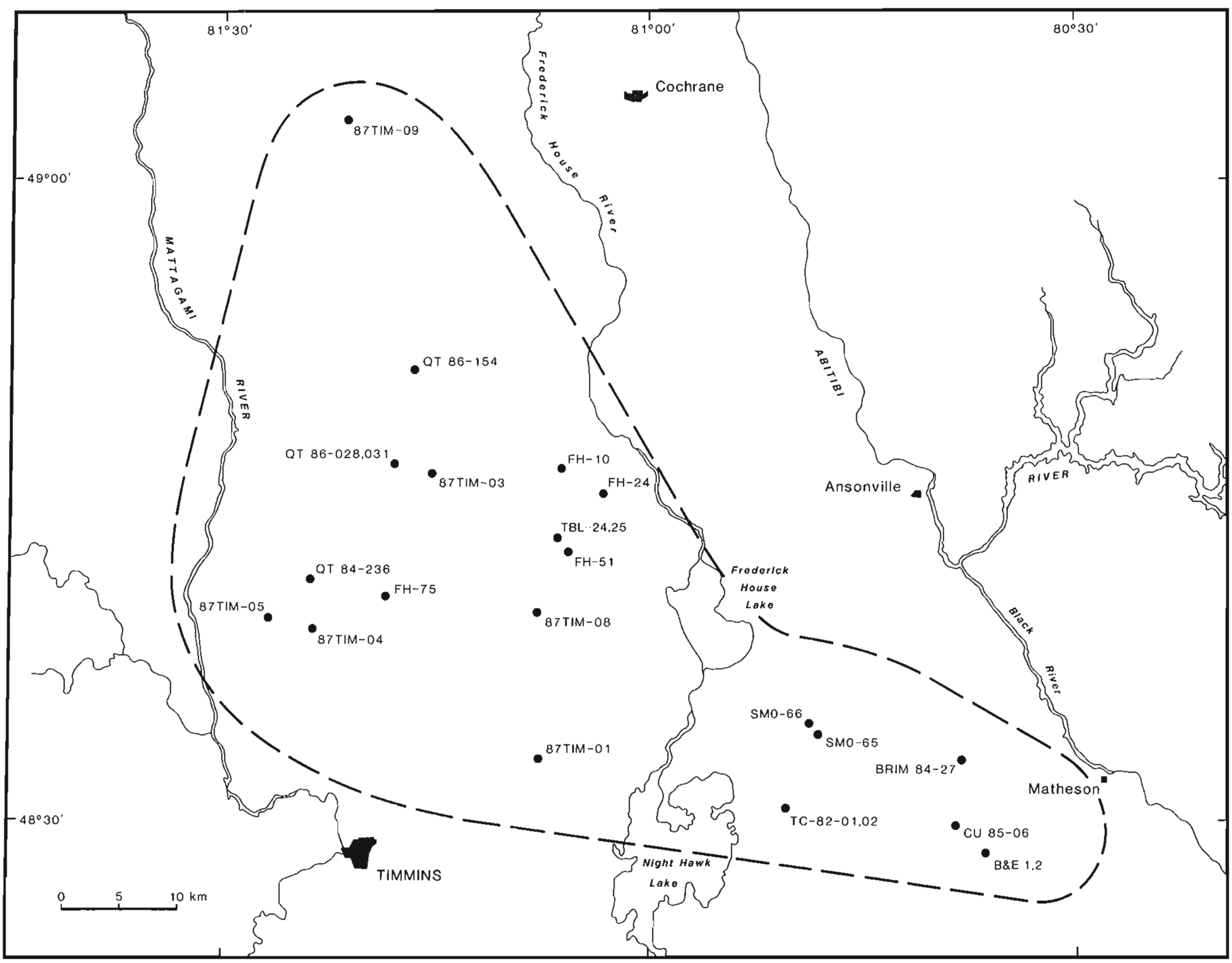

Figure 1. Location map of drillholes intersecting the Owl Creek beds. Dashed line marks the known limit of the area underlain by the unit.

These cores were 9 cm in diameter, intact, and covered only the organic-rich unit and short intervals of the units above and below it. In 1987, as part of the Canada-Ontario Mineral Development Agreement, several Rotasonic drillholes were spotted on the basis of reverse circulation drilling records to attempt to intersect multiple-till sections containing this unit. Ten cores to bedrock were recovered, five of which contained the organic-rich unit (87-TIM-03, 04, 05, 08, 09 on Fig. 1).

A variety of analyses are being performed on these cores, mainly to determine the provenance and correlations of the tills and to determine the paleoecology of the organic-bearing unit. At such an early stage of the project, we will confine this report to a description of the overall stratigraphy and preliminary results on the organic-bearing unit.

STRATIGRAPHY

The stratigraphy presented in Figure 2 is based on detailed logging of the 87-TIM cores, data from Falconbridge Ltd. and Cominco Ltd., and published data (Hughes, 1965; Skinner, 1972; Brereton and Elson, 1979; Steele et al., 1986). The upper unit of the sequence consists of varved silts and clays deposited in glacial lakes Barlow and Ojibway, and the associated clayey Cochrane Till (Hughes, 1965) which was derived from the clays during a readvance into the glacial lake. Below the clays is the Matheson Till (Hughes, 1965), a widespread pebbly silty sand till. Sands and gravel seen in contact with the Matheson Till proper have been included with it in the diagram.

Immediately underlying the Matheson Till are the Owl Creek beds, informally named after the Owl Creek gold mine in Hoyle Township (Fig. 2), where its only known exposure exists. This unit normally consists of overconsolidated organic-rich silts, clays and sands, and is an easily recognized marker unit that is often referred to as "superclay" (S.A. Averill, pers. comm., 1985). Areally, the Owl Creek beds underlie at least 2000 km^2 (Fig. 1). Because of the uneven distribution of exploration drilling, we do not know its limits to the north, northeast, and west. It is probably cut off against the rising bedrock surface to the south, and it is not known in BRIM drilling records east of Taylor and Currie Townships (Fig. 2). Drilling records supplied by Falconbridge Ltd. show that the Owl Creek beds are preserved in scattered troughs and are not present everywhere under the 2000 km^2 area outlined on Figure 1. This may partly reflect survival of the unit during the Matheson ice advance, but paleoecological evidence indicates that the sediments accumulated in a large number of small lakes or in a large lake composed of many bays, peninsulas, and islands, not in one large open lake.

Palynological results for one core of the Owl Creek beds (SMO-65, Fig. 1) indicate that an early shrub and herb dominated tundra or tundra woodland changed to a closed coniferous forest dominated by spruce and jack pine as the climate warmed to a thermal maximum similar to the present in the area. A strong late climatic cooling is not indicated by the palynological record. The composition of the beetle fauna in this core is similar to that observed in most late Quaternary assemblages observed in North America (Morgan and Morgan, 1980). The assemblage is typical of a wet environment, containing elements from a variety of habitats ranging from strictly aquatic to riparian, to drier terrestrial situations. In contrast to the pollen record, macrofossil and beetle data indicate that conditions were cooler than today at least near the top of the core. Only near the base of the core were indicators of warmer climate found, seeds of the plant *Cladium mariscoides*, which probably does not grow as far north as Timmins today. Several of the insect species that were found in core SMO-65 and in bulk samples from site TIM-09 (Fig. 1) have present distributions that are more northerly or higher altitude ones. Two of the species are now extinct in central North America: *Diacheila polita*, known as fossils in Early Wisconsinan deposits in the Scarborough Bluffs in Toronto and in the St.Pierre beds in southern Quebec (Morgan and Morgan, 1980); and *Holoboreaphilus nordenskioeldi*, known as fossil in Arctic Sangamon sites and in Middle Wisconsinan sediments at Titusville, Pennsylvania (Coope, 1970).

Clearly, the paleoenvironmental history of the Owl Creek beds requires much more work. It is expected that analysis of the other cores will build up a set of overlapping records that will show a full cycle of glacial-nonglacial-glacial environments. Because the two radiocarbon dates on the unit are infinite (>37 000 a, GSC-2148, Brereton and Elson, 1979 and >51 000 a, GSC-3875), it remains to be shown whether the Owl Creek beds represent a Middle Wisconsinan interstadial related to a nonglacial event in the Hudson Bay Lowland (Andrews et al., 1983) or whether they represent the Sangamon interglacial, correlative with the Missinaibi Formation of the Moose River basin (Skinner, 1973), although the status of the Missinaibi as Sangamon may be in doubt. Because of the proximity of Timmins to the Moose River basin, we suggest that the Owl Creek beds are correlative with the Missinaibi Formation. This correlation is also supported by the simplicity of the Matheson Till-Adam Till and Cochrane Till-Kipling Till correlations it requires with the Moose River basin, a suggestion put forward by Skinner (1973).

Below the Owl Creek beds is a sequence of at least two tills and associated stratified sediments (Fig. 2), the upper one a hard pebbly silty sand till and the lower one a compact sitly till. Very little is known about the provenance of these tills; laboratory work is under way to gather lithological data that may assist in mapping them in the subsurface and in assigning ice-flow directions to them.

ICE-FLOW DIRECTIONS

In a major study of cross-striated outcrops in the Abitibi clay belt, Veillette (1986) showed that a main set of young striae oriented towards 150-180° in the Timmins area was most likely related to the Matheson Till. A widespread earlier set oriented towards 240° is probably related to the first till below the Owl Creek beds (Bouchard et al., 1986; Steele et al., 1987; J.J. Veillette, pers. comm., 1987) because this set has been observed under that till at several sites in the clay belt. An intermediate set in the 180-220° range is probably an early Matheson flow direction. The question of ice-flow directions is one of the most important for drift prospecting in this region because the divergence of up to 90° between

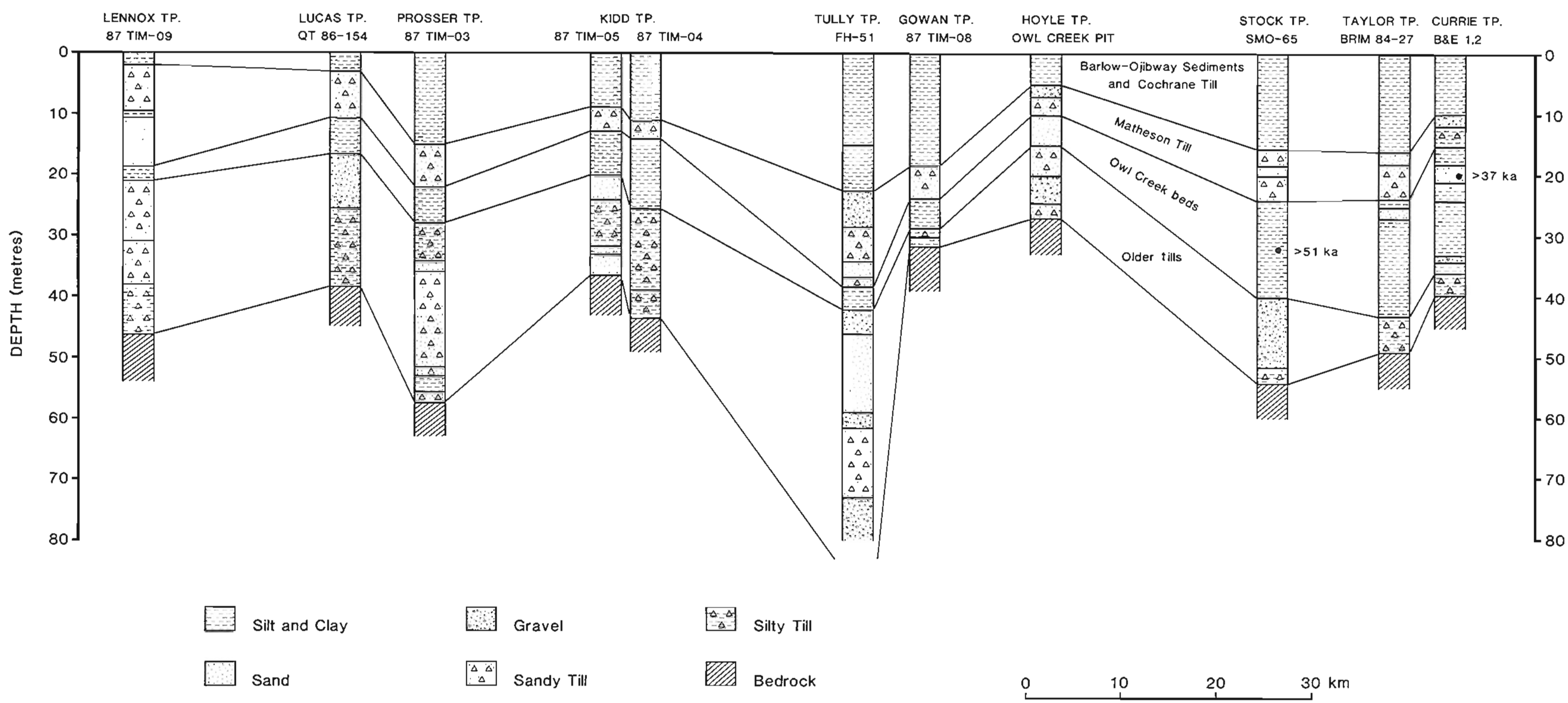

Figure 2. Schematic cross-section of Quaternary units across the area in Figure 1. Data for QT series holes and Owl Creek pit section (at 87 TIM-01 on Fig. 1) courtesy of Falconbridge Ltd. Data for FH series holes from Skinner (1972). Data for hole SMO-65 courtesy of Cominco Ltd. Data for hole BRIM 84-27 from Ontario Geological Survey (1986). Data for hole in Currie Township from Brereton and Elson (1979). Radiocarbon dates in thousands of years shown on logs of holes in Stock and Currie townships.

striae sets means that dispersal trains of ore minerals must be traced in the correct direction related to the till that is being sampled. Lithological, geochemical, and microfabric data are being amassed now in this project and in the BRIM project (Steele et al., 1986) to determine the provenance of the tills and any shifts in ice-flow direction during deposition of a given till.

CONCLUSIONS

Very thick sequences of drift, much of which has little relation to local bedrock lithology, are present in the area north and northeast of Timmins. The Matheson and Cochrane tills at many sites have essentially no contact with the local bedrock (Fig. 2), and consist of a distal component and recycled older glacial and proglacial sediments. The use of the Matheson Till in drift prospecting is limited to locales where it rests on bedrock as it does in the BRIM area (Steele et al., 1986) or where it has recycled older metalliferous till, as has been shown at the Owl Creek mine (Bird and Coker, 1987).

The fossiliferous, nonglacial Owl Creek beds may represent a complete interglacial record that will aid in the linkage between the classic sequences in the Great Lakes basins and the Moose River basin. Their paleoecological significance will be determined only after lengthy analysis. This distinctive unit can already be used as a stratigraphic marker in drift prospecting using till sampling by overburden drilling. Formal definition of this unit is planned for the near future.

The assignment of provenances and ice-flow directions to the older (pre-Owl Creek) tills and the determination of ice-flow shifts within each till, if present, is of great importance. This is because the older tills had the most contact with the local bedrock, making them the best sampling media in drift prospecting in the areas of thick drift.

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An ultramafic dispersal train and associated gold anomaly in till near Osik Lake, Manitoba[1]

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Abstract

The Osik Lake dispersal train is about 7 km wide, at least 30 km long, and is enriched in ultramafic debris derived from a previously unmapped ultramafic body that probably underlies Osik Lake. Within the train, the lithology of the till changes up-section from a local to a distal provenance, relecting either a shift in ice-flow direction or a change from basal to englacial debris during deposition. Significant levels of gold were found in the heavy mineral fraction of the distally derived till.

Other ultramafic bodies may be detectable by till sampling. A new one is postulated to exist at Baldock Lake based on the occurence of ultramafic erratics, and others may coincide with aeromgnetic anomalies northeast of the lake.

Résumé

La traînée de dispersion d'Osik Lake a environ 7 km de large, au moins 30 km de long, et elle est enrichie en débris ultramafiques dérivés d'un corps ultramafique jusque-là non levé, probablement sous-jacent au lac Osik. À l'intérieur de la traînée, la lithologie du till varie dans le tronçon d'amont et indique le passage d'une provenance locale à une provenance distale, ce qui reflète soit un changement de la direction d'écoulement des glaces, soit une variation du caractère des roches détritiques, qui de basales deviennent intraglaciaires durant la sédimentation. On a découvert des concentrations significatives d'or dans la fraction minérale lourde du till de provenance distale.

Il est possible de déceler d'autres corps ultramafiques par échantillonnage des tills. On suppose qu'il existe un nouveau corps ultramafique à Baldock Lake, étant donné la présence de blocs erratiques ultramafiques, et le fait que d'autres corps coïncident sans doute avec des anomalies aéromagnétiques répérées au nord-est du lac.

[1] Contribution to Canada-Manitoba Mineral Development Agreement 1984-1989. Project carried by Geological Survey of Canada.

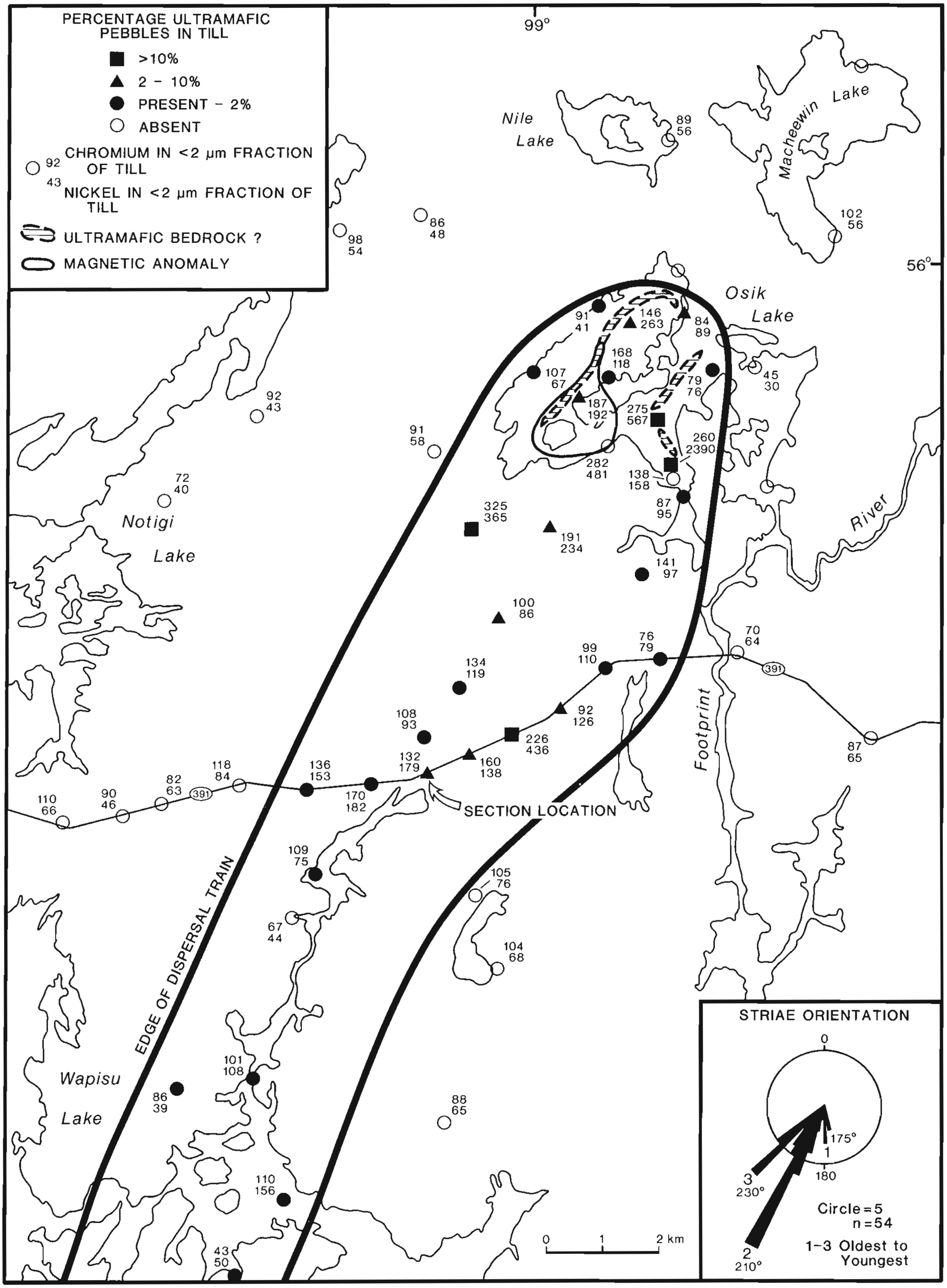

Figure 1. Map of the Osik Lake dispersal train showing abundance of ultramafic pebbles in till and the chromium and nickel contents of the till. Inset: rose diagram of trends of glacial striae in the region.

INTRODUCTION

During routine surficial geology mapping and sampling in 1986 along highway 391 between Leaf Rapids and Nelson House, ultramafic pebbles and cobbles were found in till exposed in road cuts along a 10 km stretch of the highway. There are no ultramafic units known in the local bedrock, and the only nearby ultramafic body, at Mynarski Lake, is 30 km to the northwest, an extremely unlikely direction for glacial dispersal. Using an ice flow direction derived from numerous striae measurements within the area, a detailed sampling program was carried out in the presumed up-ice direction northeast of the discovery sites. Till sampling and boulder tracing defined a classic dispersal train of ultramafic-rich till 7 km wide and at least 30 km long having its head or source in Osik Lake (Fig. 1). At the scale of sampling used here, the train has sharp lateral edges and trends about 210°, parallel to the dominant local striae direction. This direction is known to be the main direction of debris transport in this area and most dispersal trains are expected to be parallel to it (Kaszycki, 1987).

TILL COMPOSITION

Chromium and nickel contents of the till are high within the train and correlate with the presence of ultramafic pebbles and granules, which weather readily to ochreous mud. In Figure 1, the edge of the dispersal train has been placed at the up-ice and eastern and western limits of the area in which any ultramafic clasts were found in till. At almost every sample site within the train, the chromium and nickel contents of the till are two to three times higher than in the surrounding region.

Vertical compositional variability of till within the train was examined at a section exposed in a roadcut along the highway. Composition was estimated first by the weight percent of distinctive indicator lithologies present in the granule fraction (2-5.6 mm) of the till. Till composition in terms of three indicator lithologies is presented here (Fig. 2):

1. Ultramafic rocks, representing local, small scale dispersal (<30 km);
2. Paleozoic carbonates; and
3. Greywackes of the Omarolluk Formation. The last two representing long range transport (>200 km) from Hudson Bay and influx of exotic material within the region.

The abundance of ultramafic pebbles ranges from 0.6 % at the top of the section to 5.2 % at the base. Although variable, the general trend is for slightly decreasing ultramafic content up-section, with a dramatic decrease at the top. In contrast, the abundance of Paleozoic carbonate and Omarolluk greywacke exhibits a slight upward increase throughout most of the section and increases dramatically at the top. These provenance shifts may reflect the late-glacial shift in ice-flow from south-southwestward (210°) to southwestward (230°) observed in striae orientations (Fig. 1), a pattern of ice-flow shift that is widespread in this region (Kaszycki and DiLabio, 1986), resulting in the concurrent decrease in ultramafic erratics and increase in erratics of Hudson Bay (eastern) provenance. Alternatively or coincidentally, compositional variation may be related to sediment facies and reflect a shift from deposition of debris transported near the base of the ice sheet, to deposition of englacially transported debris. The section is located on the lee-side of a rock knob, around which ice flow may have been redirected and over which it may

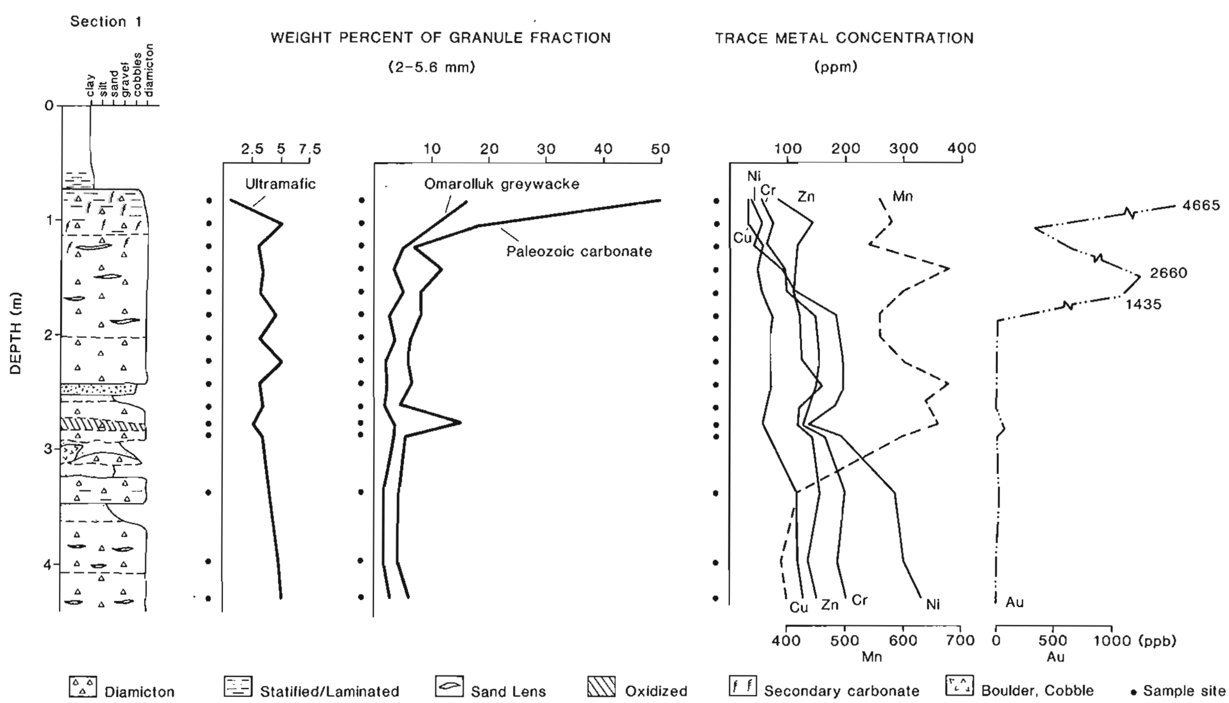

Figure 2. Section on the axis of the Osik Lake dispersal train showing lithologic and geochemical data on till in the train. Section is in a road cut on highway 391, shown in Figure 1. Trace element contents of <2 μm fraction and gold content of heavy minerals are shown.

have arched. It is difficult to determine without more detailed analyses, how much of the stratification observed at this site (Fig. 2) represents deposition by debris flow into a subglacial cavity and how much represents deposition in the ice-marginal environment. Detailed sedimentary logging revealed that the degree of internal stratification within the till increases sharply in the upper 40 cm, coincident with elevated carbonate concentrations and a much finer texture. The stratification and carbonate content may indicate that this sediment facies reflects ice-marginal debris flow sedimentation of englacially transported distal material into glacial Lake Agassiz, whereas the lower portion of the sequence represents subglacial deposition of basally transported local debris. Further studies of this type are required to more fully evaluate this problem.

A second set of compositional data consists of the trace element contents of the clay-sized fraction ($<2\ \mu m$) and the sand-sized heavy mineral fraction of the till (Fig. 2). As expected, levels of trace elements that are ultramafic indicators (Cr, Ni, Cu) decrease gradually up-section, following the trend seen in the abundance of ultramafic rocks. Manganese levels increase up-section, roughly following the trend defined by the abundance of Paleozoic carbonates, which is consistent with the higher manganese levels seen in calcareous till of Hudson Bay provenance in this region. The high gold content of several of the heavy mineral fractions was unexpected. In the upper, most calcareous part of the till section, the gold content of the heavy minerals is over 1400 ppb, whereas in the carbonate-poor lower part of the section, the gold content is at or near detection limit. Only a few rusty pyrite grains were seen in the anomalous samples. Otherwise, the heavy mineral suites are "normal", consisting mainly of garnet, magnetite, epidote, and hornblende.

Examination of the pebble and granule fractions of the gold-bearing samples showed that they contained rare clasts of a pyritic, graphitic schist. This rock type was found and sampled from bedrock and from erratics in till along highway 391 in the vicinity of the dispersal train. It is similar to that found in the transition zone between the Sickle Group and the Burntwood River Metamorphic Suite (Baldwin, 1980). Analysis of 38 samples taken in 1987 showed that this rock type and other local ones are not gold-bearing. If the scenario of late shift in ice-flow towards the west-southwest is accepted, then the calcareous till could contain sulphide-bearing clasts derived from new occurrences or the known showings at the Burntwood River-Sickle transition on and near Woweyakumaw Lake, 12 to 15 km east-northeast of the sites along the highway (Frohlinger and Kendrick, 1978). If, however, the gold was contained in englacially transported distal debris, its source will be much more difficult to find; it can only be described as somewhere to the east.

ULTRAMAFIC ROCK TYPES

The rock types present in the train are best illustrated by those collected from wave-washed till on boulder beaches in Osik Lake. The dominant rock types are apple green and pistachio green massive and brecciated serpentinized dunite or peridotite containing abundant magnetite and traces of disseminated chromite, dark green aphanitic to glassy serpentinite, red and brown hematized serpentinite, coarse grained massive tremolitic rocks, gabbro, and possible serpentinized wall-rocks or contact rocks containing abundant phlogopite. A very small number of serpentinites were collected that contain radiating acicular ghost crystals (probably not spinifex

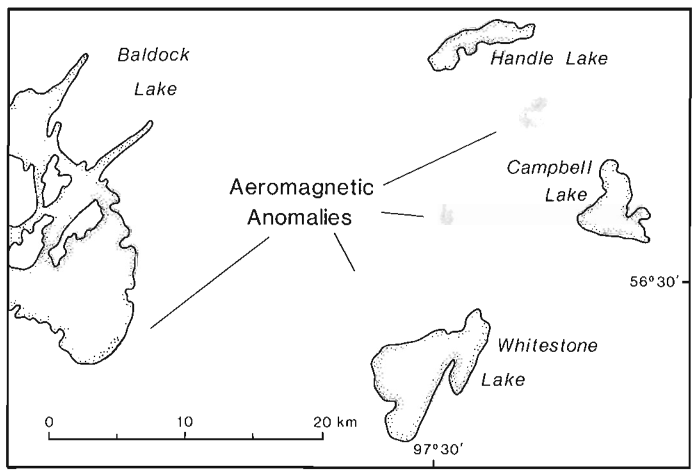

Figure 3. Aeromagnetic anomalies near Baldock Lake. After Geological Survey of Canada Map 7025G.

Table 1. Analyses of trace elements and iron in rocks from till at Osik Lake

Sample		Cr ppm	Ni ppm	Fe %	Co ppm	Zn ppm	Au ppb	Pd ppb	Pt ppb	Rock Type
86KDA	274-01	3570	1300	2.4	37	<100	2	19	<15	massive tremolite
	274-02	2950	2150	4.5	300	<100	<2	2	<15	aphanitic serpentinite
	274-03	2000	3300	17.0	160	<100	3	4	<15	serpentinized perid/dunite magnetite veinlets
	274-04	1000	1700	21.9	310	<100	3	2	<15	serpentinized perid/dunite magnetite veinlets
	274-05	410	680	69.5	120	1900	<2	<2	<15	serpentinized perid/dunite magnetite veinlets
	274-06	12600	1700	8.7	110	100	<2	2	16	hematized serpentinite
	274-07	570	200	4.0	21	<100	<2	2	<15	gabbro
	275-01	3600	1500	8.2	63	<100	<2	6	<15	hematized serpentinite
	275-02	6020	1400	11.0	50	100	4	3	<15	hematized serpentinite

(Cr, Ni, Fe, Co, Zn, and Au analyses by neutron activation; Pd and Pt analyses by fire assay-DCP) texture). Only a limited number of rocks have been analyzed for trace element geochemistry (Table 1). These contain high levels of nickel, chromium, cobalt, and iron, and low levels of Pt, Pd, and Au. Only trace amounts of sulphides have been observed. XRD analyses indicate that the main mineral components of these lithologies are serpentine, talc, and magnetite.

The known bedrock geology of the Osik Lake basin indicates that the lake sits in a canoe-shaped structural depression underlain by biotite hornblende gneiss derived from arkosic Sickle Group units. Virtually all shoreline outcrops were examined in the original bedrock mapping (1:50 000 scale) and no ultramafic outcrops were observed (Frohlinger, 1978; Frohlinger and Kendrick, 1978). Based on our study of clasts in the till, we propose that at least two lensoid bodies of ultramafic rock underlie the lake, a tremolite-rich body under the western arm and a serpentinized dunite/peridotite body under the eastern arm. This proposal is corroborated by the presence of a relatively strong aeromagnetic anomaly over the south end of the lake. Because of the preferential thickening of Lake Agassiz clays in structural basins, the full extent of this aeromagnetic anomaly cannot be considered as proven, as the clays may mask part of it.

SUSPECTED ULTRAMAFIC BODIES

Ultramafic-rich till was found at another site, on Baldock Lake, 75 km northeast of Osik Lake. In an area of no outcrop, this site is also marked by a strong aeromagnetic anomaly is one of a string of four extending northeast for a distance of 35 km (Fig. 3), we suggest that sampling till over and down-ice from these locations may reveal the presence of similar ultramafic bodies, all located within the Churchill Province, far removed from the Thompson belt. A brief search in 1987 found no ultramafic indicators along this trend northeast of Baldock Lake.

At present, we consider that the economic potential of the Osik Lake, Baldock Lake, and other possible ultramafic bodies is totally untested. They should be considered as Ni-Cu, chromite, and PGE exploration targets. Country rocks around them may be enriched in gold. Local directions of glacial dispersal and striae need to be established for the Baldock Lake and northeastern areas so that the indicator rocks may be traced to their sources.

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Reconstruction of synvolcanic alteration associated with the Linda massive sulphide deposit, Snow Lake, Manitoba[1]

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Abstract

The Linda volcanogenic massive sulphide deposit is hosted by felsic volcanic rocks on the overturned limb of the Anderson Bay anticline. Lithological units are discontinuous across the hinge region, precluding direct correlation between the Anderson Lake, Stall Lake and Rod deposits on the northwest limb of the anticline, and the Linda deposit on the southeast limb. Two synvolcanic alteration zones stratigraphically underlie the main pyritic body of the Linda deposit. The distal alteration zone is stratabound and represents a hydrothermal fluid reservoir. The proximal alteration zone, adjacent to the pyritic body, represents a discordant conduit. Thermochemical evolution of hydrothermal fluid in the reservoir, followed by expulsion into the conduit, resulted in wallrock alteration and the precipitation of sulphide minerals.

The Linda deposit was subjected to amphibolite-facies regional metamorphism and altered rocks are characterized by assemblages of metamorphic minerals. The parageneses reflect variations in bulk-rock compositions established during synvolcanic hydrothermal alteration.

Résumé

Le gisement de sulfure massif volcanique de Linda est porteur de roches volcaniques felsiques sur le flanc returné de l'anticlinal d'Anderson Bay. Les unités lithologiques sont distribuées de façon discontinue le long de la zone charnière de la région, ce qui exclut une corrélation directe entre les gisements d'Anderson Lake, de Stall Lake et de Rod situés sur la charnière sud-est. Deux zones d'altération synvolcanique supportent stratigraphiquement le corps principal pyritique du gisement de Linda. La zone d'altération distale est limitée stratigraphiquement et constituant un réservoir à fluide hydrothermal. La zone d'altération proximale, adjacente au corps pyritique, est constituée d'un conduit discordant. Le dégagement thermochimique du fluide hydrothermal contenu dans le réservoir se produit par expulsion dans le conduit ce qui provoque l'altération de la roche encaissante et la précipitation des minéraux sulphurés.

Le gisement de Linda a été soumis à un métamorphisme régional du faciès des amphibolites et les roches altérées sont caractérisées par un assemblage de minéraux métamorphiques. La paragénèse reflète des variations dans la composition des roches établies au cours de l'altération synvolcanique hydrothermale.

[1] Contribution to Canada-Manitoba Mineral Development Agreement 1984-1989. Project carried by Geological Survey of Canada, Lithosphere and Canadian Shield Division

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LOCATION AND SETTING

The Linda volcanogenic massive sulphide deposit is located in the Snow Lake area, within the Flin Flon-Snow Lake volcanic belt in the Churchill Province of the Canadian Shield. The volcanic belt is flanked on the north by the Kisseynew sedimentary gneiss belt, and overlain on the south by Paleozoic sediments. Other massive sulphide deposits in the Snow Lake area include the Chisel Lake, Ghost Lake/Lost Lake, Anderson Lake, Stall Lake and Rod mines. The Linda deposit is a mainly pyritic body with minor pyrrhotite, chalcopyrite and sphalerite. It was extensively drilled by Falconbridge Limited and Corporation Falconbridge Copper (now Minnova Incorporated), who provided access to drill core and to unpublished maps and company reports.

REGIONAL GEOLOGY

The geology, structure and metamorphism of the Snow Lake area were discussed by Froese and Moore (1980). Metavolcanic rocks of the Amisk Group, ranging from felsic to mafic, are commonly overlain by metagreywacke and shale (Fig. 1). The interpretation, by Walford and Franklin (1982), of the quartz-eye tonalite body as a synvolcanic sill was supported by Bailes (1986) who found that large areas of the pluton were subjected to synvolcanic hydrothermal alteration. Metasediments of the Missi Group consist of crossbedded lithic arenites which were deposited on Amisk Group rocks.

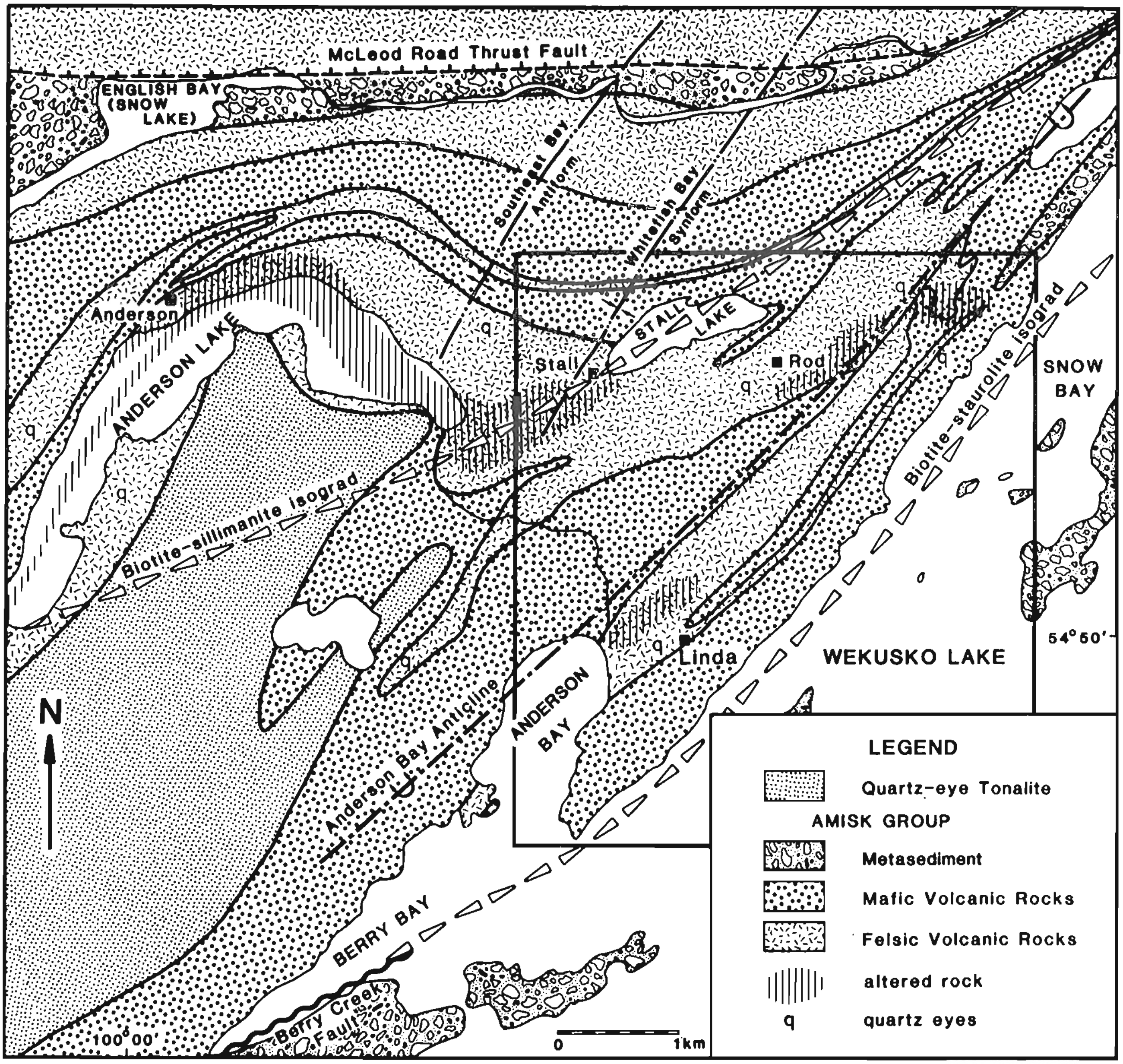

Figure 1. Geology of a part of the Snow Lake area in the vicinity of the Linda deposit. Modified from Jeffery (1982), Froese and Moore (1980) and Zaleski (1986). The trace of the proposed Anderson Bay anticline is approximately located, as constrained by opposed facing directions. The area detailed in Figure 2 is outlined.

The regional structure of the Snow Lake area is poorly understood (E. Froese, pers. com., 1986). Three main deformational events were proposed by Froese and Moore (1980). The earliest event produced tight to nearly insoclinal F1 folds, recognized mainly on the basis of opposing facing directions. Interpretation of the stratigraphic sequence of interlayered felsic and mafic volcanic rocks and the relative stratigraphic position of the massive sulphide occurrences is dependent on the recognition of F1 folds. Primary structures indicate northward facing for the succession from Stall Lake to Snow Creek (east of English Bay, Fig. 1). The existence of the Anderson Bay anticline was postulated by Falconbridge geologists (Zaleski, 1986; Jeffery, 1982, unpubl.) to account for a southward facing direction required by the disposition of the sulphide body and stringer and alteration zones at the Linda deposit. The more northerly trace of the Anderson Lake anticline, postulated by Gale and Koo (1977) and Froese and Moore (1980), may be the result of mistaken identity between the Linda I sulphide occurrence (shown on the maps of the mentioned authors), a minor stringer zone which does not yield a facing direction (Jeffery, 1982, unpubl.), and the Linda II deposit, which is the focus of this study. An S1 fabric is defined by metamorphic layering, by mineral alignment and by flattening of fragments subparallel to primary compositional layering.

Northeasterly trending open flexures of the Threehouse synform, Southeast Bay antiform and Whitefish Bay synform were attributed to a second deformational event (Froese and Moore, 1980). To the north, flanking the Kisseynew gneiss belt, these structures have been tightened by subsequent gneiss dome emplacement (D3), but they die out toward the south. D2 minor structures vary in amplitude from decametre-scale F2 folds to millimetre-scale crenulations. An S2 axial planar fabric is defined by mineral foliation, particularly of micas. A pronounced, northeasterly plunging elongation of all sulphide bodies and other primary structures, as well as a mineral lineation, was mainly attributed to L2 fabric development. However, in the Chisel Lake area, a slight angular separation (5-15°) between fragment-elongation directions and S1/S2 intersections defined by mica edges, was interpreted to indicate that regional elongation should be partly attributed to extension during D1 deformation.

The McLeod Road thrust fault was postulated to account for older Amisk volcanic rocks overlying younger Amisk and Missi sediments (Froese and Moore, 1980). The fault was interpreted to be earlier than, or contemporaneous with, D1. In the Crowduck Bay area, immediately northeast, analogous relationships led Gordon and Gall (1982) to postulate the presence of an early thrust fault. Froese and Moore (1980) stressed the difficulty of detecting such features.

The late Berry Creek fault forms a northeasterly trending lineament west of Wekusko Lake, but its eastward extension is obscure (ibid.). To the northeast, the Crowduck Bay fault also has a northeasterly trending trace and cuts metamorphic isograds (Gordon, 1981).

Metamorphic assemblages in the Snow Lake area show the transition from the low grade volcanic terrane in the south to the high grade terrane of the Kisseynew gneiss belt (Froese and Moore, 1980; Froese and Gasparrini, 1975). Three reaction isograds delineate the boundaries of four metamorphic zones characterized by the assemblages: chlorite-biotite, chlorite-biotite-staurolite, biotite-staurolite-sillimanite and biotite-sillimanite-almandine. The parageneses in muscovite-bearing pelites were regarded as definitive of metamorphic grade. The common presence of kyanite or kyanite and sillimanite, in alteration zones, was used to indicate metamorphic pressures in excess of the aluminosilicate triple-point. In the Anderson Lake and Stall Lake mines, the occurrence of the isograd assemblage, chlorite-staurolite-kyanite (sillimanite)-biotite-muscovite, was used to constrain the location of the surface trace of the biotite-sillimanite isograd. The discordant relationship of the isograds to the regional structure suggested that peak metamorphic conditions were attained after the occurrence of most of the D2 deformation. Near the gneiss domes, prograde metamorphism continued after gneiss dome emplacement (D3), resulting in static recrystallization of high grade minerals.

LOCAL GEOLOGY

The sulphide mineralization and alteration, associated with the Linda deposit, occur in Amisk felsic volcanic and volcaniclastic rocks with minor intercalated lenses of fine grained graphitic argillite. Lithological units strike northeast and dip to the northwest (Fig. 2). Alteration and stringer sulphide zones lie to the northwest of the main pyritic body, overlying the massive sulphide deposit in the present configuration. This relationship indicates that the Linda deposit is overturned and that stratigraphic facing is to the southeast (see Structure for further discussion).

Description of lithological units

The litholigical units represent a compilation of surface mapping (Fig. 2) and drillhole data (Fig. 3). Nomenclature has been based on megascopic examination of drill core and outcrops, and unit designations are intended as informal field terms. All rocks are metamorphosed, but the prefix "meta" has generally been omitted.

Unit 1. Felsic volcanic and volcaniclastic rocks. White to light grey to buff, the unit may include rhyolitic to dacitic rocks. Biotite is the most abundant ferromagnesian mineral and commonly defines a fine compositional layering of metamorphic origin (S1). The lowermost member is rhyodacitic to dacitic and variably altered in the distal semiconformable alteration zone (described in a later section) associated with the Linda deposit. Volcanic breccias (1b) dominate the uppermost 50-200 m and are intercalated with massive fine grained to aphanitic felsic rocks (1a). These upper members host the proximal alteration zone and stringer sulphide mineralization.

Unit 2. Basalt. Varies from massive to weakly layered and foliated (S1) with concentrations of plagioclase phenocrysts(?) conforming to the lensoid outer margins of evenly distributed, epidotized, alteration patches (2a). Pillowed flows (2b) are commonly sheared and slightly or intensely bleached. Hornblende is the most abundant mafic mineral and generally defines a strong lineation.

Unit 3. Graphitic sediment. Fine grained sediment occurs in lenses of limited lateral extent at two horizons in the upper part of unit 1, within the proximal alteration zone. The

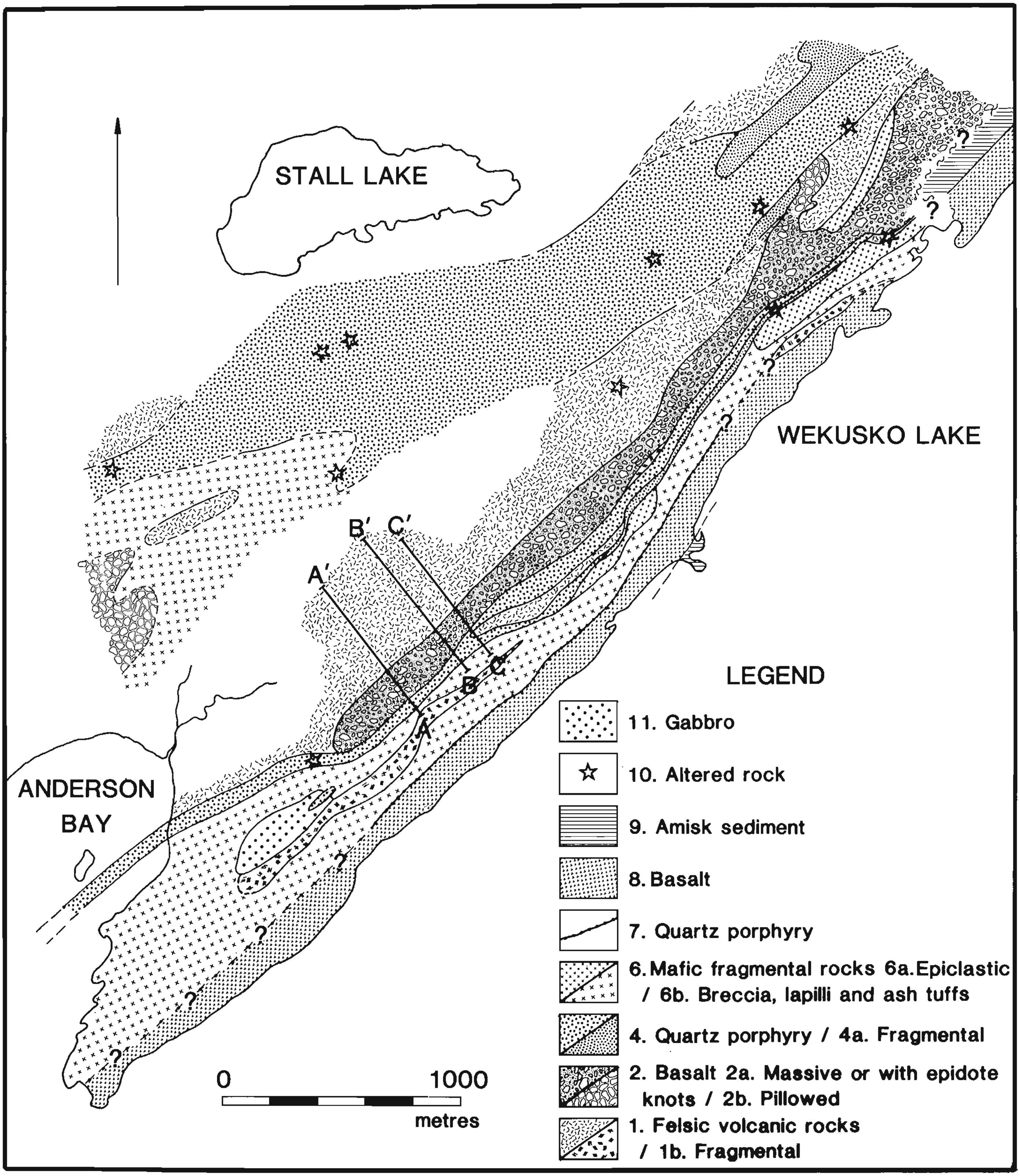

Figure 2. Geology of the Linda area. The surface projections of three composite cross-sections constrained by drill core data are indicated by A-A', B-B' and C-C'. Section C-C' is illustrated in Figure 3.

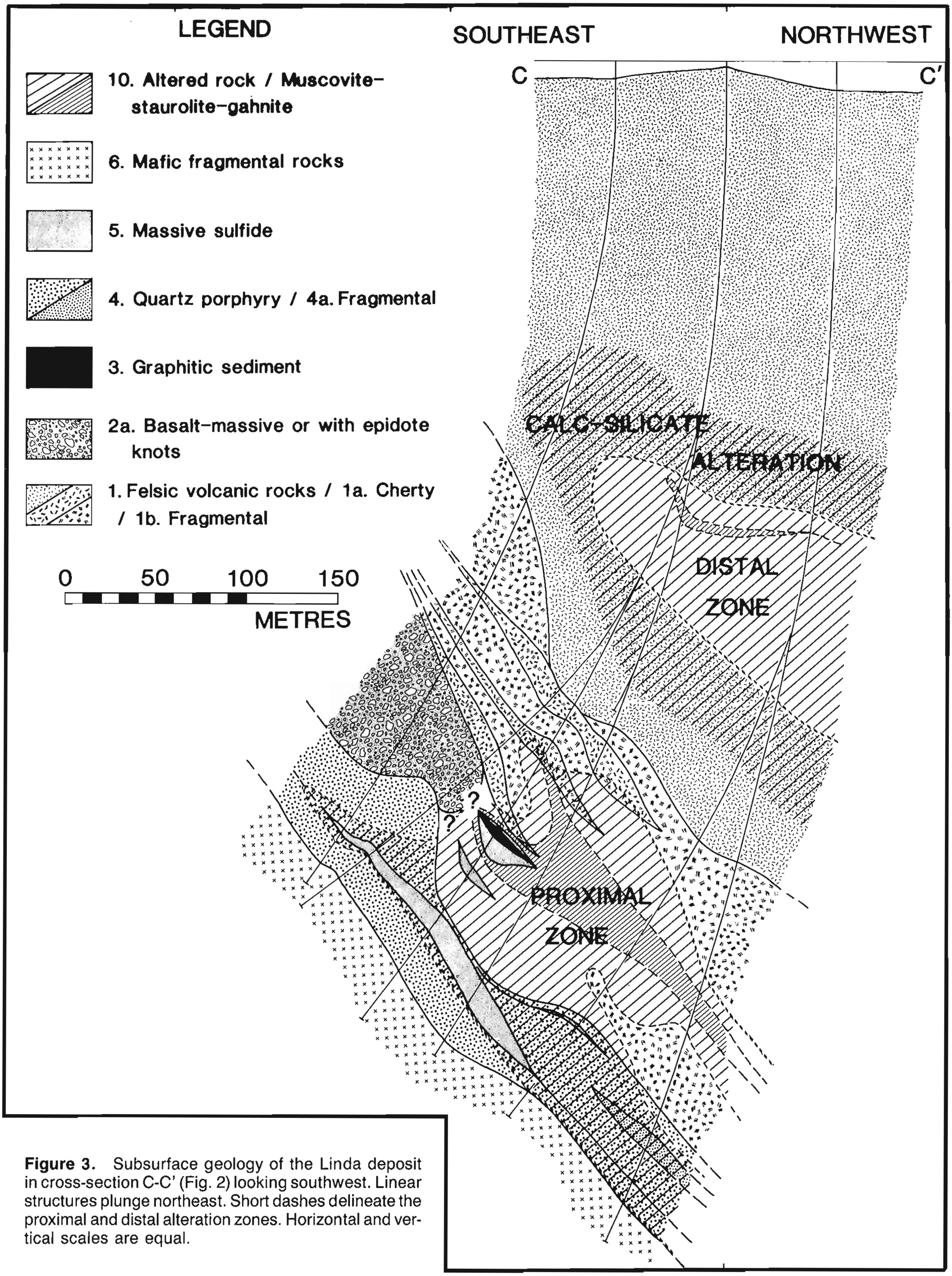

Figure 3. Subsurface geology of the Linda deposit in cross-section C-C' (Fig. 2) looking southwest. Linear structures plunge northeast. Short dashes delineate the proximal and distal alteration zones. Horizontal and vertical scales are equal.

unit is thinly bedded, dark, argillaceous, graphitic, mineralized with pyrite and pyrrhotite, and commonly associated with small massive sulphide lenses.

Unit 4. Quartz porphyry. Blue quartz eyes of 3-8 mm diameter constitute 5-25 % of the rock. A great thickness of mainly massive, locally subtly fragmental, quartz porphyry occurs northwest of the Anderson Bay anticline. On the southeast, the unit is thinner and more variable, commonly intercalated with felsic volcanic breccias (4a) containing clasts of quartz porphyry. The mineralogy and mafic mineral content are also variable, including amphibole-rich and biotite-rich types which commonly define centimetre-scale discontinuous layering. The unit hosts the main massive sulphide body, and alteration has affected both the stratigraphic footwall and hanging wall.

Unit 5. Massive sulphide. A massive to semimassive sulphide body forms a concordant lens associated with quartz porphyry (unit 4). Medium- to coarse-grained granular pyrite is intergrown with abundant calcite and/or quartz. Pyrrhotite, sphalerite and chalcopyrite are sporadically distributed in minor amounts. Several small sulphide lenses are associated with stringer and disseminated mineralization, stratigraphically underlying the main sulphide body. These minor lenses vary from dominantly pyrite-calcite, to pyrite with minor pyrrhotite-sphalerite-chalcopyrite and silicate alteration minerals.

Unit 6. Mafic volcaniclastic rocks. These consist mainly of felsic to intermediate lapilli or coarse ash and crystals(?) in a dark green mafic matrix with weak, diffuse, decimetre- to centimetre-scale layering defined by variation in clast concentrations from 0-20 % (6b). Coarser varieties, containing decimetre-sized clasts, are usually matrix-supported, heterolithic and poorly sorted. In the northeastern part of the area (Fig. 2), a distinctive epiclastic subunit (6a) occurs stratigraphically above the quartz porphyry (unit 4) and contains clasts of quartz porphyry and aphyric felsic rocks. The unit also includes some layered, coarsely recrystallized amphibolites in which primary structures are obscured.

Unit 7. Quartz porphyry. A second thin (0-10 m) unit of quartz porphyry forms a useful marker horizon higher in the stratigraphic sequence. Centimetre- to metre-scale layering is sharply defined by quartz-eye and feldspar-phyric rock intercalated with feldspar-phyric types. Although the unit apparently pinches out in one location and varies in thickness, it forms a traceable horizon for 2000 m (Fig. 2).

Unit 8. Basalt. The unit is fine grained and mainly massive but with a well developed chloritic cleavage. Of less common occurrence are coarse ash or crystal(?) tuffs, similar to those of unit 6b, or bedded tuffaceous sediment.

Unit 9. Metasediment. Two exposures of bedded greywacke have been grouped with Amisk metasediments.

Unit 10. Altered rock. Two distinct schistose alteration zones have been delineated in drill core and are described in detail in a later section. A distal, subconcordant zone occurs stratigraphically 200-250 m below the sulphide body. A proximal, discordant zone is developed in quartz porphyry (unit 4) and felsic volcanic rocks (unit 1a, 1b) directly underlying the main sulphide body. Surface exposures of altered rocks (Fig. 2), identified on the basis of the presence of staurolite with or without kyanite, muscovite, chlorite, pyrite and garnet, occur in several localities mainly within quartz porphyry (unit 4).

Unit 11. Gabbro. A small body of undeformed, coarse grained, hypidiomorphic gabbro intrudes volcanic rocks southeast of Anderson Bay.

Alteration

Two discrete alteration zones have been recognized, developed in dominantly felsic volcanic rocks (Fig. 3). The altered rocks are characterized by assemblages of metamorphic minerals resulting from regional metamorphism of synvolcanic hydrothermal alteration.

The distal alteration zone stratigraphically underlies the sulphide mineralization and proximal zone, being separated from these by apparently unaltered or slightly altered felsic volcanic rocks. The mineralogy of the altered rock is dominated by chlorite, with assemblages including staurolite, kyanite, muscovite, magnetite, biotite and gahnite. Garnet is present around the outer margins of the most intensely altered rocks. Calc-silicate minerals occur in felsic volcanic rocks on the periphery of the distal alteration zone and, to a lesser degree, between the alteration zones. Localized patches of up to 1 m length (drill-core intersection) contain tremolite/actinolite, epidote and calcite. Immediately adjacent to, and enveloping, the distal alteration zone, the development of calc-silicate minerals, especially epidote, is more pervasive, and this volume has been designated as calc-silicate alteration (Fig. 3).

The proximal alteration zone forms the footwall alteration beneath the main sulphide body and envelops minor sulphide lenses, stringer and disseminated mineralization, and mineralized graphitic sediment. Hanging wall alteration is variably developed and, in some intersections, extends to the contact with the hanging wall mafic fragmental unit (unit 6) where it abruptly dies out. Alteration affects felsic fragmental rocks, felsic massive rocks and quartz porphyry. Altered quartz porphyry can be identified by its quartz eyes, becoming a quartz-augen schist. Similarly, felsic massive rocks (unit 1a) largely retain their character, but are commonly brecciated with stringer sulphides or alteration minerals filling the fractures. They can be traced into the alteration zone, but the reason for their down-dip truncation (Fig. 3) is unclear. Felsic fragmental rocks (unit 1b) are less commonly recognizable in their altered state. However, in some cases, clasts have been more resistant to alteration and matrix and clasts display a contrasting mineralogy.

The mineralogy of the proximal alteration zone is dominanted by muscovite with assemblages including kyanite, staurolite, gahnite, biotite, chlorite, margarite, tourmaline, and rarely, magnetite, garnet, orthoamphibole and fuchsite(?). Within the intensely altered rocks is a core zone, apparently associated with the graphitic sediment and sulphide lenses, which is characterized by muscovite, staurolite and gahnite. As in the distal alteration zone, garnet occurs mainly on the periphery.

Structure

The presence of the F1 Anderson Bay anticline (Fig. 1), with an axial trace lying between the Anderson Lake, Stall Lake and Rod deposits on the northwest limb, and the Linda deposit on the southeast limb, was proposed by Falconbridge geologists (Jeffery, 1982, unpubl.) on the basis of opposing facing directions. The morphology of the Linda deposit, with stringer sulphide and alteration zones overlying the massive sulphide body, requires stratigraphic facing to the southeast. The presence of clasts of quartz porphyry in epiclastic fragmental rocks (unit 6a) in contact with quartz porphyry (unit 4) on its southeast side provides supporting evidence for this interpretation. The Anderson Lake and Stall Lake deposits face north (Walford and Franklin, 1982; Studer, 1982) and the morphology of pillowed flows observed in the vicinity of Kormans Lake (about 500 m northeast of Fig. 2) also indicates northward facing. On the basis of surface mapping, drillhole data and the analogous lithological setting of the massive sulphide deposits, a correlation was proposed between the units on each side of the Anderson Bay anticline (Jeffery, 1982, unpubl.). In particular, the quartz porphyries hosting massive sulphide deposits on each limb of the anticline were proposed to be correlative. However, outcrop mapping (this study) showed that the lithological units are discontinuous (Fig. 2) and do not define a fold nose. Therefore, although a fold closure is suggested by the opposed facing directions, the correlation of stratigraphic units on each limb of the fold is problematic.

An L-S tectonite fabric dominates the structure, as evidenced by the northeasterly trending elongation of primary features such as sulphide lenses, pillows and clasts. The main sulphide body of the Linda deposit exceeds 1500 m in down-plunge length; the maximum strike width is less than 200 m. Planar primary structures, such as lithologic contacts, bedding and layering are structurally transposed into northeasterly strikes with dips to the northwest. Two deformational events are required to explain the observed minor structures and their relationships.

D1 deformation is manifested in flattening of pillows, of clasts in coarsely fragmental units, and of quartz eyes in quartz porphyry. In felsic rocks, a fine metamorphic compositional layering is defined by segregation of micaceous minerals. In altered rocks, S1 fabric development resulted in a strong schistosity and, in some cases, in compositional layering defined by alternating staurolitic and phyllosilicate layers. The S1 fabric parallels or subparallels primary contacts, striking mainly 220-240° with dips of 50-70°. Poles to S0 and S1 define a strong maximum with a tendency to distribute along a great circle about a pole at 25°/35° (Fig. 4).

Structures assigned to D2 include planar and linear fabrics. Tight, mesoscopic F2 folds are relatively common and are observed to re-orient S1 fabrics and, less commonly, lithological contacts. They range in amplitude from millimetre-scale crenulations to hectometre-scale folds. In some localities, S0 contacts and parallel S1 metamorphic fabric have northwesterly strikes in the hinge zones of the larger scale F2 folds. The folds show a dominantly Z-asymmetry throughout the area. S2 schistosity is commonly defined by biotite aligned at an oblique angle to S1 compositional layering and by a crenulation cleavage imparted to S1 foliated muscovite and chlorite. It is observed to be axial planar to mesoscopic F2 folds. The strike is similar to that of S0 and S1 fabrics, but with nearly vertical dip (Fig. 4).

A lineation is defined by elongation of clasts, pillows and sulphide lenses, as well as by mineral lineations (quartz eyes, amphibole, micas) and by mineral aggregates (quartz eyes, garnet). These linear elements are apparently coaxial, but a small angular deviation would likely go undetected, especially as measurement of lineation was possible only on

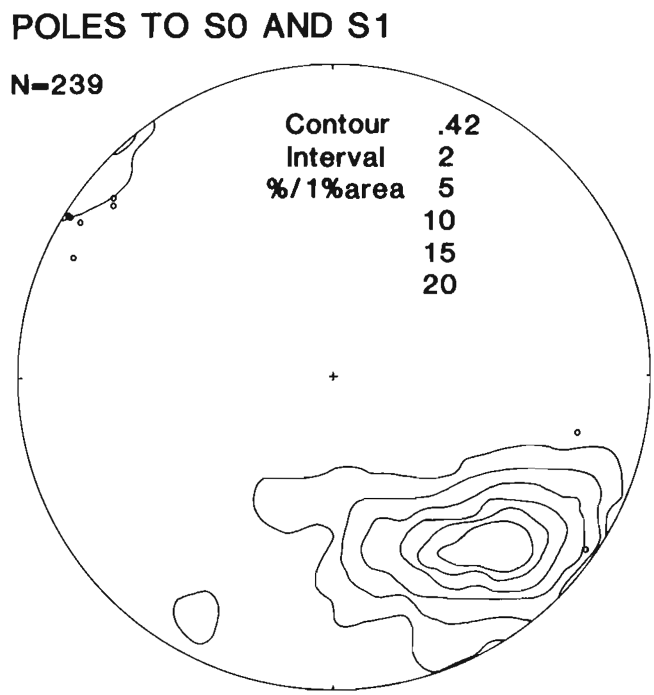

Figure 4. Stereographic projection of contoured poles to S0 and S1. Eight poles to S2 are indicated by circles.

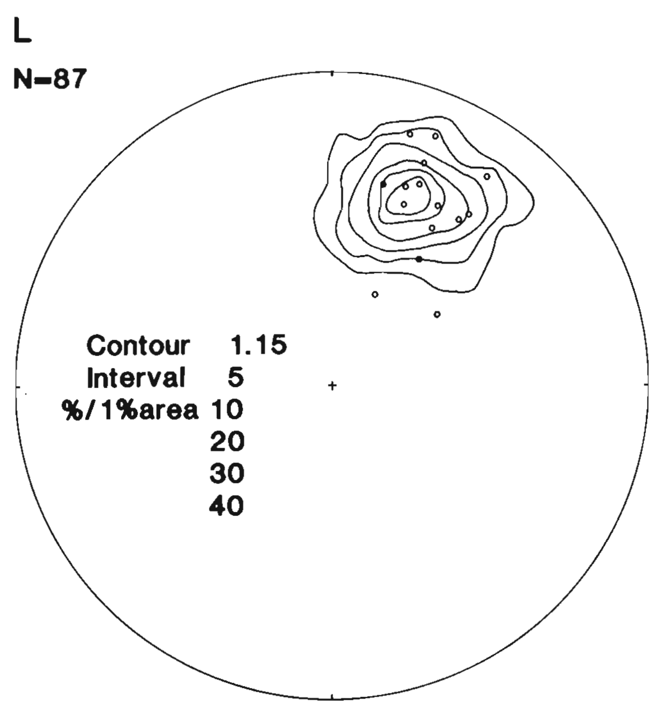

Figure 5. Stereographic projection of contoured mineral lineations. Fifteen fold axes of minor F2 folds are indicated by circles.

minerals and mineral aggregates. The lineation has previously been assigned to L2 (Froese and Moore, 1980). However, the principal flattening direction is defined by S1, and the lineation was measured on S1 surfaces, both points implying that the lineation should be assigned to L1 (J.R. Henderson, pers. comm., 1987).

The lineation shows a strong maximum centred on a trend of 23° plunging at 35° (Fig. 5). It is also coaxial with minor F2 fold axes and with a pole defined by the S0 and S1 partial girdle. In the case of pre-existing mega- and mesoscopic structures, such as sulphide lenses, clasts and pillows, L represents a stretching lineation and, as previously noted, small angular deviations from the mineral lineation direction could go undetected. Mineral lineations were commonly measured on quartz eyes, the elongation of which is the result of a combination of strain and recrystallization. Even if an angular difference exists between stretching and mineral lineations, this could reflect the difference between the principal elongation direction and the principal kinematic direction in a simple-shear regime (Escher and Watterson, 1974). It is not necessary to invoke two deformations to account for the lineations, but neither is it possible to unequivocally discriminate between these possibilities.

Metamorphism

The mapped area lies within the chlorite-biotite-staurolite zone of Froese and Moore (1980). However, altered rocks associated with the Linda deposit contain kyanite, suggesting that the metamorphic grade exceeded the biotite-aluminosilicate reaction isograd, or that the kyanite was produced by a lower grade reaction. Staurolite and kyanite porphyroblasts generally grew in S1 planes and display the same intense S2 crenulation or corrugation as the enclosing phyllosilicates. Garnet commonly has a tabular shape, parallel to S1, apparently resulting from crystallization mimetic after S1 compositional layering. Porphyroblastic minerals and crenulation fabrics commonly show an asymmetry consistent with a component of rotational strain. These observations indicate that deformation in the Linda area, in particular D2, continued after the achievement of peak metamorphic conditions.

Significant gradients of metamorphic pressure or temperature are unlikely within the small volume represented by the three cross-sections of the Linda deposit. The variety of parageneses mainly reflects differences in bulk-rock composition established during synvolcanic alteration.

Reconstruction of the Linda hydrothermal system

The distal and proximal alteration zones, and the main pyritic body, have down-plunge lengths in excess of 1500 m. These units are open-ended in both up-plunge and down-plunge directions, i.e. they intersect the present erosional surface and drilling has not defined their down-plunge limits. The main pyritic body is 150-200 m in dip-length, with a lensoid cross-section varying from a flattened ovoid (Fig. 3) to a broad flat-topped U (Fig. 2.3 *in* Zaleski, 1986). A thin (<3m) layer commonly extends up to 100 m laterally beyond the main lens. The down-dip limits of the alteration zones were not defined by drilling, although the proximal alteration narrows and bifurcates (Fig. 3). The proximal alteration envelops the main pyritic body and its thin pyritic lateral equivalents. The bifurcation and the disposition of alteration within the quartz porphyry is suggestive of a deformed funnel-shape, widening toward the pyritic body, with the narrow base transposed into parallelism with the wide top. However, in three dimensions, this cross-sectional shape continues along-plunge resulting in an overall sheeted aspect.

The disposition of the alteration zones at the Linda deposit is similar to that reported for deposits on the northwest limb of the Anderson Bay anticline. Beneath the orebodies at the Anderson Lake and Stall Lake mines, alteration pipes cut stratigraphy at an oblique angle (Walford and Franklin, 1982; Studer, 1982). At the Anderson Lake mine, the pipe extends down to intersect a lower conformable alteration zone. Walford and Franklin (1982) interpreted the lower zone as the location of a hydrothermal fluid reservoir of regional extent, intersected by several fluid conduits which allowed fluid uprise and which are now manifested by the discordant alteration pipes. The spatial relationships at the Linda deposit are consistent with this interpretation, although an intersection between the distal and proximal alteration zones has not been documented. The distal and proximal alteration zones are tentatively proposed to represent a hydrothermal fluid reservoir and a conduit, respectively.

The position of the proximal alteration zone, immediately down-dip from the truncations of the basalt (unit 2) and the three cherty rhyolites (unit 1a), and the association with thinly bedded graphitic sediment (unit 3), suggest that a primary structure, such as a synvolcanic fault, controlled the location of fluid uprise (Fig. 3). The existence of such a feature, and its nature, might be extremely difficult to demonstrate definitively. If the path of fluid ascent was controlled by a synvolcanic fault, the fault itself may be entirely within the proximal alteration zone and completely obscured by hydrothermal alteration. Structural transposition has effectively obscured primary geometrical relationships. The discontinuities in the volcanic and sedimentary units remain equivocal unless a displacement can be documented.

The reconstructed morphology of the hydrothermal system at the Linda deposit implies the development of a compositionally evolved fluid in a reservoir represented by the distal alteration zone and the venting of the fluid through a conduit, now the proximal alteration zone. This morphology, characteristic of the Snow Lake area, seems incompatible with a simple flow-through hydrothermal convective-cell model (Heaton and Sheppard, 1977; Spooner, 1977). An episodic or single-event model, such as that proposed by Lydon and Jamieson (1984) for the Cyprus deposits, involving the thermochemical evolution and explusion of trapped fluids, may be more appropriate.

Some qualitative geochemical relationships can be inferred from the mineralogy of the alteration zones. The presence of chlorite, staurolite and magnetite in the distal alteration zone would indicate enrichment in Mg, Fe and Al, and depletion in Na and Ca. The calc-silicate envelope, particularly the pervasive epidote, indicates high Ca and Fe^{+3} contents. The proximal alteration zone is enriched in K, Fe and

Al, and depleted in Na and Ca, as shown by the presence of muscovite, staurolite and kyanite. Quartz veins in both alteration zones may contain Si derived from the host rock.

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Geology of the Chapleau, Groundhog River and Val Rita blocks, Kapuskasing area, Ontario.[1]

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Leclair, A.D. and Nagerl, P., Geology of the Chapleau, Groundhog River and Val Rita blocks, Kapuskasing area, Ontario; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 83-91, 1988.

Abstract

High grade rocks of the Groundhog River and northern Chapleau blocks consist predominantly of northeast-striking belts of mafic gneiss, paragneiss and tonalite gneiss. They are separated from amphibolite facies rocks of the Wawa and Abitibi belts by major north-northeast-trending faults. A major zone of cataclasis and faulting, the Saganash Lake fault, coincides with an aeromagnetic lineament and juxtaposes granulite gneiss of the Groundhog River block against metavolcanic rocks, granite and orthogneiss of the Val Rita block to the west. Tonalite gneiss containing mafic xenoliths occurs on both sides of the northward projection of the fault trace. In the Wakusimi River area, the positive aeromagnetic anomaly of the Groundhog River block fades towards the northeast, and mafic and tonalitic gneiss are in apparent fault contact with massive to foliated granodiorite of the Abitibi Belt to the east. A northeast-trending tholeiitic dyke of the Preissac swarm is cut by brittle faults associated with the Kapuskasing structure.

Résumé

Les roches fortement métamorphisées des blocs de la rivière Groundhog et du nord de Chapleau consistent surtout en zones de gneiss mafique, paragneiss et gneiss à tonalite orientées vers le nord-est. Elles sont séparées des roches à faciès des amphibolites des zones d'Wawa et de Abitibi par des failles majeures d'orientation nord-nord-est. Une zone majeure de cataclase et failles, la faille Saganash Lake, coïncide avec un alignement aéromagnétique et juxtapose le gneiss à granulite du bloc de la rivière Groundhog contre les roches métavolcaniques, granite et orthogneiss du bloc de Val Rita à l'ouest. Des gneiss à tonalite contenant des xénolites mafiques se trouvent de chaque côté de la projection nord du tracé de la faille. Aux alentours de la rivière Wakusimi, l'anomalie aéromagnétique positive du bloc de la rivière Groundhog disparaît progressivement vers le nord-est, et les gneis mafiques et tonalitiques sont mis en contact de faille avec du granodiorite dont la structure varie de massive à foliée de la zone d'Abitibi à l'est. Un dyke tholéitique de l'essaim de Preissac orienté vers le nord-est est recoupé par des failles associées à la structure de Kapuskasing.

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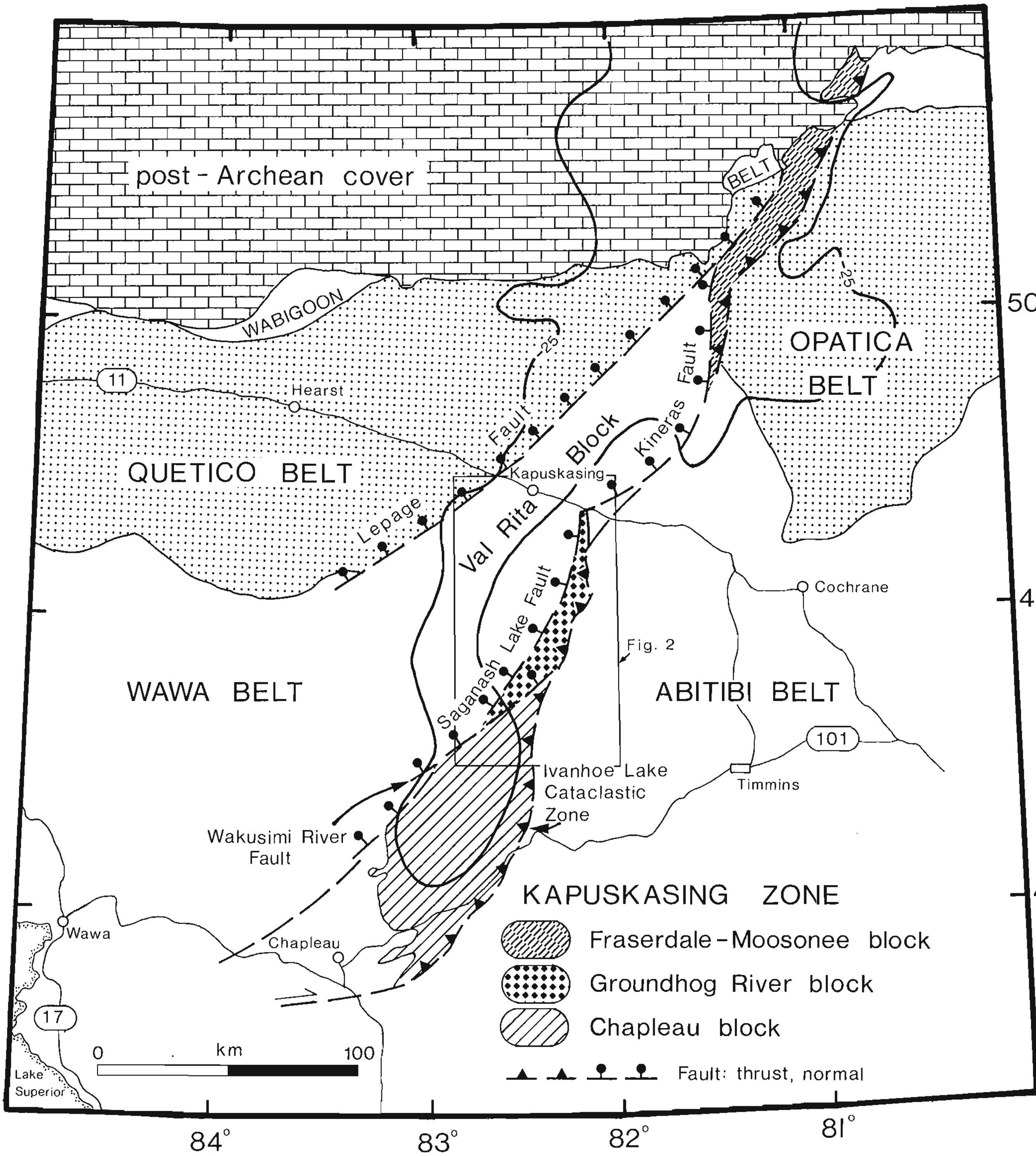

Figure 1: Regional tectonic elements of the central Superior Province showing distribution of fault-bounded tectonic blocks of the Kapuskasing uplift. The regional gravity high is outlined by the -25 mGal Bouguer anomaly contour. Adapted from Percival and McGrath (1986).

INTRODUCTION

The Kapuskasing structure is an elongate northeast-trending zone of heterogeneous high grade metamorphic rocks and geophysical anomalies that transects the east-west metavolcanic- and metasedimentary-plutonic belts of the central Superior Province (Thurston et al., 1977; Gibb, 1978; Card, 1979). It is made up of the Chapleau, Groundhog River and Fraserdale-Moosonee blocks (Fig. 1); each with distinct geological and geophysical character and, for the most part, bounded by a thrust fault on the east and late normal faults on the west (Percival, 1985; Percival and McGrath, 1986). Comprehensive assessment of regional geological data, in conjunction with gravity modelling, have provided an interpretation of the Chapleau block as the basal part of a crustal-scale slab of Archean crust that was uplifted along a northwest-dipping thrust fault, expressed at the surface as the Ivanhoe Lake cataclastic zone (Percival and Card, 1983, 1985). In the northern Kapuskasing structural zone, a recent study by Percival and McGrath (1986) suggested that the Groundhog River and Fraserdale-Moosonee blocks are "perched thrust tips", based on geophysical interpretation and limited geological information. Further mapping is needed in this area to subject this hypothesis to rigorous geological tests.

The region south of Kapuskasing, including the Groundhog River, Val Rita and northern Chapleau blocks, represents one of the largest gaps where many problems of regional significance cannot be resolved from the existing reconaissance-scale mapping. For example, the gravity high which coincides with high grade rocks of the Chapleau and Fraserdale-Moosonee blocks extends into the Val Rita block of the Wawa Belt, up to 40 km west of the Groundhog River block. This puzzling relationship raises questions regarding the nature of the Groundhog River and Val Rita blocks and their role in the Kapuskasing uplift. As an improved geological data base is required to address these and other critical questions, a program of 1:100 000 mapping was initiated in 1987 with the purpose of gaining an understanding of the character, evolution and interrelationships of the Groundhog River, Val Rita and northern Chapleau blocks, and western Abitibi Belt. This report, based on three months of fieldwork in parts of the Kapuskasing (NTS 42G) and Foleyet (NTS 42B) map areas, mainly outlines the lithology, internal structure and general metamorphic conditions of the various fault-bounded blocks, and discusses their boundary zones.

Within the regional framework of this project, the tectonometamorphic history of the tectonic blocks and the geochronology of their boundary faults will be studied by Paul Nagerl as part of a graduate thesis. Also, an isotopic $^{40}Ar/^{39}Ar$ study of the Cargill carbonatite complex by Steven Butler will form the basis of a B.Sc. thesis at Queen's University.

ABITIBI BELT

In the area of the Groundhog River, the westernmost Abitibi Belt is dominated by metavolcanic rocks of the Belford-Strachan belt, homogeneous granodiorite and hornblende-biotite tonalite gneiss (Fig. 2). This lithological package resembles other greenstone-granite terranes farther east and south in the Abitibi Belt (cf. Goodwin and Ridler, 1970).

The Belford-Strachan belt consists mainly of mafic to intermediate metavolcanic flows and layered amphibolite, with a few discrete units of pyroclastic felsic metavolcanic rocks, and minor intercalated metasediments of volcanogenic affinities (Bennett, 1969). It is intruded by an arcuate, convex to the south, body of magnetite-rich gabbro and associated dioritic and ultramafic rocks that causes a positive aeromagnetic anomaly. Exposures on the Groundhog River indicate that the gabbro and diorite are generally medium grained, homogeneous and massive.

The remainder of the Abitibi Belt in the vicinity of the Groundhog River is underlain by felsic plutonic rocks of variable compositional and structural characteristics, even at outcrop scale. The metavolcanic rocks of the Belford-Strachan belt and associated gabbro are intruded to the east by massive, medium grained trondhjemite and quartz monzonite (Bennett, 1969). Farther north, massive to foliated, locally gneissic, leucocratic hornblende-biotite $\pm$ epidote granodiorite and minor tonalite (or trondhjemite?) are medium- to coarse-grained, sparsely xenolithic and contain up to about 20 % discordant pegmatite. They are separated from a more heterogeneous unit of weakly foliated to gneissic tonalite and granodiorite by a gradational contact and the Nansen Creek fault. The most abundant rock type is hornblende-biotite tonalite; however, biotite granodiorite, biotite tonalite and mafic hornblende tonalite to diorite are also common. Generally these rocks are medium grained and leucocratic and contain less than 2 % mafic xenoliths. Xenoliths are mainly hornblende-plagioclase amphibolite with local clinopyroxene; however, some are ultramafic (hornblendite). Also present within this unit are migmatitic biotite-plagioclase-quartz $\pm$ hornblende gneiss of uncertain parentage.

Planar structures in the felsic plutonic rocks are defined by compositional layering and alignment of biotite and/or hornblende. They have an overall chaotic orientation although, north to northwest strikes and moderate dips to the northeast or east are common. The dip tends to steepen close to the Nansen Creek fault. Rare lineations, defined by axes of small folds in gneiss, plunge shallowly to the northeast or east. In metavolcanic rocks the foliation strikes north-northeast and dips steeply to the west, and a mineral lineation plunges steeply to the south (Bennett, 1969).

Metamorphic grade in the westernmost Abitibi Belt ranges from greenschist to amphibolite facies. Except for rare pillow structures, the regional metamorphism has obliterated any original features in volcanic and sedimentary rocks of the Belford-Strachan belt. The lowest grade mafic metavolcanics consist of plagioclase and clinopyroxene, with common chlorite and epidote; whereas higher grade metavolcanics are composed mainly of hornblende and plagioclase (Bennett, 1969; Thurston et al., 1977). To the north, mafic inclusions in felsic plutonic rocks are composed of amphibole-plagioclase $\pm$ clinopyroxene. Epidote is common in these rocks as a replacement of mafic minerals and late veinlets.

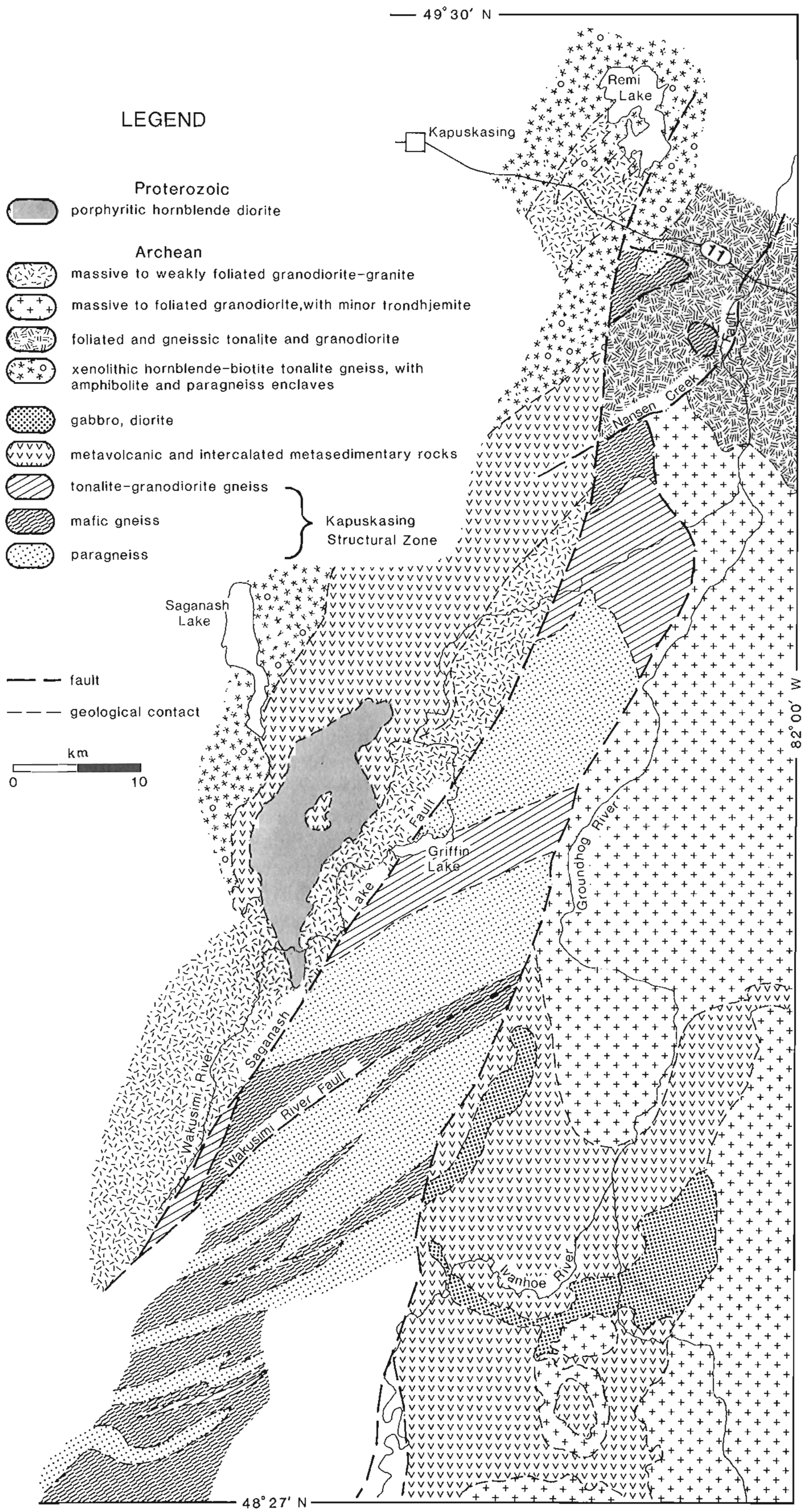

Figure 2: Generalized geological map of the Kapuskasing area including the southeastern part of Kapuskasing (NTS 42G) and the northeastern part of Foleyet (NTS 42B) map areas (modified after Bennett et al., 1967, and Thurston et al., 1977).

VAL RITA BLOCK

The Val Rita block of the Wawa Belt is defined as a normal-fault-bounded wedge consisting almost entirely of amphibolite facies tonalitic gneiss and granite (Percival and McGrath, 1986) with similar lithological and structural characteristics to the Wawa gneiss terrane (cf. Percival and Card, 1985) to the south. The easternmost part of the Val Rita block is underlain by four major north-northeast-trending units: 1) metavolcanic and associated metasedimentary rocks of the Saganash Lake belt, 2) xenolithic hornblende-biotite tonalite gneiss with amphibolite and paragneiss enclaves, 3) massive to weakly foliated granodiorite to granite and 4) porphyritic hornblende diorite (Fig. 2).

Mafic to intermediate metavolcanic and intercalated metasedimentary rocks of the Saganash Lake belt represent the largest concentration of supracrustal rocks in the Val Rita block. They form an arcuate, convex to the west, belt which terminates against the Saganash Lake fault to the north. Most exposures occur between Saganash Lake and Wakusimi River and consist dominantly of fine- to medium-grained, massive to foliated amphibolite interpreted as the metamorphic derivative of mafic to intermediate flows (e.g. McMurchy, 1960; Thurston et al., 1977). Garnet porphyroblasts are present in some exposures. Sparse primary features include pillow structure and tuffaceous beds of mafic to intermediate composition. Intermediate to felsic pyroclastic breccias with subangular to lensoidal porphyritic clasts have been observed at two localities. Intercalated metasedimentary rocks in the volcanic assemblage comprise iron-formation and minor metagreywacke. In addition, the existence of quartzite and arkose was reported by McMurchy (1960) and Thurston et al. (1977). The iron-formation consists of thinly interbedded iron minerals and quartz and is responsible for a strong anomaly on the aeromagnetic map.

The hornblende-biotite tonalite of the Val Rita block is characterized by generally well developed gneissic layering and the presence of rounded mafic xenoliths. It may also include granodiorite gneiss and is injected by concordant and discordant light grey to pink granite, pegmatite and aplite. Mafic xenoliths commonly make up at least 5 % of individual outcrops and are composed of hornblende and plagioclase, with clinopyroxene as a common constituent. Their internal structures are locally discordant to layering in enclosing gneiss. Garnet amphibolite of possibly volcanic origin and rare paragneiss occur as discrete narrow units in the orthogneiss. The paragneiss is a garnet-biotite-plagioclase-quartz rock. Biotite-plagioclase-quartz $\pm$ hornblende gneiss with 10 to 25 % tonalitic and pegmatitic leucosome, and mafic hornblende-bearing tonalite to diorite are also present.

In the Remi Lake area xenolithic tonalite gneiss occurs on both sides of the Saganash Lake fault. These rocks resemble the relatively xenolith-poor unit of weakly foliated to gneissic tonalite and granodiorite of the Abitibi Belt to the east and south. Although a structural discontiniuty that matches the Kineras fault has yet to be found north of Highway 11, these observations suggest the existence of similar lithological packages to both the east and west.

Massive to weakly foliated, leucocratic biotite $\pm$ hornblende and biotite $\pm$ epidote $\pm$ magnetite granodiorite to granite are exposed southwest of Remi Lake and west along the Saganash Lake fault. These white to pink granitic rocks and associated pegmatite are probably younger than the tonalite gneiss. They are relatively homogeneous and xenolith-free, with less than 1 % hornblende-plagioclase inclusions in a few outcrops. A local planar fabric, defined by the alignment of mafic minerals, is concordant to the gneissosity in adjacent gneiss.

A dioritic pluton intrudes both metavolcanic rocks of the Saganash Lake belt and their surrounding granitic rocks. The dominant phase is a medium- to coarse-grained, massive hornblende diorite or quartz diorite which commonly displays plagioclase phenocrysts. Mineralogically, the rock comprises hornblende, biotite, epidote, plagioclase, sphene and rare quartz. A Proterozoic age for the diorite was reported by Thurston et al. (1977).

Planar structures in the eastern Val Rita block generally strike north-northeasterly with variable dips to the east or west. They consist of compositional layering in gneiss and a foliation defined by alignment of biotite and hornblende. Fabrics in metavolcanic and metasedimentary rocks of the Saganash Lake belt are steeply dipping; pillow top determinations indicate that the volcanic assemblage is steeply overturned to the west, at least locally. A common mineral lineation plunges 35 to 75° to the north or northeast. The foliation and gneissossity in adjacent gneiss to the west dip 15 to 45° towards the east-southeast and are locally subhorizontal. Recumbent isoclinal folds with shallowly northeast plunging axes occur in paragneiss on the east shore of Saganash Lake. Planar structures in the Remi Lake area have generally consistent northeast strike and moderate northwest dips; but, locally they form chaotic and swirly patterns.

Metamorphic grade in the eastern Val Rita block is generally amphibolite facies. The assemblage hornblende-plagioclase, with or without garnet, is ubiquitous in mafic metavolcanic rocks of the Saganash Lake belt. The same assemblage, with or without clinopyroxene, characterizes mafic enclaves in tonalite gneiss. Paragneiss consists of garnet-biotite-plagioclase-quartz $\pm$ hornblende. On the south shore of Remi Lake a narrow unit of mafic gneiss flanked by granitic rocks contains garnet-clinopyroxene-hornblende-plagioclase-quartz.

GROUNDHOG RIVER BLOCK

The Groundhog River block is the most poorly exposed of the three tectonic segments of the Kapuskasing structural zone. Insofar as exposure indicates, it is made up of three major lithological components: 1) mafic gneiss, 2) tonalite gneiss and 3) paragneiss (Bennett et al., 1967; Thurston et al., 1977; Percival, 1985).

Mafic gneiss predominates in the northern part of the Groundhog River block. It is characterized by prominent compositional layers up to 15 cm thick made up of varying proportions of hornblende and plagioclase with combinations of garnet, clinopyroxene and quartz. Concordant quartz lenses and tonalitic leucosome veinlets constitute less than 5 % of many outcrops. South of the Nansen Creek fault the mafic gneiss is associated with homogeneous, foliated metagabbro composed of hornblende-plagioclase ± clinopyroxene. Migmatitic metagabbro contains 10 to 30 % hornblende- and locally clinopyroxene-bearing tonalitic leucosome. Immediately south of Highway 11, a tongue of mafic gneiss with local garnet-biotite metasedimentary gneiss represents the most northerly exposure of high-grade rocks of the Groundhog River block (Fig. 2). It is interpreted to be in fault contact with tonalite gneiss and granitic rocks of the Wawa and Abitibi belts. To the southeast, an isolated outlier of shallowly dipping mafic gneiss forms a ridge apparently surrounded by granitic rocks. Garnetiferous mafic gneiss occurs at the southwestern tip of the Groundhog River block.

The Groundhog River block is underlain by two main units of tonalite to granodiorite gneiss. These rocks locally contain elongated and rounded mafic xenoliths composed of hornblende-clinopyroxene-plagioclase ± biotite. Concordant pink pegmatite layers and pods are ubiquitous. The westernmost exposures of tonalite to mafic tonalite gneiss commonly contain orthopyroxene in addition to hornblende, biotite and clinopyroxene. They are intruded by mafic dykes and show brittle deformation. On the north shore of Griffin Lake tonalite gneiss and flaser-textured granodiorite are in sharp contact with 2 pyroxene granulite gneiss. In the Wakusimi River area, tonalite and granodiorite resemble adjacent rocks of the Abitibi Belt to the east. This makes the boundary between the Groundhog River block and Abitibi Belt difficult to define.

A few scattered outcrops of paragneiss occur in the Groundhog River block south of Wakusimi River. The paragneiss is compositionally layered on the 1 to 10 cm scale, with variable proportions of clinopyroxene, orthopyroxene, biotite, plagioclase and quartz. It is typically light brownish green and quartz-rich. Coarse grained orthopyroxene-bearing leucosome with quartz-plagioclase-magnetite ± biotite-K-feldspar(?) occur in the form of layers, pods and lenses which constitute up to 25 % of the rock.

The orientation of gneissosity and foliation varies along the length of the Groundhog River block. South of Highway 11, gneissic layering in mafic gneiss strikes northeasterly and dips at 45 to 75° to the northwest. In the Wakusimi River area, planar structures generally have north-northwest to north-northeast trends with steep dips to the east. Farther south, they strike northeasterly again, with moderate northwesterly dips.

CHAPLEAU BLOCK

The Chapleau block is characterized by linear northeast-trending belts of paragneiss, mafic gneiss, tonalite gneiss, foliated diorite to mafic tonalite and anorthosite-suite rocks (Thurston et al., 1977; Percival and Card, 1983, 1985). Under consideration is the northernmost Chapleau block where the dominant lithologies are paragneiss and mafic gneiss, with minor tonalite gneiss and anorthosite. These rocks form northeast-striking elongate units that are interfingered on a decametre to kilometre scale; complex interfingering on the metre scale is also common.

Paragneiss of the Chapleau block is migmatitic, layered, medium grained, garnetiferous biotite-plagioclase-quartz rock with some orthopyroxene. It has an overall quartz-rich composition and commonly includes 10 to 40 % concordant tonalite leucosome within which orthopyroxene is common. Melanosome consists predominantly of assemblages of garnet-biotite-plagioclase-quartz ± orthopyroxene and rare garnet-biotite-kyanite-plagioclase-quartz-K-feldspar. Clinopyroxene and hornblende are also present in some layers. Leucosome is plagioclase-quartz ± K-feldspar(?) with some orthopyroxene, garnet or biotite. Garnet porphyroblasts are commonly pink and constitute up to about 25 % of the rock. Paragneiss units typically weather yellowish to reddish brown and those at the north end of the Chapleau block appear more severly altered.

Dark green, medium- to coarse-grained, migmatitic, garnetiferous mafic gneiss represents the second largest lithological component of the northern Chapleau block, after the paragneiss. It is composed mainly of garnet, clinopyroxene, hornblende, plagioclase and quartz, producing 1 to 15 cm thick layers with differences in modal proportions. It contains 5 to 35 % concordant coarse tonalitic leucosome which includes clinopyroxene, hornblende, garnet or rare orthopyroxene. Deep red garnet porphyroblasts are almost ubiquitous, constituting up to about 20 % of the rock. Foliated, homogeneous mafic gneiss is present locally. Rare enclaves and layers with composition similar to the mafic gneiss occur within some paragneiss units.

Subordinate rock types in the northern Chapleau block include strongly foliated to gneissic, leucocratic hornblende-biotite tonalite to granodiorite and coarse gabbroic anorthosite. Only one mappable unit of tonalite-granodiorite occurs within the paragneiss/mafic gneiss sequence. The same rock type also occurs as narrow discrete units intercalated with paragneiss and mafic gneiss. So far as is known, the gabbroic anorthosite (not shown in Fig. 2) is restricted to a few shoreline exposures. It contains plagioclase, hornblende, garnet and biotite, and is cut by a mafic dyke striking 063°.

The prominent northeast structural grain of the northern Chapleau block is produced by the orientation of lithological contacts, compositional and migmatitic layering, and mineral alignment. Planar structures strike northeast and dip northwest. Locally, the gneissic fabric is isoclinally folded by minor southeast-verging folds. A mineral lineation defined by quartz streaks trends east-northeast with subhorizontal plunge to the northeast or southwest.

BOUNDARY ZONES OF TECTONIC BLOCKS

The three tectonic blocks described above have been defined based on their overall geological and geophysical characteristics (Percival, 1985; Percival and McGrath, 1986). The major northeast-trending structures that bound these blocks (Fig. 1) are recognized from aeromagnetic lineaments and from juxtaposition of rocks of contrasting lithological, structural and metamorphic character. Fieldwork has revealed that boundaries of tectonic blocks usually coincide with extensive zones of cataclasis and faulting. These zones, approximately 1 km wide, are characterized by impressive networks of brittle faults, discontinuous veinlets and pods of cataclasite to pseudotachylyte, and epidote-filled fractures. They are common in high grade rocks along the western boundary of the Groundhog River block. On the outcrop, fault-rock veinlets have random orientation and crosscut one another, and brittle faults commonly display both sinistral and dextral offsets with dip-slip components. A detailed study would be required to unravel the strain history and to determine if more than one event of brittle deformation has occurred.

A distinct positive aeromagnetic anomaly, which fades out near Highway 11 in the north, outlines the general extent of the Groundhog River block (Geological Survey of Canada, 1984). The prominent aeromagnetic lineaments in the boundary zone of this block correspond to major fault zones (Percival and McGrath, 1986; Fig. 1), relatively well exposed in a few localities.

The Saganash Lake fault, which juxtaposes granulites of the Groundhog River block against amphibolite-facies rocks of the Val Rita block to the west, is marked by conspicuous cataclastic zones at several localities along its length. For example, orthopyroxene-bearing tonalite and crosscutting mafic dykes, exposed on a discontinuous north-trending ridge in the Wakusimi River area, are riddled with pseudotachylyte veinlets and brittle faults. They display a westward increase in the percentage of pseudotachylyte, becoming cataclasites with unrecognizable protoliths near the recessive-weathering Saganash Lake fault to the west. The fault loses its aeromagnetic expression northward and xenolithic tonalite gneiss appears to extend across the northern extension of the fault trace in the Remi Lake area. Rare pseudotachylyte veinlets occur in tonalitic gneiss exposed in a roadcut just south of the lake. An identical situation south along the fault has prompted Percival and McGrath (1986) to suggest that displacement is minimal in the Chapleau area. A working hypothesis is that the Saganash Lake fault is a spoon-shaped listric normal fault with maximum displacement near Griffin Lake that progressively decreases both to the north and south.

The Ivanhoe Lake cataclastic zone is a zone of cataclasis and faulting that separates high grade rocks of the Chapleau block from low grade metavolcanic and granitic rocks of the Abitibi Belt to the east (Percival and Card, 1983, 1985). It coincides with a linear positive aeromagnetic anomaly that extends north-northeasterly as the eastern boundary of the northern Chapleau block and Groundhog River block. Bennett (1969) noted cataclastic features in the Ivanhoe River area and suggested that a major northeast-trending fault zone marks the boundary between greenschist- to amphibolite-facies metavolcanic rocks of the Belford-Strachan belt and granulite gneiss of the Kapuskasing zone. In the north, for about 35 km, the geophysically-defined eastern boundary of the Groundhog River block is not exposed, and thus no geological evidence of a fault zone is available to support the geophysical interpretation.

In the Wakusimi River area, this boundary is not well defined either geophysically or geologically. The central positive anomaly fades towards the east and northeast through a series of steps. The boundary is also not readily defined on the basis of the local geology as no obvious disparity in structural trend or lithology could be detected. A contact interpreted as a fault, based on sparse cataclastic effects, separates commonly gneissic tonalite from generally massive to foliated granodiorite to the east. This contact could correspond to the eastern boundary zone of the Groundhog River block. South of the Nansen Creek fault, mafic gneiss of the Groundhog River block is in contact with foliated leucogranodiorite to the east.

The concave eastern contact of mafic gneiss south Highway 11 is interpreted as a fault based on local mylonitic and cataclastic effects. Mafic gneiss on the east side of the body occurs as recrystallized, communited, fine grained, flinty rock with subtle compositional layering, offset by small brittle faults. Thin, elongated quartz streaks parallel the main fabric. Farther south, an outcrop of mafic gneiss and crosscutting diabase dyke contains pseudotachylyte and several sets of brittle faults. Locally, adjacent tonalite gneiss is fractured and contains pseudotachylyte-type veinlets.

The boundary zone between the Groundhog River and Chapleau blocks is defined as the Wakusimi River fault, a poorly exposed structure which coincides with a strong northeast-trending linear aeromagnetic anomaly. Meager geological data indicate that the fault mainly juxtaposes garnetiferous mafic gneiss to the north against garnet-biotite metasedimentary gneiss to the south. Cataclastic veinlets are present at one locality near the locus of the fault.

CORE SAMPLES

A drilling program was conducted in 1987 by the Terrain Sciences Division of the Geological Survey of Canada to study Pleistocene channel deposits in the Kapuskasing area. The procedure involved drilling through the thick glacial overburden and into the underlying bedrock. The recovered core samples of bedrock were made available for rock identification, courtesy of S.L. Smith. They provide valuable geological information in a region where exposure is minimal. Locations of boreholes are shown in Figure 3. The rock type and depth of each core sample is listed in Table 1.

An interesting discovery is the mylonite recovered at site 18. A possible protolith could be the metavolcanic rock of the Saganash Lake belt. The structural fabric in the core sample suggests a steeply dipping shear zone, the strike of which is uncertain.

MAFIC DYKES

Fresh, fine- to medium-grained, black weathering mafic dykes are present throughout the map area. Generally, they are more resistant to erosion than the country rocks and thus are the only rock type exposed at many localities. Three distinct swarms of diabase dykes are recognized on the basis of orientation. 1) The most common are north- to north-northwest-trending dykes, 30cm to 50m wide, which may belong to the late Archean Matachewan swarm. They cut rocks of the Groundhog River block and adjacent Wawa and Abitibi belts and are crosscut by brittle faults and pseudotachylyte veins. Dips determined on some of these dykes are vertical to steeply westward. 2) Some 010 to 020° dykes, perhaps also part of the Matachewan swarm, occur north of the Wakusimi River.

Post-intrusion re-orientation may explain the deviation from the common trend of the swarm. 3) A few northeast-trending dykes, at least 1m wide and with probable olivine, transect high grade rocks of the northern Chapleau block. These could be members of the Abitibi or Kapuskasing swarms. The definitive assignment of the dykes to major swarms must await further work.

A quartz diabase dyke, cutting foliated hornblende leucogranodiorite near the northernmost mafic gneiss unit of the Groundhog River block, is injected by cataclastic veinlets and cut by brittle faults associated with the Kapuskasing structure. The trend of the dyke is approximately 055° although the margin is offset by numerous small faults. Petrographically, the diabase displays subophitic texture and is composed of quartz (2-5 %), plagioclase (50-60 %), augite (35-40 %) and opaques (< 5 %) with minor apatite and biotite. The plagioclase is slightly saussuritized and augite is altered to green amphibole. These characteristics, including orientation and petrography, match those of tholeiitic dykes of the Preissac swarm (K.D. Card, pers. comm., 1987). These dykes have yielded ages of about 2150 Ma (Gates and Hurley, 1973; Hanes and York, 1979). A radiometric date on this dyke could impose important constraints on the age of exhumation of the Kapuskasing zone.

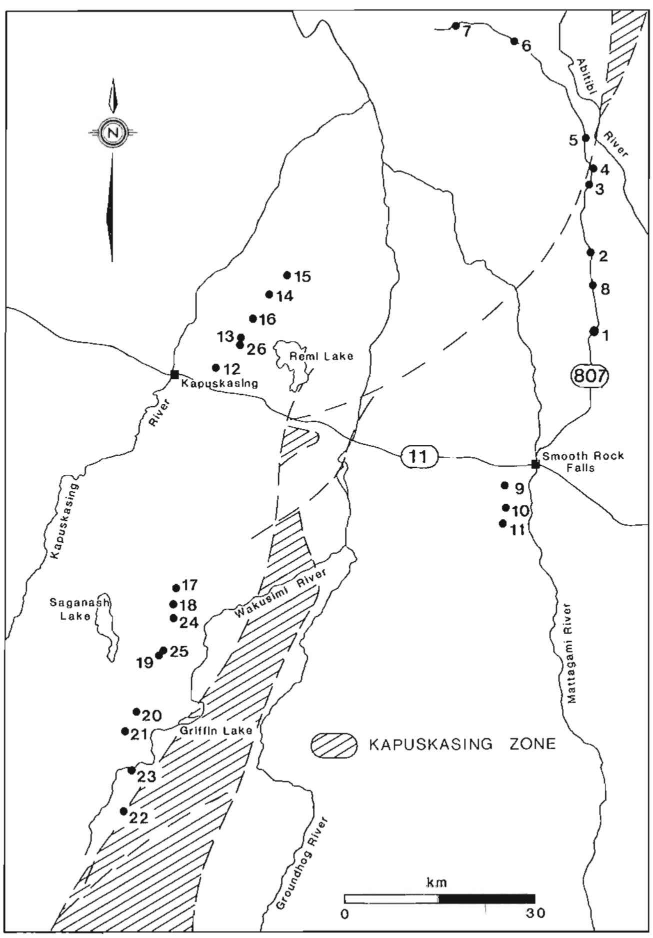

Figure 3: Locations of boreholes. See Table 1 for description of core samples. No basement rock was recovered at sites 6, 15 and 16.

Table 1: Borehole data

Core Sample	Depth(m)	Rock type
1	40.8-41.1	hornblende-biotite granodiorite
2	45.4-46.0	garnet-biotite metasedimentary gneiss
3	39.3-39.6	garnet-biotite metasedimentary gneiss
4	36.6-37.2	metavolcanic (?) fault rock
5	35.7-40.0	biotite tonalite gneiss
7	83.8-84.1	biotite tonalite gneiss
8	40.8-41.1	hornblende-biotite diorite gneiss
9	13.4-13.7	biotite tonalite gneiss
10	6.1-6.4	hornblende-biotite granodiorite
11	41.9-42.1	syenite
12	29.3-30.8	hornblende tonalite gneiss
13	35.1-36.6	hornblende-biotite granodiorite gneiss
14	12.2-12.3	hornblende-biotite granodiorite gneiss
17	33.5-35.1	mafic gneiss
18	35.4-36.9	mylonite
19	37.5-38.1	saprolite
20	25.6-25.8	porphyritic hornblende diorite
21	26.8-27.1	porphyritic hornblende diorite
22	53.0-53.3	gabbro
23	33.8-34.1	biotite granite
24	8.5-10.4	iron-formation (sheared)
25	24.1-28.6	intermediate gneiss
26	29.9-30.2	hornblende granite

MINERAL OCCURRENCES

Mineral occurrences noted in the course of mapping include: graphite in paragneiss of the northern Chapleau block, oxide/sulphide facies iron-formation in metavolcanic rocks of the Saganash Lake belt, phosphate and sulphide in the Cargill carbonatite intrusion, and disseminated sulphides at scattered localities in metavolcanic rocks, mafic gneiss and paragneiss. A drilling program by Noranda Exploration Co. Ltd. in the northern Chapleau block has revealed some anomalous gold values in sulphide-rich rocks. Exploration programs are currently underway at both the iron-formation and Cargill carbonatite intrusion.

DISCUSSION

The occurrence of xenolithic hornblende-biotite tonalite gneiss and granitic rocks on both sides of the Saganash Lake fault implies that displacement is minimal in the Remi Lake area. The presence of similar lithological packages to both the east and west supports the contention that the Wawa and Abitibi belts are part of a once continuous belt. It would be interesting to find out if radiometric ages of granitoid rocks can be correlated across the transverse Kapuskasing structural zone.

The Groundhog River block is a unique tectonic element of the Kapuskasing structural zone, in that its dense granulites are not associated with a positive gravity anomaly. Although it is the apparent northern extension of the Chapleau block, the Groundhog River block has different characteristics such as: 1) no gravity anomaly; 2) a stronger aeromagnetic expression; 3) the presence of northerly trending dykes of the Matachewan swarm; 4) the lack anorthosite-suite rocks; 5) the rocks have an apparently higher percentage of magnetite and are less migmatitic; and 6) veinlets of pseudotachylyte occur sporadically throughout the block. Some of these geological and geophysical disparities possibly indicate that granulites of different crustal levels (and thickness) have been carried to the surface, in a manner described by Percival and McGrath (1986). Thus, it is crucial to assess P-T-aH_2O conditions in granulites of both blocks. Combined with geothermobarometry on adjacent granitoid rocks of the Val Rita block and Abitibi Belt, this would provide estimates on the amount of vertical displacement along boundary faults.

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Structure of the Wawa gneiss terrane near Chapleau, Ontario

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Abstract

Orthogneisses in the Wawa gneiss terrane adjacent to the southern Kapuskasing structural zone have a complex geological history. At least five periods of deformation are recognized, the third of which involved pervasive regional extension, manifest as packages of extensional "lozenges" and small-scale, listric, ductile, normal faults. Extensional features are best developed in a 60 x 20 km belt extending westward from south of Chapleau.

Résumé

Les orthogneiss du terrain gneissique de Wawa, adjacent à la partie sud de la zone structurale de Kapuskasing, ont une histoire géologique complexe. On reconnait au moins cinq périodes de déformation, dont la troisième implique un épisode d'extension pénétrante à caractère régional, dont les résultats se manifestent sous forme de paquets de « losanges » d'extension et de failles listriques normales et ductiles de petite échelle. Les structures d'extension sont mieux développées dans une zone de 60 × 20 km qui s'étend du sud de Chapleau vers l'ouest.

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INTRODUCTION

This paper reports field observations and preliminary interpretations of the structural characteristics and history of the Wawa gneiss terrane in the Chapleau area of Ontario.

Located in the Superior Structural Province, the amphibolite facies Wawa gneiss terrane occurs between the low-grade Michipicoten greenstone belt to the west and high-grade rocks of the Kapuskasing Structural Zone to the east (Percival and Card, 1985). It is an Archean complex of regionally metamorphosed tonalitic and mafic orthogneisses thought to have crystallized at mid-crustal levels (500-600 MPa (5-6 kb)) (Percival, 1986a). In this area (Fig. 1), the Wawa -Kapuskasing boundary is a complex, gradational transition zone of lithological, structural, metamorphic (Percival, 1983), density and seismic velocity changes, thought to be analogous to the Conrad discontinuity (Percival, 1986b; Percival and Fountain, in press). Thurston et al.(1977) and Percival (1981) conducted reconnaissance mapping in the region. However the structural history of the complexly deformed gneisses and its correlation with that in adjacent terranes is poorly understood.

Recently, Percival (1987) observed horizontal high-strain gneisses in roadcuts southwest of Chapleau which he attributed to extensional deformation at mid-crustal levels. Such structures have been postulated to be the cause of prominent mid- to lower-crustal reflections observed in many seismic reflection profiles (e.g. Oliver et al., 1983; Brodie and Rutter, 1987). Hence, an understanding of the surface distribution, geometry, chronology and kinematics of these structures is crucial to the interpretation of the forthcoming Kapuskasing LITHOPROBE seismic reflection profiles.

Figure 1. Arrow indicates location of study area in the Wawa gneiss terrane at the southern end of the Kapuskasing Structural Zone (KSZ).

LITHOLOGY

The Wawa gneiss terrane in the Chapleau area consists dominantly of tonalitic orthogneisses and plutons of xenolithic tonalite. Included within these is an older suite of mafic gneisses. The compositional and textural character of each suite is detailed below.

Mafic gneiss

This suite is made up almost entirely of fine grained hornblende-plagioclase-biotite-quartz gneiss. Rare garnet and clinopyroxene are also present in the Nagasin Lake area. The rock is dark grey-green on the weathered surface and has a gneissosity defined by 3mm thick quartz-plagioclase layers. The average spacing of leucosome segregations is 4 cm and they are continuous at the 1 m scale.

Bodies of mafic gneiss generally occur as narrow discontinuous bands within tonalitic rocks at both the outcrop and map scales (Fig. 2). Rare units of fine grained diorite and medium grained metagabbro make up the remainder of the suite.

Tonalitic suite

Unlike the mafic gneisses, the field appearance of the tonalites is highly variable due to variations in composition and texture.

Composition varies according to the proportion and content of the mafic minerals which, in relative order of abundance, are hornblende, biotite, epidote and magnetite. Together these minerals rarely make up more than 15 % of the rock. Conversely there are no mafic-free tonalites and all four may co-exist in one rock. In general, biotite dominates in fine grained gneisses, whereas hornblende and epidote are prevalent in medium grained, xenolithic rocks. The latter relationship is due in part to secondary hornblende which is derived from disaggregated xenoliths of mafic gneiss. Magnetite content is generally higher (up to 2 %) in younger tonalite intrusive rocks and trace amounts of titanite occur in all tonalitic rocks.

Tonalitic rocks range in texture from strongly lineated, sugary and gneissose to massive, medium grained and homogeneous. They can be divided into two broad textural categories; gneissose and massive to foliated. The gneissic fabric appears to result from either deformation of inherited inhomogeneities (xenoliths, pegmatite veins) or segregation/injection of leucocratic layering. On this basis the tonalitic suite was subdivided into the four units which, with the other map units, are described in detail below.

Heterogeneous tonalite gneiss - a white/grey-weathering, fine grained gneiss with layering composed of 2 to 10 cm thick bands of deformed xenolith and/or intrusive phases. Layering is irregular, spaced at a 5 to 50 cm scale and continuous over 1 to 2 m.

Segregated tonalite gneiss - a grey and white banded, fine- to medium-grained-gneiss. Layering is regular at a 2 cm scale, continuous over 30 cm and defined by a lighter coloured, slightly coarser grained leucosome phase.

Xenolithic tonalite - a white and dark green-weathering rock which is massive to foliated with a mafic gneiss xenolith content between 5 % and 50 %. Xenoliths are blocky to rectangular, rounded or angular and have a size range of 5 cm to 1 m in maximum dimension.

Interlayered tonalite/mafic gneiss - a unit of both tonalitic and mafic rocks interlayered at a scale of 2 m or greater. This unit represents large-scale interlayering which is not at a mappable scale.

INTRUSIVE PHASES

Numerous phases of late granitoid intrusives occur in the Wawa gneiss terrane; two are present in Figure 2. The earlier phase, the Nebskwashi granite, is a xenolithic, fine- to medium-grained, massive to lineated hornblende granite with minor sulphides. A larger, more typical body of xenolithic, medium grained, massive to foliated granodiorite occurs near Chapleau.

REGIONAL STRUCTURAL TRENDS

Planar fabric elements strike predominantly west-northwest and dip steeply to moderately to the north-northeast. Anomalous, highly variable fabric orientations occur in the Nagasin - Highbrush Lake area. Linear elements have consistent orientation throughout the area, plunging subhorizontally at an average trend of 100°.

STRUCTURAL HISTORY

Field evidence suggests at least five deformational and seven intrusive events . The third deformational phase is the most complex and pervasive, rendering identification of early structures tenuous. The following outlines the tectonic and intrusive history; chronology is summarized in Table 1.

Deformational events:

D_1 – Produced gneissic layering in the oldest lithological unit, mafic gneiss.

D_2 – Formed tight to isoclinal minor folds of D_1 mafic gneiss layering.

D_3 – A regional, subhorizontal extensional event that produced shear-bounded lozenges and ductile normal faults (detailed description below).

D_4 – A transcurrent, ductile, west-northwest-striking, 3 km wide shear zone.

D_5 – Brittle faulting possibly related to the Saganash Lake fault (Percival and McGrath,1986).

Magmatic events:

I_1 – A period of repeated tonalite intrusion which field relationships show to be pre-D2 to post-D3.

I_2 – Early potassium feldspar-rich pegmatite dykes which now define gneissic layering in tonalites.

I_3 – Local evidence of deformed and metamorphosed mafic dykes which crosscut early phases of tonalite.

I_4 – Polyphase intrusion of granitic and granodioritic plutons.

I_5 – Intrusion of plagioclase-rich and potassium feldspar-rich pegmatite dykes.

I_6 – Hearst diabase dyke swarm (Ernst and Halls,1984).

I_7 – Kapuskasing diabase dyke swarm.

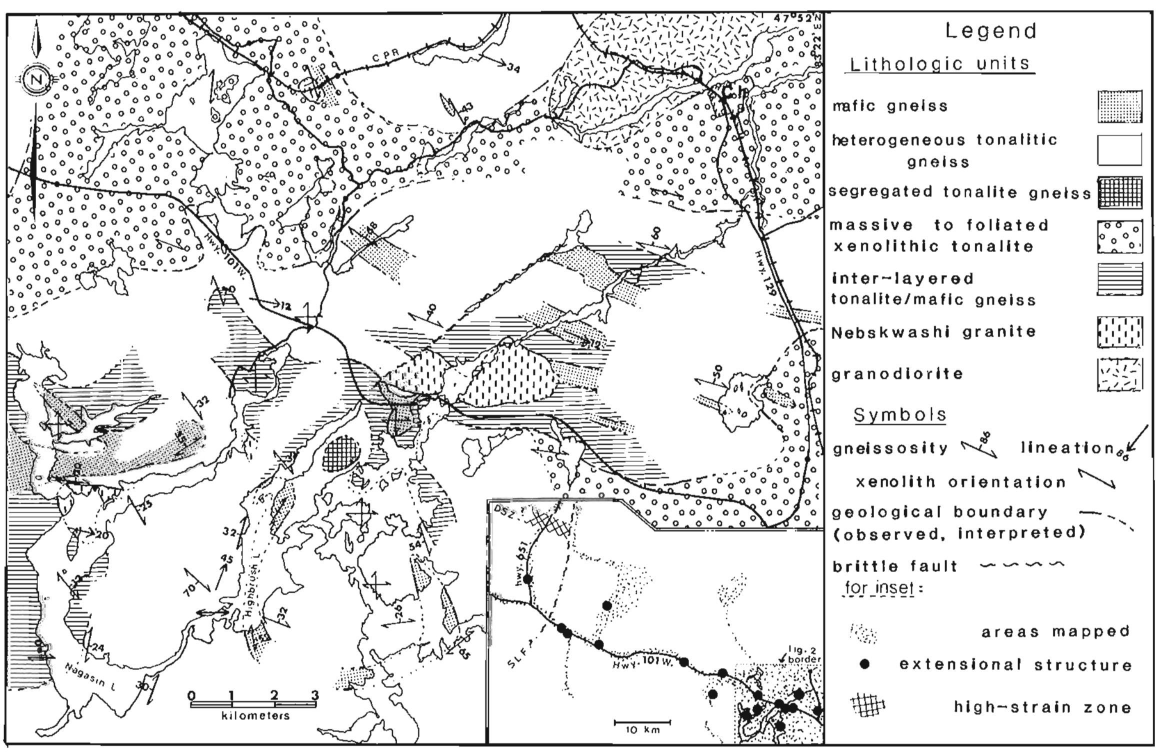


Figure 2. Geological map of a portion of the Wawa Gneiss Terrane southwest of Chapleau (Ch). Lower right inset shows distribution of extensional structures at a larger scale. (DSZ: Dalton Shear Zone; SLF: Saganash Lake Fault).

Table 1. Chronology of the tectonic and intrusive history.

Intrusive event	Deformation event	Chronologic field evidence
	D_1	Layering in mafic gneiss.
I_1 (begins)	D_2	Isoclinal folding in mafic gneiss with involved tonalite (cross-cut by extensional structures).
I_2, I_3		
I_1 (ends)	D_3	Extensional structures cross-cut by pegmatites, younger tonalites and mafic dykes.
I_4, I_5		
	D_4	Dalton shear zone deforms all pegmatites but is cross-cut by mafic dykes.
I_6, I_7		
	D_5	Brittle faulting and fracturing offsets mafic dykes.

D3: regional extension

Three different types of structures, all interpreted to result from ductile, subhorizontal extension, are present in the Wawa gneiss terrane in the Chapleau area. The geometry, kinematics, scale and lithological setting of each type is described below.

Lozenges

The term "lozenge" is used here to describe a three-dimensional, asymmetric lens of layered rock. Internal layers either mimic the elliptical shape of the lozenge (Fig. 4a, 4e) or have a low amplitude "s" or "z" asymmetry (Fig. 3c). Boundaries to these structures are curved, subhorizontal surfaces defined by low angle discontinuities in gneissic layering. Deflection of layering along these boundaries invariably indicates a normal sense of displacement. Such surfaces are commonly characterized by a slightly schistose fabric and/or a strong mineral lineation, curved in the horizontal plane. The curvature of the lineations can be as much as 20° over 0.5 m. Lineation trends in adjacent lozenges can differ by as much as 80°.

Of the lozenges which occur at outcrop scale, the average cross-section has a long axis of 1.0 to 2.5 m and a short axis (height) which ranges from 0.8 to 1.5 m.

Lozenges are developed in heterogeneous assemblages of tonalitic and mafic gneisses intercalated at a scale estimated to be several metres prior to deformation. Mega-boudins of tonalite gneiss are observed within assemblages of mafic gneiss lozenges (Fig. 3b). However, adjacent to lozenge structures in mafic gneiss, massive bodies of tonalite exhibit folds or ductile, normal faults.

Ductile, listric, normal faults

Small-scale, ductile, normal faults are developed in homogeneous and xenolithic tonalite bodies. Listric in geometry, these discontinuities extend 0.5 to 2.0 m in length above their sub-horizontal sole. A consistently normal sense of displacement is indicated by deflected layering at the fault zone. The faults form as solitary structures, imbricate sets (Fig. 4b) or conjugate sets (horst-graben geometry) (Fig. 4c) and are regionally pervasive in a belt extending 60 km west of Chapleau.

The trend of the axis of extension indicated by normal faults changes abruptly from east-west in the eastern half of the field area to roughly north-south in the western half. Several ages of faults are evident from crosscutting fault relationships (Fig. 4c, 4d). Segregated tonalite gneisses seem to be susceptible to this style of deformation. In such rocks the fault surface commonly coincides with secondary leucocratic material.

Folded tonalite gneisses

In the eastern half of, particularly surrounding Nagasin and Highbrush Lakes, folds occur in tonalite gneisses adjacent to lozenges. Folds are upright, disharmonic and noncylindrical, as illustrated in Figure 3a. Interlimb angles are generally moderate and fold noses are rounded. Where measurable, fold axes are variable in trend yet consistently subhorizontal in plunge. As well, folds are sporadic rather than pervasive throughout a given horizon. For example, layering in gneisses exposed to the left of the folds in Figure 3a is oriented more or less horizontally for 50 m.

The gneisses that exhibit this folding have fabrics characteristic of high strain rocks. They are fine grained, well-lineated and have a gneissosity defined by porphyroclastic, potassium feldspar-rich layers of grain-size-reduced pegmatite. This high-strain fabric predates folding.

DISCUSSION

Interpretation of D_3

There are numerous ductile fault structures in this part of the Wawa gneiss terrane. They consistently indicate normal sense of displacement and are interpreted to be a response to an extensional event of unknown duration, timing or significance. Even where lozenges and small-scale normal faults are in proximity, crosscutting relationships are absent, suggesting lithological control on style and contemporaneous formation. Zircon dating of tonalitic rocks which are pre- and post-D_3 is planned in an attempt to bracket the age of this event.

The change in orientation of extensional axes 30 km west of Chapleau is not yet understood. The change could be representative of variable extension directions during D_3 or a result of rotation by D_5 brittle deformation. Regardless of trend variation, the axis of extension remains subhorizontal throughout the area.

Extensional folding?

Of the numerous problems which remain to be interpreted, the most problematic is the folding documented in the Nagasin Lake area (Fig. 3a). Although clearly affecting mineral lineations and layering discontinuities that seem to have been produced during extensional deformation, the folds are also spatially associated with extensional fabrics that have not been subsequently deformed. The orientation of fold axes is highly variable although the plunge is always subhorizontal.

This structure is included with D_3 fabrics although its relationship to the extensional deformation is uncertain. Further mapping and detailed stereonet analysis of the folds is planned in order to determine their position and significance in the structural record.

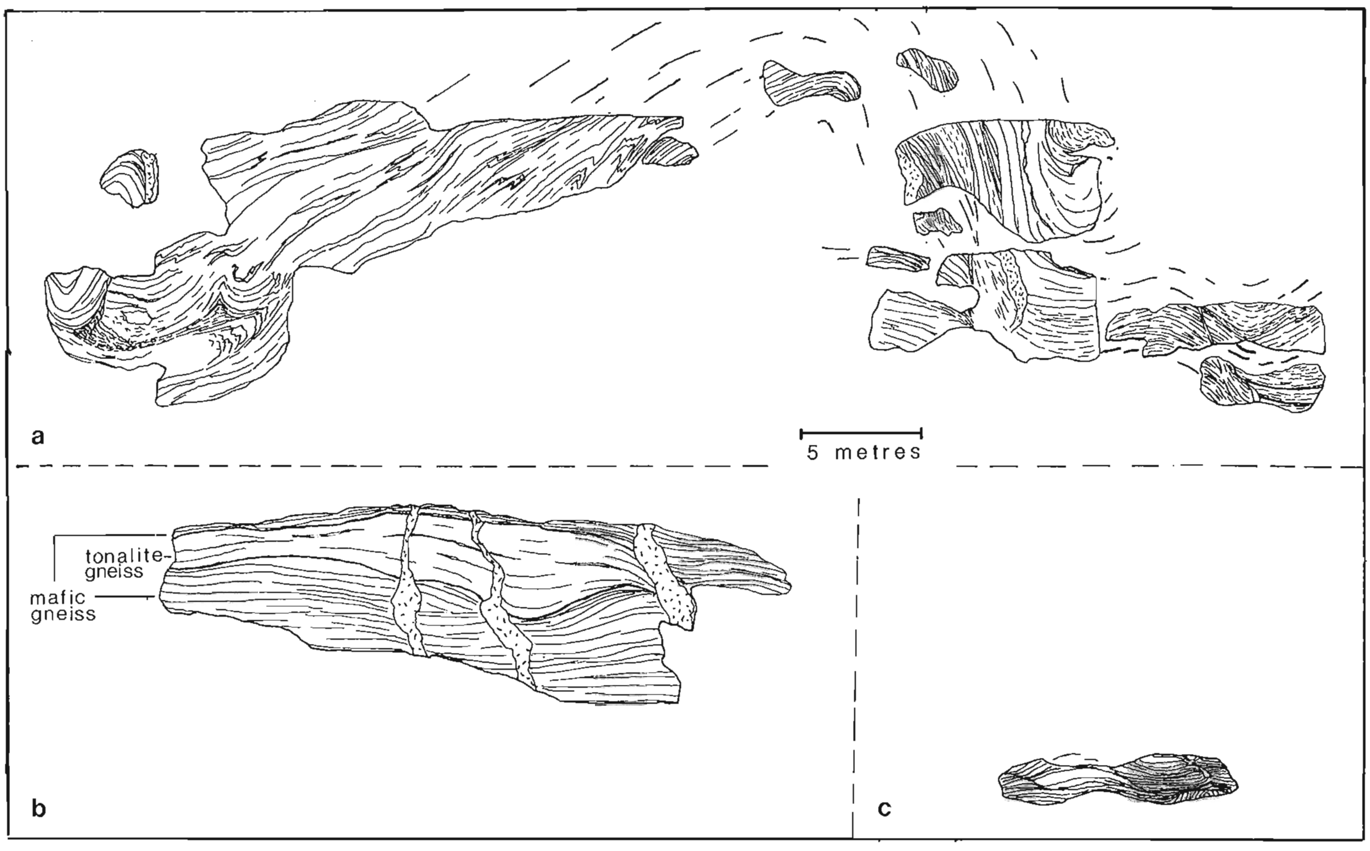


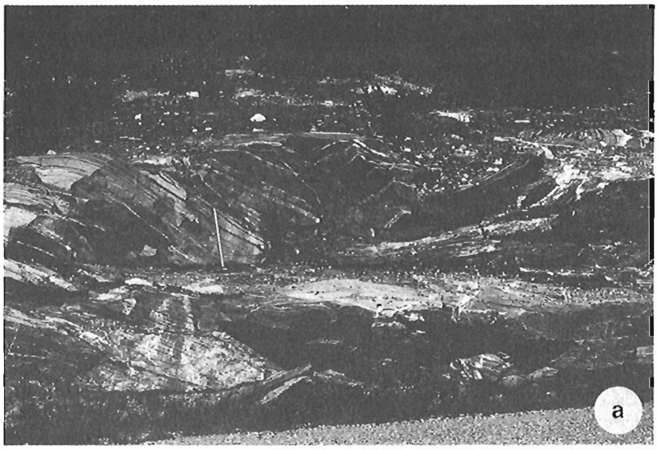
Figure 3. Sketches of three separate, near-vertical outcrops on the southwest side of a steep ridge. Note the difference in structural style between the upper and lower panels. The panels are oriented to retain the relative spatial distribution of these structures. a) Note the layering, discontinuities and thickness variations in the fold limbs of this well-lineated (dashed pattern) tonalite gneiss. The discontinuous, minor-folded layers are grain-size-reduced pegmatite. b) The third dimension of this outcrop shows this tonalitic gneiss body to be a mega-boudin. It has the same textural and compositional characteristics as the folded gneisses located 20 m above it. c) An example of a mafic gneiss lozenge; note the inclined, pegmatite-filled pressure shadow to the right.

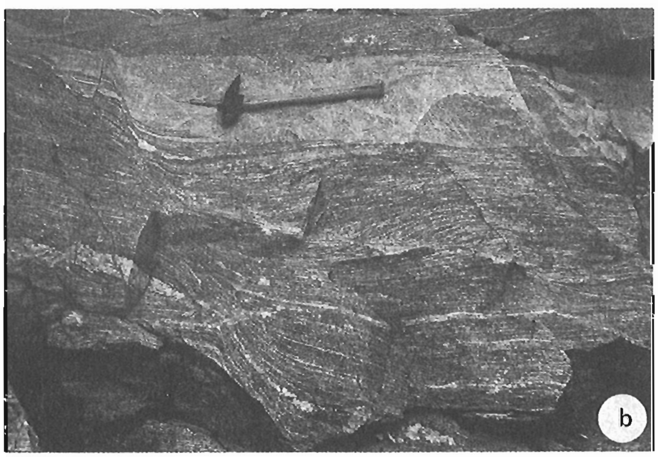
Variable lozenge lineations

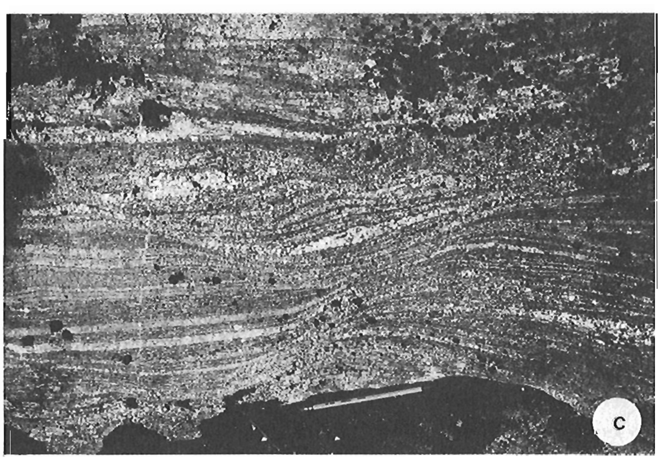
At well exposed outcrops of nested lozenges there are examples of highly variable mineral lineation trends, possibly the result of relative movement of lozenges on curved surfaces. However, the kinematics of such movement are still unknown. Some possibilities are: a) simultaneous extension in more than one direction; b) several extensional events along different axes, creating several ages of lineations; and c) extension along one axis with components of movement to either side due to jostling. In view of the prevalent regional east-west lineation, b) and c) are more viable hypotheses than a).

Regional distribution of extensional structures

The inset of Figure 2 shows the distribution of extensional structures in the area that has been mapped. Outlined is a roughly west-northwest striking band which is bounded to the south and north by large bodies of xenolith-rich intrusive rocks. The contacts appear to be intrusive rather than structurally gradational and therefore the shape of the belt characterized by extensional structures is probably unrelated to the extensional process. The western boundary, however, is structurally gradational to rocks which show no evidence of ductile extension.







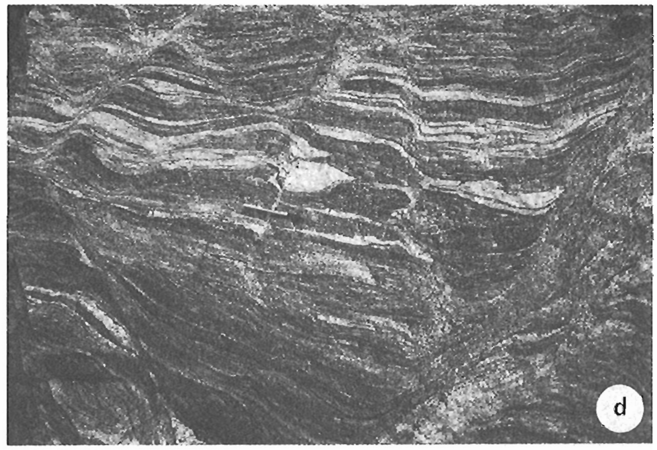


Figure 4a. Vertical cross-section through a mafic gneiss lozenge.

Figure 4b. Ductile, normal faults showing imbricate structure (note shallow, truncated fault within the hangingwall block).

Figure 4c. Conjugate, ductile, listric fault structure.

Figure 4d. Normal fault-bounded block with several internal cross-cutting minor faults.

Figure 4e. Horizontal section through circular structure on flat outcrop. Layering dips gently toward centre of structure The exposure is interpreted as a cross-section through a flat-lying mafic gneiss lozenge.

In general, lozenge structures are dominant in the east half of the area shown in the inset, while the converse is true for ductile, listric, normal faults in the west. Similar lozenge structures are present to the northeast in the Kapuskasing Structural Zone (S. Hanmer and J.A. Percival, pers. comm. 1986).

CONCLUSIONS

Orthogneisses of the Wawa gneiss terrane in the Chapleau area have a complex history of deformational and magmatic events. The most pervasive deformation occured at mid-crustal levels and resulted from synmagmatic, subhorizontal regional extension. The response of the rocks to this event produced a range of structural fabrics. Docu-mentation of the geometry, scale and kinematics of these structures will be useful in correlating surface features with depth information provided by LITHOPROBE reflection profiles and may ultimately shed light on the origin of deep crustal reflectors.

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Results of fieldwork in Foxe Fold Belt near Dewar Lakes, Baffin Island, N.W.T.

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Abstract

The Foxe Fold Belt near Dewar Lakes, central Baffin Island, is made up of Early Proterozoic supracrustal rocks; quartzite, pelitic schist, meta-igneous mafic and ultramafic rocks and metagreywacke, which overlie an Archean sialic basement. Both basement and cover have been deformed into dome and basin structures.

An initial extension (N-S) followed by several episodes of extension and compression with almost coaxial E-W axes caused the present dome and basin pattern. The peak of the first metamorphism coincided with the first extensional episode.

Résumé

La zone de plis sement de Foxe, près des lacs Dewar, aù centre de l'île de Baffin se compose de roches supracrustales datant du Protérozoïque inférieur; quartzite, schiste pélitique, roches métaignées de nature mafique et ultramafique et métagrauwacke reposent sur le socle sialique archéen, le tout ayant été déformé de façon à créer une série de dômes et de bassins.

Une phase initiale d'extension (N-S) suivie de plusieurs épisodes d'extension et compression presque co-axiales en direction E-O ont causé la configuration actuelle en dômes et bassins. Le premier maximum de métamorphisme a coincidé avec la première phase d'extension.

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INTRODUCTION

Bedrock mapping in the vicinity of Dewar Lakes (Fig. 1) was begun in July-August, 1987. This first field season was devoted to detailed studies of the structure, stratigraphy, metamorphism and metallogeny of part of the southern margin of Foxe Fold Belt in central Baffin Island. The project is a follow-up of reconnaissance mapping published by Henderson (1985a, b) and the work of Tippett (1979, 1980, 1984).

SUMMARY OF REGIONAL RELATIONS IN FOXE FOLD BELT, MELVILLE PENINSULA AND BAFFIN ISLAND

No satisfactory tectonic model has been proposed for the Early Proterozoic Foxe-Rinkian fold belt of Melville Peninsula, Baffin Island and Greenland (Fig. 1). The belt is made up of early Proterozoic supracrustal rocks overlying Archean sialic basement. Cover and basement have been deformed and metamorphosed several times between 1.9 and 1.8 Ga ago (Pidgeon and Howie, 1975; Henderson and Loveridge, 1981). The supracrustal rocks defining the fold belt (but not the orogen) form an eastward-opening wedge from Melville Peninsula to the inland ice of Greenland (Fig. 1).

Southwest of Melville Peninsula the fold belt is engulfed by calc-alkaline late- to post-kinematic batholiths comprising a probable continental magmatic arc (LeCheminant et al., 1987). To the northeast of the batholiths the fold belt is defined by the Penrhyn Group, a dominantly platformal sedimentary facies of quartz- and carbonate-rich rocks metamorphosed to upper amphibolite facies. The major structures in the belt in Melville Peninsula are nappes which have placed Archean on Proterozoic rocks over extensive areas (Henderson, 1983). Large components of longitudinal east-west extension and transverse north-south shortening imposed

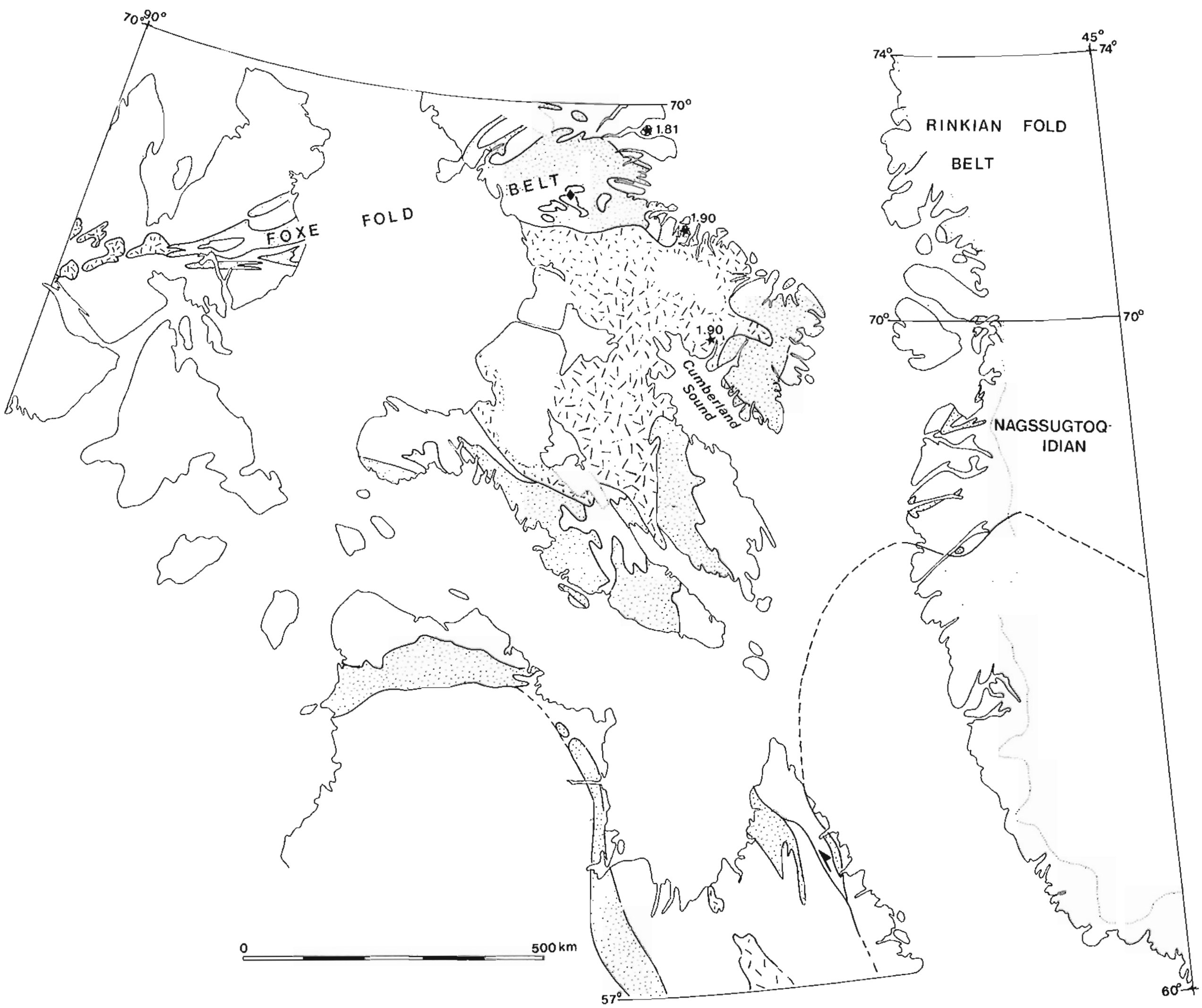


Figure 1. Location map showing the distribution of early Proterozoic supracrustal rocks (random dot pattern), and some early Proterozoic granitic terranes (random dash pattern) in northeastern Canada and western Greenland. The Dewar Lakes area (Fig. 2) is indicated by the small diamond in Foxe Fold Belt, central Baffin Island. Stars indicate the location of U-Pb zircon dates in Ga. Greenland and northeastern Canada are shown in their relative positions prior to opening of Davis Strait.

on the early recumbent structures produced sheath folds and upright folds which Henderson (1984) related to sinistral transpression of the supracrustal belt. The extension and shortening postdating early nappe formation apparently obliterated structural evidence bearing on the original polarity of the Foxe Fold Belt in Melville Peninsula.

On Baffin Island major differences occur in stratigraphy, structure and metamorphism within the fold belt compared with Melville Peninsula. There is a marked decrease in thickness of the platformal sedimentary sequence in the Early Proterozoic Piling Group eastward from the vicinity of Flint Lake, west Baffin Island (Morgan, 1983). The dominant lithologies of Piling Group are psammitic to pelitic metagreywacke with locally significant thicknesses of impure sulphide facies iron formation and mafic/ultramafic flows and sills. Sheath folds and nappes have not been documented on Baffin Island; the major folds are upright east-west striking, and deform an early bedding-parallel foliation. Both east-west (longitudinal) and north-south (transverse) mineral lineations occur with the early foliation. An elliptical basin and dome structural pattern is evident in the Dewar Lakes area (Fig. 1, 2) which Tippett (1979) attributed to active gneiss doming as well as lateral compression of the basement-cover interface and igneous intrusion. An extensive greenschist (biotite grade) terrane in metagreywacke, north of Dewar Lakes in the centre of the fold belt, indicates a post-metamorphic downwarping of the youngest supracrustal rocks. South of Dewar Lakes, Piling Group and its Archean basement rocks are progressively metamorphosed, migmatized, intruded and ultimately assimilated into the immense batholithic complex centred at the head of Cumberland Sound (Fig. 1). The northern margin of the fold belt on Baffin Island, like Melville Peninsula, is marked by the northern extent of preserved Early Proterozoic supracrustal rocks.

PURPOSE OF THE STUDY

Mapping by Morgan et al. (1976), Tippett (1979, 1980, 1984) and Henderson (1985a) indicated that in the area around Dewar Lakes a complete stratigraphy of Piling Group as well as Archean basement and younger granitic intrusive rocks are exposed at metamorphic conditions varying from biotite grade to sillimanite-K-feldspar grade. Some lake sediments in the area contain more than 1000 ppm arsenic (Cameron, 1986) and sulphides have been reported associated with metamorphosed mafic to ultramafic rocks as well as metasedimentary rocks (Tippett, 1979). Sangster (1981) considered graphitic sulphidic schist in the Piling Group a potential host for stratiform lead-zinc mineralization.

Our purpose in 1987 was to concentrate our mapping efforts in the region of lowest metamorphic grade in order 1) to define the sequence, geometry and kinematics of meso- and macroscale structures, 2) to determine the stratigraphic sequence of metasedimentary units within the Piling Group, 3) to determine the distribution, tectonic setting and mineral potential of mafic to ultramafic rocks and graphitic sulphidic schist in the Piling Group, and 4) to determine the bedrock source of arsenic in lake sediments.

SUMMARY OF RESULTS OF FIELDWORK

1. A bedding-parallel foliation (S_1) is a penetrative structural fabric element in supracrustal rocks throughout the area mapped (Fig. 2). In many areas S_0 and S_1 are subhorizontal. The S_1 fabric was not seen to be an axial-plane foliation nor to deviate from parallelism with bedding (S_0). In places S_1 is also a differentiated layering (S_d) which shows double boudinage, but the possibility cannot be eliminated that in these areas the planar fabric is a composite of S_0, S_1 and S_2.

2. A sillimanite-fibre lineation (L_1) is widespread in the region. Two trends are evident: N-S and E-W. North of the crest of South Jackson Lake anticline stepped sillimanite slickenfibres denote top-to-the-north shear (locality A, Fig. 2). South of Tirrul Lake on Boas River the N-S-trending L_1 sillimanite lineation is rotated around the E-W hinge of an F_2 fold (locality B, Fig. 2). The E-W-trending sillimanite lineations as well as striping lineations apparently are not rotated (Fig. 3), probably because the younger folds are nearly coaxial with the lineation. This raises the possibility that E-W-trending lineations on S_1 may actually be younger than D_1 (i.e. they may be D_2, D_3 or D_4 fabric elements).

3. In supracrustal rocks near Archean basement contacts, S_2 crenulation cleavage and intrafolial F_2 folds *consistently* verge away (see Weijermars, 1982 for vergence definition) from the major axis of the basement-cored domes (Fig. 4, 5). This asymmetry indicates a component of normal (extensional) shear away from the crests of the domes during D_2. However, asymmetrical fabric elements (shear bands, asymmetrical foliation boudinage) delineate D_2 mylonite zones on the flanks of the Dewar Lakes dome near the basement-cover contact that exhibit dextral shear on the south limb and sinistral shear on the north limb of the dome (Fig. 2); this may mean that the supracrustal rocks were transported from east to west during D_2 if the mylonite zone was folded over the crest of the dome.

4. A synform-antiform pair on the north limb of the east-plunging Varley Lake synform south of Tirrul Lake (Fig. 2) are macroscale F_3 folds as shown by vertical S_3 axial planar crenulation cleavage (Fig. 6) which verges *towards* the F_3 antiform and refolds an intrafolial F_2 fold which verges *away* from the F_3 antiform.

5. The Varley Lake synform (Fig. 2) folds the axial surfaces of macroscopic F_2 and F_3 folds, and therefore must be an F_4 fold. Considering that the Varley Lake synform is the linking structure between the F_2 Dewar Lakes and Varley Lake domes it seems that we are left in a quandary at the present time regarding the relative age of the large E-W-trending domes and basins which is exacerbated by their coaxial macroscopic geometry (Fig. 3).

6. Although the grade of regional metamorphism increases over several tens of kilometres from biotite to sillimanite-K-feldspar migmatite towards the south and east of Dewar Lakes (Henderson and Tippett 1980; Tippett, 1980, 1984), there is an apparent steep local metamorphic gradient in the Dewar Lakes area from the basal quartzite-schist into the overlying metagreywacke. This local gradient is probably a syn D_1 phenomenon attested by N-S oriented sillimanite lineations

71°45'
68°45'
THOMSON LAKE
VARLEY
CASSON LAKE
NORTH JACKSON LAKE
SOUTH JACKSON LAKE
MAPPING LIMIT
HARRIS LAKE
68°25'
71°45'

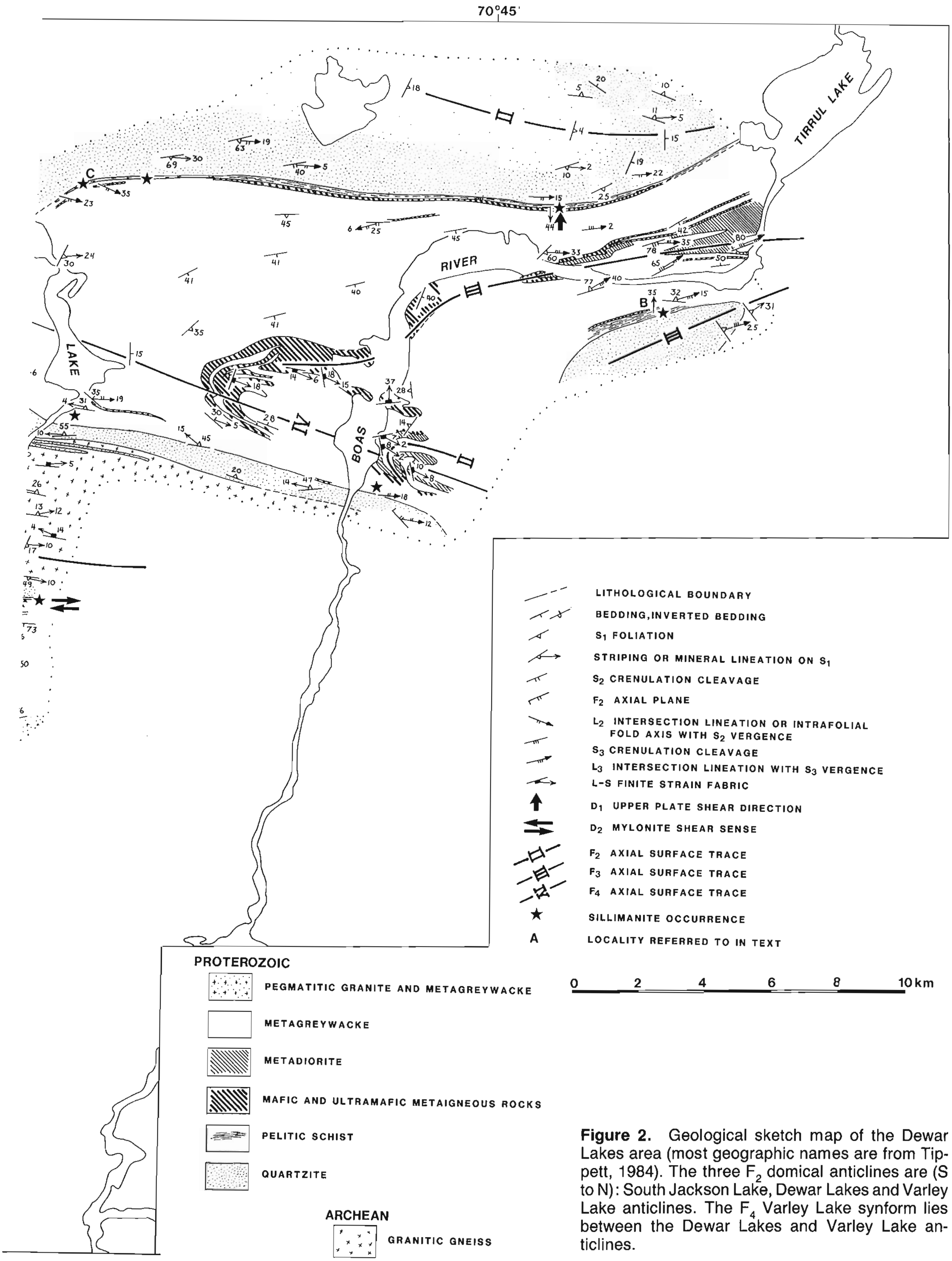
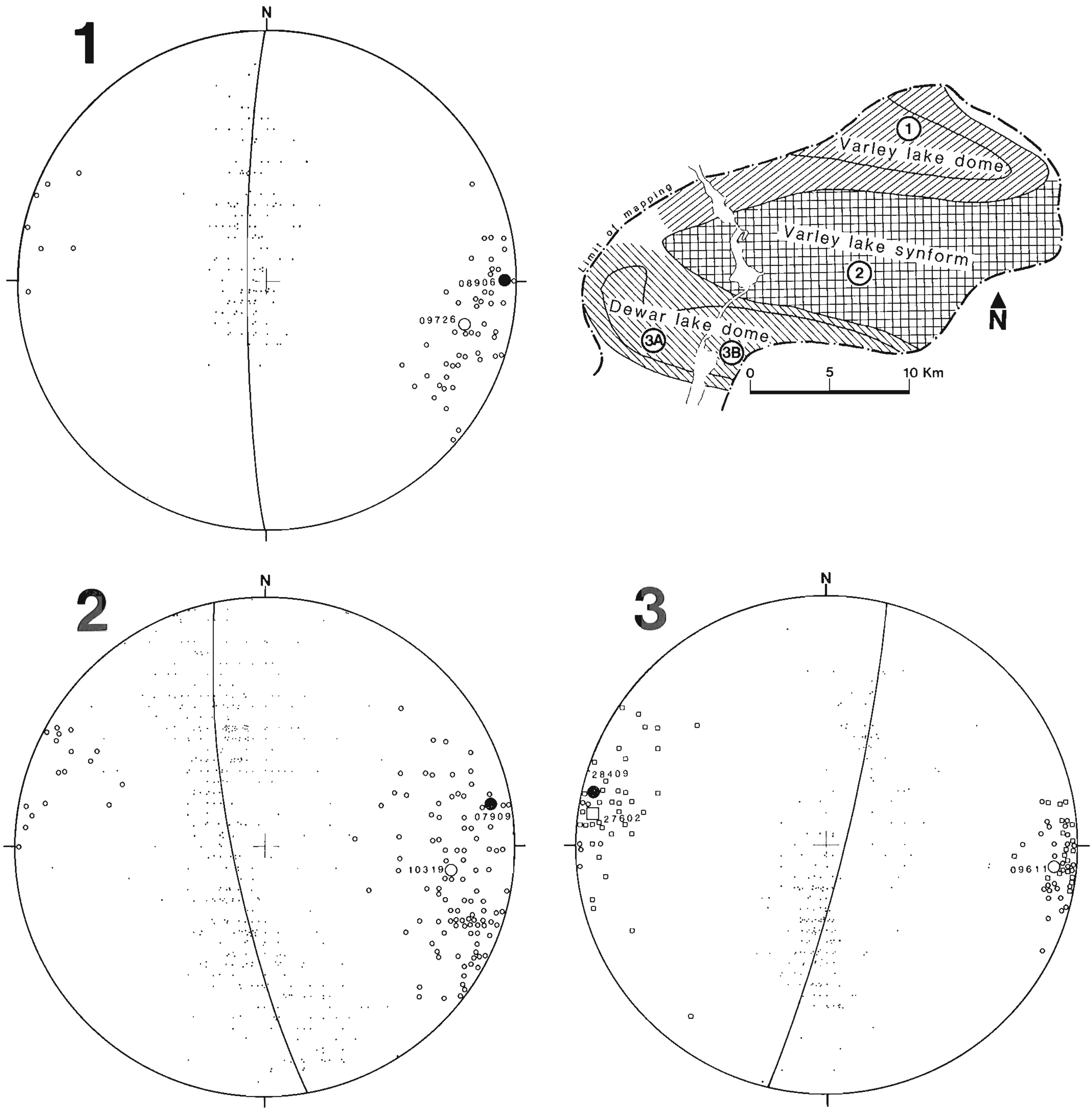


Figure 2. Geological sketch map of the Dewar Lakes area (most geographic names are from Tippett, 1984). The three F_2 domical anticlines are (S to N): South Jackson Lake, Dewar Lakes and Varley Lake anticlines. The F_4 Varley Lake synform lies between the Dewar Lakes and Varley Lake anticlines.



KEY

. POLE TO S_0, S_1, OR FINITE STRAIN PLANAR FABRIC

● POLE TO BEST FIT GREAT CIRCLE OF S_0, S_1, AND FINITE STRAIN PLANAR FABRIC

∘ STRIPING, MINERAL LINEATION, OR FINITE STRAIN LINEAR FABRIC (DOMAIN 1, 2, 3B)

○ BEST-FIT LINEATION ORIENTATION (DOMAIN 1, 2, 3B)

▫ STRIPING, MINERAL LINEATION, OR FINITE STRAIN LINEAR FABRIC (DOMAIN 3A)

□ BEST-FIT LINEATION ORIENTATION (DOMAIN 3A)

Figure 3. Macroscopic geometry of structural fabric elements around Dewar Lakes dome, Varley Lake synform and Varley Lake dome. Note the gentle east-west arch along the Dewar Lakes dome shown by fabric domains 3A and 3B.

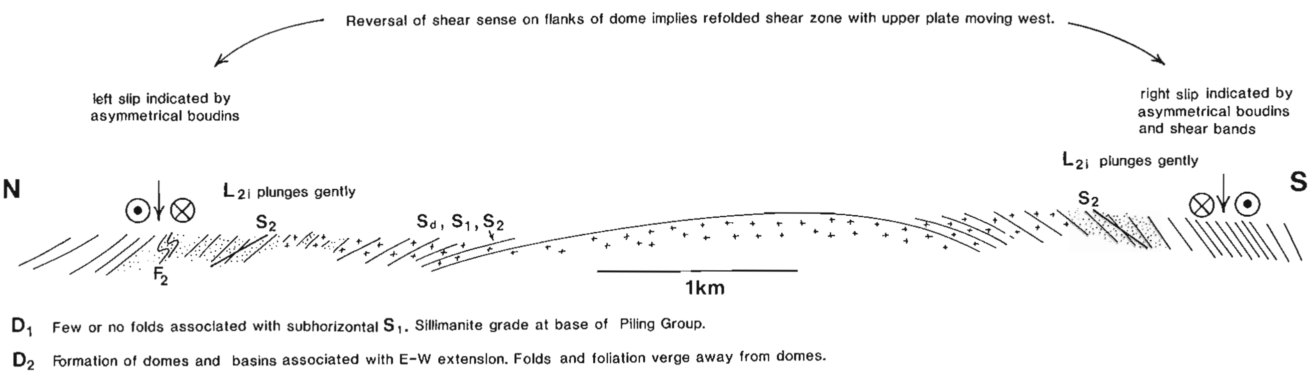


Figure 4. Schematic north-south cross section through the Dewar Lakes anticline (S_d is differentiated layering, L_{2i} is S_1/S_2 intersection lineation).

on S_1 foliations. A minimum temperature and pressure for the basal supracrustal rocks is about 560°C and 3.3 MPa based upon the occurrence of sillimanite-garnet-staurolite (locality C, Fig. 2). Pelitic beds in overlying metagreywacke are biotite grade, suggesting lower grade conditions. Abundant deformed muscovite granite pegmatite sills and dykes that intruded metagreywacke in a 2-km-wide zone around South Jackson Lake anticline (Fig. 2) may also reflect a higher metamorphic grade near the base of the Piling Group.

7. Field relations indicate that as many as eight mafic to ultramafic sills and flows may be present in the area around Casson Lake (Fig. 2). They are separated by metagreywacke that is interlayered with calcsilicate and carbonate-rich beds. Schistose, graphite-sulphide-rich gossans also occur between flow units. The mafic-ultramafic rocks have recrystallized to lower amphibolite grade and are intensely sheared in places, making their identification difficult. Field descriptions are therefore limited to grain coarseness and proportion of amphibole relative to plagioclase. Rare flattened volcaniclastic rocks were recognized.

The observed stratigraphy of a differentiated sill (up to 600 m thick) east of Casson Lake can be summarized as follows, from top to bottom: diorite (not always present), 100 m; metagabbro, coarse-grained, 200 m; ultramafic rock, fine-grained, schistose, 100 m; and ultramafic rock, 200 m.

8. Samples of metagreywacke were collected from outcrops as close as possible to lakes which have bottom sediment arsenic values ranging from 250 to more than 1000 ppm (Cameron, 1986). These greywacke samples are being analysed for arsenic and gold. No visible mineralization was apparent in any of the samples. Rusty weathered graphite-sulphide-quartz schist samples are also being analysed for arsenic and gold.

DISCUSSION OF STRATIGRAPHY, STRUCTURE AND METAMORPHISM

Our mapping so far has confirmed a basically simple stratigraphy with no evidence of significant imbrication of map units and a generally upward-facing sequence.

Figure 5. Photograph looking east of folded S_1 and S_d on the north limb of the F_3 antiform south of Tirrul Lake (near loc. B, Fig. 2) showing north vergent F_2 folds. Note hammer for scale beside folded quartz lens in centre of photo. GSC 204260-N.

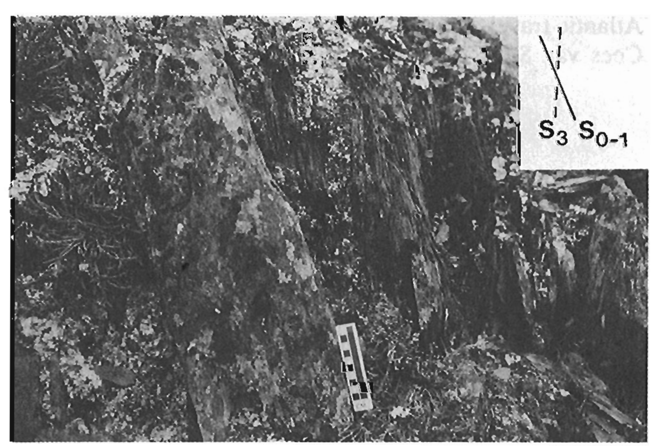


Figure 6. Photograph looking east showing vertical S_3 crenulation cleavage in schist above a 25-cm-thick metagreywacke bed on the north limb near the core of the macroscale F_3 synform southwest of Tirrul Lake (Fig. 2). GSC 204260-K.

In addition to the absence of obvious thrusting the abundance of boudinage and absence of folds associated with D_1 leads to our present preliminary conclusion that D_1 is the result of layer-parallel extension responsible for the development of the observed bedding-parallel S_1 and N-S sillimanite L_1. The generally shallow dips of bedding and S_1 foliation would be the result of listric extensional shearing at mid-crustal levels now exposed by uplift and erosion.

Later events (D_3, D_4...) are responsible for the gentle domes and basins. These folds have low amplitude and are not pervasive, commonly dying along their E-W axes which are almost coaxial (Fig. 3).

D_2 apparently was a low angle shearing event with upper plate displacement from east to west. Thus far we have not been able to determine whether D_2 involved detachment or imbrication but we favour a detachment model because of the apparent absence of stratigraphic duplication in the area. The vergence of F_2 intrafolial folds away from the domes is inconsistent with our favoured D_2 kinematic model, but the resolution of the inconsistency awaits further fieldwork.

The grade of metamorphism of the lower quartzite-pelite unit, characterised by garnet-cordierite and garnet-staurolite-sillimanite, appears significantly higher than that of the overlying metagreywacke. Additional work should determine if we are dealing with a steep isograd pattern or a metamorphic hiatus. However, the presence of a D_2 extensional shear zone at the top of the quartzite-pelite unit and bottom of the metagreywacke unit would allow for either case if the metamorphism is D_1.

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We are grateful to the FOX-3 Dew Line personnel and especially Carl van Buren and Brian Yaworski for their priceless help. Our helicopter crew, Michael Brocklebank and Shawn Coulter, provided excellent service and accepted our primitive living conditions with great humour. Grocott gratefully acknowledges the Anglo-Canadian Scientific Exchange Scheme for funding his study visit. Heÿke gratefully acknowledges the Dr. Schürmann Fonds for providing trans-Atlantic travel expenses. We thank Tony LeCheminant and Cees van Staal for reviewing the manuscript.

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Glacial studies in Labrador[1]

R.D. Klassen and F.J. Thompson

Klassen, R.D. and Thompson, F.J., Glacial studies in Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 109-116, 1988.

Abstract

Regional studies in Labrador support earlier work which indicated a complex history of shifting ice flow based on glacial striae and dispersal of glacial erratics. In the Wabush region, youngest striae indicate flow towards the southeast, consistent with glacial moulding of bedrock. Older striae record flow to the south-southwest in the west, and to the south-southeast in the south. Erratics derived from Labrador Trough, north of the Grenville Structural Province, are widespread. To the west they are rare and are associated with transport during the earliest (SSW) phase of flow. To the south they are abundant and appear to have been transported during the two last phases of flow (SSE, SE). Recently-discovered stratigraphic sections contain distinct till units and interglacial organic deposits that promise a record to match the complex glacial history recorded by surficial geology.

Résumé

Des études régionales effectuées au Labrador confirment des travaux antérieurs qui ont établi une histoire complexe d'écoulement variable des glaces en se basant sur l'observation des stries glaciaires et de la dispersion des blocs erratiques. Dans la région de Wabush, les stries les plus récentes indiquent un écoulement vers le sud-est, conséquent avec le modelé glaciaire du socle. Des stries plus anciennes indiquent un écoulement vers le sud-sud-ouest à l'ouest, et vers le sud-sud-est au sud. Les roches erratiques dérivées de la fosse du Labrador, au nord de la province structurale de Grenville, sont très nombreuses. À l'ouest, elles sont rares et associées au transport de matériaux durant la phase d'écoulement la plus ancienne (SSO). Au sud, elles sont abondantes et semblent avoir été transportées durant les deux dernières phases d'écoulement (SSE, SE). Des coupes stratigraphiques récemment découvertes contiennent des unités distinctives de till et des dépôts organiques interglaciaires qui permettront sans doute d'établir une chronostratigraphie tout aussi importante que l'histoire glaciaire complexe documentée par la géologie des formations en surface.

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[1] Contribution to the Canada-Newfoundland Mineral Development Agreement 1984-1989. Project carried by Geological Survey of Canada, Terrain Sciences Division.

INTRODUCTION

The preliminary results of Quaternary field studies conducted in Labrador during the 1987 field season are presented here as part of a project funded under Canada-Newfoundland Mineral Development agreements since 1982. The objectives of the work are to determine the history of ice movement and the composition of glacial drift as a basis for developing techniques of drift prospecting.

Based on glacial striae, till composition and distribution of glacial erratics a complex history of ice flow has been identified in Labrador (Klassen and Bolduc, 1984; Thompson and Klassen, 1986; Klassen and Thompson, 1987). For example, near the ice sheet margin in eastern Labrador dispersal from the Flowers River intrusive suite is simple, forming a train as wide as the source outcrop extending northeast for over 30 km, parallel with the last phase of ice flow (Fig. 1). In contrast, in central Labrador fan-shaped dispersal from the Red Wine Complex is the product of early flow towards the northeast and later flows to the east-northeast and east-southeast. In western Labrador, near a centre of outflow of the ice sheet, dispersal from a nepheline syenite complex is a product of early ice flow to the southwest and later flows to the north and northwest. These studies also identified areas with anomalous levels of base metals, rare earth elements and uranium in till (Klassen et al., 1986; Thompson et al., 1986).

The objectives of the 1987 field studies were to: confirm the geochemical anomalies identified during previous work, extend the regional sampling program into the Wabush region of southwestern Labrador and to develop examples of glacial dispersal at a detailed scale (<5 km) from known mineral occurrences.

METHODS

A helicopter was used for field transportation except in the Wabush/Labrador City area where roads were used for more detailed sampling and for access to Quaternary sections exposed in the open pit mines. At observation sites, till was sampled for geochemical and lithologic analysis in the laboratory and, where possible, striae and clast lithologies were noted. The work was concentrated in areas of Labrador with the greatest mineral potential but was extended to adjoining areas of Quebec and areas of lesser mineral potential in Labrador to provide a more complete understanding of the history of ice flow and of regional variations in drift composition. Sample densities were generally one to two per 100 km^2. Sample sites are preferentially located on hilltops because valleys provide few landing sites or till sampling sites due to swamp and forest cover. For this reason the 1987 sampling program was concentrated in the highlands in the

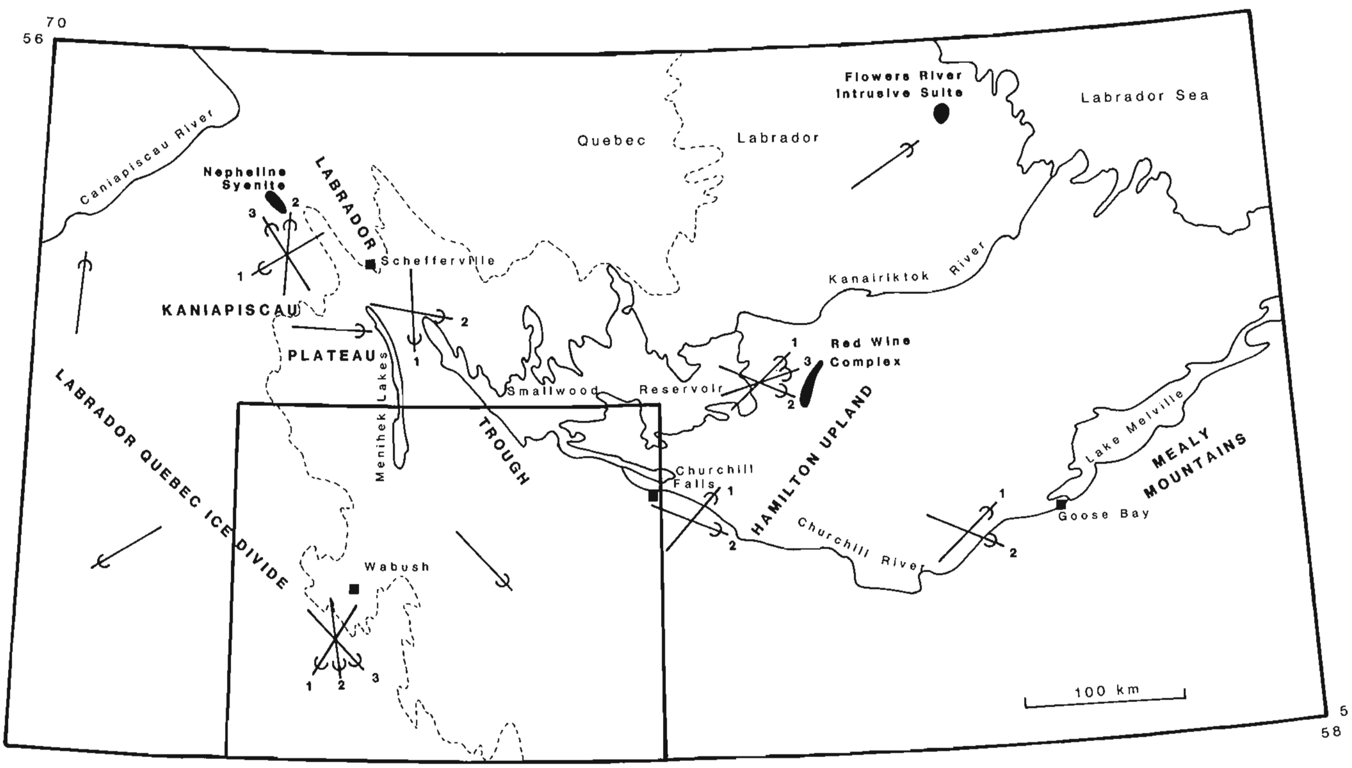

Figure 1. Location map showing bedrock (1 oldest). The Wabush study area is outlined.

western part of the Wabush area and limited work was conducted in the swampy area to the east. Pits were dug by shovel to depths of 30 to 70 cm depending upon depth to bedrock and on difficulties in penetrating stony till. Samples were collected from the least oxidized and most organic-free section of the modern soil profile near the base of the pit. In areas of tundra vegetation on hilltops, pits were dug in mudboils where soil zonation was absent or poorly developed due to periglacial churning.

REGIONAL STUDIES

Striae measured in 1987 west of Goose Bay indicate that iceflow toward the northeast preceded east and southeast phases of flow (Fig. 1). Evidence of older northeast flow has also been found in the areas of Churchill Falls (Klassen and Bolduc, 1984) and east of the Smallwood Reservoir (Thompson and Klassen, 1986) and is considered to reflect a major phase of ice movement that occurred relatively early during the course of glaciation.

In Central Labrador, erratics of Red Wine Complex were mapped along the southern margin of a regional-scale dispersal train that had been previously defined (Thompson and Klassen, 1986) and were found up to 60 km east-southeast of the bedrock source. Their distribution indicates that the ice that transported them overtopped the Hamilton Upland, 200 m above the regional plateau. Ice flow towards the east-northeast noted west of the highlands is younger than the east-southeast flow and may reflect deflection of the ice by the highlands.

South of Lake Melville along the margin of the Mealy Mountains and south of the Churchill River, well-rounded quartzite cobbles in till are thought to have been derived from quartzite cobble conglomerate of the Double Mer Formation. This bedrock type has recently been found along the Churchill River 50 km east of Goose Bay (R. Wardle, pers. comm., 1987). The wide distribution of quartzite erratics suggests that quartzite cobble conglomerate occurs in bedrock elsewhere along the Churchill River and possibly under Lake Melville. Ice that transported these erratics flowed generally eastward and was thick enough to overtop at least the northern margin of the Mealy Mountains.

WABUSH REGION

Previous work

The Wabush area lies south of the U-shaped Labrador/Quebec ice divide (Prest, 1968; Fig. 1). Earlier Quaternary work (Henderson, 1959) indicated ice flow toward the southeast across this region. West of the Wabush area, initial ice movement was toward the southwest from a dispersal centre over or east of the Labrador Trough and late ice movement was to the north in the Caniapiscau River area and to the east on the Kaniapiskan Plateau west of Manihek Lakes (Hughes, 1964). The southern limit of northward flow defines the Labrador/Quebec ice divide. In the Schefferville region a sequence of five phases of ice flow include: ice movement to the southwest and northeast from a dispersal centre in or near the Labrador Trough (Phase I), northward-directed flow across the Kaniapiskau Plateau (Phase II), flow to the northwest, southeast and south from a dispersal centre near Schefferville (Phase III), eastward flow on the Kaniapiskau Plateau and into the Labrador Trough (Phase IV) and a late minor northeast directed phase of flow across the Labrador Trough (Phase V) (Klassen and Thompson, 1987).

BEDROCK GEOLOGY AND PHYSIOGRAPHY

The study area is underlain by gneisses of the Grenville Province in the central and south parts, granitoid rocks (Ashuanipi Complex) of the Superior Province to the northwest and granites and gneisses of the Churchill Province to the northwest. Relatively unmetamorphosed sediments of the Labrador Trough (slate, arkose, quartzite, iron formation and dolomite) occur in a 70-km-wide band between the Churchill Gneisses and the Ashuanipi Complex. Recrystallized quartzite, dolomite, iron formation and schistose rocks occur at the contact between the Grenville and Superior provinces in the Wabush Lake/Shabogamo Lake area. Scattered outcrops of Paleohelikian Sims Quartzite overlie rocks of the Labrador Trough (Fig. 2).

The Ashuanipi Complex forms a rugged highland where elevations commonly exceed 800 feet a.s.l. Bedrock highlands also occur to the south in the Grenville Province while to the east swampy flatland with limited bedrock exposure overlies rocks of the Labrador Trough, Churchill Province and Grenville Province. Quartzites of the Sims Formation form distinctive highlands exceeding 800 feet a.s.l. within the Labrador Trough.

SURFICIAL GEOLOGY

Till is generally thin (<2 m) to discontinuous in the highlands and is locally thicker (2 to >10 m) in valleys. The nature of the till is broadly dependent on the underlying bedrock. On bedrock units resistant to glacial erosion, such as granitic rocks of the Ashuanipi Complex and quartzite of the Sims Formation, till is thin and has little fine matrix component. In contrast shales of the Labrador Trough and gneissic rocks of the Grenville Province are relatively soft and easily comminuted to sand, silt and clay and till overlying these bedrock types is relatively thick and has a greater component of fines in the matrix. Streamlined landforms are commonly associated with tills developed on Grenville gneisses.

ICE FLOW TRENDS

Most of the striae observations were made on exposed surfaces on hilltops. The sense of flow was determined from the streamlined aspect of the outcrop and from small-scale crag and tail features. The relative age of striae has been established by careful determination of crosscutting relationships where multiple trends are recorded. The oldest striae occur on the most protected lee faces and the youngest on the most exposed stoss faces (Klassen and Bolduc, 1984).

The youngest striae (Fig. 2) generally indicate ice flow from northwest to southeast. These striae are consistent with the molding of the outcrop and with the streamlined landforms recorded by Henderson (1959) and are consequently considered to reflect a major phase of ice movement. Eastward flow occurs in the northwest and is considered to be a part of a narrow, elongate zone of eastward ice flow that originates on the Kaniapiskau Plateau and extends across central Labrador (Phase IV; Klassen and Thompson, 1987).

Older striae (Fig. 3) were recognized that record flow to the south-southwest across the highlands in the west and to the south-southeast in the south. At two sites south-southwest striae were found to predate south-southeast trending striae. The striae indicate that ice movement was initially toward the south-southwest, shifted later toward the south-southeast and finally toward the southeast. Although uncertain, south-southwest ice flow trends may be correlated with oldest ice flow trends in the Schefferville region (Phase I; Klassen and Thompson, 1987). Striae indicating older flow to the east are also recorded but there is insufficient data to fully evaluate their significance.

DISTRIBUTION OF INDICATOR ERRATICS

Lithologies and relative abundances of clasts were noted to define the history of ice movement by recognition of the distance and direction of glacial transport. The till is derived primarily from the underlying and immediately upice bedrock and clasts that have undergone transport of >50 km commonly represent <3 % of the coarse fraction.

Erratics derived from the Labrador Trough (Sokoman Iron Formation) occur throughout the study area (Fig. 4). To the west, trough derived erratics are rare and comprise <0.1 % of the till. Within the Labrador Trough, and south

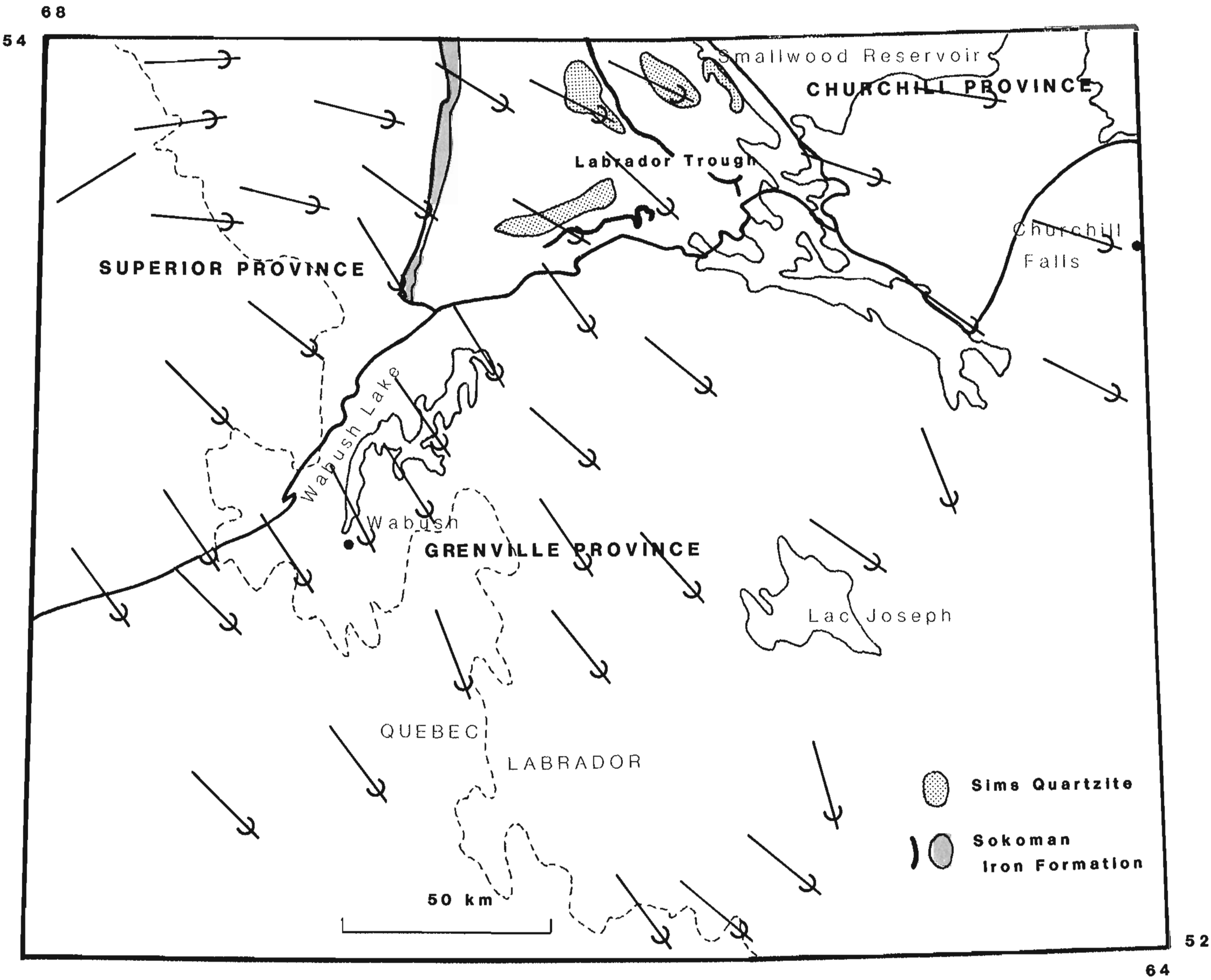

Figure 2. Youngest ice flow trends generalized after striae and streamlined landforms. Bedrock geology generalized after Greene (1970) and Avramtchev (1985).

and east of it, iron formation is abundant and generally exceeds 1 % (field estimate). The overall distribution of erratics of iron formation cannot be accounted for by the last phase of ice flow alone, which would have carried debris only toward the southeast.

The presence of a rare erratics of iron formation west and southwest of the source area is attributed to the oldest south-southwest phase of ice movement. Southwest transport of erratics of iron formation has been noted by Hughes (1964) and Klassen and Thompson (1987). The western margin of till characterized by abundant trough-derived debris extends south-southeast from the trough (Fig. 4). The trend of the dispersal margin is parallel with south-southeast trending striae of intermediate age and indicates that the transport of abundant trough derived debris occurred during the intermediate phase of flow as well as during the last phase of flow. The importance of the intermediate phase of flow in the transport of debris is also demonstrated by the distribution of erratics of Sims Quartzite which occur up to 170 km south of the bedrock source (Fig. 5). This phase of flow is associated with a dispersal centre located north of the Wabush area and may be related to southward ice flow in the Schefferville area (Phase III; Klassen and Thompson, 1987).

STRATIGRAPHY

Stratigraphic sections in open pit mines at Schefferville and Wabush comprise several distinct till units. Because some of the tills are separated by waterlain sediments, which at one site contain organic material, there is good indication that these sections represent a long period of glacial history and contain an important compositional record of change in

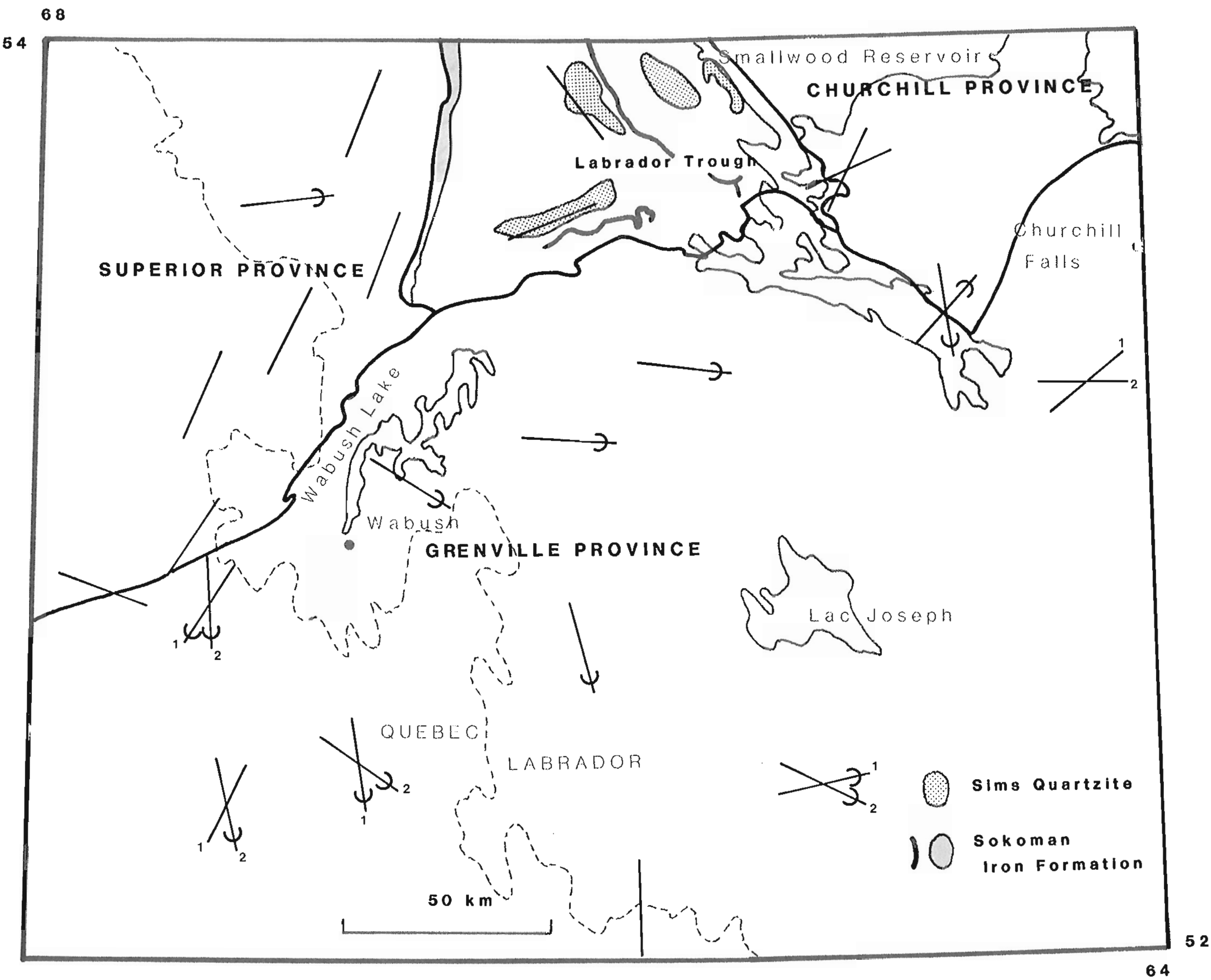

Figure 3. Older glacial striae showing direction and relative ages (1 oldest).

ice flow. The tills would appear to be related to the multiple phases of flow described here and by Klassen and Thompson (1987).

In most of the sections at least two till units can be distinguished by colour, composition and texture (field observation). The most prominent colour contrasts appear to reflect differences in the proportions of debris derived from either iron formation or red shale, both of which can impart strong red coloration at low levels of concentration. Pending analyses of composition in the laboratory, the significance of the differences in terms of change in ice flow direction and in bedrock provenance is not established. The most important of the stratigraphic sections occurs within the walls of the East and South pits of the Scully Mine, Wabush, where till units are separated by waterlain deposits containing fragments of spruce and willow that are up to 10 cm in length and 1-2 cm in width. Because of the central position of Wabush within Labrador-Quebec the growth of trees is possible only when the ice sheets have disappeared, and the presence of wood fragments is considered to demonstrate interglacial conditions.

A composite section of the Quaternary stratigraphy in the Scully Mine includes, in an ascending stratigraphic sequence (Fig. 6):

1. Grey till (muddy sand diamicton) containing well rounded clasts and small (< 30 cm) bodies of well sorted stratified sand; vivianite (?) common, particularly on joint surfaces; well defined upper contact.
2. Dark grey muddy sand, probably lake sediment, containing sandy crossbeds near the base, poorly defined laminae throughout; well defined upper contact. Contains organic matter that is finely divided and wood fragments.

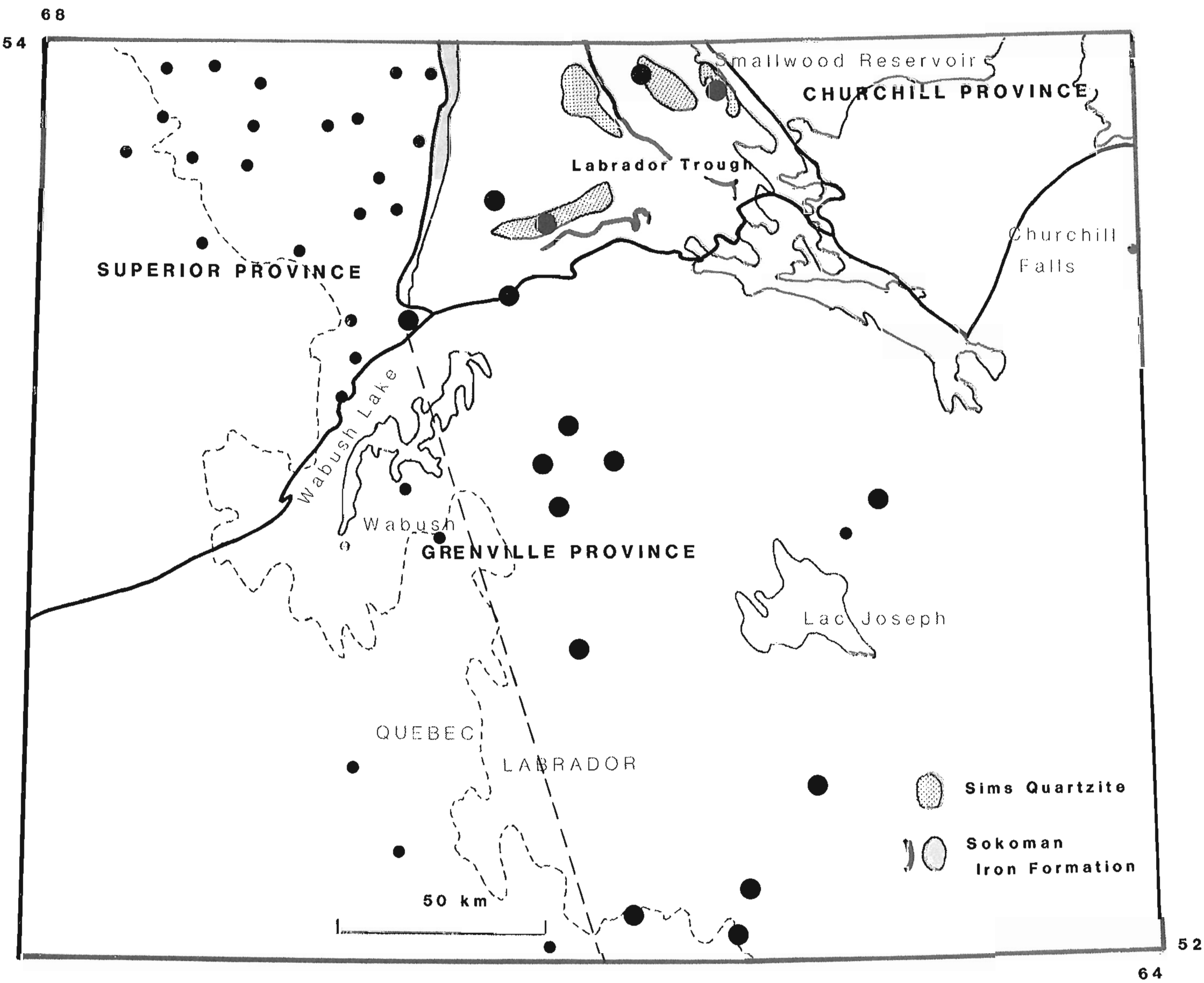

Figure 4. Distribution of glacial erratics of Sokoman Iron Formation (large dot >1%, small dot <0.1 by field estimate).

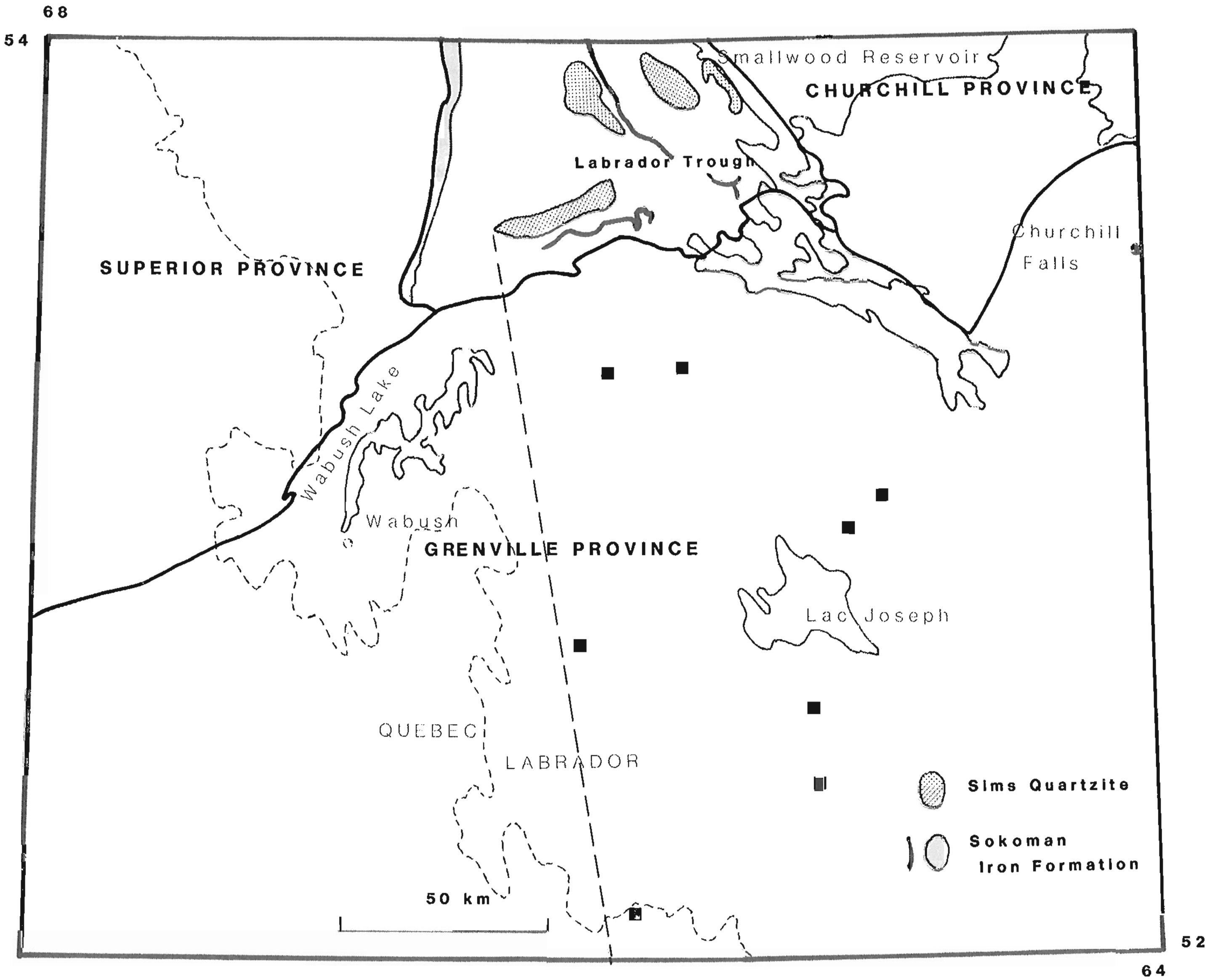

Figure 5. Distribution of glacial erratics of Sims Quartzite.

3. Olive grey till (muddy sand diamicton), well defined upper contact.
4. Brown till (muddy sand diamicton), containing numerous large angular, striated clasts; lowermost metre characterized by laminated brown mud and sandy muddy gravel, interbedded and can contain glacially deformed blocks of underlying units at lower contact.
5. Grey till (diamicton) interbedded with well sorted coarse sand and gravel, interbedded; unit occupies a channel structure and is discontinuous laterally; well defined upper contact.
6. Red brown till (muddy sand diamicton), characterized by rotten gneissic clasts.
7. Well sorted sand and gravel.

It is emphasized that the stratigraphic descriptions are preliminary and may be subject to change. The exposed sections have not been examined in detail.

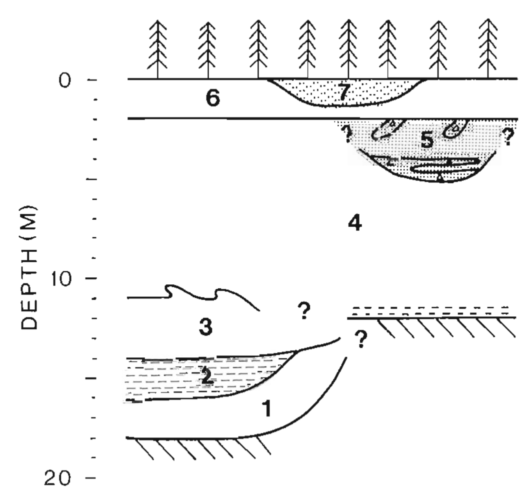

Figure 6. Composite Quaternary stratigraphic section at the Scully Mine, Wabush Labrador. Text offers a description of numbered units.

DISCUSSION AND CONCLUSIONS

The 1987 fieldwork both supports and expands on earlier presentations of glacial history in Labrador that describe complex change in ice flow trends with time. The widespread distribution of striae of several distinct ages, and the consistent patterns that they present over large (1000 $_s$ km^2) areas, along with the significant (> 100 km) distances of glacial transport demonstrated by indicator erratics indicate that flow trends represent important phases of the Laurentide Ice Sheet which could relate to independent and/or shifting dispersal centres. The sections represent an opportunity to develop a stratigraphic basis for the glacial history represented by surficial evidence and to establish a chronological control for it. Because the organic materials are central within Labrador-Quebec, and lie near the location of one or more major ice divides of the Laurentide Ice Sheet, they demonstrate interglacial conditions and indicate that a long period of glacial history is recorded.

In the Wabush area striae trends and the distribution of Labrador Trough debris indicates that surficial sediments are the composite product of more than one phase of ice flow and that the last direction of flow is not necessarily the predominant influence on drift composition.

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Uraniferous pyritic quartz pebble conglomerate and layered ultramafic intrusions in a sequence of quartzite, carbonate, iron formation and basalt of probable Archean age at Lac Sakami, Quebec

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Roscoe, S.M. and Donaldson, J.A., Uraniferous pyritic quartz pebble conglomerate and layered ultramafic intrusions in a sequence of quartzite, carbonate, iron formation and basalt of probable Archean age at Lac Sakami, Quebec; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 117-121, 1988.

Abstract

A quartz arenite unit at Lac Sakami near the La Grande hydroelectric site was traced 8 km by INCO geologists in the early seventies. The unit is in a siliciclastic succession up to 1700m thick that contains uraniferous pyritic quartz pebble conglomerate beds in its upper part. It is overlain by 700m of calc-silicate rocks and oxide-silicate iron formation capped by basalt. Compositionally banded ultramafic bodies within the quartzite were intersected in drillholes. These bodies are likely similar to a layered ultramafic intrusion exposed in a new waterline outcrop 5 km northeast and along strike from quartzite outcrops. The succession is bounded on the south by gneiss. Crossbeds, scours, pillows, and possible stromatalite domes in the sedimentary succession face northward, indicating that the gneiss probably is basement. Steep dips, interlayering with the volcanic strata, and near-vertical stretching lineations suggest that these platformal sedimentary rocks are part of the Archean Superior Province.

Résumé

L'unité de quartz sédimentaire au Lac Sakami près de l'aménagement hydroélectrique de la Grande a été suivie sur 8 km par des géologues de l'INCO au début des années 70. Il s'agit d'une succession silico-clastique dont l'épaisseur peut atteindre 1700 m et qui renferme des couches de conglomérat à galets de quartz pyritique uranifère dans sa partie supérieure. Elle est recouverte de 700 m de roches à silicates calciques et d'une formation ferrifère à oxydes et à silicates coiffée de basalte. Des massifs ultramafiques rubanés au sein du quartzite ont été recoupés par des forages. Ces massifs seraient semblables à une intrusion ultramafique litée affleurant au-dessus d'une nouvelle ligne d'eau à 5 km au nord-est et selon la direction d'affleurements de quartzite. La succession est limitée au sud par du gneiss. Des couches obliques, des marques d'érosion, des coussinets et des dômes possibles de stromatalite dans la succession sédimentaire sont orientés vers le nord, indiquant que le gneiss constitue probablement le socle. De forts pendages, des intercalations de couches volcaniques, ainsi que des linéations s'étendant presqu'à la verticale, semblent indiquer que ces roches sédimentaires de plate-forme font partie d'un terrain archéen de la province du lac Supérieur.

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INTRODUCTION

The number of recorded occurrences of extremely mature Archean quartz arenites in close association with ultramafic sills and/or flows has increased rapidly in recent years (e.g. Ashton, 1982; Donaldson and de Kemp, 1987; Covello et al., 1987). As a consequence, the ages of similar associations previously designated as Proterozoic on the basis of quartz arenite maturity warrant reconsideration. One such association, for which geochronologic data are as yet lacking, occurs within the Superior Province on the west side of Lac Sakami, Quebec (Avramtchev, 1982). This locality (53°11′ N, 76°55′ W) is about 80 km south of the La Grande hydroelectric site and the village of Radisson.

The quartzite unit (Fig. 1) contains numerous extensive lenses of uraniferous pyritic quartz pebble conglomerate that were discovered and explored jointly by INCO Limited and James Bay Development Corporation from 1971 to 1975. Managed by INCO, the exploration project included magnetic, electromagnetic and radiometric surveys and drilling totalling about 14 000 m in 65 holes distributed along a strike length of 8 km. A central 1 km section of the belt with an average width of 6 m was estimated to contain 4482 tonnes of uranium in 8.5 million tonnes of conglomeratic rock grading 0.052 % uranium distributed through 7 steeply-dipping zones to a depth of 300 m (Robertson et al., 1986).

The writers spent four days in August 1987 reviewing geological characteristics of the Lac Sakami pyritiferous conglomeratic uranium deposit, guided by detailed geological maps and drill log information provided by INCO Limited.

GENERAL GEOLOGY

Eastwardly-trending strata containing the Lac Sakami conglomeratic uranium deposit include up to 1700 m of siliciclastic sedimentary rocks bordered to the south by gneiss and to the north by calc-silicate rocks, banded iron formation and mafic volcanic rocks (Fig. 1). The calc-silicate rocks and iron formation are intercalated within a lens-like unit, predominantly calc-silicate to the south, that is up to 700 m thick in its central part but which thins westward to a few tens of metres. The sedimentary belt, which can be traced for more than 8 km to the west shore of Lac Sakami, is characterized by steeply dipping beds that generally face northward, although isoclinal folds are abundant in the iron formation.

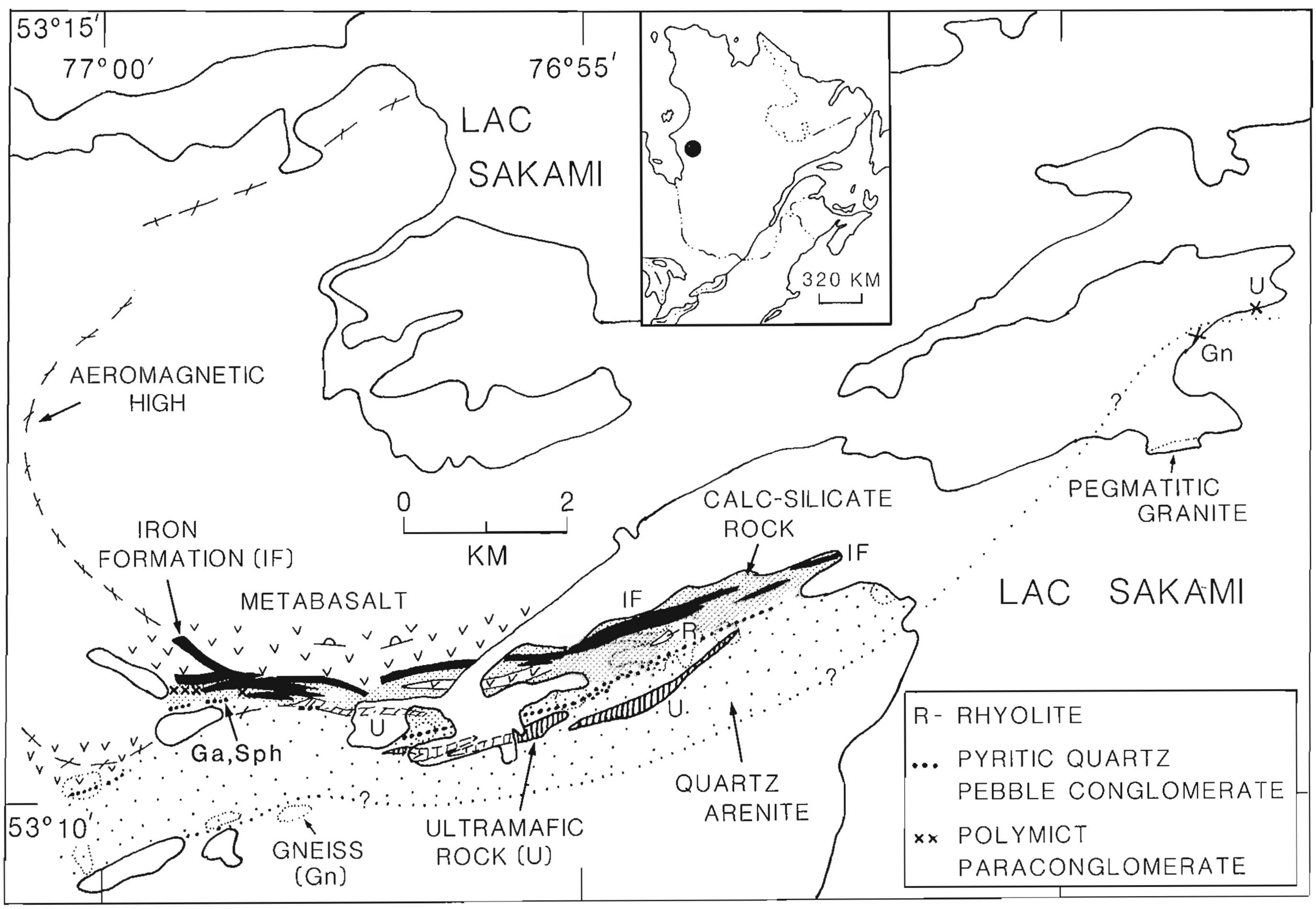

Figure. 1. Quartz arenite-ultramafic association at Lac Sakami, Quebec. Compiled from detailed geological maps prepared in 1972 by P.E. Fischer, INCO Limited.

SILICICLASTIC SEDIMENTARY ROCKS

Quartz arenite is a major component of the sedimentary succession. Fine- to medium-grained clastic textures are locally well preserved, and primary sedimentary structures such as crossbedding, scours amd ripple marks are abundant. These rocks are typically buff to greyish white on weathered surfaces, but fuchsitic mica imparts a brilliant green colour to parts of some outcrops. The quartz arenites generally are well sorted and rarely contain more than 5 percent matrix, now recrystallized to white mica. Heavy minerals include zircon, tourmaline, rutile and pyrite. Quartz-pebble conglomerates occur in thin beds and lenses interlayered with the quartz arenites. The conglomerates contain abundant pyrite and host most of the uranium mineralization (Fig. 2). Galena and sphalerite are locally present in quartz arenite and conglomerate near the western end of the belt. A pyrite-rich specimen was found to contain 96 ppb gold[1]. Closely packed pebbles define a stretching lineation that locally exceeds a ratio of 6:1, but the clasts are typically well rounded in views perpendicular to the stretching (Fig. 3). Many of the conglomerate beds occupy channels, and trough crossbedding is particularly abundant in the quartz arenite beds

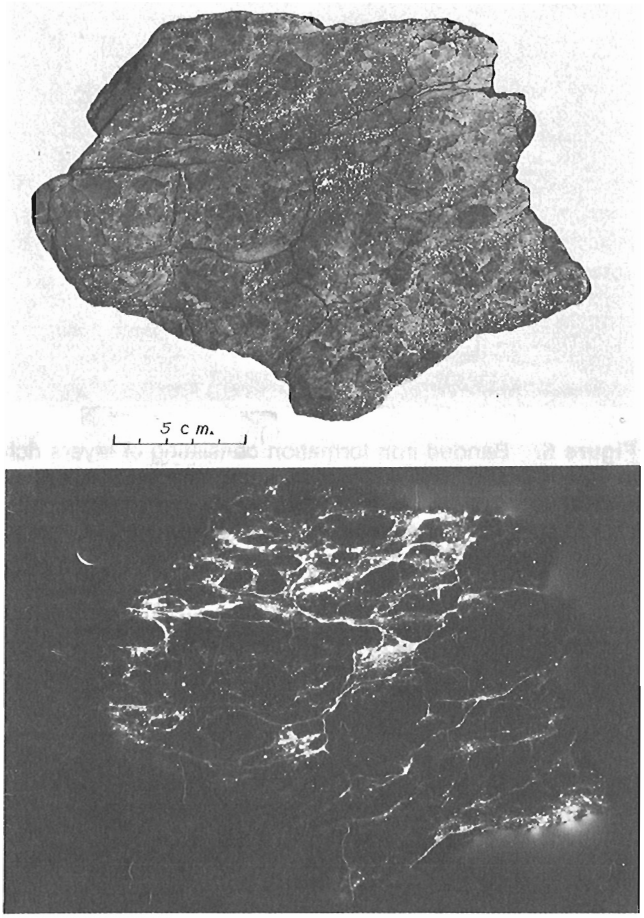

Figure 2. Pyritic uraniferous conglomerate as seen in a polished slab cut perpendicular to the stretching lineation shows an intact framework of well-rounded quartz clasts. The lower photograph is a radiograph of this slab.

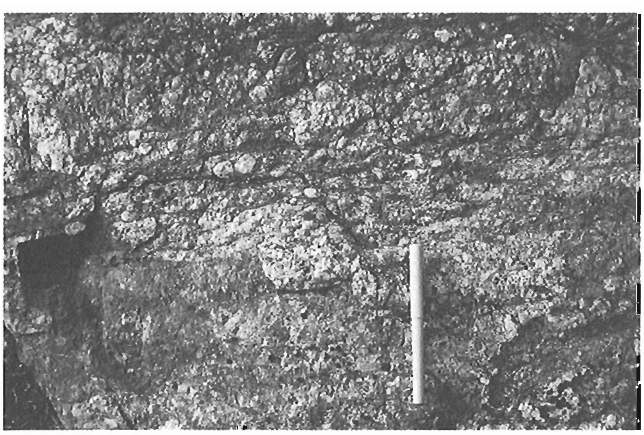

Figure 3. Clasts are distinctly elongate perpendicular to this outcrop view of pyritic uraniferous conglomerate. Pen is 15 cm long.

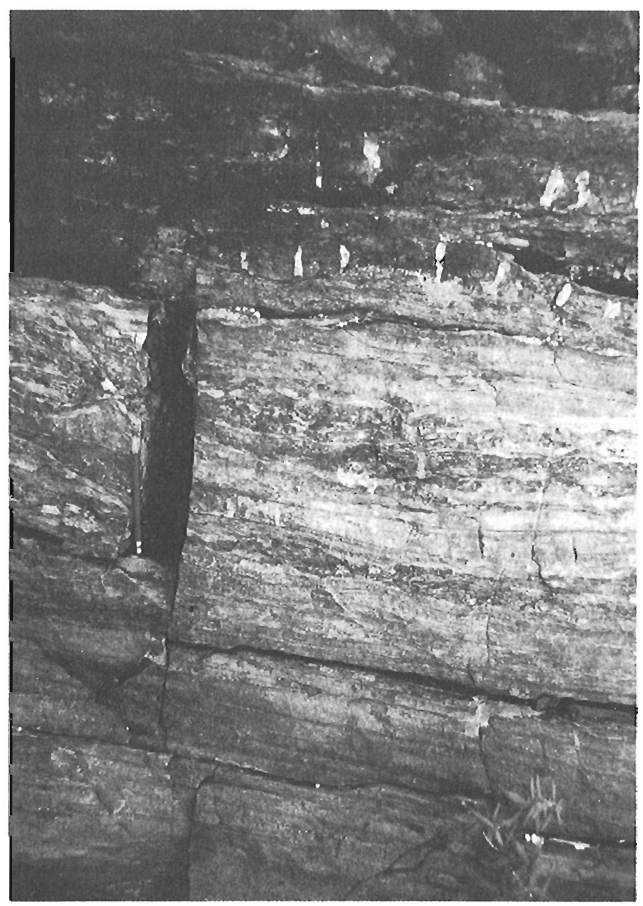

Figure 4. Quartz arenite overlain by calc-silicate strata containing crenulate laminations inferred to represent original stromatolitic bedding. The section seen here is about 1 m thick. Note quartz-filled tension gashes in upper part of outcrop.

[1] Neutron activation analyses, Bondar-Clegg, Ottawa

interbedded with the quartz-pebble conglomerates. Both sets of structures indicate a facing direction to the north. Quartz-phyric felsic flow rocks are locally intercalated with the siliciclastic strata.

CALC-SILICATE ROCKS AND METAPELITES

Metamorphosed calcareous strata containing abundant garnet and diopside directly overlie the conglomerate-bearing quartz-arenite unit. The calc-silicate porphyroblasts are generally parallel to bedding. Some clots of diopside may have been locallized by siliceous concretions in the protolith; other bedding-parallel clots show a distinct asymmetry of convex forms less than 10 cm in diameter and less than 5 cm in amplitude, and may represent metamorphosed domal stromatolites (Fig. 4). Amphibole has extensively replaced the pyroxene, and locally occurs in distinctive radial clusters on fracture surfaces. Intercalated schistose units rich in biotite, staurolite, quartz and feldspar probably were derived from siltstones and shales.

IRON FORMATION

Algoma-type banded iron formation is intercalated with both the calc-silicate strata and the overlying volcanic rocks. Bedding is well displayed by alternating silica-rich and iron-rich beds 0.5 to 5 cm thick (Fig. 5). Boudinage and pull-aparts are common, and at least two generations of folding can be recognized. Zones of intricate bedding-parallel folding may represent soft-sediment deformation. Iron silicates are abundant, but iron oxides and iron sulphides form distinct beds in some outcrops. Interdigitation with the calc-silicate rocks is reflected by locally abundant garnet and diopside. Polymictic conglomerate also is intercalated with the iron formation and calc-silicate rocks in the western part of the area, where these rock units thin and possibly wedge out in close proximity to the volcanic rocks. The polymictic conglomerates, which contain granitoid and mafic clasts as well as quartz pebbles, typically display a distinct stretching lineation (Fig. 6).

VOLCANIC ROCKS

The uppermost unit in the supracrustal succession consists of both massive and pillowed metabasalt flows and thin tuffaceous interbeds. Pillows, commonly as much as 2 m in diameter, display shapes indicative of northward facing in near-vertical flow units. This facing direction is substantiated by sparse drainage cavities (now filled by quartz) in the upper parts of some pillows. A stretching lineation plunging steeply to the southwest is well developed, and shearing has been localized in the tuffaceous layers. Interpillow hyaloclastite is abundant in some of the pillowed flows, and feldspar-phyric dykes up to 2 m thick cut several of the volcanic outcrops.

ULTRAMAFIC ROCKS

The quartz arenites contain numerous elongate ultramafic lenses, up to 200 m thick, that are known mostly from drillcore intersections. These bodies have been extensively

Figure 5. Banded iron formation consisting of layers rich in iron silicates and magnetite alternated with layers of recrystallized chert. Isoclinal folds are steeply plunging, parallel to stretching lineations in the associated strata. Pen is 15 cm long.

Figure 6. Polymictic conglomerate containing vein quartz, mafic clasts and granitoid clasts. Pen, 15 cm long, is aligned parallel to stretching lineation.

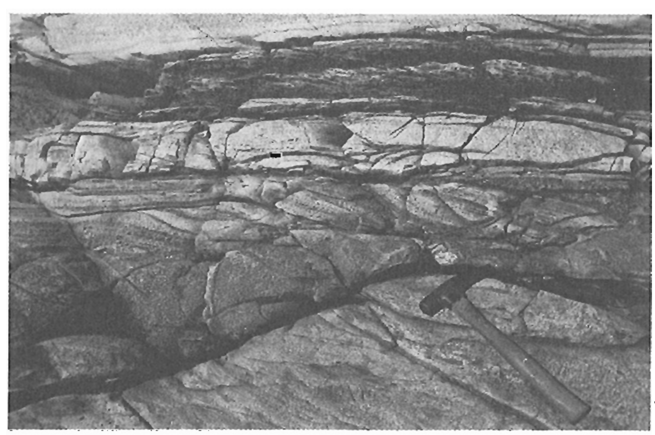

Figure 7. Shoreline outcrop of layered ultramafic rock, near east boundary of area examined, showing distinct scours and crossbedding in cumulate layers.

altered to serpentine-talc-magnetite assemblages, but relicts of olivine with unaltered cores are locally preserved. One shoreline outcrop exposed near the east boundary of the area examined displays abundant scours and crossbedding (Fig. 7). These remarkably well-preserved structures indicate convection of crystal laden magma within a body of appreciable thickness. This supports the view that most of these bedding-parallel bodies are sills, although some drillcore samples display possible spinifex textures suggestive of a flow origin.

DISCUSSION

The possibility that the extremely mature quartz arenites and intercalated uraniferous conglomerates at Lac Sakami might be of Proterozoic age was suggested by Robertson et al. (1986). However, as summarized here, these strata appear to be an integral part of an Archean greenstone belt. Their structural conformity, including a well-developed near-vertical stretching lineation, and marginal intercalation of the sedimentary succession with volcanic flows, support this interpretation. The Lac Sakami occurrence can thus be regarded as a strong candidate for addition to the growing number of Archean quartz arenites that are associated with ultramafic rocks.

ACKNOWLEDGMENTS

We are indebted to the geological staff of INCO Canada Limited for freely providing access to unpublished maps, reports and drill logs related to their exploration program at Lac Sakami. Travel and field expenses for JAD were funded in part by NSERC Grant A5536. Figure 1 was draughted by Kim Nuguyen. The paper was reviewed by O.R. Eckstrand.

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Stratigraphy of the Warren Creek Formation, Moran Lake Group, Central Mineral Belt of Labrador[1]

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North, J.W. and Wilton, D.H.C., Stratigraphy of the Warren Creek Formation, Moran Lake Group, Central Mineral Belt of Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 123-128, 1988.

Abstract

The early Proterozoic Moran Lake Group unconformably overlies Archean granitoid basement in the Croteau Lake area of central Labrador. The group is composed of the Warren Creek Formation and the Joe Pond Formation. The Warren Creek Formation can be divided into two members. The lower member overlies the basement granitoids and consists mainly of green sandstone and siltstone with interbedded dolostone, chert and quartz arenite. The upper member contains graphitic black shale with greywackes and grey siltstone, and minor banded iron-formation, arkose and vesicular basalt. The shale contains some syngenetic pyrite interbeds with lesser sphalerite, chalcopyrite and pyrrhotite. The uppermost unit in the upper member is a buff to white pebbly arkose. The massive and pillowed basalt of the Joe Pond Formation conformably overlies the Warren Creek Formation. A basal conglomerate of the middle Proterozoic Bruce River Group unconformably overlies both formations of the Moran Lake Group.

Résumé

Le groupe de Moran Lake du Protérozoïque inférieur recouvre en discordance le socle granitoïde de l'Archéen dans la région du lac Croteau au centre du Labrador. Le groupe est composé de la formation de Warren Creek et de la formation de Joe Pond. La formation de Warren Creek peut être divisée en deux membres. Le membre inférieur recouvre les roches granitoïdes du socle et consiste principalement de grès et de microgrès verts auxquels sont intercalées des roches sédimentaires formées de dolomie détritique du chert et du quartz sédimentaire. Le membre supérieur contient du schistes argileux noir de nature graphitique avec des grauwackes et du microgrès gris et, dans une moindre mesure, une formation ferrifère rubanée, de l'arcose et du basalte vésiculaire. Le schiste argileux contient des interstratifications sulfurées syngénétiques et, de petites quantités de sphalérite, de chalcopyrite et de pyrrhotite. L'unité supérieure du membre supérieur est une arkose cailloutteuse dont la couleur varie de chamois à blanc. Les basaltes massifs et en coussins de la formation de Joe Pond recouvrent en concordance la formation de Warren Creek. Un conglomérat basal du groupe de Bruce River du Protéropzoïque moyen recouvre de façon discordante les deux formations du groupe de Moran Lake.

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INTRODUCTION

Bedrock mapping, on a scale of 1:20 000, was undertaken in the Moran Lake Group of the Central Mineral Belt (CMB) of Labrador during summer of 1986. The area comprises about 100 km^2 north of Croteau and Pocket Knife Lakes (54° 27' N, 61° 10' W) and is located approximately 135 km NNW of Goose Bay (Fig. 1). Access is gained by float-equipped fixed wing aircraft, or by helicopter, from Goose Bay.

The Moran Lake Group has been the focus of intermittent exploration activity, since the late 1920s, by adventurous prospectors and mining companies searching for a variety of economic mineral deposits. The earliest explorationists recognized a thick sequence of mineralized black shales north of Pocket-knife Lake (Halet, 1946). This sequence, which was later defined as part of the Moran Lake Group (Smyth et al., 1975, 1978), and named the Warren Creek Formation (Ryan, 1984), has often been regarded as a potential host for pyritic, sediment-hosted Zn-Cu-Pb(Ag, Au) deposits. Mineralized outcrops of Warren Creek Formation shales are numerous in the study area; however, previous attempts to delineate economic concentrations of precious and/or base metals in the shales by stripping, trenching, and diamond drilling have been unsuccessful.

Previous work

Halet (1946) described the Croteau Series, dividing it into a lower marine succession (later renamed the Moran Lake Group by Smyth et al., 1978), disconformably overlain by an upper red quartzite and volcanic succession (later renamed the Bruce River Group by Smyth et al., 1975).

Fahrig (1959) redefined the Croteau Series as the Croteau Group. Fahrig's Croteau Group consisted of a lower succession of pyritic black shales, greywacke, quartzite, dolomite, and basalt, and an upper succession of conglomerate, sandstone, and porphyritic volcanic rocks. No mention was made of a possible unconformity between the upper and lower Croteau Group.

In 1974-75 the Newfoundland Department of Mines and Energy conducted a regional mapping survey in the region of the Croteau Group. Smyth et al. (1975, 1978) subdivided the Croteau Group into the Moran Lake Group and the Bruce River Group on the basis of the angular unconformity between the lower marine and upper continental successions. The Archean-Aphebian unconformity at the base of the Moran Lake Group was recognized as a depositional contact.

Ryan (1984) summarized the geology and economic mineralization of the Moran Lake and Bruce River groups,

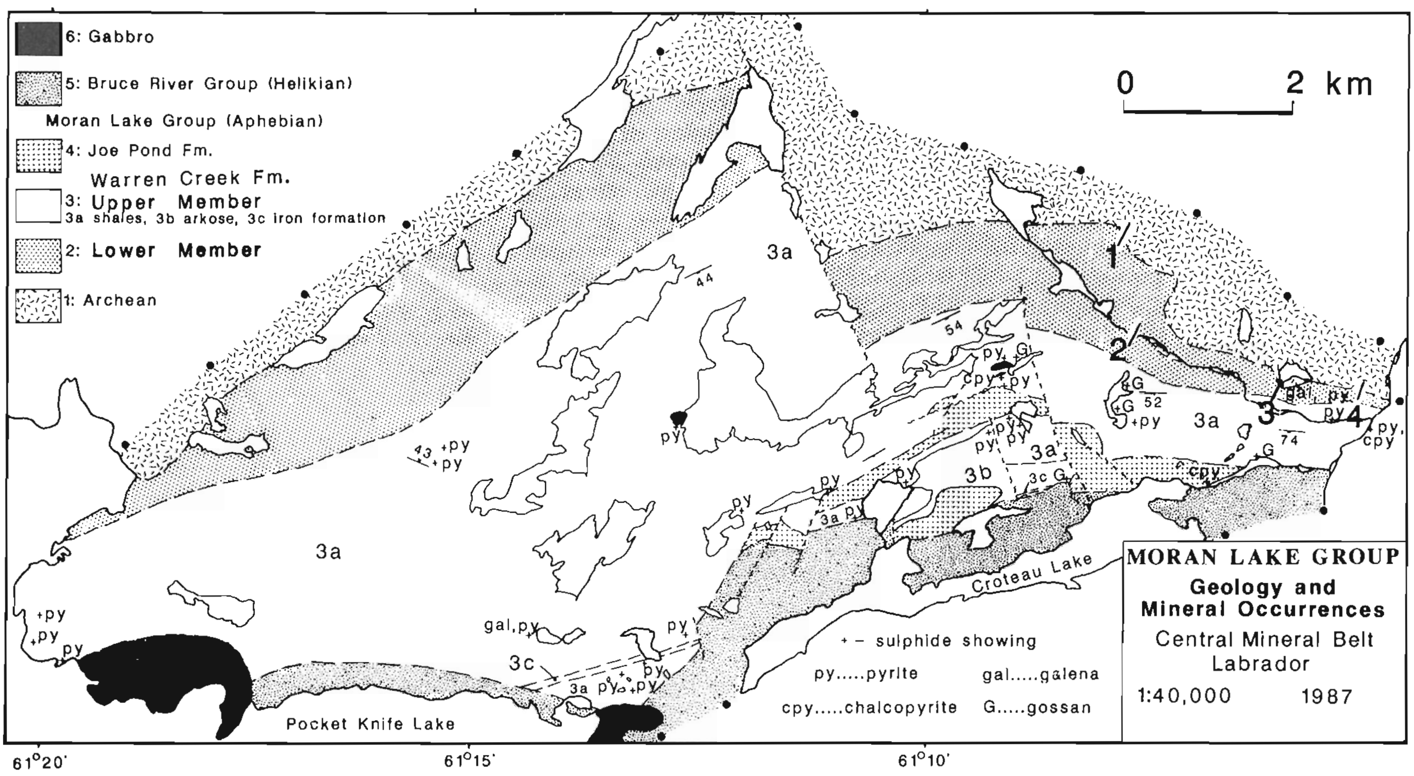

Figure 1. Geology and mineral occurrences of the Moran Lake Group in the Croteau Lake — Pocket Knife Lake area, central Labrador. The numbers on the eastern portion of the map are the locations of the stratigraphic sections from Figure 2, viz. No.1 = Meathook Lake section; No. 2 = Archean Window; No. 3 = Moon pond; and No. 4 = Warren Creek.

subdividing the Moran Lake Group into two formations: the Warren Creek Formation, consisting mainly of slate, mudstone and siltstone, and the overlying Joe Pond Formation, consisting of pillowed and massive mafic volcanics.

The present study (the basis of a MSc. thesis by North at Memorial University of Newfoundland) is to describe the setting of the mineralization within the Warren Creek Formation and to quantify the economic potential of the formation by presenting stratigraphic, petrographic, structural, and lithogeochemical data on previously known and newly discovered areas of exposed mineralization. Some results from the study have been reported in North and Wilton (1987a, b) and in unpublished reports by Wilton.

REGIONAL SETTING

The Moran Lake Group is a 3 to 5 km wide, 85 km long belt of Aphebian supracrustal strata, which extends from Pocket-knife Lake (54° 25' N, 61° 20' W) to the Kanairiktok River (54° 55' N, 60° 27' W) (Ryan, 1984). The basal Moran Lake Group strata were deposited in a marine shelf type environment, unconformably on massive to foliated Archean granite and tonalite (the Kanairiktok Intrusive Suite) of the Nain Structural Province. The Moran Lake Group is in turn unconformably overlain by the Paleohelikian Bruce River Group, a terrestrial sequence marked by a thick basal conglomerate and numerous red tuffaceous sandstone beds (Ryan, 1984).

The Moran Lake Group underwent polyphase deformation prior to the deposition of the Bruce River Group. The deformation has been assigned to the Hudsonian Orogeny (Smyth et al., 1975, 1978) about 1750 Ma, and is characterized by east-trending isoclinal F_1 fold axes and an associated penetrative slaty cleavage (S_1), which are overprinted by moderate to steeply plunging F_2 folds with axes trending southwest to southeast. The D_1 event pervasively deformed the Moran Lake Group; the D_2 strain fabric is not as obvious, forming more of a simple crenulation which is particularly well developed in areas of contrasting lithologies.

Recently (U. Scharer pers. comm., 1987) a U-Pb zircon age of 1649 Ma has been obtained from a volcanic member of the Bruce River Group. Since this unit unconformably overlies the Moran Lake Group and is also unaffected by the "Hudsonian" deformation in this latter group, the Moran Lake Group was deposited prior to 1649 Ma.

MAP UNITS

Archean basement (unit 1)

Massive to weakly foliated, medium grained, beige tonalite and granodiorite underlie the Moran Lake Group (Fig. 1). These plutonic rocks, the Kanairiktok Intrusive Suite, are part of the Kanairiktok Valley Complex of the Nain Structural Province (Ryan, 1984). They consist of anhedral to subhedral interlocking medium grained quartz and feldspar, mottled with fine- to medium-grained interstitial muscovite and chloritized amphibole. Sulphides occur both as accessory minerals (generally less than 1 %), and as secondary mineralization, along with calcite, in fractures. Outcrops of basement rocks near the unconformity with the overlying Moran Lake Group are commonly carbonatized and limonitic. Partial to total replacement of feldspar with rusty weathering carbonate is ubiquitous, and the rock has a "rotten" (saprolitic) appearance indicating the development of a regolith below the unconformity.

Warren Creek Formation

The Warren Creek Formation is a 1200 to 4600 m thick sequence of clastic and chemical metasediments, and minor basalt, which unconformably overlies Archean basement and is conformably overlain by basalts of the Joe Pond Formation (Ryan, 1984). The estimated thickness of the Warren Creek is probably exaggerated by as much as 50 % due to east-trending overturned isoclinal F_1 folds, and to some extent by buckle folding and concomitant thickening of less competent lithologies. Therefore the stratigraphic widths given, represent measurements from contact to contact without compensation for tectonic redistribution.

We subdivided the formation into informal upper and lower members.

Lower member (unit 2)

The basal member of the Warren Creek Formation consists of 180 to 1300 m of beige to light green, laminated, fine-grained feldspathic sandstone and siltstone, with minor red to mauve varieties, and intercalated quartz arenite, chert, and dolostone. The sandstones strike roughly east northeast and dip 60 to 80° S. With the exception of a few steeply overturned beds, the formation consistently faces south. Graded and cross bedding on a 1 mm to 5 cm scale are well preserved throughout the sequence. Calcareous and cherty beds and lenses are common, with dolostone and associated calcareous siltstone proximal to the Archean-Aphebian unconformity.

Upper member (unit 3)

A decrease in relief and the number of outcrops mark the contact between the basal siliceous epiclastics and the conformably overlying recessively weathered black shales of the upper member which occupy the southern and central part of the map area. The sequence is 543 to 3350 m thick, and consists primarily of fine grained and finely bedded graphitic black shale, black to brown poorly sorted fine grained greywacke, and grey siltstone. Massive grey quartz arenite, buff to white arkose, banded iron-formation, and dark green vesicular basalt flows occur near the top of the sequence. The shales are composed of aphanitic clay minerals, chert, graphite, and fine detrital quartz. Pyrite is ubiquitous, and as a result shale outcrops are nearly always covered with a distinctive rusty gossan. Massive syngenetic pyrite beds (± pyrrhotite, chalcopyrite and sphalerite) were observed in a few localities. These beds were complexly deformed by penecontemporaneous slumping and regional folding and reach thicknesses of over 1 m.

Preliminary stratigraphic reconstruction of the upper member indicates a gradual change in depositional environment and tectonic setting during deposition of the upper portions of this formation. Accumulation of thick deposits of black pyritic and graphitic shales occurred in deep reducing waters at the base of the sequence in a period of quiescence. This changed to the deposition of thin vesicular basalt flows, buff to white arkose, and jasperoidal iron-formation in relatively shallower water and an active tensional tectonic regime.

Pebbly arkose bed (unit 3b)

Buff to white arkose outcrops in the core of a mesoscopic anticline north of Croteau Lake. Previous workers (Hansuld, 1958; Smyth et al., 1975; Ryan, 1984) interpreted these rocks as basal Warren Creek Formation sediments which were upwarped into the core of an anticline, but the present study revealed that the correlation is unfounded. When compared to the basal arkose, the rocks in the core of the anticline are a little coarser grained and massively bedded, and contain quartz and feldspar pebbles . Primary sedimentary structures are also uncommon and no cherty or calcareous transitional units have been observed in this upper unit.

A re-interpretation of the rocks in the core of the anticline is warranted based on tha above data. These rocks were deposited during a change in depositional and tectonic environments at the top of the upper Warren Creek member. Vesicular basalt flows outcrop stratigraphically below the arkose and thus attest to a dynamic tensional tectonic regime with the accumulation of sediments and volcanics in relatively shallow water. Subsequent deformation of the Moran Lake Group folded the rocks in this area into a broad east-trending anticline.

Joe Pond Formation (unit 4)

A 100 to 300 m thick sequence of light to dark green, fine grained massive and pillowed basalt conformably overlies the Warren Creek Formation. Remnant fine grained plagioclase phenocrysts are rarely preserved due to the pervasive alteration of the basalts to chlorite and/or assemblages containing quartz, carbonate, and sulphides. Pillow structures and interstitial clastic and chemical interflow sediments are common. The basalts are highly fractured and carbonatized near the Aphebian- Helikian unconformity to the extent that highly fractured basalt is locally exposed as a regolith directly below the basal conglomerate of the Brown Lake Formation.

Bruce River Group (unit 5)

Red to buff boulder conglomerate of the Brown Lake Formation, the middle subdivision of the Bruce River Group (Ryan, 1984), is in contact with highly altered Joe Pond basalt on the northeast shore of Croteau Lake, defining the Aphebian-Helikian unconformity. The conglomerate is also in unconformable contact with black shales of the upper Warren Creek member. The conglomerate is a maximum of 100 m thick, varies from clast to matrix-supported, and contains poorly sorted angular to rounded boulders and pebbles of iron-formation, quartz, arkose and siltstone from the underlying Moran Lake Group. Basalt fragments were not observed. Felsic volcanic rocks, identified as dust tuff or porcellanite beds by Ryan (1984), are intercalated with the conglomerate in intervals up to 30 m thick. The tuff contains 1 to 15 cm beige and purple bands, which may represent primary beds, with 0.1 to 2 mm feldspathic sericitized lapilli or spherulites in a brittle siliceous matrix.

Fine- to medium-grained red, mauve, and buff tuffaceous sandstone overlies the basal conglomerate and felsic volcanic rocks of the Brown Lake Formation. These rocks dip gently and form the steep, rocky, southern shoreline of Croteau Lake. The rock is finely bedded on a scale of 1 mm to 1 cm, and contains well developed cross stratification and graded bedding. Selectively sericitized light green to buff horizons a few millimetres to centimetres thick, are present throughout the sandstones.

Gabbro (unit 6)

Dark green, massive, medium grained gabbro plugs and dykes intrude all other units in the area. The dykes are 20 to 80 m thick, but rounded gabbro plugs are up to 1400 m in diameter. The gabbros are fresh, but may contain secondary quartz-carbonate-sulphide mineralization in fractures and dyke margins. They consist of medium- to coarse-grained, interlocking amphibole laths in a finer matrix of plagioclase, biotite, and quartz. One biotite-phyric gabbro was observed.

GEOLOGICAL RELATIONSHIPS

Laminated sandstone and siltstone of the lower member of the Warren Creek Formation unconformably overlie Archean basement granitoids. The unconformity is exposed in three locations in the eastern map area (see Fig. 1, 2):

(1) Along Warren Creek, dark green-grey silty shale is in contact with buff, massive, limonitic granite. The siltstone grades into calcareous siltstone and dolostone which reach a maximum thickness of 40 m. The calcareous sediments are overlain by 20 m of thickly bedded grey quartz arenite, which is in contact with fissile black shale.

(2) The unconformity is well exposed on the southern shore of Moon Pond. Leucocratic, carbonatized and fractured Archean granite basement is in contact with 2 m of massive grey quartz arenite, overlain by 5 to 10 m of dark green to grey siltstone and chert, and 3 to 7 m of laminated brown silty dolostone and calcareous siltstone. A regolith is developed in the underlying Archean intrusive which is characterized by carbonatized feldspars, and secondary carbonate-sulphide mineralization in fractures. The regolith penetrates the basement rock for 1 m.

(3) Four hundred metres south of Meathook Lake, 20 m of exposure gap separate red calcareous slate and siltstone, and brown dolostone from leucocratic Archean intrusive rocks. The contact is interpreted to be unconformable. The Archean granitoid is cut by pegmatitic dykes and veins which do not penetrate into the overlying Moran Lake Group.

Conglomerate in the basal Brown Lake Formation of the Bruce River Group is in unconformable contact with the upper member of the Warren Creek Formation and Joe Pond Formation. The angular nature of the unconformity was observed at a number of locations in the eastern part of the area. The unconformable relationship is evident from:

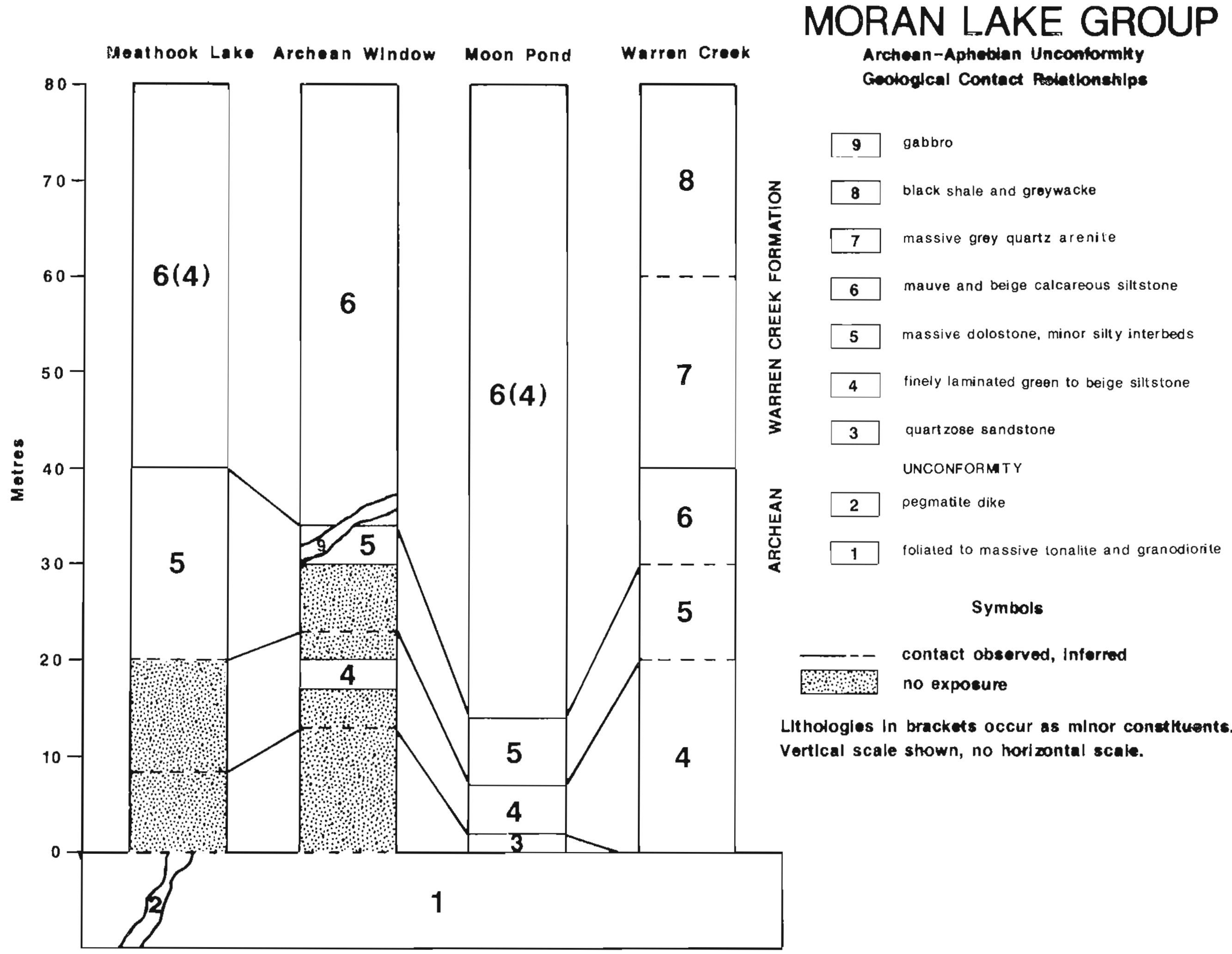

Figure 2. Stratigraphy and basement relationships of the lower part of the Warren Creek Formation, Croteau Lake — Pocket Knife Lake area, Labrador. Locations of the sections are shown on Figure 1.

(i) The presence of deformed clasts of Moran Lake Group sedimentary rocks within the Brown Lake Formation conglomerate (Smyth et al., 1978)

(ii) An angular bedding relationship between ENE-trending iron-formation and quartz arenite in the Warren Creek Formation which in one area is in fault contact with the overlying northeast-trending conglomerate of the Brown Lake Formation.

(iii) The difference in the bedding attitude of the groups. The Moran Lake Group strikes 58-88° and dips steeply south at 59-71° below the gently tilted Bruce River Group which strikes 66° and dips 38° SE.

MINERALIZATION

Scattered sulphide showings occur throughout the Moran Lake Group, but the most pervasive style of mineralization occurs within the Warren Creek Formation shales. Newfoundland Department of Mines Mineral Inventory Files list over 16 black shale-hosted pyrite occurrences within the Warren Creek Formation in the area. Massive sulphide horizons within the shales are an important rock-forming constituent of the formation, and exhibit remarkable consistency in form and mineralogy. Individual occurrences are similar, and contain pyrite with traces of pyrrhotite, chalcopyrite, and sphalerite. Two massive sulphide zones were investigated in detail by Brinex Limited in 1958, after the discovery of two total heavy metal geochemical anomalies north of Croteau Lake. Two geochemical grids were cut and limited geological follow-up work was completed (Hansuld, 1958). These sulphide showings were resampled during the present work and the samples are being analyzed. Preliminary results indicate that some contain in excess of 750 ppm zinc (North and Wilton, 1987a, b).

New sulphide horizons were discovered while completing the fieldwork in 1986; one (the Teuva Showing) contains up to 2.69 % zinc in a stratabound sphalerite horizon (North and Wilton, 1987a, b).

CONCLUSIONS

The Warren Creek Formation of the Moran Lake Group is divisible into two members. The lower member, which rests with marked unconformity on Archean tonalite-granodiorite, probably represents marine shelf deposition. The graphitic shale in the upper member contains numerous syngenetic massive sulphide interbeds which contain significant Zn concentrations; vesicular basalt flows were also discovered near the stratigraphic top. The uppermost pebbly arkose bed occurs within the core of a large anticline. The Joe Pond Formation conformably overlies the Warren Creek Formation and both these units are in turn unconformably overlain by the Bruce River Group.

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Georgian Bay geological synthesis: Key Harbour to Dillon, Grenville Province of Ontario[1]

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Abstract

Four major rock associations have been recognized in a corridor across tectonic strike of Britt domain, Central Gneiss Belt. Individual associations are dominated by gneiss groupings of contrasting compositions and are separated by younger, foliated metagranitoid plutons and in some cases by ductile shear zones. Metasedimentary rocks and gneiss of probable supracrustal origin are more abundant than previously recognized in two of these associations whereas tonalitic orthogneiss is a characteristic component of a third. The following history is valid throughout the transect: early metamorphism and deformation of unknown age; emplacement of mid-Proterozoic (ca. 1.45 Ga) granitoids along or close to association boundaries; regional emplacement of mafic dykes (lithospheric stretching ?); formation of northwest-trending folds parallel to southeast-trending stretching lineation, ductile thrusting and regional high grade metamorphism in the Grenvillian compressional event. It is possible that the four rock associations represent terranes that were assembled before intrusion of the mid-Proterozoic plutonic rocks.

Résumé

Dans un corridor traversant la direction tectonique du domaine de Britt, dans la zone gneissique centrale, quatre grandes associations rocheuses ont été déterminées. Les associations individuelles sont dominées par des regroupements gneissiques à compositions contrastantes et elles sont séparées par des plutons métagranitoïdes foliés plus jeunes et, dans certains cas, par des zones de cisaillement plastique. Les roches et les gneiss métasédimentaires d'origine probablement supracrustale sont plus abondante qu'il avait été antérieurement constaté dans deux de ces associations tandis qu'un orthogneiss tonalitique constitue une composante caractéristique d'un troisième. La chronologie suivante est valable sur toute la longueur du transect: métamorphisme ancien et déformation d'âge indéterminé; intrusion de granitoïdes datant du milieu du Protérozoïque (ca. 1,45 Ga) le long des limites d'association ou à proximité de celles-ci; une mise en place régionale de dykes mafiques (étirement lithosphérique ?) formation de plis à direction nord-ouest parallèlement à une linéation d'étirement à direction sud-est, charriage plastique et métamorphisme régional de forte intensité dans l'épisode de compression grenvillien. Il est possible que les quatre associations rocheuses représentent des terranes qui avaient été assemblées avant l'intrusion des roches plutoniques du milieu du Protérozoïque.

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INTRODUCTION

This report outlines the preliminary results of the first of three seasons' mapping of a corridor across tectonic strike in the Central Gneiss Belt of the Grenville Province, Ontario. The work was funded under Project C.2.3, Georgian Bay Geological Synthesis, part of the Canada-Ontario Mineral Development Agreement. The corridor extends from Key Harbour, situated close to the southeast boundary of the Grenville Front Tectonic Zone, southeast to Paleozoic cover near Port Severn. The traverse takes advantage of excellent rock exposure along the shores of Georgian Bay. An important aim of the traverse is to explore further the tectonics of this part of the Grenville Province, building primarily on the reconnaissance work of Davidson et al. (1982). In 1987, mapping was completed in the northeast part of the corridor, a section some some 60 km long and 10 to 15 km wide extending from Key Harbour to Dillon. This section, described below, is entirely within the lithotectonic unit referred to as Britt domain.

SUMMARY OF FINDINGS

The part of Britt domain covered by the traverse is divisible into four distinct rock groupings, here termed gneiss associations. These are separated spatially from one another by foliated plutons of large feldspar-bearing granitoid rocks of which the most prominent is the Britt pluton (Fig. 1). The associations are composed of gneissic rocks that generally have a more complex history than the younger, bounding granitoids. Some of the associations contain a higher proportion of demonstrably metasedimentary rock than was previously recognized as well as much gneiss of probable supracrustal origin. These are in striking contrast to gneiss associations having demonstrably plutonic protoliths. Both the host gneiss associations and the younger granitoid plutons were widely intruded by a swarm of mafic dykes and all were affected by pervasive, southeast-trending upright folding with axes parallel to a regionally developed, low-angle, southeast-plunging stretching lineation. These folds are both cut by and deform northeast-striking shear zones.

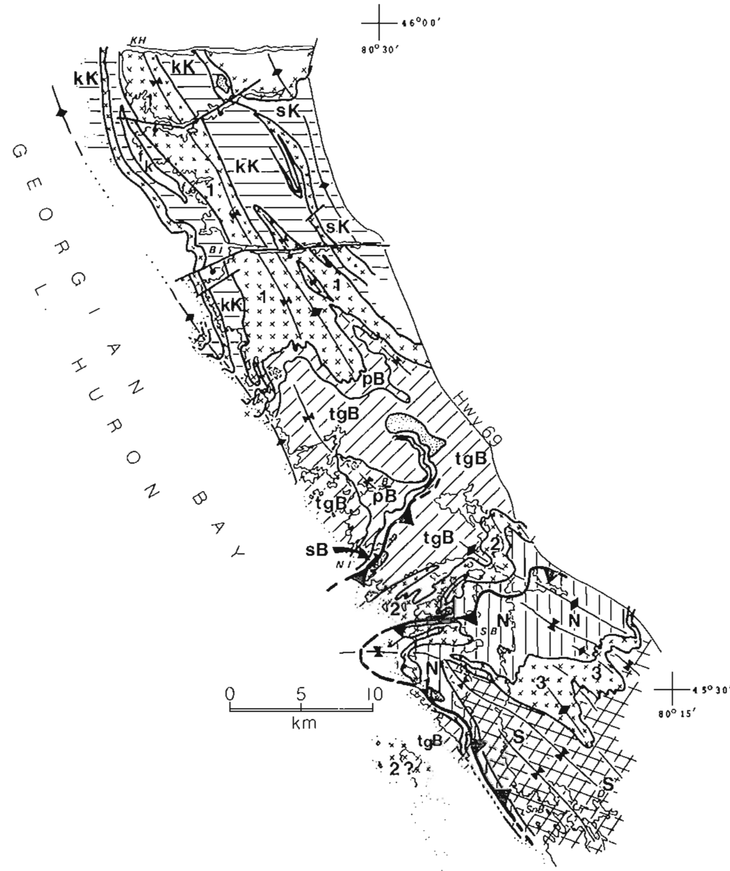

Figure 1. Geological map of part of Britt domain, coast of Georgian Bay between Key Harbour and Dillon, Ontario. Key Harbour gneiss association: kK — Key Harbour gneiss, fK — Free Drinks mafic gneiss, sK — Still River mafic gneiss; Bayfield gneiss association: tgB — tonalite and granite orthogneiss, pB — pink leucogneiss, sB — metasedimentary gneiss; N — Nadeau Island gneiss association; S — Sand Bay gneiss association. Single cycle granitoid rocks (crosses): 1 — Britt granodiorite, 2 — Pointe-au-Baril complex, 3 — Shawanaga granodiorite. Stipple — metagabbro; heavy line with teeth — mylonitic gneiss belt, teeth on upper plate; heavy line — fault, downthrown side indicated. Fine dotted line — limit of islands in Georgian Bay. KH — Key Harbour, BI — Byng Inlet, B — Bayfield, D — Dillon, NI — Nares Inlet, SB — Shawanaga Bay, SNB — Sand Bay.

GNEISS ASSOCIATIONS

Four associations of gneiss have been recognized, named from northwest to southeast: Key Harbour (KHA), Bayfield (BA), Nadeau Island (NIA) and Sand Bay (SBA) (Fig. 1). Of these, the KHA is lithologically the most varied, and together with the SBA contains much metasedimentary and suspected metasupracrustal gneiss.

Key Harbour Association

The Key Harbour association is formed of several gneiss assemblages, each of which is quite varied but contains constantly associated gneiss types. These are: Key Harbour gneiss, Free Drinks mafic gneiss and Still River mafic gneiss. The KHA extends as far south as the structurally lower (northern) boundary of the Britt pluton, itself beneath the BA (Fig. 1). Key Harbour gneiss consists of gneiss of indisputable sedimentary origin together with much biotite gneiss (with more plagioclase than K-feldspar) of probable supracrustal origin and some granitoid orthogneiss. The distinctive metasedimentary rocks are composed of lithologic combinations unlike any found elsewhere this summer (1987). The most characteristic rock types are micaceous quartzite, leucocratic quartzofeldspathic gneiss with sillimanite or late muscovite in K-feldspar-bearing leucosome, rare pods or layers of calc-silicate rock, grey biotite leucogneiss with or without garnet that lacks any hint of deformed igneous texture, and minor hornblende gneiss and amphibolite. The granitoid orthogneiss includes pink, leucocratic biotite granite as well as minor amounts of rock with more intermediate composition. The relationship of orthogneiss and metasedimentary gneiss is not clear; both rock types may contain mafic inclusions or metamorphosed dykes.

The Free Drinks mafic gneiss is an important plutonic grouping lying along the west side of the Britt pluton (Fig. 1). It is composed of gabbroic, leucogabbroic, tonalitic and minor granitic gneiss of unmistakeable plutonic aspect, although it has pervasive metamorphic textures and, notably, lacks relict igneous texture. Its heterogeneous character results from complex internal crosscutting relationships. It is cut by leucogranite that resembles a granitic phase of the Key Harbour gneiss. Locally these textures and relationships are smeared out by regional transposition (D_2), forming mafic straight gneiss. The Free Drinks gneiss is intimately associated with metasedimentary gneiss not separable at map scale.

The Still River mafic gneiss lies in the extreme east of the KHA (Fig. 1). It is composed of amphibolite and intermediate, mesocratic gneiss (hornblende-biotite-epidote-plagioclase, plagioclase-biotite) in which plutonic textures are rare. All types, including variable amounts of pink leucogneiss, are commonly interlayered and, together, strongly deformed resulting in a straight gneiss of polymodal composition. These gneisses generally display polycyclic deformational (D_1 plus D_2) and metamorphic histories which in part distinguishes them from granitoid metaplutonic rocks with single cycle histories (D_2 only).

Bayfield association

The Bayfield association extends from the upper contact of the Britt pluton south to Shawanaga Bay where it is in contact with the Pointe-au-Baril complex (Fig. 1). At the upper (southern) boundary of the Britt pluton there is a large swath of pink and grey biotite leucogneiss of dominantly supracrustal origin. This is replaced southward by the Bayfield gneiss which is characteristic of this association. It is a migmatitic metatonalite (plagioclase-hornblende-biotite-quartz$\pm$garnet) with pervasive, streaky, deformed plutonic texture. The metatonalite is cut by small bodies of pink, leucocratic, alaskitic biotite granite. It is also associated with two varieties of larger bodies of leucocratic pink granitic rocks: biotite granite with large, recrystallized feldspar which is also cut by pink alaskite and intermixed leucocratic alaskitic biotite granite of plutonic aspect and a leucocratic equigranular biotite-magnetite gneiss (similar to that overlying the Britt pluton). The latter forms a body in the centre of the association and underlies a belt of metasedimentary gneisses that are quite different to those of the KHA. They comprise garnet-rich quartzofeldspathic rock (perhaps with kyanite?) interlayered with (semi-)pelitic gneiss (garnet-biotite-quartz-feldspar). This unit has been traced inland and northward; it forms a useful marker unit that outlines the structure in this region. The metasedimentary unit also coincides with an important shear zone (Fig. 1).

Nadeau Island association

This association is separated from the Bayfield association to the north by the large Pointe-au-Baril plutonic complex of single cycle plutonic gneiss, itself enclosing gneiss of Nadeau Island type (Fig. 1). It is bounded on the south in part by the Shawanaga granodiorite pluton and in part by the northwest-trending extremity of the Sand Bay association. The Nadeau Island gneiss is a leucocratic granodiorite gneiss intermixed with gneissic metatonalite; these two components resemble, respectively, components of the KHA and the BA. Pink leucocratic gneiss is also present, although less voluminous than in the BA. Metasedimentary rocks form a substantial component of this association, comprising pelitic and semipelitic types, calc-silicate rocks and a volumetrically important, amphibole-bearing garnet-biotite gneiss (calc-pelite?) characteristic of the metasedimentary rocks of the association.

Sand Bay association

The Sand Bay gneiss association (Fig. 1) is a remarkable group of rocks quite distinct from all others in the map area in that it appears to be almost entirely formed of gneisses of supracrustal origin. These include biotite-plagioclase migmatites, quartz-rich muscovite-plagioclase-quartz gneiss, quartzite, pink leucocratic gneiss, calcareous gneiss, rare marble and substantial units of hornblende gneiss and amphibolite.

MAFIC DYKES

Small bodies of mafic rock, here referred to loosely as mafic dykes, fall into two broad groupings, the second of which is the more widespread. The first group is composed of small bodies of amhibolite that resemble mafic dykes or sills that have been broken into angular fragments, possibly by remobilization of the host rocks (or could this form be one of primary, intrusive origin?). These pre-date(?) the earliest migmatization of the polycyclic leucocratic granitoid gneisses of the KHA. Similar bodies occur in NIA gneisses but their status is less certain in the BA and SBA gneisses. It is possible that they could be used as time markers within the polycyclic gneisses to separate events that pre-date rocks such as the Britt-type granodiorite and related orthogneisses which do not contain such dykes.

The second set of mafic dykes is widespread within the map area and forms part of a group present throughout the Central Gneiss Belt. They crosscut the single cycle granitoid orthogneisses and are foliated together with them. They are therefore monitors of D_2, the event causing the prominent regional structural grain and northwest-trending folds. The dykes are variably metamorphosed (commonly containing hornblende-garnet-plagioclase$\pm$clinopyroxene) and foliated/lineated bodies which commonly have extensive areas of recognizable, though recrystallized, primary igneous texture; for the present they are regarded as belonging to the 'coronitic' type (Davidson and Grant, 1986). At a few places there are crosscutting relationships between successive dykes, but no consistent pattern has yet been recognized.

A very common feature of members of this set and occurring exclusively within the KHA is the presence of large plagioclase phenocrysts (1 cm or larger, very locally in excess of 20 cm long). These are usually arranged in a single, planar zone close to one boundary of the body and are readily recognizable even where the state of recrystallization and deformation of the body is advanced. The largest layer is an impressive, coarse cumulate more than 1 m thick. Absence of phenocrysts in dykes of this group south of the Britt pluton indicates a subdivision of the crust which is related to but of a different order than that defined by the gneiss associations.

GABBRO

There are several rounded bodies of gabbro of kilometre scale scattered throughout the map area (Fig. 1). Some of these lie along or close to boundaries between associations or along prominent internal boundaries. They are heterogeneously recrystallized and contain substantial areas where primary igneous texture is recognizable. This feature distinguishes them from the older metagabbroic phase of the Free Drinks gneiss of the KHA which is completely recrystallized. The occurrence of 'pseudoeclogite' (garnet-clinopyroxene rock; Culshaw et al., 1983) as patches within these bodies appears to be, in contrast, a metamorphic feature. So far they have been observed no further north than within or close to the boundary between the SBA and NIA (Frederick Inlet gabbros of Needham, 1987).

MEGACRYSTIC GRANODIORITE AND RELATED SINGLE CYCLE GRANITOIDS

These metaplutonic rocks belong to the suite of plutonic rocks described by Davidson et al. (1982) with compositions ranging from megacrystic granodiorite (e.g. Britt pluton) through granite (s.s.), quartz syenite to diorite (Fig. 1; not all are shown). These form extremely elongate, folded plutons spread throughout the present map area and several of the largest bodies lie along or close to association boundaries (Britt pluton, Pointe-au-Baril complex and Shawanaga pluton). These bounding plutons are all dominated by megacrystic (K-feldspar) granodioritic compositions of the Britt type. Indeed this composition is by far the most voluminous, possibly representing a 'plutonic event' of considerable magnitude. Intrusion of the Britt pluton has been dated at 1456 Ma (van Breemen et al., 1986) and it possible that all have similar mid-Proterozoic age.

As outlined in previous sections these metaplutonic rocks have important differences in structural and metamorphic history to the enclosing gneisses. They are termed 'single cycle' because they only show the effects of a single structural/metamorphic episode, D_2, which is also responsible for the regional northwest structural grain. This event post-dates the intrusion of the second (coronitic) dyke set and is commonly accompanied by transposition of an earlier foliation in the marginal country rock gneiss associations.

PEGMATITE

Pegmatites were generated and emplaced at several distinct times. The earliest are not widespread, being formed during the first stages of the polycyclic history of the old leucocratic granitoid gneisses of the KHA. A second, widespread group accompanied D_2; these were emplaced in a variety of structural settings, e.g., at a low angle to the new foliation of the single cycle plutonic gneisses or the transposed foliation of the older gneisses, within southeast-dipping, low-angle ductile shears or within strain-shadow-like tails of competent metamafic rocks. The third group comprises large bodies accompanied by folding of the wall rocks, although they were emplaced at a high angle to the northwest strike of the D_2 foliation. These tend to occur in clusters and were clearly emplaced relatively early in post-D_2 times while the rock was capable of brittle-ductile deformation. The fourth important group, common in the northern part of KHA, comprises straight-walled dykes, lacking ductile wall rock deformation and also trending at a high angle to D_2 foliation.

OUTLINE OF STRUCTURAL GEOLOGY

The principal major structures are the northwest-trending, horizontal upright folds which are part of a set extending from the southern margin of the Grenville Front Tectonic Zone to the Parry Sound domain (Fig. 1; Davidson et al., 1982; Schwerdtner, 1987). This folding (F_2) was accompanied by formation of a strong planar-linear fabric in the single cycle metaplutonic rocks and transposition of earlier foliations in the polycyclic host gneiss associations. The lineation invariable plunges shallowly in the southeast quadrant. Well-exposed minor sheath folds with axes parallel to this lineation occur within the SBA, itself disposed about a southeast-trending and low-plunging synform (Fig. 1). It is tempting to speculate that the larger structure has a similar origin to the smaller ones. In addition there are two prominent belts of mylonitic gneiss, readily distinguishable from the surrounding rocks principally on the basis of their finer grain size. That within the BA is accompanied by northeast-trending folds which deform northwest-trending foliations lying beneath the shear zone. The opposite is true for the zone within the NIA which appears to be strongly folded about northwest-trending axes (Fig. 1). Although these ductile shear zones locally pre- or post-date the regional northwest-trending foliation, they are not readily distinguished on the basis of metamorphism.

The structural history is closed by east- to northeast-trending brittle normal faulting. Minor displacement with separations in the order of tens of metres can be documented in many places. The two largest faults cut the Britt pluton and have separations along strike of several hundred metres with south side down displacements (Fig. 1).

METAMORPHISM

Although the area is dominated by metamorphic assemblages suggesting pervasive syn-D_2 upper amphibolite facies conditions, there is field evidence for two important metamorphic events. This is best displayed in the polycyclic gneisses of the KHA where D_1 leucosome was transposed during D_2, accompanied by the formation of minor new leucosome. There is one location where D_1 leucosome (with K-feldspar), developed in metasedimentary rocks, contains coarse blades of kyanite that are in turn overprinted by D_2(?) sillimanite. Throughout the map area there are scattered occurrences of orthopyroxene(?) forming cores to retrograde clots, hinting that granulite facies conditions may have been attained during this early event. The single cycle granitoids contain one set of syn-D_2 leucosomes. Whereas the younger mafic dykes post-date the single cycle plutons and therefore only record D_2, they commonly display several discrete mineral assemblages and textural heterogeneity which presumably records parts of D_2 P-T-time paths.

SUMMARY OF STRUCTURAL, METAMORPHIC AND PLUTONIC HISTORY

The segment of Britt domain studied in 1987 was originally a lithologic mosaic now represented by the distribution of distinctive gneiss associations. It is possible that the associations represent terranes that were assembled in the mid-Proterozoic or earlier: the major single cycle granodiorites, including the Britt pluton, lie along or close to the boundaries of the associations which, therefore, must have existed as discontinuities at the time of pluton emplacement. The earliest metamorphism, producing partial melting (locally with kyanite) and possibly attaining granulite facies, likely post-dated the early mafic dykes and pre-dated the intrusion of the single cycle plutonic rocks. Whether or not this event occurred during terrane assembly, or even during pluton emplacement, is not known. Mid-Proterozoic intrusion of the single cycle plutons was followed by emplacement of the younger mafic dykes; this event may constitute an episode of lithospheric stretching to which pluton emplacement and the earlier metamorphism may possibly also be assigned.

During the second metamorphism and deformation (D_2), the dominant northwest-trending folds and regional foliation as well as the northeast-striking shear zones were formed. These structures presumably accompanied convergence during the orogeny, associated with, though not necessarily accompanying, crustal thickening (Davidson, et al., 1982; Culshaw et al., 1983). The problem of the mechanism of formation of the northwest-trending folds has been addressed by Schwerdtner (1987) and elsewhere by Rivers (1983) and Coward (1984).

The post-D_2 history is recorded by pegmatites emplaced under brittle-ductile and brittle crustal conditions and later by south-side-down normal faulting.

SOME SALIENT PROBLEMS

1) Pink leucocratic granitoid rocks of both supracrustal (volcanic?) and plutonic origin are part of the lithologic makeup for all of the gneiss associations. Do these rocks therefore record a single event that post-dated the suggested assembly of the terranes represented by the different gneiss associations (crustal partial melting)? What is their age, and could they be related to lithospheric stretching?

2) Is the shear zone bounding the BA and NIA contractional, extensional or both, and what is the age of its formation?

3) What is the mechanism of formation of the northwest-trending folds? A hypothesis similar to those of the previously quoted authors seems reasonable in some cases, i.e., that they formed because thrust sheets are either constricted by irregularities in their soles or are subject for some other reason to transcurrent strains. Explanations such as these may be considered for the northernmost folds, the southeast boundary of the Grenville Front Tectonic Zone (Davidson and Bethune, 1988), which strikes obliquely to the Grenville Front, being a candidate irregularity.

4) The most easily solved problem and probably the most significant is the question of the age of the different single cycle plutonic rocks and of each of the components of the subdomains.

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Geology of the north shore of Georgian Bay, Grenville Province of Ontario

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Abstract

The Grenville Front Tectonic Zone (GFTZ) along the north shore of Georgian Bay, Ontario, contains several southeast-dipping ductile shear and mylonite zones. In its western part, gneisses probably equivalent to the ca. 1.74 Ga Killarney complex beyond the Grenville Front appear to wrap infolds of metasediments of questionable Huronian affinity. To the east, metaplutonic units in various states of preservation are separated by zones of annealed mylonite. The easternmost mylonite zone defines the southeast margin of the GFTZ, beyond which, in Britt domain, stretched granitoid plutons in an older metasedimentary gneiss host are folded about shallow southeast-plunging axes. Deformed olivine metadiabase, considered equivalent to the ca. 1.24 Ga Sudbury swarm northwest of the front, occurs throughout the GFTZ and may be represented in Britt domain by small coronitic masses. Easterly dipping seismic reflectors beneath Georgian Bay 80 km to the south probably represent mylonite zones like those exposed in the GFTZ.

Résumé

La zone tectonique du front de Grenville qui s'étend le long de la rive nord de la baie Georgienne, on Ontario, renferme plusieurs zones à mylonites et de cisaillement ductile au pendage incliné vers le sud-est. Dans la partie ouest de la zone, des gneiss équivalents au complexe de killarney datant d'environ 1,74 Ga et gisant au-delà du front de Grenville, semblent entourer des plis métasédimentaires d'affinité huronienne douteuse. À l'est, des unités plutoniques métamorphisées plus ou moins bien conservées, se trouvent séparées par des zones à mylonites recuites. La zone à mylonites située le plus à l'est délimite la limite sud-est de la zone tectonique du front de Grenville au-delà de laquelle, dans le domaine de Britt, des plutons granitoïdes étirés, logés dans un gneiss métasédimentaire plus ancien, se trouvent pliés autour d'axes gisant à faible profondeur et inclinés vers le sud-est. Des métadiabases à olivine déformées, que l'on considère équivalentes à l'essaim de Sudbury datant d'environ 1,24 Ga et gisant au nord-est du front, se manifestent d'un bout à l'autre de la zone tectonique du front de Grenville et peuvent correspondre, dans le domaine de Britt, à de petites masses concentriques. Des réflecteurs sismiques à pendage est gisant sous la baie Georgienne à quelque 80 km au sud, représentent tout probablement des zones à mylonites semblables à celles mises à nu dans la zone du front de Grenville.

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INTRODUCTION

Superb and essentially continuous exposure on the peninsulas, islands and reefs along the north coast of Georgian Bay, Ontario, reveals details of Grenville Province geology that are not available inland. This coastal strip (Fig. 1), extending from Key Harbour to Killarney, inaccessible from the land, was examined briefly in 1980 (Davidson and Morgan, 1981) and the eastern part of it in more detail in 1981 (Davidson et al., 1982). It lies within two 1:50 000 scale map areas, Collins Inlet (41H/14) and Key Harbour (41H/15). Both areas were geologically mapped at 1:63 360 scale in the mid-1920s (Quirke and Collins, 1930; Quirke, 1930a). The Collins Inlet area was remapped between 1964 and 1967 (Frarey and Cannon, 1969; Frarey, 1985); the Key Harbour area has not been remapped, although the Delamere map area (Quirke, 1930b), adjacent to the north, was remapped as part of the 1:126 720 scale Burwash map area (Lumbers, 1975).

The coastal strip provides an oblique transect of the northwest part of the Grenville Province, comprising from west to east the Grenville Front adjacent to the Southern Province, the Grenville Front Tectonic Zone (Wynne-Edwards, 1972) and part of Britt domain (Davidson et al., 1982) of the Central Gneiss Belt. The two Grenville terranes are characterized respectively by northeast-striking, lenticular map units of orthogneiss, layered gneiss, migmatite and intervening mylonitic zones, all dipping moderately southeast, and by continuous, sheet-like units of migmatitic orthogneiss and layered gneiss that are folded about gently southeast-plunging axes, giving this part of Britt domain a distinct northwesterly grain. Metasupracrustal rocks (quartzite, feldspathic and pelitic schist and gneiss) in the Grenville Front Tectonic Zone (GFTZ) near the Grenville Front have some resemblance to Huronian Supergroup lithologies in the adjacent Southern Province, although direct correlation has proven difficult (Frarey, 1985). Supracrustal gneisses farther east do not show this resemblance. Granites along the margin of the Southern Province, known to be of at least two ages (ca. 1740 and ca. 1470 Ma; van Breemen and Davidson, in press)

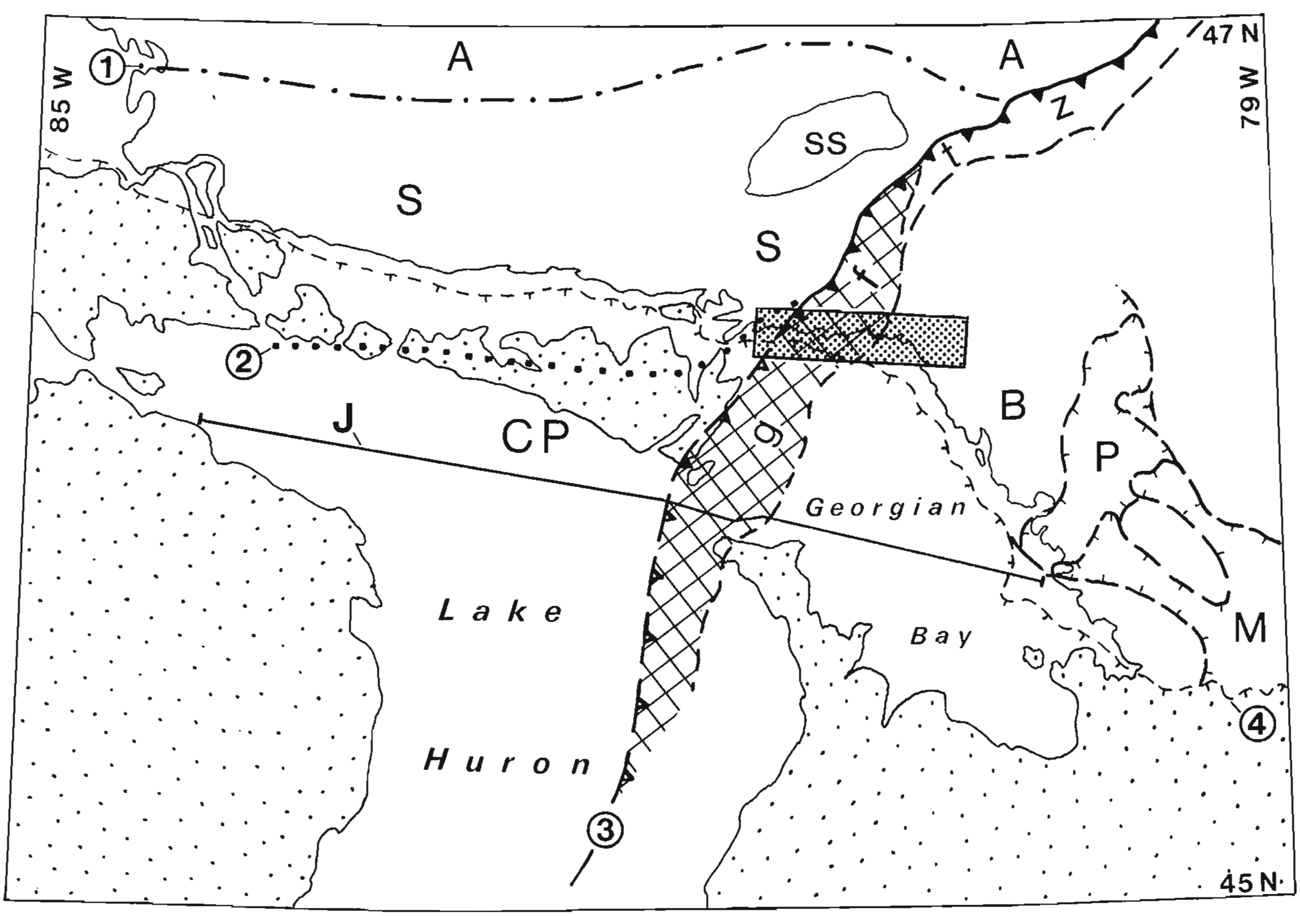

Figure 1. Location of report area, north shore of Georgian Bay. A - Superior Province; 1 - northern limit of effects of Penokean folding; S - Southern Province; ss - Sudbury structure; 2 - 'mid-Manitoulin line', assumed southern limit of Huronian Supergroup; CP - assumed northeast extension of Central Plains orogen of the midcontinent; 3 - Grenville Front; gftz - Grenville Front tectonic zone (Wynne-Edwards, 1972), with patterned part indicating aeromagnetically positive anomalous zone; B, P, M - Britt, Parry Sound, Muskoka domains of Central Gneiss Belt; 4 - northern limit of Paleozoic cover (stippled). Line J is the site of a seismic reflection survey carried out in September, 1986, as part of the GLIMPCE program (Green et al., 1987).

are represented in the GFTZ by foliated and gneissic granitoid rocks of similar composition, but some of the orthogneiss units of Britt domain appear to be petrologically different although they likely have similar ages (e.g., Britt pluton, 1457 Ma; van Breemen et al., 1986).

The Grenville Front itself has been placed in two different positions. In the Lake Panache map area (41I/3), Frarey (1985; Frarey and Cannon, 1969) placed it along the northwest margin of lenticular granite bodies that extend from Killarney 85 km northeast to Coniston, east of Sudbury. Lumbers (1975; Card and Lumbers, 1977) identified an intense, continuous mylonite zone as the Grenville Front Boundary Fault a few kilometres farther southeast. Davidson (1986a) favoured the latter position on the basis that although the northwest contact of the granite bodies with Huronian sedimentary rocks is faulted in the Lake Panache map area, to the southwest (Killarney granite) and northeast (Chief Lake granite) the contacts are intrusive and the granitic rocks are not deformed; he also argued that, from an age point of view, a continuous structural feature separating regions that have different ages and orientations of deformation, and representing a common though perhaps protracted event related to Grenvillian orogeny, is a more logical location for the Grenville Front than the margins of granites of different age that are not related to that orogeny.

Aeromagnetic maps of this region (Geological Survey of Canada, 1965, 1969) show a pronounced, northeast-trending zone of positive aeromagnetic anomaly coinciding with the GFTZ and tapering northward to die out against the Grenville Front east of Sudbury. A recent aeromagnetic survey of Georgian Bay and Lake Huron (Geological Survey of Canada, 1987) showed that this anomaly extends for at least 250 km to the southwest, defining the buried extension of the front (Fig. 1). To the west, oval magnetic patterns are interpreted as marking granitoid plutons unaffected by the Grenvillian orogeny. With the identification of pre-Grenvillian plutonic and associated felsic volcanic rocks in the Killarney area that are likely correlative with buried mid-Proterozoic terranes of the mid-continent (Davidson, 1986a; van Breemen and Davidson, in press), it seems probable that this large aeromagnetic anomaly represents the reworked equivalents of these terranes in the Grenville Province.

A seismic reflection survey along an east-west line through Georgian Bay and northern Lake Huron, located approximately 80 km south of the north coast of Georgian Bay (GLIMPCE; Green et al., 1987), passed across Britt domain, the GFTZ and the Grenville Front into the granite terrane to the west. The survey revealed a remarkable set of moderately eastward-dipping reflectors that accord well with the known disposition of structures at the surface in the exposed part of the GFTZ and probably represent a number of parallel, southeast-dipping shear zones or sheared boundaries between major rock units. One of the main purposes of this summer's transect was to provide a geological basis for interpreting both the aeromagnetic and seismic data outlined above, with the working hypothesis that the GFTZ in this region may contain mid-Proterozoic granites and rhyolites equivalent to those that lie south of the Penokean fold-belt in the mid-continent (Bickford et al., 1986) but affected by the later and perhaps unrelated Grenvillian orogeny.

GEOLOGICAL SUMMARY, NORTH COAST OF GEORGIAN BAY

The following summary is based on an almost complete examination of the coastal exposure along the north coast transect in the Collins Inlet and Key River map areas; with a few exceptions, time did not permit tracing of geologic units or structures inland, and their continuity shown in Figure 2 is based mainly on airphoto and aeromagnetic interpretation as well as on pre-existing maps.

Killarney to Beaverstone Bay

On the northwest side of the Grenville Front, which is represented by a single major mylonite zone on Philip Edward Island that extends north through Carlyle Lake and beyond (O'Donnell, 1986), the ca. 1740 Ma Killarney volcanic-plutonic complex displays upright, east-northeast-trending and -plunging folds and associated axial planar foliation and axial lineation that are cut by non-deformed ca. 1400 Ma pegmatite dykes and northwest-trending ca. 1240 Ma olivine diabase dykes of the Sudbury swarm (Davidson, 1986a; van Breemen and Davidson, in press; Krogh et al., in press). All of these are truncated by the mylonite zone, within and east of which foliations dip and lineations plunge moderately southeast. The Grenville Front Boundary Fault in this area, then, represents a major break in age and orientation of structure. An abrupt change in metamorphic grade also coincides with the Front, namely from greenschist facies to at least middle amphibolite facies on the Grenville side.

Two transects from the Front to the east shore of Beaverstone Bay, one along Collins Inlet and the other along the south coast of Philip Edward Island (Fig. 2), expose a series of lenticular map units, oriented northeast, that are not all present in both transects (Frarey, 1985). On the Collins Inlet transect granitoid and minor layered quartzofeldspathic gneisses prevail between the Front and Mill Lake and gneissic to protomylonitic granodiorite between Mill Lake and Beaverstone Bay. Along the shores and on the islands of these two bodies of water, however, are exposures of metasedimentary schist and gneiss. These rocks, characterized by quartzite, pelitic schist and minor amphibolite, are not exposed on the coastal transect where the main rocks units are variously strained granitoid orthogneisses and intervening fine grained, pink to grey, locally well-layered feldspathic gneisses, in places migmatitic. Eastward from the Grenville Front (and numbered in Fig. 2) the map units are: 1), a zone of ultramylonitic rocks on either side of a protomylonitic augen granitoid that is probably the southwest continuation of the Bell Lake granite (Frarey, 1985); 2), grey and pink, fine grained, strongly lineated biotite and hornblende-biotite gneiss derived from equigranular plutonic rock, perhaps the equivalent of the hypabyssal igneous rocks of the Killarney complex; 3), predominantly pink, fine-grained, lineated, sugary felsic to streaky quartzofeldspathic gneiss within which highly attenuated clasts, reminiscent of volcaniclastic rocks of the nearby Killarney complex, are preserved locally suggesting that this unit as a whole was derived from felsic volcanic rocks possibly correlative with the Killarney complex; 4), a body of pink, equigranular, lineated, uniform biotite leucogranite orthogneiss that is well exposed on the Fox Islands and tapers northeastward; 5), a narrow strip of pink, sugary, leucocratic quartzofeldspathic gneiss, similar to unit 3; 6), between West Desjardins Bay and Solomon Bay, a broad zone of pinkish grey, fine-grained biotite-hornblende gneiss with locally well-developed centimetre- to decimetre-scale layering showing complicated fold patterns and characterized by alignment of metamorphic hornblende parallel to the axial planes of the youngest folds; flattened clasts in layered gneiss were identified at one place; 7), a uniform, reddish pink hornblende-biotite quartz monzonite orthogneiss, exposed from Solomon Bay to Hamilton Island, that has a strongly mylonitic northwest margin flanking a complexly folded and locally migmatitic core; this is in sharp contact with 8), another unit of pink, fine-grained, sugary, migmatitic gneiss that extends northward to the west side of Mill Lake; 9), grey, uniform, protomylonitic hornblende-biotite granodiorite orthogneiss with an intensely mylonitic western margin; 10), another complexly folded, layered and variable quartzofeldspathic gneiss, granulite and migmatite unit, flanked on its northwest side by a mylonite zone within which a few metres of biotitic and quartz mylonite is exposed at the head of Moose Bay; 11), a variably deformed plutonic suite composed mainly of reddish granite orthogneiss like that of unit 7, and containing a younger, also deformed, monzodiorite to gabbro orthogneiss unit; 12), fine-grained, pink feldspathic gneiss similar to components of units 3, 5 and 10; this unit lies along the west shore of Beaverstone Bay and appears to be partly occluded to the south by a mylonite zone south of Toad Island. The mylonite zones described above vary from tens to hundreds of metres in width and are characterized by strain gradients at their margins.

Most rocks in the transect described above carry a well developed lineation, generally formed by quartz rods or blades and streaked-out mafic minerals. Lineations plunge moderately southeast and are axial to minor folds of the latest generation. The fine grained gneisses between orthogneiss units commonly display tight mesoscopic folds whose enveloping surfaces trend at a high angle across the regional northeast strike of the units, suggesting a compressed, earlier, easterly trend of layering between originally discordant plutons. Common in both supracrustal and orthogneissic units are tightly to isoclinally folded and disrupted mafic dykes; pods and lenses with remarkably well preserved diabasic texture are present locally along otherwise highly attenuated fold limbs. The regionally characteristic rib-like nature of many coastal outcrops, particularly in the orthogneiss units, is apparently due to weathering out of biotitic amphibolite, derived from mafic dykes, on the limbs of high amplitude isoclinal folds.

The metasedimentary rocks exposed at Mill Lake and Beaverstone Bay differ from any of the layered gneisses of

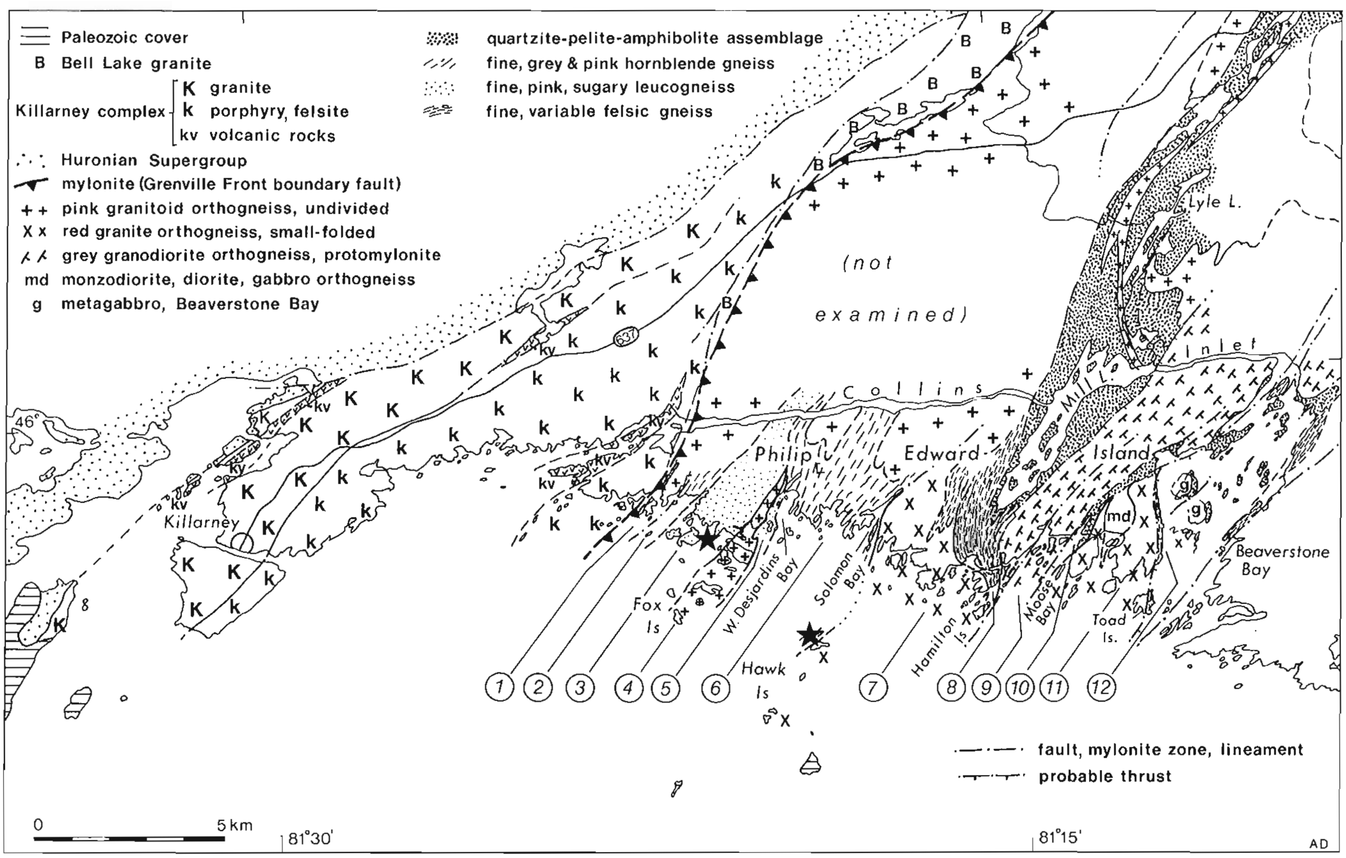

Figure 2. Western part of transect in the Grenville Province, north shore of Georgian Bay, Killarney to Beaverstone Bay. Geology in Lyle Lake area is after Frarey (1985). Circled numbers refer to map units described in the text. Stars indicate locations of deformed clastic structure in gneisses that are interpreted to have a clastic volcanic protolith.

the coastal transect. At Mill Lake they comprise pelitic schist (garnet-sillimanite-biotite-quartz-plagioclase), semipelitic schist and gneiss (garnet-biotite-quartz-2 feldspars), orthoquartzite, aluminous gneiss (biotite-muscovite-sillimanite-feldspar-quartz), less common calcareous schist and gneiss (biotite-garnet-hornblende-quartz-plagioclase) and amphibolite. Much of the quartzite is glassy, but where not so recrystallized it has relict bedding and rare semblances of crossbedding. The metasedimentary rocks for the most part are not migmatitic. They display folds at meso- and macroscopic scale whose form and enveloping surfaces, like those in the layered gneisses of the coastal transect, suggest an earlier easterly trend now transposed to northeast. These folds and associated mullions plunge moderately to steeply south rather than southeast. On the northwest side of Mill Lake pelitic and semipelitic rocks are in abrupt though seemingly gradational contact with pink, fine-grained feldspathic gneiss of uncertain origin. These rocks wrap around the pelite-quartzite assemblage at the southwest end of the lake, forming a large fold closure. To the northeast the pelite-quartzite assemblage is continuous through Lyle, Mahzenazing and Tyson lakes and beyond (Frarey, 1985). The southeast limb of this fold at Mill Lake is truncated at the mylonitic margin of a pluton of grey, uniform granodiorite orthogneiss, unit 9 of the coastal transect. The orthogneiss has a prominent moderately southeast-plunging stretching lineation; mylonitic fabrics indicate that it has been thrust northwest over the Mill Lake metasediments, the same sense as is found to the northwest in the vicinity of the Grenville Front (O'Donnell, 1986; Davidson, 1986a).

Similar metasedimentary rocks in the northern part of Beaverstone Bay lie on the opposite side of the granodiorite orthogneiss just mentioned. They too are disposed about northeast-trending folds with steep south-plunging axes and have enveloping surfaces trending easterly across the grain imposed by their attenuated northeast-trending limbs. Among the metasediments, some biotitic schists contain 1-3 cm garnet metacrysts and stubby prisms of sillimanite that may be pseudomorphs of kyanite. Amphibolite gneiss, commonly associated with orthoquartzite, is of two origins: gabbro or diabase dykes or sills, and calcareous sediments. The latter is characterized by mesoscopic compositionally layered units intercalated with biotitic schist and gneiss, whereas the meta-igneous amphibolite locally retains a relict igneous texture and is more uniform in appearance. Islands in the south part of Beaverstone Bay display deformed intrusive relationships between metasediments and metagabbro or gneissic granitoid rocks. A major fold closure like that at Mill Lake is not apparent, but the pelite-quartzite assemblage does not extend as far south as the mouth of the bay. A narrowing sliver of pelitic schist and quartzite projects southwest between the mylonitic granodiorite of unit 9 and the granite of unit 11 and, although not exposed, likely links with the narrow quartz and biotitic mylonite that flanks the granodiorite orthogneiss unit at the head of Moose Bay. Kinematic indicators along the southeast side of the granodiorite and in shear zones flanking and within the granite (unit 11) all show up-lineation northwest thrust sense; in this case, the Beaverstone Bay metasediments are thrust over the granitoid rocks on their west and northwest sides. Northeast from the bay, a narrow zone of discontinuous lenses of metasedimentary gneiss is on strike with similar metasediments in the vicinity of Broker Lake. To the southeast the pelite-quartzite assemblage is flanked by fine-grained, pink, sugary, flaggy feldspathic gneiss of uncertain affinity, or by grey, migmatitic gneiss and orthogneiss derived from tonalite and granodiorite.

Beaverstone Bay to French River, Main Outlet

Immediately east of Beaverstone Bay and extending to Batt Bay (Fig. 3) lies a complicated terrane composed of a variety of metaplutonic rocks in relatively small units within a matrix of highly and irregularly deformed, varied migmatitic gneiss, at least some of which can be shown to have been derived by intense ductile deformation and syntectonic migmatization of the plutonic rocks. Gneisses of unequivocal sedimentary origin, that is, with diagnostic composition such as quartzite or pelite, were not observed. Ranging in size from a few kilometres long and several hundred metres wide down to a few tens of centimetres, metaplutonic rocks of gabbroic, tonalitic, granitic and syenitic compositions occur as non-migmatitic lenses or 'hard lumps' in a convoluted migmatite host. Their contacts with migmatite are gradational through strain gradients and accompanying increasing development of leucosome. Much of the layering in the migmatitic gneisses has been formed by attenuation and plastic flow of a suite of multiple intrusions of different compositions. Narrow mafic dykes also feature prominently in parts of this complex, particularly around Point Grondine; they are relatively little deformed in the orthogneiss cores, but are isoclinally folded, attenuated and disrupted in the derived migmatitic gneisses, contributing to the layered aspect of the rocks. The most advanced state of degradation from formerly igneous plutonic rocks occurs in a northeast-trending 'straightening zone' passing through Chaugis Bay where irregular migmatitic gneiss with small masses of preserved orthogneiss has been further stretched into streaky, lenticular gneiss. As a whole, this complex appears to have deformed in a highly mobile way, giving rise to rocks and structures that differ markedly from those west of Beaverstone Bay.

Between Batt Bay and the main outlet of the French River to the east lies a large granitoid massif. It is composed of three parts, each separated by relatively narrow zones containing very fine-grained, sugary quartzofeldspathic annealed mylonite that in places encloses thin lenses of less grain-comminuted granitoid rock. The western part is composed predominantly of uniform, pink, non- migmatitic, protomylonitic, biotite leucogranite orthogneiss with a well developed southeast-plunging stretching lineation and a weak northeast-oriented flattening foliation. It has a light grey phase along its northwest margin in Batt Bay that is mildly migmatitic (5 per cent leucosome), in strong contrast to the migmatites that form its contact rock. The contact, however, is concordant and intrusive relationships were not established. Grey granodiorite orthogneiss also flanks the southeast side of the leucogranite orthogneiss along Fort Channel. The central part of the massif, extending east to the Bad River Channel, is fine- to medium-grained, pink, equigranular quartz syenite, again strongly lineated and weakly foliated. Its foliation defines a large fold, open to the southwest, whose core is occupied by brown biotite-hornblende syenite orthogneiss containing minor clinopyroxene. This rock grades northward to pink quartz syenite. Between Bad River Channel and the main outlet of the French River, the eastern part is occupied largely by an oval pluton of well-preserved, mildly megacrystic (K-feldspar) hornblende-biotite granodiorite that is little deformed in places, although transected by narrow ductile shear zones; xenoliths are locally abundant and are weakly elongate, defining a moderate southeast-plunging lineation in common with the strong lineations prevalent regionally. On its east side are three strongly foliated and lineated granitoid orthogneiss units: brick-red, fine-grained, sugary syenitic orthogneiss, brownish to greenish grey, medium- to fine-grained hornblende quartz syenite orthogneiss and pink, augen granite orthogneiss.

French River, Main Outlet, to Key River

The main outlet of the French River is marked by intensely flattened ductile gneisses, including mylonite and 'straight gneiss' (Davidson et al., 1982) in a broad zone parallel to the channel and extending southwest along the Bustard Rocks, west of the main Bustard Island archipelago (Fig. 4). This shear zone marks the beginning of a gradual eastward change in the orientation of regional foliation, gneissosity and disposition of map units from northeast through north to northwest; dip direction throughout this transition in invariably easterly and prominent southeast-plunging lineations prevail. It marks the broad boundary between the GFTZ and Britt domain of the Central Gneiss Belt, characterized in this sector by large-scale folds involving metasedimentary migmatitic gneiss and sheet-like plutonic orthogneiss units (Davidson et al., 1982). The most prominent fold is a huge, open synform plunging gently south-southeast whose axial trace lies just inland from the Georgian Bay coast. At Key Harbour, the core of this synform is occupied by the Britt pluton, composed of variably migmatitic garnet-hornblende-biotite orthogneiss of quartz monzodiorite to granodiorite composition. To the west lie three sheets, each less than 500 m thick, of grey biotite granodiorite orthogneiss with flattened augen of K-feldspar. Between each of these metaplutonic units lie complexly deformed gneisses that are at least in part supracrustal in origin, as shown by the presence of aluminous, calc-silicate and quartz-rich gneisses. At the coast, however, the westernmost of the grey granodiorite orthogneiss sheets cuts at a shallow angle across the eastern contact of a thick sheet of pink, garnetiferous, migmatitic granite to quartz syenite orthogneiss that has a central zone of meta-anorthosite and metagabbro. Where this readily identified unit passes around the hinge of the synform to the north it was referred to as the Pickerel complex by Lumbers (1975). Its lower contact zone is characterized in many places by included rafts of orthoquartzite, even though metasedimentary gneisses do not everywhere form its underlying contact rock. The region west of the Pickerel complex is underlain mainly by dark orthogneisses ranging through gabbro, diorite, quartz diorite and granodiorite in composition, with

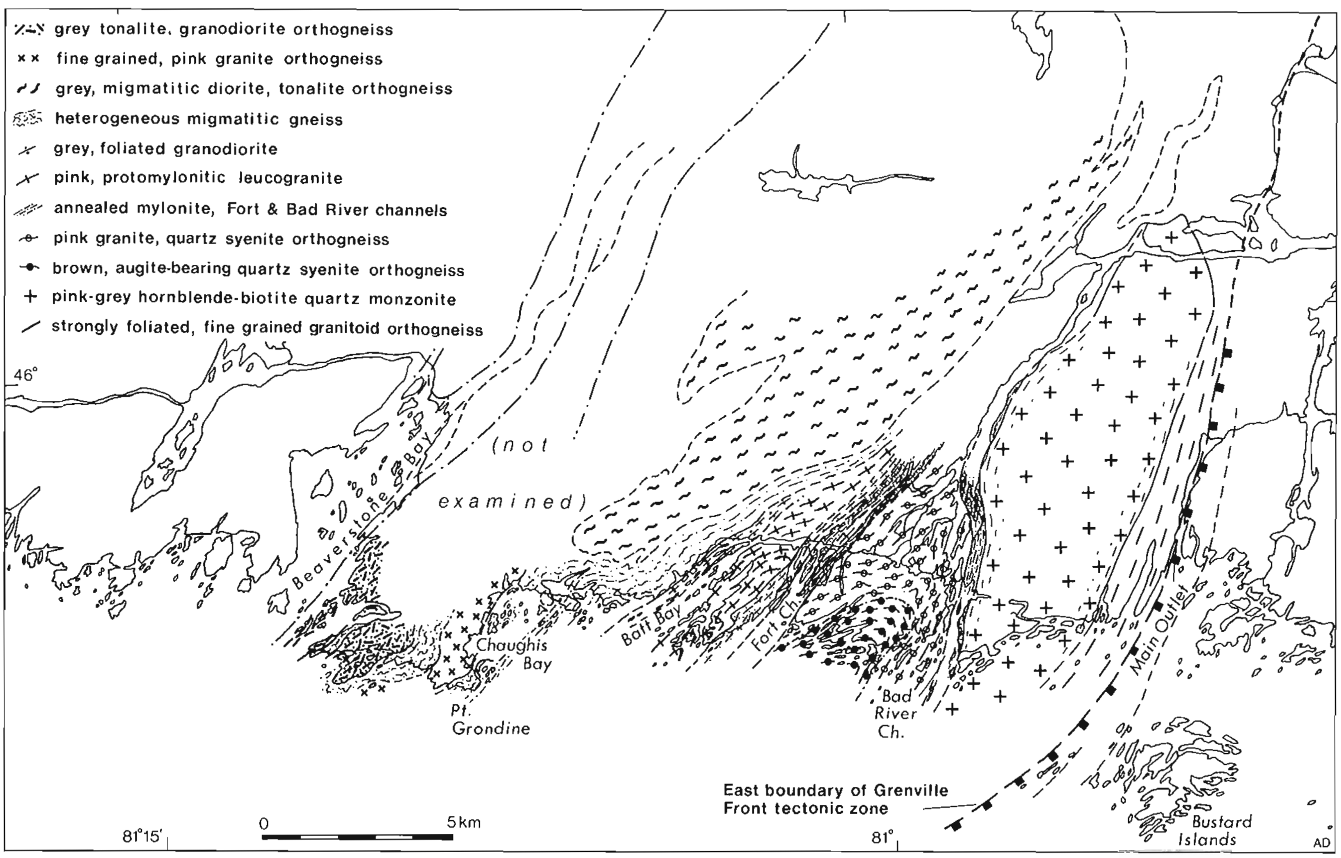

Figure 3. Central part of transect, Beaverstone Bay to the main outlet of French River. Geology north of latitude 46°N after Frarey (1985) and Lumbers (1975).

small lenticular plutons of pink leucogranite locally. Migmatitic gneisses of supracrustal origin underlie Fox Bay and part of the Bustard Islands and are also present in the shear zone that defines the northwest edge of Britt domain. Fox Bay is the site of tight, south-plunging folds that lie within the western limb of the major synform. These folds contain earlier isoclines in their limbs and are themselves warped around open, southeast-plunging folds of low amplitude.

Many of the same plutonic sheets can be traced around the nose of the Key Harbour synform and are recognized on the southwest-facing limb along Key River, but some pinch out and other, different lenses of orthogneiss appear. The Pickerel complex displays a number of tight, internal folds within the eastern limb and its extent to the southeast has not been fully elucidated. A complementary, southeast-plunging antiformal closure crosses the Key River just west of Highway 69.

DIABASE AND METADIABASE

Diabase dykes of two ages cut Huronian rocks and the Killarney complex in the transect area northwest of the Grenville Front (Palmer et al., 1977). The youngest swarm, termed Grenville dykes (Fahrig and West, 1986), are tholeiitic, east-trending and post-Grenvillian; their distribution is shown in Figure 5. The older swarm, the Sudbury dykes, are enriched in iron and alkalis and contain plentiful olivine, trend northwest and are brittly deformed at the Grenville Front. Equivalent dykes in the GFTZ are metamorphosed and variably deformed, depending on their distance from the front (see Bethune and Davidson, 1988), but in many places can be seen to have cut 'Grenvillian' fabrics (southeast-dipping foliation and southeast-plunging stretching lineation) in their host rocks. Their suggested age of ca. 1220 Ma (Fahrig and West, 1986), recently confirmed by U-Pb baddeleyite dating at 1238 $\pm$ 4 Ma (Krogh et al., in press) leads to the inference that Grenvillian-style tectonism was underway before intrusion of the Sudbury swarm (but not before ca. 1400 Ma, the age of non-deformed pegmatite in the Killarney complex; van Breemen and Davidson, in press). Deformation in the GFTZ thus began earlier than in major ductile shear zones in the interior of the province (van Breemen and Hanmer, 1986; van Breemen et al., 1986), though not necessarily before deformation in the intervening lithotectonic domains. Metamorphic reaction between olivine and plagioclase to produce coronas of pyroxenes and garnet in the GFTZ within a few kilometres of the front suggest marked uplift adjacent to the front, perhaps in the order of 20 km.

In Britt domain, coronitic olivine metadiabase was recognized in several places but does not have the form of dykes. Rather, it occurs as equant masses, rarely more than a few tens of metres across, within highly ductile gneisses, in the

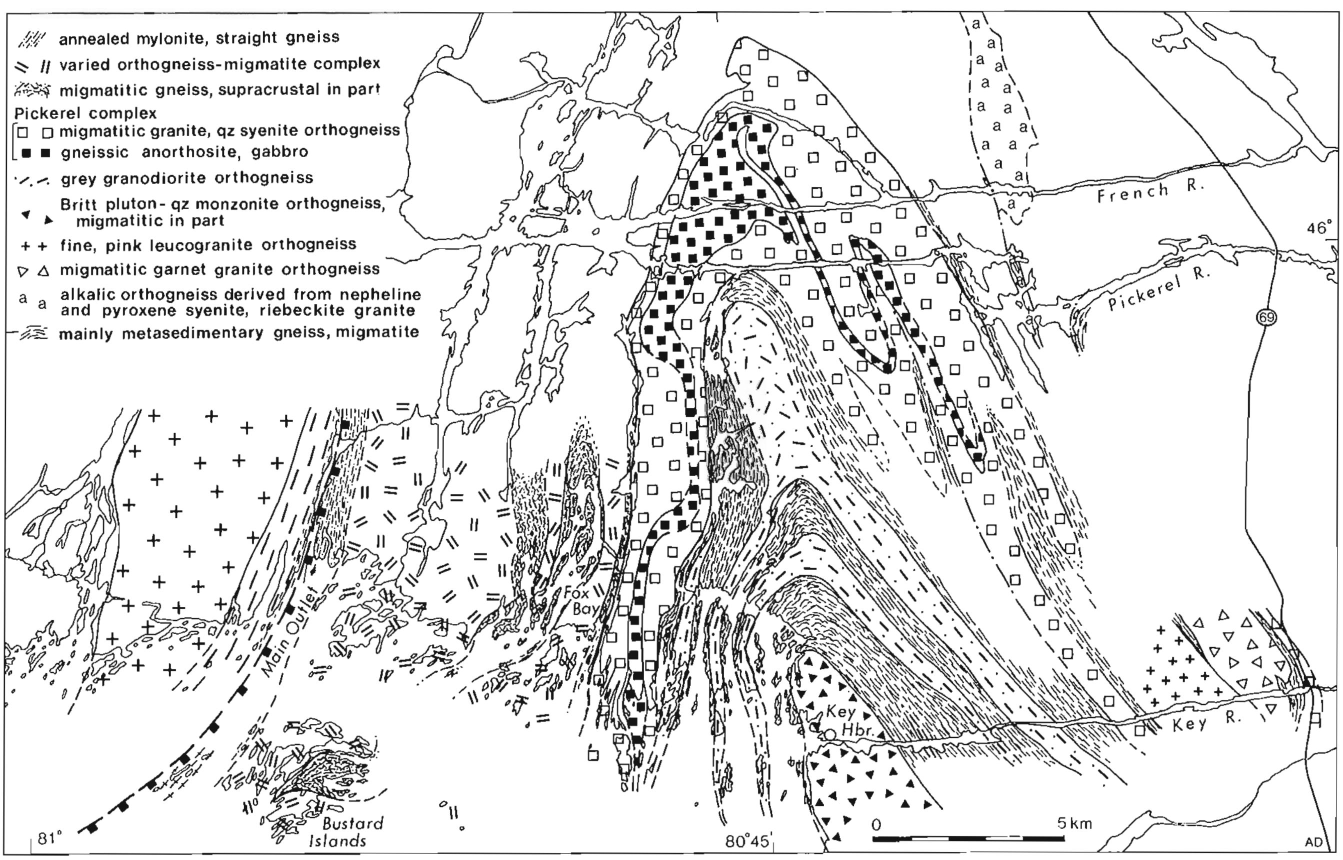

Figure 4. Eastern part of transect, main outlet of French River to Key River. Geology north of 46°N after Lumbers (1975).

same manner as is found farther southeast (Davidson and Grant, 1986). If the coronitic metadiabase in the Key Harbour area (Fig. 5) can be shown to be equated with the Sudbury dykes, this would provide elegant proof that, were the rocks of this region once exotic with respect to the mid-Proterozoic North American craton, assemblage occurred before ca. 1240 Ma. It has yet to be determined whether these coronitic masses are tectonic boudins derived from formerly larger, pre-tectonic dykes, whether they reflect the original form of intrusion into a ductile medium, or whether they are simply tectonic inclusions of unknown origin.

Another type of metadiabase is common in Britt domain and occurs both as equant masses like the coronitic metadiabase and as dykes generally less than 5 m wide that are folded and become attenuated on fold limbs. This metadiabase is thoroughly recrystallized (plagioclase-hornblende-hypersthene-augite-garnet), yet possesses relict igneous texture. Characteristic are scattered, large plagioclase phenocrysts with metamorphic garnet rims. Lack of corona structure suggests that this type of metadiabase did not originally contain olivine. Small dykes of this type cut coronitic metadiabase masses in the Bustard Islands and south of Key Harbour. Such dykes are comparatively rare in the GFTZ, although they locally abound in the migmatite-orthogneiss complex at Point Grondine (Fig. 3) and in parts of the granite body southeast of Solomon Bay (Fig. 2).

OBSERVATIONS AND INTERPRETATIONS

The transect along the north coast of Georgian Bay covers the whole width of the Grenville Front Tectonic Zone and the northwest part of the Central Gneiss Belt of Wynne-Edwards (1972). Never formally established in the field, the southeast margin of the GFTZ is described only on a regional structural basis, for example, as the line or zone where structural orientations "... change from dominant Front- parallel trends to other orientations farther southeast ..." (Davidson, 1986b, p. 70). The relatively abrupt change in structural style, however, from northeast-oriented lenticular rock units separated by mylonitic zones west of the main channel of French River to much more continuous, tabular metaplutonic units with intervening supracrustal gneiss and migmatite on the flank of major north to northwest- trending folds to the east in Britt domain, allows definition of this boundary as the locus of the easternmost of several southeast-dipping high-strain zones in the GFTZ (Fig. 3). For more information on the characteristics of Britt domain, the reader is referred to Culshaw et al. (1988).

High-strain zones in the GFTZ are characterized by mylonitic equivalents of adjacent rocks. In some places these zones, which invariably dip moderately (30-50°) to the southeast and carry a pronounced down-dip stretching lineation, truncate structures in the underlying rocks: for example, the mylonitic granodiorite east of Mill Lake (Fig. 2) truncates folds in the underlying metasedimentary rocks. In other zones, such as at Fort Channel (Fig. 3), mylonite encloses lenses of protomylonitic granitoid rock and appears to be discordant to earlier structure in both under- and overlying panels. Preliminary examination of thin sections of oriented specimens of mylonitic rocks from several high-strain zones in the area west of Beaverstone Bay, where the mylonitic rocks are generally inequigranular (porphyroclastic) and whose mineral grains, particularly quartz, show much internal strain (Fig. 6A), has revealed a consistent reverse sense of displacement on the basis of orientation of C-and-S fabric, shear-band foliation and rotated porphyroclasts or augen. In contrast, mylonites between Beaverstone Bay and the main outlet of French River are characterized by extremely elongate or platy quartz grains, seen in thin section to be composed of equant grains with a high degree of crystallographic orientation and no internal strain (Fig. 6B). Referred to here as 'annealed' mylonites (rather than 'blastomylonites'), these rocks do not contain obvious kinematic indicators, although subsequent quartz petrofabric analysis may prove useful in this regard.

Metasedimentary rocks of the quartzite-pelite-amphibolite association such as are found at Mill Lake and Beaverstone Bay were not encountered farther east. Indeed, the next undoubtedly metasedimentary unit east of Beaverstone Bay lies along the GFTZ boundary, for example, calc-silicate and garnet-sillimanite quartzofeldspathic gneiss in the

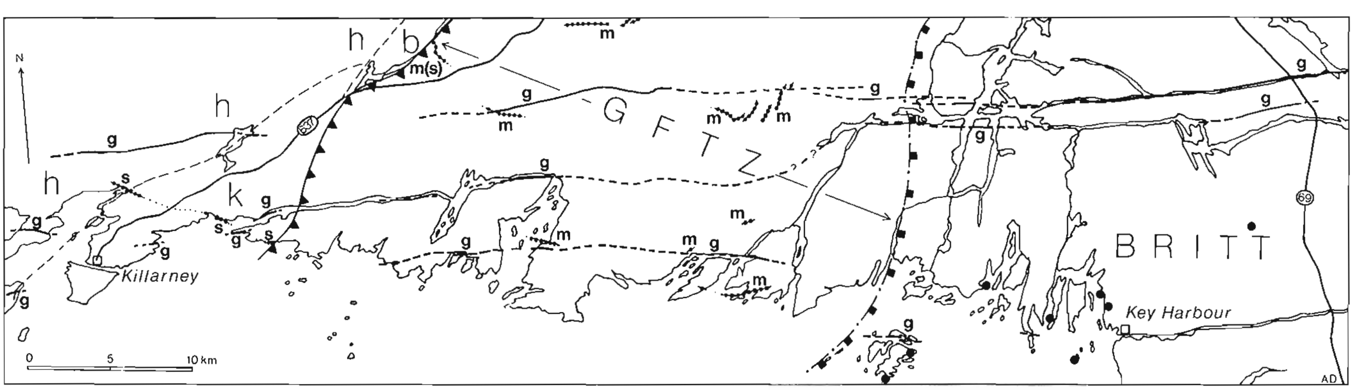

Figure 5. Distribution of diabase and metadiabase in the transect area. S - olivine diabase dykes of the Sudbury swarm (ca. 1240 Ma); m - metamorphosed olivine diabase dykes, coronitic in part (GFTZ); black dots - tectonic inclusions of coronitic olivine metadiabase (Britt domain); g - post- Grenvillian tholeiitic diabase dykes of the Grenville swarm; h - Huronian; k - Killarney complex; b - Bell Lake granite.

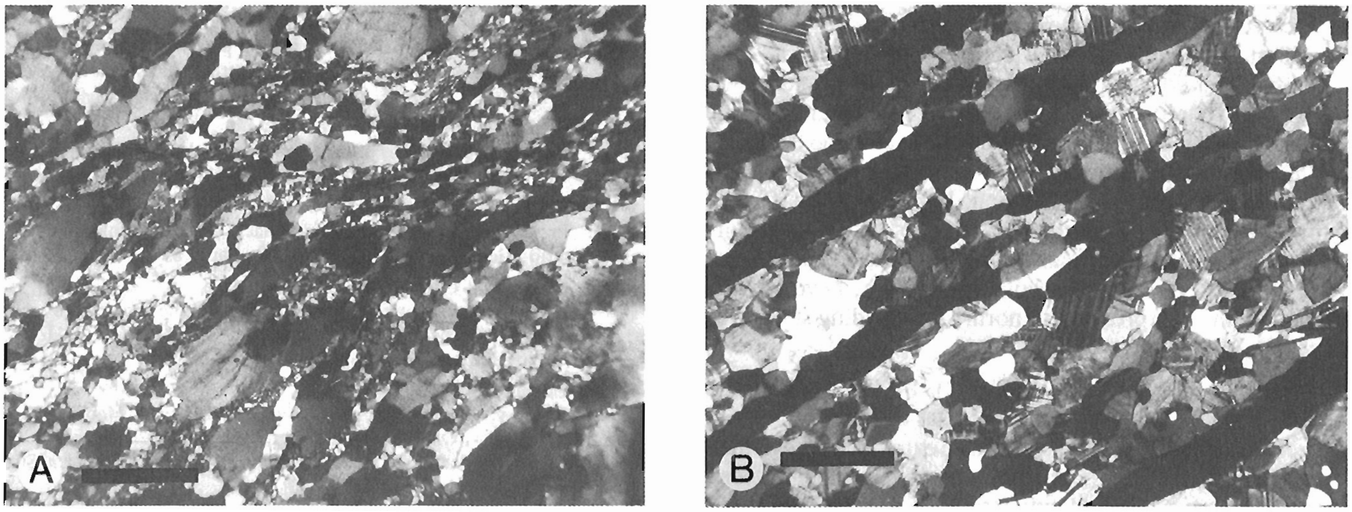

Figure 6. Photomicrographs, crossed polarizers, scale bar 500 microns, showing types of mylonitic texture typical of shear zone rocks west and east of Beaverstone Bay respectively. A - mylonitic texture in deformed granodiorite, northwest side of Beaverstone Bay (Fig. 2); B - annealed mylonitic texture from deformed granite, Fort Channel (Fig. 3).

western Bustard Islands. Some of the mylonitic rocks in the Bad River Channel and Fort Channel zones may have been derived from supracrustal screens between individual granitoid plutons, but because most of these rocks have compositions similar to the adjacent orthogneisses it is at this stage preferred to interpret them as the extreme products of annealing accompanying deep-crust mylonitization. Rocks of obvious metasedimentary protolith were not observed in the varied and complicated migmatitic gneisses in the area around Chaughis Bay (Fig. 3), for which development from plutonic protoliths through intense and repeated ductile deformation and migmatization is the preferred interpretation at present.

The identification of highly deformed coarse clasts of probable volcanic origin within fine grained feldspathic gneisses in the GFTZ west of Beaverstone Bay, together with the presence of suitable protoliths among the clastic volcanic rocks of the Killarney complex beyond the Grenville Front (unit kv in Fig. 2; Davidson, 1986a), suggests that the gneisses and metagranitoid rocks of Philip Edward Island may be the reworked equivalents of the felsic volcanic and plutonic association of the Central Plains orogen, of which the Killarney complex is considered to be the northeastern most representative (van Breemen and Davidson, in press; see also McGrath et al., 1988). The relationship of fine-grained feldspathic gneisses to the quartzite-pelite-amphibolite association at Mill Lake and Beaverstone Bay, marginal to them and apparently wrapping the southwest closure of these sediments in Mill Lake, tempts the interpretation that their contact may be the folded equivalent of the intrusive or faulted contact between Huronian sediments and volcanic-plutonic rocks of the Killarney type. The affinity of the Mill Lake and Beaverstone Bay metasedimentary rocks to the Huronian, however, is not clear. Although the limited thickness of lithologic units in these metasediments may be due to tectonic thinning, some of these units, particularly the aluminous pelites, and the close association of units, namely aluminous and calcareous pelite with thin orthoquartzites, and orthoquartzite with numerous amphibolite intercalations, do not have obvious counterparts in the Huronian succession, as was pointed out by Frarey (1985). The alternatives are that the metasediments represent either different facies equivalents of Huronian formations or units higher in the Huronian succession than are exposed in the Southern Province, or that they are not related to the Huronian Supergroup but perhaps to younger Proterozoic successions such as are found farther west in the Penokean fold belt. At present, however, derivation through a process of tectonic interleaving of thin slivers, out of sequence, of different Huronian formations cannot be ruled out.

INTERPRETATION OF GEOPHYSICAL DATA

There is a general correspondence between contacts of major rock units and gradients in the total magnetic field. From west to east along the coast of Georgian Bay, the felsic gneisses and metaplutonic masses between the Grenville Front and Beaverstone Bay are associated with positive anomalies, particularly the shear zone-bounded granite orthogneiss of unit 11 (Fig. 2), although the similar granite of unit 7 is magnetically less prominent. The metasedimentary rocks of Mill Lake and Beaverstone Bay, in contrast, correspond to magnetic lows. A major magnetic break coincides with the southeast shore of Beaverstone Bay, separating the magnetically 'noisy' terrane just described from a more subdued region to the southeast, corresponding with the metaplutonic migmatite complex, itself magnetically positive, to the southeast. The granitoid and syenitic orthogneiss units between Batt Bay and Bad River Channel (Fig. 3) are magnetically low, but the little-deformed granodiorite body between Bad River Channel and the main outlet of French River is expressed by a pronounced positive anomaly that tapers out southwestward in Georgian Bay. In Britt domain (Fig. 4), the change in structural trend from northeast through north to northwest is faithfully recorded by the orientation of magnetic gradients, the Pickerel complex coinciding with a prominent magnetic low.

The continuity of aeromagnetic anomalies in the GFTZ northeast and southwest of the coastal transect suggest considerable lateral continuity of the felsic gneisses and granitoid rocks exposed west of Beaverstone Bay, extending northeast and then north-northeast to terminate against the Grenville Front southeast of Sudbury, and southwestward parallel to the front. Southeast of this region, however, the different expressions, both positive and negative, of individual granitoid bodies makes it difficult to interpret the magnetic patterns beneath northern Georgian Bay which, beyond Britt domain with its expression of northwest-trending structure, is rather subdued.

Seismic reflection line J (Fig. 1), 80 km south of the coastal transect, shows a prominent set of easterly-dipping reflectors across the GTFZ, the surface extrapolation of which coincides with the aeromagnetically positive anomalous zone discussed above. The identification of several southeast-dipping high-strain zones containing mylonite and straightened rocks that juxtapose rock units of different lithologies and magnetic response prompts the interpretation that the reflectors correspond to similar zones in the GFTZ beneath Georgian Bay to the south.

A gravity survey was undertaken along the coast of Georgian Bay in August, 1987, preliminary findings of which are reported by McGrath et al. (1988); the survey showed that a major gravity low coincides with the Killarney complex and the felsic and granitoid gneisses southeast of the Grenville Front, and that an inflection in the eastward-rising gravity gradient corresponds to the southeast margin of the GFTZ.

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An extension of the Killarney complex into the Grenville Province based on a preliminary interpretation of a new gravity survey, Georgian Bay, Ontario

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Abstract

Regional Bouguer gravity data were augmented with 135 gravity observations measured along the northern and northeastern shorelines of Georgian Bay using a 2-km station spacing. An elongate, northeast-trending, 25 mGal amplitude negative Bouguer gravity anomaly with a length of 170 km, extends 60 km inland from the northern Georgian Bay shoreline where it is coincident with granitoid plutons and associated felsic volcanic rocks of middle Proterozoic age (1.75-1.47 Ga). The anomaly, therefore, is older than and unrelated to the Grenvillian Orogeny (1.3-1.0 Ga). The southeastern boundary of the Killarney complex (1.74 Ga), as interpreted from the gravity data, occurs within the Grenville Province approximately 12 km to the southeast of the Grenville Front Boundary Fault. Therefore the southeastern part of the Killarney complex was deformed during the Grenvillian Orogeny.

Résumé

Aux données gravimétriques régionales de Bouguer, sont venues s'ajouter 135 observations gravimétriques mesurées le long des lignes de rivage nord et nord-est de la baie Georgienne et obtenues en respectant un intervalle de 2 km entre les stations. Les auteurs ont constaté la présence d'une anomalie gravimétrique de Bouguer négative à amplitude de 25 mGal, étirée sur 170 km dans une direction nord-est et s'étendant sur 60 km vers l'intérieur à partir de la rive nord de la baie Georgienne où elle coïncide avec des plutons granitoïdes et où elle est associée à des roches volcano-felsiques datant du milieu du Protérozoïque (1,75-1,47 Ga). Par conséquent, l'anomalie est plus ancienne que l'orogenèse grenvillienne et sans parentée avec celle-ci (1,3-1,0 Ga). La limite sud-est du complexe de Killarney (1,74 Ga), telle qu'interprétée à partir des données gravimétriques, fait partie de la province de Grenville à environ 12 km au sud-est de la faille limitrophe du front de Grenville. Par conséquent, la partie sud-est du complexe de killarney était déjà déformée pendant l'orogenèse grenville.

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INTRODUCTION AND GEOLOGICAL SETTING

Recent geological and geophysical surveys in the vicinity of Lake Huron have focused new interest in the area. A crustal reflection profile extending from the east side of Georgian Bay to the northwest end of Lake Huron (east end of profile location shown in (Fig.1) was surveyed in 1986 under the Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE), (Green et al., in press). For results of new aeromagnetic surveys of Georgian Bay and Lake Huron see Geological Survey of Canada., (1987a, 1987b). Ongoing geological studies include shoreline sections in Georgian Bay extending from Killarney to Port Severn (Davidson and Bethune, 1988; Culshaw et al., 1988). Of interest in the present context are new gravity data collected in 1987 along two shoreline profiles, A-B and C-D, in conjunction with the geological mapping programs of Davidson and Culshaw. These profiles transect two prominent Bouguer gravity anomalies that occur over northwestern and south central Georgian Bay and over the adjacent land areas to the northeast in Ontario, K and P in Figure 1.

The Killarney gravity low, K, occurs at the Grenville Front, a geological province boundary extending from Labrador (Sharpton et al., 1987) to Alabama (Denison et al., 1984), and possibly as far as west Texas (King, 1975). The

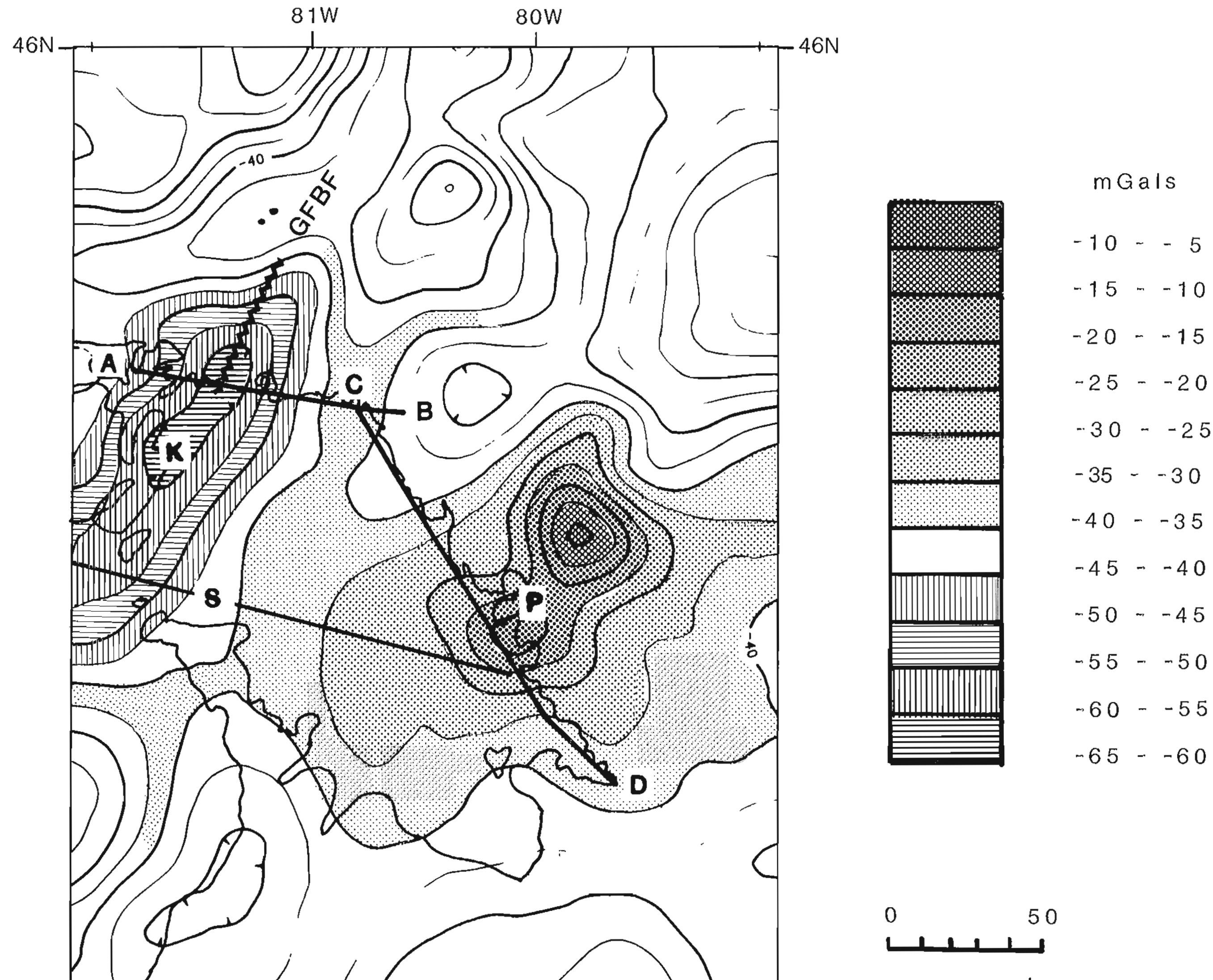

Figure 1. Bouguer gravity map with shaded contour intervals about the Killarney gravity low, K, and the Parry Sound gravity high, P. The map was derived from gravimeter observations spaced at 8 to 12 km intervals. Lines A-B and C-D represent the 1987 gravity surveys with stations at a two km station spacing. The survey results are illustrated in Figures 2, 3. The heavy line, S, marks the location of the east end of the crustal seismic profile surveyed in 1986 under the Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE). GFBF — Grenville Front Boundary Fault.

Grenville Front delimits the northwestern extent of the Grenville orogen (1.3-1.0 Ga), and separates the Grenville Province on the east and southeast from the Southern (> 1.83 Ga), Superior (>2.5 Ga), Churchill (< 1.8 Ga) and Nain (>2.5 Ga) provinces in Canada, and from the granite-rhyolite terranes (1.48-1.34 Ga) in the United States (Bickford et al., 1986). The occurrence of negative gravity anomalies is not, in general, typical of the Grenville Front in Ontario, although strong negative gravity anomalies are prominent at the front in parts of Quebec and Labrador.

The Southern Province of the Canadian Shield lies on the northwest flank of the Killarney low. It is composed of lower Proterozoic sediments of the Huronian Supergroup which were intruded by sills of Nipissing diabase (2.2-2.1 Ga), and folded during the Penokean orogeny (1.89-1.83 Ga) which imprinted east-west tectonic trends. Within a wedge-shaped area extending 80 km northeast of Georgian Bay, several elongate granitic plutons (1.75-1.47 Ga) have intruded these deformed Southern Province rocks. One of these, the Killarney granite (1.74 Ga), intruded at a high crustal level (Davidson, 1986a), and is part of a felsic volcanic-plutonic complex, the Killarney complex. Subsequent to intrusion, the southeast side of the complex was deformed about east- and northeast-trending axes. The deformed complex was transgressed by pegmatite dykes dated at 1.4 Ga (van Breemen and Davidson, in press), and by northwest-trending olivine diabase dykes of the Sudbury swarm dated at ca. 1.24 Ga (Palmer et al., 1977; Krogh et al, in press). The 1.74 Ga age suggests that the Killarney complex is an extension of the Central Plains Orogen which occurs southwest of the Michigan Basin under Phanerozoic cover (Bickford et al., 1986; Davidson, 1986a), and the lack of deformation of the 1.4 Ga pegmatite dykes establishes a northeast-trending pre-Grenvillian deformation for the complex. Other Proterozoic plutons, e.g. the Bell Lake granite (1.47 Ga) immediately to the northeast of the Killarney complex, are age-correlative with the granite-rhyolite terrranes of the United States.

The southeast edge of the Killarney complex is truncated by a northeast-trending mylonite zone, the Grenville Front Boundary Fault (Lumbers, 1975), which also deformed and displaced the pegmatite and olivine diabase dykes. The lack of significant Grenvillian-age deformation to the northwest establishes the mylonite zone as the favored site for the Grenville Front east of Killarney (Card and Lumbers, 1977; Davidson, 1986a). It has been argued that the early history of the Grenville Front includes the intrusion and pre-Grenvillian deformation of the Killarney complex, and that the Grenville Front is properly placed along the northwest side of the Killarney complex separating it from the Huronian metasedimentary rocks and Nipissing diabase (see Frarey, 1985). It seems reasonable however, to define the Grenville Front as marking the limit of Grenvillian-age deformation, as noted previously, and not to include older terranes possessing older structural trends which parallel those in the adjacent Grenville Orogen.

Line C-D transects the Central Gneiss Belt of the Grenville Province. New interpretations of this orogenic belt (Davidson, 1986b) suggest that the southwestern Grenville Province in Ontario is composed of a series of blocks and slices which were uplifted from mid-crustal depths along low-angle ductile shear zones by northwest-directed thrusting. Grenvillian tectonic activity produced no plutonism; deformation affected only older crust (see Easton, 1986, Fig. 10) and hence appears to be the only manifestation of the Grenvillian event.

The Parry Sound gravity high, P, is associated with one of the thrust blocks, the Parry Sound domain. Rocks within the domain are at a granulite grade of metamorphism and consist of predominantly mafic and intermediate gneisses with minor metasedimentary units intruded by gabbroic and granodioritic rocks. Parry Sound domain is surrounded on all sides by amphibolite grade metamorphic rocks consisting of para- and orthogneisses intruded by granitoid bodies now themselves deformed and metamorphosed. Interpretation of the Bouguer gravity anomaly led Lindia et al. (1983) to suggest that the Parry Sound domain extends to a depth of 13 km in its central region.

GRAVITY SURVEY AND PRELIMINARY RESULTS

In order to obtain detailed profiles across the Killarney gravity low and the Parry Sound gravity high for modelling purposes, the regional gravity data were supplemented by additional gravity measurements made by two of the authors (DWH and BF) along two profiles, A-B and C-D (Fig. 1) at a two-km station spacing. A LaCoste-Romberg Geodetic gravimeter was utilized for the survey, and observations were tied to base stations of the National Gravity Net. The elevation of each gravimeter station was tied to the surface elevation of Georgian Bay (178 m AMSL) using, where necessary, a hand-held level and survey rod. Bouguer anomalies were computed using a uniform rock density of 2.67 g/cm^3 and the sea level datum. Terrain corrections were not computed but are unlikely to exceed 1 mGal.

The Bouguer anomalies along profiles A-B and C-D are illustrated in Figures 2, 3, respectively. The geological cross-section for A-B is based on Davidson and Bethune (1988) and Card (1976).

For profile A-B, if it is assumed that the two sides of the Killarney gravity low, which is located between 10 and 50 km in Figure 2, can each be modelled using shallow depth contact models (Paul et al., 1966), then the positions of maximum horizontal gradient (Grauch and Cordell, 1987) can be utilized to make inferences regarding the location of density boundaries along the profile. Such an inflection point occurs on the northwest side of the Killarney gravity low at approximately the 20-km position on traverse A-B, and appears to be associated with the northwest contact of the Killarney complex. A second inflection point occurs on the southeast side of the Killarney low at approximately 42 km. These data may be interpreted to suggest 1), the Killarney gravity low is produced by rocks of the Killarney complex which are less dense than either the Southern Province rocks to the northwest or the Grenville Province gneisses to the southeast, 2), since there is a good correlation between the west side of the Killarney gravity low with the Killarney complex, the fact that the anomaly extends eastward 12 km across the Grenville Front Boundary Fault might suggest the rocks

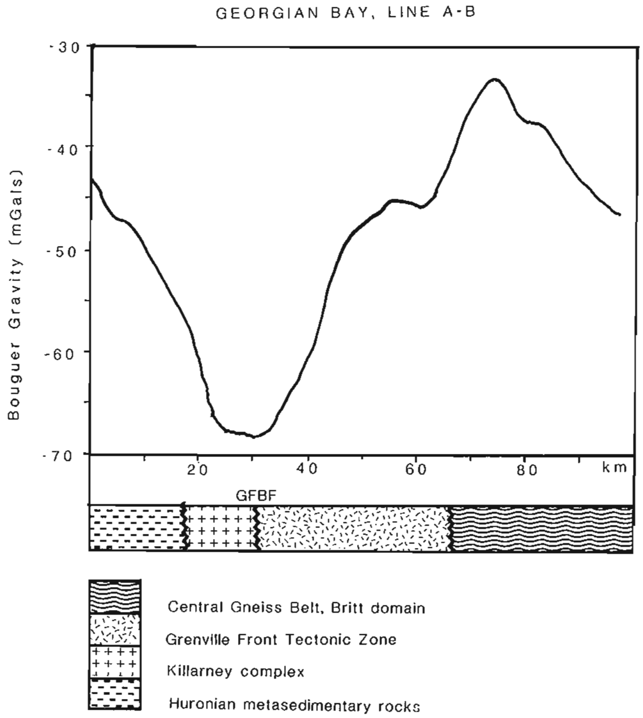

Figure 2. Profile along Line A-B of Bouguer gravity anomalies. The profile extends from west to east along the north shore of Georgian Bay. The geological cross-section is based on Card (1976) and Davidson and Bethune (1988). The Killarney gravity low anomaly between 10 and 50 km is associated with the Killarney complex. The positive Bouguer anomaly between 65 and 90 km is probably caused by a zone of more mafic gneisses occurring immediately to the southeast of the Grenville Front Tectonic Zone. GFBF — Grenville Front Boundary Fault.

up to 12 km east of the Grenville Front are deformed equivalents of the Killarney complex within the Grenville Province, and 3), given that the Killarney gravity low is produced by rocks older than the Grenville Orogeny, then caution must be exercised in utilizing the gravity anomaly to make inferences regarding the position of the Grenville Front in Georgian Bay unless it can be demonstrated that the Proterozoic granites were a controlling factor for the location of the Grenville Front east of the town of Killarney. The interpretation of deformed equivalents of Killarney complex to the southeast of the Grenville Front lends support to the observation of Davidson and Bethune (1988), who note the occurrence of felsic gneisses which are perhaps equivalent to the hypabyssal igneous and felsic volcanic rocks of the Killarney complex. These altered rocks are situated in a northeast-trending zone which extends 13.5 km eastward along the north shore of Georgian Bay from the Grenville Front Boundary Fault to Beaverstone Bay.

The 15 mGal positive anomaly occurring between 65 and 90 km on Line A-B occurs over a zone with a larger component of mafic gneisses than is typical in the adjacent rock units (Davidson, pers. comm., 1987). An inflection point

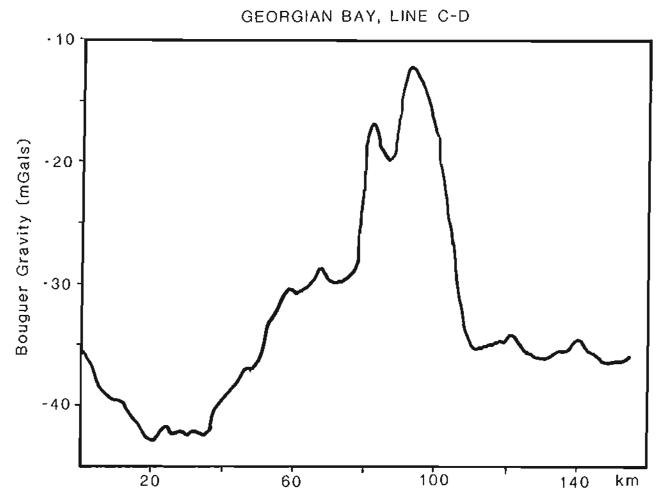

Figure 3. Profile along Line C-D of Bouguer gravity anomalies. The profile extends from northwest to southeast along the northeastern shoreline of Georgian Bay. The Parry Sound gravity high anomaly between 80 and 105 km is caused by dense rocks within Parry Sound domain. Britt domain occurs to the northwest and Go Home domain to the southeast of Parry Sound domain. The source of the negative anomaly between 0 and 50 km is unknown but may reflect an increased abundance of granitic plutons within the denser Grenville gneisses.

at 65 km may be related to the eastern margin of the Grenville Front Tectonic Zone (Davidson and Bethune, 1988), with Britt domain lying to the east.

The Parry Sound gravity high on Line C-D between 68 and 100 km is caused by the Parry Sound domain as has already been established by Lindia et al. (1983). The 20 mGal amplitude of the anomaly at the shoreline suggests that Parry Sound domain extends into Georgian Bay at least 5 to 10 km. The cause of the negative anomaly between 0 and 50 km over Britt domain on Line C-D is unknown. A preponderance of granitic plutons within the metasedimentary gneisses would produce such a negative gravity response.

The implications of the gravity survey, namely the lower density of the Killarney complex and other plutons as contrasted with the surrounding terranes, and the occurrence of deformed equivalents of the Killarney complex to the southeast of the Grenville Front will be examined by measuring the densities of field samples provided by Davidson and Culshaw, and by modelling the gravity data on Lines A-B and C-D.

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Diabase dykes and the Grenville Front southwest of Sudbury, Ontario

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Abstract

Between Sudbury and Georgian Bay to the southwest, two swarms of diabase dykes cut Huronian rocks and Proterozoic granites in the vicinity of the Grenville Front. They may be distinguished by appearance, petrography and chemistry. The younger east-trending, tholeiitic Grenville dykes cross into the Grenville Province without disruption. The older, olivine diabase dykes of the Sudbury swarm, trending between east and southeast, undergo progressive brittle to ductile deformation as they cross into the Grenville Front Tectonic Zone. This deformation is accompanied by metamorphism involving reaction between plagioclase and both olivine and Fe-Ti oxide. Within 8 km southeast of the Grenville Front, deformed and dispersed coronas of hypersthene-augite-garnet in place of olivine suggest a minimum pressure of around 600 MPa, intimating considerable uplift of the Grenville block relative to the Southern Province. Lateral offset of individual dykes is small relative to uplift which implies, along with the presence of pervasive down-dip lineation in the southeast-dipping gneissic rocks southeast of the front, that strike-slip displacement is of minor importance in this part of the Grenville Front zone.

Résumé

Entre Sudbury et la baie Georgienne vers le sud-ouest, deux groupes de dykes de dolérite recoupant des roches uroniennes et des granites protérozoïques se situent dans le voisinage du front de Grenville. Ils se distinguent par leur apparence, leur pétrographie et leur chimie. Les plus jeunes dykes de Grenville, tholéiitiques et à direction est, traversent la province de Grenville sans la déranger. Les dykes de dolérite à olivine, plus anciens, du groupe de Sudbury, d'orientation est à sud-est, subissent une déformation progressive qui va de fragile à plastique lorsqu'elle s'exerce à travers la zone tectonique du front de Grenville. Cette déformation s'accompagne d'un métamorphisme au cours duquel se produit une réaction entre le placioclase d'une part et l'olivine et l'oxyde de Fe-Ti d'autre part. Dans les 8 km au sud-est du front de Grenville, la présence de couronnes réactionnelles à la fois déformées et dispersées d'hypersthène-augite-grenat qui remplacent l'olivine suggère une pression minimale d'environ 600 MPa, ce qui indique un soulèvement considérable du bloc de Grenville par rapport à la province du Sud. Le déplacement latéral de dykes individuels est faible par rapport au soulèvement qui implique, en même temps que la présence d'une linéation pénétrante en aval-pendage dans les roches gneissiques que l'on trouve au sud-est du front et qui présentent une inclinaison également sud-est, que le rejet horizontal est d'une importance secondaire dans cette partie de la zone du front de Grenville.

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INTRODUCTION

On a recently published geological map of the Lake Panache — Collins Inlet area (Frarey, 1985), diabase dykes are shown to cross the Grenville Front. Palmer et al. (1977) identified two ages of diabase immediately northwest of the front on the basis of K-Ar age determinations and paleomagnetic pole positions, and suggested that the two are chemically different. Although Frarey (op. cit.) identified one dyke of the younger age group, other dykes of this set are not distinguished on his map from those of the older swarm. In the adjacent Burwash map area to the east, Lumbers (1975) made a distinction on the basis of whether or not the diabase contains appreciable olivine. On a map of diabase dyke swarms of the Canadian Shield (Fahrig and West, 1986), the northwest-trending olivine diabase dykes, which are termed the Sudbury swarm and assigned an age of 1220 Ma, are shown to terminate against the Grenville Front in the region southwest of Sudbury; a newly reported U-Pb baddeleyite age of 1238 ± 4 Ma (Krogh et al., in press) probably reflects more accurately the time of crystallization and intrusion than do the K-Ar ages. Dykes of the other swarm, termed the Grenville swarm and assigned an age of 575 Ma (Fahrig and West, 1986), trend east-west and are present on both sides of the front.

Lumbers (op. cit.) recognized a third category, termed 'cataclastic metadiabase', that occurs only in the Grenville Front Tectonic Zone (GFTZ) adjacent to the front. In the vicinity of Tyson Lake (Fig. 1), what appear to be folded dykes of olivine diabase in the GFTZ occur southeast of and on strike with straight-coursed dykes of the Sudbury swarm to the northwest; Frarey (op. cit., p. 48) described the diabase of the dykes at Tyson Lake as 'cataclastic'. Thin sections of samples collected from both Lumbers' and Frarey's map areas show that the Sudbury dykes, dykes at Tyson Lake and 'cataclastic metadiabase' bodies have many similarities; they could thus all have the same primary age and belong to the same swarm. The deformed ('cataclastic') diabases southeast of the front are also metamorphosed, showing reactions between plagioclase and both olivine and Fe-Ti oxide, in places with formation of corona texture, that suggest relatively high pressure conditions. This paper is the first report of a Ph.D. study being undertaken by the first author at Queen's University to test the kinship hypothesis, to elucidate the change in style of deformation within the diabase dykes as they enter the GFTZ, and to determine what metamorphic conditions prevailed during and/or after deformation in order to assess the amount of uplift along the Grenville Front.

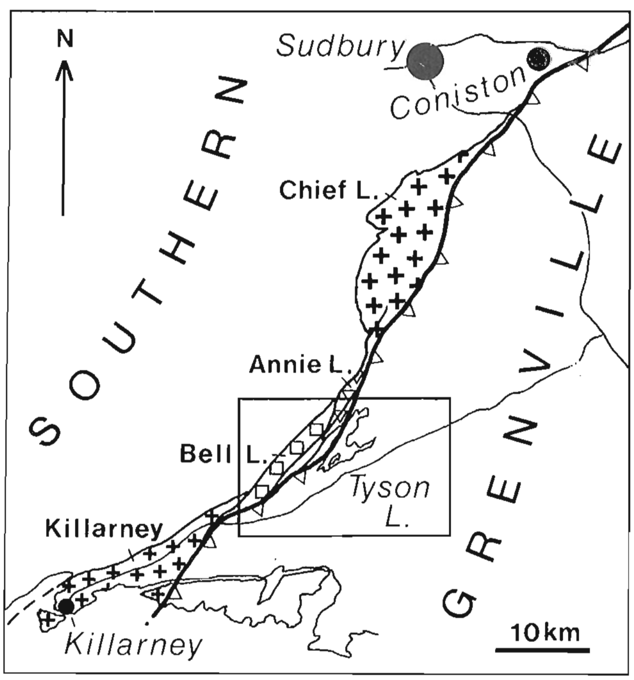

Figure 1. Location map of the study area. The Proterozoic granite plutons at the margin of the Grenville Province mentioned in the text are named.

REGIONAL GEOLOGY

Between Killarney and Coniston, Ontario (Fig. 1), a number of elongate granitoid plutons separate sedimentary rocks of early Proterozoic age (Huronian Supergroup, Southern Province) from predominantly high-grade (middle to upper amphibolite facies) gneisses of the Grenville Province to the southeast. For details of the geology of this region, see Lumbers (1975), Frarey (1985) and Davidson (1986), on whose work the following summary is based.

A thick succession of Proterozoic supracrustal rocks, the Huronian Supergroup, whose minimum age is given by cross-cutting gabbro sills (Nipissing diabase) dated at ca. 2.22 Ga (Corfu and Andrews, 1986), underlies the Southern Province. These rocks have been folded about east-trending axes and metamorphosed mainly to greenschist grade, effects attributed to the Penokean Orogeny (ca. 1.85 Ga). Penokean structures are clearly truncated by granitic plutons along this section of the front. Contacts with Huronian strata to the northwest are either intrusive or faulted. To the southeast the plutonic rocks are only locally in contact with recognizeable Huronian strata (Davidson, 1986); elsewhere they are in either sharp or gradational contact with gneisses or mylonites, presumed to have been formed during the Grenvillian Orogeny. The granitoid plutons are of at least two distinct ages, ca. 1.74 Ga and ca. 1.47 Ga (van Breemen and Davidson, in press). The older ones (Killarney, Chief Lake) are similar in age to deformed granites and rhyolites that make up the Central Plains orogen of midcontinental North America (Bickford et al., 1986); the younger (Bell Lake) is contemporaneous with granites and rhyolites that lie mainly south of the Central Plains orogen. The granitic rocks were deformed before mylonitization along the Grenville Front and related faults and before the development of moderately southeast-dipping foliation and southeast-plunging stretching lineation in the GFTZ to the southeast. For example, fabrics in the Killarney complex, a volcanic-plutonic rock association that includes the Killarney granite, are clearly different in orientation to those that characterize the Grenville Province (Davidson, 1986); they are also cut by non-deformed, pre-Grenvillian pegmatite dykes dated at 1400 ± 50 Ma (van Breemen and Davidson, in press). Steep, southeast-trending, pre-Grenvillian fabric (foliation, flattened xenoliths) is also present locally in the Bell Lake granite. The gneissic rocks in the GFTZ adjacent to the elongate granite bodies are in part deformed equivalents (augen gneiss, protomylonite, orthogneiss) of similar granites and perhaps felsic volcanic rocks (Davidson and Bethune, 1988), and in part metasediments whose affiliation to the Huronian Supergroup is equivocal (Frarey, 1985, p. 34).

The age-span of the development of 'Grenville-style', southeast-dipping planar and southeast-plunging linear structure along and adjacent to the front has not been adequately determined. K-Ar ages document uplift around 1.0 Ga; only one U-Pb zircon age of pegmatite close to the front east of Sudbury, 1150 Ma, has been obtained so far (Krogh and Wardle, 1984), although somewhat younger pegmatites (1025 Ma) are present farther southeast (Krogh and Davis, 1972). U-Pb zircon dating of gneissic granitoid rocks in the GFTZ have given only pre- Grenvillian ages of their protoliths. Rb-Sr whole rock dating of mylonitic rocks near Coniston has yielded an age-span that has been interpreted to imply long-lived tectonic activity along the front (LaTour and Fullagar, 1986), extending back in time beyond the generally accepted age-span for the Grenvillian Orogeny (1.2 — 1.0 Ga).

The Grenville Front northeast of Coniston is marked by a major fault along which Huronian strata at greenschist or even anchizone metamorphic grade are juxtaposed against middle to upper amphibolite facies gneisses in the Grenville Province. Anovitz (1987; Anovitz and Essene, in press) have obtained paleopressures between 810 and 1030 MPa within a few kilometres of the front in this region. Southwest of Coniston, Lumbers (1975; Card and Lumbers, 1977) identified the 'Grenville Front Boundary Fault' (GFBF) as a single feature in an area where there are several parallel or subparallel, anastomosing faults or mylonite zones. In the study area, Frarey (1985) showed a different position for the front, namely along the crosscutting northwest contact of the elongate granite bodies, up to 4 km northwest of Lumbers' boundary fault. Davidson (1986) has relocated part of the GFBF exposed near the coast of Georgian Bay and has argued against using the northwest contact of the pre-Grenvillian granites to define the Grenville Front (also see Davidson and Bethune, 1988).

As outlined in the introduction, deformed and metamorphosed diabase dykes in the GFTZ of the Lake Panache — Collins Inlet map area (Frarey, 1985) may be equivalent to the ca. 1240 Ma Sudbury swarm in the Southern Province. Individual olivine diabase dykes cross the Grenville Front as it is defined by Frarey (Fig. 2). They also continue with only minor offset across the GFBF as located by Card and Lumbers (1977) between Bell and Tyson lakes. East of the west shore of Tyson Lake, which lies east of Lumbers' GFBF, the apparently equivalent dykes have different geometry and orientation and show internal strain (see Frarey, 1985, Fig. 35), but nevertheless clearly cut across structure of Grenville style and orientation. Farther east in the GFTZ the same dykes are more highly deformed and carry a southeast-plunging lineation. It thus appears that, if 'Grenville-style' structure is to be equated with Grenvillian Orogeny, the dykes are syntectonic in an overall sense and orogeny began before ca. 1240 Ma. One of the aims of the present study is to clarify this relationship in order to understand the respective roles of pre- and post-dyke deformation. The region centred around Tyson Lake and straddling the Grenville Front, however defined, has been chosen for detailed study (Fig. 2). In this area several Sudbury dykes in the Southern Province lie opposite numerous, less regular bodies of deformed and metamorphosed olivine diabase in the GFTZ.

GEOLOGY OF THE TYSON LAKE AREA

The map area (Fig. 2) lies almost entirely within the Lake Panache — Collins Inlet map area of Frarey (1985); a narrow strip at its eastern margin is within Lumbers' (1975) Burwash map area. In the northwest corner the Serpent, Gowganda and Lorrain formations of the Huronian Supergroup, all of which are intruded by Nipissing diabase, dip steeply and face toward the south, being in the north limb of the La Cloche syncline. They are in faulted, intrusive contact with the Bell Lake granite to the southeast, a northeast-oriented, lenticular mass whose maximum width does not exceed 2 km. To the east, the northeast part of the Bell Lake pluton is described by Frarey to be in contact with variable and mylonitic, younger granitoid rocks referred to as the Annie Lake complex. Farther southwest, orthoquartzite with feldspathic and muscovitic quartz gneiss, correlated with the Lorrain Formation, are the contact rocks. These in turn wedge out between the Bell Lake granite and a mass of gneissic granite that lies southeast of Johnnie Lake.

Metasedimentary gneiss and schist including quartzite, pelite and minor amphibolite form a belt passing northeast through Tyson Lake. In places interlayered with narrow bodies of foliated granitoid rocks, this package dips moderately southeast. South of Tyson Lake these gneisses lie on the southeast flank of a large granitoid gneiss mass that widens southwestward and that is in mylonitic contact with the Bell Lake granite along Carlyle Lake. Metaplutonic rocks predominate southeast of the Tyson Lake metasedimentary belt and are interleaved with slivers of paragneiss.

Several northeast-trending faults or mylonite zones form an anastomosing network in the plutonic and gneissic rocks of the map area. They are shown both within map units and along many of their contacts (Frarey, 1985). The northwest contact of the Bell Lake granite, much of which is faulted, is the locus of Frarey's Grenville Front (Fig. 2). The southeast contact corresponds to Lumbers' Grenville Front boundary fault, except near the north edge of the map area where it is shown to step east and to lie along the east margin of the Annie Lake complex (Card and Lumbers, 1977). Intensely mylonitic rocks are exposed along the shores of the western part of Tyson Lake (Frarey, 1985, p. 54), extending northward along the east side of the Annie Lake complex and southwestward within granitoid rocks. Mylonites of this zone exposed on the west side of the peninsula in southern Tyson Lake have well-developed down-dip lineation and rotated feldspar porphyroclasts indicating thrust uplift to the northwest. A similar and consistent sense of displacement has been documented in mylonitic and protomylonitic rocks in the vicinity of Carlyle and Johnnie lakes (O'Donnell, 1986) and at the coast of Georgian Bay (Davidson, 1986).

There is a pronounced change in metamorphic grade across the map area. In the northwest, beyond the contact effects of the Bell Lake granite, siltstones of the Gowganda

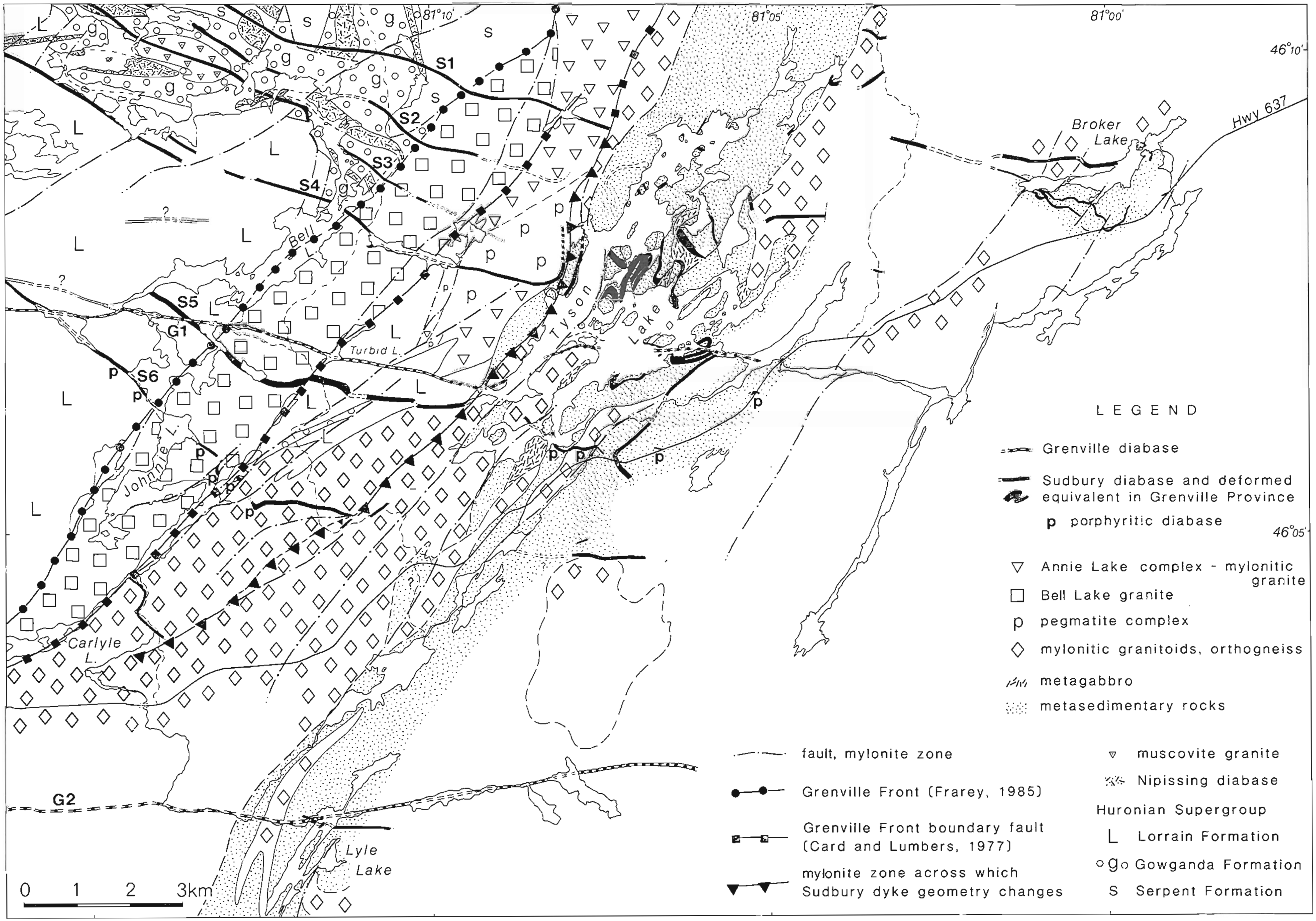

Figure 2. Diabase dykes in the Tyson Lake area. Geology modified after Frarey (1985).

Formation show greenschist facies assemblages. On the southeast side of this granite, micaceous quartzofeldspathic schist derived from arkose and feldspathic sandstone of the lower and middle parts of the Lorrain Formation contains quartz and muscovite stable together, and coarse, two-mica schist probably derived from Gowganda siltstone contains minor garnet, suggesting lower to middle amphibolite facies. Metasedimentary gneiss in the southeast part of the Tyson Lake belt generally lacks muscovite and contains sillimanite in association with K-feldspar, suggesting upper amphibolite facies. Relicts of orthopyroxene within aggregates of secondary biotite are present in metamorphosed gabbroic, granitoid and metasedimentary gneisses just southeast of Tyson Lake. The across- strike distance between rocks in greenschist and in granulite facies is at most 8 km. Rather than being a continuous gradient, the change in metamorphic grade is probably the result of progressive differential uplift of discrete blocks along faults and mylonite zones associated with the Grenville Front.

DIABASE DYKES

Distinction between Grenville and Sudbury dykes

Northwest of the Grenville Front where the Sudbury dykes are not metamorphosed, the Grenville and Sudbury diabases examined between Sudbury and Killarney both have a sparkling crystalline appearance on broken fresh surfaces. They may be distinguished, however, on the basis of the colour of the fresh rock and of the texture as seen on clean weathered surfaces. Grenville diabase has a green hue in comparison to Sudbury diabase, which is plain grey. Grenville diabase tends to have a thin, smooth, reddish brown rind and to show a pronounced diabasic texture; Sudbury diabase on the other hand has a characteristic grey-brown, speckled or spotted, rough surface with equant, brown-weathering mafic minerals set in a cream to tan-coloured matrix of interlocking plagioclase laths. In addition, Sudbury diabase commonly carries widely scattered plagioclase phenocrysts and, locally, crowded porphyritic zones parallel to the dyke margins, whereas Grenville diabase seems to be entirely aphyric. In thin section, Grenville diabase shows plagioclase ophitically intergrown with colourless to pale tan augite that surrounds cores of pigeonite; rare olivine grains are almost invariably serpentinized, biotite is rare or absent, and micrographically intergrown quartz and albite is present in the mesostasis between plagioclase laths. Sudbury diabase contains fresh, usually abundant olivine; its augite is pink to violet and has no associated pigeonite; biotite generally occurs in small amounts around Fe-Ti oxide grains and quartz is absent. On the whole, dykes of the Grenville swarm trend east-west whereas those of the Sudbury swarm are oriented more northwesterly; parts of individual dykes of one type, however, may locally have the orientation typical of the other type, so this criterion alone is not everywhere appropriate to distinguish them.

Whole rock analyses of five samples of Sudbury diabase from west of Tyson Lake and of three samples of Grenville dykes show a pronounced and consistent difference in rock chemistry (Table 1, columns 1 and 2; Fig. 3). As previously reported by Palmer et al. (1977), Sudbury diabase is strongly olivine-normative, iron- and alkali-enriched, and contains more Ba than Sr. Grenville diabase is tholeiitic; it is weakly quartz-normative and contains more Mg, Ca and less alkalis than Sudbury diabase.

Grenville dykes

Of eight Grenville dykes so far identified between Sudbury and Killarney, two occur in the map area (Fig. 2). Neither exceeds 50 m in width and neither is particularly well exposed; low outcrops generally occur widely spaced along the sides of pronounced east-west lineaments. One of these (G1 in Fig. 2), first identified by Frarey (1985) and sampled and dated by both Frarey and by Palmer et al. (1977) (K-Ar on biotite; seven age determinations between 400 and 500 Ma), passes from within the Lorrain Formation through the Bell Lake granite and the metamorphosed Lorrain to the east. At Tyson Lake it either changes direction northeastward until it reaches the east-southeast fracture occupied by Tyson Channel which it then follows, or it switches to this fraction in en echelon fashion. The other (G2) occurs in a prominent east-trending lineament passing through Lyle Lake in the south part of the map area. This dyke is in line with segments of Grenville dykes in the Huronian west of the Killarney complex and in the channel of the French River to the east beyond the GFTZ (Davidson and Bethune, 1988, Fig. 6). Both dykes cut metamorphosed olivine diabase dykes in the GFTZ.

Table 1. Whole rock chemical analyses of Grenville diabase, Sudbury diabase olivine metadiabase, Tyson Lake area.

Wt. per cent	1	2	3	4
SiO_2	48.7	47.2	47.7	48.1
TiO_2	2.20	2.88	3.04	2.64
Al_2O_3	13.7	14.6	14.8	14.3
Fe_2O_3	3.8	2.7	3.1	2.3
FeO	10.5	12.9	11.8	12.9
MnO	0.21	0.21	0.21	0.21
MgO	6.7	5.7	5.3	5.7
CaO	10.63	7.69	8.15	7.71
Na_2O	2.1	3.2	3.3	3.1
K_2O	0.53	1.36	1.24	1.24
P_2O_5	0.16	0.59	0.56	0.61
S	0.02	0.09	0.11	0.12
CO_2	0.3	0.1	0.2	0.3
$H_2O_{(T)}$	1.1	0.6	0.7	0.5
Total	100.63	99.82	100.21	99.73
ppm				
Ba	107	662	675	589
Sr	182	324	365	317
Rb	6	39	33	38
Zr	108	221	190	207
Y	31	50	37	48
Nb	nd	17	11	17
norm An	59	42	42	43
norm Qz/Ol	1.6 Qz	14.8 Ol	9.9 Ol	9.4 Ol

1 — Grenville diabase, average of 3 analyses; 2 — Sudbury diabase, average of 5 analyses; 3 — olivine metadiabase, Tyson Lake, average of 3 analyses; 4 — olivine metadiabase, Broker Lake, one analysis. XRF and rapid chemical analyses by Geological Survey of Canada laboratory.

Sudbury Dykes West of Tyson Lake

Frarey (1985, map) has identified seven northwest-trending diabase dykes in Huronian strata immediately next to the northwest contact of the Bell Lake granite. Six are shown to cross this contact without apparent offset. All were examined during the past field season, and all are olivine diabase dykes of the Sudbury type except the dyke that passes through Turbid Lake (G1 in Fig. 2), which is the northern of the two Grenville dykes already described; in addition, the seventh (S3) was also found to continue into the Bell Lake granite. Frarey shows three of the Sudbury dykes (S1, S4 and S5) to continue undisturbed across the east contact of the Bell Lake granite and through the rocks beyond. Detailed mapping of one of these (S5, Fig. 4) shows that as this contact is approached from the west the dyke curves toward the northeast, yet is not precisely in line with its continuation on the east side. At Johnnie Lake, dyke S6 was traced farther southeast than it was previously known to exist. Near the southeast contact of the Bell Lake granite it is offset southward along steep northeast-trending faults and cannot be traced to the shore (Fig. 4). A short segment of the same dyke is present east of the contact, but is again cut out by a fault and is not exposed in the fault block of mylonitic granite to the east. Beyond yet another fault, the dyke shows a sinuous trace at the west end of a relatively straight segment 2 km long. Adjacent to these faults the diabase shows cataclasis. That all these dyke segments belong to the same dyke is strongly suggested by the presence of zones of coarse plagioclase phenocrysts near its margins, a feature not noted in any of the other Sudbury dykes in this part of the map area. A narrow, porphyritic dyke exposed at the shore of Johnnie Lake, probably an offshoot of the larger dyke (S6) to the northwest, displays elongation of altered plagioclase phenocrysts (Frarey, 1985, Fig. 36).

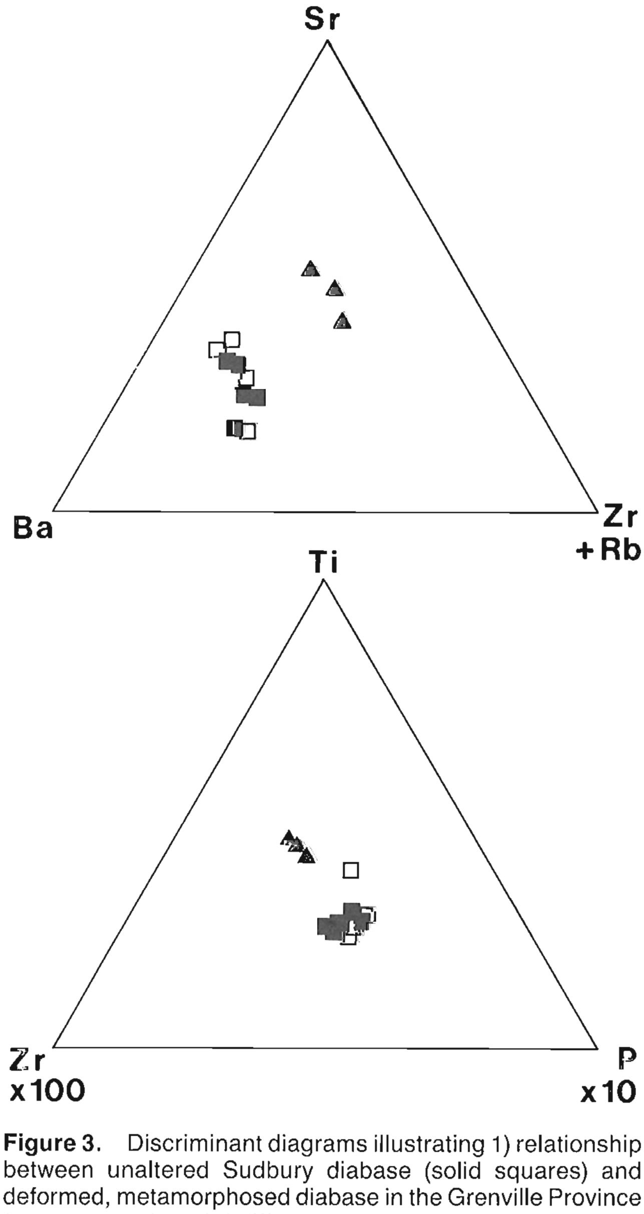

Figure 3. Discriminant diagrams illustrating 1) relationship between unaltered Sudbury diabase (solid squares) and deformed, metamorphosed diabase in the Grenville Province (open squares), and 2) difference between Sudbury and Grenville diabase (solid triangles).

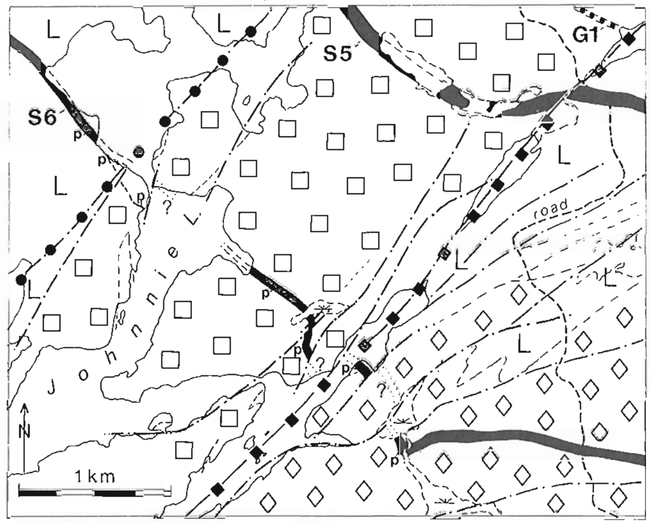

Figure 4. Relationship between Sudbury dykes S5 and S6 and faulting in the Johnnie Lake area. Map symbols are the same as in Figure 2.

Thin sections of Sudbury dykes west of Tyson Lake show pristine primary mineral assemblages and texture (Fig. 5A). However, in some samples taken close to northeast-trending faults or lineaments the olivine is partly or wholly altered to mixtures of chlorite, antigorite, magnetite dust and pale, fibrous amphibole. Samples taken near faults commonly show brittle deformation in the form of microbrecciation.

Olivine diabase dykes at Tyson Lake

Along the northwest shore of Tyson Lake and in line to the southwest, Sudbury dykes turn northeastward. Among the islands and peninsulas of Tyson Lake, olivine diabase has either the form of lobe- and hook-shaped masses that give the appearance of having been folded, or of straight or sinuous segments in various orientations that individually can rarely be traced for more than 1 km. The diabase in this area is darker grey than is typical of Sudbury diabase to the west, and is dull, lacking the crisp crystallinity of fresh diabase.

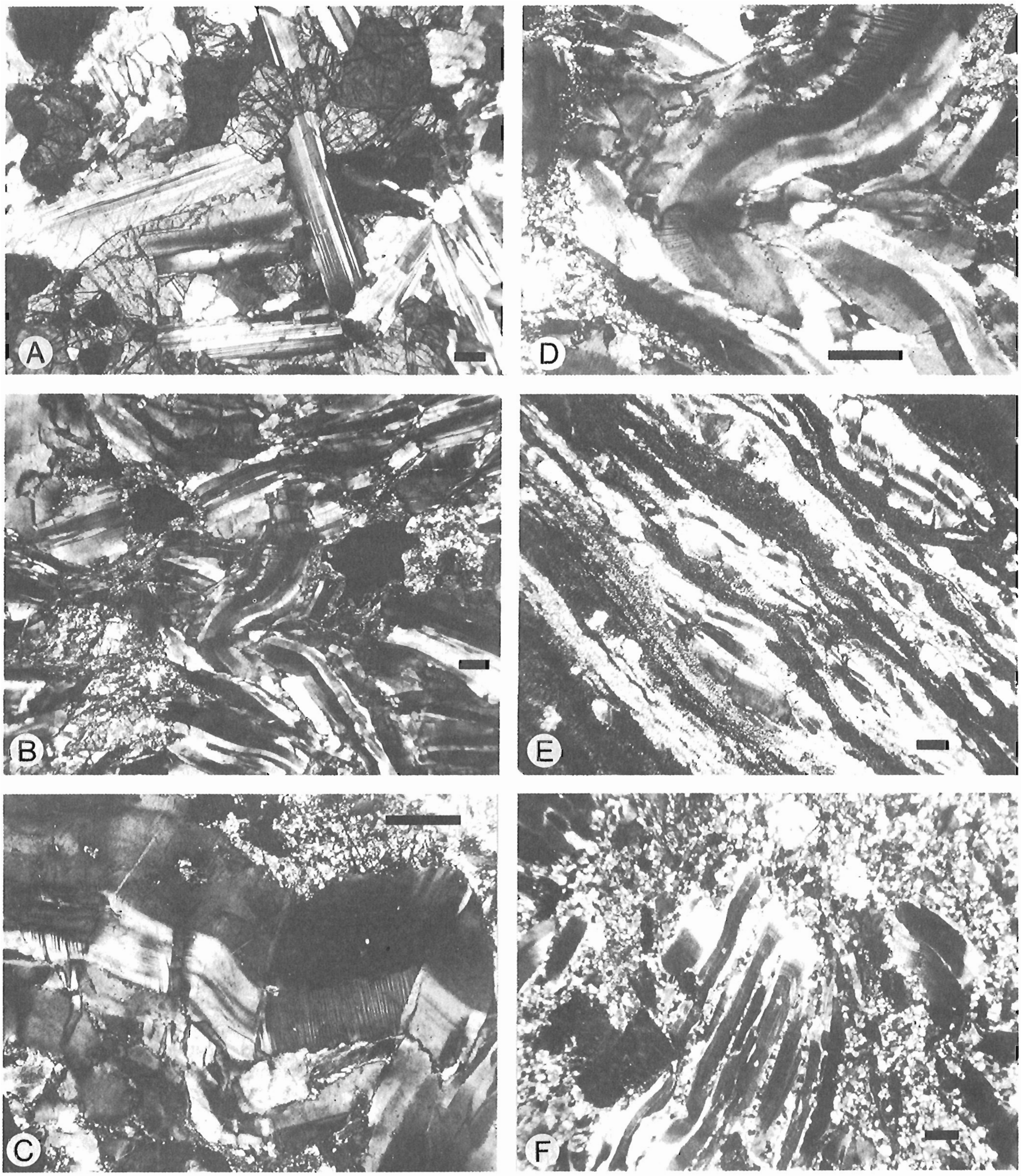

Figure 5. Photomicrographs, crossed polarizing prisms:

A. Typical texture of fresh Sudbury diabase, Southern Province;

B. Olivine metadiabase with deformed plagioclase laths and corona minerals around opaque oxide grains, southeast of Tyson Lake;

C, D. Details of bending and kinking of plagioclase laths in B;

E. Mylonite developed in shear zone in metadiabase, Lyle Lake.

F. Metadiabase showing relics of zoned plagioclase laths penetrated by fine recrystallized material, south of Broker Lake.

Scale bar in all photomicrographs is 200 microns

On the whole it also appears more fine grained, even though many of its masses are at least as thick as the widest Sudbury dyke (120 m). In some places it carries a faint fabric, locally seen to be associated with small mylonitic shears. Most of these masses have steep contacts that cut across foliation and layering in the metasedimentary gneiss country rocks. In two places exposed contacts show centimetre-scale folds with axes parallel to the southeast-plunging regional lineations.

Thin sections of diabase from the Tyson Lake vicinity show the same primary mineralogy as typical Sudbury dykes. Whole rock analyses of three samples from Tyson Lake are indistinguishable from Sudbury diabase (Table 1, column 3; Fig. 3). It is important to note, however, that contacts between olivine and plagioclase grains are the sites of growth of new minerals, namely orthopyroxene and pale green amphibole intergrown with green spinel, forming narrow coronas. In addition, fine-grained biotite forms narrow, irregular rims on Fe-Ti oxide grains and in some rocks is associated with brown hornblende and outer rims of spongy (symplectitic?) garnet. Garnet has also been noted as porphyroblastic patches in some chilled margins.

As noted by Frarey, and indicating a change at least as important as the onset of metamorphic reactions, thin sections of diabase from the Tyson Lake area show internal deformation in the form of fractured, bent and kinked plagioclase laths that have undulose extinction and impart a 'shattered' look to the rock (Fig. 5B, C, D); hence Lumbers' term 'cataclastic metadiabase'. This feature is apparent microscopically even in diabase that is macroscopically massive. Some of these samples have a mylonitic structure and, where present, large plagioclase phenocrysts may be wrapped by mylonitic foliation. Narrow mylonite and ultramylonite zones mark discrete shears in places (Fig. 5E). The internal deformation noted within diabase at Tyson Lake is consistent with the hypothesis that dyke geometry is the result of deformation.

Diabase dykes in the Broker Lake area

In the Broker Lake area, dull, dark grey olivine metadiabase occurs as east- trending dykes that appear to be more continuous than those just described at Tyson Lake. Parts of these dykes show remarkably sinuous courses. In addition, much of the diabase exhibits an obvious linear and planar fabric, both penetrative and restricted to shear zones; lineations plunge southeast parallel to those in the country rocks. The dykes cut across country rock foliation; observed contacts are commonly the sites of shearing.

Thin sections show further advances in both metamorphic mineral production and in style of deformation. Where least deformed and recrystallized, narrow coronas of orthopyroxene, clinopyroxene and rare garnet are present around primary olivine grains in contact with plagioclase, and garnet rims are well developed around biotite and hornblende adjacent to Fe-Ti oxide grains. Plagioclase laths, invariably bent regardless whether the dykes are straight or sinuous, show intense clouding by spinel dust that accentuates original zoning; this cloudiness is not present in plagioclase in diabase from the western part of Tyson Lake, but becomes increasingly apparent eastward. Where foliation is well developed, the metamorphic minerals are strung out in fine grained lenticles; plagioclase laths are penetrated by veinlets of finely recrystallized and nonclouded plagioclase (Fig. 5F). Mylonitic fabric is well developed in discrete shear zones.

One sample of olivine metadiabase collected at the highway south of Broker Lake was analyzed and is chemically indistinguishable from Sudbury diabase (Table 1, column 4; Fig. 3).

Porphyritic diabase

A distinct variety of diabase is characterized by large (up to 5 cm) plagioclase phenocrysts (or xenocrysts?) crowded together in a fine grained matrix. This type is found in narrow dykes (1 to 5 m wide) that are reasonably interpreted as off-shoots of larger dykes in which similarly crowded plagioclase crystals occur close to the margins. Apart from one narrow dyke at the north margin of the map area, porphyritic diabase has been found so far only in association with dyke S6 at Johnnie Lake (Fig. 2), in a dyke exposed at the south end of Tyson Lake and in several small dykes extending for 4 km to the east. In the Johnnie Lake area the phenocrysts are usually cream-coloured due to clouding by secondary low-grade alteration minerals, but locally unaltered, clear cores remain. In the Tyson Lake area they are black due to clouding by exceedingly fine opaque dust, in part with dust distribution controlled by the feldspar host lattice; no examples of low-grade alteration were observed. So far, porphyritic diabase has not been identified in the Broker Lake area.

Olivine grains are commonly found included in plagioclase phenocrysts and show coronas of secondary minerals due to reaction between these two primary phases. Those in the Johnnie Lake area show incomplete coronas of chlorite, antigorite, pale fibrous amphibole and opaque oxide. In the south Tyson Lake area, coronas of hypersthene surrounded by pale green amphibole with symplectitically intergrown green spinel are particularly well developed. The corona minerals around olivine thus reflect increase in metamorphic grade eastward within the Grenville Province, as do the products of reaction between primary minerals in the groundmass.

It is possible that the occurrences of porphyritic diabase between Johnnie Lake and south Tyson Lake are related to the same 'master' dyke. If so, their distribution suggests relatively little lateral displacement on faults and mylonite zones in the Grenville Front region. Metamorphic change in all the olivine diabase dykes suggests considerable uplift across this 5 km distance.

SUMMARY AND CONCLUSIONS

Of two sets of diabase dykes present in the Southern Province immediately northwest of the Grenville Front, the younger tholeiitic dykes of the Grenville swarm pass undisturbed into the Grenville Province. The older alkaline olivine diabase dykes of the Sudbury swarm (ca. 1240 Ma) become progressively deformed and metamorphosed as they pass into the Grenville Province. That deformed and in part discontinuous metadiabase dyke segments in the GFTZ are correlative with Sudbury diabase dykes in the Southern Province is suggested not only by their primary igneous mineralogy and by preliminary whole rock chemical analyses, but also by the on-strike continuity of a particular dyke with characteristic plagioclase phenocrysts near its margins and in spatially associated, narrow dykes that are likely offshoots from the main dyke.

Metamorphic assemblages of orthopyroxene, clinopyroxene and garnet derived by reaction between olivine and plagioclase in metadiabase in the Broker Lake area are probably indicative of pressure in the order of 600 MPa (Whitney and McLelland, 1973). That this reaction has taken place within a distance of 8 km across regional strike of unaltered olivine diabase west of Tyson Lake suggests considerable uplift across the northeast-striking faults and mylonite zones associated with the Grenville Front in this region. Continuity of diabase and metadiabase across the front in turn suggests very little lateral post-dyke displacement.

The most marked change in the olivine diabase occurs along a line that passes close to the northwest side of Tyson Lake and extends southwest toward Carlyle Lake (Fig. 2). This line is in part the Grenville Front Boundary fault of Lumbers (1975) and in part lies to the southeast. Northwest of this line the Sudbury dykes are cataclased, dislocated and show low-grade alteration locally at faults, between which they follow relatively straight courses and are neither internally deformed nor metamorphosed. Southeast of this line, reactions between plagioclase and both olivine and Fe-Ti oxide are evident, along with internal deformation shown particularly by twisted and kinked plagioclase laths; these effects increase in intensity toward the southeast, in parallel with increasing metamorphic grade in nearby metasedimentary gneisses.

Olivine metadiabase dykes, regardless of their state of deformation and metamorphism, cut across pre-existing foliation and lineation in their host rocks. This fabric southeast of Tyson and Carlyle lakes has the typical 'Grenvillian' orientation. Southeast-dipping foliation and especially southeast-plunging stretching lineation of this orientation is variably developed in the metadiabase, suggesting that intrusion of the Sudbury dykes was syntectonic with respect to Grenville-style deformation which, therefore, must have begun before ca. 1.24 Ga along this part of the Grenville Front Tectonic Zone.

ACKNOWLEDGMENTS

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Deformation and plutonism in the western Contwoyto Lake map area, central Slave Province, District of Mackenzie, N.W.T.[1]

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Abstract

The two turbiditic formations of the western Contwoyto Lake area (NTS 86E) are distinguished by characteristically thin (Contwoyto Fm) and thick (Itchen Fm) bedding as well as by the presence (Contwoyto Fm) or absence (Itchen Fm) of iron-formation. The Itchen Fm, which is interpreted to overlie the Contwoyto Fm, interfingers with a basaltic to intermediate volcanic belt, the Central volcanic belt. These volcanics are not considered correlative with the Point Lake Fm. Intrusive units include, by order of relative chronology, synvolcanic gabbros (C1), porphyritic granodiorite (C2), granodiorites and tonalites (C3), diorites (C4), tonalites (C5), and monzo- and syenogranites (C6). Metasediments have experienced four phases of Archean deformation (D_1-D_4), two phases of Proterozoic folding (D_5, D_6) and one phase of Proterozoic brittle faulting (D_7). The most intense phase of folding and cleavage formation (D_3) coincided with the metamorphic thermal peak. The multiple deformation produced complex structural patterns that can be defined using bedding-cleavage and cleavage-cleavage (S_2-S_3) relations together with the trace of marker beds. Two structural domes have been produced by fold interference.

Résumé

Les deux formations turbiditiques de la région ouest du lac Contwoyto (SNRC 86E) se distinguent par la présence de strates particulièrement minces (formation de Contwoyto) ou épaisses (formation d'Itchen) de même que par la présence (formation de Conwoyto) ou l'absence (formation d'Itchen) de formations ferrifères. La formation d'Itchen, qui serait sus-jacente à la formation de Conwoyto, s'interdigite avec une zone volcanique de nature basaltique à intermédiaire, soit la zone volcanique centrale. Ces roches volcaniques ne sont pas considérées comme étant corrélatives à la formation de Point Lake. Les unités intrusives comprennent, par ordre chronologique relatif, des gabbros synvolcaniques (C1), de la granodiorite porphyritique (C2), des granodiorites et des tonalites (C3), des diorites (C4), des tonalites (C5) ainsi que des monzo- et syénogranites (C6). Les métasédiments ont connu quatre phases de déformation archéenne (D_1-D_4), deux phases de plissements protérozoïque (D_5, D_6) ainsi qu'une phase de formation de failles cassantes du Protérozoïque (D_7). La phase la plus intense de plissement et de clivage (D_3) a coïncidé avec la pointe des conditions thermales associées au métamorphisme. La déformation multiple a produit des configurations structurales complexes qui peuvent être définies à l'aide des rapports stratification-clivage et clivage-clivage (S_2-S_3) conjugués à la trace laissée par les lits-repères. L'interférence des plis a engendré deux dômes structuraux.

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INTRODUCTION

The Contwoyto-Nose lakes mapping project (Fig. 1) was initiated in 1987 with the aim of: 1) completing a geological transect of the Slave Province between latitudes 65°N and 66°N through 1:50 000 to 1:100 000 scale mapping of the Contwoyto Lake (86E) and the west half of the Nose Lake (86F) map sheets and 2) furthering the understanding of gold mineralization in the area. In the course of the project the following topical studies are being undertaken: a study of the petrogenesis, geochemistry and isotopic signature of the igneous units by W.J.D., and a detailed structural, metamorphic and stratigraphic analysis of the low- to medium-grade supracrustal units by C.R. Relf, whose study is being supported by Canada-Northwest Territories Mineral Development Agreement (Project C1.1.1). O. Van Breemen (GSC) will continue geochronological studies in the eastern Slave Province in association with the project.

Previous work in the area by the Geological Survey of Canada includes 1:500 000 reconnaissance mapping during "Operation Bathurst Inlet" (Fraser, 1964), 1:50 000 scale mapping of the northwest corner of the Contwoyto Lake sheet (Tremblay, 1976), and 1:250 000 scale mapping of the west half of the Contwoyto Lake sheet (Bostock, 1980). Because of longstanding economic interest in the area, there are numerous geological maps of claim groups available in the assessment files of Indian and Northern Affairs Canada. Mapping

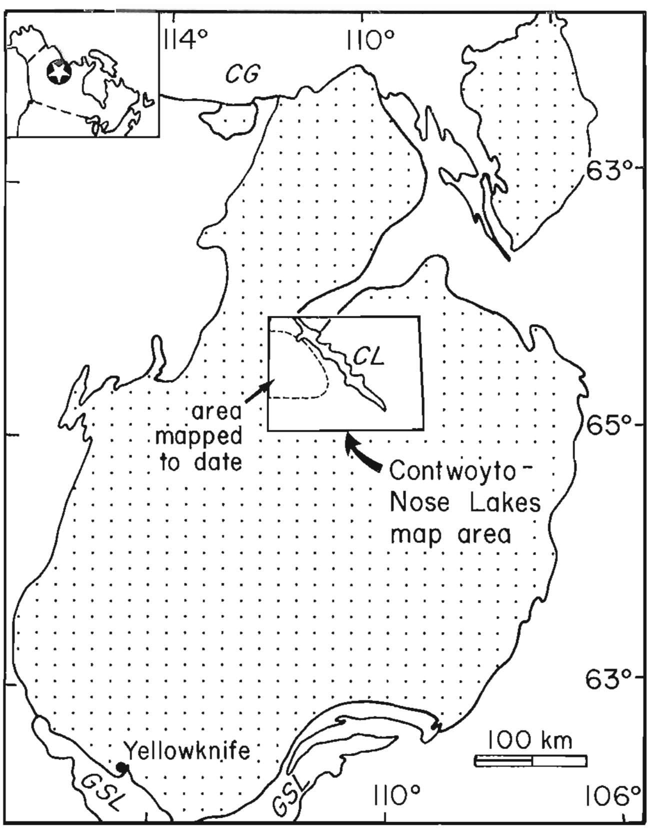

Figure 1. Location of the Contwoyto-Nose lakes project area. Patterned area is the Slave Structural Province. Inset shows position of the Slave Province in Canada. CG -Coronation Gulf; CL - Contwoyto Lake; GSL - Great Slave Lake.

for the present project started in the west half of the Contwoyto Lake sheet in order to re-examine some critical relations implied by Tremblay's (1976) and Bostock's (1980) maps, to subdivide the intrusive units outlined by Bostock (1980) and to carry out a structural analysis of the metasedimentary units. A compilation of this summer's mapping is presented in Figure 2. Work next summer will extend this database eastward where only reconnaissance maps are presently available.

The following report contains 1) a brief discussion of important characteristics of the Yellowknife Supergroup recognized this summer, 2) a description of the plutonic rocks in the map area, 3) a summary of the deformational history and metamorphism 4) a discussion of the relationship between deformation and plutonism in the map area, and 5) speculations on the possible tectono-magmatic environment. The character and distribution of the Quaternary surficial deposits are also briefly described. The report concludes with a summary of the implications of the results to date for gold exploration in the area.

YELLOWKNIFE SUPERGROUP

The Archean supracrustal rocks of the map area are, by the definition of Henderson (1970), part of the Yellowknife Supergroup. Bostock (1980) assigned local formation names as used below. Tremblay (1976), Bostock (1980), and Bubar and Heslop (1985) provided detailed descriptions of the units. The present report will discuss several important aspects of the Yellowknife Supergroup in the map area that are pertinent to the tectonic interpretation.

Central volcanic belt

Volcanics of the Central volcanic belt (CVB), interpreted by Bostock (1980) to be part of the Point Lake Formation (Fig. 2), range from basalt to rhyolite, but the dominant compositions are basaltic to intermediate. A well exposed sequence of CVB volcanics near Gondor lake (Fig. 2) was characterized by Bubar and Heslop (1985) as subalkaline to calc-alkaline. The basalts are characterized by massive and pillowed flows, with associated pillow breccia. Rhyolites form rare, small discontinuous flows. Both the basalts and rhyolites also form units that are locally identifiable as tuffs. In most places, however, these units can only be classified as volcaniclastics (Fisher, 1966). The intermediate units are most commonly volcaniclastics bedded at the centimetre scale, but locally may be massive flows. Heterolithic, coarse fragmental units and debris flows occur at several horizons within the CVB. The horizons of volcaniclastic units, although dominated by intermediate compositions characteristically contain finely interlayered mafic and felsic beds. From the variety of compositions throughout the CVB it appears that a range of magmas was erupted throughout the build-up of the volcanic pile. The volcanic structures indicate that the extrusive sequence was at least in part subaqueous. Stratigraphic facing indicators are too scarce to denote regional way-up or to define macroscopic folds.

Anthophyllite-garnet-cordierite schist (Fig. 3), previously included in the tuff units of the Point Lake Formation, was recognized this summer as a distinct unit (Fig. 2). Bedding was locally observed, but is generally obliterated by coarse recrystallization. Minor amounts of metasandstone and calcareous metasediments occur within the formation. The protolith to this units was probably a Mg-rich metasediment (cf. Tremblay, 1976). A primary relationship to the CVB of the Mg- rich sediments is postulated on the basis of its present spatial relationship to the CVB.

Contwoyto and Itchen formations

The Contwoyto and Itchen formations are both greywacke-shale turbiditic sequences, but are distinguished by the occurrence of iron-formation in the Contwoyto Formation and calcareous concretions in the Itchen Formation (Bostock, 1980). Detailed lithological descriptions of the turbidites and the iron-formation are provided in Tremblay (1976) and Bostock (1980).

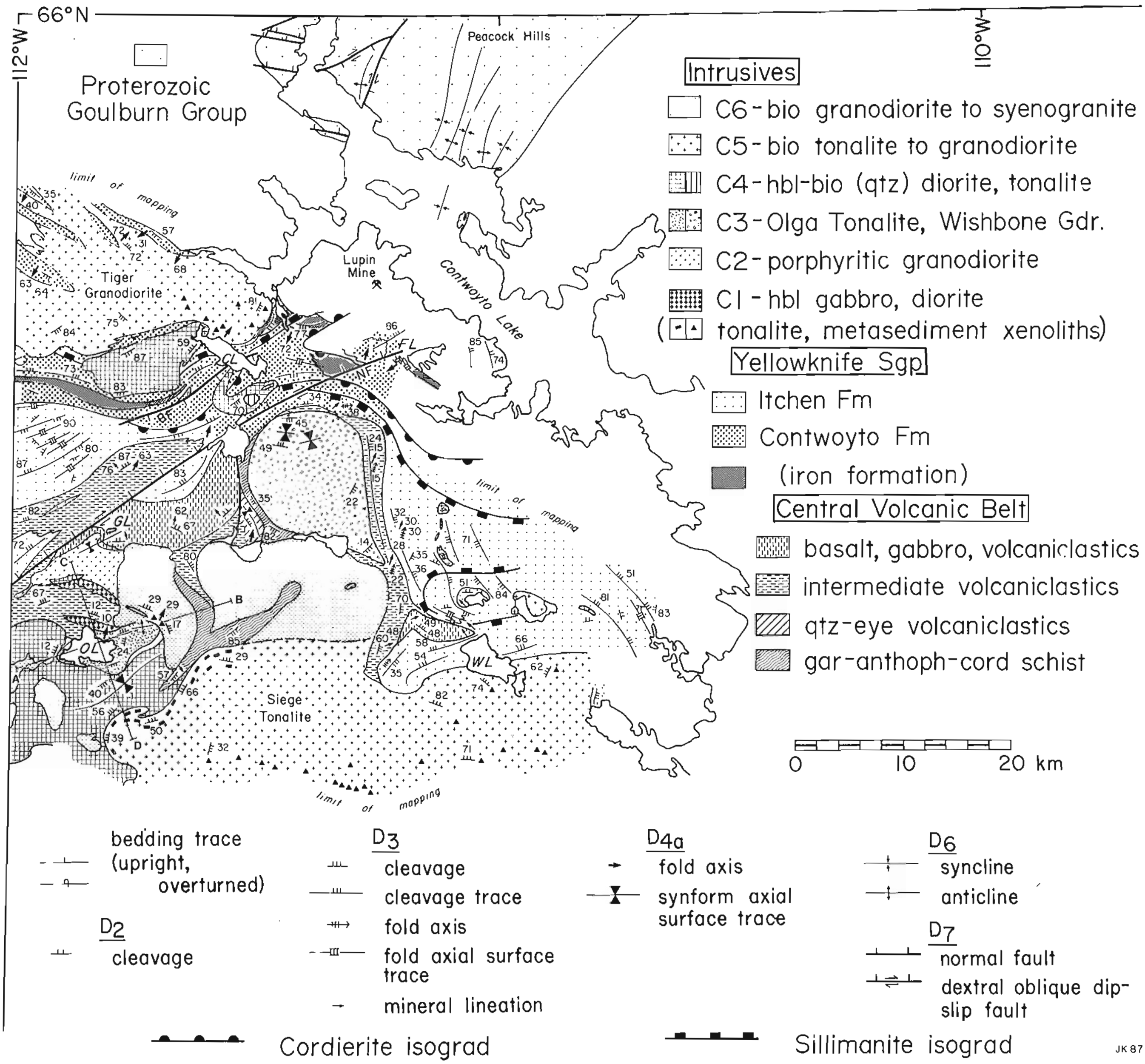

Figure 2. Simplified geological map of work to date in the western Contwoyto Lake area. "Iron-Formation" represents a stratigraphic horizon enriched in iron-formation, not a single bed. D_1, D_5 and D_6 structural elements were not observed at the map scale and are not included here. Units in the CVB are simplified. Trace of Proterozoic unconformity and structure northeast of Contwoyto Lake from Tremblay (1976). Isograd ornamentation is on the high-T side. Abbreviation: CL -Concession Lake; FL - Fingers Lake; GF - Grid fault; GL - Gondor lake; NF - Norma Fault; WL -Windy lake.

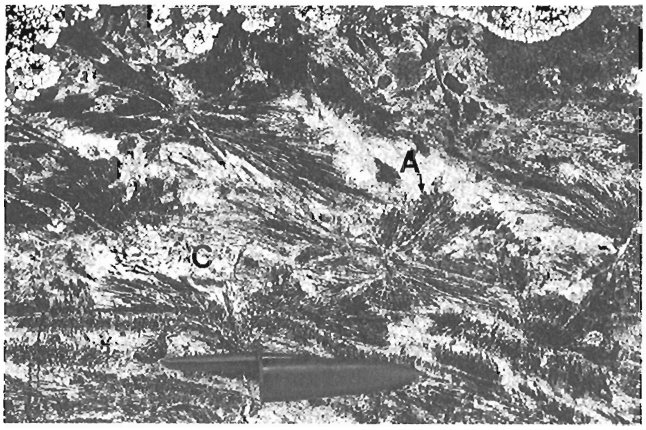

Figure 3. Anthophyllite (A) - cordierite (C) - garnet (G) schist. Pen cap is 6 cm. (GSC 20448-D)

Based on regional relationships west of the present map area, Bostock (1980) suggested that the Itchen Formation overlies the Contwoyto Formation and that the Contwoyto in turn overlies the Point Lake Formation. The map pattern in the western Contwoyto area, in which the Itchen Formation occurs between the Contwoyto Formation and the Central volcanic belt, is apparently not compatible with this interpreted sequence of units. There are two alternatives to resolving this incompatibility: 1) the CVB is not part of the Point Lake Formation, but is a second volcanic formation enveloped by the Itchen Formation; or 2) deformational processes (thrusting?) have duplicated the Point Lake Formation. No evidence for tectonic contacts has been found to date and the first alternative is therefore currently the more reasonable.

Two important differences, beside the presence or absence of iron-formation, between the two turbiditic formations were noted this summer. Firstly, the Itchen Formation is relatively thick-bedded, with beds ranging from 2 cm to at least 2 m, (averaging 15-20 cm, Fig. 4a) (cf. Bostock, 1980). In contrast, the Contwoyto Formation is thin-bedded with beds characteristically 0.5-10 cm (Fig. 4b) although thicker greywacke beds occur locally in the section. The second difference between the formations is that the Contwoyto Formation contains a number of tuffaceous and volcaniclastic beds while the Itchen Formation does not (cf. Gardiner, 1986). Active volcanism must therefore have been occurring during deposition of the Contwoyto Formation.

PLUTONIC UNITS

Six phases of plutonism (C1 to C6), have been distinguished on the basis of modal composition and interpreted chronology of emplacement (Fig. 2, 5). The phases, their identifying characteristics and the interpreted chronology of emplacement are described below. The C3, C5 and C6 phases are in large part refinements of Bostock's (1980) Central belt batholith, Contwoyto batholith and hybrid units. Nomenclature follows that of Streckeisen (1976). In the course of the following descriptions, references are made to several generations of structural elements. These elements are discussed more completely in the succeeding section.

C1 - Hornblende gabbros and diorites

These bodies, found only within the CVB (Fig. 2), occur as small dykes, sills and irregular plutons that have intrusive contacts with compositionally similar volcanic units. They are are typically well foliated and have similar metamorphic assemblages to volcanics of comparable composition. Because of their intimate spatial relationship with the volcanic belt, these bodies are interpreted to be synvolcanic, intrusive analogues to the volcanic rocks.

C2 - Porphyry dykes and porphyritic granodiorite

Porphyry dykes and sills are common in the CVB near the porphyritic granodiorite at the southwest margin of the belt (Fig. 2). The dykes and sills contain variable proportions of quartz and plagioclase phenocrysts in a groundmass that is too fine grained to classify on the basis of mineral assemblage in the field. The granodiorite is characterized by notably blue, rounded to euhedral quartz, and by a variable content of K-feldspar, plagioclase and biotite phenocrysts. Individual bodies are generally well foliated. The C2 bodies have intrusive contacts with the C1 plutons, dykes and sills,

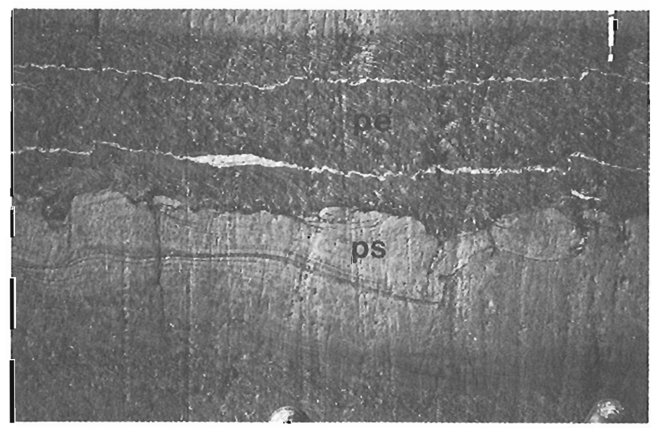

Figure 4a. Typical thick-bedded turbidites of the Itchen Formation. The lighter grey bed is a cordierite-bearing greywacke with primary structures still preserved. The darker grey bed is a cordierite-andalusite pelitic mudstone. Pen is 13.5 cm. (GSC 20448-F)

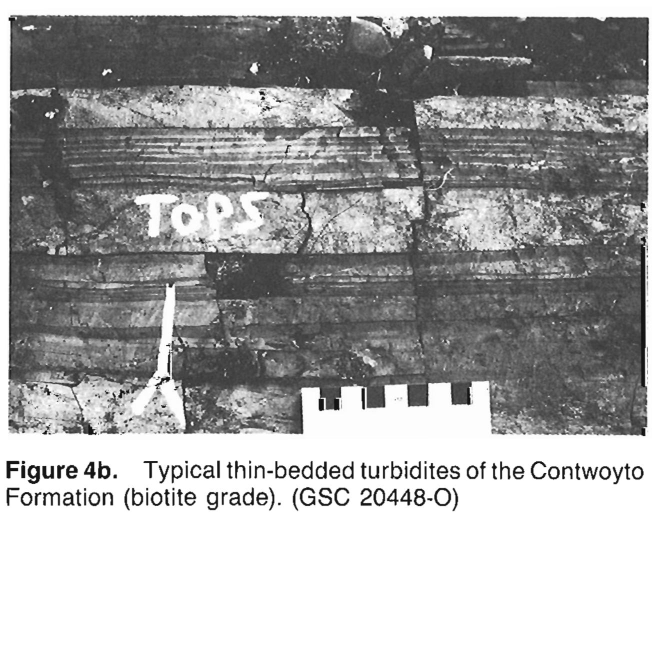

Figure 4b. Typical thin-bedded turbidites of the Contwoyto Formation (biotite grade). (GSC 20448-O)

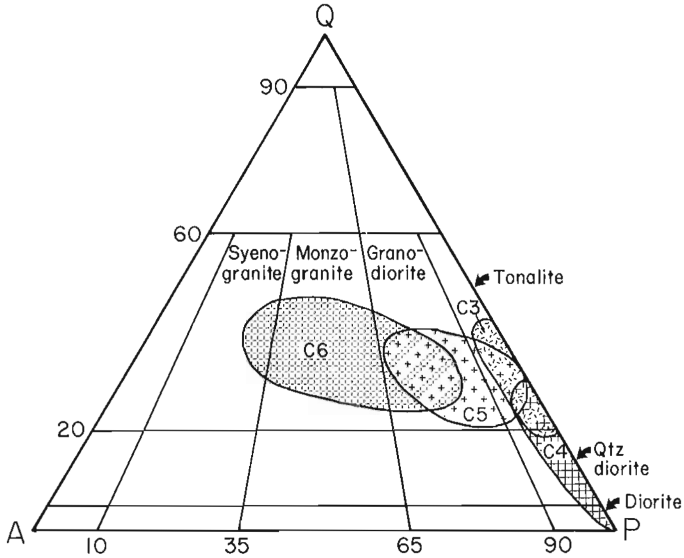

Figure 5. Compositional ranges of intrusive units in the western Contwoyto Lake area plotted on a Streckeisen diagram (classification based on field observation). Patterns as in Figure 2 except for the basalt-gabbro- volcaniclastic unit of the Central volcanic belt (CVB) which is left unpatterned.

and the structurally lowermost volcanics. Given the close spatial association with the volcanic belt, and the similar phenocryst assemblage to those in the intermediate to felsic volcanics, these bodies are interpreted to be genetically related to the volcanics, possibly generated from the same magma source.

C3a - Olga tonalite

The Olga tonalite is best exposed in the core of a structural dome (Olga dome) that is centred on Olga lake. It is composed of numerous (minimum of 4) texturally and mineralogically distinct intrusive units that range in composition from the predominant biotite tonalite to biotite-quartz diorite. Magnetite is a ubiquitous accessory, forming grains of less than 1 mm to 5 mm. The body appears to have been emplaced as multiple intrusions at the 10 cm to 10 m scale. Internal intrusions are generally sill-like resulting in a compositional layering which imparts a gneissic aspect to the body. All of the internal units are well foliated. The foliation generally parallels, but locally transects the intrusive contacts. Foliation is concordant with S_3, the dominant cleavage (see below) in adjacent supracrustal rocks. The contact between the Olga tonalite and the structurally overlying rocks of the CVB is typically sharp and concordant. North of Olga lake, narrow (metre-scale) sills of at least one phase of the

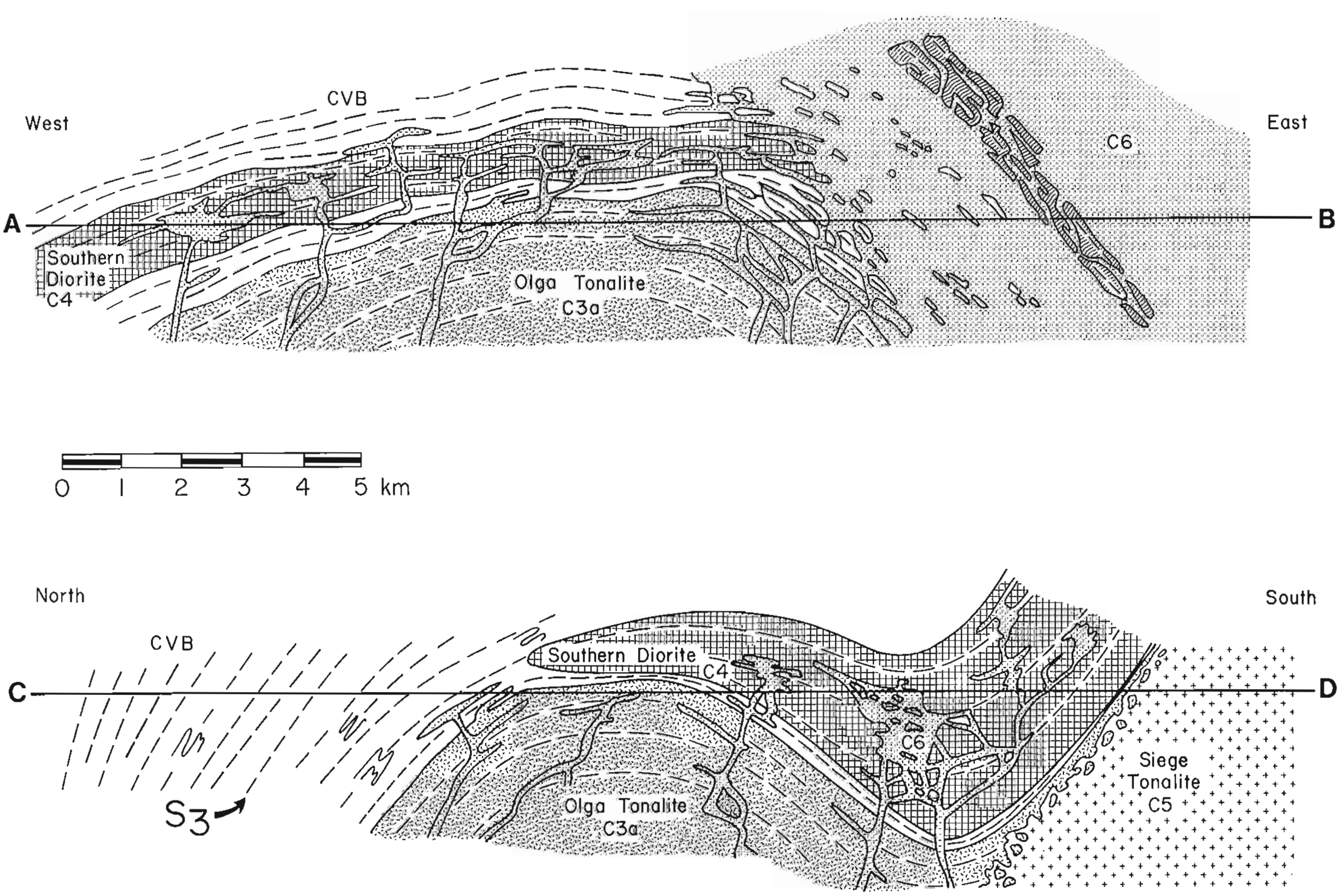

Figure 6. Schematic cross section of the Olga dome. Section lines shown on Figure 2. Patterns as in Figure 2.

Olga tonalite intrude the supracrustal units within 20 m of the unexposed regional contact. Large xenoliths of a tonalitic gneiss identical to the Olga tonalite are present in the marginal zone of a pluton (C5 Siege pluton, see below) that occurs south of the (C4) diorite-cored synform south of Olga lake (Fig. 2, 6). If extrapolated beneath the synform, the Olga tonalite-diorite contact resurfaces at about this contact. These xenoliths may therefore be the southward extension of the Olga tonalite. If so, the tonalite is likely a sheet-like body.

C3b - Wishbone granodiorite

The Wishbone granodiorite is exposed in the core of a structural dome (Wishbone dome) similar to that of the Olga dome (Fig. 2) and is composed of numerous texturally distinct, equigranular, fine- to medium- grained phases of biotite granodiorite to tonalite. Intrusive units are probably decametres to kilometres in scale but outcrop is too poor and the rocks are commonly too oxidized to trace contacts. The foliated granodiorite of the body corresponds in part to the 'Leucocratic granoblastic gneiss' (unit 1) of Tremblay (1976), but is interpreted to be a deformed granitoid rather than a quartzitic or tuffaceous rock (cf. Bostock, 1980) (the 'massive homogeneous red granite' described by Tremblay (1976) as a second unit in the Wishbone granodiorite body, largely corresponds to the late monzogranite (C6) of the present study). A foliation defined by alignment of biotite and, in places, flattened quartz and feldspar grains, is pronounced in the Wishbone granodiorite and is generally concordant with the S_3 foliation in the mantling supracrustal units. The foliation is locally crenulated coaxially with F_4 in the supracrustals, however, much of the outcrop is frost-heaved and this is only possible to document in the north part of the granodiorite. Intrusive contacts with the surrounding supracrustals were not observed. One kilometre-sized block of intermediate metavolcanic rock, similar to the mantling supracrustals, occurs within the granodiorite. It is not clear whether this represents a xenolith, roof pendant or an infolded portion of the overlying supracrustals (cf. Bostock, 1980). In the first two cases, the Wishbone granodiorite would be clearly intrusive into the volcanic-sedimentary package. In the latter case, the relationship of the Wishbone granodiorite to the supracrustals would remain uncertain.

C4 - Hornblende-biotite gabbro, diorite, quartz diorite and tonalite

The C4 intrusions are volumetrically dominated by diorite but include a range of compositions. The bodies are sill- or laccolith-like. The Southern pluton (Bostock, 1980), exposed south and west of Olga lake (Fig. 2), is the largest and most compositionally homogeneous of the bodies, being composed almost exclusively of a medium grained hornblende-biotite diorite to quartz diorite. The base of the Southern pluton is exposed south and west of Olga lake where its shallow-dipping, sharp contact is arched over the Olga dome (Fig. 2, 6). There is, however, no evidence that the contact of the pluton is necessarily intrusive. The topographic basin that contains Olga lake provides a window through The Southern pluton into the structurally lower Olga tonalite (Fig. 2, 6). A foliation, parallel with S_3 in the adjacent host rocks and concordant with the dome form, is variably developed within the pluton. The foliation also defines a synform on the southeast side of the dome (south of Olga lake) with 2 km of amplitude (Fig. 6). The top of the pluton is not observed and the minimum thickness of the body at this locality is therefore 2 km. The Southern pluton was a preferred site of emplacement for monzogranites and pegmatitic syenogranite of the C6 suite (see below), to the extent that a large part of the pluton is essentially an agmatite complex (Fig. 6).

Two spatially separate, lithologically distinct mafic plutons were mapped south of Concession Lake (Fig. 2). Although Bostock (1980) included both of the bodies in his definition of the 'Concession pluton', we retain the name 'Concession pluton' only for the larger western body. This body is a relatively homogeneous, medium grained, hornblende-biotite quartz diorite to diorite except along its northwestern margin, where a medium grained hornblende gabbro phase occurs. The gabbro may be either an earlier cognate phase intruded by the diorite, or a more mafic phase that intruded still liquid diorite resulting in magma mixing. The smaller, eastern body has a thick rim of medium grained biotite tonalite and a small core of medium grained hornblende-biotite diorite (Fig. 2). The diorite phase has clearly intruded the tonalite and there is abundant evidence of stoping and mechanical disaggregation of the tonalite by the diorite. Seven other bodies of the dioritic suite have been mapped to date (Fig. 2). These are small, sill-like, single or multiple intrusions similar in compositional range and intrusive relations to the pluton southeast of Concession Lake.

The dioritic plutons are heterogeneously, but generally well foliated with the exception of the set of sills north of Windy lake (Fig. 2), which are poorly foliated to nonfoliated. The foliation in each pluton is concordant with S_3 in the adjacent country rocks while the sills are also parallel to S_3. This suite is therefore interpreted to have been emplaced pre- to syn-D_3.

C5 - Biotite tonalite to granodiorite

These white weathering plutons are generally fine- to medium-grained, equigranular and contain accessory magnetite, apatite and muscovite. The plutons are large bodies (complete size not yet mapped out) (Fig. 2) whose shape in the third dimension is not known. Subtle internal variations in grain size, colour index and composition may reflect internal magmatic processes and/or a composite origin for the bodies. However, internal contacts have not been documented. Two major plutons have been defined to date. The southernmost, the Siege tonalite (Fig. 2), is characterized by brownish, subhedral quartz and euhedral, equant, white plagioclase. Biotite-rich xenoliths, schlieren and wispy folia are both broadly dispersed throughout the pluton (less than 5 %) and locally abundant in screens and patches. At the southern extent of present mapping, the pluton is essentially an intrusive migmatite composed of the tonalite plus xenoliths of paragneisses, orthogneisses and some diorite (possibly C4). The northwestern contact is a complex zone rich in xenoliths of strongly foliated tonalite. The xenoliths have been folded during ductile deformation, although the intruding tonalite is not itself deformed (Fig. 7). The xenoliths are therefore interpreted to have been deformed during intrusion.

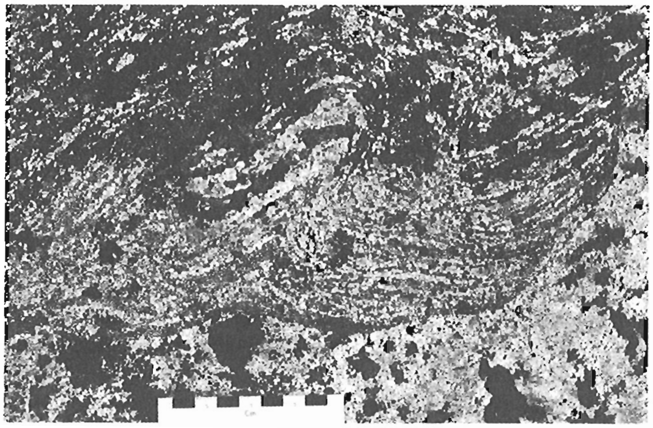

Figure 7. Xenolith of strongly foliated tonalite (Olga tonalite?) in the northwestern boundary zone of the Siege tonalite. This type of folding does not occur elsewhere and the intruding tonalite is not folded. The xenoliths are interpreted to have been deformed during intrusion of the Siege tonalite. (GSC 20448-M)

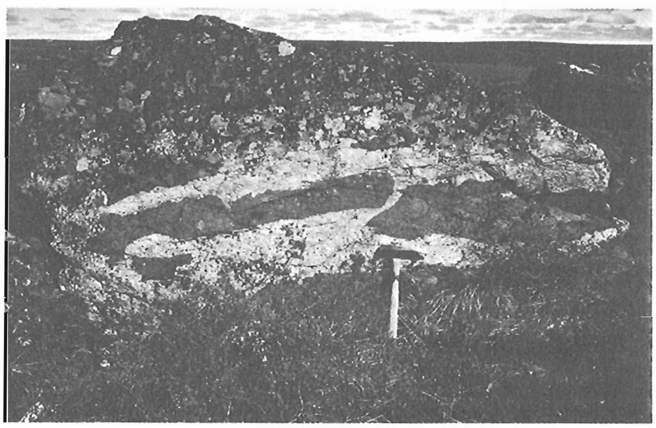

Figure 8. Angular xenoliths of basalt in a C6 monzogranite. Photo is from the northeast flank of the Olga dome. The dip of the foliation in the xenoliths is shallow, parallel with that in the non-intruded part of the northeast flank. There appear to be less than 20° rotation of the xenoliths from their pre-intrusion orientation. Because of this, trains of xenoliths such as these can be used to map structures and stratigraphy preserved within the C6 plutons. Hammer is 45 cm. (GSC 20448-P)

The more northern C5 pluton, the Tiger granodiorite (Fig. 2), is distinguished by the inclusion of abundant metre- to kilometre-scale schlieren, xenoliths, and screens of Contwoyto Formation metasediments, including some iron-formation. The xenoliths are less abundant and generally smaller on its eastern side, and increase in abundance and size range to the northwest where the body is a complex of intimately associated tonalite and metasedimentary xenoliths. The intrusive mechanism for the C5 suite appears to have been intense sill injection with assimilation of the host rocks. Local deformation of the wall rocks occurred due to continued movement of magma in marginal zones.

The C5 tonalites and granodiorites are weakly foliated to nonfoliated. Both the foliation and the contacts are broadly concordant with S_3 in the adjacent supracrustal rocks. The intrusion of this suite of plutons is interpreted to have occurred late in D_3.

C6 - Biotite-muscovite granodiorite to synenogranite and associated pegmatites

The granitoid and pegmatitic bodies of this suite intrude all other plutonic units. The granitoids are generally equigranular, medium- to coarse- grained, and contain accessory tourmaline, garnet and apatite. The most common composition is between granodiorite and monzogranite. Some bodies are locally K- feldspar porphyritic, and, more rarely, megacrystic. Associated pegmatites are common and are dominated by large, perthitic K-feldspar, quartz, muscovite and biotite. They commonly contain abundant, large tourmaline crystals and locally contain apatite and clusters of garnet or radiating sprays of sillimanite, muscovite and quartz. There is an apparent regional variation in pegmatites which is not reflected in major minerals but in the accessory phases garnet, tourmaline and sillimanite. The accessory phases are found in pegmatites intrusive into the turbiditic metasediments while they are generally absent where pegmatites intrude other host rocks. This variation is most likely related to assimilation processes. The intrusions are typically metre- to decametre-scale dykes and sills that mutually crosscut and coalesce. Map-scale bodies are in fact amalgamations of multiple dyke-and-sill intrusions, metres to hundreds of metres in scale (Fig. 6), that are compositionally similar but texturally distinct. Stratigraphic units and large-scale structures defined by stratigraphy and orientation of foliation (e.g. Olga dome) can be traced through C6 plutons as trains of xenoliths (Fig. 6, 8). There has been remarkably little (less than about 20°) rotation of the xenoliths during intrusion. This and the abundance of fractured, angular blocks of country rock within these intrusions (Fig. 8) suggest an intrusive mechanism of magmatic stoping at this level. The sharply bounded dykes and fractured blocks of country rock suggest that deformation during intrusion was dominantly brittle. Local, small deflections of pre-existing foliations in the country rock, however, indicate that minor ductile deformation also occurred during intrusion. The orientation of the dyke and sill system is strongly controlled by tectonic layering in the country rock (Fig. 6) in that the bodies tend to be sill-like along the foliation and to cut across the foliation at high angles.

The units of this suite are very weakly foliated to nonfoliated. They clearly crosscut S_3. Where thinner dykes or sills are favourably oriented, they are folded by F_4 (Fig. 9), This suite was therefore intruded post-D_3 and pre-to early in D_4. Although this timing infers that this suite is post-metamorphic, no contact aureoles were recognized, implying that the host rocks were still relatively warm when the C6 plutons were emplaced.

Discussion

In summary, the relative sequence of intrusion is interpreted to be (from oldest to youngest), synvolcanic hornblende gabbros (C1) and porphyritic granodiorite (C2), the strongly deformed tonalites (C3a, C3b) and diorites (C4), the weakly foliated tonalites (C5) that intrude C3 and C4 bodies and finally the weakly to nonfoliated monzo- to syenogranites (C6) that intrude C3 and C5 bodies. The compositional variation in time from early hornblende-diorites and tonalites through biotite tonalites and granodiorites to monzogranite and

Figure 9. Small veinlet, branching from a C6 pegmatite, that is folded by F_4. Pen cap is 6 cm. (GSC 204431-B)

syenogranite (Fig. 5) is a classic calc-alkaline trend. The discontinuity of the Olga tonalite (C3a) and the Wishbone granodiorite with this subsequent trend may be evidence that these bodies formed in a somewhat different tectonic setting than the later set of intrusions. The Olga tonalite is strikingly similar in composition and structural setting to the Hanimoor granitioid complex in the Hackett River Dome (Frith and Hill, 1975; Frith and Percival, 1978; Fyson and Frith, 1979) 140 km east of the present area. Radiometric dating has shown that at least one phase of the Hanimoor Gneiss Complex has an age of 2666 ± ca. 24 Ma (Frith and Loveridge, 1982). It is possible that the Olga tonalite is of similar age.

DEFORMATION HISTORY

Observations primarily in the metasediments have established a preliminary deformation history for the map area. The sequence of deformation events as defined to date include: 1) an early, enigmatic phase of folding (D_1); 2) two phases of

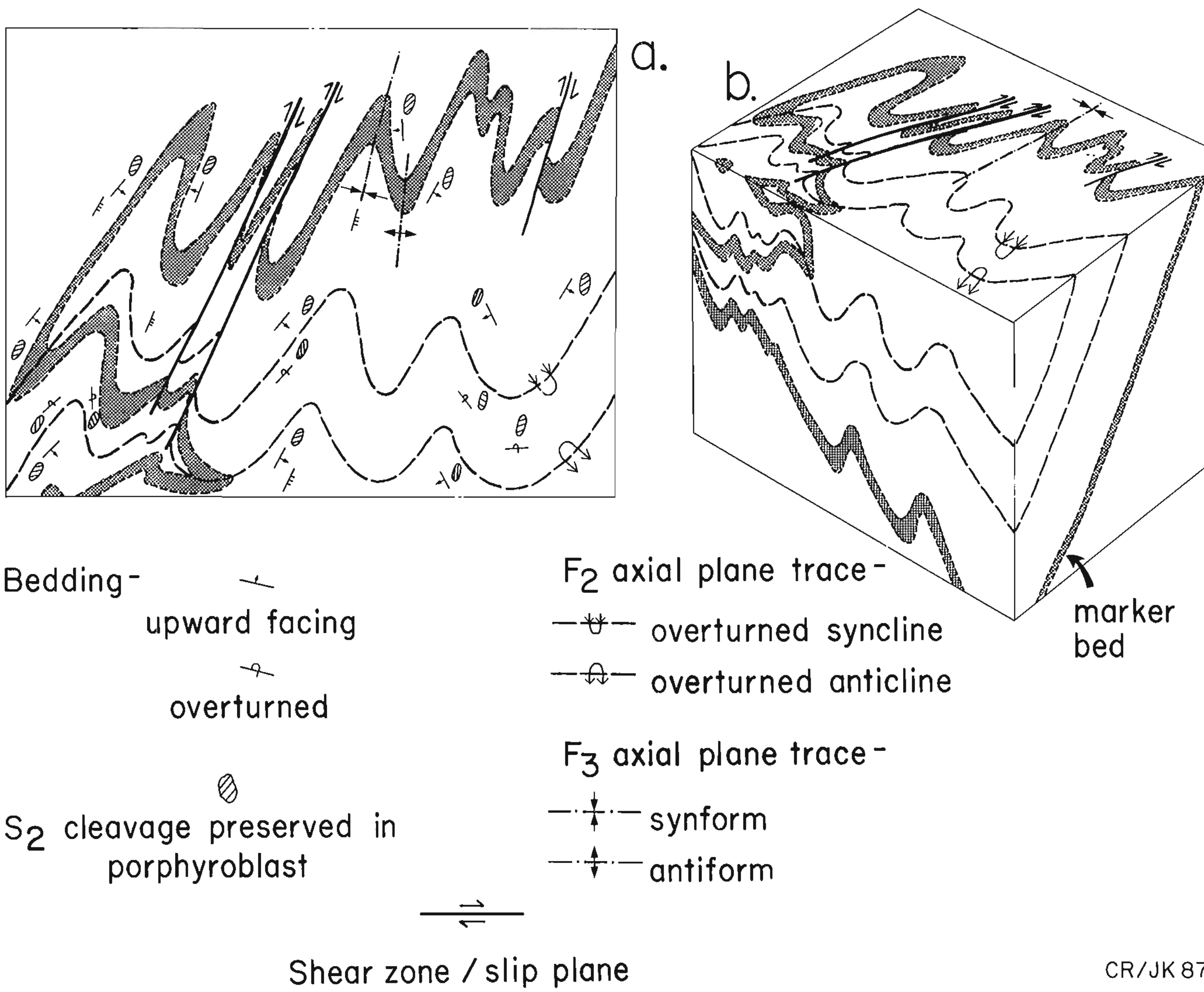

Figure 10. Schematic illustration of the style of, and interference between, bedding, F_2, S_2 (preserved in cordierite porphyroblasts), F_3, S_3 and syn-F_3 faults. a - map view; b - three dimensional view. The F_2-F_3 interference pattern approximates the type 3 of Ramsay (1967). Sketch is a composite from a number of outcrop areas.

isoclinal folding with associated penetrative cleavage (D_2, D_3); 3) superposed open folding with associated spaced cleavage (D_{4a}), discrete shear zones (D_{4b}), a conjugate fracture cleavage (D_{4c}); 4) two sets of Proterozoic folds (D_5, D_6) defined by regional relations, and 5) a system of joints and normal faults (D_7).

D_1

A single, isoclinal pre-S_1 fold, 10 cm in half-wavelength, was outlined by a 5 mm wide quartz vein. It is not known whether this fold is representative of the D_1 set of structures. The variable, sometimes steep plunges of the F_2 folds may in part result from pre-D_2 tilting of bedding. However, the D_2 folds may have been periclinal.

D_2

Folds of the D_2 generation (F_2) are isoclinal with first-order (recognized) wavelengths of ca. 100 m. Second-order F_2 folds, metres in wavelength are observed in outcrop. Where marker beds are lacking, the axial traces of macroscopic folds were defined using bedding-S_2 relationships (Fig. 10). Where marker beds, such as iron-formation, are well exposed, F_2 closures can be walked out. In their present orientation, the F_2 folds plunge moderately to steeply and are steeply overturned. Their original axial trend was at high angles to that of the superposed F_3 structures (Fig. 2, 10). The axial planar cleavage associated with F_2 (S_2) is a slaty biotite cleavage in the biotite zone. At cordierite grade it is a coarse biotite foliation that is often preserved within cordierite porphyroblasts and in microlithons between the S_3 crenulation cleavage. S_2 is not recognized in the sillimanite zone where it has probably been obliterated by coarse recrystallization. D_2 occurred during biotite grade metamorphism but before the thermal peak.

D_3

Folds of the D_3 generation are the most obvious in the map area, forming tight to isoclinal similar folds with wavelengths ranging from centimetres to hundreds of metres. Distinct orders of folds have not been distinguished. The D_3 structures plunge steeply down the limbs of the F_2 folds. Many of the attenuated limbs are disrupted by shearing or slip along S_3 surfaces, resulting in an imbrication of the fold limbs (Fig. 10, 11). Both bedding and S_2 are deformed by F_3; the resulting F_2-F_3 interference pattern approximates the Type 3 of Ramsay (1967) (Fig. 10). 'Z' asymmetries dominate the F_3 pattern west of the Wishbone dome whereas 'S' asymmetries dominate east of the dome. An S_3 foliation is in most places axial planar but locally transects the F_3 axial surface at a small angle. This is probably due to rotation of strain axes during fold and cleavage development, or a time lag between inception of folding and cleavage development (Borradaile, 1978). At biotite grade the S_3 foliation is a slaty cleavage. At higher grades it is a biotite-muscovite schistosity or a spaced crenulation cleavage. Where present, cordierite and andalusite are aligned along the S_3 plane. The long axes of porphyroblasts are either random on the S_3 plane or aligned parallel to F_3 axes. The porphyroblasts overgrew crenulation microfolds in various stages of development and are enveloped by S_3. S_3 is therefore interpreted to have developed during thermal-peak porphyroblast growth. Within quartz veins deformed by F_3, recrystallized linear streaks, interpreted as a stretching lineation, are also coaxial with F_3. By comparison with experimental models (Watkinson, 1975), such axis-parallel extension implies that F_3 folds are cylindrical.

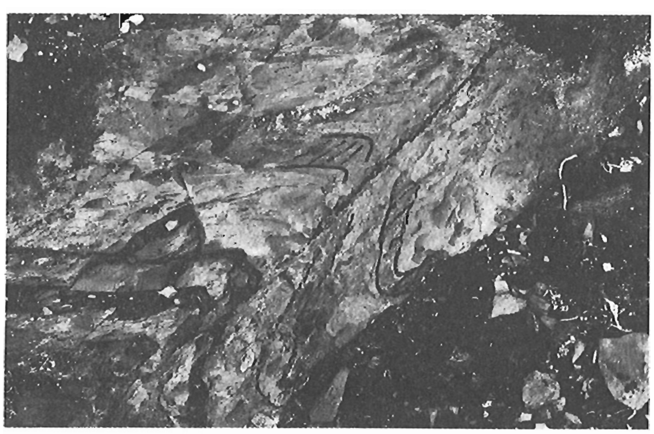

Figure 11. Displacement of an F_3 fold limb by syn-F_3 slip along the S_3 cleavage. Beds and fault trace are highlighted. Beds do not correlate across the fault. Photo is approximately 50 cm across. (GSC 204448-L)

D_4

Three sets of structures that post-date D_3 and predate the late fault set have been recognized. These sets do not overlap in space and their relative timing is therefore not well known. To facilitate discussion but to avoid relative time connotations they are designated as D_{4a}, D_{4b} and D_{4c} and the Proterozoic deformations as D_5, D_6 and D_7.

S_3 is folded by E- to NE-trending chevron folds (D_{4a}) that range in wavelength from centimetres to metres and plunge down the dip of pre-existing structures. In places there is an axial-planar cleavage defined by the alignment of biotite. The fold in which the Gondor massive sulphide deposit is located (Bubar and Heslop, 1985) is interpreted as a D_{4a} fold.

Local, narrow (10 cm to 5 m wide) shear zones (D_{4b}) transect S_3 at a small angle in sedimentary, volcanic and C4 and C6 plutonic units. In most cases a coarsely schistose fabric characterizes the zone but in two localities 10-50 cm wide shear zones are mylonitic. No kinematic indicators were observed. Visual inspection of synkinematic mineral assemblages suggest that shear zone development was syn- to post-thermal peak.

A third post-D_3 set (D_{4c}) comprises a heterogeneously developed conjugate fracture system with fractures spaced at 5-20 cm and oriented at at 010 and 095°. Where both of the conjugates are well developed they form a braided system. Where only one of the conjugates is developed its trace is serrated and apparently composed of linked conjugate segments. This fracture set is probably correlative with Tremblay's (1976) 'strain-slip' second cleavage.

D_5 and D_6 Proterozoic folding

Previous work has shown that the Slave Province has been involved in N-trending folds during the early Proterozoic Calderian deformation (King, 1986). This fold set is generally not recognizable at the outcrop scale within the Archean rocks (except in Exmouth Anticline beneath the metamorphic-internal zone of Wopmay Orogen (King et al., 1987)). The trace of the sub-Proterozoic unconformity, however, outlines a train of possible N-trending folds across the Slave Province (Fig. 12). These include mapped folds on the west flank of the Slave, and folds interpreted from the map pattern of the Rockinghorse outlier, and the cross-arched form of the Peacock Hills synclinorium. (Fraser, 1964; McGlynn, 1977; King, 1986). These arches and basins are spaced at progressively increasing wavelengths, from 20 km to 80 km, west to east across the Slave (Fig. 12). This wavelength of folding implies crustal-scale shortening. As such, these folds probably represent thick-skinned shortening of the Slave Province during E-W directed compression initiated on its west flank.

It is also clear from published maps (Fraser, 1964; Tremblay, 1976; Bostock, 1980) that the Early Proterozoic Goulburn Group at the north end of Contwoyto Lake is folded together with its Archean basement (Slave Province) about shallowly NE-plunging axes (Fig. 2, 12). Unfortunately the folds plunge too shallowly and are generally too open for their effects to be recognized in the already multiply deformed Archean basement away from the trace of the unconformity. However, the involvement of at least the northern Slave Province in a NE-trending Proterozoic fold set is documented by the trace of the sub-Proterozoic unconformity around the large culminations and saddles on the west flank of the Slave Province (King, 1986; Hoffman et al., in press) and around the SW-extending arm of Goulburn Group (Fraser, 1964; McGlynn, 1977). Wavelengths of 80-140 km and amplitudes of up to 15 km characterize this fold set. The D_6 folds are considered to be related to Tree River deformation (Hoffman et al., 1984; King, 1986).

If the correlation of the fold sets with the Calderian and Tree River deformations is correct, the N-trending fold set predates the NE-trending set (ie. N-trending is D_5 and NE-trending is D_6 in the present structural nomenclature). The two sets of Proterozoic folds must interfere in a macroscopic Type 1 (Ramsay, 1967) dome-and-basin pattern. Although it is difficult to recognize the effects of the Proterozoic folding within the basement, there is potential at the scale of the Slave Province for recognizing the domes and basins. Geobarometry and careful examination of structural-stratigraphic relations, viewed from a regional perspective, will be the tools with which to identify these fold domains within the Archean.

Part of the southeastern limb of the Peacock Hills synclinorium (Fig. 2, 13) has previously been interpreted as a shallow thrust that placed Archean over Proterozoic (Tremblay, 1976). Reconnaissance this summer has shown, however, that basal Proterozoic strata are present adjacent to the

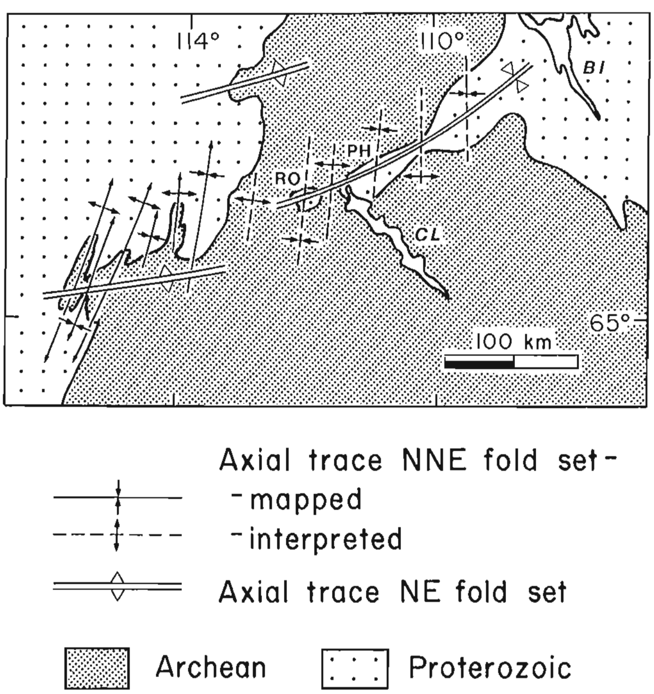

Figure 12. Mapped and interpreted Proterozoic folds that involve the Archean Slave Province. Abbreviations: RO - Rockinghorse Outlier; PH - Peacock Hills; BI - Bathurst Inlet; CL - Contwoyto Lake.

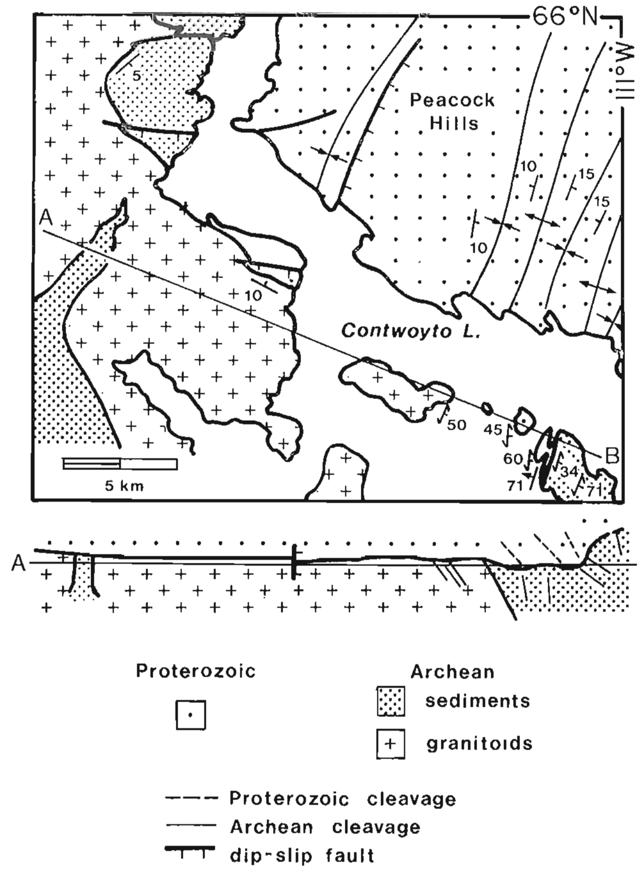

Figure 13. Map and cross-section showing the fold pattern of the Proterozoic-Archean contact near the north end of Contwoyto Lake. The southeastern limb of the Peacock Hills synclinorium is interpreted to be a steeply up-turned unconformity. Map modified from Tremblay (1976).

basement (J. Grotzinger, pers. comm., 1987), and dip 70° to the NW. (Fig. 13). Based on these observations, there is no stratigraphic need to invoke a thrust at the contact. Rather, the sub-Proterozoic unconformity is interpreted to be preserved, inclined at a 70° dip. The shallowly dipping cleavage in the Archean rocks at the contact is a coarse muscovite schistosity and is similar to the Archean S_3. It is distinct from the much lower grade, sericitic(?) phyllitic cleavage that occurs in the basal Proterozoic units. The shallow cleavage appears to be simply an overturned Archean cleavage (Fig. 13).

D_7 - Proterozoic faulting

The final set of structures recognized in the area includes ubiquitously developed joints and three major faults. The description by Tremblay (1976) of the joint set in the northwest corner of the area is generally valid for joints throughout the map area. Although the development of the joint set cannot be unambiguously related to that of the major faults, similarities in orientation, brittle deformation, and timing imply that they are related.

The trace of the Norma fault, initially identified by Tremblay (1976), has been extended southwestward through the Central volcanic belt to the edge of the present map sheet (Fig. 2). Its trace is continuous with that of the post-metamorphic "Point Lake fault" that extends northeastward from Point Lake (King, 1981). The name 'Point Lake fault' should therefore be dropped as "Norma fault" has precedence in the literature (cf. Tremblay, 1976). The total length of the fault, from Point Lake to the NE corner of the present map area, is 100 km. In the immediate vicinity of the fault, discontinuous pseudotachylytes, cataclastic zones, centimetre to decametre-scale kink folds and quartz- and carbonate-filled fractures are present. These features indicate that the fault is a relatively brittle structure. Deflection of country rock foliations adjacent to discrete minor faults of the system indicate that minor ductile deformation also occurred, probably related to strain accumulation before rupture.

The apparent displacement on the Norma fault in the Contwoyto Lake area, as derived from the apparent offset of the moderately E-dipping cordierite isograd. is 3 km (Fig. 2). This displacement can be accounted for by either dip-slip or strike-slip displacement, or a combination of the two. Displacement on the westward extension of the fault in the Point Lake area has been interpreted as dip- slip (King, 1981). In addition, units in this area do not exactly match across the fault, indicating that displacement was not pure strike-slip. Dextral transcurrent displacement in the fault system in the present map area is attested to by deflections and Riedel shear patterns associated with the Norma fault and with related, parallel, minor faults. The overall displacement sense of the Norma fault is probably dextrally oblique dip-slip (N-side-down).

South of Concession Lake, a second fault is required by abruptly truncated metamorphic mineral zones (trace modified from that of Bostock, 1980) (Fig. 2). The fault, informally referred to as the 'Grid fault', is parallel to the Norma fault but has S-side-down post-metamorphic dip slip. A third, N-trending fault is defined by the abrupt truncation of stratigraphy and D_2 and D_3 structures on the west flank of the Wishbone dome (Fig. 2). Displacement is not known but it does not offset the Norma fault, toward which it trends, and it is possible that this fault is a splay of the Normal fault.

The orientation and timing of this set of Proterozoic faults in the Contwoyto Lake area suggest that they are part of the NE-NW conjugate transcurrent fault system that affects that northwestern Canadian Shield (Hoffman, 1984; Tirrul, 1984). Tirrul (1984) and Hoffman et al., (1984) have shown that deformation on this system in the foreland of Wopmay Orogen approximated plane strain with NE-trending dextral and NW-trending sinistral displacement. Dip-slip is minor and restricted to areas of minimum fault rotation (ie. azimuths of 60-120°). The 065°-trending Norma fault is suitably oriented to have experienced dextral strike- slip with possible dip-slip movement.

Quartz veins

Numerous generations of quartz veins have been distinguished in the map area on the basis of their state of deformation. The quartz veins are most abundant in the metasediments of the Contwoyto and Itchen formations but also occur within metavolcanics. The earliest generation of quartz veins recognized are generally thin and parallel to bedding. One of these is isoclinally folded by F_1. Veins of the second generation are the most abundant in the map area. Their development spans the D_3 event and also, therefore, peak temperatures. These veins are of millimetre- to metre-size and are steeply dipping. They commonly form arrays of conjugate, en echelon veins and gashes (Fig. 14) that are symmetrically disposed about S_3. However, they are also locally foliation-parallel or foliation-orthogonal. In a given outcrop one member of the conjugate set is generally dominant, although all orientations of veins may be present. The locus of emplacement of the quartz veins is strongly controlled by the mechanical properties of the host unit (ie. thickness and competency). Second generation veins are: 1) best developed in

Figure 14. En echelon quartz vein gashes folded by F_3, When the enveloping surface of the veins is oriented clockwise to S_3, the folds of quartz veins are asymmetric in a 'Z' sense. Pen is 13.5 cm. (GSC 204448-B)

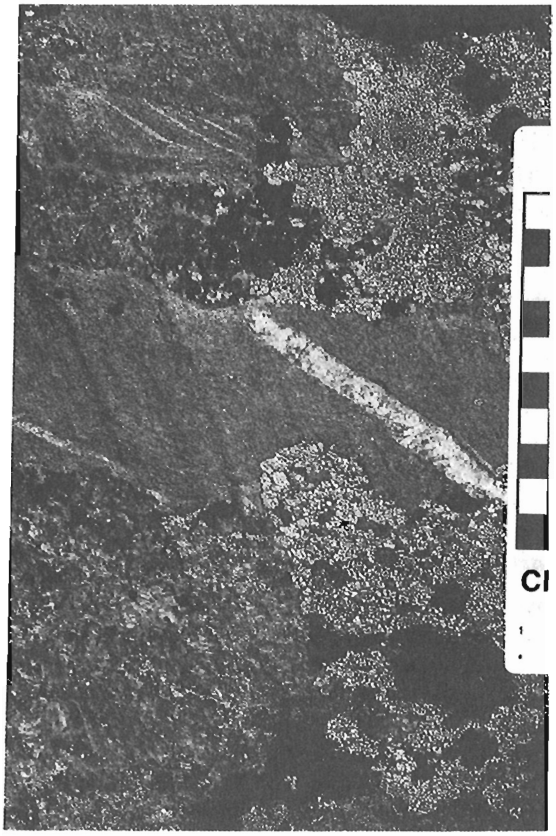

Figure 15. Quartz veins stratabound in psamite beds. The more competent psammite beds were apparently the mechanically more favourable medium for fracturing during the episode of quartz veining than were the adjacent pelitic beds. Iron-formation beds, or sequence of beds, are similarly favourable sites of emplacement for quartz veins. (GSC 204431-A)

relatively competent beds (e.g. iron-formation and greywackes); 2) are stratabound within more competent beds (Fig. 15) (or set of beds as in iron-formation horizons); 3) terminate slightly beyond the boundaries of the competent bed(s), or 4) transect all layers, refracting 5-25° in trend at contacts between units of contrasting competencies. D_3 deformation of this vein set depends on the orientation of the veins with respect to S_3. The set whose enveloping surface is clockwise (CW) to S_3 is 'Z' folded (e.g. Fig. 14), the set oriented counterclockwise (CCW) to S_3 is 'S' folded, those nearly orthogonal to S_3 are 'M' folded, and the bedding-parallel set are boudinaged along S_3. Where not folded, the conjugate veins are commonly deformed in asymmetric pinch-and- swells. Younger, less deformed veins cut older more deformed veins, and are always at a higher angle to S_3 than the older ones. These relationships can be accounted for by a model involving ongoing intrusion of quartz into conjugate fractures sets during progressive D_3 shortening. As peak metamorphism was syn- D_3, the dehydration of hydrous silicates in the sedimentary rocks is a probable candidate for the source of the quartz.

The latest members of this generation of quartz veining are similarly arranged in conjugate sets about S_3, with one of the conjugates dominating in any one area, but are little deformed to not deformed, and tend to be wider. In the localities where these veins are not folded their relationship to progressive D_3 deformation is inferred by their orientation.

All of the forementioned generations of quartz veins are recrystallized. This contrasts with the final phase of quartz veining related to the Proterozoic faulting. The Proterozoic veins are distinguished by their non-recrystallized quartz and spatial association with breccia zones, carbonate veins and hematitic alteration.

RELATIONSHIP BETWEEN DEFORMATION AND PLUTONISM

It does not appear that the emplacement of any of the plutons had a significant role in the deformation history of the area. The intrusion of the C1 and C2 (synvolcanic?) plutons presumably deformed their country rocks to some extent but there is no preserved evidence of that deformation. There is also no identifiable intrusion-related deformation associated with the C3 and C4 intrusions. The C1, C2, C3, and C4 bodies have all experienced D3 deformation; their relation to D1 and D2 is not known.

Structures in the supracrustal rocks bend around the C5 tonalite plutons but there is no obvious change in intensity, asymmetry, or character of D_3 folds with proximity to the plutons. There is also no observed change in the number of generations of structural elements with proximity to either of these plutons, or overlap of distinct strain patterns between the plutons. All of the structural elements present throughout the belt appear to simply have been rotated near the plutons. Only a small part of the deformation history is therefore apparently related to the intrusion of the C5 plutons. The C6 set of intrusions also had very little effect on the bulk deformation of their country rocks. As described above, their intrusion occurred mainly by brittle processes. The NE-trending D_{4a} folds deform the C5 and C6 plutons and occur across the mapped area. They must therefore have resulted from a late, NW-SE directed shortening related to an event of a scale greater than the present map area.

ORIGIN OF THE OLGA AND WISHBONE DOMES

The Olga and Wishbone domes are significant features of the map pattern (Fig. 2) and the deformation history. The Wishbone dome was recognized by Tremblay (1976), who suggested it was diapiric. The Olga dome is newly recognized. Several features of the structures that are key to understanding their origins are: 1) there is no change in intensity, number of fabric elements (ie. generations), character of the folds or foliations, or relative angle between sets of structures near the domes; 2) S_3 is arched over the domes; 3) upright F_4 NNE- trending, chevron folds and crenulations and open, E-W-trending (F_6?) minor folds occur on the limbs of the Olga dome and in the crest of the Wishbone dome; 4) mineral lineations that plunge obliquely to steeply down S_3 are coaxial with F_3 fold axes and have no apparent relationship to the domes; 5) the synforms flanking the Wishbone dome plunge north in the southern corners of the dome.

Based on these features we have concluded that the domal structures formed after the formation of the S_3 surface, that the deformation history in the supracrustal rocks is not specifically related to the domes and the pre-dome dip of S_3 around the Wishbone dome was shallow to moderate northward. Based on present information, the domes are interpreted to be the product of interference between F_4 and Proterozoic folds (D_4 and/or D_6) (the two domes are located along strike from the arch that flanks the southern limb of the Peacock Hills synclinorium) (Fig. 2). The uplift of the Hackett River gneiss dome, a structure 140 km to the east, that is very similar to the Olga and Wishbone domes, and also interpreted to have developed late in or after a folding event similar to our D_3 (Fyson and Frith, 1979) is interpreted by Fyson and Frith (1979), on the basis of temporal correlation with buoyantly rising plutons, to have been buoyantly uplifted. To date there is no evidence to support an origin of buoyant uplift (diapirism) of the core of the Olga and Wishbone domes.

METAMORPHISM

Metamorphic mineral zones mapped in the field include biotite, cordierite, andalusite and sillimanite (Fig. 2) Visible staurolite was too limited to define a mineral zone. However, it was always recognized together with cordierite, implying that the staurolite zone occurs at higher temperature than the first occurrence of cordierite. Garnet is present in iron-formation units in all mineral zones and also in some relatively Fe-rich pelitic and psammitic beds of the Contwoyto Formation that are within the cordierite and sillimanite mineral zones. The occurrence of garnet shows a strong dependence on composition and a distinct garnet zone was not distinguished.

The trace of the cordierite isograd, similar to that defined by Tremblay (1976) and Bostock (1980), forms an open arc northeast of the Wishbone dome, concave to the east (Fig. 2). Northwest of the Wishbone dome, biotite-zone rocks and the bounding cordierite isograd, have been downfaulted on the Grid fault (Fig. 2). An extension of the north-facing sillimanite isograd has been mapped northeast of the Grid fault to where its trace is truncated against a C6 pluton (Fig. 2). Around the Wishbone dome, the south-facing sillimanite isograd is redefined to be east of, rather than coincident with, the contact between volcanogenic and turbiditic units (Fig. 2) (cf. Tremblay, 1976). In addition, the sillimanite zone extends away from the dome in an E-W-trending "thermal high". Grade does not, therefore, increase concordantly downward through the tectono-stratigraphic sequence and the dome is not a distinct area of maximum temperature as suggested by Tremblay (1976). The undulating metamorphic zonation is suggestive of originally undulating mineral zones and/or post-equilibration deformation of mineral zones (original dip unknown). As the metamorphic minerals were growing during D_3, it is probable that the mineral zones have been deformed during D_3 and D_4, The overall pattern is one of an E-W trending syn-D_3 thermal zonation that increases in grade toward the E-W trending, syn-D_3 tonalites (C5). The metamorphic pattern is therefore interpreted to be related to the thermal regime extant during emplacement of the C5 tonalites.

TECTONO-MAGMATIC ENVIRONMENT

The most plausible tectono-magmatic environment for the accumulation and emplacement of the igneous rocks of the Contwoyto Lake area is one of a magmatic arc. This interpretation is suggested by 1) the dominance of intermediate volcanic compositions, 2) the physical character of the volcanics (abundant fragmental and tuffaceous units), 3) the apparent stratigraphic sequence (volcanics overlain by turbiditic sediments) and 4) the apparent calc-alkaline trend in time of the plutonic suites (Fig. 5). All of these features are considered characteristic of modern magmatic arc environments. In this model, the Central volcanic belt and the C1, C2 and C3(?) intrusions are viewed as remnants of an early phase of arc evolution, while the subsequent C4 to C6 intrusions are considered elements of a succeeding, prograding phase of arc evolution. Turbiditic sedimentation and the main phase of deformation occurred between the early and final phases of magmatic activity. The proposed tectono-magmatic environment is compatible with either the accretionary complex (Hoffman, 1986; Kusky, 1986) and back-arc basin (Folinsbee et al., 1968; Helmstaedt and Padgham, 1986) models for the evolution of the Yellowknife Supergroup. However, the model presented here is not compatible with the ensialic rift hypothesis (McGlynn and Henderson, 1970). Radiometric dating of the magmatic units will document the incremental and total time span for initial arc magmatism, subsequent drowning by turbiditic sediments, the main shortening and thickening (and transport?) and final magmatism. From data throughout the Slave Province this timespan is typically 40-180 Ma (Hoffman, 1986).

QUATERNARY SURFICIAL DEPOSITS

Quaternary surficial deposits west of Contwoyto Lake are extensive, covering up to 90 % of the area. Mapping of these deposits this summer by R.W.A. has outlined northwest-trending glaciofluvial corridors (St-Onge, 1984), spaced at 10-12 km intervals across the area, and the intervening till deposits (Fig. 16). Previous work on the surficial deposits of the present area includes that of Blake (1963), who defined the regional glacial features and the direction of glacial flow, and that of Bostock (1980) who outlined areas of moderate and minimal drift cover and provided more detailed information on ice flow direction.

Deposits within the glaceofluvial corridors were subdivided into three map units. 'Bouldery sands, gravels and diamicton' occur as series of irregular ridges and conical hummocks 1-5 m high. This unit is interpreted as being deposited by meltwater and debris flows within crevasses and unoriented depressions along the ice margin. 'Deltaic sediments', characterized by sands and gravels that contain lenses of pebbles and boulders, occur as flat-topped aprons of sediment 5-20 m thick with occasional collapse fronts. The deltas are generally associated with esker complexes at the margin of glacial lakes. 'Washed till' includes concentrations of sand, gravel and boulders on the bedrock surface and are located along the margin of glaciofluvial corridors. It is thought to originate from the removal of fines by meltwater flow.

Two major units of morainal deposits (ie. deposited directly by glacial ice) were mapped (Fig. 16). 'Till blanket' is a lodgment or basal meltout till, 2-10 m thick, with a gently rolling, locally fluted surface, which masks bedrock topography. 'Till veneer' is a lodgment and ablation till, less than 10 m thick, with a surface that mimics the bedrock topography.

The glaciofluvial corridors represent the former beds of major meltwater streams in which processes of both sediment deposition and removal occurred. The corridors thus contain stretches of bare bedrock alternating with ice contact features (St-Onge, 1984). Continued mapping of the surficial deposits of the area will provide further information on the glaciofluvial system and its relationship to major ice movements in the area.

The extensive glacial cover in the vicinity of Contwoyto Lake provides an ideal environment for carrying out systematic geochemical sampling of drift. Sampling of till and esker systems in similar terrain in District of Keewatin yielded geochemical anomalies related to concealed mineralization (Ridler and Shilts, 1974; Shilts, 1973). As in Keewatin, however, it will be necessary to evaluate the compositional influence of far-travelled debris on local anomalies. The drift cover of the Contwoyto Lake region is anomalously continuous for this generally sediment-poor part of the Laurentide Ice Sheet. Shilts et al. (1987, p. 134) attributed this to trains of debris shed northwestward along topographic lowlands from easily eroded terrain to the south and east.

IMPLICATIONS FOR GOLD EXPLORATION

Results to date that are of special interest to gold exploration are:

1. As the iron-formation-hosted exploration targets in the area are stratabound, it is critical to be able to trace individual layers through complicated structures. The proposed model of fold interference and cleavage relationships to the fold sets can be used to predict or to interpret changes in facing direction of bedding, strike, dip or plunge of structures (Fig. 10).

2. Syn-F_3 faults significantly affect the structural geometry in that they result in imbrication of fold limbs and consequent abrupt loss of units along strike on fold limbs (Fig. 10, 11). As in result '1', an understanding of such potential changes in strike or position of individual units is critical to exploration in the area.

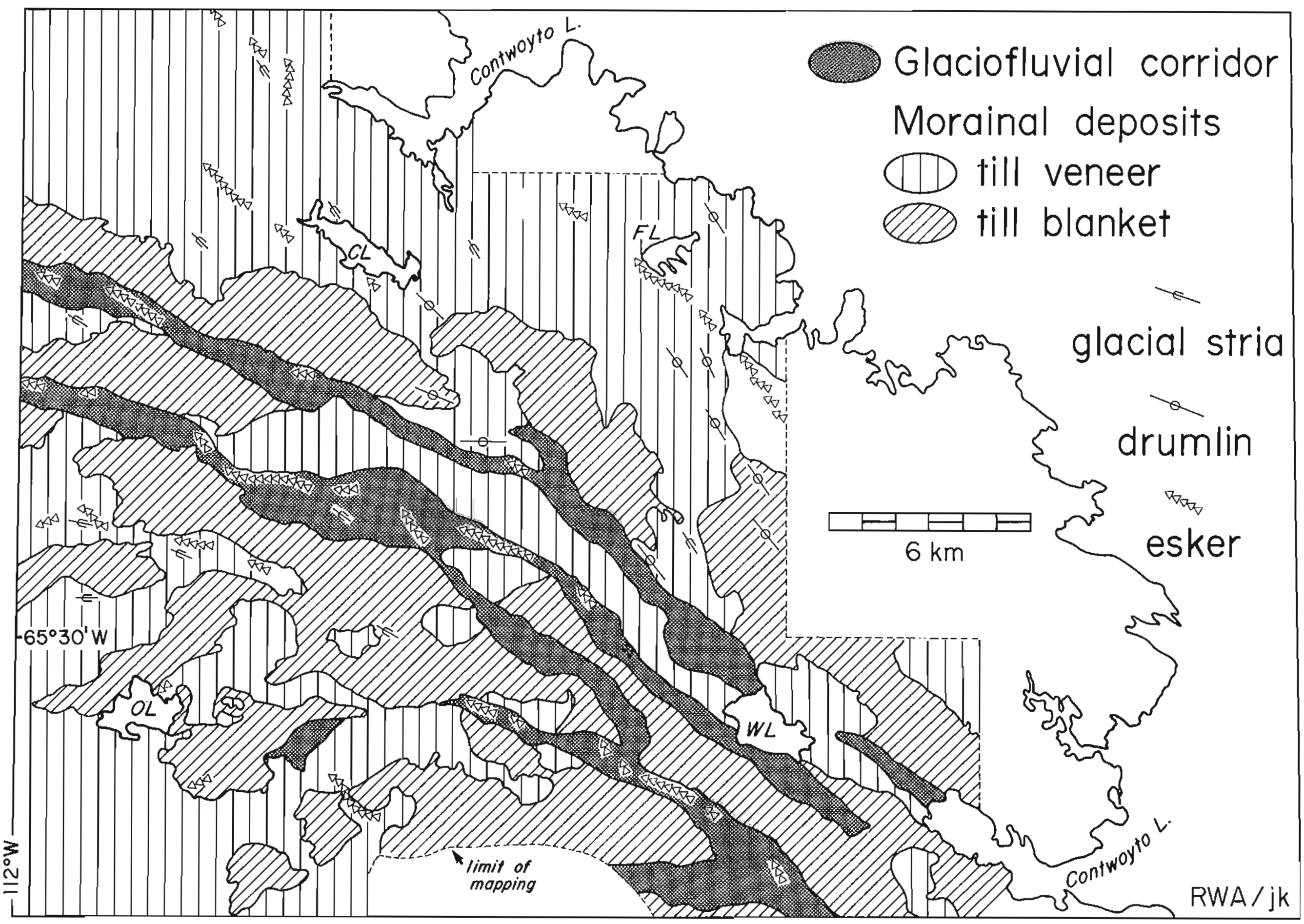

Figure 16. Simplified map of the distribution of the Quaternary glaciofluvial corridors and morainal deposits in the western Contwoyto Lake area. Bodies of water: CL - Concession Lake; FL -Fingers Lake; OL - Olga Lake; WL - Windly Lake.

3. The intensity and style of folding in the vicinity of Lupin Mine is similar to that developed throughout the metasediments of the map area. The fold structure itself is therefore not necessarily an exploration target.

4. Some of the metasediments previously designated as Itchen Formation (Bostock, 1980) have been identified to be Contwoyto Formation (particularly east of Concession Lake) (Fig. 2), thus considerably increasing the area of high exploration potential.

5. Most of the rocks in the Wishbone dome and the Olga lake area that were previously interpreted as felsic metavolcanics (units Avat in Bostock, 1980), and thus potential exploration targets were found to be plutonic units (compare Fig. 2 with Bostock, 1980).

6. Although some large areas designated by regional mapping (Bostock, 1980) as 'hydbrid rocks' are primarily paragneisses with a lesser intrusive component (e.g. at the east end of Point Lake (King, 1981), much of the area northwest of Concession Lake, similarly designated as a 'hybrid rock' comprises dominantly intrusive units that contain abundant, discontinuous screens and xenoliths of Contwoyto Formation. Some of the larger screens may be potential exploration targets.

7. Despite marked differences in interpretation concerning the importance of quartz veins in controlling gold distribution in strat form, iron-formation-hosted gold deposits (Kerswill, 1985, 1986; Kerswill and Caddey, 1987; Lhotka and Nesbitt, 1987a, b) it is clear that the veining is related temporally to the thermal peak of metamorphism (i.e. syn-D_3, C5).

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Volcano-plutonic setting of U-Cu bearing magnetite veins of FAB claims, southern Great Bear magmatic zone, Northwest Territories[1].

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Gandhi, S.S., Volcano-plutonic setting of U-Cu bearing magnetite veins of FAB claims, southern Great Bear magmatic zone, Northwest Territories; in Current Research, Part C, Geological Survey of Canada Paper 88-1C, p. 177-187, 1988.

Abstract

Pitchblende, chalcopyrite, pyrite and fluorite occur in magnetite-rich veins and breccia-fillings in an intermediate to felsic volcanic sequence at FAB claims, 12 km east of the Rae Lakes settlement. The showings are up to 100 m long and 5 m wide, and contain subeconomic concentrations of U and Cu. They are hosted by a dark grey, magnetic, volcaniclastic unit dipping steeply to the northeast, and by a discordant, light pink feldspar porphyry on its east side. The U-Cu bearing veins represent only a few of the magnetite-apatite-actinolite veins that are widely distributed in a broad north-trending zone over 7 km long.

The magnetite-rich veins are interpreted as the product of hydrothermal activity related to a younger quartzmonzonite pluton. Textures indicate deposition of pitchblende and chalcopyrite contemporaneously with magnetite. The mineralization is similar to that in a volcanic breccia pipe at the Sue-Dianne Cu-U deposit 40 km to the southeast.

Résumé

Les claims FAB, situés à 12 km à l'est de la localité de Rae Lakes, renferment de la pechblende, de la chalcopyrite, de la pyrite et de la fluorine qui se présentent sous forme de veines riches en magnétite et de remplissages bréchiques dans une séquence qui va du type intermédiaire au type volcanofelsique. Les venues atteignent 100 m de longueur et 5 m de largeur, mais leurs concentrations en U en Cu sont toutefois inférieures au seuil de rentabilité. Elles sont encaissées dans une unité volcanoclastique magnétique et gris foncé qui est fortement inclinée au nord-est et dans un porphyre à feldspath rose pâle disposé en discordance sur son côté est. Les veines recélant de l'U-Cu ne représentent que quelques-unes des veines à magnétite-apatite-actinolite qui sont abondamment réparties dans une vaste zone à direction nord sur 7 km de longueur.

Les veines riches en magnétite résulteraient d'une activité hydrothermale liée à l'existence d'un pluton plus jeune de quartz-monzonite. Les textures indiquent la mise en place de pechblende et de chalcopyrite contemporainement à la magnétite. La minéralisation est analogue à celle que l'on rencontre dans une cheminée de brèches volcaniques située sur les lieux du gisement Sue-Dianne de Cu-U à 40 km au sud-est.

[1] Contribution to Canada - Northwest Territories Mineral Development Agreement, 1987-1991.

INTRODUCTION

The study area is located in the south central part of the Great Bear zone, which is a 1900-1850 Ma old continental magmatic arc (Fraser et al., 1972; Hoffman, 1980; Fig. 1). U-Cu occurrences were discovered in the area during 1969 (Curry, 1969b). The geological setting of these occurrences, however, was inadequately known from the earlier regional mapping (1 :253 440 scale) and the limited area covered by the exploration work, and the present study was undertaken in 1987 to define it better and to conduct comprehensive metallogenic investigations. The occurrences were briefly examined by the writer in 1982.

PREVIOUS WORK

Mineral exploration interest in the Faber Lake region dates back to 1934 when officers of the Geological Survey of Canada first discovered radioactive minerals in the southern part of the region (Fig. 1). This eventually led to the discovery of the Rayrock uranium deposit which is in a giant quartz vein system, and produced 150 t of uranium metal during 1957-59 (Lang et al., 1962; Gandhi, 1978).

Early reconnaissance mapping by Kidd (1936) was followed by mapping at a scale of 1 :253,440 which recognized the presence of dacitic feldspar porphyry intruded by granite in the study area (Fraser, 1967). Adjacent map sheets on the same scale outlined a belt of the porphyry trending south-southeast (Lord, 1942; Wilson and Lord, 1942; Fig. 1). More detailed mapping of the belt south of latitude 63°30′N, and between latitudes 63°40′ and 64°00′N was carried out by McGlynn (1968, 1979). He recognized abundant dacitic to rhyodacitic ash-flow tuffs in the central part of the belt near Faber Lake. In view of this and the recognition of andesite-rhyodacite sequence in the study area during the present work, the belt is referred to here as the Faber Lake volcanic belt (Fig. 1) The belt has a distinctly high magnetic response (Geological Survey of Canada, 1963, 1969a, b).

An airborne scintillometer survey over the study area in 1968 by Precambrian Shield Resources Limited and Numac Oil and Gas Limited resulted in the discovery of uranium occurrences and staking of FAB claim group, of which claims 1, 5 and 9 cover the main showings area (Curry, 1969a, b,). The FAB property was optioned by the Rayrock-Ryowa Joint Venture in 1977, and 3 radioactive zones were trenched (Morris, 1977). An aeromagnetic anomaly 3 km to the northwest, staked as the RON claims, was found to be caused by a distinctly magnetic mafic volcanic unit (Morris, 1977). Eventually the claims were allowed to lapse. In 1979 the FAB 1 and 5 claims were restaked by Noranda Exploration Company Limited, but only limited work was done (Bryan, 1981). All the claims in the area lapsed during August 1987.

The Sue-Dianne Cu-U deposit is the most promising prospect in the belt (Fig. 1). It was discovered as a result of regional airborne gamma ray spectrometer survey conducted in 1973 by the Geological Survey of Canada (Richardson et al., 1974). The strong radiometric anomaly over the deposit led to staking of SUE and DIANNE claims which were optioned by Noranda Exploration Company Limited in 1974 (Charbonneau, 1988). Exploration by the company revealed that a magnetite-rich breccia pipe approximately 400 m in diameter and carrying notable values in Cu and some U, Ag and Au, to a depth of over 200 m, occurs in dacitic ash flow tuff (Climie, 1975, 1976).

GENERAL GEOLOGICAL SETTING

The host volcanic sequence of the Fab U-Cu occurrences comprises andesite to rhyodacite flows, pyroclastics and related epiclastic sediments (units 1 to 6, Fig. 2; Table 1). It is more than 2 km thick, and dips steeply to the northeast. Its stratigraphy is not yet well defined. Metamorphism is in the sub-greenschist facies. On its east side there is a discordant

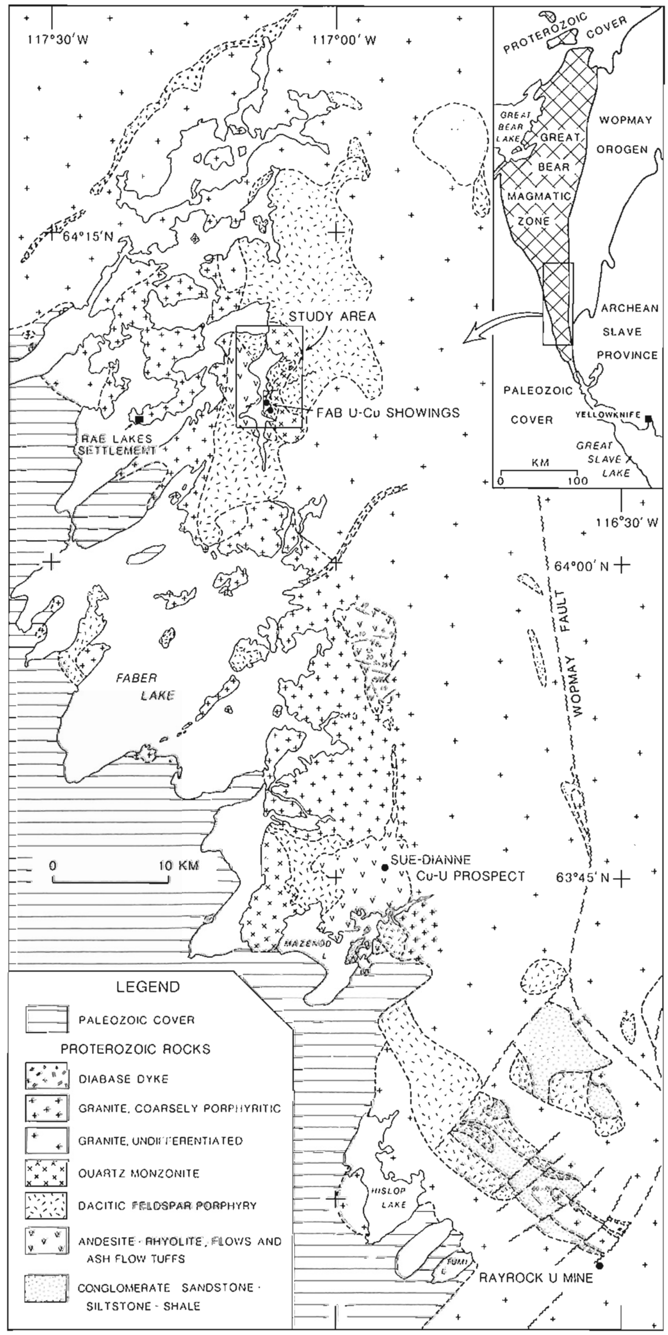

Figure 1. General geology of the Faber Lake volcanic belt and location of the study area.

dacitic feldspar porphyry which is likely a contemporaneous sub-volcanic intrusion. Magnetite-apatite-actinolite veins occur near the boundary of the volcanic sequence and the porphyry. The porphyry is intruded by a quartz monzonite pluton. Magnetite-apatite-actinolite veins cut the pluton on an island in Rae Lake (Fig. 2). A set of coarse quartz feldspar porphyry dykes that cut the veins occur in the north-central part of the study area and are chemically similar to the quartz monzonite (Table 1). A granite stock in the centre of the area is younger than the quartz monzonite. It is cut by some of the giant quartz veins that are related to northeast striking right-lateral faults common throughout the Great Bear magmatic zone. A few diabase dykes seen in the study area are fresh-looking, and are probably the youngest rocks here.

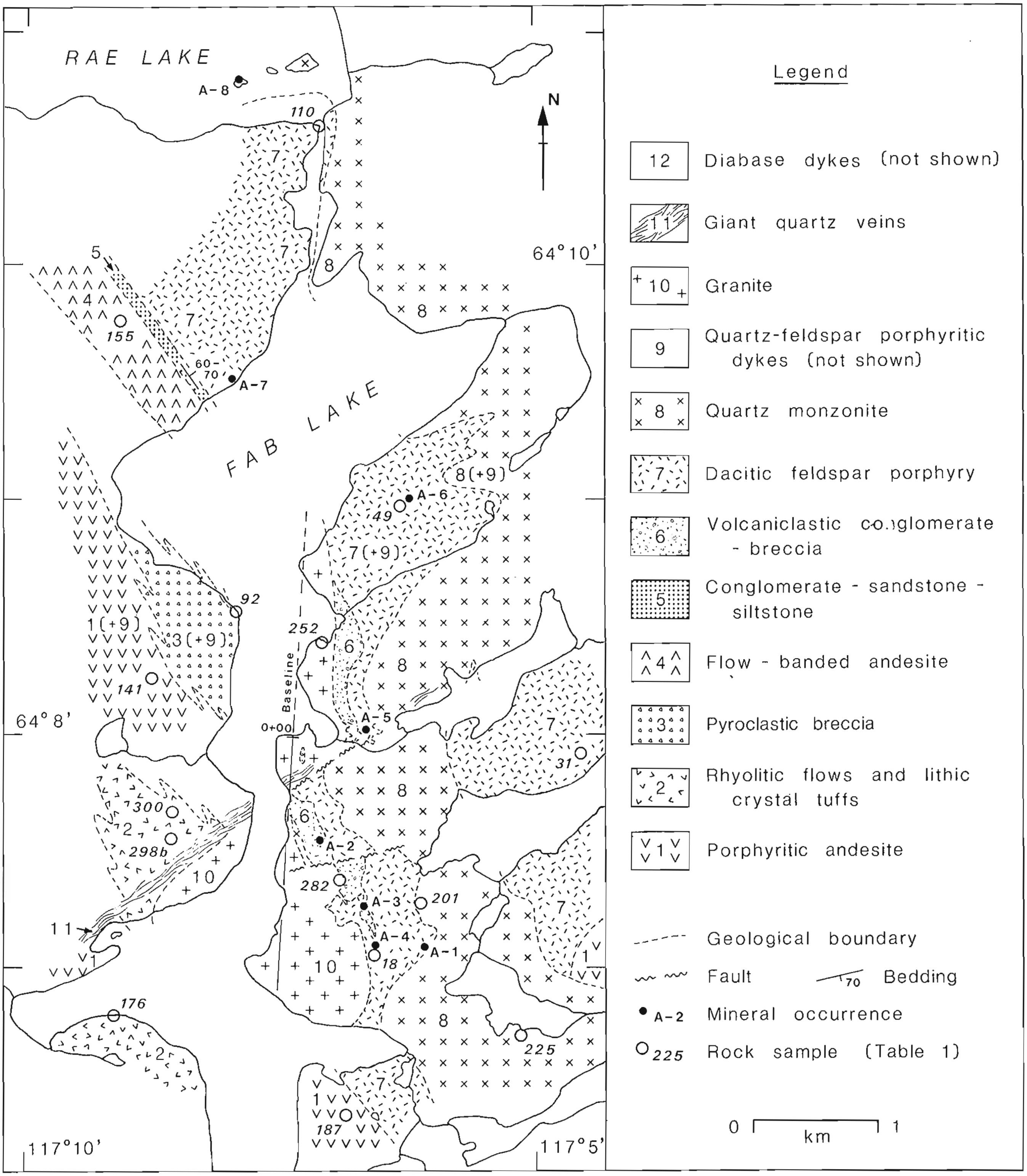

Figure 2. Geology and mineral occurrences of the Fab Lake area, Northwest Territories.

LITHOLOGICAL UNITS

Porphyritic andesite (unit 1)

Outcrops of this unit occur mainly west of Fab Lake. An occurrence east of the lake is probably much higher stratigraphically (Fig. 2). Rocks of unit 1 are characterized by euhedral to subhedral phenocrysts of andesine-labradorite and pyroxene, which are 2 to 4 mm long and are set in aphanitic matrix. The plagioclase and pyroxene phenocrysts make up close to 40 and 15 per cent of the rock respectively and are randomly oriented. Magnetite occurs as scattered octahedral crystals, making up to 3 per cent of the rock which is only weakly magnetic. Some inclusions of magnetite in pyroxene, and of pyroxene in plagioclase were observed. A few plagioclase phenocrysts show zoning at their margins. Matrix contains abundant microlites of feldspar and interstitial mafic silicates and oxide. A few grains of apatite are present. Alteration is pervasive; feldspars are cloudy in thin section and pyroxene is partly or wholly chloritized. The unit is more variable in the south, where it is affected by shear zones dipping moderately to the northeast.

Rhyolitic flows and crystal tuffs (unit 2)

These rocks are found in the southern part of the study area. They contain abundant phenocrysts of plagioclase, quartz and potash feldspar, up to 4 mm in length, set in a fine grained siliceous matrix. Some of the phenocrysts are angular, and appear to be fragments of larger crystals. Plagioclase phenocrysts predominate, and are sodic andesine to albite in composition. Mafic grains and aggregates, including some magnetite, are sparsely distributed in the rock.

In several places lapilli of similar composition but slightly darker colour occurs in the unit. Thus at least part of the unit is lapilli crystal tuff. Chemical analyses of two samples from this unit show its felsic character, and thus a notable compositional range in the volcanic sequence of the area.

Table 1. Chemical analyses of selected volcanic and intrusive rocks from the Fab Lake area, southern Great Bear magmatic zone, Northwest Territories

Analysis Number		1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Sample Number		141	187	176	300	298b	92a	92b	155	18	31	110	201	225	49	282	252
Lithological Unit		1	1	2	2	2	3	3	4	7	7	7	8	8	9	10	10
SiO_2	per cent	59.90	61.30	69.30	70.90	72.20	65.70	38.70	57.50	66.00	66.60	66.00	62.00	62.00	66.80	71.90	77.50
Al_2O_3	"	15.20	15.00	15.30	14.50	14.10	13.10	17.80	16.30	15.60	16.10	15.80	17.30	17.80	15.90	13.80	11.90
TiO_2	"	0.74	0.66	0.29	0.26	0.26	0.61	0.95	0.79	0.49	0.52	0.53	0.60	0.58	0.34	0.24	0.11
FeO	"	3.70	3.30	1.50	1.90	2.00	2.40	8.20	5.40	1.80	2.10	2.00	2.90	2.60	1.70	1.00	0.90
Fe_2O_3	"	3.00	2.80	1.00	0.50	0.00	3.20	5.80	3.60	1.30	1.10	1.40	1.70	1.30	1.10	0.80	0.30
MnO	"	0.18	0.11	0.06	0.05	0.05	0.14	0.33	0.12	0.02	0.06	0.05	0.06	0.07	0.03	0.02	0.02
MgO	"	3.21	3.43	0.38	0.55	0.89	2.28	9.69	3.22	0.81	0.88	0.91	1.58	1.43	1.02	0.64	0.24
CaO	"	4.98	5.15	1.47	1.76	0.52	8.29	9.65	5.82	2.00	1.80	2.71	3.19	3.27	2.67	0.66	0.42
Na_2O	"	2.70	2.40	3.20	3.50	3.70	2.00	0.20	2.80	2.70	3.40	3.10	3.40	3.10	2.80	2.90	2.90
K_2O	"	4.34	4.15	5.48	4.94	4.80	1.34	3.47	3.41	7.08	5.74	5.78	5.01	6.23	5.52	6.23	5.54
P_2O_5	"	0.08	0.08	0.04	0.03	0.04	0.07	0.09	0.08	0.05	0.05	0.05	0.10	0.08	0.04	0.02	0.00
H_2O	"	1.60	1.30	0.80	0.60	1.10	1.30	4.50	1.20	1.10	1.20	1.30	1.70	1.40	1.30	0.90	0.50
Co_2	"	0.20	0.20	0.20	0.10	0.10	0.10	0.60	0.40	0.10	0.20	0.30	0.20	0.30	0.10	0.10	0.10
S	"	0.00	0.01	0.00	0.04	0.00	0.00	0.00	0.00	0.02	0.03	0.02	0.04	0.00	0.00	0.01	0.00
Total	"	99.83	99.89	99.02	99.63	99.76	100.53	99.98	100.64	99.07	99.78	99.95	99.78	100.16	99.32	99.22	100.43
Ba	ppm	659	598	768	592	785	372	561	769	983	822	814	1243	1451	1183	436	137
Be	"	2.80	2.69	3.14	3.25	2.76	1.83	2.06	2.14	3.29	3.91	3.14	2.50	2.29	2.67	4.77	3.16
Co	"	24	22	5	5	5	16	47	26	9	8	8	12	10	9	4	2
Cr	"	100	91	15	18	19	110	130	34	20	20	23	20	23	25	13	8
Cu	"	18	31	10	8	8	25	20	17	52	37	28	26	13	20	4	10
La	"	47	49	60	62	63	33	65	35	73	60	55	56	43	51	53	53
Nb	"	0	0	0	0	0	0	0	0	2	0	9	0	0	0	1	10
Ni	"	40	42	8	12	11	37	83	18	12	12	12	14	14	15	8	5
Rb	"	210	204	248	247	218	46	181	156	310	296	250	205	199	214	302	270
Sr	"	264	266	239	242	157	333	381	212	176	178	201	345	363	371	121	53
V	"	120	110	21	18	17	120	190	160	32	33	37	47	44	26	11	4
Y	"	22	27	19	8	17	17	41	28	30	23	29	15	0	3	34	33
Yb	"	2.50	2.40	2.64	2.37	2.43	1.96	3.25	2.35	2.96	2.87	2.93	2.16	1.91	2.11	2.87	3.21
Zn	"	160	82	40	26	36	73	340	34	6	27	28	29	47	19	9	1
Zr	"	206	188	217	167	182	140	271	142	277	268	266	335	374	227	179	118

Notes:

1) Analyses by the Analytical Chemistry Section, Mineral Resources Division, Geological Survey of Canada, Ottawa.
2) All analyses by x-ray fluorescence method except FeO, H_2OT, CO_2T, C and S by rapid chemical methods.
3) Fe2O3 is calculated using the formula:
   Fe_2O_3 = Fe2O3(XRF) — 1.11134 * FeO(Volumetric)
4) Sample Locations in Figure 2 (GFA-'87 series of samples).

Pyroclastic breccia (unit 3)

This unit is well exposed on a ridge west of Fab Lake in the central part of the study area, for a distance of 750 m along northeast direction. It is characterized by very mafic fragments in highly felsic matrix, both of which contain abundant feldspar phenocrysts in fine grained groundmass (Fig. 3, 4). Fragments of other lithologies, mainly andesite and rhyodacite, are present but are quantitatively insignificant. The boundaries of mafic fragments and the felsic matrix are well defined, but margin of the fragments tend to be more felsic than the core, indicating some reaction with the felsic matrix. The compositional differences are apparent on the weathered surface, and contrasting chemistry is seen from the separately analyzed mafic fragment and felsic matrix in one sample (Table 1 ; Fig. 4). Texturally however the difference is mainly in the relative concentration of mafic and felsic constituents in their groundmass, and to a lesser extent in the relatively higher abundance of the groundmass in the fragments and somewhat more calcic character of plagioclase in them. Mafic silicate phase or phases are completely altered to chlorite amphibole and epidote in the thin sections examined. Plagioclase phenocrysts in the mafic fragments are andesine labradorite in composition. Magnetite grains are finely disseminated in their matrix. A few grains of quartz are also present.

The breccia has a fairly sharply defined boundary with andesite of unit 1 in the west. Andesite also occurs on its east side on the shore of Fab Lake. Its field relations with unit 2 however are not known. Its general character suggests that it is an explosion breccia derived from a partially differentiated magma chamber.

Flow-banded andesite (unit 4)

This unit forms a high northeast-trending ridge with a steep escarpment on the southwest side and a gentler slope on the northeastern side. Its distinctly magnetic character is reflected in an aeromagnetic anomaly which attracted exploration interest but no mineral occurrence has been found in it (Morris, 1977). The rock is dark grey, vitrophyric, and contains sparsely distributed phenocrysts of plagioclase and pyroxene that make up approximately 10 and 5 per cent of the rock respectively. The phenocrysts are commonly euhedral, randomly oriented, and 2 to 4 mm long. In addition there are smaller crystals of magnetite, and a few grains of quartz. The magnetite crystals make up about 2 per cent of the rock, but the mineral also occurs as abundant disseminated very fine grains in the matrix, which accounts for its strongly magnetic nature. Accessory minerals include apatite and zircon. Inclusion of apatite in a pyroxene crystal and of zircon in a magnetite crystal were observed in thin section.

The matrix contains abundant feldspar microlites, which are oriented subparallel to each other, and indicate some flowage around phenocrysts. Mafic silicate and magnetite occur interstitially with the feldspar microlites. Flow banding is due to relative abundance of mafic minerals in the matrix. Mafic bands and lenses are commonly 1 cm or so thick. Chlorite is a common alteration mineral and has anomalous birefringence colours.

The unit has a flinty character which is most pronounced at the eastern margin where phenocrysts are scarce or absent. Autoclastic features are seen at some places along that margin. The contact and flow banding dip 50 to 70 degrees to the northeast. The flow banding however in some places shows open folding on a scale of a few metres. Foliation is moderately well developed in some narrow zones, up to a few metres wide, and dips steeply to the northeast. Foliated rock in thin section shows mafic streaks in the matrix.

Conglomerate-sandstone-siltstone (unit 5)

A 50 m wide zone of sediments is exposed along the eastern margin of unit 4 for a strike length of several hundred metres. It comprises interbedded polymictic conglomerate, sandstone and siltstone ranging in thickness from a fraction of a metre to several metres. Bedding is well developed in sandy and silty sediments, and dips 50 to 70 degrees to the northeast. Crossbedding, seen in some beds, indicates tops facing to the northeast. This is also supported by graded bedding and

Figure 3. Pyroclastic breccia (unit 3), Fab Lake area, Northwest Territories. Note the larger mafic fragments have highly chloritized core which is lighter in colour on weathered surface than the outer zone. Location half-way between sample sites 92 and 141 (Fig. 2). GSC 204398-S

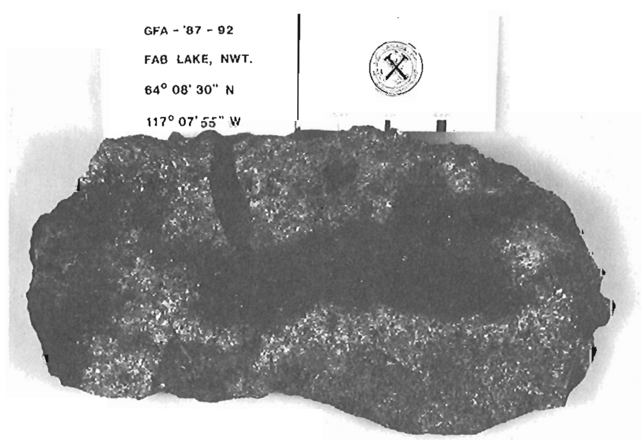

Figure 4. Mafic fragments in felsic matrix of a sample of pyroclastic breccia (unit 3), prior to cutting of the larger fragment and the matrix material on its sides for separate analyses reported in Table 1. Sample site 92 (Fig. 2). GSC 204132-H

channel scour features observed at a few places. The western boundary with unit 4 is sharp with local irregularities, indicating deposition on a flow top. The eastern boundary with dacitic feldspar porphyry is sharp, irregular and in general discordant to the bedding in the sediments.

The sediments contain a high proportion of volcanic debris. Polymictic conglomerate contains many pebbles and cobbles of porphyritic andesite, dacite and rhyodacite (Fig. 5). Sandstones contain abundant plagioclase grains, and a few magnetite-rich layers up to a centimetre thick. Siltstones range from thick massive beds to thinly laminated beds. Sandstones and siltstones are generally drab grey. The sediments are mainly volcanogenic, and an integral part of the volcanic sequence.

Volcaniclastic breccia-conglomerate (unit 6)

This is the main host of U-Cu occurrences of FAB claims (Fig. 9), and was mapped by Curry (1969b) as volcanic fragmental unit containing interbedded tuff, fragmented porphyry and volcanic breccia. He noted rounded clasts of feldspar porphyry in tuffaceous matrix lacking bedding (Fig. 6), and their differential weathering in relation to the matrix. He also noted dark green to black aphanitic clasts as occurring mainly in the southwestern part of the unit. The porphyritic clasts predominate and are of pebble to cobble in size range, but a few are larger. They contain plagioclase phenocrysts in an aphanitic dark grey to chocolate coloured matrix. The phenocrysts make up approximately 25 per cent of the rock, are 2 to 4 mm long and are randomly oriented. Thin sections show that the clasts are similar to andesite of unit 1, except for a lower proportion of phenocrysts. The clasts and the matrix are both weakly to strongly magnetic.

The unit has sharp, irregular boundaries with the dacitic feldspar porphyry on the northeast and south sides, and in the west it is intruded by quartz monzonite and granite as shown by Curry (1969b) and represented in Figure 9. Remnants of the unit isolated by various younger intrusions, were found during the present study north of Curry's map area near the east margin of the granite stock (Fig. 2). The relation of this unit to the other units in the volcanic sequence is thus not clear. Its conglomeratic character and position approximately along the strike of the sedimentary unit however, suggest that it may be stratigraphically close to the latter (Fig. 2).

Figure 5. Polymictic conglomerate containing volcanic clasts (unit 5), near northwest shore of Fab Lake. GSC 204398-U

Figure 6. The host volcaniclastic conglomerate (unit 6) of zone A-2, exposed on blasted surface near west end of the northernmost trench (Fig. 9; Morris, 1977). GSC 204398-O

Dacitic feldspar porphyry (unit 7)

This porphyry unit is areally extensive and uniform in composition. The regional map by Fraser (1967) shows this and younger granite as the only rocks in the study area. Observations by Curry (1969b) and the writer show discordant contacts with the volcanic sequence. The porphyry remains uniform right up to its sharp contact with the sequence. Offshoots of it into the sequence are rare, and so are inclusions of any kind in it. The unit is probably a shallow seated intrusion, possibly in part extrusive, in a volcanic complex. It is cut by magnetite-apatite-actinolite veins (Fig. 7).

Phenocrysts of oligoclase or andesine, up to 5 mm long, make up about 25 per cent of the rock. They are randomly oriented, although at a few localities a crude alignment, plunging moderately to southeast can be discerned. A few phenocrysts of potash feldspar and quartz are also commonly present. Scattered aggregates of mafic minerals, up to the

Figure 7. Magnetite-apatite-actinolite vein and subsidiary veinlets in dacitic feldspar porphyry (unit 7) at the mineral occurrence A-6 (Fig. 2); the vertical vein strikes 340 degrees; looking north-northwest towards outlet of the Fab Lake. GSC 204398-P

size of plagioclase phenocrysts, contain mostly chlorite, some pyroxene, iron oxide and sphene. Felty aggregates of chloritized biotite, and zircon and apatite crystals are present. The matrix is fine grained equigranular containing approximately equal amounts of quartz and saussuritized feldspar.

Quartz monzonite (unit 8)

A large pluton of quartz monzonite composition intrudes the dacitic feldspar porphyry east of Fab Lake. Quartz monzonite also occurs along the west margin of unit 6 in a narrow zone truncated by younger granite in the west. In this zone it does not differ much texturally from the equigranular granite, and Curry (1969b) did not distinguish the two. The larger pluton in the east however is distinctly porphyritic except near its contact with the dacitic feldspar porphyry, and also has coarse porphyritic dyke phases in the dacitic feldspar porphyry. Curry (1969b) mapped these as porphyritic granite, and regarded them as gradational to the equigranular granite. The distinction between quartz monzonite and younger granite made here is based on the field relations and compositional differences (Table 1), and is important from the standpoint of metallogeny.

Quartz monzonite is characterized by well formed plagioclase crystals and interstitial medium to coarse grains of feldspar, hornblende, biotite, quartz, magnetite, sphene, and traces of apatite and zircon, and alteration minerals chlorite, sericite, actinolite and leucoxene. Quartz content varies from 5 to 15 per cent. Some of it occurs as myrmekite. Potash feldspar occurs as orthoclase, microcline, and perthite, and in granophyric intergrowth with quartz. The coarsely porphyritic phase has phenocrysts of potash feldspar up to 2.5 cm long, which make up approximately 5 per cent of the rock, and more abundant smaller phenocrysts of plagioclase.

Quartz-feldspar porphyritic dykes (unit 9)

Numerous dykes containing coarse phenocrysts of quartz, plagioclase and potash feldspar in a fine grained matrix occur in a zone trending northeast just north of the centre of the study area. The phenocrysts in the dykes make up about 30 per cent of the rock. Plagioclase phenocrysts predominate over quartz and potash feldspar phenocrysts. Matrix is fine grained granitic. Mafic minerals include biotite, less abundant chloritized hornblende, and magnetite as octahedral crystals in clusters of mafic minerals. Apatite and zircon with radioactive halos are common in the clusters. The dykes range in width from a few metres to a few tens of metres, and have chilled margins up to 10 cm wide. They strike in various directions, although a northeasterly trend is most common. Their dips are also variable. They are too small to show in Figure 2. They cut across magnetite-apatite-actinolite veins in dacitic feldspar porphyry (Fig. 8). Since similar veins cut the quartz monzonite on an island in Rae Lake (mineral occurrence A-8, Fig.2), the dykes of unit 9 are regarded as younger than the quartz monzonite pluton. The dykes were not seen in contact with the granite, hence their relationship to it is not certain. Mineralogy and chemical analysis of one of the dykes (Table 1) show close affinity to porphyritic quartz monzonite pluton. Furthermore, offshoots of the granite stock in the study area are aplitic and do not resemble the rocks of this unit.

Granite (unit 10)

A stock of granite is exposed on the east shore of Fab Lake and on the hills nearby in the central part of the study area. The rock is massive, equigranular to porphyritic, and pink on weathered surface. It contains 25 to 30 per cent quartz, phenocrystic plagioclase and potash feldspar with hornblende as the main mafic mineral. Chloritic veinlets are numerous at some localities, and hematite occurs along some fractures.

The southwestern contact of the stock is exposed on a hill, and the rocks of unit 2 at the contact have developed a distinct red colouration with specularite along some of the fractures. Radioactivity on both sides of the contact is nearly twice the background. The northeastern contact with units 6, 7 and 8 is very irregular, and offshoots of granite and aplitic dykes extend into them. A xenolith of unit 6, more than 50 m long, occurs in the granite approximately 700 m north of mineral occurrence A-2 (Fig. 2).

Giant quartz veins (unit 11)

A northeast-trending giant quartz vein system cutting unit 2, is well exposed on the hill a few tens of metres from the southwest contact of the granite stock. The vein zone is approximately 60 m wide, and is exposed over 1 km. It also extends across Fab Lake to the east shore granite outcrops (Fig. 9). A northeast trending fault is exposed here. It is several metres south of the veins, across a small gully where the steep northwest-facing wall has highly chloritized unit 7, with nearly vertical chlorite-lined fractures. The close spatial relationship and parallel trend of the giant quartz vein zone and the fault indicate that the vein zone is localized along tensional fractures related to the fault.

Another smaller quartz vein system trending northeast is observed 20 m north of zone A-1 (Fig. 9; Curry, 1969b). All the quartz veins contain milky white quartz, generally barren of other minerals. Stringers of chlorite and hematite

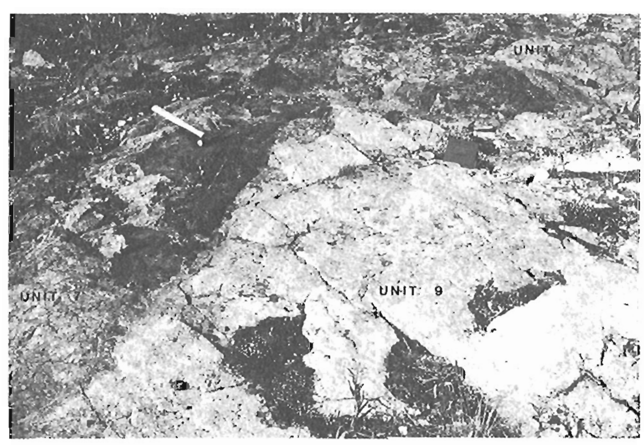

Figure 8. Quartz-feldspar porphyritic dyke (unit 9) cutting magnetite-apatite-actinolite veins and pods in dacitic feldspar porphyry (unit 7), approximately 100 metres west of mineral occurrence A-6 (Fig. 2). GSC 204398-Q

along fractures, however, are common at the margins of the veins. Some younger, crystalline comb quartz veins are associated with the milky quartz veins.

Diabase dykes (unit 12)

Diabase dykes encountered in the study area are only a few metres thick, and hence too small to show in Figure 2. They are fine to medium grained with chilled margins. Their attitude is varied. A few of them have minor deviations in strike. They likely belong to more than one set. A few of them cut across aplitic dyke, but none were observed in contact with giant quartz veins in the study area. Elsewhere in the Great Bear magmatic zone however, many diabase dykes younger than the giant quartz veins are known.

The magnetic survey reported by Curry (1969a) shows a well defined linear anomaly striking north northeast along the narrow central part of Fab Lake.It is probably caused by an unexposed diabase dyke.

PETROCHEMISTRY

Chemical analyses of 16 samples from the volcanic and plutonic rocks of the study area are presented in Table 1. The

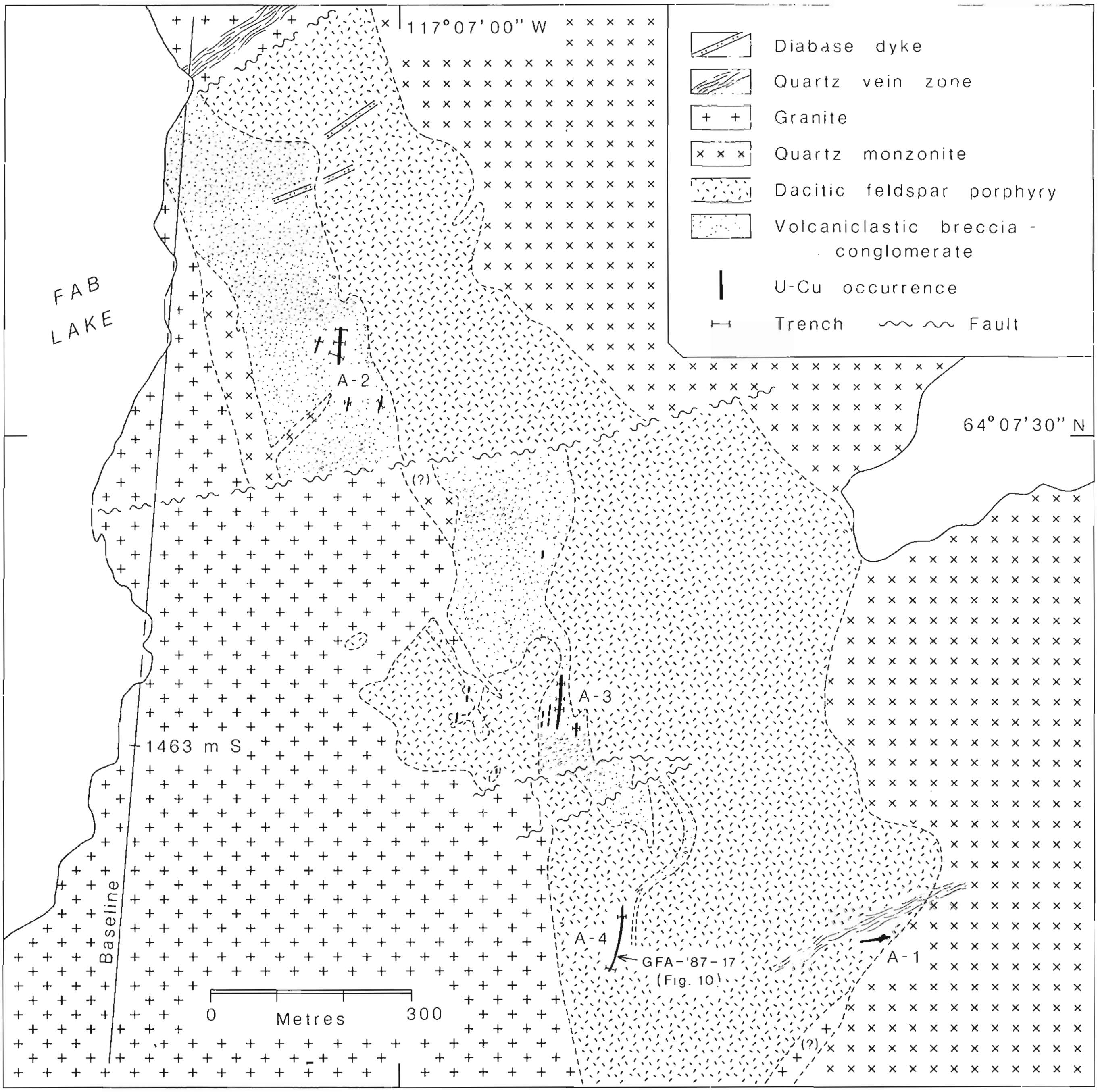

Figure 9. Geology and the main mineralized zones (A-1 to A-4) of the FAB claim group (after Curry, 1969b and Morris, 1977).

volcanic rocks of units 1, 2 and 4 show a range of andesite to rhyolite (analyses nos. 1 to 5, and 8). This compositional range is consistent with their tectonic setting in the continental magmatic arc of the Great Bear zone (Hoffman, 1980; Hildebrand, 1986). The dacitic feldspar porphyry is uniform in character as reflected in analysis of three samples from widely separated localities. It is believed to be a shallow seated intrusion related to the volcanic rocks and part of it may be extrusive.It fills the compositional gap between the andesite (units 1 and 4) and the rhyolite (unit 2). Furthermore, 5 km south-southeast of Fab Lake, McGlynn (1979) has mapped dacite to rhyodacite ash flow tuffs with interbedded sediments.

The chemical data thus reflect the calk-alkaline nature of the Faber Lake volcanic belt as a whole. A somewhat potassic character of the volcanic belt is apparent from the dominance of potash over soda. Iron content in the rock representing unit 4 is relatively high, as can be expected from the abundance of magnetite noted in it earlier. Trace elements show variations corresponding to the major elements; Cr, Cu, Zn, Co, Ni and V are enriched in relatively more mafic volcanic and plutonic rocks and particularly in the mafic fragments of unit 3; Ba is notably enriched in intermediate intrusive rocks (units 7, 8, and 9). In case of the pyroclastic unit 3, the contrast in the chemistry of the mafic clast and felsic matrix is significant (Table 1, analyses 5 and 6; Fig. 4). If these clasts and the matrix are comagmatic as they appear to be, they represent strong differentiation of a magma originally intermediate in composition. The precise mechanism of differentiation is not clear at this stage and will require further study, but it is interesting to note that the mafic fraction so generated may have been tapped by the hydrothermal fluids that formed the magnetite-apatite-actinolite veins.

The distinction between the quartz monzonite and granite made in the field by the writer is borne out by the chemistry and petrography (Table 1, analyses 14, 15 and 16). The chemical character of the quartz monzonite reflects its genetic link with the intermediate volcanic sequence, as is the case with the quartz monzonites near the Great Bear Lake (Hoffman, 1980; Hildebrand, 1986). The rock is somewhat less siliceous than the dacitic feldspar porphyry and the coarse quartz-feldspar porphyritic dykes but otherwise the three are chemically similar (Table 1). Relative ages of the three units are however well established from the fieldwork. It appears therefore that magma of this composition intruded in pulses, probably over a short time period following the volcanic activity. The granite, on the other hand, represents chemically different magma that could not have evolved from the quartz monzonitic magma by differentiation in situ, and may have intruded much later than the quartz monzonite. One of the two granite samples analyzed is representative of the stock (Table 1, no. 16), but the other one, is located at the contact of unit 6 is less siliceous and has textural similarities with the quartz monzonite that made distinction in the field difficult. This may be the reason for regarding the two units as gradational to each other in the earlier work.

URANIUM AND COPPER OCCURRENCES

Several occurrences of anomalously high radioactivity were reported in the grid area of FAB 1 to 9 claims mapped on scale 1 inch to 200 feet by Curry (1969b). Three of the larger ones, referred to as zones A-2, A-3 and A-4 (Fig. 9) are over 100 m long and up to 5 m wide, and were trenched and sampled as reported by Morris (1977). Assays across widths of 0.75 to 1.50 m ranged from 0.010 to 0.065 % U and 0.01 to 0.13 % Cu. Some zinc values and traces of gold were also reported. The mineralization is discontinuous along strike, and high grade pockets occur sporadically in the mineralized zones.

The mineralized zones contain 10 to 50 per cent magnetite as veins, aggregates and breccia-fillings. A ground magnetic survey carried out in 1969 indicates a close correspondence of strong magnetic anomalies and highly anomalous radioactivity, but there are some high magnetic anomalies where radioactivity is either only weakly anomalous or at background level (Curry, 1969a, b). On the other hand, zone A-1 (Fig. 9) has several highly radioactive but non-magnetic spots at the margins of what appears to be an altered mafic dyke. During the present study, numerous magnetite-apatite-actinolite veins were found in dacitic feldspar porphyry on a ridge at the north end of the magnetic survey grid, near the mineral occurrence A-6 (Fig. 2). One of the veins or this occurrence is 150 m long in a north-northwest direction and is up to 1.5 m wide (Fig. 7). Radioactivity along it however is generally at background level, with local highs up to 4 times the background. This is also the case with a similar but nearly horizontal vein, located near the northwest shore of Fab Lake at the mineral occurrence A-7 (Fig. 2). A large number of veinlets and stringers carrying various proportions of magnetite, apatite and actinolite are associated with these large veins and elsewhere in a broad northerly trending zone over 7 km long. They form anastomosing networks and some of them occur in small subparallel groups. The three minerals tend to form elongate clusters enriched in one of them and subparallel to the walls. Chalcopyrite and pyrite are sparsely distributed in all the veins, and form aggregates at a few places which tend to be relatively more anomalous in radioactivity (up to 10 times background). The majority of the veins are steep and strike within a few tens of degrees from the north. It is apparent that the main radioactive zones in the FAB claims are part of this major system, but are somewhat special in that they contain abundant magnetite and relatively less apatite and actinolite than the veins elsewhere.

In highly radioactive zones, pitchblende occurs as thin veinlets, stringers, and finely disseminated grains that are intimately associated with magnetite and some actinolite. The disseminated character as shown in Figure 10, indicates that it was most probably precipitated at the time of formation of magnetite. It also tends to be concentrated along the sides of magnetite veinlets, and in the stringers and aggregates of pyrite, chalcopyrite and purple fluorite. Some quartz, chlorite and calcite are associated with the pitchblende-fluorite stringers. Anatase and sphene occur with quartz locally. The paragenetic sequence appears to be magnetite, actinolite, pitchblende, pyrite, chalcopyrite, fluorite, quartz, chlorite and calcite. Hematite is developed in some places but is not conspicuous. The felsic minerals in the host rock have a reddish coloration, probably due to hematization. The mineralized rocks are generally dark grey in the hand specimen however. Secondary yellow uranium minerals are rare, and occur in highly radioactive spots.

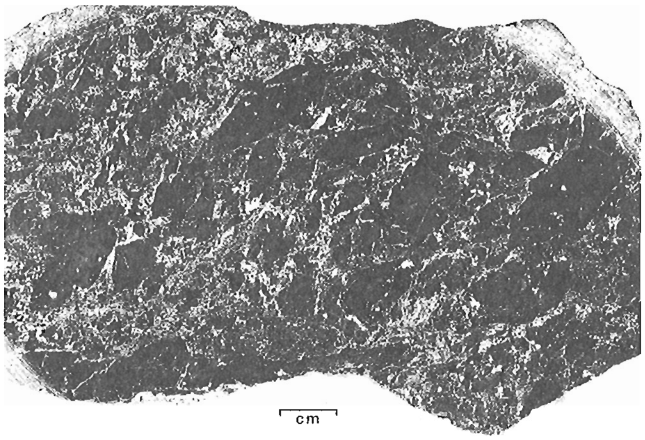

Figure 10a. Photograph of polished slab of a mineralized sample from zone A-4 (Fig. 9, sample GFA-'87-17). Note fragments of dacitic feldspar porphyry (light grey) and abundant magnetite (white) as breccia-filling. GSC 204130-B

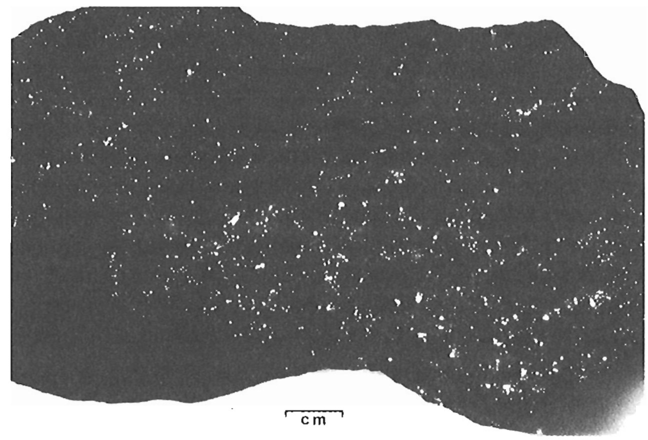

Figure 10b. Autoradiograph of the polished slab shown above. Note distribution of pitchblende as disseminated grains in the magnetite-rich breccia-filling.

The smaller magnetite-rich radioactive occurrences are similar in character to the main zones. A new pitchblende occurrence (A-5, Fig. 2) apparently unrelated to the magnetite-apatite-actinolite veins, was found during the present study. It has high radioactivity along fractures within a 15 x 10 m area of dacitic feldspar porphyry. There is one veinlet of pitchblende and chalcopyrite up to 0.8 cm wide and 10 cm long. No magnetite, apatite or actinolite are associated with these radioactive fractures, but some dark grey alteration is developed along them, and appears to be due to hornblende ± epidote. The fractures are steep, and a majority of them trend north-northwest, but the one with visible pitchblende has a westerly bend.

ORIGIN OF MINERALIZATION

The geological setting of the Fab occurrences is similar to that of the magnetite-apatite-actinolite veins found at and near the margins of the quartz monzonite intrusions in the Great Bear and Great Slave Lakes regions (Badham and Morton, 1976; Badham, 1978; Gandhi and Prasad, 1982; Hildebrand, 1986). Uranium, copper, and rarely cobalt-nickel arsenides are associated with some of the veins in these regions. The veins all occur in the tensional fractures, which can be interpreted as related to cooling of the epizonal quartz monzonite intrusions. In the Fab Lake area, the broad tensional zone occurs in the volcanic sequence and adjacent dacitic feldspar porphyry which dip steeply towards the quartz monzonite intrusion in the east. Tilting of the sequence may have occurred prior to or during the intrusion of quartz monzonite. The complex set of events could have resulted from caldera collapse.

The hydrothermal fluids may have acquired some of the iron and other metals from the country rocks during their ascent. In this respect, it must be noted that the mafic fragments in unit 3 are indicative of possible source of the metals, particularly iron, in differentiated magma chamber. There appears to be a regional zonation in the Fab Lake area with relatively magnetite-rich occurrences in the south where the host rocks are virtually surrounded by the quartz monzonite. This suggests possible thermal gradient decreasing towards the north. Relatively higher temperature in the south may have favoured greater magnetite concentration relative to that of apatite and actinolite. Higher magnetite deposition in turn may have created more favourable Eh-pH conditions for the precipitation of uranium. Sulphides were deposited during this main stage of mineralization.

The FAB claims occurrences bear close similarities to the Sue-Dianne Cu-U prospect which is a magnetite cemented breccia-pipe in dacitic to rhyolitic ash flow tuffs or ignimbrites (Climie 1975, 1976). McGlynn (1979) recognized a quartz monzonite — granodiorite pluton within 2 km west of the prospect. Thus it can be postulated here that the breccia-pipe is related to an unexposed quartz monzonite pluton in the vicinity which may or may not be connected to the pluton to the west.

The breccia-pipe is approximately 400 m in diameter, and 4 drill holes through it intersected copper values over core lengths of 3 m to 129 m to a depth of 200 m. The assays ranged from 0.46 to 1.12 % Cu, and showed up to 90 ppm U over 32m, and traces of Ag and Au (Climie, 1976). It is suggested that this disseminated uranium was deposited contemporaneously with copper and iron.

In addition, uranium occurs in high grade younger veins which were not intersected in drill holes, and which contain Bi-Cu-Pb-S-Se-Te minerals (Climie, 1976; Miller, 1981,1982). These have similarities to the polymetallic veins at Great Bear Lake (Lang et al, 1962). It is apparent from the above that the Fab Lake area has potential for simple and polymetallic veins in addition to the Sue-Dianne type magnetite-rich deposits.

CONCLUSIONS

The mineralization in the FAB claims occurs in a volcano-plutonic complex of calc-alkaline character. Pitchblende and chalcopyrite were deposited contemporaneously with abundant magnetite and subordinate apatite, actinolite, pyrite and fluorite, in tensional openings. The mineralization is interpreted as hydrothermal, related to a quartz monzonite intrusion. Their geological setting is similar to the magnetite-apatite-actinolite occurrences associated with other quartz monzonite intrusions in the Great Bear magmatic zone (Hildebrand, 1986) and in the Great Slave Lake region (Badham, 1978; Gandhi and Prasad, 1982), which define a petrographic province related to a subduction event at about1870 Ma.

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Geology of the north half of the Taltson Lake map area, District of Mackenzie

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Abstract

East of the Taltson Magmatic Zone, retrograded metadiorite, mafic and granitic gneisses and amphibolite of inferred Archean age are overlain by fine to coarse clastics of the Nonacho Group of probable early Proterozoic age. Within the zone two batholithic plutons comprising the Slave Monzogranite (1955 Ma) and the Konth Syenogranite (1922 Ma) and a minor charnockite pluton intrude granulite facies gneisses likely of Archean age. Granodiorite, diabase, quartz latite and feldspathic dykes with associated fluorite intrude these rocks.

Zones of early Proterozoic ductile shear have developed extensive mylonite belts in quartzofeldspathic rocks and a major belt of low grade slate and schist in gneisses of sedimentary-volcanic origin. Cataclasites are prominent locally. These zones provide a complex pattern of sinistral and dextral shear which may contribute to a better understanding of early Proterozoic regional plate tectonics if the sequence and absolute age of the various shears can be determined.

Résumé

À la zone magmatique de Taltson, de la diorite métamorphisée et rétrogradée, des gneiss mafiques et granitiques ainsi que de l'amphibolite de l'Archéen déduit sont recouverts par des roches clastiques fines à grossières du groupe du Nonacho datant probablement du début du Protérozoïque. À l'intérieur de la zone, deux plutons batholitiques comprenant de la monzogranite de la province des Esclaves (1955 Ma) et de la syénogranite de Konth (1922 Ma) ainsi qu'un pluton mineur de charnockite pénètrent des gneiss à faciès de granulite datant probablement de l'Archéen. Des dykes de feldspath, de latite à quartz, de dolérite et de granodiorite avec de la fluorine associée pénètrent ces roches.

Des zones de cisaillement plastique remontant au début du Protérozoïque ont formé d'immenses zones de mylonite dans des roches quartzofeldspathiques ainsi qu'une grande zone d'ardoises et de schistes à faible teneur dans des gneiss d'origine volcano-sédimentaire. Des roches cataclastiques sont proéminentes par endroits. Ces zones fournissent une configuration complexe de cisaillement sénestre et dextre susceptible de contribuer à une meilleure compréhension de la tectonique régionale des plaques du début du Protérozoïque à condition de pouvoir déterminer la séquence de l'âge absolu des divers cisaillements.

SURFICIAL GEOLOGY

Although exposure of bedrock is excellent over most parts of the Taltson Lake sheet, (NTS 75E) overburden increases eastward and northward. Over most of the area ice movement was from the northeast, but in the northeast near Gagnon Lake, more westerly directed striae may represent late draw-down of ice into the basin of Great Slave Lake. A thin veneer of silt was deposited in local ephemeral lakes as the ice retreated. Intervening terrain is dominated by thin till and eratics. Erratics derived from the Nonacho Group at the east margin of the area are present throughout.

GENERAL GEOLOGY

Archean(?)

Mixed gneiss and plutonic rocks (1)

Rocks of Unit 1 (Fig. 1) predominate in the basement beneath Nonacho Group along the east margin of the map area and they extend westward to a younger plutonic front (Taltson Magmatic Zone, Bostock et al., 1987) marked by intrusion of unfoliated granite (Unit 5). They include rocks of dioritic composition intruded by at least two ages of granite and pegmatite.

Dioritic rocks of Unit 1 locally consist of massive, medium- to fine-grained, biotite or hornblende-biotite diorite; but mostly the rock is foliated to gneissic rocks. Bands to irregular intrusive bodies of both foliated and massive, fine- to medium-grained biotite or chlorite granite contain local amphibolite bands. In places the dioritic rocks persist only as remnants within more granitic rocks. Rarely biotite- chlorite schists with minor muscovite suggest that a metasedimentary component may be present.

One body (unit 1b) of strongly deformed megacrystic granite, characterized by extensively granulated potash feldspar megacrysts up to 3 or 4 cm long in a medium-grained quartz-feldspar-biotite (chlorite) matrix, occurs within the gneisses west of Borrowes Lake. Part of this body forms a thick, gently dipping sheet deformed with the enclosing gneisses. It therefore appears to be significantly older than unit 5 (which is largely undeformed) and may be only slightly younger than the enclosing gneisses. Elsewhere within the gneisses local irregular bands of similar though less granulated megacrystic chlorite-biotite granitoid rocks appear to form part of the gneiss sequence.

Crosscutting relations indicate that the dioritic rocks are older than the granite batholith (Unit 5) which intrudes them on their western border. Some of the granitic phases present within the mixed gneisses of Unit 1 may be correlative with this batholith. The dioritic rocks lithologically resemble some rocks within the Hill Island Lake area far to the south which extend southward into the Archean Noland Block (Van Schmus et al., 1986).

Paragneiss (2)

Paragneiss With Minor Amphbibolite

Paragneiss (2) containing minor amphibolite, hornblende biotite schists (2a) is the most widespread component of Unit 2. It occurs as remnants included throughout most parts of the syenogranite (Unit 6) and the monzogranite (Unit 4) from Rutledge Lake west to the west border of the map area.

The most common lithology is fine grained garnet biotite gneiss. Sillimanite and cordierite are recognizable locally. In places grey, fine-grained plagioclase-rich bands contain hypersthene suggesting that the metamorphic grade has reached granulite facies. Quartz-rich and calc-silicate gneisses form a subordinate component. At Rutledge Lake some thin bands (up to a few metres thick) of metabasite may be metamorphosed iron formation whereas others appear to be metamorphosed mafic intrusions. Graded bedding observed in similar metasediments in the southern part of the Taltson Lake area (Bostock, 1987) suggest that the gneiss is derived from turbidites.

Although the paragneiss within the Taltson Lake and the Fort. Smith areas are not dated the gneiss is lithologically similar to rocks in the Fitzgerald area in northeastern Alberta (lying along strike) which are cut by 2500 Ma pegmatites (Baadsgaard and Godfrey, 1972) indicating that the gneisses are of Archean Age. It remains possible however that more than one gneiss sequence may be represented in this vast area.

Paragneiss With Amphibolite

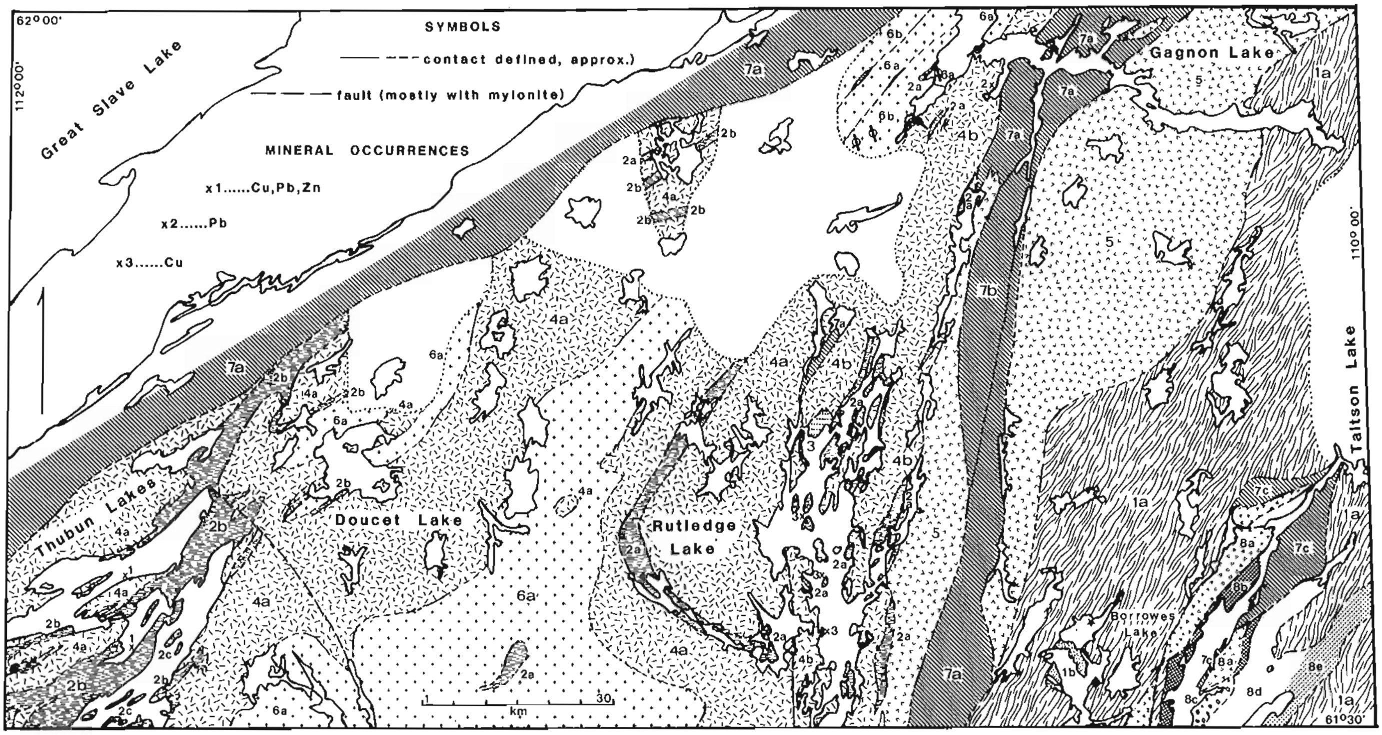
Paragneiss with an extensive mafic volcano-plutonic component (2b) is extensive only near Thubun Lakes. The sedimentary component is similar to that of Unit 2a but includes a higher proportion of calc-silicate gneiss and marble.

The mafic component of the paragneiss at Thubun Lakes consists of mafic and intermediate amphibolites and metagabbros. Other lithologies are so sheared that interpretation of their origin cannot be made with any certainty. Felsic bands reduced to mylonite are common and some sugary textured felsic gneisses may be recrystallized mylonites. Metamorphic grade of the mafic rocks is typically epidote amphibolite facies but the pelitic gneiss with which they are associated appears to pass by decreasing retrogression into middle or upper amphibolite facies within the surrounding area.

Plagioclase-rich and garnetiferous gneisses with bands of quartz dioritic composition are present within monzogranite (Unit 5) in the isolated area of mapping some 25 km east-southeast of Gagnon Lake. These may represent a calcareous metasedimentary protolith with a smaller component of mafic volcanic material than is present at Thubun Lakes. Small remnants of similar plagioclase-rich gneisses are found locally in the monzogranite south of Gagnon Lake.

Low Grade Tectonite

Schist, slate, gneiss and mylonite (2c) form islands in south Thubun Lake and merge with a mylonite belt which cuts the Deskenatlata Granodiorite of the Fort Resolution area (85H) to the southwest. The unit comprises low grade schists, slate and gneiss with prominent, fine-grained chlorite, muscovite and/or biotite, local garnet and minor quartzofeldspathic mylonite bands. Rocks on some islands at the northeast end of the lake having a high quartz-feldspar content and gneissic texture are also associated with minor mylonite.



LEGEND

PROTEROZOIC

HYPABYSSAL INTRUSIVES

feldspathic dykes

diabase dykes (not shown)

* quartz latite dykes

♦ granodiorite dykes

NONACHO GROUP (8)

a. monomict conglomerate
b. slate, schist
c. polymict conglomerate
d. sandstone with local pebble beds
e. undivided Nonacho Group including arkose

SHEAR ZONE LITHOLOGIES (7)

a. mylonite, some sheared granite and amphibolite
b. mixed mylonite and cataclasite
c. cataclasite

SYENOGRANITE (6)

a. garnet-cordierite-bearing megacrystic syenogranite with common paragneiss inclusions
b. granite, pegmatite, mylonite, cataclasite

GRANITE (5)

massive to megacrystic unfoliated monzogranite with local included amphibolite concentrations

MONZOGRANITE (4)

a. massive to foliated, garnet-cordierite-bearing monzogranite, some fine grained granite, common paragneiss inclusions
b. foliated to mylonitic garnet-cordierite-bearing monzogranite with common paragneiss inclusions

ARCHEAN

CHARNOKITE (3)

hypersthene-bearing granitic rock

PARAGNEISS (2)

a. paragneiss with minor amphibolite
b. paragneiss and amphibolite, mylonite, dioritic gneiss
c. low grade tectonite of sedimentary origin

MIXED GNEISS AND PLUTONIC ROCKS (1)

a. dioritic rocks, granite, amphibolite, minor paragneiss
b. foliated megacrystic granite-granodiorite.

Figure 1. Geology of the north half of Taltson Lake area (75E) south of Great Slave Lake Shear Zone (northwestern unit of 7a).

The intimate association of Unit 2 with higher grade schists and gneisses along their margins, their high state of shear, and the fact that they are directly along strike with a mylonite belt to the southwest, suggest that they are tectonites derived from the high grade paragneiss of Unit 2. To the northeast the shearing with which they are associated splays piecemeal eastward into the monzogranite (unit 5), but may in part step north into the valley of north Thubun Lake.

Charnockite (3)

Charnockite (hypersthene-bearing granite) was described by Culshaw (1984) on the northwest shore of Rutledge Lake. The body is oval shaped being about 4 by 1.5 km in length and breadth.

The charnockite is grey to white, fine- to medium-fine-grained, has a colour index of 5 to 10, and mafic minerals are largely altered to biotite and/or chlorite. Fine-grained mafic bands on the eastern side of the charnockite may be deformed dykes, or boudins derived from banding of unknown origin. Locally unfoliated pegmatites cut the charnockite.

Lineation tends to be better developed than foliation. At its northeastern margin the charnockite appears to be interleaved with paragneiss (Unit 2) and both are cut by strong foliation. The high degree of deformation and metamorphism in this body suggests that it intruded the paragneiss before much or all of Proterozoic plutonism.

Proterozoic (Aphebian)

Monzogranite (4)

Massive To Foliated Monzogranite

Monzogranite forms part of a northerly trending major batholith between Rutledge and Thubun Lakes. On the northwest it is transected by the Great Slave Lake Shear Zone (GSLSZ) (Hanmer and Lucas, 1985). In the central part it is extensively intruded by megacrystic syenogranite (Unit 6).

The batholith comprises grey, white, buff, or pink, medium-grained equigranular to slightly megacrystic, biotite monzogranite varying locally to granodiorite or syenogranite. Garnet and cordierite or their alteration products are common. Remnants of paragneiss and locally of metabasite are widespread within the monzogranite but are mostly of submappable size. A few occurrences of fine-grained granite are thought to be consanguineous with the coarser grained monzogranite.

Rocks from the northwestern part of the Fort Smith area (NTS 75D), considered correlative with the monzogranite, have been dated at 1955 $\pm$ 2 Ma (U/Pb zircon, Bostock et al., 1987). Discontinuous remnants of similar granite, albeit of widely different degrees of deformation and metamorphic overprint, extend southward into Alberta where they have been called the Slave Granite (Godfrey and Langenberg, 1978).

Foliated To Mylonitic Monzogranite

Fine- to medium-grained mostly strongly deformed monzogranite (4b) occurs on islands in and along the shores of Rutledge Lake where it is extensively intersheared with remnants of paragneiss (Unit 2).

Most rocks are fine grained and either strongly lineated or strongly foliated. Mafic minerals are commonly partly altered to chlorite. Garnet is evident in places. Tough, particularly fine-grained varieties resemble mylonite except they are rarely porphyroclastic. A narrow mylonite belt marks the western boundary of the unit beyond which foliation decreases and grain size increases gradationally into typical monzogranite (Unit 4a). At the east margin the unit is in highly sheared and veined contact with a major belt of paragneiss (Unit 2, Mama Moose Complex of Culshaw, 1984). Metabasite on the east side of this belt is intruded by less deformed granite (Unit 5).

The weakly foliated monzogranite of Unit 4b passes southward into more strongly foliated monzogranite of Unit 4a.

Granite (5)

Unit 5 forms a lenticular granitic batholith centred about 10 km south of Gagnon Lake. The granite is mostly massive and of monzogranite composition ranging in grain size from two to five centimetres. Large areas within the pluton are potash feldspar megacrystic. Biotite, the most common mafic mineral, is partly or entirely altered to chlorite. Inclusion and partial assimilation of gabbro and amphibolite have extensively contaminated the southeastern margin of the batholith. Much of the central part of the intrusion is as yet unmapped.

The batholith intrudes the mixed dioritic gneiss complex (Unit 1) along its eastern margin, but unfoliated granite and pegmatite within this complex may be of similar age. Near its western margin the batholith is cut by a major mylonite belt, leaving a fringe of foliated granite to the west of this belt. This fringe locally intrudes metabasite of Unit 2 along Rutledge Lake whereas contacts between Unit 2 and the monzogranite (Unit 5) are much more highly sheared. The granite batholith is assumed to be younger than the monzogranite (Unit 4) although no absolute radiometric dating has yet been done.

Syenogranite (6)

Garnet-cordierite-bearing syenogranite

Syenogranite (Unit 6a, Konth Granite, Bostock et al., 1987) extends as a southward expanding belt along the axis of the monzogranite batholith (Unit 4) through the central part of the map area. Outlying lenses occur about Doucet Lake. The rock is pink to buff or grey and is characterized by megacrysts of potash feldspar in variable abundance which locally reach 3 or 4 cm in length in a matrix which may be fine- or medium-grained. Garnet and cordierite, though commonly partly altered to chlorite, are widely recognizable. Most of the rock is only weakly to moderately foliated. A U/Pb concordia zircon-monazite date from the central part of the pluton in the Fort Smith area along strike to the south suggests an age of emplacement at 1922 $\pm$ 2 Ma (Bostock et al., 1987).

Pegmatite, Granite, Mylonite, Cataclasite

Pink to salmon coloured, foliated to massive, fine- to coarse-grained, locally finely fractured leucocratic granite (6b) with remnants of paragneiss occupies an unusually heavily drift-covered area immediately west of Gagnon Lake. Coarse patches of grey to white quartz are common. Mafic minerals, predominantly biotite-chlorite, form 2-3 per cent of the rock. Garnet and possibly altered cordierite are present locally both in the granite and in included metasediments. Some exposures are megacrystic in zones with potash feldspar subhedra reaching 2 to 3 cm in length. Elsewhere potash feldspar porphyroclasts occur in fine-grained granitic mylonite. It is possible that the granitic rocks in this unit consist of a mixture of Units 2, 4 and 6 which have been variably sheared, intruded by pegmatitic granite, recrystallised and subject to cataclasis.

Shear Zone Lithologies (7)

Major shear zones occur along the west, north and eastern margins of the map area. In the west about Thubun Lakes, where three zones of shearing merge (Fig. 2), the shear zone lithologies are derived in large part from a paragneiss-amphibolite protolith which has produced slates and schists with a relatively small proportion of mylonite. These rocks have been included with Unit 2 because the amount of shear which they represent is commonly uncertain. In the east, where the protolith is predominantly quartzofeldspathic gneiss and granite, major belts of mylonite and cataclasite have formed. In the north the map area is truncated from southwest to northeast by the dextral Great Slave Shear Zone (Hanmer and Lucas, 1985; Hanmer and Connelly, 1986). Lithologies within the southeastern part of this zone are predominantly quartzo-feldspathic mylonite.

Mylonite

Mylonites (7a) include protomylonite, mylonite proper, and ultramylonite. Scattered bodies of amphibolite and zones of sheared granitic rocks occur within the mylonite belt north and south of Gagnon Lake. Bands of mylonite from protomylonite to ultramylonite occur within the gneisses and schists at Thubun Lakes, and suggest that most of these latter rocks have undergone a similar degree of shear. Age, correlation and kinematics of the various shear belts are considered under structural geology.

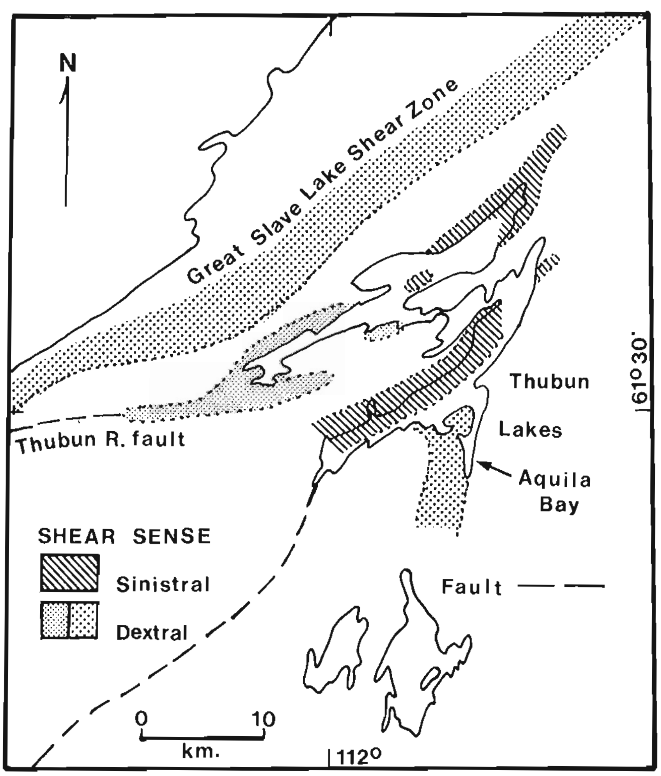


Figure 2. Shear sense in rocks near Thubun Lakes (dextral shear in Great Slave Shear Zone after Hanmer and Lucas, 1985).

Mixed mylonite and cataclasite

Mixed mylonite, amphibolite and cataclasite (7b) occur in a large as yet poorly defined area within the north-south belt of mylonite immediately south of Gagnon Lake. Much of the rock is fine grained to aphanitic and of granitic composition with common feldspathic porphyroclasts. Bodies of amphibolite appear to have been partially assimilated within these granitic rocks and are therefore at least in part probably older than the granitic rocks. Quartz and epidote veining is widespread and alteration of mafic minerals to chlorite is evident. Outcrops are commonly red stained and magnetite is evident in places. Foliation is discontinuous and patches of breccia are common.

Mylonite within the zone of cataclasis is intruded by diabase dykes presumably of the Sparrow Dyke Swarm (about 1700 Ma, McGlynn et al., 1974). The dykes are broken up and altered suggesting that at least a part of the cataclasis is later than they are.

Cataclasite

A major zone of cataclasite (7c) is developed along the southeastern shore of Taltson Lake, and similar rocks occur at the north end of the lake. Cataclasis is most severe in a belt along the southeastern shore. The cataclasite consists primarily of variably fractured to brecciated, medium-grained, megacrystic to equigranular granitic rocks intruded by hypabyssal mafic bodies and by granite pegmatites. Intimate brecciation of the granite, alteration of mafic minerals to chlorite-epidote, and quartz veining are well exposed in the granitic cataclasite along the shores of the lake. Small irregular shaped to dyke-like, fine-grained mafic intrusive bodies are altered but commonly appear less brecciated and granite pegmatites show least brecciation and alteration. Along the southeast shoreline of Taltson Lake cataclasis is acccompanied by faulting with juxtaposition and repetition of Nonacho Group and its granitic basement together with rocks of unknown protolith. In some of these fault slivers masses of pink granite are veined with pink ultracataclasite (less than 10 per cent of granite fragments in a comminuted unfoliated granitic matrix). In others irregular masses of green ultracataclasite up to 2 or more meters across occur within foliated altered greenish granitic rocks. Zones of oxidation and silicification are present. Flinty intrusive siliceous rock with minute splinters of quartz occurs along a fault at one point on the shore. In places within this zone masses of vaguely foliated cataclasite up to 100 m wide separate slivers of Nonacho Group, or lie between Nonacho Group and granitic rocks. These cataclasites contain lenses of granitic rocks and partially dismembered minor, fine grained mafic bodies in a chloritic medium- to fine grained matrix of granitoid composition. A pegmatite was seen to cut across one of these cataclasites perpendicular to the adjacent faulting, but at other places pegmatites appear to have been deformed by cataclasis.

Nonacho Group (8)

Nonacho Group in the southeast corner of the map area occupies part of a remnant of a much larger sedimentary basin which extends to the northeast, east and south. Within the map area it consists of five lithologic units: monomict conglomerate (8a); slate, schists, argillite and siltstone (8b); polymict conglomerate (8c); and arkose (8d). It is not clear from the present mapping whether these units are discrete, or whether each may represent parts of two or more cycles separated by other units or by unconformities. The group has been studied in detail by Aspler (1985) whose work has suggested that such cycles do exist. The Group lies unconformably on rocks thought to be mostly of Archean age.

Monomict conglomerate

The monomict conglomerate (8a) is most extensively exposed along the northwest margin of the group, but it is also repeated by faulting in the central part on the southeast shore of Taltson Lake. It is monomict in the sense that almost all of the clasts are derived from basement (igneous or metamorphic lithologies or vein quartz). Occasional clusters of mafic volcanic clasts have likely provenance in nearby irregular mafic dykes which are intruded by pegmatite. Clasts tend to be most abundant, most angular and reach one metre or more in diameter near the basal unconformity. Thin silt, sand and pebble beds occur in the southeastern (upper?) part of the unit northwest of Taltson Lake. In places the unit appears to pass conformably upward into siltstone-shale, but the actual contact is rarely seen and is commonly the locus of increased deformation.

Slate, argillite, siltstone, schist

A grey to greenish, commonly finely bedded to laminated siltstone-slate with local isolated granite pebbles or cobbles (8b) appears to lie above the monomict conglomerate along the northwest shore of Taltson Lake and inland on the opposite shore. On the southeast shore of the lake, where bedding and cleavage are not parallel, cleavage is refracted at intersections between fine and coarser clastic beds. Near bedding planes left lateral deflection of cleavage occurs in the finer component in several places.

Polymict conglomerate

Polymict conglomerate (8c) occurs at the south central extremity of the group within the map area. It consists of clasts, usually pebbles and cobbles, of a wide variety of lithologies in a sandy matrix. Basement lithologies typically predominate but are accompanied by variable proportions of fine- to medium-grained clastic sediments. Interbeds of sandstone are common. Along its northern contact with the monomict conglomerate, exposures with intermediate numbers of sediment and basement clasts occur. This may reflect changing source areas within laterally equivalent units, or the polymict conglomerate may be the younger as suggested by limited indicators of stratigraphic sequence and its apparent position within the central part of the group.

Sandstone with local conglomerate interbeds

This unit (8d) occurs in the central part of Nonacho Group and appears to grade southward into polymict conglomerate. It consists of fine- to medium-grained sand with local coarse sand or siltstone interbeds mostly less than 1 m thickness, and isolated rounded cobbles. Rare distinctive, cobbles of white orthoquartzite in the sandstone are of unknown provenance.

Undivided Nanacho Group

The area shown on the map as undivided Nonacho Group (8e) has been examined only at its northern periphery where schist, siltstone and a pinkish brown arkose not previously seen, have been observed. Air photo interpretation of other parts of the area suggest that it is also underlain by Nonacho Group.

Age relations

The Nonacho Group is intruded by the Sparrow Diabase Dyke Swarm dated by Ar^{40}/Ar^{39} at about 1700 Ma (McGlynn et al. 1974). The maximum limiting age is poorly constrained by the likely Archean age of basement upon which it rests unconformably. Within the map area the unconformity is best exposed on the central southeast shore of Taltston Lake about 10-15 m back from the waterline. There the basement consists of granite cataclasite intruded by pegmatites, and is overlain by an outlier of siltstone and conglomerate. At the unconformity there occur 4 cm of granite breccia, 50 cm of granite pebble conglomerate and then granite sand interleaved with siltstone grading upward (over 6 or 7 m) into slate. Pegmatites do not intrude the unconformity and cataclasis at least locally does not appear to affect rocks above it. On the other hand pegmatites are intrusive into cataclasites of unknown protolith that appear to merge with the conglomerate along faults farther south. These latter cataclasites may include a sedimentary component. It is possible therefore either that the pegmatite and cataclasite are distinctly older than Nonacho Group, or that the cataclasite (and pegmatite) accompanied uplift and faulting which preceded and accompanied deposition of Nonacho Group. By the latter hypothesis it is possible that pegmatite emplacement was taking place at depth (not within Nonacho Group) during part of the interval over which Nonacho Group was being deposited.

Hypabyssal intrusives

Hornblende granodiorite. A hornblende-bearing granodioritic dyke at least 150 m wide and possibly as much as 9 or more km long occurs about 300 m west of Borrowes Lake at the south margin of the map area. Grain size reaches up to 1.5 mm but phenocrysts of hornblende are locally up to 2.5-3 mm. Some parts of the dyke contain scattered feldspar phenocrysts reaching 2 cm. Feldspar, hornblende, chlorite and locally quartz are evident in the matrix. Angular inclusions of amphibolite up to 1 mm diameter are widespread, inclusions of hornblende gabbro, medium grained granite and coarse grained hornblendite are present locally. The dyke intrudes megacrystic granite but its relationship to other dykes is unknown.

Quartz latite

Two small irregular bodies of grey to white aphanitic to fine grained porphyritic quartz latite intrude biotite gneiss on adjacent islands in a lake northwest of Borrowes Lake. These bodies have irregular chilled margins, contain bipyramidal quartz, plagioclase and biotite phenocrysts up to 2 to 3 mm in diameter and are unfoliated and apparently unaltered.

Sparrow diabase

Typically northwesterly trending, aphanitic to fine-grained diabase dykes (not shown in Fig. 1), reaching about 5 m in thickness but mostly less than 1 m, are present throughout the map area. Dykes which depart from the northwesterly trend tend to occur in zones which have been deformed by cataclasis and have themselves been fractured and altered. The swarm has been dated by McGlynn et al. (1974) at about 1700 Ma by the Ar^{40}/Ar^{39} method.

Feldspathic dykes

Eight steeply dipping aphanitic to fine grained, equigranular, feldspar-rich dykes up to 12 m wide occur on either side of the large western arm of Gagnon Lake and in the surrounding country. Small inclusions of country rock are common in some of these dykes, and many of them have purple fluorite lining fractures along their margins. Although contacts are irregular in detail, the dykes are tabular form and appear undeformed.

A feldspathic dyke has been emplaced in gneiss (Unit 2a) alongside and apparently parallel to the contact of an irregular body of Sparrow diabase on an island in the southern part of the west arm of Gagnon Lake. Both dykes appear to be chilled locally at their contacts, but the diabase is penetrated by several felsic veins up to 1 cm in width for a distance of up to 1 metre from the contact. These may indicate that the feldspathic dykes are the younger, but it is also possible that emplacement of the diabase has remobilized felsic material along its contacts.

METAMORPHISM

Pelitic rocks of Unit 2 contain sillimanite-cordierite-microcline assemblages over most of the map area. Thin sections of similar rocks in the southern half of Taltson Lake area show that plagioclase-rich bands within Unit 2 commonly contain hypersthene. Hypersthene also occurs locally in the granitic rocks of Units 4 and 6. Thus Unit 2 may represent an older granulite facies terrain which has been either intruded by younger granites which have then reached similar metamorphic conditions, or within which the intruded granites have been relatively dry permitting xenocrystic hypersthene to persist. East of a north-south line approximately along the east margin of Rutledge Lake mineral assemblages are characterized by chlorite-epidote-hornblende and rarely garnet. Although pelitic gneiss is absent or much less common in the eastern terrain where the rocks are typically more mafic, it is clear that there is a substantial lowering of maximum metamorphic grade present from granulite facies to upper greenschist-lower amphibolite facies across this line. The age of high grade metamorphism is as yet undetermined.

A pronounced zone of retrograde metamorphism is evident along the valley of south Thubun Lake. There upper amphibolite or lower granulite facies gneisses have been reduced to schists and slates of greenschist to lower amphibolite facies grade in association with strong sinistral shearing.

STRUCTURAL GEOLOGY

Structure of rocks up to and including Unit 6 (syenogranite) has been profoundly affected by Middle Proterozoic shearing. This shearing is nevertheless concentrated along the east and west margins of the monzogranite (Unit 4) and is thus in large part coincident with the east and west margins of early Proterozoic (S-type) granite plutonism within the Taltson Magmatic Zone (Bostock et al., 1987). Field examination of kinematic indicators related to this shearing has produced the structural patterns discussed below (Fig. 2, 3). Because the presently known distribution of kinematic indicators is sparse and uneven, the deduced preliminary pattern is tentative. The kinematic indicators used are predominantly shear bands, and to a lesser extent rotated feldspars. That the shear has been predominantly horizontal is indicated by the generally shallow plunge of stretching lineations.

In the western part of the area near Thubun Lakes (Fig. 2), the shear pattern is complicated by intersection of at least three distinct shear zones. The oldest of these, and the one that is regionally least well defined, is a zone of dextral shear which trends in a northerly direction through Acquila Bay (Fig. 2). Dextral kinematic indicators occur within the gneisses along this zone, and both dextral shear and shallow plunging lineations go around the axis of a steeply plunging major fold immediately south of Thubun Lakes. The southward extent of this zone is not yet adequately established, but it is possible that it is part of a more extensive zone that continues south to Deskenatlata Lake where north-south oriented dextral shear has been observed in the gneisses (about 61°N 112°W). Alternatively it may represent the preserved part of a major fold structure of which one limb was pinned during folding.

Cutting across the north-south zone from the southwest is a zone of sinistral shear that follows the valley of south Thubun Lake and splays eastward into the granite terrain to the east. In the eastern part of the lake sinistral shearing extends north into the northeastern half of north Thubun Lake and continues for a presently unknown distance beyond to the northeast. Sinistral kinematic indicators are abundant and well exposed along most of the lake shores indicated (Fig. 2). Minor sinistral mylonite bands, which strike roughly parallel to this shear zone, cut through the nose of the major fold at Acquila Bay which has deformed the dextral shear zone described above. Sinistral shear therefore clearly postdates the southern dextral shear zone.

A third zone of shearing enters north Thubun Lakes from the west at about Thubun River and appears to merge with the south Thubun shear zone at and to the north of the creek joining the two lakes. At the western extremity of north Thubun Lakes dextral kinematic indicators are abundant in rocks of this zone and are present along the northwest shore almost as far as the central part of the lake. Along the contact between the second and third zones kinematic indicators are less common and may be of opposite sense in different exposures. Age relations between the northern dextral and the sinistral shear zones are currently indeterminate.

Shearing related to the northern two zones appears to die out eastward; however, the sinistral belt may continue southwestward as a narrower mylonite belt (about 100 m wide) which transects the Deskenatlata Granodiorite as far as the Paleozoic overlap at Rat River (Bostock, 1986). Dextral shear bands related to the northern dextral shear zone have been found along Thubun River to a point some 10 km west of Thubun Lakes and the zone likely extends at least to its projected intersection with Great Slave Lake Shear Zone. It is perhaps possible that two sets of conjugate shears are present: an early dextral-sinistral set represented by the Thubun River fault-south Thubun shear zone, and a later dextral-sinistral set represented by pairing dextral movement on GSLSZ with movement on the northeast Thubun Lakes sinistral zone (see Fig. 1, 2). Such an interpretation would suggest explanations for the north-stepping sinistral shear pattern, and for the presence of a complex interference structure on the northwest coast of north Thubun Lakes. Further work may permit a test of this hypothesis.

Proterozoic shearing (Fig. 3) is complex along the east margin of Taltson Magmatic Zone. At Lady Grey Lake south of the map area (just south of the 61st parallel) major sinistral shear zones accompanied by mylonite belts up to several km across draw together. Sinistral kinematic indicators are widely present within the gneisses between mylonite belts and in places have been found in the megacrystic syenogranite batholith west of the major mylonite belts. Sinistral shear, representing a continuation of this shear belt, is locally evident in the eastern gneisses of the southern half of the Taltson Lake sheet. Kinematic indicators are best exposed, but are nevertheless discontinuous, along Taltson River valley south of Taltson Lake. Deformation of the gneisses becomes more brittle northwards, and faults splay northeastward from the river valley. A northern section of this sinistral shear zone (Fig. 3) cuts across the southeastern corner of the map area where a part of its northwestern edge culminates in the zone of cataclasis (Unit 7c) described above. Cleavage-bedding intersections in the Nonacho slates along the margins of the uplifted and cataclastically deformed basement are left lateral.

Farther west, between Taltson and Rutledge lakes, a second major belt of mylonite and cataclasite (initially reported by Culshaw, 1984) trends northward through Gagnon Lake (Fig. 3). It is broadest where it follows the western margin of the unfoliated granitic batholith (Unit 5); but just beyond the southern limit of the batholith, it splays southeastward into mafic gneisses as a series of thin granitic protomylonite bands. Evolution of the mylonite belt may therefore to some degree have been controlled by the presence of the batholith. Only a thin band of mylonite up to a few 100 m across extends south to the 61st parallel. Dextral kinematic indicators have been observed at three localities in this belt.

At Benna Thy Lake south of the map area (60°40′N; 110°34′W) a dextral mylonite belt up to several kilometres wide (possibly continuous with the dextral belt described above) is tangent to, and lies along the west side of, the major sinistral belt. A small elongate granitic pluton between them contains a dextral C and S fabric but no evidence of sinistral shear. The mylonite belt to the east however, is deformed by dextral kink bands. Current data suggest therefore that dextral shear occurred significantly later than the more widely extensive sinistral shear.

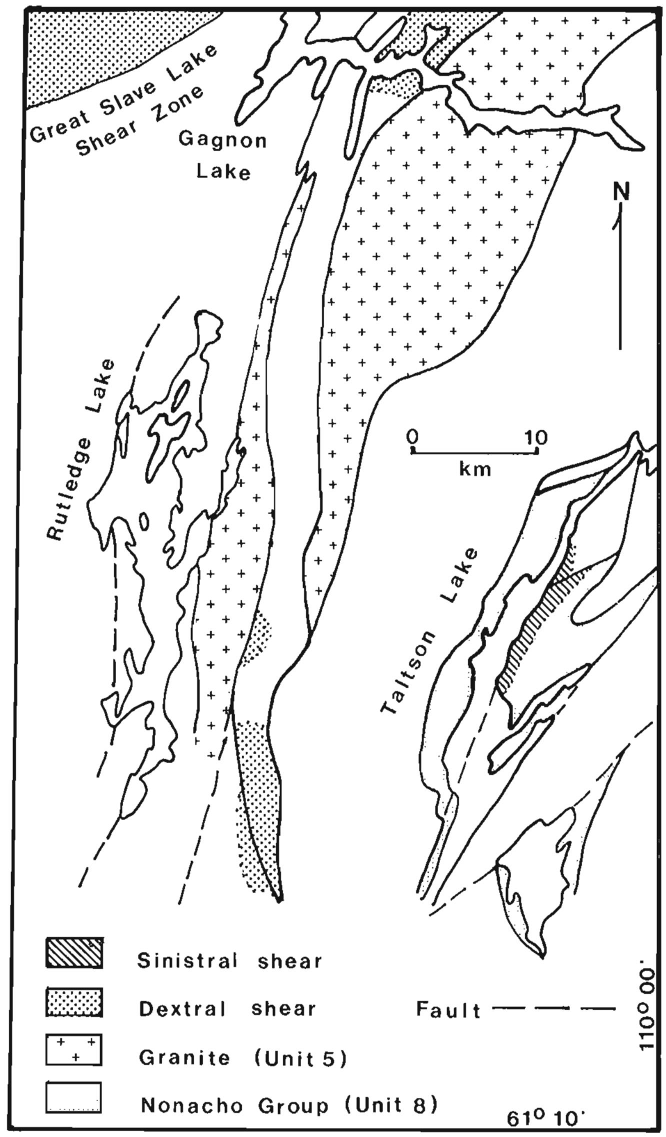


Figure 3. Shear sense in rocks along the east margin of Taltson Magmatic Zone (dextral shear in Great Slave Lake Shear Zone after Hanmer and Lucas, 1985).

A third major zone of north-south oriented shear is evident to the west of the major belt of dextral shear. This zone encompasses the valley of Rutledge Lake and extends from the western limit of the granite (Unit 5) to a narrow mylonite belt that follows the west margin of the lake. Mylonite at the western contact of the zone appears to die out both to the north and south within Unit 4 (monzogranite) and Unit 6 (syenogranite). The sense of shear within it is currently indeterminate.

Gibb (1978) suggested that major north-south oriented sinistral shear might be expected south of MacDonald fault as a result of a plate tectonic collision between Churchill and Slave provinces. The data collected here document the existence of such a shear zone, although the northeastern end of it remains unknown beyond the present project area. The data also suggest that at least part of the sinistral shear occurred within a period between 1922 Ma, the age of emplacement of the megacrystic syenogranite (Unit 6, which is the youngest dated unit to show the sinistral shearing), and emplacement of the Sparrow dykes (about 1700 Ma, McGlynn et al., 1974). This period is younger than that of emplacement of syntectonic granites in the Great Slave Lake Shear Zone (2.15-1.94 Ma, S. Hanmer, pers. comm., 1987).

ECONOMIC GEOLOGY

On the southeast shore of the large bay at the outlet of *South Thubun Lake* are the remnants of a camp with wooden buildings and a heap of discarded drill core. A few hundreds of metres to the southwest along the hillside a number of small blast holes have been excavated in a sequence of biotite gneiss with interlayered calc-silicate and minor marble bands. Traces of malachite, chalcopyrite and bornite are evident in some of these holes. In addition sphalerite, galena and pyrrhotite are reported by Reinhardt (1969). A pegmatite, which occupies a fold nose or lineation-parallel lens, is exposed for about 3 m down plunge parallel to the hillside at one of these holes. Fluorite, tourmaline and coarse muscovite are present in and along the margins of the pegmatite and selvedges of bright green feldspar, possibly amazonite (lead-rich potash feldspar), are present locally.

On the southeast shore of the central part of north Thubun Lake a quartz breccia vein up to 1 m thick and about 19 m in length cuts strongly foliated monzogranite. Scattered cavities contain cocks comb quartz with patches of brown sphalerite, galena, malachite and azurite.

Tourmaline is commonly found in lenses of pegmatite and in crosscutting granitic veins in the gneisses of Unit 2 about Thubun Lakes. It appears to be much less common in this unit than in most other parts of the map area.

Fine-grained mafic bands mostly less than one metre wide occur conformably within the paragneiss (Unit 2) at many places along the shore of *Rutledge Lake*. Rusty zones containing iron sulphide are associated with some of these bands. No clearly crosscutting relations were seen although both paragneiss and mafic bands are sheared. Some bands show slightly differeng colour and texture between rims and cores of bands. These minor bodies resemble iron formation seen at rare intervals in other parts of the Taltson and Fort Smith areas (Bostock, 1982, 1986), but no microscopic examination has yet been made of these rocks.

Fine- to medium-grained mafic bands and lenses also occur within and along the margins of the charnockite on the northwest shore of Rutledge Lake. These bodies less commonly have associated minor gossans. The largest mafic body occurs in the gneiss just off the northwest margin of the charnockite where it extends intermittently along shore for about 70 m and has a maximum exposed thickness of about 2 m. About 1 m of metapyroxenite rises above water level, the remainder consisting of metabasite. Other bands and lenses are finer grained and of uncertain mineral compositon.

Small stratiform bodies of sulfides (dominant chalcopyrite and pyrrhotite with important nickel content) were reported by Culshaw (1984) in association with minor ultramafic bodies and are marked on the map (Fig. 1). Examination of the mineral and chemcial composition of representative samples of these bands will be of particular interest.

The large western arm of *Gagnon Lake* lies along a zone of cataclasis and shearing within which numerous fine grained to aphanitic feldspathic dykes have been emplaced. Small pockets of purple fluorite are commonly present along the margins of these dykes and locally within small quartz-chlorite veins in the country rock. On the west shore of the next major southerly-trending bay to the east a quartz vein up to 5 cm thick cuts monzogranite and carries local pockets of galena. The vein meanders in and out of a narrow irregular breccia zone up to 15 cm across trending 104 degrees west from the shore. Minor galena and pink carbonate are associated with patches of cocks comb quartz along the vein.

DISCUSSION

Hoffman (1987), in a modification and extension of the indentor concept of Gibb (1978), proposed that, "the Great Slave Lake Shear Zone is a continental transform structure related to oblique collision and indentation of an Andean margin by the southeast corner of Slave microcontinent." Conjugate shears at Thubun Lakes and sinistral shear along the east margin of Taltson Magmatic zone likely represent secondary deformation related to shear along this major plate margin. Likewise southward displacement of a crustal block at the west margin of Churchill Province, caused by indentation of Slave Province along GSLSZ, may be responsible for extensive north-south sinistral shear along the east margin of Taltson Magmatic Zone. It is less clear how one or possibly two periods of north-south directed dextral shear may be directly related to the indentor model. Rather, all of these shear zones may be considered to have evolved in response to the broader sequence of plate tectonic events that occurred at the west margin of the early Proterozoic craton.

Acceptance of GSLSZ as a transform fault in the manner envisaged by Hoffman (1987) suggests that an active plate margin may have existed along the western edge of Churchill Province (now beneath Paleozoic overlap) when movement along GSLSZ began. Such a margin would presumably have been responsible for the evolution of Taltson Magmatic Zone within its adjacent sector. The sequence of sinistral-dextral shear, here supported by field work, and the observation that dextral shear does not cut GSLSZ but rather appears to bend into tangency with it (at least north of Gagnon Lake), allows that dextral shear was related to events along the plate margin west of Taltson Magmatic Zone.

Determination of absolute ages for the various shearing events involving Taltson Magmatic Zone, through dating of granitic dykes both intrusive into, and deformed by, the shear zones, will refine the position of at least some of these events within the local evolutionary sequence. Hopefully they will permit correlation of deformation with events in surrounding terrains leading to a broader understanding of plate tectonic history along the west margin of Churchill and Slave Provinces.

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Geology of parts of the Calder River map area, central Wopmay Orogen, District of Mackenzie

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Hildebrand, R.S. and Bowring, S.A., Geology of parts of the Calder River map area, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 199-205, 1988.

Abstract

Supracrustal rocks of the eastern Great Bear magmatic zone unconformably lie upon rocks of the Hepburn metamorphic-plutonic zone over the entire length of the map area. Rocks of both zones are folded about northerly-trending axes in the region of the Wopmay fault zone. The northerly-trending folds form a central corridor in a regional sigmoidal pattern of folds and reflect local increased shear strain perhaps due to the occurrence of a buried suture. Metasedimentary rocks of the Grant Group lie unconformably on Archean gneisses. At the southern margin of the map area and in the adjoining sheet, the Grant Group and its probable Archean basement are tectonically repeated on a thrust fault. Plutonism within the Great Bear magmatic zone ranges both compositionally and temporally from more intermediate to more siliceous.

Résumé

Les roches supracrustales de la zone magmatique orientale de Great Bear reposent en discordance sur des roches de la zone métamorpho-plutonique de Hepburn couvrant toute la longueur de la carte. Les deux groupes de roches sont plissés suivant des axes à direction nord dans la région de la zone de failles de Wopmay. Les plis orientés au nord forment un corridor central dans une configuration sigmoïdale régionale de plis et reflètent la contrainte accrue localement par suite, probablement, de l'occurence d'une ligne structurale principale. Les roches métasédimentaires du groupe de Grant reposent en discordance sur des gneiss archéens. À la marge sud de la zone représentée par la carte ainsi que sur les feuilles contiguës, le groupe de Grant ainsi que son soubassement probablement archéen se répètent tectoniquement sur une faille chevauchante. Le plutonisme qui règne à l'intérieur de la zone magmatique de Great Bear varie à la fois sur le plan de la composition et sur le plan temporel, allant d'un caractère plus intermédiaire à plus siliceux.

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INTRODUCTION

This paper summarizes field work within the Calder River (86F) map area completed during the summer of 1987. The map area encompasses parts of three tectonic zones of Wopmay orogen, from east to west: Hepburn plutonic-metamorphic zone, Wopmay fault zone, and Great Bear magmatic zone. Work done previous to the current project within the area was of a reconnaissance nature and is listed in Hildebrand et al. (1987a).

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WOPMAY FAULT ZONE

Wopmay fault zone is a 7-18-km-wide north-trending belt of lenticular brittle-fault bounded blocks that lie near the western margin of the Hepburn metamorphic-plutonic zone (Hoffman, 1972; 1984; Easton, 1981: King et al., 1983; King, 1985). Mapping in 1986 (Hildebrand et al., 1987a) showed that rocks of the zone are folded about north-south axes; that supracrustal rocks of the Great Bear magmatic zone unconformably overlie rocks of the Hepburn metamorphic-plutonic zone; and that the folding post-dates the main phase of magmatism within the Great Bear magmatic zone. Part of the southward continuation of the Wopmay fault zone was mapped during 1987.

In the Wopmay Lake area (Fig. 1) supracrustal rocks of the Great Bear magmatic zone unconformably overlie rocks of the Grant Group (Easton, 1981) and probable, but undated, Archean basement. The unconformity is well-exposed east of central Wopmay Lake (Fig. 2) where, low grade sedimentary rocks of the basal Great Bear magmatic zone, the Dumas Group (Hoffman and McGlynn, 1977), sit upon higher grade deformed psammites, carbonates, and mafic intrusions. One kilometre farther south the erosional surface cuts obliquely downward through rocks of the Grant Group and rocks of the Dumas Group lie directly on mylonitic granitoids of probable Archean age.

Directly east of central Wopmay Lake (Fig. 2), 10-15 m of quartzite, semi-pelite, and psammite of the Grant Group unconformably overlie the mylonitic granitoids. The supracrustal rocks are not mylonitized. Thus, the age of mylonitization must be older than the Grant Group: about 1.9 Ga (Bowring, 1984). The lower part of the sedimentary sequence is cut by numerous gabbroic dykes and sills. Overlying the siliciclastic sequence are 5 m of marl, containing conspicuous porphyroblasts of zoisite, which in turn is overlain by a coarsely recrystallized carbonate unit up to 10 m thick. The carbonate unit is overlain by several hundred metres of sillimanite-garnet-biotite schist (Fig. 3). The section is capped by pillowed basalt flows intruded by abundant gabbro bodies.

The structure of the Wopmay fault zone is complex but in the Calder River map area is dominated by the north-south folds. The folds, as stated earlier, are clearly younger than the main phase of magmatism within the Great Bear zone. The folds appear to form the central corridor in a regional set of folds that occur within the Great Bear zone to the west (see Hildebrand et al., 1987b) and the Hepburn metamorphic-plutonic zone to the east (see King, 1986). In both areas the folds trend northwest; however, as they approach the central corridor the orientation of their axial traces curves progressively into a northerly trend and they become much tighter. Thus, the overall pattern of the axial traces of the folds is sigmoidal and reflects, at least in the central area, dextral strike-slip movement (Fig. 4). The reason for the localization of increased shear strain within the central corridor is unknown but may relate to the contact between Hottah terrane and Hepburn metamorphic-plutonic zone buried beneath the Dumas Group within the corridor (Hildebrand et al., 1987a).

Most of Wopmay Orogen is cut by a conjugate set of transcurrent faults (Hoffman et al., 1984; Tirrul, 1984) which are younger than the folds. The north-south corridor along the Wopmay fault zone forms a domain boundary between northwest-trending, left-lateral faults to the east and northeast-trending, right-lateral faults to the west. Although the transcurrent faults are younger, only a few northwest-trending, left-lateral examples cut the corridor with most of the faults bending, splaying and dying as they approach it. Since rocks on both sides of the corridor are megascopically deformed by the faults and those within it are not, rocks of the zone must be penetratively strained. Because this deformation is superimposed over older north-south oriented fabrics related to the folding we have not been able to separate the effects of the two deformations at the grain scale. However, in many places within the corridor rocks of the Dumas Group have a well-developed moderately-plunging northerly oriented mineral lineation.

HEPBURN METAMORPHIC-PLUTONIC ZONE

In the southeast corner of the map area is a small part of an interesting structure which we found during a brief reconnaissance to continue to the southeast onto the adjoining map sheet as shown by Wilson and Lord (1942). Metamorphosed supracrustal rocks of the Grant Group unconformably overlie granitic gneisses of inferred Archean age and are in turn overlain by granitic gneisses of similar lithology to those beneath the unconformity (Fig. 5). The structurally higher slab of gneisses has been dated as 2.5 Ga (U-Pb zircon) and is, like the lower gneisses, unconformably overlain by metasedimentary rocks of the Grant Group. The Grant Group, sandwiched between the gneisses, comprises pyritic schist, orthoquartzite, volcanic cobbly conglomerate, and sillimanite-potassium feldspar gneisses and migmatite. They are cut by a variety of leucocratic granitic intrusions ranging from strongly deformed to non-deformed. Rocks of the Grant

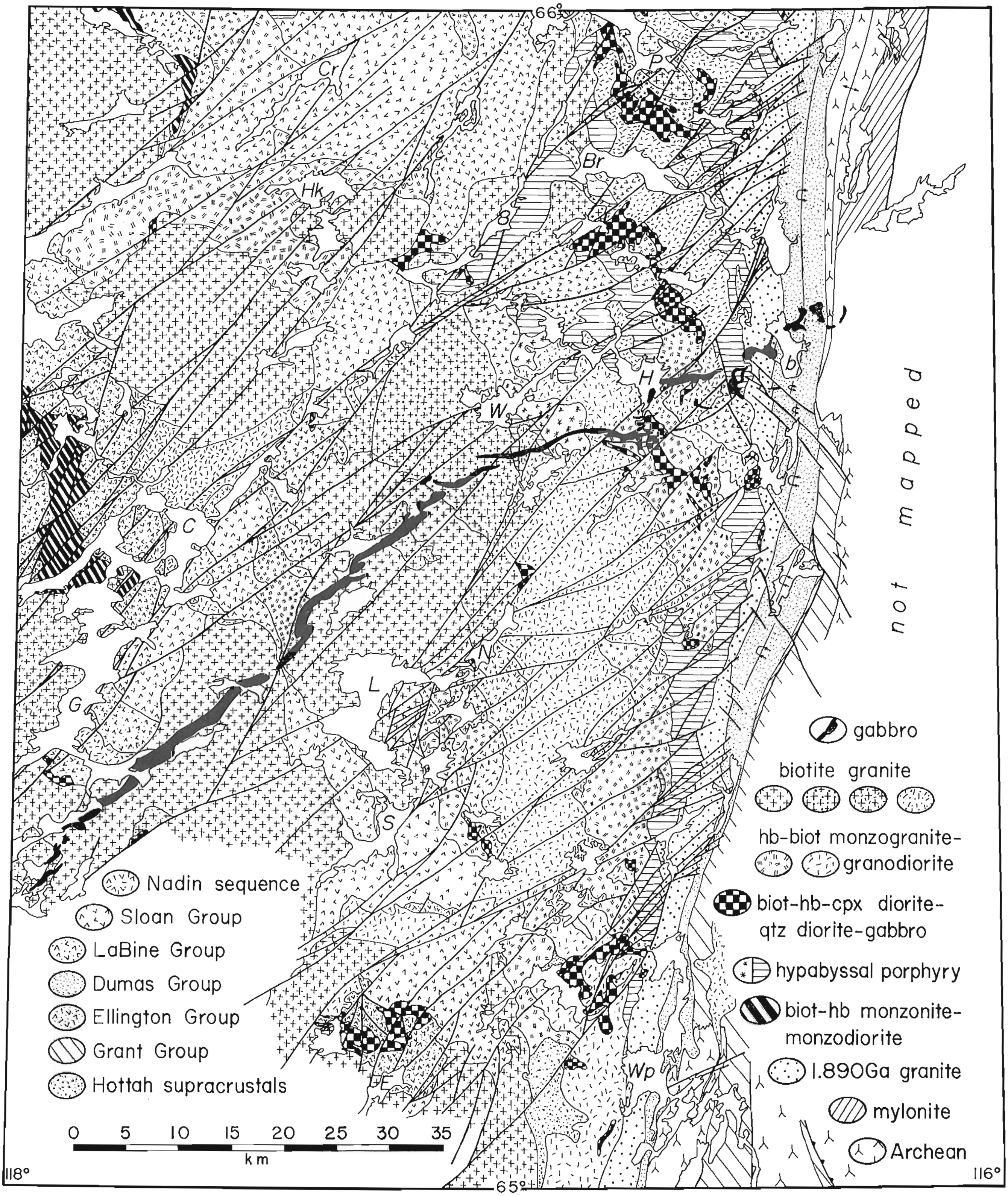

Figure 1. Generalized geological map of the Calder River map area showing distribution of major rock units mapped to date. Geology in the Ellington Lake area was simplified from Pelletier (1986). *C* = Clut Lake; *G* = Grouard Lake; *W* = Wiley Lake; *H* = Hansen Lake; *Br* = Breadner Lake; *b* = Brain Damage Lake; *Cr* = Cruickshanks Lake; *Hk* = Hooker Lake; *L* = Lever Lake; *S* = Self Lake; *N* = Nadin Lake; *E* = Ellington Lake; *Wp* = Wopmay Lake.

Group are recumbantly folded about axial planes that parallel both the unconformity and the overlying contact. Deformation, in the form of mylonitic foliation within the overlying gneisses increases downward toward the contact. On the basis of tectonic repetition and the occurrence of highly strained rocks at the contact, we interpret the contact to be a thrust fault along which granitic gneisses were transported and emplaced above rocks of the Grant Group. Thus, the Grant Group and its basement consitute a window beneath Archean granitic gneisses. This implies that major areas of the orogenic hinterland may be Archean basement rather than high-grade Grant Group: a possibility recognized by King (1986). The oval shape of the structure is probably due to the refolding of older northerly-oriented Calderian structures by the northwesterly-trending Great Bear folds.

GREAT BEAR MAGMATIC ZONE

Sloan Group

Rocks mapped as Sloan Group (Hoffman and McGlynn, 1977) are dominantly intermediate ash-flow tuffs. The largest area of rocks belonging to the Sloan group is located in the Cruickshanks Lake area (Fig. 1). There, greater than 2 km of densely-welded, lithic-rich, dacitic ash-flow tuff, containing 35-45 % broken phenocrysts and abundant round to oval-shaped cognate inclusions, is overlain by conglomerate, thinly-bedded tuff and siliclastic sedimentary rocks, rhyolitic lava flows, and rhyodacitic ash-flow tuff. The conglomerate is clast-supported and monomictic comprising dacitic cobbles and boulders in a matrix of dacitic debris. Minor sandstone beds within the conglomerate are crossbedded and also composed of dacitic material. The conglomerate unit is wedge-shaped in cross-section and thickens rapidly toward the northwest until it intersects a major strike-slip fault. It is not present northeast of the fault and presumably is displaced northeastward out of the map area. The base of the thick dacitic tuff unit is not exposed within the map area because it is intruded by a hypabyssal porphyry consisting of euhedral-subhedral phenocrysts of quartz and feldspar in a red to pink aphanitic groundmass.

Overall, the lithic and crystal-rich nature of the tuff coupled with its extreme thickness and dense welding suggest that it is intracauldron facies tuff. The composition of the overlying conglomerate indicates that portions of the tuff were uplifted and eroded shortly after cooling. Such a process occurs during resurgent doming of the central portions of calderas. Resurgence would provide the necessary monolithological source as only thick intracauldron facies tuff is typically exposed on resurgent domes. The porphyritic pluton intruding the lower part of the tuff could represent a resurgent intrusion. The diameter of the caldera is unknown because we were unable to find either the structural or topographic walls within the map area; however, as the unit was traced for about 15 km along strike, the caldera must be at least that wide.

Additional ash-flow tuff units assigned to the Sloan Group occur as a northeasterly dipping strip extending from Clut Lake southeastward to Self Lake (Fig. 1). In the Clut Lake area the tuffs clearly overlie rocks of the LaBine Group (Hildebrand et al., 1987a) and in the Self Lake area they overlie rocks termed the Ellington Group by Pelletier (1986); thus the diverse sequence of interbedded lavas, tuffs, and sedimentary rocks of the lower Ellington Group may be correlative with the LaBine Group (Hoffman and McGlynn, 1977) to the west and the Dumas Group to the east.

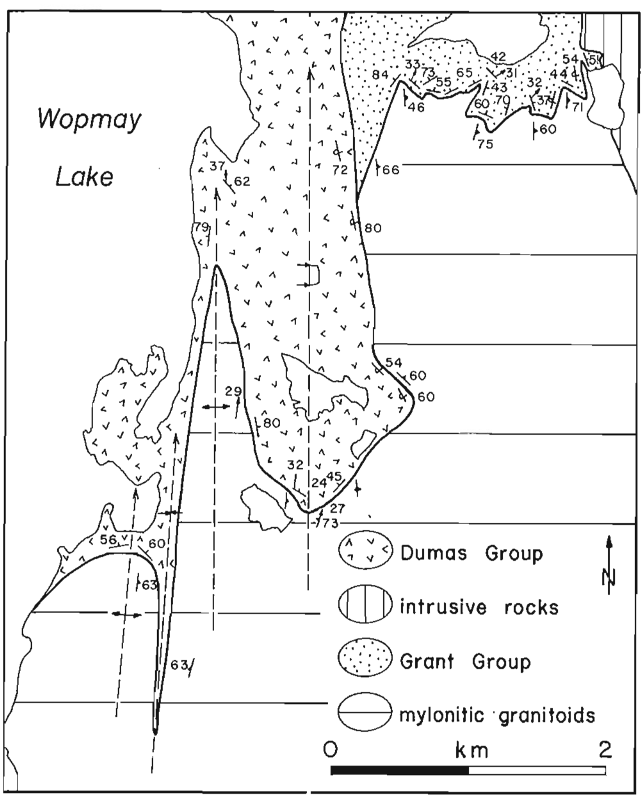

Figure 2. Geological sketch map showing northerly-plunging folds of the Wopmay fault zone and the unconformities between the various units.

Figure 3. Sillimanite-garnet-biotite schist of the Grant Group, east of Wopmay Lake. Note pen in top centre for scale. GSC photo 204401-A.

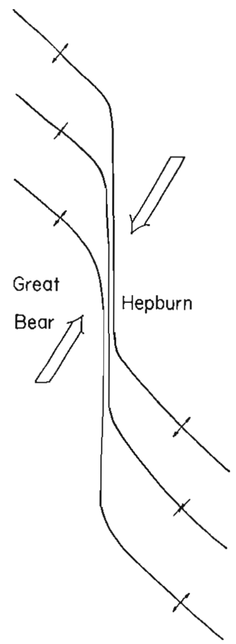

Figure 4. Model for the origin of the northerly-oriented folds in the Wopmay fault zone and the regional sigmoidal form outlined by their axial traces.

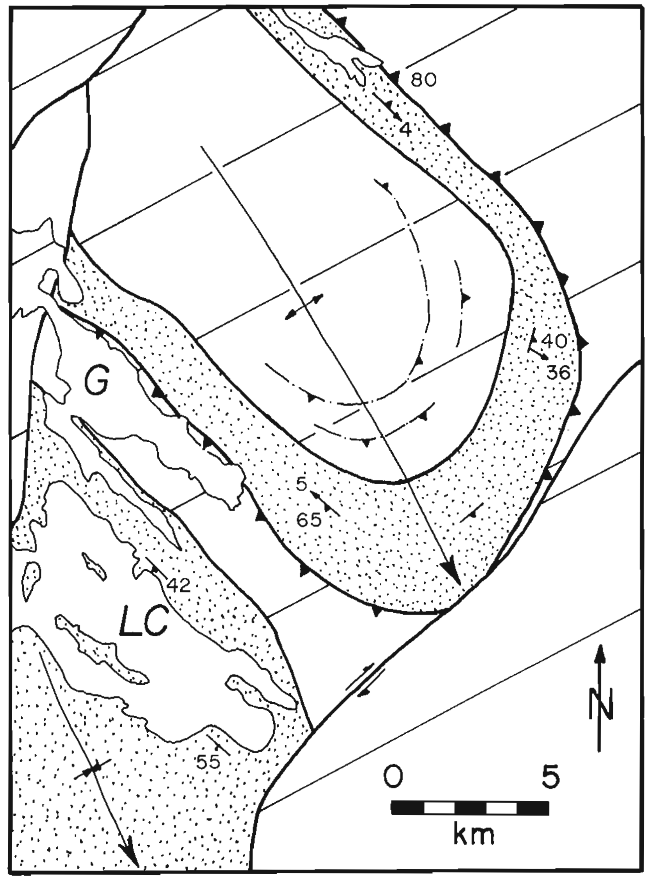

Figure 5. Geological sketch map showing window through possible basement-involved thrust. Cover sequences are stipled and basement gneisses are ruled. The dashed and dotted lines are trend lines of foliation. Geology taken in part from Easton (1981) and Wilson and Lord (1942). *G* = Grant Lake; *LC* = Little Crapeau Lake.

Dumas Group

The Dumas Group mapped during the past field season occurs as a continuation of the northward trending, overturned syncline previously mapped to the north (Hildebrand et al., 1987a). The syncline narrows southward but it is continuous over the entire length of the map area. The lithologies vary little from those described in Hildebrand et al. (1987a) except that in the extreme southern part of the sheet at Wopmay Lake there are considerable more intermediate and mafic intrusions within the group. Also, the degree of deformation is greater in the south with most of the rocks displaying a conspicuous shallow- to moderately-plunging mineral lineation.

In the northern part of the map area rocks of the Dumas Group within the syncline lie unconformably on 1.890±5 Ga (U-Pb zircon age) granite on the west and rocks of the internal zone to the east. During 1987 we mapped the southward continuation of the contacts and found the unconformities on both sides of the syncline to be well-exposed in many places. In the Wopmay Lake area the eastern limb of the syncline is overturned and low-grade interbedded sitstones and mudstones of the Dumas Group (Fig. 6) unconformably overlie sillimanite-garnet-biotite psammites of the Grant Group (Fig. 3).

Farther to the north where the major syncline of Dumas Group narrows markedly, basaltic and gabbroic pebbly conglomerate of the basal Dumas Group lie unconformably on metabasites of the Grant Group. The contact is exposed on the side of the prominant west-facing scarp. In fact, we have now examined and mapped the escarpment over the entire length of the map area and it is interpreted to be an unconformity wherever it has been observed.

The western limb of the fold is vertical to overturned nearly everywhere along strike and in several places we observed a well-developed regolith up to 5m thick developed from the underlying 1.890 Ga granite (Hildebrand et al., 1987a). The granite is visibly altered for an additional 20 m beneath the regolith.

Other supracrustal sequences

Along the southeastern side of Nadin Lake (Fig. 1) is an interesting sequence of supracrustal rocks whose stratigraphic position is unclear due mainly to its isolated location and tight folding. The sequence consists of thinly bedded to laminated mudstones and siltstones, granite bouldery conglomerate, and densely-welded rhyolitic to dacitic ash-flow tuff. The conglomerate occurs between the tuffs and sedimentary rocks. Although facing directions within the sedimentary rocks were found we were unable to determine the overall facing direction of the pile because the sedimentary rocks are tightly folded. However, the lack of any clasts of ash-flow tuff within the conglomerate suggests that the tuffs lie stratigraphically above the conglomerate.

Prefolding supersuite

Members of this suite mapped during the field season are diverse in composition and texture, ranging from equigranular granite to orbicular quartz diorite, diorite and gabbro. In most cases the mafic-intermediate rocks are older than the more siliceous rocks.

The largest pluton of the suite mapped is a variably foliated and lineated porphyritic granite that trends southeast from Wiley lake (Fig. 1). Potasssium feldspar, quartz, and plagioclase phenocrysts are surrounded by a fine-grained groundmass of feldspars, quartz, and biotite. Because it intrudes volcanic and sedimentary lithologies typical of the Great Bear magmatic zone; has contacts that trend southeast parallel to the regional folds; and is intruded by intermediate composition plutons it is considered to be a formation of the prefolding suite.

Many other plutons were mapped during the field season and they are, for the most part, typical of those described by Hildebrand et al. (1987a, b). A notable exception is the occurrence of orbicular diorite, quartz diorite, and gabbro in the east central parts of the map area (Fig. 7). The bodies are very heterogeneous and occur as remnants, or roof pendants, within younger granodiorites and monzogranites. Associated with the orbicular rocks are coarsely pyroxene porphyritic gabbros and minor pyroxenite.

Postfolding suite

Several plutons which most likely postdate folding were mapped during the field season. Two oval shaped bodies are located south and west of Hooker Lake and another occurs in the Lever Lake area (Fig. 1). The three plutons are

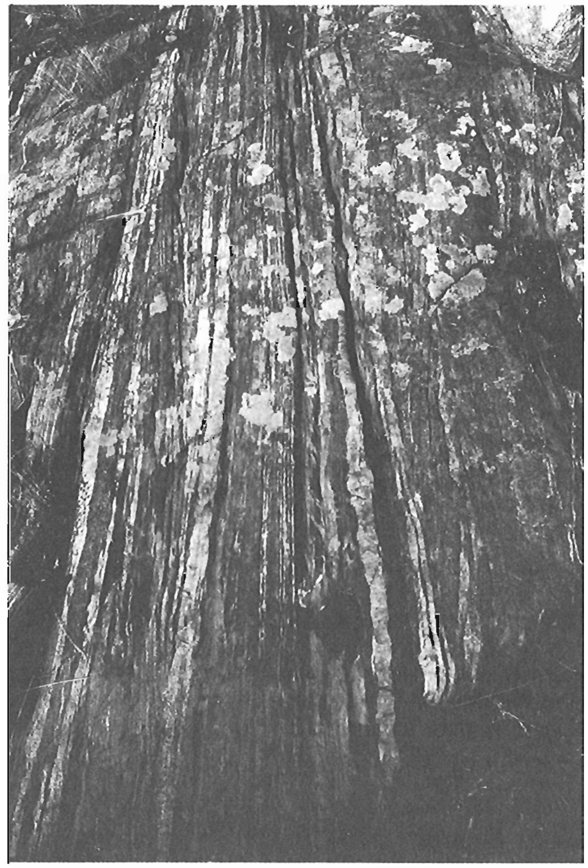

Figure 6. Interbedded volcanogenic siltstones and mudstones of the Dumas Group, east of Wopmay Lake. The rocks shown in this photo unconformably overlie those of Figure 3. Note pen for scale. GSC photo 204401.

Figure 7. Orbicular quartz diorite. GSC photo 204400-Y.

dominantly potassium feldspar porphyritic rocks with no visible fabric. Typically, quartz, potassium feldspar and plagioclase sit in a much finer groundmass of feldspars, quartz and biotite. Contacts with wall rocks are sharp and local pegmatites are common. The pluton at Lever Lake locally contains abundant miarolytic cavities and on the large peninsula that protrudes from the eastern side of the lake there is a considerable area of miarolytic granophyre attesting to its epizonal character.

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Quaternary stratigraphy of overburden drill cores, Timmins to Smoky Falls, Ontario[1]

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Smith, S.L. and Wyatt, P.H., Quaternary stratigraphy of overburden drill cores, Timmins to Smoky Falls, Ontario; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 207-216, 1988.

Abstract

The multiple till sequences of the Timmins-Noranda region have been extensively studied in mineral exploration programs. However, depositional models and associated ice flow directions of tills described largely from drill samples, have remained uncertain because of a lack of surface exposures. To clarify the Timmins area stratigraphy, which is critical for drift prospecting, stratigraphy from intact overburden cores will be correlated with the well-exposed sequences in the James Bay Lowland.

During 1987, twenty-six boreholes were drilled between Timmins and Smoky Falls. Preliminary correlations indicate that up to four tills are preserved in the area. They are tentatively interpreted as two old tills below an organic-bearing bed, possibly equivalent to the Missinaibi Formation from the Moose River basin and/or the Owl Creek beds to the south, overlain by Matheson Till which in turn is overlain by laminated sediments of glacial Lake Ojibway, Cochrane Till, and postglacial sediments.

Résumé

Les multiples séquences de till de la région de Timmins-Rouyn-Noranda ont été abondamment étudiées à l'occasion de programmes d'exploration minière. Toutefois, les modèles sédimentaires et les directions d'écoulement glaciaire qui leur sont associés et qui sont le plus souvent décrits à partir d'échantillons de forage, sont demeurés incertains en raison de l'absence d'affleurements de surface. Afin d'expliquer la stratigraphie de la zone de Timmins, qui est d'une importance fondamentale pour la prospection du drift, la stratigraphie établie à partir de la couverture intacte de dépôts meubles sera corrélée aux séquences bien exposées dans les basses-terres de la baie de James.

Au cours de l'année 1987, 26 sondages ont été forés entre Timmins et Smoky Falls. Des corrélations préliminaires indiquent que jusqu'à quatre tills sont préservés dans le secteur. Il pourrait s'agir de deux tills anciens situés au-dessous d'une couche à teneur organique, soit l'équivalent probable de la formation de Missinaibi à partir du bassin de la rivière Moose ou des couches du ruisseau Owl du côté sud, les deux tills étant recouverts par du till de Matheson, qui est à son tour recouvert par des sédiments laminés du lac glaciaire Ojibway, du till de Cochrane et de sédiments postglaciaires.

[1] Contribution to Canada-Ontario Mineral Development Agreement 1985-1990.
Project carried by Geological Survey of Canada.

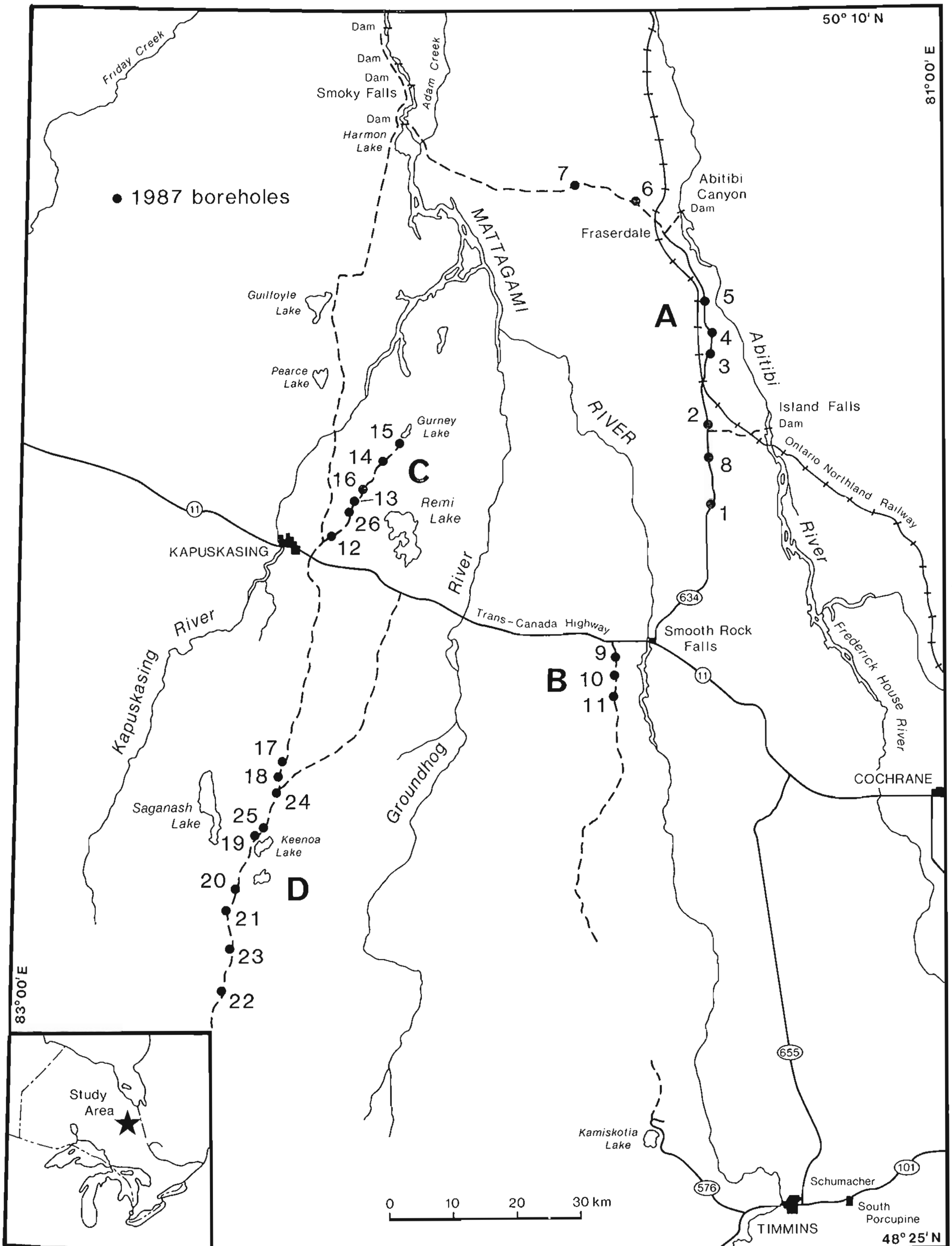

Figure 1. Map of study area and borehole locations. For convenience of drawing tie-lines, four groupings of the holes have been made as indicated: A = Smooth Rock Falls transect N; B = Smooth Rock Falls transect S; C = Kapuskasing transect N; D = Kapuskasing transect S.

INTRODUCTION

During August of 1987, twenty-six overburden boreholes were drilled in the area to the north and northwest of Timmins, Ontario (Fig. 1). The purpose of this project is to provide stratigraphic control in a large area of minimal exposure of Quaternary sediments between the James Bay Lowland and Timmins area.

This summer was the first major field season for the Timmins Stratigraphic Drilling Transect component of the Canada/Ontario Mineral Development Agreement and was geared toward reconnaissance drilling. The Rotasonic drill operated by Midwest Drilling of Winnipeg, was utilized to retrieve relatively intact overburden cores of 10-cm diameter to bedrock. Drilling sites were chosen on the basis of areal distribution, access, and both airborne and ground resistivity mapping. The results of these electromagnetic (EM) techniques were used to estimate overburden thickness. Based on the interpreted bedrock topography, deep, narrow valleys, likely to contain the most complete sequence of Quaternary sediments, were targeted preferentially over broad depressions.

The nature of drilling and retrieving core can be more restrictive than working with natural exposures in the sense that; 1) drill core represents 10 cm-wide "sections" and gives little indication of the lateral continuity or the significance of stratigraphic "units", and, 2) the quality of core depends on the cohesiveness of the sediment and its ability to retain sedimentologic structure. Obvious advantages are that the whole sequence of sediments to bedrock can be retrieved, and the Rotasonic method produces relatively intact core as compared to reverse circulation slurries or sediment "chips".

The cores have been sampled and described in detail using a descriptive lithofacies code (Eyles et al., 1983). This paper presents preliminary correlations between units in the cores based on relative stratigraphic position and field observable characteristics such as colour, percent clast content, matrix texture, and nature of contacts. Detailed study of granule-pebble-fraction lithology, heavy minerals, and geochemistry, along with development of depositional models, will allow correlations to be better constrained. Ultimately, efforts aimed at solving this problem will enhance the usefulness of drift prospecting in the Timmins mining camp.

PREVIOUS WORK

The stratigraphy of the Hudson Bay Lowland as proposed by Andrews et al. (1983), comprises up to six distinct till sheets, separated by waterlain sediments. In the Moose River Basin immediately to the north of the study area, Skinner (1973) documented five till units (Fig.2). The interglacial Missinaibi Formation, consisting of marine, fluvial, forest-peat, and lacustrine members, forms a significant marker horizon.

South of the James Bay Lowland, Hughes (1955, 1965) described, from oldest to youngest, the Matheson Till, varved clays of the Barlow-Ojibway Formation, and massive, pebbly clay, diamicton, and laminated sediments of the Cochrane Formation. Matheson till is characterized by pebble fabric orientations toward the southeast, and highly variable pebble lithologies that range from up to 32 per cent Paleozoic carbonates in the Smooth Rock Falls map area, to one per cent or less in the Iroquois Falls map area about 100 km to the southeast (Hughes, 1965).

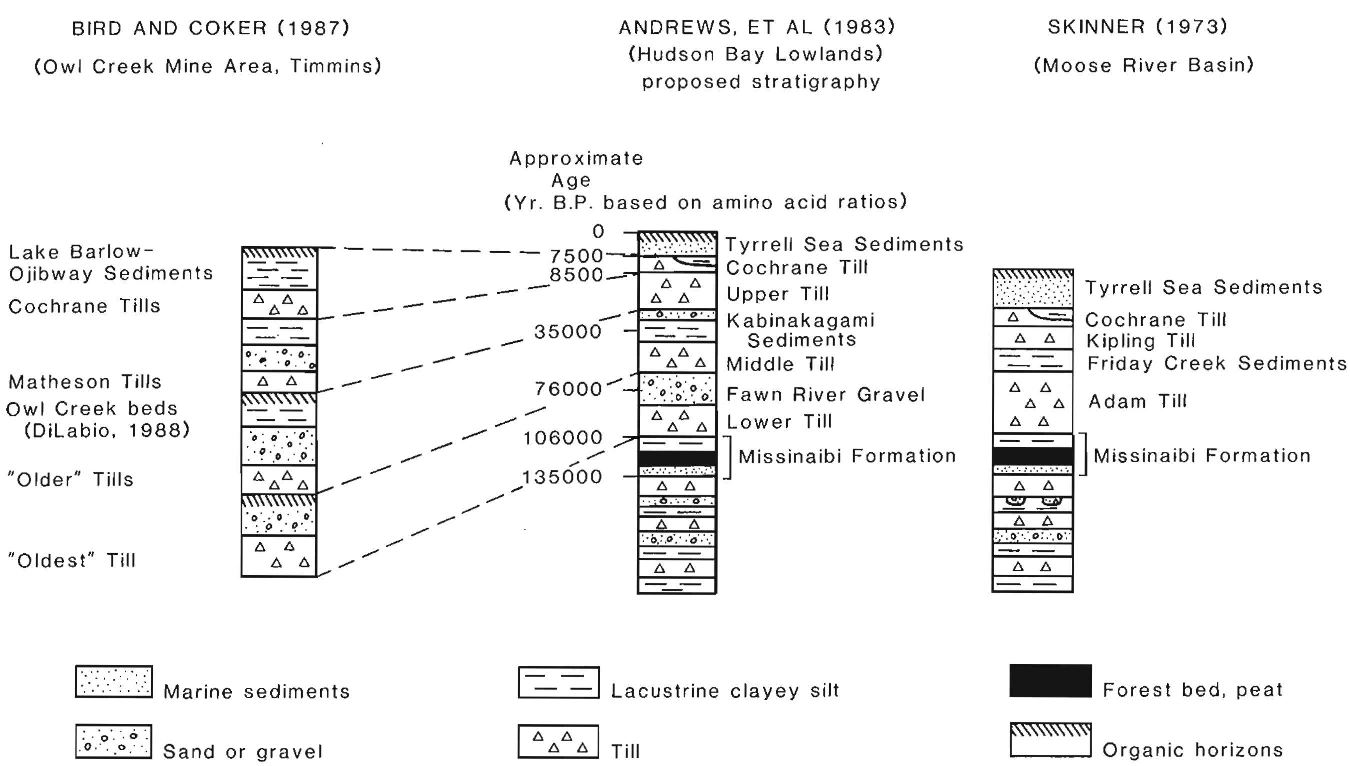

Figure 2. Stratigraphic columns and correlations.

Boissonneau (1966) in the Cochrane-Hearst area of northwestern Ontario recorded striae, eskers, and drumlinoid landforms that indicated a fan-like regional ice movement oriented toward the southwest in the western part of the area, southward in the central part, and to the southeast in the east. This glacial event deposited a sandy till which Boissonneau (1966) correlated with Matheson Till of Hughes (1955). Boissonneau (1966) also described an earlier ice movement based on striae, which moved normal to the overall trend, or roughly east-west, but which deposited no recognizable till.

The Pinard Moraine located immediately south of Smoky Falls, as defined by Boissonneau (1966), records the limit of recession of the fan-shaped regional ice mass. Varved sediments were deposited in glacial Lake Barlow-Ojibway to the south of this limit. Based on a clay till cap on esker and morainic deposits, the Barlow-Ojibway sediments were inferred to have been overridden during a late readvance of this ice mass that deposited Cochrane Till. Post-readvance sediments were deposited in shallow, short-lived lakes, which indicates very rapid retreat of the ice (Boissonneau, 1966).

Forty-two overburden drill cores from the Black River-Matheson area northeast of Timmins (Averill et al., 1986), commonly intersected only Matheson Till below laminated silt and clay. The direction of ice advance associated with the Matheson Till in this area is 165°. In a few cores, an older till is associated with striae and canoe-sized grooves from a pit exposure, striking 240° (Averill et al., 1986).

In the Timmins area, Bird and Coker (1987) suggested that three glacial advances occurred prior to the Cochrane readvance in the vicinity of the Owl Creek Mine (Fig. 2). The three Wisconsinan, pre-Cochrane tills are associated with striae from ice movements from oldest to youngest, of 240°, 150°, and 170° respectively. Organic sediments cover the two lowermost till-bearing "stratigraphic packages" (Bird and Coker, 1987). DiLabio et al. (1988) described a traceable marker unit, equivalent to the uppermost organic-bearing horizon of Bird and Coker (1987) as the Owl Creek beds. Brereton and Elson (1979) described a similar wood-bearing unit in an adjacent area.

DESCRIPTION OF SEDIMENTS

In this logging scheme, "diamicton" refers to poorly sorted sediments irrespective of genesis. The term could equally apply to both glacial and nonglacial materials. "Till is applied to sediment deposited directly from ice.

The majority of the boreholes intersected only one regionally traceable diamicton immediately above bedrock. This unit is considered a subglacially deposited till on the basis of poor sorting, moderate compaction, and the presence of far-travelled faceted and striated carbonate and other erratics. However, at nine sites, three and up to four till units were intersected. The tills are separated vertically by either waterlain sediments or by a well-defined contact.

Above the widespread till, rhythmically laminated sediments occur ubiquitously above sorted sand and pebbly sand in the area south of Kapuskasing. In most cores, the silt/clay rhythmites exhibit primary convolutions or chaotic bedding.

The uppermost sediments in most of the boreholes, typically consist of massive clay with few clasts. The clay grades upward into pebbly mud (1-2 % pebbles/granules) which is locally overlain with a sharp contact by a diamicton with a greater than 5 % clast component. In the Kapuskasing area, the sequence is capped by a thin unit of laminated silt and clay, or by sand north of Smooth Rock Falls. In some sandy units in the study area, distinctive black, Mesozoic lignite fragments are common. However, in a few cases, brown, woody and presumed younger organic fragments have also been transported as detritus in fluvial deposits.

Six graphic core logs have been chosen to illustrate not only specific stratigraphic information, but also the amount of sedimentological detail that can be obtained from this type of drilling (Fig. 3a, b).

Borehole 11 (Fig. 3a) contains two diamictons. The uppermost unit is light grey, clast-rich, and clay-rich as compared to the lower thicker, olive to dark grey, silt-rich diamicton. Borehole 26 contained three distinctive diamicton units separated by waterlain sediments. They are differentiated by stratigraphic position and field observable characteristics. Borehole 8 exhibited only one diamicton at the base capped with rhythmites and a thick sequence of pebbly clay and diamicton. The pebbly clay overlies contorted laminated sediment in all three of the cores illustrated in figure 3a. The structureless clay coarsens upward in increasing pebble content, and was labelled a diamicton in this logging scheme only if the clast content exceeded about 5 % and it had sharp contacts.

Borehole 22 (Fig. 3b) illustrates the complexity of the sediments in the cores from south of Kapuskasing. The only diamicton intersected lies on bedrock and is overlain by a thick sequence of coarse- to fine-grained waterlain sediments. Borehole 21 shows a similar sequence of sorted sediments that grade upward into finer silt and clay. Underlying the sand is a 0.5-m-thick pebbly silt horizon, which is laminated in the lower half of the unit and massive at the top, containing abundant disseminated organic fragments throughout. Wood fragments and peat chunks are visible. The unit has a peculiar dark olive-grey colour and an upper and lower erosional contact.

Borehole 19 also contains multiple diamictons as in borehole 26. A contact at 25-m-depth between superposed diamictons interpreted as different units, is defined by thin silt laminations between the upper dark grey, compact, clay-matrix, clast-rich unit, and the lower light grey, less compact, silt/sand matrix, clast-poor unit. Below the overburden, the drill intersected greenish saprolite at least 3.3 m thick (the hole was not completed to fresh bedrock), which may represent a preglacial or interglacial weathering deposit that was protected from glacial erosion. The lowermost diamicton was differentiated from the underlying similar but greenish saprolite by the presence of carbonate erratics. About 100 m to the north of borehole 19 on the same EM anomaly, borehole 25 (Fig. 4b) did not intersect the greenish diamicton and only the upper 0.5 m of the fine-grained bedrock was altered.

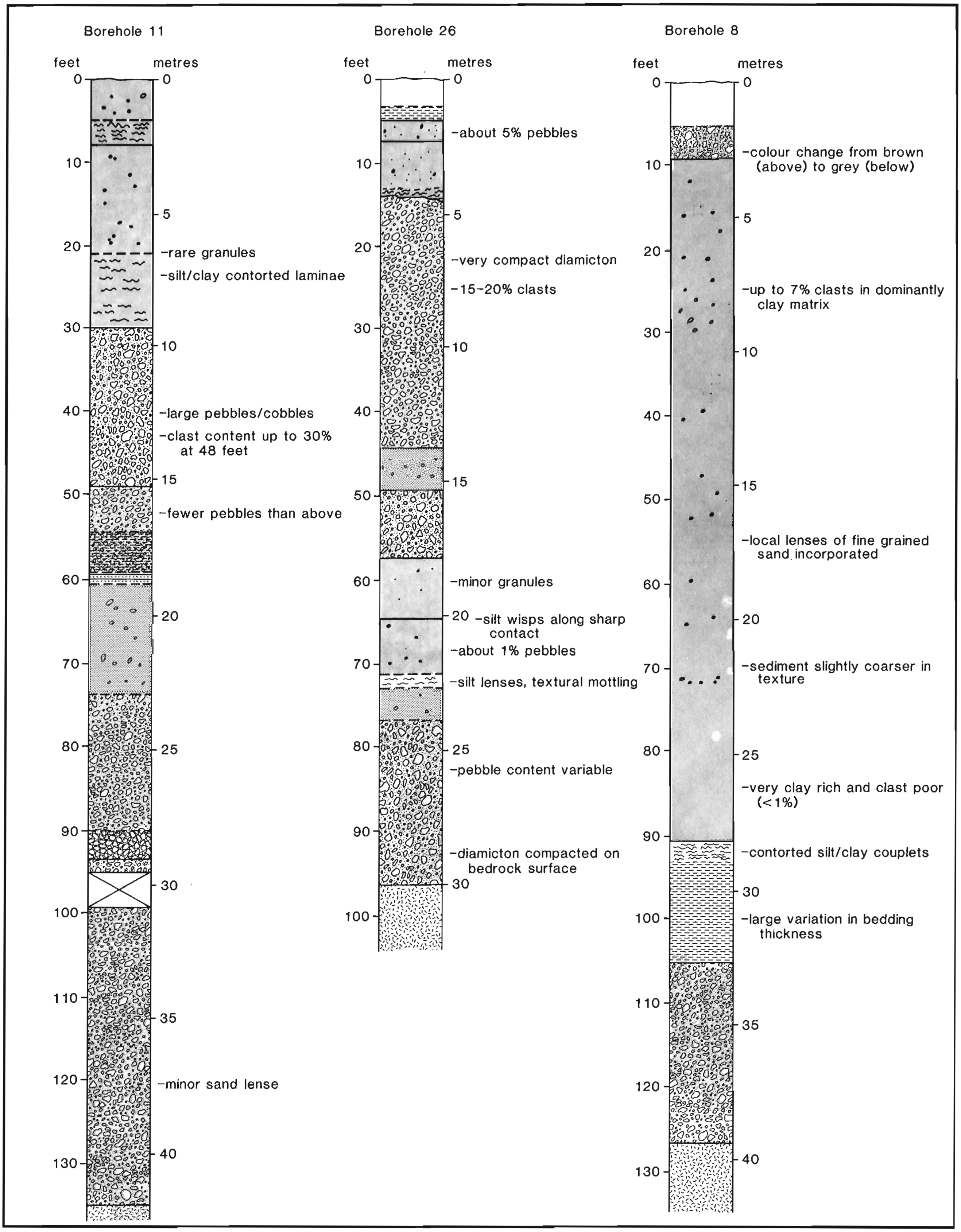

Figure 3a. Detailed graphic logs of selected boreholes. See Figure 4 for sediment legend.

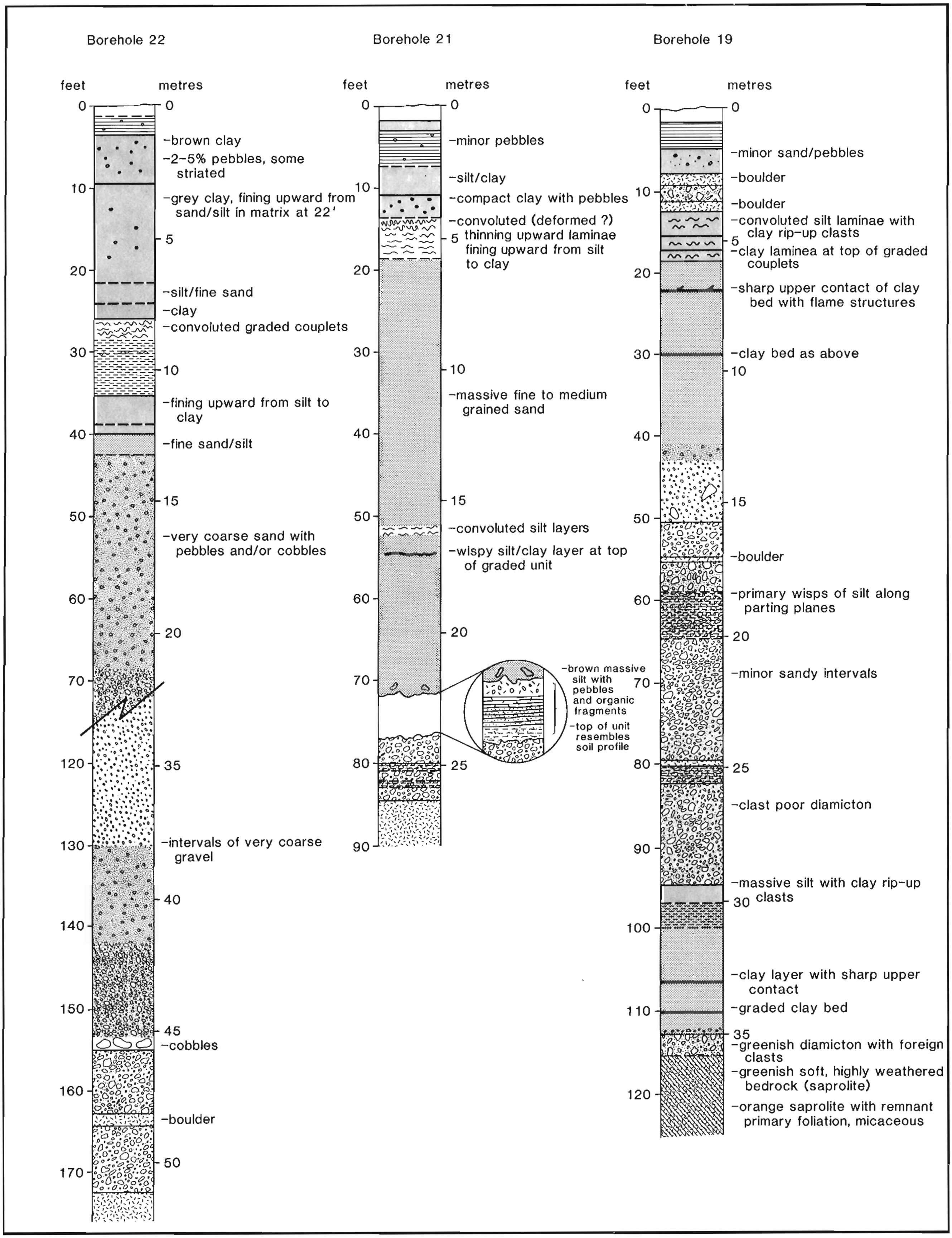

Figure 3b. Detailed graphic logs of selected boreholes. See Figure 4 for sediment legend.

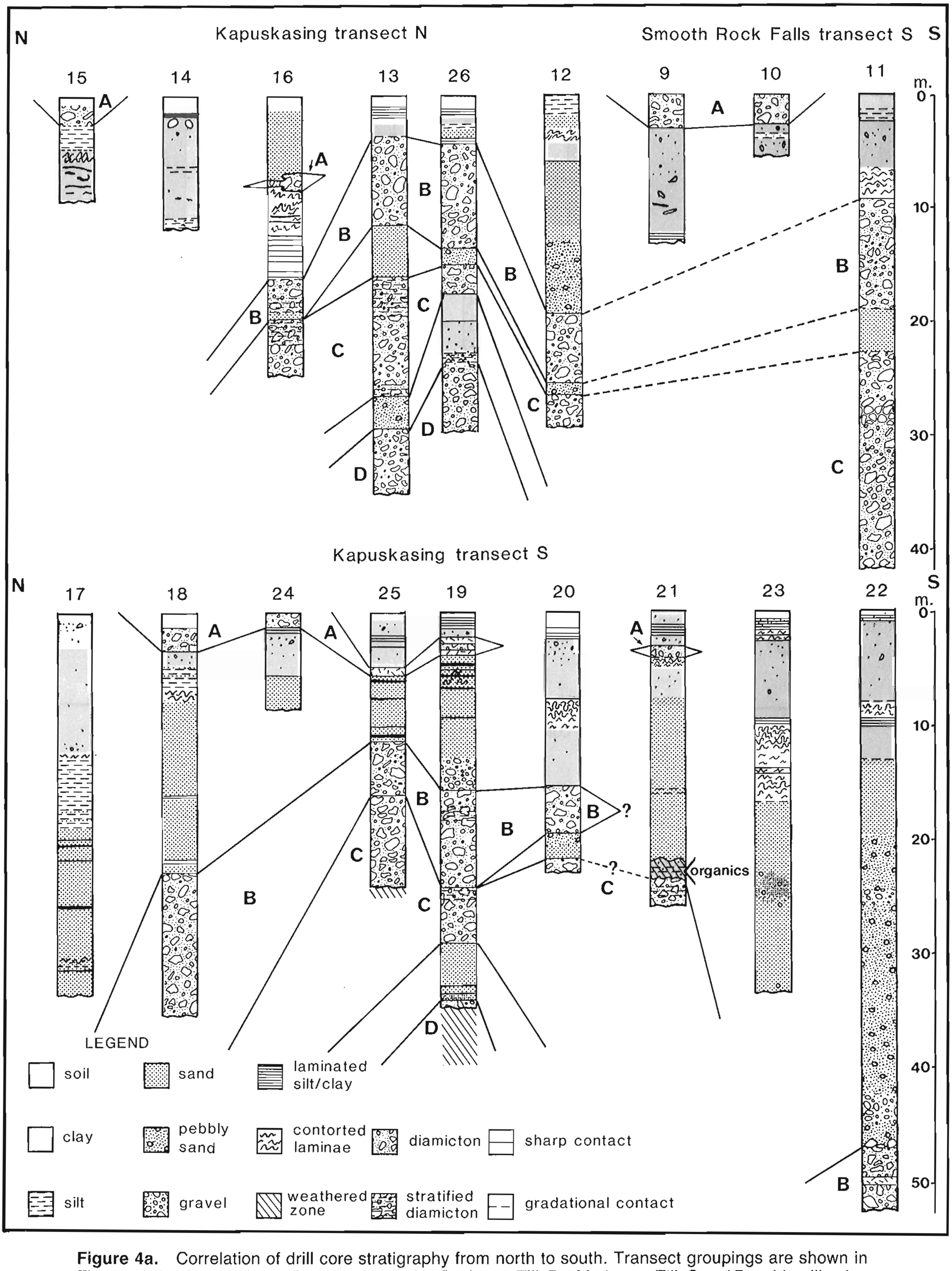

Figure 4a. Correlation of drill core stratigraphy from north to south. Transect groupings are shown in Figure 1. Diamicton units are interpreted as; A = Cochrane Till, B = Matheson Till, C and D = older till units.

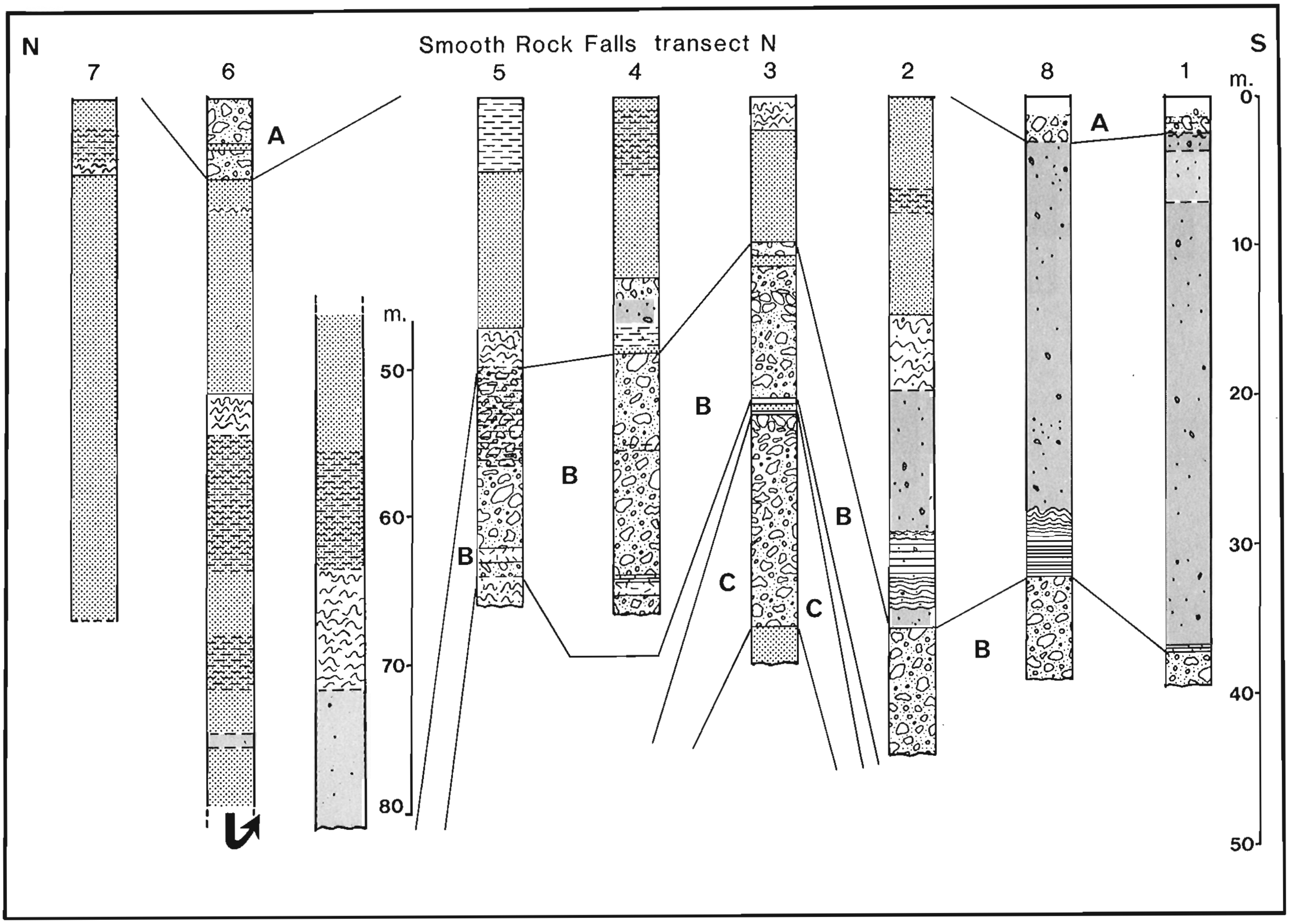

Figure 4b. Correlation of drill core stratigraphy. See Figure 4a for explanation of units.

DISCUSSION

A tentative correlation of the four diamicton units in the twenty-six boreholes was made on the basis of stratigraphic position and field observable criteria (Fig.4). The sequence apparently consists of as many as three distinct tills beneath Lake Ojibway sediments and one unit equivalent to Cochrane Till deposited during a readvance of ice into Lake Ojibway. The three lowermost till units are preserved in sequence at only two sites. It is possible that they correlate with the three pre-Cochrane tills in the Timmins area as documented by Bird and Coker (1987).

Pre-Matheson Sediments

Waterlain sediments and tills that occur stratigraphically below interpreted Matheson Till are not well represented in the cores. Pre-Matheson sediments were intersected in eleven cores of which only three exhibit both older till units. No Matheson Till was apparently preserved in borehole 21 so the lowermost till was interpreted as pre-Matheson because of the overlying organic-bearing unit. Organic-bearing, laminated to massive, pebbly clay from borehole 21, may be equivalent to the Owl Creek beds which may or may not be equivalent to the Missinaibi Formation. Paleoecological analysis of core samples will hopefully allow a better comparison to be made. No relative age has been established for these beds although DiLabio (1988) has published a recent date of >51,000 years, which could represent either a Sangamon or early-Wisconsinan non-glacial episode.

In the Moose River basin, Skinner (1973) suggested ice flow directions for his earliest tills (pre-Missinaibi) as from the northeast, which is supported by observations of Veillette (1987). If this is also true for the pre-Matheson tills, then provenance studies should show a significant deviation from pebble lithologies etc. compared to tills that were deposited regionally from the northwest.

Matheson Till

In the Timmins area, Bird and Coker (1987) report a narrow fabric range of 170° ± 5° for Matheson Till. Potentially older ice flow directions in the field area, indicated by east/west trending striae in adjacent areas (Veillette, 1987), suggest that flow patterns of the so-called Matheson Till, as well as earlier advances, may be more complex that presently understood. If the older tills, or early Matheson Till, are associated with ice flowing in an east/west orientation they should have a lithologic composition reflecting their Precambrian provenance. The name "Matheson" for the single till sheet as labelled by Boissonneau (1966) and Hughes (1965) fails to take into account these complex ice flow directions. Defined as such, Matheson Till may have a complex depositional history spanning the majority of the Wisconsin Stage. Pebble lithology studies of the core samples will help to constrain provenance areas.

Cochrane Formation

A sequence of Cochrane Formation sediments similar to those described from the type section (Hughes, 1965) occurs in several of the boreholes. Cochrane Till does not form a continuous unit across the area according to the core records. Cochrane Till (Fig. 4a, b) frequently overlies a massive silty clay with a clast content of granules and pebbles >2 %. The lower contact of the massive silty clay with the underlying varved clays appears conformable. The dropstone-rich silty clay likely represents continuous deposition in an increasingly proximal lacustrine environment.

It remains to be resolved how glacial overriding would deform varved sediments and yet preserve high relief features such as eskers.

APPLICATIONS TO DRIFT PROSPECTING

As indicated by Bird and Coker (1987), the glaciers that produced till units most useful for drift prospecting are those that had the most contact with bedrock and thus potential mineralized deposits. From this point of view, the two tills that lie below Lake Ojibway sediments would be the best candidates for further research. The Matheson and the next oldest till are reasonably widespread based on frequency of intersection in the cores. However, effective drift sampling programs depend on correct identification and regional correlation of the till units being investigated, their corresponding ice flow directions, and a knowledge of their areal extent as provided by regional drilling. Matheson Till as a stratigraphic unit will have to be better constrained.

It is doubtful that Cochrane Formation sediments will be useful in drift prospecting programs because of the complexity of the late-glacial sequence and the presence of abundant waterlain and far-travelled sediments that would be difficult to trace back to source.

SUMMARY

Sediments cored at a number of sites over such a large study area provide a unique opportunity to draw correlations in an area of minimal exposure. The twenty six overburden drill cores revealed four tills that were traceable over the study area based on field descriptive criteria. They have been tentatively correlated with the Cochrane Till as the youngest, Matheson Till, and two older unnamed units.

An organic-bearing horizon in one of the cores, is tentatively correlated with the Missinaibi Formation and/or the Owl Creek beds, of possible Sangamon or younger age, and may be the only marker stratum on which to base stratigraphic interpretation, in this area. Postdating the Matheson glaciation, a complex package of sediments represents deposition in glacial Lake Ojibway and records a possible readvance or fluctuating ice margin that deposited Cochrane Till and postglacial sediments.

Both the Matheson Till and the next oldest till, may be locally useful for drift prospecting, although ice flow directions must be better defined. Regional drilling is essential to interpreting drift sampling programs in the region blanketed by thick glacial cover in northeastern Ontario.

ACKNOWLEDGMENTS

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Stratigraphy and visible gold content of till in the Beardmore-Geraldton area, northern Ontario

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Thorleifson, L.H. and Kristjansson F.J., Stratigraphy and visible gold content of till in the Beardmore-Geraldton area, northern Ontario; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 217-221, 1988.

Abstract

The increased use of till sampling in the search for potential ore bodies obscured by glacial overburden has produced a requirement for a better understanding of the nature of glacial sediments, especially till, in the vicinity of potential gold mineralization. An extensive sheet of till is well exposed throughout the Beardmore-Geraldton area, but the extent to which the composition of this till reflects local mineralization varies greatly, particularly in response to till thickness. Visible gold in till has been encountered frequently in areas of thin till in the Beardmore area, but similar occurrences are lacking at the surface of thicker deposits near Geraldton. Drilling in this area of thick till has yielded evidence for greater local derivation at the base of thick till. A separate, more locally derived lower till unit was encountered at several sites.

Résumé

L'utilisation accrue des méthodes d'échantillonnage du till dans la recherche de gîtes minéraux potentiels dont la présence est obscurcie par des morts-terrains glaciaires, a créé le besoin de mieux comprendre la nature des sédiments glaciaires, particulièrement le till, au voisinage de minéralisations potentielles en or. Une importante couche de till affleure largement dans toute la région de Beardmore et Geraldton, mais le degré auquel la composition de ce till reflète la minéralisation locale varie grandement, particulièrement en ce qui a trait à l'épaisseur du till. On a trouvé très fréquemment de l'or visible dans le till des zones de till mince de la région de Beardmore, mais des manifestations semblables sont absentes en surface des dépôts plus épais près de Geraldton. Le forage dans cette région de till épais a fourni des indices de l'existence d'une dérivation locale plus grande à la base du till épais. Une unité plus basse de till, distincte et dérivée de matériaux plus locaux, a été trouvée à plusieurs endroits.

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Contribution to Canada-Ontario Mineral Development Agreement 1984-1989.

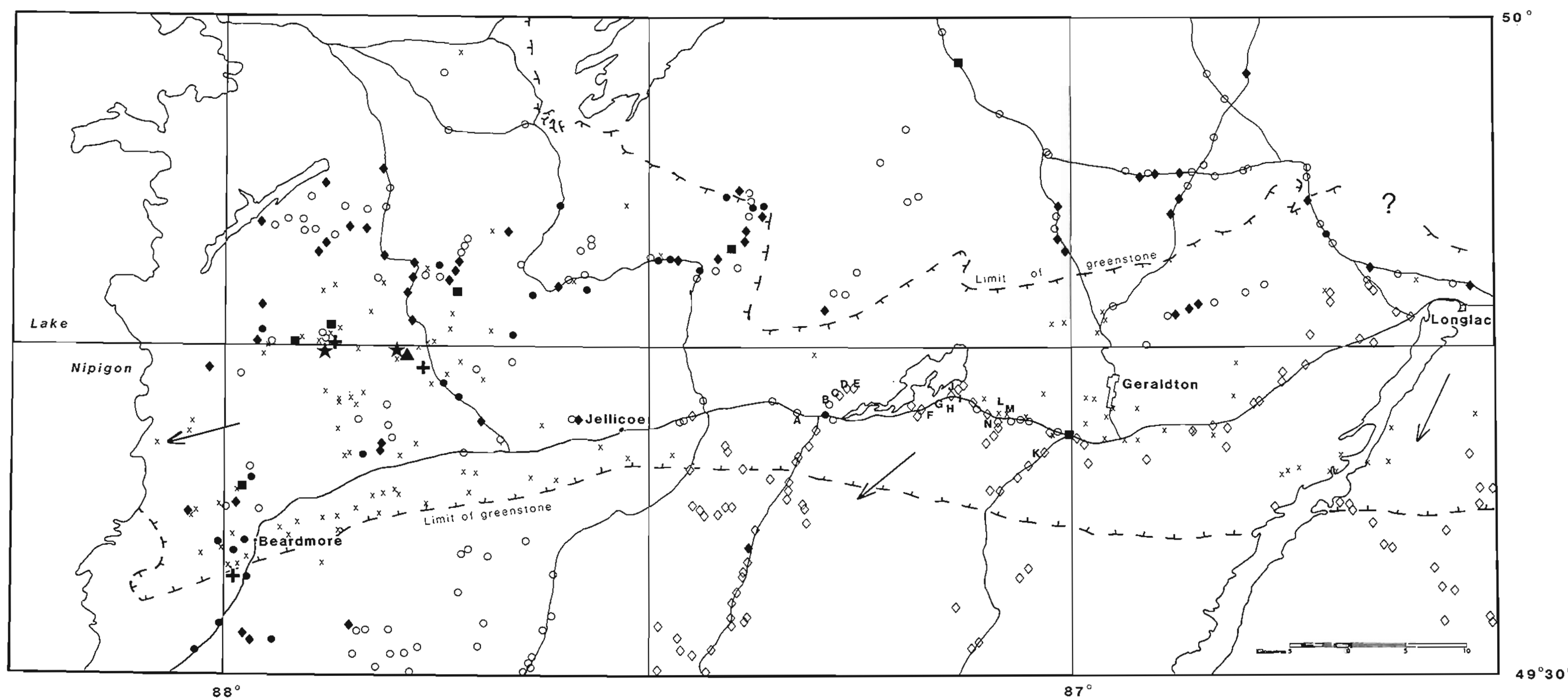

Figure 1. The distribution of visible gold grains in surface till samples. Symbols indicate in a 2 kg sample, no gold (open diamond) and, in a 10 kg sample no gold (open circle), one gold grain (closed diamond), two to four grains (closed circle), five to ten grains (square), eleven to twenty grains (triangle), 21 to 50 grains (cross) and 51 to 100 grains (star). Known sites of gold mineralization (x) are taken from Pye et al. (1966). Arrows indicate the direction of the youngest ice flow.

INTRODUCTION

The Canada-Ontario Mineral Development Agreement is a subsidiary agreement to the Economic and Regional Development Agreement signed by the governments of Ontario and Canada in November of 1984. The Beardmore-Geraldton area is one of several areas in which geological surveys meant to stimulate further growth in Ontario's mining industry have been initiated. Past gold production from this belt has been concentrated in areas of abundant outcrop. Excellent potential for further discoveries exists where glacial overburden conceals the rock. Increased recognition of the usefulness of till sampling as an exploration tool in such areas (Shilts, 1984; DiLabio and Coker, 1987) has necessitated the acquisition of a more thorough knowledge of these sediments. The glacial geological research program in the Beardmore-Geraldton study area (Fig. 1) here described has been designed to address the stratigraphy, sedimentology, composition and source of glacial sediments, primarily till, in the vicinity of known and potential gold mineralization.

PRECAMBRIAN GEOLOGY

The Precambrian geology and gold mineralization of the Beardmore-Geraldton area have recently been described by Mason and McConnell (1983) and Mason and White (1986). The Wabigoon Subprovince is subdivided in this area into two greenstone belts, the Onaman-Tashota Belt in the northwestern portion of the study area and the Beardmore-Geraldton Belt which parallels and flanks the highway linking the communities of Beardmore, Jellicoe, Geraldton and Longlac (Fig. 1). The Onaman-Tashota Belt consists predominantly of intermediate to felsic volcanics intruded by felsic plutonic bodies. The Beardmore-Geraldton Belt is a predominantly metasedimentary sequence intercalated with mafic to intermediate metavolcanic rocks. Late Precambrian diabase dykes and, in the western portion of the study area, large tabular sills intrude the older rocks. Nineteen past-producing mines in the area yielded over four million ounces of gold and 300 000 ounces of silver (Mason and White, 1986). Copper, nickel, zinc, lead, molybdenum and iron mineralization are also present (Pye et al., 1966).

QUATERNARY GEOLOGY

Prior to the present study, the Quaternary geology of the Beardmore-Geraldton area was examined by Zoltai (1965, 1967) and Sado (1975).

The bedrock surface in the area is, with the exception of the central portion of the study area, generally well exposed and displays abundant streamlined roches moutonees and whaleback forms. These features parallel the youngest set of striations, a radiating pattern ranging from about 250° at Beardmore, 230° at Geraldton, to 210° at Longlac. Rare occurrences of preserved older striations indicate a former ice flow direction of 210° at Beardmore and 190° at Geraldton.

Till forms an extensive to, in the central portion of the study area, continuous sheet which constitutes the most common surficial unit in the area. The most distinctive feature of this deposit is the abundance, particularly in areas of thick till, of granule and pebble-sized clasts of Paleozoic carbonates and Proterozoic clastic and chemical sediments derived from the Hudson Bay Lowland 150 km to the northeast. This far-travelled material has undergone surprisingly little dilution by more locally derived debris. This is attributed to a zone of vigorous ice flow emanating from Paleozoic terrane, to the high erodability of Paleozoic carbonates combined with the low erodability of Archean granites occurring between the greenstone belt and the Paleozoics, as well as to the short distance over which glacial ice flowed over greenstones prior to the deposition of till in the study area. The drumlinized surface of the till, its massive and compact structure, and the ubiquitous presence of faceted and striated clasts, including pavements, indicate that this deposit is a lodgement till deposited by actively sliding ice.

Numerous well developed eskers constitute belts of thick sand and gravel crossing the area. Extensive glaciolacustrine deposits obscuring underlying till and bedrock occur only as sand in low areas near Lake Nipigon in the western portion of the study area.

FIELD AND ANALYTICAL METHODS

The surficial geology of the southernmost three-quarters of the area depicted in Figure 1 has been mapped in detail over the summers of 1986 and 1987.

Till samples have been collected in shallow excavations dug by hand as well as with the use of a backhoe. An area of thick and continuous till in the Wildgoose Lake area, between Jellicoe and Geraldton, was sampled during September 1987 using a rotasonic overburden drill operated by Midwest Drilling of Winnipeg. Although smaller samples were taken during the earliest phase of fieldwork, a standardized procedure in which a till sample of approximately 14 kg is collected has been established. Four subsamples are taken for: 1) analysis of grain size and matrix carbonate content, 2) fine grained sediment geochemistry, 3) supplementary analyses, and 4) an archive subsample. Chemex Labs Ltd. has undertaken the preparation and analysis for precious metal and trace element geochemistry of the less than 63 μm (-230 mesh) silt plus clay fraction as well as trace element geochemistry only on the less than 2 μm clay fraction. The remaining sample of approximately 10 kg has been processed by Overburden Drilling Management Ltd. of Nepean, Ontario. This procedure includes the preparation for lithologic analysis of the gravel fraction, the separation for mineralogical and geochemical analysis of the heavy mineral concentrate with a specific gravity greater than 3.3, as well as the identification and description of visible gold grains on a shaker table and by panning. Chemical analysis of the heavy mineral concentrate is being carried out by Chemex Labs Ltd. Descriptive mineralogy of this fraction has been analyzed by Consor Mines of Hull, Quebec as well as by Overburden Drilling Management Ltd.

VISIBLE GOLD CONTENT OF SURFACE TILL SAMPLES

The occurrence of visible gold grains in till samples collected in the study area from the C horizon of soil profiles at a depth of 0.6 to 1.0 m has been reported by Thorleifson and Kristjansson (1987) and is summarized in Figure 1. Background values of one and two grains occur sporadically throughout the area, including sites north of the mineralized belt. A lack of visible gold grains in till samples collected in the southeastern portion of the study area, down-ice flow from known gold mineralization in the Geraldton area may be due to the greater thickness of till in this area. While the upper metre of till generally lacks visible gold, the lower portion of a compositionally stratified sequence would better reflect nearby mineralization. A stratification of this nature could have been produced by the gradual elimination of local debris sources as subglacial bedrock highs were buried by the accreting till. Alternatively, a completely homogenized till could yield negligible indications of local mineralization due to excessive dilution by far-travelled debris.

In contrast, visible gold grains in near-surface till samples occur frequently in the western and northwestern portions of the study area. Till is thinner and bedrock outcrops are much more frequent in this area. Glacial ice flow would in this case have had a much greater opportunity to disperse locally derived material into till now exposed at the surface. The lesser quantity of far-travelled debris in this area results in some combination of less burial of locally-derived till along with a lesser degree of dilution.

STRATIGRAPHY IN AREAS OF THICK TILL

In order to investigate the nature of the thick till sheet occurring between Jellicoe and Geraldton, 14 boreholes, labelled A to N in Figure 1, were drilled during September 1987. Thick, massive and compact silty till, rich in Paleozoic granules and pebbles, was encountered in all but one hole (Fig. 2). At five of these sites, a distinctly separate till unit of variable texture but with few Paleozoics and a high concentration of locally derived debris was encountered at the base of the sequence. These occurrences of a second till may or may not all be correlative. Futhermore, the composition of this till may be a reflection of a shorter distance of transport by the same ice flow which deposited the overlying till, or it may have been the product of an earlier, perhaps substantially different, ice flow direction. Elsewhere, Paleozoic carbonate-rich till extends to bedrock, but angular metasedimentary and metavolcanic clasts recognized at the base

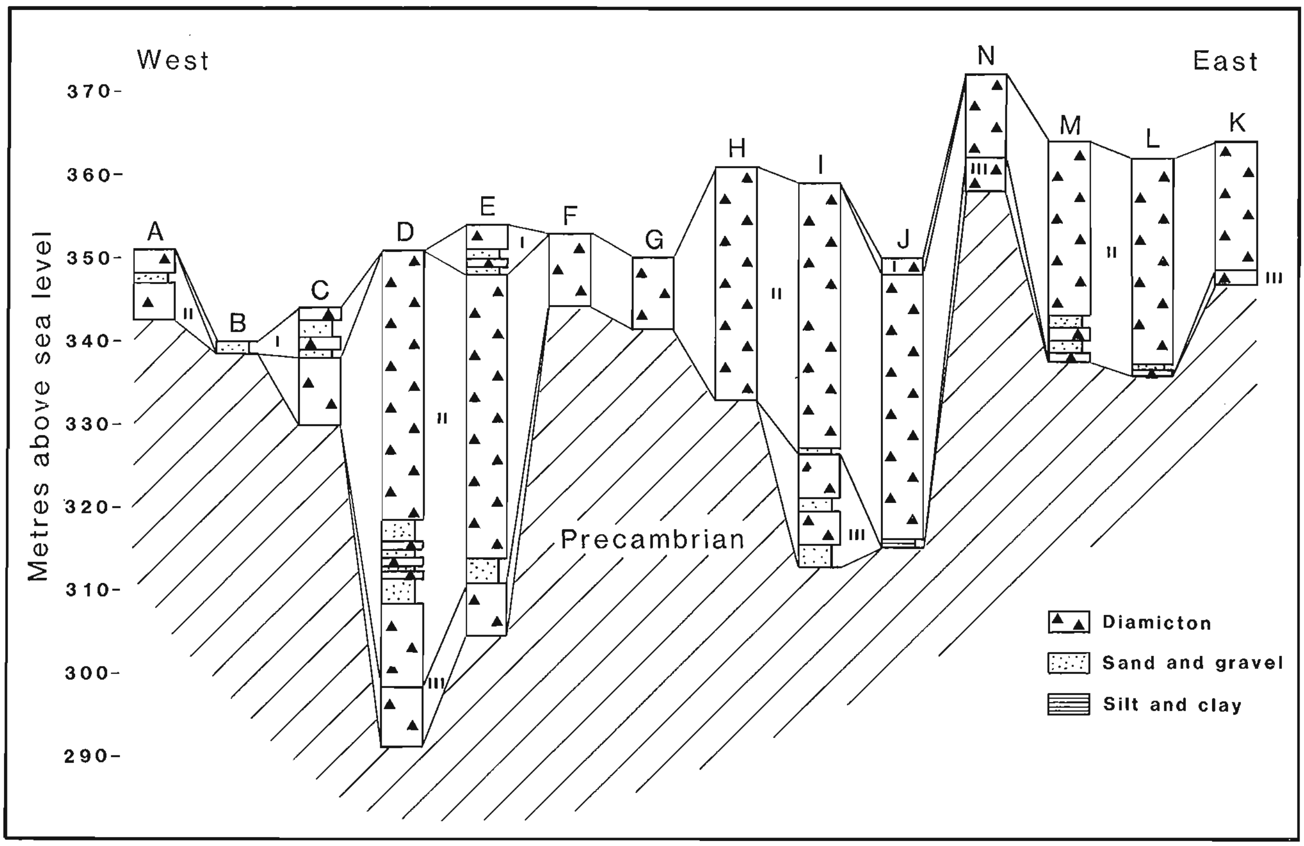

Figure 2. Stratigraphy of thick overburden from boreholes (A-N) between Jellicoe and Geraldton. A discontinuous cover of proglacial sand and sediment flows (unit I) overlies a thick and extensive silty till rich in granules and pebbles of Paleozoic carbonate derived from the Hudson Bay Lowland (unit II). At five sites, a second till unit (unit III) with few Paleozoic carbonates was encountered. Locations are given in Figure 1.

of this till may indicate a locally derived component which is potentially useful for mineral exploration. At four sites, the uppermost sediments consist of graded diamictons interpreted to have been deposited as sediment flows in a proglacial environment interbedded with sand and gravel.

SUMMARY

An extensive and well exposed till sheet in the Beardmore-Geraldton area shows promise as a useful tool in mineral exploration. In areas of thin and discontinuous till, surface samples appear to contain detectable locally derived debris indicative of gold mineralization located in an up-ice flow direction. In areas of thick till, surface samples lack a detectable local component, but drilling may yield locally derived material at the base of the sequence. The distinction between a lower till and the lower zone of a thick till rich in locally derived material is likely to be crucial in deciphering dispersal direction. Thorough analysis of glacial sedimentological features and hence the use of non-destructive drilling techniques is therefore recommended.

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Archean quartz arenite and ultramafic rocks at Beniah Lake, Slave Structural Province, N.W.T.[1]

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Covello, L., Roscoe, S.M., Donaldson, J.A., Roach, D., and Fyson, W.K., Archean quartz arenite and ultramafic rocks at Beniah Lake, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 223-232, 1988.

Abstract

Orthoquartzite and ultramafite occur at Beniah Lake (85P/8) in the Beaulieu River volcanic belt. These occurrences, along with volcanic-hosted chert-magnetite iron formations, copper-nickel prospects and other ultramafites distinguish this Archean belt from others in the Slave Province. The quartzite, ultramafite, and some metabasalt are between two northeasterly-converging shear zones. The ultramafite is altered to serpentine, tremolite, talc, dolomite, and magnetite, but cumulate texture can be recognized in places and layers of chromite up to 15 cm thick are present. Quartzite, generally white but locally green due to fuchsitic mica, displays abundant crossbeds scoures, and ripple marks. Beds strike about 120°, dip nearly vertically, and face southerly across 1.3 km, away from ultramafite exposed 1.7 km to the northeast and towards pillow lavas that also face south. Siltstone, paraconglomerate and orthoconglomerate lenses are present in the southernmost quartzite outcrops. The strata contain sills and abundant north-striking dykes of gabbro a few metres thick.

Résumé

On trouve des orthoquartzites et des roches ultrabasiques à Beniah Lake (85P/8) dans la zone volcanique de la rivière Beaulieu. La présence de ces gisements ainsi que de formations ferrifères à chert et magnétite logées dans des roches volcaniques, de zones d'intérêt de cuivre et nickel et d'autres roches ultrabasiques distinguent cette zone de l'Archéen des autres situées dans la province des Esclaves. Les quartzites, les roches ultrabasiques et certains métabasaltes se trouvent entre deux zones de cisaillement convergeant vers le nord-est. Les roches ultrabasiques ont été transformées en serpentine, trémolite, talc, dolomie et magnétite mais on peut observer une texture à cumulats à certains endroits ainsi que des couches de chromite de plus de 15 cm d'épaisseur. Le quartzite, en général blanc mais vert à certains endroits dû à la présence de mica fuchsitique, présente une stratification oblique, est décapé et est marqué par des rides de plage par endroits. Les couches, dont la direction est d'environ 120°, plongent presque verticalement et font face au sud sur une distance de 1,3 km, dans le sens opposé aux roches ultrabasiques exposées à 1,7 km vers le nord-est et en direction des laves en coussins qui font aussi face au sud. Des lentilles de microgrès, de paraconglomérats et d'orthoconglomérats se manifestent dans les affleurements de quartzite situés à l'extrême sud. Des filons-couches et de très nombreux filons intrusifs de gabbro de quelques mètres d'épaisseur et à direction nord envahissent les roches.

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[1] Contribution to Canada-Northwest Territories Mineral Development Agreement 1987-1991. Project carried by the Geological Survey of Canada.
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INTRODUCTION

A discovery in 1986 and 1987 of orthoquartzite and ultramafic rock at Beniah Lake (85P/8), is noteworthy as a new example of an enigmatic lithological association not previously known in the Slave Structural Province.

As reviewed by Eriksson and Donaldson (1986), chemically and tecturally mature quartz arenites, rudites, lutites and carbonate rocks are more common in Archean terranes than had been realized until recently. Some, such as the 3.5-3.1 Ga Beit Bridge strata in the Limpopo Province and the 2.9-2.6 Ga, Witwatersrand Supergroup in South Africa, must have accumulated on extensive supracratonic platforms during prolonged periods of crustal stability. Their depositional and subsequent tectonic histories are entirely comparable to those of many Proterozoic and Phanerozoic successions. The histories of some Archean platformal-type successions, however, are more difficult to interpret. The Woodburn Lake Group, 80 km north of Baker Lake, N.W.T. (Ashton, 1982), strata at North Spirit Lake in western Ontario (Donaldson and Ojakangas, 1977), quartzite containing uraniferous, pyritic oligomict conglomerate at Sakami Lake in Quebec (Roscoe and Donaldson, 1987) and the quartzite at Beniah Lake exemplify the problems. These are cases where the 'stable' platformal-type strata are nowhere seen to nonconformably overlie basement rocks, where the strata are closely associated with Archean volcanic rocks and turbiditic sediments deposited in tectonically unstable environments, and where the two types of successions have been intensely deformed and metamorphosed together.

GEOLOGICAL SETTING

The Beniah Lake sector of the Beaulieu River volcanic belt of Archean age is 145 km northeast of Yellowknife, N.W.T. (Fig. 1). The belt extends 70 km north (Miller et al., 1951) and 80 km south of the study area along the east flank of a structural high that contains a gneissic terrane, the Sleepy Dragon Complex. This terrane is evidently older than, and basement to, the steeply-dipping metavolcanic rocks of the Beauliu River belt (Davidson, 1972; Henderson, 1985; Lambert and van Staal, 1987). It is likely the source for some of the detritus in metasedimentary rocks in the belt. To the south and west, the Beaulieu River belt is contiguous with the Cameron River volcanic belt that extends 90 km north along the west flank of the structural high. The volcanic rocks are overlain by a greywacke-mudstone succession. These turbidites underlie extensive areas south of the Cameron-Beaulieu volcanic belts and westward 50 km to the Yellowknife volcanic belt. Volcanic belts that can be regarded as segments of the Beaulieu River belt extend from Beniah Lake at least 40 km northeast into the MacKay Lake area (Miller et al., 1951) and 70 km east into the Camsell Lake area (Henderson, 1944).

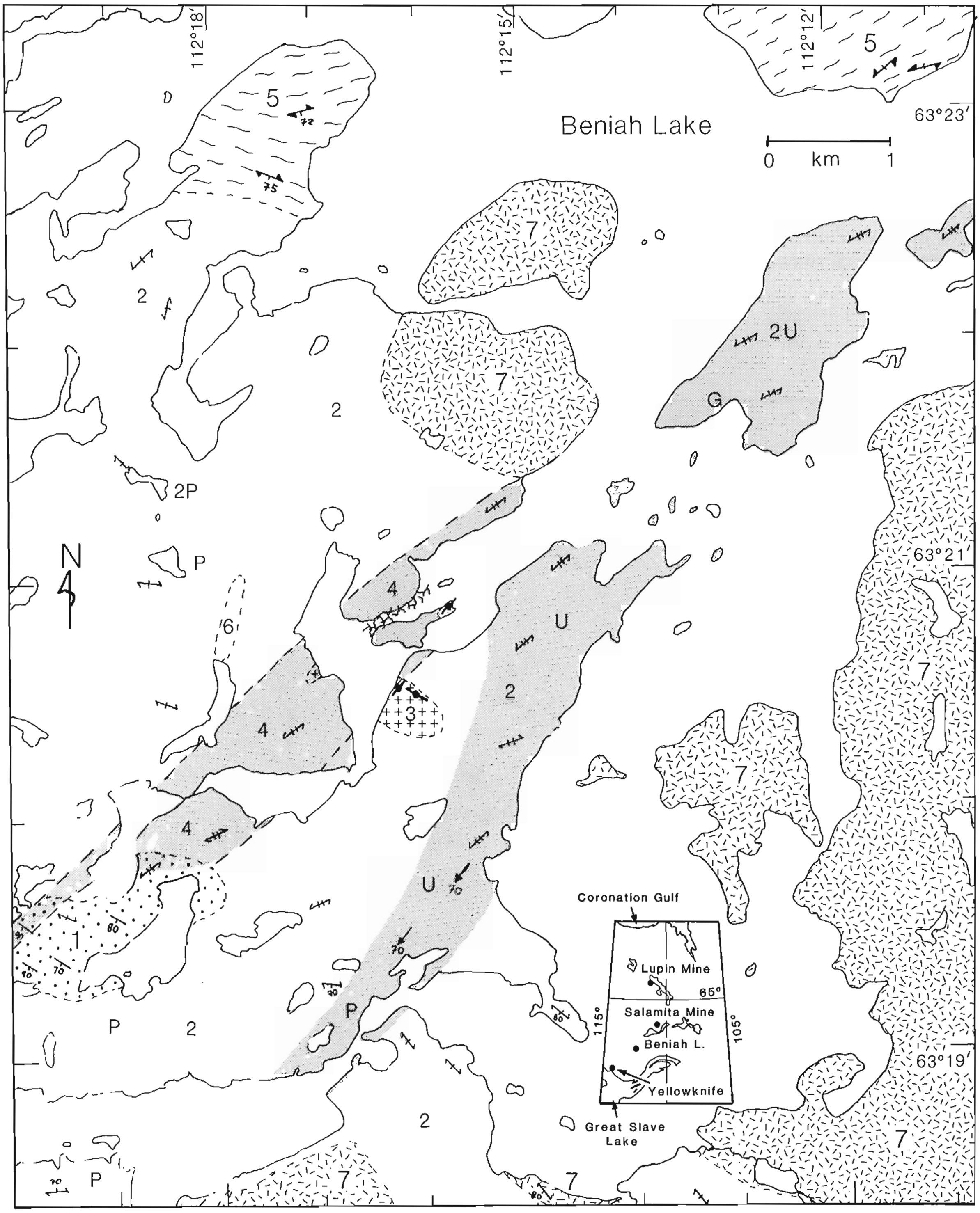
Several lithologic and metallogenic features of the Beaulieu River belt are rare or absent in other Archean volcanic belts within the Slave Province, although they are common in belts of the Superior Province and in Archean domains of the Churchill Province. Ultramafic rocks have been mapped recently by Lambert and van Staal (1987) along a sheared north-striking basement-metavolcanic contact at Amacher Lake 50 km south of the occurrence at Beniah Lake. They also may be present in an anorthosite-bearing intrusion in the Camsell Lake sector where copper-nickel concentrations have been explored. Another copper-nickel mineral occurrence has been reported 40 km north of Beniah Lake. Extensive beds of chert-magnetite iron formation are present within volcanic strata at Amacher Lake and Beniah Lake. It is interesting that polymictic orthoconglomerate has been mapped at Amacher Lake in view of occurrences reported herein of polymictic paraconglomerate associated with oligomictic orthoconglomerate and very well sorted quartz arenites at Beniah Lake. As in many other Slave Province volcanic belts, the Beaulieu River belt contains many concentrations of gold in quartz veins and in sulphide-rich zones within iron formations. Numerous volcanogenic 'massive' sulphide deposits containing subeconomic concentrations of zinc, lead, copper, silver and gold are also present.

Rocks near and on both sides of the sheared contact between the Sleepy Dragon Complex and supracrustal rocks are intruded by abundant thin, north-striking gabbro dykes that predate the shearing and regional amphibolite grade metamorphism. Several generations of folds and foliations have been recognized in different parts of the belt (Henderson, 1985; Lambert and van Staal, 1987). Lineations commonly plunge steeply but those in the northerly-striking regional shear zone at Amacher Lake plunge shallowly. Numerous plutons of foliated to massive granitiod rocks have intruded the belt, and along with older rocks are cut by unmetamorphosed Proterozoic diabase dykes.

GENERAL GEOLOGY OF PART OF BENIAH LAKE AREA

Gneissic rocks at Beniah Lake (Fig. 1) are similar to the basement gneisses of the Sleepy Dragon Complex as described by Davidson (1972). They are mainly quartzofelspathic migmatites with thin (< 1 cm) segregations of fine grained, well-foliated hornblende. A pronounced lineation of quartz aggregates is visible on foliation surfaces. Locally, the gneisses are blastomylonitic. Gneissic laminations display interfolial isoclinal folds (Fig. 3) as well as large, north-trending open folds. The gneiss and variably schistose mafic rocks to the south are intruded by ubiquitous, metamorphosed, but little deformed, metagabbro dykes and sills.

A map unit consisting mainly of fine-grained mafic rock, presumably metabasalt, contains numerous deformed pillows. The relationship of this mafite to gneissic rocks bordering it to the north is obscured by drift cover. Relationships revealed by mapping to the south (Lambert and van Staal, 1987), however, suggest that the mafite is likely younger than the gneiss which was likely basement to the supracrustal rocks. A major belt of shearing up to 3 km wide transects the mafite from southwest to northeast a few kilometres south of the gneiss. Quartzite and ultramafite occur within relatively little-deformed blocks between anastomosing shear zones that converge towards the northeast. A small lens of chert-magnetite iron formation is responsible for a very strong northerly-trending aeromagnetic anomaly several hundred metres north of the intensely sheared rocks. The ultramafic



1 - Metaquartzite
2 - Mafite (P-pillows, G-gabbro, U-undivided)
3 - Metadunite
4 - Interbanded felsite and mafite
5 - Hornblendic quartzo-feldspathic gneiss
6 - Banded iron formation
7 - Granite
Shear zone deformation

Bedding
S1
S2
Gneissosity
60 Stretch lineation
Quartz stockwork
Chromite bands

Figure 1. Geology of part of the Beniah Lake area (85P/8)

rocks within the shear zone belt are also distinctly magnetic. Quartzite beds face south away from the ultramafite. Pillows in lava south of the quartzite unit (Fig. 2), but north of the southern shear zone, also face south.

Rocks least affected by the northeast-striking shearing show an earlier, east to southeast-trending, steep foliation subparallel to the general strikes of bedding. Rocks in the shear zones have a well-foliated , steeply-dipping, locally mylonitic fabric. Lineations plunge 70° southeast. The mylonitic fabric displays Z-shaped vertical folds with north-northeast-trending axial surfaces. The quartzite block with beds striking about 120° appears to have been rotated clockwise within the belt of major shearing. This shearing predates post-dyke shearing in the regional, northerly-striking major shear zone west of the map area.

Metamorphosed mafic dykes, similar to those described by Lambert and Ernst (1987) and Lambert and van Staal (1987), cut the gneiss, supracrustal rocks, and sheared rocks (Fig. 4). Most are 0.5 to 10 m thick and commonly show chilled margins, some of which are foliated. Sills are recognized within quartzites but could not be distiguished within mafites. All rocks except Proterozoic diabase dykes are intruded by plutons of generally undeformed granitic rock.

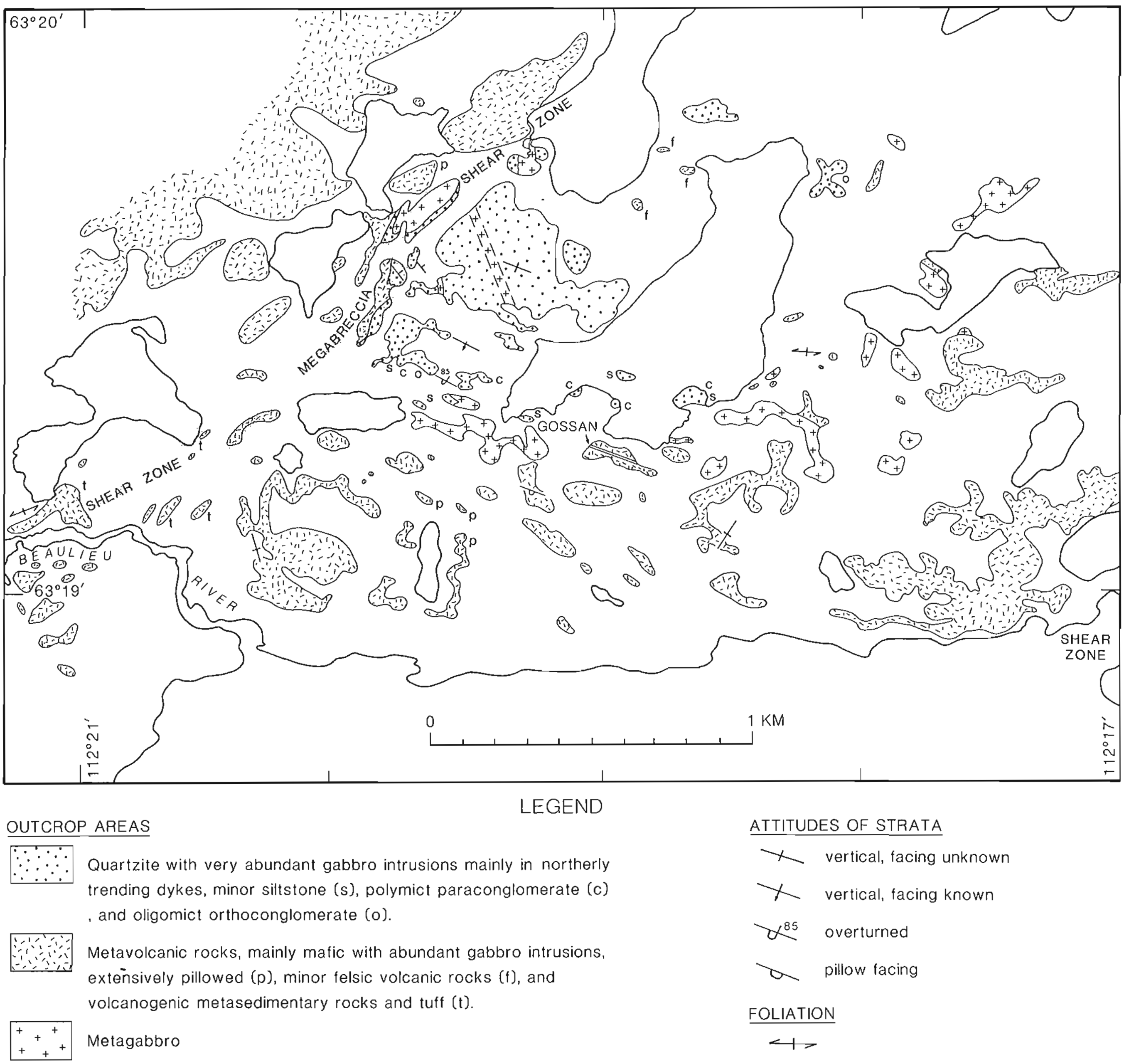


Figure 2. Geological map of outcrops in area of orthoquartzite, Beniah Lake area

ULTRAMAFIC ROCK

Little-deformed ultramafic rocks outcrop in an area about 400 by 500 m at the south side of a bay near the south end of Beniah Lake (63°20.4'N, 112°16.2'W, Fig. 1). More deformed and altered ultramafite is recognizable on the large island in the bay, and magnetic surveys with a Conimag instrument show that the ultramafite extends some distance northeast of the island and is probably also present in the bay and on the mainland southwest of the island.

The ultramafic rocks have been altered to aggregates of serpentine, tremolite, talc, dolomite and magnetite, but well-preserved primary structures and textures and apparent ranges of primary composition can be discerned in many outcrops. Some occurrences of metadunite exhibit polygonal fracture patterns ('elephant hide structure') common in ultramafic rocks (Fig. 5) and well-preserved pseudomorphs of olivine (Fig. 6). Some outcrops exhibit primary igneous layering including, most spectacularly, layers up to 15 cm thick of cumulate chromite grains (Fig. 7). The layering dips steeply and strikes about 120° . Veins of fibrous amphibole and flexible chrysotile asbestos are present.

Attenuated and folded layers of chromitite (Fig. 8) and banded talcose rock mark extensions of the ultramafite into zones of intense shearing but elsewhere protoliths of the interbanded fissile to mylonitic felsic and mafic rocks (unit 4, Fig. 1) in the shear zone are uncertain. Along both shores of the channel north of the large island, the sheared rocks are invaded by veins, stockworks and replacements of quartz (Fig. 9).

Chromite grains in some but not all chromitite occurrences have altered rims displayed by a distinct difference in reflectance (Fig. 10). Table 1 gives microprobe analyes of cores and rims of a zoned and an unzoned grain. The zoned grain (DR 14-87) shows a pronounced decrease in chrome and increase in ferric iron in its altered rim; the unzoned grain (DR 100-87) shows little difference in compositions of its rim compared to its core. Table 2 summarizes analytical data on a sheared ultramafite containing a chromitite layer (DR-87-132), sheared ultramafites lacking visible chromite (DR-87-134a and 135a), and sheared silicified ultramafite (DR-87-133) containing minor concentrations of sulphides. The latter contains the highest concentration of

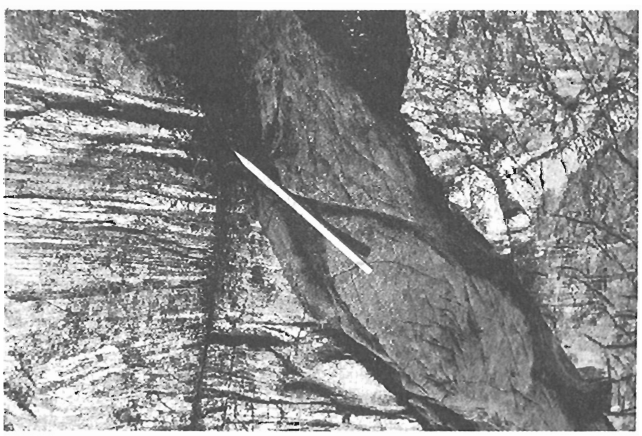


Figure 3A. Metagabbro dyke cutting faulted fold in blastomylonitic migmatitic gneiss.

Figure 4. Sheared silicic rock intruded by metagabbro dyke.

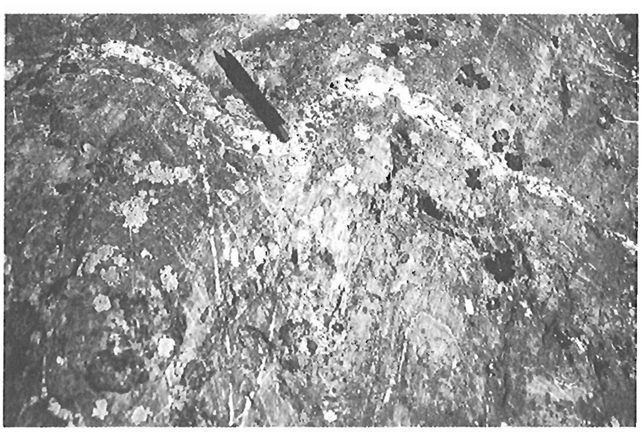


Figure 3B. Folded granite dyke in contorted mylonite (pencil points north).

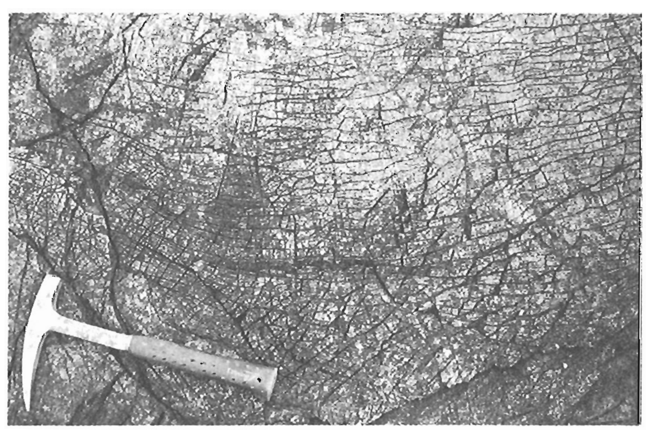


Figure 5. Polygonally fractured metadunite.

platinum group elements (PGE) and has distinctly anomalous palladium content probably related to hydrothermal transport attendant upon the movement of silica. Sulphides are present in some chromitite layers. Table 3 presents results of neutron activation analyses and fire assay/DC plasma analyses for 36 elements. in ultramafic rock containing sulphides as well as thin chromitite layers. No significant enrichments of PGE or other elements were detected.

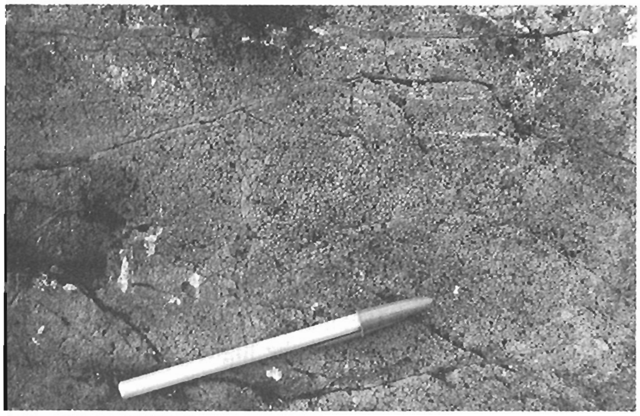


Figure 6. Olivine psuedomorphed by serpentine in metadunite.

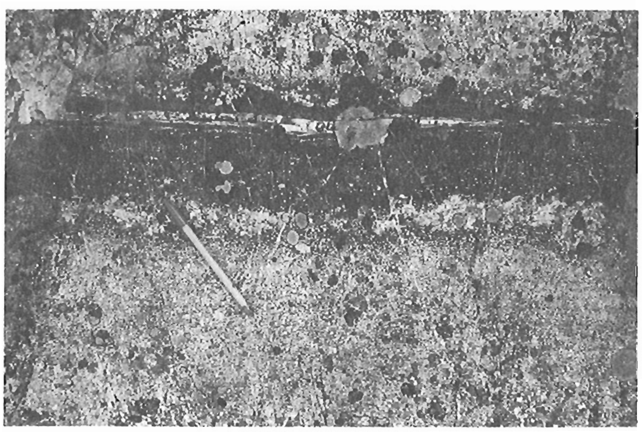


Figure 7. Chromitite layer 15 cm thick in metadunite.

Figure 8. Chromitite layer attenuated and folded in sheared ultramafite.

Table 1. Microprobe analyses of cores and rims of an altered and an unaltered chromite grain

	ALTERED DF14-87		UNALTERED DR100-87	
	CORE	RIM	CORE	RIM
TiO_2	0.26	0.48	0.25	0.25
Al_2O_3	18.12	1.21	15.44	15.36
Cr_2O_3	44.03	39.74	46.24	46.11
Fe_2O_3	4.80	26.84	4.05	4.39
FeO	25.26	28.58	31.91	32.00
MgO	4.61	1.04	0.69	0.62
MnO	1.32	1.71	1.58	1.58
ZnO	1.52	0.4	0.3	0.28
TOTAL	99.95	100.04	100.46	100.39
Cr/Cr+A1	0.62	0.96	0.67	0.67
Mg/Mg+Fe	0.25	6.25	3.71	3.34
$Fe3/A1+Cr+Fe_3$	6.04	0.38	5.27	5.46

Table 2. Analyses of sheared and altered ultramafic rock

SAMPLE	DR-87-132	DR-87-133	DR-87-134a	Dr-87-135a
SiO_2 %		61.3	46.8	46.6
$A1_2O_3$ %		3.2	3.2	3.2
Fe_2O_3 %		17.7	11.1	10.0
Mn O %		0.58	0.13	0.12
Mg O %		10.7	36.9	36.3
K_2O %		0.051	0.004	0.002
Na_2O %		0.266	0.003	0.003
TOTALS		93.2	98.1	96.2
Cr %	11.8	0.67	0.39	0.51
Pt ppb	<20	60	<20	40
Pd ppd	<10	230	10	10
Ni ppm		2600	1500	1540
Co ppm		540	82	94
Cu ppm		52	8	25

DR-87-132 sheared ultramafite with chromite layer
DR-87-133 sheared silicified ultramafite
DR-87-134a undeformed ultramafite
DR-87-135a undeformed ultramafite

Analyses by Terramin Research Labs Ltd., Calgary, Alberta

Table 3. Minor elements in an ultramafite specimen (87RF 38) containing sulphides and chromite layers.

Na % 0.44	Cd ppm <5	Eu ppm <1
Fe % 14.0	W ppm 2	Sm ppm 0.91
Cr % 2.18	Sc ppm 20.9	Tb ppm <0.5
Ni ppm 2770	Br ppm <2	Yb ppm <2
Co ppm 320	Rb ppm <5	Lu ppm 0.3
Zn ppm 530	Cs ppm <0.5	Th ppm 0.7
As ppm 5.4	Ba ppm 86	U ppm <0.2
Sb ppm 1.2	Zr ppm <200	Ag ppm <2
Se ppm <2	La ppm 3	Au ppb 9
Te ppm <10	Ce ppm <5	Ir ppb <50
Sn ppm <100	Hf ppm <1	Pt ppb <15
Mo ppm <1	Ta ppm <0.5	Pd ppb 5

Analyses by Bondar-Clegg & Company Ltd. Ottawa, Canada
by neutron activation except Pd and Pt by fire assay and D.C. plasma

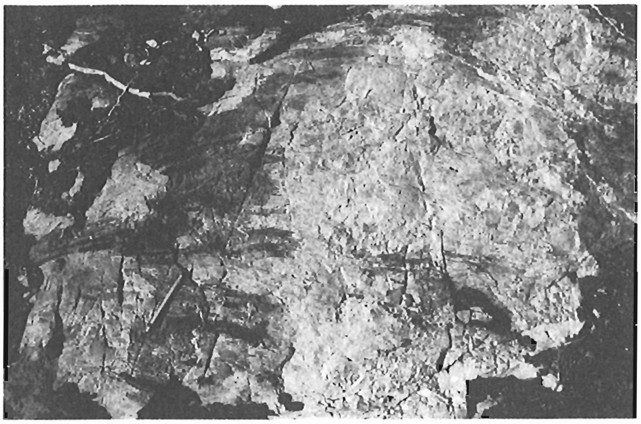


Figure 9. Quartz replacement of highly sheared rock in vicinity of ultramafite.

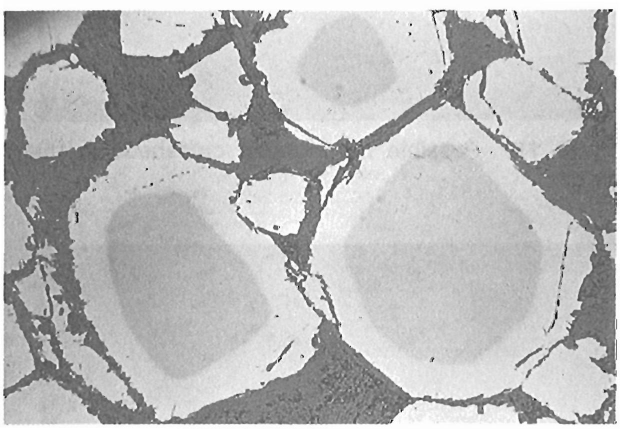


Figure 10. Zoned chromite grains in chromitite layer.

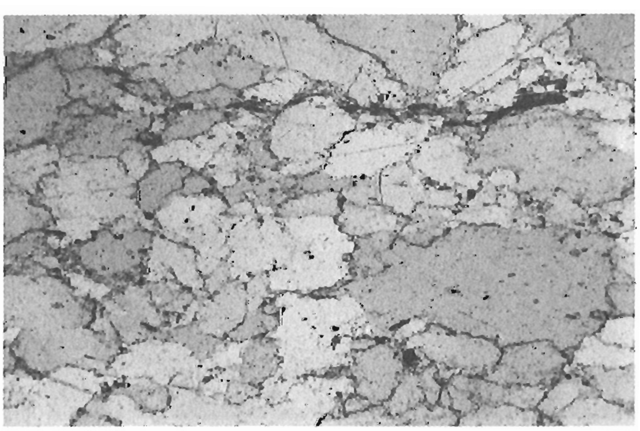


Figure 11. Photomicrograph of orthoquartzite; largest grain is 1 mm long. Nicols partially crossed

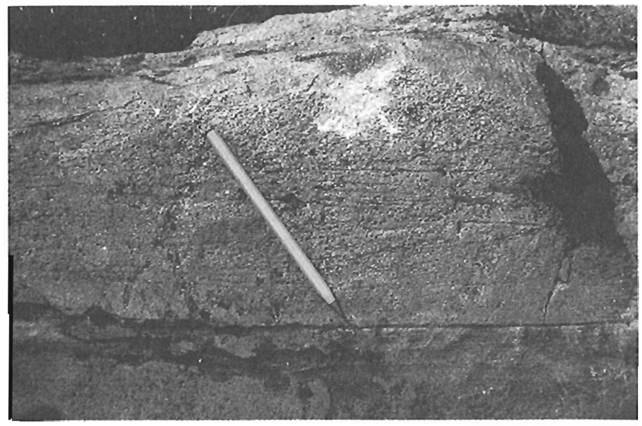


Figure 12. Crossbedded quartzite.

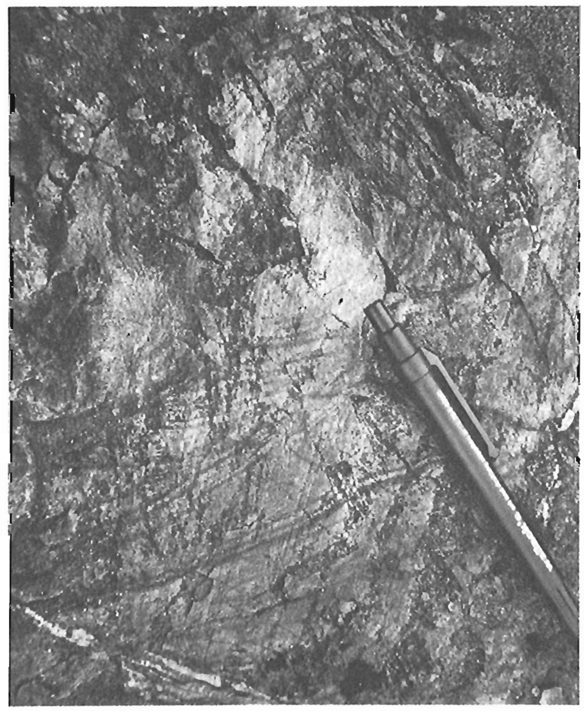


Figure 13. Small-scale channel crossbeds in fine-grained quartzite.

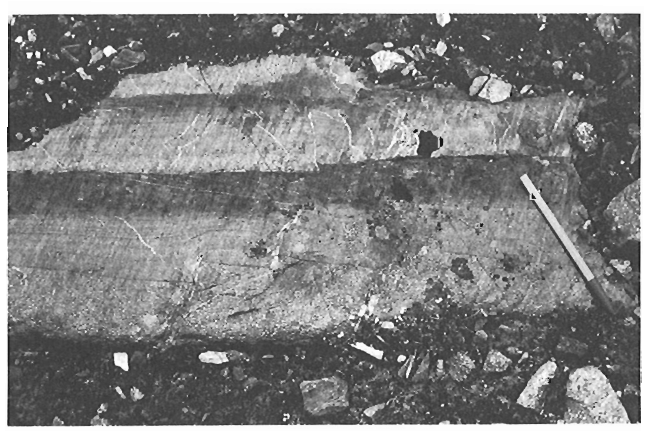


Figure 14. Waning-current graded quartz arenite-quartz siltstone beds.

ORTHOQUARTZITE, SILTSTONE, AND CONGLOMERATE

Outcrops and boulders of white, pale grey, pale cream, and green orthoquartzite (Fig. 11), together with equally abundant gabbro, occur in an area centred at 63°19.5′N and 112°19.0′W (Fig. 2), 1.7 km southwest of the main outcrop area of ultramafic rocks. Virtually the only constituents other than quartz in most samples are a few percent of muscovite and smaller amounts of minute grains of amphibole, pyrite, zircon and other heavy minerals which, in some cases, outline original positions of detrital grains of quartz and bedding lamination. Primary structures are well displayed in many places. These include crossbedding (Fig. 12, 13), scours, waning-current graded beds (Fig. 14), ripple marks and reactivation surfaces. Herringbone crossbed patterns possibly indicative of tidal current reversals were observed in one place (Fig. 15). Beds are 20 to 80 cm thick and are parallel to layering in the ultramafite. They strike about 120°, dip steeply south and top south away from the outcrop area of ultramafic rocks. Siliceous siltstone, polymictic paraconglomerate (Fig. 16), pebbly grit (Fig. 17), and a lens about 70 cm thick of clast-supported quartz pebble conglomerate are present in southernmost, (stratigraphically uppermost) exposures of the unit. Pebbles are stretched as much as 4:1 in a near vertical direction. Quartz grit, pale-coloured siltstone, and black siltstone are present locally in northern outcrops. The cross strike exposed width of the unit is about 1300 m. There is no evidence of stratigraphic repetition, but abundant sills as well as northerly-striking dykes of gabbro are present so the stratigraphic section represented may be somewhat less than 1000 m.

Parts of the quartz pebble conglomerate bed are rusty, suggesting that it may originally have been pyritiferous. There is no evidence for this, however, and the bed shows no radioactivity significently above the low backgound levels that prevail at orthoquartzite outcrops. Siltstones (black siltstone in particular) are distinctly radioactive. The most radioactive rocks are poorly sorted quartz grit and pyrite-bearing conglomerate with matrix-supported clasts. Two 1 cm bands of slightly radioactive black and pale grey cherty rocks occur at the western side of the map area, 150 m north of Beaulieu River at the south end of a small lake. The radioactive bands are in a laminated felsic-mafic rock (Fig. 21). The original rock, possibly mylonite, and its relationships to the quartzite and to the nearby mafic volcanic rocks is uncertain.

Neutron activation analyses of 9 samples of rocks within and near the quartzite area (Table 4) show that the impure slightly pyritic and radioactive conglomerate (87RF-83) contains more uranium, and a higher ratio of concentration of uranium relative to thorium, than other samples. It is the only sample in the group that contains a geochemically significent gold. enrichment. These concentrations and that of chromium might be related to slight enrichments in heavy minerals.

The quartzite-bearing block is truncated on the northwest by a zone containing megabreccia (Fig. 20) and sheared rock which separates it from mafic lavas. The position and character of other borders of the quartzite block are unclear due to paucity of outcrops and abundant gabbro intrusions.

Figure 15. Possible herringbone crossbeds in quartz arenite.

Figure 16. Polymict paraconglomerate.

A small, apparently irregular-shaped, diatreme intrudes the quartzite and gabbro near the centre of the northwest border of the quartzite block. It contains abundant angular to rounded clasts of unfoliated red granite and amphibolite clasts set in a fine grained mafic matrix (Fig. 19).

Figure 18. Quartz pebble conglomerate with well-rounded, well-sorted, tightly packed pebbles (lens cap is 5 cm in diameter).

Table 4. Minor elements in orthoquartzite and associated rocks

		SAMPLE NUMBERS 87RF SERIES								
		82	78	73	81	72	83	75	76	71
	D.L.									
	%									
Fe	0,5			3.1	1.5	3.2	3.5		2.3	8.7
Cr	0.005		220	130	200	280	220			200
	ppm									
Co	10			13			11			32
Ni	50									
Zn	200									
Cd	10									
As	1		37	14	6	18	14	5		12
Sb	0.2	0.2	0.3		0.5	0.6	0.4	1.2	0,8	0.6
Se	10									
Sc	0.5		2.3	10	16	20	10	1.7	2.2	32
Rb	10		44	62	120	150	20	88	40	24
Mo	2		3		4	3				
W	2			4	4					
Ba	100		210	340	480	510		300	170	
La	5			12	43	46	9	7	31	8
Yb	5							6	6	
Eu	2									
Ta	1			2	2		1	1	2	
Hf			2	7	8	7	5	5	9	4
Th	0.5	1.8	14	18	22	23	13	18	30	1.4
U	0.5		1,7	3.4	4.3	4,7	8.4	4.5	4.9	
Ag	5									
	ppb									
Au	5						30			
Ir	100									

D.L. — detection limit; 86RF 82 — white orthoquartzite; 78 — greem quartzite; 73 — laminated quartzite; 81 — pale siltstone; 72 — dark grey argillite; 83 — paraconglomerate; 75 and 76, respectively — dark and light lamellae in sheared rock near Beaulieu River; 71- mafite at north edge of qurtzite area.

Neutron activation analyses provided by Bondar-Clegg% Company Ltd., Ottawa

Figure 19. Diatreme containing red granite clasts.

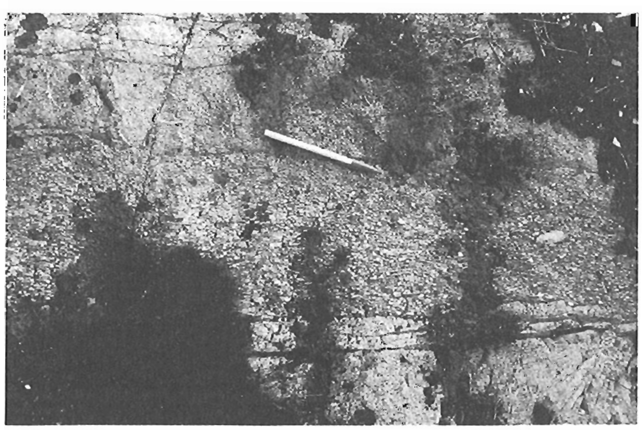


Figure 17. Graded grit- microconglomerate.

Figure 20. Megabreccia of quartzite and amphibolite blocks along fault contact between quartzite and metavolcanic rocks.

Figure 21. Mylonite with a singleradioactive lamella under centre of guage on scintillometer.

DISCUSSION

Ultramafic rocks are present at the base of the Beaulieu River volcanic belt at Amacher Lake 50 km south of their occurrence at Beniah Lake, so it is reasonable to postulate that the Beniah Lake occurrence is in an allocthonous block of basal Beaulieu River rocks that contains, in stratigraphically ascending order, ultramafite, supermature quartz arenite, and basalt.

The association of the ultramafite with quartzite, and the excellent igneous layering within it, suggests an analogy with anorogenic layered mafic-ultramafic intrusions rather than with komatiite flows, alpine-type serpentinites, or ophiolites. The relationships are compatable with deposition of the quartzite and intrusion of the ultramafite under stable tectonic conditions that were likely to be widespread and long-lived. There is no direct evidence, however, that this occurence and other isolated occurrences of Archean quartz arenites are remnants of regionally extant supracratonic successions or that there was an important time gap between their deposition and the deposition of associated strata under unstable conditions. Perhaps the processes of rock weathering and development of supermature sediments were more rapid in Archean time than subsequently. Perhaps Archean quartz arenites were preferentially preserved where they were deposited on metastable areas of crust that foundered soon after the accumulation of the platformal sediments.

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The Grenville Province boundary in the Burnt Lake area, Central Mineral Belt of Labrador[1]

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Abstract

The felsic volcanic and sedimentary rocks of the Upper Aillik Group and post-tectonic granites in the Burnt Lake area of Labrador have variously been included within and outside of the Grenville Tectonic Province. Some authors have even suggested that the Grenville Front passes through the Burnt Lake area (and consequently through several radioactive occurrences in the region). This paper presents new structural and Rb-Sr isotopic data which indicate that the Burnt Lake area has been strongly affected by the Grenvillian Orogeny and as such belongs within the Grenville Province.

Résumé

Dans des études antérieures, les roches volcaniques et sédimentaires felsiques du groupe supérieur d'Aillik et les granites post-tectoniques de la région du lac Burnt dans le Labrador ont été, selon le cas, inclus ou exclus de la province tectonique de Grenville. Selon certains auteurs, le front de Grenville traverserait la région du lac Burnt (et par conséquent traverserait les gisements de roches radioactives de cette région). Dans le présent exposé, de nouvelles données structurales et isotopiques (Rb-Sr) indiquent que la région du lac Burnt a été très touchée par l'orogénèse de Grenville et que, de ce fait, elle fait partie de la province de Grenville.

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INTRODUCTION

The Burnt Lake area of the Central Mineral Belt of Labrador has been the focus of numerous detailed studies (e.g. Kontak, 1980; White and Martin, 1980; and MacKenzie and Wilton, 1987) aimed primarily at the geochemical and petrographic aspects of the uranium mineralization in the area. The recognition of a precious and base metal mineralization association (Wilton et al., 1987) enhances the economic potential of the area.

The main objective of this report will be to discuss the location of the Grenville Front in the Burnt Lake area with respect to new field and isotopic data gathered in the past year. The lithogeochemical relations in the Burnt Lake — Emben area have been discussed in an earlier paper (MacKenzie and Wilton, 1987), and MacKenzie will expand these studies as a MSc. thesis at MUN.

The lithologies of the Burnt Lake area are predominantly felsic volcanic and sedimentary rocks of the Aphebian Upper Aillik Group which have been intruded by leucocratic granite (Fig. 1). Zircon separates from rhyolites of the group near the Michelin Deposit have been dated by U/Pb isotope systematics at 1855 Ma (Scharer et al., in press). As defined by Gower and Ryan (1986), the area is within the Makkovik Structural Province of the Canadian Shield and was involved in the Makkovikian-Ketilidian orogenies, ca. 1810-1790 Ma (Wardle et al., 1986).

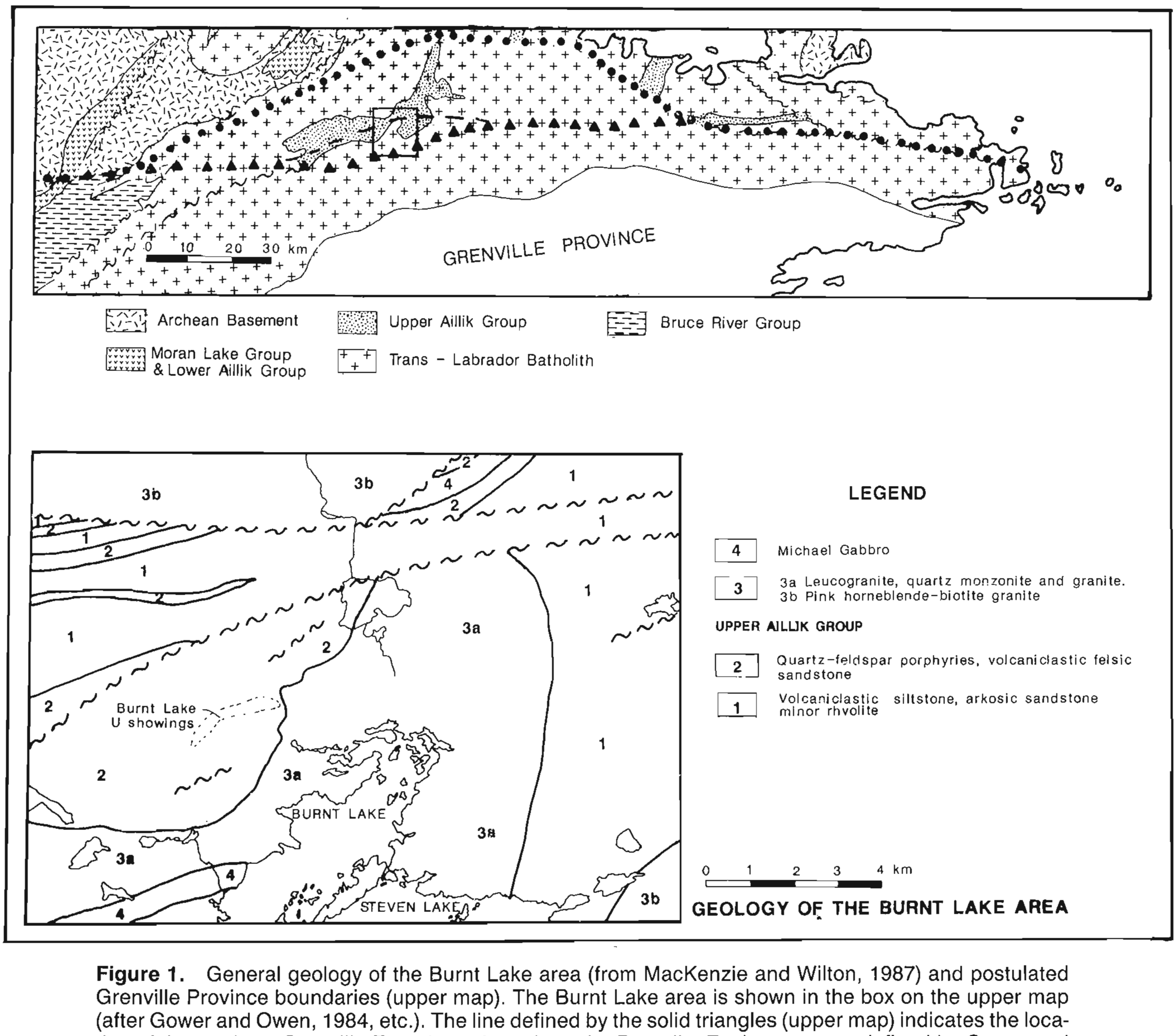

Figure 1. General geology of the Burnt Lake area (from MacKenzie and Wilton, 1987) and postulated Grenville Province boundaries (upper map). The Burnt Lake area is shown in the box on the upper map (after Gower and Owen, 1984, etc.). The line defined by the solid triangles (upper map) indicates the location of the northern Grenville Front boundary along the Benedict Fault system as defined by Gower and Ryan, 1986), Owen et al., 1986 and Kerr, 1987; the dashed line (through the middle of the Burnt Lake area) indicates the boundary of Gower and Owen (1984) and Scharer et al. (1986); and the line of solid circles indicates the boundary (along the Adlavik Brook Fault) of Gower et al. (1980; 1982) and Wardle et al. (1986). Based on our isotopic data, the latter boundary (ie. along the Adlavik Brook Fault) may be more suitable.

Granitic rocks, including the Burnt Lake Granite (MacKenzie and Wilton, 1987), in the area are part of a Middle Proterozoic (ca. 1680-1645 Ma) granitoid terrane termed the Trans-Labrador Batholith (Wardle et al., 1986; Kerr, 1986; 1987), a portion of the ca. 1.65 Ga Labrador Orogen (Thomas et al., 1986).

The dominant structural fabrics of the Upper Aillik Group in this area were developed during the Hudsonian orogeny (Bailey, 1979; Gower et al., 1982) which pre-dated the Burnt Lake Granite. A Grenvillian deformational fabric (south plunging stretching lineations) has been recognized (Bailey, 1979) as the latest fabric developed in the Burnt Lake area.

THE GRENVILLE FRONT IN LABRADOR

The location and nature of the Grenville Front in Labrador has long been debated (see Gower et al., 1980). The most recent subdivison of the Grenvillian orogen is based upon the recognition of allochthonous elements through the southern portions of eastern Labrador and western Labrador Quebec (Gower and Ryan, 1986; Rivers and Chown, 1986; Wardle et al., 1986).

The criteria used to define the northern boundary of the front (Fig. 1) have been structural and lithological features, along with metamorphic and geophysical characteristics. The boundary was defined by Gower et al., (1980) as the northern extent of widespread Grenvillian brittle deformation coinciding with major regional faults. The boundary has also been described as the northernmost limit of the Michael Gabbro (eg. Gower and Ryan, 1986). Gower et al. (1980;1982), and Wardle et al. (1986) placed the boundary along the Adlavik Brook Fault system north of the Burnt Lake area. Stevenson (1970), Kerr (1987), Gower and Ryan (1986), and Owen et al. (1986) described the boundary as being along the Benedict Fault System south of Burnt Lake (i.e. south of the Upper Aillik Group) within the Trans-Labrador Batholith.

Gower and Owen (1984), and Scharer et al. (1986) describe the northern boundary of the Grenville tectonic front as passing through the Upper Aillik Group proximal to the Michelin, Burnt Lake and Emben uranium deposits. The spatial juxtaposition of the front and uranium deposits as defined by these authors would therefore be suggestive of a genetic link between the mineral-izing event and Grenvillian tectonism. The aim of this paper is to show that tectonism affected all of the Upper Aillik Group within the area and should not be considered to have ended along the traces of the uranium occurrences.

STRUCTURAL DATA FROM BURNT LAKE

Changes in lithology as mapped in the Burnt Lake area (Fig. 1) are usually attributable to faulting, however, the extent of the deformation affecting the lower sedimentary and upper volcanic sequences is similar. In general, the sedimentary rocks in the Burnt Lake area have well-developed northeasterly-striking, southeast-dipping bedding while the volcanic rocks have a northeasterly-striking, south-dipping cleavage with southeasterly-plunging mineral lineations.

The main phase of Hudsonian-Makkovikian deformation which affected the Burnt Lake area, produced isoclinal, southwesterly plunging antiform-synform pairs with the antiforms generally sheared along their axes (Bailey, 1979). Bailey also mentions the presence of L-fabrics (mineral lineations) in granitic rocks. Bailey (1978) states that lineations are developed along fault (mylonitic) and/or shear planes perpendicular to the shear fabric. Bailey (1979) and Gower et al.(1982) ascribe L-fabric development to shearing/faulting during Grenvillian deformation.

Figure 2 is an equal area contoured stereographic projection of mineral stretching lineations as measured in the Burnt Lake area. The diagram indicates that the lineations mainly have a steep southeastern plunge with minor examples of northwestern plunges. Gower et al. (1982) stated that folds were overturned to the north-northwest during the Grenvillian deformation concurrent with the L-fabric development. If the linear fabrics as shown in Figure 2 were produced during the Grenvillian orogeny, then this event obviously had a great effect on rocks in the Burnt Lake area; enough of an effect to suggest the area lies within the Grenville Province.

Rb-Sr ISOTOPIC DATA FROM THE BURNT LAKE GRANITE

A new Rb/Sr isochron which was generated for whole rock samples of the Burnt Lake Granite (Fig. 3) indicates that isotopic signatures have been influenced by some superimposed event in the Burnt Lake area . The age of 1548 $\pm$ 90 Ma is much younger (by about 100 Ma) than would be expected if the granite were part of the Trans-Labrador Batholith which is dated at ca. 1620-1650 Ma on the coast and ca.1654 Ma in central Labrador (Thomas et al.,1986). In particular, the Strawberry Granite with a K-Ar biotite age of 1610 $\pm$ 55 Ma and the Monkey Hill Granite with a K-Ar biotite age of 1620 $\pm$ 60 Ma (Wanless et al., 1970). (These two granites occur within coastal exposures of the Upper Aillik Group near Makkovik).

This young date agrees with a 1550 Ma Rb/Sr date obtained by Kontak (1980) for the Walker Lake Granite, which he defined as part of the same pluton. The younger age for the granite might be correct if the granite represents a later (or terminal) phase of the Trans-Labradorian granitoid magmatism (1660-1650 Ma — after Thomas et al., 1986). However, we wish to suggest that the 1548 Ma age may represent a partial re-setting of the Rb/Sr isotope systematics (i.e. loss of radiogenic Sr or gain of Rb) giving the Burnt Lake Granite an apparent 100 Ma younger age.

A similar Rb-Sr isotope "re-setting" has occurred in Upper Aillik Group felsic volcanic rocks in this same Michelin zone (ie. that portion of the Aillik Group which occurs south of the Adlavik Brook Fault). Kontak (1980) reported a well fitted Rb/Sr whole rock isochron of 1767 $\pm$ 4 Ma for a rhyolite unit on Michelin Ridge (just to the northwest of the Michelin deposit). Kontak later revised the age to 1786 $\pm$ 38 Ma (in Ryan, 1984 p. 177-179) and stated that this date reflected Hudsonian orogenic events rather than the true volcanic age. The interpretation that the isochron does not indicate the primary magmatic age has been borne out by a U/Pb zircon age of 1855 Ma (Scharer et al., in preparation) from the same outcrop area sampled by Kontak.

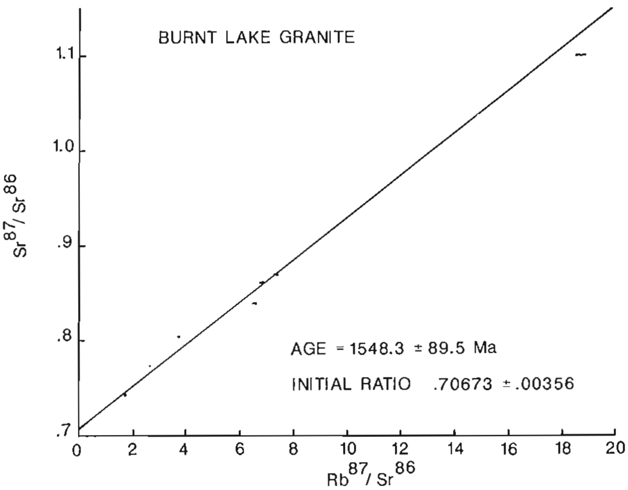

Figure 2. Equal area contoured stereographic projection of mineral elong-ation lineations in the Upper Aillik Group rocks in the Burnt Lake area.

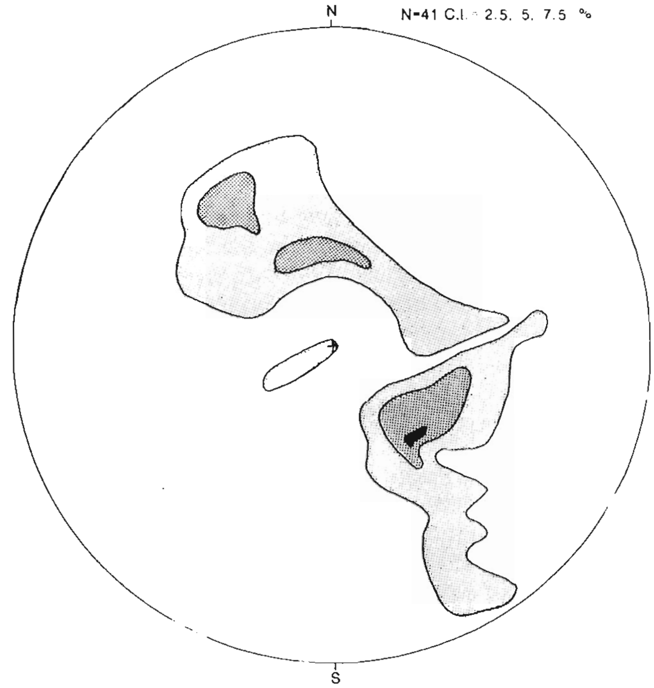

Figure 3. Rb/Sr whole rock isochron for the Burnt Lake Granite (data available upon request).

Recently, Heaman et al., (1986) studied the U-Pb zircon, U-Pb sphene and Rb-Sr whole rock ages from granites in the Central Metasedimentary Belt of Ontario and found similiar isotopic complexities. They concluded that the Rb-Sr disturbance giving an incorrect younger age (compared with the more accurate U-Pb zircon age) was the result of late stage-fracturing and alkali metasomatism. The most interesting feature of the apparent isotopic re-setting in Labrador is that it has apparently thoroughly affected both the older volcanic rocks and post-tectonic granitiods on a very broad regional scale (i.e. the isochron re-setting is not attributable to single samples, thus, all samples must have undergone the same isotopic alteration), and the effect in both cases has been to reduce the "real age" by about 60-90 Ma. Since the post- Hudsonian granites were equivalently affected, the isotopic disturbance in the felsic volcanic rocks was not a result of Hudsonian deformation. The only major subsequent deformation of rocks in the area was Grenvillian.

Walraven et al. (1986) reported similar "disturbances" in Rb-Sr isotope systematics that have produced a consistent lowering in the whole rock Rb/Sr isotopic dates of granites from the Bushveld Complex. These authors described the phenomenon as resulting from the preferential removal of radiogenic Sr (^{87}Sr) from potassium feldspar. We suggest that the widespread, cross-unit age date lowering in the Burnt Lake area might have resulted from a similar mechanism. The Sr removal could conceivably have been initiated by Grenvil-lian tectonics. (We plan to undertake Rb-Sr isotope analyses of mineral separates from the Burnt Lake Granite in order to test the hypothesis of radiogenic Sr removal from distinct mineral sites).

Kontak (1980) reported that two pitchblende separates from the Burnt Lake Showing gave discordant Pb isotope ages of 1770 and 1680 Ma. He thought that these dates resulted from Grenvillian disturbance of the uranium-lead isotope system. Gandhi (1986) reported Pb isotopic analyses of a galena sample from the Michelin uranium deposit located 20 km west-southwest of the Burnt Lake Showing. The galena convained a very high proportion of radiogenic lead, which he interpreted as the result of redistribution of element during the Grenville time. A similar Grenvillian remobilization can also be seen in the isotopic analyses of a galena sample from the Burnt Lake Showing which has yielded the following ratios: $^{206}Pb/^{204}Pb = 762.02$ and $^{207}Pb/^{204}Pb = 126.80$. A two stage calculation based on a regression line at a slope of about 0.15 (controlled by the highly radiogenic isotopic ratios) and a maximum host rock age of 1855 Ma (after Gandhi, 1986 and Schares et al., in preparation) yields a calculated minimum age of about 900 Ma for the extraction of radiogenic lead from uranium and it's deposition as galena (R.I. Thorpe, personal communication, 1986).

CONCLUSIONS

Although far from being conclusive, the data presented appears to show that the Burnt Lake area has undergone Grenvillian deformation as evidenced by the numerous faults and south-trending mineral-stretching lineations. Such a deformational overprint would also explain the isotopic signatures of the relevant rock types and mineral occurrences; including the "young" Rb-Sr whole rock ages of the Burnt Lake Granite and Upper Aillik Group. The intensity of the deformational overprint and the presence of Michael Gabbro intrusions north of the Burnt Lake Showings (Fig. 1) are further evidence that the Burnt Lake area lies within the Grenville Province and that the northern boundary of this province is more logically defined along the Adlavik Brook Fault.

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Stratigraphy and lithological composition of Quaternary sediments from five boreholes, Kipling Township, Ontario

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Kettles, I.M. and Wyatt, P.H., Stratigraphy and lithological composition of Quaternary sediments from five boreholes, Kipling Township, Ontario; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 239-244, 1988.

Abstract

A sequence of waterlaid sediments overlies and underlies a clay till interpreted as Kipling or Cochrane Till, which in turn overlies up to three other tills separated by beds of massive and laminated silt-clay. Pebble lithology data do not exhibit consistent patterns from core to core due to the variabiltiy of sediment types between boreholes. Few stratigraphic correlations can be drawn between boreholes. Organic layers in an intertill bed of laminated silt-clay indicate that some units date back at least to the early Wisconsinan.

Major lithological groups identified in the granule-pebble (2-5.6 mm) fraction of tills are carbonate, rounded quartz, Precambrian crystalline, and dark fine grained clastic. The key indicator erratics identified are: rounded quartz, lignite, white sandstone, iron oxide encrusted dark fine grained clastic and red fossiliferous siltstone.

Résumé

Des sédiments quaternaires ont été décrits et échantillonnés dans cinq trous de sondage forés dans le socle crétacé de la rivière Mattagami dans le comté de Kipling. Une séquence de sédiments déposés dans l'eau est sus-jacente et sous-jacente à un till argileux que l'on considère comme appartenant au till de Kipling ou de Cochrane, lequel repose à son tour sur trois autres tills séparés par des couches de limon et d'argile massifs et laminés. La configuration lithologique des galets n'est pas cohérente d'une carotte à l'autre en raison de la variabilité des types de sédiments qui se trouvent entre les trous de sondage. Peu de corrélations stratigraphiques peuvent être établies entre les trous de sondage. La présence de couches organiques dans une couche mise en place entre deux tills et composée de limon et d'argile laminés indique que certaines unités datent au moins du début du Wisconsinien.

Les principaux groupes lithologiques identifiés dans la fraction formée par les graviers et les galets (2-5 à 6 mm) des tills sont des roches carbonatées, des quartz arrondis, et des roches clastiques foncées à grain fin provenant des roches cristallines du Précambrien. On a identifié les blocs erratiques d'origine connue suivants: quartz arrondi, lignite, grès blanc, roche clastique foncée à grain fin et à incrustations d'oxyde de fer et microgrès fossilifère rouge.

INTRODUCTION

In August 1985, five boreholes were drilled in Kipling Township, Ontario (Fig. 1) to evaluate the mineral potential of the Douglas silica-kaolin-mica property controlled by Carlson Mines Limited. Examination and analyses of the Quaternary sediments recovered during this project provide information about the composition and stratigraphy of glacial sediments in this part of the James Bay Lowland. A better understanding of Quaternary geology in this area is needed to link the results of work undertaken in the Hudson Bay Lowlands with that undertaken by Averill et al. (1986), and Bird and Coker (1987) in the Timmins mining district. Present knowledge of the Quaternary sequence in this region is based on river traverse work carried out by Skinner (1973) and aminostratigraphy conducted by Andrews et al. (1983).

In this report, the lithologic composition of the different till units in this area of Phanerozoic bedrock is quantified by examining the granule-pebble fraction (2-5.6 mm). These counts will provide background data for comparison with the lithological composition of tills recovered from ongoing stratigraphic studies immediately to the south on shield terrane (Smith and Wyatt, 1988; Bird and Coker, 1987; DiLabio et al., 1988). Mapping regional variation in till lithology across the Precambrian-Paleozoic boundary should help constrain interpretations of ice flow directions based on erratic provenance, and may provide information about transport distances.

METHODS

The boreholes were drilled using a high frequency vibration Rotasonic drill operated by Midwest Drilling of Winnipeg, Manitoba. The drilling system and methods of core recovery are described in Averill et al. (1986). The core was described and sampled in detail on site.

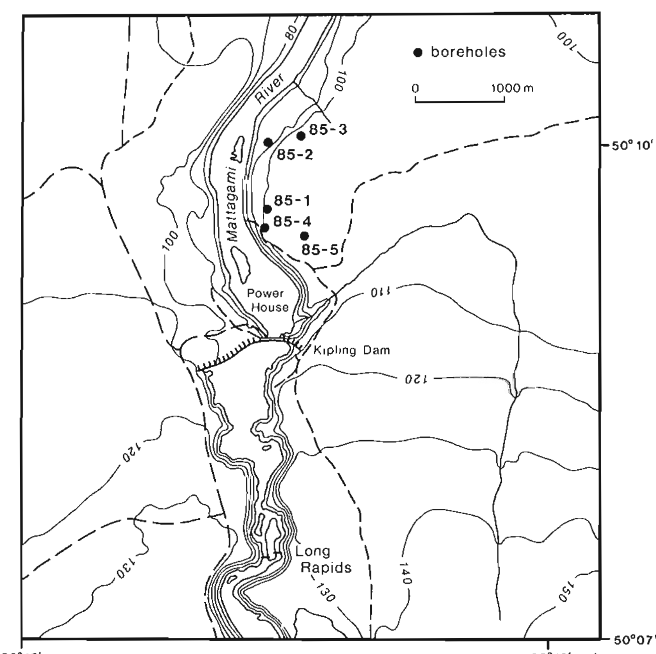

Figure 1. Location map of study area and borehole locations, Kipling Township, Ontario.

The granule-pebble fraction was separated from diamicton samples, yielding 25 to 50 g sample separates. Different lithological groups were identified and calculated as a weight per cent of the total fraction using the technique of Shilts et al. (1979).

BEDROCK SETTING

The property is located just north of the Kipling Dam on the Mattagami River near the contact between underlying Cretaceous sediments and the Canadian Shield to the south (Fig. 2). At the contact, just south of the drilling site, Precambrian rocks form an escarpment which rises more than 60 m over 3 km. North of the contact, Cretaceous deposits overlie Devonian limestones and shales which dip gently towards the escarpment (Vos, 1982). Here the terrain is flat except for deep incisions along the rivers and streams.

The Cretaceous deposits are poorly consolidated terrigeneous sediments predominantly composed of clay, sand and lignite coal. They are incompletely mapped due to the thick cover of Quaternary sediments. Deposits of silica sand are predominantly composed of transparent to transluscent quartz, with larger grains slightly frosted and pitted (Vos, 1982). Grains larger than 2 mm are rounded or subrounded. Carbonization of Cretaceous wood has formed lignite which appear glassy under magnification, although plant cell structure is still evident. The soft lignite grains fracture conchoidally, and may not survive the pebble extraction procedure. The lignite examined during this project would not likely be confused with Quaternary wood. To date, several inliers of Paleozoic and Precambrian rock have been delineated, some of which are close to the drilling site.

As shown in Figure 2, Cretaceous deposits are completely encircled by numerous Paleozoic sedimentary rock formations except along their southern edge where they abut the Precambrian escarpment (Sanford and Norris, 1975). Paleozoic rocks include limestones, dolomites, and calcareous siltstones, as well as noncalcareous, varicoloured cherts, shales, siltstones, mudstones and arkosic sandstone. Precambrian rocks in this region are composed of Archean migmatitic metasediments and minor metavolcanics.

DESCRIPTION OF THE BOREHOLE SEDIMENTS

The sequence of Quaternary sediments found in each borehole is summarized in the column diagrams (Fig. 3). The lithofacies code used to describe the sediment is modified after Eyles et al.(1983). Nearly all units designated clay matrix-supported diamicton on the the column diagrams were interpreted in the field as till. In the unit descriptions and other discussions, they are referred to as clay till unless indicated otherwise. Colours used to characterize the sediments describe how they appeared when described fresh at the site. When till sample colours were later determined in the laboratory using the Munsel colour chart, they were all coded one of three colours — 5Y 4/1 (dark grey), 5Y 4/2 (olive grey), or 5Y 4/3 (olive).

The upper part of each core consists of a sequence of interbedded sand, gravel, and diamicton and both massive and laminated silt-clay. The sequence is 19.8, 13.1, 29.3, 9.8 and 12.5 m in cores 85-1 to 85-5, respectively. Modern organic layers and fragments are usually present in the top 3 m of the core.

In cores 85-3 and 85-5, grey and light brown clay till is present at depths of 9.5 — 11 m and 5.5 — 7.9 m, respectively. In core 85-3 the till is overlain by blue-grey, laminated silt-clay grading to grey massive clay. It is underlain by blue-grey, laminated silt-clay. The clay till in core 85-5 is overlain by massive silt-clay and underlain by beds of sand and grey laminated silt-clay within which organic layers were found at 8.8 m. Organic fragments were also found at 15.3 m in core 85-3.

Below 29.3 m in core 85-3, there are three units of compact, olive, clay till at 29.3 — 31.4 m, 33.8 to 40.5 m and 43.6 to 46.3 m. The lowest unit is overlain by 0.9 m of green pebbly gravel interbedded with layers of olive-coloured clayey diamicton. Between the three till units just described are two beds of fine grained sediment. The upper one is composed of compact, olive, laminated silt-clay grading down to gravel interbedded with clay diamicton. The lower is composed of laminated silt-clay overlying green pebbly clay.

In core 85-2, there is an upper, grey, clay to sandy clay till (13.1 — 14.6 m), a gravelly diamicton (24.6 — 31.7 m) and a lower unit that contains one bed of very dense, olive till (35.7-37.5 m), which overlies another bed of gravel diamicton and sand within which there is a layer of blue grey pebbly clay diamicton, 0.3 m thick.

The upper intertill bed (15.6 — 24.6 m) in core 85-2 is laminated sand, coarse silt and clay, interbedded with a layer of pebbly silty sand (23.8-24.2 m). The laminated unit becomes sandier by 23.8 m but again increases in silt content below the layer of coarse sediment. There are also two zones (20.7 — 21.0 m and 22.7 — 23.8 m), where individual laminates contain wood chips and other organic matter. The lower intertill bed (31.7 - 33.6 m) consists of grey to blue grey, massive silt-clay with occasional pebbles. This is underlain by very dense, green, laminated silt-clay (33.6 — 35.1 m) and massive, pebbly clay (35.1 — 35.7 m).

In the lower part of core 85-4, there is a till unit between 9.8 and 16.2 m within which four different subunits were distinguished on the basis of colour and texture. They are, from top to bottom: from 9.8 to 12.2 m, a blue-grey clay-rich pebbly unit which becomes much sandier towards the base; from 12.2 to 13.4 m, a very hard blue grey till that is less pebbly than the overlying unit; from 13.4 to 14.8 m, a layer of blue-grey clay till; and from 14.8. to 16.2 m, a pebble-poor clay till. This sequence is in turn underlain by interbedded sand, silt and clay, 4.0 m thick.

There are two till units separated by a gravel bed 0.5 m thick, below 12.5 m in core 85-5. The upper unit is light grey brown and clay-rich and the lower is grey and sandier

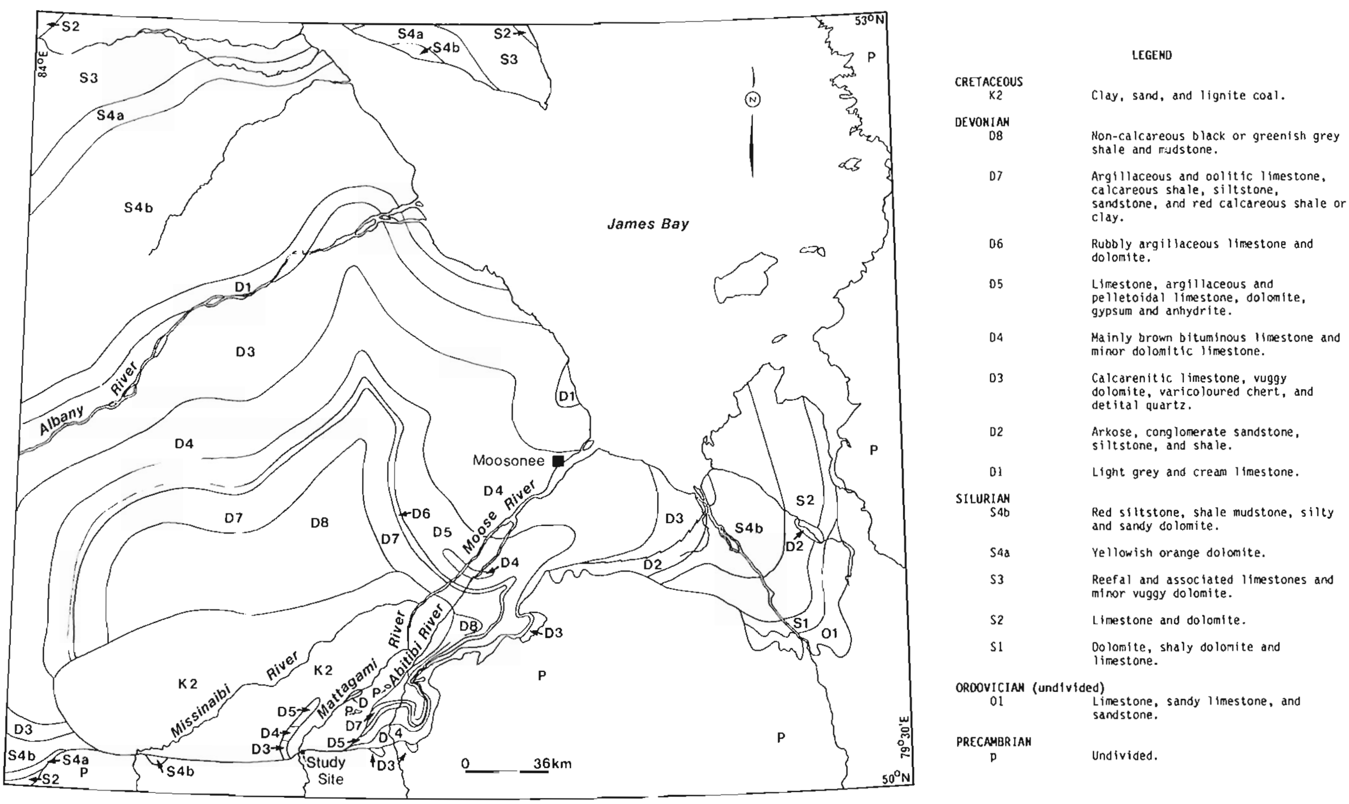

Figure 2. Bedrock map of James Bay Lowland (modified after Sanford and Norris, I975).

in texture. In core 85-1, there is only one unit of sandy till between the overlying fine grained unit and underlying Cretaceous silica sand.

DISCUSSION

Considering the close proximity of the boreholes, the variation in sediment facies within individual cores and from one core to the next is marked. In all cores except core 85-1 there are two to four till units interbedded with massive or laminated silt and clay. Variation in the sequence preserved from core to core is presently unexplained but there are at least two possible explanations. In this area there may have been a complex history of fluvial erosion and deposition during previous interglacial episodes. It is possible that some boreholes have intersected abandoned river channels incised in older glacial deposits which were subsequently filled with younger glacial and nonglacial sediment. Variation in the sequence of sediments preserved at each site may also reflect local changes in the patterns of subglacial erosion or deposition which developed as the ice overrode the escarpment during different glaciations.

It is difficult to evaluate the significance of the different intertill fine grained units. Because ice from northerly source areas advances up the topographic gradient, it is always fronted by a proglacial lake during advance and the early stages of retreat. As a result, beds of laminated and massive silt and clay may be deposited during minor fluctuations of the ice front during a single glaciation, as well as during major interglacials. When the ice margin is close to the escarpment, there may be local thickening of lacustrine sediments since deposition is favoured in bathymetric lows. The laminated unit beginning at 18 m in core 85-2, contains two zones where disseminated organic matter and wood fragments occur. It is likely that these lacustrine or paludal sediments are interglacial deposits and are possibly correlative with the Owl Creek Beds of the Timmins area (Dilabio et al., 1988) or with the nearby Missinaibi Formation sediments of the James Bay Lowlands (Skinner, 1973). Skinner observed forest beds which he correlated with Missinaibi Formation on Adam Creek, 4 km north of the drilling site.

Despite their proximity, there are few obvious stratigraphic correlations that can be drawn among the boreholes. The massive and laminated fine sand to clay interbedded with a thin layer of clay till above 29.3 m and 12.5 m in cores 85-3 and 85-5, respectively are likely equivalent to the sequence of sediments designated Friday Creek sediments, Kipling Till and postglacial sediments by Skinner (1973). The clays of this upper unit are grey or blue-grey and noncompact as opposed to the subtill laminated clays of 85-2, and 85-3 which are compact and have a distinct olive colour.

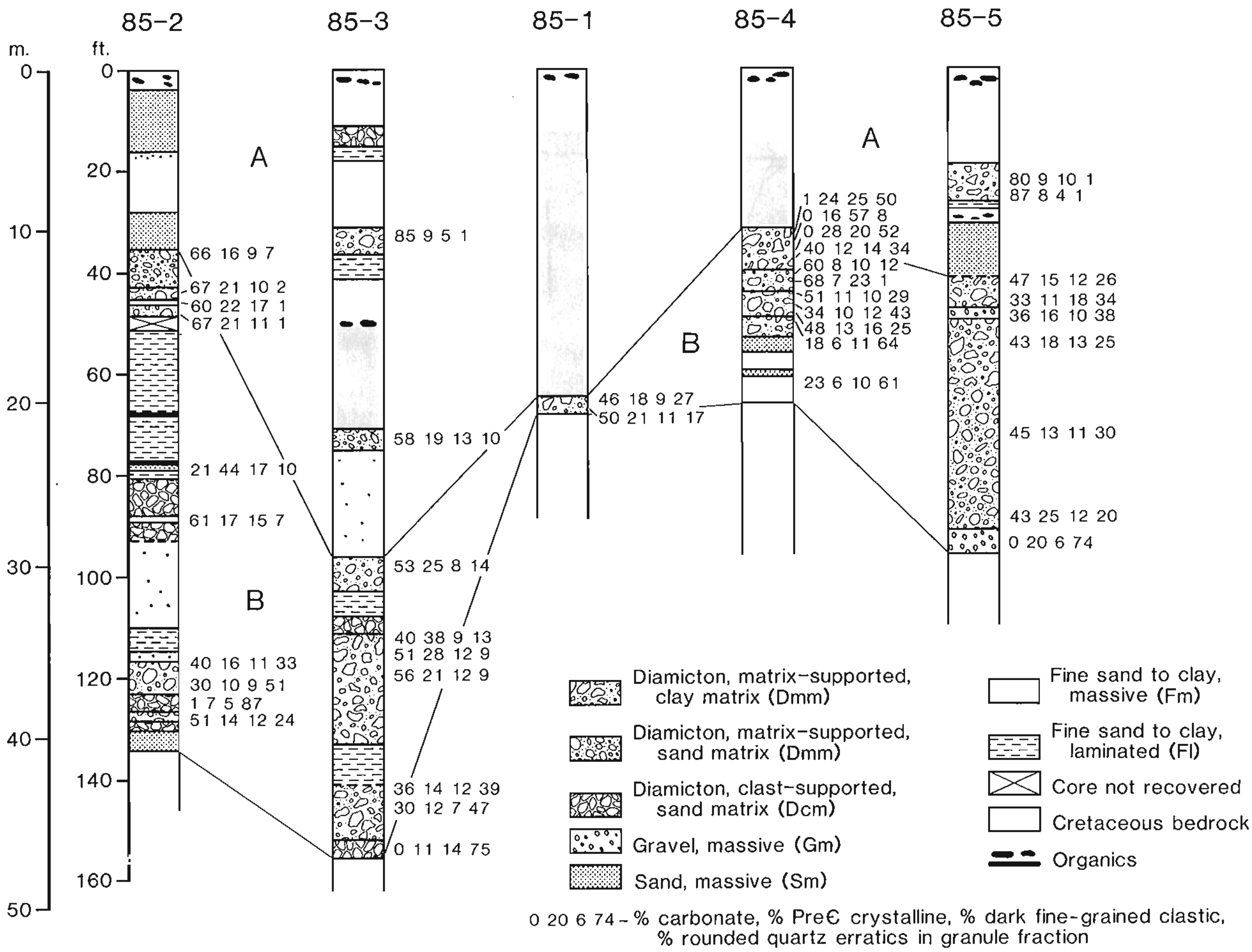

Figure 3. Stratigraphic columns and correlations. Units are interpreted as: A = Friday Creek and overlying sediments, and B = pre-Friday Creek sediments.

Table 1. Weight per cents of different lithological groups in the granule-pebble fraction (2-5.6 mm) of till and other diamicton samples.

LOCATION AND DEPTH (FEET)	SAMPLE WEIGHT grams	Carbonate -includes clastics	Grey Chert	Dark Non-calc. Clastics	Iron Form	Wood Frags	Met-allic	Red Sand-stone	Other Sand-stone	Round Quartz	Mafic	Other Cryst-alline	Dia-base	White Sand-stone	Lignite	Oxide Coated Sand-stone
85-1-65	34.55	45.79	0.06	8.80	0.12	0.00	0.00	0.00	0.29	26.53	0.00	18.27	0.00	0.14	0.00	0.00
*85-1-65	46.66	50.39	0.00	8.78	0.06	0.00	0.00	0.00	0.11	17.47	0.00	21.32	0.00	1.88	0.00	0.00
*85-2-44	21.32	67.35	0.52	9.01	0.00	0.00	0.00	0.00	0.00	1.97	0.00	20.12	0.00	1.03	0.00	0.00
*85-2-45	35.44	60.33	0.00	13.05	0.00	0.00	0.00	0.48	0.00	1.33	0.17	21.54	0.00	3.07	0.03	0.00
*85-2-47	23.62	67.12	0.00	8.77	0.00	0.04	0.00	0.00	1.81	1.26	0.00	20.99	0.00	0.00	0.00	0.00
*85-2-78.5	42.02	21.20	0.09	12.97	0.00	0.00	0.00	0.40	0.07	18.45	0.00	43.43	0.00	3.37	0.00	0.00
*85-2-90	21.25	60.77	0.00	15.00	0.00	0.00	0.00	0.00	0.24	6.63	0.00	17.36	0.00	0.00	0.00	0.00
*85-2-118	46.15	40.11	0.00	9.67	0.00	0.00	0.00	0.11	0.07	33.36	0.11	15.73	0.22	0.63	0.00	0.00
*85-2-122	30.98	29.91	0.00	8.82	0.00	0.00	0.00	0.23	0.00	50.74	0.00	10.01	0.00	0.00	0.00	0.00
*85-2-125	15.96	0.13	0.00	5.33	0.00	0.00	0.00	0.00	0.00	87.41	0.00	7.14	0.00	0.00	0.00	0.00
*85-2-127.5	48.75	50.62	0.00	11.12	0.00	0.00	0.00	0.00	0.10	24.20	0.00	13.57	0.00	0.39	0.00	0.00
*85-3-33	29.66	84.82	0.00	4.41	0.00	0.00	0.00	0.17	0.10	1.14	0.00	9.36	0.00	0.00	0.00	0.00
*85-3-73	42.22	57.51	0.50	11.72	0.00	0.02	0.00	0.73	0.21	9.90	0.00	19.14	0.00	0.24	0.02	0.00
*85-3-74.6	20.04	68.00	0.00	7.70	0.00	0.00	0.00	0.00	0.00	10.70	0.00	15.00	0.00	0.47	0.00	0.00
*85-3-98	17.36	52.78	0.00	6.53	0.00	0.00	0.00	0.00	1.15	13.98	0.34	25.21	0.00	0.00	0.00	0.00
*85-3-112	17.61	39.47	0.00	8.73	0.00	0.00	0.00	0.00	0.17	13.23	0.00	38.06	0.34	0.00	0.00	0.00
*85-3-115	27.25	50.57	0.00	11.47	0.00	0.00	0.00	0.00	0.00	8.98	0.84	28.14	0.00	0.00	0.00	0.00
*85-3-120	32.44	56.32	0.00	11.25	0.00	0.03	0.00	0.31	0.00	7.83	3.14	21.12	0.00	0.00	0.00	0.00
*85-3-142	38.73	35.88	0.00	11.66	0.00	0.00	0.00	0.00	0.00	38.68	0.00	13.78	0.00	0.00	0.00	0.00
*85-3-145	44.74	29.53	0.00	6.96	0.00	0.00	0.00	0.00	0.00	46.69	0.00	12.37	0.00	0.47	0.00	0.00
*85-3-153	42.88	0.00	0.14	13.60	0.00	0.00	0.00	0.00	0.07	75.70	0.00	10.49	0.00	0.00	0.00	0.00
*85-4-32.5	34.99	0.34	0.00	21.82	0.00	0.00	0.00	0.00	0.54	50.23	0.00	24.50	0.00	2.57	0.00	0.00
*85-4-33.3	31.28	0.00	0.00	55.34	0.00	0.00	0.00	0.00	0.67	8.13	0.00	15.87	0.00	0.67	0.00	19.32
*85-4-35	41.44	0.00	0.00	19.21	0.00	0.00	0.00	0.00	0.00	51.44	0.00	27.54	0.92	0.89	0.00	0.00
*85-4-37.7	67.45	40.28	0.00	13.30	0.00	0.00	0.00	0.27	0.00	33.69	0.00	11.61	0.13	0.71	0.00	0.00
*85-4-41	26.02	60.00	0.00	19.00	0.00	0.00	0.00	0.25	0.00	12.00	0.00	8.00	0.00	2.00	0.00	0.00
*85-4-42	29.66	67.73	0.00	22.54	0.67	0.00	0.00	0.00	0.00	1.35	0.00	6.86	0.00	0.84	0.00	0.00
*85-4-44.5	7.10	51.00	0.00	10.00	0.00	0.00	0.00	0.00	0.00	29.00	0.20	11.00	0.00	0.00	0.00	0.00
*85-4-47	65.14	33.72	0.00	11.20	0.00	0.00	0.00	0.00	0.37	42.77	0.00	10.30	0.00	0.63	0.05	0.00
*85-4-48.5	21.40	45.00	0.00	16.00	0.00	0.00	0.00	0.00	0.00	25.00	0.00	13.00	0.00	1.00	0.00	0.00
*85-4-50	46.17	18.42	0.00	8.75	0.00	0.00	0.00	0.26	0.00	63.89	0.00	5.52	0.00	1.65	0.00	0.00
*85-4-62.3	44.24	22.95	0.00	7.36	0.00	0.00	0.00	0.05	0.36	61.00	0.00	6.16	0.00	2.12	0.00	0.00
*85-4-65	21.48	15.94	0.00	6.85	0.00	0.00	0.00	0.00	0.00	67.57	0.00	8.85	0.00	0.79	0.00	0.00
*85-5-23	21.29	79.84	0.00	6.53	0.00	0.00	0.14	0.79	0.00	0.51	0.14	9.43	0.00	2.61	0.00	0.00
*85-5-25	37.71	86.85	0.00	3.80	0.00	0.00	0.00	0.00	0.00	0.35	0.00	8.37	0.00	0.64	0.00	0.00
*85-5-42	22.21	47.46	0.00	11.33	0.00	0.00	0.00	0.00	0.00	25.84	0.00	15.01	0.00	0.36	0.00	0.00
*85-5-46	49.82	33.37	0.00	13.14	0.04	0.00	0.00	1.01	0.74	34.26	0.00	10.67	0.00	3.28	0.10	3.38
*85-5-48	41.67	35.97	0.05	8.16	0.00	0.00	0.00	0.00	0.00	38.32	0.00	15.74	0.00	1.75	0.00	0.00
*85-5-54.0	35.14	43.40	0.00	12.46	0.00	0.00	0.00	0.00	0.00	25.15	0.00	17.83	0.00	0.37	0.79	0.00
*85-5-72	44.29	45.02	0.00	10.48	0.00	0.00	0.00	0.47	0.27	29.74	0.00	13.16	0.00	0.09	0.00	0.77
*85-5-89	45.46	42.90	0.00	11.51	0.33	0.00	0.00	0.00	0.35	19.92	0.00	24.92	0.00	0.00	0.07	0.00
*85 5-94	38.40	0.00	0.42	4.34	0.00	0.00	0.00	0.47	0.55	74.17	0.00	19.74	0.00	0.31	0.00	0.00

The tie line which separates unit A from unit B, in Figure 3, marks the contact between Friday Creek and overlying sediments from underlying tills.

In an effort to correlate some of the till units, their lithological composition was analyzed, and the results are shown in Table 1. The separate lithological categories were further grouped as cumulative totals of carbonate, crystalline, dark noncalcareous argillites and greywackes, and rounded quartz. These values are included beside their respective stratigraphic units in Figure 3. Lithologies which did not fit in these categories make up less than 1 % of the total in all but 3 samples. The notable exception is the till at 10.2 m in core 85-4, where 19.3 % of the pebble fraction is iron oxide encrusted dark fine grained sandstone.

The pebble lithology data do not exhibit consistent patterns from core to core, which is not surprising given the variability of sediment types among boreholes. The drill sites are located west-southwest of outcrops of most of the major bedrock groups of the James Bay Lowland (see Fig. 2). The southwesterly flow direction has been ascribed to ice which deposited the upper two tills as well as the older pre-Missinaibi tills in the Moose River Basin (Skinner, 1973). Therefore, minor changes in flow patterns may result in pronounced lithological variability in and between units.

In general, it is not possible to correlate till units by a distinctive lithological composition in these cores. However, the following exceptions have been noted: 1) grey sandy till at 19.8 m in core 85-1, is similar in texture, appearance, and lithological composition to the till between 12.2 and 16.5 m in core 85-5; 2) the clay tills in cores 85-3 and 85-2 between 43.6 and 46.3 m and 35.7 and 37.5 m, respectively, are both compact, olive coloured, and have similar lithological composition; and 3) the uppermost clay tills in cores 85-3 and 85-5 have nearly identical lithologies, with carbonate contents greater than 80 %, and may be equivalent to Skinner's (1973) Kipling Till or Hughes' (1965) Cochrane Till.

The percentages of rounded and frosted quartz grains derived from the underlying Cretaceous silica sands, generally diminish upwards through each core as carbonate percentages increase. One exception is core 85-4 where the upper 0.8 m of the till unit between 10.4 and 11.6 m is devoid of carbonate erratics and is brown in contrast to the grey till below. In addition, the uppermost portion of the till contains 19.3 % iron oxide encrusted sandstone. The lack of carbonate and presence of oxides may be interpreted as evidence for an interglacial weathering surface, or more likely it reflects the glacial erosion and redeposition of material derived from a Mesozoic surface. This would explain the high percentages of rounded quartz relative to crystalline rocks, and the restriction of heavy oxide encrustation to only one pebble type. The oxide encrusted sandstones are also found to a lesser extent in the till at core 85-5. The till throughout this core has a relatively uniform lithological composition and is probably one conformable unit.

The granule-pebble category of dark noncalcareous clastic rocks (Table 1) includes greywackes, arkose, siltstones, shales, and argillite. The uppermost units of the Belcher Fold Belt, along the east side of Hudson Bay, are the Omarolluk and Loaf formations, which consist of weakly metamorphosed Proterozoic arkose, grit, argillite, greywacke, sandstone, and greywacke sandstone (Dimroth et al., 1970). Dispersal of these distinctive erratics through and beyond the Hudson Bay Lowland has been documented by several authours (i.e. Shilts, 1980; Prest and Nielsen, 1987). Soft, dark fine grained clastic Devonian rocks also occur immediately north and east of the Cretaceous sediments (Fig. 2), in the Long Rapids (D8), Williams Island (D7), and Sextant formations (D2) in the James Bay Lowland (Sanford and Norris, 1975). Due to the inherent difficulties of working with the relatively small clast size of the granule-pebble fraction, no attempt was made to distinguish between Phanerozoic and Proterozoic dark fine grained clastic rocks.

SUMMARY

The stratigraphic sequence revealed in cores from these five boreholes on the Mattagami River is complex. There are up to four units of till separated by massive and laminated silt-clay, and coarser sediment facies. The amount of variation in sediment facies within sediment cores collected in such close proximity is unexplained. It suggests that caution be excercised when making regional correlations based on sparse borehole data.

A clay till in the upper part of cores 85-3 and 85-5 is underlain and overlain by units of massive and laminated silt-clay interbedded with diamicton and sand. If these are late glacial deposits, they are likely equivalents to those designated from older to youngest — Friday Creek sediments, Kipling Till and postglacial and modern sediments by Skinner (1973). The clay till in these two cores may also be equivalent to Cochrane Till of Hughes (1965), in which case the waterlain units both above and below the till represent sediments deposited in glacial Lake Ojibway. If the latter is the correct interpretation, it marks the northern limit of the lake about 30 km north of the Pinard Moraine (Boissoneau, 1966) at Smoky Falls.

There are at least 3 older till units separated by massive and laminated silt-clay and other coarser facies. The presence of organic layers in intertill laminated silt-clay unit in core 85-2 indicate that some units date back at least to the early Wisconsinan.

The authors have identified and calculated relative weight percentages of major and minor lithological groups present in the granule-pebble faction of till and other diamicton samples. Dominant groups are carbonate, rounded quartz, Precambrian crystalline, and dark fine grained noncalcareous clastic erratics. Some erratics derived from Cretaceous bedrock - lignite, rounded quartz, and rusty sandstone- are likely unique indicator erratics for this area. Non-calcareous white siltstone, arkose and red calcareous siltstones observed in sediment from these cores may also have restricted source areas within Paleozoic bedrock terrain. These data show relative frequencies of Precambrian versus Paleozoic lithologies in tills immediately north of the Canadian Shield and as a result will serve as a reference to which lithological data obtained from James Bay Lowland and Timmins area drilling transects can be compared.

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The Grenville Front thrust belt in western Labrador[1]

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van Gool, J., Brown, D., Calon, T., and Rivers, T., The Grenville Front thrust belt in western Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 245-253, 1988.

Abstract

The Grenville Front in western Labrador is the northwestern margin of a thin-skinned thrust belt with considerable basement involvement. The belt consists of a stack of thrust sheets, with Archean basement tectonically interleaved with Lower Proterozoic cover rocks, emplaced on the adjacent Superior foreland in a northwest directed thrust movement during the Grenvillian Orogeny.

The present level of exposure forms an oblique section through the belt. The metamorphic grade increases across the strike of the belt from lower greenschist in the northwest to middle amphibolite facies in the structurally higher thrust sheets in the southeast, and also along strike from northeast to southwest.

The style of deformation varies, depending upon lithology, metamorphic grade and position in the stack. Thrust sheets some tens to several hundreds of metres in true thickness are separated by high strain ductile shear zones. The thrust sheets are internally deformed by folds and imbricate thrusts, some of which have been identified as duplex structures.

Résumé

Le front de Grenville dans l'ouest du Labrador correspond à la bordure nord-ouest d'une zone de charriage à croûte mince au sein de laquelle le socle joue un rôle inportant. La zone est composée d'une superposition de nappes de charriage dans lesquelles des roches archéennes sont tectoniquement intercalées avec des roches de couverture de Protérozoïque inférieur qui ont été mises en place sur l'avant-pays contigu de la province du lac Supérieur dans un mouvement de charriage dirigé vers le nord-ouest survenu au cours de l'orogenèse grenvillienne.

La partie exposée forme une coupe oblique qui traverse la zone. Le degré de métamorphisme des roches augmente à travers la direction générale suivie par la zone, passant du faciès des schistes verts inférieur dans le nord-ouest au faciès des amphibolites intermédiaire dans les nappes de charriage structuralement plus élevées dans le sud-est, et aussi le long de la direction, du nord-est au sud-ouest.

Le mode de déformation varie selon la lithologie, le degré de métamorphisme des roches et l'emplacement au sein de la superposition. Des zones de cisaillement ductiles très déformées séparent les nappes de charriage mesurant quelques dizaines à plusieurs centaines de mètres d'épaisseur réelle. Ces dernières sont déformées à l'intérieur par des plis et des nappes de charriage imbriquées dont quelques-unes sont duplex.

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INTRODUCTION

Four key areas were mapped in detail (1: 10 000 scale) in a 40 km segment of the Grenvillian thrust belt adjacent to the Grenville Front in western Labrador, together with some reconnaissance fieldwork in the intervening areas (Fig. 1). This work is part of a two year mapping project in the Grenville Front area. The results of the mapping in the Bruce Lake area (Fig. 1) carried out in the 1986 field season were reported by van Gool et al. (1987). The areas for this year's detailed mapping project were selected on the basis of regional structural and metamorphic relationships, accessibility and outcrop density, covering along and across strike geological variations within the belt. Correlation between the areas was by the geological maps by Rivers (1980a, b, 1985a, b, c). Other previous work in the area, mostly of a reconnaissance nature, has been published by Gastil and Knowles (1960), Fahrig (1967), Roach and Duffel (1974), Rivers (1983a, b) and Rivers and Nunn (1985). The emphasis in the project was placed on the structural and metamorphic characteristics of the belt, while attention was also paid to the economical significance of the area, especially with respect to the major iron ore mines in the region. Techniques introduced by Boyer and Elliott (1982) and Butler (1982) were used for the mapping and interpretation of the thrust structures that characterize the area.

In this paper features of the area indicated in Figure 1. are discussed, including data from both summers' fieldwork. Two of the recently mapped areas at Emma Lake and Goethite Bay (Fig. 1) are described in more detail, as examples of the geology on a local scale.

GEOLOGICAL SETTING

The study area is situated in the parautochthon of the Grenville Province in western Labrador, as defined by Rivers and Chown (1986), which is separated from the autochthonous foreland to the northwest by the Grenville front. Southeast of the front all units have been affected to varying degrees by the Grenvillian orogeny. The rock types found in the area are either Archean basement rocks of the Ashuanipi Metamorphic Complex that form the Superior foreland or unconformably overlying Lower Proterozoic metasediments of the Knob Lake Group, which form the extension of the Labrador Trough into the Grenville Province.

The basement rocks are polydeformed gneisses, granulites, migmatites and pyroxene bearing amphibolites which are crosscut by post-tectonic Archean granitic intrusions. The stratigraphic sequence of the overlying Knob Lake Group is, from bottom to top: Attikamagen Formation, interbedded psammitic and pelitic schists; Denault Formation, massive, homogeneous dolomitic marbles; Wishart Formation, massive, thickly bedded quartzites; Sokoman Formation, banded quartzitic, carbonate and silicate iron formations; Menihek Formation, graphitic schists. Gabbros of the Middle Proterozoic Shabogamo Intrusive Suite have intruded all levels of the sequence. The sequence was deposited with onlap onto a Lower Proterozoic continental shelf at the eastern edge of the Superior craton. A more extensive description of these lithologies is given by Rivers (1983a) and van Gool et al. (1987).

The parautochthon in western Labrador forms a thin-skinned thrust belt showing considerable basement involvement (Rivers, 1983a; van Gool et al., 1987). A stack of thrust sheets was emplaced on top of the Superior foreland by a northwestward directed thrust movement. The individual thrust sheets are separated by ductile shear zones, with the sole thrust of the belt forming the actual Grenville Front. It is the northwesternmost thrust recognized in the area (Fig. 1). The present level of erosion cuts obliquely through the thrust belt. Towards the southwest, rocks have been exhumed from progressively deeper crustal levels. The regional tilt of the area enables a study of the thrust belt along a range of structural and metamorphic levels, so that a three dimensional image of the belt can be reconstructed, in spite of the lack of topographic relief.

LITHOTECTONIC SUBDIVISION OF THE AREA

A large number of thrust sheets have been delineated on the basis of their structural and lithological characteristics. They are grouped into four categories, which are represented in belts parallel to the strike of their bounding shear zones. From northwest to southeast these are (see fig. 1): I) Adjacent to the foreland, thrust sheets consisting predominantly of basement rocks that are not notably affected by Grenvillian deformation on an outcrop scale, but are transected by several Grenvillian thrusts. Locally these thrusts are decorated with thin slices of Sokoman Formation, as for instance in the Lac Virot area. II) Thrust sheets containing predominantly cover rocks of the Knob Lake Group with subordinate basement rocks in lower greenschist facies. The Corinne Lake, Flatrock Lake and Wide Valley thrust sheets are in this belt. III) A belt of upper greenschist to lower amphibolite grade, dominated by intensely deformed basement rocks, with locally preserved sedimentary cover. The Elmer Lake and Bondurant thrust sheets represent the main part of this belt. IV) The easternmost unit consisting exclusively of Knob Lake Group metasediments of amphibolite grade, represented by the Goethite Bay thrust sheet and the areas to the east and south of it, on the Julienne Peninsula and the mine area west of Wabush Lake. The thrust sheets of units II to IV were internally deformed during the thrust movement, as opposed to the sheets in unit I.

To the east of Wabush Lake another distinct belt is formed by the Shabogamo Lake unit, which has been examined in reconnaissance only. It consists of rocks that are interpreted as intensely reworked Archean basement, overlain by equally deformed predominantly pelitic metasediments, which may represent the Attikamagen Formation. The unit is cut by a large foliated tonalitic pluton of uncertain age. The nature of the structural relationship between the Shabogamo Lake Unit and the belts to the northwest is not clear, as the contact has not been observed. However, it is likely that it forms a higher thrust sheet in the stack. Internal Grenvillian deformation is mainly in the form of the development of a penetrative foliation and folding (Fig. 2e), rather than thrusting. No discrete shear zones have been recognized (so far).

Farther to the southeast, outside the area indicated in Figure 1, a sudden increase in the abundance of large gabbroic intrusions suggests the probable existence of an additional lithotectonic unit.

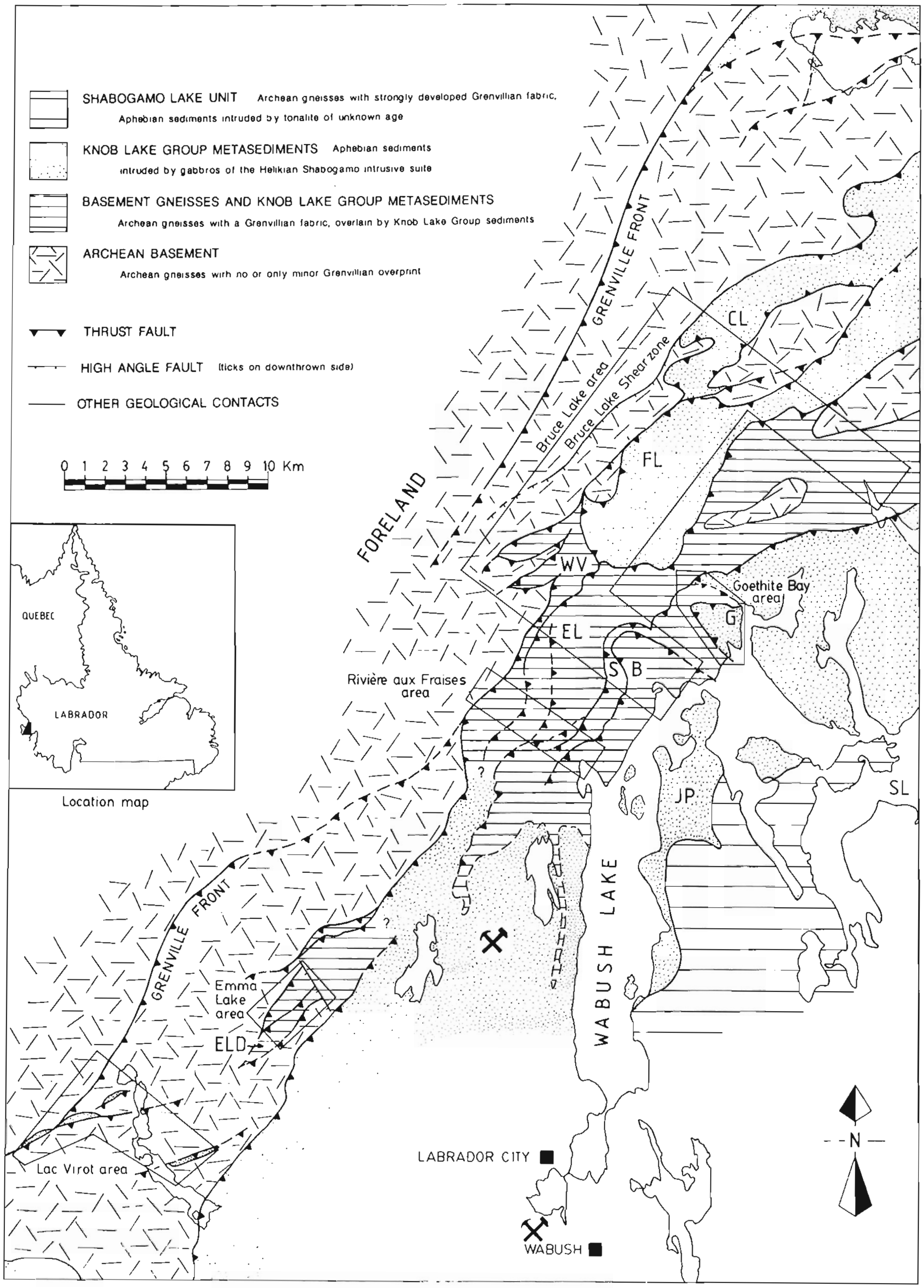

Figure 1. Location of the four map areas and the Bruce Lake area (see van Gool et al., 1987) in the Grenville Front zone in western Labrador. A distinction is made between thrust sheets containing only basement rocks, those containing only rocks of the sedimentary cover, those containing both and the Shabogamo Lake unit.
B = Bondurant thrust sheet, CL = Corinne Lake thrust sheet, EL = Elmer Lake thrust sheet, ELD = Emma Lake Duplex, FL = Flatrock Lake thrust sheet, G = Goethite Bay thrust sheet, S = Sokoman duplex, WV = Wide Valley thrust sheet. JP = Julienne Peninsula, SL = Shabogamo Lake. Map compiled and partly re-interpreted from Rivers, 1985a,b,c; van Gool et al., 1987.

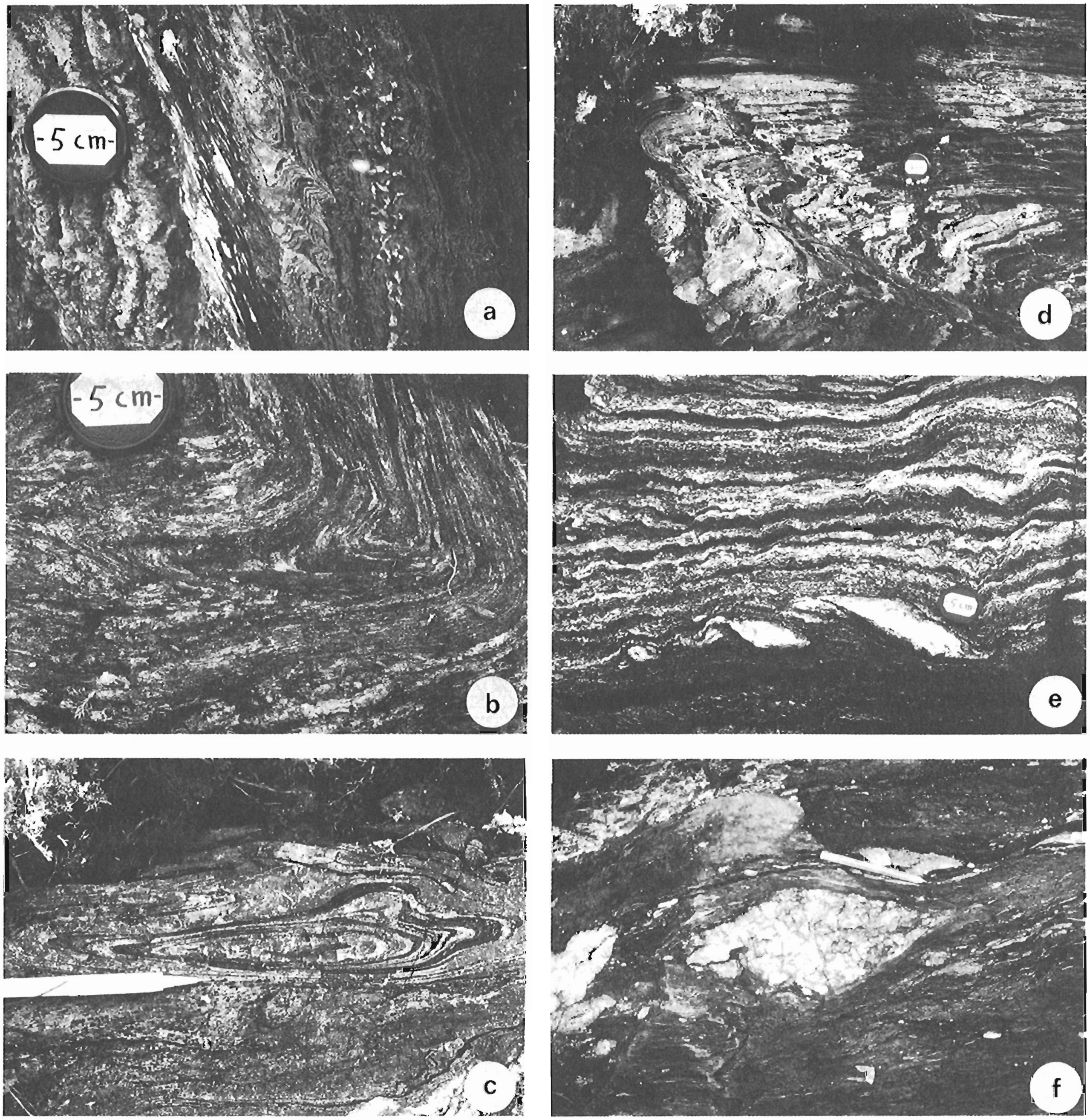

Figure 2. Examples of small scale structures. a) F_2 deformation as open crenulations, closely spaced axial plane schistosity, crenulation cleavage and tight crenulations without axial plane cleavage, in layers of different composition in the Sokoman Formation. Steep limb of an F_2 anticline (Flatrock Lake thrust sheet). b) Overprinting of two folding phases in a basement shear zone (Lac Virot area). c) Sheath fold in a shear zone in the Sokoman Formation (Lac Virot area). d) Sheared northwest verging fold, as small scale example of the fold nappes that exist on a larger scale (Flatrock Lake thrust sheet). e) Deformed metasediments (Attikamagen Formation) of the Shabogamo Lake Unit. The quartz-feldspar lenses in the bottom of the figure are isolated fold hinges. An older fabric is transposed to the presently dominant foliation. Younger deformation caused the open F_3 folds. f) Low strain eye-shaped lense of granitic composition in a basement mylonite (Goethite Bay area)

STRUCTURES

The structural development of the area is dominated by the thrust movement. All structures, with the exception of a late open folding (F_3), are caused by the northwest directed shearing. In Figure 2 some examples are shown of small scale structures. At least two generations of folding, F_1 and F_2, can be recognized using overprinting relationships (Fig. 2b). These folds are approximately of the same age as the shear zones. All folds are highly asymmetric and generally northwest-verging. In zones of high shear strain, however, folds have become highly non-cylindrical, through reorientation of their fold axes, resulting in sheath folds (Fig. 2c) and fold axes (sub-)parallel with the extension lineation. The latter situation can also arise where folds have developed above lateral ramps. Characteristics of the two fold phases are described in more detail by van Gool et al. (1987). They are interpreted to represent deformation during the early and later stages of thrust movement. They are not clearly separated in time and are not regionally correlable folding events (cf. Coward and Potts, 1985). Relationships between folds and lateral or frontal ramps can be indicated locally as, for example, in the Emma Lake area. Figure 2d shows on a small scale the relationship between folding and thrusting. Locally, especially in the southern part of the Flatrock Lake thrust sheet, fold nappes comparable to the structure in Figure 2d developed, but on a scale of tens of meters.

The thrusts in the area are developed as ductile simple shear zones, marked by the development of a penetrative foliation and a stretching lineation. High strain shear zones in the basement rocks are often developed as S-C mylonites. In this rock type eye-shaped lenses of granitic composition, more competent than the surrounding matrix, represent remains of pulled apart granitoid layers (Fig. 2f). The Sokoman and Wishart formations develop a platy fabric in the shear zones. In the Menihek Formation the thrusts are only recognizable by the development of a shear band cleavage, since these schists are already well foliated in low strain zones.

Kinematic indicators, such as folds that are related to the thrusting, tailed porphyroclasts and S-C relationships, all indicate thrust movement towards the northwest. The foliations in the shear zones typically dip between 20° and 40° to the southeast. The dips of the shear zone boundaries are somewhat shallower. Stretching lineations have a consistent strike, of between 150° and 170°. However, locally two stretching lineations at angles up to 30° are observed within one shear plane. This indicates that local deviations of the general movement direction did occur.

A distinction can be made between the floor thrusts of the thrust sheets, which have accumulated considerable strains and are often hundreds of metres wide, and the subsidiary thrusts within the sheets, which are narrower and show lower strains. The subsidiary thrusts diverge upwards from shallowly dipping floor thrusts and form imbricate stacks. Although the subsidiary thrusts are only locally observed to rejoin in a roof thrust (eg. in the Emma Lake and Sokoman duplexes), it is assumed that most of the thrust sheets are actually duplexes. In the southern part of the Flatrock Lake thrust sheet some of the thrusts die out in overturned west verging folds (Fig. 2d). The stacking within the thrust sheets generally took place in a piggy-back fashion. However, out-of-sequence thrusts also occur, generally carrying basement rocks in the hanging wall.

The structural style changes considerably through the belt. The undeformed basement sheets in belt I are overlain by thrust sheets that are internally deformed. In belt II the deformation is restricted to fairly narrow shear zones, up to a few metres in width and to strongly folded zones. Intervening zones consist of gently dipping, relatively undisturbed sequences. In the higher thrust sheets the deformation is less constrained to narrow zones. A pervasive foliation is developed throughout the units, although in places quite large low strain domains still occur. The shear zones in this part of the area are much wider and sequences of several hundreds of metres with fabric elements indicating a component of simple shear are common. In the areas of highest metamorphic grade, the Shabogamo Lake Unit and the mine area west of Wabush Lake, large scale folding seems to predominate over thrusting (Rivers, 1983a). This variation in structural style in the thrust belt appears to be a function of not only the metamorphic grade and rheology of the rocks, but also to a large extent of the position in the thrust stack, as the internal deformation signature in the basement of the frontal thrust sheets is similar west of Flatrock Lake (lower greenschist facies) and near Lac Virot (amphibolite facies) (Fig. 1.). The accumulated strain of the rocks, which is progressively larger towards the internal part of a thrust belt (Platt, 1986), therefore appears to play an important role in determining the structural style.

The Bruce Lake shear zone (Fig. 1) is a large scale extensional, high- angle fault. It is a 100 to 200 m wide ductile shear zone that can be traced for at least 15 km along strike. It affects the floor thrust of the Corinne Lake thrust sheet and must be a late feature (van Gool et al., 1987). Extensional faults have also been recognized on a smaller scale.

Large scale open folding (F_3) on southeast trending fold axes is another late structural development. The intensity of this deformation increases to the southeast (Rivers, 1983a). Only locally F_3 folds can also be recognized on a small scale (Fig. 2e).

METAMORPHISM

The reverse metamorphic zonation typical for sequences that are telescoped in a thrust stack, is expressed in this area in the increase of metamorphic grade towards the higher thrust sheets in the east. Because of the tilt of the area, the isograds cut obliquely through the belt, which causes an increase in metamorphic grade along strike to the southwest as well as across strike of the belt. The metamorphic grade in the Flatrock Lake thrust sheet is lower greenschist facies, whereas the grade in the Lac Virot area, which lies closer to the Grenville Front is upper greenschist to lower amphibolite. The Mont Bondurant and Goethite Bay thrust sheets, as well as the area east of Wabush Lake, are in amphibolite facies. Rivers (1983b) discussed the Grenvillian metamorphic zonation in western Labrador in more detail. With only a minor proportion of metapelites in the mapped areas, diagnostic mineral assemblages are scarce. Criteria that can be used in the field are the transition from the muscovite + chlorite

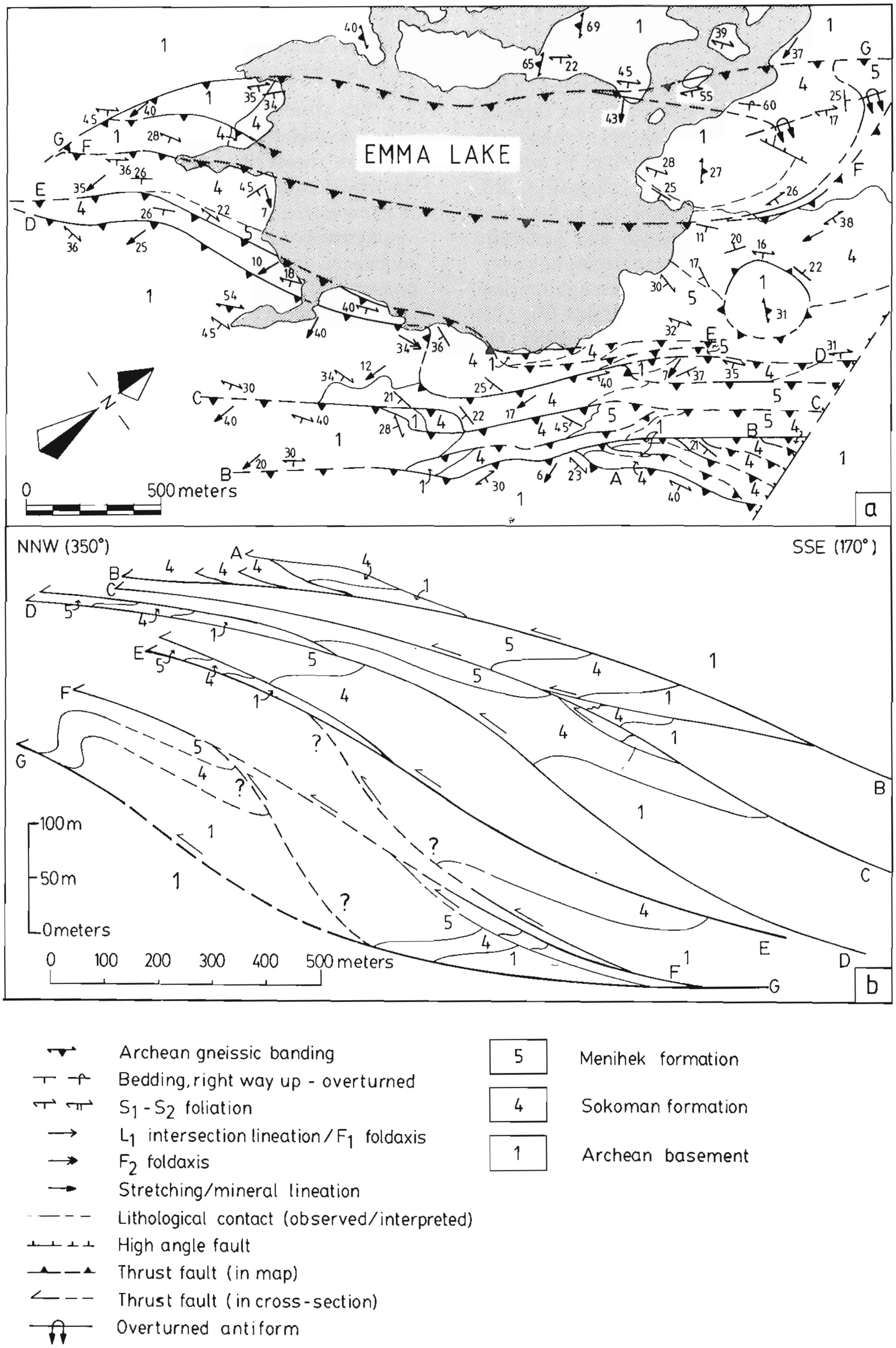

Figure 3. Structural map (a) and cross section (b) of the Emma Lake area. Cross section is a down plunge projection on a vertical plane striking 170°, parallel with the stretching lineation. Vertical exaggeration 2x. Note the difference in scale between section and map. Location of the area is indicated in fig.1. The thrusts marked « ? » in (b) are inferred and are not indicated in the map. A to G mark the main thrusts for easier comparison of cross section and map.

to the muscovite + chlorite + biotite assemblage in metapelites or retrograde, deformed basement rocks (west of the Corinne Lake and Flatrock Lake thrust sheets). At higher grades biotite and chlorite no longer co-exist and the assemblage biotite + garnet appears (Emma Lake duplex, eastern side of the Elmer Lake thrust sheet and higher units). In the Goethite Bay thrust sheet the co-existence of kyanite and garnet in pelitic layers in the Wishart Formation indicates amphibolite grade. In iron formations the reaction of quartz and siderite to form grunerite can be used to determine increases in metamorphic grade.

EMMA LAKE AREA

Figure 3 is a geological map and cross section of the Emma Lake area. The plunge of the structures is towards the northeast. The map pattern shows an oblique section through the imbricate thrust stack. Although there is very little topographic relief in the area, the oblique section provides an opportunity to reconstruct a three dimensional image of the belt. A cross section was constructed by the down plunge projection method (projection axis plunges 10° to 080°). Because data from the whole area are projected onto one plane, the result is somewhat distorted by the non-cylindricity of some of the structures, such as lateral ramps. The figure is therefore not representative for a section along any one plane, but it clearly shows the overall geometry of the duplex.

The structure involves rocks of the basement, Sokoman and Menihek formations. The horses are thickest where they contain basement rocks, but they are all less than 100 m thick, generally much thinner. The duplex itself is less than 400 m in true thickness. The duplex is sandwiched between two large thrust sheets that carry almost pristine Archean basement. To the northeast the duplex is cut off by a high-angle fault which juxtaposes the thrust stack against another large basement block. In some of the horses of the duplex folds formed above hanging wall ramps, producing a fold nappe geometry (cf. Coward and Potts, 1985). The best example is the hanging wall anticline in the front of the lowest sheet (Fig. 3b). Folds that are related to lateral ramps occur as well, in most cases as F_2 folds (eg. in the small duplex on top of the Emma Lake duplex, in the east corner of Fig. 3a).

The stretching lineations in the duplex consistently plunge towards approximately 170°. The movement on the thrusts was towards 350° at a fairly small angle to the present strike of the thrusts. All thrusts are developed as ductile shear zones with penetrative L-S and/or C-S fabrics.

GOETHITE BAY AREA

The map and a cross section of the Goethite Bay area are shown in Figure 4. The area is dominated by a repitition of Wishart and Sokoman formations, consisting of massive, pure quartzites and mainly Fe-oxide bearing quartzites, respectively, which form the Goethite Bay thrust sheet. This repitition, interpreted as an imbricate thrust stack, is cross cut by several out-of-sequence thrusts, carrying highly sheared basement rocks. Compared with the Emma Lake duplex, the rocks in this area are much less folded and deformation is concentrated in the shear zones, implying that crustal shortening was accommodated by thrusting rather than folding. Thicknesses of the thrust slices are in the order of 100 m. One of the higher slices, consisting of Wishart quartzites, seems to be much thicker. It is likely, however, that this sheet is internally cut by thrusts that are not exposed, repeating one section of quartzite several times. The main part of the thrust sheet is bounded in the west by a steep fault, which is interpreted as a lateral ramp. However, exposure is not continuous enough to reject the possibility of it being a tear fault, which also intersects the lower thrusts to the north.

ECONOMIC SIGNIFICANCE

The findings of this project can have long term implications for the iron ore mines in the region. Although folding predominates over thrusting in the mine areas, the iron formations lie within a thrust sheet and must be truncated at depth by the floor thrust of the sheet.

Abundances of iron ore minerals (hematite, magnetite and goethite) that are of economical value appear in the quartzite iron formations only. Outside the mine areas, the southern part of the Flatrock Lake and the Goethite bay thrust sheets contain ore grade rocks in large quantities. Compared to the extent of ore bodies in the mines, however, these two occurences are of minor importance.

Deposition of gold can be related to the thrusts in the area. Shear zones with a phylonitic texture can concentrate gold if large amounts of fluid pass through them. Quartz lenses are found in close proximity to the shear zones. They reach considerable sizes in the area (over 100 m in length) indicate major hydrothermal activity. In some of these bodies gold associated minerals, such as pyrite, pyrrhotite and arsenopyrite, have been concentrated. Chemical analyses are being carried out on samples from suitable rock types.

DISCUSSION

The frontal zone of the Grenville Orogen in western Labrador is developed as a foreland thrust belt. Lithologies that are equivalents of those that occur in the foreland have been affected by Grenvillian deformation and metamorphism and transported to the northwest. The four lithotectonic belts that have been distinguished form a telescoped thrust stack with an inverted metamorphic zonation. Towards the east a progressively larger part of the stratigraphic sequence is incorporated in the thrust stack. The thrust sheets of belt II contain only the upper two stratigraphic units, the Sokoman and Menihek formations. Towards the southeast the stratigraphically underlying Attikamagen, Wishart and Denault formations also appear in the stack. This variation is a result of the original onlap of the sedimentary sequence on to the basement and implies that the more distal parts of the continental margin stratigraphy are present only in the structurally higher thrust sheets.

Although many of the characteristics of the thrust model of Boyer and Elliott (1982) can be recognized in the belt, the model does not completely fit the Grenville Front thrust belt. The main difference lies in the fact the sole thrust does not climb up into the cover rocks towards the foreland in the presently exposed part of the belt. It developed completely within the basement, and delamination must have taken place at considerable depth. Furthermore, the position of the

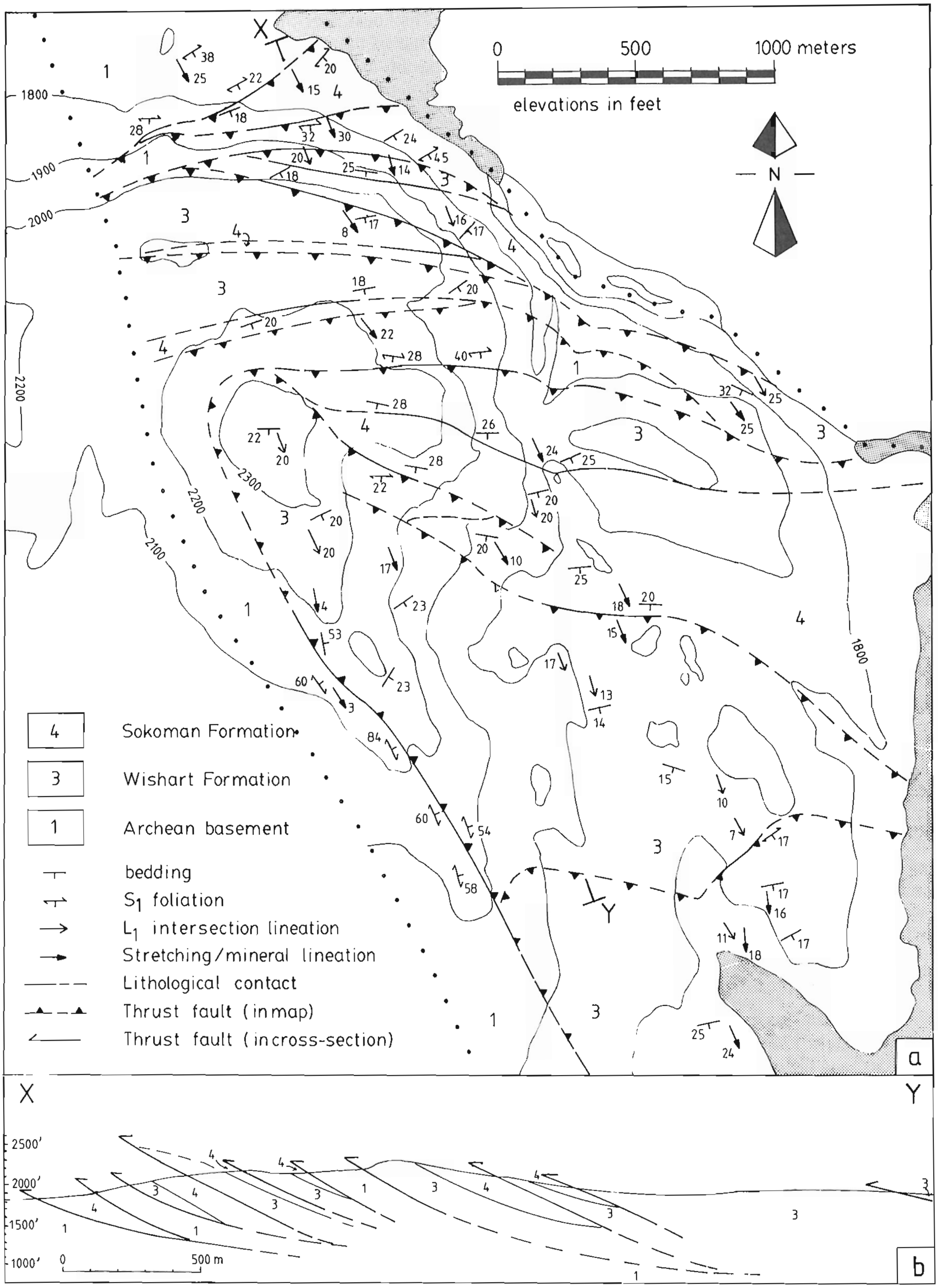

Figure 4. Structural map (a) and schematic cross-section (b) of the Goethite bay area. Location of the area is indicated in Figure 1. No vertical exaggeration in cross-section. The lowest indicated thrust forms the boundary between the Elmer Lake and Goethite Bay thrust sheets.

basement-dominated belt III (Elmer Lake thrust sheet) between the two sediment-dominated belts II and IV cannot be explained by a simple model of northwestward propagating thrusts that are stacked in a piggy-back sequence.

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Gamma spectrometric and magnetic anomalies associated with Cu-U mineralization, Faber Lake volcanic belt, District of Mackenzie, N.W.T.

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Charbonneau, B.W., Gamma spectrometric and magnetic anomalies associated with Cu-U mineralization, Faber Lake volcanic belt, District of Mackenzie, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 255-258, 1988.

Abstract

A north-trending belt of volcanic rocks occurs in the southern Great Bear Magmatic Zone in the Northwest Territories. This belt is some 100 km long by 5 km wide. Three significant mineral occurrences are located along this trend. Two of them are Cu-U in magnetite-rich veins and breccia filings, believed to be hydrothermal. One of these also contains Ag, Au, Se and Bi. The third occurrence is a minerologically simple vein deposit in a fracture zone, and is a past producer. These occurrences were discovered over a time span of four decades. Each individual occurrence was discovered principally utilizing data from one of three different programs of the Geological Survey of Canada (1) geological mapping, (2) aeromagnetics and (3) airborne gamma ray spectrometric surveying, and consequently they present interesting examples of the direct use of federal work in mineral exploration.

Résumé

Une zone de roches volcaniques se dirigeant vers le nord se trouve dans le sud de la zone magmatique de Great Bear dans les Territoires du Nord-Ouest. Cette zone mesure quelque 100 km de long sur 5 km de large. Trois gisements de minéraux importants ont été localisés le long de cette direction. Dans le cas de deux d'entre eux, il s'agit de gisements de Cu-U logés dans des filons et des brèches filoniennes riches en magnétite et d'origine probablement hydrothermale. L'un des deux contient aussi les métaux suivants: Ag, Au, Se et Bi. Le troisième gisement est un filon minéralogiquement simple situé dans une zone fracturée, et a déjà fait l'objet de travaux d'exploitatgion. La découverte de ces gisements s'est échelonnée sur quatre décennies. Chacun des gisements a été découvert en utilisant les données recueillies dans le cadre des trois programmes suivants de la Commission géologique du Canada: 1) la cartographie géologique, 2) les levés aéromagnétiques et 3) les levés aériens de spectrométrie des rayons gamma. Ils constituent donc des exemples intéressants de l'utilisation directe de résultats de travaux fédéraux en exploration minérale.

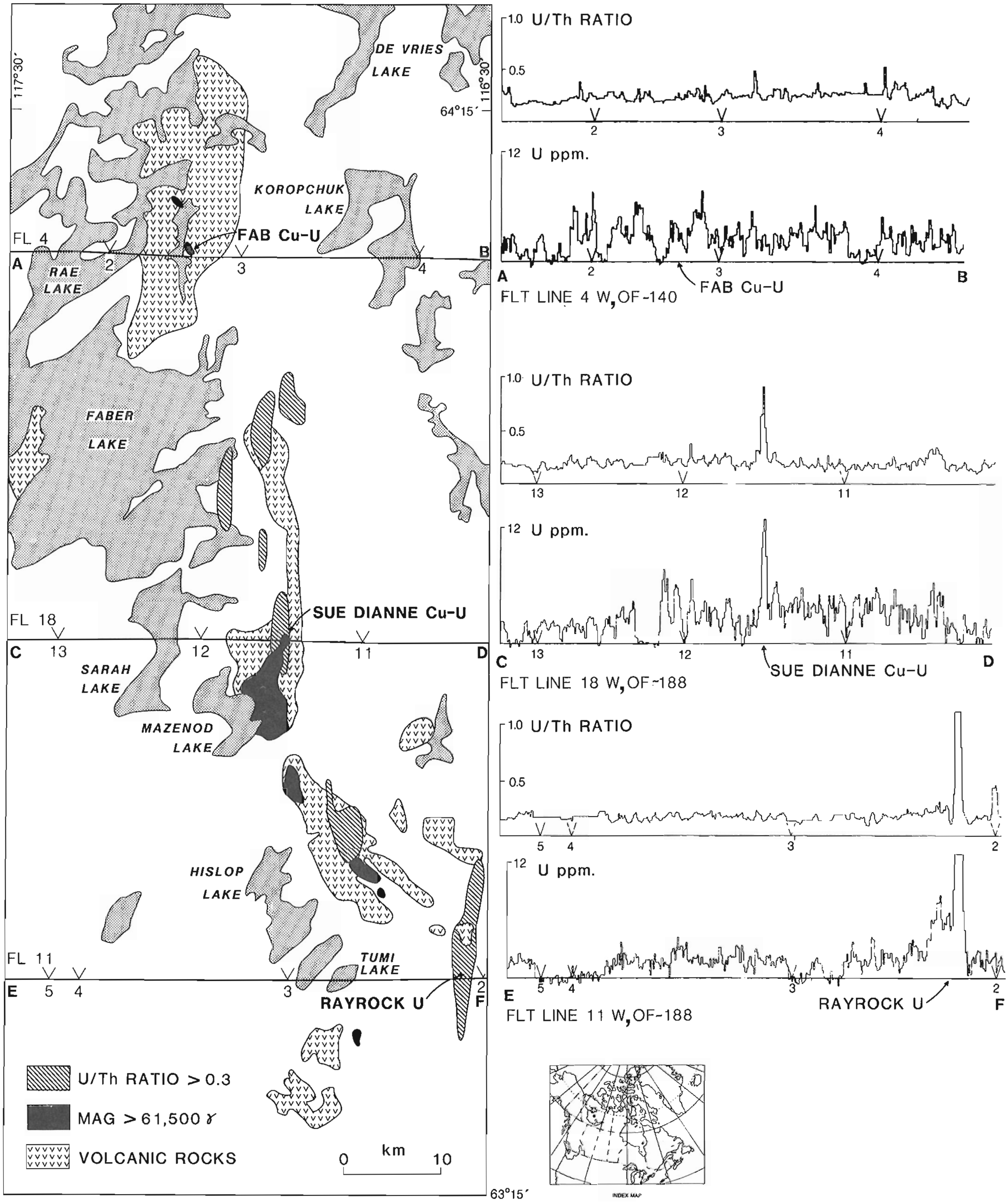

Figure 1. Geophysical sketch of the Faber Lake volcanic belt showing major mineral occurrences, Northwest Territories.

INTRODUCTION

A belt of predominantly volcanic rocks is located between latitudes 63°15′ and 64°15′ north near longitude 117°W (Fig. 1). Several Cu and U occurrences are known along this belt, three of which are relatively more significant, and their geophysical expression is of particular interest. A more detailed account of the geology and mineralization in this belt can be found in this volume (Gandhi, 1988), and hence these aspects are only briefly touched upon here.

This paper complements Gandhi's, provides a brief geophysical discussion, and also documents the part that GSC regional programs (geological mapping, aeromagnetics and gamma spectrometry) played in discovery of these mineral occurrences.

GEOPHYSICAL PATTERN - GEOLOGY - MINERAL OCCURRENCES

Figure 1 summarizes the geology, aeromagnetics and gamma spectrometric pattern of the area. The geology is simplified from mapping by Kidd (1936), Lord (1942), Wilson and Lord (1942), Fraser (1967) and McGlynn (1979).

The volcanic rocks, principally quartz feldspar porphyry and fragmental volcanics, are surrounded by younger granites. The aeromagnetics on Figure 1 are derived from Geological Survey of Canada (GSC) Geophysical series map 7203G north of 64°N and 7197G south of 64°N. The gamma spectrometry patterns north of 64°N are derived from GSC Open File 140 (Richardson et al., 1973) and south of 64°N from GSC Open File 188 (Richardson et al., 1974). Airborne gamma spectrometric profiles for U and U/Th ratio along flight lines that pass over or very close to the three major mineral occurrences in the belt are also shown on Figure 1.

The volcanic rocks are in places marked on the geophysical maps by higher U/Th ratios (>0.3) approximately 50 % above the average crustal values according to Clarke et al., (1966) and distinctly high aeromagnetic response ($>61{,}500\gamma$). These values are more than 1,000γ (nT) above local background.

The RAYROCK "U" property was discovered well before the geophysical surveys were published. It is marked by a well-defined U and U/Th ratio anomaly (pitchblende) and occurs on a magnetic break corresponding to a major northeasterly fault. The sequence of events leading to the discovery of this deposit commenced when uranium staining was located by officers of the Geological Survey of Canada in 1934, as described in Lang et al. (1962).

The FAB Cu-U occurrence (Morris, 1977), farthest north on the belt, was staked in 1968 by a joint venture of Shield Resources Limited, and Numac Oil and Gas Limited, based on their radiometric survey that showed anomalies coincident with an aeromagnetic anomaly shown on Geophysical Map 2939G which is part of Aeromagnetic Compilation Map 7203G (Geological Survey of Canada, 1969). As shown on Figure 1, there is no expression on the radiometric profile for this occurrence because the nearest flight line did not pass close enough to the exposed mineralization. This circumstance can be expected based on the regional 5 km spaced flight lines. More detailed airborne scintillometer surveys employed by industry however did detect a radiometric response and the knowledge of the aeromagnetic anomaly did play an important part in the exploration strategy (Curry, 1969a, b).

The SUE DIANNE Cu-U occurrence was staked privately and transferred to Noranda Exploration Company Limited in 1974 (Climie, 1975, 1976). The staking was based on the airborne gamma spectrometric anomaly which was in fact indicated as especially noteworthy in the marginal notes of GSC Open File 188. This sharp, well-defined U and U/Th ratio anomaly corresponded with an aeromagnetic anomaly as well. This occurrence has been drilled, and substantial tonnages containing Cu and U are indicated (Climie, 1976). Presence of small amounts of Ag, Au, Se and Bi are reported (Climie, ibid; Miller, 1982).

The geophysical signature indicated for the FAB and SUE DIANNE occurrences is consistent with their genetic nature involving hydrothermal magnetite concentrations containing principally uranium and copper (Gandhi, 1988).

DISCUSSION

The fact that geological, geophysical, geochemical programs provide a regional data base that greatly facilitates mineral exploration is well accepted. This brief report presents an interesting record of discovery of three significant mineralizations each based on a different program of the GSC - namely geological mapping, aeromagnetics and gamma spectrometry, in a belt which otherwise may not have attracted exploration interest. The time frame is also extremely interesting in that the three discoveries took place over four decades pointing out the long term viewpoint that must be taken when evaluating earth science data bases.

ACKNOWLEDGMENTS

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Geophysical expression of the carbonatites and fenites, east of Cantley, Quebec

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Charbonneau, B.W. and Hogarth, D.D., Geophysical expression of the carbonatites and fenites, east of Cantley, Quebec; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 259-269, 1988.

Abstract

Integrated airborne geophysical surveys (gamma spectrometry, magnetometry, VLF) over an area of carbonatites and fenites east of Cantley, Quebec present an excellent example of the varied geophysical signature which can be expected in these rocks.

The Quinnville carbonatite and surrounding fenite zone is only slightly above the background of uranium and thorium in the regional Precambrian gneisses and marble. However a well-defined potassium alteration zone of several square-kilometres characterizes the local radioactivity. Bedrock potassium values attributed to a rock composed almost exclusively of microcline often exceed 10 % K, with maximum values of about 12 % K or 15 % K_2O. This feature is nonmagnetic due to the presence of hematite rather than magnetite. In contrast, the nearby Templeton Fenite Zone is anomalously magnetic because of magnetite which is present in both fenite and carbonatite. However these rocks are not appreciably radioactive.

Résumé

Les levés géophysiques aériens intégrés (spectrométrie des rayons gamma, magnétométrie, TBF) d'une zone de carbonatites et de fénites à l'est de Cantley (Québec) présentent un excellent exemple des signatures géophysiques variées auxquelles on peut s'attendre de ce type de roches.

Dans la zone de carbonatites et de fénites environnante de Quinnville, on n'enregistre des signaux que légèrement supérieurs au bruit de fond en uranium et thorium enregistrés dans les gneiss et marbres précambriens régionaux. Cependant, une zone de transformation en potassium bien définie, couvrant plusieurs kilomètres carrés, caractérise la surface radioactive locale. Les valeurs en potassium du socle, attribuées à une roche composée presque exclusivement de microcline, dépassent souvent 10 % en K avec des valeurs maximales d'environ 12 % de K ou de 15 % de K_2O. Il s'agit de roches non magnétiques à cause de la présence d'hématite plutôt que de magnétite.

Au contraire, la zone de fénite voisine de Templeton présente une anomalie due à la présence de magnétite dans les fénites et les carbonatites. Ces roches sont toutefois peu radioactives.

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INTRODUCTION

Precambrian rocks near the Ottawa-St. Lawrence Valley are mainly metasedimentary and include biotite gneiss, pyroxene gneiss, feldspathic quartzite, marble and calc-silicate rock (''skarn'' or ''pyroxenite''). The rocks have been intensely folded and intruded by granite, monzonite, diorite, syenite and gabbro, and thermally metamorphosed during the Grenville orogeny. The orogeny appears to have been polyphase with major granulite metamorphism at about 1100 Ma and a lower grade event at 950 Ma (Baer, 1981; Bell, 1981; Doig, 1977).

These rocks are transected by numerous faults, the result of repeated activation which extended from Precambrian time at least into the Ordovician. Most of these faults are nearly vertical and, in the Ottawa region, belong to west-northwest, east-west and northeast sets. A complex of down-dropped blocks defines the Ottawa Valley Fault System and its northeast-trending arm, the Ottawa-Bonnechere Graben (Fig. 1; Kumarapeli, 1978; Kumarapeli and Saul, 1966).

The availability of airborne gamma ray spectrometric and aeromagnetic data along much of the St. Lawrence System, provides an ideal data base for study of the fracture system and the associated intrusive rocks, in particular the carbonatites (Currie, 1976; Lumbers, 1982). Gravity surveys are also available covering the St. Lawrence Fault System (Earth Physics Branch, 1980) and regional anomalies can be correlated with part of the structure (Thompson and Garland, 1957; Forsyth, 1981).

Other carbonatites along the Ottawa-St. Lawrence Fault System have been investigated geophysically by the Geological Survey. At Allan Lake in Algonquin Park near Pembroke, Ontario a previously unknown carbonatite was discovery by ground follow-up of an airborne gamma spectrometric anomaly (Ford et al., in press).

GEOPHYSICAL SURVEYS

During 1983 an airbone geophysical survey (gamma ray spectrometry with magnetometer and VLF-EM) was flown over NTS map sheet 31G, the Ottawa sheet, (Geological Survey of Canada, in press), at standard reconnaissance line spacing of 5 km. The location of this survey is shown on Figure 1. This survey shows a pronounced potassium anomaly which parallels the St.Lawrence Fault System in the area (Fig. 2).

Figure 3 is a stacked profile (flight line 12) which shows the nature of the potassium highs that can be found along this feature. The anomaly (arrow) near Buckingham, Quebec, overlaps the potassic Cambro-Proterozoic volcanic rocks described by Lafleur and Hogarth (1981). An area east of Cantley, Quebec, (Fig. 4) was also surveyed in 1983. Flight lines were spaced at 400 m over an area of 40 km^2 (10 $\times$ 4 km) where carbonatite and fenite have been described (Hogarth and Rushforth, 1986).

The results of the detailed survey are shown on Figures 5, 6 and 7. The aeromagnetic map (Fig. 6A) is the standard aeromagnetic map available for the area (Geological Survey of Canada, 1955). However, the aeromagnetic data shown on the profile (Fig. 7) are from the detailed 1983 survey.

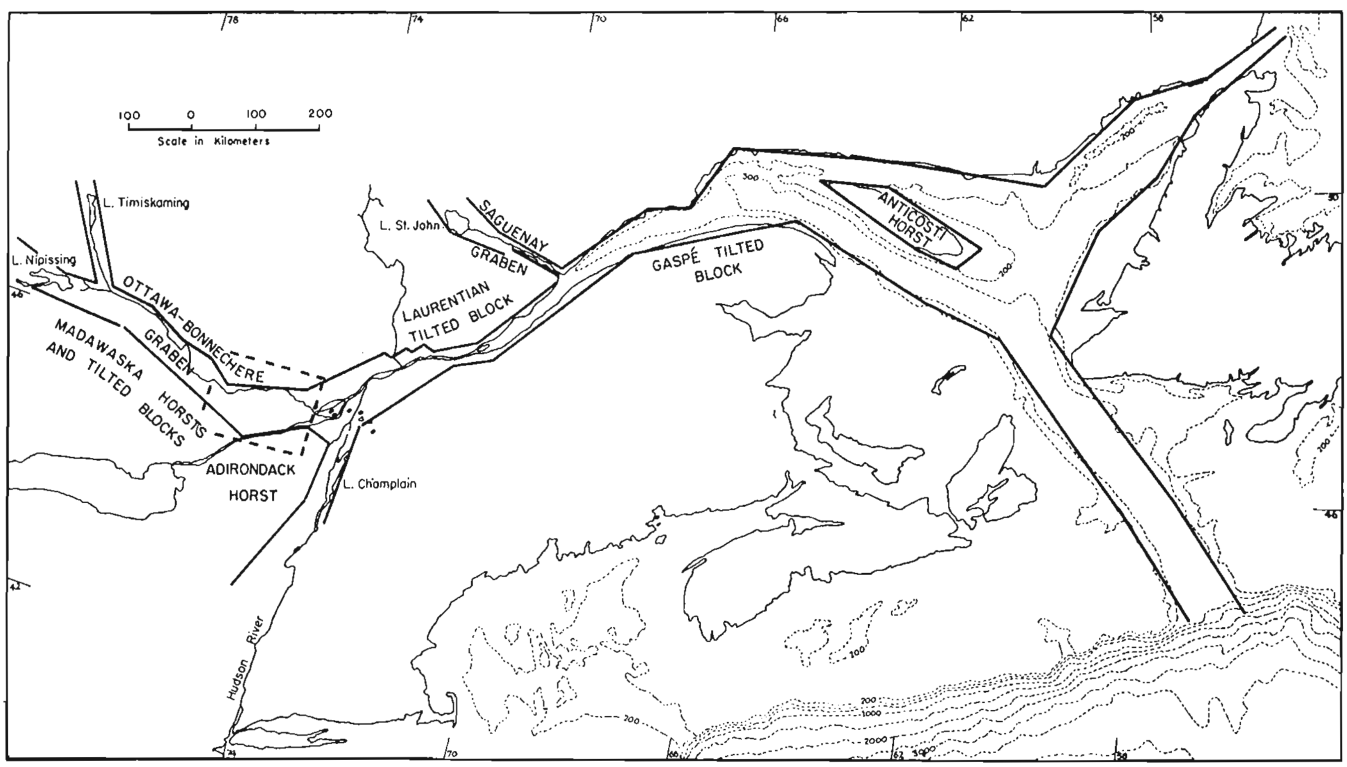


Figure 1. Ottawa — St.Lawrence Valley fault system (after Kumarapeli and Saul, 1966) showing location of map sheet NTS 31G (dashed line).

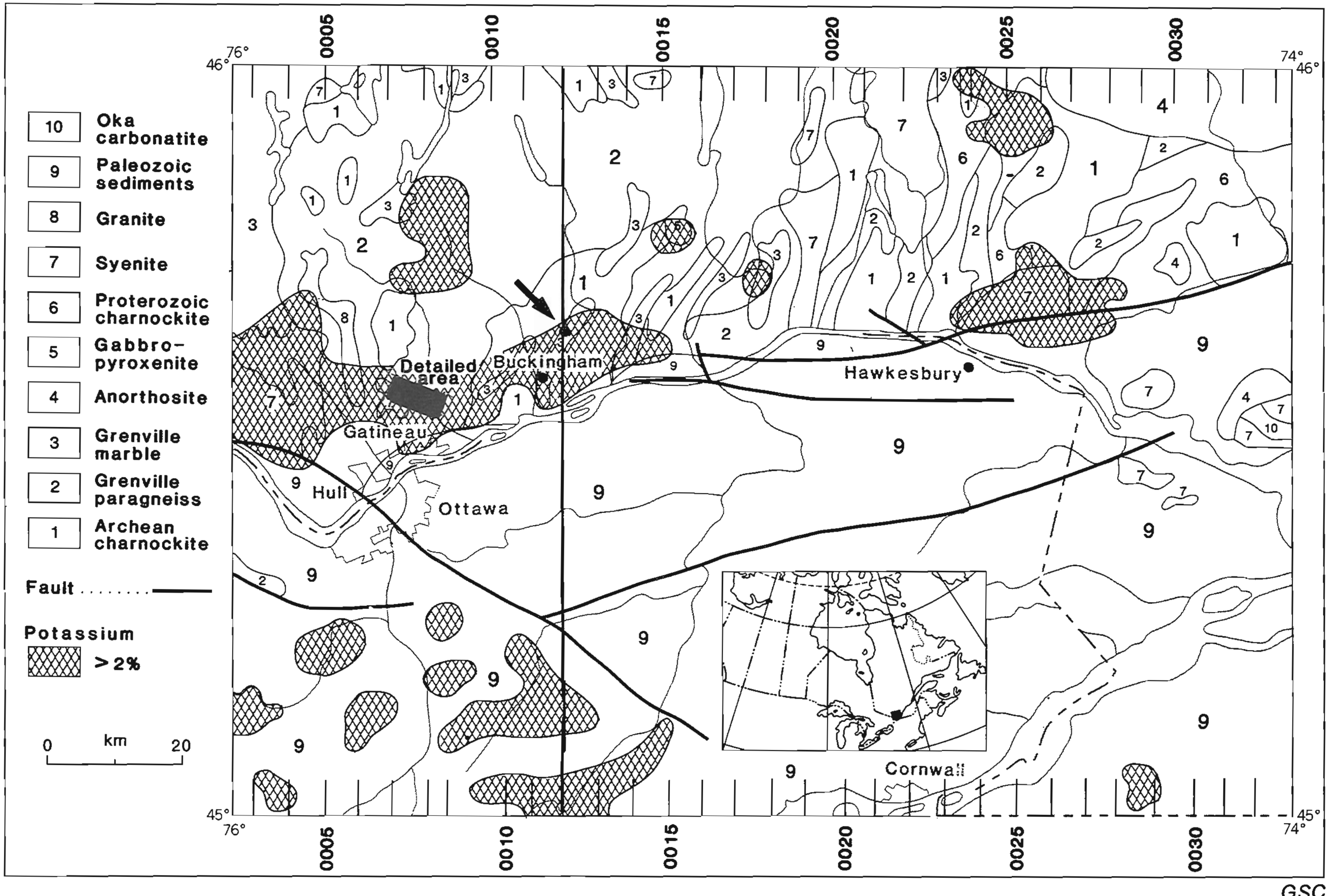


Figure 2. Potassium map from airborne gamma spectrometry survey. Geology from Avramtchev, 1985.

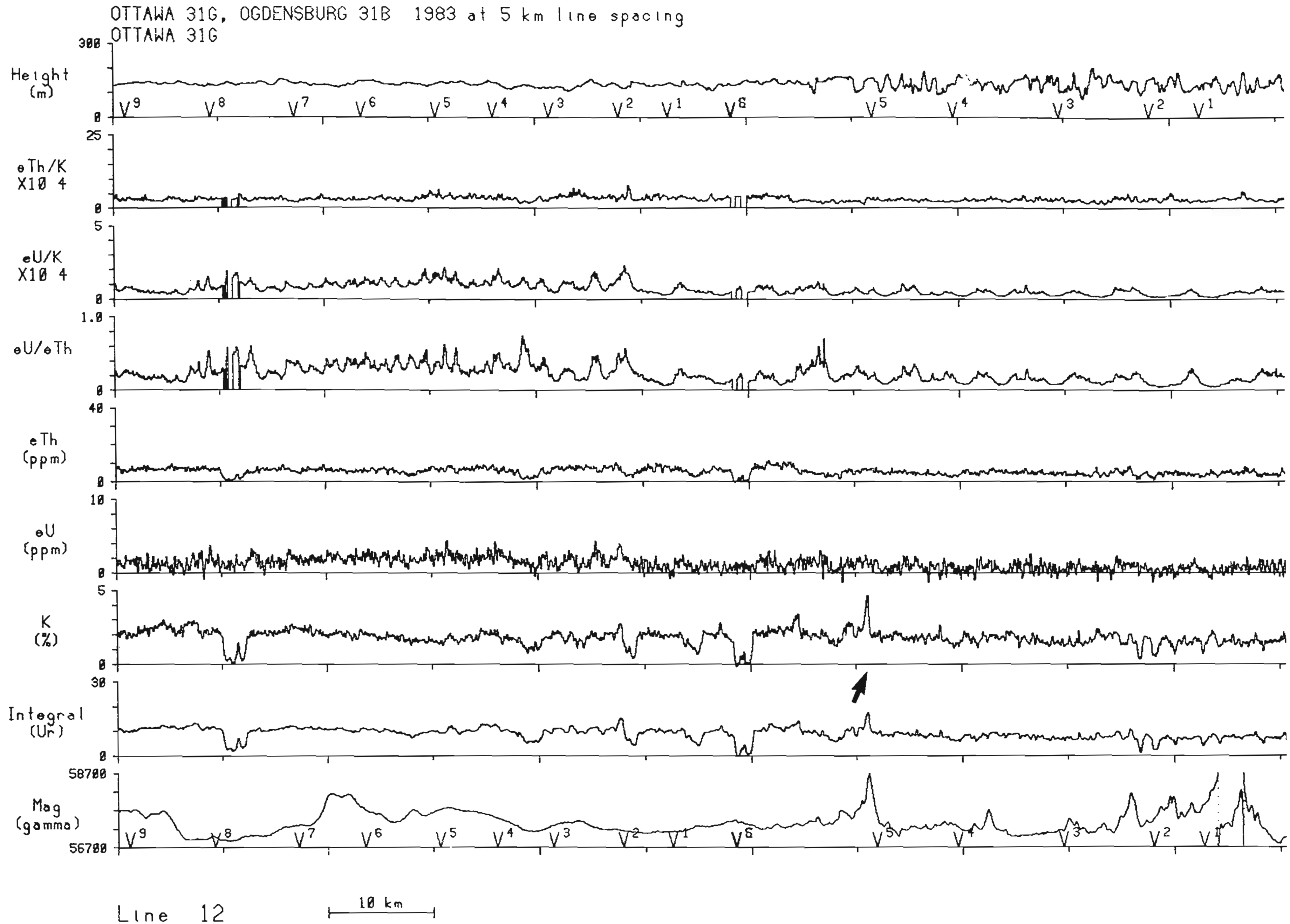


Figure 3. Stacked profile for line 12 from gamma spectrometry survey for NTS 31G.

The results of the detailed survey indicate:

1) an intense potassium anomaly which occupies an area of several square kilometres surrounding the Quinnville carbonatite (Fig. 5A, 5b);

2) the area of potassium high correlates with an aeromagnetic low (Fig. 5A, 5B, 6A, 7);

3) the magnetometer high correlates with the Templeton Fenite Zone (Fig. 5A, 6A, 7);

4) an increase in thorium near and within the carbonatite (Fig. 5A, 7).

VLF data indicate the major faults in the area, particularly fault 'F' on Figure 5A, which cuts the Templeton Fenite but shows no definite correlation with the fenite or carbonatite. The most prominent anomaly on Figure 6B relates to a power line.

Analysis of the magnetometer profile in Figure 7, using the MAGRAV program of Broome (1986) has produced the

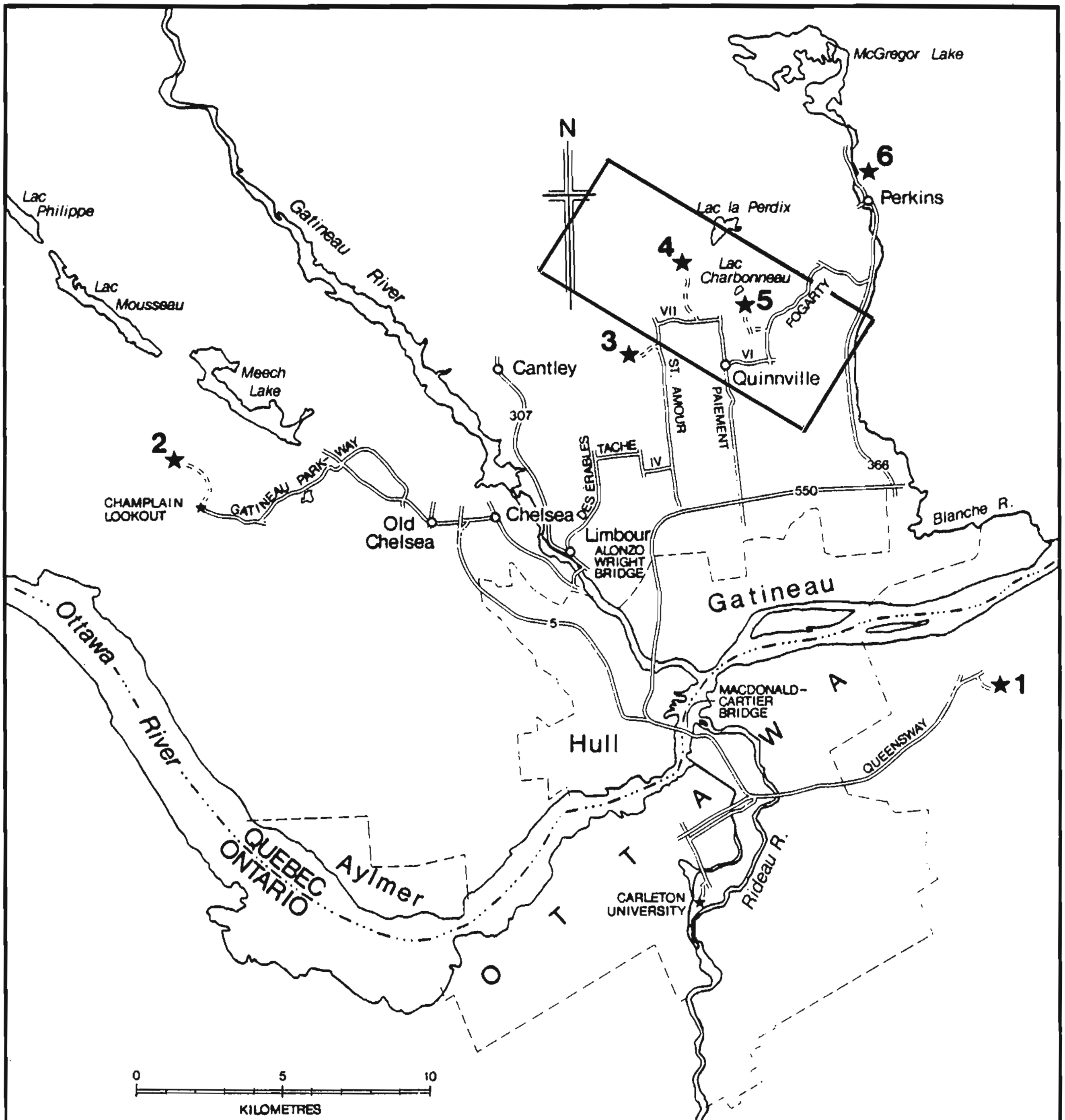
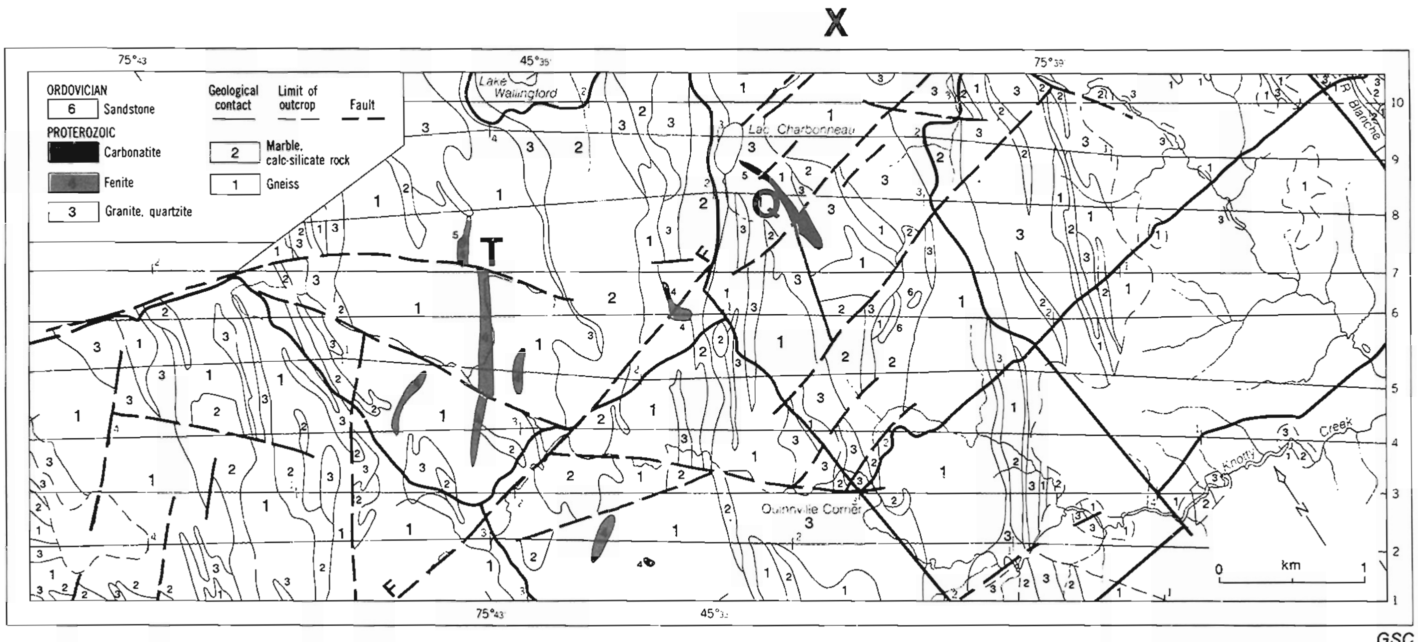


Figure 4. Location of Cantley detailed airborne geophysical survey showing location of (1) Blackburn carbonatite, (2) Meech Lake carbonatite, (3) Haycock fenite, (4) Templeton carbonatite-fenite, (5) Quinnville carbonatite, (6) Perkins fenite. Base map from Hogarth and Rushforth (1986).



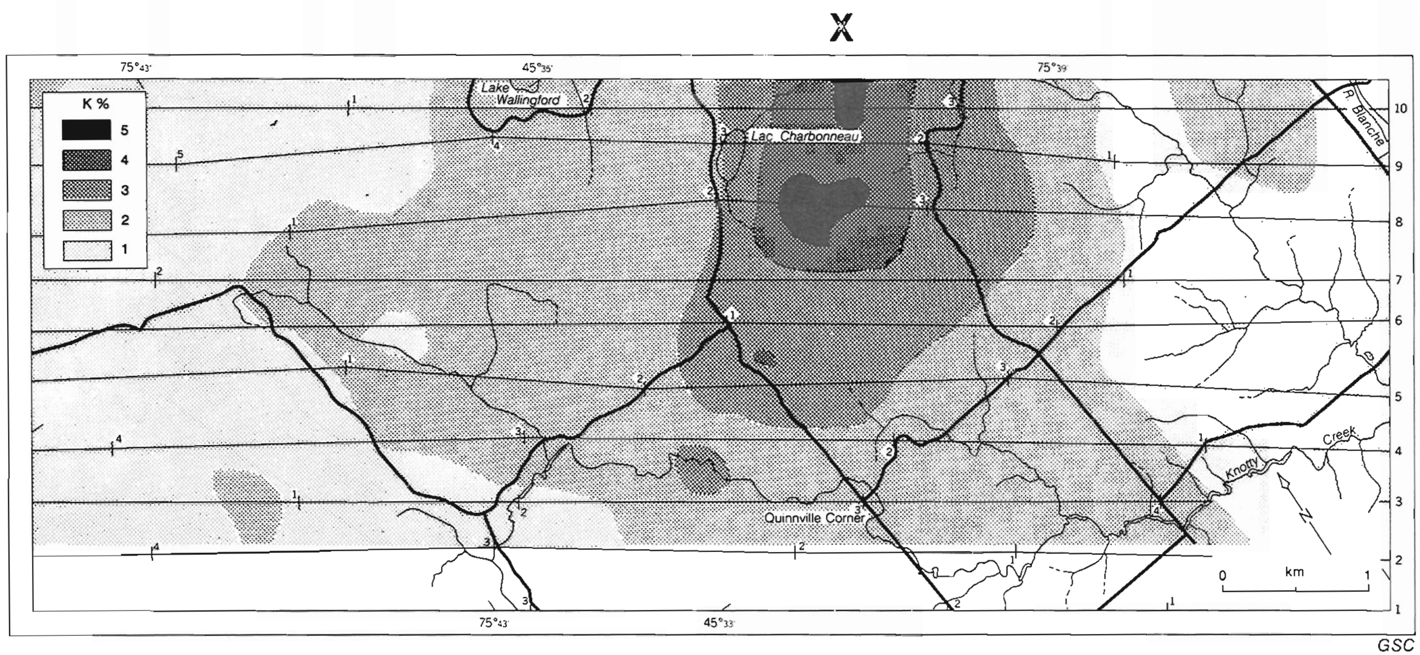


Figure 5(A). Geology of detailed survey area near Cantley, Quebec. Geology by D.D. Hogarth. Q Quinnville carbonatite, T Templeton carbonatite-fenite, H Haycock fenite, X Quinnville north. **(B)** Potassium map of Cantley detailed survey area.

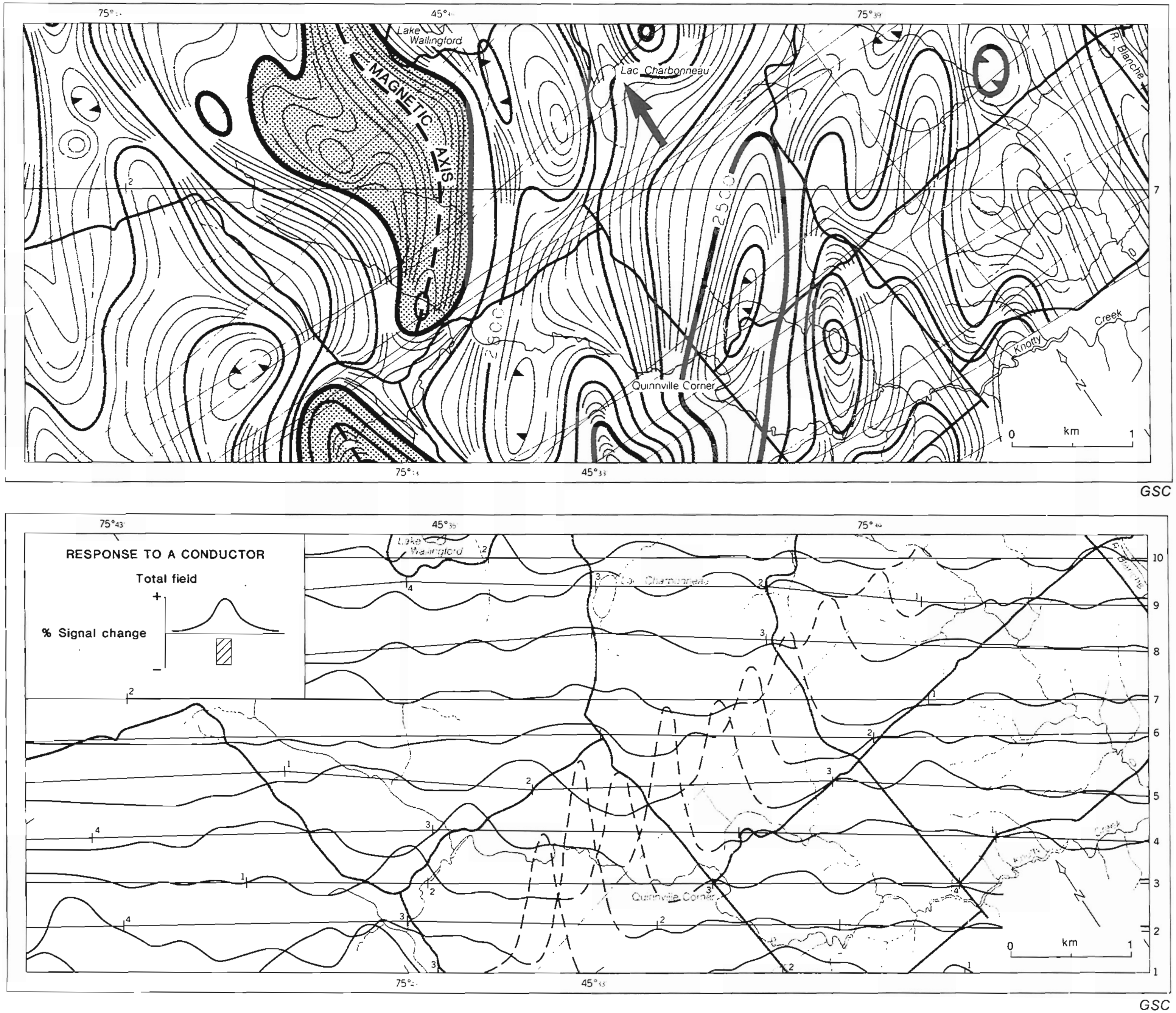


Figure 6. **(A)**Aeromagnetic map Cantley detailed survey area. **(B)** VLF map for Cantley detailed survey area.

interpretation in Figure 8 (courtesy of L.J. Kornik, GSC). The interpretation shows steeply west-dipping Templeton Fenite Zone 'A', and the axis of the carbonatite 'B' correlated with a sharp magnetic trough within the broad magnetic low.

GROUND STUDY AND INTERPRETATION

Ground studies during 1986 and 1987 provided interpretation and verification of the geophysical pattern.

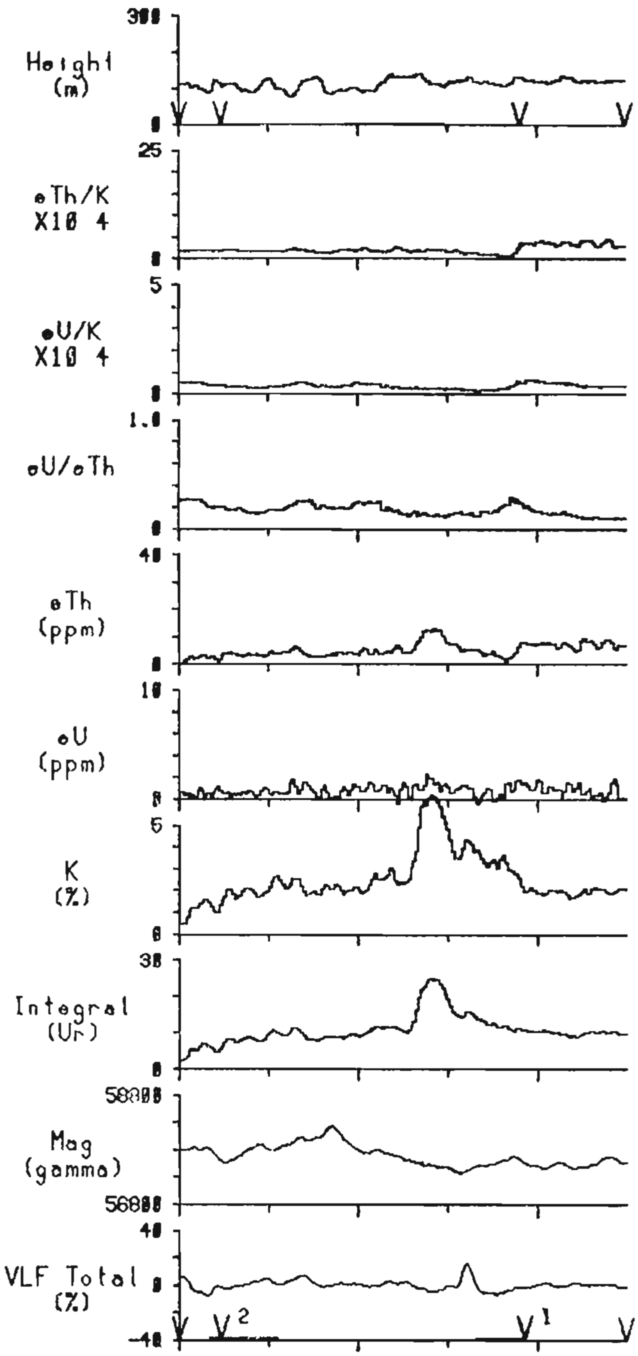


Figure 7. Stacked profile for line 7 from Cantley detailed survey area.

1. The potassium high correlates with bedrock highly enriched in potassium, often exceeding 10 % K, mainly in microcline. The highest levels observed (in situ gamma spectrometry) are in excess of 12 % K or nearly 15 % K_2O (S. Leclair pers. comm.). The fenite appears to be the type described by Heinrich (1966), characterized by K-enrichment dominance of alkali feldspars, but absence of alkali amphiboles and soda pyroxenes.

2. The restricted airborne thorium anomaly relates to bedrock concentrations of up to 50 ppm. The carbonatite itself contains patches exceeding 300 ppm. The thorium sources are very fine grained and occur mainly within wispy hematite-rich areas although there are a few more well defined sources. Figure 9 is an autoradiograph (one week) designed to show the fine wispy nature of the thorium sources in the carbonatite. The radioactive mineralogy was not identifiable in the

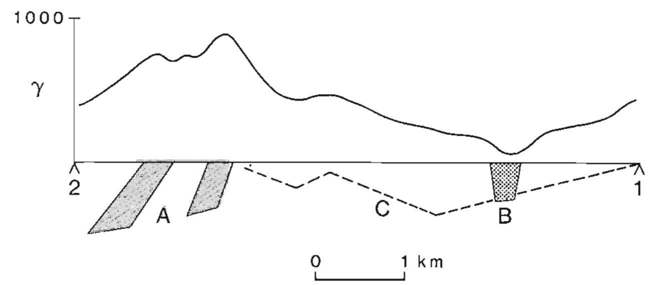


Figure 8. MAGRAV 2 analysis of magnetometer profile from line 7 Cantley detailed survey. A) Templeton fenite, B) Quinnville carbonatite, C) aeromagnetic low.

Figure 9. Autoradiograph (one week exposure) of a thoriferous phase (300 ppm) of the Quinnville carbonatite. Arrow points to wispy hematite rich radioactive area.

wispy areas. Some of the more "discrete" radioactive sources were shown to be mixtures of two or more La-Ce fluorocarbonates (Hogarth et al, 1985).

3. The rocks underlying the potassium high were found to be extensively hematized, thus accounting for the magnetic low with many samples taken from fenitized wall rock near the Quinnville carbonatite showing alteration of original magnetite grains (Fig. 10) however, most of the hematite in these rocks (often present as specular hematite) results from secondary iron introduction.

4. Some fenitess are characterized by appreciable uranium (> 50 ppm) in addition to thorium. In the Quinnville North fenite ('X' in Fig. 5A) uraninite and monazite are found with hematite in an interlacing network of hematite veinlets (Fig. 11). A potassium-rich alkali amphibole (Fig. 12), apparently similar to those described by Hogarth et al; (in press) is closely associated. The large potassium anomaly at Quinnville North may be caused by undiscovered potassium fenites. The significant amount of sulphide, occuring mainly as pyrite along fractures in the fenite (at location 'X') was probably originally present in the rocks and remobilized during fenetization (Fig. 13).

5. The Templeton Fenite Zone, (Fig. 14) characterized by magnesio-arfvedsonite and aegirine (Fig. 14), locally contains disseminated magnetite, which produces a weak magnetic anomaly. The fenite is close to background radiation except on approach to the carbonatite where thorium increases by a factor of 20, and uranium by a factor of 3, compared to the biotite gneiss. Radioactivity within the carbonatite decreases from the contact to the core.

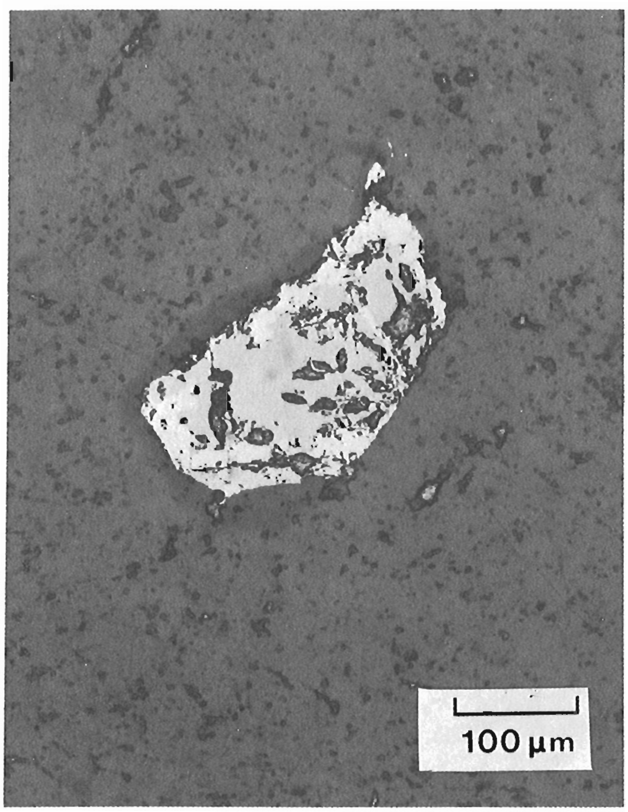


Figure 10. Martization of original magnetite grain — light area surrounding darker magnetite core.

The Haycock Fenite Zone to the southwest, the possible dextral offset of the Templeton, has a similar silicate composition. Some occurrences are cored by solid specularite-rutile-magnetite lenses (Hogarth and Lapointe, 1984) and therefore give a distinct magnetic response. However, except for local monazite-bearing specularite-apatite and magnesio-arfvedsonite lenses and a single parisite-bearing carbonatite dyke (site 4, Hogarth et al; 1985), the Haycock fenites have low radioactivity relative to country rock biotite gneiss.

The occurrence in the same area of two quite distinctive types of fenite the "dark" Na-Mg-Fe and the K rich makes the Cantley area a scientifically interesting and unusual one.

ECONOMIC ASPECTS

In addition to the increased uranium (50 ppm) and thorium (300 ppm) noted above, the Quinnville carbonatite contains high concentrations of REE, with maximum values of lanthanum in excess of 1.33 %, barium >3 %, apatite and traces of copper in samples analyzed in this study. In addition, small lenses in the Haycock fenite contain about 10 % iron-titanium.

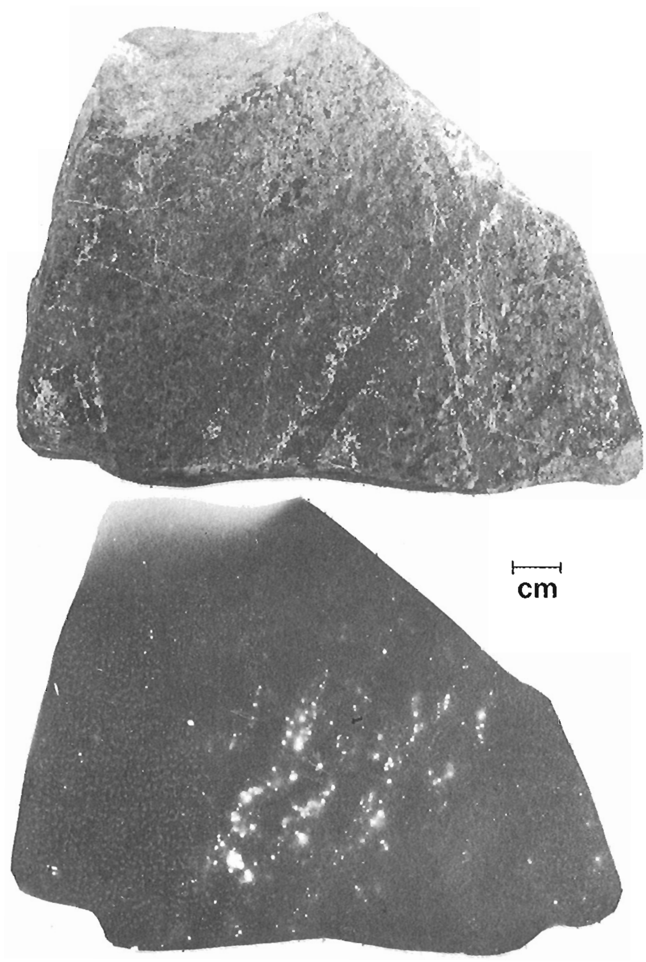


Figure 11. Network of hematite veinlets in the potassium anomaly at Quinnville North located by X on Figure 5A. Accompanying autoradiograph (one week exposure) shows a marked correlation of radioactive areas. Uraninite, monazite with hematite-rich areas.

CONCLUSION

This study is an example of utilization of integrated geophysical surveys in a carbonatite area.

Carbonatites are highly variable in their geophysical signatures. Thus in the Cantley region, fenite surrounding the Quinnville carbonatite produces an unusually high potassium anomaly, but neither carbonatite or fenite have magnetic signature. On the other hand the Templeton fenite zone produces a distinct magnetic anomaly but is virtually nonradioactive.

ACKNOWLEDGMENTS

P.B. Holman was responsible for the airbone geophysical surveys and compilations. M. Hudon drafted many of the figures and G.M. LeCheminant provided mineralogical data. S. Leclair of Ottawa University is currently studying the potassium anomaly surrounding the Quinnville carbonatite for a Bachelor's thesis and some of his observations are included in the above text. Thanks are extended to the Geological Association of Canada for allowing presentation of Figure 4. Many individual property owners in the area provided access and information to the authors for which they are grateful. The manuscript was critically read by K.L. Ford and K.A. Richardson.

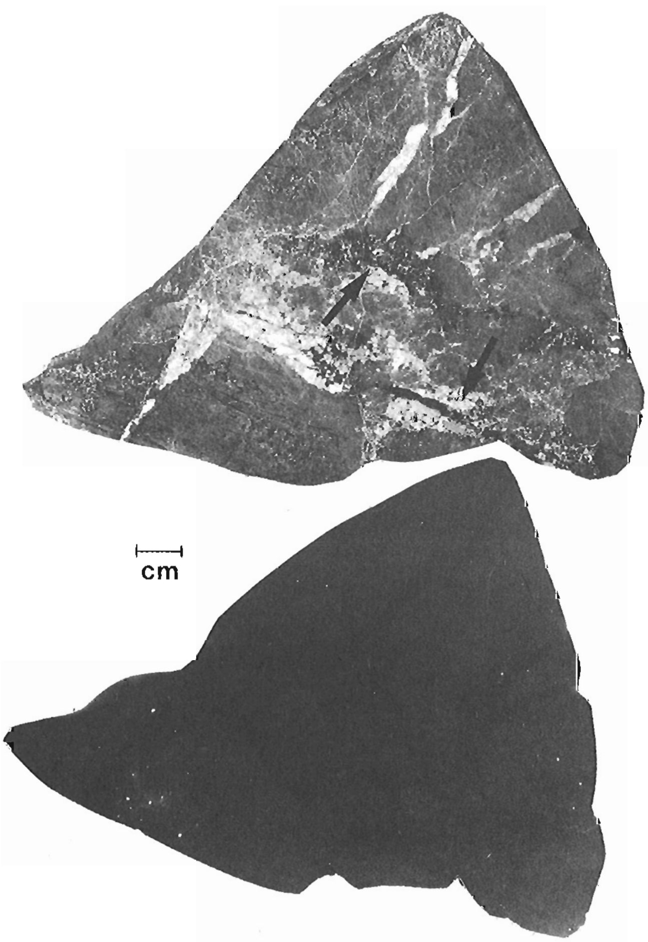


Figure 12. Fenitized rock north part of potassium anomaly, locality X, Figure 5A. Darker mineral is potassium-rich amphibole, ligher minerals carbonate, barite, quartz.

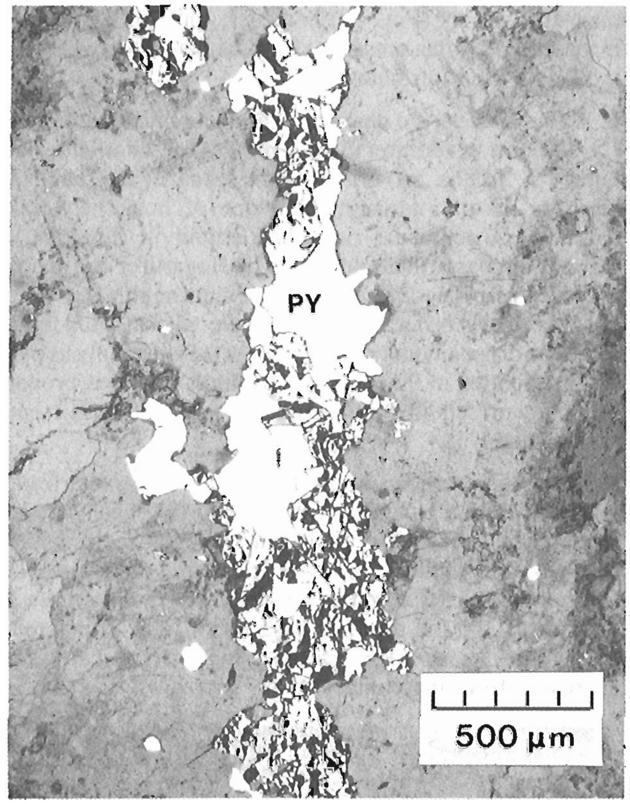


Figure 13. Pyrite along a fracture in rock from northern part of potassium anomaly locality X, Figure 5A.

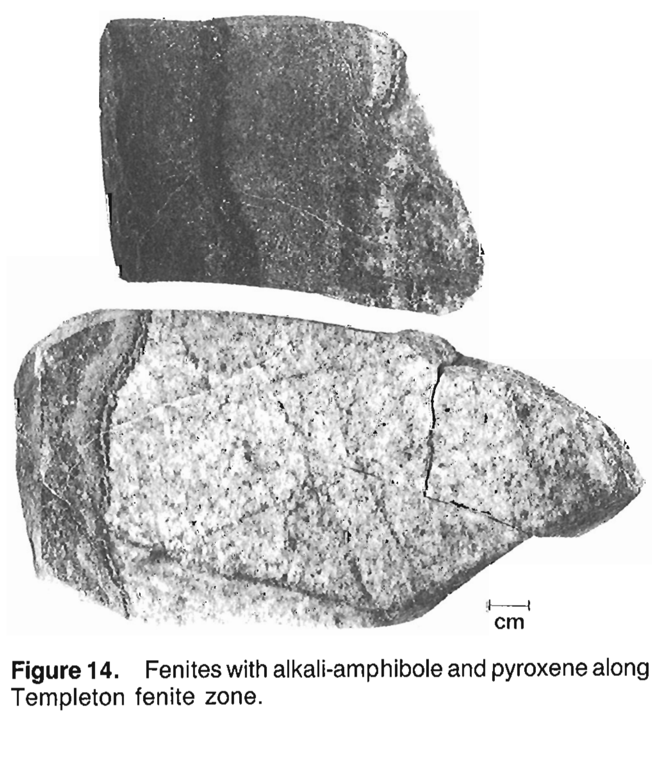


Figure 14. Fenites with alkali-amphibole and pyroxene along Templeton fenite zone.

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Geology of radioactive zones in the Round Pond area, Labrador[1]

Craig S. MacDougall[2] and Derek H. C. Wilton[2]

MacDougall, C.S. and Wilton, D.H.C., Geology of radioactive zones in the Round Pond area, Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 271-275, 1988.

Abstract

Narrow, linear radioactive zones with sphalerite, and minor galena, chalcopyrite, molybdenite, fluorite and precious metals occur in the Round Pond area of the Central Mineral Belt of Labrador within the early Proterozoic Upper Aillik Group. Uranium contents are generally less than 1 % and occur as pitchblende inclusions in amphibole crystals and/or as grains associated with apatite, sphene and Fe-Ti oxides. Mineralization occurs in fractures and breccia zones both along lithological contacts between, and within, rhyolite and amphibolite. The mineralizing event was apparently epigenetic with respect to the hosting Upper Aillik Group.

Résumé

D'étroites zones radioactives linéaires contenant de la sphalérite, ainsi que, dans une moindre mesure, de la galène, de la chalcopyrite, de la molybdénite, de la fluorite et des métaux précieux se trouvent dans la région de Round Pond, dans la zone minérale centrale du Labrador, au sein du groupe supérieur d'Aillik datant du Protérozoïque inférieur. Les teneurs en uranium atteignent en général moins de 1 % et se manifestent sous forme d'inclusions de pechblende à l'intérieur des cristaux d'amphibole ou de grains associés à l'apatite, la sphène et les oxydes de Fe et Ti, ou les deux. La minéralisation s'est produite dans des fractures et des zones de brèche le long de contacts lithologiques entre la rhyolite et l'amphibolite et à l'intérieur de celles-ci. Quant à ce qui a trait au groupe supérieur encaissant d'Aillik, il semble qu'il se soit agi d'une minéralisation de nature probablement épigénétique.

[1] Contribution to Canada-Newfoundland Mineral Development Agreement 1984-1989. Project carried by Geological Survey of Canada, Mineral Resources Division.

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INTRODUCTION

The Round Pond area (Fig. 1), located within the coastal region of the Central Mineral Belt near Makkovik, is underlain by early Proterozoic felsic volcanic and volcaniclastic rocks of the Upper Aillik Group (Gower et al., 1982). The rocks have been intruded by two high-level, satellite stocks of the ca. 1620 ± 60 Ma (Wanless et al., 1970), Monkey Hill Granite (MacDougall and Wilton, 1987). The area is characterized by numerous, widely distributed, pyritiferous (± molybdenite) gossans; magmatic-hydrothermal veins hosting Mo-Cu-Pb-Zn-U-F mineralization; and linear radioactive zones with associated base metal-molybdenite mineralization (the subject of this paper). Enrichments in Au and Ag contents are typically associated with all styles of mineralization (Wilton et al., 1987). Details concerning the general geology, styles of mineralization, and a discussion of the metallogeny in the Round Pond area have been previously reported by MacDougall and Wilton (1987).

The locations of the radioactive zones sampled within the Round Pond study area are shown in Figure 1. Gandhi (1978) called the uranium occurrences in this area the Falls Lake — Shoal Lake — Bernard Lake Uranium Belt. Some of these showings have been previously reported by M.J. Piloski of Brinex Limited (in 1955) and in 1980 by J.G. Burns of Placer-Brinex joint venture exploration programs. However, individual descriptions of smaller showings apparently do not exist; a number of exploration reports dealing with more extensive showings remain confidential; and there is often considerable discrepancy between the locations of the various showings on the mineral inventory maps published by the Newfoundland Department of Mines and exploration maps of the companies.

PREVIOUS WORK

The discovery of uranium mineralization in 1954 at Pitch Lake, 15 km south of Makkovik, by M.J. Piloski working for Brinex Limited, initiated extensive mineral exploration in the area. In 1955, geologists with this company discovered uranium mineralization in three radioactive zones east of Falls Lake. These zones known as Showings Nos. 16, 17, and 18, were investigated at that time through detailed mapping, trenching and shallow drilling. Due to low grades and discontinuous distributions, exploration of these showings was curtailed in 1956.

Exploration of the area was resumed by Brinex Limited late 1960s. In 1967, an extensive radioactive zone was discovered southeast of Round Pond. A 230 kg (500 lb) bulk sample was removed, but returned a grade of only 0.025 % U (S.S. Gandhi, unpublished report, 1968). No further work appears to have been done, although further exploration was recommended. In 1978, a re-evaluation of known radioactive occurrences, and further assessment work, including VLF-EM and magnetic surveys, resumed under a Placer-Brinex joint venture. No new showings were reported.

URANIUM MINERALIZATION

The radioactive zones range from less than 10 m to over 600 m in length, and from less than 1 to 2 m in width. Grades of uranium mineralization averaged over sample widths (0.5 to 0.75 m) are less than 1 % U in better sections (M.J. Piloski, 1955; S.S. Gandhi, 1968, unpublished reports). Mineralization is generally sporadic along strike, and is localized in one of the following ways:

1) in fractures and breccia zones along lithologic contacts (eg. Showing Nos. 16, 16 North, and southeast Round Pond),

2) fractures and zones of shearing and brecciation within rhyolite (eg. Showing No. 18), and amphibolite (eg. Showing No. 19),

3) discordant zones within permeable felsic volcanic conglomerate (eg. northeast Falls Lake).

Uranium occurs as:

1) individual grains of pitchblende,

2) pitchblende inclusions in secondary amphibole minerals,

3) pitchblende associated with apatite, sphene, and Fe-Ti oxides. Uranophane staining is common in areas of high radioactivity. Sphalerite is often associated with the uranium mineralization, with lesser amounts of galena, chalcopyrite, molybdenite and fluorite. Anomalous enrichments in gold and silver are also typical.

Alteration associated with the radioactive zones is dominated by the hematitization of the felsic volcanic country rock. Feldspars are totally altered to albite as a result of Na-metasomatism. Abundant hornblende, biotite, carbonate, Na-Ca pyroxene, sphene, apatite and fluorite also occur as fracture-fillings and disseminations.

Showing No. 16

Showing No. 16 (Fig. 1) consists of a discontiuous, moderately radioactive zone, 0.30 m to 2.0 m wide, and roughly 600 m long. The radioactivity is concentrated along the contact between a porphyritic to nonporphyritic banded rhyolite and a fine-grained mafic unit of probable volcanic origin, and forms zones with distinctive red to pink hematitic staining within the rhyolites. The uranium occurs as pitchblende concentrated in steeply dipping amphibole-carbonate veinlets and fracture-fillings which strike at low angles to the contact zone, and as pitchblende disseminations throughout red hematized zones in the rhyolitic rocks. Yellow uranophane staining is common. The contact zone commonly exhibits evidence of brecciation and extensive fracturing. Gandhi (1978) stated that most of the mineralization occured as veinlets within the mafic unit. However, samples for this study have revealed that mineralization at this showing appears to be restricted to the hematized rhyolitic rocks, and that the mafic metavolcanic lithology is barren of radioactive mineralization. Abundant sphalerite and pyrite, together with minor chalcopyrite and fluorite, are typically associated, and galena has been observed in some carbonate veinlets.

Showing No. 16 North

During the sampling program, a series of trenches approximately 400 m north of Showing No. 16 were found and sampled. The trenches were most probably blasted during exploration by the Placer-Brinex joint venture. The mineralization in the trenches is similar to Showing No. 16 and is regarded here as representing a strike extension of the latter. At Showing No. 16 North, the mineralized zone is 0.3 to 0.5 m wide, about 30-40 m long, and occurs at the contact of a rhyolite and a thin, amphibole-rich mafic unit. Extensive hematization and yellow uranophane staining occur along the contact together with amphibole-carbonate veinlets and fracture-fillings, and associated brecciation. Galena is also present. Scintillometer readings indicate locally very high radioactivity.

Showing No. 17

Showing No. 17 consists of sporadic uranium mineralization in a zone 5 m wide, and over 400 m long. The zone is located several hundred metres east of Showing No. 16. M.J. Piloski, stated that Showing No. 17 was similar to No. 16 in that uranium mineralization occurs in the contact zone between rhyolite and a mafic unit. However, examination of the trenches has revealed that mineralization is hosted entirely within the mafic unit, with no contact zone having been observed within the trenches. Uranium mineralization appears to be concentrated in fractures and narrow, rusty breccia zones. An interesting feature of the showing is the occurrence of calc-silicate and carbonate pods and veinlets similar to those associated with the Mo-Cu-Pb-Zn-U-F showings near Round Pond (MacDougall and Wilton, 1987). Abandoned

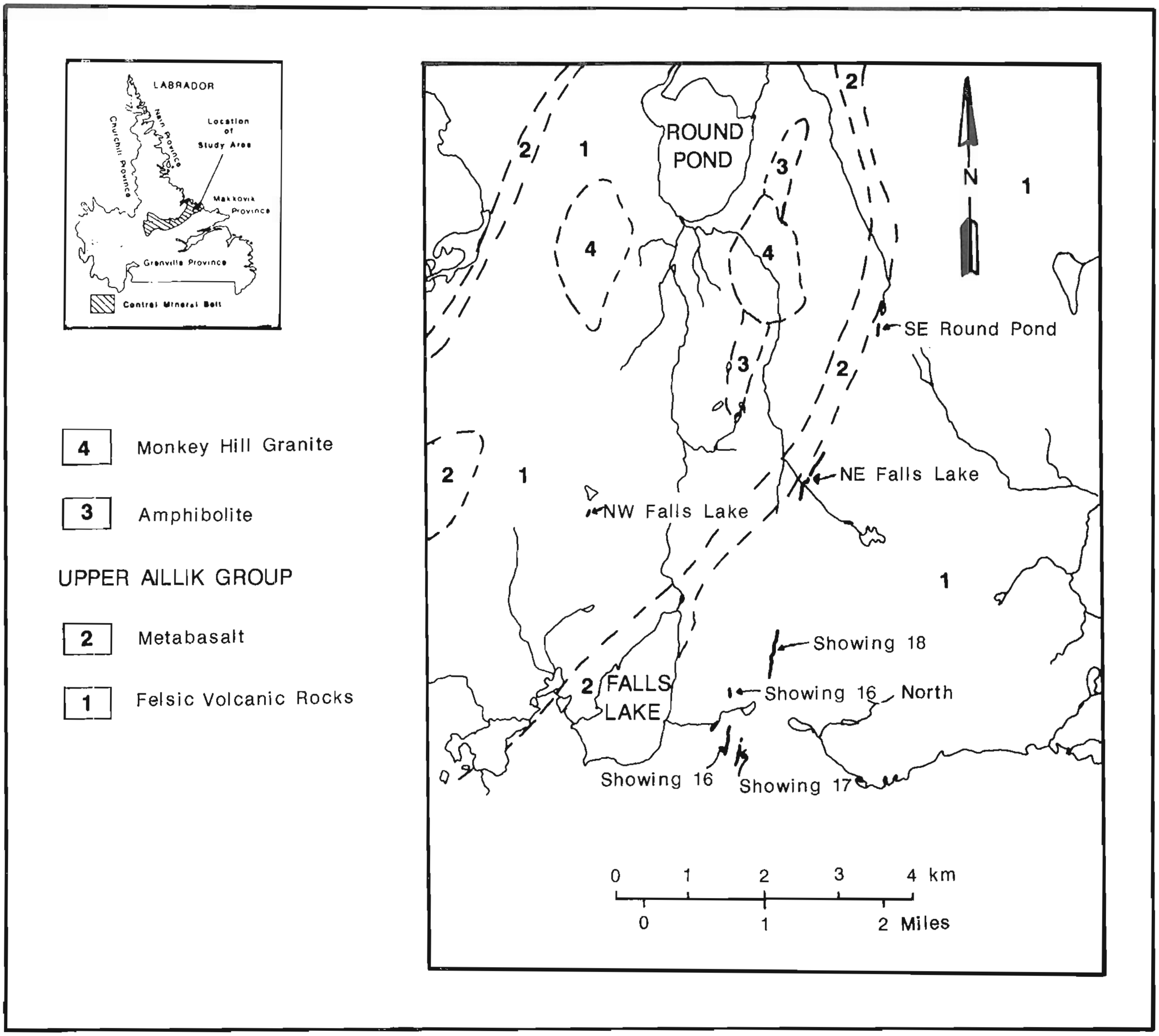

Figure 1. Geology of the Upper Aillik Group south of Round Pond (after MacDougall and Wilton, 1987) and location of radioactive zones. Inset map indicates location of study area in Central Mineral Belt of Labrador.

drill core nearby contains carbonate-calcsilicate veinlets carrying molybdenite and chalcopyrite.

Showing No. 18

Showing No. 18 consists of a moderately radioactive zone, 0.3 to 2.0 m wide with a minimal length of 600 m, reported in 1955 by M.J. Piloski to be open to the south. The rhyolitic rocks exhibit the characteristic red hematitic alteration, but the mafic rock associated with the other showings described above was not observed. M.J. Piloski, reported that the mineralization was concentrated within weak fractures along ''bedding planes'' of sandy quartzites. However, observations this past field season indicate that the radioactive fractures are slightly discordant to the regional foliation of the banded rhyolites. In addition, lensoid clasts were observed that may be the product of brecciation. Reported uranium grades were generally below 0.06 % U_3O_8, although grades as high as 0.13 % and 0.299 % U_3O_8 were obtained over narrow widths. Yellow uranophane staining commonly occurs with areas of moderate radioactivity. Minor pyrite, chalcopyrite, malachite staining, fluorite, and rare molybdenite were observed, and are typically associated with carbonate.

Northwest Falls Lake Showing

The showing northwest of Falls Lake consists of a small, 0.3 to 1.0 m wide and 10 m long radioactive zone. Numerous pyritiferous, molybdenite-bearing veins and gossans occur nearby. The showing is hosted in a felsic volcaniclastic conglomerate that exhibits the characteristic red hematitic staining and amphibole alteration minerals associated with the other rhyolite-hosted radioactive showings. The mineralized zone is discordant with the orientation of the conglomeratic clasts in the area. Associated with the uranium mineralization is abundant Cu mineralization occurring as chalcopyrite, malachite, and lesser bornite. Minor sphalerite, molybdenite, and fluorite are also present. Geochemical analysis reported by Wilton et al. (1987), indicate a significant enrichment in Ag (68 ppm) and Au (1250 ppb). Along strike from the mineralized zone, a hematitic boulder was observed which had moderate radioactivity and high grade molybdenite mineralization. The occurrence of molybdenite mineralization with hematized, radioactive rock has not been previously reported in Labrador.

Northeast Falls Lake Showing

The showing located northeast of Falls Lake, discovered by Brinex Ltd. in 1967, was referred to as the Round Pond radioactive zone. S.S. Gandhi (in an unpublished report) stated that the zone was continuous for more than 350 m, with a width of 15 m. Sampling carried out during this study revealed that the mineralization is very sporadic and discontinuous, and, thus, is not easily traced. He described the mineralization as fine disseminations in a grey, fine-grained ''feldspathic quartzite'', associated with aggregates of calcite, hornblende and biotite. Gandhi (1978) later referred to this unit as a massive to amygdaloidal andesitic flow. A 230 kg (500 lb) bulk sample, taken from a stream bed, returned an assay of 0.025 % U. The site of the bulk sample site could not be located with any confidence, but a sample taken from an anomalously radioactive zone near the stream is very similar to previous descriptions. In addition, a small red hematitic radioactive patch occurring in a quartz porphyritic rhyolite was sampled along the apparent trend of the radioactive zone. Gandhi (1978) reported notable amounts of Ag and Se from a fairly high grade radioactive sample found in rhyolite from the area.

Southeast Round Pond Showing

The showing southeast of Round Pond is not located on any mineral inventory or Brinex exploration maps, but may have been discovered by the Placer-Brinex joint venture (the assessment reports on this joint venture are still confidential). The showing, although small, may represent a significant northeast extension of the ''Round Pond radioactive zone'' discussed above. The showing consists of a small, red (hematized), moderately radioactive zone hosted in a quartz porphyritic rhyolite, and a close 10-15-cm-wide hydrothermal vein exhibiting much higher radioactivity. Exposure is poor, so dimensions of the showing are not readily apparent. The red hematized zone in addition to uranium mineralization, contains considerable sphalerite, pyrite, minor chalcopyrite, and anomalous Ag.

The small hydrothermal vein is unique among uranium occurrences in the Round Pond area. It is located about 8 m west near the contact of the rhyolite with an amygdaloidal metabasalt. The vein is characterized by large, 10 cm long euhedral, hornblende crystals, with intergrown fluorite, carbonate, garnet, and feldspar. Uranium mineralization occurs as small individual grains in brown patches within the hornblende crystals. Significant molybdenite, chalcopyrite, minor galena, and anomalous Ag values have been reported from these occurrences (Wilton et al., 1987).

CONCLUSIONS OF THE GENESIS OF THE URANIUM MINERALIZATION

The uranium mineralization occurs in distinctive, hematized, linear zones which can be traced discontinuously for up to several hundred metres. The mineralization is localized in fractures and breccia zones along lithological contacts, and within rhyolite and amphibolite, as well as in discordant permeable zones in conglomerate. Thus, the mineralization appears to represent a later, post-tectonic mineralizing event, and is not syngenetic with respect to the Upper Aillik Group volcanism as suggested by Gandhi (1978) and Gower et al. (1982).

The most extensive uranium mineralization in the Round Pond area occurs east of Falls Lake, and is localized along the contact between a felsic volcanic rhyolite and a thin mafic unit of probable volcanic origin; or within the mafic metavolcanic unit itself. The association of uranium mineralization with the mafic metavolcanic is rather problematical. It may indicate a stratigraphic control with respect to the uranium mineralization, and thus a possible syngenetic origin as suggested by Gandhi (1978). However, other radioactive showings in the area do not appear to be associated with such a lithology. As well, evidence of brecciation, and extensive, low angle, crosscutting fracturing along the contact, and within the mafic metavolcanic suggests that the lithology and contact zone may have acted as a zone of structural weakness, allowing for the deposition of uranium during a later mineralizing event. Other uranium occurrences in the area appear to support the idea of a later mineralizing event by exhibiting mineralized veins and zones discordant with regional trends (eg. Showing No. 7-11). The concentration of pitchblende into amphibole-rich veinlets and as disseminations in discordant hematized radioactive zones, often associated with fluorite, suggests an epigenetic-hydrothermal origin for the uranium mineralization in the Round Pond area.

The association of molybdenite and base metal sulphides with uranium and fluorite is typical of a granophile ore mineral assemblage, and clearly suggests a possible magmatic source for the mineralized hydrothermal fluids and the uranium mineralization. Another possibility is that magmatic fluids were able to leach sufficient uranium from the felsic volcanic units of the Upper Aillik Group to form the radioactive showings in the Round Pond area. However, post-tectonic granite stocks are present in the vicinity of the occurrences and there a number of uraniferous pegmatites are related to them; which serves to indicate that a granitic source of the mineralization is indeed possible. The observations presented herein support a metallogenic model based on granite-related mineralization as suggested by MacDougall and Wilton (1987) and Wilton and Wardle (1987).

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Stratigraphic and metallogenic relationships along the unconformity between Archean granite basement and the early Proterozoic Moran Lake Group, central Labrador[1]

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Abstract

A well exposed unconformity marks the contact between the Warren Creek Formation of the early Proterozoic Moran Lake Group and the underlying basement granitoids of the Archean Kanairiktok Intrusive Suite. The contact zone is the locus for a number of epigenetic quartz (±carbonate) galena-sphalerite chalcopyrite vein systems. A series of transects was completed across the contact zone in order to further define the zone and examine the potential for additional lead-zinc mineralization. Galena-bearing quartz veins were found at only one new locality; the unconformity itself was encountered at four localities.

Résumé

Une discordance bien exposée délimite le contact entre la formation de Warren Creek du groupe de Moran Lake datant du Protérozoïque inférieur et les granitoïdes de socle sous-jacents de la série de roches intrusives de Kanairiktok d'âge archéen. Dans la zone de contact se trouvent un nombre de systèmes filoniens épigénétiques de galène, sphalérite et chalcopyrite à quartz (+carbonate). Une série de transects a été tracée à travers la zone de contact afin de mieux définir la zone et d'analyser le potentiel de minéralisations supplémentaires de plomb et de zinc. On n'a trouvé des filons de quartz renfermant de la galène qu'à un seul endroit nouveau: la discordance elle-même a été observée à quatre endroits.

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INTRODUCTION

The Warren Creek Formation of the Early Proterozoic (?) Moran Lake Group rests with marked unconformity on the Archean granite basement of the Kanairiktok Intrusive Suite (Ryan, 1984). Northeast of Moran Lake the Archean-Proterozoic contact zone contains four previously documented galena (-sphalerite)-bearing quartz-carbonate vein systems (Fig. 1, 2) at the Ellingwood, CANICO Anomaly No. 8, Green Pond and Kanairiktok No. 14 Showings (Ryan, 1984; Wilton *et al.*, 1987, unpublished reports). Lead isotope data, derived for galena separates collected from these showings during the 1985 field season (Wilton, 1986), produced a secondary isochron indicating a common Archean source (t_1 = 3127 Ma) for the lead in the vein systems and mineral deposition of probable Grenvillian age (t_2 = 1270 Ma).

In order to evaluate the potential for other lead-zinc quartz vein lode showings along this contact and to examine in detail the stratigraphic and structural context of the showings, a series of mapping/sampling transects were conducted across the contact zone during the 1986 field season.

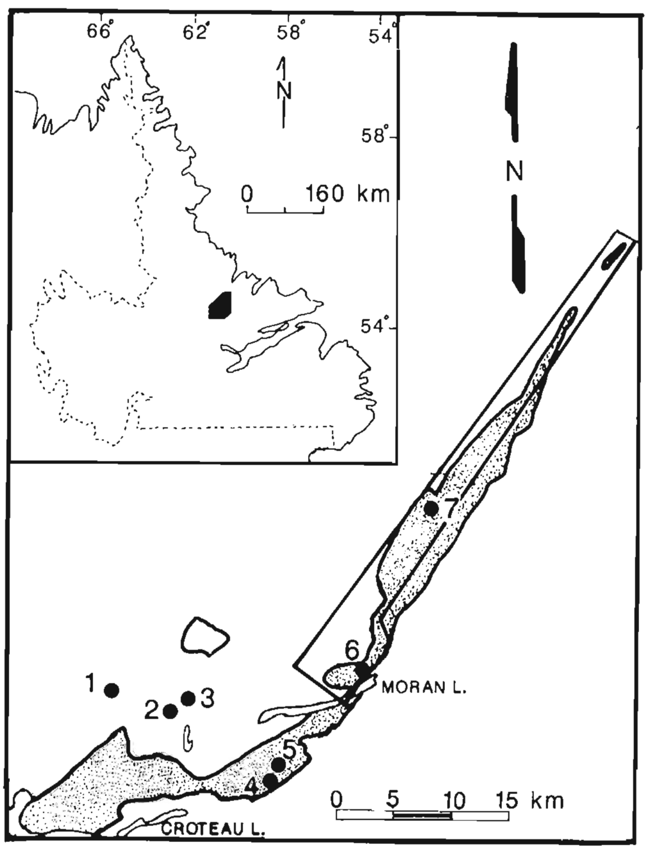

Figure 1. Location of the Archean granite basement — Early Proterozic (?) Moran Lake Group contact studied in this report and location of some radioactive occurrences discussed in the text.

TRANSECTS ACROSS THE ARCHEAN – MORAN LAKE GROUP (EARLY PROTEROZOIC) CONTACT

Twenty-one transects were completed across the contact between the Archean Kaniariktok Intrusive Suite basement and the Moran Lake Group between Moran Lake and the northeastern Kanairiktok River, a strike distance of 60 km (Wilton *et al.*, unpublished report, 1987). The locations of the transects are shown on Figure 2, based on a geological map compiled from the authors' field work and Ryan (1984). In the contact zone, the Kanairiktok Intrusive suite ranges from a coarse-grained granodiorite-diorite, to a fine- to medium-grained leucogranite. In the region northeast of the Kanairiktok River, the basement granite was not seen along Transects 1, 3; amphibolitized mafic volcanic units were found instead.

The Archean granitoid is relatively undeformed with only minor, variably developed S fabrics and shear/fracture sets. Chloritization of the biotite and/or amphibole are the predominant alteration effects found in the granitoids. Within 1-2 m of the Moran Lake Group contact, the granitoids exhibit a paleo-weathered regolithic surface, a feature developed prior to deposition of the Proterozoic Moran Lake Group (Ryan, 1984). Pervasive sericitic and/or brown carbonate alteration are developed in this weathering zone, and closest to the contact, the granite has a « bleached » grey colour. Identical relationships are found at the Archean granitoid Moran Lake Group contact near Croteau Lake (North and Wilton, 1988).

The Archean granite has intruded portions of the Weekes Amphibolite which are part of the Kanairiktok Valley Complex and represent the oldest rock types in the area (Ermanovics and Raudsepp, 1979; Ryan, 1984). This amphibolite underlies the area to the northwest of Embee Pond and was encountered in Transect 16 and 17. The foliated amphibolite has feldspar porpyroblasts (augen) set in a micaceous matrix and is cut by numerous dykes and veinlets of granite near the contact. On Transect 15 isolated xenoliths of amphibolite were found in the Archean granite.

The Warren Creek Formation consists mainly of black to grey (with rare green) shale and siltstone with local interbeds of dolostone, chert and black conglomerate (with siltstone and shale fragments, though A.B. Ryan (pers. comm., 1987) suggests the conglomerates are volcaniclastic). Minor, small, undeformed gabbroic dykes and sills intrude the Warren Creek Form-ation northeast of Moran Lake. The dykes and sills have not as yet been classified (petrographically or geochemically), but seemingly postdate the Moran Lake Group. Pillowed basalts of the Joe Pond Formation, green with a strong siliceous alteration, conformably overlie the Warren Creek Form-ation in the transect zones.

A series of stratigraphic columns, representing each individual transect, is illustrated on Figure 3. The actual unconformity between the Archean Kaniariktok Intrusive Suite and the Warren Creek Formation was seen at only four localities; viz. (1) immediately north of Moran Lake (Transect No. 22), (2) Transect No. 13 (west of Island Pond), (3) Transect No. 7 (northwest of Ballet Pond), and (4) Transect

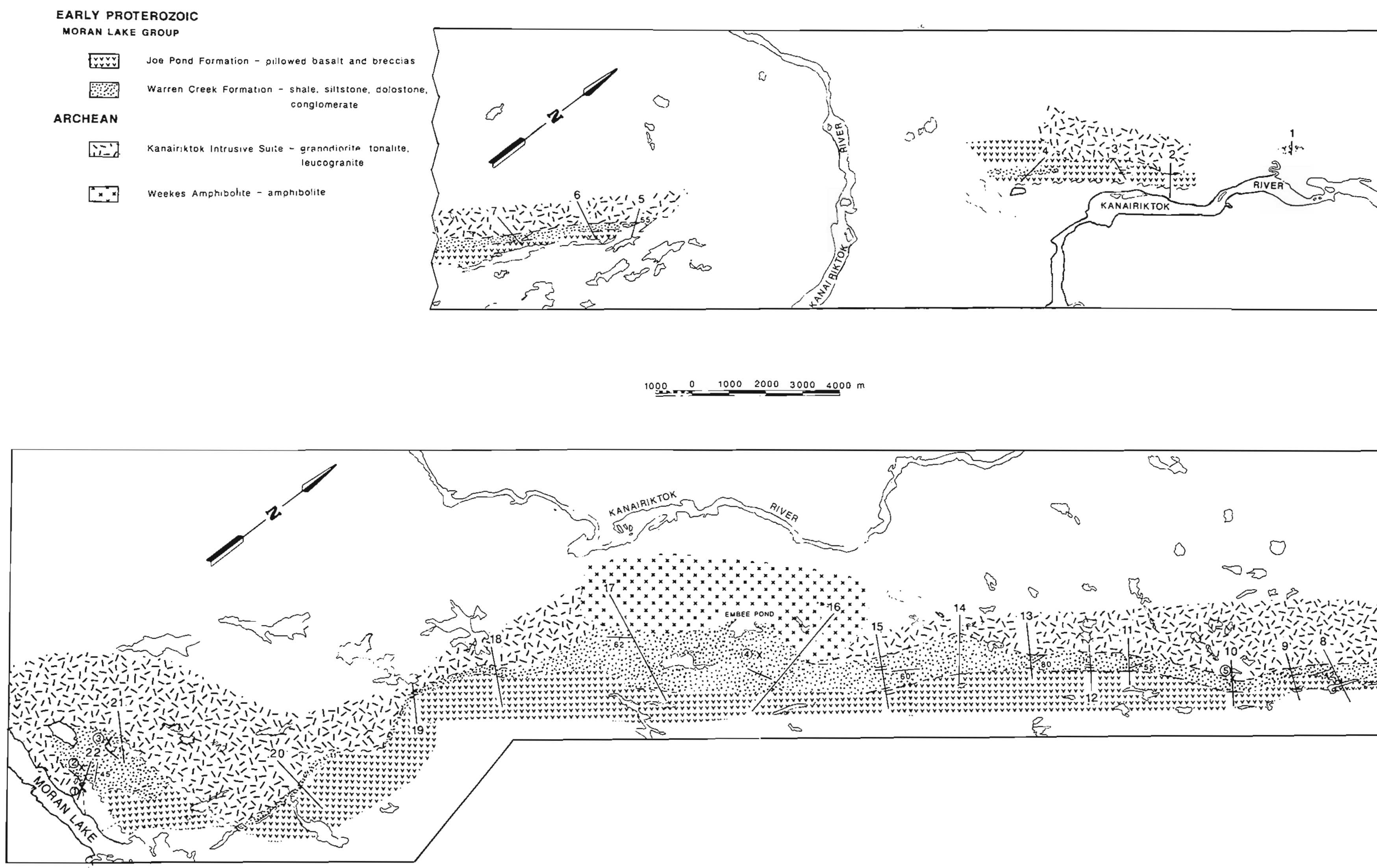

Figure 2. Geological map of the Moran Lake Group — basement contact (some geological information after Ryan, 1984) and location of the transects examined in the present study. Circled numbers refer to locations of quartz (carbonate) vein systems with galena and sphalerite, θιζ.; 1 = the newly discovered galena veins; 2 = the Ellingwood Showing; 3 = the CANICO Anomaly No. 8; 4 = the Green Pond Showings; and 5 = the Kanairiktok No. 14 Showing.

No. 6 (the Kanairiktok No. 14 Showing). At locality (1) (i.e. north of Moran Lake), a regolith developed on the granite is overlain by dolostone (Fig. 4); at (2), a finely laminated black and grey shale overlies foliated granite; at (3), a basal conglom-erate, with interbedded, isolated dolostone pods, rests on leucogranite (Fig. 5), and at (4) sheared siltstone is infolded within a leucogranite. A basal conglomerate was observed just southwest of the Kanairiktok No. 14 Showing, but its contact with the granite is not exposed. The unconformity at these localities was also reported by Ryan (1984) except for that along Transect 22.

The siltstones and shales of the Warren Creek Formation are overlain by basalt of the Joe Pond Formation. In some areas the Warren Creek Formation also contains interbeds of arkosic sandstone, phyllite, chert and black conglomerate. Similar lithologies are seen to a much greater extent in the area 8 km southwest of Moran Lake at Croteau Lake (North and Wilton, 1988).

Transect 5 started in the Archean and ended in a schist unit termed the Ballet Pond Schists by Ryan (1984). The actual contact between the schist and granite was not seen. Ryan (1984) stated these schists were derived by Proterozoic reworking and reconstitution of Archean gneiss, granite and diabase.

MINERAL OCCURRENCES IN THE ARCHEAN — MORAN LAKE GROUP CONTACT ZONE

Uranium showings within the Archean basement

The only mineral occurrences found to date in the Kanairiktok Intrusive Suite of the contact zone are three uraniferous showings north and northwest of Meathook Lake (8 km west of Moran Lake — see Fig. 1). These occurrences, called the Anomaly No. 7, Anomaly No. 7A, and Anomaly No. 17 Showings, were discovered and trenched by in 1979 by CANICO Limited, and Anomaly No. 7 was drilled in 1980.

At Anomaly No.7 the main lithology is a foliated granodiorite which is transected by a series of shear zones and joints. The youngest fractures are filled with hematite and pitchblende and are up to 2 cm wide. Around the radioactive fractures, as a selvage in the granodiorite, are zones of intense hematization (red). There are also some pegmatite dykes within the granodiorite cut by radioactive fractures.

The host rock to Anomaly No. 7A is hematized granodiorite with pitchblende fracture-fillings. Chalcopyrite also occurs in fractures at this showing with minor malachite staining.

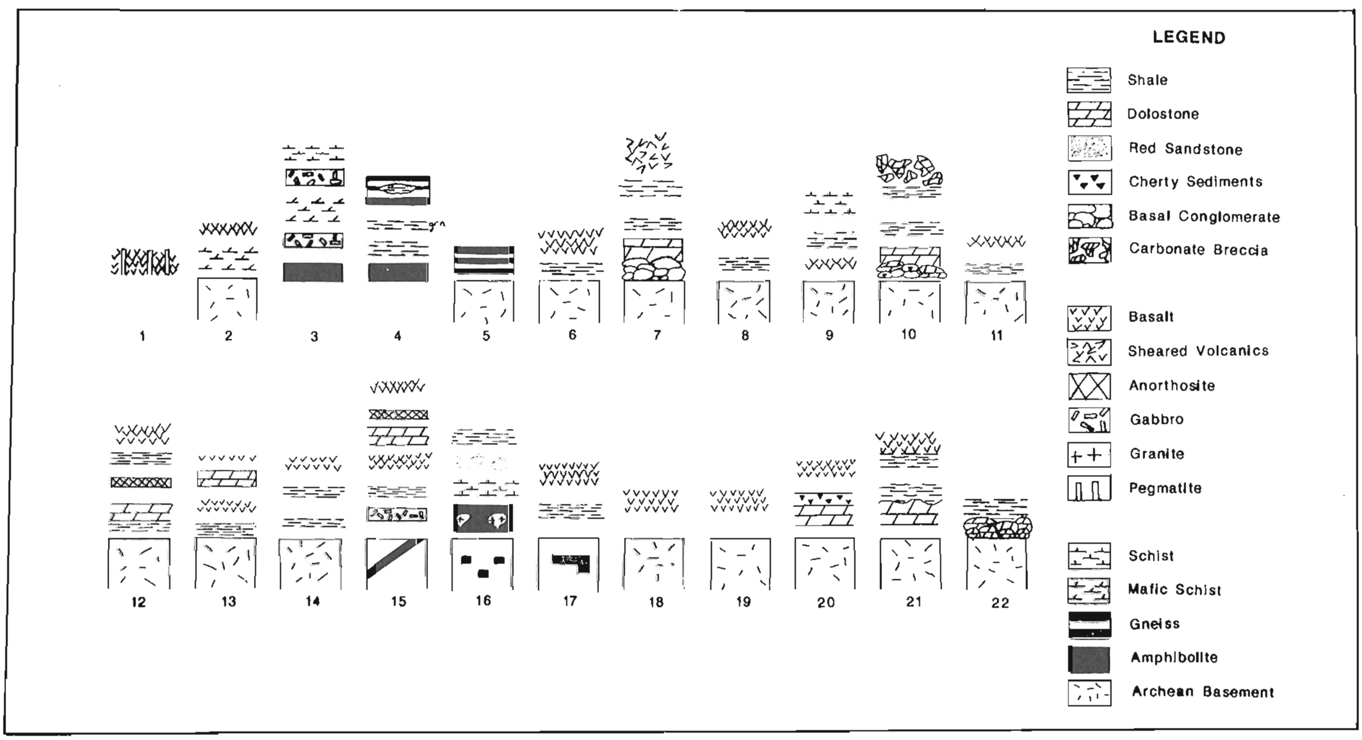

Figure 3. Composite stratigraphic section of transects across the Archean Proterozoic boundary, northeast of Moran Lake.

Anomaly No. 17 occurs within a very schistose granodiorite with thin ($\leq$ 2 mm thick) pitchblende coatings on fractures. There is a strong hematitic stain developed around the fractures.

The Joe Pond Formation pillow lava unit contains shear zone-hosted uranium occurrences (i.e. definitively epigenetic). The CANICO Anomaly No. 15 Showings (8 km south-southeast of Moran Lake; see Fig. 1) are rusty, radioactive patches in sheared Joe Pond Formation basalt. At the CANICO Anomaly No. 16 Showing (6 km south-southeast of Moran Lake) there are several small radioactive zones developed within highly fractured and/or sheared basalt. Intense rusty (gossan-like) alteration is usually associated with these zones which can contain up to 20 % pyrite. At the base of the outcrop ridge above the Moran Heights uranium boulder field (about 1.25 km north of Moran Lake), highly sheared and fractured (with chloritized slickensides) Joe Pond Formation basalts contain numerous carbonate-quartz veins with chalcopyrite, bornite and pyrite. Hematization of the host rock selvages is associated with those veins which are also radioactive.

U/Pb isotopic studies and geochemical analyses of pitchblende separates from these various occurrences are underway in order to determine the age(s) and source(s) of the hydrothermal fluids which produced the radioactive mineralization. Ryan (1984) suggested that the proximity of the uraniferous occurrences to unconformities (the Warren Creek Formation in the case of the Anomaly Nos. 7, 7A and 17; and the Bruce River Group for the Joe Pond Formation-hosted occurrences), and hence potential loci for fluid flow, may have had a genetic implication for the mineralizing systems.

The Warren Creek Formation contains no known radioactive occurrences.

Epigenetic galena-sphalerite quartz-carbonate vein showings occur along the Archean basement — Warren Creek Formation unconformity at the Ellingwood, CANICO Anomaly No. 8, Green Pond and Kanairiktok No. 14 Showings. The Ellingwood Showings comprise 9 trenches (Piloski, 1958) in which quartz-carbonate veins (up to 1 m thick) contain galena, sphalerite (red and honey brown varieties), chalcopyrite and pyrite. Grab samples from the veins at this occurrence had concentrations of up to 125 ppb Au and 25 ppm Ag (Wilton et al., 1987).

At Anomaly No. 8 (J. Perry reported) a small outcrop of brown weathering dolostone, from the basal portions of the Warren Creek Formation stratigraphically just above the unconformity with the Archean granite, contains numerous quartz veinlets some of which contain galena.

According to J.C.G. Moore the Green Pond Showing consists of two showings, the Green Pond No. 1 and the Green Pond No.2 Showings. The No.1 is the main occurrence and contains ten trenches and blasted outcrops in which quartz ($\pm$ carbonate) veins (up to 1 m thick) brecciate the host black siltstone. The veins contain layers of both red and honey-coloured sphalerite, up to 12 cm thick, and/or irregular masses of galena and chalcopyrite. The best assay values reported by Moore (in an unpublished 1954 report) were 0.84 % Pb, 17.87 % Zn and 0.002 oz/t Ag over 1 m and 0.69 % Cu, 3.25 % Pb, 5.25 % Zn and 0.73 oz/t Ag over 1.2 m. The No. 2 Showing is approximately 300 m north of the No. 1, and consists of three small trenches with quartz and carbonate veins that contain minor amounts (compared to Showing No. 1) of galena (+ chalcopyrite and sphalerite).

The Kanairiktok No. 14 Showing comprises a small trench in which quartz veins with variable thicknesses (up to 10 cm wide) contain large sulphide masses (up to 8 cm across) of sphalerite with galena and pyrite. The host rocks are sericitized leucogranite of the Archean basement and black siltstone of the Warren Creek Formation.

The only new galena occurrence found during this study is present in quartz veins cutting the basement granodiorite north of Moran Lake along Transect 22 (and southeast of the Ellingwood Showing). This new occurrence is interesting because it is located about 10-15 m above a chalcocite-chalcopyrite-pyrite quartz vein showing (NDM 13K/10 Cu018) in the basement, and thus, may indicate a metal

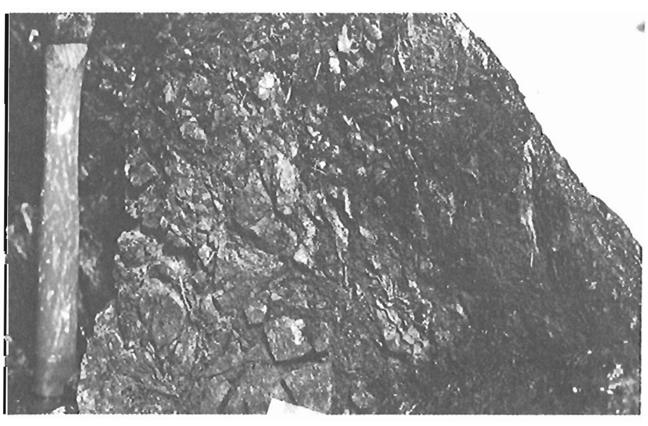

Figure 4. Regolith developed within Archean granite of the Kanairiktok Intrusive Suite along transect No. 22. The carbonate-cemented regolith is overlain by Warren Creek Formation dolostone.

Figure 5. Basal conglomerate composed of Archean granite fragments overlain by Warren Creek Formation shales along transect No.7. (see also Ryan, 1984, Plate 4-1).

zoning within the vein systems. Furthermore, the proximity of the galena occurrence to the Ellingwood Showing (hosted by Warren Creek Formation shales), suggests the mineralizing fluids at least travelled through the basement rocks, if not actually derived from therein. At other locations along the unconformity both the basement granite and overlying shales are cut by abundant quartz veins; these veins are being examined for possible lead-zinc minerals.

CONCLUSIONS

The unconformity between the Proterozoic Moran Lake Group and the Archean Kaniariktok Intrusive Suite is well defined north of Moran Lake and is exposed in at least four localities. The basement granitoids often exhibit regolithic paleo-weathering just below the Warren Creek Formation. The contact zone contains a series of epigenetic galena- and sphalerite-bearing quartz veins along and near the unconformity which crosscut the structural fabrics within the host rocks. The vein systems presumably resulted from the mobilization of ore fluids from the Archean basement into the unconformity zone during the Grenvillian Orogeny. Pitchblende occurs in shear zones cutting the Archean basement and the Joe Pond Formation basalts of the Moran Lake Group, but radioactive occurrences are unknown within the Warren Creek Formation. The source of the uranium and age of mineralization are at present unknown.

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Hanbury Island Shear Zone, a deformed remnant of a ductile thrust, District of Keewatin, N.W.T.

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Tella, S. and Annesley, I.R., Hanbury Island Shear Zone, a deformed remnant of a ductile thrust, District of Keewatin, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 283-289, 1988.

Abstract

The Hanbury Island Shear Zone (HISZ), exposed on the shore of Hudson Bay east of Chesterfield Inlet, is an ENE to NE trending, ENE plunging synformal ductile shear zone formed under amphibolite to granulite facies conditions. The zone overlies a lower grade granitoid gneiss terrane, and has a strike length of 40 km and a width of 5 km. The protoliths include two-pyroxene granulite, layered tonalite gneiss, garnet-biotite paragneiss, megacrystic granitoid, metamafic dykes, and coronitic gabbro and anorthositic gabbro. The mylonitic layering contains disrupted and boudinaged mafic and quartzofeldspathic layers, and heterogeneously strained coronitic gabbro and anorthosite. A shallow (10-30°) ENE plunging mineral-stretching lineation is well developed. Kinematic data indicate sinistral and dextral senses of shear. The zone is probably an integral part of ductile high-strain zones associated with the uplift of granulite-anorthosite complexes between Baker Lake and Daly Bay. The HISZ is interpreted as a deformed remnant of a ductile thrust composed of deep-crustal rocks emplaced into its present position during the late Archean or early Proterozoic.

Résumé

La zone de cisaillement de Hansbury Island (ZCHI), exposée sur le littoral de la baie d'Hudson à l'est de l'inlet Chesterfield, est orientée de l'est-nord-est au nord-est; il s'agit d'une zone de cisaillement ductile synforme, plongeant vers l'est-nord-est et formée sous des conditions de faciès variant des amphibolites aux granulites. La zone recouvre un terriane de gneiss granitoïde de qualité pauvre et s'étend sur 40 km de long et 5 km de large. Les protolithes comprennent des granulites à deux pyroxènes, du gneiss tonalitique lité, du paragneiss à grenat-biotite, des granitoïdes mégacristiques, des dykes métamafiques et du gabbro coronitique et anorthositique. Les couches de mylonites contiennent des interruptions et du boudinage de roches mafiques et quartzofeldspathiques, du gabbro et de l'anorthosite sous tension hétérogène. Une linéation composée de minéraux étirés plongeant peu profondément en direction est-nord-est (10°-30°) est bien développée. Les données cinématiques indiquent des sens sénestres et dextres du cisaillement. Cette zone est probablement une partie intégrale des zones ductiles sous hautes tensions accompagnée du soulèvement des complexes granulite-anorthosite entre Baker Lake et Daly Bay. La zone de cisaillement de Hansbury Island est interprétée comme un vestige d'une poussée ductile composée de roches crustales profondes mises en place au cours de l'Archéen et du Protérozoïque inférieur.

INTRODUCTION

Bedrock mapping in the northern parts of the Chesterfield Inlet map area (NTS 55∅), at a scale of 1:250 000, was completed during the 1987 field season. The fieldwork delineated a 40 km long and 5 km wide, folded ductile shear zone herein referred to as the Hanbury Island Shear Zone (HISZ). This paper describes the geometry and tectonic setting of this zone, and summarizes the field observations and preliminary conclusions. A brief outline of the regional geology, structure, and metamorphism precedes the observations on the HISZ. The map area is part of a region previously mapped by several workers on a reconnaissance scale (Bell, 1885, 1887; Lord, 1953; Wright, 1955, 1967). Results of recent bedrock mapping in the adjoining region to the south was reported by Tella and Annesley (1987) and Tella et al. (1986).

REGIONAL GEOLOGY, STRUCTURE, AND METAMORPHISM

The distribution of rock units in the northern half of the Chesterfield Inlet map area is shown in Figure 1. The Archean and/or early Proterozoic lithologies in the region are dominated by polydeformed and regionally metamorphosed granulite gneiss, orthogneiss, and migmatite, with subordinate amphibolite, gabbroic anorthosite, anorthositic gabbro, and pelitic gneiss (units 1 to 6). Type I and III interference fold patterns (Ramsay, 1967) are present throughout the region, and the metamorphic grade is within the amphibolite to granulite facies. Post-tectonic Proterozoic intrusive activity is recorded by a fluorite granite (unit 7) and by northwest-trending diabase dykes (unit 8). At least three sets of late brittle faults (east-, northwest-, and north-trending) transect the region on all scales.

The granulite suite (unit 1, Fig. 1) is dominated by well-layered and compositionally banded, quartzofeldspathic granulites interlayered with minor proportions of mafic granulites, paragneiss, granitic gneiss, layered anorthosite, and anorthositic gabbro. The suite is widely distributed in the northwestern portion of the Chesterfield Inlet map area. The quartzofeldspathic granulites are predominantly tonalite and contain orthopyroxene-clinopyroxene-garnet-hornblende-plagioclase-quartz-opaques assemblages. Most paragneiss layers contain abundant garnet-biotite $\pm$ cordierite + quartz + plagioclase. The granulites commonly show a waxy-green lustre and granoblastic textures, and the lithological contacts with the overlying amphibolite grade grey gneiss are either gradational or fault controlled. Metamorphic transitions from granulite to amphibolite grade, both along and across the strike of the compositional layering, are present throughout the region. The transitions parallel to the layering suggest that at least locally, the isograds are at a high angle to the layering. Assessment of prograde and retrograde affects awaits further detailed petrographic study of the granulite-gneiss terrane. The suite is compositionally and structurally similar to the Archean Kramanituar (Schau et al., 1982) and the Daly Bay complexes (Gordon, 1980; in press) that underlie Baker Lake and Daly Bay regions, respectively. The Kramanituar complex is a layered granulite-gabbro-anorthosite complex (Schau, 1980; Schau et al., 1982). The Daly Bay Complex, composed of granulite gneiss, gabbroic anorthosite and migmatite, is interpreted as a belt of Archean lower- and middle-crustal rocks that have been emplaced into higher crustal levels in a north-northwesterly transport direction during the early Proterozoic (Gordon, 1980; in press). The outer margin of the Daly Bay Complex is marked by two complex marginal ductile shear zones (inner and outer) that are characterized by textural, metamorphic, and strain gradients (Gordon, in press; pers. comm., 1987). The inner zone is characterized by granoblastic textures, higher grade assemblages, and high ductile strain.

Rocks of unit 2 comprise polydeformed and regionally metamorphosed (amphibolite facies), well layered grey orthogneiss, migmatite and biotite schist with minor proportions of iron-rich metasedimentary rocks, and garnetiferous amphibolite. Numerous discontinuous, narrow (100-200 m thick) belts of garnet- biotite paragneiss (unit 3) occur within the grey gneisses (unit 2) in the south-central portions of the map area (Fig. 1). Garnet porphyroblasts range from less than 1 mm to 5 cm in diameter. Sillimanite, cordierite, and kyanite were noted locally. The lithology and structure of these units (units 2 and 3), which extend to the south of the map area, are similar to those described previously (Tella and Annesley, 1987).

A number of granitoid plutons (unit 4), distributed along a northeasterly trend (Fig. 1), intrude the layered gneiss and biotite paragneiss units. Although the individual plutons are compositionally homogeneous, there is a range in composition from quartz diorite in the northeast to granite in the southwest. These granitoid plutons form the cores of domical gneiss structures and contacts between cores and margins are gradational. The plutonic rocks are coarse grained, equigranular, light pink to dark grey, homogeneous bodies that contain abundant inclusions of paragneiss, layered gneiss, and amphibolite. Field relations suggest that the regional metamorphism of the gneisses was probably contemporaneous with the syntectonic emplacement of granitoid rocks. Similar relations prevail in the Gibson Lake region to the southwest (Reinhardt and Chandler, 1973; Reinhardt et al., 1980).

The Hanbury Island Shear Zone (HISZ) (unit 5) is composed of mixed lithologies which include rocks of units 1 to 4 described above. Its detailed tectonic setting and structural aspects will be described in a later section.

An enequigranular, megacrystic (K-feldspar), leucocratic granite pluton (unit 6) that contains less than 10 % mafics (hornblende and biotite) occupies most of the northeastern portion of the map area (Fig. 1). The large pluton is mylonitized along the southern and southwestern margins and becomes a part of the HISZ. Magnetite is sporadically distributed in individual crystals and in crystal aggregates up to 5 cm across. The magnetite-rich character of the pluton is reflected in a pronounced aeromagnetic signature (Geological Survey of Canada, 1966). The granite contains abundant inclusions (rafts and xenoliths) of older rocks (grey gneiss, amphibolite, paragneiss) near the margins and the contacts are commonly gradational, and to a lesser extent fault bounded. Granitic and pegmatite dykes and veins occur within the gneiss terrane adjacent to the granite contacts. Smaller satellite stocks occupy the cores of domal structures to the west. The granite is commonly massive at the cores and foliated at the margins of the stocks. The satellite stock on the northern shore of the Chesterfield Inlet (Fig. 1) contains large xenoliths of well foliated to massive metagabbro and metapyroxenite.

Massive to weakly cleaved, and equigranular to porphyritic, fluorite granite (unit 7, Fig. 1) is exposed in the southwestern corner of the map area. The field relations and textural aspects of this pluton have been reported previously (Tella and Annesley, 1987). The granite is compositionally similar to some of the early Proterozoic (1.75-1.85 Ga) intrusions that are widely distributed in the Baker Lake region to the west (LeCheminant et al., 1987), and a similar age is suggested.

Numerous east-trending metadiabase dykes (not shown in Fig. 1), commonly less than 1 m thick, occur throughout the region. They cut the units 1 to 5 quartzofeldspathic and granulite gneiss terrane and are in turn displaced by late north- and northwest-trending faults. Deformed, sheared, and discontinuous biotite lamprophyre dykes of uncertain age are

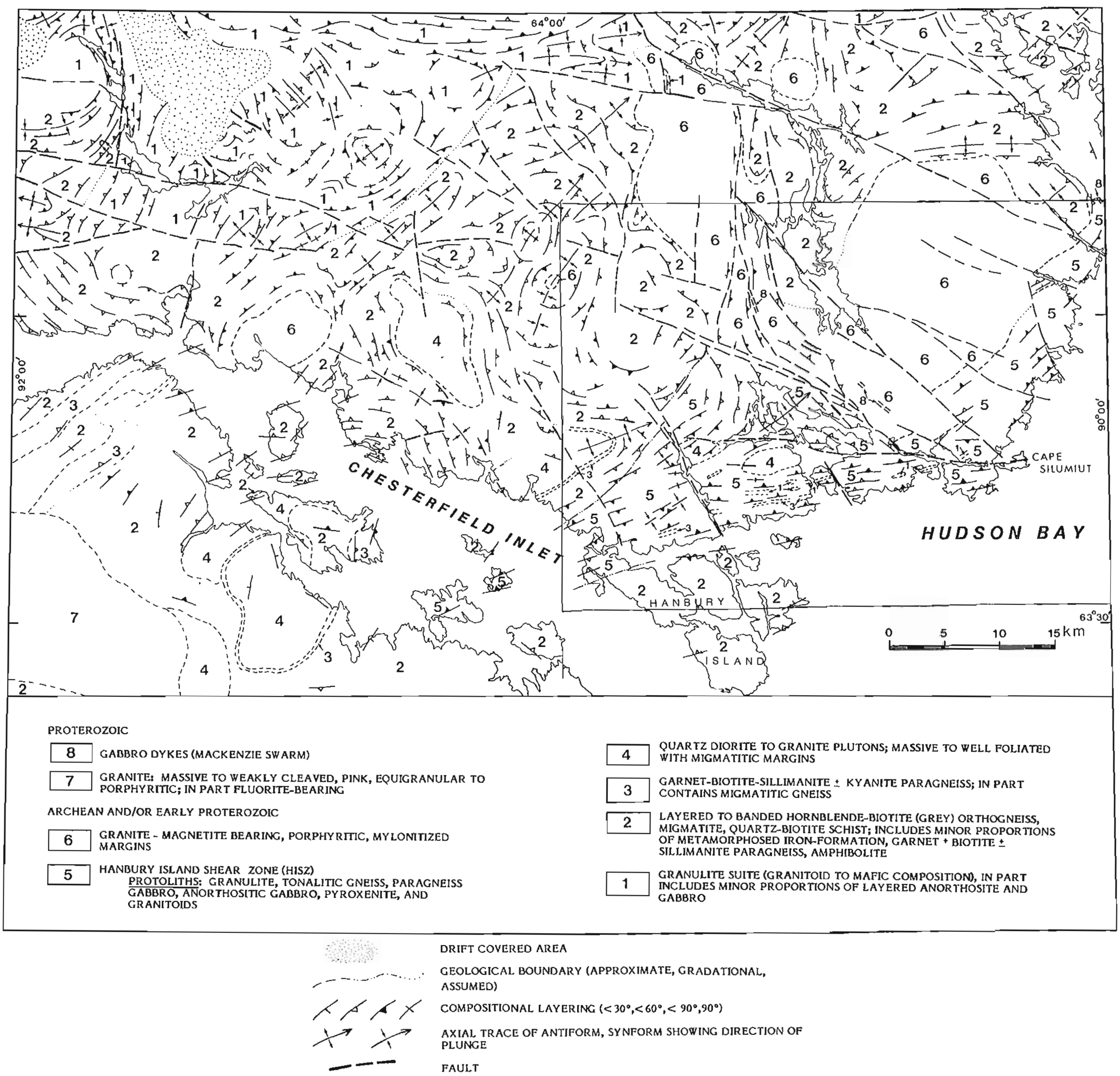

Figure 1. Simplified geological sketch map of the northern half of Chesterfield Inlet map area, District of Keewatin.

present in the eastern half of the area. They are dark grey to black, medium-to fine-grained rocks consisting of phenocrysts of biotite/phlogopite. They range up to 20 cm in thickness and have irregular trends.

Northwest-trending, fine- to medium-grained, massive diabase dykes (unit 8), probably related to the Mackenzie swarm, were noted at several localities in the eastern half of the area. Only the widest dykes are shown on the map (Fig. 1). They are up to 60 m thick and locally contain abundant xenoliths of gneissic country rocks. These dykes record the youngest intrusive event in the region. They do not have the characteristic magnetic expression of the Mackenzie swarm on aeromagnetic maps.

Numerous north-, northwest- and east-trending late brittle fault sets transect the region on all scales and affect all units with the exception of diabase dykes (unit 8). The northerly

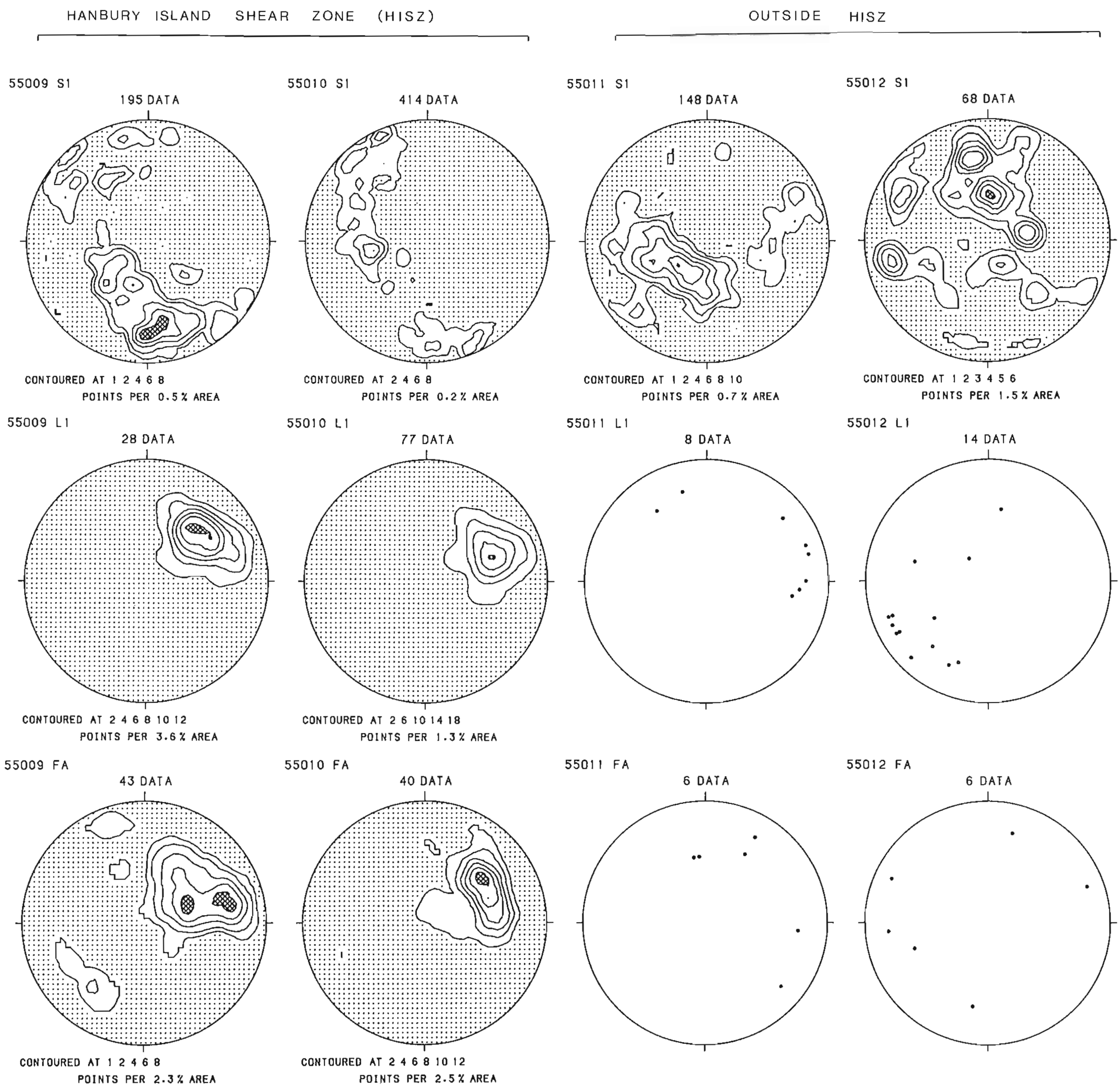

Figure 2. Orientation of planar and linear fabric elements from within and outside the Hanbury Island Shear Zone. Data are restricted to the region outlined by the rectangle in Figure 1. Contoured according to the method of Starkey (1969, 1977).
Sl - compositional layering parallel to mylonite foliation.
Ll - quartz and feldspar mineral stretching lineations.
FA - fold axes which define the orientation of the HISZ synform.

trending faults appear to be the oldest, and the northwest- and east-trending sets appear to be coeval as they are in part curviplanar. The northwesterly fault sets are pervasive structures that extend from Rankin Inlet (Tella et al., 1986) through Gibson Lake (Reinhardt et al., 1980) and Chesterfield Inlet (Tella and Annesley, 1987), to the Baker Lake region (Schau et al., 1982). The net displacements on the fault zone appear to be relatively minor.

HANBURY ISLAND SHEAR ZONE (HISZ; UNIT 5)

The HISZ is an east-northeast- to northeast-trending, shallow (10-30°) east-northeast plunging synformal ductile shear zone characterized by amphibolite to granulite grade mylonites. The zone is well exposed between the northwestern tip of Hanbury Island and Cape Silumiut (Fig. 1), and has an exposed strike length of 40 km and an average width of 5 km.

The amphibolite to granulite grade HISZ structurally overlies a lower grade (amphibolite facies) granitoid gneiss terrane. The northern and southern margins of the zone are commonly gradational with the underlying gneisses, but a structural break is clearly indicated by the discordant relations in the orientation of fabric elements (compositional/mylonite layering, fold axes, lineations) from within and outside the HISZ (Fig. 2). The northeastern margin is a faulted intrusive contact with a large granite pluton (unit 6). This granite is, in part, mylonitized and becomes a part of the HISZ at the margins, suggesting that its emplacement into the shear zone is syn-to late-tectonic with respect to the HISZ development. Zircon geochronology is planned to determine the absolute age of the granite that would provide a minimum age for the development of the HISZ.

The HISZ is made up of a multitude of rock types which include two-pyroxene granulite, well layered quartzofeldspathic (tonalitic) gneiss, garnet-biotite paragneiss, megacrystic granitoid, metamafic dykes, coronitic gabbro, anorthositic gabbro, amphibolite, and pyroxenite - all of uncertain age, but lithologically similar to those of units 1 to 4. The mylonitic layering contains abundant, highly disrupted, attenuated, and boudinaged mafic and quartzofeldspathic layers, and heterogeneously strained coronitic gabbro/anorthosite. Boudinaged quartzofeldspathic and mafic layers attest to ductile strain at high metamorphic grade (Fig. 3).

A penetrative, shallow (10-30°) east-northeast plunging mineral-stretching lineation is well developed within the HISZ. The orientation of the lineations (Fig. 2) suggests an overall northeast-southwest transport direction. Most pull-apart structures (Hanmer, 1986) in mafic and quartzofeldspathic layers exhibit symmetrical layer parallel extension (Fig. 3, 4), but a few asymmetrical drag folds and pull-aparts in mafic layers show conflicting senses of shear (sinistral and dextral). Although the contrasting senses of shear (viewed parallel to the lineation) may be explained by the folded nature of the shear zone, the true transport direction (NE or SW) is not clear. This is because of the uncertainty involved in correlating the observed kinematic indicators across tight minor folds associated with the major synform.

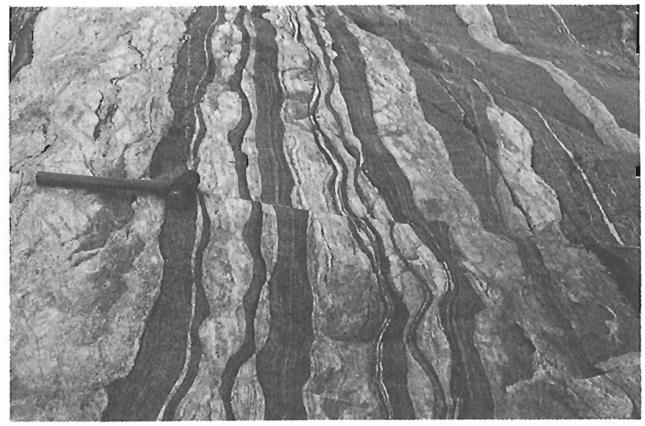

Figure 3. Boudinaged quartzofeldspathic layers alternating with mafic gneiss, Hanbury Island Shear Zone. Horizontal erosion surface, view to the west-southwest. (GSC 204040-Y).

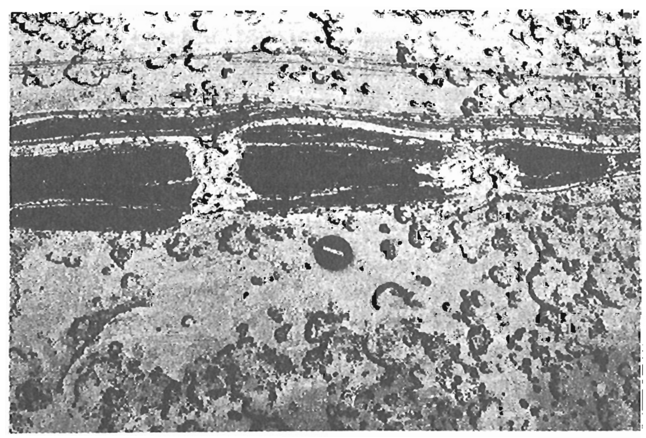

Figure 4. Symmetrical pull-aparts in an amphibolite layer, HISZ. Subhorizontal erosional surface, view to the north-northwest. (GSC 204423-N).

Figure 5. Tectonized fragments and blocks of coronitic gabbro/anorthositic gabbro within quartzofeldspathic gneiss, HISZ. Subhorizontal erosion surface, view to the east. (GSC 204423-K).

Highly strained and dismembered layers, rafts and fragments of gabbro and gabbroic anorthosite (Fig. 5), ranging in thickness up to 100 m with strike lengths of up to 0.5 km, are sporadically distributed throughout the length of the HISZ. They are relatively more abundant in the southern limb than in the northern limb of the synform, and provide good marker horizons within the zone. In many places the progressive transformation of primary igneous textures into well lineated mylonites (Fig. 6), can be demonstrated on a mesoscopic scale. Shear senses derived from these dismembered layers, however, cannot be correlated with those determined from the adjacent gabbro/anorthosite layers because of the irregular orientaiton of the layers within the shear zone.

The HISZ appears to link with the marginal shear zones (inner and outer) of the Daly Bay Complex exposed approximately 25 km to the northeast of the present area. The lithologies in the HISZ are strikingly similar to those noted in the inner shear zone of the Daly Bay Complex, but the northwest tectonic transport direction of the Daly Bay Complex is incompatible with the overall NE-SW transport direction of the HISZ. Lithologies in the HISZ are also comparable to those observed in the Kramanituar Complex (Schau et al., 1982), an Archean layered gabbro-anorthosite-granulite complex exposed approximately 75 km to the west-northwest of the study area. The complex is thought to have been emplaced along a late Archean ductile strain zone, the Chesterfield Fault Zone (Schau and Ashton, 1980; Schau et al., 1982). The HISZ may represent a remnant of a number of discontinuous high-strain zones that link the Kramanituar and the Daly Bay complexes. Based on field observations, on lithological similarities to the Kramanituar and Daly Bay complexes, and on the structural position of the high grade gneisses over a lower grade terrane, the HISZ is tentatively interpreted as a deformed remnant of a ductile thrust composed of deep-crustal rocks emplaced into its present position during the late Archean or early Proterozoic. Detailed petrography, geochronology, and thermobarometric studies are planned to document the tectonometamorphic evolution of the HISZ.

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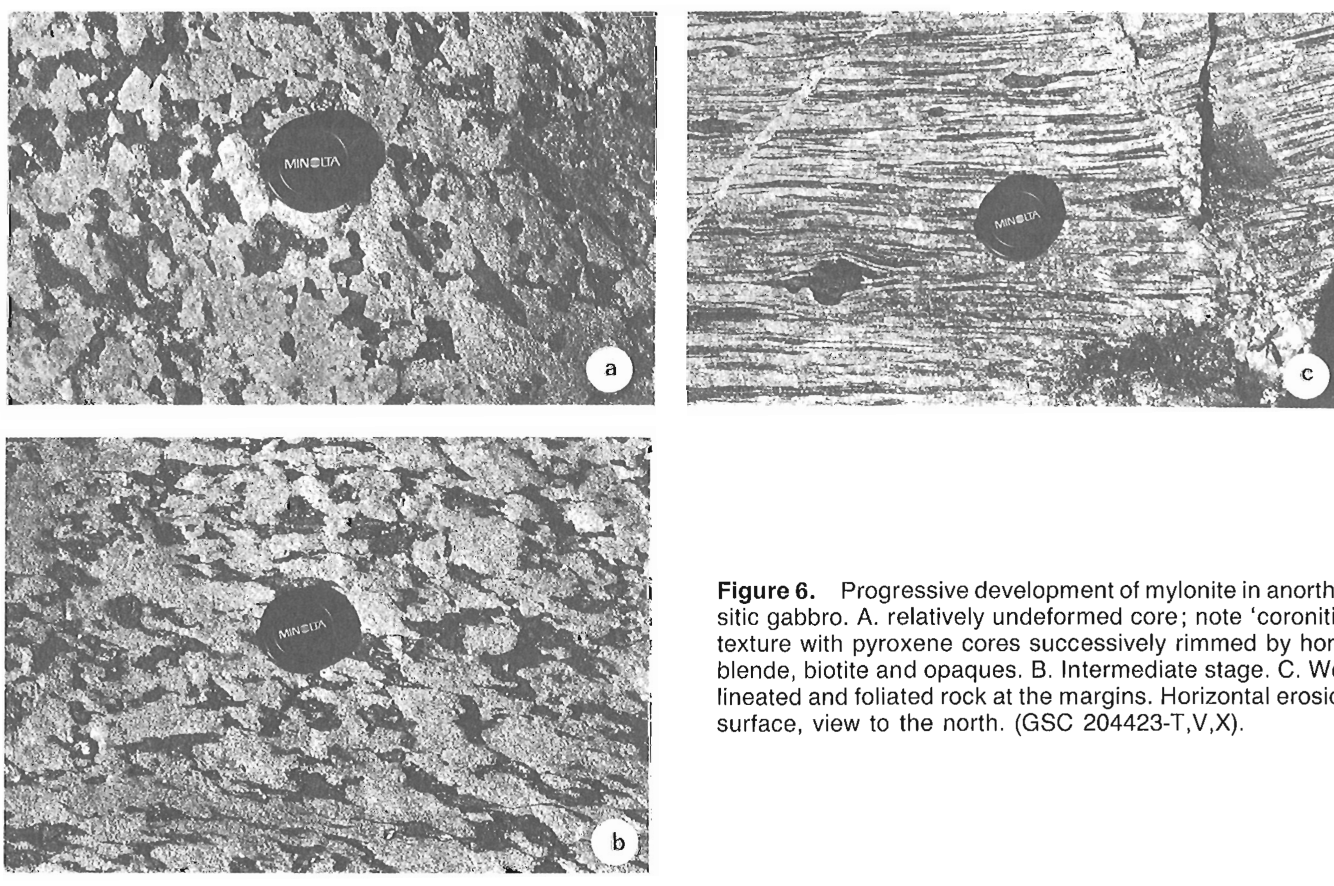

Figure 6. Progressive development of mylonite in anorthositic gabbro. A. relatively undeformed core; note 'coronitic' texture with pyroxene cores successively rimmed by hornblende, biotite and opaques. B. Intermediate stage. C. Well lineated and foliated rock at the margins. Horizontal erosion surface, view to the north. (GSC 204423-T,V,X).

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Copper occurrences in the Bruce River Group, Central Mineral Belt of Labrador[1]

Derek H. C. Wilton[2]

Wilton, D.H.C., Copper occurrences in the Bruce River Group, Central Mineral Belt of Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 291-297, 1988.

Abstract

The middle Proterozoic Bruce River Group supracrustal sequence contains copper (± zinc and/or lead) occurrences localized within faults, fractures and/or shear zones. The most intensive mineralization occurs within the stratigraphically lower sedimentary units of the Bruce River Group due to the more inhomogeneous rheological properties of these units compared to the upper, dominantly felsic, volcanic rocks. The brittle deformation which produced pathways, and hence depositional sites, for base metal-bearing fluids was presumably produced during the Grenvillian Orogeny. Chemical data indicate that the fluids were enriched essentially in Cu, Zn, Pb, Fe, S, SiO_2, and $CaCO_3$.

Résumé

La séquence supracorticale du Protérozoïque moyen du groupe de Bruce River contient des manifestations de cuivre (±zinc et /ou plomb) localisées dans des failles, des fractures et des zones de cisaillement ou les deux. La minéralisation la plus forte se produit dans les unités sédimentaires stratigraphiquement inférieures du groupe de Bruce River due aux propriétés rhéologiques les plus hétérogènes de ces unités comparées à celles des roches volcaniques, surtout felsiques supérieures. La déformation fragile qui produit des passages et donc des sites de dépôt pour des fluides contenant des métaux communs, a probablement été produite pendant l'orogénie de Grenville. Des données chimiques indiquent que les fluides ont été enrichis essentiellement de Cu, de Zn, de Pb, de Fe, de S, de SiO_2 et de $CaCO_3$.

[1] Contribution to the Canada — New Foundland Mineral Development Agreement 1984-1989. Project carried by Geological Survey of Canada, Mineral Resources Division.
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INTRODUCTION

This sampling project was undertaken to re-examine the numerous (over eighty-five according to Newfoundland Department of Mines (NDM) Mineral Inventory Files) copper occurrences in the Bruce River Group. The ultimate aim of the study is to determine a genetic model for the mineralization, including sources of the components, fluid composition and transport mechanisms; to determine the relationships between ore-forming structures and Grenvillian tectonism; and to provide elemental concentration data on the showings.

At the time of writing (August 31, 1987) eighty-eight samples from the Cu vein systems and host rocks have been analysed geochemically. The material collected represents 'grab samples' of the showings rather than systematic 'chip samples'. Geochemical data collected to date consist of X-Ray fluorescence analyses of whole rock pressed powder pellets.

Fluid inclusion sections have been prepared and will be examined at Memorial University (MUN). Sulphide and galena separates are awaiting sulphur and lead isotope analyses respectively at the Geological Survey of Canada. Carbonate separates from the vein systems will undergo C (and O) isotope analyses at MUN. Further laboratory work to be carried out include precious metal, major element and REE analyses. These will be coupled with ore microscopy, electron microprobe and SEM examination of sulphide specimens. On a larger scale, the structural settings of the vein systems will also be evaluated to resolve stress regimes.

GEOLOGICAL SETTING OF THE Cu OCCURRENCES

The Bruce River Group (Baragar, 1981; Ryan, 1984; Ryan et al., in press) is a Middle Proterozoic supracrustal assemblage of mixed sedimentary and volcanic lithologies in the central part of the Central Mineral Belt of Labrador (Fig. 1). The group contains three formations: the Heggart Lake Formation, a fluvial sequence of conglomerates and arkosic sandstones; the Brown Lake Formation, a sequence of conglomerate and volcaniclastic sandstone; and (the stratigraphically highest unit) the Sylvia Lake Formation, a mixed felsic (dominantly) and mafic volcanic sequence with flows, welded ash flow tuffs, agglomerates and subvolcanic intrusives.

Both the Heggart Lake and Brown Lake formations rest with marked angular unconformity on the Early Proterozoic Moran Lake Group. The Sylvia Lake Formation is in turn unconformably overlain by the ca. 1300 Ma Seal Lake Group (Ryan, 1984). Scharer et al. (in prep.) have obtained a 1649 Ma U-Pb isotope age for zircon separates from a felsic volcanic unit of the Sylvia Lake Formation.

COPPER OCCURRENCES IN THE BRUCE RIVER GROUP

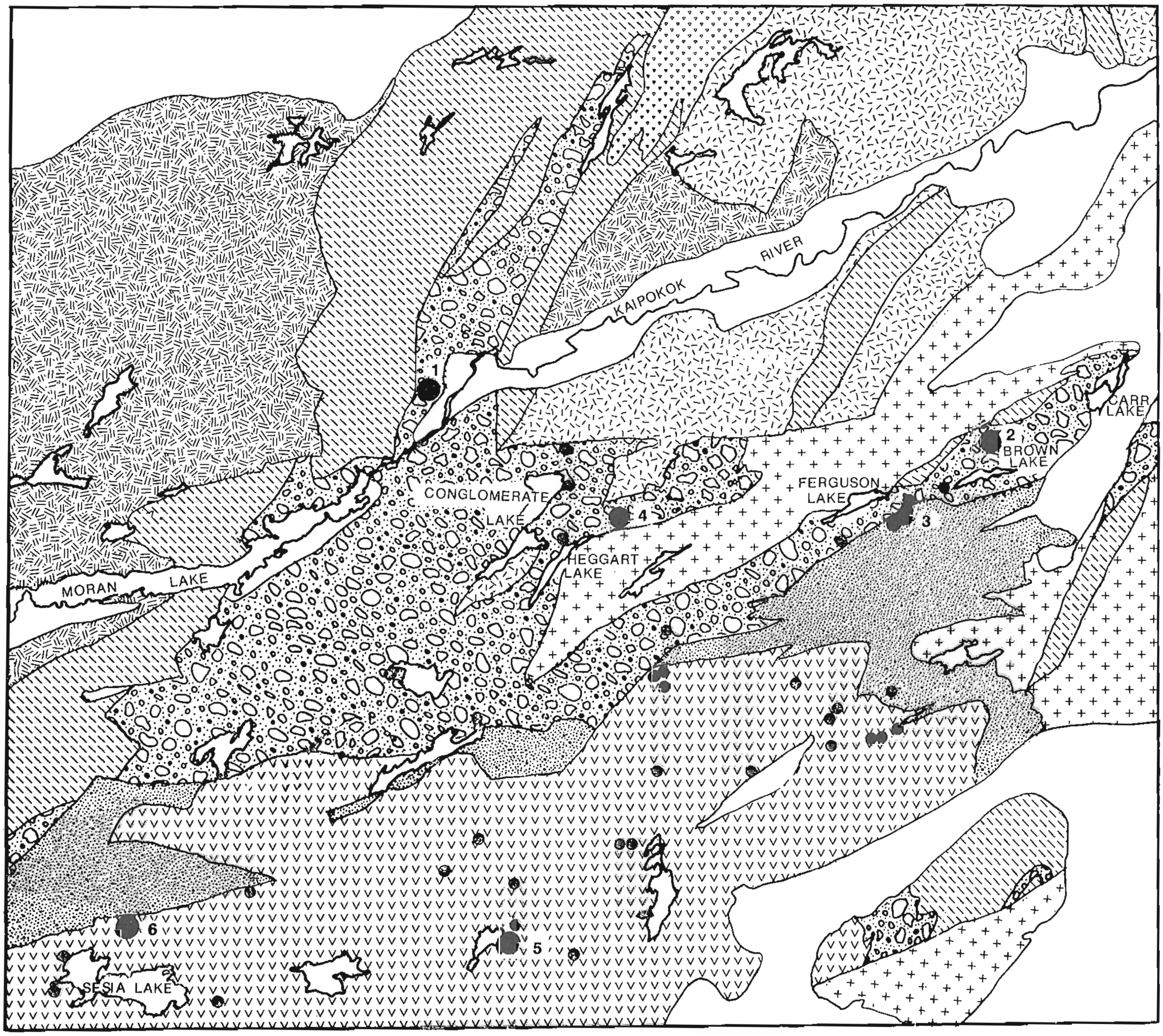
Most of these showings were discovered by prospectors and geologists with AMCO Limited during exploration from 1952-1954 and are documented in unpublished company report by H.G. Macpherson, 1954; and J.C.G. Moore, 1952, 1954, and, except for the limited re-mapping of some by Brinex Ltd. geologists J.E. Collins, and M.J. Piloski in 1958 and NDM geologists in 1974 (Ryan , 1984), have not been investigated in detail since. Other small occurrences in the area have also been reported by geologists from Brinex (M.J. Piloski, 1959), Mokta Canada Limited (J. Bernazeud, 1965), Shell Canada Resources Limited (W.L. McKenzie, 1978), and CANICO Limited (J. Perry, 1980).

The copper showings occur at all stratigraphic levels within all members of the Bruce River Group (Ryan, 1984), *viz.* the basal sedimentary Heggart Lake and Brown Lake formations, and the uppermost, dominantly felsic volcanic, Sylvia Lake Formation. The relatively more significant showings, however, occur within the two lower sedimentary units. The sulphide minerals form stringers in fracture/shear-fillings or in quartz-carbonate (+epidote) vein systems that fill brittle structures ranging from faults to fractures. The inherent anisotropic response to imposed stress of the lower sedimentary (especially conglomeratic) units within the Bruce River Group (compared to the more rheologically homogeneous felsic volcanic members), resulted in more fracture/shear channelways and hence the more significant showings. R. Wardle (pers. comm., 1987) suggests the structural control may be a function of the basement directly underlying the sedimentary formations. At a number of showings (*eg.* NDM 13K/10/Cu012, and 13K/07/Cu021) the quartz veins occur in conjugate sets; in other occurrences (*eg.* the Brown Lake and Heggart Lake Showings) chalcopyrite occurs as conjugate fracture-filling stringers enclosing lithic fragments.

Mineralization within most occurrences consists mainly of chalcopyrite and pyrite with malachite (± azurite) staining; bornite and/or chalcocite are present in a few areas. The hosting vein material is either quartz, or carbonate, or a combination of quartz and carbonate; where paragenetic sequences are visible (*eg.* NDM 13K/07/Cu004) it appears that quartz precipitated first in the opening shear/fracture followed by the sulphide and carbonate which precipitated after slight deformation of the pre-existing quartz. Sericite and/or brown weathering carbonate are developed as alterations of the wall rocks (*i.e.* vein selvages).

The most significant occurrences are the Brown Lake, Ferguson Lake and Heggart Lake Showings (Fig. 1). The Heggart Lake Showing consists of eight trenches in sheared conglomerate of the Heggart Lake Formation. Chalcopyrite and bornite occur in both carbonate veinlets and stringers. The conglomerate has a red hematitic alteration and also some sericitization around the sulphide zones. In portions of the zone, alteration is so intense that clasts within the conglomerate are not recognizable. AMCO Ltd. drilled 18 holes with a cumulative depth of about 1500 m on this showing in 1953; the highest assay obtained was 2.52 % Cu over 5 feet (1.5 m).

The Ferguson Lake Showings comprise eight small showings occurring over a distance of 3.3 km. The largest showing is the Ferguson Lake No. 4 in which the best assay according to J.C.G. Moore was 2.14 % Cu over 5 feet (1.5 m). This showing is exposed in four trenches and in 1954 AMCO drilled a single hole through it. Two smaller occurrences, Ferguson Lake Nos. 3 and 1, have small trenches. The host rock to the Ferguson Lake No. 4 Showing is



Glacial Till

Granitoid Intrusives

BRUCE RIVER GROUP

Sylvia Lake Formation

Brown Lake Formation

Heggart Lake Formation

Moran Lake Group and Equivalents

Ballet Pond Schists

Archean Metasediments

Archean Granite/Gneiss

Copper Occurrence

0 1 2 3 4 miles

0 1 2 3 4 5 km

Figure 1. Geological map of the Bruce River Group and surrounding rocks in the Moran Lake area, Labrador Central Mineral Belt (from Ryan, 1984). The locations of some of the copper occurrences examined are indicated by dots, the numbered occurrences refer to those specifically mentioned in the text; No. 1 = (NDM Showing 13K/10/Cu012), No. 2 = Brown Lake Showing, No. 3 = Ferguson Lake Showings (esp. #'s 1, 3, and 4), No. 4 = Heggart Lake Showing, No. 5 = (NDM Showing 13K/07/Cu017), and No. 6 = (NDM Showing 13K/07/Cu004).

sheared grey arkose of the Brown Lake Formation that is altered to sericite and/or chlorite, and is cut by numerous carbonate veinlets. The mineralization consists of chalcopyrite-pyrite (± bornite) veins and stringers zones (up to 1 m wide) with malachite staining.

The Brown Lake Showing consists of over a dozen stripped outcrops and four blasted trenches. The host rock is sheared/fractured and chloritized Heggart Lake Formation conglomerate. Chalcopyrite (plus pyrite and minor bornite) fills a conjugate fracture set and often the sulphide stringers wrap around lithic clasts. In one trench, a chalcopyrite-bornite vein (up to 7.5 cm thick) is apparently tightly folded. According to J.C.G. Moore, these trenches were the only ones, of all Cu showings in the Bruce River Group, that were analysed for Ag and Pb. One sample had 1.16 % Cu, 0.02 % Pb and 0.52 oz/t Ag over 3 feet (0.9 m); while another contained 1.15 % Cu, 0.02 % Pb and 0.22 oz/t Ag over 5 feet (1.5 m).

Regional fault structures cutting the Bruce River Group contain large masses of quartz vein filling, indicative of the very large scale post-deformation fluid flow through the cataclastic structures. These vein systems have been sampled. The rather small chalcopyrite-bearing quartz vein at NDM showing 13K/07/Cu021 occurs at the intersection of two large faults east of Sylvia Lake and is associated with a massive, barren, quartz vein stockwork over 10 x 15 m in size. Approximately 6 km north of Brown Lake there is a northeasterly-trending linear ridge (up to 1 x 0.25 km) made up of 2-3-m-thick quartz veins cutting a leucogranite. This vein zone occurs along the trace of a fault. The vein material is dominantly quartz with very minor feldspar and virtually no sulphide, though there are small rusty patches.

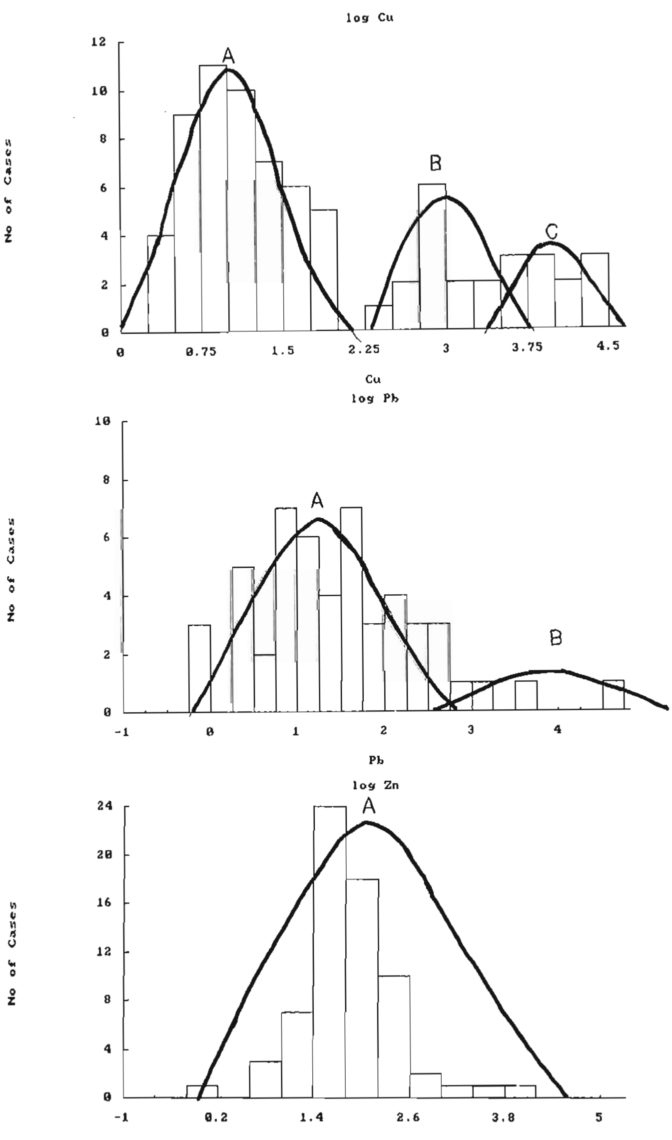


Figure 2. Log normal histograms for the Cu, Pb and Zn contents in samples from the copper occurrences hosted by the Bruce River Group.

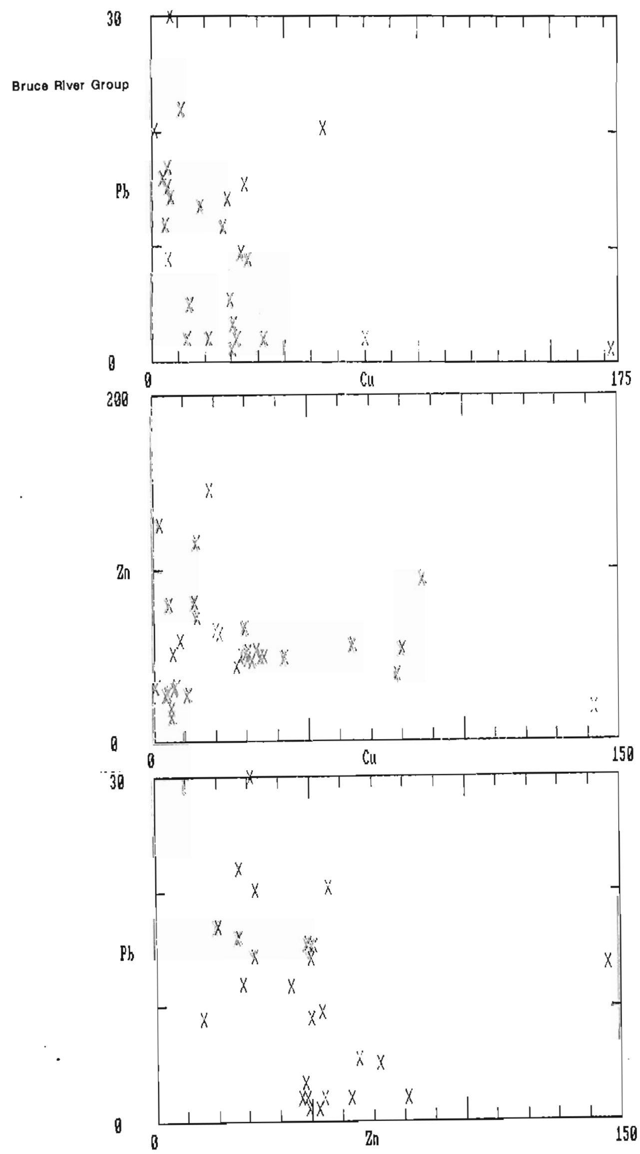


Figure 3. Base metal variation diagrams for samples from the copper occurrences in the Bruce River Group.

Since most of the fault systems and shear zones in the region postdate the Hudsonian orogeny (as the Bruce River Group has been dated at 1649 Ma) and have a general northeast to east strike, they are presumably the products of Grenvillian deformation. More work remains to be completed on the structural setting of the occurrences. In the following section, the implications of the geochemical data to probable sources of the metallic components are discussed.

GEOCHEMISTRY OF THE Cu OCCURRENCES

Base metal concentration data for 88 samples from the Bruce River Group Cu occurrences indicate lognormal distributions of Cu (max. 41868; min. 0; mean 3035 ppm), Pb (max. 60382; min. 0; mean 1490 ppm), and Zn (max. 219896; min 0; mean 3100 ppm). In the case of each element, concentrations are very strongly skewed to the background end; 53 (or 60 %) samples had < 100 ppm Cu, 65 (or 74 %) had < 100 ppm Pb and 52 (or 59 %) had < 100 ppm Zn.

When the Cu, Pb and Zn data are log transformed, the values have more normal Gaussian distributions (Fig. 2). Zn values have a unimodal dis-tribution with a geometric mean (i.e. visually estimated) of about 100 ppm. Log Cu values (Fig. 2a) seem to indicate the presence of at least two, and probably three, populations. The population at the lower end of the Cu value spectrum (i.e. A) contains most of the samples (mean of about 10 ppm), and can be described as containing all those values < 100 ppm. The second population (B) contains values between < 178 and < 1700 ppm with a mean of about 1000 ppm. The final, and most anomalous, population (C) contains all of those samples with Cu contents of > 1700 ppm and has a geometric mean of about 10 000 ppm. Samples in population C are restricted to those showings hosted by the basal sedimentary units (i.e. the Heggart Lake and Brown Lake formations) of the Bruce River Group, and not those in the volcanic Sylvia Lake Formation (excepting samples CM86-71 and -72 from around NDM Showing 13K/7Cu008).

Pb contents have a skewed unimodal distribution that may actually indicate the presence of two populations; the dominant (with lower values) population (A) has a mean of about 18 ppm and values are < 316 ppm, the second (anomalous) population (B) would include those values > 316 ppm. Significant Pb concen-trations within the anomalous second population were only found in samples from the Ferguson Lake no. 1, no. 7, Lead, Sesia Lake and West Sylvia Lake Showings (refer to Table 1).

Data obtained by the author and others on the regional background abundances of Cu, Pb and Zn (compiled from 37 samples collected in the Bruce River Group) indicate Cu values are log normally distributed ranging from 0 to 172 ppm, with a mean of 37.5 ppm, and with a geometric mean at about 11 ppm. Pb contents are likewise log normally distributed (max. 30; min. 0; mean 8 ppm). Zn data (max. 146; min. 0; mean 56 ppm) have a normal distribution at lower concentration levels and four samples with "anomalously higher" values (> 110 ppm Zn); the geometric mean of the distribution is about 50 ppm.

Table 1 lists those samples which have the maximum Cu (>500 ppm), Pb (>200 ppm) and Zn (>200 ppm) concentrations, and the showings in which they occur. The concentration limits for this table are much higher than the "background" values defined in the Bruce River Group (i.e. < 175 ppm Cu, < 35 ppm Pb and < 175 ppm Zn) and therefore represent truly anomalous "ore" values.

When the base metal distributions are compared to each other in the samples from the Cu occurrences (Fig. 3), the only definitive correlation is the positive one between Pb and Zn. There is a slight positive relationship between Cu and Zn. When Cu and Pb are compared, there is apparently a slight positive correlation with Cu values < 1000 ppm, but in samples with > 1000 ppm Cu, there appears to be a negative correspondence. In contrast, within the Bruce River Group regional (i.e. unmineralized) samples (Fig. 4), Pb and Zn have a negative relationship and there are no correlations between Cu-Pb and Cu-Zn.

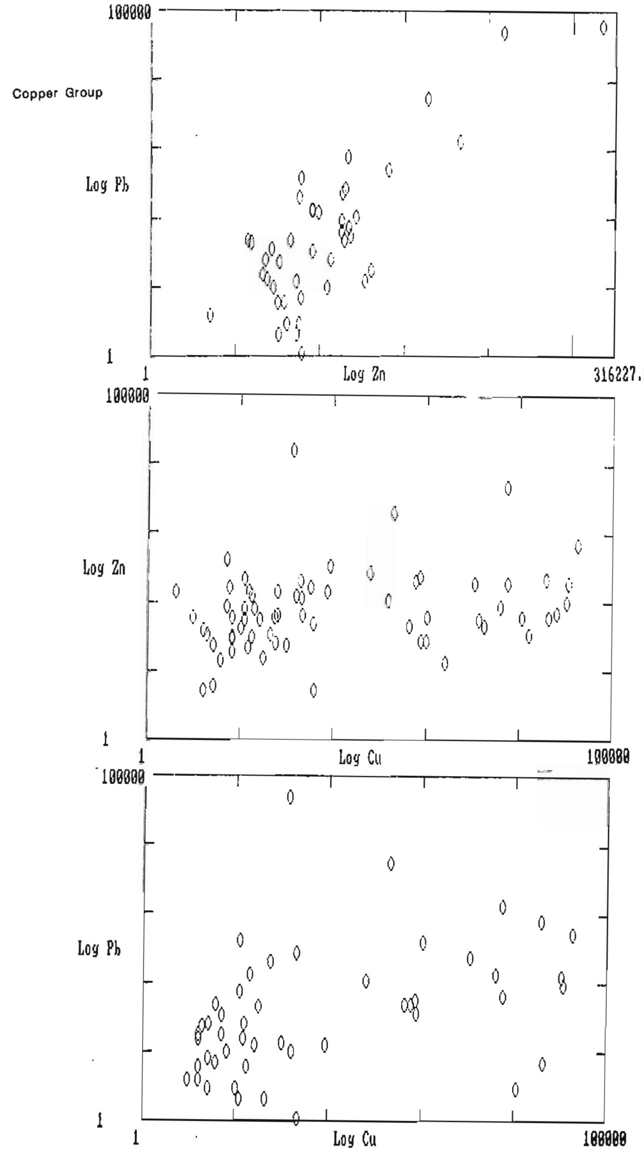


Figure 4. Base metal variation diagrams for unmineralized (background) lithogeochemical samples from the Bruce River Group.

On ternary elemental plots (Fig. 5) the compositions within the Cu occurrences contrast with those of the Bruce River Group. The Bruce River Group samples are Cu (in terms of the Cu apex), U and Pb-poor, and Zn-rich when compared to the Cu occurrences.

CONCLUSIONS

Among the numerous fracture/fault-controlled copper occurrences in the Bruce River Group, the greatest concentrations of economically interesting elements are within the lower sedimentary formations rather than the upper volcanic members. Zn and/or Pb enrichments are found in some of the occurrences but no anomalous concentrations of other elements were observed in any of them in this geochemical study.

The Pb and Zn enrichments suggest affinity of the copper occurrences to the epigentic galena-sphalerite veins that occur along the Moran Lake Group - Archean basement unconformity (Wilton et al., 1988). These latter vein systems have been modelled as the products of a Grenvillian orogeny-induced mobilization from the Archean basement. Further data must be collected from both vein systems to synthesize, compare and contrast the genetic models for these two different epigenetic vein systems.

Table 1. Analyses of selected anomalous grab samples from copper occurrences in the Bruce River Group, Labrador

Sample	Showing Name	NDME #	Cu(ppm)	Pb(ppm)	Zn(ppm)
LM86-32A	Ferguson Lake #4 (Main)	13K/10Cu009	41686	b.a.l.*	647
LM86-32B	〃	〃	b.a.l.	b.a.l.	407
LM86-31A	Ferguson Lake #1	13K/10Cu004	19577	758	216
LM86-33A	Ferguson Lake #7	13K/7Cu007	7355	1264	4581
LM86-33B	〃	〃	b.a.l.	263	200
CM86-19	Ferguson Lake #5	13K/7Cu010	3249	233	b.a.l.
CM86-21	〃	〃	b.a.l.	b.a.l.	221
W86-23	Ferguson Lake #8	13K/10Cu011	32302	b.a.l.	b.a.l.
W86-24	〃	〃	763	〃	〃
W86-26	〃	〃	33972	〃	〃
CM86-22.1	Ferguson-Brown Lake U	13K/10U003	b.a.l.	b.a.l.	269
LM86-35T5	Heggart Lake	13K/10Cu005	25355	b.a.l.	b.a.l.
LM86-36T6	〃	〃	4160	〃	〃
LM86-37	〃	〃	20855	〃	〃
LM86-39	〃	〃	3703	〃	〃
LM86-40	〃	〃	12749	〃	〃
W86-19	Brown Lake	13K/10Cu002	6251	b.a.l.	b.a.l.
W86-20	〃	〃	7440	〃	〃
W86-22	〃	〃	849	b.a.l.	229
CM86-72	n.n.**	13K/7Cu008	10554	b.a.l.	b.a.l.
CM86-71	n.n.	〃	2242	b.a.l.	b.a.l.
W86-34	Sesia Lake	13K/7Cu050	1022	380	b.a.l.
LM86-42	Sesia Lake	13K/7Cu003	648	b.a.l.	b.a.l.
CM86-26A	N of Sesia Lake	n.n.	873	b.a.l.	b.a.l.
CP86-1	Conglomerate Lake #3	13K/10Cu026	1592	b.a.l.	b.a.l.
W86-12	Moore's (1954) #1	13K/10Cu012	1884	b.a.l.	b.a.l.
LM86-46	Chaulk Lake	13K/7Cu007	992	b.a.l.	b.a.l.
W86-36A	n.n.	13K/7Cu005	641	b.a.l.	b.a.l.
LM86-34B	Ferguson Lake Pb	13K/10Cu050	b.a.l.	60382	219896
LM86-34C	〃	〃	b.a.l.	5458	1903
W86-46	〃	〃	b.a.l.	48039	15051
CM86-73A	West Sylvia Lake	n.n.	b.a.l.	405	b.a.l.
LM86-77	Vee Lake	13K/7Cu002	b.a.l.	199	b.a.l.
W86-32	Cecil Lake	13K/10Cu013	b.a.l.	b.a.l.	339

*b.a.l. = below anomalous limit **n.n. = no name or no number

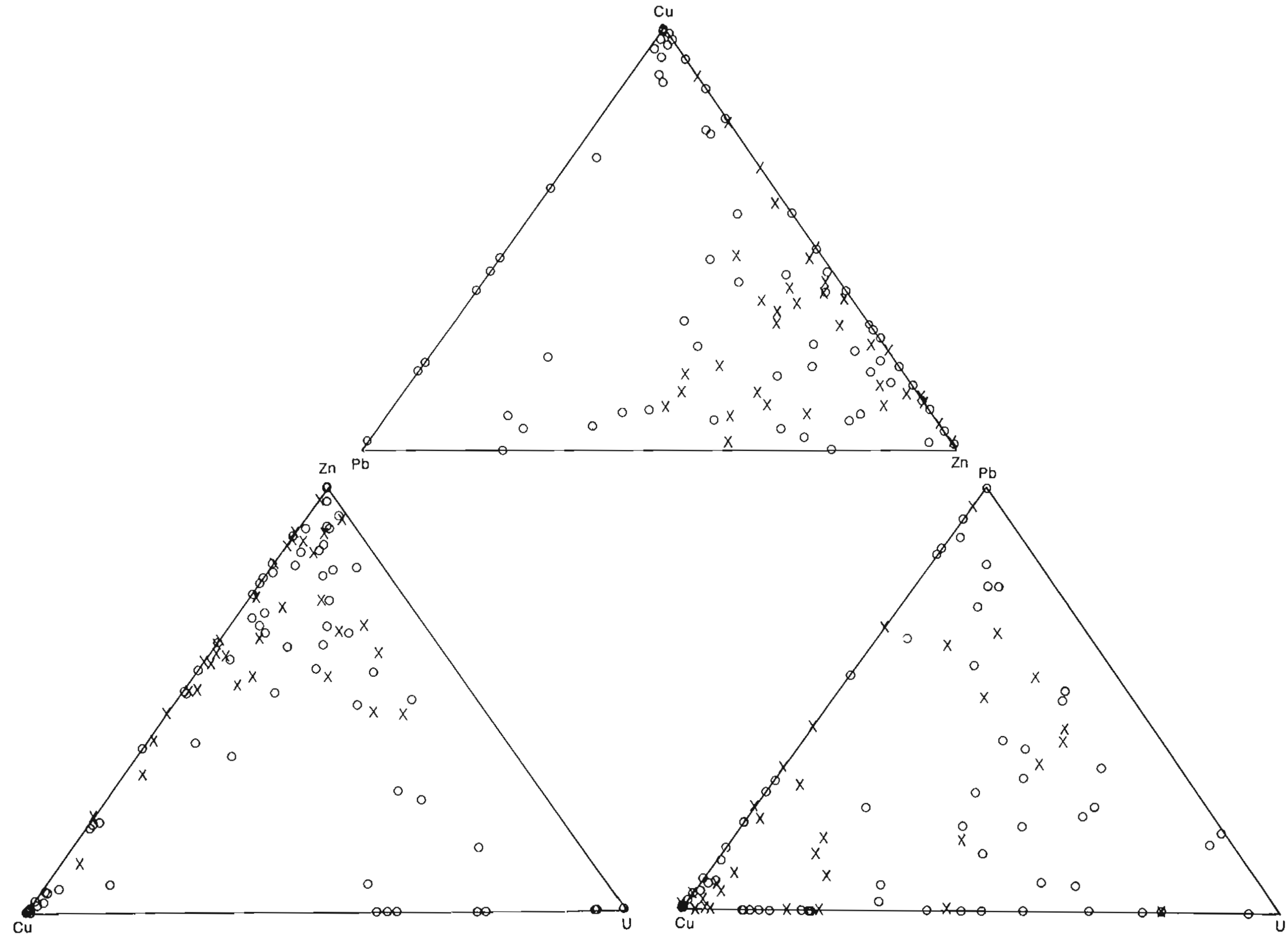


Figure 5. Ternary metal plots for samples from copper occurrences in the Bruce River Group (o) and unmineralized Bruce River Group (x).

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A preliminary analysis of middle Proterozoic karst development and bitumen emplacement, Parry Bay Formation (dolomite), Bathurst Inlet area, District of Mackenzie

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Pelechaty, S.M., Grotzinger, J.P., Goodarzi, F., Snowdon, L.R., and Stasiuk, V., A preliminary analysis of middle Proterozoic karst development and bitumen emplacement, Parry Bay Formation (dolomite), Bathurst Inlet area, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 299-312, 1988.

Abstract

A middle Proterozoic paleokarst profile in the Elu Basin is developed within the upper part of a sequence of peritidal cyclic carbonates (Parry Bay Formation). Spectacular collapse sinkholes, caves, grikes, and associated collapse breccias define the paleokarst profile along a 65-km discontinuous outcrop belt within the Bathurst Inlet area. During subaerial exposure the Parry Bay Formation underwent a period of karstification whereby carbonate dissolution occurred predominately along a northeast-southeast paleojoint system. During a subsequent period of transgression, the karst topography was reworked in part and overlain by shallow marine sediments (Kanuyak Formation). Development of sheet cracks, pisolites, and void-filling cements on both the Parry Bay and lower Kanuyak formations suggests modification of the sediment during later, minor events of subaerial exposure.

Bitumen was discovered within Parry Bay dolomites near Walker Bay on Kent Penninsula. Petrographic and Rock-Eval/TOC analysis indicates the bitumen was initially gilsonite and was later subjected to sudden pressurized thermal shock. The temperatures were unobtainable during "normal" burial processes and were probably associated with high temperature hydrothermal fluids related to the extrusion of the Ekalulia flood basalts.

Résumé

Dans le bassin d'Élu, un profil de paléokarst du milieu du Protérozoïque s'est développé dans la partie supérieure d'une séquence de carbonates cycliques péritidaux (formation de Parry Bay). Des dolines spectaculaires, des grottes, des lapiés et des brèches d'effondrement associées définissent le profil de paléokarst le long d'une zone d'affleurement discontinu de 65 km dans la région de l'inlet Bathurst. Lors de l'exposition subaérienne, la formation de Barry Bay a subi une période de karstification pendant que la dissolution des roches carbonatées se produisait surtout le long d'un réseau de paléodiaclases nord-est-sud-est. Plus tard, la topographie karstique a été retravaillée en partie et recouverte par des sédiments marins peu épais (formation de Kanuyak) pendant une période de transgression. La formation de diaclases horizontales, de pisolites et de ciments remplissant des interstices sur les deux formations de Parry Bay et de Kanuyak inférieur suggère une modification du sédiment pendant des événements moins importants de l'exposition subaérienne ultérieure.

On a découvert du bitume dans les dolomites de Parry près de Walker Bay sur la péninsule Kent. Les analyses pétrographiques et d'évaluation des roches/COT indiquent que le bitume était à l'origine de la gilsonite soumise plus tard à un choc thermique soudain sous pression. Les températures étaient impossibles à obtenir pendant des processus d'enfouissement «normaux» et ont été probablement associées à des fluides hydrothermiques de températures élevées en ralation avec l'épanchement des basaltes du plateau d'Ekalulia.

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INTRODUCTION

The middle Proterozoic Parry Bay Formation is currently being studied in order to document the development of a major karst horizon of regional extent at the top of the formation, and also to examine the occurrence of void-filling bitumen deposits located stratigraphically below the karst horizon. The Parry Bay Formation is preserved in the Elu Basin, northeastern Slave Province (Fig. 1).

The paleokarst profile is best exposed along the eastern shorelines of the three largest islands within Bathurst Inlet. These islands define a discontinous northwest-trending outcrop belt, extending from Bear Island, in the southeastern part of Bathurst Inlet, to Kanuyak and Ekalulia islands in the north-central part of the Inlet (Fig. 2). The islands are capped by columnar jointed flood basalts of the Ekalulia Formation which produces excellent cliff exposures of the underlying Kanuyak and Parry Bay formations. These exposures allow measurement of complete vertical sections of the karst profile, as well as the opportunity to trace laterally all pertinent contacts in order to define the stratigraphic relationships between component lithofacies.

The main objectives of the karst study (SMP, JPG) are to 1) document the stratigraphic relationships between the paleokarst profile and the underlying cyclic carbonates, and the siliciclastic deposits of the overlying Kanuyak Formation, and 2) to interpret the variety of fabrics developed within the Parry Bay collapse breccias and within the basal part of the overlying Kanuyak Formation. Field studies will be supplemented by petrographic and geochemical work in order to unravel the diagenetic history associated with the karstification.

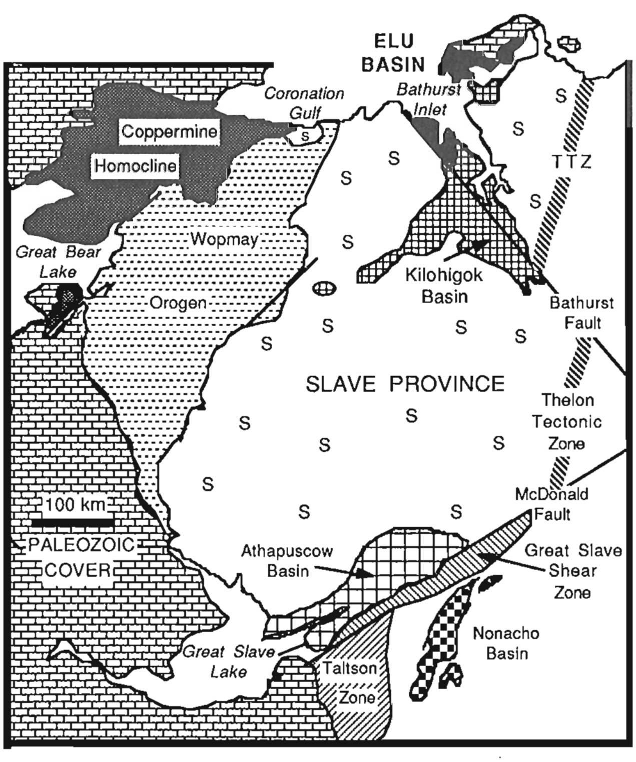

Figure 1. Location of Elu Basin (black) relative to other main tectonic elements within the northwest Canadian Shield.

PREVIOUS WORK

Initial mapping of middle Proterozoic rocks within the Bathurst Inlet area was conducted by O'Neill (1924). Subdivision and description of the main stratigraphic units was initiated during various follow-up reconnaissance programs (Lord, 1953; Wright 1957, Fraser 1960, 1963; Tremblay, 1967; Fraser and Tremblay, 1969; Fraser et al., 1970; Tremblay, 1971).

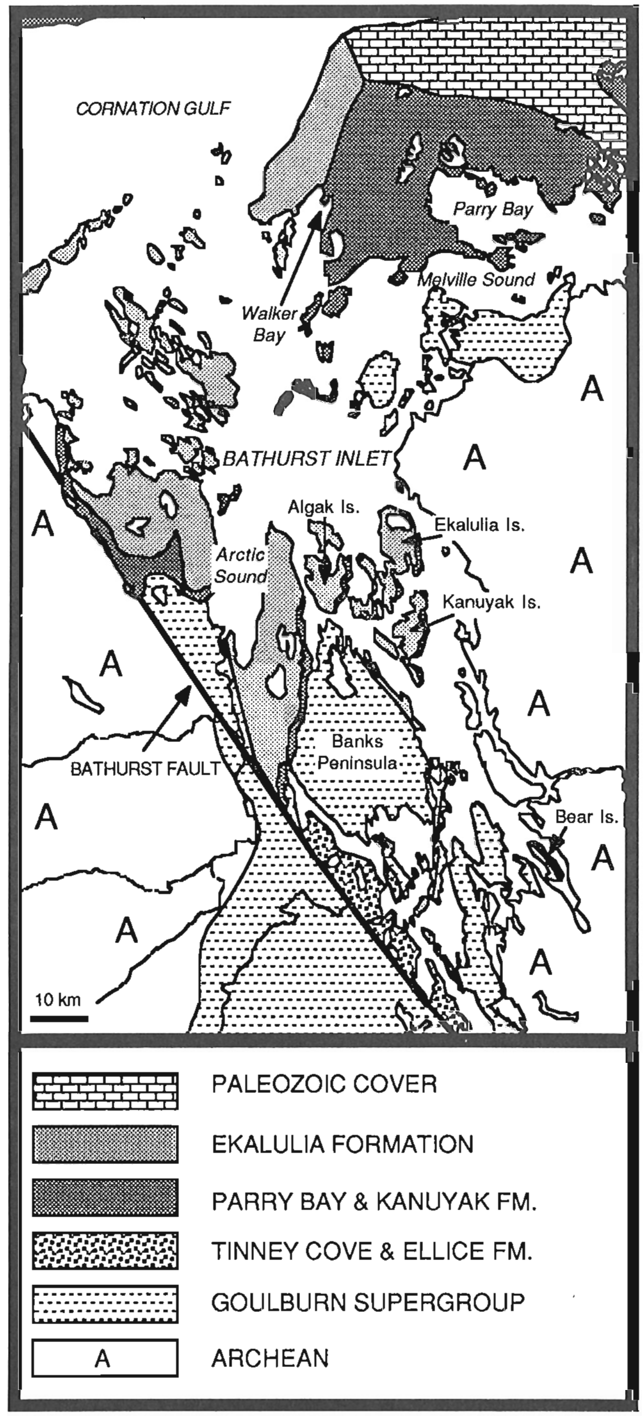

Figure 2. Location of Parry Bay, Kanuyak and Ekalulia formations, Bathurst Inlet area, Northwest Territories. Note location of Bear Island, Kanuyak Island, and Ekalulia Island. Modified after Campbell (1978).

Campbell (1978, 1979) first documented the occurrence of large-scale oligomictic breccias on top of the Parry Bay Formation. In addition, thin (< 2m) breccia zones (grikes) up to 75 m below the base of the Kanuyak Formation were recognized. These were interpreted as solution collapse deposits produced during a prolonged period of pre-Kanuyak uplift and erosion, whereby, solution collapse was initiated by dissolution of evaporites within the Ellice Formation.

Kerans (1982) documented a karst collapse breccia developed within the Dismal Lakes Group in the Coppermine Homocline, 300 km to the west, which is believed to be in part correlative with the Parry Bay Formation (Fraser and Tremblay, 1969).

REGIONAL GEOLOGY

Elu Basin is located in the northeastern portion of the Slave Province (Fig. 1). Sedimentary and volcanic rocks of the Elu Basin are exposed in the northwestern part of Bathurst Inlet area just east of Bathurst Fault, at the southern and eastern parts of Victoria Island, Kent Penninsula, and on islands within Bathurst Inlet. Figure 3 illustrates the stratigraphy of Proterozoic units within the Bathurst Inlet area. The basal unit (Ellice Formation) rests unconformably on older (early Proterozoic) fanglomerates of the Tinney Cove Formation, which in turn unconformably overlie sediments of the early Proterozoic Goulburn Supergroup (Fig. 3; Fraser, 1963; Campbell, 1978).

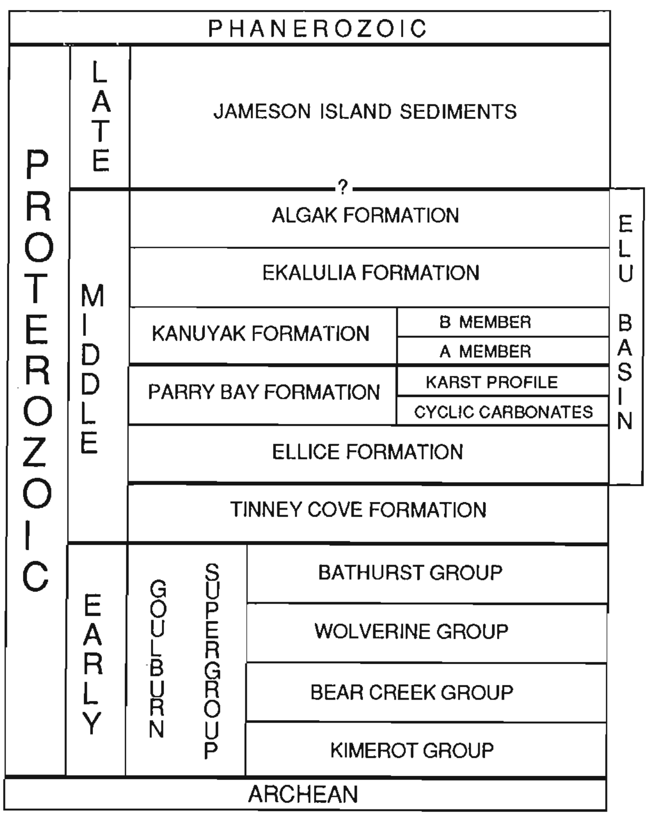

Figure 3. Stratigraphic nomenclature for Proterozoic rocks within the Bathurst Inlet area. Darkened lines represent major unconformities. Modified after Campbell (1978, 1979).

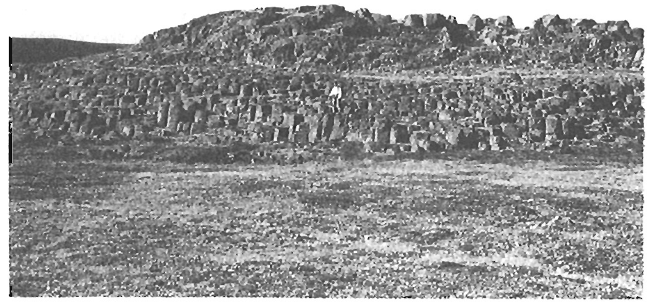

Figure 4. Ekalulia Formation flood basalts with well developed columnar jointing, South end of Kanuyak Formation.

The Parry Bay Formation conformably overlies the Ellice Formation (Campbell, 1978, 1979) and the contact is gradational at the southern end of Bear Island . Bedded carbonates within the upper parts of the Parry Bay Formation have been karsted and are characterized by sinkholes, caves, and grikes with associated collapse breccias. The collapse breccias at the top of the Parry Bay Formation are unconformably overlain by predominately siliciclastic sediments of the Kanuyak Formation. The paleokarst profile is developed within the upper part of the Parry Bay Formation. To date, the karst-related breccias have been shown to extend from the southern tip of Bear Island to the northern end of Ekalulia Island (Fig. 2). Karst is not known from the sections of Kent Penninsula which are thicker than those of the more southerly Bathurst Inlet areas. Bitumen-bearing dolomites were found approximately 100 m below the contact with overlying units.

A succession of flood basalts (Ekalulia Formation), up to 500 m thick, overlies the Kanuyak Formation and rests directly on the Parry Bay Formation where the Kanuyak sediments are absent (Campbell, 1979). These basalts are most likely correlative with the Coppermine Group flood basalts of the Coppermine Homocline (Campbell, 1978). Spectacular columnar jointing and large amygdules filled with quartz, calcite, and copper mineralization are commonly associated with these basalts (Fig. 4). The middle Proterozoic rocks for the most part are unmetamorphosed, excepting contact metamorphism related to outpouring of Ekalulia Formation flood basalts and the intrusion of Late Proterozoic diabase sills and dykes. Within the south and central parts of Bathurst Inlet the rocks are block faulted placing, in areas, middle Proterozoic sediments adjacent to Archean basement rocks. Sediments within the Parry Bay Formation rarely dip greater than 20°.

STRATIGRAPHIC RELATIONSHIPS: PARRY BAY KARST

Three cross-sections were measured along the eastern shorelines of Bear Island (Fig. 5a), Kanuyak Island (Fig. 5b), and Ekalulia Island (Fig. 5c). The base of the Ekalulia Formation was used as the datum in all three cross-sections.

The paleokarst profile is mainly expressed by large sinkhole features developed along the upper part of the Parry Bay Formation and filled with collapse breccias and siliciclas-

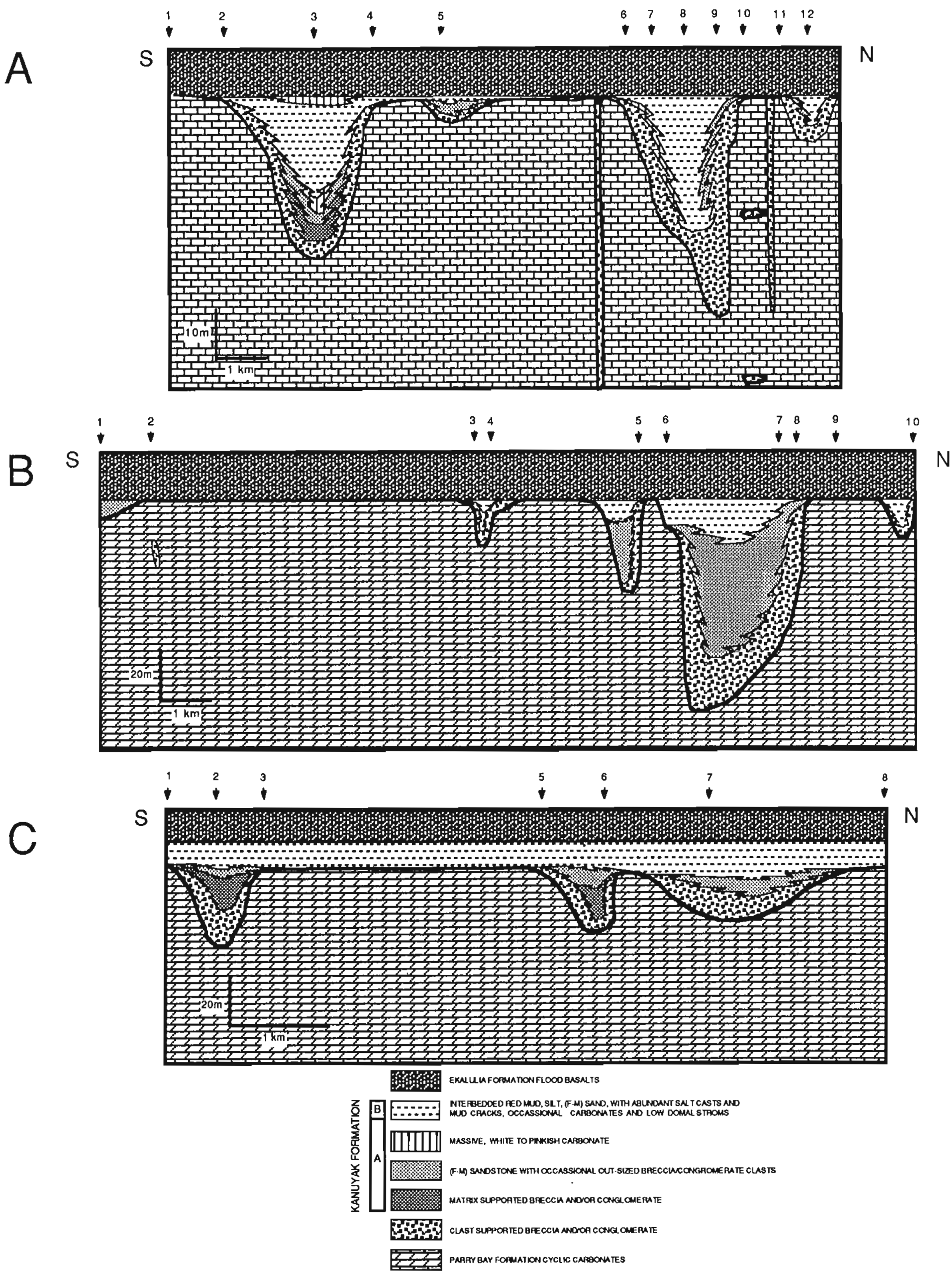

Figure 5. Stratigraphic cross-sections of Parry Bay karst profiles relative to underlying cyclic carbonates, and overlying Kanuyak and Ekalulia formations. Base of flood basalts used for datum. A) Bear Island cross-section: sections 1 and 12 located at the southern and northern most ends of the island respectively. Note: grikes near section 6 and section 11 approximately 4.5 m wide. B) Kanuyak Island cross-section: Sections 1 and 10 located at extreme ends of island. C) Ekalulia Island cross-section: section 8 located approximately 7.5 km south of northern end of island. See Figure 2 for location of islands.

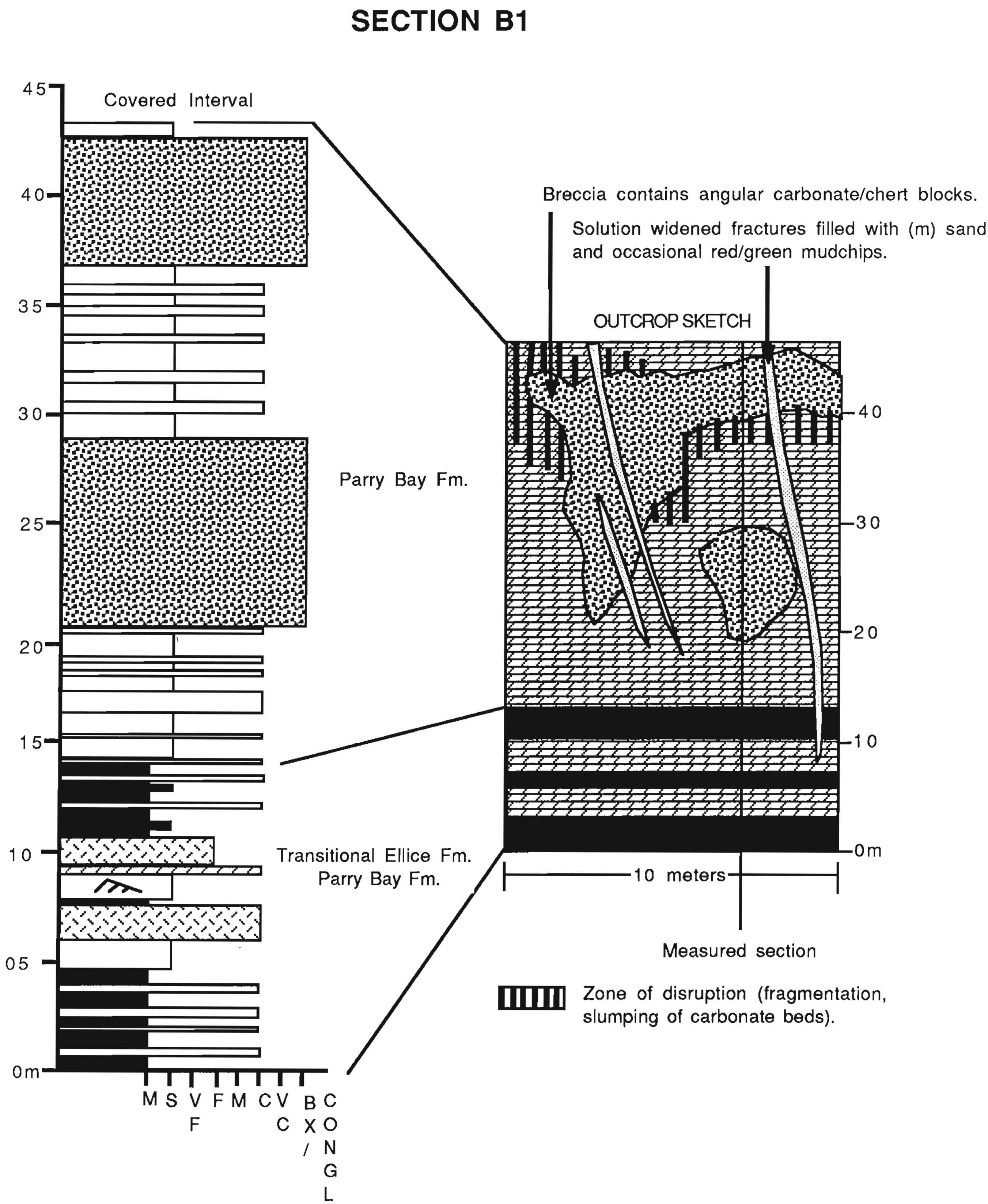

Figure 6. Stratigraphic column located at section 1, south end of Bear Island. Note gradational contact between Parry Bay Formation and Ellice Formation. See Figure 7 for key to lithological symbols.

tic sediments of the Kanuyak Formation. Other karst features such as caves and deep fractures filled with sediment occur up to 180m below the upper Parry Bay contact. The bedded carbonates and breccias are unconformably overlain by siliciclastic and minor carbonate material of the Kanuyak Formation. A disrupted zone is developed within the upper part of the collapse breccia and within the basal parts of the Kanuyak Formation. This zone is characterized by sheet cracks, pisolite, and cement fabrics.

Parry Bay Formation

The Parry Bay Formation is subdivided into two main stratigraphic units: lower cyclic carbonates, and overlying karst profile (Fig. 3). Figures 6 and 7 show detailed sections of the Parry Bay Formation peritidal carbonates measured near the south end of Bear Island; section B1 (Fig. 6) is located 1 km southwest of section B2 (Fig. 7). The carbonates now are composed of fine to medium-grained dolomite. The dolostones consist of cyclically interbedded stromatolites, cryptalgalaminites, massive to laminated dolosiltites, dololutites and grainstones. Individual cycles generally shallow upward, forming grainstone/stromatolite/cryptalgalaminite sequences. Grainstones often contain ooids, intraclasts, ripups of stromatolite and cryptalgalaminites, and occasional silicified salt casts. In outcrop, the laminites are generally recessive. Stromatolite beds consist of low relief (< 15cm) laterally linked domes less that 1 m in width . Commonly, stromatolites are associated with ooid and intraclast grainstones that occur between individual domes. Domal stromatolites locally pass laterally into stratiform stromatolites. Cryptalgalaminites are finely laminated and often exhibit paper-like partings and laminoid to irregular fenestral fabrics (Fig. 8).

Locally, continuous horizons of chert nodules up to 2 m thick occur preferentially within the laminite beds. Nodules are generally yellow with minor black nodules; fine laminations are often preserved within the chert nodules. Chert horizons are bedding-parallel and make excellent local marker horizons, useful in documentation of local relief developed as the result of karstification. Conical stromatolites have been observed on Kanuyak and Ekalulia islands in the upper part of the stratigraphy.

Paleokarst Profile

Parry Bay Formation karst facies occur at the top of and within the cyclic carbonates (Fig. 5). Work to date shows that the karst profile extends from the southern tip of Bear Island to an area approximately 7.5 km south of the northern end of Ekalulia Island (Fig. 2). Karst-related collapse breccias are best developed within large sinkhole features that are up to 80 m deep and 2.5 km wide. Other karst features such as caves and grikes occur within the upper parts of the Parry Bay Formation. The paleokarst profile is discussed below in more detail.

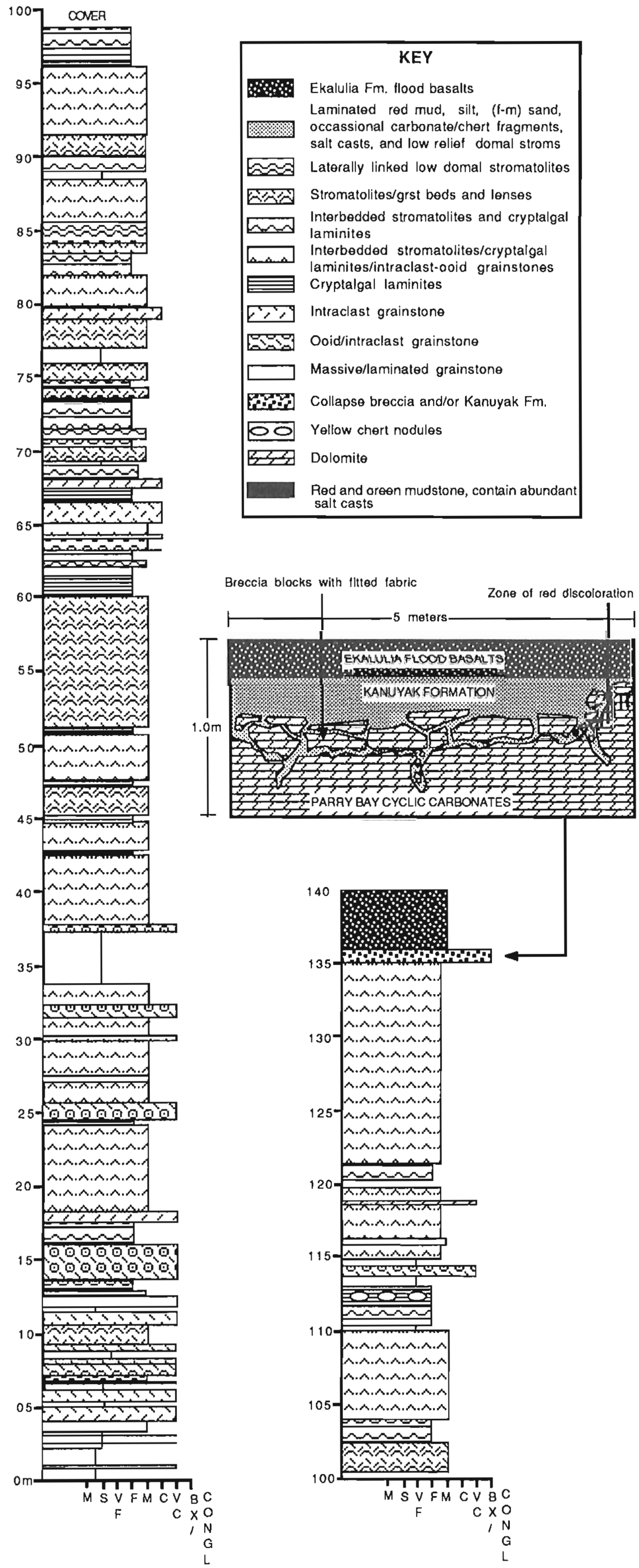

Figure 7. Stratigraphic column of Parry Bay Formation peritidal carbonates. Note poor development of breccia and thin aspect of Kanuyak Formation. Column located at section 2, south end of Bear Island.

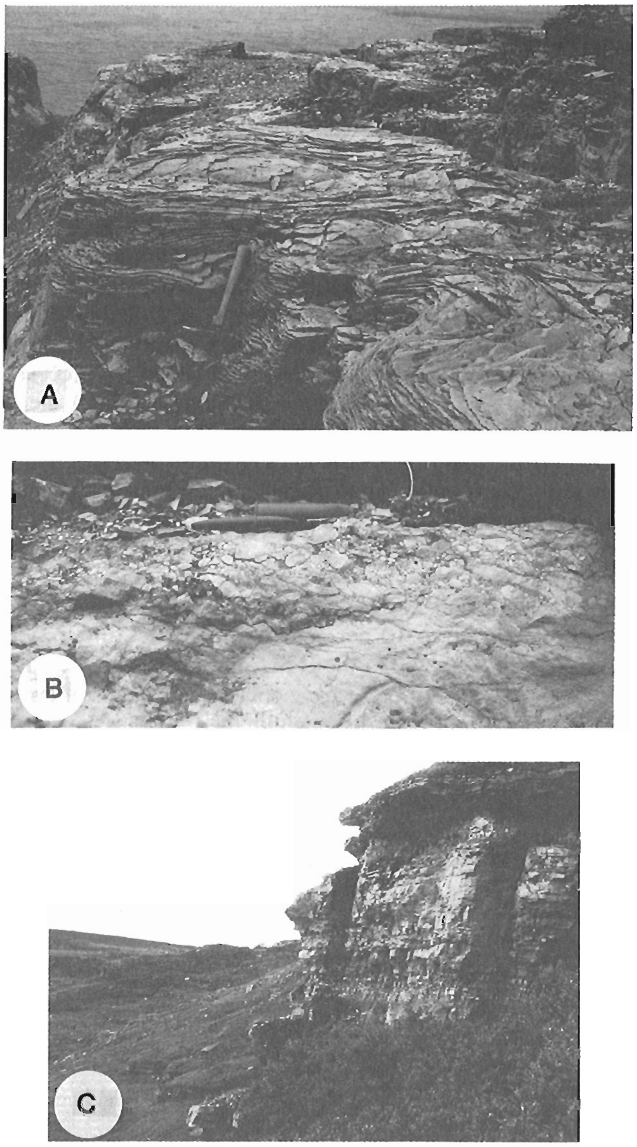

Figure 8. A) Laterally linked low domal stromatolites within Parry Bay Formation. South end of Ekalulia Island. B) Massive dolosiltite capped by thin intraformational conglomerate bed. Pen cap for scale. C) Differential weathering of Parry Bay carbonates exhibit cyclicity. Stromatolites form ridges and cryptalgal laminites and grainstones weather recessively. South end of Ekalulia Island.

Kanuyak Formation

Siliciclastic and minor carbonate sediments of the Kanuyak Formation unconformably overlie the Parry Bay Formation (Fig. 5). Sediments within the Kanuyak Formation occur primarily within karst sinkholes (Fig. 5a, 5b). The formation thickens northward and at Ekalulia Island it covers the entire Parry Bay Formation with up to 20 m of sediment in addition to filling sinkholes (Fig. 5c). The Kanuyak Formation can be subdivided into two distinct members; the A and B members.

The A member consists of sandstones, siltstones, minor carbonates and sand-supported breccias to conglomerates. These sediments fill sinkholes and are often interbedded with conglomerates derived by reworking local collapse breccias (Fig. 10a,b). Sand-supported breccias and conglomerates consist of large blocks up to 1.5 m in size. Blocks are composed of Parry Bay dolostones whereas finer material consists of both chert, carbonate, and medium to coarse sand. Sandstone beds are both tabular and lenticular-bedded and may exhibit trough crossbedding. Rare carbonates within the A member occur within the two major sinkholes at Bear Island (Fig. 5a, 9). These carbonates are typically thick (up to 5 m), massive beds of white to slightly pinkish dolomite.

The B member was deposited within the upper parts of sinkholes (Fig. 5 and 9), and as a discreet unit overlying the entire Parry Bay Formation at Ekalulia Island (Fig. 5c). Member B sediments typically consist of interbedded red mudstones and siltstones, fine-grained sandstones, and occasional (less than 10 %) stromatolitic and clastic carbonate (Fig. 10c). Individual beds are tabular and less than 10 cm thick. Abundant mudcracks, ripples, planar laminations, and mudchips are present within siliciclastic sediments. Rare salt casts occur. Stromatolitic carbonates generally occur near the basal parts of the unit while mechanically deposited carbonate with common ripples and occasional planar lamination and hummocky cross-stratification occurs near the top.

Disrupted Zone

This zone is characterized by well-developed sheet cracks, pisolite, and void-filling cement fabrics. These features are developed within the uppermost parts of the collapse breccias, and within the basal parts of the Kanuyak Formation. Detailed descriptions and stratigraphic relationships of these fabrics are discussed below.

PALEOKARST PROFILE

Description of Paleokarst Features

Various karst-related features of the Parry Bay Formation were documented on the three Islands that have been studied to date (Fig. 5). These features are primarily developed on top of the preserved Parry Bay Formation. However, other features have been documented up to 180 m beneath the upper contact of the Parry Bay Formation. Karst-related features include sinkholes developed on top of the Parry Bay Formation, carbonate breccia and sand-filled caves that occur within the Parry Bay Formation, and grikes or solution-widened fractures.

Sinkholes

Conventionally, a sinkhole is a North American term for a doline: a tub-shaped, commonly circular, depression with varying dimensions, formed by surficial limestone dissolution (Sweeting, 1973; Cvijic, 1893). The features in this paper are similar in morphology to dolines but probably developed as the result of collapse over a pre-existing cave that formed during a period of subsurface dissolution. Sinkholes or collapse sinkholes will be used to denote these features.

Collapse sinkholes were the major feature developed during karstification of the Parry Bay Formation. In cross-section individual sinkholes are up to 80 m deep and 2.5 km wide

(Fig. 5b, 12a). The outcrop belt generally provides only a two-dimensional perspective of the sinkholes, however, local deviations in outcrop trend indicate that many sinkholes are elongate. Figure 11 illustrates the orientation of sinkholes and smaller-scale solution-widened fractures; northeast and southeast trends are apparent.

Individual sinkholes are commonly lined with breccia (Fig. 5, 9) composed of carbonate and minor chert clasts. Individual clasts are stromatolite, laminite, and grainstone fragments derived from Parry Bay cyclic sediments. The breccia matrix is typically comparatively finer carbonate and chert material. Fine- to coarse-grained quartz sand is also present within the breccia matrix, and is sparse within the basal parts of deposits and abundant in the upper parts. Generally, breccias fine upward and blocks near the base are poorly sorted, usually angular to subangular and are up to 5 m in size. Upward, individual clasts are subangular to rounded, showing a higher degree of reworking. The uppermost parts of collapse breccias exhibit vague bedding and are often interbedded with matrix-supported breccias and sandstones. Between sections 6 and 10 along the Bear Island cross-section (Fig. 5a) the breccia is up to 40 m thick. The basal 10 m of the breccia contain large blocks up to 3-4 m in size. Large cavities between those blocks are filled with well-bedded fine- to medium-grained sandstone. Near the base of the sinkholes the breccia matrix is characteristically red in colour. It is composed mainly of mud, silt, and fine-grained sand. The muddy matrix breccia occurs as a lens up to 10 m thick and is in relatively sharp contact with overlying white to slightly yellow sandy matrix breccias.

Breccia deposits that occur in the sinkholes are overlain by sediments of the Kanuyak Formation. At Bear and Kanuyak islands (Fig. 5a,b) sediments of the A member and B member overlie breccias and fill in the sinkholes. At Ekalulia Island (Fig. 5c) the A member is the primary fill, whereas the B member deposits overlie both the filled sinkholes and the Parry Bay Formation in areas of less spectacular karst development. Within the sinkholes, sand-supported breccias, sandstones, siltstones, and minor mudstones of the A member are commonly interbedded with conglomerates and breccias along the margins of sinkholes (within 10-20 m). In this area sediments contain up to 60 per cent carbonate and chert fragments generally less than 3 cm in size. The sand-supported breccias contain fragments up to 1 m in size. Towards the centre and vertically upwards from the base of the sinkhole the percentage and size of fragments decreases. These sediments contain up to 10 per cent fragments less than 2 cm in size. Sediments within the B Member often cap the sinkhole fill. These sediments rarely contain fragments of Parry Bay lithology. Carbonate beds within the sinkholes have been

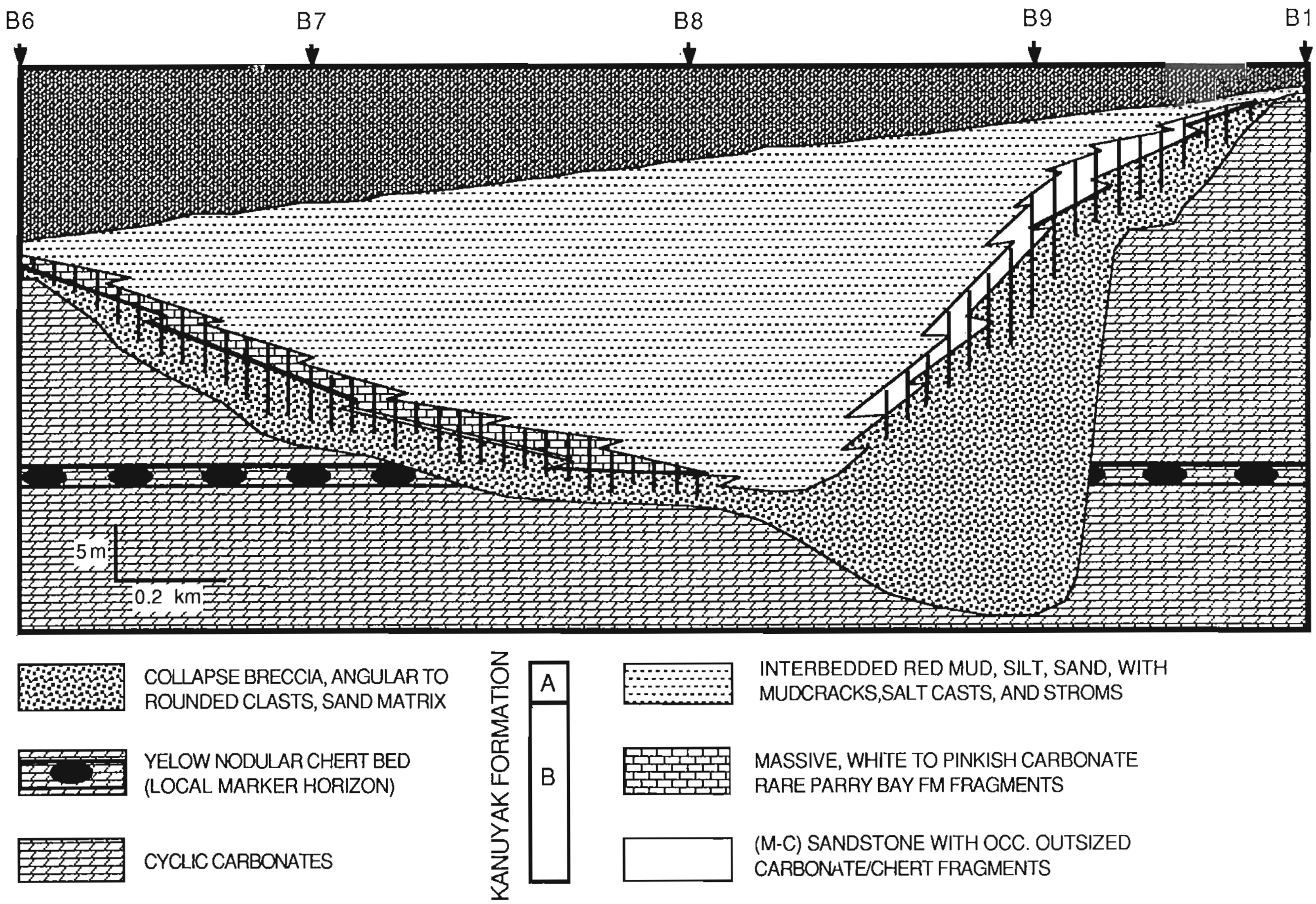

Figure 9. Detailed stratigraphy within sinkhole near the north end of Bear Island. Note local marker horizon. Vertical bars represent disrupted zone containing sheet cracks, pisolite, cements, etc. Disrupted zone is superimposed on both the karstic breccia breccia and overlying Kanuyak Formation.

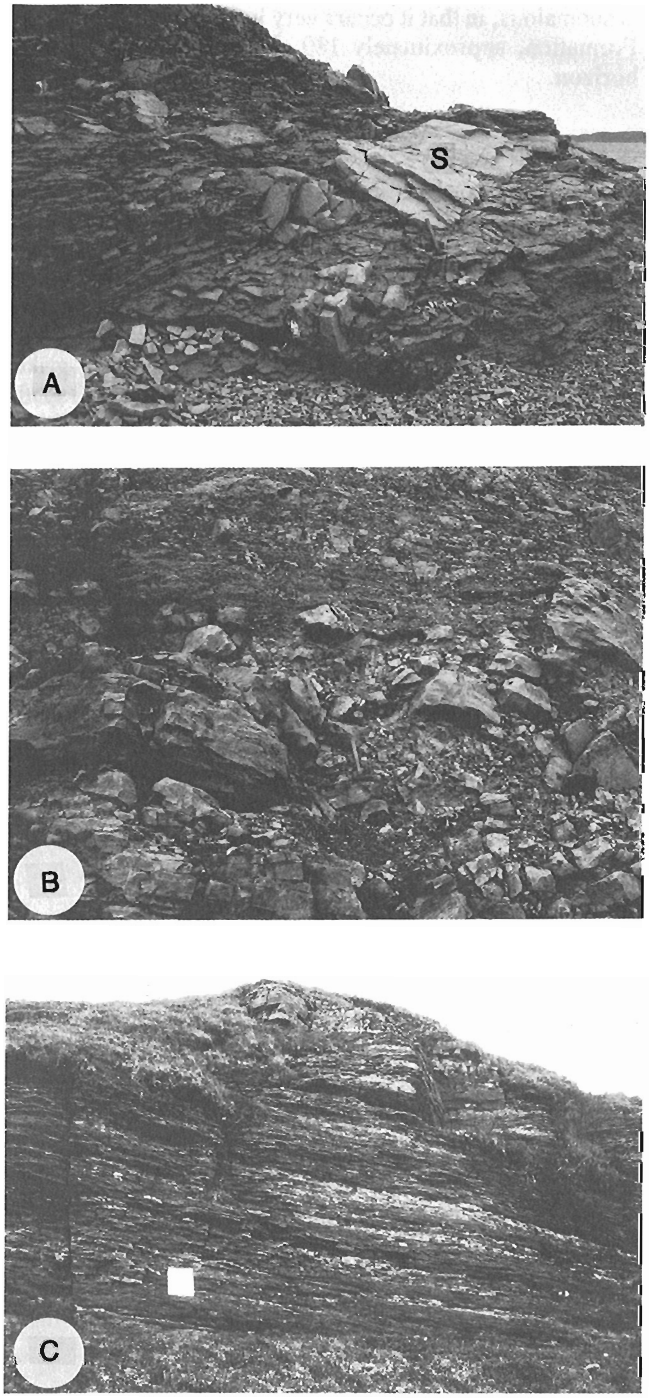

Figure 10. A) Interbedded siltstones and sandstones (Kanuyak Formation, A member) onlap collapse breccia. Note large angular carbonate blocks floating within bedded siliciclastics (S). Section 5, Ekalulia Island. Hammer for scale. B) Sandstones of the A member (S) interbedded with collapse breccia. Section 9, Bear Island. Hammer for scale. C) Interbedded red mudstones and siltstones, sandstones, and carbonates (light coloured beds) of the B member. Section 3, Bear Island. Clip board for scale.

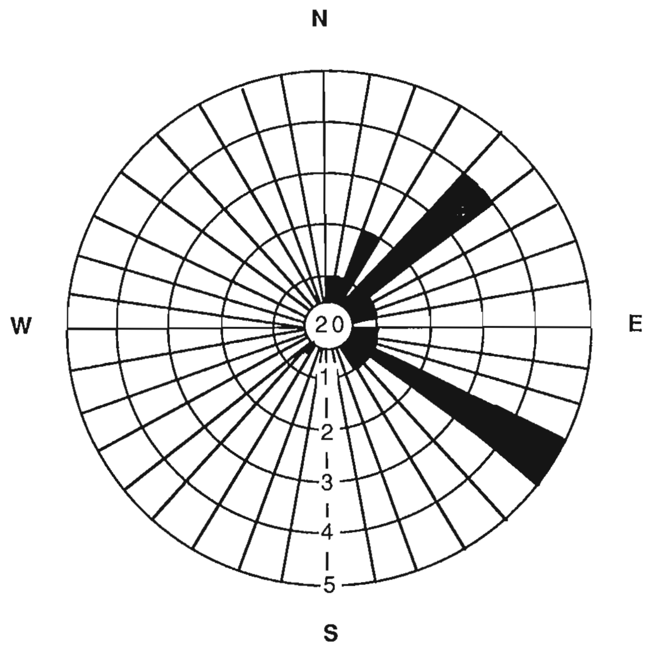

Figure 11. Rose diagram of sinkhole and grike trends. Note northeast and northwest trends. 20 features were measured.

observed within the two major sinkholes at Bear Island (Fig. 5a). These beds also contain minor Parry Bay fragments.

In areas of less spectacular karst development, for example, between sections 1 and 2 at Bear Island (Fig. 5a), cyclic carbonates are capped by less than 1 m of breccia, mud, and sand (Fig. 7, 12b,12c). The upper contact of the bedded carbonates is usually irregular, with fractures up to 2 m in depth and a few ten's of centimetres wide. Fragments within the basal portion of the breccia often exhibit a fitted fabric. Individual blocks are similar to the lithology of the underlying carbonates. The matrix between the blocks and within the fractures consists of mud, silt, fine sand, and smaller carbonate and chert fragments

Caves

Caves occur entirely within the Parry Bay Formation cyclic carbonates. Caves have been documented along Bear Island (Fig. 5a, 6) and at the south end of Kanuyak Island (Fig. 4b). The caves near section 10 at Bear Island (Fig. 5a) are up to 55 m below the upper unconformity. However, a cave at section 1 at the south end of Bear Island (Fig. 6) was discovered approximately 180 m below the unconformity near the Parry Bay — Ellice contact. Most of the caves are filled with mud, silt, very fine-to medium-grained sand and occasional outsized fragments less than 3 cm in size. Layering in these deposits is parallel to cave floors. The uppermost cave at section 10 (Fig. 5a) is approximately 5m wide and 1m thick. It is surrounded by undisrupted dolostones along the sides and base of the cave. However, the cave is overlain by disrupted bedded carbonates dipping up to 45°. This zone represents the preserved walls and roof of a collapsed cave. The lowermost cave at section 10 (Fig. 5a) is within undisrupted cyclic carbonates. This cave is 10 m wide and up to 2 m high and completely filled with sediment. Near the base the sediments are medium- to coarse-grained bedded sandstone with occasional carbonate and chert fragments (< 3 cm) fining upwards to mud and fine-grained sand within the upper 10-15 cm of the cave fill.

At section 1 near the south end of Bear Island (Fig. 5a, 6) a cave occurs near the Parry Bay — Ellice contact. The cave is filled with poorly sorted, unstratified breccia with angular to subangular clasts up to 25 cm in size. Individual clasts are composed chiefly of dolostone fragments and minor chert derived from the cyclic carbonates. The matrix contains finer fractions of the larger blocks. No detrital quartz was observed within the matrix. Thin (<5 cm wide) sediment-filled fractures crosscut the carbonates, breccias, and mudstones within the gradational zone between the Parry Bay and Ellice formations. These fractures are filled with medium-grained allochthonous quartz sand and minor red and green mudchips derived from the Ellice Formation. This breccia-filled cave is anomalous, in that it occurs very low within the Parry Bay Formation, approximately 180 m below the main breccia horizon.

Grikes

Grikes, or kluftkarren, are solution-widened joints that commonly are infilled with later sediment (Sweeting, 1972; Bogli, 1980). Grikes penetrate up to 180 m into Parry Bay Formation dolostones (Fig. 6). Commonly, they are less than 10 cm wide, and therefore are too thin to plot on the cross-sections in Figure 5 (Fig. 12). Typically they are filled with sand and minor carbonate and chert fragments. In plan view, grikes are relatively straight and elongate trending northeast to southeast (Fig. 11). Near sections 6 and 10 at Bear Island (Fig. 5a) the grikes are up to 4.5 m wide and penetrate the substrate up to 50 m below the upper unconformity. At section 6 the grike is filled with major amounts of carbonate and chert clasts up to 60 cm in size. The percentage of sand within the matrix increases toward the base of the overlying Kanuyak Formation. The second grike at section 10 is filled with planar-bedded silt and sand. Grikes are most common near the sur-

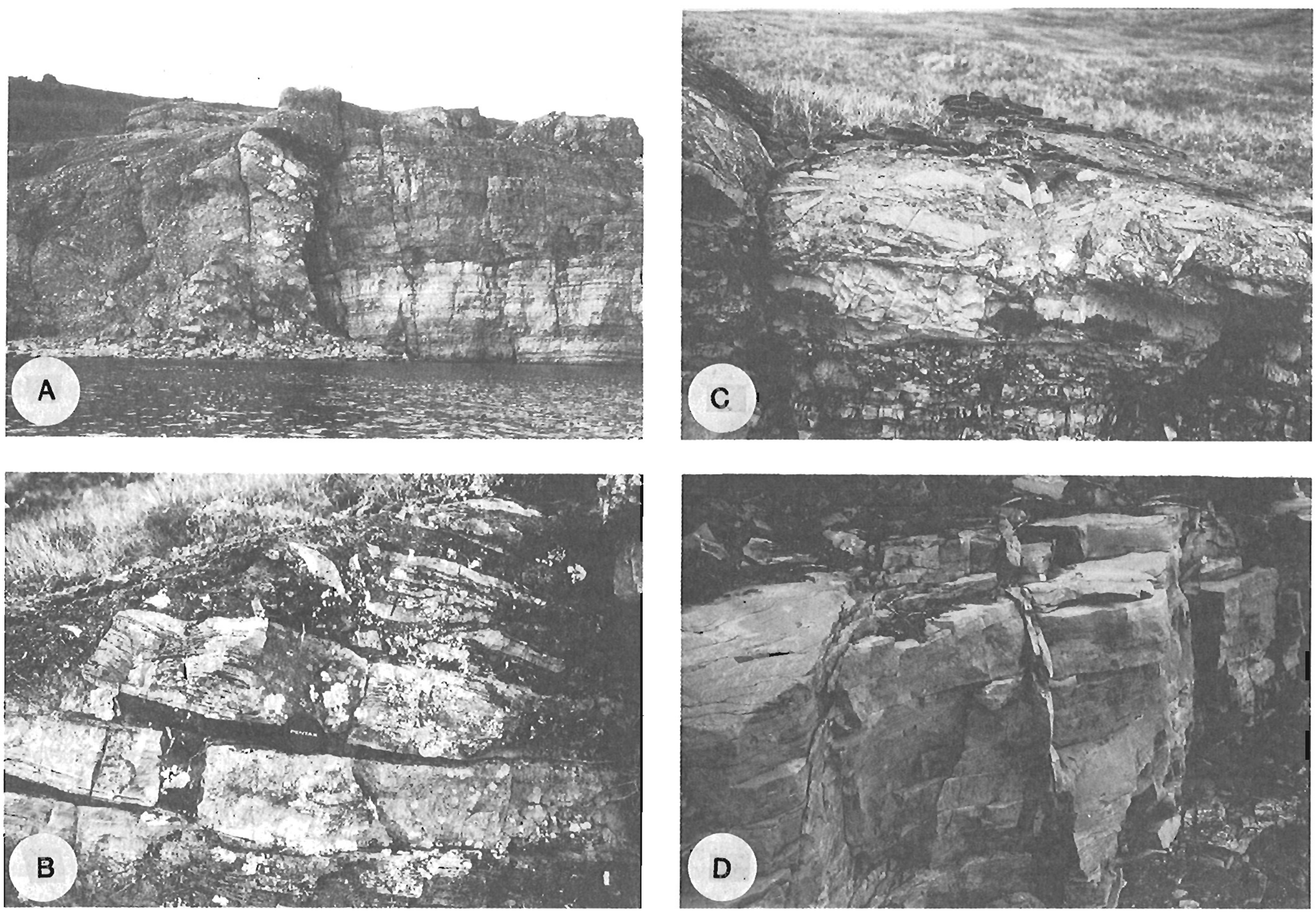

Figure 12. A) Spectacular matrix to clast supported karst breccia adjacent vertical sinkhole wall with up to 48 meters of relief. Section 8. Bear Island. Day pack at base of breccia. B) Breccia developed near section 2, Bear Island. Carbonate blocks near base exhibit fitted fabric. Dark colored matrix consist of red mud, silt, sand, and smaller chert and carbonate fragments. C) Thin (< 40cm) clast to matrix supported breccia developed ontop of Parry Bay nodular carbonate bed. Breccia overlain by 10 cm thinnly bedded red siltstone and sandstone (B member). Section 9, Bear Island. D) Two sand filled grikes approximately 64 meters below main paleokarst profile. Grikes trend southeast. Section 4, Bear Island.

face of the unconformity in areas of less spectacular karst development. In these areas the grikes penetrate up to 2 m into the substrate and are filled with allochthonous terriginous material. Rarely, the grikes are filled with laminated carbonate cements that contour grike walls.

Interpretations of karst features

Karst facies capping the Parry Bay Formation suggest that prolonged subaerial exposure and dissolution occured along this boundary. Carbonate dissolution occured along lines of weakness, probably a paleojoint system as suggested by elongate sinkholes (Fig. 11). Collapse of caves produced large sinkholes lined with angular breccia blocks. The textural maturity of breccia blocks suggests that the relative distance of transport during collapse increased upwards. Blocks further from the sinkhole walls are progressively more rounded and reworked . Individual blocks were rounded and reworked during subsequent agitation in probably shallow marine environments. Waters flowed between the large blocks throughout the collapse breccia depositing mud, silt, sand, and occasional coarser carbonate and chert fragments. Red muddy waters percolated through the basal parts of the collapse breccias depositing sediment between individual blocks.

A few caves that have not collapsed are preserved within the Parry Bay section. These may have formed either within the phreatic zone or vadose zone. Regardless of the zone in which the cave first developed, allochthonous sand and coarser debris was transported through the cave systems via underground currents. The cave systems were complexly interconnected, and in communication with the surface as indicated by the presence of siliciclastic sediment. In one case, a cave was completely filled with upward fining sediments (section 10, Fig. 5a). In another example, well bedded sediment approximately 1 meter thick was overlain by disrupted host carbonates that collapsed onto the cave floor deposits.

The presence of abundant salt casts within the Ellice Formation and the development of breccias at a great depth within the Parry Bay Formation may be the result of evaporite dissolution rather than karstification (Fig. 6). However, smaller karst features including grikes are present throughout the Parry Bay Formation at various stratigraphic levels and some of the grikes are up to 180 m below the unconformity. Thus, it may well have been possible that the relative sea level drop that exposed the Parry Bay platform was on the order of 200 m.

DISRUPTED ZONE

Discription of Disrupted Zone

The disrupted zone is an association of fabrics that overprints the upper Parry Bay — lower Kanuyak stratigraphy. Three main fabrics are developed within the disrupted zone: sheet cracks, pisolite, and void-filling cements. Commonly, the fabrics occur within the upper parts of the collapse breccia, and in the basal parts of the sand supported breccias, sandstones, and massive carbonates of the Kanuyak Formation. The degree of fabric development appears most intense near the margins of sinkholes. Figure 9 represents the relationship of the disrupted zone to the various lithofacies deposited within a sinkhole near the north end of Bear Island. The rocks in which the fabrics are developed will be referred to as the host material. The fabrics that comprise the disrupted zone are not host-selective (Fig. 9). However, different types of fabrics are better developed within the collapse breccias as compared to the carbonates or sandstones.

Sheet cracks

These features are cracks or fractures filled with isopachous carbonate cements. White cements are common and interlaminated with red, pink, or brown cements. Individual sheet cracks are randomly orientated from horizontal to vertical (Fig. 13a). They are formed in various host materials, although best developed at the base of the Kanuyak Formation. When these features are developed within sandstones they often contain sand material. Horizontally-oriented sheet cracks may exhibit teepee structures which are up to 50 cm high. Sheet cracks range in thickness from less than 1 cm to 30 cm thick and are up to a few metres in length. The thinner cracks are developed within the collapse breccias and do not crosscut clasts (Fig. 13b). In one example, the laminated cements within the cracks were folded and brecciated.

Pisolite

Pisolitic layers are closely associated with development of sheet cracks and cavities within the basal parts of the Kanuyak Formation. Individual pisoliths are spherical in shape and exhibit fine concentric laminations as well as occasional radial textures. They range from a few millimetres to 8 cm in size (Fig. 13c,d). Pisolites associated with the sheet cracks and cavities commonly are composed of very fine to medium dolomite and may exhibit inverse grading. Breccia clasts containing pisolites have been observed within host sand-supported breccias and sandstones.

Void-filling cements

Void-filling cements are developed between blocks within collapse breccias and cavities formed in the basal parts of the Kanuyak Formation. Cements occur as isopachous rims along walls of cavities and as coarsely crystalline percipitates completely filling cavities. Individual isopachous layers are 1 mm to 1 cm thick and are composed of fine to coarsely crystalline dolomite.

Fabric zonation

One example of vertical succesion of fabrics within the disrupted zone was documented. Near section 3 at Bear Island (Fig. 4a), the disrupted zone is developed within the lowermost carbonate bed. The bed is 3.5 m thick and can be divided into three main units. The basal unit is approximately 1.5 m thick and is undisrupted. The second unit (less than 1 m) contains a network of sheet cracks oriented in all directions with associated pisolite (Fig. 13a). Cavities up to 40 cm in size are developed with pisoliths at the base which exhibit inverse grading. Isopachous cements and other unidentified cement fabrics coat the peripheries of the cavity. This is overlain by a unit, up to 1 m thick, characterized by a 'pitted' fabric which may represent incipient pisoliths.

Discussion of Disrupted Zone

The fabrics of the disrupted zone are superimposed on both the Parry Bay and lower Kanyuyak formations. Thus it is likely that they represent sediment modification that post-dates karst development as well as the earlier phases of subsequent karst infill by terrigenous sediments. The fabrics are similar to those that are developed during brief periods of minor subaerial exposure on other carbonate platforms of both Phanerozoic (eg. Dunham, 1972) and Proterozoic (eg. Grotzinger, 1986) age. Thus exposure and fabric development does not necessarily have to relate to the main phase of karst development.

BITUMEN IN THE PARRY BAY FORMATION

Occurrence

Dark grey to black bitumen was discovered in biostromal dolomites of the upper Parry Bay Formation near Walker Bay, Kent Penninsula (Fig. 3). Outcrop is poor but biostromal stromatolite units are locally well exposed as low-lying benches.

The bitumen occurs in coarsely crystalline dolomite with well-developed, interconnected vuggy porosity. The porosity is almost exclusively restricted to the axial zones of conical-shaped stromatolites that form bioherms which are up to 1 m thick and laterally continuous for up to 50 m. Component stromatolite columns exhibit uncommon linkage and well-developed axial zones, but are elongate in plan and therefore cannot be considered as Conophyton (sensu strictu). Columns are up to 1 m tall and 30 cm wide.

Vuggy pores within columns are irregularly laminoid in shape (outlining relict lamination) and in the longest dimension are up to 1.5 cm. Voids are characteristically incompletely filled and lined with thin coatings of black bitumen. In rare instances porosity is entirely occluded by the bitumen.

Analysis

Proterozoic rocks commonly contain bitumen and pyrobitumen along with mineral deposits (Bogdanova et al; 1977; Goodarzi and Ghandi, in prep; Goodarzi et al; in prep). Natural bitumen is grouped into two main catagories based on their solubility in common solvents, optical properties, molecular structure and their ability to form graphitizing or

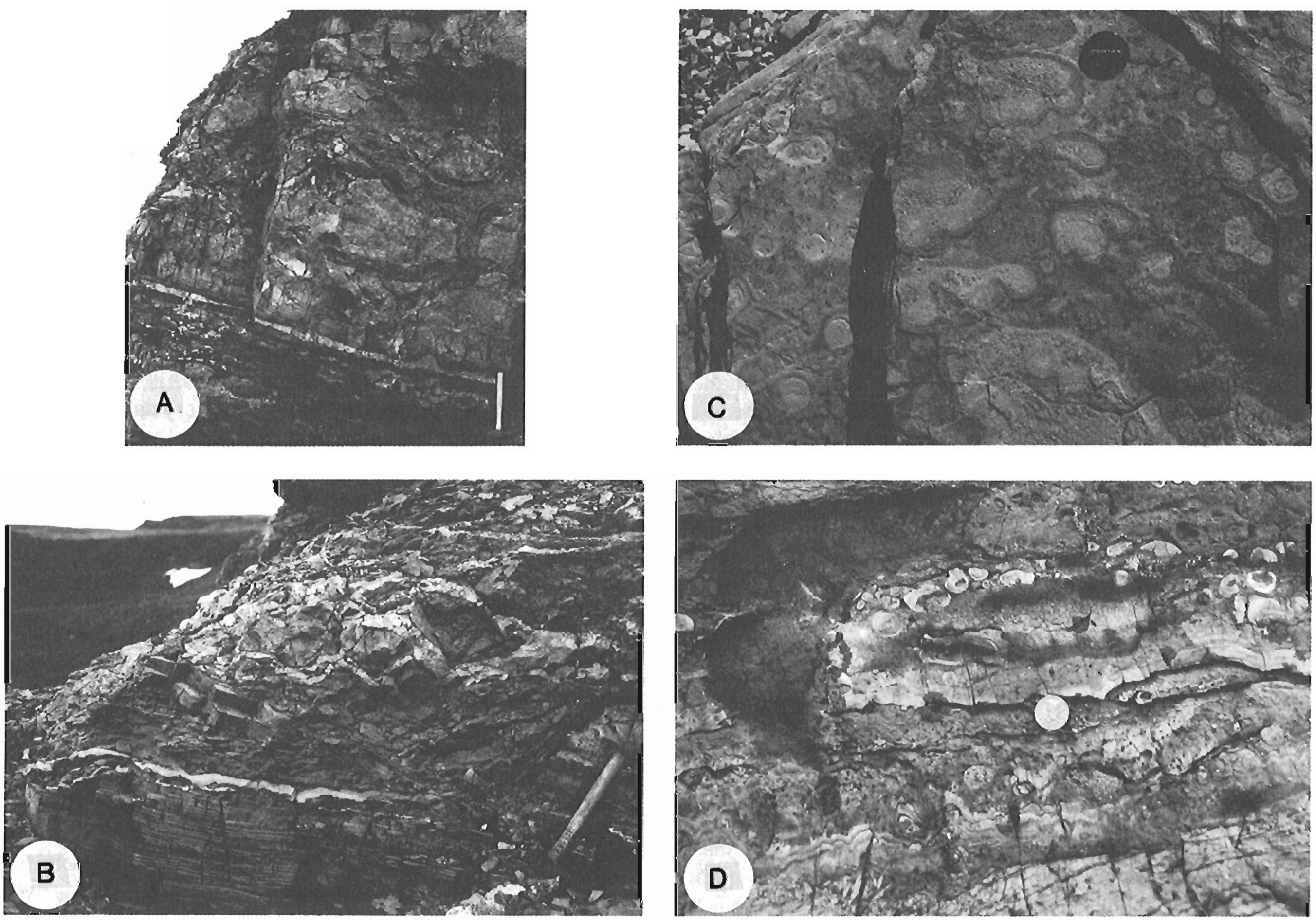

Figure 13. A) Vertically and horizontally oriented sheet cracks within carbonate host rock (A member). Note sediments of B member near top of photograph. Section 3, Bear Island. B) Sheet cracks composed of finely laminated white carbonate cement developed within basal part of collapse breccia. Section 5, Ekalulia Island. C) Pink and red laminated pisolites composed of sand and defined by liesegang banding. Near section 5, Bear Island. D) Pisolites and finely laminated cements developed within the basal Kanuyak Formation. South end of Ekalulia Island.

non-graphitizing carbon (Jacob, 1975; Khavari-Khorassani, 1975; Williams and Goodarzi, 1981; Goodarzi and Williams, 1985). The soluble bitumen forms the asphaltic group of natural bitumen and they are graphitizing, that is their ordering changes from two dimensional turbostatic ordering to three dimensional graphitic ordering at elevated temperatue and pressure (Khavari-Khorasani, 1975; Goodarzi, 1985).

A small portion of Parry Bay Formation sample containing bitumen was pulverized and analyzed in a Rock-Eval — TOC instrument. No pyrolyzable hydrocarbons were recovered from the sample. The total organic carbon was determined to be 0.22 % and 0.23 % on duplicate runs. This must be considered to be a minimum value due to the high level of thermal alteration of the sample. The Rock-Eval oxidation oven operates at 600 degrees C in air and thus high rank carbon may not be completely oxidized.

Organic petrographic analysis of a small block of polished sample was carried out using reflected white light on a Zeiss MPM II microscope. Flow structures, mosaic structure and mesophase were all observed. The sample shows an optical texture which is coarse-grained with very strong anisotropy (cf Goodarzi and Murchison, 1978; Grint and Marsh, 1981) and high bireflectance (%Ro max — %Ro min, Table 1; Fig. 14). These properties indicate that a) the sample was

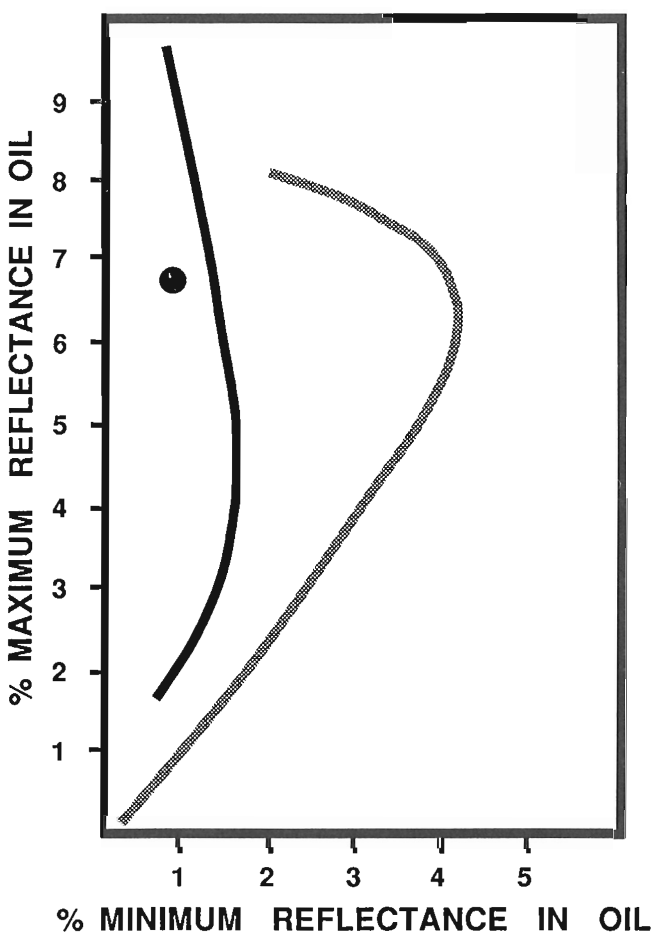

Figure 14. Maturation and bireflectance trends presented by Goodarzi and Norford (1985) for bitumen subjected to normal, burial-related maturation (grey line), and for bitumen altered by significant thermal events (black line). The sample from this study (dot) clearly falls in the region of heat-affected bitumen.

Table 1

Sample	%Ro max	%Ro min	bireflectance
graphite (Ragot, 1977)	15.75	0.60	15.15
graphite (Gooderzi et al, in prep)	17.53	0.65	16.88
bitumen (present study)	6.75	0.67	6.05

initially graphitizing bitumen (probably gilsonite which has a high fluidity during carbonization), and b) it has been subjected to a sudden thermal shock under pressure. The high bireflectance of the sample indicates a high level of structural ordering in aromatic carbon (Goodarzi and Murchison, 1975), which can be induced in material during laboratory heat treatment under elevated pressure and high heating rates (Marsh et al; 1971; Goodarzi, 1985; Goodarzi and Marsh, 1980; Goodarzi and Norford, 1985; Khavari-Khorasani and Murchison, 1978.

Discussion

The properties of the bitumen indicate that the vuggy pores originally contained gilsonite or some other form of asphaltic bitumen. This bitumen was highly reactive when it was "heat affected", that is, subjected to a rapid rise in temperature. Heating rates associated with burial of sediments even in high heat flow regimes are insufficient to cause the formation of what is essentially coke. Sufficiently high heating rates are usually observed to have resulted from the flux of hydrothermal fluids, the contact metamorphism of an igneous event, or high temperature laboratory experiments. In general, it is also reasonable to assume that the temperature reached a minimum of about 120 degrees C during the heat alteration episode. Metal sulphide mineralizations are also commonly observed in the same samples because the presence of the organic matter allows the reduction of sulphates associated with the hydrothermal fluids (Powell and Macqueen, 1984).

The conclusion that the sample was abruptly heated to high temperatures, unobtainable during "normal" burial, suggests that outpouring of the Ekalulia flood basalts may have been responsible for the set-up of a hydrothemal convection cell that operated at depth in the Parry Bay Formation. If so, the trace element and isotopic signature of earlier diagenetic environments (eg. karst-related meteoric) may be reset by this later, probably regional event. This potential complication deserves consideration in future diagenesis studies.

ACKNOWLEDGMENTS

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Stratigraphy of a 1.9 Ga foreland basin shelf-to-slope transition: Bear Creek Group, Tinney Hills area of Kilohigok Basin, District of Mackenzie

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Abstract

The stratigraphy of units in the Bear Creek Group (Goulburn Supergroup) beneath the Burnside Formation is discussed in this report. Work during the field season of 1987 concentrated on relationships in the Tinney Hills area located along the eastern coastline of Bathurst Inlet. The Tinney Hills spans a major stratigraphic transition between northwesterly shelf facies and southeasterly slope and basinal facies. The actual shelf-to-slope transition is superbly exposed in the southern Tinney Hills and the pinch-out of shelf sand bodies can be directly observed on a bed-for-bed basis.

Résumé

Ce rapport traite de la stratigraphie des unités dans le groupe de Bear-Creek (supergroupe de Goulburn) au-dessous de la formation de Burnside. Les travaux sur le terrain pendant la saison 1987 se sont concentrés sur les relations dans la région de Tinney Hills située le long de la côte est de l'inlet Bathurst. Les collines Tinney couvrent une transition stratigraphique importante entre un faciès de plateau nord-ouest et une pente sud-est et un faciès de bassin. La transition réelle plateau à pente affleure superbement dans le sud des collines Tinney et on peu observer directement le rétrécissement et la disparition des masses de sable de la plate-forme couche par couche.

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INTRODUCTION

Field work during the 1987 field season concentrated primarily on documentation of the stratigraphy and detailed facies relationships that characterize the lower Bear Creek Group in the Tinney Hills area of eastern Bathurst Inlet, Northwest Territories (Fig. 1,2). This report discusses only the Link, Rifle and Hackett formations of the Bear Creek Group (Fig. 3), as the intervening Beechey Formation has been described previously (Grotzinger et al., 1987), and the overlying Burnside Formation is discussed as a complementary paper within this volume (McCormick and Grotzinger, 1988).

The sub-Burnside Bear Creek Group is an important stratigraphic package in that it contains the record of basin subsidence, beginning with the transition from passive-margin to foredeep sedimentation associated with platform drowning. Basinal sedimentation, followed by shallow-marine sedimentation, terminated with the onset of molasse-wedge progradation. Within this framework, the Tinney Hills area is paleogeographically significant in that unconformity-bounded shelf successions pass basinward into correlative deeper-water successions. Data summarized here demonstrate that the shelf-to-slope transition of the Bear Creek Foredeep was located in the southern part of the Tinney Hills. The superb exposures that highlight this region permit very detailed stratigraphic and sedimentologic work to be carried out, and in several of the formations the actual pinch-out of shelf sand units can be documented on a bed-by-bed basis.

In order to document the critical facies relationships in as much detail as time constraints allowed, all units were first mapped at 1:25 000 scale. Contacts were traced laterally in order to identify the location of all major facies changes. Subsequently, detailed stratigraphic sections (see Fig. 4, for

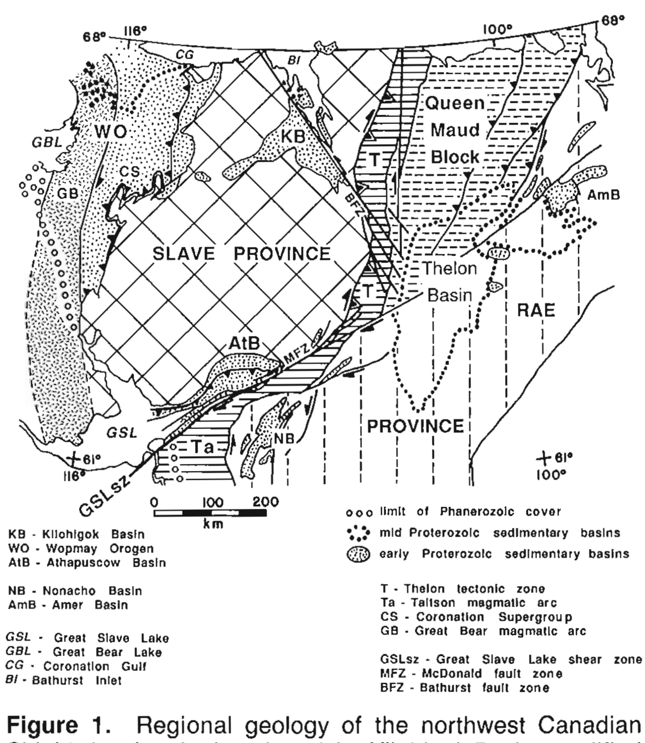

Figure 1. Regional geology of the northwest Canadian Shield showing the location of the Kilohigok Basin; modified after Hoffman (1987).

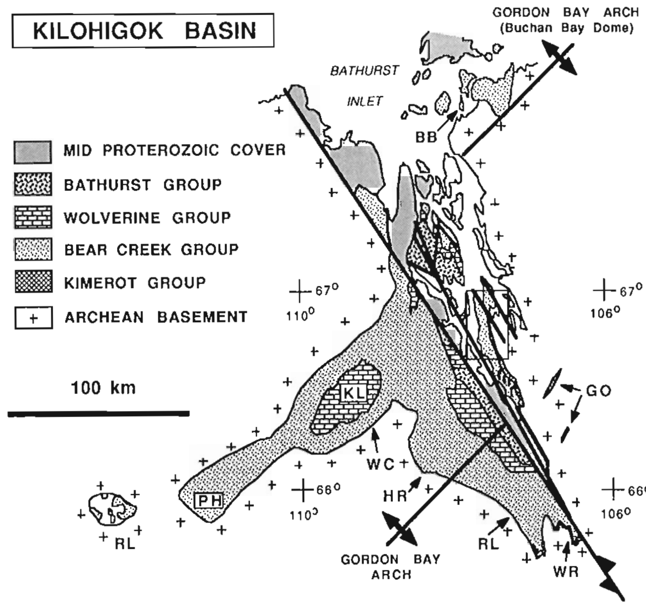

Figure 2. Simplified geology of Kilohigok Basin. RL, Rockinghorse Lake; PH, Peacock Hills; KL, Kuuvik Lakes; WC, Wolverine Canyon; HR, Hackett River; RL, Rifle Lake; WR, Western River; BB, Buchan Bay area; GO, outliers of Goulburn Supergroup. Box outlines Tinney Hills area of Figure 5. After Campbell and Cecile (1976) and Tirrul (1985).

GOULBURN SUPERGROUP	BATHURST GROUP	AMAGOK FM.
		BROWN SOUND FM.
	WOLVERINE GROUP	KUUVIK FM.
		PEACOCK HILLS FM.
	BEAR CREEK GROUP	QUADYUK FM.
		MARA FM.
		BURNSIDE FM.
		LINK FM.
		BEECHEY FM.
		RIFLE FM.
		HACKETT FM.
	KIMEROT GROUP	PEG FM.
		KENYON FM.

Figure 3. Stratigraphic nomenclature for Goulburn Supergroup (Grotzinger et al., 1987).

example) were measured at strategic positions (Fig. 5) in order to illustrate the major features of the shelf-to-basin transition. The results of this stratigraphic work are presented below as summary cross-sections utilizing map relationships and individual stratigraphic sections compiled at at scale of approximately 2.5 cm inch to 100 m (Figs. 6,7,8). This report will be followed up in the future with a more comprehensive presentation including 1:25,000 scale maps and stratigraphic sections complied at 2.5 cm inch to 50 m and 2.5 cm to 20 m.

Finally, this report is generally restricted to description of relationships observed in the field. Comprehensive interpretations of the sequences discussed below in terms of their environmental, stratigraphic, and paleotectonic significance are forthcoming.

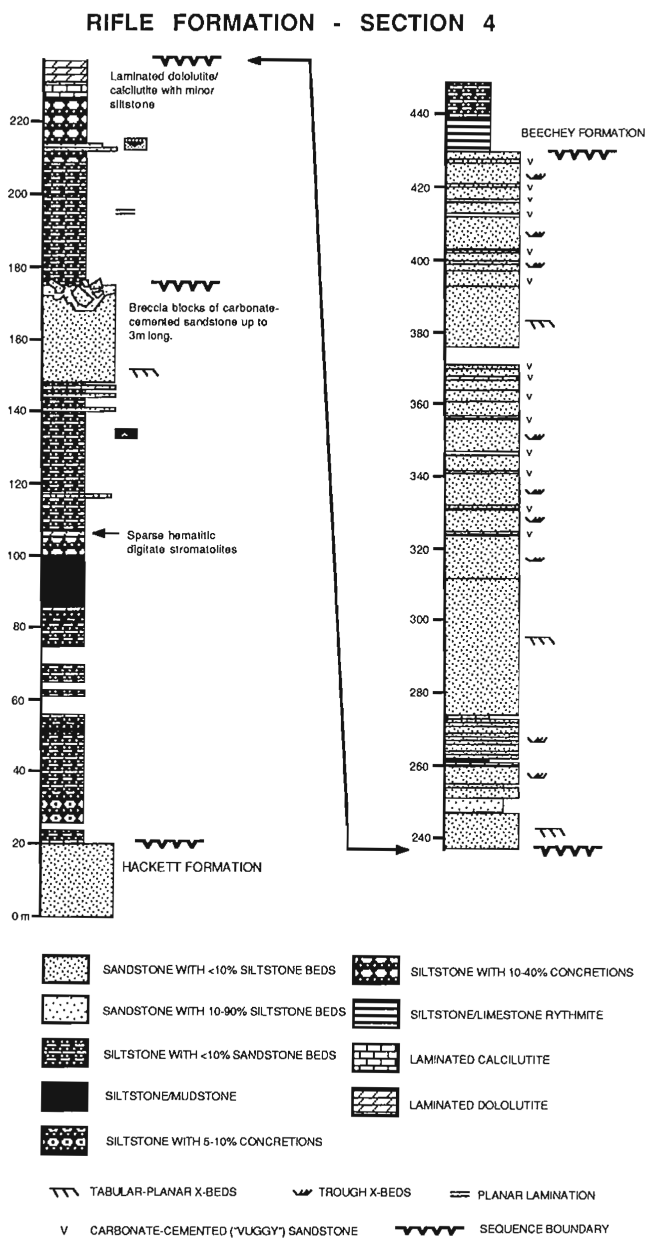

Figure 4. Detailed measured section of Rifle Formation. See Figure 5 for section location.

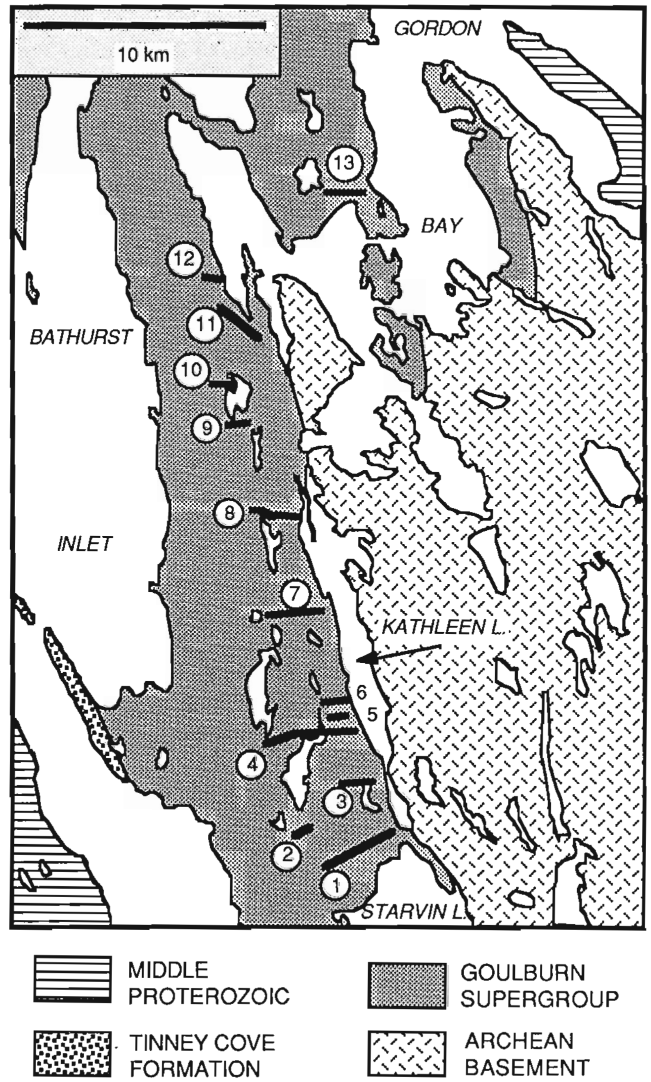

Figure 5. Location of measured stratigraphic sections referred to in this report for Tinney Hills area. See Figure 2 for location of Tinney Hills.

STRATIGRAPHY OF THE SUB-BURNSIDE BEAR CREEK GROUP, TINNEY HILLS

The Hackett, Rifle, Beechey and Link formations (Fig. 3; Grotzinger et al., 1987) can all be recognized in the Tinney Hills area. The Hackett Formation is simple, upward-shallowing offlap unit ranging in thickness from 160 m at the north end of Kathleen Lake (Section 8, Fig. 6) to >300 m at the south end of Kathleen Lake (Section 1, Fig. 6). It consists of a lower siltstone member and an upper sandstone member. The lower member rests abruptly on carbonates of the Kimerot Group and the upper member is unconformably overlain by siltstones of the Rifle Formation.

The Rifle Formation (Fig. 3,6) is a more complex upward-shallowing offlap sequence with two minor internal disconformities. The sequence thickens from 280 m to 420 m between the north and south ends of Kathleen Lake (Fig. 6); it contains a lower siltstone member and an upper sandstone member. The lower member rests unconformably on sandstones of the Hackett Formation and the upper member is unconformably overlain by siltstones of the Beechey Formation.

The Link Formation (Fig. 3,7,8) is a complex, highly variable sequence of sandstone wedges and interfingering siltstones that records the transition from dominantly shallow marine units of the sub-Burnside Bear Creek Group to dominantly fluvial units of the Burnside Formation. It contains four members which are designated, in ascending order, the L1, L2, L3 and Tinney Hills members; all are conformable with the exception of the Tinney Hills member which exhibits locally spectacular erosional down-cutting along its lower and upper contacts. Siltstones of the lower Link Formation rest unconformably on the Beechey Formation and siltstones of the upper Link Formation grade conformably into the lower Burnside Formation.

Hackett Formation

The stratigraphy of the Hackett Formation in the Tinney Hills area is shown in Figure 6. The formation is a southeast-thickening offlap sequence in which shallow-water, cross-bedded sandstones and silty sandstones overlie relatively deeper-water laminated siltstones. The base of the sequence contains several volcanic ash layers up to 60 cm thick; euhedral zircons in these beds have yielded a preliminary U-Pb age of 1.95 Ga (S.A. Bowring, pers. comm., 1987).

Basal siltstones are laminated with minor lenticles, starved ripples, and "streaks" of very fine to fine sand. Upwards, the proportion of sand increases and the dominant sedimentary structures include wavy, thin bedding, planar lamination, and hummocky cross-stratification. Sand beds are interbedded with layers of siltstone on the scale of a few tens of centimetres. The upper part of the sequence consists almost exclusively of fine to medium, locally coarse sandstone with uncommon interbeds of sandy siltstone. Sandstones are medium- to thick-bedded and commonly contain abundant trough crossbeds.

Sandstones of the upper Hackett Formation pass laterally southeastward into slope and basinal siltstones. This transition is inferred on the basis of stratigraphic work south of Starvin Lake which covers the transition zone. To the northwest, the siltstone and sandstone members of the Hackett Formation extend across the the Bathurst Fault and are continuous at least as far as the Hackett River (Fig. 2) which is the type area of the formation.

The upper contact of the Hackett Formation is unconformable. Subaerial exposure fabrics are best developed over the Gordon Bay Arch near the Hackett River area (Grotzinger et al., 1987), but the amount of local erosional relief appears to be greatest near the south end of Kathleen Lake (Fig. 6). Figure 6 reveals that the lower siltstone member of the overlying Rifle Formation is extremely thin at Section 3 suggesting significant paleorelief (50 m) along the top of the Hackett Formation. Although the actual contact is not continuously exposed along strike, it can be located within 10 m in several cases and mapping supports this interpretation.

Rifle Formation

Stratigraphic relationships within the Rifle Formation for the Tinney Hills area are illustrated in Figures 4 and 6. The Rifle Formation is bounded by two regionally extensive disconformities (major sequence boundaries) at the top of the Hackett Formation and the base of the Beechey Formation. These surfaces are continuous at least as far to the northwest as the Hackett River area (Fig. 2). Internally, the Rifle Formation contains two local disconformities (minor sequence boundaries) that are developed southeast of section 6 (Fig. 6) and converge to form a single boundary northwest of section 6. The relationship between the single minor sequence boundary northwest of section 6 and the major boundary at the top of the Rifle Formation is not yet clear. However, previous work suggests that in the vicinity of the Western River area, the two boundaries converge such that no sandstones are preserved at the top of the formation northwest of Rifle Lake (see Grotzinger et al., 1987); northwest of Rifle Lake siltstones of the lower Beechey Formation rest directly on siltstones of the lower Rifle Formation with no intervening sandstones. On this basis, combined with the evidence for locally significant erosional relief on top of the Rifle siltstones, the sequence boundary at the top of the Rifle Formation was interpreted as one of the most significant in the basin (Grotzinger et al., 1987).

Figures 5 and 6 show that southeast of section 6 the lower of the two minor sequence boundaries cuts down through sandstones of the lower sequence. Although breccias are locally developed along this boundary suggesting subaerial exposure of the sandstones, detailed work reveals that the apparent "cut-down" is not just erosional. In large part, this surface also delineates the original depositional relief on a major sand wedge prograding southeastward toward the shelf edge; as sea level fell the upper surface of this depositional clinoform was diagenetically modified. Local slumping of the upper surface also took place as shown by a layer of breccia debris that can be trace along the sequence boundary all the way to section 1 (Fig. 6). The "pinch-out" of a major sandstone body in this location in part defines the position of the depositional shelf-break of the Bear Creek Foredeep.

Southeast of section 6 a minor depositional sequence occurs between the two minor sequence boundaries. Although it is only of limited local extent (Fig. 6), the pinchout of the thin sandstone wedge helps further define the foredeep shelf edge.

Sandstones of the uppermost sequence in the Rifle Formation did not prograde appreciably further than in underlying sequences (Fig. 6). This sequence is the thickest of the three developed in the Rifle Formation, and only thin sandstone tongues are developed southeast of section 3. The actual "pinch-out" of the sand body can be mapped out in much detail and the relationships are accurately portrayed in Figure 6.

Regardless of which sequence the sandstone bodies occur in, they are consistently trough or planar-tabular cross-bedded, medium- to thick-bedded, and generally medium- to coarse-grained. Locally sands are very coarse-grained with outsized granules and pebbles, and contain up to 25 % feldspar. Generally, trough crossbedding is more common in the more northwesterly sections and tabular-planar crossbedding is more common to the southeast. Tabular-planar crossbedding is particularly abundant near section 3 (Fig. 6) where reactivation surfaces are well-developed on foresets. A major channel sequence (up to 20 m) consisting of thick-bedded lensoidal sandstones is developed in the sandstone unit of the lower sequence at section 5 and pinches out within 1 km in both directions.

Generally sandstones are quartz-cemented. However, northwest of section 4 (Fig. 6) quartz-cemented sandstones are increasingly interstratified with carbonate (dolomite)-cemented sandstones (Fig. 4). Contacts between sand sheets with varying cement types are usually abrupt and a vertical zonation defined by the percentage of carbonate to silica is often present on the scale of a few metres. Within carbonate-cemented units the amount of carbonate generally increases upwards and in an extreme case is associated with pisolite and cement-filled, solution enlarged (?) irregular voids. It is possible that this may be a burial phenomena, but the stratigraphic distribution of the cements and the local vertical zonation of carbonate fabrics in some layers suggests that it may be early diagenetic. One possible explanation, among others, is that carbonate cementation occurred during intermittent subaerial exposure of the sand shelf. If related to sea level, then it would predict an increasing number of exposure events of increasingly greater duration in the direction of decreasing subsidence. The increase in carbonate toward the northwest (Gordon Bay Arch) is consistent with this. A similar relationship has been documented in sandstones of the Beechey Formation (S.M. Pelechaty, unpublished data) and in the L3 member of the Link Formation (see below).

Stratigraphic relationships suggest thinning of the Rifle Formation onto the Gordon Bay Arch (Grotzinger and Gall, 1986; Grotzinger et al., 1987). The high feldspar composition and coarse grain size of the sands suggest that a granitic source may have been exposed at this time. In this respect it is noteworthy that in the Buchan Bay area of Bathurst Inlet (Fig. 2), the basement rocks are granitic and that this area coincides with the inferred position of the arch at that time. If this area was exposed during Rifle time, then it implies that the arch was more active in this location than along strike to the southwest in the Hackett River area where the underlying Hackett Formation is preserved. Interestingly, the Burnside Formation rests directly on granitic basement in the Buchan Bay area (Campbell and Cecile, 1981; McCormick and Grotzinger, 1988), and the relationships described above suggest that this region was probably subaerially exposed during deposition of most of the sub-Burnside Bear Creek Group. This discounts the alternative interpretation which would allow deposition of condensed sequences over the arch in the Buchan Bay area prior to Burnside time, but would then require uplift and erosion of those units prior to Burnside deposition. Consequently, the data suggest that the Buchan Bay area may have been a dome which, on the average, had a greater amplitude than the arch along strike to the southwest. This is not surprising given the dynamic setting of foreland basins, the along strike variability in the configuration of collisional margins, and the spatial migration of sediment and thrust loads with time (Quinlan and Beaumont, 1984).

Link Formation

The Link Formation is divided into four members (Fig. 7, 8). In ascending order these are the L1, L2, L3, and Tinney Hills members. Figure 8 specifically illustrates relationships within the Tinney Hills member. Each of the members is a discrete sand body that can be mapped at 1:25 000 scale (Fig. 7, 8). They are interstratified with siltstone/mudstone units that contain minor sandstone sheets up to a few metres thick, interbedded with siltstones, that form sequences up to 50 m thick. These units are highly lensoidal; larger scale sandy units as well as individual sand sheets are discontinuous over a few kilometres and could not be continuously mapped out.

The Link Formation in the Tinney Hills area is highly complex and includes several important stratigraphic transitions of regional tectonic and paleogeographic significance. This is shown by the advance of major clastic wedges from the southeast (L1, L3, and Tinney Hills members) over what previously was the shelf-break of the foredeep. This restructuring of Bear Creek foredeep stratigraphic architecture from northwesterly, craton-sourced clastic wedges to southeasterly, hinterland-sourced clastic wedges culminating in the Burnside wedge, represents infilling of the basin axis during terminal uplift and unroofing of the hinterland (Thelon Tectonic Zone and/or Queen Maud Block, Fig. 2). The L2 member is the final clastic wedge derived from the craton and is similar in facies to the sand bodies of the underlying Rifle Formation.

L1 Member.

The L1 member is the lateral continuation of a thick sequence (up to 300 m) of axially-deposited turbidites developed in the allochthons of the Bear Creek Hills Thrust/Fold Belt. Sedimentologically the unit changes character along strike to the northwest, and in the southern part of the Tinney Hills (Section 1) shares more features in common with a shelf sequence than a turbidite sequence. It is strongly diachronous to the northwest and pinches out abruptly between sections 1 and 8 (Fig. 7).

The L1 member is conformably enclosed in laminated dark green-grey mudstones with common sandy partings. Units within the L1 member are sequences of thin-to medium-bedded sandstone up to 50 m thick alternating with sequences of silty sandstone up to 40 m thick. Sands are medium- to coarse-grained with conspicous pink and white stripes that enhance trough and tabular-planar crossbed foresets. These units generally contain 5 to 10 % feldspar and locally abundant glauconite. Units with abundant planar lamination and possible hummocky cross-stratification are also developed.

L2 Member.

The L2 member is a northwest-thickening wedge of medium to coarse sandstone. It is a conformably bounded sequence that is 60 m thick at section 13, pinching out systematically to the southeast (Fig. 7). Sedimentologically, it is very similar to the sandstone units of the underlying Rifle Formation, containing abundant trough crossbedding and up to 15 % feldspar.

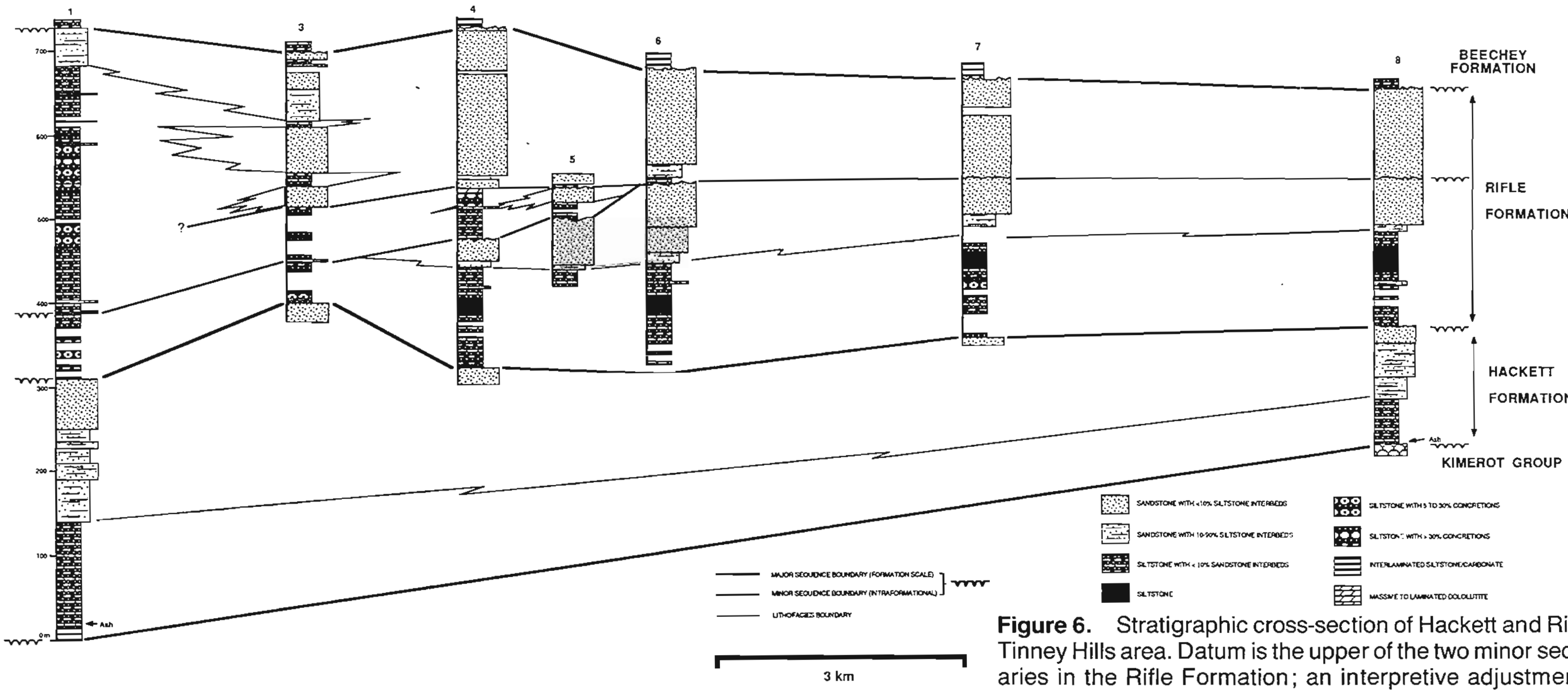

Figure 6. Stratigraphic cross-section of Hackett and Rifle formations, Tinney Hills area. Datum is the upper of the two minor sequence boundaries in the Rifle Formation; an interpretive adjustment is made for sections 1 and 3. See Figure 5 for location of measured sections.

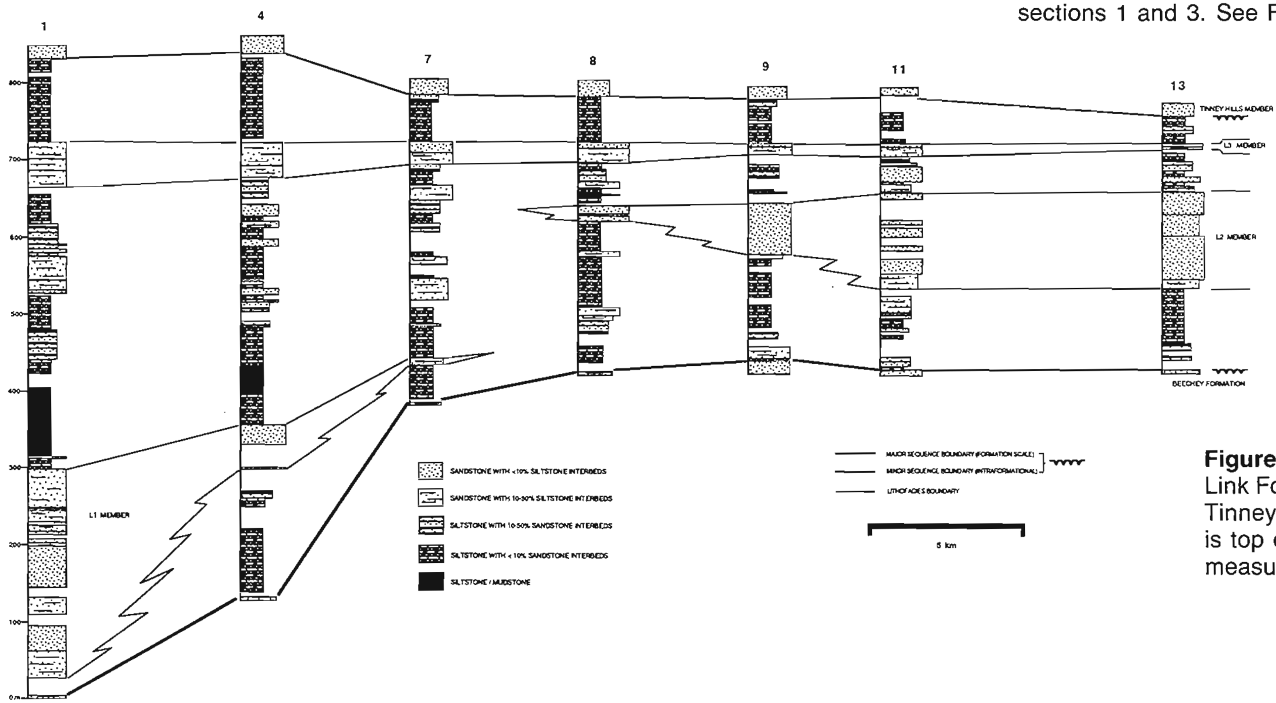

Figure 7. Stratigraphic cross-section of the lower Link Formation including L1, L2, and L3 members. Tinney Hills member is shown in Figure 8. Datum is top of L3 member. See Figure 5 for location of measured sections.

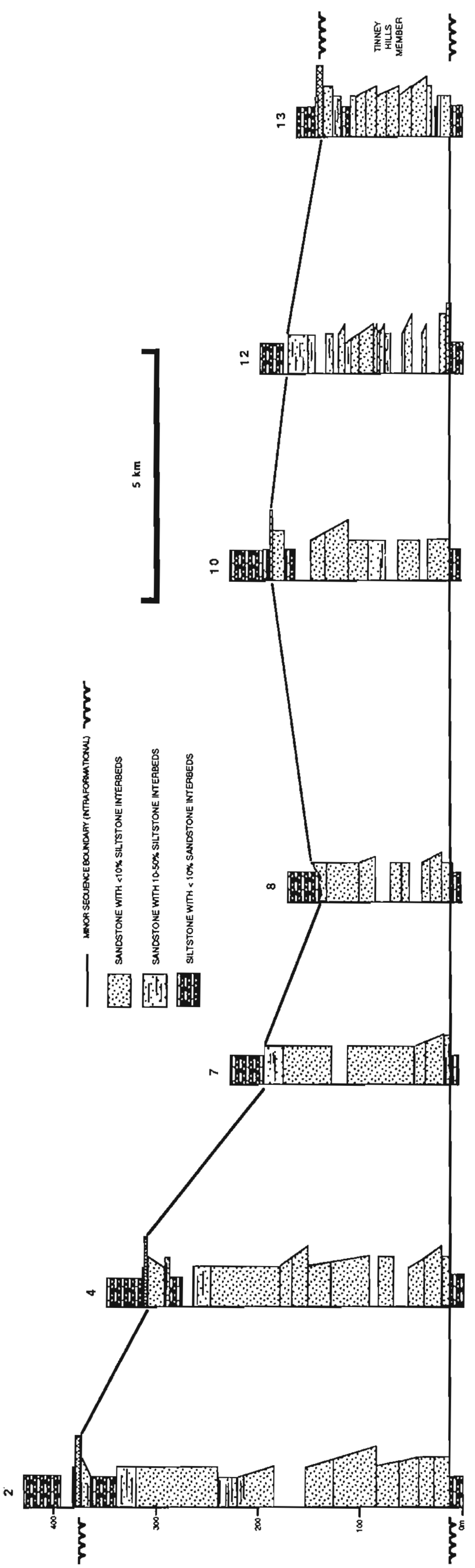

Figure 8. Stratigraphic cross-section of Tinney Hills member of Link Formation. Datum is base of member. Note local erosional cut-out at top of member. See Figure 5 for location of measured sections.

Sandstones are predominantly quartz-cemented, but as in the sandstones of the underlying Beechey and Rifle formations, the amount of carbonate cement increases systematically to the northwest. Occasionally, the carbonate is concentrated in pisolith-like accretionary structures and as laminar-fibrous cements that fill voids. As above, one possible interpretation is that the carbonate cements were precipitated in voids during intermittent subaerial exposure of the shelf in northerly areas proximal to the Buchan Bay area of Gordon Bay Arch.

L3 Member.

The L3 member is northwest-thinning wedge (Fig. 7) of medium to coarse-grained, trough cross-bedded sandstone with 5-10 % feldspar. It contains two distinctive submembers that can be correlated over the length of the Tinney Hills.

The lower submember in the region of the southern Tinney Hills is a thin sequence (<40 m thick) of grey/green/white sandstones with abundant thin interbeds of green/grey siltstone. Sandstones have scoured bases and contain abundant mudchips which are a diagnostic feature of the L3 member. The upper submember is also a thin sequence (<20 m) of sandstone but is characteristically coloured maroon with abundant maroon mudchips. The unit is thin- to medium-bedded with minor thin siltstone layers; sands have scoured bases and are trough crossbedded. Mud layers draping crossbedded sand layers are characteristically mudcracked.

Both the lower and upper contacts of the L3 member are conformable. The upper contact of this member is used as the datum in Figure 7 because the member is very distinctive in the field, and also because it does not change in thickness across strike as abruptly as other members in the Link Formation.

Tinney Hills Member.

The Tinney Hills member is a major clastic wedge with a maximum thickness of almost 400 m at section 1, tapering northwestward to 130 m near section 13 (Fig. 8). It is composed dominantly of braided fluvial subarkosic sediments that increase in feldspar content to the southeast. As the first hinterland-derived molasse wedge, it represents an important phase in the evolution of the Bear Creek Foredeep. It is succeeded in time and magnitude by the regionally extensive Burnside Formation which represents the main phase of molasse sedimentation in the basin (see McCormick and Grotzinger, 1988).

Figure 8 reveals that the Tinney Hills member is bounded by disconformable sequence boundaries. In some localities the lower contact appears gradational but in others it is sharply disconformable, cutting down into the underlying siltstones for up to 10 m. The upper contact is everywhere disconformable in the Tinney Hills area and up to 35 m of the upper part of the member are locally cut out beneath this surface. The Tinney Hills member is overlain by a sequence of marine siltstone, mudstones and sandy siltstones up to 250 m thick that grades up into the fluvial sandstones of the Burnside Formation. The upper contact of the member is commonly marked by an erosional surface underlain by extensive brecciated and comminuted sandstone and siltstone clasts, and characteristically draped by laminated carbonate up to 50 cm thick that isopachously drapes the breccia. The carbonate is best developed along the basal and marginal parts of stratigraphic ''cut-outs''. The carbonate is commonly overlain by marine siltstones of the uppermost Link Formation.

Internally, the Tinney Hills member is characterized by numerous fining-upward channel sequences. These are on the order of a few metres thick, and each sequence is characterized by a scoured base, a variably developed channel lag, followed by trough crossbedding and then planar beds/laminae capped by current ripples. Channel dimensions vary, but are commonly on the order of tens of metres in width and up to 3 metres in depth. The flat channel geometry combined with the absence of overbank deposits suggests that the fluvial system was braided with variable discharge.

The bulk of the member is characterized by one or two larger scale thinning-upward and fining-upward sequences up to 100 m thick In the upper third of the member, one sequence is overlain by a thin siltstone sequence that lacks desiccation cracks and contains abundant wave ripples (see Section 2 of Fig. 8, for example). This thin unit is interpreted as marine, but passes upwards into a final fluvial sequence just below the upper contact of the member.

CONCLUSIONS

The stratigraphy of the sub-Burnside Bear Creek Group in the Tinney Hills area records the spatial transition from shallow shelf environments fringing the outer trench slope of the foredeep (Slave Craton) to deeper slope and basinal environments of the foredeep trench axis. The shelf-to-slope transition is covered in the Hackett Formation, but is superbly exposed for the three sequences in the Rifle Formation and the L2 member of the Link Formation. Results presented here, combined with previous work (Grotzinger et al., 1987) show that the position of the ''shelf edge'' fluctuated only 10-15 km between units in the Rifle, Beechey and Link formations. Recognition and mapping of sequence boundaries in addition to lithofacies boundaries permits primary depositional (time) surfaces to be identified.

The L1 member of the Link Formation records filling of the trench axis and the lateral transition of southeasterly turbidite facies into northwesterly shelf facies. The Tinney Hills member of the Link Formation represents the final filling of the trench axis, the destruction of paleobathymetric relief in the basin, and the progradation of fluvial molasse across the zone of maximum subsidence. On a larger scale, this probably correlates with the initiation of major unroofing of the Thelon Tectonic Zone and/or Queen Maud Block, and decreased basin subsidence related to isostatic recovery of the lithosphere in response to load reduction (erosion).

ACKNOWLEDGMENTS

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Aspects of the Burnside Formation, Bear Creek Group, Kilohigok Basin, District of Mackenzie, N.W.T.

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McCormick, D.S. and Grotzinger, J.P., Aspects of the Burnside Formation, Bear Creek Group, Kilohigok Basin, District of Mackenzie, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 321-339, 1988.

Abstract

The Burnside Formation is a thick (1.5 to >3.0 km) northwest-tapering wedge of predominantly fluvial subarkose which dominates the upper portion of the upward-shallowing early Proterozoic Bear Creek Group. Detailed sedimentological and stratigraphic studies reveal that the Burnside Formation represents a prograding braid-delta/braided-stream system which flowed west-northwest out of the present Thelon Tectonic Zone. Lithofacies geometry, paleocurrent patterns, stratigraphic position with respect to other units in the Goulburn Group, and conglomerate clast composition are most consistent with a foreland basin setting, with the Burnside a culminating molasse. Correlations between distal Burnside sections and nearby sections in the Epworth Group suggest that the Burnside fluvial system interfingered with marine shelf facies of the contemporaneous passive margin of Wopmay Orogen. The large volume of siliciclastic material in the Burnside wedge requires that the fluvial system drained an area larger than the present Thelon Tectonic Zone, suggesting that the Queen Maud block may have been uplifted during Burnside sedimentation.

Résumé

La formation de Burnside est constituée d'un épais biseau (1,5 à >3,0 km) composé surtout d'arkose d'origine fluviatile qui va en s'amincissant vers le nord-ouest; ce biseau domine la partie supérieure du groupe de Bear Creek datant du Paléozoïque inférieur et mis en place dans un milieu d'eaux progressivement moins profondes, vers le haut. Des études sédimentologiques et stratigraphiques détaillées révèlent que la formation de Burnside représente un réseau hydrographique anastomosé de type deltaïque ou fluvial en progression vers la mer; ce réseau se serait écoulé vers l'ouest nord-ouest, hors de la zone connue aujourd'hui sous le nom de zone tectonique de Thelon. La géométrie des lithofaciès, les tracés des paléocourants, l'emplacement stratigraphique par rapport aux autres unités du groupe de Goulburn, et la composition des fragments de conglomérat, sont cohérents avec les conditions particulières à un milieu de bassin d'avant-pays, la formation de Burnside jouant le rôle de molasse sommitale. Des corrélations établies entres des sections distales de la formation de Burnside et des sections avoisinantes du groupe d'Epworth semblent indiquer que le réseau fluvial de Burnside se soit mêlé au faciès de milieu marin de plate-forme correspondant à la marge passive contemporaine de l'orogène de Wopmay. Le volume considérable de matériaux silico-clastiques contenu dans le biseau de Burnside aurait exigé que le réseau fluvial draîne une surface plus vaste que la zone tectonique actuelle de Thelon; ce phénomène porte à croire que le bloc de Queen Maud aurait pu faire l'object d'un soulèvement au cours de l'épisode de sédimentation responsable de la mise en place de la formation de Burnside.

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INTRODUCTION

The Burnside Formation is a northwest-thinning wedge of subarkosic foredeep molasse, deposited mostly in a braided stream environment, that caps of the upward-shallowing Bear Creek Group in the early Proterozoic Kilohigok Basin (Fig. 1,2,3; Grotzinger et al., 1987). A sedimentologic and stratigraphic study of the Burnside Formation is being conducted to document (1) the evolution of and controls on a Proterozoic fluvial system, (2) the provenance of the Burnside Formation and its relationship to the Thelon Tectonic Zone (TTZ), and (3) the relationship of the Burnside Formation to the Epworth Group of Wopmay Orogen.

This report summarizes the results of field studies conducted during the summers of 1986 and 1987; it builds on and complements previous results of this study (Grotzinger and Gall, 1986; Grotzinger et al., 1987), and forms part of

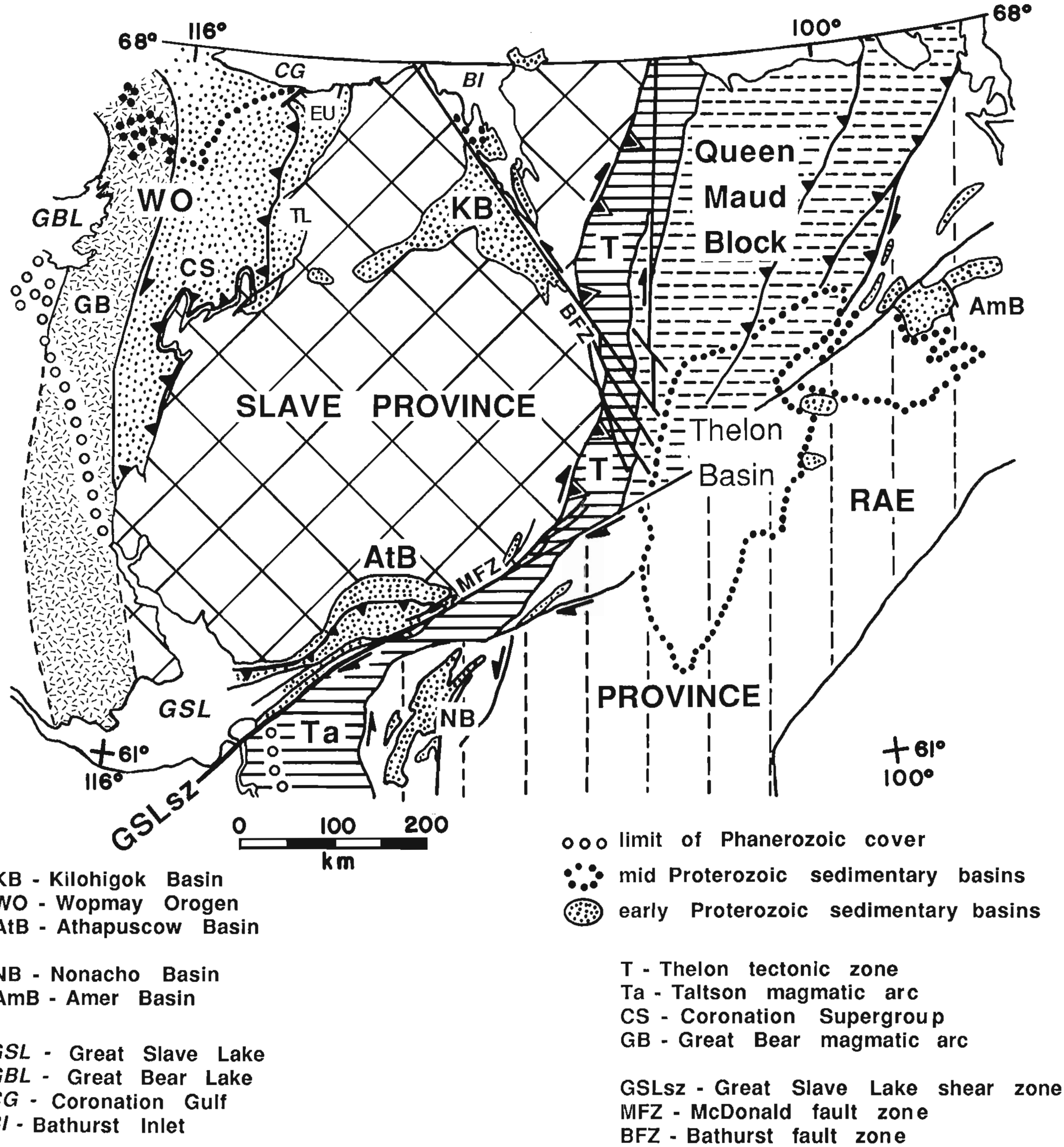

Figure 1. Location of Kilohigok Basin within the northwestern Canadian Shield and related Archean and Proterozoic areas. Figure modified after Hoffman (1987). TL — Takiyuak Lake; EU — Eokuk Uplift.

a larger study documenting the sedimentologic and tectonic evolution of the Kilohigok Basin. This study of the Burnside Formation is the subject of a Ph.D. dissertation for one of us (DSM). Thirteen stratigraphic sections were measured, informal members were defined locally for mapping and correlation purposes, and members were mapped at 1:50 000-scale in some locations. Areas where the Burnside Formation was studied include the Contwoyto Lake (76E), Beechey Lake (76G), Tinney Hills (76J), Mara River (76K), Arctic Sound (76N), Rideout Island (76O), and Point Lake (86H) map areas. In addition, two stratigraphic sections were measured in the Epworth Group of Wopmay Orogen (Hoffman et al. 1984) in the Takiyuak Lake (86I) and Kikerk Lake (86P) map areas.

Work thus far refines conclusions of previous structural and stratigraphic studies of the Goulburn Supergroup in that the thickness, lithofacies, and paleocurrent trends within the Burnside Formation are most consistent with deposition in a foreland basin which formed in response to convergence, northwest-directed thrusting, and probable flexural loading of the southeastern margin of the Slave Province by the Rae Province (northwest Churchill Province) along the TTZ (Tirrul, 1985; Grotzinger and Gall, 1986; Hoffman, 1987; Grotzinger and McCormick, in press). The Burnside Formation is a culminating molasse that records the unroofing of an igneous-metamorphic terrane that lay to the southeast in the TTZ (Grotzinger et al., 1987; Grotzinger and McCormick, in press) and possibly the Queen Maud Block. Furthermore,

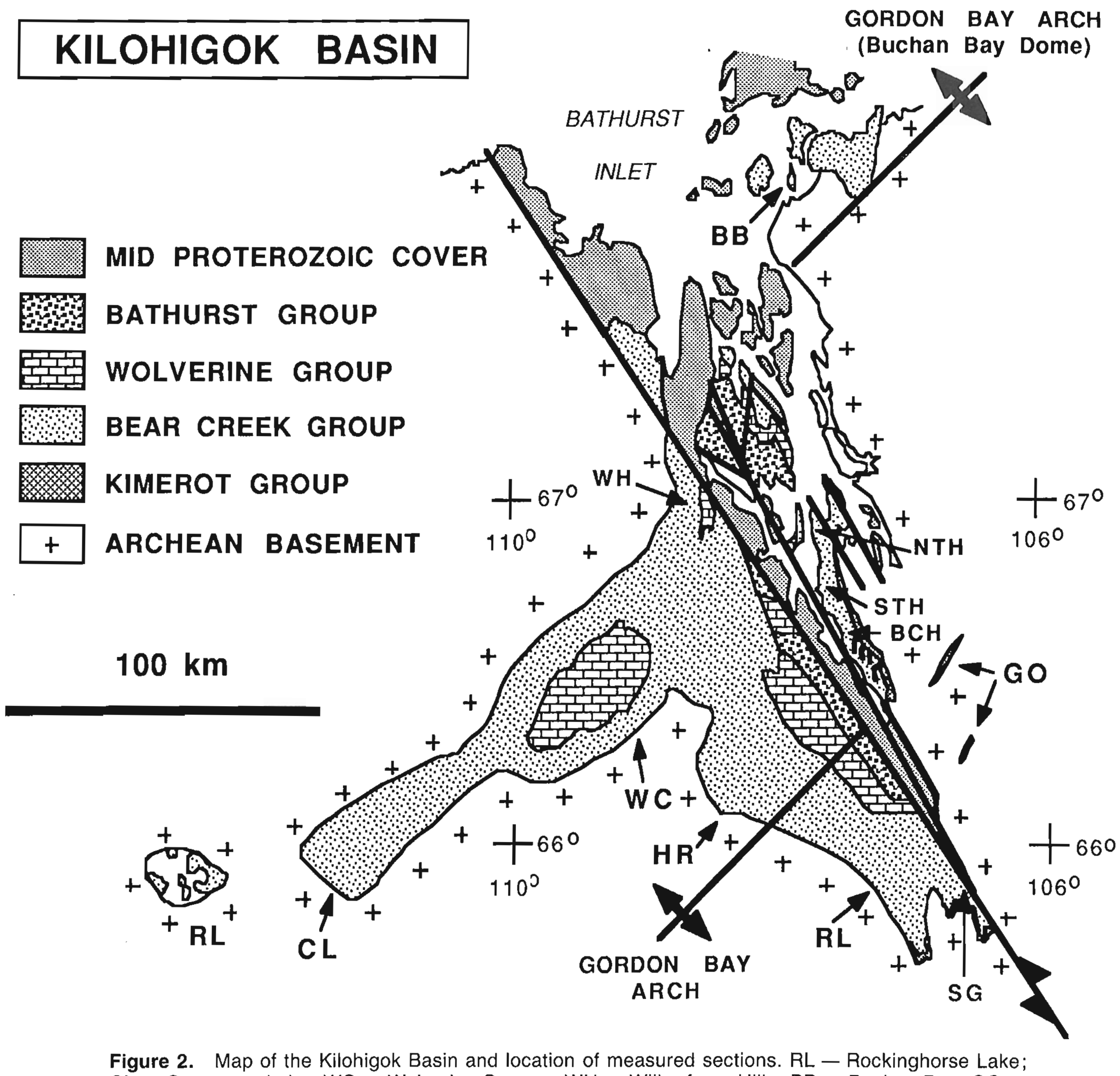

Figure 2. Map of the Kilohigok Basin and location of measured sections. RL — Rockinghorse Lake; CL — Contwoyto Lake; WC — Wolverine Canyon; WH — Wilberforce Hills; BB — Buchan Bay; GO — Goulburn Supergroup outliers; HR — Hackett River; LL — Link Lake; SG — Straight Gorge; NTH — northern Tinney Hills; STH — southern Tinney Hills; BCH — Bear Creek Hills.

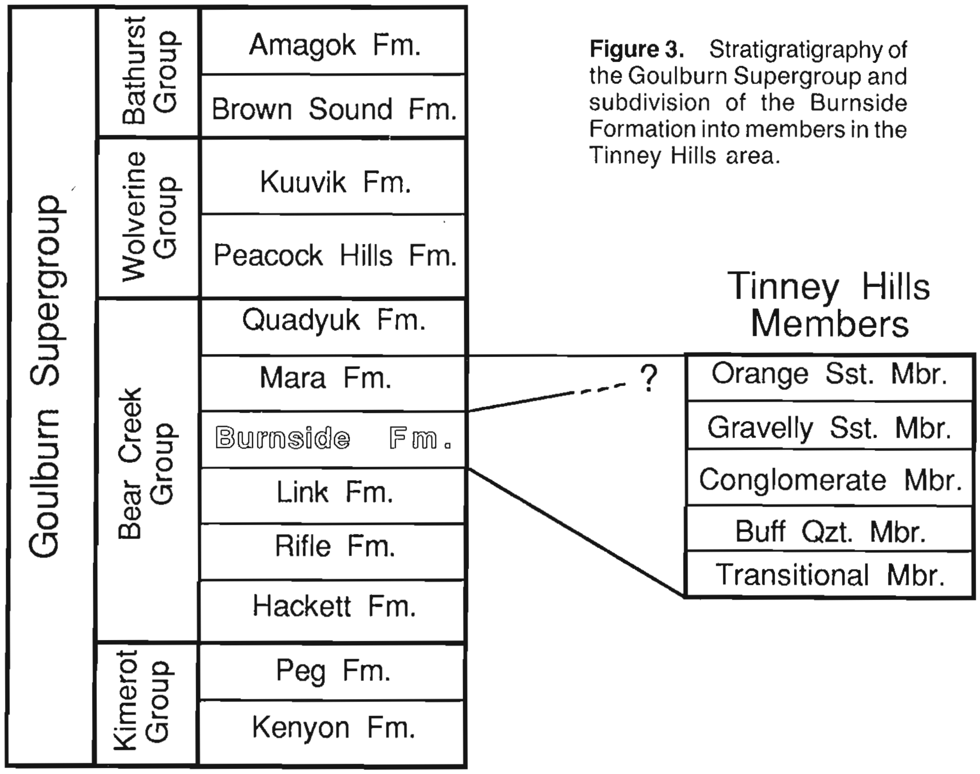

Figure 3. Stratigratigraphy of the Goulburn Supergroup and subdivision of the Burnside Formation into members in the Tinney Hills area.

this study strengthens previous correlations that show that the Burnside braid-delta/braided-stream system prograded to the northwest across the Slave Province, into the in-part coeval passive margin of Wopmay Orogen (Hoffman et al. 1984; Grotzinger and Gall, 1986; Grotzinger et al., 1987).

In particular, this report addresses the following findings.

(1) The subdivision of the Burnside Formation into siliciclastic lithofacies groups by Grotzinger et al. (1987) is refined and extended to include marine shelf facies found in the distal part of the basin. The spatial and temporal distribution of Burnside lithofacies allows interpretation of marine, deltaic, and fluvial environments. It also constrains the evolution of the fluvial system and the subsidence history during the latter stages of Bear Creek Group sedimentation. The proximal Burnside lithofacies, dominated by braid plain facies, reflect initial aggradation of the fluvial system behind the Gordon Bay arch, followed by progradation over the arch and across the Slave craton as the basin filled. The presence of the Burnside Formation lying directly on granitic basement to the north suggests that the Gordon Bay arch may have been a positive topographic feature locally until the latter stages of Bear Creek Group sedimentation. The distal Burnside lithofacies represent the intertonguing of marine shelf, braid delta, and distal braid plain environments. Eustatic sea level changes may have influenced the fluvial system during the latter half of Burnside sedimentation.

(2) The changes in quartzite and conglomerate composition upward in the stratigraphic section reflect increasing contribution from three sources which lay to the southeast. In order of their introduction, these sources are: (i) metasediments possibly derived from Archean basement; (ii) intraformational clasts cannibalized from the Burnside Formation itself; and (iii) an igneous terrane which supplied increasing amounts of feldspar, and felsic volcanic and granitic clasts.

(3) The correlation between distal sections of the Burnside Formation and the Epworth Group of Wopmay Orogen, as suggested by Hoffman et al. (1984), is strengthened by comparison between distal Burnside lithofacies and those in the Odjick Formation. Furthermore, the link between the Rocknest Formation and the dolomite member of the Burnside Formation also appears strong. These correlations imply synchronous evolution of a foreland basin on one side of the Slave craton with a passive margin on the other.

LITHOFACIES GROUPS AND SEDIMENTARY ENVIRONMENTS

Siliciclastic lithofacies and their characteristic groupings in the Burnside Formation have been defined previously (Grotzinger et al., 1987, p. 228-234). More detailed studies in the most proximal and distal ends of the Kilohigok Basin warrant the refinement of previous lithofacies groups and the addition of new groups. The first two lithofacies groups described below have tabular or unchannelized bed geometries, while the latter five contain abundant scours and channels of various scales. The siliciclastic lithofacies designations used here are modified after Miall (1977, 1978) and Rust (1978).

Sr and Shcs

These lithofacies are ripple crosslaminated sandstones and hummocky cross-stratified sandstones, respectively (Fig. 4a, b). Sr comprises 5-15 cm tabular, laterally continuous beds of very fine- to fine-grained maroon to pink

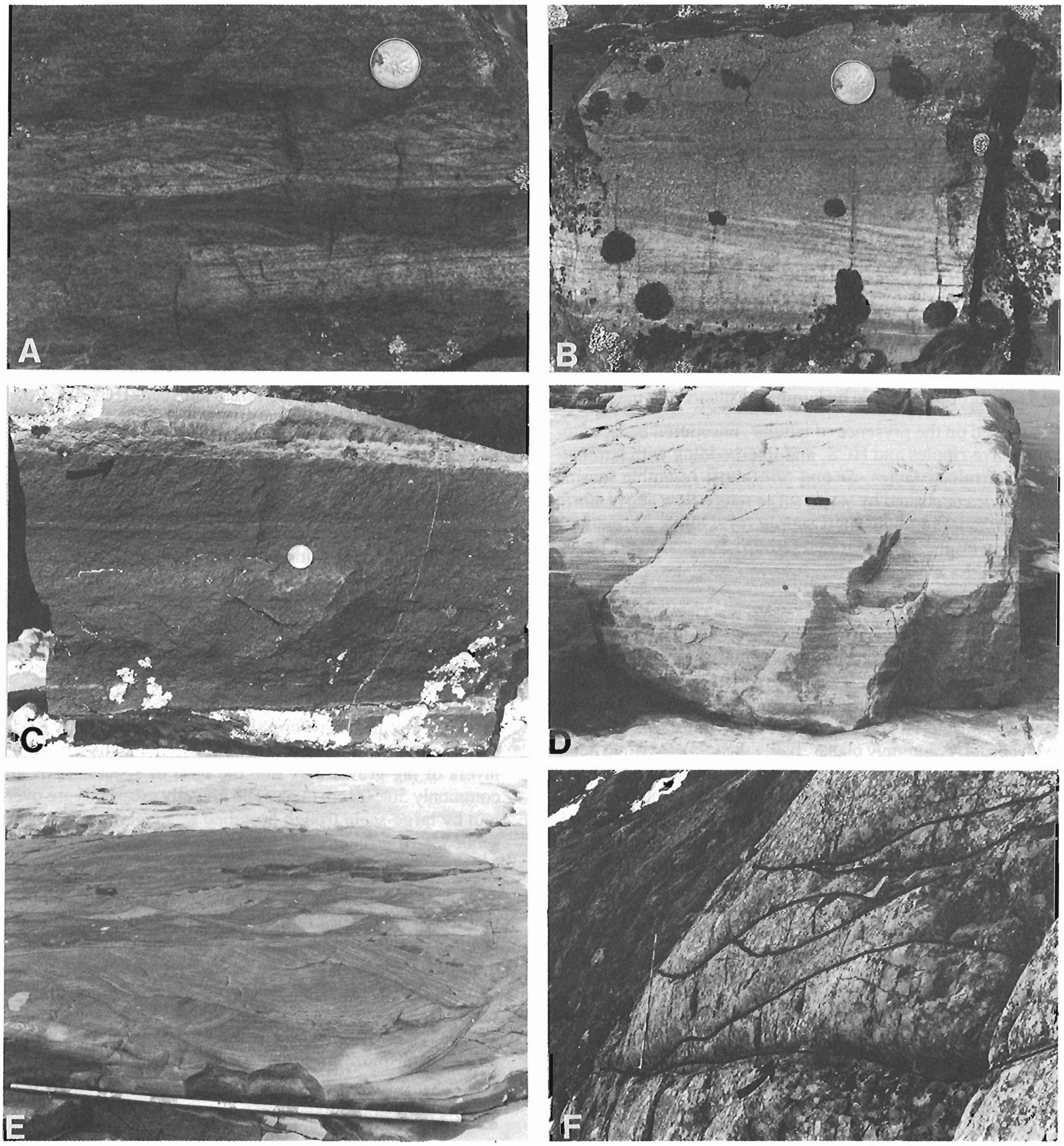

Figure 4. Common lithofacies of the Burnside Formation. A) Sr lithofacies with symmetrical wave ripples in upper bed interbedded with laminated mudstone (dark). Coin for scale. B) *Shcs* lithofacies with low-angle crosslaminae and internal truncations. Coin for scale. C) *Sh* lithofacies with diffuse horizonal laminae. Bed is capped by wavy-laminated green siltstone (arrow). Coin for scale. D) *St-Sh* lithofacies showing thick bed of horizontally laminated medium-grained sand. Knife for scale, about 10 cm long. E) *St* lithofacies with two scales of trough crossbedding and horizontal bounding surfaces between crossbed sets. Staff is 1.5 m long. F) *Buff Qzt* lithofacies seen at an oblique angle. Numerous internal scours are seen, and a master bedding surface with very low relief is seen at base (arrow). Staff is 1.5 m long.

subarkose. Beds are planar-based, commonly contain current ripples or horizontal laminae at the base, and grade to wave ripples at the top. Minor loads and flame structures are present. Thin (millimetre to a few centimetres) beds of laminated green or red siltstone or mudstone commonly cap these beds. This facies occurs at the base of the Burnside Formation, is gradational with underlying laminated or massively-bedded siltstone of the Link Formation, and grades upward to *Sh* lithofacies.

Shcs consists of 5-30-cm beds of white to maroon, fine-to medium-grained subarkose. The base of beds are planar and tops are hummocky over 30-100 cm; internal low-angle truncations of low-angle lamina sets are present. These features are considered diagnostic of hummocky cross-stratification (Allen, 1981; Dott and Bourgeois, 1982; Soegaard and Eriksson, 1985). Bed tops are locally wave-rippled, and are abruptly overlain by laminated siltstone or mudstone.

Based on the presence of tabular, unscoured beds, common wave ripples and HCS, and interbedding with laminated fine-grained facies, *Sr* and *Shcs* most resemble marine shelf facies deposited by storm surge return flow above storm wave base in a middle shelf setting (cf. Odjick Formation: Hoffman et al., 1984; Ortega Group: Soegaard and Eriksson, 1985).

Sh

This lithofacies group consists of horizontally-laminated sandstone which is moderately- to well-sorted, very fine- to fine-grained, pink, and subarkosic (Fig. 4c). Beds are 10-60 cm, tabular, and laterally persistent over tens to hundreds of metres with abrupt, commonly planar, bases. Scours of < 5 cm depth may be present both at the base and internally, and are commonly overlain by layers or lenses of green mud clasts up to 5 cm. Internally, beds are mostly horizontally-laminated, though it is commonly indistinct; a few centimetres of current- and wave-rippled silty sandstone with mudstone and siltstone laminae cap most beds. Minor fining occurs in the upper few centimetres of beds. In coarsening-upward sequences, especially in the southeastern belt of the Tinney Hills, Sh is gradational from underlying *Sr*, and scouring increases with grain size.

The *Sh* group is similar to distributary mouth shoals and swash bars of a braid delta environment described by Vos (1981). This interpretation is based on the lateral persistence of beds, textural maturity, fine grain size and the common occurrence of horizontal laminae and ripples on bed tops. Single beds are interpreted as the product of a single flood event supplied from braid delta distributary channels which were subsequently reworked by storm and wave action.

Channelized lithofacies

The remaining five lithofacies contain evidence of various scales of channelization and erosion between depostional units, all of which contain mostly trough crossbedding. These lithofacies comprise most of the Burnside Formation. They are essentially subdivisions of the *St* assemblage of Grotzinger et al. (1987, p. 232), and are distinguished by the presence of gravel and associated sedimentary structures. The last three lithofacies are largely restricted to the source-proximal southeastern outcrop belt at Straight Gorge and along the east side of Bathurst Inlet.

Some of these lithofacies resemble those described by Campbell and Cecile (1981). However, the current study distinguishes the various depositional environments in detail, and constrains the stratigraphic distribution of these lithofacies.

St and gSt

The *St* lithofacies forms the majority of the Burnside Formation especially in the outcrop belt to the west of Bathurst Inlet. This lithofacies consists of 50-120-cm beds of fine-to coarse-grained, pink subarkose with medium- and large-scale trough crossbedding (10-30 cm and 30-80 cm, respectively) that show vertical decrease in size within beds (Fig. 4e). Lower contacts are undulatory 5-50 cm scours and are uncommonly overlain by a layer of mud clasts or gravel. Gravel along foresets is rare. Horizontally-laminated or ripple cross-laminated very fine- to fine-grained sand may cap the upper 5-20 cm of beds, but are commonly scoured by overlying beds. Paleocurrents in this facies are commonly unimodal with low dispersion within and between beds in sequences of tens to hundreds of metres. This facies is commonly gradational with the *Sh* as scouring, grain size, and crossbedding increases. In the upper part of the section at the northern Tinney Hills troughs are commonly lined with heavy mineral laminae. *St* is gradational with all other lithofacies.

The *gSt* lithofacies is identical to the *St*, except that the basal large-scale trough crossbed sets contain gravelly medium to coarse sand along crosslaminae and that discontinuous layers of lag gravel may line the base of scours. Beds are commonly 50-500 cm thick. The gravelly crossbeds are overlain by large-scale trough crossbed sets lacking gravel; the sequence of structures is otherwise identical to *St*.

The upward decrease in grain size, the scale of trough crossbedding, and unimodal paleocurrent patterns suggest that the St lithofacies represents vertical infilling of broad, shallow channels by migrating sinuous-crested subaqueous dunes in a sandy delta plain to lower braid plain environment subject to rapid channel switching and cannibalization (cf. Eriksson and Vos, 1979; Vos, 1981; Walker and Cant, 1984). The *gSt* lithofacies represents somewhat deeper braid channels on more proximal reaches of the braid plain.

Buff Qzt lithofacies

The Buff Quartzite Member of the Burnside Formation was distinguished for mapping purposes in the Tinney Hills (Fig. 3). It is a massive, very thick-bedded, prominent cliff-forming lithofacies which dominates the eastern crest of the Tinney and Bear Creek Hills east of Bathurst Inlet. It is composed of medium- to coarse-grained subarkose in beds 120-300 cm thick, bounded by nearly planar surfaces, especially at the base of bedsets (Fig. 4f). Bedsets are 8-20 m thick, composed of 5-10 beds. It is distinguished by a monotonous repetition of large-scale trough crossbedding, with 5-10 crossbed sets per bed. Minor medium-scale trough cross-bedded, horizontal-laminated, and ripple crosslaminated sand appear only in the upper metre or two of a bedset. Individual beds are traceable over hundreds of metres laterally; the bed- or bedset-bounding surfaces could not be traced into larger-scale channel forms, but the presence of very large scale channels or lateral accretion surfaces cannot be dismissed. The *Buff Qzt* culminates the three prominent coarsening-upward sequences of the Transitional Member in the Tinney Hills. Vertically there is a common increase in gravel and it is overlain by the *Gm-St* lithofacies.

The monotonous repetition of trough crossbedding, lateral persistence of bedding surfaces, and the incomplete fining-upwards sequences in the *Buff Qzt* resemble the Westwater Canyon Member of the Morrison Formation, a Jurassic-age braided fluvial sheet sandstone (C. Campbell, 1976). One dissimilarity is that distinct enclosing channel bodies cannot be identified. This may be due to the very large width-to-depth ratio of the channels in the Westwater Canyon Member (up to kilometres across) and the fact that outcrops in the Tinney Hills are oriented at an acute angle to the predominant transport direction in the Buff Qzt Member, thus obscuring across-strike geometries. Despite these differences, the *Buff Qzt* lithofacies is interpreted to be the product of vertical aggradation of deep (metres) fluvial channels during flood stage or perennial flow conditions by sinuous crested dunes. The *Buff Qzt* also resembles certain braided fluvial facies described from the Archean Moodies Group (Eriksson, 1978).

St-Sh

This lithofacies differs from St in that thick (50-300 cm) beds of horizontally-laminated, medium-grained sand cap the trough crossbedded facies (Fig. 4d). It differs from Sh lithofacies because it is coarser grained, contains upper flow regime bedforms, and is associated with trough crossbedded sand. Individual laminae are 0.5-2.0 cm thick and are traceable over metres to tens of metres. Locally horizontal lamina sets pass into current ripple crosslaminated or small-scale trough crossbedding for a few centimetres. Locally a few isolated scours up to 30 by 200 cm are filled with trough crossbedded sand that grades up to, and is entirely enclosed within, horizontally-laminated sand. This facies is best developed at the top of the Gravelly Sandstone Member in the Tinney Hills.

This group is interpreted as vertical infilling of braid channels by the migration of sinuous-crested dunes (*St*) until flow depth decreased so that horizontally-laminated braid-bar-top sand flats formed. Scours with troughs are interpreted as scoured chute channels on the bar tops where flow depth was deep enough to form dunes. This resembles the sand flat facies sequence in the South Saskatchewan River (Cant, 1978). A notable difference is that Cant indicates that planar crossbedded sand (*Sp*) is a common component, which he interprets as cross-channel bars formed at low-flow stage. The lack of *Sp* in this Burnside lithofacies group suggests that river flow and channel aggradation were relatively continuous until flow depth decreased to form *Sh*, rather than having low-flow lateral accretion off bars. This is an indication that the Burnside fluvial system operated under humid perennial flow conditions.

Sp-St

This lithofacies differs from *St* by the presence of broad scours or channels 30-120 cm deep and 3-8 m wide that are infilled with planar-tangential crossbedding which is commonly highly oblique to both channel scours and trough crossbedding (Fig. 5). Beds are 120-500 cm thick with a few centimetres of relief on beds. Scour infilling is commonly incomplete, leaving residual topography which is overlain by medium- or large-scale trough crossbedding; upward decrease in scale of trough crossbedding is the same as in *St*. This facies is gradational with *St*. It is also gradational upward from *Gm-St* lithofacies where it commonly contains gravel along cross-laminae in both planar and trough crossbedded sands.

The scale of bedding, larger scale of crossbedding and divergent paleocurrent patterns within beds suggests a combination of vertical infilling during flood-stage and lateral accretion during low-flow stage of deeper fluvial channels on a sandy braid plain (Fig. 5). Planar crossbeds represent lateral infilling of scours off the margin of braid bars (cf.

Figure 5. *Sp-St* lithofacies in the southern Tinney Hills. Large set of planar-tangential crossbeds carrying polymictic gravel in lower 30-50 cm of set. Scour at base of crossbed set cuts down about 80 cm. Set is overlain by horizontally-laminated sand. Planar tangential crossbeds are oriented about 240°, whereas the predominant trough crossbed paleocurrent azimuth is directed into picture (about 300°). Staff is 1.5 m long.

Proterozoic Waterberg Group: Eriksson and Vos, 1979; South Saskatchewan River and Battery Point models: Walker and Cant, 1984).

Gm-St

This lithofacies group is unchanged from previous descriptions, except to note that the trough crossbedding above gravel layers is commonly gravelly and shows both coarse-tail grading and decrease in trough size upwards. Beds are commonly 50-150 cm between scoured irregular basal surfaces which are lined with lag gravel layers up to 50 cm thick (Fig. 6a). Scours are commonly 50-200 cm deep and 5-20 m wide; the largest clasts commonly occur in lenses at the base of these scours (Fig 6b). Distinctive conglomerate horizons can be mapped laterally up to 20 km along the Tinney Hills based on the presence of unique clast types.

This lithofacies is interpreted to represent repeated dissection of the proximal braid plain by migrating channels which carried gravel at high flood stage, planing off and covering resulting topography. After peak flood stage, sinuous-crested dunes migrated through the channels carrying finer gravel. The upward decrease in gravel and size of trough crossbedding indicates waning flow strength. The incomplete fining-upwards in beds results from the deep scouring of the next flood event. Individual stream channels were probably on the order of 80-150 cm deep, but based on the high transport energy of this lithofacies, channels may have been much deeper (cf. Proterozoic Witswatersrand Group: Minter, 1978; Proterozoic Waterberg Group: Eriksson and Vos, 1979).

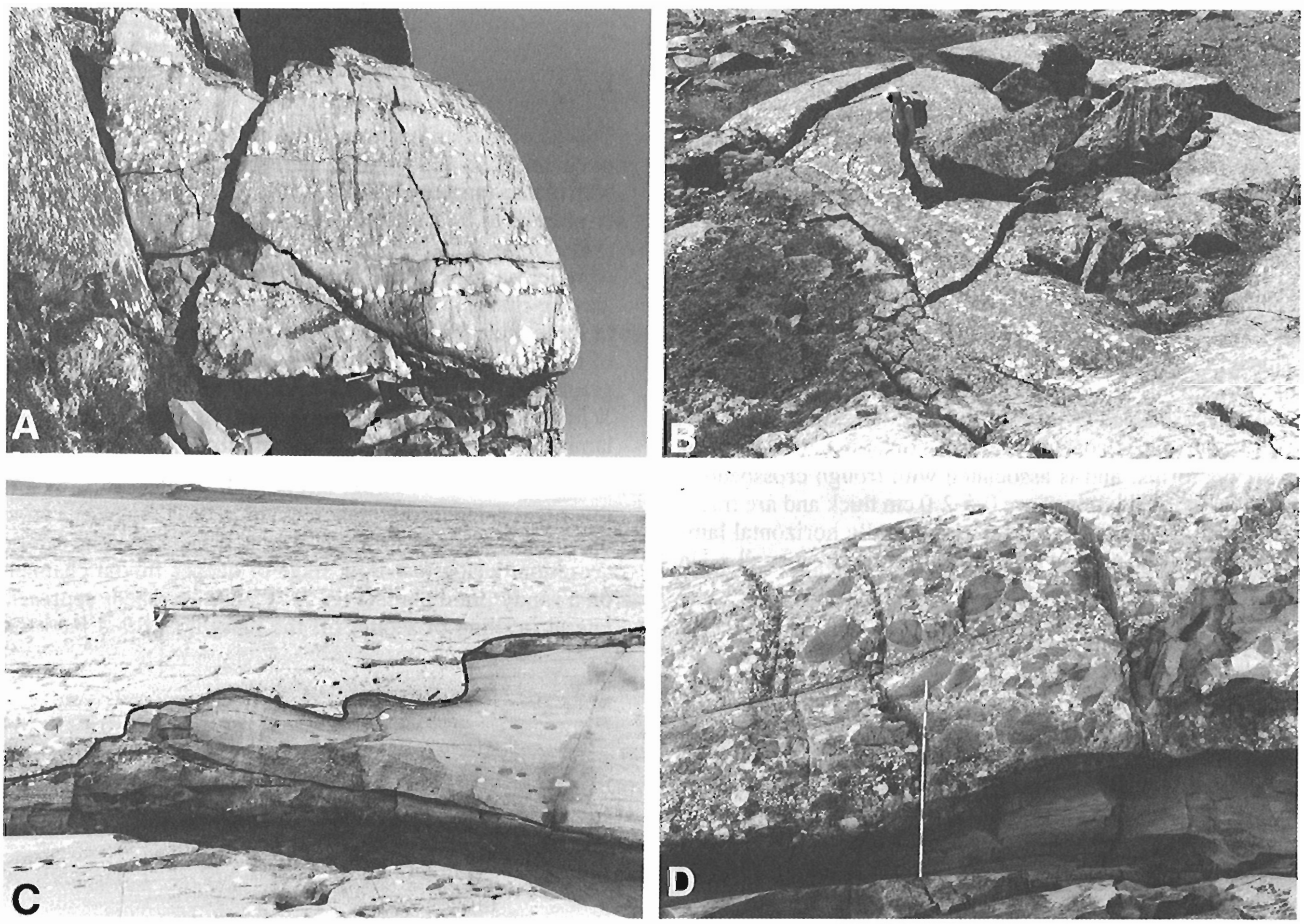

Figure 6. Conglomeratic lithofacies of the Burnside Formation in the Tinney Hills. A) Base of Conglomerate Member showing crudely bedded gravel with coarsest clasts at base of beds. Upper parts of beds are trough crossbedded sandstone. Hammer for scale (arrow). B) Oblique view of same horizon about 30 m along strike. View shows the broad scours (about 5 m) and lateral amalgamation of conglomerate layers. Coarsest clasts line the base of scours. Person for scale. D) Uppermost conglomerate horizon at southern Tinney Hills at the top of the Gravelly Sandstone Member. Lower bed is volcanic clast horizon and upper bed is polymictic horizon with Burnside intraformational clasts. Note steep scour relief on channel margins and nearly-horizontal upper surface. Staff is 1.5 m long. D) Manning Point conglomerate horizon at northern Tinney Hills. Largest dark clasts are Burnside lithofacies which define deep channels. White clasts are vein quartz. Smallest clasts are the background polymictic population, similar to those in the previous photo. Staff is 1.5 m long.

MEASURED SECTION AND LITHOFACIES GEOMETRY

The areas investigated can be divided into two main outcrop belts based on thickness and lithofacies present, (1) a Southeastern Belt from Bear Creek Hills to Tinney Hills, and (2) a Northwestern Belt from Straight Gorge to Rockinghorse Lake.

Southeastern Belt

In the proximal Southeastern Belt, the Burnside Formation thins significantly to the northwest (Fig. 7), towards the Gordon Bay arch of Grotzinger and Gall (1986), matching trends observed in the underlying lower Bear Creek Group (see Grotzinger and Gall, 1986; Grotzinger et al, 1987; Grotzinger et al., 1988). The datum for the three sections are a distinctive quartz pebble conglomerate layer. Although this horizon is clearly time-transgressive to the northwest, the sections span only about 30 km across-depositional-strike and each section is more than 1000 m thick. It is noteable that the use of this datum will reduce the appearance of differential subsidence because this boundary is younger in the direction of decreasing subsidence.

The base of each of the sections comprise the Transitional Member which contain three thick (> 100 m) mappable coarsening-upwards sequences, present only in the Southeastern Belt, that pass from *Shcs+Sr* to *Sh* to *St* lithofacies groups (Fig. 7) The lower two sequences are capped by an abrupt transition (<25 m) to finer grain size and decreased scale of sedimentary structures, while the last one is overlain by the Buff Quartzite Member.

The most striking feature of this cross-section is the pronounced thickening towards the axial (southeast) part of the basin. A simple extrapolation predicts that the Burnside Formation may have been over 3.8 km thick in the Bear Creek Hills but only 1.5 km thick in the northern Tinny Hills, only 30 km further onto the craton. The area of maximum thickening (STH to BCH) spatially coincides with the shelf edge

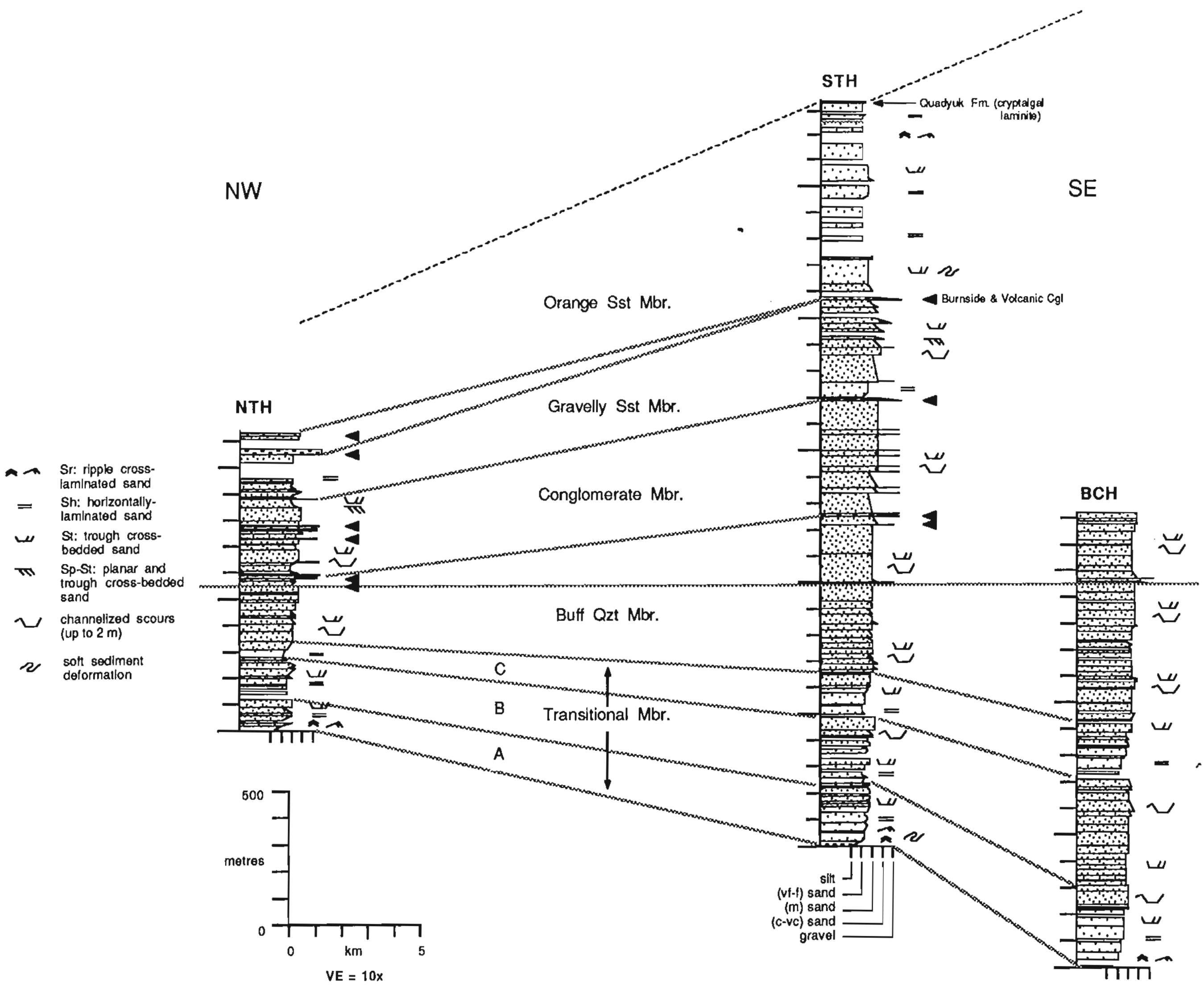

Figure 7. Cross-section of the Burnside Formation in the Southeast Belt from Bear Creek Hills to North Tinney Hills. Note pronounced thinning to northwest. Arrows indicate stratigraphic level of conglomerate clast counts described in text and in Figure 12 and Table 1.

Table 1. Abundance of clast lithologies in the Tinney Hills

							Clast Lithologies							
Location	No. Clasts	Vein Qz %	G/W Qzt %	Foliated Qzt %	Blk Meta Sltst%	Burnside Qzt %	Volcanic %	Colored Qzt %	Arg/ Siltst %	Jasper/ Chert %	Amph/ Gneiss %	Breccia %	Granitoid %	Other %
STH-4b	250	57.6	15.2		1.2	8		0.8		5.2	0.8	3.2	7.2	0.8
STH-4a	100	37	6			9	24	2	1	1		1	18	1
STH-3	248	71.3	12.5	5.1	1.1	7.7				1.1	0.4	0.8		
STH-2	230	48.2	18.3	14.8	5.2			7.4	5.2	0.9				
STH-1	274	93.4	6.2					0.4						
NTH-MPb2	136	43	3.5		0.8	7.5	44.4			0.8				
NTH-MPb1	133	57.4	3.7		1.5	5.9	26.4	2.9		0.7	1.5			
NTH-MPa2	132	51.6	7.6		0.8	27.2		8.3	3	1.5				
NTH-MPa1	160	57.2	6.3			26.5		6.3		3.1			0.6	
NTH-5	100	39	6	3	2	47				1			2	
NTH-4	100	92	5	1	2									
NTH-3	101	77	8	6	6					3				
NTH-2	100	98	2											
NTH-1	100	98	2											

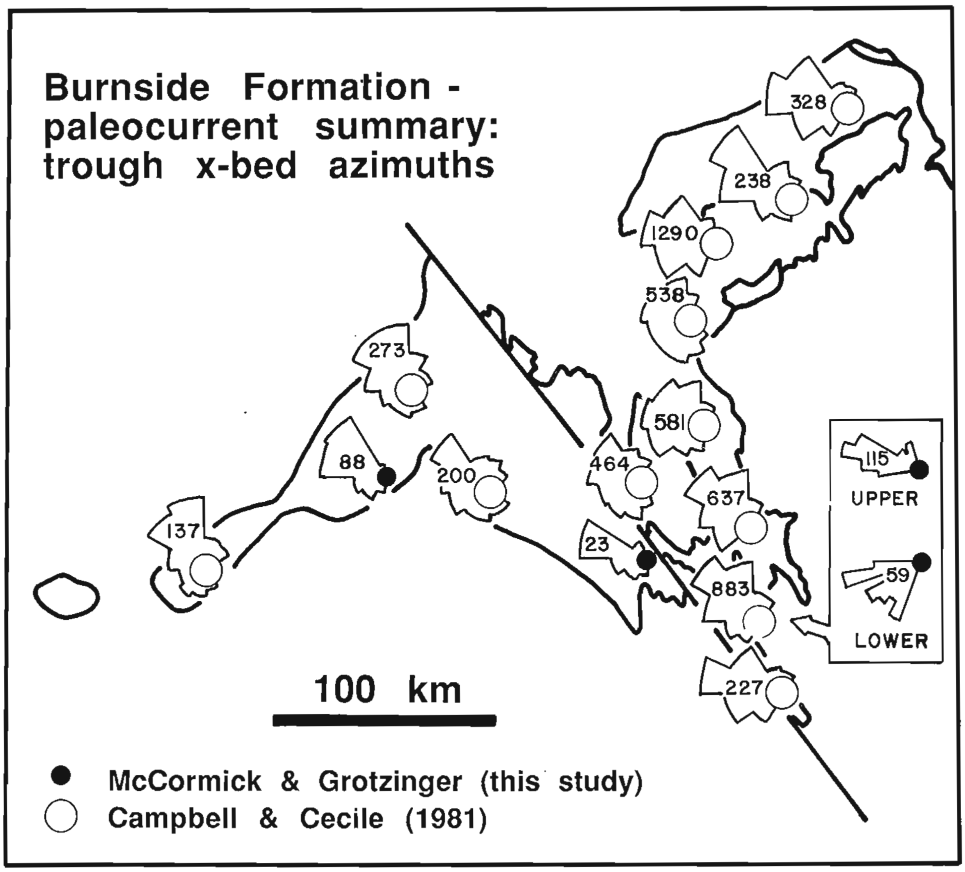

Figure 8. Paleocurrent summary of the Burnside Formation. Note the distinctive southwest mode in the Transitional Member in the Tinney Hills ("lower") as compared with upper members ("upper"). At all other locations the paleocurrents are west-northwest-directed with low-dispersion. Note that this figure show the basin restored for approximately 130 km of left-lateral slip on the Bathurst Fault zone, as recommended by Campbell and Cecile (1981) and Tirrul (1985).

and area of maximum gradient of thickening of the underlying lower Bear Creek Group (Grotzinger and Gall, 1986; Grotzinger et al., 1988). This suggests that the mechanical property of the basement which localized the position of the underlying lower Bear Creek shelf edge (Grotzinger et al., 1988) exerted a similar control on the patterns of proximal foredeep subsidence throughout Burnside sedimentation as well.

The coarsening-upward sequences are interpreted to represent the marine-to-fluvial transition with the incursion of a fluvial-dominated sandy braid delta system (cf. Elliot, 1987, and McPherson et al., 1987). The cause for the abrupt upper contact of these sequences could be related to three causes: (1) distributary lobe abandonment (e.g. Elliot, 1986), (2) eustatic sea level rise (Vail et al., 1977), or (3) increased subsidence in the proximal area of the basin, for instance a thrusting event (Jordan et al., in press).

Paleocurrents from trough crossbeds in the lower transitional member are unimodal with low-dispersion towards the west-southwest ($238° \pm 29°$, n=59) (Fig. 8), which parallels paleocurrents in underlying lower Bear Creek Group turbidites (Grotzinger and Gall, 1986). This mode indicates that paleoflow initially was restricted behind and was parallel to the Gordon Bay arch. From the Buff Quartzite Member upward, paleocurrents abruptly shift to the west-northwest ($294° \pm 27°$, n=115), roughly transverse to the arch, and they remain west-northwest-trending throughout most of Burnside sedimentation. This shift indicates that the Burnside fluvial system had aggraded to the point where sediments prograded across the arch and flowed uniformly to the northwest over the Slave craton. Trough crossbed paleocurrents at the base of the formation beyond the Southeast Belt are uniformly directed to the west-northwest or Northwest. This suggests that the Conglomerate Member (Fig. 7) is the probable time-equivalent of the basal Burnside sections in the Northwest Belt, because at Straight Gorge conglomerates lie 200 m above the basal contact, and pebbly pink subarkose lies within 20 m of the base at Link Lake (see Grotzinger et al., 1987, Figs. 23.8 and 23.10).

The middle members of the Burnside Formation in the Southeast Belt are characterized by a coarsening-upwards from the *Buff Qzt* to *Gm-St* lithofacies with the introduction of gravel. The composition of the gravels changes upwards from mostly vein quartz and quartzite clasts to metamorphic clasts to intrabasinal and igneous clasts. This change probably reflects both the intrinsic compositional sorting process caused by abrasion in a high-energy braided stream system which tends to destroy less durable clasts (Jordan et al., in press; Paola, in press), and the increasing contribution of clasts of the Burnside-type lithofacies and an igneous terrane (see below). Down-slope fining and sorting trends are indicated in the Conglomerate and Gravelly Sandstone members in the Tinney Hills (see below).

The top of the Conglomerate Member is marked by an abrupt shift to fine-grained Sh and St lithofacies which lack feldspar and gravel and which coarsen upwards to *gSt* and *Sp-St* lithofacies characteristic of the Gravelly Sandstone Member. This abrupt fining superimposed on an overall fining in the midst of a conglomeratic interval is suggestive of an external base level control, as shifts in the Burnside fluvial facies are otherwise characterized by gradual changes over tens to hundreds of metres of section. This fining could signal either a rapid proximal subsidence event in the source area or a distal sea level rise, either of which would cause backstepping of fluvial and deltaic facies belts (Jordan, 1981; Schedl and Witschko, 1984; Jordan et al., in press).

Two distinctive conglomeratic intervals occur at the top of the Gravelly Sandstone Member, one dominated by felsic volcanic and granitic clasts, and one with a highly polymictic composition with major proportion of intraformational Burnside-type lithofacies. These will be discussed below.

Above these upper conglomerates the section fines considerably and gravel disappears. This part of the section, the Orange Sandstone Member, is only exposed at the southern end of the Tinney Hills near Tinney Cove, where the section is measured in two overlapping sections. This member is probably the sandier proximal equivalent of the « Br » mapping unit of the Burnside Formation used by Campbell and Cecile (1976b) which they later named the Mara Formation (Campbell and Cecile, 1981). This inferrence is based on the much finer grain size and the distinctive colour of the unit. However, in the Tinney Cove area, Campbell and Cecile (1976b) mapped Burnside Formation continuously up to the Peacock Hills Formation and did not recognize the Mara or Quadyuk formations. Detailed work in the area shows that the Quadyuk Formation is developed directly below the Peacock Hills Formation, and that the Mara Formation may be present as a Burnside-type lithofacies equivalent (Orange Sandstone Member, Fig. 3,7).

The lower portion of the Orange Sandstone Member is dominated by an upward-decreasing scale of crossbedding from *Sp-St* to *St* lithofacies and the appearance of common mud chip on scours and trough crossbeds overturned foresets. These features resemble ones found in the Proterozoic Skadduvarri Formation (Bergh and Torske, 1986) interpreted as resedimented braid delta distributary mouth deposits.

The upper portion of the Orange Sandstone Member contains tabular *Sh* facies with minor coarsening-up sequences capped by *St* facies. Feldspar is abundant in this upper unit (12-25 %), and heavy mineral laminae are present. Within 150 m of the top, a sandy coarsening-upwards sequence is abruptly overlain by brown silty sandstone and siltstone that are composed of thin beds of *Sr* lithofacies. Over 25 m this coarsens up to silt-free *Sh* and minor *St*. This pattern is characteristic of the Mara Formation sections examined at Wilberforce Hills and Wolverine Canyon, except that distal northwestern sections contain less sand. The abrupt lower contact of this silty facies present across the Kilohigok Basin suggests that a eustatic rise in sea level may have been superimposed upon waning Burnside sediment supply.

The Burnside Formation in the southern Tinney Hills is capped by three thin (< 3 m) cycles of sandstone which pass up to dolomite which are interpreted as the lateral equivalent of the Quadyuk Formation in that they lie immediately below the Peacock Hills Formation. These units are cryptalgal laminites and low-relief stromatolite domes which are interpreted as peritidal facies. This contrasts with the thicker subtidal biohermal facies of the Quadyuk Formation found to the northwest along the west side of Bathurst Inlet (Campbell and Cecile, 1981). This suggests that the Quadyuk Formation interfingered with the waning Burnside-Mara deltaic system and passed southeastward from bioherms to tidal flat facies fringing the delta front. The persistance of fluvial sediment input and the interfingering relationship in the proximal Tinney Hills area is also suggested by the presence of siliclastic sand in the Peacock Hills and Kuuvik formations in that area, unknown in other parts of the basin (cf. Cecile and Campbell, 1978; Campbell and Cecile, 1981).

Buchan Bay

Approximately 100 km to the north of the Tinney Hills, just east of Buchan Bay, a thick (> 1 km), continuous, gently-dipping stratigraphic section of pink or white subarkosic quartzite outcrops that is lithologically identical to the Burnside Formation except that it rests directly on a pink syenogranite basement (Fig. 2). Fraser (1964) and Campbell and Cecile (1976a) correlate this quartzite with the Burnside Formation. A stratigraphic section was measured, and confirms the previous correlations (Fig. 9).

The upper 15-20 m of the syenogranite is weathered, and the basal contact with the subarkose is lined with a quartz pebble conglomerate interval <20 cm thick. The lithofacies patterns seen in this section are quite comparable to the section at north Tinney Hills, except that the Buchan Bay section is dominated by the *St* lithofacies from the base. Thick-bedded *Buff Qzt*, *gSt*, and minor *Gm-St* lithofacies occur upwards and Burnside clasts appear toward the top of the section. The lithofacies patterns mimic those in the northern Tinney Hills from the Buff Quartzite to the upper Gravelly Sandstone Member. This suggests that the Buchan Bay area was a positive topographic area until at least the deposition of the Buff Quartzite Member in the Tinney Hills; significant aggradation of the fluvial system had to occur before this area began to subside and was blanketed by fluvial siliciclastic sediments.

The significance of this section is two-fold. First, the absence of stratigraphically-lower units of the Bear Creek Group (Grotzinger et al., 1988) indicates that subsidence patterns were complex over the basin. Campbell and Cecile (1981) interpreted this as evidence for the eastern margin of Kilohigok Basin, which was inferred to trend south of the Buchan Bay area. We believe that these patterns are more easily explained in the context of northwest-thinning foreland basin stratigraphy, in which the Buchan Bay section occurs directly over the flexural arch. Evidence for this includes: (1) The stratigraphic cutout of the Kimerot Group, which defines the Gordon Bay arch, trends southwest (Grotzinger et al., 1987). The arch crest (Fig. 2), defined in the Hackett River area directly coincides with the Buchan Bay area when extrapolated along a co-parallel southwest trend, given the approximately 130 km of left-lateral slip on the Bathurst Fault zone defined by independent methods (Campbell and Cecile, 1981; Tirrul, 1985; Grotzinger and McCormick, in press). (2) This trend is parallel to paleocurrent trends in the Transitional Member of the Burnside Formation and lower Bear Creek turbidites (Grotzinger and Gall, 1986). (3) The arch trend is perpendicular to upper Burnside Formation paleocurrents throughout the basin. (4) The arch trend is perpendicular to finite extension lineations in thrust-nappes in the Bear Creek Hills Thrust-Fold Belt (Tirrul, 1985).

Such stratigraphic and structural patterns are more consistent with foreland basin settings. Complex patterns of basins and domes are common in foreland basins with interfering patterns of basement flexure due to discontinuous domains and times of overthrusting, as in the mid-Paleozoic foreland of the Appalachians (e.g. Quinlan and Beaumont, 1984) and the Cretaceous of the North America Cordillera (e.g. Jordan, 1981), and we interpret the Buchan Bay area as a possible dome (Fig. 2). This interpretation is supported by stratigraphic work in the sub-Burnside Bear Creek Group which shows that the area was positive during the deposition of those units (Grotzinger et al., 1988).

Northwest Belt

The Burnside outcrop belt west of the Bathurst Inlet is characterized by gently-dipping outcrops (< 10°), resulting in generally discontinuous and poor outcrops above the lower hundred

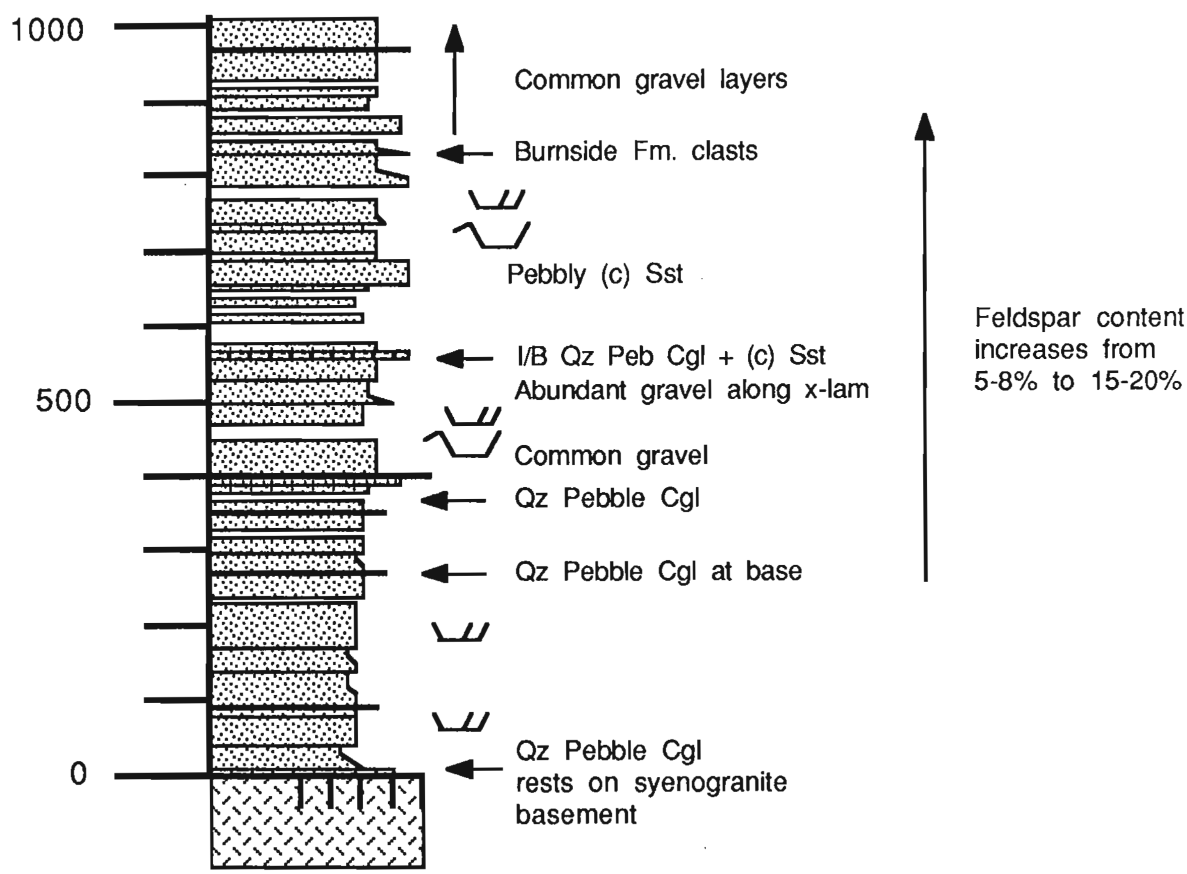

Figure 9. Stratigraphic section at Buchan Bay. Note that the section rests directly on granitic basement with no intervening lower Bear Creek or Kimerot groups. Although this section lacks a lower Transitional Member, lithofacies patterns are otherwise similar to those seen at northern Tinney Hills.

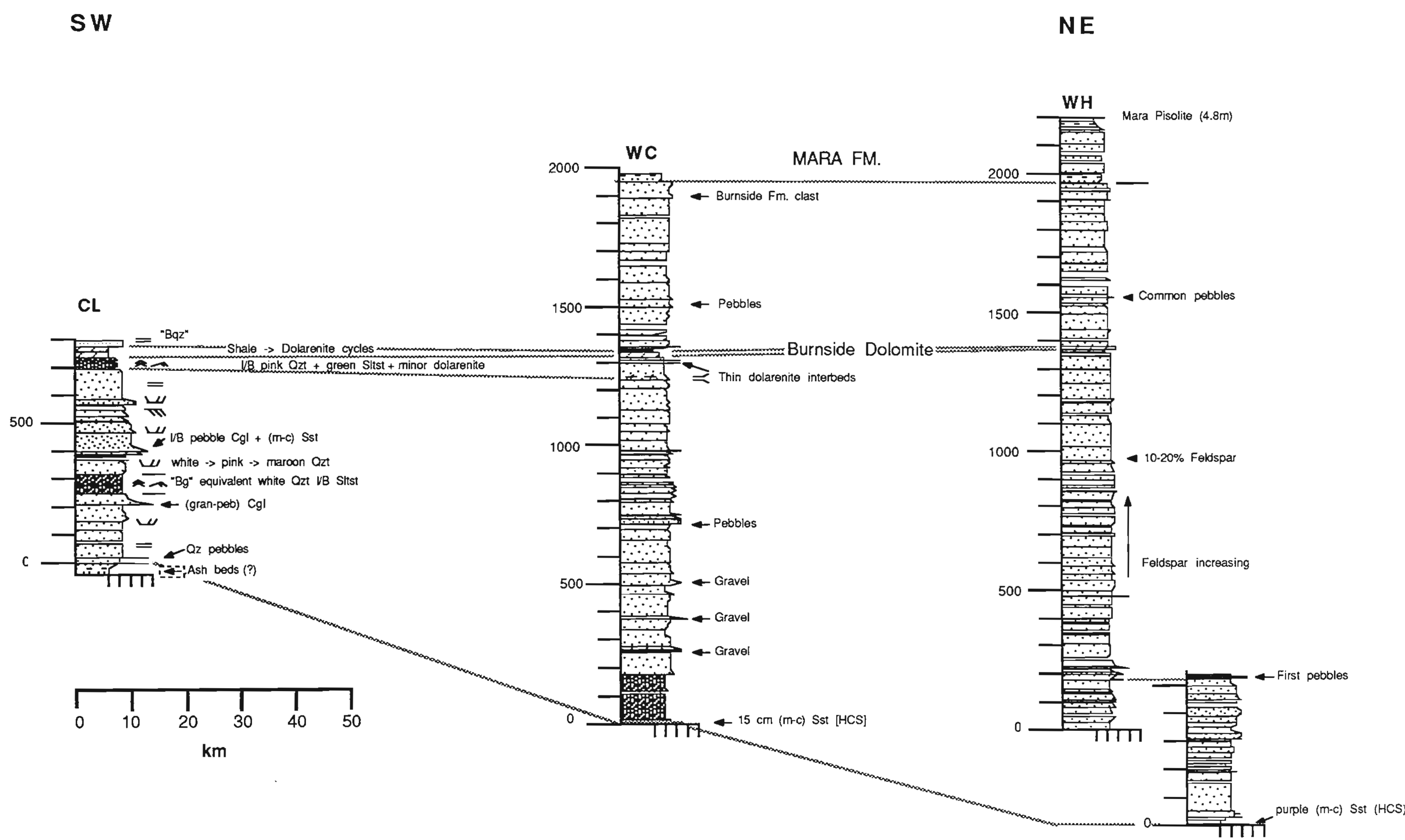

Figure 10. Cross-section of the Burnside Formation in the Northwest Belt from Wilberforce Hills to Contwoyto Lake. Note the presence of a distinctive dolomitic horizon (Burnside Dolomite Member) at comparable stratigraphic levels and its westward thickening.

metres of the basal contact. Consequently, viable stratigraphic sections are widely separated by many ten's of kilometres and individual members cannot be mapped as in the Southeast Belt. However, measured sections exhibit the general vertical grain size trends seen in the Southeast Belt, but lithofacies are finer to the northwest (Fig. 10). The most prominent facies is St, with subordinate *Sp-St* and *Sh*. The sections from Straight Gorge to Wolverine Canyon were described previously (Grotzinger et al., 1987). The sections show decreasing gravel abundance and clast size down paleoslope west-northwest, the progradational nature of the lower conglomeratic unit, thinning of the basal pink quartzite lithofacies over the Link Lake-Hackett River area, and thickening again to the northwest, and lateral facies changes. Quartz pebbles and pink medium-grained sandstones were noted by Tremblay (1967, 1968) at Contwoyto Lake and Bostock (1967) at Rockinghorse Lake. The presence of *St* and *Sp-St* lithofacies in all sections of the NW Belt indicate that fluvial facies spanned the Slave craton as far as Rockinghorse Lake. The presence of thick intervals of *Sr* and *Shcs* lithofacies in distal sections indicates that the marine and fluvial systems interfingered on the distal margin of the basin.

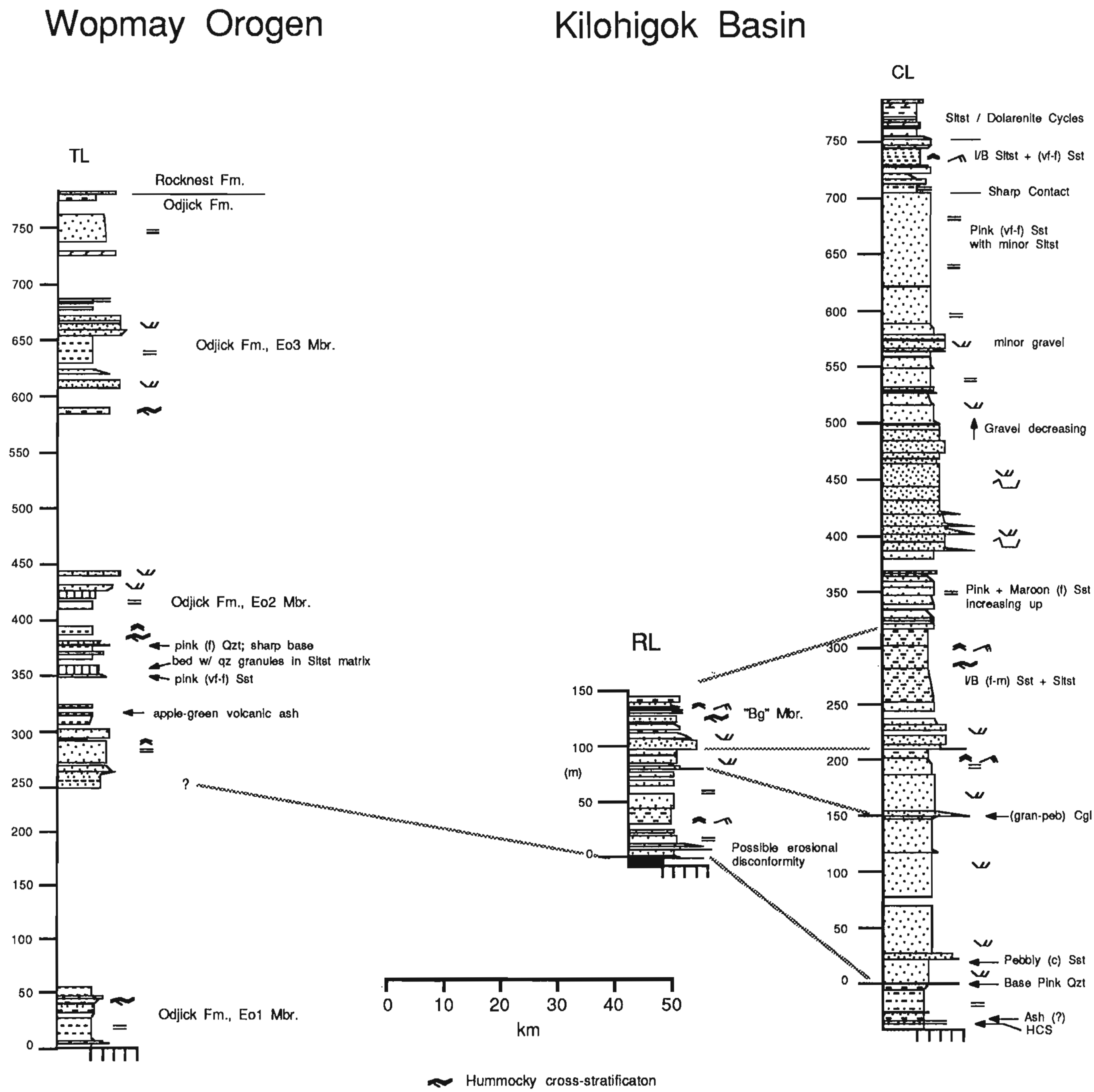

Figure 11. Cross-section of the Burnside Formation from Contwoyto to Rockinghorse lakes with a comparison to the Odjick Formation at Takiyuak Lake, about 100 km away. Note the presence of pink sand and quartz gravel in all sections. Note change in vertical scale from other cross-sections

EXTENT AND SIGNIFICANCE OF THE BURNSIDE DOLOMITE MEMBER

The Wolverine Canyon section contains a distinctive 21-m-thick dolomite-argillite interval that is dominated by clastic-textured dolomite and wave- and current- produced sedimentary structures which resemble the lagoonal facies of the Rocknest Formation (Grotzinger, 1986; Grotzinger, et al., 1987) (Fig. 10).

A thinner (13 m) dolomitic interval occurs at Wilberforce Hills at a similar stratigraphic level and they are believed to be correlative. This interval at Wilberforce Hills is characterized by thin interbeds of quartz sand of *Sh* lithofacies and laminated dolomite interbeds that have been tectonically flattened and sheared sufficiently to obscure most sedimentary structures. This unit is overlain by 800 m of crossbedded quartz arenites before encountering the Mara Formation pisolite horizon, the true contact with the Peacock Hills Formation.

At Contwoyto Lake, the Burnside section is incomplete, but based on comparison to the section at Wolverine Canyon, the thick (ca. 157 m) dolomitic interval at the top of the section is correlated with the Burnside Dolomite Member, as suggested by Hoffman et al. (1984) and Grotzinger (1985) (Fig. 11). Tremblay (1967, 1968) originally assigned this interval to the Peacock Hills and Kuuvik formations. Campbell and Cecile (1976b, 1981) maintained this correlation by defining the units as unique members of the Peacock Hills and Kuuvik formations (''P5'' and ''K4'' members, respectively). However, it is suspected that this is incorrect based on the following. First, the lower silty siliciclastic portion of this interval (ca. 25 m) contains distinctive tabular pink sandstone beds which are identical to the Burnside *Sh* lithofacies. The Peacock Hills Formation lacks sand except in the southeastern proximal part of the basin (southern Tinney Hills), where it contains a few thin very fine to fine sand beds. Cecile (1976) does not note the presence of sand beds, and inspection of the section at Wolverine Canyon, Kuuvik Lakes, and Wilberforce Hills show it to be exclusively siltstone or finer. Based on the distal locality of this section the Peacock Hills Formation would be expected to consist of even finer facies.

Secondly, the carbonate portion of this interval is characterized by cyclic alternation of laminated siliciclastic siltstone or mudstone passing up to clastic-textured dolomite with wave- and current-ripples, minor syneresis cracks, and abrupt upper contacts. This resembles the lagoonal facies of the Rocknest Formation (Grotzinger, 1986) rather than any lithofacies of the Kuuvik Formation as described by Cecile (1976) or Cecile and Campbell (1978).

Thirdly, this carbonate interval is overlain by pink and white tabular bedded quartzite. Campbell and Cecile (1976b) name this the ''Bqz'' member of the Brown Sound Formation, which only appears on their map (Campbell and Cecile 1976b) and is not discussed in any report. This lithofacies resembles the *Sh* lithofacies of the Burnside Formation ex-

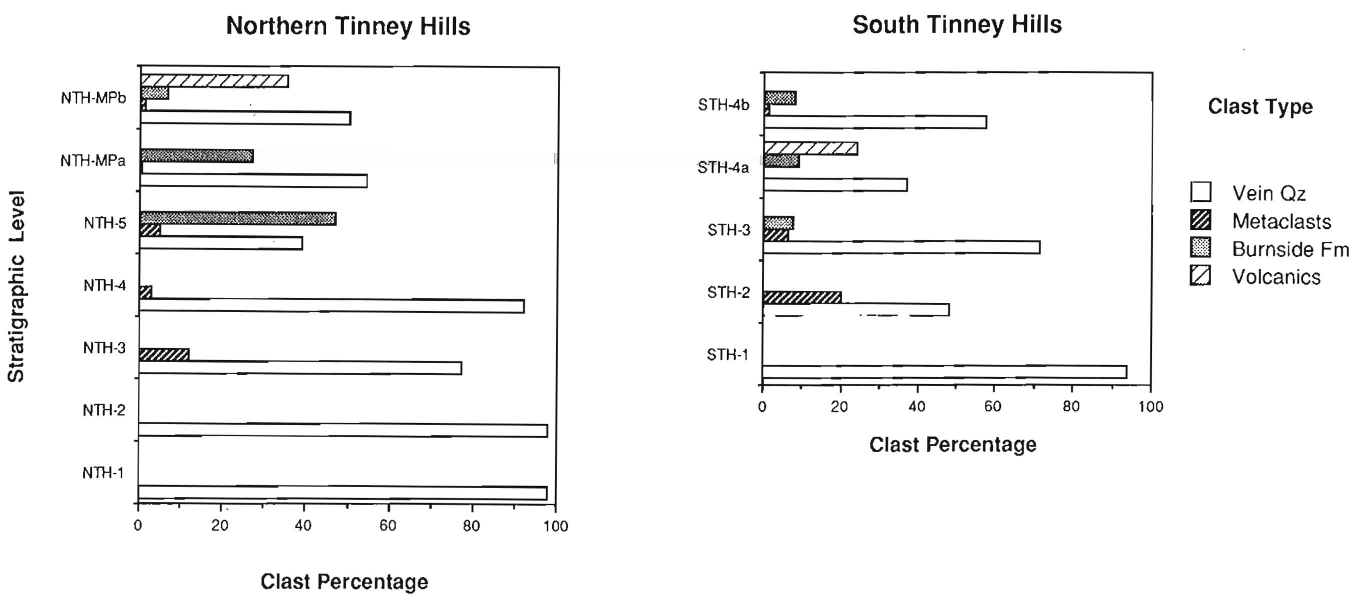

Figure 12. Vertical trends in major conglomerate clast types in the Tinney Hills with appearence of distinctive clast at certain horizons. Metaclasts refer to metagreywacke and other metamorphic clasts.

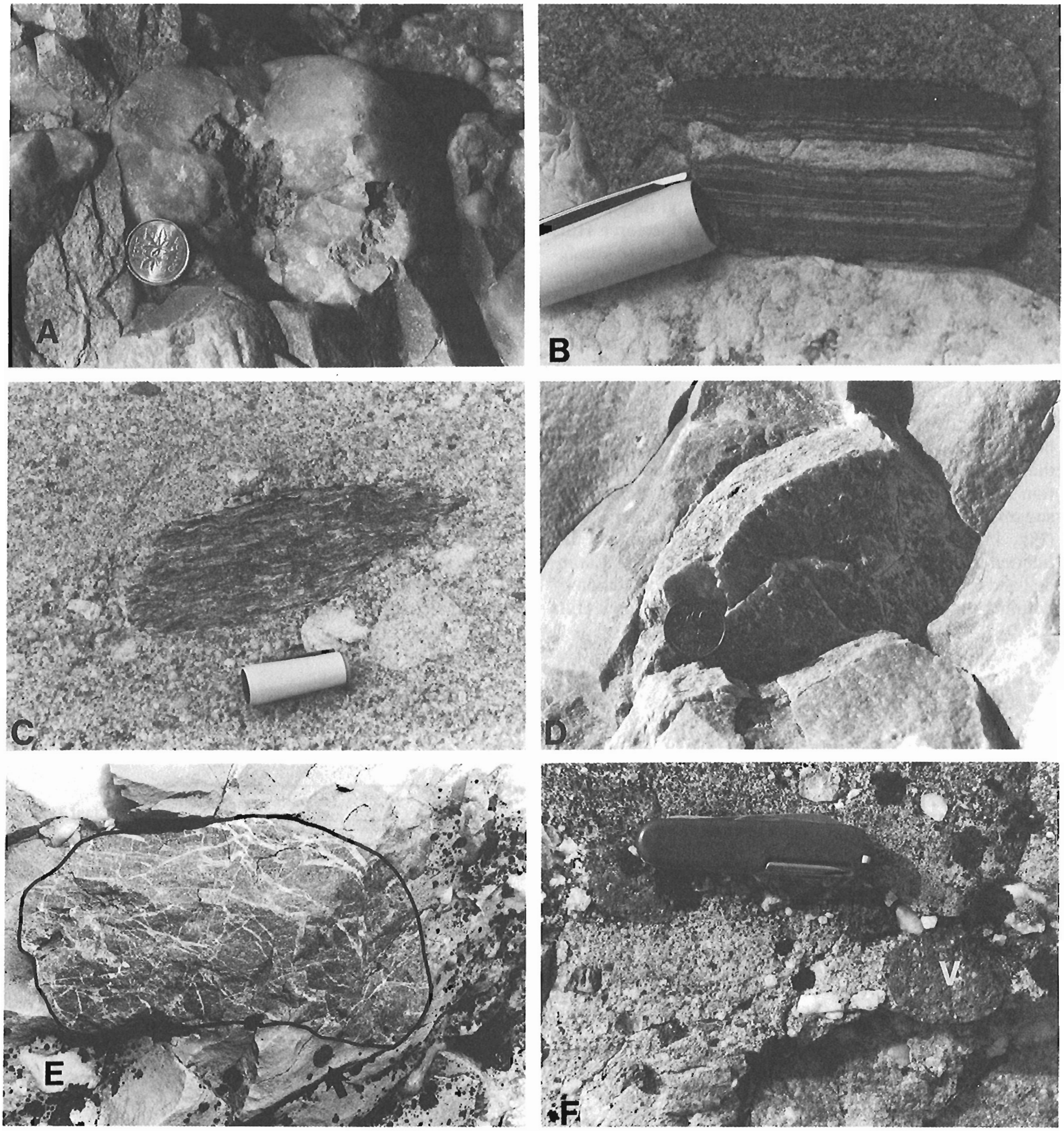

Figure 13. Representative conglomerate clast lithologies found in the Tinney Hills. A) Vein quartz clast, the most common lithology. This clast contains K-feldspar grains and is probably derived from pegmatite. Coin for scale. B) Metagreywacke clast, presumably derived from Archean basement. Pen cap for scale. C) Schist clast. Pen cap for scale. D) Foliated quartzite clast. Note intersections lineation. Coin for scale. E) Fractured Burnside-type clast. Vein quartz fracture filling which does not extent into the matrix sand. This clast must have been fractured and filled at depth before being re-incorperated into the Burnside fluvial system. Cannibalization of syn-orogenic sediments is common in foreland basins. Coin for scale (arrow). F) Undeformed felsic volcanic clast (V). Knife is about 10 cm long.

cept for its white colour. The quartzites appear to be a well-sorted marine lithofacies that is similar to previously unrecognized white quartzite beds in the Burnside Formation just below their ''P4 Member''. Thus the ''P4/K5'' members are sandwiched between well-sorted, clean, probably marine lithofacies, in which the overall sequence is compatible with a wedge of marine siliciclastic/carbonate facies in the dominantly fluvial Burnside Formation.

Finally, white, tabular-bedded quartzites are not present in lower members of the Brown Sound Formation. Everywhere else in the basin the basal Brown Sound Formation contains cm- to dm-scale fining up beds with red mudstone caps which commonly contain mud cracks and minor halite casts (Campbell and Cecile, 1981).

All of these features at Contwoyto Lake are comparable to the Burnside Dolomite Member seen at Wolverine Canyon, and are inconsistent with features documented in the Peacock Hills, Kuuvik, and Brown Sound formations seen at other localites. Campbell and Cecile (1976b, 1981) erected new members of each formation to account for the observed lithofacies. The new results discussed above suggest that these lithofacies are most consistent with a westward-thickening interval of the Burnside Dolomite Member, which passes laterally into a tongue of the Rocknest Formation carbonate platform. A consequence of this revised correlation is that the influence of the ''Rockinghorse Arch'' during sedimentation is substantially diminished, and that subsidence may have been more continuous in this area than was previously thought (cf. Campbell and Cecile, 1981).

The presence of the Burnside Dolomite Member in all three distal sections provides a crude marker of the limit of marine transgression during and indicates that the paleoshoreline trended roughly southwest, perpendicular to fluvial paleocurrent trends and parallel to the Gordon Bay arch.

CORRELATIONS WITH THE EPWORTH GROUP

Several aspects of the distal sections in the Contwoyto and Rockinghorse Lake areas suggest that the Burnside Formation is correlative with parts of the Odjick and Rocknest formations in Wopmay Orogen (Fig. 11). First, paleocurrent trends in fluvial lithofacies consistently are west-northwest-directed, and the most distal section is less than 50 km from the autocthon of Wopmay Orogen (Fig. 1). Secondly, marine shelf siliciclastic lithofacies (*Sr* and *Shcs*) are common in the distal part of the Kilohigok Basin and are identical to those seen by us in the Odjick Formation at Takiyuak Lake. These lithofacies correspond to the ''Bg Member'' of the Burnside Formation of Campbell and Cecile (1976b) which they map only at Rockinghorse Lake, but are identical to the middle part of the Contwoyto Lake section (Fig. 11). In addition, pink subarkosic sand and minor quartz pebbles, which are identical to the basal Burnside Formation at Rockinghorse Lake, are seen in the midst of the predominantly white marine sand lithofacies of the Eo2 Member of the Odjick Formation. Furthermore, the white quartzite intervals in the Burnside Formation at Contwoyto Lake, including the ''Bqz Member'' of Campell and Cecile (1976b), are identical to quartzites in the Eo2 Member. This indicates that regional interfingering of the ''Burnside fluvial'' and « Epworth marine » facies occurred in the area between Takiyuak and Contwoyto lakes.

Finally, the Burnside Dolomite Member bears a very strong resemblance to the lagoonal facies of the Rocknest Formation, as discussed above. The main difference between the Burnside Dolomite and shale-based cycles of the Rocknest inner-shelf is that the base of Burnside Dolomite cycles contain more siliciclastic silt and sand, consistent with the facies trends documented in the Rocknest Formation (Grotzinger, 1986). Grotzinger (1985) argues that the constant influx of siliciclastic debris into the Rocknest lagoon explains the relatively narrow width (ca. 150 km) of the Rocknest platform when compared to other major passive-margin carbonate platforms (> 250 km). Grotzinger (1985) suggested that Burnside Dolomite at Contwoyto was correlative with the lower non-cyclic member of the Rocknest Formation which he attributed to a eustatic rise in sea level, causing back-stepping of lagoonal facies to the east. This would account for the abrupt fining and appearance of carbonate and white texturally-mature quartzite in the Burnside system.

The co-existence of the Burnside fluvial system and the Epworth passive margin explains the presence of relatively coarse siliciclastic detritus to the Epworth marine shelf. A volcanic ash bed was discovered in the Odjick Formation at Takiyuak Lake and just below the base of the Burnside Formation at Contwoyto Lake; zircon dating may test this correlation.

CONCLUSIONS

Stratigraphic and sedimentologic study of the Burnside Formation indicate that it represents a westward-prograding braid-delta/braid-plain system which blanketed the northwestern Slave Province and interfingered with the passive margin of the Epworth Group of Wopmay Orogen. The paleoshoreline probably faced northwest, and the paleoclimate was probably humid. The laterally-persistent, westward-thinning lithofacies patterns, restriction of conglomerates largely to the Southeast Belt, west-northwest directed paleocurrent trends, and upward occurrence of Burnside intraformational clasts are most consistent with deposition in an asymmetrically-subsiding foreland basin that drained the Thelon Tectonic Zone and possibly the Queen Maud Block.

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Gold and arsenic in till, Wheatcroft Lake dispersal train, Manitoba[1]

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Kaszycki, C.A., Suttner, W., and DiLabio, R.N.W., Gold and arsenic in till, Wheatcroft Lake dispersal train, Manitoba; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 341-351, 1988.

Abstract

In 1984, a zone of high As values in till was outlined south of Wheatcroft Lake. This anomaly was based on 9 samples taken over an area of approximately 600 km^2. In 1986, 114 additional till samples were collected, to define the anomaly more clearly, as well as to delineate possible source areas and lithologies.

Well defined As and Au distribution patterns have been established. Arsenic-enrichment in till (20ppm) occurs within a well defined northeast-southwest-trending zone extending approximately 60 km southwest of Wheatcroft Lake. Distribution patterns, in part, reflect glacial dispersal along known ice flow directions. Trace element concentrations as high as 227 ppm As and 69 ppb Au have been recorded. No single source area or lithology has been identified, and the anomaly may have multiple sources. It is possible, however, that Au mineralization is associated with the margins of post-Sickle intrusions, and with sulphide mineralization in the Burntwood River Metamorphic Suite.

Résumé

En 1984, une zone de valeurs élevées en As a été délimitée au sud du lac Wheatcroft. Cette anomalie se fonde sur neuf échantillons prélevés dans une zone d'environ 600 km^2. En 1986, on a recueilli 114 échantillons de till supplémentaires pour définir plus clairement l'anomalie et pour délimiter les zones et les lithologies d'origine possibles.

Des configurations de répartition bien définies des éléments As et Au ont été établies. Le till riche en arsenic (>20 ppm) a été localisé dans une zone bien définie à direction nord-est située à environ 60 km au sud-ouest du lac Wheatcroft. Les configurations de répartition réflètent en partie la dispersion des dépôts glaciaires le long des directions connues d'écoulement des glaces. Des concentrations d'éléments en traces aussi élevées que 227 ppm pour l'arsénic et 69 ppm pour l'or ont été enregistrées. On n'a identifié aucune zone ou lithologie originelle de sorte qu'il se peut que l'anomalie provienne d'origines multiples. Il se peut, cependant, que la minéralisation aurifère soit associée aux bordures des intrusions postérieures au groupe de Sickle et à la minéralisation de sulfures dans la série métamorphique de Burntwood River.

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INTRODUCTION

Since 1983, a program of systematic till sampling and compositional studies has been carried out in northwestern Manitoba, as part of the Canada-Manitoba Mineral Development Agreement. One of the primary objectives of this project is to define the lithological and geochemical components of till that reflect areas of known mineralization, and bedrock units favourable for mineralization, as an aid to mineral exploration within the region.

In 1984, till compositional studies in the Granville Lake region (NTS 64C) defined an area of 600 km^2, characterized by anomalous As concentration (>20 ppm) at nine sample sites south of Granville Lake (Kaszycki and DiLabio, 1986a) (Fig. 1). At this scale, sampling density was insufficient to determine whether the anomaly was formed by glacial dispersal from a zone of sulphide mineralization along the south shore of Granville Lake, or whether it, in part, represented a local increase in background As concentration associated with unmapped massive and disseminated sulphide mineralization within the area (Kaszycki and DiLabio, 1986a, b). In 1986, an additional 114 samples were collected within the Wheatcroft Lake area, in an effort to define the shape and internal composition of the anomaly, as well as to delineate possible source areas and lithologies related to high As concentrations. In addition, regional sampling was carried out over map area NTS 63N, south of Wheatcroft Lake, in part, to delineate the southern extension of the regional As anomaly. Preliminary results of the follow-up study are summarized in this paper.

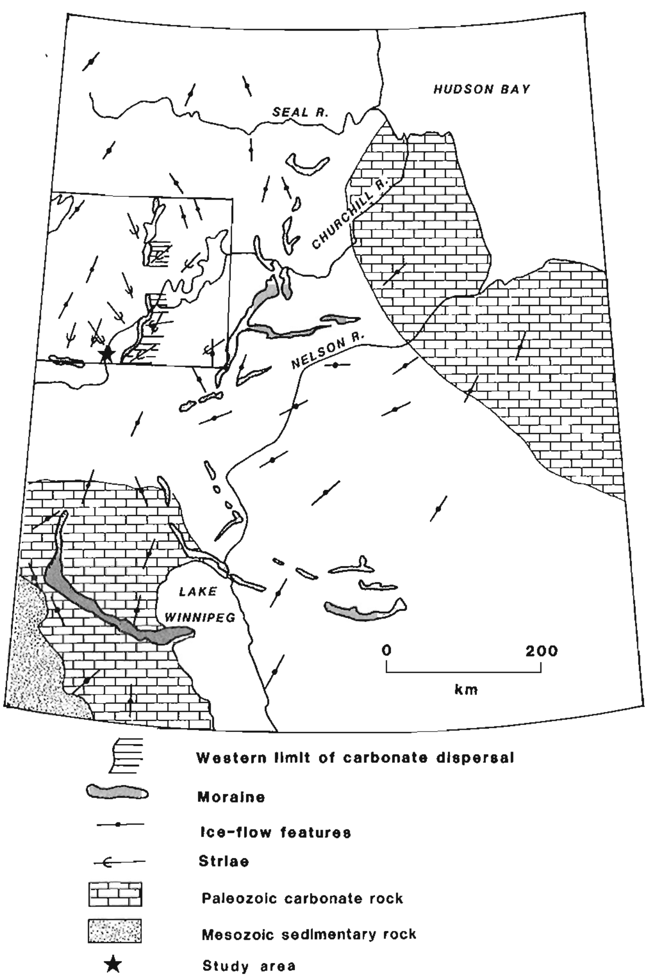


Figure 1. Major glacial landforms of northern Manitoba (taken in part from Klassen, 1983). Area for which detailed surficial mapping has been completed is outlined. Striae represent general trends within the region (after Kaszycki and DiLabio, 1986a).

REGIONAL SETTING

The study area is located within the Churchill Structural Province of the Canadian Shield (Fig. 1). Regional till sampling and surficial geological mapping at a scale of 1:100 000 have been completed for four map areas (NTS 64B, C, F & G) (Fig. 1). Regional till sampling has also been completed for map area NTS 63N, south of the area shown on Figure 1.

Shield terrane in this region is flanked to the east-northeast and to the south-southwest by Paleozoic carbonate rocks. Throughout the extreme western part of the province ice-flow was from a Keewatin ice centre, at approximately 190-210°, and till is derived from crystalline shield lithologies. In the eastern part of the province ice-flow was from a more easterly source, with striae directions ranging from 225-260°. Till in this area contains abundant Paleozoic carbonate erratics indicating ice-flow out of Hudson Bay (Kaszycki and DiLabio, 1986a).

Inundation by glacial Lake Agassiz, during ice retreat, resulted in deposition of calcareous laminated silt and clay throughout the central and eastern parts of Manitoba, as well as along major river valleys extending from inflow channels draining into the lake from the west.

SURFICIAL GEOLOGY

Morphological evidence, coupled with clast composition of tills within the region, suggest that a zone of convergence between Keewatin lobe ice and Hudson lobe ice existed in the Leaf Rapids area (Fig. 2). This zone is marked by a discontinuous north-south trending esker-like ridge termed the Leaf Rapids interlobate moraine. Striae to the east of the moraine record a clockwise rotation of ice flow from approximately 190° (oldest) to approxiately 260° (youngest). Till in this area was deposited by ice of Hudson Bay provenance and contains abundant exotic Paleozoic carbonate. To the west of the moraine, till was deposited by ice of Keewatin provenance and is derived predominantly from crystalline shield lithologies. The earliest ice flow recorded in this area is approximately 190 to 210°. The youngest striae to the west of the ridge record a late glacial shift of ice flow to the southeast at 165°, indicating a late glacial zone of convergent ice-flow existed along the Leaf Rapids interlobate moraine. Eskers within this region also exhibit a convergent pattern in the vicinity of the Leaf Rapids interlobate moraine (Fig. 2).

In the Wheatcroft Lake area, which is located approximately 20km to the west of the Leaf Rapids interlobate moraine, till is composed primarily of local shield lithologies. The oldest striae in this area indicate ice flow to the southwest at approximately 190 to 210°. However, the youngest striae, trending to the southeast at approximately 165°, are the dominant ice-flow indicators in the area.

Bedrock exposure is limited, except in the southwest corner of the study area. Till cover is thin and discontinuous, and bedrock was intersected at a depth of 45 cm or less at over 25 % of sample sites. During ice retreat, inundation by glacial Lake Agassiz resulted in deposition of laminated silt and clay as a discontinuous blanket over till and/or bedrock. Particularly extensive lacustrine deposits, up to several metres thick, occur in areas of low relief flanking the Churchill River. Where Lake Agassiz clay and/or organic deposits exceed 1 m in thickness, till is inaccessible in hand-dug pits, providing a major obstacle to till sampling in this area.

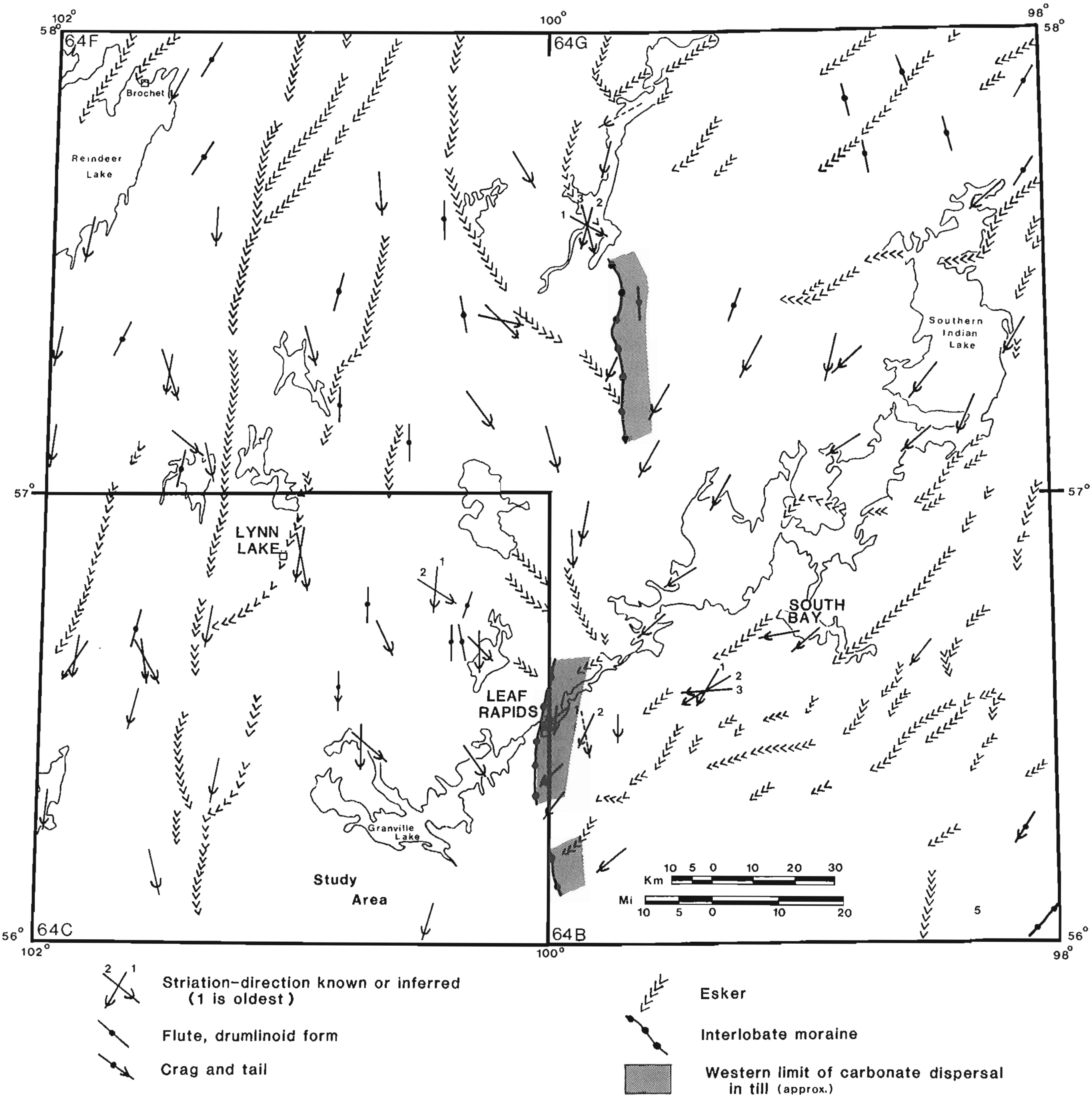


Figure 2. Generalized surficial geology of the Lynn Lake — Leaf Rapids region. Wheatcroft Lake area shaded (after Kaszycki and DiLabio, 1986a).

BEDROCK GEOLOGY

In the Wheatcroft Lake area, the major lithostratigraphic units, from oldest to youngest, include: metagreywackes and derived gneisses and migmatites of the Burntwood River Metamorphic Suite (BRMS); meta-arkoses, meta--conglomerates and derived gneisses and migmatites of the Sickle Group; and mafic to felsic post-Sickle intrusives (Fig. 3). In the upper part of the BRMS (the BRMS transition zone), metagreywacke gives way to amphibolite, calc-silicate and quartz lithologies, correlated with metasedimentary and metavolcanic rocks of the Wasekwan Group, that form the Lynn Lake greenstone belt to the north (Baldwin, 1980). Because local relief rarely exceeds 30 m, much of the terrain is swamp covered, and in these areas bedrock geology is poorly known.

The region has been classified as having high to moderate potential for massive sulphide mineralization (Gale et al.,1980). Disseminated and minor massive sulphide mineralization (mainly pyrite) occurs as thin stratiform deposits within the upper part of the BRMS (Wasekwan equivalent) and within the lower 200 m of the Sickle Group, which conformably overlies the BRMS in this region (Baldwin, 1980). Several sulphide occurrences located on the south shore of the Laurie River (Fig. 3), between Granville Lake and Trophy Lake, include lenses of massive sphalerite, pyrite, chalcopyrite and galena (Barry, 1965). Disseminated pyrrhotite has also been found in amphibolite at the top of the BRMS (Barry, 1965; Pollack, 1966), and may occur along the entire length of the unit. In 1950, gold was discovered north of Wheatcroft Lake. At this site, a metasediment containing disseminated arsenopyrite was sampled over a zone several metres wide and returned gold values of up to 0.15 oz/ton (Barry, 1965).

TILL GEOCHEMISTRY

A total of 114 till samples were collected in the Wheatcroft Lake area, at 106 sites evenly distributed over an area of approximately 1200 km^2. Sampling was helicopter-supported except for an east-west trending traverse along a power line crossing the central part of the area. Sample holes were hand-dug to a depth of between 15 and 120 cm, with an average depth of about 40-50 cm. Where possible, samples were collected from below the postglacial solum to minimize the effects of post-depositional weathering on till geochemistry (Shilts, 1984; Rencz and Shilts, 1980; Nikkarinen et al., 1984).

In this study, trace element composition was determined for several size fractions in each sample. The clay fraction ($<2\mu$m) was analyzed for Ag, Cd, Co, Cr, Cu, Fe, Mn, Mo, Ni, Pb, and Zn using standard atomic absorption techniques after a hot (HNO_3-HCl) acid leach. Arsenic was analyzed by colourimetry and uranium by delayed neutron activation. Concentrations of most trace elements are greater in this fraction than in coarser size fractions due to primary enrichment of metals within the structure of clay-sized minerals (phyllosilicates), and adsorption of trace elements released by weathering of labile minerals (Shilts, 1984).

The silt plus clay size fraction ($<63\mu$m) was analyzed for Au, As, Cd, Co, Cr, Fe, Mo, Ni, U, and Zn by neutron activation. Because the silt plus clay fraction commonly contains variable amounts of inert quartz and feldspar, the concentration of most trace elements is much lower in this size range than within the clay fraction ($<2\mu$m) (Table 1). It has been demonstrated, however, that Au is preferentially enriched in the fine fractions of oxidized till (DiLabio, 1985), and because use of the $<63\mu$m size fraction enables recovery of a sample large enough for Au analysis (>10gr), this size fraction was analyzed for Au plus other trace elements listed above. The heavy mineral fraction is also scheduled for analysis of Au and related elements.

Summary statistics comparing till geochemical data for the Wheatcroft Lake area (NTS 64C/2) and the Granville Lake-Lynn Lake region (NTS 64C), including the Lynn Lake Greenstone Belt (Kaszycki and DiLabio, 1986b), are presented in Table 2. In the Wheatcroft Lake area, average and anomalous values for all trace elements, except Fe and Mn, are significantly higher than regional values. For example, over 65 % of the samples in the Wheatcroft Lake study contain As in concentrations in excess of regional background (15ppm), and the 90th and 95th percentile values for As are approximately six to seven times higher than regional values. Similar but less pronounced relationships are observed for Au, Cu, Zn, Ni, and Cr, in the northcentral part of the area.

BEDROCK GEOCHEMISTRY

In addition to till , 43 samples of bedrock were analyzed for the same suite of trace element, using standard atomic absorption and neutron activation techniques. Samples were selected to represent the range of bedrock lithologies in the area, particularly within the mineralized belt south of Granville Lake (Fig. 3). Some samples were collected also from areas of known mineralization.

Table 1. Average trace element concentration in till and bedrock samples. Concentrations below detection limit assigned a value of 1/2 detection limit. For this reason some averages are approximate only. Heavily mineralized bedrock samples (#11 & #12) were not included in averaging.

	Till		Bedrock	
Element	$<2\mu$m (AA)	$<63\mu$m (NA)	AA	NA
Au ppb	-	6.3	-	10.2
As ppb	44	10.5	15.7	15.7
Cu ppm	76	-	51.6	-
Zn ppm	134	68.4*	93.1	<200
Pb ppm	14.6	-	10.8	-
Ni ppm	51.5	19.1*	36.2	50*
Cr ppm	107	77.8*	48.0	82*

*Value approximated.

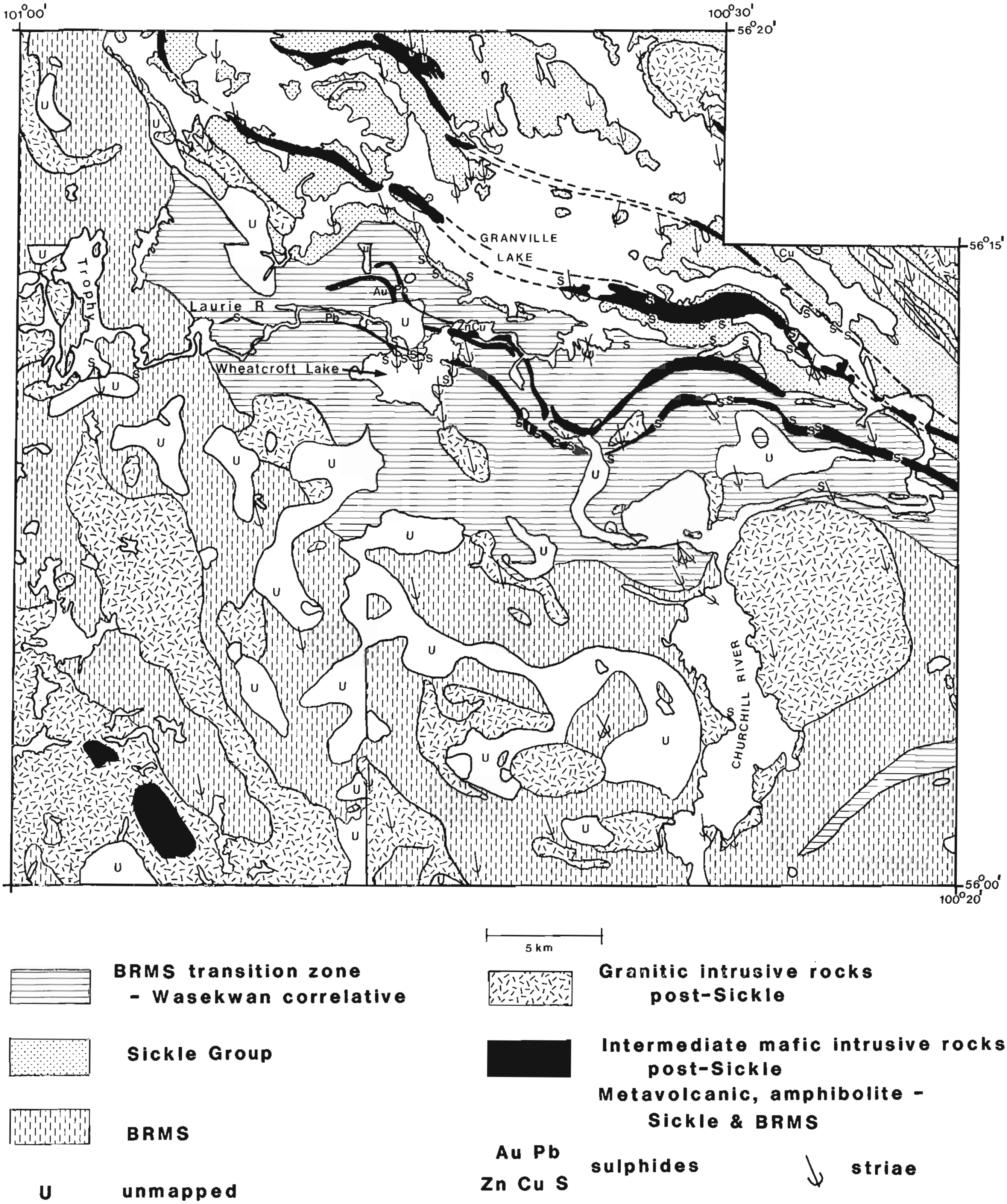


Figure 3. Bedrock geology simplified from Barry (1965), Barry and Gait (1966), Cranstone (1968), Godard (1966), and Pollack (1966). Unmapped areas are extensively swamp or drift covered. Lithostratigraphic units have been interpreted following Zwanzig (1980) and Baldwin (1980).

Average trace element concentration in bedrock samples are included in Table 1. Trace element concentrations in bedrock samples are significantly less than concentrations in the clay size-fraction of till. This reflects the dilution of metal-bearing phyllosilicates and sulphides by inert rock forming minerals (mainly quartz and feldspar). Average trace element concentrations in the $<63\mu m$ size fraction are generally less than average concentrations in bedrock samples, possibly reflecting dilution of till by nonmineralized lithologies, and/or exclusion of metal-bearing lithic fragments (usually $>63\mu m$ in diameter) in analysis of the $<63\mu m$ size fraction of till.

Trace element concentrations for bedrock samples considered to be anomalous with respect to As and/or Au are presented in Table 3. Of these, only sample 11 contains significant concentrations of Au and As, or any other trace element. This sample, containing .06 oz/ton Au, was collected at the Cu-Zn showing on Wheatcroft Lake (Fig. 3),

Table 2. Summary statistics for till geochemical data in the Wheatcroft Lake area (NTS 64C/2) and Granville Lake region (NTS 64C).

Element	Minimum		Maximum		Arithmetic mean		Geometric mean		90th percentile		95th percentile		n	
	64C/2	64C	64C/2	64C	4C/2	64C	4C/2	64C	4C/2	64C	4C/2	64C	4C/2	64C
Au ppb(1)	1	1	70	92	6.3	4.2	4.0	2.3	10	7	19	11	134	354
As ppm(1)	.9	.2	41	17.0	10.5	1.92	7.3	1.25	22	3.6	29	5.4	134	354
As ppm(2)	1	1	227	110	44	8.1	24.2	5.1	110	15	136	24	136	374
Cu ppm(2)	13	13	440	202	76	61.6	62	51	129	109	187	134	136	374
Zn ppm(2)	49	17	338	1350	134	96.9	127	83.5	108	166	231	190	136	374
Pb ppm	3	3	41	57	14.6	12.1	13.7	11.4	20	18	22	20	136	374
Cr ppm(2)	39	20	230	253	107	77.3	102.5	72	148	108	162	123	136	374
Ni ppm(2)	16	7	99	446	51.5	37.2	49	33	70	56	78	65	136	374
Mn ppm	134	50	1606	1467	395	433	355	381	606	730	889	805	136	374
Fe %	1.2	1.1	9.5	9.5	4.2	4.0	4.1	3.8	5.3	5.7	6.0	6.1	136	374

(1) analysis by neutron activation on <63 μm size fraction of till
(2) elements for which local background concentration (64C/2) greatly exceeds regional values (64C)

Figure 3. Bedrock geology simplified from Barry (1965), Barry and Gait (1966), Cranstone (1968), Godard (1966), and Pollack (1966). Unmapped areas are extensively swamp or drift covered. Lithostratigraphic units have been interpreted following Zwanzig (1980) and Baldwin (1980).

Sample #	1	5	7	9	11	12	16	17	24	38	39
Rock type	MS	A	A(r)	A	A(r)	A(?)	M	Gr	M	A(r)	Gt
Au ppb	150	<5	24	<5	1860	8	41	<5	<5	18	5
As ppm	20	34	45	48	8760	356	7	43	67	72	54
Fe %	13	9.4	11.0	7.2	26.0	5.1	5.2	<.5	5.0	8.8	4.2
Co ppm	39	20	31	55	35	20	17	<10	16	<10	13
Ni ppm	*10*	*<2*	*<2*	*74*	*14*	*33*	*66*	*2*	*28*	*<2*	*37*
Cr ppm	*3*	*2*	*3*	*160*	*6*	*130*	*150*	*4*	*120*	*2*	*190*
Cu ppm	*24 10*	*12*	*56*	*11,600*	60	*14*	*4*	*84*	*12*	*24*	
Zn ppm	28078	*50*	*50*	<30,000	880	*150*	*26*	*96*	*24*	*88*	
Ag ppm	*<.1.1*	*<.1*	*<.1*	210	*2.3*	*0.6*	*<.1*	*<.1*	*<.1*	*<.1*	
Cd ppm	*<.2<.2*	*<.2*	*<.2*	500	*1.8*	*<.2*	*<.2*	*<.2*	*<.2*	*<.2*	
Pb ppm	*32*	*6*	*4*	*7*	*18,100*	*176*	*12*	*6*	*6*	*7*	*4*

Italic indicates analyses by atomic absorption after a hot (HNO_3-HCl) leach; all other analyses by neutron activation.

where Barry (1965) reported 0.03 oz/ton Au, also from a grab sample. Elevated background concentrations of Ni and Cr in till are not observed in heavily mineralized bedrock (e.g. sample 11), but do occur within some amphibolites and metasediments associated with mineralized zones (samples 9, 12, 16, 28, and 37). This suggests that the mineralized belt south of Granville Lake may be a source for elevated Ni and Cr concentrations in till, as well as other trace elements commonly associated with sulphide mineralization (Cu, Zn, Pb). In general, till geochemistry reflects bedrock geochemistry within the mineralized belt south of Granville Lake, being enhanced in the $<2\mu m$ size fraction of till and slightly subdued in the $<63\mu m$ size fraction.

REGIONAL DISTRIBUTION OF AU AND AS IN TILL

Arsenic

The regional As anomaly in the $<2\mu m$ size fraction of till is shown in Figure 4. Arsenic data have been compiled from regional sampling in both the Granville Lake (NTS 64C) and Kississing (NTS 63N) map areas, as well as detailed sampling in the Wheatcroft Lake area. In general, As-enrichment in till (>20 ppm) occurs within a well defined northeast-southwest trending zone, extending approximately 60km southwest of the mineralized belt at Wheatcroft Lake, and encompassing an area of about 1150 km^2.

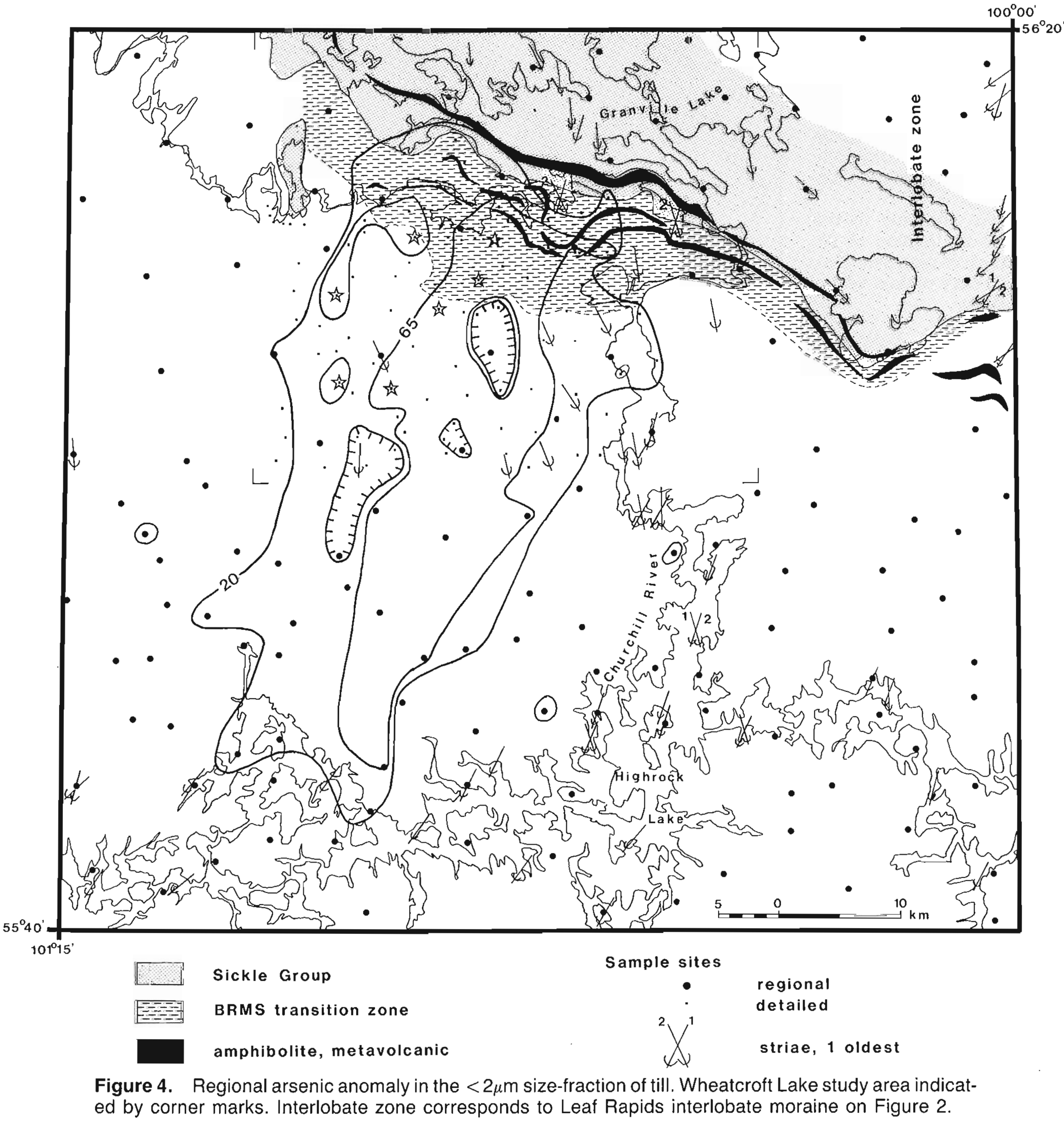
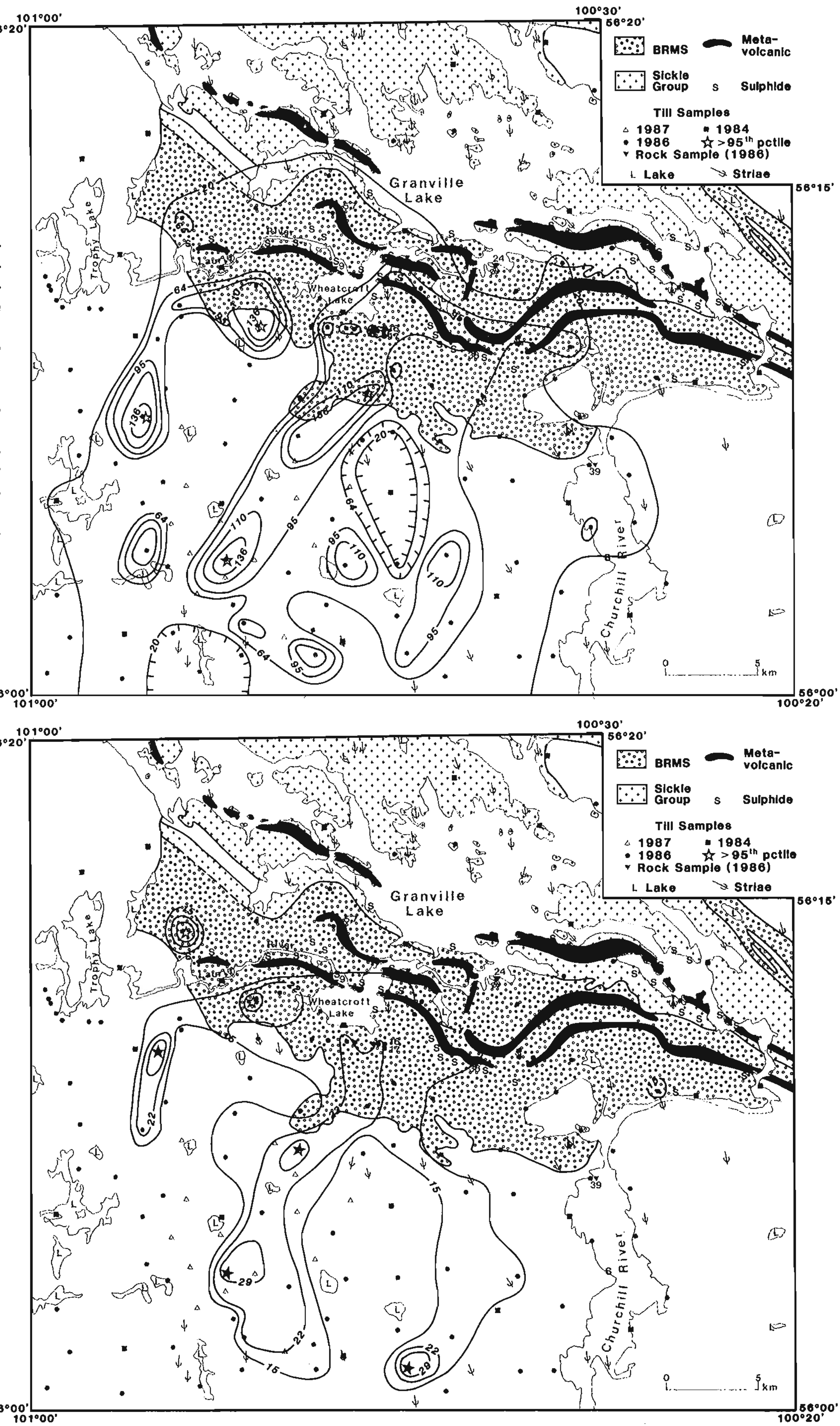


Figure 4. Regional arsenic anomaly in the $<2\mu m$ size-fraction of till. Wheatcroft Lake study area indicated by corner marks. Interlobate zone corresponds to Leaf Rapids interlobate moraine on Figure 2.

Figure 5. Detail of the northern half of the arsenic dispersal train superimposed on outcrop area of mineralized Sickle Group and BRMS lithologies.
a Arsenic anomaly in the $<2\mu m$ size fraction of till. Contour values represent 75th, 85th, 90th, and 95th percentile values respectively. 20 ppm contour represents the regional anomaly as originally mapped by Kaszycki and DiLabio (1986a).
b Arsenic anomaly in the $<63\mu m$ size fraction of till. Contour values represent 75th, 90th, and 95th contour values respectively.



The trend of the regional anomaly is approximately 200°. This direction is parallel to the oldest striae observed at the northern end of the anomaly, and oblique to the youngest and dominant striae orientation in that area (Fig. 4). Farther south, the relative age of striae reverses, and the trend of the anomaly is parallel to the youngest striae. This suggests that the ice flow at about 200°, was regionally pervasive throughout the period of glacial erosion and transport of As-enriched debris. Flow events recorded by older striae in the southern part of the train, and by younger striae in the northern part of the train did not significantly influence glacial dispersal. These relationships demonstrate the complexities associated with inferring dispersal direction on the basis of local striae orientation, and emphasize the importance of deciphering regional ice flow history when undertaking drift prospecting programs at any scale.

Detailed sampling in the Wheatcroft Lake area (NTS 64C/2) revealed the internal composition of the northern part of the dispersal train (Fig. 5a). The area of highest As concentration forms a prominent linear zone south of Wheatcroft Lake, defined by several samples containing As concentrations in excess of 95 ppm (85th percentile). This zone, as contoured, is approximatley 4 km wide and 18 km long, and trends approximately 210°, also parallel to the earliest ice-flow within the region. Several other single and multisample As anomalies also occur within the regional dispersal train (>20ppm contour).

The head of the linear anomaly lies south and east of Wheatcroft Lake within the east-west trending belt of sulphide mineralization (Fig. 5a). Glacial erosion and transport of debris from this belt may have produced the long finger of As-enriched till extending southwestward from the vicinity of Wheatcroft Lake. The observed trends, however, may also represent the cumulative effect of glacial dispersal from a number of sources (Shilts, 1975). Several single and multisample anomalies (Fig. 5a) may reflect contribution from unmapped sources within the regional dispersal train. No direct link between As concentrations in the $<2\ \mu$m fraction and any specific bedrock lithology at individual sample sites has been recognized, but due to continuous drift cover in parts of the study area, outcrop is extremely limited and much of the bedrock remains unmapped (Fig. 3). Because of the abundance of sulphide showings in the BRMS near Wheatcroft Lake, we suggest that As-enriched till was eroded from several such occurrences, some of which may remain unmapped.

The distribution pattern reflecting large scale glacial dispersal of As in the clay fraction of till (Fig. 4) can not be clearly identified in the $<63\mu$m size-fraction. This is, in part, a function of lower trace element concentration within this size-fraction and rapid dilution by nonmineralized lithologies. In general, however, the distribution of As in this size fraction is similar to that observed in the $<2\mu$m fraction (Fig. 5b). The zone of anomalous As concentration extends southwestward from near Wheatcroft Lake, parallel to the earliest ice flow direction. The southern end of this anomaly exhibits a subtle shift to the southeast, possibly corresponding to the late glacial shift in ice-flow in that direction. In general, the resolution of areas of anomalous As is better in the $<2\mu$m size fraction than in the $<63\mu$m fraction. This suggests that analysis of the clay fraction of till, at least for As, is more effective for detecting areas of potential mineralization on a regional level, despite the significantly higher cost of sample preparation.

Gold

Contoured distribution patterns of Au in the $<63\mu$m fraction of till in the Wheatcroft Lake area are presented in Figure 6. The lowest value used in contouring is the 75th percentile (7 ppb). Elsewhere, DiLabio (1982) has considered gold levels of >10 ppb in the $<63\mu$m fraction of oxidized till as anomalous. Anomalous Au concentration was not observed in any samples comprising the southern extension of the regional As dispersal train (Fig. 4), and this area is not considered in the following discussion.

In the Wheatcroft Lake area, the distribution of Au in till, although similar to that observed for As, is different in many aspects. The area of highest gold concentration is located south of Wheatcroft Lake and the Laurie River (Fig. 6). In this area, Au concentration is at, or above, the 90th percentile value (10 ppb) at seven sites forming a coherent north-south trending zone approximately 13 km long and 5 km wide. This contains two points having Au concentrations in excess of 19 ppb (95th percentile).

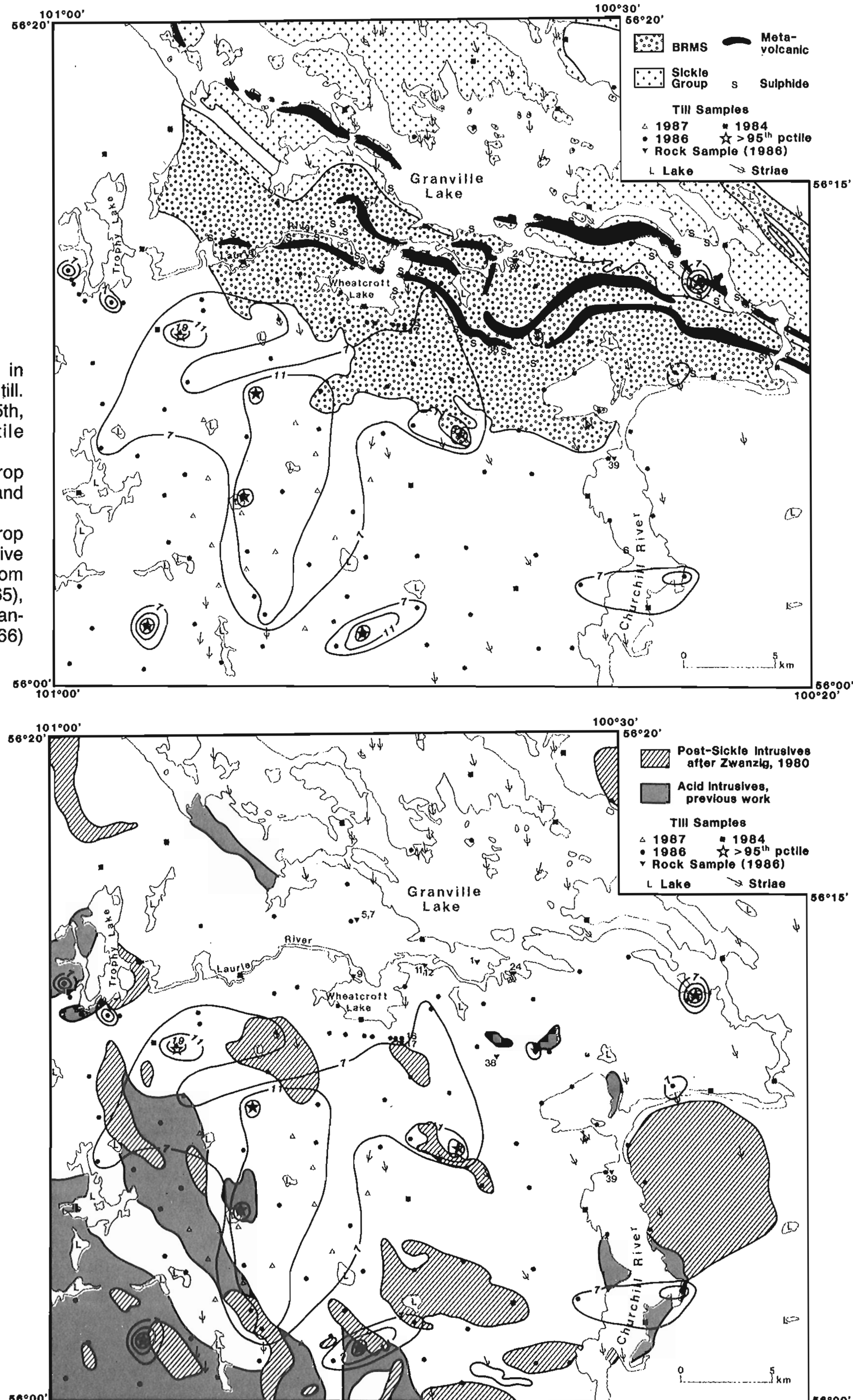
The trend of this anomalous zone, although parallel to observed ice flow directions (190°), is significantly different from that observed for As in the $<2\ \mu$m size fraction (210°) (Fig. 5). It is also significant that samples containing the highest Au concentrations are not anomalous with respect to As in either the $<63\mu$m or $<2\mu$m size fractions, although samples containing the highest As concentration are anomalous with respect to Au. This suggests that the patterns observed for As and Au may be indirectly related in terms of bedrock sources, and glacial dispersal.

Samples containing anomalous Au are located at least 3 to 4 km to the south of the mineralized belt south of Wheatcroft Lake, and approximately 8 km south of the only known Au showing in the region (Fig. 6a). In contrast, several samples containing Au values in excess of 19 ppb (95th percentile) are located on, or within, 100 m of post-Sickle intrusions (Fig. 6b). This may indicate that gold mineralization is associated with quartz veining or fracture systems near the margins of these intrusive bodies. However, not all samples collected near granitic intrusions exhibit this relationship.

Arsenic anomalies, >95th percentile, in both the clay, and silt plus clay size fractions occur within the north-south trending Au anomaly, suggesting that unmapped sulphide occurrences may contribute to Au concentration within the region. The association of As and Au within this zone and within mineralized bedrock, strongly support the use of As as a pathfinder for Au, particularly on a regional scale. Because the only known Au showing in the area is associated with sulphide mineralization in the BRMS, similar sources for the Au anomaly should also be considered.

The degree to which glacial dispersal has influenced Au distribution is not clear. The general trend of the anomaly suggests dispersal parallel to the earliest ice flow direction within the region (190°). However, potential sources for Au

Figure 6. Gold anomaly in the <63μm size fraqction of till. Contour values represent 75th, 90th, and 95th percentile values respectively:
a superimposed on outcrop area of mineralized BRMS and Sickle Group lithologies;
b superimposed on outcrop area of post-Sickle intrusive rocks. Geology complied from Zwanzig (1980), Barry (1965), Barry and Gait (1966), Cranstone(1968), Godard (1966) and Pollack (1966).



within the train are varied and extremely poorly known, and the extent of dispersal, if any, is difficult to determine.

CONCLUSIONS

The regional distribution of As in till defines a northeast-southwest trending zone of As enrichment (>20ppm), which extends approximately 60 km to the southwest from the south shore of Granville Lake. The trend of the anomaly is parallel to a regionally pervasive ice flow event. Flow events recorded by older striae in the southern part of the train, and by younger striae in the northern part of the train, did not significantly influence glacial dispersal. These relationships demonstrate the complexities associated with inferring dispersal direction on the basis of local striae orientation, and emphasize the importance of deciphering regional ice flow history when undertaking drift prospecting programs at any scale.

In detail, the area of highest As concentration forms a dispersal train extending southwest from the mineralized belt near Wheatcroft Lake. Several single and multi-sample anomalies also occur within the area, and may reflect contribution from similar unmapped bedrock sources within the dispersal train.

The distribution of Au in the $<3\pi$m size fraction is similar to those observed for As in both size fractions, but differs in some aspects. Several samples containing anomalous Au are located on, or close to, post-Sickle intrusions, suggesting local Au sources related to quartz veining or fracture systems near the margins of these intrusive bodies. However, the association of As and Au in till and in bedrock, suggests that local unmapped sulphide occurrences within the BRMS may also have contributed to Au concentration within till in the region.

The pattern of gold and arsenic disperal in the Wheatcroft Lake area does not indicate either a single source area or a definite source lithology. The association of As and Au within the dispersal train strongly supports the use of As as a pathfinder for Au, particularly at a regional scale. Arsenic in either size fraction may be considered a useful pathfinder for gold in this area, although resolution of As anomalies reflecting areas of potential mineralization on a regional basis is much greater in the $<2\mu$m size-fraction.

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Geological and structural context of the Grenville Front, southeast of Chibougamau, Quebec

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Ciesielski, A., Geological and structural context of the Grenville Front, southeast of Chibougamau, Quebec; in Current Research, Part C, Geological Survey of Canada, Paper 88-1C, p. 353-366, 1988

Abstract

The Grenville Front is a litho-structural boundary that separates the Superior and Grenville provinces, here considered as autochthonous and parautochthonous with respect to Grenvillian deformation. In the Chibougamau area, Archean volcanic and sedimentary supracrustal sequences and intrusive rocks of the Superior Province are separated from Archean orthogneisses reworked during the Grenvillian Orogeny by a structural and lithological break. This front is marked by faults of the Mistassini system, by rotation of east-west Kenoran foliations to the north-northeast, by an overprint of Grenvillian north-northeast schistosity and by southeast-plunging lineations. An eastward prograde metamorphism of the Superior Archean sequences is related to Grenvillian deformation. Southeast of the Grenville Front, the orthogneisses are relatively homogeneous and contain numerous inclusions correlated with the autochthonous Archean sequences. Southeast-plunging Grenvillian lineations are well developed, but the gneisses preserve Kenoran fabric elements.

Résumé

Le front de Grenville est un accident litho-structural séparant la province du lac Supérieur de celle de Grenville considérées ici comme autochtone et parautochtone par rapport àà la déformation grenvillienne. Dans le secteur de Chibougamau, les roches supracrustales et intrusives archéennes de la province du lac Supérieur sont séparées des orthogneiss archéens tectonisés au Grenvillien, par une rupture lithologique et structurale. Ce front est marqué par les failles du réseau de Mistassini, par une rotation des foliations kénoréennes de l'est-ouest vers le nord-nord-est, par une schistosité grenvillienne surimposée nord-nord-est et par des linéations sud-est. Dans les séquences archéennes du Supérieur, un métamorphisme prograde vers l'est est relié au tectonisme grenvillien. Côté sud-est du front, les orthogneiss sont relativement homogènes et contiennent de nombreuses inclusions correlées avec les séquences archéennes autochtones. Les linéations sud-est grenvilliennes sont bien développées mais les gneiss préservent des éléments structuraux kénoréens.

INTRODUCTION

The region adjacent to the Grenville Front 50 km southeast of Chibougamau, Quebec, has been investigated since the summer of 1984 (Ciesielski et Ouellet, 1985; Ciesielski et al., 1986). The 2500 km^2 study area lies north of Highway 167 and covers part of the Superior Province to the northwest, the Grenville Front, and the Archean gneisses of the Grenville Province to the southeast (Fig. 1). The Grenville Front is a litho-structural boundary that separates two different geological provinces: (1) the Superior Province of Archean age to the northwest, mainly composed of volcanic and sedimentary supracrustal rocks and syn- to late-tectonic intrusions of variable compositions, and (2) the Grenville Province to the southeast, composed of Archean gneisses, tectonites of variable compositions, and intrusive rocks, all affected by the Grenvillian Orogeny (Martignole, 1986; *see* isotopic ages for intrusive rocks in Easton, 1986).

REGIONAL GEOLOGICAL SETTING

Three lithological and chrono-structural entities and two structural and lithological boundaries constitute the regional setting (Fig. 2). These three major entities namely (1) the Su-

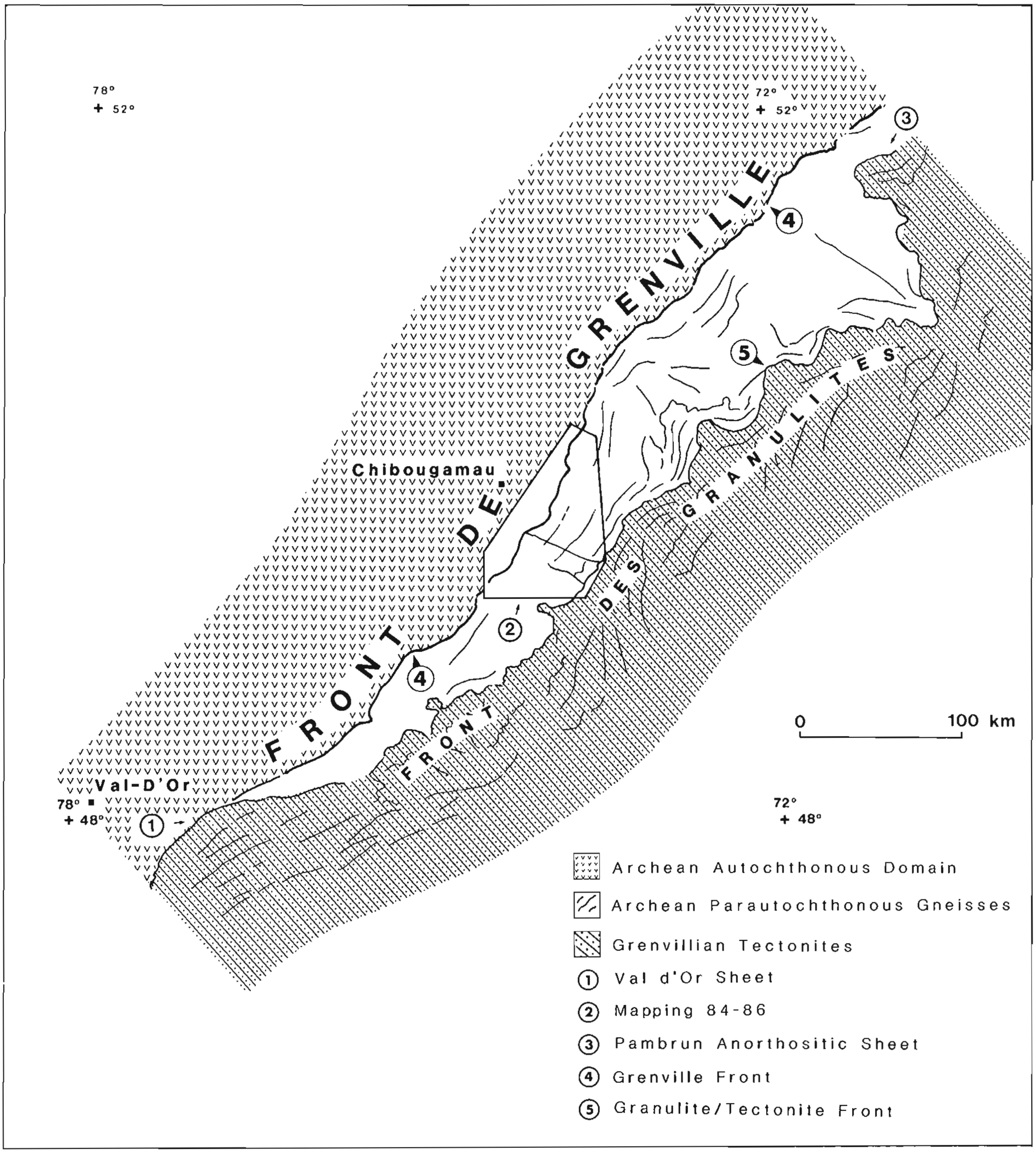


Figure 1. Schematic map showing the autochthonous Superior Province domain separated from the Grenville Province by the Grenville structural front and the adjacent parautochthonous orthogneisses separated from the Grenvillian allochthonous tectonites by the granulite lithological front.

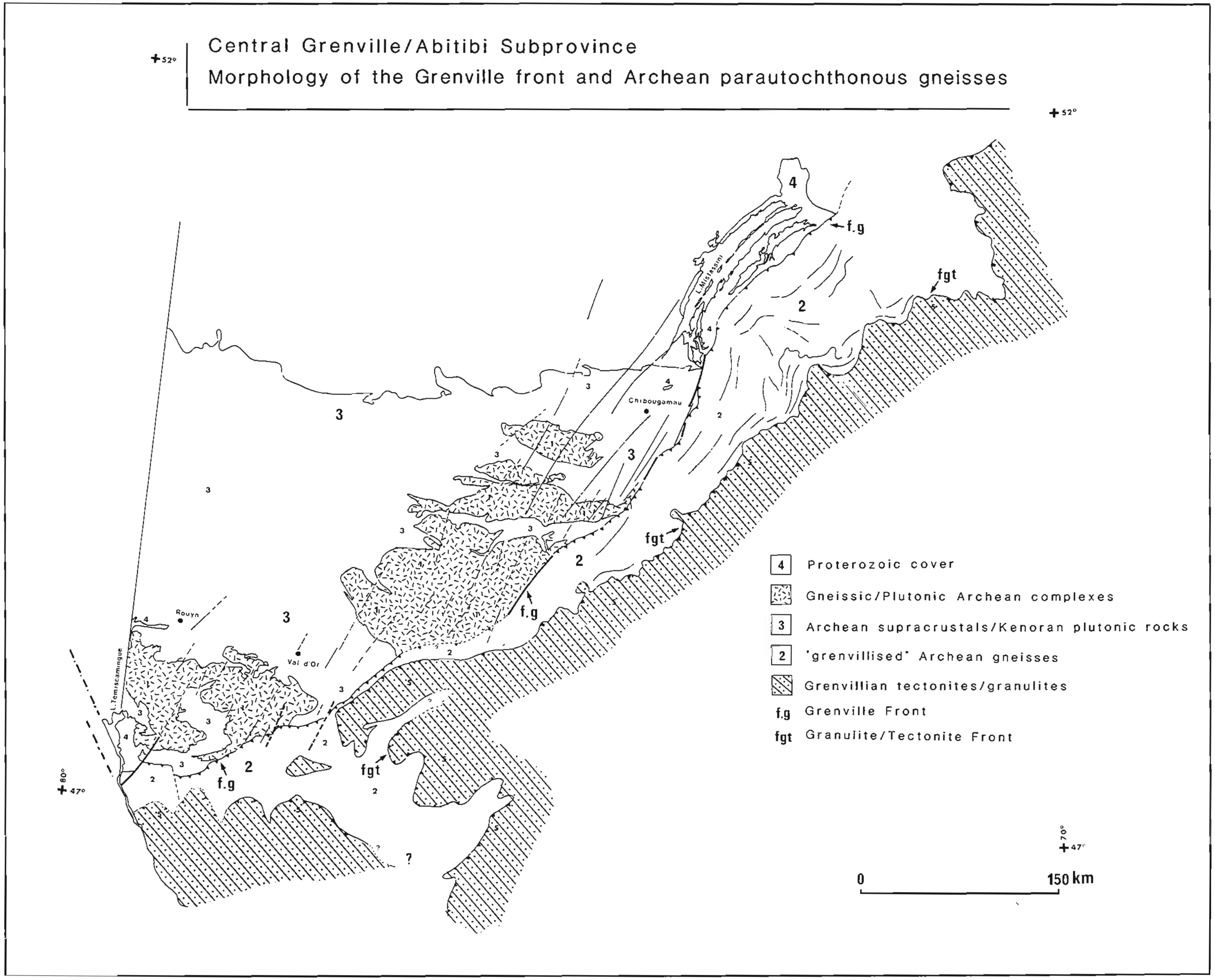


Figure 2. Regional map showing the location and shape of the Grenville Front and the adjacent parautochthonous domain. Note that the faults (heavy and broken lines) are restricted parallel to the front. North-northwest faults in the Temiscaming lake region may represent a conjugate set to the Mistassini system.

perior Province, (2) the reworked grey orthogneisses and (3) the Grenvillian tectonites, are considered to be respectively autochthonous, parautochthonous and allochthonous (Rivers and Chown, 1986) with respect to the Grenvillian Orogeny by analogy to younger alpine tectonic interactions. The Grenville Front is an extra-provincial litho-tectonic boundary, and the granulite front is an intra-Grenvillian lithochronological break.

The Grenville Front itself is a major boundary that truncates the Superior Province over two third of his strike length of 2000 km from Georgian Bay to the Labrador Sea, and defines the northwestern limit of the Grenville Province (Davidson 1986a, b; Wardle et al., 1986). The term "Grenville Front" was proposed by Derry (1950) but the concept was first elaborated in the Chibougamau district by Norman (1936, 1940). Southeast of Chibougamau, local similarities in lithological character and/or metamorphic grade on both sides of the front, lead to the use of several parameters for definition of the Grenville Front, as follow:

(1) lithotectonic break and/or
(2) tectonometamorphic break and
(3) lithological/age/structural break.

The criteria used to define the Grenville Front therefore change with the different geological settings encountered.

The Granulite/Tectonite Front appears as an abrupt linear magnetic break in Figure 3 (Geological Survey of Canada, 1984). Already described to the southeast by Charbonneau (1973), the discontinuity in the total magnetic field southeast of the front is due to (1) the appearance of free magnetite in the rocks,(2) the presence of intermediate to basic intrusive rocks and (3) to some intermediate granulitic rocks. This front is a major lithological break; the grey gneisses of the parautochthonous domain are juxtaposed against a much more heterogeneous terrane to the southeast. It is characterized by a southeastward positive change of relief, by the appearance of pink granitic gneisses, mangeritic rocks, metasediments, by intermediate to basic intrusive rocks and by the presence of clinopyroxene in pegmatites. There is no major southeastward break in the structure, as the grey gneisses are already transposed to a northeast trend before getting into the lithological break. The only major difference would be the presence of thrusted sheets (the tectonites) and subhorizontal (tangential) tectonics southeast of the granulite front that characterize the Grenville Province.

THE AUTOCHTON

Two Archean groups of metasedimentary and metavolcanic rocks and associated intrusions and two Proterozoic sedimentary groups composes the geological setting (Fig. 4a). The Archean Roy Group comprises from base to top, (1) the Obatogamau formation, essentially composed of basic volcanics, (2) the Caopatina formation, composed of volcanogenic greywackes and conglomerates, (3) the Waconichi formation, composed of acidic pyroclastics, (4) the Gilman formation, composed of basic volcanics, and finally (5) the Blondeau formation, composed of intermediate volcanic and acidic pyroclastic rocks intruded by the Cummings Complex formed of three different ultrabasic sills. The overlying Opemisca Group, mainly composed of sedimentary rocks, comprises (1) the Stella formation, (2) the correlative Chebistuan and (3) the basic volcanic Hauy formation (*see* Dimroth et al., 1985 for details).

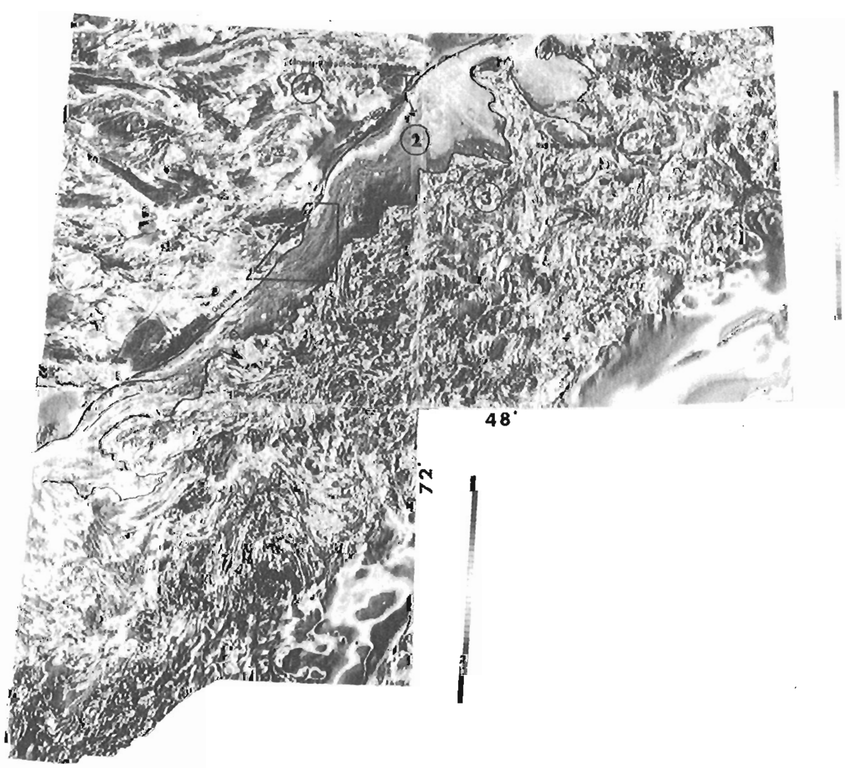


Figure 3. Aeromagnetic total field and shaded relief map (reduced from 1:1 000 000 scale). Each domain is characterized by a different magnetic signature that is related to the grade of metamorphism, structures and lithologies.
(1) Autochthonous Archean supracrustals and intrusive rocks
(2) Parautochthonous orthogneisses
(3) Grenvillian tectonites

Three Proterozoic formations overlie the Archean sequences (1) the Chibougamau formation, correlative with Huronian glacial sediments, (2) the Mistassini clastic sediments, and (3) the Albanel carbonates that form the Aphebian Mistassini Group.

The intrusive rocks, i.e. the Doré Lake complex, the Chibougamau, La Dauversière and Frances Lake plutons (Fig. 4a) are believed to be intrusive in the upper part of the Roy Group. The first two plutons are located in a major east to northwest anticline that forms the centre of the main deformation axis in the Chibougamau district. The Chibougamau anticline is bordered to the north by the Chibougamau syncline, and to the south by the Chapais syncline (Fig. 4a). The Waconichi formation is dated at 2.730 $\pm$ .002 Ga (U/Pb zircon, J. Mortensen, pers. comm., 1987). The Kenoran Orogeny affecting the Archean sequence dated circa 2.7 Ga (Gariépy and Allègre, 1985; Mortensen, 1987) is composed of two phases of Archean deformation D1 and D2 (Daigneault et Allard, 1984). D2 (or S2) is the major phase that has affected all the roks and created the east- west trending series of anticlines and synclines. D1 is a local phase that is related to pluton ascent (R. Daigneault comm. pers., 1987).

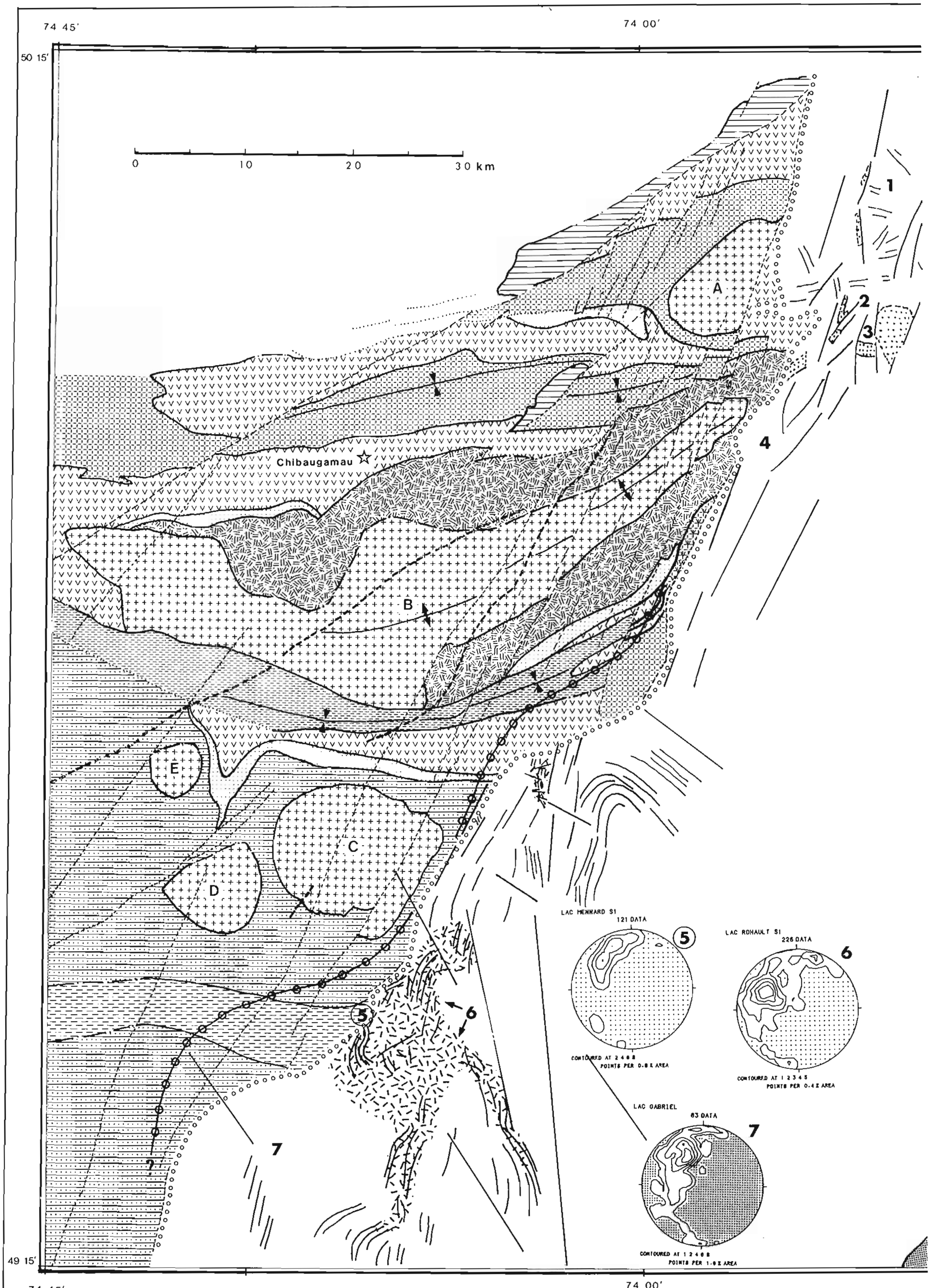
The Archean sequence is affected by three systems of faults. (1) The east-west Kapunapotagen fault system, parallel to the formation boundaries, is believed to be the result of D2 deformation. (2) The northeast Gwillim fault system comprises two small faults, Lamark and McKenzie, and a major one, the Gwillim. This system also cuts the Proterozoic rocks and is in turn transected by the Mistassini fault system. The field relations suggest a late Archean age and a subsequent reactivation during Proterozoic times (Dimroth et al., 1984). (3) The north-northeast Mistassini fault system is well developed near the boundary with the parautochthonous gneisses all along the Grenville Front (Fig. 3). In the Chibougamau district, the Mistassini system extends from a few kilometres in the parautochthonous gneisses westward to 40 km into the autochthonous Archean sequence (*see* maps in Allard, 1976, and Daigneault et Allard, 1984). This fault system is involved in the deformation of the Archean sequence near the front during the Grenvillian Orogeny (Allard 1981). The field relations suggest a post-Aphebian age for the Mistassini faults. Moreover, Ar-Ar (hornblende) isotopic studies by Baker (1980) on metamorphic assemblages developed in the Archean volcanics on foliations parallel to the Mistassini fault system yield Grenvillian ages.

Many diabase dyke swarms occur in the autochthon of the Chibougamau district. The northeast-trending Preissac dykes have been dated at 2.15 Ga and the east-northeast-trending Abitibi dykes at 1.15 Ga (Fahrig and West, 1986). The northwest-trending Mistassini dyke swarm in the Mistassini lake region, is dated at 1.95 Ga (Chown and Archambault, 1987). A correlation between fault systems and dyke swarms suggests a minimum Grenvillian reactivation age for the Gwillim fault. In the vicinity of the Grenville Front, some diabase dykes are parallel to the north-northeast Mistassini fault system (Allard, 1976; Avramtchev, 1975; Hébert, 1980; Deland et Grenier, 1959). The shaded relief aeromagnetic map clearly shows anomalies parallel to the fault system near the Grenville Front (Geological Survey of Canada, map NM 18). Most dykes are unmetamorphosed but some contain garnet (Hébert, 1980) and are affected by the Grenvillian metamorphism that took place in the Archean autochthonous sequence (Baker, 1980). Although no geochronological data are available, Chown et Archambault, (1987) considered that these dykes belong to the Preissac swarm. However, based on their geological setting, they may be grouped into a different family, the "Grenville Front dyke swarm". A recent study by Ranalli and Ernst (1986) relates the Abitibi swarm to a boundary load caused by the Grenvillian Orogeny in the Mistassini lake region. No arguments are given for the origin of the "boundary load" except that it is related to the presence of the so called Mistassini Bulge; the hypothesis is not supported by radiometric ages.

All supracrustal and intrusive rocks of the Superior Province are affected by Grenvillian deformation in the vicinity of the Grenville Front. The first effect observed is the gradual rotation of the Kenoran S2 fabric from its east-west orientation to the north-northeast. The geological map (Fig. 4a) shows a general northeast flexure of formation boundaries towards the Grenville Front (*see* also Gobeil et Racicot, 1983) that become parallel to the shape of the front (Allard, 1976; 1981; Daigneault, 1987). In addition to the rotation of foliation planes, penetrative concordant faulting (Mistassini fault system) controls the morphology of the Grenville Front (Allard, 1981; Daigneault et Allard, 1984). Figure 5 illustrates the penetrative character of the Mistassini fault system, the concordant relationship between the foliation trend and the morphology of the front, and the orientation of various planar and linear fabric elements.

In addition to the rotation of the Archean S2 fabric, a new foliation overprinted the rocks in the vicinity of the Grenville Front (Fig. 6a, b). This foliation (Z1) is roughly parallel (but not everywhere) to the north-northeast-trending Mistassini fault system, and contains the Grenvillian metamorphic mineral assemblages developed in the autochthonous rocks. This foliation may have various orientations as the rotation of S2 is not always completed. The angle between the Archean and the new Grenvillian foliation S2 vs Z1 is variable, and as the Grenville Front is approached both fabrics may become juxtaposed. A locally developed northeast-trending crenulation appears to be linked with Z1 (Fig. 7). A southeast-plunging lineation is well developed almost everywhere in the vicinity of the Grenville Front and overprints most of the pre-existing structures (including folds). There is evidence of a link between Mistassini faulting and rotation of Kenoran foliations but not with the lineations. Figures 4a and 5 show the consistent orientation of the lineations as opposed to the various foliation trends.

Kenoran greenschist grade metamorphism affected the volcanic and sedimentary rocks of the autochthonous sequences. Zircon ages circa 2.7 Ga for the Kenoran metamorphism are given by Mortensen (1987) and Gariépy and Allègre (1985). The metamorphic overprint linked with the north-northeast foliations gives a Grenvillian age (Argon ages on amphiboles in Baker, 1980). A few kilometres west of the Grenville Front, garnet and hornblende are present in the chilled margins of pillowed mafic volcanic and intermediate pyroclastic rocks (Fig. 8a, b). Chlorite disappears progressively to the east, and appearance of staurolite and kyanite near the front record amphibolite facies metamorphism. This


74 45'
74 00'
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0 10 20 30 km
Chibaugamau
A
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LAC ROHAULT S1
226 DATA
LAC GABRIEL
83 DATA

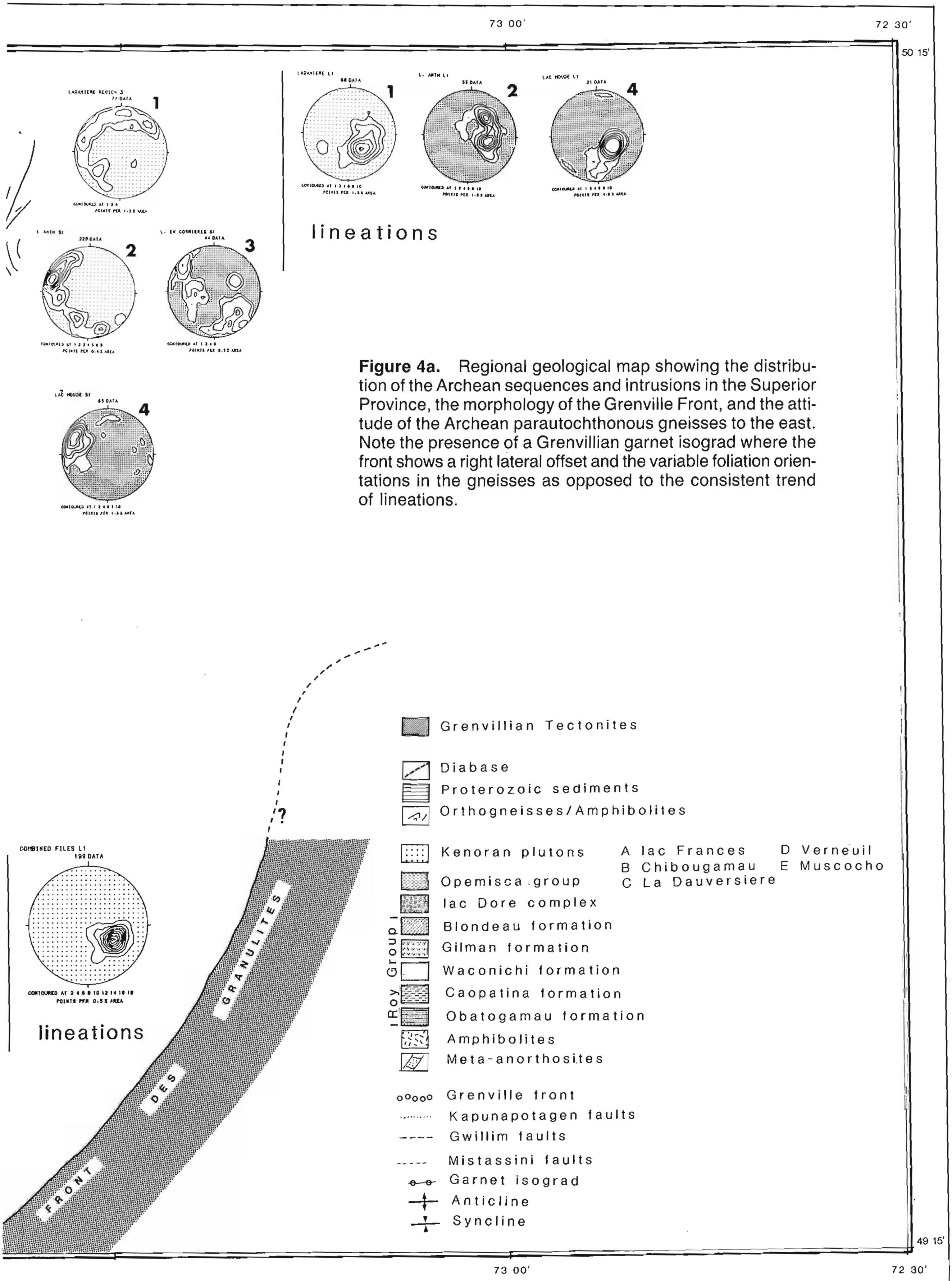

Figure 4a. Regional geological map showing the distribution of the Archean sequences and intrusions in the Superior Province, the morphology of the Grenville Front, and the attitude of the Archean parautochthonous gneisses to the east. Note the presence of a Grenvillian garnet isograd where the front shows a right lateral offset and the variable foliation orientations in the gneisses as opposed to the consistent trend of lineations.

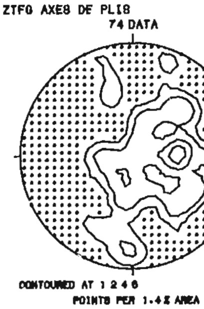

Figure 4b. Distribution of the fold axes along the Grenville Front. Compare with lineations on Figure 4a and 5.

eastward prograde Grenvillian metamorphism was recognized where a right lateral offset of the Mistassini fault system resulted in a flexure of the Grenville front (see the "garnet isograds" on Fig. 4a). In Dollier township (Fig. 5), the metamorphosed Archean sequence stops at the front and is in contact with tonalitic orthogneiss that forms the parautochthonous domain. In Rohault Township, south of La Dauversire pluton (Fig. 4a), the metavolcanic and metasedimentary sequences (the Obatogamau and Caopatina formations) attain amphibolite facies at the front (E. Ouellet, pers. comm., 1987). Here the front forms a structural break, and the rocks are preserved in the parautochthonous domain east of it for over 20 km where the metamorphism increases to clinopyroxene-amphibolite sub-facies.

THE GRENVILLE FRONT

The role of the north-northeast faulting in the morphology of the front is predominant, as faults in places define the front and are linked by tectonic contacts. A right lateral offset pattern of the Mistassini fault system can be seen both south and north of La Dauversière pluton (Fig. 4a). In the field, no important east-west faults are visible that could be related to this offset. In many places, the front is defined by an abrupt lithological change from metavolcanic rocks to tonalitic orthogneisses. This lithological break can coincide with major faults and may show cataclasites, breccias, protomylonites and magnetite or clinopyroxene bearing pegmatites (Fig. 13). Tectonic contacts between metavolcanic rocks and gneisses without specific chronological or structural relations are also noted. South of La Dauversière pluton (Fig. 4a), the Grenville Front is defined by the structural flexure to the north-northeast as the supracrustal rocks are preserved across the autochthonous and parautochthonous domain boundary and the metamorphic grade remains constant. In certain cases, the front is a zone of imprecise width having in places almost the same orientation as the original Kenoran foliations. Distributions of fold axes along the front are not concordant with the Grenvillian trend (Fig. 4b); they are likely to be related to preservation of ancient (Kenoran?) structures.

THE PARAUTOCHTHON

The extent of this geological entity is shown on Figure 2. The geological setting near Chamouchouane River and on western Gouin Reservoir and recent studies southeast of Val d'Or, allow extension of the parautochthonous gneisses to near the Pontiac schists south of Val d'Or (Charbonneau, 1973; Chown, 1971; Laurin, 1965; Indares and Martignole, in press; Ciesielski et al., in press; A. Indares, pers. comm., 1987). In the study area, the parautochthonous gneisses extend from the Grenville Front to the granulite/tectonite front to the southeast over a distance of 60 km (Fig. 2).

The parautochthonous domain (Fig. 2), is mainly composed of orthogneisses of tonalitic and trondhjemitic composition. Although heterogeneous at outcrop scale, the orthogneisses present a relative chemical homogeneity (Ciesielski et al., in press) (Fig. 9). Southeast of the Grenville Front, they show variable texture and grain size and different kind of inclusions (Fig. 10). The rock can gradually vary from a homogeneous fine grained tonalite to a coarse brecciated diatexitic granodiorite. Continuation of Archean sequences from the Superior Province (autochthonous domain) eastward through the Grenville Front is well documented (Gilbert, 1959; Laurin et Sharma, 1975). East of the Lac Frances pluton, Allard (1978, 1979) and Lacoste (1986) described anorthositic gneisses that have been correlated with the lac Doré complex and amphibolitic, ultrabasic and felsic gneisses correlated with the Blondeau formation (Fig. 11a, b). Alhtough problematic, since the Grenville Front in this region is principally marked by faults, these associations are realistic mainly because of similar mineralogical compositions and space distribution of the different types of inclusions. Anorthosite remnants (Fig. 4a), are found very close to the front; they show penetrative ductile deformation in places and are separated from the lac Dor complex by a thin (5 m) band of grey gneiss. Farther to the northeast, the meta-anorthosites form a large homoclinal succession, where they are associated with garnet amphibolitic rocks. The felsic, basic and ultrabasic layered inclusions are either intimately deformed with the gneisses or form large rafts and maintain their original mineralogical layering; no primary volcanogenic structures are preserved. In Rohault Township, south of la Dauversière pluton (Fig. 4a), the gneisses are mixed with amphibolites of volcanic origin and metasedimentary rocks that may be correlated with the Archean Obatogamau and Caopatina formations (Fig. 12a, b). Volcanogenic structures, presence of phenocrysts in the amphibolites and the presence of amphibole and garnet in most of the metasediments give enough evidences for the correlation.

The structure of the parautochthonous gneisses reflects two phases of deformation that cannot be discriminated easily. Large tight folds, overturned to the east having a shallow plunge to the south, probably represent early folds related to Kenoran deformation. Other large folds with northwest axial traces that are seen in the gneisses northeast and south of La Dauversière pluton can either be related to Kenoran or to Grenvillian drag folding along the Grenville Front. Fold axes along the front reflect preservation of Kenoran structures (Fig. 4b). The lineations in the gneisses have a consistent plunge to the southeast (east in certain places) and are not related to most of the folds encountered within and along the front; moreover they cannot be linked to any crossing schistosities (Fig. 4a). The Kenoran deformation was partly obliterated by the Grenvillian event. The rotation of the Kenoran foliations and fold axes to the north-northeast during Grenvillian deformation is not complete, and explains the remnant ancient orientations.

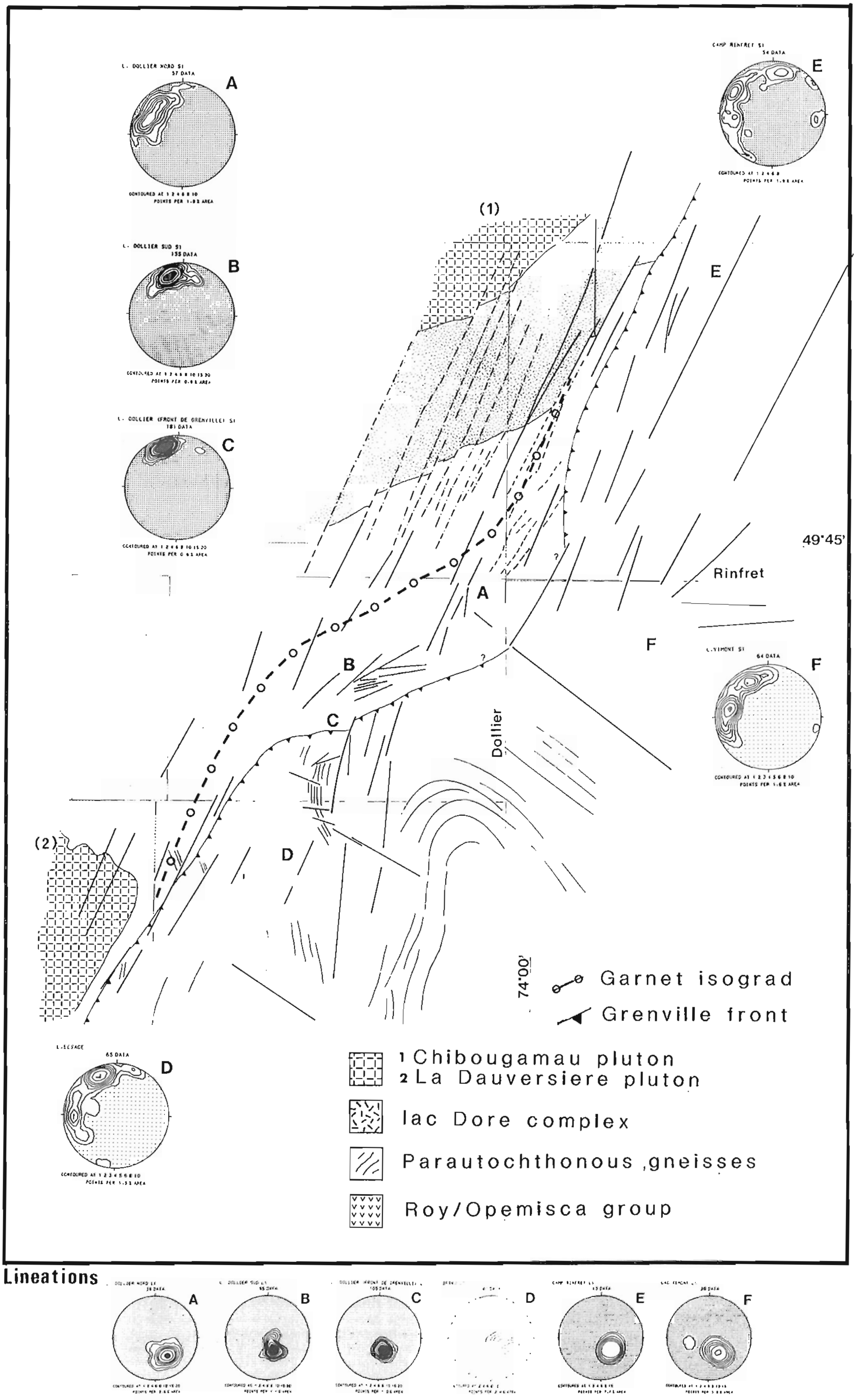

Figure 5. Geological map of the Dollier Township area showing the importance of the Mistassini fault system (heavy and broken lines), the relative parallelism of the garnet isograd with the Grenville Front, and the discordance between foliations and lineations in the gneisses.

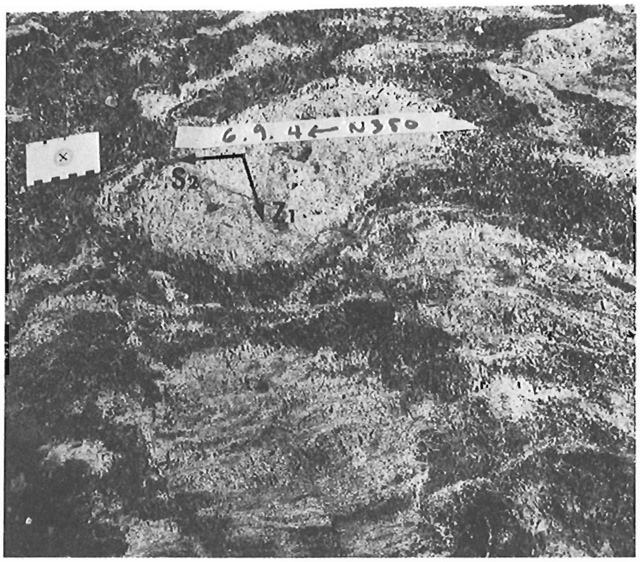

Figure 6a. Superimposed Grenvillian foliation (Z1) on elongated pillow lavas (S2) of the Gilman formation.

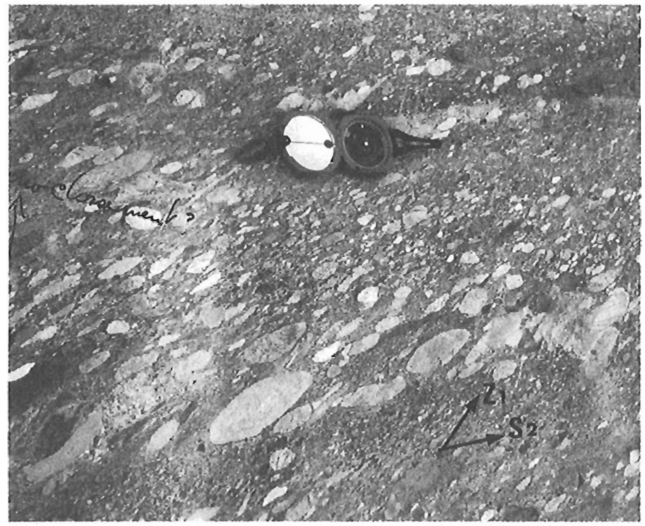

Figure 6b. Superimposed Grenvillian foliation (Z1) resulting in stretching of the conglomerate cobbles of the Stella formation already deformed by Kenoran S2 (parallel to S0).

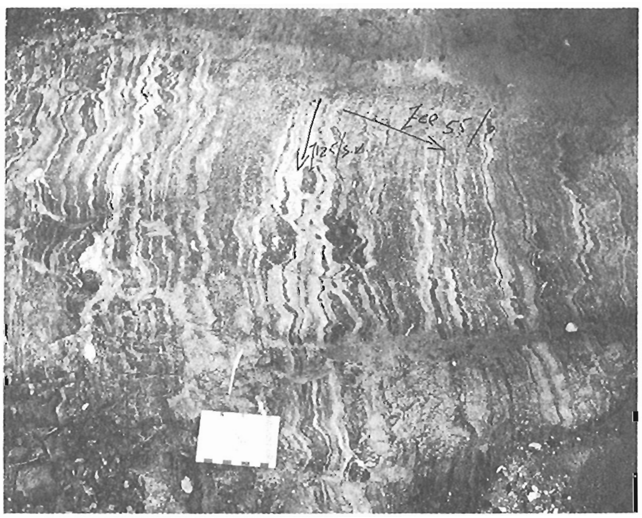

Figure 7. Crenulations formed near the Grenville front in pyroclastic rocks of the Blondeau formation.

East of the Grenville Front all the inclusions or mappable Archean supracrustal and intrusive rocks have reached the amphibolite facies. Epidote-garnet-hornblende assemblages are present throughout the region. Clinopyroxene is also commonly encountered near the front southeast of La Dauversière pluton (Fig. 13). North-northeast-trending diabase dykes having a maximum width of 30 m show coronitic assemblages of orthopyroxene-clinopyroxene-garnet-hornblende. The dykes are deformed into boudins several hundred metres long and show marginal deformation and a strong amphibolitization not always parallel to the host-rock contacts. It is not clear whether or not the metamorphism that has affected the supracrustal rocks is responsible for the coronitic textures (Neale, 1959). A possible magmatic origin was also suggested by Davidson and Grant (1986) and it raises the problem of the age of the dykes versus the

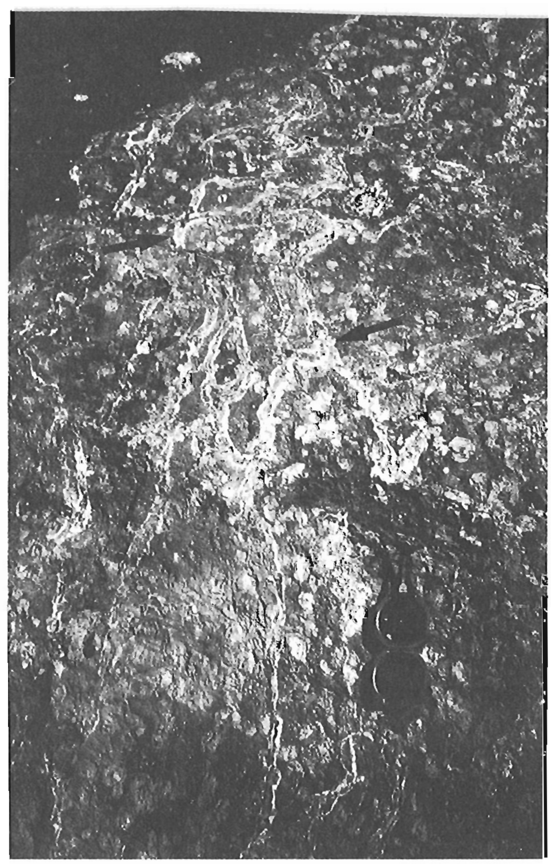

Figure 8a. Garnet developed at the chilled margins (arrows) and inside the pillows within basalts of the Gilman formation.

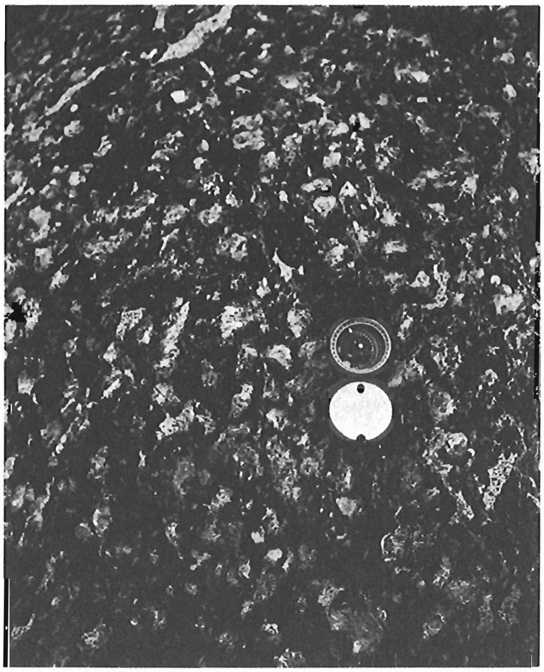

Figure 8b. Basic pyroclastic rock of the Blondeau formation, containing large garnets and amphiboles.

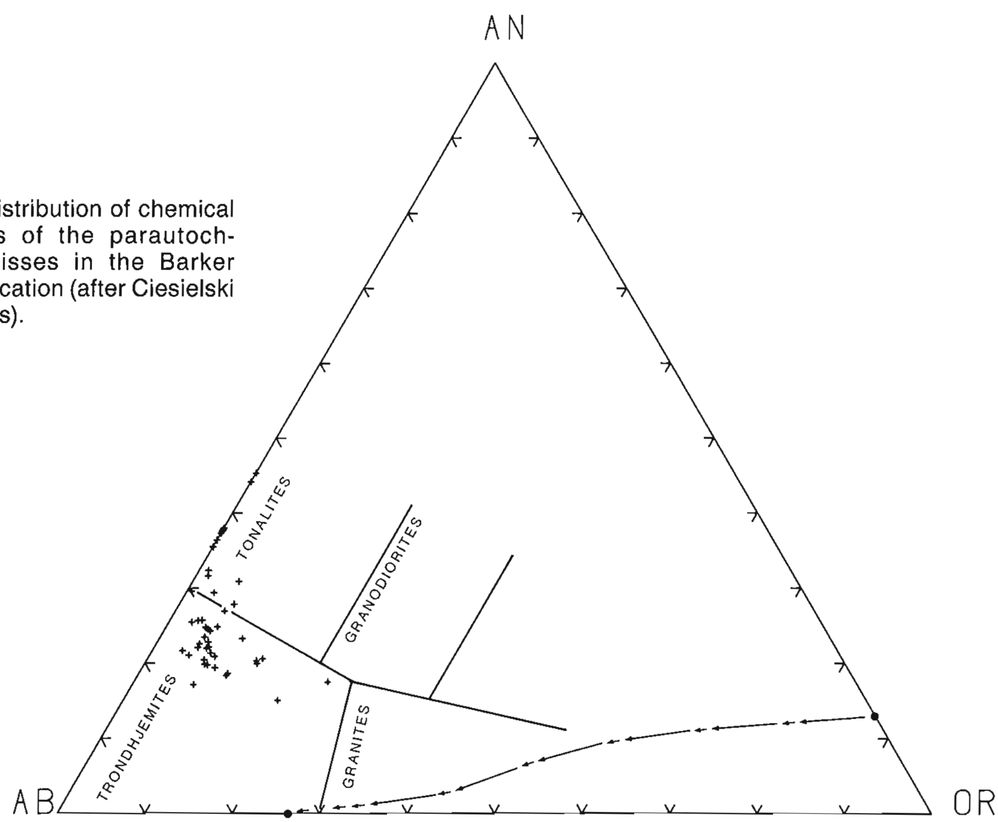

Figure 9. Distribution of chemical compositions of the parautochthonous gneisses in the Barker (1979) classification (after Ciesielski et al., in press).

Figure 10. Brecciated orthogneiss containing migmatized metasedimentary inclusions.

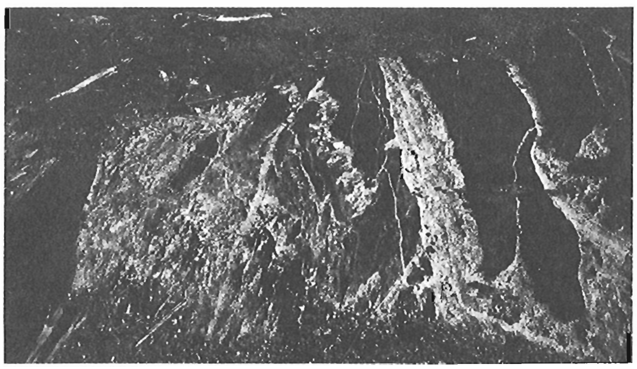

Figure 11a. Meta-anorthosite and gabbro forming large elongated xenoliths in the parautochthonous orthogneisses. Correlated with the Lac Doré complex.

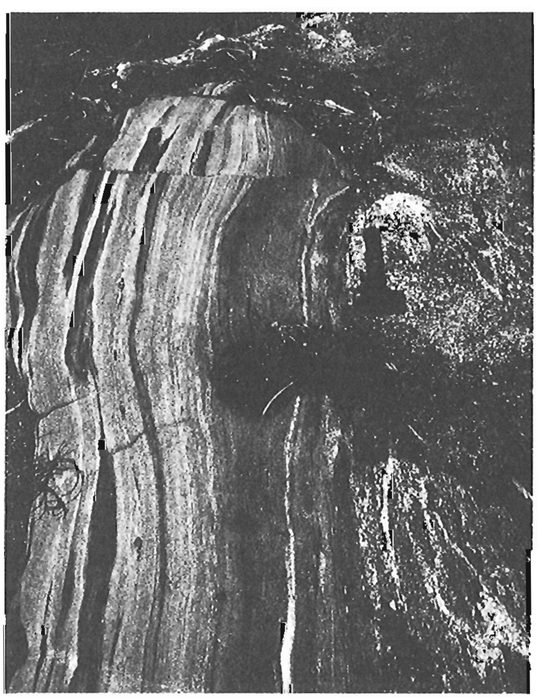

Figure 11b. Intermediate and acidic layered inclusion in brecciated orthogneiss. Correlated with Blondeau formation.

Figure 12a. Amphibolite of volcanic origin showing pillow lavas and phenocrysts in the parautochthonous gneisses. Correlated with Obatogamau formation.

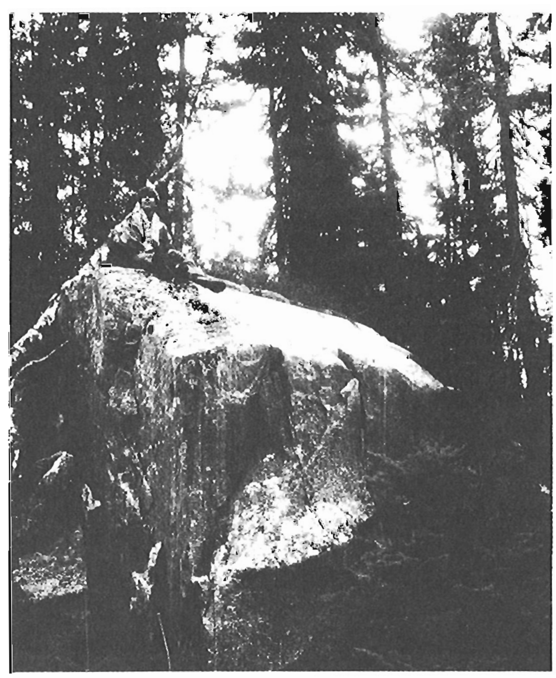

Figure 12b. Subvertical quartzitic metasediment in the parautochthonous gneisses. Correlated with Caopatina formation.

Figure 13. Pegmatite with large clinopyroxene, located at the Grenville Front.

coronites, and its importance in the uplift model of the parautochthonous domain. This uplift model suggests that an upward motion of the parautochthonous domain during Grenvillian deformation was reponsible for the overprinted Grenvillian metamorphism in the autochthonous domain. Still speculative, the uplift and its extent should be calculated according to the PT conditions preserved by the mineral assemblages present in the rocks, providing a simple upward gravity model. Described by Neale (1952) 100 km to the north, the uplift model appears realistic considering that the parautchthonous gneisses carry Archean sequence inclusions at much higher grade than their autochthonous equivalents and are overprinted by lineations that are not related to schistosities interaction; in the model they would have a stretching origin. This upward northwest movement could also account for the rotation of the Archean supracrustals to the north-northeast in the autochthonous domain representing an apparent sinistral horizontal component of movement along the Grenville Front.

CONCLUSIONS

The principal geological aspects of the Grenville Front in the Chibougamau area are: (1) the vergence to the north-northeast of the Archean supracrustal sequences and multiple plutonic intrusions and the synchronous development of the Mistassini fault system, (2) the superimposed Grenvillian schistosity and lineations that accompany the eastward prograde Grenvillian metamorphism, (3) the structural heterogeneity of the Grenville Front, made of north-northeast faults and right-lateral offsets linked by tectonic contacts between metavolcanics and orthogneisses, (4) the relative homogeneous chemical compositions of the parautochthonous gneisses, (5) the apparent continuity of the Archean sequences across the Grenville Front and persistence of Archean intrusive rocks east and pre-Grenvillian foliations and folds that are partly obliterated and rotated as a result of the Grenvillian deformation east of the front.

Important questions remain to be answered in the area along and southeast of the Grenville Front. Baker (1980) documented Grenvillian Ar-Ar ages on amphiboles for the eastward prograde metamorphism in the autochthonous domain. The precise age of this metamorphic event and its relationship with the Mistassini fault system will have to be confirmed. Frith and Doig (1975) found pre-Kenoran Rb-Sr ages for the gneisses southeast of the Grenville Front. A recent study yielded a U-Pb zircon age of 2.620 $\pm$ .002 Ga for the parautochthonous gneisses (M. Joly, pers. comm., 1987). This age raises the question of crustal level origin for the gneisses. The question remains that if this 2.620 Ga age is dating the emplacement of the gneisses protolith, then the deformation preceding the Grenvillian event is post-Kenoran and it does not fall within the presently known regional deformation history. It may also reflect a metamorphic resetting of the zircons U/Pb ratios. Moreover, the absolute age relation between these gneisses and their supracrustal inclusions should be established prior to any petrogenetic model, since the field observations alone do not allow the problem to be resolved.

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Stratigraphic setting of gold concentrations in Archean supracrustal rocks near the west side of Bathurst Inlet, N.W.T[1]

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Abstract

Exploration targets for gold in the Bathurst Inlet area include auriferous pyrrhotitic silicate iron-formations, fracture-controlled sulphide-rich sections within magnetite-quartz iron-formations and sulphide and arsenide-bearing altered zones in massive amphibolite and in felsic tuff. The two iron-formation-hosted types occur a short distance stratigraphically above the other in some places, as at the deposits near Pistol Lake , or along strike from each other within the same general stratigraphic zone. East of the inlet, both occur in rocks up to and including sillimanite grade and in migmatitic gneisses, but they are much disrupted in the higher grade rocks. An interesting coherent stratigraphic zone about 2 km wide has been mapped for 40 km from James River south to Hood River. It contains long sections of clast-supported conglomerate, oxide and silicate iron-formations, felsic tuff, micaceous quartzite and amphibolite possibly derived from ultramafite.

Résumé

Des cibles présumément aurifères de la région de l'inlet Bathurst contiennent des formations ferrifères à silicates pyrrotitiques aurifères, des sections contrôlées par des fractures et riches en sulfure dans des formations ferrifères à magnétite et quartz, et des zones altérées contenant des sulfures et des arséniures dans une amphibolite massive et un tuf felsique. Les deux types logés dans des formations ferrifères gisent, par endroits, à une courte distance stratigraphiquement au-dessous du troisième, comme dans le cas des gisements près du lac Pistol ou en direction l'une de l'autre dans la même zone stratigraphique générale. À l'est de l'Inlet, les deux gisent dans des roches dont la qualité s'étend jusqu'au niveau de la sillimanite et dans des gneiss migmatitiques, mais se trouvent très perturbés dans les roches de qualité supérieure. Une zone stratigraphique cohérente d'environ 2 km, particulièrement intéressante, a été cargographiée sur 40 km depuis la rivière James, au sud, jusqu'à la rivière Hood. Elle contient de longues sections de conglomérat supporté par des roches clastiques, des formations ferrifères à oxydes et à silicates, un tuf felsique, du quartzite micacé et de l'amphibolite qui proviendrait d'ultramafite.

[1] Contribution to Canada-Northwest Territories Mineral Development agreement 1987-1991. Project carried by the Geological Survey of Canada

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INTRODUCTION

Archean metasedimentary rocks east and west of Bathurst Inlet contain numerous concentrations of gold in metamorphosed iron-formations. The first discovery of one of these in 1958 predated the discovery at Contwoyto Lake of the iron-formation — hosted deposit presently mined at the Lupin gold mine of Echo Bay Mines Limited 190 km southwest of Bathurst Inlet. This prospect near the east side of the inlet was found by H. Vuori employed by Inco Limited. Several, including the Pistol Lake prospects, were found west of the inlet in 1963 and 1964 by Noel Avadluk and others prospecting for Roberts Mining Company. New discoveries have been made since 1983 in both the eastern and the western metasedimentary belts. Many of those in the eastern belt resulted from investigations of occurrences of gossaned garnet-amphibole-quartz gneiss (iron-formation protoliths) found in highly metamorphosed sedimentary rocks by Geological Survey of Canada mapping parties (Thompson and Culshaw, 1985; Thompson et al., 1986).

The only published information on the belt of dominantly metasedimentary Archean suprecrustal rocks west of the inlet has been its generalized outline derived from helicopter-borne reconnaissance geological mapping (Fraser, 1964). Geologists employed by Silver Hart Mines Ltd. and by Echo Bay Mines Limited have recognized and mapped in detail an interesting package of rocks that forms a coherent core of the belt, extending 50 km south from James River to the Booth River area. This work, mainly by Silver Hart Mines who hold mineral rights through most of the core zone, forms the base for the synopsis of regional geology of the belt presented here. We believe that this zone presents special opportunities to study environments of deposition of two distinct types of iron-formations, effects of metamorphism on these, and timing of gold enrichment within them.

GENERAL GEOLOGY

The main units in the Turner Lake supracrustal belt are amphibolite grade metasedimentary rocks and gneisses derived from greywacke, mudstone and argillaceous quartzose arenite (Fig. 1). The strata, throughout most of the belt, are believed to face east away from an extensive area of tonalite and other granitoid rocks. Pillowed mafic volcanics are present in a small area (67°09'N, 109°10'W) along the margin of the tonalite. They are overlain by felsic volcanic rocks and conglomerate. These may be remnants of a once-extensive volcanic sequence subjacent to turbiditic arenites — a relationship common in supracrustal belts of the Slave Structural Province. A stratigraphic zone about 2 km wide, containing extensive thin iron-formations, long lenses of conglomerate, micaceous quartzite, sparse layers of felsic tuff, mafite of possible extrusive origin, and intermediate to mafic intrusive rocks (Fig. 2), is central within the belt of dominantly arenaceous clastic sedimentary rocks.

Strata dip steeply. Locally, some beds are tightly folded and may be duplicated by faulting, but no evidence has been found of large scale folds or repetitions of major portions of the stratigraphic pile. Clasts in conglomerate are markedly to extremely flattened parallel to steeply-dipping bedding and, to a much lesser extent, are stretched steeply to the north. An important shear zone (not shown in Fig. 1 and 2) strikes parallel to beds west of Turner Lake.

The supracrustal rocks and early intrusions are intruded by two tonalite pluton, one small and one large, in addition to the the extensive granitiod complex to the northwest. The post-orogenic Booth River Intrusive Complex, including mafic to ultramafic rocks and alkaline granite (Roscoe,1985), intrudes the southern part of the belt. Sedimentary rocks of the Aphebian Goulburn Group nonconformably overlie the Archean rocks and the Booth River intrusions. Sheets of gabbro intrude the base of the Goulburn Group and in places extend down into the underlying Archean rocks. The youngest consolidated rocks in the area are diabase dykes. There are several sets and probably several ages of these — a northwesterly, a northerly and a westerly-trending set. The northwest-trending Mackenzie dykes of Helikian age are most numerous and are conspicuously reflected on aeromagnetic maps. Only a few Mackenzie dykes are shown in Figure 1. The Bathurst Fault truncates the north end of the belt.

The Turner Lake belt shows a number of features that are atypical of other Slave Province metasedimentary belts as mapped to date. These include: distinctive conglomerate formations that have been traced 36 km; minor, but widespread, volcanic rocks; associations of two types of iron-formations along very extensive strike lengths; a varied suite of intrusive rocks; several types of gold concentrations; minor concentrations of copper-nickel sulphides; late nickel arsenide-bearing veins that cut the copper-nickel concentrations. The numerous mineral prospects have been omitted from figures in this report.

JAMES FALLS CONGLOMERATE

A conglomerate unit, first noted by Roscoe (1983, Fig.5), is best exposed at the falls on the lower part of the James River. It has been traced southward 36 km from the Bathurst Fault, a few kilometres north of the falls, nearly to the Hood River. Extensive overburden cover in the south part of the belt may hide further extensions of the unit. It is structurally disrupted and in places several lenses of conglomerate are present across sections up to 2 km wide. Some of these lenses may represent structural repetitions, but others are more likely to be channel fillings. Widths of the main band vary greatly, due in part to differences in tectonic thinning which is reflected by flattening of clasts which in exteme cases may appear as ribbons a few centimetres wide and several metres long. The greatest width of conglomerate observed is about 300 m but the original thicknesses of the formation likely exceeded this.

Clasts in all conglomerate exposures are predominently cobble-size and are mainly pale grey, folliated, quartzose, micaceous clastic sedimentary rocks and tonalite. They are squeezed into mutually-accommodating elongate forms, and the less competent sedimentary clasts are moulded around more competent tonalite clasts. Even in outcrops of the least deformed conglomerate (Fig. 3), some sedimentary clasts are difficult to distinguish from compositionally similar, sparse matrix material that originally filled interstices between tightly packed cobbles and boulders. Clasts of dark, recessive weathering meta-argillite, black cherty rock, diorite, mafite, and quartz are also present. The black cherty clasts are less deformed and more angular than others. A few are elongate with long axes that strike northerly, in contrast to the formation-parallel north-northeasterly strike of long axes and foliation of all other clasts. Yet they appear to have been rotated clockwise toward the foliation plane. Granular quartz pebbles less than 2 cm in diameter, believed to have been vein quartz originally, are the smallest clasts. Many sedimentary clasts are rusty-weathering due to contents of iron sulphide. Matrix rock is also sulphidic in many areas. In places, slight increases in radioactivity seem to be associated with these sulphide concentrations. Tourmaline has been noted by one of us (C.F.S.) in conglomerate matrices.

No bedding or other internal primary structures are found within the conglomerate bodies. West of Turner Lake, however, there is an extensive 1 m layer of quartz arenite within the conglomerate, 2 m from its east border (and presumed top). The proportion of tonalite to other clasts decreases from north to south. Clast sizes probably decrease southward also, but this is difficult to determine due to variations in widths and lengths of squeezed clasts. Clast elongations plunging about 70° north were noted at James Falls but clasts everywhere are flattened much more than they are stretched.

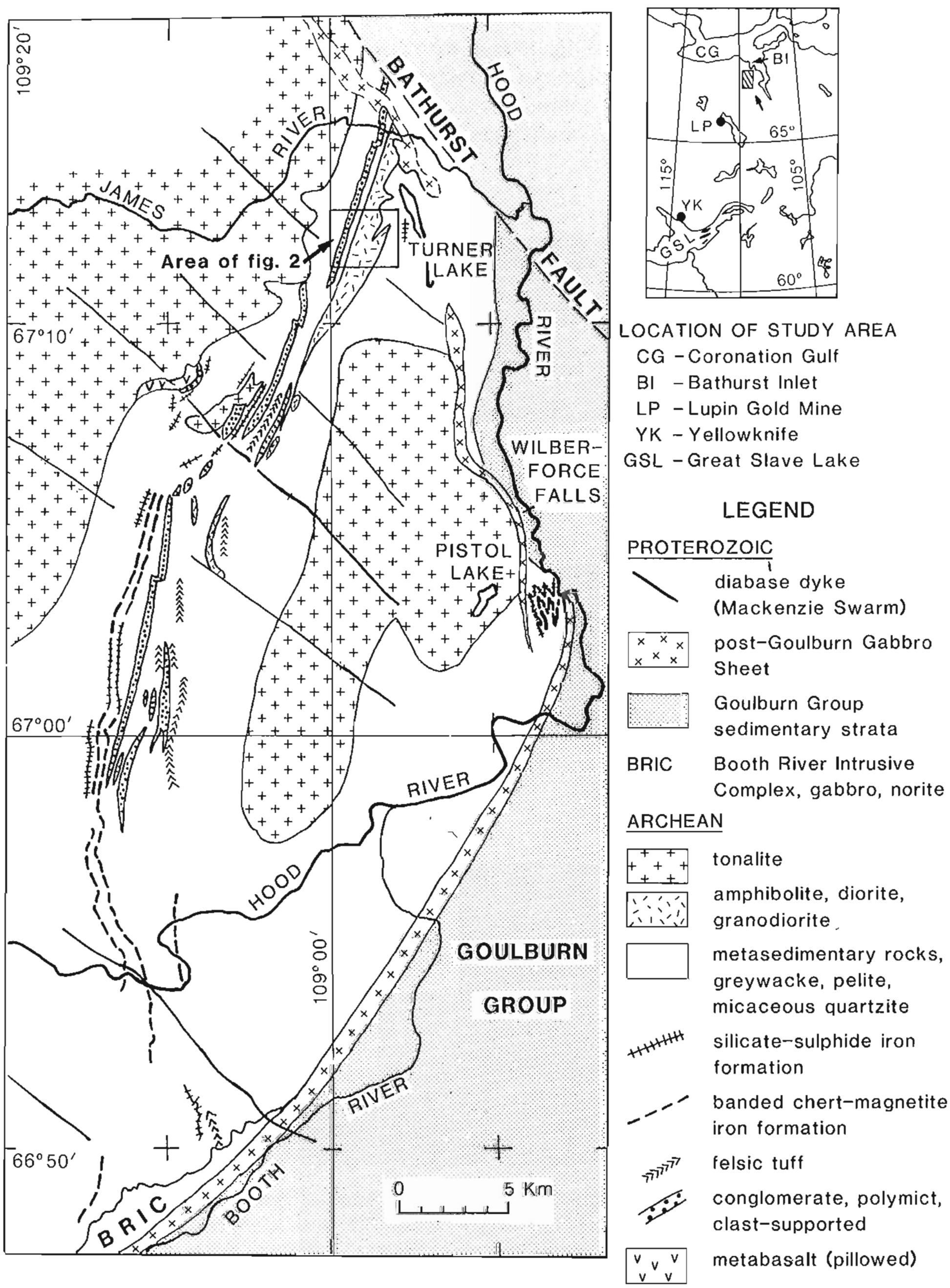

Figure 1. Geology of the Turner Lake belt of Archean supracrustal rocks west of Bathurst Inlet, modified from Geological Survey of Canada Map 45-1963, unpublished maps of Silver Hart Mines Ltd. properties by E.A. Hardy and C. Clode, and a map of Echo Bay Mines Limited claims between latitudes 67°07' and 67°11'.

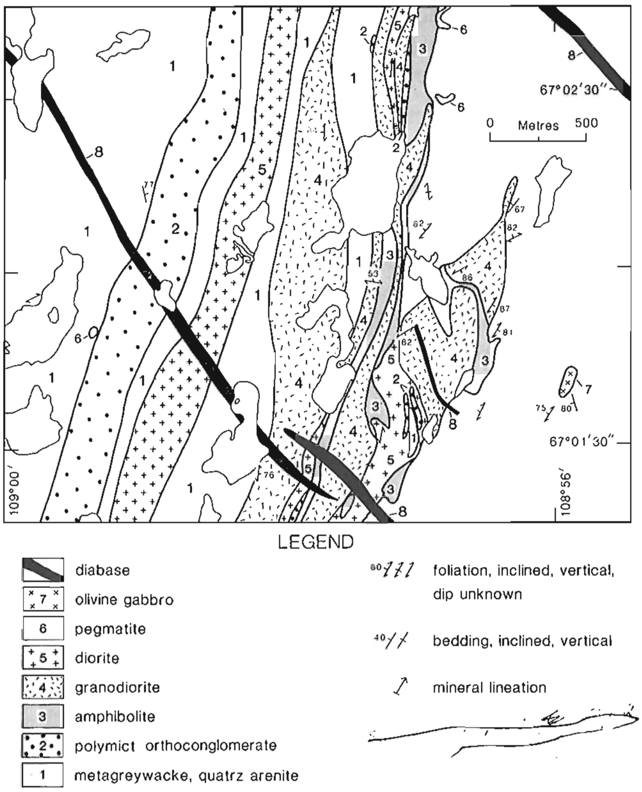

Figure 2. Geology of an area near the north end of the Turner Lake including intermediate and mafic intrusions east of the James Falls conglomerate unit, from detailed map by C. Clode for Silver Hart Mines Ltd.

FELSIC VOLCANIC ROCKS

Thin beds and lenses of light-coloured crystal tuff are common within metasedimentary rocks in the central part of the belt. They are characterized by rounded to elongate white plagioclase crystals that resist weathering, resulting in distinctive outcrop surfaces. The crystals are set in a fine-grained, mottled matrix that contains scattered, irregular patches of biotite. Cherty varieties commonly contain quartz eyes in a pale grey or pink matrix. Lapilli tuff is also present. White aphanitic fragments are stretched (or flattened) to lengths of as much as 50 cm in a greenish-grey matrix containing abundant feldspar crystals. Thin cherty bands containing minor amounts of pyrite are present locally. Euhedral crystals of arsenopyrite are present in tuff in many places and in amounts up to 15%.

IRON-FORMATION

Iron-formations are present in clastic sediments along the entire length of the belt. The most extensive of these are two bands of oxide facies iron-formations that stratigraphically underlie the main polymictic cobble orthoconglomerate unit. Despite their narrow widths (rarely greater than 20 m) they extend apparently continuously along a distance of at least 25 km. Locally, isoclinal folding has produced thicknesses of over 100 m. The oxide iron-formations consist of rythmically interlayered beds millimetres to decimetres thick, of massive or disseminated magnetite with yellowish-green grunerite and recrystallized chert. Thin layers rich in amphibole (other than grunerite) and bands containing small almandine garnets are also present. Proportions of these components vary along strike. The silicate facies increases northward and southward at the expense of the oxide component. One or both iron-formations may grade laterally into chlorite quartz schist at their northern extremities.

Pyrite or pyrrhotite, or both together, are present locally in oxide iron-formation in amounts that may be as great as 10 volume percent. They occur as fine disseminations, veinlets, and, less commonly, as isolated blebs and euhedral crystals in magnetite or in other minerals. Rare arsenopyrite may be associated with the iron sulphides, in most cases as small euhedral crystals and blebs.

Silicate iron-formations, composed dominantly of amphibole and recrystallized chert, are generally thinner and less continuous than oxide iron-formations. They are compositionally banded on a millimetre to centimetre scale.

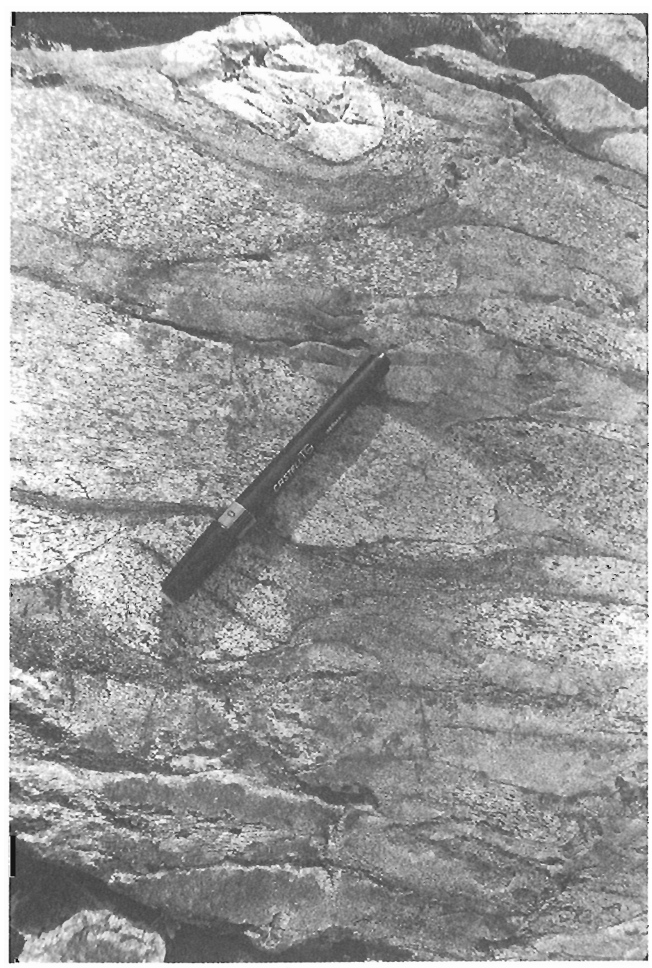

Figure 3. Polymictic cobble orthoconglomerate at James Falls clasts of tonalite, diorite, and quartzose and argillaceous metasedimentary rocks at this locality are amongst the least deformed of any present within the conglomerate unit.

Grunerite is the main amphibole. It is generally fine grained but in places forms radiating clusters more than 2 cm in diameter. Almandine garnets are abundant and in places occur as layers of garnetite. Biotite is a minor component of some silicate iron-formations. Iron sulphides, principally pyrrhotite, are common and may be concentrated in layers. A silicate iron-formation immediately west of the main oxide units at latitude 67°00' is an example of a silicate iron-formation containing sulphide-rich sections. Up to 20 % finely disseminated pyrrhotite is widespread in fine grained amphibolite bands. Pyrite occurs in lesser quantities as fine disseminations, blebs and stringers. Small euhedral arsenopyrite crystals, are present, in amounts not exceeding 5 %, in some sulphide-rich sections.

Near Pistol Lake, silicate-sulphide iron-formation overlies oxide iron-formation and both are folded into an anticline overturned to the east and flanked by synclines. Gold-bearing zones have been found, and partially explored by drilling, in both iron-formations (Silver Hart Mines Ltd.,1985). In the F-zone within the silicate-sulphide iron-formation, the best gold concentrations are believed to be present where there are secondary folds with amplitudes of 50 to 130 m. The G-zone is at the base of the oxide iron-formation. The best widths and gold grades within it occur in discrete shoots that are spatially related, apparently, to fractured zones developed in secondary warps or folds.

AMPHIBOLITE AND ASSOCIATED INTRUSIVE ROCKS

The field term 'amphibolite' has been applied by Silver Hart geologists to a massive amphibole-rich mafic rock associated with intrusions of 'diorite' and granodiorite near Turner Lake in the northern part of the belt. It is finer grained and generally more foliated than these intrusions and in isolated outcrops it might be considered likely to be of extrusive origin (see Roscoe, 1983, Fig.5) It evidently crosscuts other rocks, however, and it is believed to be an intrusion correlative with mafic gabbro (originally pyroxene phyric) that is present farther south. A sample analyzed for Silver Hart Mines Ltd. contained: SiO_2 — 49.21 %; Al_2O_3 — 15.21 %; Fe_2O_3 — 10.59 %; MgO — 10.60 %; CaO — 7.04 %; Na_2O — 2.91 %; K_2O — 0.82 %; TiO_2 — 0.69 %; MnO — 0.22 %. Near the north end of Turner Lake, the amphibolite is altered where it hosts a gold prospect. A sample of the altered rock contained: SiO_2-45.14 %; Al_2O_3-9.09 %; MgO-19.29 %; Na_2O-0.90 %.

Concentrations of pyrrhotite containing small amounts of copper and nickel sulphide are associated with the amphibolite and other mafic intrusive rocks. One of these, north of Turner Lake, is associated with a 1 m diameter concentration of tourmaline surrounding a quartz core. Late carbonate veinlets cut the sulphide-bearing rocks and in three of the prospects they contain small amounts of niccolite. Galena is present in the northernmost of these which is near the James River.

DISCUSSION

The extensive layers of iron-formation and associated formations of well-sorted conglomerate, the presence of beds of homogeneous quartz-rich arenites, and, perhaps, a scarcity of discernible grain gradations in banded argillaceous arenites indicate that great thicknesses of sediments in the Turner Lake belt were deposited under conditions incompatible with deposition of turbidites.

Two distinctly different types of iron-rich sediments were deposited in the area; under similar imposed environments of diagenesis and metamorphism, one developed into oxide iron-formation, the other into silicate-sulphide iron-formation. Conditions favourable for deposition of one or the other apparently alternated through short time intervals and across short distances in the depository.

Epigenetic gold concentrations are present in intrusive rocks and volcanic rocks near two types of iron-formation-hosted gold concentrations. Epigenesis probably played a major role in the formation of concentrations in structurally controlled, presumably fractured zones, in oxide iron-formations. Its role in the formation or modification of gold concentrations in sulphidic layers in silicate iron-formations is less certain. Similar styles of iron-formation-hosted gold mineralization (non-stratiform and stratiform) occur at the global scale (Kerswill and Caddey, 1987) and have been recognized in the Contwoyto Lake area (Kerswill, 1986).

Once formed, whether under low grade or moderate grade metamorphic conditions, the two types of iron-formation-hosted gold deposits are apparently stable chemically during further increases in metamorphic grade.

ACKNOWLEDGMENTS

We thank Silver Hart Mines Ltd. and Echo Bay Mines Limited for permission to publish information on the geology of the Turner Lake belt that was obtained from internal company reports. Mining claims held by Silver Hart Mines cover the bulk of the geologically most interesting sections of the belt. Their properties were mapped mainly by E.A. Hardy and C. Clode assisted by S. Bishop. Roscoe's field studies in 1987 were funded by the Canada-Northwest Territories Mineral Development Agreement. He is indebted to his co-authors and to Glen and Trish Warner of Bathurst Inlet Lodge for help and courtesies that made his field work possible. Figures 1 and 2 were draughted by Kim Nuguyen. The paper was reviewed by O.R. Eckstrand resulting in a number of changes that were necessary or useful.

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