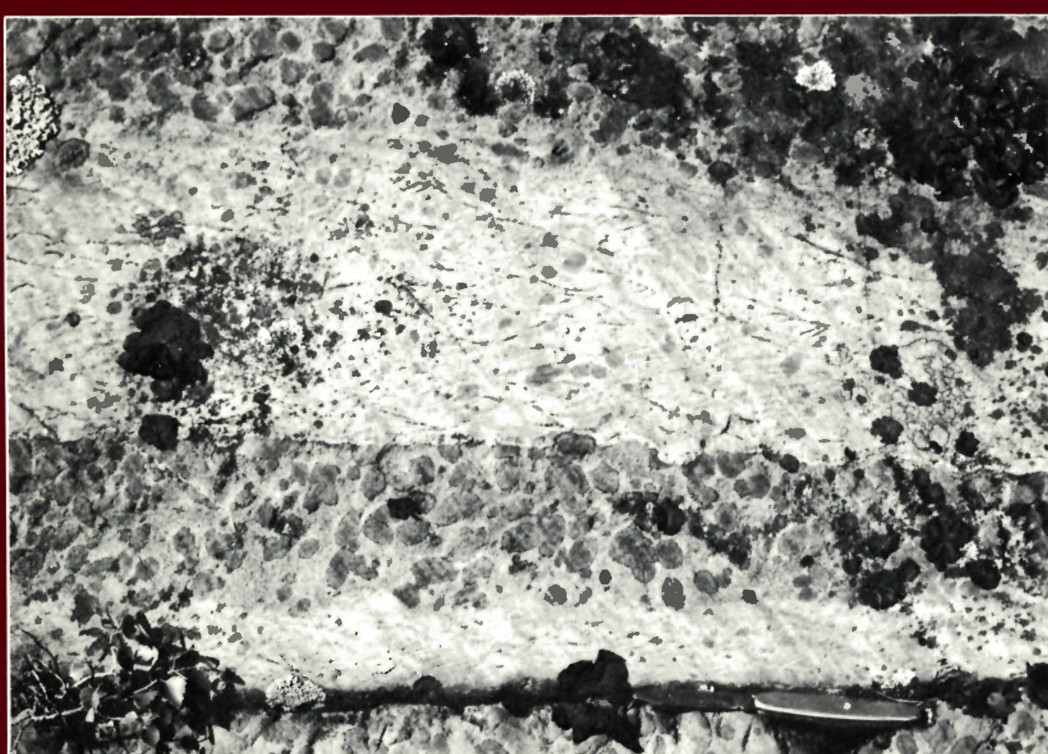
Geological Survey of Canada
Commission géologique du Canada

MEMOIR 414

GEOLOGY OF THE YELLOWKNIFE-HEARNE LAKE AREA, DISTRICT OF MACKENZIE: A SEGMENT ACROSS AN ARCHEAN BASIN

J.B. HENDERSON



Canada

1985

GEOLOGICAL SURVEY OF CANADA
MEMOIR 414

GEOLOGY OF THE YELLOWKNIFE-HEARNE LAKE AREA, DISTRICT OF MACKENZIE: A SEGMENT ACROSS AN ARCHEAN BASIN

J.B. HENDERSON

1985

Available in Canada through

authorized bookstore agents and other bookstores

or by mail from

Canadian Government Publishing Centre
Supply and Services Canada
Ottawa, Canada K1A 0S9

and from

Geological Survey of Canada offices:

601 Booth Street
Ottawa, Canada K1A 0E8

3303-33rd Street N.W.,
Calgary, Alberta T2L 2A7

A deposit copy of this publication is also available
for reference in public libraries across Canada

Cat. No. M46-414E Canada: $13.50
ISBN 0-660-11942-0 Other countries: $16.20

Price subject to change without notice

Cover photo

Geological evolution of the Yellowknife-Hearne Lake area in microcosm: delicate climbing ripple laminations in Archean turbidite beds overgrown by metamorphic cordierite porphyroblasts in an outcrop striated during Quaternary glaciation. (GSC 178141)

Critical Reader

K.C. Card

Original manuscript received: 1981 - 02 - 23
Final version approved for publication: 1983 - 04 - 22

Preface

The Yellowknife area is one of the major gold producing areas in Canada. Gold was first discovered in 1898 but it was not until 1934 that serious exploration commenced. It was first mined in 1938 and is currently produced at mines on shear zones in mafic volcanic rocks near Yellowknife. There has also been considerable interest in lithium, tungsten, rare-earth elements, base metal and nickel deposits, although there has been only limited if any production of these. This region has been the most productive in the Northwest Territories due almost entirely to the production of gold.

The Geological Survey has been working in this southernmost part of the Slave Province since the 1930s. This report gives the results of a study of the 22 000 km^2 Yellowknife-Hearne Lake region and incorporates some results of detailed studies by other workers.

Although in part covered by Phanerozoic and Proterozoic rocks this complex area has remained essentially stable since the close of the Archean era 2500 million years ago. Magnificent exposures provide a unique opportunity to study the development of Archean sediments deposited in an ensialic rift basin that is bordered by thick accumulations of volcanic rocks.

The detailed stratigraphic, structural and metamorphic relationships and the economic geology described in this memoir will be of great value to those concerned with the geology and mineral potential of this part of the Canadian Shield.

R.A. Price
Director General
Geological Survey of Canada

Préface

La région de Yellowknife est l'une des grandes zones productrices d'or du Canada. C'est en 1898 qu'a été découvert le premier gisement aurifère de la région, mais l'exploration minière à proprement parler n'y a commencé qu'en 1934. Extrait pour la première fois en 1938, l'or provient actuellement de mines exploitées dans les zones de cisaillement des roches volcaniques mafiques situées près de Yellowknife. Les gisements de nickel, de métaux communs, de terres rares, de tungstène et de lithium suscitent également beaucoup d'intérêt bien que la production de ces minéraux, si tant soit telle, a été limitée jusqu'à maintenant. Cette région est la plus productive des Territoires du Nord-Ouest, ce qu'elle doit presque entièrement à sa production d'or.

Dans le cadre de ses travaux, la Commission géologique oeuvre dans cette zone, partie la plus méridionale de la province des Esclaves, depuis les années 30. Le présent rapport donne les résultats d'une étude portant sur la région de Yellowknife et du lac Hearne (superficie de 22 000 km^2) et intègre certains résultats d'études détaillées réalisées par d'autres chercheurs.

Même si elle est en partie recouverte de roches du post-Précambrien et de l'Algonkien, cette région à structure complexe affiche une très grande stabilité géologique depuis la fin de l'Archéen, il y a 2 500 millions d'années. Les affleurements de toute beauté qu'on y trouve offrent une occasion unique d'étudier le processus d'accumulation des dépôts sédimentaires de l'Archéen dans un bassin de failles ensialiques délimité par d'épaisses accumulations de roches volcaniques.

Les relations stratigraphiques, structurales et métamorphiques de la région et la géologie économique que décrit le mémoire deviendront un précieux outil de travail à tous ceux qui s'intéressent à la géologie et au potentiel minéral de cette partie du Bouclier canadien.

Le directeur de la
Commission géologique du Canada
R.A. Price

CONTENTS

Tables

Illustrations

GEOLOGY OF THE YELLOWKNIFE-HEARNE LAKE AREA, DISTRICT OF MACKENZIE: A SEGMENT ACROSS AN ARCHEAN BASIN

Abstract

The Yellowknife-Hearne Lake area is situated in the southernmost Slave Structural Province and underlies an area of about 22 000 km^2.

The oldest rocks are metamorphosed and deformed granitoid plutons and gneisses that are basement to the Archean Yellowknife Supergroup. The Yellowknife supracrustal rocks occur in the remnant of an ensialic rift basin, whose margins are marked by thick accumulations of mainly basaltic to andesitic, subaqueously extruded volcanic rocks. Smaller felsic volcanic complexes occur both separately, and in close association with, the mafic volcanics. The main part of the basin is filled with greywacke-mudstone turbidites of dominantly mixed felsic volcanic and granitoid provenance. Conglomerate and shallow-water lithic to quartz-rich sandstones occur locally at the basin margins. Minor limestone, dolomite, and silicate and oxide iron formation are associated with some volcanic complexes. The Yellowknife rocks were complexly folded during at least two phases of deformation and have been metamorphosed to varied degrees in a low pressure series from greenschist to middle amphibolite grade. Both the basement and Yellowknife supracrustal rocks have been intruded by a varied assemblage ranging from synvolcanic epizonal plutons to weakly foliated, concordant plutonic complexes, to massive, sharply discordant intrusive bodies. Compositions range from quartz diorite to potassic granite. Gabbroic sills contemporaneous with the volcanics occur in abundance in the mafic volcanic complexes and locally in the metasediments.

Intrusive into the Archean rocks is the early Proterozoic Blachford Lake Intrusive Suite, a series of alkalic plutons ranging from gabbro to granite. Unconformably overlying the Archean is part of the early Proterozoic Great Slave Supergroup shelf carbonate to basinal sediment transition that has been telescoped by thrust faulting. Four sets of early Proterozoic diabase dykes or sheets and one set of middle Proterozoic diabase occur in various parts of the area. Lying unconformably on the Precambrian south of Great Slave Lake is a gently dipping homocline of sandstones, shales, carbonates and evaporites of Cambrian, Ordovician and Devonian age, each separated by unconformities.

Gold is the most important mineral commodity of the area with two major mines operating on shear zones in the mafic volcanics at Yellowknife. Smaller quartz-vein gold deposits occur in the metasediments. Other commodities that have attracted interest include lithium, beryllium, tantalum and columbium in pegmatites; tantalum, columbium and rare-earth elements in a Blachford Lake alteration zone; tungsten in quartz veins in the metasediments, base metals and nickel.

Résumé

La région de Yellowknife et du lac Hearne, d'une superficie de quelque 22 000 km^2, est située à l'extrémité sud de la province des Esclaves.

Les roches les plus anciennes sont des gneiss et des intrusions ignées granitoïdes métamorphisés et déformés sous-jacents au supergroupe de Yellowknife datant de l'Archéen. Les roches de Yellowknife en surface se trouvent dans les vestiges d'un bassin d'effondrement ensialique qui est bordé d'épaisses accumulations de roches volcaniques mises en place sous l'eau et constituées principalement de basalte et d'andésite. Des complexes volcaniques silico-feldspathiques moins importants existent à titre distinct ou sont intimement liés à des formations volcaniques mafiques. La partie principale du bassin est composée de turbidites à grauwackes et à mudstones qui proviennent de formations constituées principalement de roches volcaniques et granitoïdes silico-feldspathiques. Des conglomérats et des grès d'eau peu profonde qui varient du grès lithique au grès riche en quartz forment des dépôts circonscrits sur les bords du bassin. Des formations mineures composées de calcaire, de dolomite et de fer silicaté et oxydé sont associées à quelques complexes volcaniques. Les roches du Yellowknife ont été déformées par des plis complexes, à deux reprises au moins, et métamorphisées à des degrés divers sous l'effet d'une faible pression: le métamorphisme varie du faciès des schistes verts au degré moyen de l'amphibolite. Diverses associations, allant d'intrusions ignées épizonales contemporaines de phénomènes volcaniques à des complexes plutoniques concordants à feuilletage peu développé, en passant par des massifs intrusifs nettement discordants, se sont introduites dans les roches du socle et dans les roches en surface du Yellowknife. La composition varie de la diorite quartzifère au granite potassique. Il y a de nombreux filons-couches gabbroïques de même âge que les roches volcaniques dans les complexes volcaniques mafiques et, par endroits, dans les roches sédimentaires métamorphisées.

Les roches de l'Archéen ont été pénétrées par la série intrusive de Blachford Lake qui date du début du Protérozoïque et est composée d'intrusions ignées basiques variant du gabbro au granite. Sur l'Archéen, repose en discordance une partie des formations du supergroupe de Great Slave datant du début du Protérozoïque. Ces formations, déformées par des failles de chevauchement, représentent la transition entre les faciès carbonatés du plateau continental et les faciès sédimentaires profonds. Quatre séries de dykes à diabase ou de silts datant du début du Protérozoïque et une série à diabase datant du Protérozoïque moyen se retrouvent à plusieurs endroits dans la région. Au sud du Grand lac des Esclaves, repose en discordance sur le Précambrien une structure monoclinale à pendage faible constituée de grès, de schistes argileux, de roches carbonatées et d'évaporites datant du Cambrien, de l'Ordovicien et du Dévonien, chaque période étant séparée par des discordances.

L'or est au premier rang des matières premières minérales dans la région: deux mines importantes sont exploitées dans les zones de cisaillement des formations volcaniques mafiques à Yellowknife. Il y a des gisements d'or de moindre importance dans les veines de quartz qui sillonnent les roches sédimentaires métamorphisées. Parmi les autres matières premières présentant un intérêt, mentionnons le lithium, le béryllium, le tantale et la columbium que l'on retrouve dans des pegmatites; le tantale, le columbium et des éléments des terres rares qui se trouvent dans la zone d'altération de la série de Blachford Lake; le tungstène contenu dans les veines de quartz des roches sédimentaires métamorphisées des métaux communs et le nickel.

INTRODUCTION

Within the borders of the Yellowknife-Hearne Lake area a well preserved segment across an Archean supracrustal basin provides an unusual opportunity to examine the evolution of an Archean basin within the Slave Structural Province.

The Yellowknife-Hearne Lake area includes NTS map sheets 85 J (Yellowknife) and 85 I (Hearne Lake), and is situated in the District of Mackenzie of the Northwest Territories along the north shore of Great Slave Lake, between 62 and 63° north latitude and 112 and 116° west longitude (Fig. 1, and map, in pocket). The area, mapped at 1:250 000 scale, covers about 22 000 km^2.

Yellowknife, the capital city of the Northwest Territories, is located on the west shore of Yellowknife Bay. There is good access into the area with frequent scheduled airline service to Yellowknife from Edmonton, and Winnipeg. An all-weather road west and north of Great Slave Lake joins Yellowknife to Hay River on the south shore of Great Slave Lake and continues south to Edmonton. There is also a road east from Yellowknife to Tibbit Lake, a distance of about 60 km. Access to other parts of the area on the ground is possible on winter roads but at other times of the year is best accomplished by float-equipped aircraft available at Yellowknife. Of the rivers in the area only the Snare system (Russell Lake) and the Yellowknife River are possible canoe routes, although many portages are involved in the latter. Both these river systems extend north to the height of land where one can portage into the Coppermine system to the Arctic coast. Other rivers in the area are small and contain too many rapids to be practical canoe routes.

The vicinity of Yellowknife is one of the major gold producing areas in the country. Gold was first mined in 1938 and is currently (1983) produced at the Con and Giant Yellowknife mines at Yellowknife. Since the late 1930s over a dozen gold mines have operated within the map area for varied lengths of time. There has also been considerable interest in lithium, tungsten, rare-earth elements, base metal and nickel deposits in the area, although there has only been very limited if any production of any of these commodities.

The terrain is typical of much of the Canadian Shield – generally rather flat and featureless. The areas underlain by volcanic rock, particularly the more felsic volcanic rock units as at Russell Lake, tend to have greater relief than the granitic terrane or areas underlain by sedimentary rocks. The area underlain by flat lying Paleozoic rocks southwest of North Arm, Great Slave Lake, is particularly flat and featureless. The elevations range from a minimum of 156 m above sea level on Great Slave Lake and gradually increase toward the northeast corner of the area to a height of about 400 m and toward the southwest to a height of just over 200 m. Within the area underlain by Precambrian rocks, maximum local relief is of the order of 10 to 20 m.

Drainage throughout the area is south into Great Slave Lake, which is drained by the Mackenzie River. The main rivers from west to east are the Snare (Russell Lake) which empties into the end of North Arm, Great Slave Lake, and the Yellowknife and Cameron rivers which empty into Yellowknife Bay on North Arm, Great Slave Lake. Lakes are abundant with approximately 27% of the area underlain by water (Fremlin, 1974). Some of the lakes that appeared normal in air photos taken in 1945 had completely dried up prior to 1970. Others are greatly reduced in size and are rimmed by a white deposit of monohydrocalcite (Fig. 2). Upland Lake 32 km east of Yellowknife is an example of this sort of lake.

Previous work and sources of information

Various parts of the area have been mapped, mainly in the 1930s to early 1950s, at scales ranging from 500 feet to the inch to 8 miles to the inch. Subsequent to the mapping discussed in the present report, the Exploration and Geological Services Unit of the Department of Indian and Northern Affairs has undertaken a detailed mapping program in the immediate Yellowknife area. Previous and recent geological mapping in the Yellowknife-Hearne Lake area is outlined and listed in Figure 3 and Table 1.

Field methods and extent of mapping

Fieldwork was undertaken between 1970 and 1972 with brief visits to the area in 1974 and 1976. Most of those parts of the area underlain by Archean rocks were traversed on foot from the larger lakes in the area and from roads where available, with a line spacing of approximately 3 km. Areas where previous, more detailed mapping had been done (Fig. 3, Table 1), although not systematically remapped, were traversed for purposes of familiarization. The more extensive granitoid terranes remote from the larger lakes, such as the area north of the Great Slave Highway, the vacinity of Defeat Lake, Meander Lake, Payne Lake and northeast of Morose Lake, were mapped with the aid of a helicopter with landings of about 4 km spacing. Areas underlain by younger rocks such as the Aphebian Blachford Lake Intrusive Suite north of Hearne Channel and the Aphebian Great Slave Supergroup to the south, which were being mapped in greater detail than the present scale at the time, and the Paleozoic rocks southwest of North Arm, Great Slave Lake, which had been mapped previously, were not mapped in the course of this project.

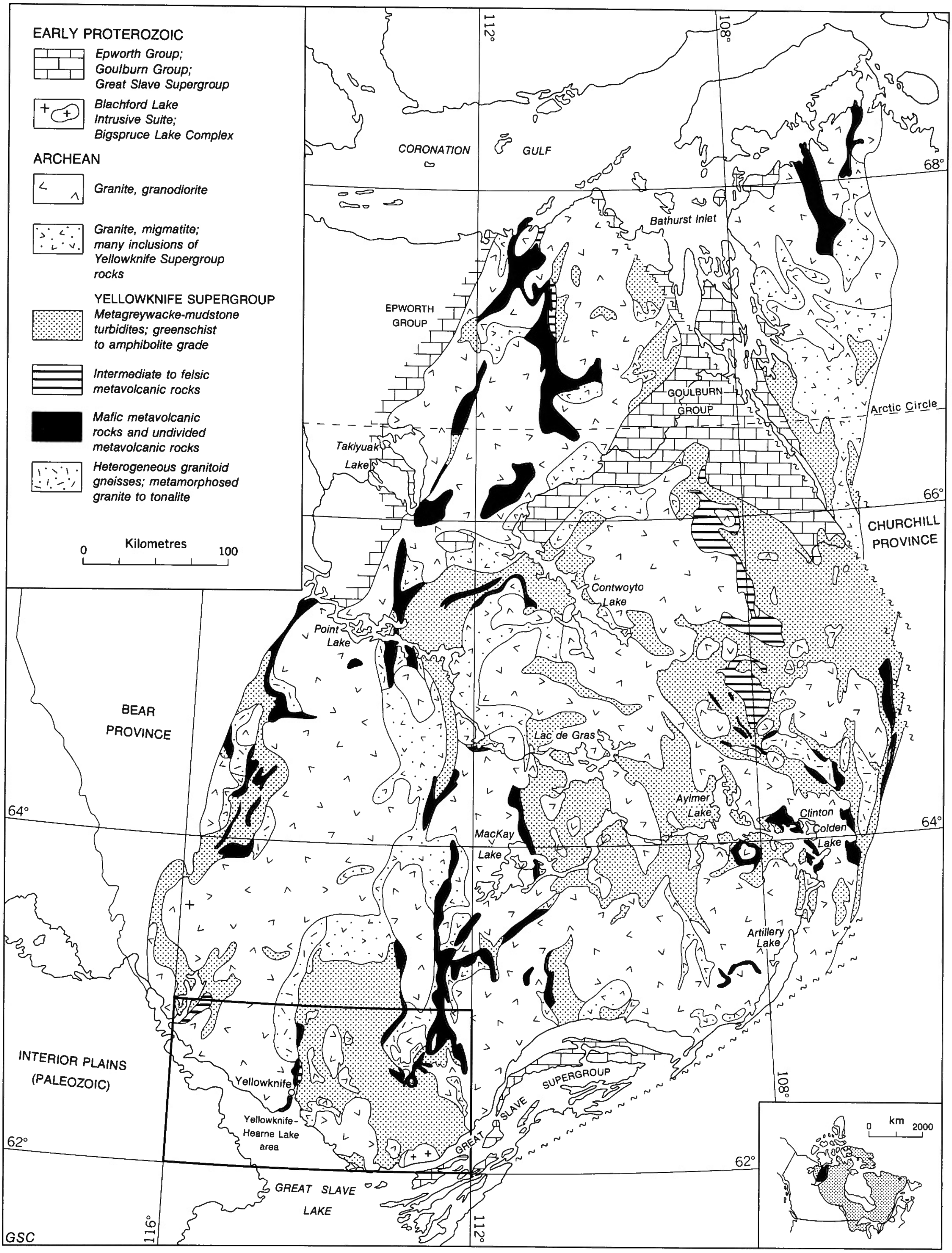


Figure 1. General geology of the Slave Structural Province. The Yellowknife-Hearne Lake area is outlined in the southern part of the province. Modified after McGlynn (1977).

Figure 2. *A white deposit of monohydrocalcite rims shoreline rocks and covers beaches of a poorly drained small lake between Hearne and Campbell lakes. Similar deposits are seen around several other poorly drained lakes in the area whose level has dropped significantly over the last few decades. GSC 177715*

Acknowledgments

The author was ably assisted in the field by A.J.M. Elliot (geologist), D. Cook and G. Dodds (geological assistants) and R. Laing (cook) in 1970, by M.P. Cecile and D.C. Kamineni (geologists), L. Hartwig, Y. Legault and H. Moore (geological assistants) and M. Curley (cook) in 1971, and by M.B. Lambert, G.R. Peeling (geologists), P. Nadeau, R. Troyer and G. Wright (geological assistants) and L. Moss (cook) in 1972. Camp moves in all field seasons were by Wardair Canada Ltd. and in 1972 the helicopter provided by Liftair International Ltd. was flown by G. Causey, with E. Godleski as engineer.

The author and this report have benefited from numerous discussions with colleagues, in particular Mikkel Schau, concerning Slave Province geology. B.G. Craig (Quaternary geology), A. Davidson (Blachford Lake Intrusive Suite), P.F. Hoffman (Great Slave Supergroup), M.B. Lambert (Yellowknife metavolcanics), P.H. McGrath (geophysics), S.M. Roscoe (economic geology), P.H. Thompson (structural and metamorphic geology), and G.K. Williams (Paleozoic geology), who read various parts of the report in which they had particular expertise, are thanked for their interest and assistance. P.H. Thompson collected metamorphic assemblage data in the course of a regional review of metamorphism in the Slave Province. O. van Breemen recalculated, where necessary, much of the previously published geochronological data used in this report using the presently accepted decay constants proposed by Steiger and Jaeger (1977) for potassium-argon, rubidium-strontium and uranium lead systems. K.D. Card critically read the complete report which has been significantly improved as a result.

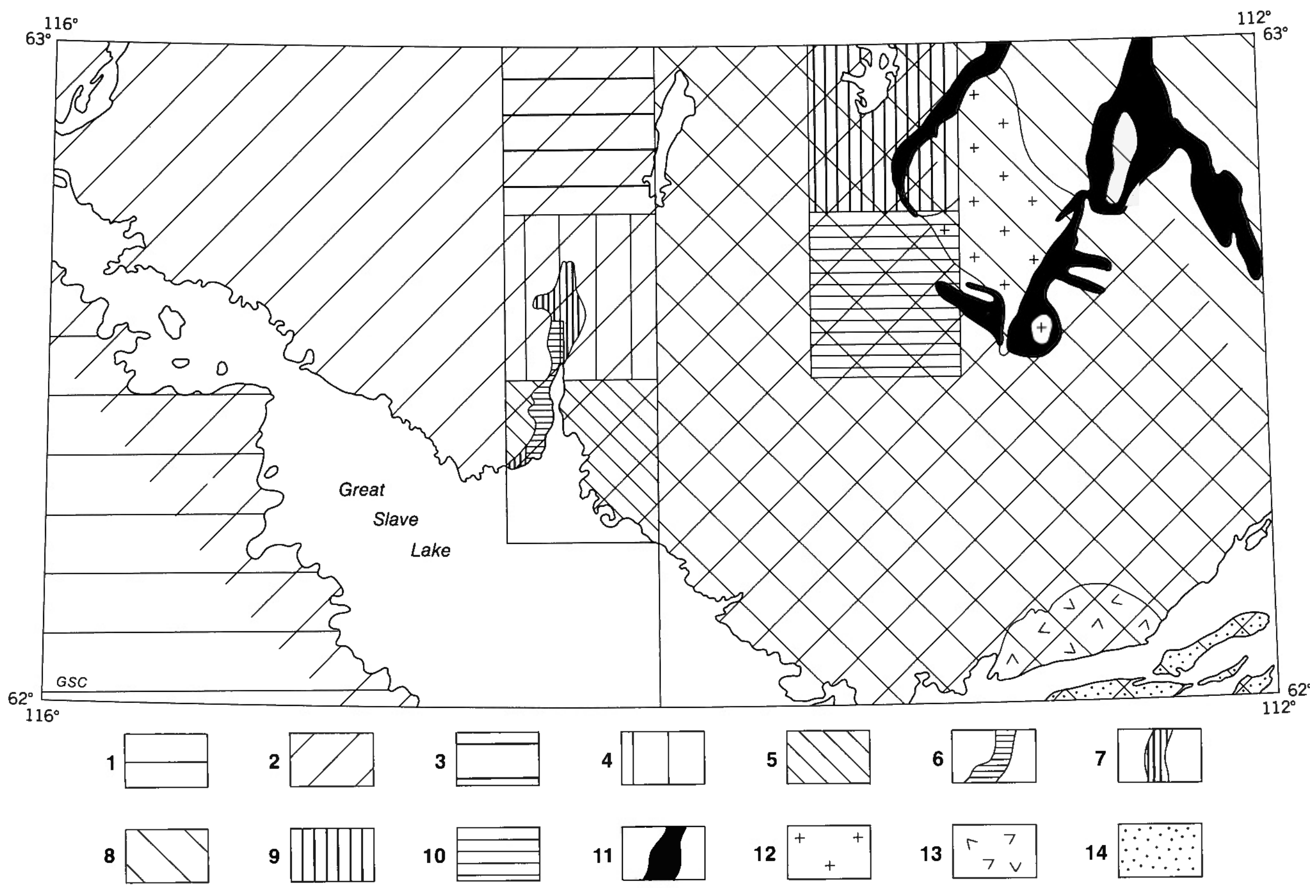


Figure 3. *Location of previous and recent geological mapping within the Yellowknife-Hearne Lake area. See Table 1 for identification, authorship and scale of the various areas.*

Table 1. Previous and recent mapping in the Yellowknife-Hearne Lake area

No.*	Area	Reference	Scale
1	Southwest of North Arm, Great Slave Lake	Douglas and Norris (1960) Richmond (1965)	1 in = 8 mi 1 in = 8 mi
2	Most of Yellowknife and Hearne Lake areas	Jolliffe (1936)	1 in = 8 mi
3	Quyta Lake map area	Jolliffe (1939)	1 in = ½ mi
4	Prosperous Lake map area	Jolliffe (1946)	1 in = 1 mi
5	Yellowknife Bay map area	Jolliffe (1942)	1 in = 1 mi
6	Yellowknife greenstone belt	Henderson and Brown (1966, 1948, 1949, 1950, 1952a, b)	1:12 000 1 in = 500 ft
7	Yellowknife greenstone belt	Helmstaedt et al. (1979) Hauer (1979) Helmstaedt et al. (1980) Easton and Jackson (1981)	1:10 000 1 in =700 ft 1:9600 1:10 000
8	Hearne Lake area	Henderson and Jolliffe (1941)	1 in = 4 mi
9	Gordon Lake area	Henderson (1941b)	1 in = 1 mi
10	Ross Lake area	Fortier (1947)	1 in = ½ mi
11	Cameron River and Beaulieu River volcanic belts	Lambert (in press)	1:50 000
12	Plutonic and gneissic area between Cameron and Beaulieu rivers	Davidson (1972) (and unpublished)	1:300 000 (1:50 000)
13	Blachford Lake area	Davidson (1981)	1:50 000
14	Blanchet Island area	Stockwell (1936) Hoffman (1977)	1 in = 4 mi 1:50 000

*Numbers keyed to areas on Figure 3

Finally acknowledgment is made to the previous and current workers in the area whose maps, reports and in some cases field notes, formed an important source of information for the compilation of the present report.

GENERAL GEOLOGY

The Yellowknife-Hearne Lake area is situated in the southernmost part of the Slave Structural Province which, aside from Proterozoic faulting, folding as indicated by the broad open structure of the early Proterozoic Goulburn Group (Fraser, 1964) and minor local intrusions, has remained essentially stable since the close of the Archean era 2.5 Ga ago (Fig. 1). The supracrustal rocks of Archean age are part of the Yellowknife Supergroup which includes all Archean supracrustal rocks within the Slave Province (Henderson, 1970; see Table of Formations). Within the map area, the Yellowknife Supergroup consists of major units of dominantly mafic volcanic rocks and minor units of felsic volcanics that occur as separate bodies or are intermixed with the mafic volcanics. Metasedimentary rocks of the Yellowknife Supergroup underlie the largest area of any map unit. The vast majority of these sediments are greywacke-mudstone turbidites that are the main basinal fill. In addition there are units of conglomerate, crossbedded lithic sandstone, dolomite, limestone and silicate and oxide iron formation. The Yellowknife rocks represent deposition under tectonically unstable conditions. Within the map area two probable basin margins are preserved, as shown by the presence of alluvial sediments. Others may be present as well, although the evidence is less definitive. Only a segment across the basin is preserved within the map area as the basin can be followed to the north (Fig. 1) and presumably its remnants exist to the south underneath Proterozoic and Phanerozoic rocks and Great Slave Lake.

There is evidence that the Yellowknife rocks lie unconformably above a granitic to gneissic basement with a long complex intrusive and deformational history (see Table of Formations).

The Yellowknife rocks have been complexly folded during at least two phases of deformation and have been regionally metamorphosed to varied degrees ranging from greenschist to middle amphibolite facies. The metamorphism is of a lower pressure facies series.

The supracrustal and basement rocks have been intruded at different times and under different conditions by a varied assemblage of granitic rocks (see Table of Formations). Archean intrusions range from massive to weakly foliated concordant plutonic complexes to massive, sharply discordant, intrusive bodies. Composition of these intrusions ranges from granodiorite to potassic granite. In addition, there are a few small plutons of gabbro to diorite intrusive into a granodioritic intrusive complex. Extensive thick contemporaneous gabbroic sills have intruded the Yellowknife sedimentary rocks locally as well as the volcanic rock where in places they are very abundant.

TABLE OF FORMATIONS

EON	ERA	PERIOD / EPOCH	GROUP / SUITE	FORMATION / LITHODEME	MAP SYMBOL	LITHOLOGY
PHANEROZOIC	CENOZOIC	QUATERNARY			Qu	Unconsolidated gravel, sand and silt
	Unconformity					
	PALEOZOIC	MIDDLE DEVONIAN		HORN RIVER	DHR	Dark brown shale and grey limestone
				LONELY BAY	DLB	Brown to grey limestone
				CHINCHAGA	DC	White to dark grey gypsum
				FITZGERALD	DF	Light brown dolomite
				MIRAGE POINT	DMP	Dark red dolomite
		Unconformity				
		ORDOVICIAN		CHEDABUCTO LAKE	OCL	Brown dolomite
		Unconformity				
		CAMBRIAN		LA MARTRE FALLS	ЄLMF	Green shale
				OLD FORT ISLAND	ЄOFI	Quartzose sandstone
Unconformity						
PROTEROZOIC	MIDDLE PROTEROZOIC			MACKENZIE	Pm	Diabase dykes
	Intrusive contact					
	EARLY PROTEROZOIC		COMPTON INTRUSIVE SUITE		PCI	Diorite and quartz monzonite laccoliths
			Intrusive contact			
			GREAT SLAVE SUPERGROUP			
			CHRISTIE BAY	STARK	PS	Mudstone and carbonate breccia
			PETHEI (PLATFORMAL FACIES NORTH)	HEARNE	PH	White to light grey laminated limestone
				WILDBREAD	PW	Grey stromatolitic limestone
				UTSINGI	PU	Blue-grey discontinuously laminated limestone
				TALTHEILEI	PT	Light brown stromatolitic dolomite
			PETHEI (BASINAL FACIES SOUTH)	PEKANATUI POINT	PPP	Light grey aphanitic limestone
				BLANCHET	PB	Greywacke-mudstone with limestone
				McLEAN	PM	Limestone and mudstone
			PETHEI	DOUGLAS PENINSULA	PDP	Marlstone with argillaceous limestone and mudstone
			KAHOCHELLA	CHARLTON BAY	PCB	Dark green shale
				McLEOD BAY	PMB	Red shale
			SOSAN	SETON	PSE	Basalt and rhyolite with sandstone and shale
				KLUZIAI	PK	Pink to grey feldspathic sandstone
			Stratigraphic position relative to Great Slave Supergroup not defined			
					Psc	Sandstone, arkose, conglomerate
			Unconformity			
				HEARNE	Ph	Diabase dykes
			Intrusive contact			
			BLACHFORD LAKE INTRUSIVE SUITE	THOR LAKE SYENITE	PTL	Dark green fayalite-hedenbergite and ferrorichterite syenite
				GRACE LAKE GRANITE	PGL	Light grey to greenish grey riebeckite granite
				MAD LAKE GRANITE	PML	Pink granite
				HEARNE CHANNEL GRANITE	PHC	Pinkish brown hornblende granite
				WHITEMAN LAKE QUARTZ SYENITE	PWL	Green hornblende-quartz syenite

GSC

TABLE OF FORMATIONS (Continued)

EON	ERA	PERIOD / EPOCH	GROUP / SUITE	FORMATION / LITHODEME	MAP SYMBOL	LITHOLOGY
PROTEROZOIC	EARLY PROTEROZOIC		BLACHFORD LAKE INTRUSIVE SUITE	CARIBOU LAKE GABBRO	ƤCL	Gabbro
			Age of dykes relative to Great Slave Supergroup and Blachford Lake Intrusive Suite not defined			
				INDIN	Ƥi	Diabase dykes
				MILT	Ƥm	Diabase sheets
				DOGRIB	Ƥd	Diabase dykes
Intrusive contact						
ARCHEAN			Relative ages of following six granitoid units not defined			
				DUCKFISH GRANITE	ADU	Mauve-pink, equigranular biotite granite
				MOROSE GRANITE	AM	White to pink to brown inequigranular biotite-muscovite granite
				REDOUT GRANITE	AR	Pink, equigranular, heterogeneous biotite-muscovite granite with pegmatite
				PROSPEROUS GRANITE	AP	White to buff, equigranular, homogeneous biotite-muscovite granite with abundant pegmatite
			MEANDER LAKE PLUTONIC SUITE		AML	Heterogeneous biotite, locally biotite-muscovite, granite to granodiorite
			AWRY PLUTONIC SUITE		AA	Light pink, heterogeneous equigranular, biotite granite to granodiorite; pegmatite common
			STAGG PLUTONIC SUITE		AS	Grey to dark pink, equigranular to coarsely porphyroblastic, biotite, locally hornblende granite to tonalite
			DEFEAT PLUTONIC SUITE		AD	White to pink, homogeneous, equigranular, biotite rarely hornblende granodiorite to tonalite
				WOOL BAY QUARTZ DIORITE	AWB	Dark grey to black, equigranular quartz diorite
				DETOUR GRANODIORITE	ADT	Grey, metamorphosed, porphyritic granodiorite
			The following two units are approximately contemporaneous with units of the Yellowknife Supergroup			
				AMACHER GRANITE	AAM	Pale pink to grey, homogeneous, metamorphosed biotite granite
					Aa	Dark green amphibolite sills and small plugs
			Intrusive contact			
			YELLOWKNIFE SUPERGROUP (Stratigraphic order not implied in following sequence of units)			
			DUNCAN LAKE	BURWASH	ABg, ABa	Metagreywacke, siltstone and mudstone at greenschist grade ABg and psammitic to pelitic schist at amphibolite grade ABa
				JACKSON LAKE	AJL	Sandstone and conglomerate
				RAQUETTE LAKE	ARQ	Quartzite sandstone and conglomerate
			BEAULIEU		Acb	Dolomite limestone and calc-silicate gneisses
				PAYNE LAKE	APA	Undifferentiated volcaniclastic metasedimentary rocks
				BANTING, SHARRIE TOWNSITE FLOWS, TURNBACK, OTHER UNNAMED UNITS	Af	Felsic volcanic rocks
				WEBB LAKE, ALICE	AWi, AALi	Intermediate volcanic rocks
				KAM, DUCK, DOME LAKE, CAMERON RIVER, TUMPLINE, SUNSET LAKE, OTHER UNNAMED UNITS	Am	Mafic volcanic rocks
			Unconformity and / or intrusive contact			
			ANTON COMPLEX		AAN	Foliated granodiorite, tonalite granite and gneisses with unmetamorphosed intrusive phases
			SLEEPY DRAGON COMPLEX		ASD	Weakly to strongly foliated granodiorite, tonalite, granite and gneisses

GSC

The Blachford Lake Intrusive Suite (Davidson, 1972, 1978a, 1982; see Table of Formations) is an alkalic complex that consists of a series of intrusions that range in composition from gabbro to granite. These plutons intruded the stabilized Slave craton during the early Proterozoic.

To the southeast the Archean rocks are unconformably overlain by gently dipping early Proterozoic Great Slave Supergroup rocks of the Athapuscow Aulacogen (Hoffman, 1973a,b; see Table of Formations). The Precambrian Shield in the southwest part of the area is unconformably overlain by Paleozoic sediments (Douglas and Norris, 1960; Douglas et al., 1974; see Table of Formations).

Within the map area there are four major Proterozoic diabase dyke swarms (see Table of Formations). They consist of the east-northeast-trending Dogrib dykes which are the oldest, the northeasterly trending and complementary northwesterly trending Indin dykes, the northeasterly trending Hearne dykes and the youngest set, the north-northwesterly trending Mackenzie swarm. All the Archean rocks, the Blachford Lake Intrusive Suite and the Aphebian diabases, but not the Mackenzie dykes, are offset by a series of short displacement north-northwesterly trending left lateral transcurrent faults.

ARCHEAN PRE-YELLOWKNIFE SUPERGROUP GNEISSIC AND PLUTONIC ROCKS

Sleepy Dragon Complex

The Sleepy Dragon Complex defined here consists of a heterogeneous assemblage of generally metamorphosed, and in some cases highly deformed, intermediate to mafic rock units that are considered to be older than the adjacent Yellowknife supracrustal rocks. The complex occupies a roughly rectangular area 25 km wide and 45 km long between the northern segment of the Cameron River and Turnback Lake and between Upper Ross Lake and Victory Lake and the northern border of the map area. The unit is incompletely known. The northern part has been mapped only from isolated helicopter landings, whereas other parts, notably in the vicinity of Sleepy Dragon Lake and northwest of Turnback Lake, have been mapped in some detail (Davidson, 1972; Fig. 4). On the basis of this detailed work the

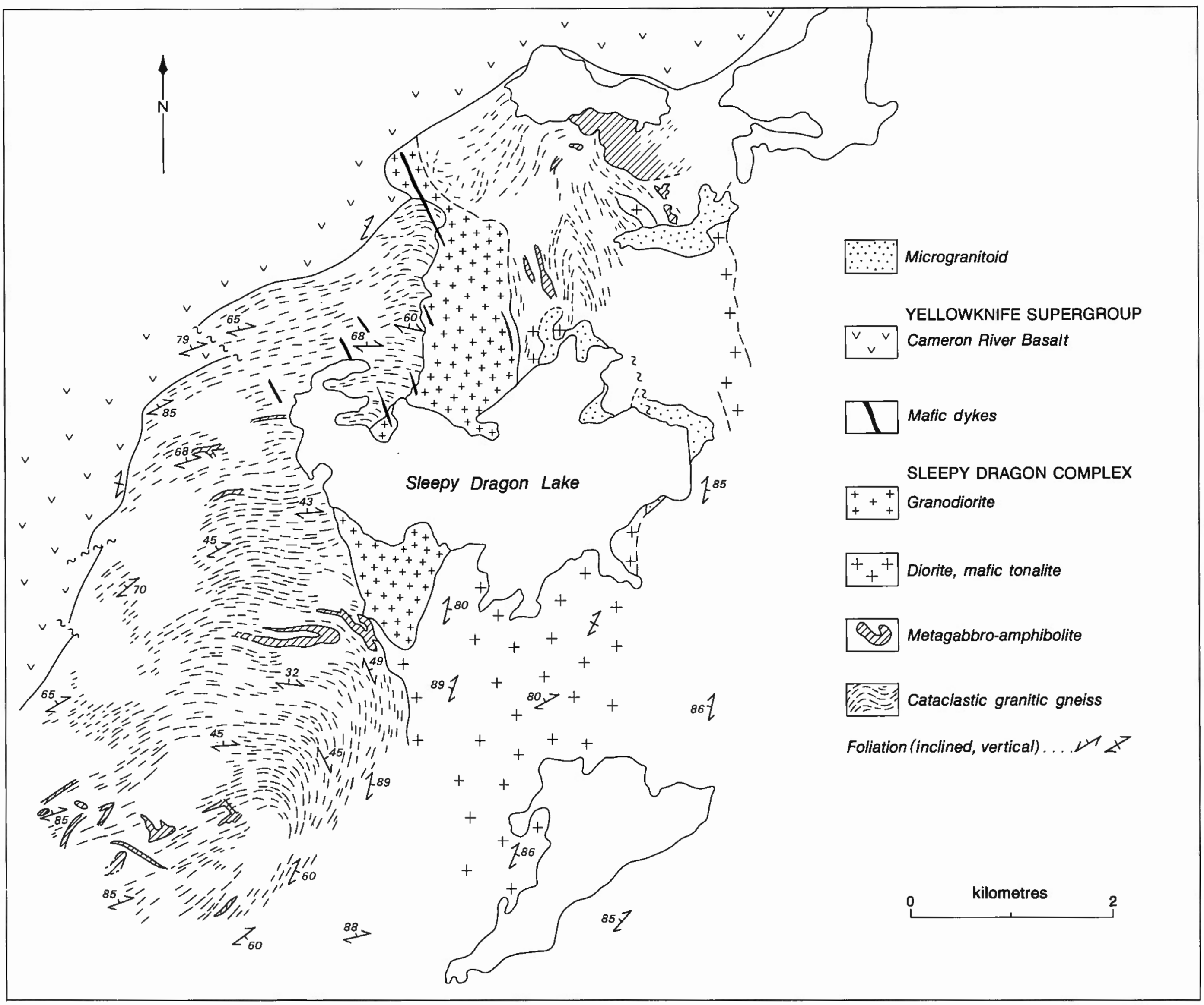


Figure 4. *Geology of part of the Sleepy Dragon Complex in the vicinity of Sleepy Dragon Lake showing the structural and lithological complexity. Geology mapped at 1:50 000 scale by A. Davidson (1972, and field maps).*

heterogeneous nature of the unit became evident and it was realized that detailed mapping of the entire unit would be required before formational-scale units could be outlined. The Sleepy Dragon Complex then is a "group" rank unit (North American Commission on Stratigraphic Nomenclature, 1983) in which there are a number of as yet undefined "formational" rank rock units. It is intended to include in the complex those units metamorphosed and deformed prior to the deposition of the Yellowknife Supergroup. Thus the unmetamorphosed plutons of the Morose, Redout and Prosperous granites that intrude it are not considered part of the complex. This distinction may prove difficult to apply, particularly in the southeast part of the complex where the older rocks are evidently involved in post-Yellowknife deformation and metamorphism.

The complex is not accompanied by a metamorphic aureole that can be related to the complex per se. Where the sediments and volcanics adjacent to the complex are metamorphosed, the metamorphism is associated with the emplacement of younger intrusions such as the Redout Granite which has a wide metamorphic envelope as outlined by the cordierite isograd west of Victory Lake. Regional metamorphic grade is generally high east of the complex although no gradient has been recognized that would relate the metamorphism to the complex. On the other hand the complex is itself metamorphosed as will be discussed later.

Basement unconformity

Contact relations between the complex and the adjacent supracrustal rocks are particularly interesting as this is one of the few examples where basement to Archean supracrustal rocks has been recognized. Baragar (1966) first suggested that the granitic rocks of the complex underlay the volcanics in the vicinity of Cameron River on the basis of the extensive swarm of mafic intrusions that occur both within the complex and the volcanic sequence which were probably feeders to the volcanics. This would require the granitic rocks to be older than the volcanics. Lambert (1982) has similarly described the dykes along part of the contact near Patterson Lake. Davidson (1972), concurring with this interpretation, cited as additional evidence the contrast in deformational style and locally developed discordance in tectonic foliation between the volcanics and granitoid rocks.

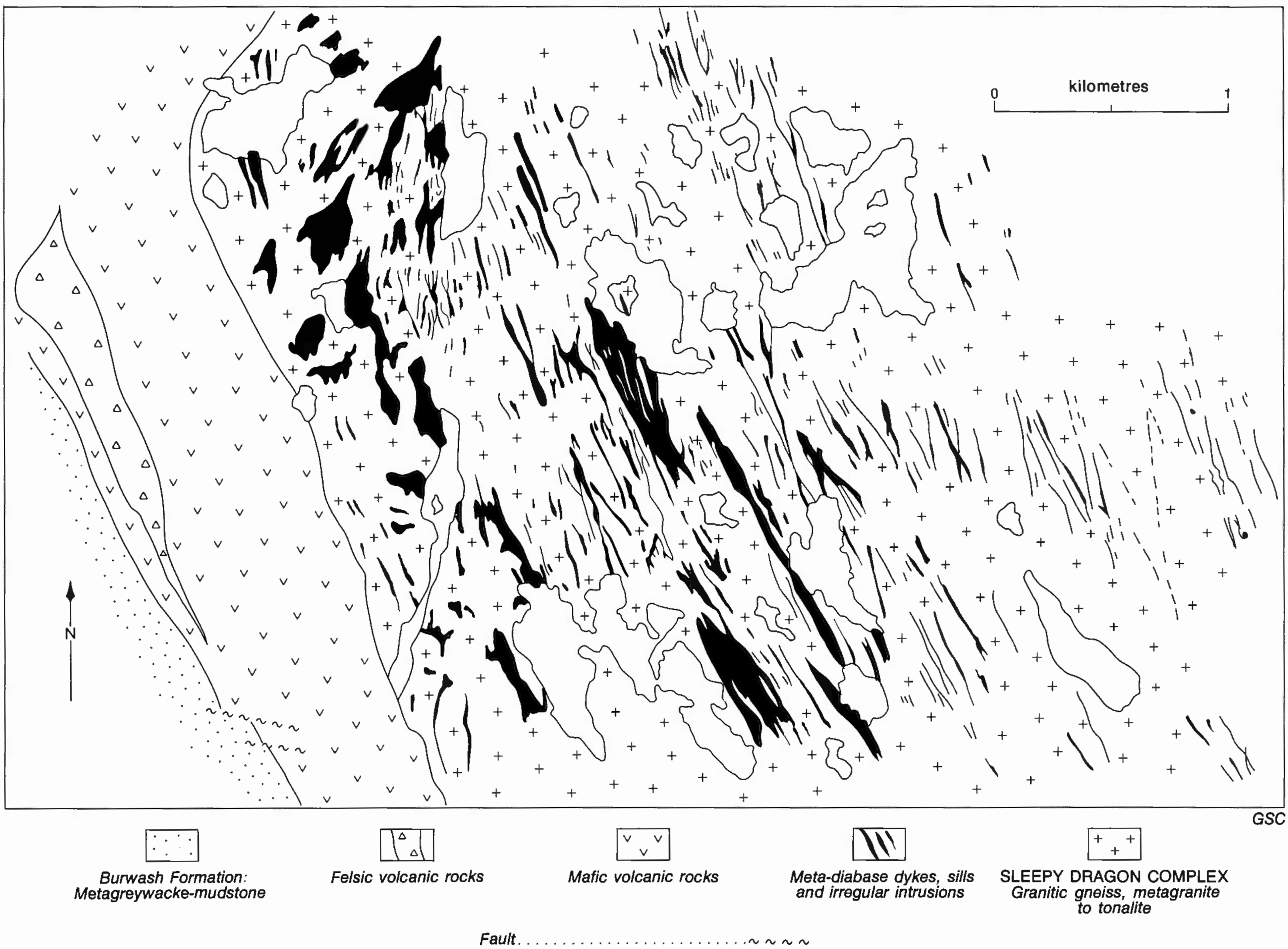


Figure 5. *A swarm of sills and irregular metamorphosed mafic intrusions occurs in the basement complex 4 km north of Upper Ross Lake. The Cameron River volcanics also contain similar intrusive bodies. These mafic bodies may represent a part of the subvolcanic conduit system in the basement below the volcanics. Dykes perpendicular to the unconformity are present locally but cannot be traced across the unconformity due to movement along it during deformation.*

The contact is best preserved and exposed along the west side of the complex north of Victory Lake through to Sleepy Dragon Lake. Evidence supporting the presence of an unconformity is abundant. There is a strong structural contrast between the commonly highly deformed granitic rocks and the mafic volcanics (Davidson, 1972). This is particularly evident in the vicinity of Sleepy Dragon Lake and Patterson Lake where the granitic rocks are typically cataclastic to mylonitic in texture, whereas the adjacent volcanics, although somewhat more sheared in the immediate vicinity of the contact, contain easily recognizable primary structures such as pillows within a metre or so of the contact.

The swarms of mafic dykes and sills within the complex at its margin, thought by Baragar (1966) to represent a feeder system to the adjacent volcanics, are particularly abundant immediately east of Upper Ross Lake and north to Webb Lake (Henderson, 1941b; Lambert, 1982) with smaller concentrations to the northeast near the contact with the volcanics. These mafic bodies typically occur as near vertical elongate lensoid bodies more or less parallel to the contact and the foliation in the complex (Fig. 5). They are highly variable in shape and size ranging from a few centimetres to a few tens of metres in thickness and up to several hundred metres in length. Also present are a few dykes that strike directly towards the volcanics although none were found that could be traced across the contact. Similar dykes and sills occur within the Cameron River volcanic sequence (Lambert 1982, in press). In some places such as south of Webb Lake, the mafic intrusions are so abundant that the granite occurs as blocks or inclusions within a network of dykes and sills. Primary textures such as chilled margins and coarse plagioclase phenocrysts are commonly preserved in the intrusions (Fig. 6). The intrusions are at the same metamorphic grade as the volcanics. The margins of the dykes commonly consist of a thin layer of biotite schist that is probably due to minor shearing after their emplacement. Clearly, however, the main deformation of the granitic complex took place before the emplacement of the dykes.

Figure 6. *Metamorphosed diabase dyke in basement Sleepy Dragon Complex east of Upper Ross Lake that is thought to be a feeder to the Cameron River volcanics. Note plagioclase phenocrysts concentrated in centre of dyke opposite scale. GSC 177750*

The proposal that these intrusions represent a subvolcanic conduit system (Baragar, 1966) seems a reasonable explanation of the relations observed. The lensoid sills of the amphibolite represent horizontal sheets injected into the then flat lying foliation of the granitic gneisses below the unconformity while the much rarer dykes at a high angle to the foliation were conduits to the Archean surface. Lambert (1982) on the other hand has suggested the majority of intrusions within the Sleepy Dragon Complex represent vertical feeders to the mafic volcanic sequence. Similarly, above the contact within the volcanic section, mafic sills, some dykes and other intrusive bodies can be seen intruding flows lower in the sequence that are part of the feeder system to the upper part of the seqence. The fact that individual dykes have not been traced across the contact suggests that there has been some movement along it. The regional irregularity of the contact between the volcanics and the Sleepy Dragon Complex would suggest any movement along it was minor and little if any of the volcanic section above the contact has been lost.

At some localities the unconformity surface has been preserved as, for example, east of the central part of Upper Ross Lake. There the minimally deformed volcanic sequence starting with a mafic volcanic breccia overlies 15 to 30 cm of black biotite schist that contains scattered granitoid pebbles up to 5 cm in diameter. As can be seen in thin section, over half the schist consists of biotite with the remainder being mainly fine quartz and some plagioclase less than 0.1 mm in size. Scattered clasts of polycrystalline quartz up to 1 mm are also present as well as the scattered granitic pebbles seen in outcrop. This schist may represent a metamorphosed, iron-rich, aluminous regolith presumably derived from the weathering of the adjacent granite that accumulated prior to the deposition of the mafic volcanics. The granite contact is sharp with local relief of 5 to 10 cm over half a metre. The granite within half a metre of the contact is highly altered but is better preserved 2 m from the contact. The texture of some of the granite is suggestive of weathering with the feldspars in some cases completely broken down to a fine aggregate of quartzofeldspathic material with muscovite, chlorite and biotite while the quartz remains largely intact. The texture is reminiscent of that of the weathered granite at Point Lake (Fig. 7, 8) and some of the weathered granitic cobbles that occur in the conglomerates both at Point Lake and at Yellowknife as well as some of the granitic cobbles above the unconformity at Ross Lake.

Another exposure of the unconformity occurs at the east end of the small lake northeast of Paterson Lake. There the steeply dipping mylonitic foliation in the granitic basement is more or less parallel to the contact with the similarly steeply dipping mafic volcanics. Local relief on the unconformity surface is up to a metre with the layered volcaniclastic units at the base of the volcanic sequence truncated against it. Scattered granitoid clasts occur in the volcanic section at one point a few metres above the unconformity surface.

East of the southernmost bay of Upper Ross Lake the deformed granitoid rocks are extensively brecciated in the vicinity of the contact. The breccia is filled with carbonate to a depth of about 25 m below the contact. The overlying Raquette Lake Formation which consists of both conglomerate and sandstone has a carbonate matrix but is not brecciated. This breccia zone may represent a hydrothermal spring system associated with the volcanism. In this regard Zn-Pb mineralization is associated with the carbonate at the unconformity. The sporadic mineralization is marked by a gossan 70 m long and about 7 m wide. Carbonate units occur locally, and in some cases over considerable distances, with the volcanics between Victory Lake and the lake on the

Beaulieu River south of the Amacher Granite. Zn-Cu-Pb mineralization is associated with these carbonates at Victory Lake and Turnback Lake.

Parts of the unconformity southeast of Cameron River are faulted. The best documented example is the northerly trending fault west of Paterson Lake, although others occur in the vicinity of Webb Lake and also to the northeast (Fig. 9). Although the contact between the granitic basement and the volcanic sequence is clearly offset, the faults cannot be traced through the volcanic sequence and there is no discernible offset on the volcanic-sediment contact. This would imply that these faults are contemporaneous with volcanism but ceased movement before the end of volcanism (see Lambert, in press). In addition the volcanic sequence west of the fault at Paterson Lake contains large blocks, some in excess of 5 m, of mainly mafic volcanics but also granite which could be a result of material sliding off a rising fault scarp. These faults are thought to represent relatively minor splays off the main fault system (not preserved at the present erosion level) along which the basement block represented by the Sleepy Dragon Complex rose, and along which mafic volcanism was concentrated.

Other contact relations

The contact along the northern part of the eastern side of the complex is less well known but probably similar, if less well preserved, to that on the west side. As on the west, as the contact with the Sunset Lake volcanics is approached, the granitic rocks become cataclastic to mylonitic. At the contact west of Amacher Lake the granite is strongly mylonitic and contains severely deformed mafic bodies. The degree of deformation is such that it is not possible to tell if these are inclusions of volcanics, mafic dykes or older amphibolite bodies as is the case on the west side. There is no evidence of dykes from the Sleepy Dragon Complex intruding the mafic volcanics.

To the southwest in the vicinity of Turnback and Detour lakes the contact relations with the supracrustal rocks are much more complex. As mapped by Davidson (1972) the granitic rocks are migmatitic gneisses of several generations that are cut by younger granitic phases including that mapped as the Redout Granite. The adjacent Burwash metasediments are at sillimanite grade. Within the granitic gneiss complex, amphibolite is locally abundant as elongate highly deformed bodies. These could include Yellowknife mafic volcanics as well as mafic inclusions of other ages. The nature of the

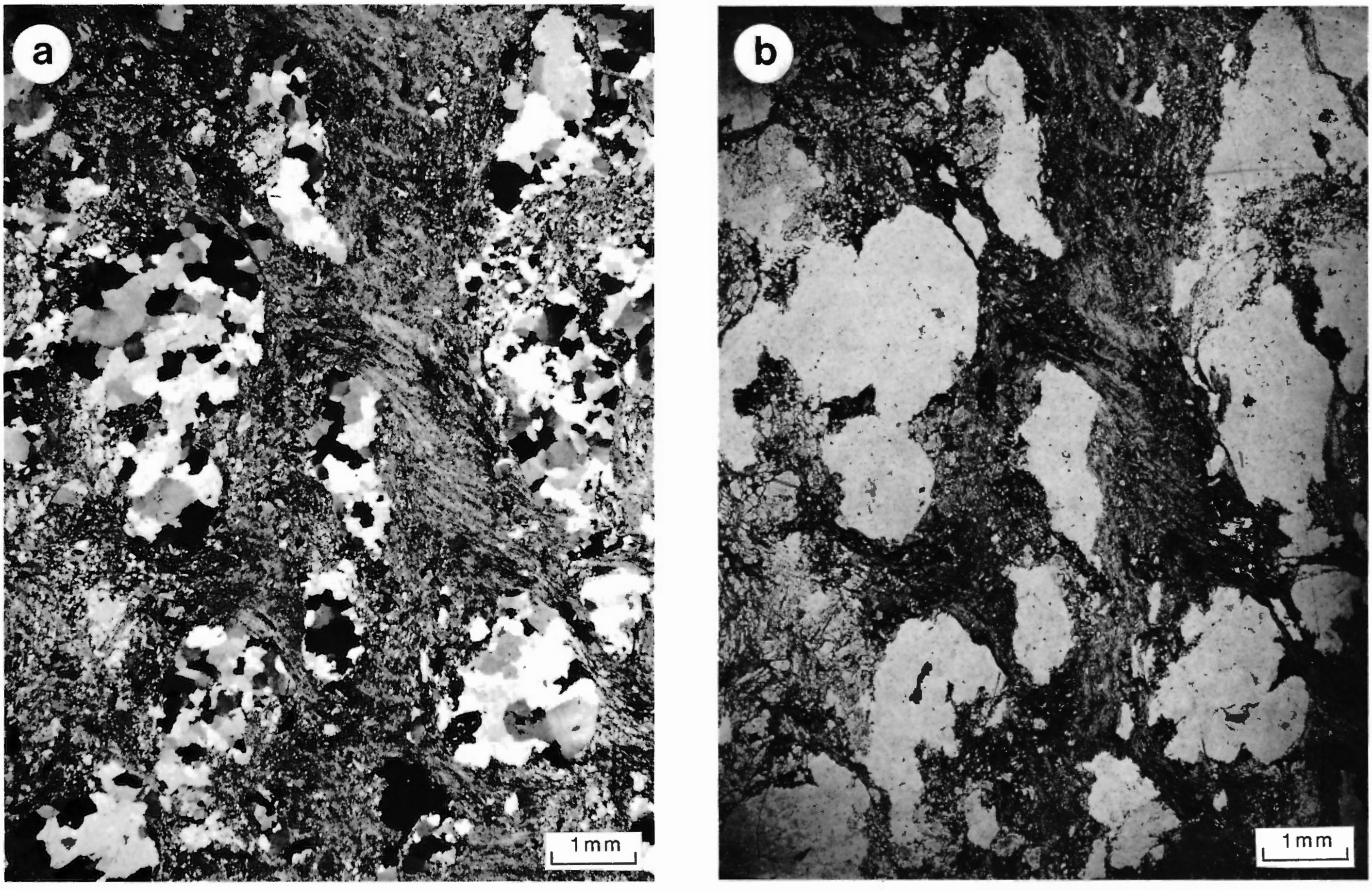


Figure 7. Extensively weathered and metamorphosed Sleepy Dragon Complex granitoid rock from near the unconformity with the Cameron River volcanics 3 km north of Upper Ross Lake. Most of the feldspars and mafic minerals have been completely altered due to weathering and have subsequently been metamorphosed to a fine grained aggregate of quartz, feldspar, muscovite, biotite and chlorite. A few broken coarser relict plagioclase grains remain. The quartz, although highly polycrystalline, has not been significantly elongated due to deformation. Compare with less metamorphosed weathered granite in Figure 8. (a) Crossed polarizers. GSC 203660-N. (b) Plane light. GSC 203660-J

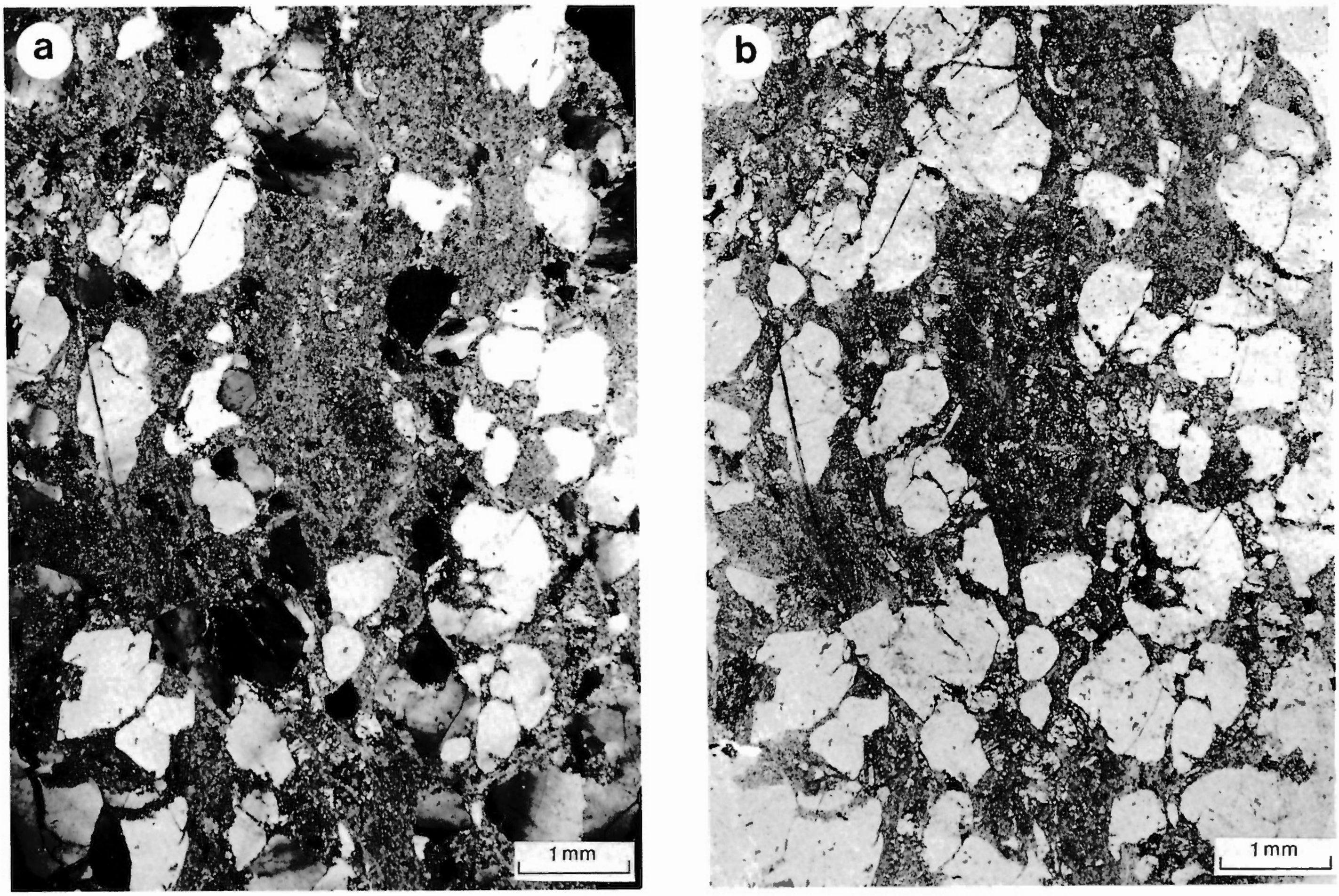


Figure 8. *Weathered 3.1 Ga granite from immediately below the unconformity with Yellowknife Supergroup rocks at Point Lake in the west-central Slave Province. The mafic minerals and most of the feldspars are altered to a fine aggregate of white mica and quartzofeldspathic material although the remaining quartz is essentially unaffected. Texturally this rock is very similar to the weathered Sleepy Dragon granitoid in Figure 7 although this granite at Point Lake is at a significantly lower grade of metamorphism. (a) Crossed polarizers. GSC 203660-T. (b) Plane light. GSC 203660-U*

original contact has been obscured by later events and very detailed mapping would be required to separate the various units.

The central part of the Sleepy Dragon Complex is intruded by large plutons of Morose and Redout granite and by two small two mica granite plutons that have been included with Prosperous Granite. The contact relations are described in detail in the section on the granites, but in brief the western and northwestern contact of the Morose tends to be sharp while the southeastern contact is more diffuse. The contact between the Redout and the Sleepy Dragon rocks is migmatitic while the nature or even the precise location of the Prosperous Granite contact is not known.

Lithology

The Sleepy Dragon Complex consists of several intrusive units that have been only partly defined. Davidson (1972) in the course of more detailed mapping of parts of the complex has defined about 8 discrete units. The various units of the complex are metamorphosed, deformed, and in general tend to be more mafic in composition than the younger intrusive units (Table 2), but show a wide range of composition (Fig. 10).

The Ross Lake granodiorite, a part of the Sleepy Dragon Complex named by Green and Baadsgaard (1971), occurs east of Upper Ross Lake and is contained on the east by the Redout and Morose granites which intrude it. Its extent to the northeast is not defined but it is probably gradational into units described about Sleepy Dragon Lake by Davidson (1972). In general the unit is granodioritic in composition but appears to be quite heterogeneous. A north-northwesterly to northerly trending steeply dipping foliation is commonly present. The rock ranges from massive to gneissic with the degree of deformation increasing towards the west; near the contact with the Yellowknife supracrustal rocks the fabric is cataclastic to mylonitic. The crushed rocks occur in narrow zones separated by less deformed rocks. Faults approximately parallel to the foliation occur locally near the contact, and the rocks to the west of these faults tend to be more cataclastically deformed and heterogeneous than those to the east which are more recrystallized and homogeneous. An example of such a fault occurs along the east shore of Paterson Lake and extends to the south (Fig. 9).

Where best preserved the granodiorite is a massive, equigranular, coarse grained rock composed largely of altered, subhedral, 3 or 4 mm plagioclase laths. Distribution of the relatively coarse alteration products suggests that the

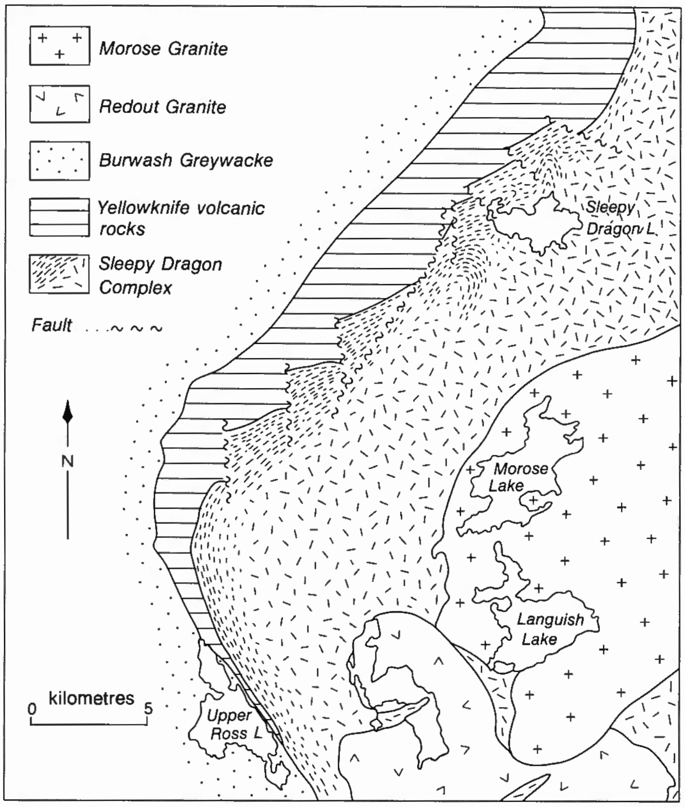


Figure 9. *Archean faults in the Cameron River volcanics and Sleepy Dragon Complex that were active during, but did not continue after, volcanism as there is no displacement of the upper contact of the volcanics. These faults are possibly splays off a master fault that controlled basin development but is not exposed at this level of erosion.*

Table 2. Modal analyses, Sleepy Dragon Complex

	1	2	3	4	5
Quartz	34	45	41	24	26
Plagioclase	20	43	53	45	49
Microcline	31			22	3
Biotite		12	1	7	21
Muscovite				1	X
Hornblende			6		
Chlorite	16			X	X
Epidote			X		

original plagioclase was zoned. Microcline is present but much less abundant as subhedral, equant, clear grains with inclusions of plagioclase. It tends not to occur as interstitial material. Quartz, however, has an interstitial habit and is typically polygonized. Biotite occurs both as large scattered randomly oriented flakes on the order of a millimetre or more in size and also as fine grained (0.05 mm) oriented aggregates between the larger feldspar grains. Where more metamorphosed, as in the vicinity of the younger Redout and Morose granites, the coarser biotite is recrystallized into patches with a fine grained decussate texture (Fig. 11). As the rock becomes increasingly deformed a cataclastic to mylonitic fabric is developed with the coarse feldspars retaining their size in general but developing increasingly milled margins. A matrix forms consisting of broken fragments of the coarser feldspar and quartz with, for example, the matrix in the immediate vicinity of large

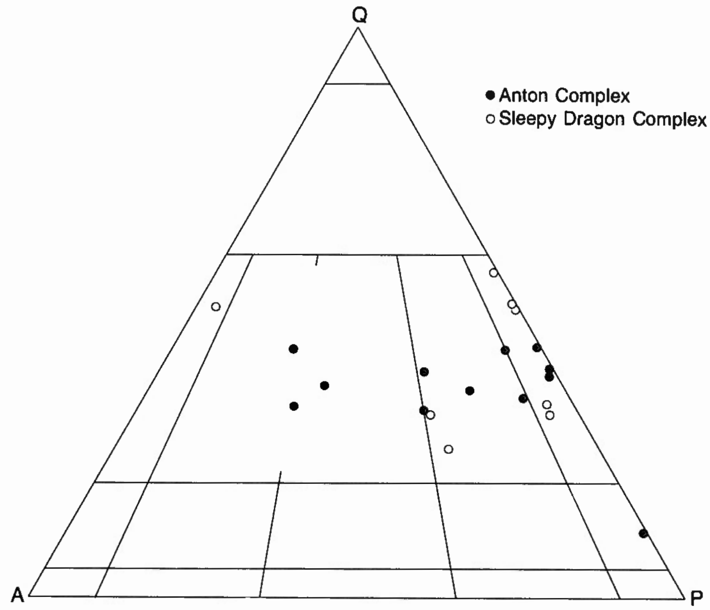


Figure 10. *Plot of modal analyses of the Anton and Sleepy Dragon complexes. The triangle is subdivided following the granite classification system of Streckeisen (1976). Q = quartz, A = alkali feldspar, P = plagioclase.*

microcline clasts being relatively richer in broken, angular microcline than elsewhere. The quartz is increasingly polygonized and drawn out into long ribbons (Fig. 12). Biotite becomes shredded and commonly is recrystallized, overgrowing the cataclastic fabric due to subsequent metamorphism. A gneissic fabric is developed in the granodiorite away from the western margin, for example southeast of Paterson Lake. There the rock is more even grained, recrystallized into anhedral grains with the biotite, which best defines the foliation, tending to be wrapped around the quartz and feldspar.

Mafic inclusions occur in the Ross Lake granodiorite. In most cases they are relatively small but 1 km east of the north end of Upper Ross Lake there is a lens shaped mafic body about 3 km long and up to 300 m wide that is intruded both by the mafic dykes and the Ross Lake metagranodiorite. The mafic body is distinctly layered on few centimetre to 30 cm scale with the layering defined by variations in texture and composition. It consists of varied proportions of biotite, chlorite, anthophyllite, cordierite and plagioclase. Some layers are very coarse grained with rosettes of anthophyllite several centimetres in diameter.

The Ross Lake granodiorite is cut by the metamorphosed mafic dykes thought to be feeders to the Cameron River volcanics to the west. There is also an abundance of pegmatites, locally rare-element bearing, related to the younger Redout Granite. A more massive granite east of Webb Lake whose extent is not known but which is also thought to be older than the Cameron River volcanics, intrudes the more metamorphosed granodioritic to tonalitic rocks. Northwest of Languish Lake in the contact area between Redout, Morose and Ross Lake granitoid units, Davidson (1972) mapped a metamorphosed diorite body of undefined extent, whose relationship to the Ross Lake granodiorite is not known, but which is intruded by the Morose Granite.

To the north, in the vicinity of Sleepy Dragon Lake, Davidson (1972) mapped several units of the Sleepy Dragon Complex (Fig. 4). Adjacent to the contact with the Yellowknife volcanics is a strongly foliated cataclastic to

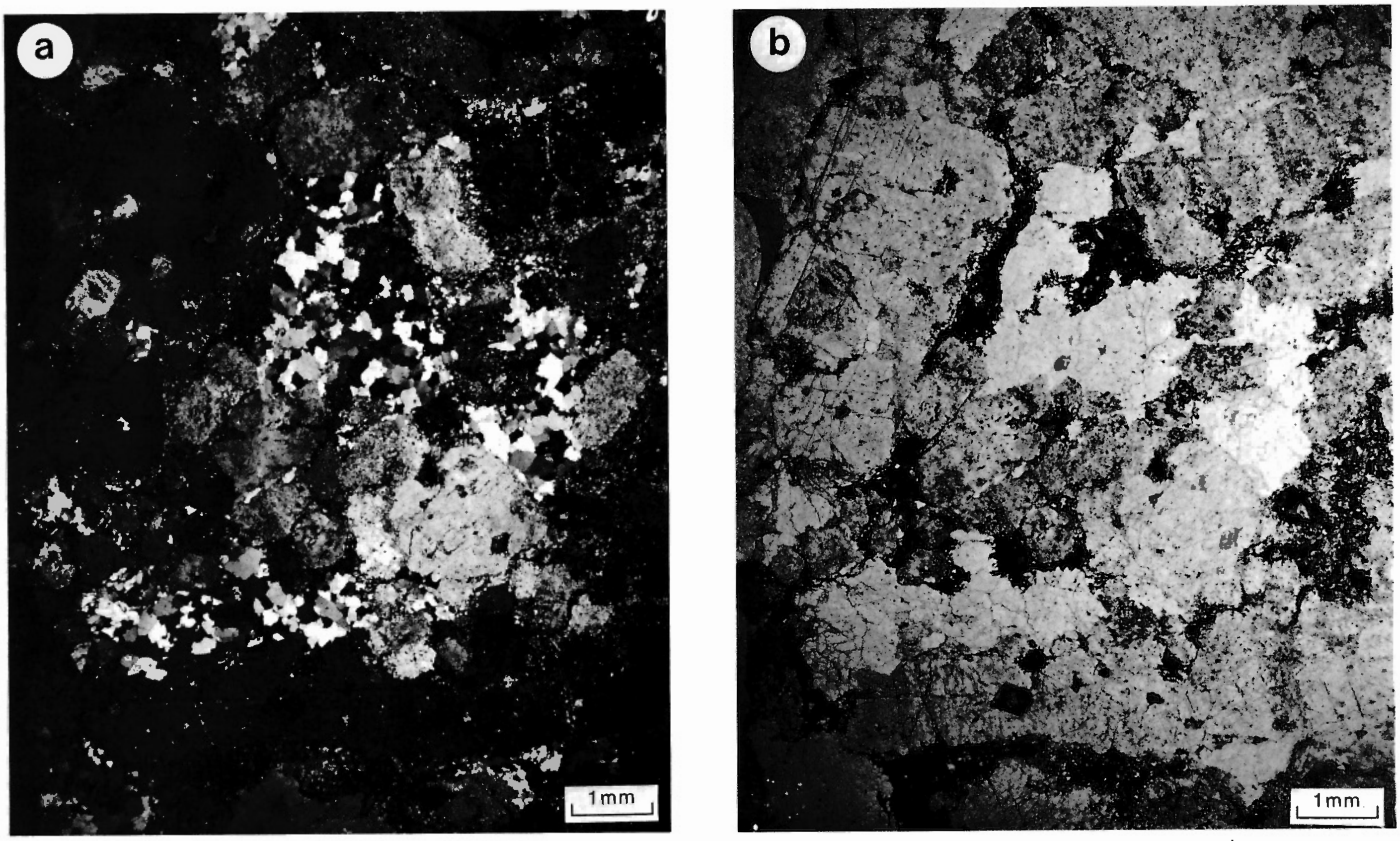


Figure 11. *Massive metamorphosed granite from the Sleepy Dragon Complex 2 km south of the south end of Webb Lake. Note the extensive alteration of the plagioclase and the irregular granulated grain boundaries of the feldspars. The quartz is highly polygonized and the biotite occurs as fine grained decussate aggregates. Little, if any, planar fabric has been imposed on this metagranite. Compare with Figure 12. (a) Crossed polarizers. GSC 203660-R. (b) Plane light. GSC 203660-Q.*

mylonitic granitoid rock that is probably correlative with the previously discussed Ross Lake granodiorite to the southwest. The unit is texturally heterogeneous, ranging from augen gneiss to mylonite. Pegmatitic phases within it are also deformed. Within the cataclastic granitoid are generally concordant thin pods or lenses of amphibolite that in some cases appear to be tightly folded. These are distinct from the relatively lower grade mafic Yellowknife feeder dykes and the sills in the Ross Lake body. Southwest of Sleepy Dragon Lake the gneisses occur in a northeasterly trending open asymmetric fold that plunges moderately to the northeast. The contact between the volcanics and the cataclastic granitoid gneisses east of the lake is one of the few places where there is a significant structural discordance at the unconformity. It is possible that this contact is faulted, with the faulting active during volcanism as previously discussed (Fig. 9). North of Sleepy Dragon Lake the gneisses are tightly folded about northerly trending axes. South of the lake the cataclastic gneiss is intruded by a rather massive diorite to mafic tonalite that contains inclusions of the older gneiss and towards the east is intimately associated with grey granodiorite phases. To the south the diorite is increasingly gneissic towards the Morose Granite. The cataclastic gneiss is also intruded by a metamorphosed and weakly foliated granodiorite that occurs both north and south of the lake. Both the cataclastic gneiss and the foliated granodiorite are intruded by mafic dykes that are probably of the same generation as those east of Upper Ross Lake thought to be feeders to the volcanics. North of Sleepy Dragon Lake, near the contact, the gneisses are intruded by a metamorphosed gabbro body considered by Davidson (1972) as a possible subvolcanic intrusion due to its chloritic composition. Northeast of the lake a few small bodies of uniform, grey, fine grained, biotite-muscovite granite contrast strongly with the highly deformed, metamorphosed rocks they intrude.

The northernmost part of the complex is known only from helicopter landings. It consists mainly of weakly foliated granodiorite, tonalite and quartz diorite intruded by thin dykes of granite and pegmatite. Inclusions of more mafic material are common. Towards the eastern contact with the Sunset Lake mafic volcanics, as on the west side, the rocks become increasingly cataclastic.

Northwest of Turnback Lake the metamorphic grade is much higher and both the granitic rock of the presumed basement complex and the adjacent Yellowknife supracrustal rocks are strongly deformed. The granitic rocks consist of migmatitic gneisses of several generations cut by younger granitic phases. The gneisses are streaky and irregular with locally more massive and homogeneous granitic phases. Inclusions of preserved Yellowknife supracrustal rocks are very elongated and highly deformed.

Geochronology

The only geochronological data available for the Sleepy Dragon Complex are from the Ross Lake granodiorite. Green and Baadsgaard (1971) have done some preliminary work and on the basis of a single zircon sample determined a

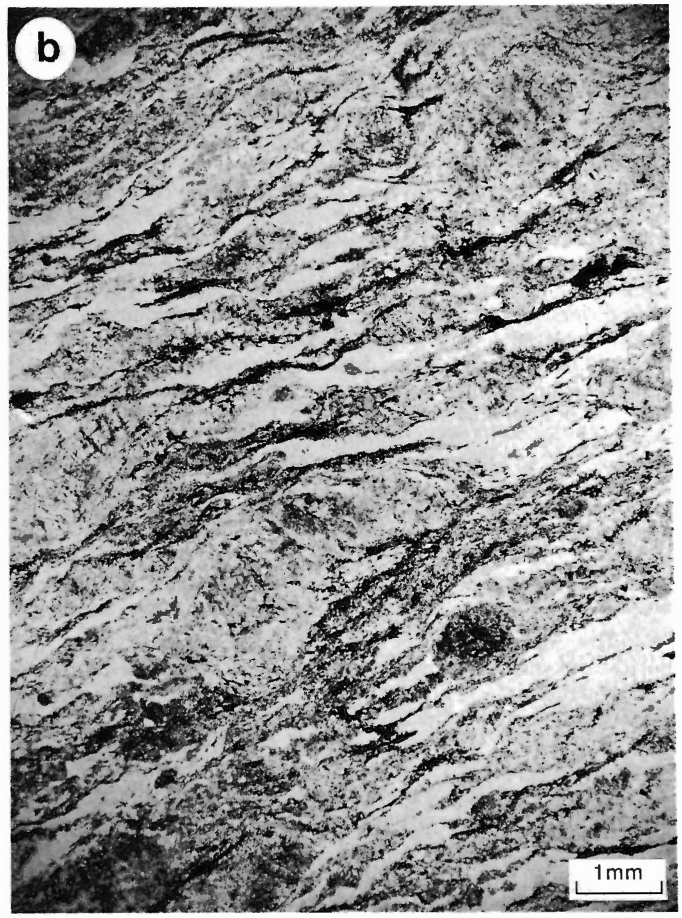


Figure 12. *Mylonite from the Sleepy Dragon Complex near the Cameron River Basalt contact west of Sleepy Dragon Lake. Original granitoid rock has been crushed and comminuted so that the grain size grades from the 3 mm clasts of feldspar to quartzofeldspathic material less than 0.01 mm. Note the elongate ribbons of quartz which accentuate the strong fabric. (a) Crossed polarizers. GSC 203660-K. (b) Plane light. GSC 203660-L.*

^{207}Pb-^{206}Pb age of 2595 Ma. A Rb-Sr whole-rock isochron indicated an apparent age of 2457 ± 120 Ma (Sr_i = 0.707 ± .005, λ^{87}Rb = 1.42) although they pointed out (Green and Baadsgaard, 1971, p. 185) that the sample material was probably "affected to a small, but variable, extent by the pervasive K metasomatism" related to the close proximity of Redout Granite pegmatites. Three biotite K-Ar dates on the Ross Lake granodiorite are 2360 ± 27, 2490 ± 28 and 2440 Ma (Green and Baadsgaard, 1971; Burwash and Baadsgaard, 1962). Although recognizing that additional data would be desirable they concluded that the Ross Lake intrudes the mafic volcanics which Green (1968) dated at 2574 ± 200 Ma (Sr_i = 0.706 ± 0.005, λ^{87}Rb = 1.42). The significance of these data is not clear. Elsewhere in the province the Yellowknife supracrustal rocks have zircon ages of about 2.7 Ga – Yellowknife, 2640 and 2610 Ma, R.K. Wanless (personal communication, 1969), Green and Baadsgaard (1971); Back River, 2667 Ma, Lambert and Henderson (1980); Point Lake, about 2.7 Ga, Scharer and Allègre (1982). Basement ages are about 3 Ga (Point Lake, 3.15 Ga, Krogh and Gibbins (1978); Indin Lake, 2.9 Ga, Frith et al. (1977); Yellowknife, 2.9-3.2 Ga, Nikic et al. (1980). The existence of a basement-supracrustal relationship between the Sleepy Dragon Complex and the Yellowknife volcanics are not particularly well supported by the presently available geochronological data.

Significance

The Sleepy Dragon Complex is considered to be the largest area of pre-Yellowknife basement known to date in the Slave Province. Its extent to the north beyond the map area remains undefined. The Anton Complex to the west is similar in many respects, although its relationship to the supracrustal rocks is not as clearly defined. Other examples of basement in the Slave Province have been described by Baragar and McGlynn (1976).

The complex is considered to be a block of the continuous sialic basement that underlay most, if not all, of the area presently outlined by the Slave Province 2.7 Ga ago. Following the model of McGlynn and Henderson (1970), Lambert (1977), and Henderson (1981, and this report), it is envisaged that the sialic crust underwent rifting resulting in a series of grabens and horsts, one of which is represented by the Sleepy Dragon Complex. The grabens became repositories for the Yellowknife supracrustal rocks. Thick mafic volcanic sequences were extruded along the margins of the grabens where the faulting due to rifting provided zones of weakness along which conduits to the deep source of the mafic volcanics could form. Although the main fault zone has not been preserved or at least not recognized at the present erosion level, possible splays from it that were active during volcanism have been mapped. The Sleepy Dragon Complex basement block was subsequently deformed along with the Yellowknife supracrustal rocks, with the emplacement of younger intrusive units, the Redout and

Morose granites. Thus we have preserved at the margin of the complex a vertical section extending from the basement complex into the Yellowknife supracrustal rocks.

Similar horsts of basement occur elsewhere in the province. These include the basement block at Point Lake (Henderson and Easton, 1977; Easton et al., 1982), at Healey Lake (Henderson et al., 1982; Henderson and Thompson, 1982) and the Hanimor Complex, 120 km east of Contwoyto Lake (Frith and Percival, 1978; Frith, 1981). Within the map area, as has previously been suggested, the lithologically similar Anton Complex is another example. Here again mafic volcanics are located between the Anton Complex and the sedimentary rocks. Because of the major accumulation of mafic volcanics to the south, in the vicinity of Yellowknife Bay and Walsh Lake, it can be postulated that a similar basement lay to the west but was subsequently lost with the emplacement of the large Defeat plutons. Possible remnants of this basement occur as blocks of gneissic tonalite in a diatreme within the Kam mafic volcanic sequence (Nikic et al., 1980). Zircons from these blocks have ^{207}Pb-^{206}Pb ages that range between 2900 and 3210 Ma.

Anton Complex

The Anton Complex consists of metamorphosed granodiorite and quartz diorite to gneiss with intrusive phases of granodiorite and granite. It occurs west of the Yellowknife River and north of Duckfish Lake in a roughly rectangular area 30 km long and 15 km wide. On the basis of similar aeromagnetic pattern it may extend over 50 km north of the area. It is similar in many respects to the Sleepy Dragon Complex to the east between Cameron River and Turnback Lake.

Little is known about this complex as only a relatively small proportion of it has been traversed on foot, the main part being covered by isolated helicopter landings. Its origin is somewhat enigmatic, but at least part of the unit is thought to be older than the supracrustal rocks although there are clearly phases within it that are younger.

The Yellowknife supracrustal rocks adjacent to the complex are metamorphosed to amphibolite grade although this metamorphism is at least in part attributable to the broad thermal ridge associated with the Prosperous Granite to the east. At the north border of the map area is another Prosperous Granite pluton with an associated thermal aureole. Southwest of Clan Lake the metamorphic isograds are parallel to the contact with the complex. The aureole, as defined by the cordierite isograd, is about 1.5 km wide.

Contact relations

The Anton Complex is older than the adjacent plutonic units. The contact between the Anton and the Awry granite to the west is gradational over a poorly defined zone up to 15 km wide consisting of a highly varied assemblage of inclusions of the more mafic Anton granodioritic and quartz dioritic gneisses in a matrix of the more leucocratic Awry granitoid rocks. The inclusions range in size from a few centimetres to large areas of several hundred square metres and vary from angular bodies with sharp clean contacts to rounded bodies with diffuse contacts. The proportion of inclusions in the contact zone ranges from none to over 80%. Inclusions are found west of the contact zone as mapped but are much less abundant. To the east of the contact zone, Awry granite occurs as extensive dykes and small bodies. To the south the contact between the Anton gneisses and the Defeat Plutonic Suite is irregular and gradational; the contact as indicated on the map is very generalized. Over a distance of 10 or 15 km the gneissic granodiorite and quartz diorite becomes less gneissic, more homogeneous and less mafic southward where it is mapped as part of the Defeat. Both the Anton and Defeat in this area are intruded by dykes of granite that are presumably related to the Awry Plutonic Suite. The Duckfish Granite intrudes both the Anton and Defeat at this contact zone. Inclusions of Anton gneiss occur within the Duckfish along its north border, and dykes and sills of the granite intrude the gneisses in the contact area which is relatively poorly defined. The contact between the gneiss and the intrusive Prosperous Granite pluton at the north border is gradational and contrasts strongly with the sharp, clean, intrusive contact of the same body with the metasediments and metavolcanics to the east. Over a distance of about half a kilometre the massive, medium grained, biotite-muscovite Prosperous Granite becomes finer grained, somewhat foliated with inclusions and vaguely defined layers of a more mafic gneiss. The Anton gneiss in this contact area commonly contains pegmatites and aplitic bodies.

Much of the contact between the Anton Complex and the Yellowknife supracrustal rocks to the west is faulted. South of Bell Lake between the gneisses and the adjacent volcanic rocks is a mixed zone of dominantly mafic volcanic inclusions in a pink granitic matrix. The volcanic inclusions are typically angular with sharp contacts and are up to several tens of metres in size. To the west in this zone a more mafic granodiorite occurs that is also cut by the granite. North-northwest of Nelson Lake the Anton Complex is in contact with the Burwash Formation metasediments. The contact is sharp with scattered inclusions of biotite schist in a light pink intrusive granitic phase of the complex. No evidence has been recognized to suggest the Anton granodiorite and quartz diorite gneisses intrude the Yellowknife supracrustal rocks although younger granitic phases clearly intrude both.

Lithology

The Anton Complex consists of at least one unit of metamorphosed granodiorite to quartz diorite and one or more younger intrusive granitic phases. The orthogneiss is a weakly to moderately foliated dark brownish weathering grey to dark grey rock. The gneiss is varied texturally and compositionally (Fig. 10) from massive homogeneous units to units with layers a few centimetres to several metres thick due to slight gradational variations in composition and texture. Texturally they are medium to fine grained, most commonly equigranular but locally with plagioclase augen. The foliation generally has a northerly to north-northeasterly trend and is steeply dipping although locally it can be very irregular even on an outcrop scale. These rather mafic orthogneisses are intruded by dykes and small massive intrusions of pink, equigranular, medium grained, leucocratic granite. The latter are probably related to the adjacent intrusive granitic units, mainly the Awry Granite to the west. In the vicinity of the muscovite-bearing Prosperous Granite pluton near Clan Lake the Anton gneisses are intruded by muscovite-bearing granite dykes and pegmatites. The proportion of mafic gneiss to intrusive granite is not known but is probably quite varied.

In thin section the gneisses consist of a varied proportion of plagioclase, quartz, biotite, microcline, hornblende and epidote (Table 3). The texture is also varied from gneissic with weakly to moderately oriented biotite randomly distributed throughout the rock or concentrated into vague zones, to fairly massive and homogeneous. In some cases the gneisses are clearly cataclastic with zones of crushing, but more commonly the grain margins are irregularly interlocking and not broken or milled. Plagioclase occurs as large grains, up to a centimetre or more in length, that in some cases have been completely recrystallized into polycrystalline aggregates with the outline of the original coarser grains

Table 3. Modal analyses, Anton Complex

	1	2	3	4	5
Quartz	30	30	35	34	31
Plagioclase	51	40	55	51	42
Microcline	5	21	1	X	13
Biotite	12	4	9	12	11
Muscovite	1	2			
Hornblende			1		3
Chlorite		4		1	
Epidote				1	1

defined by biotite. Most typically plagioclase occurs as subhedral to broken irregular grains with a wide range of grain size. They are not zoned or have only very simple zoning and have a patchy distribution of alteration products. In some cases they are antiperthitic. Quartz, also with a wide range in grain size, occurs as irregular to polycrystalline aggregates and as irregular lensoid masses parallel to the foliation of the gneiss. Microcline is present in the felsic phases both as irregular grains and as coarse porphyroblasts that are slightly perthitic but typically contain inclusions of plagioclase. Biotite occurs as scattered small (<1 mm) flakes or aggregates, more or less oriented to define the fabric of the rock and, where abundant, tends to wrap around the coarser feldspars. It is commonly partly altered to chlorite. Fine anhedral hornblende locally occurs with the biotite, as does epidote. The granitic dykes and intrusions are more massive and if deformed have mylonitic foliation rather than gneissosity.

Geochronology

Preliminary geochronological data on the Anton Complex have been obtained from 2 samples collected at the central part of Narcisse Lake in the southern part of the complex (R.K. Wanless, personal communication, 1978). Potassium-argon ages on biotite are 2197 ± 48 and 2107 ± 46 Ma and on hornblende in one of the samples 2506 ± 52 Ma. Zircon concordia intercept ages based on two nonmagnetic size fractions from the two samples are 2648 ± 12 and 2707 ± 24 Ma.

Significance

It is suggested that the oldest gneisses in the Anton Complex may be older than the Yellowknife Supergroup. No recognizable unconformity is preserved between them but the relationship is suggested on the basis of the metamorphic and structural contrast between orthogneiss of the Anton Complex and the adjacent, relatively well preserved Yellowknife metasediments. The metamorphic grade of the Yellowknife at the contact is amphibolite facies but drops to greenschist grade within 2 km of the contact. It is also suggested the intrusive Defeat granodiorite to the south may have been derived from the Anton as their compositions are similar and the contact between the two is gradational. The Anton zircon concordia intercept ages at 2648 and 2707 Ma, although somewhat similar to the zircon ^{207}Pb-^{206}Pb age from the Ross Lake granodiorite of the Sleepy Dragon Complex (2595 Ma) which is also considered to be basement to the Yellowknife, are much younger (200-400 Ma) than the basement ages elsewhere in the province. They are, however, similar to the lead isochron ages of 2650 ± 10 Ma (Thorpe, 1971) and 2635 Ma (Cumming and Tsong, 1975) from the Defeat Plutonic Suite to the south and indeed are not that different from the age of the presumed volcanogenic zircon from the Yellowknife Supergroup at several places in the province (2600-2700 Ma).

ARCHEAN SUPRACRUSTAL ROCKS – YELLOWKNIFE SUPERGROUP

All the supracrustal rocks of Archean age in the Slave Province are part of the Yellowknife Supergroup. These rocks were originally included as part of the Point Lake – Wilson Island Group (Stockwell, 1933). The term Yellowknife Group was first used by J.F. Henderson (1938) in the original mapping of the Hearne Lake map area and since that time others mapping in other parts of the province have used the term for any Archean supracrustal rocks (e.g. Lord, 1942; Fraser, 1964; Tremblay, 1976). The rank of the Yellowknife Group was raised to supergroup to better reflect the diverse nature of the Yellowknife supracrustal rocks (Henderson, 1970).

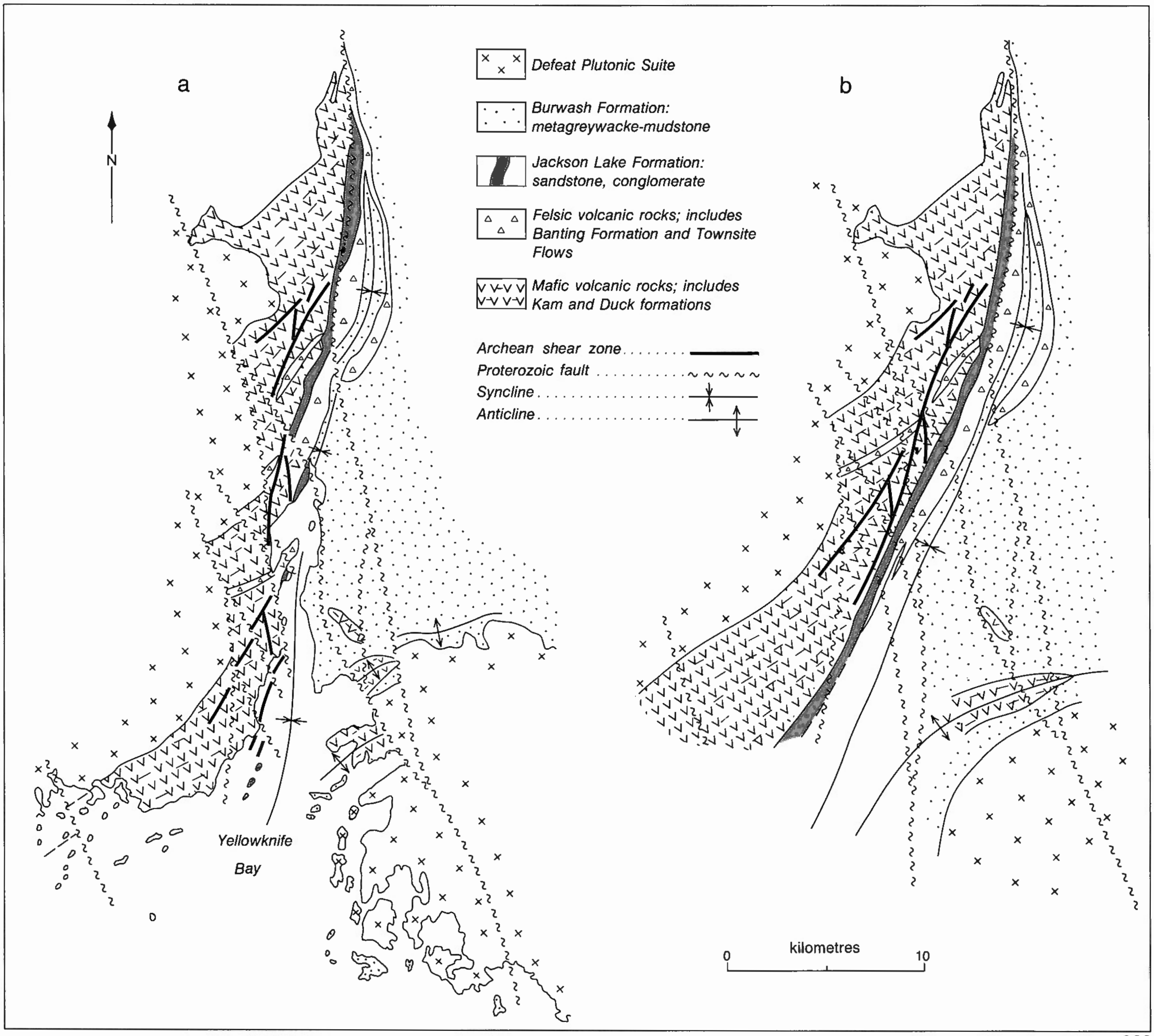
The rocks of the Yellowknife Supergroup are similar to many Archean supracrustal successions in other shield areas of the world in that they consist of thick sequences of volcanics and immature sediments. Thick sequences of quartzites, carbonates and other rock types characteristic of stable environmental conditions are conspicuous by their absence. Throughout the province the Yellowknife differs from other Archean successions in that the area underlain by sediments is much greater than that underlain by volcanic rocks (McGlynn and Henderson, 1970). Whereas the vast proportion of Yellowknife sediments, consisting dominantly of greywacke mudstone turbidites, are similar to those seen elsewhere in the world, the volcanics differ to some degree. In most Archean terranes the volcanics are typically mafic in composition, as for example in the Superior Province, where about 85% of the volcanics are of mafic and 15% of felsic composition (Goodwin, 1977). In the Slave, however, the proportion of felsic volcanics appears to be higher with extensive centres of dominantly intermediate to felsic composition (e.g. Lambert, 1976, 1978; Fraser, 1964; Padgham 1974), although thick, dominantly mafic sequences are also present (e.g. Bostock, 1980). In the Yellowknife-Hearne Lake map area this generally higher than normal proportion of felsic volcanics is evident.

Throughout the province the areas of supracrustal rocks are surrounded by large areas of intrusive granitic rocks. In a few places, however, granitic basement to the supracrustal rocks has been preserved (Baragar, 1966; Davidson, 1972; Stockwell, 1933; Henderson and Easton, 1977; Frith et al., 1977; Nikic et al., 1980). The basement rocks are tonalitic to granitic and are about 3.1 Ga old where dated by U-Pb zircon methods.

Given the amount of subsequent granitic intrusion and deformation these rocks have undergone, it is difficult if not impossible to define with any degree of confidence the outlines of original depositional basins although it is clear that segments of basinal margins are preserved locally. Also, it is not yet possible to demonstrate unequivocally whether the various supracrustal terranes were originally interconnected or developed separately at different times at different places. In some cases the supracrustal terranes are linked by areas of granitic rock with extensive zones of highly metamorphosed supracrustal rocks, suggesting the terranes may have been originally joined. Also supporting the idea of contemporaneity of the Yellowknife basins is the fact that no major unconformities have been recognized within the Yellowknife Supergroup. If the basins evolved at different times one might expect at least somewhere that a younger basin would develop over an older basin with a structural and probably a metamorphic discordance between. This has not been recognized. Evidence supporting contemporaneity includes preliminary geochronological data from two widely separated areas of Yellowknife rocks, at Yellowknife and the Back River area in the east part of the province, which have similar ages at about 2670 Ma (Henderson 1981). More recently Schärer and Allègre (1982) reported a similar age for volcanogenic zircons found in low grade metasediments in the Point Lake area of the west-central Slave Province.

If the basins are approximately contemporaneous then the depositional rock record of the Archean in the Slave Province is very short. Folinsbee et al. (1968) have suggested that the supracrustal rocks at Yellowknife could have accumulated in about 15 million years on the basis of comparison of the Yellowknife rocks with the well dated rocks of similar character and thickness of the Miocene Fossa Magna of Japan. Between the oldest known rocks in the province, the granitic basement rocks dated at about 3.1 Ga, and the end of the Archean eon 2.5 Ga ago, a period of 600 million years, the depositional rock record is only about 15 million years or about 5% of that time. Is then the Yellowknife Supergroup representative of conditions during this significant period of earth history?

At Yellowknife the Yellowknife Supergroup has been divided into two groups and six formations (Henderson, 1970). The sedimentary rocks within the contiguous supracrustal terrane within the Yellowknife-Hearne Lake area are assigned to the Duncan Lake Group while the volcanic units are part of the Beaulieu Group. The supracrustal rocks in the western part of the map area are not considered part of the



GSC

Figure 13. *Location of early shear zones in, and the unconformity above, the Kam Formation mafic volcanics. (a) is a geological map of the Yellowknife Bay area. The early shear zones are displaced by post-Archean faults. In (b) the offset of the late faults has been replaced. The felsic volcanic marker horizon in the mafic volcanic sequence is displaced along the early shear zone. The unconformable contact between the Kam and the overlying Jackson Lake Formation, however, is not displaced along the shear zone, suggesting that movement on the early shear zones took place before deposition of the Jackson Lake Formation above the unconformity.*

groups defined east of Yellowknife as their relationship to those rocks remains to be demonstrated. The formations defined at Yellowknife occur only in the immediate Yellowknife area, with the exception of the Burwash Formation greywacke-mudstones which are continuous throughout the supracrustal terrane east of Yellowknife. The other formations are of local extent and no useful purpose would be served by correlating them with other similar though physically separated units. Thus the Kam mafic volcanics occur only at Yellowknife, whereas, for example, the mafic volcanic units east of the Cameron River, although very similar to the Kam, have been given a different formational names: the Cameron River, Webb Lake and Dome Lake formations (Lambert, in press). The stratigraphic relationship of the Banting Formation felsic volcanics to several other members of the Yellowknife Supergroup was not clear when the formations were originally defined (Henderson, 1970). More recent work by Baragar (1975) has shown that the Banting Formation is conformable with the other adjacent units of the Yellowknife Supergroup and hence is part of the Beaulieu Group. A thin quartz sandstone – conglomerate unit, the Raquette Lake Formation, has been defined in the northeastern part of the area at Upper Ross Lake and Lambert (in press) has formally defined 9 volcanic units in part of the same area.

In the descriptive section on the Yellowknife Supergroup that follows the rocks are described on a lithological basis in the various geographical locations in which they occur. The order of description has no basis for any stratigraphic succession.

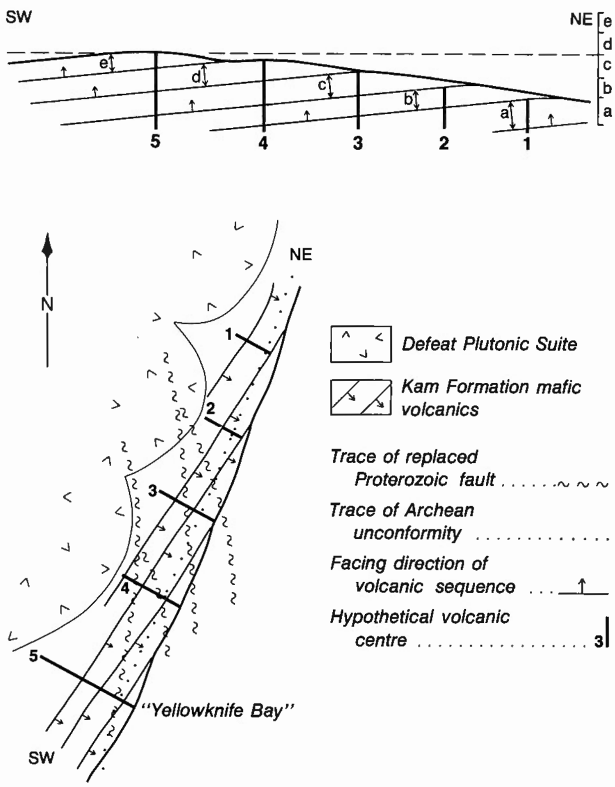


__Figure 14.__ In the Kam Formation the thickness of the sequence has commonly been determined by measuring the aggregate thickness by correlating across fault blocks along marker horizons within the sequence. In the following model it is shown that this may represent a considerable overestimate. In (a) a hypothetical section through the Kam Formation demonstrates how the volcanic pile may have accumulated. The section shows a series of diagrammatic volcanic centres (1,2,3...) where each successive centre occurs south of and is younger than the previous centre, with feeder dykes intruding the flanks of the older centres. The result is a southward progradation of the volcanic complex. In the section the maximum height of the volcanic complex is less than the total aggregate thickness of its component parts (a+b+c...); in this case about a third. In (b) this hypothetical section is rotated and superimposed on the geological map of the Kam Formation on which the late faults have been replaced (see Fig. 13). The dotted line represents the approximate position of the unconformity between the Kam and Jackson Lake formations. Thus, although the unconformity on the geological map appears to cut completely through the volcanic section, if this model is correct it may only cut along the top of the complex such that only a relatively small volume of the mafic volcanics was actually removed.

Mafic volcanic rocks

Kam Formation – Yellowknife Bay

One of the major mafic volcanic rock units in the area is the Kam Formation at Yellowknife Bay. This formation extends north from Great Slave Lake over a distance of 50 km and consists of a homoclinal, steeply to vertically dipping succession of dominantly mafic volcanic flows. Within the sequence are units of felsic volcanics including the Townsite flows, that are described separately. The Kam Formation is of considerable interest as the two major gold mines in the region, as well as several smaller mines are located within it.

The map unit has a generally northern trend, due mainly to the left lateral offset of the formation by a series of northwesterly trending post-Archean faults. Flows within the formation, however, have a northeasterly strike and are truncated on the east by a north-trending unconformity. The western contact with a major granodioritic plutonic complex is intrusive although Padgham (1981a) has suggested small remnants of sialic basement to the volcanics occur at the north and south extremities of the formation. If the late faults are returned to their original pre-fault position (Fig. 13) it can be seen that the volcanics are truncated at an angle of about 20° by the unconformity between the Kam and the overlying Jackson Lake Formation. The cuspate intrusive contact to the west roughly parallels the unconformity so that at any one place the volcanics rarely have a thickness in excess of about 6500 m, although if correlations are made along marker horizons within the volcanic pile an aggregate thickness in excess of 10 000 m can be measured. In Figure 14, following the suggestion of Davidson (1967), a model is presented that shows how the original thickness of the volcanic sequence may be seriously overestimated if the accumulated thickness of correlated sections is used. It is suggested that the maximum thickness of the Kam Formation preserved at any point below the unconformity is probably closer to 6500 m than the more commonly accepted thickness of 10 000 m. The base of the volcanic section is not preserved due to intrusion by later granitic plutons and the top of the formation is either truncated by the unconformity or covered by Great Slave Lake.

Gibb and Thomas (1980) have suggested on the basis of a gravity anomaly that the Kam Formation extends to the south a short distance under Great Slave Lake before turning to the southeast and continuing for a distance of about 60 km (Fig. 15). This part of the formation is exposed only on the West Mirage and East Mirage islands at the mouth of Yellowknife Bay, where the mafic volcanic rocks have the same general trend as the anomaly as a whole. According to the model of Gibb and Thomas (1980), based on a detailed gravity survey (stations spaced at 3 km in the northern 2/3 of the anomaly and 6 km for the southern third), the formation

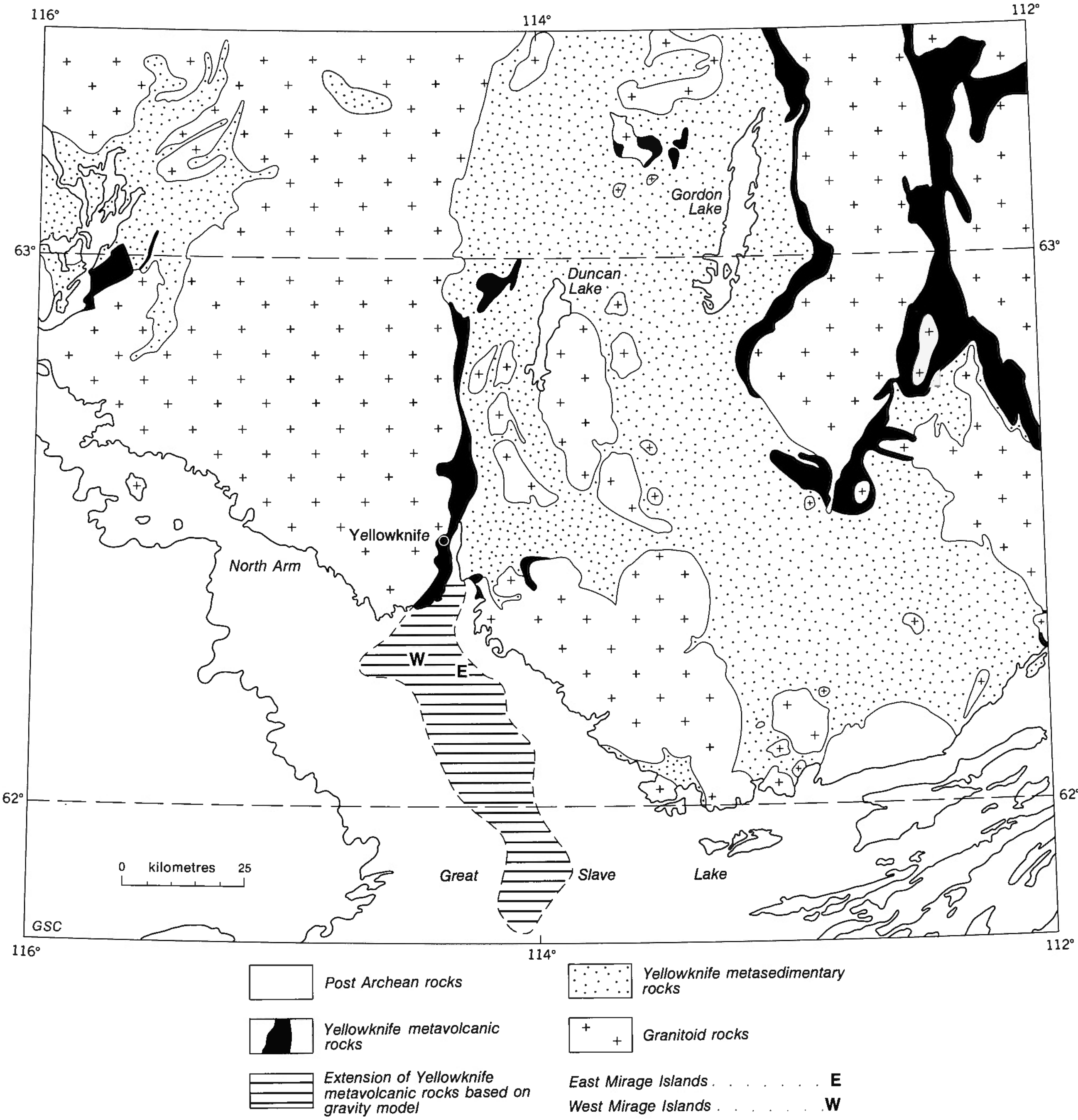


Figure 15. *Distribution of major volcanic units within and north of the Yellowknife-Hearne Lake area (outlined by broken line). The horizontally patterned area represents the southern extension of the Kam Formation mafic volcanic rocks southeastward under Great Slave Lake as modeled by Gibb and Thomas (1980) from detailed gravity data. The East Mirage and West Mirage islands south of the mouth of Yellowknife Bay, consisting of mafic volcanic rocks with a similar trend to that of the gravity anomaly, are the only exposed part of the southeasterly trending part of the Kam Formation. According to the gravity model of Gibb and Thomas (1980), this part of the formation has a significantly wider horizontal distribution (up to 15 km) and thinner present day vertical thickness (1-3 km) than the northerly trending part (less than 5 km wide and at least 7 km deep).*

occurs over a width of up to 15 km and a depth of 1 to 3 km which is significantly wider and shallower than the northern exposed part of the formation.

Along the north shore at the mouth of Yellowknife Bay 8 km southeast of Octopus Lake, Helmstaedt et al. (1981) report a unit of metamorphosed mafic volcaniclastic rocks and flows with metasediments including granitoid-bearing conglomerate at the top. The unit occurs as inclusions in the granodiorite and on islands. Because the southerly trend of this unit, as exposed on the island, differs from the northeasterly trend of the Kam Formation to the northeast, Helmstaedt et al. (1981) suggest there is an angular unconformity between this unit and the Kam. On the other hand, the southerly trend at this location over a distance of little more than 4 km may represent part of the change in trend of the Kam from northeasterly to southeasterly. A similar almost 90° change in trend is seen in the trend of a similar sequence of mafic volcanic rocks east of Gordon Lake (Fig. 15).

The Duck Formation east of Yellowknife Bay may be an eastward lateral extension of the main Kam volcanic pile. It is described in a later section. To the north between the Yellowknife River and Bell Lake another unit of mafic volcanics may be closely related to the main volcanic sequence at Yellowknife, but is separated from the Kam by a narrow zone of mixed gneiss, granitic rocks and mafic inclusions.

Various parts of the Kam volcanics have been mapped at 1:6000 scale by Henderson and Brown (1948, 1949, 1950, 1952a, b, 1966) at 1:8400 by Hauer (1979) and 1:10 000 by Helmstaedt et al (1979). The geochemistry of the volcanics has been studied by Baragar (1966) and the geochemistry of the extensive shear zones in the volcanics by Boyle (1961). Campbell (1949), Brown and Dadson (1953) and Brown et al. (1959) have discussed the geology of the mines in the volcanics. The following account is, for the most part, abstracted from the foregoing works.

Lithology

The volcanic sequence consists of both contemporaneous extrusive flows and intrusive dykes and sills of similar composition, with relatively minor amounts of volcaniclastic material throughout the sequence. The metamorphic zonation in the pile parallels the intrusive contact to the west (Boyle, 1961). Most of the volcanics are metamorphosed to amphibolite grade with a narrow zone about 0.5 km wide west of the unconformity at the north end of Yellowknife Bay at greenschist grade. Boyle (1961) divided the amphibolite grade rocks into a 0.5-1.5 km wide hornblende-andesine zone next to the granodiorite contact, and a fibrous amphibole-epidote-oligoclase zone to the east. The low grade rocks consist of a chlorite-albite assemblage. At all grades, however, primary structures are well preserved; the main difference in outcrop being the much darker and harder nature of the higher grade rocks. Primary structures such as pillows are commonly preserved, even within inclusions in the intrusive granodiorite to the west. Although the rocks dip vertically, structures in horizontal section are commonly only minimally distorted, although in vertical section there has been some degree of elongation. In the extensive shear zones in the formation all primary features are completely obliterated.

Within the volcanic pile massive and pillowed flows occur in about equal abundance (Henderson and Brown, 1966). In composition about two-thirds of the mafic part of the pile is basalt, while the remainder is mainly andesite (Baragar, 1966). Toward the top of the pile at the southern end of its outcrop area volcaniclastic breccia and tuff are increasingly abundant. Individual flows vary in thickness from about a metre to as much as 150 m and are similarly variable in lateral extent, although individual flows are commonly difficult to trace laterally unless they have some distinctive characteristic. Flow contacts are difficult to distinguish unless marked by flow-top breccia or tuffaceous material. Individual flows are either pillowed or massive; only rarely are individual flows made up of both a pillowed and a massive component.

In the pillowed flows the pillows range in diameter from about 10 cm to more than 1.5 m, with the average diameter about 0.5 m. The largest pillow reported is 9 m long and a metre high (Fig. 16, 17). The shape of the pillows varies from equidimensional to elliptical, with the larger pillows tending to be more flattened. In their detailed mapping of the flows, Henderson and Brown (1966) stressed that, with rare

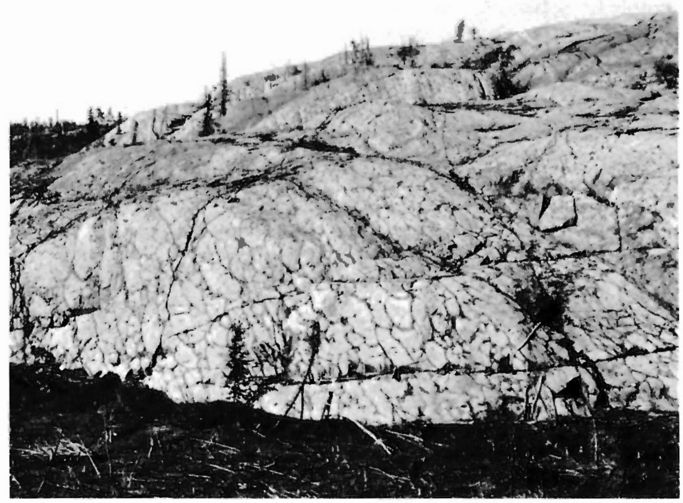


Figure 16. *Thick flow of mafic pillow lavas of the Kam Formation north of Yellowknife. The vertically standing flows striking into the photograph are facing towards the left (southeast). GSC 178134*

Figure 17. *Pillows in the mafic volcanic Kam Formation north of Yellowknife. The basal part of the pillow conforms to the space between the underlying pillows. The margin or pillow selvage, formed due to quenching of the molten lava, is clearly defined about most of the central pillow except where it is in contact with the pillow to the lower right. This might be explained if the "pillows" are actually tubes with, in this case, the two adjacent "pillows" being lava conduits active at the same time. The contact between them would cool more slowly and resulting selvages there would be different in aspect from the margins that were quickly chilled against seawater or other cold pillows. Vesicles in the lava are concentrated towards the upper margin of the pillow. The dark structure near the top of the pillow is a primary void space formed as a result of the partial draining of the tube. They commonly have flat basal surfaces that represent the horizontal at the time of solidification of the remaining lava within the pillow. GSC 178132*

exceptions, each pillow is separated from its neighbours by complete selvages, and so thought that the pillowed flow formed by the deposition of detached globules of lava while still in the molten, or at least plastic stage, and not by a process of budding. On the other hand, if the volcanic belt more or less marks the margin of a depositional basin (see later section) it is conceivable that the flows, extruding from a series of fissures parallel to the basin margin, flowed out into the basin (Fig. 17). The presently exposed section would then be more or less perpendicular to the direction of flow and so would not show the linked or budded nature of the structures, if indeed that is how they formed. The presence of some examples of very elongate pillows might suggest that lava tubes are present. Padgham (1980) has illustrated one such large, elongate, complexly budded pillow from the formation.

Massive flows commonly tend to be coarser grained than the pillowed flows, and are commonly difficult to distinguish from the more or less contemporaneous sills of similar composition, except where amygdules, ropy flow tops or other structures characteristic of extrusive flows are present.

In addition to the common pillowed and massive flows eight groups of variolitic flows occur within the sequence that are useful as distinctive marker horizons in the volcanic pile. These flows are almost always pillowed with the varioles consisting of centimetre-scale, spherical, lighter weathering, fine intergrowths of epidote, hornblende and a somewhat higher proportion than normal of plagioclase, locally with radiating texture (Fig. 18). Varioles are smaller and more widely spaced in the outer parts of the pillow and become coarser and commonly coalesce in the central part of the structure. The flows occur both as continuous flows that can be traced from one fault block to another, and as a series of lenses occurring at the same stratigraphic position.

Fragmental volcanic rocks occur throughout but are particularly abundant in the upper part of the formation. Breccia and agglomerate units vary in thickness from a metre to several hundred metres, although their average thickness is between 2 and 3 m. In some cases they grade along strike into finer grained thinner bedded tuffs. The clasts are similar in composition to the flow rocks. In "pillow" breccias the coarser clasts commonly have chilled margins and, except for the more irregular outlines, are very similar to the pillows in the flows. Other clasts are more angular and range up to 15 cm or more in size (average 6 cm). The fragmental units are typically unsorted and massive, although in some cases there is crude stratification.

Figure 18. Large variolitic pillow in the Kam Formation, 4 km north of Yellowknife. Spherical to ellipsoidal variolites are coarser and more widely spaced in the centre of the pillow and become finer and more concentrated toward the margins where their individual identity is lost. Hammer is about 50 cm long. GSC 178135

Tuffaceous beds also occur throughout the pile and in several instances form distinctive continuous marker horizons that can be traced across the pile. They range from sand-sized lithic to crystal tuffs to very fine grained cherty material. They vary from a few centimetres to as much as 30 cm in thickness. Individual tuff layers are commonly graded, with the uppermost part being cherty. Some of the tuffs have a high content of graphite and pyrite that Boyle (1961) has suggested may be biogenic.

Aside from the breccias and finer grained sediments there is little in the way of sedimentary material within the volcanic sequence. Rarely, thin lenses of discontinuous beds of crossbedded volcaniclastic detritus of local origin can be seen (Fig. 19). Minor thin oxide iron formation occurs in the northernmost part of the Kam Formation 1 km west of Greyling Lake. A prominent aeromagnetic anomaly north of Likely Lake in the same region would suggest a second oxide iron formation locality although it was not seen in the field.

The flows are extensively intruded by dykes, sills and irregular bodies of similar composition to the flows. These intrusive bodies, which locally make up to 25% of the formation, appear to be penecontemporaneous with the flows and fragmental deposits and probably represent feeders to them, although Henderson and Brown (1966) in their detailed mapping of the volcanic sequence could find no direct evidence of this. The major sills, which are up to 200 m thick, occur in two groups, one near the top of the pile and the other associated with the felsic Townsite flows in the central part of the formation. The irregular-shaped bodies are much more variable in size. Contact relations are also varied, ranging from sharp chilled contacts to gradational contacts, but in which the volcanic country rock has obviously been affected by the intrusion. The dykes, for the most part, are younger than the other intrusions and are concentrated in the northern and southern parts of the belt. Low in the section in the northern part of the pile the dykes parallel the strike of the flows and dip moderately east, whereas in the upper part of the pile in the south their strike is more discordant to that of the flows and the dip is to the west. Their attitude in the pile corresponds approximately to the general attitude of the shear zones in that part of the pile.

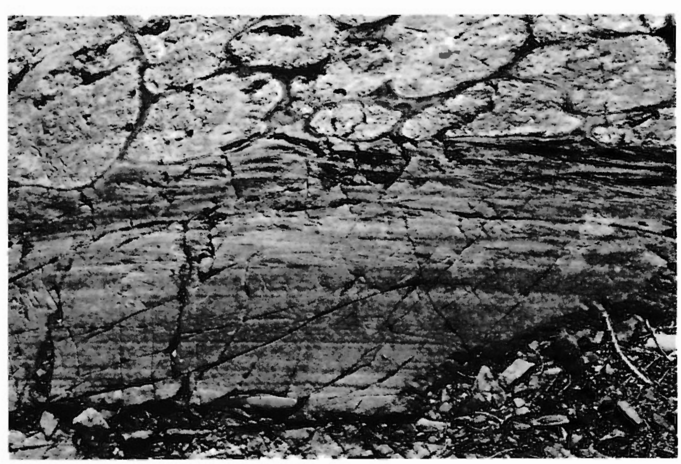


Figure 19. Rare volcanic sandstone bed in the Kam Formation north of Yellowknife, formed from hyaloclastic debris from local volcanic extrusions. Note climbing ripple structures just below the overlying pillowed flow. GSC 178139

One other type of intrusion in the Kam Formation is the irregular network of felsic porphyries that are particularly abundant southeast of Ryan Lake (Fig 20). Their origin and significance has been regarded as somewhat of an enigma (Boyle, 1961, 1976; Henderson and Brown, 1966), but, it is suggested that they may represent feeders to a felsic volcanic centre similar to that that resulted in the felsic Townsite flows or the felsic centre at Berry Hill, but that was eroded prior to the deposition of Jackson Lake Formation.

Petrographically, both the mineralogy and texture of the higher grade volcanics and intrusive bodies is secondary. In the lower grade volcanic rocks primary textures are locally

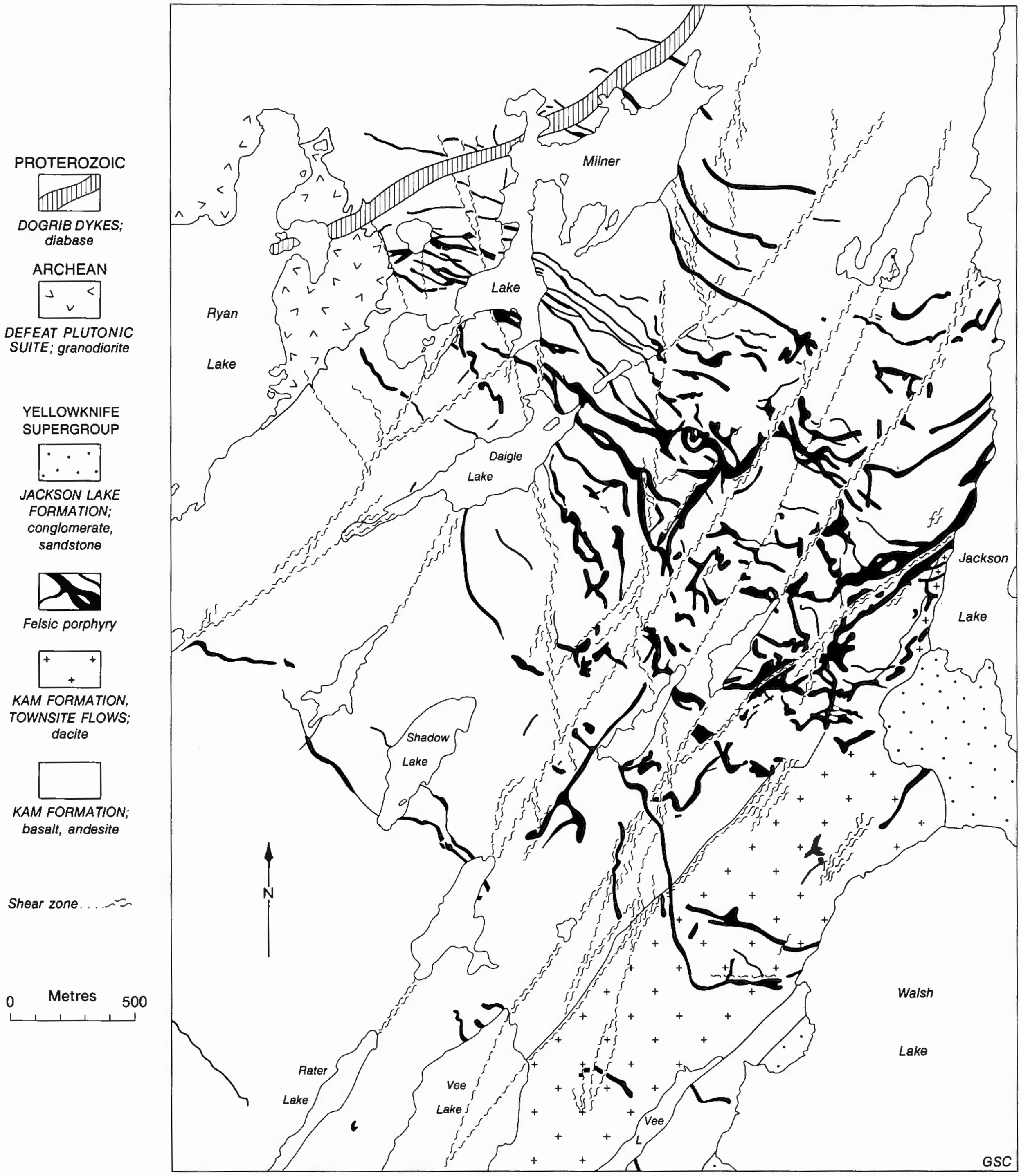


Figure 20. *Part of the Kam Formation between Ryan Lake and Walsh Lake where a swarm of dykes, sills and irregular bodies of felsic porphyry intrude the mafic volcanics. These felsic rocks both cut and are cut by the shear zones in the mafic volcanics but are truncated by the unconformity below the Jackson Lake Formation. They are thought to represent a conduit system feeding a felsic volcanic centre that is more or less contemporaneous with the formation of the shear zones but which was eroded prior to the deposition of the Jackson Lake Formation. After Henderson and Brown (1966).*

well preserved as for example shown by Baragar et al. (1979). Relict ophitic textures are preserved in some of the lower grade dykes. Approximately two-thirds of the rock consists of amphibole and chlorite minerals and about one-third plagioclase and epidote, with minor amounts of carbonate, white mica, biotite, apatite, zircon and quartz. At higher metamorphic grades near the intrusive contact with the granodiorite, the amphibole is hornblende and the plagioclase andesine. At progressively lower grades amphibole is paler and of fibrous habit, chlorite is more abundant, and plagioclase is oligoclase. The similarities of the extrusive and intrusive rocks supports the suggestion that they are closely related.

Baragar (1966), who sampled a section through the formation (Table 4), has shown that on the basis of their chemistry, the volcanic rocks can be divided at the Townsite flows into two cycles. The upper parts of each cycle are enriched in felsic components. At the top of the first cycle are the felsic Townsite flows, which mark an abrupt change in the chemistry (Table 5). Felsic volcanics (the Banting Formation) also occur above the dominantly mafic second cycle, although the mafic and felsic volcanics are separated by an unconformity and the Jackson Lake lithic sandstones. Superimposed on these two calc-alkalic differentiation cycles are a series of short (kilometre-scale) iron-enrichment cycles. There is, however, no overall iron-enrichment trend through the whole formation. Baragar (1966) suggested the series of iron-enrichment trends is due to the differentiation of the original tholeiitic magma in a series of chambers that were successively tapped as volcanism at the surface took place. The calc-alkalic trends were interpreted as due to contamination of the differentiating tholeiitic magma by sialic crustal material in which it was contained.

Condie and Baragar (1974), in a rare earth element study of the Kam volcanic rocks, reported that the tholeiitic basalts have a flat pattern to a pattern showing slight enrichment in light REE with the upper cycle basalts tending to show a somewhat greater light REE enrichment than the lower. In the andesites there is a light REE enrichment and a moderate negative europium anomaly. They suggested these sorts of patterns are what could be expected from either progressive shallow fractional crystallization of tholeiitic magma or by decreasing amounts of equilibrium melting of a plagioclase bearing mantle source.

Geochronological data from the Kam Formation consists of a Rb-Sr age estimate of 2570 ± 160 Ma (Sr_i = 0.7022 ± 0.0023, $\lambda^{87}Rb$ = 1.42) and a ^{207}Pb-^{206}Pb age of 2610 Ma from zircons from the Townsite felsic volcanic flows within the Kam Formation (Green and Baadsgaard, 1971).

Nature and origin of gold-bearing shear zones

The Kam mafic volcanics are displaced by faults of two ages (Fig. 13). The youngest are a set of north-northwesterly post-Archean faults, movements on which have caused the originally northeast-trending volcanic belt to assume a northerly trending aspect. The earlier faults are of considerably greater interest, both from an economic as well as a geological point of view, as they are the locus of the

Table 4. Averages of chemical analyses, Kam Formation

	1	2	3	4	5
SiO_2	50.4 (0.8)	50.7 (2.4)	50.7 (2.0)	49.4 (0.9)	56.0 (4.3)
TiO_2	.86 (0.12)	1.18 (0.57)	0.92 (0.14)	1.02 (0.29)	1.25 (0.27)
Al_2O_3	14.7 (0.8)	13.8 (1.0)	14.6 (1.0)	13.9 (0.7)	14.5 (0.6)
Fe_2O_3	2.7 (2.0)	3.0 (0.8)	2.1 (0.6)	2.3 (0.8)	2.3 (0.7)
FeO	9.4 (1.4)	9.2 (2.6)	9.6 (1.2)	9.8 (0.7)	7.9 (2.2)
MnO	.18 (0.03)	.20 (0.07)	0.17 (0.02)	0.19 (0.05)	.12 (0.03)
MgO	5.5 (1.0)	4.3 (1.8)	5.2 (1.4)	5.5 (0.9)	4.2 (0.8)
CaO	9.1 (1.4)	8.2 (3.4)	9.7 (0.9)	8.8 (1.0)	6.3 (1.5)
Na_2O	2.5 (0.5)	2.8 (0.9)	2.4 (0.6)	2.7 (0.7)	4.0 (0.9)
K_2O	.2 (0.2)	0.3 (0.2)	0.4 (0.2)	0.4 (0.2)	0.6 (0.3)

Data from Baragar (1966) Table II

Bracketed value is the standard deviation

Columns 1 through 5 represent averages of a series of analyses from increasingly younger parts of the Kam Formation mafic volcanic sequence

1. Average of 13 analyses for a section between Chan Lake and Oro Lake (Average 1 of Baragar, 1966)
2. Average of 18 analyses for a section between Landing Lake and Vee Lake (Average 2 of Baragar, 1966)
3. Average of 19 analyses for a section between Frame Lake and Negus Point (Average 4 of Baragar, 1966)
4. Average of 10 analyses for a section between lake southeast of Kam Lake and Kam Point (Average 5 of Baragar, 1966)
5. Average of 8 analyses for a section on islands east of Octopus Lake (Average 6 of Baragar, 1966)

Table 5. Averages of chemical analyses, Banting Formation and the Townsite flows (Kam Formation)

	1	2
SiO_2	62.9 (2.0)	66.3
TiO_2	0.49 (0.10)	0.62
Al_2O_3	15.0 (1.0)	15.5
Fe_2O_3	1.2 (0.6)	1.4
FeO	2.9 (0.8)	2.5
MnO	0.19 (0.20)	0.20
MgO	0.9 (0.4)	0.03
CaO	2.8 (1.1)	3.6
Na_2O	3.4 (1.1)	3.7
K_2O	1.9 (0.7)	2.1

Data from Baragar (1966) Table II

Bracketed value is the standard deviation

1 Average of 7 analyses for a section in the Banting Formation west of the Yellowknife River at the north end of Yellowknife Bay (Average 7 of Baragar, 1966).

2 Average of 2 analyses from the Townsite flows (Kam Formation) east of Frame Lake (Average 3 of Baragar, 1966).

major gold deposits of the Yellowknife area. These structures are described in detail by Campbell (1949), Boyle (1961, 1979), Brown and Dadson (1953), Brown et al. (1959) and Henderson and Brown (1966) and are only briefly reviewed here, although a different interpretation as to their origin and its implications is suggested.

The shear zones consist of two types. Small, narrow zones parallel to the attitude of the flows occur in three distinct systems. They are commonly associated with tuffaceous units within the pile. The much larger discordant structures occur in several separate systems and contain the economic gold deposits. Individual shear zones within these systems range from distinct breccia with well-preserved clasts up to 15 cm to schist zones in which the original nature of the rocks is obliterated. In some cases there is a gradation from breccia to schist. The contact between relatively undeformed country rock and the shear zone is commonly gradational. The shear zones, particularly the larger discordant zones, are complex consisting of an interlacing network of sheared material separated by blocks of only weakly deformed volcanics. For example, in the Con system, Henderson and Brown (1966) describe from the surface a part of the system consisting of two shear zones 3 m wide separated by about 17 m of pillowed flows, with a third shear zone 13 m wide 30 m away. However, at the 500 foot level in the mine the three individual zones have joined to form a single zone 35 m wide. In the largest system, the Giant-Campbell, the system is up to 400 m wide, and again consists of several shear zones separated by less or minimally deformed blocks. The concordant shear zones are similar, although they are much narrower and more simple in general. The ore bodies occur as narrow discontinuous bodies within the shear zones near the ends of unsheared blocks where deformation was particularly intense and complex, and dilatent zones were available for quartz vein accumulation.

The flows face southeasterly and are vertical or dip steeply to the southeast. The discordant shear zones intersect the strike of the flows at a moderate angle in the southern or stratigraphically higher part of the section, but are more concordant in the northern or lower part of the formation. In general the shear zones dip to the west between 40 and 60° although some parts of the system dip east and indeed appear to be folded into a complex syncline-anticline pair (Brown and Dadson, 1953; Brown et al., 1959). Shear planes within the shear zones have a similar strike but a generally steeper dip than the zones themselves. Movement on the zones, where it can be demonstrated, is west side up. This corresponds to the movement sense indicated by cleavage relations in most zones. Campbell (1949) determined a 350 m displacement on the Con system; Boyle (1961) determined a 200 m displacement on part of the Negus Rycon system, and Brown et al. (1959) reported movement of over 450 m on part of the Giant Campbell system.

As far as the origin of these structures is concerned, most authors have concluded that the structures are the locii of thrust faults and that the faulting was related to the folding of the supracrustal rocks and intrusion of the granodiorite plutons that lie to the west (Brown et al., 1959; Boyle, 1961; Henderson and Brown, 1966). There seems to be little doubt that the shear systems involve some displacement. However, another possibility might be considered regarding the time and sense of displacement. It is suggested that initial movement took place <u>during</u> the accumulation of the Yellowknife supracrustal rocks, <u>before</u> they were folded into their present configuration, and that the displacement was normal rather than reverse. There may well have been later movement on these structurally weaker zones during subsequent deformational and intrusive events.

Supporting this suggestion, the following should be considered. The shear zones of concern occur only within the mafic volcanic pile. That is to say the apparent displacement of marker horizons in the volcanic sequence (Henderson and Brown, 1966) along the Giant-Campbell shear zone does not extend into the overlying sediments above the unconformity. In the Prosperous Lake map area of Jolliffe (1946) the contact between the Kam and Jackson Lake formations is not offset where the shear zones (not plotted on Jolliffe's map) would intersect the contact. Thus movement on the shear zones is older than the unconformity. In this regard the relationship between the shear zones and a set of felsic porphyry dykes is of interest (Fig. 20). In the north-central part of the Kam mafic volcanics there is an extensive network of anastomosing crosscutting dacite porphyry dykes that may represent the subvolcanic conduit system to a felsic centre. According to Henderson and Brown (1966) the porphyry bodies are also truncated at the unconformity between the Kam and the overlying Jackson Lake volcanic lithic sandstones. Thus the extrusive part of the centre, if it existed, was eroded prior to deposition of the Jackson Lake sandstones which are derived in part from a felsic volcanic source (see later section). The relationship between these porphyries and the major shear zones is equivocal as the porphyries both cut and are cut by the shear zones (Henderson and Brown, 1966). An interpretation of this is that initial shear zone movement was contemporaneous with the emplacement of the dyke complex (and possibly felsic volcanism?) with the shear zones cutting earlier dykes and being cut by later dykes, but with everything being truncated at the unconformity.

Since the Banting felsic volcanics, the Burwash greywackes and mudstones, and the equivalent of the Kam mafic volcanics (the Duck Formation) are all conformable on the east limb of the major syncline at Yellowknife (Henderson, 1970) and the Banting volcanics and Jackson

Lake sandstones on the western limb are also conformable (Baragar, 1975; elsewhere this report), the unconformity is a local feature representing a short length of time during deposition of the Yellowknife Supergroup. The shear zones, which are truncated by the unconformity, also must have formed during the accumulation of the Yellowknife Supergroup.

It is suggested that the shear zones represent gravitational sliding or creep due to the instability of the main volcanic pile. Earliest movement was on concordant zones associated with tuff beds (Boyle, 1961; Henderson and Brown, 1966). The water-saturated tuff beds acted as planes of weakness along which the early movement could start. Later movement was generally on the discordant west-dipping slip surfaces. Movement on the shear zones as presently oriented has been shown to be west side up (Campbell, 1949; Boyle, 1961; Fig. 21). If the flows are rotated into a horizontal position about the more or less horizontal synclinal axis in Yellowknife Bay, the shear zones become a series of moderately southeastward dipping planes along which blocks of volcanics slipped downward toward the adjacent basin to the east. A recent example of this sort of collapse was reported by Van Bemmelen (1954) where 1000 years ago part of the cone of the Merapi volcano in central Java, including the crater and part of the underlying conduit, an area 18 km long and up to 10 km wide made up of several sliding masses, slumped into an adjacent valley, lowering the original 3300 m height of the volcano over 300 m, and initiating renewed eruption of the volcano. Similarly, Moore (1966) and Swanson et al. (1976) have interpreted the Hilina fault system of Kilauea volcano on the island of Hawaii as being due to gravity slides that are still moving (Fig 22). Duffield et al. (1982) described and interpreted similar structures on Piton de la Fournaise, La Reunion, as well as the Hilina structures on Kilauea and suggested that the large landslide blocks form due to failure along the unbuttressed flanks of the volcanic complex as a result of inflation of the active volcanic edifice. Such a mechanism seems not unreasonable for the formation of the structures in the Kam Formation.

The slip surfaces of these structures at Yellowknife would provide conduits, certainly for meteoric water percolating downward into the volcanic pile and, if volcanism was still active as seems most likely, could also have provided a site for the passage of fumarolic material upward. Both processes would greatly speed up the alteration and devitrification of the volcanic flows in the vicinity of the movement surfaces. This early alteration would explain the width of the present shear zones and the diffuse nature of their contact with the volcanics. The altered zones would be a zone of weakness along which later movement during folding could take place. If the orientation of the zone was at an angle to the direction of later movement, the zone itself could have been folded due to shearing associated with this later movement. The folded aspect of the shear zones in the Giant Yellowknife mine is clearly illustrated by Brown et al. (1959) although they felt the formation and apparent deformation of the zones was concurrent.

Boyle (1961) has suggested that the gold and associated mineralization in the deposits in the shear zones were derived entirely from the altered volcanics in the shear zones during metamorphism of the volcanic rocks. The early formation of the shears and their alteration, together with the much greater volume of water available in the shear zones at that time, would provide an easier mechanism for the mobilization of the mineralizing elements. In a recent study by Kerrich et al. (1977) it was suggested the Yellowknife deposits are hydrothermal in origin on the basis of the reduced nature of the iron in the shear zones although like Boyle (1961) they believed that shearing and mineralization took place during folding and metamorphism of the pile.

Duck Formation – Yellowknife Bay

On the east side of Yellowknife Bay, between Akaitcho Bay and Preg Lake, is a discontinuous unit of mafic to intermediate volcanics, the Duck Formation, that extends over a distance of 20 km east from Yellowknife Bay. The Duck Formation occurs in the axis of a variably plunging anticline. The west half of the structure more or less parallels the contact with the northwest plutonic lobe of the

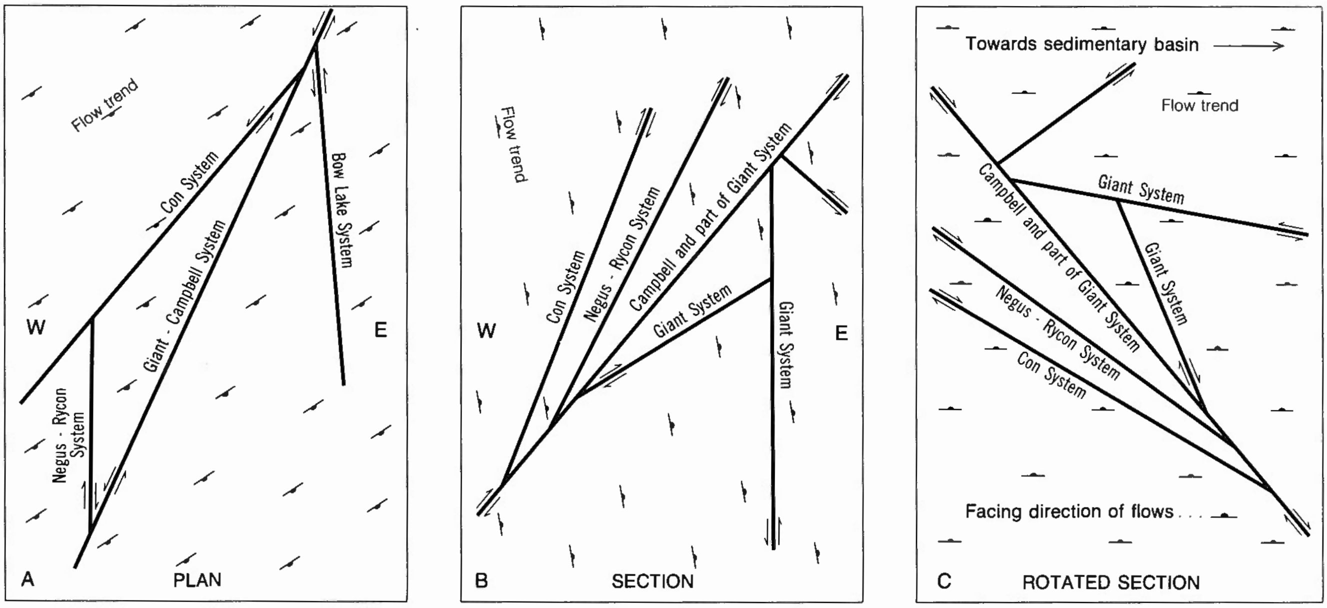


__Figure 21.__ Diagrammatic plan and section after Boyle (1961) showing sense of movement along the shear zones and their relationship to each other and to the volcanic flows of the Kam Formation. If the flows are rotated into a horizontal position the movement on the shear zones is for the most part normal, with the upper blocks moving down and out towards the adjacent sedimentary basin to the east.

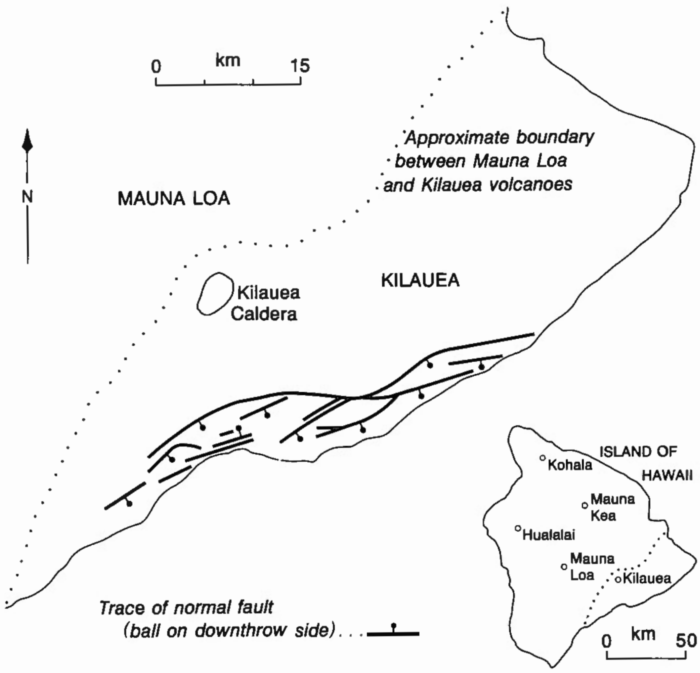


Figure 22. *On the south coast of the island of Hawaii on the flanks of the Kilauea volcano is a series of normal faults that have been interpreted as gravity-induced slumps of the volcanic pile. This would be a analogue for the origin proposed for the shear zones in the Kam Formation at Yellowknife (after Swanson et al., 1976; compare Fig. 13).*

Defeat Plutonic Suite granodiorite that lies to the south and east. The northern lobe of the granodiorite, however, intrudes the core of the anticline so that the easternmost part of the volcanic unit now mantles the plutonic lobe. The anticline plunges to the southwest in Yellowknife Bay, south of Akaitcho Bay, and to the northwest at Preg Lake. The volcanics are conformably overlain by greywacke-mudstone turbidites of the Burwash Formation. The base of the Duck Formation is nowhere exposed and in the east it is intruded by the granodiorite. The greywacke-mudstone turbidites and very vesicular volcanic flows (not conglomerate) interstratified with the volcanics on the south shore of Duck Lake are not thought to be correlative with the conglomerates and crossbedded sandstones of the Jackson Lake Formation on the west side of Yellowknife Bay as previously mapped (Unit 4 of Jolliffe, 1942).

The volcanics of the Duck Formation consist of mafic to intermediate massive to pillowed flows and breccia units. In general the volcanics tend to weather a lighter green than the mafic flows of the Kam Formation to the west, which may indicate a more intermediate composition. The flows are commonly vesicular, with irregularly distributed quartz-filled amygdules up to several centimetres in diameter.

The upper contact of the formation with the sediments is conformable although the contact itself is commonly sheared due to the contrasting competency of the two units. Several elongate lenses of mafic flows occur within the sediments, the largest of which, at the west end of Duck Lake, is about 1 km thick. Most are only a few hundred metres thick. As in the main formation itself, these smaller units consist of both flows and breccia units. Minor thin greywacke units are also present locally near the top of the formation. In the sediments above the contact are abundant mafic tuffaceous horizons that vary from a few centimetres to a metre or more in thickness. They contrast with the greywackes in that they are dark green to yellow, are vaguely to sharply layered, are composed of varied proportions of amphibole and plagioclase, and in some cases have a sharp basal contact and are compositionally and texturally graded.

The contact of the Duck volcanics with the greywackes is similar to that of the volcanics southwest of Clan Lake and differs from the rather abrupt contact between the metasediments and other mafic volcanic sequences such as east of Gordon Lake or the previously described Kam Formation, which is separated from the sediments by an unconformity. It is thought that the Duck Formation represents the much thinner, distal eastern extension of the Kam mafic volcanics which occur to the west. Although the units cannot be traced directly due to complex structure and lack of exposure in Yellowknife Bay, all islands in the mouth of Yellowknife Bay south of the Sub Islands are underlain by similar mafic volcanic rocks.

Mafic volcanic rocks – Clan Lake

North of the main mafic volcanic pile at Yellowknife and immediately west of the Yellowknife River is a smaller, more deformed homocline of mafic volcanics. It has been mapped at 1/2 mile to the inch scale by Jolliffe (1939). It is separated from the Kam volcanics to the south by a complex zone of mafic inclusions in granitoid intrusions. The granitoid rocks to the west are thought to be in part possible basement to the Yellowknife Supergroup (see Anton Complex) although there are clearly younger intrusive granitic phases present as well. The contact between the volcanics and these granitoid rocks is faulted, although similar granitic dykes and sills occur in the volcanic rocks in the contact area and increase in abundance toward the fault. The northeastern contact between the volcanics and the sediments is conformable with volcanic units generally less than 300 m thick occurring above the main sediment-volcanic contact. The eastern contact with the sediments is a major north-trending fault, which if the displacement is of the same magnitude as other faults in the region, would suggest these volcanics were originally in closer proximity to the dominantly felsic volcanics east of Clan Lake and indeed may be related. Most of the sequence is metamorphosed to amphibolite grade.

The volcanic sequence consists mainly of dark to light green weathering mafic volcanics with both pillowed and massive flows. Volcanic breccia units occur throughout and locally are as much as 10 m thick. Where the clasts are better rounded, the coarse breccias are crudely stratified. Finer volcaniclastic units are also present, notably at the contact between the volcanic sequence and the overlying sediments. The volcanic units above the contact within the sediments are mainly mafic volcaniclastics that have compositional layers 2 to 3 cm thick. Several gabbro sills intrude both the sediments and the volcanics in the contact area. Within the upper part of the mafic sequence and decreasing in abundance downward are minor felsic volcanic units, the largest of which is 300 m wide and about 3000 m long. Other felsic units are much thinner but locally make up as much as 25% of the section. Most are very fine grained, sheared, light weathering felsites in which all primary textures have been destroyed. In a few of the coarser units, primary clasts can be seen.

Oxide iron formation occurs within the volcanic sequence near the lowermost part of the section at Bell Lake, and consists of two units up to 20 m thick of thin layered (centimetre scale) recrystallized chert, magnetite and very minor iron silicate iron formation. The iron formation causes a prominent anomaly on the aeromagnetic maps of the region (Geological Survey of Canada, 1969b). Other similar prominent anomalies are present in the unit near the eastern contact near Quyta Lake and to the north at Nelson Lake, and are presumed to represent iron formation as well. Similar oxide iron formation occurs to the south in the northernmost part of the Kam Formation.

The relationship of these volcanics with the Kam Formation to the south is not clear, but they probably represent a related but separate smaller volcanic buildup. The two mafic sequences, although quite similar, differ in some respects. There are more felsic volcanic units interlayered with mafic flows in the upper part of the Clan Lake sequence than in the Kam Formation. Also, there is no unconformity between these volcanics and the overlying sediments as with the Kam Formation, suggesting the Clan Lake mafic volcanics were never raised above sea level during or soon after their extrusion. The abundant felsic layers in the upper part of the sequence may represent distal equivalents of the felsic volcanic centre east of Clan Lake (see Felsic volcanic rocks). As mentioned previously the mafic volcaniclastic and flow unit associated with the felsic volcanic centre east of Clan Lake may be correlative with the mafic volcaniclastic units interbedded with the greywacke-mudstones above the main mafic volcanic sequence.

Mafic volcanic rocks – Stagg Lake

Between Russell and Stagg lakes at the northwestern part of the area is a relatively thin unit of predominantly mafic metavolcanics. The unit is about 700 m thick and consists predominantly of thinly layered mafic volcaniclastics with local pillowed flows. The volcanics are now amphibolites at medium metamorphic grade. Pelitic to psammitic schists of similar grade occur on both sides of the volcanic unit and minor beds of argillaceous metasediment occur within the metavolcanic sequence.

Although pillow structures are preserved locally, not enough data are available to state the structural relationship of the volcanics to the adjacent sediments. The relationship of this unit to the major felsic volcanic unit to the west at Russell Lake is not clear, although intermediate to mafic volcanics are present at the base of that dominantly felsic volcanic sequence.

The mafic unit north of Stagg Lake has been traced 18 km north of the present area in the Snare River map area (Lord, 1942). To the west and just north of the present map area two other similar thin elongate units of mafic volcanics occur within the metasedimentary terrane (Lord, 1942).

Cameron River Basalt, Webb Lake Andesite, Dome Lake Basalt – northern Cameron River

A major mafic volcanic unit extends from Upper Ross Lake along the Cameron River to the northern margin of the map area, a distance of over 40 km. Just north of the map area the strike of the belt abruptly changes through almost 90° from northeasterly to northwesterly and then extends to the north another 50 km (Moore et al., 1951). The main part of the volcanic belt was mapped at 1 inch to 1 mile scale by J.F. Henderson (1941a, b) and the southernmost part by Fortier (1947). Baragar (1966) made a detailed geochemical section across the central part of the unit. The volcanics were remapped and volcanic facies analyzed by Lambert (1974, 1977, in press). The description of the units that follows is based in large part on the field data of Lambert (in press).

The unit varies in outcrop width from between 2500 to 4000 m over most of its extent, but where it changes strike from northeast to southeast it thins abruptly and pinches out over a distance of about 13 km. Structurally the volcanics occur in a steeply dipping to overturned homoclinal succession. As with most of the mafic volcanic sequences, it is conformably overlain by greywacke-mudstone turbidites. Over most of the contact relief is minimal, but east of Waite Lake relief on this surface is of the order of 550 m over 1700 m. Lambert (1977) has suggested that the relief may be due to the presence of a volcanic centre in this vicinity. The contact with the overlying sediments is typically abrupt, although local, minor thin tuffaceous units occur within the greywackes just above the contact.

There is evidence that the metagranitoid rocks to the east, part of the Sleepy Dragon Complex, are older than the volcanic rocks (Baragar, 1966; Davidson, 1972; Baragar and McGlynn, 1976). This is based on the structural and metamorphic contrast between the volcanics and the adjacent granitic rocks and the presence of swarms of mafic dykes and sills in the granite that may represent feeders to the volcanic flows. The nature and significance of the unconformity is discussed in the section on the Sleepy Dragon Complex. Although there is little doubt that the complex is older than the volcanics, the unconformity surface itself is rarely preserved. In most places there is evidence of movement on the contact, although when the relief on the surface is considered the amount of movement is probably minimal. Thus the preservation of the basal part of the volcanic unit is close to complete.

The belt has been divided into three formations by Lambert (in press). The largest is the Cameron River Basalt which forms most of the belt. It consists mainly of basaltic pillowed lavas, pillow breccias and rather minor (<5%) massive lavas. The pillows are commonly flattened to some extent and are particularly so close to the contact with the Sleepy Dragon Complex. Discontinuous, massive to poorly bedded, volcaniclastic units formed largely from hyaloclastic debris occur locally throughout the formation, but are most abundant in the upper part of the unit east of Allan and Fenton lakes and to the southeast where the belt thins and pinches out. Although no distinctive marker horizons such as the tuff units and variolitic flows in the Kam Formation at Yellowknife have been recognized, feldsparphyric flows with associated dykes and sills occur locally and some amygdaloidal flows are present at the top of the formation. Near Bambi Lake a 10-15 m thick unit of mainly black shale with grey chert, siltstone and minor oxide iron formation occurs over a distance of 5 km.

West of Webb Lake the upper part of the Cameron River Formation interfingers with the Webb Lake Andesite. The Webb Lake attains its maximum thickness of about 1400 m in this area but thins abruptly to the northeast where it conformably overlies the Cameron River and less abruptly to the south, again conformably above the Cameron River Formation. It is conformably overlain by the greywacke mudstone turbidites of the Burwash Formation. At its thickest locality there is as much as 500 m relief on domes at the sediment-volcanic contact which contrasts strongly with the almost minimal relief on the same contact to the northeast and south. The formation consists mainly of andesite in thick pillowed lava and pillow breccia sequences that are devoid of internal layering for the most part but are commonly interlayered in varied proportions. Individual units are discontinous laterally.

The Dome Lake Basalt occurs as a 300 m thick dome-like unit above the Webb Lake Andesite at the top of the western most part of the belt. It consists of highly amygdaloidal pillowed basalt that alternates with volcaniclastic units.

Two unnamed rhyolite units have been mapped within the dominantly mafic volcanic belt. The lens shaped units are up to 400 m thick and occur at the contact between the

Webb Lake Andesite and the Cameron River Basalt. They consist of white to buff, massive rhyolite with associated breccias and finer clastics. A similar rhyolite unit occurs within the Burwash sediments at Upper Ross Lake. Other smaller unmapped rhyolite to rhyolite breccia units occur within the other more mafic formations of the belt.

Throughout the belt are abundant dykes and sills of similar composition to the volcanics they intrude. They form up to 35% of the volcanic belt and are particularly abundant between Webb and Paterson Lakes (Lambert, 1982). The intrusions are massive, medium- to fine-grained, commonly chilled against the volcanics they intrude and are metamorphosed to the same grade as the volcanics. They are thought to represent feeders to the volcanic sequence as in rare cases, clearly intrusive bodies can be traced into pillowed flows. On the basis of stratigraphic correlations at various levels in the volcanic sequence Lambert (in press) has shown that the mafic dykes and sills at the base of the sequence had to be emplaced before the volcanics in the upper part were extruded.

A similar suite of mafic dykes, sills and irregular intrusions occurs within the Sleepy Dragon Complex that underlies the volcanic sequence (Henderson, 1941b). It has been suggested that these also represent feeders to the volcanic sequence (Baragar, 1966). Their distribution is variable, with the densest concentration between Webb Lake and Upper Ross Lake within about 3 km of the contact between the granitic rocks and the volcanics. In this area the intrusions, making up about 20% of the rock, are predominantly sills and lenses parallel to the foliation in the granitic rocks, which in general is parallel to the granite-volcanic contact, and irregularly shaped bodies particularly in the vicinity of the contact (Fig 5). Farther north, southeast of Paterson Lake, there are fewer intrusions and they take the form of isolated dykes perpendicular to the strike of the flows. These mafic intrusions, like the flows, have been metamorphosed so that the mineralogy is now dominantly amphibole and plagioclase. Primary textures, in many cases, have been well preserved and include chilled margins and original coarse (1-3 cm) plagioclase phenocrysts concentrated in the centre of some dykes (Fig 6). In some cases the sills have a thin biotite-rich selvage next to the foliated granitic rocks that presumably formed during metamorphism and shearing of the dyke assemblage. These intrusions are clearly distinct from the essentially unaltered post-Archean diabases and the highly altered, recrystallized and deformed amphibolite bodies also associated with the granitic rocks.

Table 6. Average and single chemical analyses of volcanic rocks in the Cameron River area

	1		2	3
SiO_2	54.7	(4.6)	79.1	54.0
TiO_2	1.11	(0.27)	.02	.95
Al_2O_3	15.0	(1.1)	12.6	15.1
Fe_2O_3	1.6	(0.7)	0.0	1.8
FeO	7.9	(2.0)	1.3	7.2
MnO	0.19	(0.06)	.01	.16
MgO	4.2	(1.5)	1.3	3.8
CaO	7.3	(2.0)	0.9	7.0
Na_2O	3.9	(1.2)	3.5	2.8
K_2O	0.4	(0.3)	2.5	1.4

Data from Baragar (1966) Table II

Bracketed value is the standard deviation

1 Average of 10 analyses from a mafic volcanic between Webb Lake (west end) and Cameron River (Average 8 of Baragar, 1966)

2 Quartz latite (Analysis 9 of Baragar, 1966)

3 Andesite (Analysis 10 of Baragar, 1966)

Lambert (in press), in a discussion of the chemistry of the volcanic rocks, reports an abrupt change from tholeiitic to calc-alkaline composition that corresponds to the contact between the Cameron River Basalt and the Webb Lake Andesite. He concludes that the volcanics were derived from mantle derived tholeiitic basalts that were extruded directly or spent some time in shallow magma chambers where they differentiated along a calc-alkaline trend. These chambers were subsequently tapped and calc-alkaline andesite to rhyolite extruded. Baragar (1966; Table 6) analyzed material from a section through the Cameron River and Webb Lake formations noting a subtle, somewhat scattered but general increase in sialic components with stratigraphic height. He considered the data similar to the more complete section through the Kam Formation to the southwest which was interpreted as representing an initial tholeiitic magma contaminated by sialic crust resulting in extrusions with calc-alkaline characteristics.

An Rb-Sr age estimate of 2574 ± 200 Ma (Sr_i = 0.706, $\lambda^{87}Rb$ = 1.42) for the volcanics of the Cameron River area was made by Green (1968).

The transition from dominantly mafic volcanics to sediments along strike of the belt is well exposed east of Upper Ross Lake. In many of the other belts this relationship is obscured by structural complications or granitic intrusion.

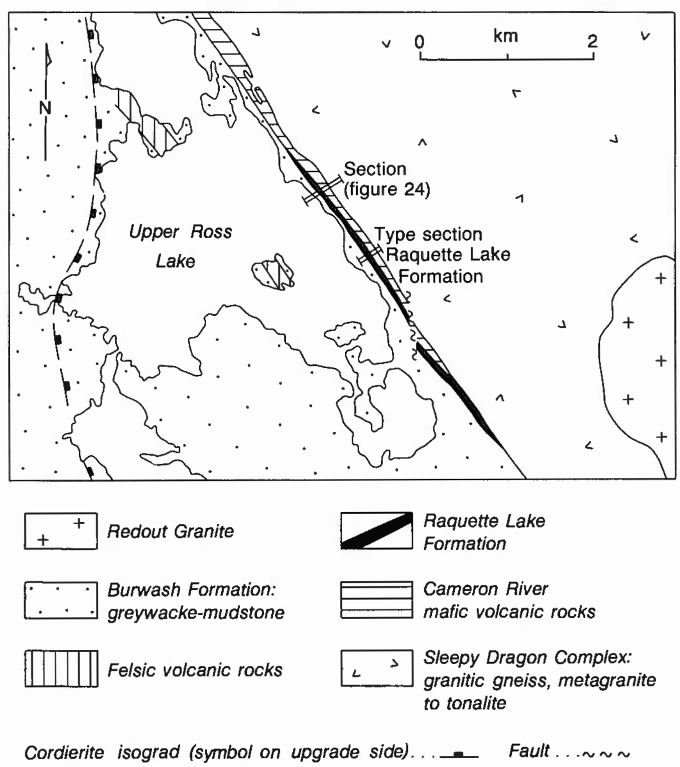


Figure 23. *Upper Ross Lake area showing the location of a section from the basement through distal Cameron River metavolcanics to the Burwash metasediments (Fig. 24), the distribution of the Raquette Lake Formation and the location of its type section.*

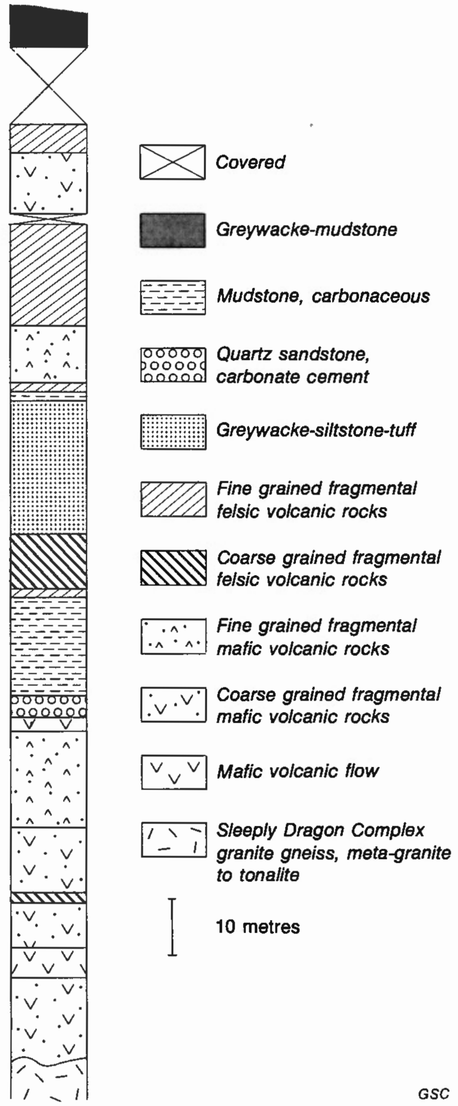


Figure 24. *Section through the distal part of the Cameron River mafic volcanic rocks at Upper Ross Lake (Fig. 23) showing the transition from volcanism to sedimentation. The section is dominated by volcaniclastic rocks in contrast to the mainly pillowed and massive mafic flows of the main part of the formation to the north and northeast. The base of the section may represent the unconformity between the Yellowknife supracrustal rocks and the basement rocks of the Sleepy Dragon Complex. The lower part of the section is dominantly mafic volcanic fragmental rocks while the upper part is mixed felsic and mafic fragmentals and sediments with minor carbonate. A 4 m thick unit of coarse sandstone containing quartz and granitic rock fragments evidently derived from the granitic basement occurs about 60 m above the unconformity. This quartz sandstone is the northernmost extension of the Raquette Lake Formation.*

The volcanic sequence thins and disappears from a normal thickness of about 3000 m over a distance of 13 km. The internal nature of the volcanic sequence changes also from dominantly pillowed flows in the main part of the unit to increasingly higher proportions of volcaniclastic material that itself becomes increasingly finer grained away from the main sequence. A section through a distal part of the sequence at Upper Ross Lake (Fig. 23) is shown in Figure 24. Here, half the sequence is mafic volcanic fragmental rocks whereas only 5% are flow rocks. Fragmental felsic volcanics make up 20% and carbonaceous mudstone 10% of the section. A 4 m thick unit of dominantly coarse grained, locally carbonate-cemented Raquette Lake Formation sandstone in 15 to 30 cm thick beds with tuffaceous interbeds, is interbedded with the volcaniclastic sediments of the Cameron River Basalt at this locality. These sandstones are clearly very different from the typical Burwash greywackes that occur above the section. Some of the clasts are small granitoid pebbles presumably from the nearby exposed Sleepy Dragon Complex basement. Three kilometres to the south the volcanics are reduced to a 20 to 30 m thickness and are overlain by Raquette Lake Formation volcanic conglomerate and quartz-rich sandstones of granitic derivation. At the south end of Ross Lake the volcanics are no longer present and the basement granitoid rocks (here rather brecciated) are overlain by sediments of the Raquette Lake Formation including basal, well-rounded, mafic volcanic boulder and pebble conglomerate followed by poorly sorted, crudely bedded, angular to well-rounded conglomerate composed mainly of granitic clasts, but with minor amounts of mafics, some chert and vein quartz.

Of note here is the lack of greywacke-mudstone turbidites within the volcanic sequence or interfingered with them. Sediments present in the volcanic sequence are not greywacke but units of black thinly laminated carbonaceous mudstone or conglomerate and sandstone derived directly from the volcanic pile or the underlying basement. Thus at the location of the growing volcanic pile the volcanics accumulated before or at least, as the pile grew, continued to remain topographically higher than the adjacent accumulating basinal sediments. There was evidently no time break between volcanism and sedimentation as there is no structural discordance, and in fact, locally mafic to intermediate tuffaceous sediments occur in the greywackes just above the contact. This relationship seems to be typical of many of the mafic volcanic sequences in the area, the only major exception being southwest of Clan Lake where there is significant interlayering of mafic volcanics and greywacke-mudstone turbidites.

Mafic volcanic rocks – Doubling Lake

On the eastern margin of the area, north of Hearne Channel, is a small unit of mafic metavolcanic rocks consisting of both pillowed flows and mafic fragmental units, although for the most part the volcanics are strongly deformed and primary structures are poorly preserved. The main part of the unit occurs outside the area where the northern part of the unit extends at least 5 km to the east, and a possible correlative volcanic unit another 5 or 6 km into the Christie Bay map area (see Brown, 1950; Fig. 25). The volcanics also appear to thicken to the east. The volcanics are in conformable contact with the Burwash greywacke-mudstones. They almost completely mantle a Defeat Plutonic Suite granodiorite pluton although one part of the unit is breached by a smaller subsidiary granodiorite pluton. The unit is offset by northerly trending dominantly left lateral faults.

The mantling of massive Defeat granodiorite plutons by mafic volcanics is a common feature in the area (Fig 25). Such relationships occur with the Duck Formation east of Yellowknife Bay, at Doubling Lake, and east and south of Tumpline Lake in the east-central part of the Hearne Lake area. In all cases the granodiorite is the same massive rock type. Also of interest is the fact that where this relationship occurs it is close to other major volcanic accumulations apparently near the postulated margin of the basin. In the central parts of the basin, where similar granodiorite plutons also occur, no volcanics are present. This would support the

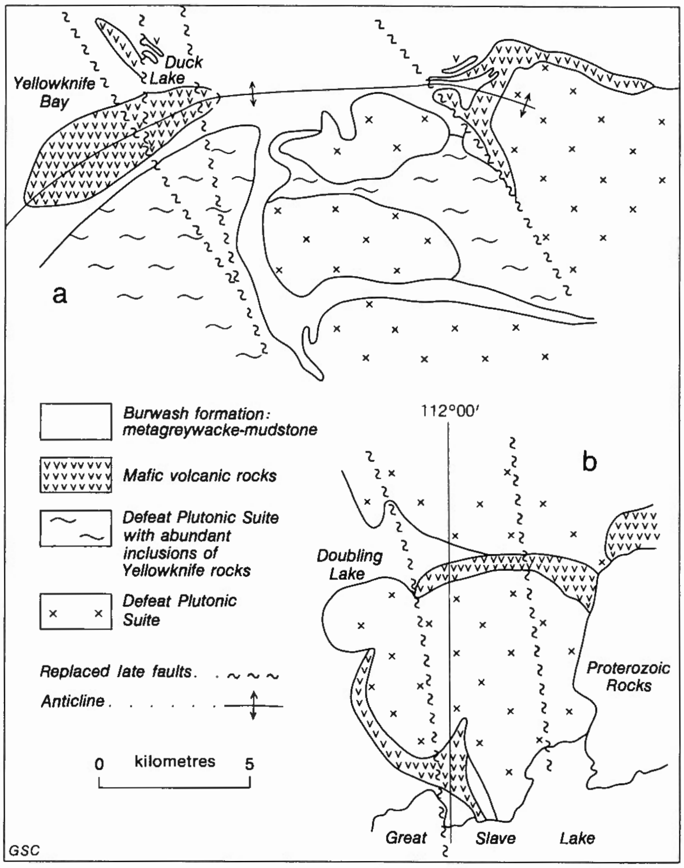


Figure 25. *In many cases mafic volcanic rocks mantle or are at least adjacent to plutons of the Defeat Plutonic Suite that are close to the margins of the Archean supracrustal basin. This is true of the Kam Formation at Yellowknife and east and south of Tumpline Lake in the east-central part of the area. In (a) the late faults in the Duck Lake area east of Yellowknife Bay have been replaced to show the close relationship between the intrusion and the volcanics. Similarly, in (b) at Doubling Lake in the southeast corner of the Hearne Lake area, the Defeat pluton which extends beyond the map area to the east is mantled by the mafic volcanics. (Geology east of longitude 112°W from Brown, 1950).*

suggestion that the volcanics did not completely underlie the basin as a whole but were restricted to certain localities along the basin margin.

Sunset Lake Basalt, Alice Formation, Payne Lake Formation – Sunset Lake

The largest single area underlain by volcanics is that about Sunset Lake in the northeastern corner of the map area. The rocks consist dominantly of mafic volcanic pillowed and massive flows in the northern or main part of the complex, but the volcanics of the southeastern part are intermediate in composition and consist dominantly of fragmental deposits. This volcanic belt was mapped at 1:50 000 scale by Lambert (1974, 1977, in press) and the description here is based for the most part on his observations. For more detailed descriptions the reader is referred to Lambert (in press).

The main part of the complex forms a northerly trending structure up to 15 km wide cored by the Amacher Granite lying between the north border of the map area almost to the east-west segment of the Beaulieu River, a distance of 30 km. A limb up to 6 km wide extends 30 km from the main part of the complex towards the southeast. To the southwest these volcanics are correlative with the volcanics in the vicinity of Turnback Lake, being separated from them by an intrusive pluton. North of the map area the dominantly mafic volcanic complex continues with a northerly trend over a distance of about 120 km (McGlynn, 1977; Moore et al., 1951) and so represents one of the greatest laterally persistent volcanic belts in the Slave Province.

Contact relations and structure

The relationship of these volcanics to the surrounding rocks is not everywhere clear due to the somewhat greater degree of structural deformation and, particularly for the southeastern arm of the complex, the high grade of metamorphism that some of these rocks have undergone compared to many other volcanic sequences in the area. On the other hand, in the central part of the complex at and north of Sunset Lake the metamorphic grade is greenschist. To the south the volcanics are in conformable contact with amphibolite grade greywacke-mudstone turbidites of the Burwash Formation. The contact is very sharp in the vicinity of the Beaulieu River but along the southeast arm of the complex, where the volcanics are mainly of fragmental origin, there is more of a gradational contact between the volcaniclastic sediments and the volcanogenic Burwash greywackes. Top direction indicators are rarely preserved, but where present indicate that the volcanics underlie the sediments, as is normally the case.

West of the complex, the volcanics are in contact with the highly deformed gneissic to cataclastic to mylonitic granitoid rocks of the Sleepy Dragon Complex that locally contain blocks or inclusions of amphibolite. The foliation in this deformed zone, which has a strong influence on the topography and is clearly visible in air photographs, is parallel to the contact with the volcanics and is up to 4 km wide. The volcanics east of this contact, although less severely deformed, are sheared up to a kilometre from the contact. The fragmental nature of some units in this region is preserved and a major chert-magnetite iron formation that occurs at the contact west of Amacher Lake is not severely deformed but locally contains only small scale low amplitude folds, and in places is broken rather than strongly sheared. The contrast in degree of deformation is striking. Within the volcanics no dykes or sills of granitic rock, that would suggest an intrusive relationship on the part of the Sleepy Dragon rocks, were observed. The significance of the blocks of amphibolite in the cataclastic gneisses is not clear. If the gneisses are older the amphibolites may represent the deformed remnants of a feeder system in a granitoid basement. On the other hand, if the contact is tectonic the gneisses may represent granitic rocks with inclusions of the volcanics emplaced at a relatively low level that have subsequently been faulted against the less deformed volcanic rocks. The contact is interpreted as tectonic and in light of the contrast in structural style between the highly deformed granitoid rocks and the volcanics, it does not seem unreasonable to suggest the granitoid rocks to the east may be older. Although not examined in the same detail, the geological situation appears similar to that on the west side of the Sleepy Dragon Complex where strongly deformed granitoid rocks are unconformably overlain by the Cameron River basalts.

The eastern contact of the volcanic complex contrasts strongly with the western contact. Here, the massive, medium grained, pink Meander Lake Plutonic Suite intrudes the volcanic sequence. The contact as mapped is gradational, consisting of a zone of varied width of mixed rocks ranging from dominantly mafic rocks with dykes and sills of granite through to mainly granite with scattered inclusions of mafic rocks. West and south of Payne Lake the contact is coincident with a cataclastic to mylonitic zone up to 1.5 km wide that strikes north 35 degrees west through the granitic rocks.

In the centre of the main part of the complex is the Amacher Granite, a massive, medium coarse grained granite that is distinct from other granitic units in the area. The granite is also metamorphosed. The contact with the volcanics that surround it is everywhere sharp. Except for blocks of mafic volcanics within the granite at the northeast end, no inclusions of the volcanics were seen within it. Locally a few small veins of granitic material intrude the adjacent volcanics. J.F. Henderson (1939) reported the occurrence of a two foot thick "peculiar fragmental phase of the granite" at its northeast contact with the volcanics. Although the granitic clasts are angular to subrounded and have the appearance of a conglomerate, he also considered the possibility that this unit may represent a brecciated zone along the contact. The nature and origin of this granite is discussed more fully in the section devoted to it, but briefly it is considered to be an epizonal subvolcanic pluton possibly related to the felsic volcanic phases in the volcanic sequence.

The structure of the volcanic rocks appears to be complex, with locally intense deformation resulting in rather poor preservation of top direction indicators. The structural framework proposed here, then, is hypothetical due to the lack of sufficient data. At its simplest, the structural style is that of a southward opening wedge of volcanics that has been compressed in an east-west direction. Such data that are available indicate that the volcanics face away from the granitic units. Thus there is a northward trending anticlinal axis through the Amacher Granite core with a synclinal axis to the east and west. The anticlinal structure and westerly syncline die out north of the Amacher Granite. The syncline through Sunset Lake appears to be the dominant structure and probably can be traced north of the map area. The axis is marked by an extensive sheared zone and a chain of elongate lakes.

The southeast arm of the complex is unusual in that the outcrop belt between Payne Lake and the fault through Meander Lake is about twice as wide as the same belt to the north or south. Due to lack of structural data it is difficult to resolve this anomaly with any confidence. One possibility is that there is an unrecognized northerly splay off the fault through Meander Lake. The required sense of displacement (left lateral) would be the same as for the other numerous strike-slip faults in the region, but the amount of displacement required would be on the order of 12 km and would be reflected in major displacements elsewhere along the fault. There is no evidence of any displacement of this magnitude; indeed, such faults typically have displacements of less than 5 km. Lambert (in press) has suggested that there is a northeast-trending fault with a right lateral horizontal component along the north margin of the wider part of the belt, and that the extra thickness of the belt is due to a different structural level being exposed between the faults. The preferred explanation in this report is that the southeast arm of the volcanic complex is folded about more or less horizontal north-trending axes slightly oblique to the belt.

Most of the volcanics have been metamorphosed to amphibolite grade with the mafic flows now being made up mainly of assemblages of plagioclase and hornblende while the fragmental rocks to the southeast locally contain garnet, biotite and cordierite as well as amphibole and plagioclase. However in the central part of the complex about Sunset Lake the volcanic rocks are at greenschist grade.

The volcanics of the Sunset Lake – Payne Lake area have been divided into three formations and an unnamed unit of rhyolite by Lambert (in press).

Sunset Lake Basalt

The Sunset Lake Basalt is volumetrically the most important, underlying most of the main part of the terrane except for the axis of the syncline through Sunset Lake. Due to structural complexities the stratigraphic thickness of the formation is unknown. It is the oldest formation in the area being overlain gradationally in the central part of the area by the Alice Formation andesites and dacites and is laterally gradational with the Payne Lake Formation volcaniclastic units to the southeast. It is also locally conformably overlain by the Burwash Formation greywacke-mudstone turbidites. As with the other major mafic volcanic sequences the Sunset Lake consists mainly of flattened mafic basaltic pillowed lavas, pillow breccias and associated minor hyaloclastic sediments. Massive flows and volcaniclastic sediments are minor. Oxide iron formation occurs within the formation near Amacher Lake. Much of the formation is at amphibolite metamorphic grade which, along with the accompanying deformation, has resulted in the obliteration of many primary structures and textures. In the south-central part of the formation where the grade is lower and deformation less, primary features are better preserved.

Along the eastern margin of the northern part of the formation is a major series of amphibolite sills that form a belt up to 2 km wide and over 20 km in length extending north of the map area. The sills are massive and coarser grained than the mafic flows with which they are associated. Like the flows they have been metamorphosed and are thought to be more or less contemporaneous with volcanism. The contact of this zone with the flows to the west is gradational. The proportion and thickness of the sills increases steadily to the east until the volcanics occur only as inclusions within the series of thick homogeneous sills. The contact of the amphibolite sills with the granitic rocks to the east is gradational over a kilometre or more, with granitic dykes and sills in the amphibolite becoming increasingly abundant until the mafic sills are represented only by minor blocks and inclusions of the amphibolite. In addition to the main series, similar mafic intrusions occur elsewhere in the Sunset Lake Basalt, the larger of which are mapped separately. These metamorphosed mafic intrusions are similar in nature and origin to the mafic intrusive bodies that are common in the other mafic volcanic sequences in the region.

Along the western margin of the complex west and southwest of Amacher Lake is a major chert-magnetite iron formation that causes a prominent magnetic anomaly extending over a strike length of 12 km on the aeromagnetic map of the region (Geological Survey of Canada, 1969a). Where seen, the iron formation consists of one to three units, 1-15 m wide (Lambert, in press) that consist of about 80% thin layered recrystallized chert-magnetite iron formation interbedded with lappilli to fine grained tuffs of mafic composition. The oxide iron formation layers range in thickness from a few millimetres to 3 cm and average about 1 cm. Magnetite makes up between 30 and 60% of the iron formation, the remainder being recrystallized chert and minor iron-rich silicates. The laminae are complexly folded on a small scale and at least part of the folding may be due to soft sediment deformation.

Alice Formation

In the central part of the belt the Alice Formation conformably overlies, and to some extent is interfingered with, the Sunset Lake Basalt. The formation consists of a lower andesite member and an upper dacite member estimated to be 800 and 400 m thick respectively (Lambert, in press). The top of the formation is not exposed but to the south is conformably overlain by Burwash Formation greywackes. The andesite member consists of pillowed lavas, breccias, and massive flows with minor pyroclastic and volcaniclastic sediments. The lavas are locally amygdaloidal in contrast to the underlying Sunset Lake basalts. As much of the formation is at only greenschist metamorphic grade, primary textures are more commonly preserved and include a variety of phenocryst types with euhedral to resorbed outlines, oriented microlites suggesting flow textures and, in the clastic rocks, some relict shards and scoriaceous blocks. The rocks themselves weather pale green to buff in contrast with the darker green Sunset Lake basalts. The dacites gradationally overlie the andesites and tend to be lighter grey, have a lower specific gravity, and consist more of tuffs, breccias and massive flows that are commonly autobrecciated in contrast to the dominantly pillowed flows of the underlying andesite member. The dacites are quite heterogeneous with individual units discontinuous. The member occurs mainly in the axis of the syncline through Sunset Lake along which there has been considerable cataclasis to shearing with the resultant loss of primary features.

Lambert (in press) has suggested that there is a transition from iron-rich mafic volcanism to more alkali rich volcanism through the volcanic sequence that in general corresponds to the change from the Sunset Lake basalts to the Alice Formation andesites and dacites. In contrast to the similar but more abrupt chemical change between the Cameron River basalts and the Webb Lake Andesite to the west, the transition in this area is gradational. Lambert (in press) suggested the more alkali-rich volcanics represent the differentiated products of the original iron-rich basaltic magma with the gradational change being due to continuous tapping of the source as it began to differentiate.

Rhyolite

Three dominantly rhyolitic units have been mapped in the volcanic complex that in all cases occur at or near the top of the sequence at any given locality (Lambert, in press). South of the Amacher Granite at the contact between the Sunset Lake Basalt and the Burwash Formation is a unit up to 1100 m thick consisting mainly of massive to layered, white, quartz phenocryst-rich rhyolite with associated breccia units. The top of the unit (towards the south) is capped with fragmental volcanic sediments that in one place consist of 150 m of thin-layered volcaniclastic sediments interbedded with 2 to 3 m thick, dominantly carbonate layers that in turn are overlain by 40 m of carbonate with scattered lenses of volcaniclastic sediments. At the south end of Sunset Lake is another felsic centre, again up to 1100 m thick, that consists of dacitic to andesitic volcanics at the base that grade up to a white rhyolite to rhyolite breccia centre with fans of volcaniclastic debris on its flanks. At the north end of Sunset Lake are three felsic units separated by Alice andesites and dacites. As with the other centres these units consist typically of white weathering rhyolite and rhyolite breccia. These felsic units are interpreted as local felsic domes that developed late in the evolution of the dominantly mafic to intermediate volcanic complex (Lambert, in press). Lambert has suggested these domes were emergent for at least part of the time during their formation. Although generally smaller, they are thought to be similar to the felsic domes in the Tumpline – Turnback lakes area to the southwest.

Black, carbonaceous, locally pyrite-rich mudstone is prevalent along the southern margin of the main part of the complex and its equivalents to the southwest and west at the transition between the volcanics and the sediments. For example, above the carbonate unit at the top of the felsic dome south of the Amacher Granite is a 20 m thick unit of black, cherty mudstone with locally abundant pyrite. Elsewhere in the Slave Province in similar situations at the contact between intermediate to felsic volcanics and the greywacke mudstone turbidites, this carbonaceous mudstone facies is commonly associated with carbonate facies iron formation (siderite). Examples include the central part of the east side of the Back volcanic complex on the Back River in the eastern Slave Province and at Snofield Lake in the High Lake area in the northern Slave Province (Henderson, 1975a). No carbonate iron formation has been reported in this region although the pyrite-rich carbonaceous mudstones can be thought of as lean iron formation.

Payne Lake Formation

The southeast arm of the complex is quite different from the main part of the complex as well as the other volcanic belts in the map area. South of Payne Lake the volcanics are dominantly a heterogeneous assemblage of volcaniclastic sediments as opposed to the pillowed flows of the main part of the complex. Lambert (in press) has named this unit of metavolcaniclastic rocks the Payne Lake Formation. The formation interfingers with, to conformably overlies, the Sunset Lake Basalt to the north and is conformably overlain by the Burwash greywacke-mudstone turbidites to the southwest. The granites to the northeast are intrusive, with, in some cases, sills of weakly foliated to massive granite emplaced in the more severely deformed volcanics within a kilometre wide zone. The sequence consists of compositionally layered units 30 cm or more in thickness ranging down to 1-5 cm. In general, the beds become thinner to the southeast. The layering varies from sharp to diffusely bounded. The rocks range from very dark green through brown, grey to light pink, reflecting the variations in composition. All primary features except compositional layering have been lost due to extensive metamorphic recrystallization, although in a few cases coarse breccia clasts are preserved as elongate lensoid structures. Commonly, individual volcanogenic layers are separated by thin biotite-rich layers or laminae, possibly indicating argillaceous sedimentation between episodes of volcanogenic deposition. These metavolcanics are fine- to medium-grained, typically with a strong foliation. Metamorphic textures predominate. Some layers contain metacrysts of poikiloblastic garnet up to 1 cm in size and in some layers altered cordierite is present.

Lambert (in press) has divided the unit into three members. About 70% of the formation is made up of mafic volcaniclastic rocks. These schists contain up to 45% hornblende and normally less than 5% biotite and are thought to be detritus derived in large part from the mafic volcanics to the north. Felsic biotite-bearing volcaniclastic rocks are the next most abundant and occur in three or four main, possibly correlative, bodies in the northern part of the formation. In some cases relict breccia clasts are preserved. These rocks are thought to be derived from a mixed felsic to mafic volcanic source. White, biotite-poor, rhyolite members occur in which both massive and breccia facies, and layered units that may represent volcanic arenite units, are locally recognizable. Lambert (in press) suggested these may represent rhyolite centres within the dominantly volcaniclastic unit.

In summary, the extensive volcanic complex in the vicinity of Sunset Lake represents mainly mafic subaqueous volcanics that evolved into more intermediate to felsic facies with time and that were extruded along a long linear northerly trending fracture system that extended over 150 km. Although the southern end of the belt is truncated by granitic intrusions, the original belt probably did not extend much farther to the southeast as this part of the belt is made up mainly of relatively thin layered fragmental deposits, probably representing a distal facies of the volcanic complex. The granitoid rocks to the west of the main part of complex may represent the sialic basement upon which the volcanics were extruded. The metamorphosed epizonal granitic body in the core of the complex may be a synvolcanic intrusion comagmatic with the felsic volcanic centres. To the east the granitic rocks are clearly intrusive. The presence of large areas of abundant supracrustal blocks of both sedimentary and volcanic origin as inclusions in the Meander Lake granites would suggest the basin extended east of the volcanic complex as presently mapped.

Felsic volcanic rocks

Felsic volcanic rocks – Russell Lake

A major unit of dominantly felsic volcanic rocks occurs east of Russell Lake between the east side of the lake and the north border of the map area. The volcanic rocks have a northeasterly trend and outcrop over an area about 5 km wide and 16 km long. The unit is offset by the Stagg River fault. The volcanic terrane has more topographic relief than the adjacent terrane underlain by metasedimentary or granitic rocks. A prominent ridge within the unit is due to more erosion-resistant felsic volcanic rocks.

The volcanics occur in a steeply dipping homocline that, on the basis of rare but consistent top indicators, faces to the northwest. The top of the unit is in sharp contact with typical amphibolite facies Yellowknife metagreywacke-mudstone turbidites. The presence of a prominent lineament parallel to this contact, the highly sheared nature of the rocks in this vicinity, and the abundance of quartz-filled fractures in the volcanics, suggests the contact may be faulted. The basal part of the unit to the southeast is intruded by granitic rocks in the west. To the northeast the volcanics gradationally overlie metasedimentary rocks that become increasingly less volcanogenic and more argillaceous away from the contact, before they too are intruded by the granitic rocks.

The southeastern two-thirds of the sequence is intruded by muscovite and tourmaline-bearing granitic to pegmatitic dykes and veins. These intrusions have a maximum width of about 12 m, are irregularly distributed, but increase in abundance to the southeast. The contact between the volcanics and the sediments to the northeast is a fault, as it is to the southwest, although the relations are obscured by Russell Lake.

Within the volcanic unit a crude stratigraphy is recognized. The volcanics near the base of the unit are intermediate to mafic with pillowed flows occurring northeast of the locality where the granite contact transects the sedimentary volcanic contact. More mafic material occurs as inclusions of amphibolite within the granitic rocks to the southeast. A ridge of more felsic than normal volcanics characterized by its more massive structure, white weathering, and greater abundance of quartz phenocrysts, occurs in the lower central parts of the unit. Despite the width of the unit, its known strike length is only about 15 km, which contrasts strongly with the mafic volcanic belts which are laterally much more extensive.

The felsic volcanic rocks weather pale green to pale brown and are grey on the fresh surface. More felsic varieties (more quartz-rich) tend to weather lighter. The rocks for the most part are sheared, with the result that primary textures are commonly lost. Textures in the coarser material are better preserved, although the clasts tend to be elongated, particularly in the vertical direction.

The volcanics are mainly of fragmental origin. They occur in fine grained, thin-bedded units that in places consist of alternating lighter quartzofeldspathic and darker amphibolitic material on a 2 to 3 cm scale. Grain size gradation is present in a few thicker beds and consistently indicates the unit faces northwest. Massive featureless beds up to 50 cm thick are also present. Volcanic conglomerate or breccia units several tens of metres in thickness are common locally (Fig. 26). The clasts are up to 20 cm in maximum dimension, although they are typically deformed. Within any breccia unit the composition of all the clasts are very similar. In the lower part of the section pillowed flows of intermediate composition occur.

Carbonate occurs locally as matrix to some of the breccias and also as thin laminated beds up to 20 cm thick that are typically highly sheared. In one locality a carbonate bed occurs between a volcanic breccia unit and overlying massive fine grained fragmental rocks (Fig. 27).

Figure 26. *Deformed felsic volcanic breccia east of Russell Lake. GSC 158294*

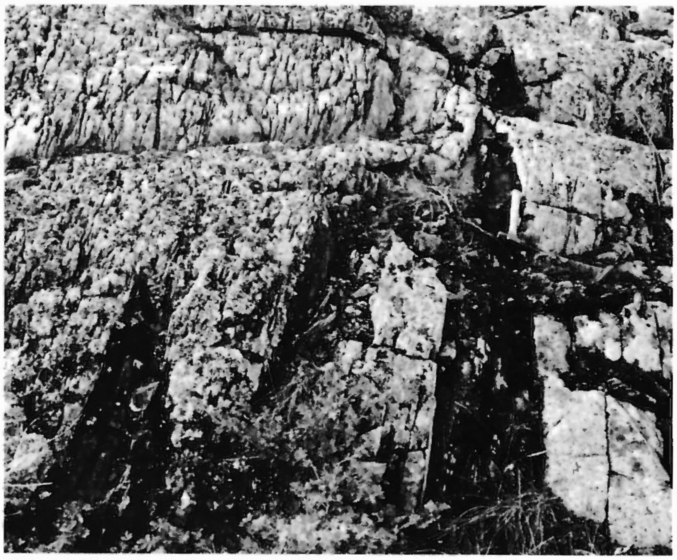


Figure 27. *Carbonate unit (dark, recessive) between overlying thick, massive, fine grained felsic volcaniclastic rocks (to right) and underlying felsic volcanic breccia (to left) east of Russell Lake. GSC 158287*

Rare calcareous concretions occur in some beds as irregular epidote-rich bodies up to 30 cm in size. At only one locality within the main part of the volcanic unit were argillaceous sediments observed.

Interbedding of volcanic units with turbidites and the presence of pillowed flows in the lower part of the section would suggest the volcanics were deposited subaqueously. Continuity of thin bedded units throughout the rest of the section also suggests subaqueous deposition although a subaerial environment cannot be excluded. For the most part, the rocks are too deformed to allow recognition of particular depositional environments.

Felsic volcanic rocks – Clan Lake

East of Clan Lake on the Yellowknife River is a major irregularly shaped body of dominantly felsic volcanic rocks 5 km wide and 8 km long. A thin unit of mafic volcanic rocks occurs along the southeast side of the felsic body and can be traced about 10 km north of the map area (Jolliffe, 1939). The felsic and mafic volcanics are surrounded by metagreywacke-mudstone turbidites.

The major part of the felsic volcanic unit consists of light grey, brownish-orange, and greenish weathering, grey dacitic fragmental volcanic rocks. For the most part the unit is homogeneous and massive, although locally layers 5 to 20 cm or more thick are defined by darker more argillaceous material. Most of the volcanics have a fragmental texture with a grain size on the order of 2 to 5 mm. Coarser breccia units with clasts up to 20 or 30 cm occur locally. Rock fragments are the most abundant component. Feldspar crystals, and to a much lesser extent, quartz crystals, are present also. Carbonate is rare but where present occurs as matrix to the fragmental rocks.

In thin section the primary texture of the volcanics is only poorly preserved. Euhedral to subhedral phenocrysts of plagioclase are typically strongly altered, such that in the extinct position the outline of the crystal fades into the matrix. The phenocrysts range in size from a maximum of 2 or 3 mm down to a tenth of a millimetre, commonly within the same section. Quartz, where present, occurs as smaller, rounded to subhedral grains with the embayments characteristic of volcanic quartz. In many cases, however, the quartz has been deformed into polygonal aggregates of irregular shape. The matrix consists dominantly of a very fine grained quartzofeldspathic material that coarsens and becomes more distinct with increasing metamorphic grade. Mafic minerals, including chlorite, biotite and hornblende, occur as scattered individual grains, local aggregates to discontinuous diffuse laminae. No relict textures representing shards or lithic fragments are apparent in the matrix but are more apparent megascopically on weathered surfaces.

Along the southwest margin of the felsic body is a thin tightly folded mafic volcanic unit that is wrapped around the southernmost lobe of the felsic volcanics and extends north beyond the map area. It consists of both mafic fragmental and flow units. Pillowed flows are present locally. As well, mafic to intermediate volcanic units occur within the felsic volcanics up to a kilometre from the mapped contact between the felsic and mafic volcanics.

The contact between the felsic volcanics and the surrounding greywackes is gradational over a narrow stratigraphic interval. As the contact with the volcanics is approached, the greywacke beds become increasingly rich in volcanogenic components, thinner, have a more siliceous aspect and change from grey to yellowish green. In the embayment on the southwest side of the volcanic body several black carbonaceous mudstone units are present, commonly with abundant disseminated pyrrhotite.

The cordierite isograd, as defined by metamorphic mineral assemblages in the surrounding pelitic metasediments, passes through the felsic metavolcanic body. The metamorphic gradient is much more obvious in the more mafic units as the mafic flows become amphibolitic at higher grades, whereas little megascopic change is apparent in the quartzofeldspathic units.

The structural position of the volcanics with respect to the adjacent metavolcanic rocks is not clear. The volcanics are clearly separated from the homoclinal sequence of dominantly mafic volcanic flows that lie to the northwest across a fault of unknown displacement. They are surrounded by isoclinally folded greywackes and mudstones that are conformable with the volcanics. The felsic volcanics, on the other hand, outcrop in a more or less equidimensional lobate shape, although the mafic volcanics on the southeast side are folded in a style more similar to that of the sediments. Because layering is obscure within the felsic unit, it is difficult to determine the structure within it. Because of mechanical differences the much more competent felsic volcanic mass may have acted as a resistant boss about which the much less competent sediments were tightly folded. If such is the case, the volcanics may only be minimally deformed with relatively open folds and gentle dips, in contrast to the tight folding in the adjacent sediments and mafic volcanics. The lobate outcrop pattern suggests interference folding. The structural situation of the Clan Lake complex may be quite similar to the much larger Back River volcanic complex in the east-central Slave Province (Lambert, 1976, 1978; Henderson and Thompson, 1980).

The complex has been recently mapped in detail and briefly described by Hurdle (1983).

Banting Formation – Walsh Lake

North of Yellowknife and west of Prosperous Lake is a major unit of dominantly felsic volcanic rocks, the Banting Formation (Henderson, 1970), that extends about 30 km along strike. The formation was originally mapped by Jolliffe (1946). The central and southern parts of the formation have been remapped by Easton and Jackson (1981) and the central and northern parts by Helmstaedt et al. (1980).

Helmstaedt et al. (1980) have proposed that the sequence from the Kam Formation on the west through the Jackson Lake Formation, Banting Formation, and Walsh Formation (a formation similar in many respects to the Burwash Formation but separated from it by the Banting Formation (Henderson, 1970)) to the Burwash Formation on the east is a homoclinal succession with the Banting volcanics considered to occur as two members separated by the Walsh Formation turbidites. In this report, following the original somewhat equivocal suggestion of Jolliffe (1946), the Banting is considered to occur on two limbs of a major synclinal structure through Walsh Lake. This structure would be the faulted continuation of the major syncline through Yellowknife Bay. The position of the axial trace through Walsh Lake is about where it would be expected, considering the apparent sense and amount of displacement on the fault through Jackson, Walsh and Duck lakes as indicated by the displacement of mafic dykes and sills of both Proterozoic and Archean age. Top indicators within the Walsh Lake zone are equivocal due to intense deformation, as might be expected in the axial zone of a structure made up of units with high contrast in competancy.

On the western limb of the structure the formation has a maximum thickness of 1300 m, whereas on the eastern limb it has a maximum thickness of 800 m. The thickness is much more varied on the eastern limb and indeed the formation is not present at all between the south end of Walsh Lake and

Yellowknife Bay. In Yellowknife Bay, Latham Island, the island to the northeast, the east half of Jolliffe Island and most of Mosher Island are all considered to be underlain by Banting Formation.

The Banting Formation is conformably overlain by the mudstone and thin greywacke-mudstone turbidites of the Walsh Formation (in this report undifferentiated from the Burwash Formation with which it is lithologically very similar). The east limb is underlain by the greywacke-mudstone turbidites of the Burwash Formation. The contact between the Banting and the Jackson Lake Formation on the west limb is very poorly exposed but Baragar (1975), after examining Giant Yellowknife Mines Ltd. drill core that intersected the contact 2 km north of Latham Island, found that the Banting volcanics are conformable and interfinger with the underlying Jackson Lake sandstones. Prior to this the relationship of the Jackson Lake (which unconformably overlies the Kam mafic volcanics) to the Yellowknife Supergroup in general or the Banting volcanics in particular was in considerable doubt (Henderson and Brown, 1952b, 1966; Boyle, 1961; Henderson, 1970) or misinterpreted (Henderson, 1973). Subsequently Helmstaedt et al. (1980) have found an exposure of the contact between the two formations 1 km north of Banting Lake in which the gradational nature is evident.

The Banting Formation is mainly of volcanic origin, consisting of felsic lithic and crystal tuff, porphyritic flows, breccias and locally pillowed andesite. The formation is sheared in the northern part of its outcrop area and primary textures and structures are best preserved in the vicinity of Walsh Lake and to the south. For example, dark fiamme possibly representing welded pumice fragments in an ignimbritic felsic flow occur on the shore of Yellowknife Bay, 2 km north of Latham Island (Fig. 28; see Padgham (1980) who considers this flow part of the Kam Formation). The volcanics in the southern part of the formation are mainly felsic volcanic fragmental deposits although units of more intermediate composition are present locally, some with pillow structures.

As described by Helmstaedt et al. (1980) and Easton and Jackson (1981) the generalized stratigraphy of the western formation west and south of Walsh Lake consists of a lower medium- to coarse-grained quartz-feldspar porphyry unit that when seen over several hundred metres in drill core (Baragar, 1975) can be seen to consist of layers of massive porphyry with thin intercalations of breccia. The porphyry is overlain by, or locally intrudes, conglomerates, tuffaceous sediments and a sequence of intermediate to mafic flows, pillow lavas and breccias. The main part of the unit follows and consists of felsic fragmental volcanics with mafic and felsic bedded tuffs to ash flow tuffs that grade into the greywacke-mudstone turbidites of the overlying formation. The fragmental units comprise thick breccias with angular clasts up to 60 cm in maximum dimension, tuffs in massive units several metres thick with grading, if present, only in the top few centimetres, and thinly layered (centimetre scale) cherty tuffs. Beds to thin members of conglomerate, similar to those of the Jackson Lake Formation, occur south of Banting Lake with one near the top of the formation up to 70 m thick containing mainly lithic sandstone cobbles with minor felsic volcanic clasts.

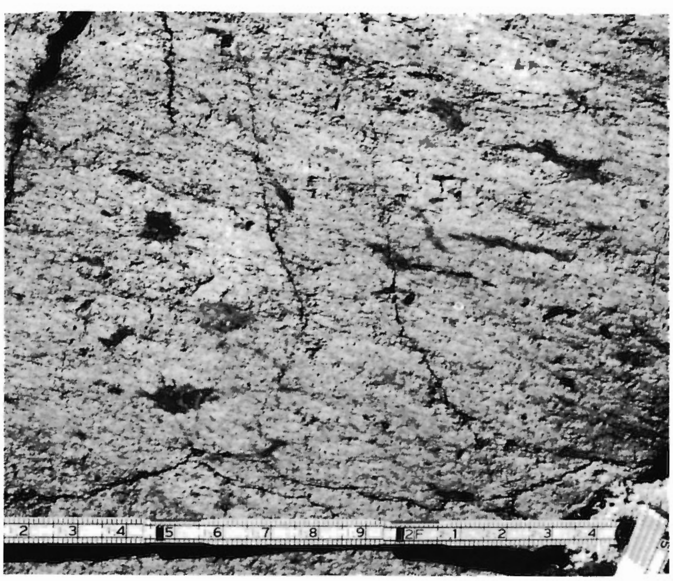


Figure 28. Massive welded tuff unit in the Banting Formation on the west side of Yellowknife Bay 4 km north of Yellowknife. Note the dark fiamme, possibly representing flattened pumice fragments. Scale in tenths of feet. GSC 157440

The volcanics on the eastern limb consist mainly of bedded dacitic to rhyolitic volcaniclastic rocks with interbedded more mafic material. The general stratigraphy seen in the formation west of Walsh Lake is not seen on the eastern limb (Easton and Jackson, 1981). The volcanics are also more altered than those to the west. In the central part of Yellowknife Bay, Latham, Mosher and Jolliffe islands are underlain dominantly by fragmental tuffs and breccias. Pillowed andesite flows occur on the island northeast of Latham Island.

In thin section the Banting fragmental rocks are completely unsorted. The most obvious clasts are typically broken plagioclase crystals. Quartz phenocrysts are present but much less abundant, making up less than one per cent of the rock. The crystals are euhedral to subhedral with the embayments typical of volcanic quartz. The matrix consists of a very fine quartzofeldspathic intergrowth in well preserved examples, in which vague outlines of primary clasts can be seen. The outlines are expressed by variation in the crystallinity of the quartzofeldspathic intergrowths with the central part of the clasts coarser than the margins. These clasts and fine grained matrix represent devitrified glass so that original textures are very susceptible to destruction during subsequent deformation. The proportion of phenocrysts to "glass" clasts is highly variable from one flow to another.

The porphyry at the base of the sequence on the west side is coarse and highly porphyritic with quartz and feldspar phenocrysts up to 5 mm in size forming up to 70% of the rock. The plagioclase is typically euhedral, commonly zoned and not broken. The quartz is also commonly euhedral and deeply embayed. Biotite and euhedral hornblende are minor constituents. There is a gradation in grain size of the major components down to matrix size. The matrix, as in the tuffs, consists of very fine quartzofeldspathic intergrowths.

An average of 7 chemical analyses from the Banting Formation is presented in Table 5 (page 25).

Evidence for a felsic volcanic centre is present in the formation at the north end of Yellowknife Bay (Easton and Jackson, 1981). There, much of the formation consists of thick units of massive quartz-feldspar porphyry that may represent a felsic dome in the form of either shallow intrusive sills or thick massive viscous flows presumably derived from a nearby vent. To the north and south the volcanics are largely clastic, presumably derived in part from the centre. Clasts in the fragmental volcanics decrease in size to the north. Possible relicts of a subvolcanic plumbing system are found to the west in the immediate underlying mafic Kam volcanics in the form of a few dykes of dacite porphyry (see maps of Henderson and Brown, 1966) which are the only ones of this type in the immediate vicinity. To the north, east of Ryan Lake, there is a more extensive network of anastomosing crosscutting dacite porphyry dykes in the Kam Formation that may represent a more complete subvolcanic conduit system to another older felsic centre that has not been preserved as the dykes are truncated at the unconformity (Fig. 20). The nature and significance of these porphyry dykes are discussed further in the section on the Kam Formation.

Townsite flows – Yellowknife

Within the Kam Formation, the main mafic volcanic sequence at Yellowknife is a unit of dacitic volcanics termed the Townsite flows (Henderson and Brown, 1966 quoting Traill, 1950). They occur just north of the city of Yellowknife and their faulted equivalents occur west of the north end of Yellowknife Bay and at Vee Lake west of Walsh Lake. Helmstaedt et al. (1981) have suggested that the faulted equivalent of the Townsite flows may occur as a felsic fragmental volcanic unit at Kam Point, based on the similarity of the rocks, the association with thick mafic sills and the similarity of distance between the felsic volcanics in each case and the contact with the intrusive Defeat granodiorite. However since this correlation is not compatible with the correlation of three marker units and two diabase dykes by Henderson and Brown (1966) across the fault in question, this suggestion is not accepted.

The unit is about 400 m thick. Throughout its exposed length the flows are intruded by major gabbro sills. As described by Henderson and Brown (1966) the unit consists of light weathering, dark grey to greenish grey, typically porphyritic (feldspar and quartz) flows and fragmental deposits. The fragmental units occur in discontinuous lenses, one of which was measured as 700 m long and 30 m thick. Clasts of dacite in the breccia units are up to 70 cm in size. Tuffaceous beds are thinner; up to 1 m thick. The flows are massive and for the most part structureless, but locally features such as flow lines and amygdules can be seen. Pillow structures are present in some of the more mafic units. An average of 2 chemical analyses from the Townsite flows is presented in Table 5 (page 25).

As this unit is situated conformably within the dominantly pillowed Kam Formation that is presumably of subaqueous origin, and as the more mafic flows within the Townsite flows are also pillowed, this would indicate that the Townsite felsic volcanics also accumulated subaqueously.

Mixed mafic and felsic volcanic rocks

Tumpline Basalt, Turnback Rhyolite, Sharrie Rhyolite – Victory to Turnback lakes area

The volcanic complex in the vicinity of Detour Lake, Tumpline Lake and Turnback Lake in the east-central part of the map area differs from other volcanic complexes in that it is made up of approximately equal proportions of mafic and felsic volcanic rocks. In other volcanic complexes in the area both mafic and felsic material are present, but one typically greatly exceeds the other. This complex also differs from the others in that it is structurally more complex. The complex is similar to some of the volcanic belts in the more northern parts of the Slave Province, such as the Hackett River area (Padgham and Bryan, 1975; Padham and Ronayne, 1976; Frith and Hill, 1975) and the High Lake area (Padgham, 1974).

The volcanics and granitic rocks in the Victory-Turnback lakes area have been mapped at 1:50 000 scale by Lambert (1974, 1977, in press) and Davidson (1972). What follows is a brief outline of the nature of the volcanics in the area based in part on their observations. For more detailed descriptions of this volcanic complex the reader is referred to Lambert (in press).

The volcanic complex is situated south of the extensive granitoid terrane at least part of which (Sleepy Dragon Complex) is older than the supracrustal rocks. The central part of the complex is situated out in the basin, in that Yellowknife sediments occur between the granitic rocks (and probable basement rocks) to the north and the main part of the volcanic complex. The volcanics extend from southeast of Victory Lake to just northeast of Turnback Lake, a distance of about 40 km in a series of subhorizontal isoclinal folds.

Plutons from three distinct granitic units intrude the volcanic complex and adjacent sedimentary rocks. The presumed oldest are the massive Defeat biotite granodiorite plutons that are concordant in general, but locally crosscut the volcanic rocks. They occupy both the dome in the southeastern part of the complex and the large open structure to the east. In the north, the Redout Granite, which intrudes the Sleepy Dragon Complex, also concordantly to somewhat discordantly intrudes the volcanics in the northern part of the volcanic complex. The third and presumably youngest intrusion is the tongue of weakly foliated, highly discordant Prosperous biotite-muscovite granite that separates the northeasterly and northwesterly trending structures in the volcanic complex.

Throughout most of its extent the volcanic complex is conformably overlain by the greywacke-mudstone turbidites of the Burwash Formation. Wherever examined, the transition is typically abrupt, with fragmental volcanics, and volcanogenic sediments giving way to the argillaceous and quartz-bearing greywackes over a distance of a metre or so. Locally a unit of black carbonaceous mudstone, commonly with disseminated pyrite and pyrrhotite, occurs at this transition point. An example of this can be seen at the east end of Goose Lake. Similar sulphide-bearing carbonaceous mudstones occur at the transition between volcanics and normal sediments elsewhere in the area and at similar situations elsewhere in the province (Henderson, 1975a).

The central and presumably thickest part of the complex is situated just southwest of Tumpline Lake. Sediments in a synclinal structure separate this main part of the complex from the northern fringe that is in contact with the metagranitoid complex to the north. In the contact area the volcanics are best described as mafic schists intruded by pegmatites and granitic dykes. The granitoids, on the other hand, are extremely heterogeneous with granite gneiss phases, granitic bodies, pegmatite and a highly variable proportion of mafic inclusions (Davidson, field notes, 1971). The original nature of the contact between the volcanics and any basement granitoids that may have been present is thus obscured by intrusion, metamorphism and deformation. Fifteen kilometres to the northwest, however, near Upper Ross Lake there is evidence that the metagranitic rocks in this complex are basement to the supracrustal rocks. Some of the older phases in this part of the granitoid complex near Tumpline Lake may also represent basement although it would be difficult to prove.

The structure of the volcanic belts between Victory and Turnback lakes has been generalized on the map due to the difficulty of determining top directions in the commonly highly deformed rocks. The major folds are outlined but no doubt are actually complex anticlinoria and synclinoria, with smaller scale folds difficult to resolve at this scale.

In the northwest the volcanics occur in two anticlinal structures that plunge to the northwest. In the south-central part of the complex the volcanics occur in a dome concordantly cored by a pluton of Defeat granodiorite.

Several small satellite plugs of similar granodiorite occur within the volcanics immediately to the north and to the southeast. To the north the volcanics strike northeasterly in two anticlinoria. In this direction the volcanics are separated from the major mafic volcanic pile at Sunset Lake by an intrusion of Prosperous Granite. Southeast of Turnback Lake is a broad anticlinal structure that trends east, perpendicular to the strike of the volcanics through Tumpline and Turnback lakes. This structure is cored by a more or less concordant Defeat granodiorite pluton similar to the intrusion that occupies the dome to the southwest. This structure ends at a northerly trending fault, although inclusions of volcanics occur in the granitic rocks on strike from the southern limb.

The volcanic complex consists of approximately equal proportions of mafic and felsic units. Although the Complex has been divided into either felsic or mafic units, individual units or parts of a unit can vary from being dominantly one composition or the other to intimate mixtures of the two. In many cases the contact between map units is abrupt and, in some cases, clearly distinguishable even on air photographs, whereas elsewhere it is gradational. The units are based on field criteria, in this case mainly colour. This is not particularly reliable as colour can be influenced by metamorphic processes and later bleaching. Lambert (in press) in his more detailed study of these rocks has shown that specific gravity is a much better indication of composition. The felsic units are mainly rhyolitic in chemical composition whereas the mafic units are mainly iron-rich basalts. As well as compositional variation there is a wide textural variation with both flow and fragmental units of both compositions. Lambert (in press) has divided the volcanic rocks into three formations: the Tumpline Basalt, which includes all the mafic volcanic rocks, and two rhyolite formations, the Turnback Rhyolite and Sharrie Rhyolite.

Tumpline Basalt

Mafic volcanics in the Victory-Turnback lakes area are all part of the Tumpline Basalt (Lambert, in press) and occur as both fragmental and flow units. In general they weather much darker than the felsic units, typically black to dark green to brownish green. The effect of metamorphism is to darken the colour of the mafic rocks. Primary textures are less well preserved in the mafic volcanics than in the felsic volcanics due to metamorphic recrystallization. Primary layering is still evident, however, and is expressed in variations in composition and metamorphic texture. Individual units vary in thickness from a few centimetres to several metres or more. Bedding planes are commonly not sharply bounded. The mafic flow units are thicker, more massive and more homogeneous in aspect and are commonly difficult to distinguish from the thinner intrusive sills although the sills tend to be somewhat coarser in grain size. The lava flows occur locally within the mafic fragmental volcanic sequences as a series of flows several tens of metres thick and also as thick sequences made up almost entirely of lava flows, particularly in the central part of the complex. The flows are both massive and pillowed, with pillowed flows being particularly common in the thick lava flow sequences. Pillowed flows, as with the other units, are highly deformed and top determinations can only be made with confidence locally. Primary flow top breccias are present in some cases. Lambert (in press) notes that in areas that have undergone extreme deformation, originally pillowed volcanics can look like layered rocks. The volcanics of the formation are dominantly iron-rich basalts although Lambert (in press) has mapped a small andesite member west of Devore Lake. As opposed to most other andesites in the region it, like the basalts, is iron rich.

The mafic volcanics are not as clearly related to particular centres as are the felsic volcanics. However, there is a general regional variation with thick sequences of lava flows more common in the central part of the complex whereas the more distal parts consist of a higher proportion of fragmental units. For example, in the western half of the complex between Devore and Sharrie lakes the volcanics are dominantly mafic and pillowed, whereas to the northwest of Sharrie Lake there is a greater mixing with felsic units and the mafic material is fragmental. South of Tumpline Lake the volcanics are mainly pillowed flows. In the easterly trending, dominantly mafic volcanic structure northeast of Turnback Lake the volcanics are dominantly pillowed flows at the nose of the structure, while to the east the volcanics have a higher proportion of fragmental rocks and become increasingly thinner bedded. Along the north edge of the complex the volcanics, mainly mafic, are too highly deformed and metamorphosed to preserve any primary textures.

Mafic intrusions, as in most of the other mafic volcanic sequences, are present in this complex. They occur mainly as more or less concordant sill-like structures and as large irregular intrusions. The larger bodies are shown on the map and are most abundant in the central part of the complex. They consist of massive medium- to coarse-grained amphibolite and have suffered varied degrees of deformation. Similar intrusions, although much fewer in number, also occur in the sediments. One such sill occurs just southwest of the discordant Prosperous Granite tongue that intersects the complex. Although the sill and the sedimentary sequence in which it was emplaced are now steeply dipping, the fact that the sill has a granophyric top would indicate that the sill was emplaced while the sedimentary sequence was still horizontal. They are probably contemporaneous with volcanism.

Turnback Rhyolite and Sharrie Rhyolite

The felsic volcanics in the main body of the complex make up the Turnback Rhyolite while those volcanics at the end of the northwest arm of the complex (northwest of Sharrie Lake) form the Sharrie Rhyolite (Lambert, in press). The felsic volcanics generally weather light and range from white to grey, through pale yellows and greens to pale pink and buff and are darker on fresh surfaces. The rocks are commonly porphyritic with embayed quartz phenocrysts most abundant but also plagioclase both as individual crystals and as aggregates of crystals. Lithic clasts to recrystallized glassy lenticles are also present.

The volcanics are dominantly fragmental, ranging from fine ash flow tuffs, bedded tuffs, arenites to coarse breccias with clasts in excess of 30 cm. The clastic nature of the rocks is commonly more easily seen in weathered outcrop than in polished surfaces or even thin section. Throughout the complex the rocks are strongly deformed, with the clasts having length to width ratios commonly in excess of 6:1. The felsic fragmental rocks are layered on all scales from less than a centimetre to more than several metres, with the thickest units occurring in the thickest parts of the formation. In many cases lapilli tuffs occur in thick massive units. Bedding or laminae tend to be straight and parallel. This is due in part to the shearing of the units.

Some thick massive fine grained equigranular and sparsely porphyritic units are interpreted as lava flows. In some cases flow layering is evident and varies from straight to highly contorted. In the thicker massive bodies layering is defined only by contact with adjacent breccias. As they are not laterally extensive they probably represent felsic domes. North of Goose Lake at the sharp contact between one of the felsic domes and a unit of mixed mafic and fragmental volcanics, large massive felsic volcanic blocks up to 2 m long and over 1 m wide occur in the mixed fragmentals, and conceivably could have slid off the adjacent dome. Although mafic fragmental units occur locally within the dominantly felsic sequences, they never occur within the large massive felsic domes.

Felsic units also occur in varied proportions within the dominantly mafic volcanic sequences. They range from thin fragmental units of a few centimetres to units several metres thick. Commonly they occur in sequences 3-5 m thick consisting of several depositional layers.

The felsic volcanics form five major centres in the complex. These include the centre at the northwest end of the complex south of Victory Lake and the lensoid structure between Devore and Goose lakes that is transected by the tongue of Prosperous Granite. This intrusion also transects the large felsic centre between Detour and Tumpline lakes. West of the pluton in the core of the dome is another felsic centre, and another is situated at the north arm of Turnback Lake. These centres probably represent felsic domes with a massive core of fine grained homogeneous lava with more distal deposits of breccia and tuff on the flanks. For example, at the felsic centre south of Victory Lake, massive homogeneous lavas occur on the southwest side while the northern part is dominantly fragmental. Similarly, between Goose and Devore lakes the volcanics are massive and homogeneous just north of Goose Lake, while towards Devore Lake and east toward the dome fragmentals dominate. These felsic centres are in the order of 10 km in diameter, although their relief is not known. Throughout the complex are smaller felsic concentrations that may represent smaller centres or the distal equivalents of other centres otherwise not exposed at the present erosion level.

The stratigraphic position of these centres relative to the mafic volcanics is difficult to assess due to the complex structure, but it would appear that the various centres of felsic volcanism both underlie and/or overlie the basalts of the Tumpline Formation (see Lambert, in press).

Carbonate units

Carbonate units occur from place to place along the northern margin of the complex. They are best developed at the east and west ends of Detour Lake and on islands in the lake, and north of and along the west shore of the north arm of Turnback Lake. Similar units occur to the northwest at Upper Ross Lake and to the northeast, southwest of the Amacher Granite. They are typically at or close to the contact between the volcanics, particularly the felsic volcanics and the sediments. In places, black carbonaceous, locally sulphide-bearing mudstone is associated with the units. The carbonate units tend to be lensoid and vary from essentially homogeneous dolomite or calcite to a carbonate-cemented felsic volcanic breccia. The carbonates or carbonate-rich rocks represent a discontinuous stratigraphic horizon that can be traced up to 50 km from south of the Amacher Granite to Upper Ross Lake where the Raquette Lake Formation sandstones and conglomerates have a carbonate matrix. Equivalents of the carbonate have been identified by Davidson (1972) as inclusions within the granitic units between Detour and Turnback lakes. Although the carbonate may represent a stratigraphic horizon, there is little likelihood that it ever was a continuous stratigraphic unit, but was originally lensoid, representing local accumulations thought to represent carbonate precipitates from fumaroles contemporaneous with volcanism. This association of the carbonate deposits with the felsic volcanics and carbonaceous sediments is common in several places in the Slave Province. Examples include the large, mainly felsic volcanic complex near the headwaters of the Back River which is discontinuously rimmed by carbonates (Henderson, 1975a; Lambert, 1976, 1978), and parts of the High Lake volcanic belt (Padgham, 1974; Henderson, 1975a,c). At both localities the carbonate occurs as local lenses along a stratigraphic horizon. Base metal mineralization is locally associated with these carbonate units. For example, at Turnback Lake the base metal mineralization is closely associated with a carbonate unit but has evidently been concentrated by later pegmatitic activity (Shegelski and Thorpe, 1972).

Carbonate cemented volcanic breccias also occur within the volcanic complex, with one such example occurring southwest of the small satellite plug southeast of the large granodiorite pluton that intrudes the central part of the volcanic complex.

Evolution of the Victory – Turnback lakes volcanic complex

The complex is interpreted as a volcanic centre in which subaqueous mafic volcanism, mainly in the form of pillow lavas, erupted from several as yet undefined vents or fissures near the centre of the complex. The more distal parts of the complex are dominated by mafic volcaniclastic rocks. Superimposed on the mafic volcanic pattern were smaller, localized, relatively high relief centres of felsic volcanism that, at least in some cases, were centred by felsic flows in the form of domes with a carapace of dominantly fragmental material that extended beyond the flow centres. Various centres were active before, during and after mafic volcanic activity. Attempts to interpret the environment of volcanism is made difficult by the high degree of deformation that commonly obscures primary textures. Certainly the fact that the complex is overlain by greywacke-mudstone turbidites, and the presence of pillowed mafic flows would indicate that during at least part of its history it was subaqueous. Lambert (1977, in press) concluded that part of the complex was emergent on the basis of his recognition of welded tuff textures in better preserved parts of some felsic units.

The only essential difference between the Victory-Turnback lakes volcanic complex and the complexes at Yellowknife, Cameron River and Sunset-Payne lakes is the proportion of mafic to felsic volcanics. In all areas the volcanic complexes consist of mafic subaqueous volcanics with local felsic centres that in general but not always occur towards the top of the mafic sequence. In the Victory – Turnback lakes complex the felsic centres tend to be both greater in number and larger in size so that the difference is more of degree than style.

Conglomerates and lithic and quartz sandstones

Jackson Lake Formation – Yellowknife

The Jackson Lake Formation consists of a locally distributed basal conglomerate overlain by crossbedded sandstone and local thin conglomerate layers. The formation occurs on the west side of Yellowknife Bay, from the Sub Islands in the south to Greyling Lake north of the bay, over a strike length of 37 km. It was first mapped at 1 inch to 1 mile scale by Jolliffe (1942, 1946) and parts of it by Henderson and Brown (1950, 1952a, 1966). The southern part of the formation was studied in detail by Henderson (1975b) and the following account is based mainly on that work.

Structurally the formation is situated on the west limb of the major northerly trending syncline through Yellowknife Bay. It does not appear on the east limb but its place is taken by the Burwash Formation greywacke-mudstone turbidites. The formation is offset by a series of left lateral faults that intersect the formation at low to intermediate angles. Throughout most of its exposed extent the formation has been metamorphosed to greenschist grade although in its northernmost exposures it reaches amphibolite grade. The metamorphic mineral assemblage at the lowest grade in the central part of the exposure area is quartz-plagioclase-chlorite-white mica-carbonate. Where the bulk composition is favourable, quartz-plagioclase-chlorite-chloritoid-white mica is a common assemblage.

The Jackson Lake Formation lies unconformably above the Kam Formation mafic volcanics. The unconformity is angular and intersects the flows of the Kam Formation at an angle of about 20 degrees (Fig. 13). The Jackson Lake Formation appears to transect most of the stratigraphic thickness of the volcanic sequence but since the thickness of the volcanic pile decreases on strike towards the unconformity (Henderson and Brown, 1966) and the Kam Formation itself evolved by progradation from north to south (see Fig. 14), the actual amount of material removed by erosion is less than would be expected if the volcanic sequence is assumed to have a uniform thickness throughout its extent. The unconformity surface, where exposed, is well preserved, with the original relief clearly evident in many places. South of Jackson Lake, for example, there is a depression on the unconformity surface that is up to 250 m deep and 700 m across.

The upper contact of the Jackson Lake Formation is conformable with the Banting felsic volcanics as first shown by Baragar (1975). Most previous workers considered the contact to be faulted (Henderson and Brown, 1952b, 1966; Boyle, 1961; Henderson, 1973, 1975b). Although Jolliffe (1942) originally implied that the Jackson Lake represents the basal unit of his Division B of the Yellowknife Group, to others the relationship of the Jackson Lake Formation to the Yellowknife Supergroup was in some doubt. For example Henderson and Brown (1952b, 1966) and Boyle (1961) suggested, on the basis of its apparent structural relationship and contrast in the nature of the sedimentary rocks to others in the Yellowknife Supergroup, that the formation might be significantly younger. Henderson (1970), on sedimentological grounds, proposed that the formation is part of the Yellowknife Supergroup and is the shallow water equivalent of the basinal turbidites (the Burwash Formation) to the east. However, Baragar (1975) in studying drill core over a thickness of about 1800 m across part of the Kam, the Jackson Lake and part of the Banting Formation near the north end of Yellowknife Bay, reported that the contact at that locality, although somewhat obscured by quartz-carbonate alteration, is abrupt but that beds of sediments like those of the Jackson Lake occur within the Banting volcanics, suggesting the contact is conformable and possibly interfingering. Helmstaedt et al (1980) have subsequently found a surface exposure 1 km north of Banting Lake on which the gradational nature of the contact between the two formations is evident.

The thickness of the formation is varied and appears to correspond inversely with the thickness of the overlying Banting felsic volcanics. The formation is thinnest (300 m) in the vicinity of Walsh Lake where the Banting Formation reaches its greatest thickness. To the north and south the formation becomes thicker, up to 600 m near Homer Lake and at the north end of Yellowknife Bay Baragar (1975) reports a similar thickness.

The formation consists basically of two parts: a discontinuously distributed basal conglomerate and overlying dominantly crossbedded sandstones with thin beds and lenses of conglomerate.

Conglomerate

The basal conglomerate occurs mainly in depressions in the unconformity surface, the best example being south of Jackson Lake. Other similar conglomerates occur on various island groups in Yellowknife Bay, such as south of Jolliffe Island and the Sub Islands, although their relationship to the unconformity is not exposed. The conglomerate consists mainly of angular clasts of volcanic rocks similar to those below the unconformity. The volcanic clasts are mainly mafic but minor amounts of felsic volcanic clasts are commonly also present and locally they can be quite important. For example where a felsic unit in the volcanic pile is intersected by the unconformity, such as south of Jackson Lake, felsic clasts dominate the overlying basal conglomerate. The angular to subrounded clasts are up to a metre in size although in most cases are less than 15 or 20 cm. Also present are relatively minor amounts of sandstone clasts although locally they can be quite important. Some of the mafic clasts are rimmed with darker zones that range from a thin millimetre scale darker rind to more rarely wider zones several centimetres thick. These represent either predepositional weathering or a postdepositional low grade metamorphic reaction with the much more siliceous matrix.

Several of the conglomerates contain pink to grey granitic cobbles. They are generally much less abundant than the volcanic clasts, are invariably much better rounded, and in some cases are the coarsest clasts in the conglomerate. These boulders are commonly trondhjemite with an average of about 55% altered plagioclase in subhedral to euhedral laths, 35% quartz, and 10% mafic minerals that are presently mainly chlorite. Potassium feldspar is rare to absent. Miarolitic cavities are present in one of the boulders. Zircons recovered from granite boulders from two localities in the formation have ^{207}Pb-^{206}Pb ages of 2555 and 2535 Ma (Green and Baadsgaard, 1971). Although the source of these granitic clasts has not been identified, it would appear that they were derived from a shallowly emplaced pluton. Texturally they are similar to the Amacher Granite near the northeast corner of the map area which is thought to be an epizonal pluton contemporaneous with Yellowknife volcanism. Also present in these conglomerates are altered granitic boulders in which the mafic minerals are completely replaced and feldspars are altered to a fine quartz-feldspar-carbonate mass, although in some cases vague outlines of the original granitic texture is perceptible. In many cases these weaker cobbles tend to be deformed, resulting in a diffusely bounded sheared boulder with abundant coarse quartz. As these altered clasts occur in the same conglomerate as the relatively fresh cobbles, the alteration must have been predepositional and a product of Archean weathering.

The matrix of the conglomerate consists of coarse quartz, felsic, and to a lesser extent, mafic volcanic rock fragments and feldspar. The amount of matrix is varied and while most of the conglomerates are clast-supported, some are matrix-supported. In general, the basal conglomerates are structureless although in some cases there is a weak size gradation and in places a compositional gradation with the granitic cobbles more abundant toward the top. Discontinuous to lensoid units of current laminated to cross-bedded sandstones, generally less than 1 m thick, occur locally within the conglomerate.

The abundant coarse angular clasts of very local derivation suggest that the conglomerate was derived from a very rapidly eroded part of the volcanic pile and may represent rapid erosion of fault scarps in the vicinity of the depositional area. The presence of the much better rounded granitic clasts would indicate the exposure of a presumably somewhat more distant granitic terrane, also possibly exposed due to faulting in the hinterland to the west. The discontinuous current laminated sandstone lenses within the conglomerate and the presence of the conglomerate only in depressions in the unconformity surface would imply movement of the material by water, perhaps by periodic floods that were controlled to a certain extent by the morphology of the erosional surface. The succeeding sandstone member was deposited on a relatively even surface above the conglomerate where present or Kam Formation.

Sandstones

The main part of the Jackson Lake Formation consists of a thick sequence of well bedded lithic wackes. Associated with these sandstones are thin lensoid conglomerates that are more common in the lower part of the sequence. The transition from basal conglomerate to sandstone is abrupt, with the sandstones immediately overlying the conglomerate being generally more thickly bedded, more varied in grain size, and somewhat more argillaceous than the generally homogeneous main part of the formation.

The conglomerate associated with the sandstones occurs as elongate lensoid layers rarely more than 1.5 m and commonly only 15 to 30 cm thick. Rarely units are up to 18 m thick. These thicker units, as can be seen on Jolliffe Island, are commonly graded. The cobbles and pebbles are typically well rounded and consist mainly of sandstone, felsic volcanics, vein quartz and chert. Mafic volcanic clasts are rare. The clasts tend to be less than 7 cm in size. These conglomerate beds contrast strongly with those of the basal conglomerate.

The sandstones typically occur in massive units 15 cm to 3 m thick with an average thickness of about 30 cm. They vary from light greyish green to pale brown on weathered surfaces and darker grey to greyish green on fresh surfaces. The sandstone beds commonly are separated by thin argillaceous siltstone or mudstone layers to partings. The sandstones are commonly parallel sided but locally pinch and swell due to scouring and channelling (Fig. 29). Mud chip conglomerates derived from the erosion of the argillaceous horizons occur locally. Large scale channels are locally present, some of which are up to 12 m deep. Sedimentary structures in these beds include parallel lamination, medium scale festoon and tabular crossbedding (about 15 cm), and rare ripple crosslaminae in some of the finer sands and silts. Locally pebbles occur either as single pebbles or in matrix-supported groups. Concretions commonly occur as diffuse dark brown recessive weathering features in many of the sandstones, and are particularly evident in crossbedded sands in which the crossbeds stand up in relief.

The sandstones of the Jackson Lake Formation contrast strongly with Burwash Formation greywacke-mudstone turbidites that are the dominant type of sediment in the Yellowknife Supergroup within the map area. The scours and channels, abundant parallel lamination in the sands, the low angle crossbedding and the coarser grain size of some of the beds imply deposition under the influence of strong currents. The presence of thin mud drapes and lensoid gravel deposits suggests a widely fluctuating current regime. The sandstones are interpreted as fluvial deposits on the lower part of an alluvial fan or alluvial plain. There is no evidence of the typical fining upward sequences characteristic of point bar deposition of meandering rivers. The sands of this formation represent the deposits of a braided river system with the depositional locus shifting randomly over the alluvial plain, resulting in no regular cyclical deposition of sediments.

The sands consist primarily of rock fragments and quartz. On the weathered surface the quartz grains are readily apparent but individual rock fragments are rarely separable from the matrix of the rock. In thin section, however, the outlines of detrital rock fragments are clearly visible in minimally deformed rocks. The rock fragments and quartz grains occur in an abundant matrix of very fine quartzofeldspathic material, chlorite, white mica and chloritoid (Fig. 30). Quartz grains form about one-quarter of the rock and are for the most part monocrystalline and angular. A few grains are euhedral in outline and have embayments characteristic of volcanic quartz. The quartz is all first cycle as any secondary quartz overgrowths on the grains are postdepositional. Rock fragments form about 30% of the sediment. They are generally well rounded and consist of a very fine intergrowth of mainly quartzofeldspathic material. There are slight variations in colour and internal grain size among the grains. Some contain quartz and feldspar phenocrysts and are clearly of volcanic origin. Some granitic rock fragments are present as well but are very rare. However as Schau and Henderson (1983) have pointed out, the similarity between highly altered feldspar in weathered granites and the fine grained quartzofeldspathic matrix of intermediate to felsic volcanics that have both undergone low grade metamorphism makes it difficult to determine with confidence the provenance of many individual lithic clasts. In a few cases intraformational mudstone clasts are present. Plagioclase feldspar is a very minor component of these sandstones and occurs as angular grains that have undergone only minimal alteration. Potassium feldspar has not been reported. The matrix of the sandstone makes up between 30 and 40% of the rock and is derived largely from the postdepositional breakdown of the fragments. In many cases a complete gradation can be seen from sharply bounded clasts to the diffuse mosaic of quartzofeldspathic material that is the major component of the matrix. As a major framework component of the sediment is felsic volcanic rock fragments that may have been largely glass at the time of deposition or altered feldspar from weathered granite, postdepositional mechanical and chemical breakdown of such unstable clasts to form matrix could be expected. Metamorphic minerals make up between 5 and 20% of the rock. Of these, the most important is chloritoid, which occurs as scattered crystals to radiating masses within the matrix. Coarse chlorite and fine muscovite are also present. In the southernmost part of the formation the rocks are extensively carbonatized.

Chemical data from this formation as determined by Jenner et al. (1981) are presented in Table 7. In general the relatively low concentrations of K, Na, Ca and high Al and Fe

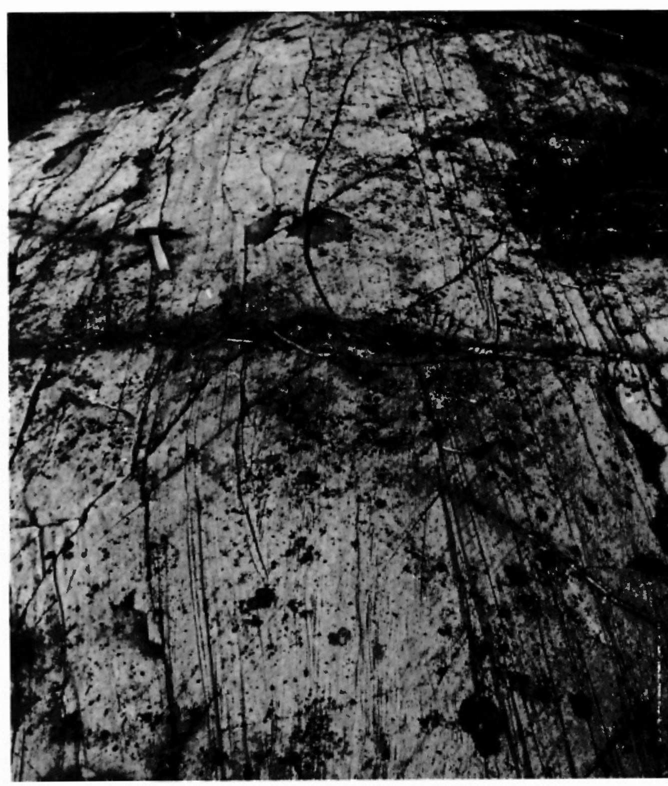
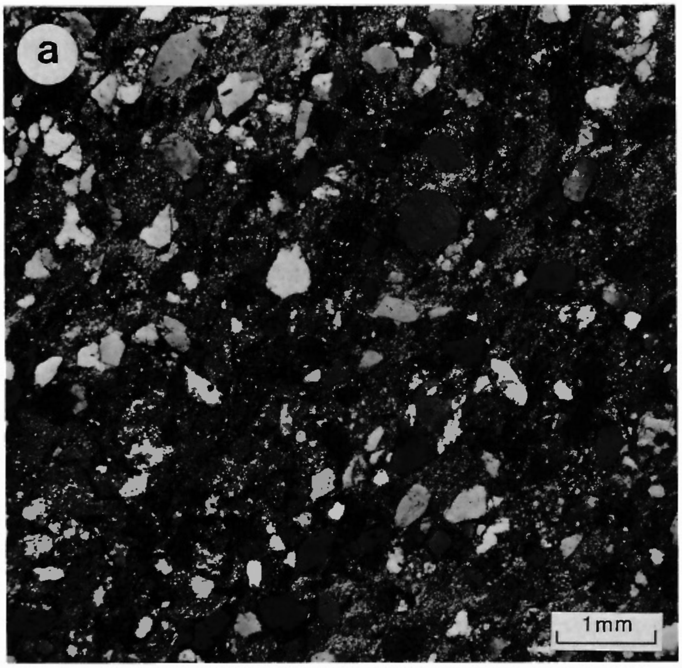


***Figure 29.** Jackson Lake lithic sandstone showing irregularly scoured bedding planes, parallel and crosslamination, and dark discontinuous mudstone interbeds that in most cases define the bedding planes. GSC 157544*



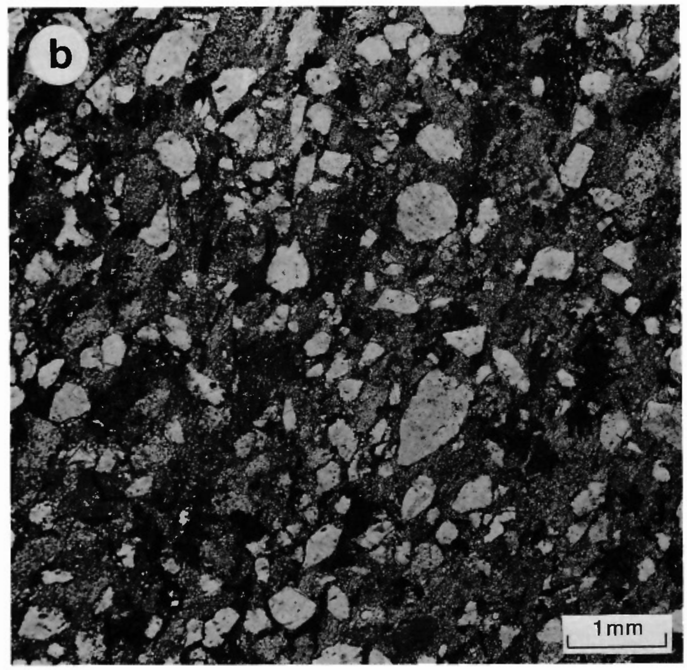


Figure 30. *Jackson Lake Formation lithic wacke. This sandstone is derived from a mixed felsic volcanic and weathered granitoid source, as indicated by the abundant quartz grains and lithic clasts composed of a fine quartzofelspathic mosaic. Postdepositional diagenesis and low grade metamorphism have commonly obscured the outlines of the lithic clasts. The black acicular minerals are chloritoid. (a) Crossed polarizers. GSC 203660. (b) Plane light. GSC 202790-Z.*

Table 7. Chemical analyses, Jackson Lake Formation

	1	2	3	4	5	6
SiO_2	60.53	67.98	67.02	58.90	58.77	79.37
Al_2O_3	18.48	14.85	17.34	12.60	9.81	9.55
Fe_2O_3	14.12	6.99	9.89	4.15	8.41	4.81
MgO	1.03	2.38	.47	1.09	3.43	1.25
CaO	.04	.70	.11	8.48	6.58	1.17
Na_2O	.12	.50	.23	4.54	1.92	.11
K_2O	.52	1.69	.95	.65	.95	.21
L.O.I.	2.92	3.96	2.53	8.09	11.29	2.23
TiO_2	1.22	.78	1.03	.34	.39	.39
P_2O_5	.05	.12	.04	.02	.09	.03
MnO	.24	.03	.04	.14	.14	.03
Total	99.27	99.98	99.65	99.00	101.78	99.15

Chemical data from Jenner et al. (1981). Trace element and rare earth element data are available for this and other units in the immediate Yellowknife area in Jenner et al. (1981).

Total iron reported as Fe_2O_3.

as well as the high Fe/Mg ratio favour the formation of metamorphic chloritoid (Hoschek, 1967) which is so common in this unit.

Source and significance

The basal conglomerates of the Jackson Lake Formation were derived mainly from a mafic volcanic terrane, the Kam Formation. With the deposition of the sandstones, however, the composition changed abruptly from mafic to felsic. On the basis of the occurrence of quartz and feldspar phenocryst-bearing rock fragments, a felsic volcanic component in the source terrane is indicated. Although no preserved sequence of felsic volcanics that is a likely source of this material has been recognized, the Jackson Lake Formation is conformably overlain by the Banting felsic volcanics, indicating that felsic volcanism was active at that time. Within the underlying Kam Formation there is a swarm of anastomosing dykes and sills (Fig. 20) that are truncated at the unconformity between the Kam and the Jackson Lake formations and could conceivably represent the subvolcanic conduit system to a now eroded felsic volcanic sequence.

The abundance of quartz in the formation requires explanation. It seems unlikely that all the quartz could be volcanogenic even though some clearly is, as shown by the morphology of the grains. Felsic volcanics that are preserved in the region are not particularly rich in quartz phenocrysts. While some concentration of this quartz may have taken place due to the breakdown of glassy volcanic rock fragments, the well-preserved state of the lithic fragments that remain indicates that such a process would likely be inadequate to produce the necessary amount of quartz. An additional source is required and could presumably be a quartz-rich granitic terrane. The general lack of feldspar would indicate that any such granitic source may have been deeply weathered. The presence of highly altered granitic boulders in some of the basal conglomerates is further evidence of a weathered granitic source.

On the basis of their chemical data, primarily the rare earth data, Jenner et al. (1981) have suggested the sediments were derived in some cases from a mixture of equal parts mafic and felsic volcanics and in others from a mainly felsic volcanic source. Petrographically no evidence of a mafic volcanic component has been recognized. Schau and Henderson (1983) pointed out the petrographic similarity between the fined grained felsic rock fragments in the Jackson Lake sandstones and the weathered granite seen unconformably below Archean supracrustal rocks as, for example, at Point Lake (see also Fig. 8). Schau and Henderson (1983) also pointed out that the major oxide data (Table 7) have a similar pattern to that of weathered granitoid rocks (low in Na and Mg, high in Al, Ti and Fe). They argued that rare earth elements are chemically differentiated during weathering and early diagenesis, and suggest the Jackson Lake rare earth data can be interpreted as supporting evidence for the existence of a weathered granitoid component in the source.

The Jackson Lake Formation and the Raquette Lake Formation near Upper Ross Lake (described in the following section) are the only units in the area indicative of subaerial or shallow water deposition. The presence of such units, in contrast to the intervening basinal turbidites of the Burwash Formation, is the best evidence indicating that the margin of the depositional basin east of Yellowknife is close to its present-day structural margin. Sediments similar to the Jackson Lake Formation are known to be present at two other localities in the Slave Province; Point Lake in the west-central Slave Province (Henderson and Easton, 1977) and Anialik River (Tirrul and Bell, 1980). At Point Lake, about 300 km north of Yellowknife, the Keskarrah Formation (Bostock, 1980) is situated at the margin of another or perhaps correlative basin. Like the Jackson Lake the Keskarrah consists of a granite and volcanic boulder conglomerate member which predominates, and a sandstone member. Both are quite similar in character to those at Yellowknife (Henderson and Easton, 1977). The Keskarrah Formation lies unconformably above both granitic basement and, locally, a major mafic volcanic sequence, although in places coarse granitic detritus occurs within the mafic volcanic sequence. It is evident that block faulting in the granitic basement was contemporaneous with mafic volcanism in the Point Lake area (Henderson, 1981). Coarse detritus, derived from both the volcanics and granitic basement of the uplifted blocks, was shed into adjacent negative areas and was succeeded by deposition of alluvial sands of dominantly felsic derivation. Although these relationships are not as clearly preserved at Yellowknife, the situation at Point Lake seems to be applicable.

In the Superior Province similar alluvial sediments, although not common, do occur. Turner and Walker (1973) have described Archean alluvial fan deposits in the Abram Group near Sioux Lookout, Ontario. Teal and Walker (1977) and Hyde and Walker (1977) have discussed similar rocks of the Manitou Group near Dryden, Ontario, and Timiskaming Group near Kirkland Lake, Ontario, respectively.

Raquette Lake Formation – Upper Ross Lake

The Raquette Lake Formation is defined herein as a heterogeneous assemblage of discontinuous sandstone and conglomerate units that occur east of Upper Ross Lake and north of Raquette Lake, after which the formation is named (Fig. 23). Although volumetrically rather minor, it is an important unit as far as interpreting the evolution of the basin in which the Yellowknife Supergroup was deposited.

The Raquette Lake Formation overlies, and to some extent is interbedded with, the southeast flank of the Cameron River mafic volcanic sequence where the volcanics have thinned from a thickness of 3000 m to a few tens of metres (Fig. 23, 24). To the southeast it continues past the most distal volcanics and there lies above the Sleepy Dragon Complex. The contact with the volcanics is conformable, commonly gradational, with the breccias of the Cameron River volcanics grading into the well rounded mafic volcanic, pebble conglomerate of the basal Raquette Lake Formation. The contact with the Sleepy Dragon Complex is unconformable although in most cases it is faulted as well. The extent of the formation to the southeast is unknown due to Quaternary cover east of Raquette Lake. West of Detour Lake a calcareous unit occupies a similar stratigraphic position. To the northwest, in the vicinity of the central part of Upper Ross Lake, the Raquette Lake Formation interfingers with the volcanics. This would indicate the formation is contemporaneous with volcanism. At no place was the contact between the Raquette Lake and the overlying Burwash Formation greywacke-mudstone turbidites observed, although to the north the contact between the Cameron River volcanics, which underlie and interfinger with the Raquette Lake Formation, is conformably overlain by the Burwash, which would suggest a conformable contact between the Raquette Lake and Burwash as well. The known lateral extent of the formation is about 5 km. Its maximum known thickness is about 60 m. The formation is metamorphosed to amphibolite facies as it lies within the metamorphic aureole about the Redout Granite as defined by the cordierite isograd in the adjacent Burwash metasediments.

The formation is quite varied along strike with the most common unit being a quartzite that occurs in most sections. No single section contains all units. At its northernmost exposure on the central part of the west side of Upper Ross

Figure 31. *Raquette Lake Formation pebbly sandstone with angular intraformational clasts of finer sandstone near the southeast end of Upper Ross Lake. The sandstone consists dominantly of quartz. The smaller light clasts are granitic while the larger, darker angular clasts are finer grained intraformational sandstone. Both the sandstone and the sandstone clasts have a carbonate matrix. GSC 177760*

Lake the formation consists of only 4 m of quartzite overlain by 15 m of carbonaceous mudstone. Below the formation at this locality is about 40 m of dominantly mafic volcanic breccia, flows and fine volcaniclastic units. It is overlain by 60 m of felsic to intermediate volcanic units. The most complete section, designated as the type section, occurs a kilometre to the southeast (Fig. 23). There, 18 m of mafic volcanic-derived conglomerate is in gradational contact with the mafic volcanic breccia to conglomerate of the Cameron River Formation. Less than 5% of the cobbles consist of vein quartz, felsic volcanics, possible metadiorite and very minor granitoid pebbles. The conglomerate is followed by 3 m of dominantly felsic volcanic detritus. This is overlain by 18 m of quartzite that occurs in rather structureless 15 to 30 cm thick, white to buff beds of poorly sorted subangular quartz-rich detritus commonly in a carbonate matrix. Coarser clasts up to a centimetre in size are granitoid pebbles. Blocky intraformational clasts of a generally darker, finer grained quartzite occur locally in some beds (Fig. 31). The uppermost part of the formation and its relationship with the Burwash greywackes that are presumed to overlie it are not exposed.

In thin section the quartz-rich sandstones are highly varied in composition and degree of preservation (Fig. 32). The dominant framework component is poorly sorted subangular to subrounded polycrystalline quartz. Altered plagioclase and microcline are much less common components but many of the coarser framework grains are aggregates of quartz, plagioclase and microcline, which clearly indicate a granitic source. The matrix of the metasandstone ranges from a fine grained quartz-feldspar-biotite aggregate in which the primary textures of the sandstone are best preserved, to a carbonate cement where reaction between the carbonate and the framework quartz has resulted in irregular to sutured margins superimposed on the overall subrounded to subangular outline of the quartz. Sandstones with a calc-silicate metamorphic mineral matrix are also common and in these metasediments primary textures are all but obliterated. The calc-silicate assemblage consists of quartz-tremolite-diopside-microcline-epidote and is a reflection of the middle amphibolite metamorphic conditions the formation has been exposed to.

East of the southernmost bay of Upper Ross Lake the Raquette Lake Formation directly overlies granitoid basement of the Sleepy Dragon Complex. The basal unit consists of a well rounded mafic pebble conglomerate with a carbonate matrix. The cobbles are varied in texture and composition from fine grained schists to medium grained amphibolites and because of the carbonate matrix are only minimally deformed. The mafic conglomerate is overlain by a crudely bedded, poorly sorted conglomerate composed of generally coarser, angular to rounded clasts of a white, medium fine grained, deformed granitic rock similar to the

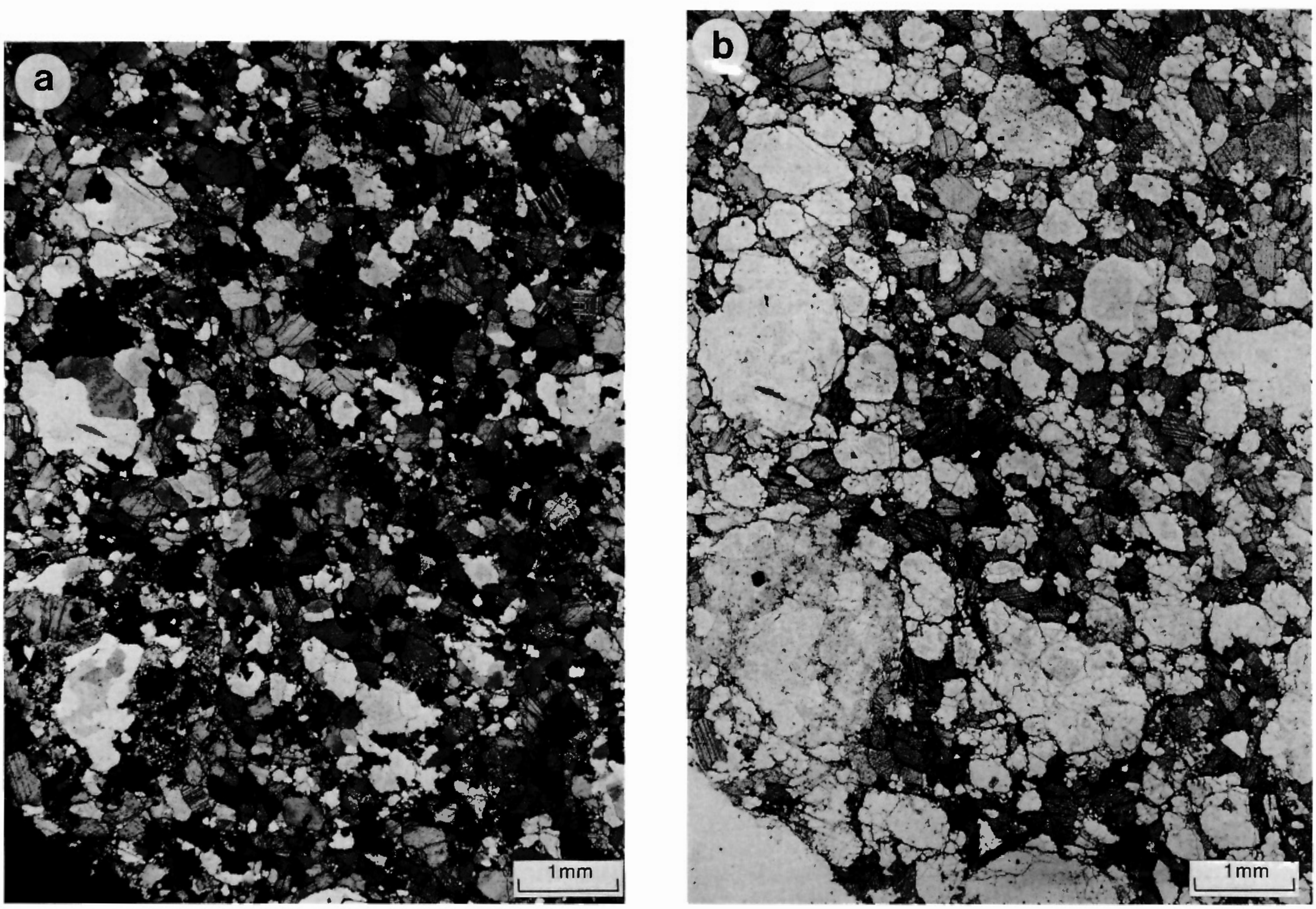


Figure 32. *Raquette Lake Formation carbonate-cemented quartz (feldspar) sandstone. Most of the original framework components are angular, irregular, polycrystalline quartz that has been partially replaced by the carbonate. A few microcline grains are also present. The coarser grains such as in the lower left corner of the photograph, are commonly aggregates of quartz and microcline, suggesting a granitoid provenance – a likely candidate being weathered detritus from the underlying Sleepy Dragon Complex basement. The sandstone is metamorphosed to amphibolite facies. (a) Crossed polarizers. GSC 201859-X. (b) Plane light. GSC 201859-U.*

deformed granitoids of the basement complex (Fig. 33). Less than 10% of the cobbles are chert, vein quartz and the occasional gabbroic or mafic metavolcanic clast. The matrix is a sandy carbonate similar in some respects to the quartzites elsewhere in the formation.

The basement underlying the conglomerate at this locality contains extensive carbonate-filled fractures and breccia zones up to 25 m below the contact with the Raquette Lake Formation. This breccia zone may represent a hydrothermal spring system that existed at the time of sedimentation and volcanism. Such springs may have been the source of the carbonate which forms the matrix to both the conglomerates and sandstones of much of the formation. In this regard, minor Zn-Pb mineralization occurs at the unconformity in the carbonate-filled breccia below the basal conglomerate, and is marked by a rusty zone 70 m long and 7 m wide.

The Raquette Lake Formation has important implications regarding the evolution of the Yellowknife supracrustal basin. The sediments provide unequivocal evidence for the derivation of the detritus from a granitoid source. The course, angular to rounded, deformed granitoid cobble conglomerate contains clasts closely resembling the underlying basement rocks and presumably derived from that basement from a locality near the site of deposition. Erosion of an adjacent rising fault scarp at the margin of the basin seems a likely environment. The quartz-rich sandstones that lie above the conglomerates are also of granitic derivation and may represent the redeposited quartz-rich lag of the weathered basement complex.

Greywacke-mudstone

Burwash Formation – east of Yellowknife

The Burwash Formation occupies almost half of the map area. The area underlain by the formation is about 60 km wide in the north between Yellowknife and Victory Lake. South of Detour Lake the basin widens to about

Figure 33. *Raquette Lake Formation breccia at the southeast end of Upper Ross Lake. The breccia is poorly to crudely bedded with angular to subrounded clasts of mainly granitoid but some mafic clasts (note large mafic clasts in centre). Carbonate matrix which is common in this formation is abundant. The breccia represents talus accumulation derived from a rising basement fault scarp. GSC 177870*

120 km between the city of Yellowknife and Francois Lake. The extensive Defeat Plutonic Suite south of Jennejohn Lake occurs within the sedimentary basin as essentially all the islands in Great Slave Lake between Mattonabbee Point and Wool Bay to the northeast are underlain by sedimentary rocks of the Burwash Formation. Gibb and Thomas (1980) suggest on the basis of gravity data that the sediments extend about 10 km to the southwest under Great Slave Lake. Their extent to the south is unknown as they are covered by water or Aphebian rocks in the East Arm of Great Slave Lake. Well preserved sediments extend 30 km to the north, beyond which the metamorphic grade, and proportion of granitic intrusions and migmatite greatly increase. Highly metamorphosed, extensively intruded and migmatitic equivalents of these sediments can be traced northward 350 km into the Point Lake-Contwoyto Lake area of the central Slave Province (Fig. 1).

The Burwash Formation consists almost entirely of interbedded greywacke-mudstone turbidites that are locally variable in character, but together form a rather homogeneous map unit.

Due largely to its comparatively incompetent nature, the Burwash Formation has been complexly folded throughout the basin. In general the beds occur in steeply dipping isoclinal folds of highly varied trend and wavelength. In many places complex interference patterns are developed. The structure is discussed more completely in the section on structure.

The sediments are everywhere metamorphosed to at least greenschist metamorphic grade. In this report sedimentary rock terminology (greywacke, mudstone, etc.) is commonly used for these rocks, particularly where the context concerns their primary nature. It should be kept in mind, however, that these, and all the supracrustal rocks, are metamorphosed. Within the basin there are two thermal ridges outlined on the map by the cordierite isograd. One about 40 km wide extends north from 8 km south of Prelude Lake to at least 30 km beyond the north border of the map area. Within this thermal ridge are extensive intrusions of Prosperous granites. The other thermal high is a dome-like feature 35 km in diameter about Buckham Lake. Higher metamorphic grades are associated with the Defeat Plutonic Suite and the Meander Lake Plutonic Suite. The metamorphism is of a low pressure facies series. In many cases primary structures are not unduly affected by the metamorphism which is discussed in greater detail later.

Contact relations

The formation is contained by Yellowknife volcanic units or intrusive granitic terrane. Several granitic bodies also intrude the Burwash. The contact between the sediments and the volcanic units, where exposed, is everywhere conformable. At the margin of the outcrop area of the formation the volcanic rocks everywhere underlie the sediments as the steeply dipping volcanics everywhere face towards the sedimentary basin. The only departure from the conformable nature of the contact between the volcanics and sediments is at Yellowknife where the alluvial sediments of the Jackson Lake Formation, thought to be the basin margin facies equivalent of the Burwash basinal turbidites, unconformably overlie the Kam mafic volcanics. However a few kilometres to the east, within the basin, the distal basinal equivalent of the Kam, the Duck volcanics, are conformably overlain by typical Burwash turbidites. The contact with the volcanics is commonly quite sharp. On the eastern side of the basin there is no repetition of volcanics above the first volcanic-sediment contact. This is also true in the area around Detour and Tumpline lakes where the apparent interlensing of volcanic and sedimentary units is due to structural reasons, with the volcanics occurring in the cores of anticlinal structures. On the west side of the basin, however, there is interbedding of the volcanics with the sediments in the vicinity of the volcanic contact. This is evident southwest of Clan Lake and also north of Duck Lake and at Preg Lake east of Yellowknife.

It is of interest to note that no volcanics outcrop in the central part of the basin; they are only present at the margin of the basin as presently exposed. On the basis of the gravity anomaly pattern on the regional gravity map of the area (Earth Physics Branch, 1969) there is little reason to believe the sediments in the central part of the basin are underlain

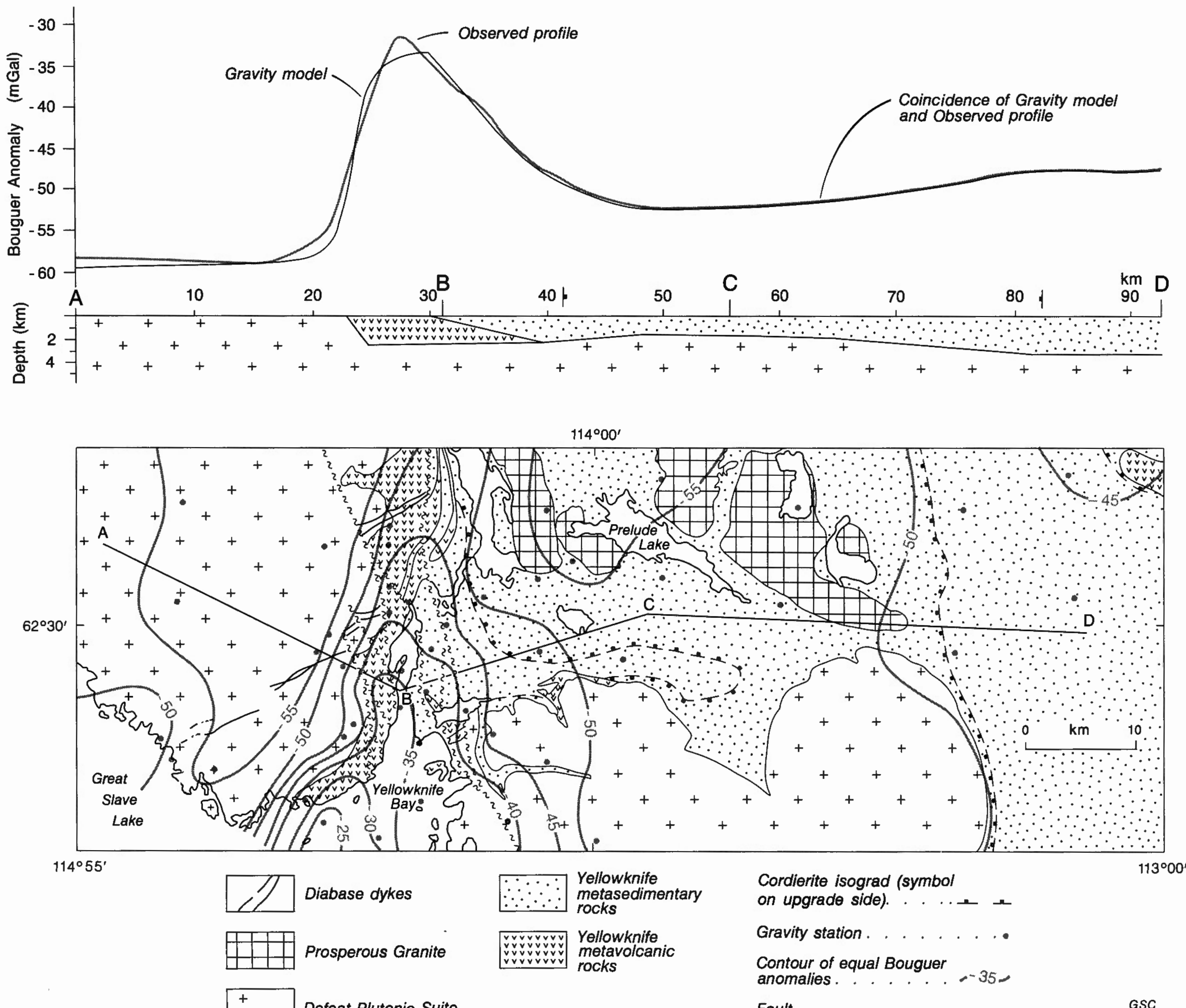


Figure 34. *Bouguer gravity anomaly profile through Yellowknife across the Kam mafic volcanics and out over halfway across the Archean basin that is filled mainly by Burwash Formation greywacke-mudstone turbidites. A gravity model that corresponds closely to the observed profile suggests the anomaly can be explained by a wedge-shaped block of Kam volcanics with a depth of about 2.75 km and a maximum width of about 10 km that is succeeded to the east by the Burwash metasediments which have a maximum depth of about 3.5 km at the east end of the profile. The supracrustal rocks have granitoid rocks both adjacent to and below them that in the central part of the profile, rises to within about 2 km of the surface. This region corresponds to the Archean thermal ridge within the basin, defined here by the cordierite isograd, that is extensively intruded by granite plutons at the present erosion level north (and south) of the profile. The model indicates that the Kam volcanics or its equivalents do not exist below the Burwash metasediments beyond about 10 km from the basin margin. While this is not proof that such a layer never existed at any time during the evolution of the basin, if such a layer did exist, there is no evidence that any part of it has been preserved in that part of the basin below the profile. After McGrath et al. (1983).*

by a thick sequence of mafic volcanics similar to that exposed at the margin. The larger mafic volcanic units have a pronounced positive anomaly and any significant thickness of mafic volcanics, if present below the Burwash, would influence the gravity anomaly pattern (Fig. 34) (McGrath et al., 1983). In this regard it is on the basis of a similar linear positive anomaly southeast of Yellowknife Bay that Gibb and Thomas (1980) proposed that the Kam mafic volcanics at Yellowknife continued southeasterly under Great Slave Lake. The volcanics that can be traced into the basin tend to become thinner away from the margins. Examples of this can be seen at Duck Lake southeast of Yellowknife where the volcanics thin towards the east and finally disappear 6 km east of Preg Lake. Similarly, at Turnback Lake, the volcanics that mantle the granodiorite pluton southeast of Turnback thin to the east away from the main volcanic centre. Thus the basin as it is presently preserved appears to be rimmed by volcanics that in general tend to be slightly older but overlap the time of sedimentation within the basin. Volcanics do not appear to underlie the entire basin.

In detail, the contact between the volcanics and the sediments is abrupt. In some places there are a few volcanic fragmental deposits interbedded with the normal quartz-rich greywackes above the contact. An example of this can be seen where the small river draining Gordon Lake enters the Cameron River. The sediments above the contact at Duck Lake are locally volcanogenic, and at the volcanic contact with the Burwash 8 km southwest of Payne Lake, where the volcanics are dominantly fragmental, the contact with the sediments is gradational. Intrusive granitic bodies occur at both the margin of the present basin and within it. Contact relations between the various granitoid units and the Burwash Formation are described more completely in the various sections on the granitic rocks but are summarized briefly as follows.

The Defeat intrusions are generally concordant, with the smaller bodies tending to be less concordant. Screens and trains of metasedimentary inclusions can be traced into the larger plutonic complexes. The Prosperous granites, on the other hand, are more discordant. In the area between Prosperous and Duncan lakes the Burwash Formation is extensively intruded by small stocks and pegmatites of Prosperous Granite. The Meander Lake Plutonic Suite north of Francois Lake tends to be discordant while to the south its contacts with the metasediments are more concordant. The extensive zones of inclusions of metasediments within the Meander Lake terrane would suggest the Burwash sediments may have extended to the east beyond the map area. The early Proterozoic Blachford Lake Intrusive Suite is sharply discordant to the Burwash sediments.

Lithology

The Burwash Formation is characteristically uniform which, together with the lack of distinctive stratigraphic units and its internal structural complexity, makes it difficult to subdivide stratigraphically. Within the unit there are no distinctive marker horizons that have any lateral continuity. Throughout most of the area there is no defined top or base to the formation. At the margins of the basin thick volcanic sequences that appear to become thinner into the basin underlie the formation. At Yellowknife, the Banting Formation felsic volcanics, which occur above the unconformity that truncates the mafic volcanics, also occur within the upper part of the sedimentary section to the east, although more greywackes and mudstones occur above these felsic volcanics (Walsh Formation see Henderson, 1970). At Yellowknife, the estimated thickness of the Burwash between the underlying mafic volcanics and the felsic volcanic unit high in the sedimentary section is approximately 5000 m. Elsewhere in the basin it is not possible to estimate the thickness as the top of the formation is nowhere recognized.

What follows is a brief descriptive summary of the nature of the unit. A reasonably representative and easily accessible part of the formation on the east side of Yellowknife Bay has been described in more detail elsewhere (Henderson, 1972, 1975b).

Throughout the formation the sedimentary units consist mainly of greywacke and mudstone couplets that give a distinctive striped aspect to individual outcrops (Fig. 35). The coarser grained sediments tend to weather lighter, generally light to medium grey through to light brownish grey, but locally can be a greenish grey, depending on composition and later alteration effects. The finer grained sediments tend to be progressively darker so that the finest mudstones are a dark grey to greyish black. Within a bed there tends to be a colour gradation from lighter to darker from the base of the bed towards the top reflecting the compositional and textural gradation in the sediments. The fresh surface of the rock is darker in all cases and there is much less colour variation.

The sedimentary units have most of the characteristics of sediments deposited by turbidity currents. Individual beds are laterally continuous and typically maintain constant thickness across the length of exposure. The bases of beds are sharply bounded. Most beds are graded, either continuously from sand size material at the base to clay size (originally) in the uppermost pelagic tops, or dicontinuously with grading restricted to part of a bed or varying abruptly at certain levels within a bed. In very rare cases there is an inverse grain size gradation at the base that normally reverts to normal a few centimetres above the base of the bed. Some beds have a very sharp grain size discontinuity within the bed, particularly between the sand sized material and the pelitic sediment. This also occurs within the coarser part of the bed (fine sand size or coarser) with the lower part being a much "cleaner" sand, while the upper part is somewhat finer grained but much more argillaceous. These features are

Figure 35. *Shoreline outcrop of vertically dipping greywacke-mudstone turbidites of the Burwash Formation, Yellowknife Bay. The beds are varied in thickness but are laterally constant and continuous. The lighter layers are greywacke while the more argillaceous layers are darker. Various current structures and graded bedding indicate the top of the sequence is to the left. Southwesterly trending glacial striae at a high angle to bedding are prominent. GSC 157291*

suggestive of grain flow deposits (Middleton and Hampton, 1973) where the moving sediment mass is kept mobile due to the grains bouncing off each other as opposed to the grains being supported by turbulence in the fluid medium, as is the case of true turbidity currents. Turbidity currents and grain flows along with fluidized sediment flows have been suggested by Middleton and Hampton (1973) as being members of a group of sedimentary mechanisms known as sedimentary gravity flows or mass flows. They suggest that gravity flows probably include more than one of these members during the evolution of a flow that perhaps started out as a debris flow or high density fluidized flow and evolved through a grain flow to a turbidity current as it progresses away from its source.

Grain size gradation is commonly apparent, particularly in the less metamorphosed material. Even at higher grades the grain size variation of quartz can be felt by scratching with a knife or hammer point, or be seen by the variations in mineralogy as the more aluminous metamorphic minerals tend to be more abundant and larger in size in the upper, finer grained parts of a bed. This is a reflection of variations in primary composition as well as texture through the bed. Cleavage at an angle to bedding is also influenced by grain size variation with the angle between bedding and cleavage being greater in the coarser grained material (Fig. 36). Thus, despite the degree of deformation and metamorphism, it is commonly possible to determine the facing direction of the sediments.

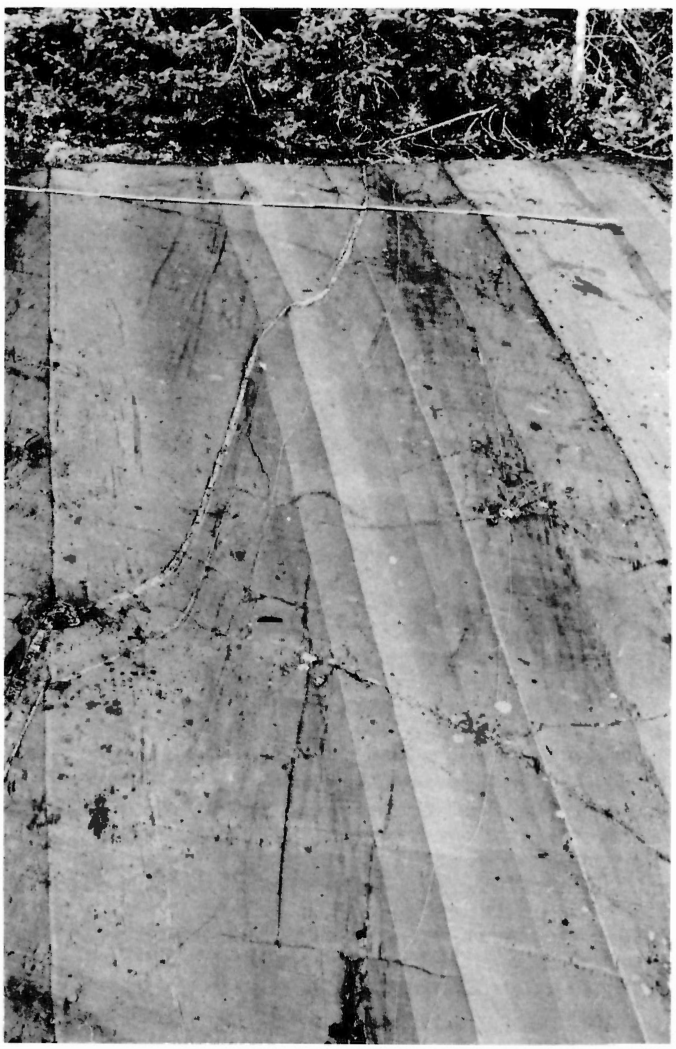


Figure 36. *Two cleavages in Burwash Formation metagreywackes at Duck Lake. The earlier cleavage is parallel to bedding and is reflected by the flattened mudstone intraclasts in the bed on the left. The later cleavage, representative of the regional north-northwesterly trending fabric throughout the region, is expressed here as quartz-filled fractures at an angle to bedding. Note the refraction of the cleavage in response to the varied grain size of the graded layers it intersects. GSC 157359*

The thickness of individual beds ranges from thick massive greywacke units in excess of 8 m to thin laminations on a millimetre scale. Detailed examination of the extremely thick sands, however, reveals that they are commonly composite as indicated by a sharp increase in grain size. They represent two or more depositional events in which the finer grained to pelitic parts of the first deposit was removed by erosion prior to deposition of subsequent deposits. The proportion of mudstone to sandstone and siltstone is also highly varied, but in general sand and silt greatly exceed mud. This is particularly evident in thick-bedded sequences, although in thin-bedded sequences the reverse is true.

The Bouma sequence of sedimentary structures that occurs in turbidity current deposits is common in the Burwash beds. The sequence consists of a graded division at the base of a bed followed by a parallel laminated division and then a ripple laminated division. The upper two divisions of the cycle are a second parallel laminated division and finally a pelitic division. This sequence of structures reflects a decreasing current regime at any given point as the bed accumulates. The five part cycle is ideal as in most cases one or more Bouma divisions are missing. The order of divisions, however, is always the same and any deviation from it is a likely indication that the bed represents the composite deposits of several turbidity currents.

Small scale scours and channels are common at the base of beds. They typically contain the coarsest grained material, in some cases significantly coarser than in the main part of the bed. These scours are commonly deformed under soft sediment conditions to form exaggerated load structures and flame structures. The bedding plane surfaces of these sediments are rarely exposed, due mainly to the effects of low grade metamorphism which has effectively welded the sands to the underlying mudstone. The morphology and orientation of bedding plane features in turbidite sequences

Figure 37. *Rare exposure of undeformed sole marks on base of greywacke turbidites at west-central part of Hearne Lake. Grooves, scours, and flutes formed due to current activity and are oriented parallel to the current. The sense of transport was from upper left to lower right. GSC 177709*

are commonly used as indicators of the direction of sediment transport. However in the rare instances where Burwash Formation bedding planes are exposed these structures are commonly reoriented parallel to local fold axes. In Figure 37, however, a rare example of a bedding plane surface with undeformed well developed and preserved sole marks can be seen. Major large scale channels that cut down through many underlying beds have not been recognized.

In addition to primary structures formed due to the accumulation of the sediment, there is an abundance of postdepositional structures. These include the previously mentioned load and flame structures at the base of beds, and various collapse features such as small scale ball and pillow structures where an originally continuous, usually rippled siltstone breaks up to form discrete nodules that in some cases sink down into the underlying, less dense sediment. Compaction of the sediment has resulted in various dewatering structures such as small dome-shaped features at the interface between the relatively porous sand and relatively impervious pelitic material in some of the grain flow deposits. Related to these are sandstone dykes and sills that form due to the expulsion of water from the coarser parts of the beds. This internal movement of water within the sequence after deposition, particularly in the thicker greywacke units, can result in the complete disruption of thin argillaceous interbeds due to excessive intrusion of sand. This results in nests of angular clasts of mudstone within the sandstone units that were never transported but were broken essentially in place.

Concretions are common in psammitic beds throughout the basin where they occur as scattered nodules to continuous layers. They tend to be more yellow weathering than the sandstone, have diffuse to sharp boundaries and are commonly compositionally zoned (Fig. 38). They are more calcareous in composition than the greywacke in which they are contained and now consist of a calc-silicate metamorphic mineral assemblage. The concretions formed soon after deposition as in some cases they are truncated by channels formed during the deposition of overlying beds.

Many of the primary features of these sediments are preserved despite the varied grade of metamorphism and degree of deformation. For example in Figure 39, delicate ripple crosslaminations are clearly evident in a metasiltstone that also displays coarse cordierite porphyroblasts.

Locally there are very thin units of tuff interbedded with the greywacke-mudstones. They are most commonly seen as angular clasts, some of which are up to several metres in length, but in a few localities occur as part of the normal sedimentary sequence. The tuff units consist of a few layers of light yellow weathering, extremely well graded, millimetre or less sized grains of dominantly felsic rock fragments, euhedral plagioclase and some quartz. The layers are generally less than 2 cm thick, never show any evidence of current activity, and grade abruptly into a very black mudstone that in general is thinner than the graded tuffs. The mudstone is more carbonaceous than the normal pelite and contains disseminated pyrite in some cases. The graded yellow tuff and black pelite couplets occur in groups of up to 7 or 8 layers and are rarely, if ever, separated by normal sediments. In one instance a thick sequence of mudstone with minor thin siltstone beds is followed by five couplets of tuff and black pelite. These are immediately followed by a series of very thick-bedded, moderately coarse grained, turbidites. This package would seem to be the product of a sequence of events that is possibly representative of the filling of the basin as a whole. A period of quiescence in this part of the basin, represented by the mudstone sequence, ended with a series of subaerial felsic volcanic eruptions expressed in the section by the water-laid tuff layers. Seismic shocks, possibly associated with and following the volcanism, may have initiated major turbidity currents that resulted in the deposition of the thick greywacke deposits immediately above the tuffs. The greywackes themselves may consist in part of epiclastic volcanic detritus derived from the same volcanic events that resulted in the tuffs. As similar tuff layers or blocks derived from them are found throughout the basin, it is evident that felsic volcanism was active during sedimentation.

Figure 38. *Zone of metacarbonate concretions in thick Burwash greywacke that have formed as discrete bodies to layers parallel to bedding. (a) These concretions have been transposed parallel to the regional cleavage (GSC 177716). (b) The concretions are compositionally zoned, which is reflected in the texture and calc-silicate mineralogy of the metamorphosed concretions (GSC 17718).*

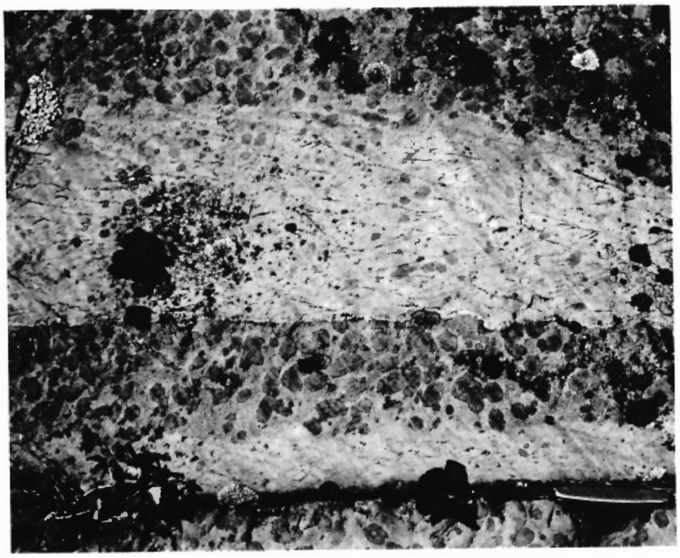


***Figure 39.** Coarse cordierite porphyroblasts in pelitic layers of metamorphosed Burwash Formation greywacke-mudstone from south of Prelude Lake. Note the undeformed climbing ripple crosslaminations in the more psammitic layers. GSC 178141*

Thick units of black mudstone are locally associated with the turbidites. They are particularly abundant in the west-central part of the basin between Prosperous Lake and Duncan Lake. The mudstones occur in units a metre or more thick and are composed entirely of fine grained argillaceous material. These units are typically massive with no sandy or silty layers. The black colour of the unit is due to a higher than normal carbon content. In some cases these carbon-rich mudstones contain disseminated pyrite. Where metamorphosed to amphibolite grade these units contain spectacular aggregates of very coarse pink chiastolite that are commonly associated with large quartz veins. These units are of interest as they represent rather long periods of quiescence in the basin that were not interrupted by the introduction of even fine silty material. Also, the carbon-rich nature of the sediment is suggestive of possible biological activity. Boyle (1961) has suggested that some of the black carbonaceous tuffs associated with the volcanics at Yellowknife may be of biogenic origin. Similarly, the carbon in these mudstones may also be due to the accumulation of the remains of single celled life forms. The occurrence locally of disseminated sulphides with these carbon-rich mudstones could be due to the activity of sulphate reducing organisms.

Petrography

The textural and chemical immaturity of the Burwash greywackes make them particularly valuable in providing an insight on the nature of the source terrane from which they were derived. This is of great value as the areas recognized as basement, let alone source area to the Archean supracrustal rocks, are small and few in number.

The coarsest grained rocks on the whole are the greywackes, with an average grain size of up to 1.5 mm, although coarser material over 2 mm in size occurs locally in scours at the base of beds. In rare instances, thick massive units have a maximum grain size of up to 1 cm. In a few cases scattered isolated cobbles up to 6 cm in size occur within the otherwise normal greywacke units. These coarse cobbles are rounded to subrounded and are mainly felsic volcanic in origin. Examples of these dispersed pebbles occur along the southwest shore of Upland Lake north of Jennejohn Lake and at Section G, Yellowknife Bay (Henderson, 1975b). From the maximum size (except for the dispersed pebbles) there is a complete gradation through silt to the presumably original clay grain size at any point in the rock. Grain size gradation through the bed is accomplished by an upward decrease in maximum grain size. There is a general uniformity of grain size distribution throughout the basin.

The greywackes consist of four main components, matrix, quartz, rock fragment and feldspar, that vary somewhat in their proportions but in general are present in this order of abundance (Fig. 40). The coarser grains are angular and tend not to be in contact with other coarser grains but are enveloped in a matrix of chlorite, muscovite and quartzofeldspathic material. In many cases the clasts, particularly the softer rock fragments, are oriented due to deformation.

Quartz is the most abundant clast component making up approximately one-quarter of the rock. In some cases distinctive volcanic crystal forms with characteristic embayments are present, but for the most part the provenance of the quartz is difficult to determine.

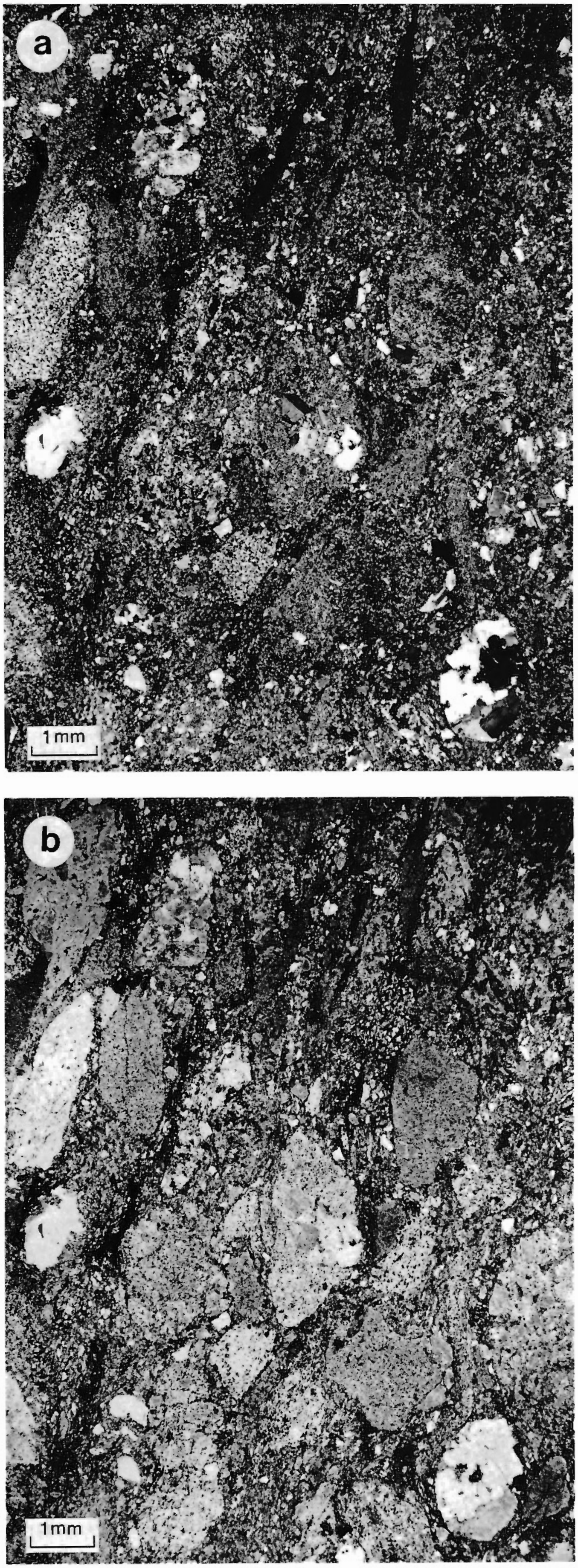
Rock fragments, making up slightly less than a quarter of the rock, tend to be rather diffuse in outline, are often gradational into the matrix, and are particularly susceptible to deformation. Many are of felsic volcanic derivation and in the coarser examples, phenocrysts of plagioclase are evident. More mafic varieties are present, but are of less abundance. Other rock fragments include recrystallized chert, fine grained argillaceous sediment of presumably intraformational origin and plutonic clasts. The granitic clasts are small in number, due no doubt to the relatively fine grain size of the sediment.

Plagioclase feldspar is the least important, volumetrically forming a little more than one-tenth of the rock. In most cases the plagioclase shows varied degrees of alteration that is a reflection of variation in original composition and its response to metamorphic equilibration as well as varied primary weathering of the source rocks.

The matrix is the most abundant component, making up over a third of the rock. It consists of a very fine grained mixture of chlorite, muscovite, biotite and quartzofeldspathic material that is derived from original argillaceous material in the sediment and from the diagenetic breakdown of coarser components, particularly the volcanic rock fragments that were presumably largely glass and hence susceptible to devitrification, and weathered feldspars.

In the finer grained sediments the proportion of matrix increases as the size of coarser components decreases. The pelagic mudstone, where best preserved, is a very fine aggregate of chlorite, muscovite, quartz and feldspar that has recrystallized from the originally highly argillaceous sediment.

The above description is of the lowest metamorphic grade Burwash sediments. As the metamorphic grade increases, the primary textures are increasingly obscured. The finer grained more argillaceous sediments are most susceptible to change. In most cases primary textures are retained through greenschist grade. Biotite becomes increasingly coarser grained and the outlines of rock fragments more obscure. Up to amphibolite grade metamorphic conditions, detrital outlines of quartz are preserved, but beyond lower amphibolite grade the rock is completely recrystallized into a homogeneous quartz-feldspar-biotite matrix with coarser porphyroblasts of cordierite, andalusite and garnet, among others. At the megascopic level, however, where there has not been



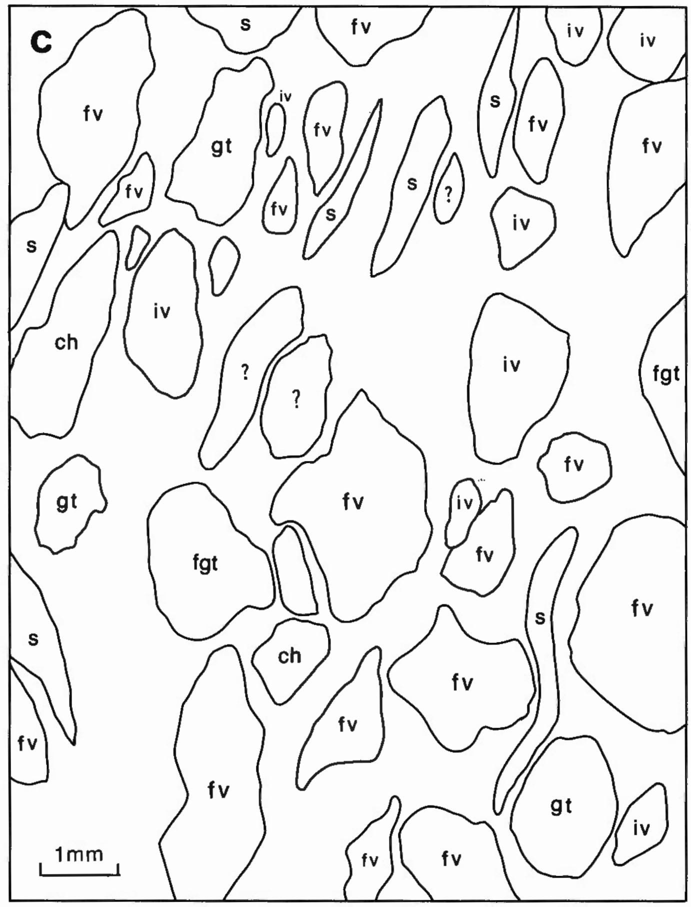


Figure 40. *A coarse grained phase of the Burwash Formation metagreywacke in which the lithic clasts are particularly well preserved because of their large size. The rock consists of volcanic lithic clasts of mainly felsic composition (fv) but also lesser amounts of more intermediate phases (iv) and relatively rare mafic volcanic clasts. Also present are granitoid clasts (gt), consisting of aggregates of polycrystalline quartz and plagioclase in some cases fine grained (fgt), a few sedimentary clasts of chert (ch), and intraformational metamudstone (s). The matrix consists of quartz, quartzofeldspathic material and biotite. The rock has been metamorphosed to upper greenschist grade. The source terrane contained both a mainly felsic volcanic and granitoid component. Most greywackes in the Burwash Formation are finer grained and this, together with the effects of diagenesis, metamorphism and deformation commonly make the identification of particular lithic clasts rather difficult. Nevertheless the same components as seen here are normally found in most greywackes of the formation. (a) Crossed polarizers. GSC 203660-B. (b) Plane light. GSC 203660-A.*

pervasive shearing, very delicate sedimentary structures such as ripple crosslamination in siltstones can be seen through an overprint of cordierite porphyroblasts (Fig. 39). On the other hand, in the more argillaceous beds, particularly where there has been some deformation, the original fabric of the rock can be completely obscured with even bedding difficult to see.

Source and depositional pattern

The textural and compositional characteristics of greywackes indicate their components have undergone a minimal degree of alteration and sedimentary differentiation since their erosion from the source area. Thus greywackes provide a good indication of the nature of the source terrane. From the composition and texture of the coarser components, it is evident that the material was derived from both a granitic and volcanic source terrane. The importance of the volcanic terrane is most easily appreciated from the make-up of the rock fragment component, consisting in large part of clasts of felsic volcanic parentage with lesser contributions from intermediate to mafic volcanic sources as well as an unknown proportion of reasonably fresh to highly altered feldspathic material from a weathered granitoid source. In addition, the fine grained quartz aggregates interpreted as recrystallized chert may represent siliceous sinter detritus associated with hydrothermal springs in the volcanic terrane. Fresh recognizable granitic detritus is generally present but in very small quantities. This is due largely to the grain size of the greywackes which, being generally less than 1.5 mm, makes it unlikely that granitic rock fragments will be preserved as such but will tend to have broken down into their component minerals. Thus much of the quartz and feldspar in the greywackes may be granitic detritus. In the coarsest grained material the proportion of granitic clasts is much higher. Some of the fine grained quartzofeldspathic rock fragments that look so similar to the felsic volcanic rock fragments (except for the lack of quartz and feldspar phenocrysts) may be altered feldspar derived from a weathered granitoid terrane. These clasts are like similar clasts in the Jackson Lake Formation, thought to be the alluvial facies equivalent of the Burwash greywackes (Henderson, 1970, 1975, this report). Because of the petrological similarity of these clasts to Archean weathered granite that occurs unconformably below Yellowknife supracrustal rocks elsewhere, and the similarity of the chemical pattern of the sediments as a whole to that of weathered granitic rocks, led Schau and Henderson (1983) to suggest a weathered granitoid provenance for some of the clasts. While it is clear both granitic and felsic volcanic source terranes were involved, it is difficult to estimate the proportional contribution of each.

The volcanic detritus is likely derived from volcanics contemporaneous with the Yellowknife supracrustal rocks. There are extensive felsic volcanic accumulations at several places within the basin, such as at Yellowknife, Clan Lake and the Detour Lake-Tumpline Lake- Turnback Lake areas, that are contemporaneous with sedimentation. The rare thin felsic tuffs interbedded with the sediments throughout the area are also indicative of contemporaneous felsic volcanism. The granitic source is more difficult to define. It may be in part also contemporaneous with the felsic volcanism in the form of subvolcanic plutons that were exposed and provided detritus for the basin. The presence of granitoid cobbles with epizonal textures in the Jackson Lake Formation, the alluvial facies equivalent of the Burwash, is an indication of such a source. The Amacher Granite, which is mantled by mafic volcanics, may be an example of such a subvolcanic pluton although there is no direct evidence that it was unroofed at that time. Felsic detritus derived from a pre-existing basement terrane is volumetrically perhaps a more important source. The rock units within the area that have been interpreted as probably older than the Yellowknife supracrustal rocks (Sleepy Dragon and Anton complexes) are commonly granitoid gneiss, or at least deformed granitoids, and range in composition from tonalite to granite. Of interest is the fact that no detrital metamorphic minerals have been recognized in the lower grade sediments. This might suggest that old, highly metamorphosed pelitic units in the basement gneisses were rare or nonexistent or that such minerals if present were not able to withstand the retrogressive effects of the low grade metamorphism.

The Burwash Formation is a major unit and in area greatly exceeds that of the preserved volcanic and other sedimentary units in the area. The source from which it was derived must have been a major sialic terrane. As it seems unlikely that the required high volume of material could have been derived by the erosion of essentially contemporaneous felsic differentiates in a dominantly mafic or "oceanic crust", the pre-existence of a substantial sialic source is indicated. Chemically the Burwash greywackes are not unlike greywackes of other ages (Table 8). Since greywackes are mineralogically, chemically and texturally immature sandstones, their chemistry represents a reasonably close approximation of the composition of their source terrane. This was clearly sialic in composition and most closely resembles an average composition of a granodiorite (Table 8). They also resemble Eade and Fahrig's (1971) estimate of the average composition of the surface of the presently exposed Canadian Shield. Thus the source terrane, whatever its textural nature, may not have been much different in composition from the presently exposed Shield.

Unfortunately, very little of the source terrane has been preserved or at least recognized to date. Indeed, due to structural complexities there is little information available as to transport direction of the sediments into the basin from the source terrane. At Yellowknife, at a preserved margin of the basin, paleocurrent directions have been determined that indicate that the source lay to the west, in the area now dominated by intrusions of the Defeat Plutonic Suite (Henderson 1972, 1975b). The original source terrane that provided the sediment to the basin has been lost due to erosion at the time of accumulation of the sediments and subsequent intrusion, uplift and further erosion of the terrane.

It has been suggested previously that the Burwash Formation turbidites at Yellowknife accumulated on a series of submarine fans that extended into the basins from its margin (Henderson, 1972, 1975b). Subsequent mapping of the Burwash throughout the map area would indicate that this model is appropriate throughout the basin.

Briefly, the model is based on the fact that the sedimentary sequences consist of background sedimentation of thin-bedded, fine grained turbidites that typically consist only of the upper divisions of the Bouma cycle, indicating deposition under low current velocity conditions. Cyclically superimposed on this pattern are groups of much thicker, coarser grained grainflow beds deposited under much higher current velocities. It was suggested that the sediments were introduced into the basin on a complex of submarine fans that extend out into the basin. The main part of any sediment-charged current would be contained within the depositional valleys on the fan surface, although the upper part of the turbidity current containing more highly dispersed finer grained material presumably extended above and beyond the confines of the active depositional valley. The passage of an individual turbidity current would result in a relatively thick, coarse grained deposit within the valley, as well as a much thinner finer grained deposit outside the valley on the adjacent fan surface that may also have extended into neighbouring, at that time inactive, fan valleys. Small slumps and slides on the sides of these valleys would result in

Table 8. Averages of chemical analyses pertaining to the Burwash Formation

	1	2	3	4	5	6	7	8
SiO_2	66.07	66.8	66.75	65.2	66.09	53.47	53.6	64.02
Al_2O_3	15.24	15.0	13.54	15.8	15.73	20.56	20.8	15.82
Fe_2O_3	0.70	1.0	1.60	1.2	1.38	1.25	1.4	7.10
FeO	4.52	4.6	3.54	3.4	2.73	7.10	7.4	Fe_2O_3
MgO	2.73	2.6	2.15	2.2	1.74	4.79	4.3	3.00
CaO	1.70	1.6	2.54	3.3	3.83	1.24	1.6	1.09
Na_2O	3.10	2.8	2.93	3.7	3.75	2.19	2.8	2.83
K_2O	1.91	2.0	1.99	3.23	2.73	3.51	3.2	1.92
H_2O^+	2.48	2.0	2.42	.8	.85	4.61	3.7	2.89
H_2O^-	0.08	n.d.	0.55		.19	0.11	n.d.	
CO_2	0.38	-	1.24	.20	.08	0.10	-	n.d.
TiO_2	0.64	0.6	0.63	.57	.54	0.93	0.8	0.70
P_2O_5	0.12	-	0.16	.17	.18	0.16	-	.08
MnO	0.06	0.1	0.12	.08	.08	0.09	.10	.06
Other	-	-	0.48	.11	-	-	-	
Total	99.73	99.1	100.61	100.0	99.90	10.11	99.7	99.51

1. Average of 3 Burwash greywacke composite samples, Yellowknife Bay (Henderson, 1975b)
2. Average of 5 Burwash greywackes, Prosperous Lake (Ramsay and Kamineni, 1977)
3. Average of 61 greywackes of all ages (Pettijohn, 1957)
4. Average composition of the Canadian Shield (Eade and Fahrig, 1971)
5. Average of 885 granodiorites (LeMaitre, 1976)
6. Average of 3 Burwash mudstone composite samples, Yellowknife Bay (Henderson, 1975b)
7. Average of 7 Burwash metapelites, Prosperous Lake (Ramsay and Kamineni, 1977)
8. Average of 6 Burwash greywacke-mudstones at Yellowknife (Jenner et al., 1981)

the initiation of small turbidity currents within the valley as well. Thus both the thin, fine grained beds and thick, coarse grained beds would collect within the depositional valley, while out on the fan surface only thin-bedded fine grained deposits would accumulate. Eventually the fan valley would become unstable by rising too high relative to the adjacent fan surface and subsequent turbidity currents would shift laterally to another part of the fan surface, form a new fan valley, and the fan complex would continue to grow.

It has not been possible to define individual fans in the basin or trace a given surface of a fan any distance into the basin. This is due in part to the scale of mapping, but also to the structural complexity, and lack of marker horizons and time lines within the formation.

There do not appear to be significant regional variations in the turbidites, reflecting different environments of deposition within the basin. For example, near the apparent margin of the basin at Yellowknife one might expect the turbidites to be more "proximal" in character, with thicker beds, coarser grain size and more abundant erosional channels than farther out in the basin. This does not appear to be the case. The problem may be one of scale. The segment of the basin preserved may only represent a small part of an originally much larger basin. One basin margin is recognized at Yellowknife and paleocurrent directions there indicate the turbidites were derived from the west (Henderson, 1972, 1975b). Remnants of other margins, or perhaps islands within the basin, are present in the map area. For example, detritus from the basement block represented by the Sleepy Dragon Complex was shed into the basin east of Ross Lake (Raquette Lake Formation) although there have been no paleocurrent studies in the Burwash Formation in this area to confirm the greywackes were also derived from the same region. Fans from different parts of the basin may have coalesced, resulting in a more uniform, homogeneous deposit than would be the case, for example, if sediment had been contributed only from one side of the basin. As well, the topography of the basin itself may have been very irregular and indeed could have been constantly changing during sedimentation due to minor block uplifts and faulting. This also would have contributed to very complex sediment distribution patterns, including anomalously thick-bedded sequences where the sediment was ponded well away from the upper parts of the fans.

Metagreywacke-mudstone – Russell Lake

Yellowknife supracrustal rocks occur at Russell Lake in the northwest corner of the map area. Along with a similar area of metasedimentary rocks at Stagg Lake to the east, these areas are the southernmost extension of a large, dominantly metasedimentary terrane that occurs to the north of the map area in the Snare River area (Lord, 1942). These Yellowknife supracrustal rocks can be traced continuously from Russell Lake through the Indin Lake map area (Lord, 1942) to the west end of Point Lake, a distance of over 300 km along the western margin of the Slave Structural Province (McGlynn, 1977).

The supracrustal rocks at Russell Lake, consisting of both sediments and volcanics, occupy a triangular-shaped area along 13 km of the northern border of the map area and extends 13 km south. The large, mainly felsic volcanic unit that occupies the east-central part of the area, and the smaller mafic volcanic unit to the east, have been described previously.

The Yellowknife rocks are surrounded by generally massive intrusive bodies of the Stagg and Awry plutonic suites and are intruded by at least three small plutons of Stagg granite and Defeat granodiorite. A small complex of felsic porphyry dykes and sills on the east side of the large bulbous peninsula in Russell Lake also intrudes the supracrustal rocks. The northwestern contact between the metasediments and the underlying large felsic volcanic unit is sharp, whereas the southeastern contact with the same volcanic unit is gradational over several hundred metres; the greywacke-mudstone becoming increasingly volcanogenic with a concomitant increase in the number of tuffaceous beds upwards towards the volcanic sequence. The contact between sediments and the mafic volcanics to the east is sharp.

The sediments are steeply dipping to overturned and throughout the area strike consistently in a north-northeasterly direction, but are isoclinally folded, as shown by the common reversal of top direction across the strike of the beds. They are everywhere metamorphosed to amphibolite grade with assemblages that include quartz, plagioclase, biotite, chlorite, muscovite, cordierite, andalusite and garnet. In many parts of the area there has been a pervasive retrograde event shown by the extensive breakdown of biotite and cordierite to chlorite.

These sediments are for the most part greywacke-mudstone turbidites similar to those of the Burwash Formation east of Yellowknife. The beds are laterally continuous and generally maintain a constant thickness ranging between a few millimetres and several metres, but averaging about 25 cm. In many cases there has been a background sedimentation of relatively thin bedded layers consisting of 3 to 5 cm thick siltstones to fine greywackes that grade up to 10-15 cm thick mudstone, with the occasional coarser grained greywacke bed up to 60 cm thick. Superimposed on this background are packages of thick-bedded massive greywackes up to several metres in thickness. These are also much coarser, with grains up to 2 or 3 mm in scour channels at the bases of the beds. In many cases these thick massive sands are amalgamated units representing two or three separate influxes of relatively coarse sediment such that any fine grained siltstone or pelagic mudstones that may have been present were completely eroded by the subsequent major turbidity current.

The sediments range from light to dark grey through grey brown to greenish grey, with the coarser grained sediment lighter in colour than the fine material. In many places the greywackes have a distinct greenish aspect. Beds of this nature occur in zones several tens of metres thick between greywackes of more normal appearance. In the field these sediments appear to have a greater mafic volcanic component than the normal coloured greywackes, but in thin section it is apparent that the colour is due to a postmetamorphic alteration with all the cordierite metacrysts and biotite in the rock having been retrograded to chlorite. In all other respects these rocks are similar to the normal greywackes.

Despite deformation and metamorphism up to amphibolite grade, primary sedimentary structures are commonly well preserved. Grading is the most common and is shown by the normal gradation of original detrital grains, particularly quartz, the "reverse" grain size gradation shown by metamorphic minerals which have grown preferentially and larger in the more argillaceous, finer grained sediment, the colour gradation, and the refraction of cleavage where the angle of intersection of cleavage with bedding increases towards the base of the bed in response to the changing grain size. Other sedimentary structures common in turbidite beds include parallel lamination, ripples and climbing ripples, and various sole marks such as channels, scours, crosslaminated scour fills and flame and load structures. The Bouma sequence of sedimentary structures (Bouma, 1962) reflecting a waning current regime can be seen in these beds although they are normally incomplete. Calcareous metaconcretions within the greywacke beds occur locally as zoned elliptical to irregular patches with calc-silicate mineralogy.

Despite the metamorphism of the sediments to amphibolite grade, primary textures in some cases are preserved in the coarser grained, less argillaceous material. Original detrital grains can be seen in these coarser rocks and even volcanic rock fragments are identifiable in some cases. As elsewhere, the major components are quartz, plagioclase and rock fragments in a now thoroughly recrystallized matrix. The quartz is angular, with corroded margins due to metamorphism and is always strained or, in the coarser grains, polygonized. The feldspar is recrystallized, resulting in a patchy appearance with little or no albite twinning, although in places the relict twinning pattern can be seen in the distribution of alteration products. The rock fragments are the most poorly preserved, although their outlines in some cases are clearly visible in plane light. They consist mainly of felsic volcanic rock fragments consisting of tightly interlocking equant networks of quartzofeldspathic material with relict plagioclase phenocrysts in some cases. Fine argillaceous clasts of probable intraformational origin and some recrystallized chert clasts are preserved as well. In the coarsest material many of the coarser quartz clasts have attached or included grains of plagioclase, indicating that they are granitic rock fragments. Thus much of the quartz and feldspar grains could also have been derived from a similar source. The matrix is recrystallized into a quartz-feldspar-biotite aggregate.

In the more argillaceous sediments the effects of metamorphism are more pronounced. Any primary detrital texture is lost due to recrystallization and the growth of coarse metacrysts commonly disrupts the fabric. The metacrysts include cordierite and to a lesser extent andalusite. Both are commonly highly poikiloblastic, are up to 2 cm long, and in many places are severely altered to chlorite and/or white micas. This retrogressive metamorphism is not pervasive or equally developed throughout the area, but as mentioned earlier, occurs in wide zones where at its most extreme all the biotite and coarse metacrysts are completely altered, leaving a quartz-feldspar-chlorite-white mica assemblage. In many cases the porphyroblasts have a wide margin of chlorite with a small fresh core of the original mineral. This retrograde event had some effect on rocks in most of the map area but nowhere to the extent seen in the Russell Lake area.

These sediments are very similar to the Burwash Formation east of Yellowknife, both in the mode and probably environment of deposition and in the nature of the terrane they were derived from. The abundance of quartz, plagioclase and silicic rock fragments indicates the sialic nature of their source whereas the abundance of felsic volcanic rock fragments and the presence of granitic rock fragments in the coarser grained, least recrystallized material indicates a mixed granitic plutonic-felsic volcanic provenance. This is the same type of terrane that is indicated at Yellowknife.

Iron formation

Unlike the Burwash Formation east of Yellowknife, the metagreywacke-mudstone at Russell Lake contains iron formation units. These occur along the east side of the main peninsula and on the three islands west of the cabin on the east side of the lake. Minor iron formation also occurs in the sediments east of the small peninsula on the east side of the lake southeast of the large island. The first two occurrences are more or less on strike with each other.

On the three small islands the iron formation is interbedded with the normal greywacke-mudstones and consists of units a few centimetres to several metres thick. The contact between the greywackes and the iron formation is generally sharp. The iron formation is of the silicate facies. Each unit is layered on the scale of a few millimetres to centimetres and layering is due to both compositional and textural variation. The layers are composed mainly of hornblende and members of the cummingtonite-grunerite series that vary from flat aligned lath-like crystals to radiating clusters up to 2 cm in diameter. Some layers contain abundant coarse garnets in poikiloblastic masses that have overgrown and partly replaced iron oxides and amphiboles, the outlines of which can still be seen within the garnet. There are also biotite-rich layers, light coloured siliceous layers and layers rich in magnetite and hematite. The various layers range from almost monomineralic to mixtures of two or more of the above phases. Minor thin sulphide layers (mainly pyrite) are also present. Finely disseminated pyrite also occurs in the other layers, particularly with the amphibole layers. Weathering of the sulphide has resulted in the rusty colour of the otherwise very dark green to black outcrop surface. The iron formation units are associated with generally thin-bedded silty turbidites. Associated with the iron formation units in a few exposures in this area are thin rippled siltstone turbidites, each of which grades up into silicate iron formation instead of mudstone. This suggests that the normal, presumably very slow background pelagic sedimentation in this part of the basin that consisted mainly of aluminous clays was replaced by the deposition of hydrous iron oxides, iron-bearing silica gels and colloidal particles of iron silicate which may have been the primary precipitate that eventually became a silicate iron formation (French, 1973).

Five kilometres to the south-southwest, more or less along strike from the three islands, are at least two units of carbonate that can be followed for about 3 km along the shoreline of Russell Lake. Silicate iron formation is associated with the carbonate at its southernmost exposure. The carbonate is dominantly an orange-brown weathering dolomite occurring as units about 1.5 m thick interbedded with the greywacke-mudstone turbidites. The carbonate units are made up of 5-10 cm thick layers that are defined by

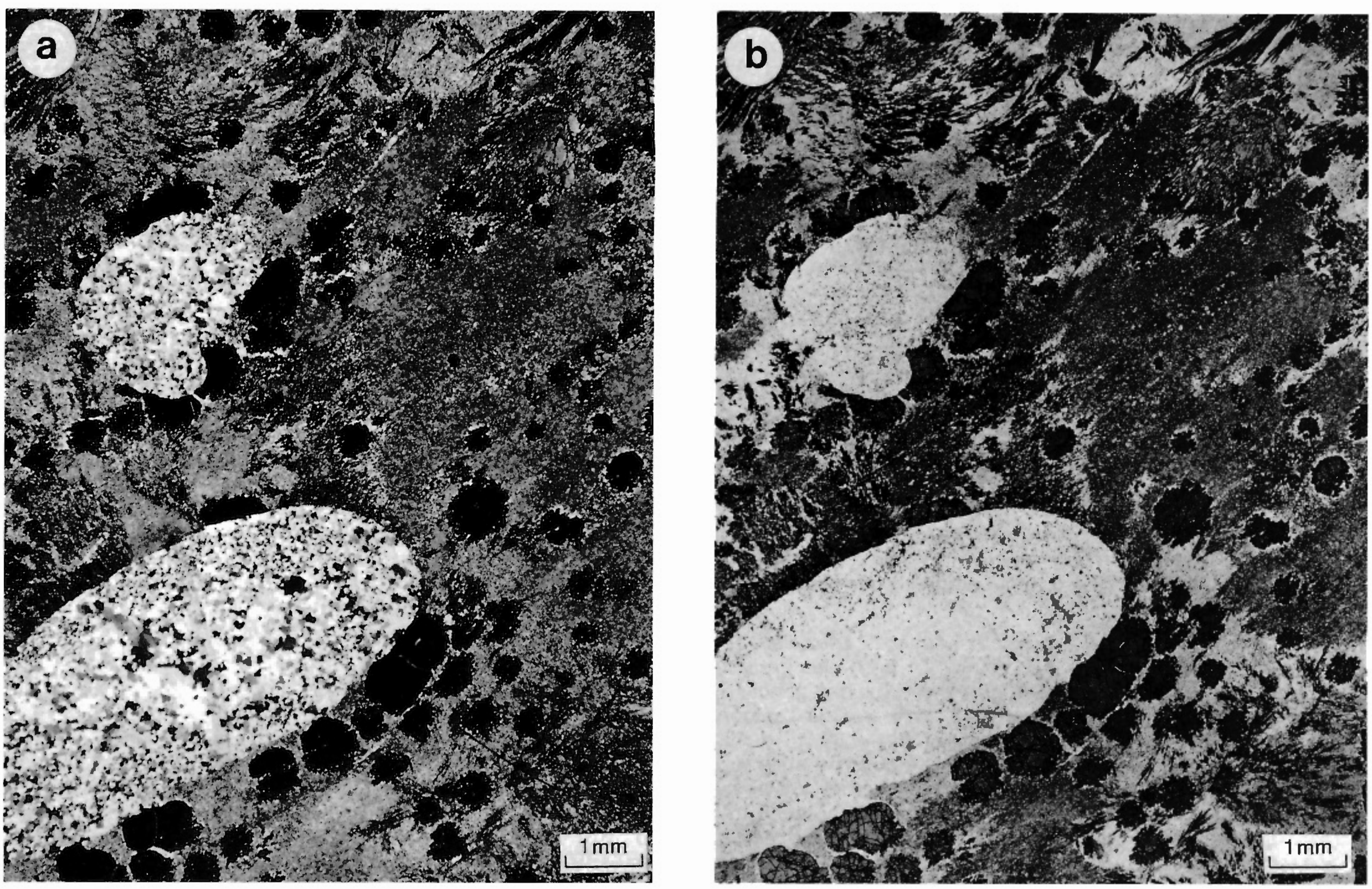


Figure 41. *Silicate iron formation at Russell Lake. The iron formation consists mainly of altered iron-rich amphiboles, garnet, quartz, plagioclase and very coarse polycrystalline quartz grains that represent recrystallized chert nodules. Where sulphides are present these iron formations can be anomalously high in gold. This silicate iron formation is metamorphosed to amphibolite facies. (a) Crossed polarizers. GSC 202790-Y. (b) Plane light. GSC 201859.*

siliceous laminae and minor textural and compositional variations. There is a certain amount of interbedding between the greywacke and the carbonate units, particularly at the margin of the units. Locally, the carbonate contains large bladed crystals of siderite up to 1 cm in length. At the northernmost exposure of the carbonate there is considerable pyrite mineralization. There is also a network of dykes, sills and irregular intrusions of a grey felsic feldspar porphyry in the area. The silicate iron formation that occurs locally with the carbonate is similar to that described to the north, but in this locality also contains ellipsoidal quartz nodules up to 2 cm long that are interpreted as original chert nodules (Fig. 41).

Although these occurrences of iron formation cannot be traced continuously they do lie approximately on strike with each other. To the north beyond the map area, Lord (1951, p. 72) described a similar silicate iron formation about 3 km north of the north end of Russell Lake that is also approximately along strike with the occurrences at Russell Lake. If these iron formations are related and occur along the same stratigraphic horizon, they represent a marker horizon up to 30 km long, which is a rare occurrence in these sedimentary basins. The variation in composition of the iron formation from dominantly silicate iron formation in the north to mixed silicate and carbonate in the south is similar to the variation in facies seen in the iron formation on the east side of the major felsic volcanic complex at the headwaters of the Back River in the east-central part of the Slave Province. There the iron formation at the contact between the volcanics and the normal greywacke-mudstones includes four facies of iron formation over a strike length of at least 15 km (Henderson, 1975a). The silicate iron formations at Russell Lake are similar in many respects to those found in similar sediments between Contwoyto Lake and Point Lake 350 km to the north that have been described by Bostock (1968, 1977, 1980) and Tremblay (1976). In the Contwoyto Lake area these iron formations are of considerable economic interest for their gold content. Similarly, the iron formation north of Russell Lake contains a gold prospect (Lord, 1942, p. 72) and, in addition, the sulphide-rich parts of the iron formation on the islands in Russell Lake are anomalously high in gold (up to 235 ppb).

Metagreywacke-mudstone – Stagg Lake

An arm of metasedimentary rocks extends south through Stagg Lake from the large area of metasediments that lie to the north in the Snare River map area (Lord, 1942). The northerly trending belt extends 13 km south from the north boundary of the map area and is about 5 km wide. The belt is composed entirely of sediments that have been metamorphosed to amphibolite and upper amphibolite grade.

This belt of Yellowknife metasediments transects the regional structural grain of the country. Both the granitic units and the supracrustal rocks elsewhere have a generally northeasterly trend whereas the Stagg Lake belt has a more northerly trend that transects the Stagg Plutonic Suite that engulfs it at a moderate angle. Within the belt the structure of the sediments is parallel to the trend of the belt, with the beds dipping moderately to steeply. Because of the generally higher grade of metamorphism and greater degree of recrystallization, top direction indicators are not abundant. Such data that are available suggest the sequence is east-facing but there is insufficient data to preclude the possibility that the sediments are isoclinally folded, as is so commonly the case elsewhere. A pronounced lineation plunging 10-30° to the north, formed by oriented metacrysts, is common in the belt. Lineations with this orientation occur elsewhere in the map area, in particular in the vicinity of Duncan Lake.

The metasedimentary belt is everywhere intruded by the surrounding granitic rocks. Granites of the Stagg Plutonic Suite bound the east and west sides of the belt while the Awry Plutonic Suite cuts off the southern portion of the belt. Several typically foliated stocks and small plutons of Defeat granodiorite occur within the belt and a larger body of more massive Defeat granodiorite occurs at the southwestern margin. In the contact area the Stagg granite normally contains only scattered inclusions of sedimentary schists, although at the northernmost part of the eastern contact there is a zone of large blocks to smaller inclusions of metasedimentary schist within the Stagg granite. Larger zones of metasedimentary material occur in the granite terrane east and northwest of the main belt and in addition, small zones of schist occur in the Stagg granite north of the central part of the east-west segment of the Stagg River. In the metasedimentary terrane as the contact is approached, there is typically an increasing amount of intrusive material, particularly on the western side. Remote from the contact, pegmatites are prevalent and become thicker and more abundant closer to the granite. Before the contact is reached, the pegmatites are replaced by increasing amounts of dykes, sills and irregular bodies of the granite. The contact with the Awry Plutonic Suite to the south is also sharp but inclusions of the metasedimentary schist in the granite are much more abundant. A large area of metasedimentary inclusions occurs 4 km south of the end of the belt. The abundance of metasedimentary schist inclusions remote from the schist belt would suggest the large terrane of Yellowknife supracrustal rocks, most of which is north of the map area, was more extensive than indicated by the distribution of the present-day belts.

The sediments within the belt are metamorphosed to amphibolite and upper amphibolite grade with the loss of most primary sedimentary features. What remains is alternating layers of pelitic and semipelitic to psammitic schists. The more psammitic beds retain relict primary structures best, and in a few places some grading or cross-laminated basal scours are preserved. Where best preserved, these sediments are similar to the Yellowknife sediments that occur elsewhere in the region and are interpreted as originally being an alternating greywacke-mudstone turbidite succession. The schists are various tones of grey through brown to black and consists of alternating lighter coloured psammitic layers 8 to 60 cm thick and darker pelitic layers 5 to 10 cm thick, commonly with thin quartz to pegmatitic segregations and veins, indicating incipient melting, that are commonly ptygmatically folded. Coarser pegmatites and granitic bodies up to 3 m or more in width are common, particularly close to the contacts. In places lighter weathering yellow-brown layers between 5 and 15 cm thick with more amphibole than biotite occur and are interpreted as intermediate to mafic volcaniclastic layers within the original greywacke-mudstone turbidite sequence. The metamorphic foliation, though present, is commonly weak with the dominant planar feature being compositional layering.

The sediments are recrystallized with the complete loss of any relict detrital outlines of the original grains. This has resulted in a more or less granoblastic assemblage of generally monocrystalline quartz, unaltered plagioclase and biotite with coarser porphyroblasts of cordierite and andalusite. Garnet is also more common in these rocks than elsewhere. Locally, needles of sillimanite are associated with the biotite and in places near the contact with the granitic rocks on the central west side, coarse sillimanite is pseudomorphous after andalusite porphyroblasts. Retrograde alteration has severely altered the cordierite to white mica and chlorite and the biotite to chlorite.

Evolution of the Yellowknife Supergroup

The model considered to best portray the evolution of the Yellowknife Supergroup is the rift basin model previously used by McGlynn and Henderson (1970), Lambert (1977) and Henderson (1981). According to this model the supracrustal terrane within the Slave Province represents the remnants of depositional basins (Fig. 42). In the Yellowknife – Hearne Lake area, the major mafic volcanic sequences such as the Kam Formation at Yellowknife, the Cameron River Formation southeast of Gordon Lake and the Sunset Lake Basalt in the northeastern part of the area, among others, occur at the margins of the supracrustal terrane. Although much of this terrane is bounded by intrusive granitoid rocks, what are thought to be basement rocks older than the Yellowknife locally occur immediately adjacent to the Yellowknife rocks. The Anton Complex north of Yellowknife is one example and in the northeast the Sleepy Dragon Complex is a basement block that is almost completely mantled by Yellowknife volcanic rocks. It has been suggested that the volcanic rimmed margins of the supracrustal terranes are a close approximation to the margins of the original basin. Walker (1978) has warned that the definition of Archean basin margins is highly subjective and consideration should be given to the possible time transgressive nature of such boundaries and the difficulties in temporally correlating basin margin facies in different parts of the basin in question. Nevertheless, for the purpose of this discussion the margins of the Archean basin east of Yellowknife are considered to be the transition from subaerial or shallow water to deep water deposition. This may be considered a somewhat unrealistic definition as major shallow water deposits may have accumulated beyond the area occupied by the deep water sediments and volcanics. To date however such deposits have not been recognized; the only shallow water sediments known occur only in close association with some of the major mafic volcanic sequences.

In the basin east of Yellowknife two segments of the margin, one at Yellowknife and one in the vacinity of Upper Ross Lake, have been relatively well preserved. The best preserved example of a Yellowknife basin margin complex, however, occurs not within the Yellowknife – Hearne Lake area but at Point Lake, about 300 km north of Yellowknife (Henderson and Easton, 1977; Henderson, 1977). There the same supracrustal facies that are seen in the Yellowknife – Hearne Lake area occur, but the relationships between them and the basement rocks to them are much more clearly displayed. At Point Lake a major linear belt of mafic volcanics occurs at the edge of an extensive supracrustal terrane adjacent to a gneissic terrane that is considered to be older than the volcanics (Henderson and Easton, 1977; Bostock, 1980; Easton et al., 1982). The volcanics unconformably overlie a basement granite that has an age of 3.15 Ga (Krough and Gibbins, 1978; Henderson et al., 1982). As the unconformity is followed out into the basin, the volcanics can be seen to thin until they disappear, at which point the basement granite is overlain by greywacke – mudstone turbidites similar to those that overlie the volcanics nearer the basin margin and essentially identical to the Burwash Formation east of Yellowknife. Locally overlying, but in some cases also interbedded with, the underlying volcanics are granitoid and volcanic clast bearing conglomerates and sandstones analogous with to those of the Jackson Lake and Raquette Lake formations east of Yellowknife. The various basement gneissic lithologies that occur outside the basin as well as the basement granite that occurs within the basin have been recognized within the conglomerates (Easton et al., 1982). Similar conglomerates also occur above the basement granite out in the basin beyond the extent of the volcanics but there contain no volcanic clasts.

This complex of Yellowknife volcanic and sedimentary rocks has been interpreted as representing the margin of a fault bounded basin (Henderson, 1981). Mafic volcanism was concentrated along the faulted margin of the basin and thinned out into the basin away from the fracture zone. The uplifted rim of the evolving basin shed granite gneiss and granitoid detritus as well as volcanic detritus from earlier stages of the volcanism into the basin. Erosion of the weathered gneissic and granitoid basement rocks (Schau and Henderson, 1983) as well as contemporaneous felsic volcanic centres in the region contributed to the main greywacke-mudstone fill of the deepening basin.

So complete and well exposed a marginal complex does not occur in the Yellowknife – Hearne Lake area but various aspects of the Point Lake situation can be seen at the two localities previously mentioned. The Sleepy Dragon Complex represents an uplifted basement block (Lambert, 1977)

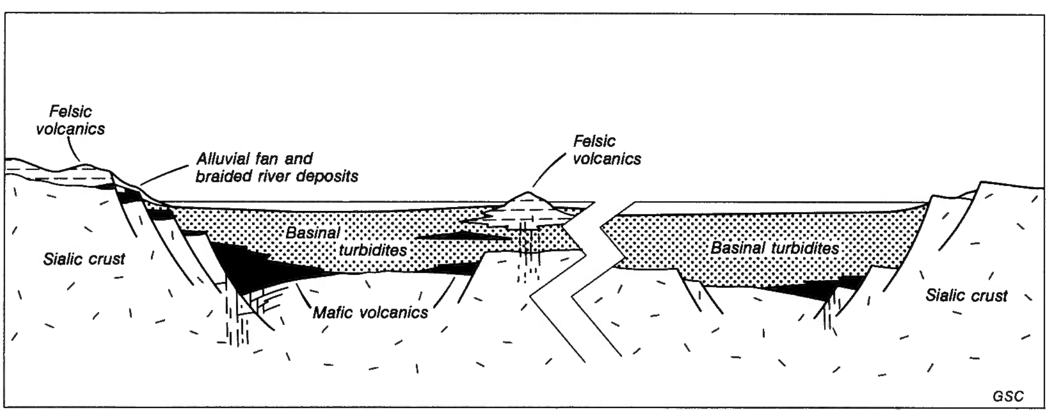


***Figure 42.** Cross-section through a fault-bounded Archean basin such as that partially preserved in the Yellowknife-Hearne Lake area. Greywacke-mudstone turbidites, the Burwash Formation, derived both from the basement and felsic volcanics on the basin margin are the main basinal fill. Dominantly mafic volcanics such as the Kam, Cameron River and Tumpline formations, are restricted to basin-marginal fault zones or fault blocks within the main basin. Felsic volcanics such as the centre at Clan Lake, the Banting Formation, and parts of the Tumpline Formation, occur both within the basin, where they tend to be preserved, and outside where they tend not to be but were a major contributor to the sediments that filled the basin. Fluvial sediments such as the Jackson Lake Formation and Raquette Lake Formation occur only at the margins of the basins (from Henderson, 1981).*

analogous to the basement rocks at Point Lake. There, largely mafic volcanic sequences and, in the south, a mafic sequence with several associated felsic centres, discontinuously mantle the basement block. In the vicinity of Upper Ross Lake the major sequence of mafic volcanics (Cameron River, Webb Lake and Dome Lake formations) developed along the northeast-trending basin margin, thins and ends along the southeast-trending margin at Upper Ross Lake. There conglomerates derived from the basement and locally from the volcanics, and quartzites and quartz-rich sandstones derived from the weathered granitoid and gneissic basement rocks (represented by the Raquette Lake Formation) occur above the basement and interfinger with the thinning volcanic sequence.

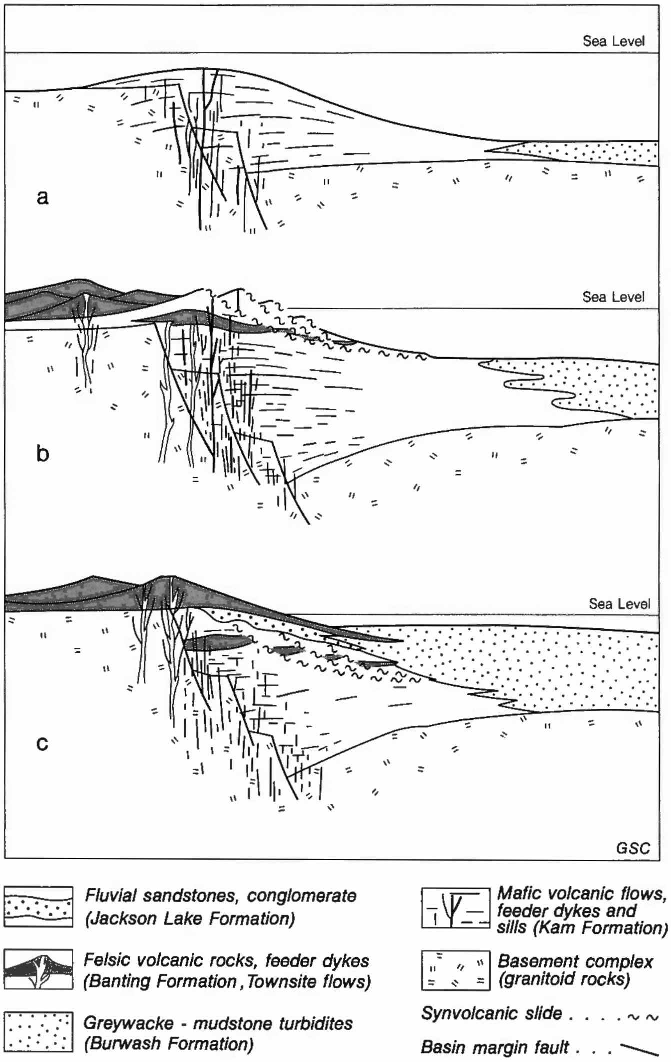


Figure 43. *Evolution of the Kam volcanic sequence began with (a) fracturing of the sialic crust and the intrusion of mafic dykes and extrusion of flows of mafic volcanics. This was related to an extensional event that resulted in the formation of the major sedimentary basin to the east as a down-dropped block of a graben and horst system. Volcanics continued to accumulate subaqueously as the developing basin continued to deepen. The distal part of the volcanic complex interfingered with the sediments accumulating in the basin. At one point (b) the upper part of the volcanic edifice tilted and rose, perhaps as a result of earlier fault blocks becoming sealed to the rising rim of the basin. As a response to the gravitational instability of the uplifted sequence, local slumping of the volcanic edifice toward the basin took place and where above sea level, part of the volcanic pile was eroded. The felsic volcanism (c), mainly remote from the Kam volcanics, contributed to the erosional detritus that together with weathered detritus derived from the basement granitoid rocks, formed the fluvial sandstones of the Jackson Lake Formation above the Kam, as well as the basinal Burwash sediments, and are themselves represented in the basin as the Banting Formation felsic volcanics.*

At Yellowknife, the basement rocks are not preserved but a major, linear, thick belt of volcanics, the Kam and Duck formations, occurs at the basin margin. On the basis of gravity modeling it would appear that there is little evidence that the volcanics exposed at the present erosion level occur out into the basin underneath the sediments much beyond where they are seen to locally outcrop between Yellowknife Bay and Pregg Lake (Fig. 34; McGrath et al., 1983) – a situation similar to that seen at Point Lake where the thick mafic volcanic sequence thins out into the basin.

In the earlier section it was suggested that the Kam Formation evolved as a series of southerly prograding volcanic centres (Fig. 14) which, according to the fault basin model, may have been controlled by a zone of boundary faults along the basin margin (Fig. 43). At some point the upper part of the volcanic edifice tilted and rose above sea level, possibly as a result of earlier fault blocks becoming sealed to the rising rim of the basin, producing an unstable situation that ultimately resulted in the collapse and sliding of part of the volcanic pile towards the basin. The erosion of the upper part of the sequence and the subsequent deposition of the shallow water Jackson Lake conglomerate and sandstone followed. Felsic volcanism also presumably related to basinal rifting took place both within the basin (ie. the felsic volcanic centre at Clan Lake), at the basin margin (Townsite flows and Banting Formation) and perhaps beyond the margin. Paleocurrent data from the Burwash Formation greywacke – mudstone turbidites in the basin just east of Yellowknife indicate that the sediments were derived from the west, from the nearby basin margin, where felsic volcanic centres outside the basin and the rising rim of the basin itself provided a source of felsic volcanics and granitoid detritus that filled the basin in a series of coalescing submarine fans.

Considering the basin as a whole, both that part within the Yellowknife – Hearne Lake area and that part to the north as shown in Figure 44, the outline of the basin as defined by the distribution of the volcanic rocks and basement block remnants is quite apparent. If the extension to the Kam Formation mafic volcanics under Great Slave Lake as proposed by Gibb and Thomas (1980) is included, the basin is more complete and the regular zig-zag pattern of the basin margin is more obvious. The northeasterly trends are shown by the Kam, Cameron River and parts of the Tumpline and Sunset Lake formations; the northwesterly trends by the extension of the Kam Formation, the Payne Lake Formation and parts of the Tumpline and Sunset Lake formations, and in the northern part of the basin the general northerly trend of the volcanic units is approximately parallel to the trend of the basin as a whole. This zig-zag pattern of the basin margin is a characteristic feature of many rift valleys (Reches, 1978) and the generally rhomboid pattern of faults in such structures is evident in such examples as the Rhine Graben (Illies, 1981), the Rio Grande Rift of the Basin and Range Province (Kelley, 1978), and the Kenya Rift (Griffiths, 1980) and, somewhat closer to home, the Ottawa Valley rift. There would appear to be little consensus as to the reason for such a pattern, with explanations ranging from the various tends having different ages, to changes in the direction of stress with time, to control by older structures, to the suggestion that this pattern is what might be expected of a terrane undergoing three-dimensional strain (Reches, 1978).

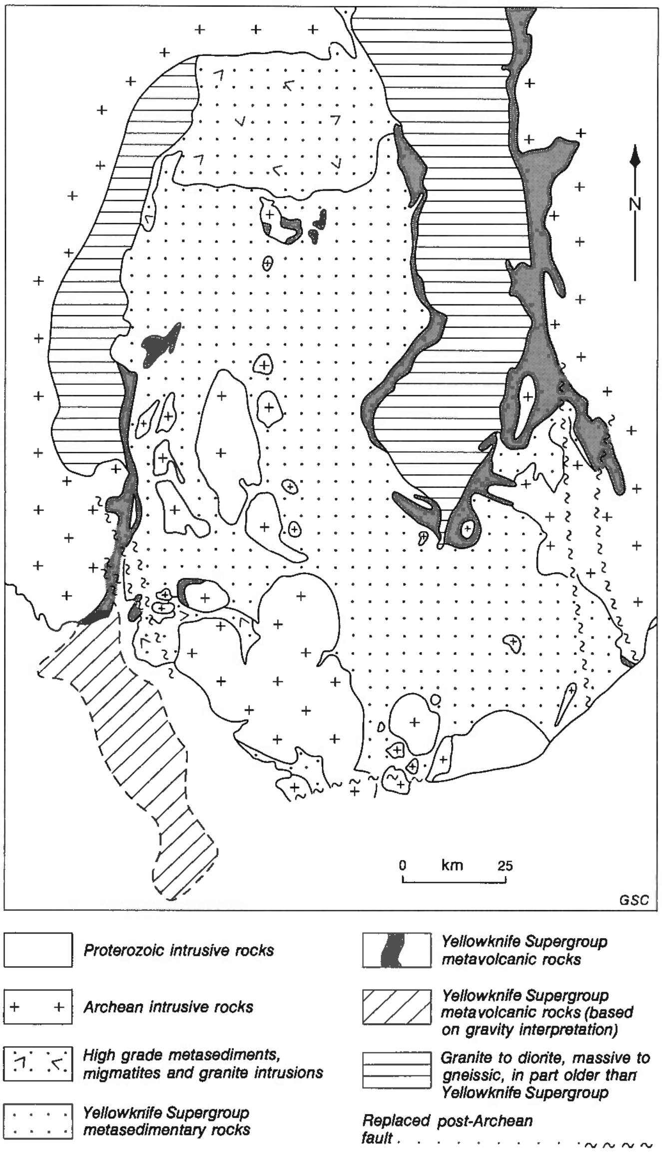


Figure 44. *Archean basin segment within and north of the Yellowknife-Hearne Lake area. The exposed basement blocks are bounded in part by dominantly mafic volcanics and the central part of the basin is filled with greywacke-mudstone turbidites. Southeast of the volcanic sequence at Yellowknife underneath Great Slave Lake is a prominent Bouguer gravity anomaly that has been interpreted by Gibb and Thomas (1980) as being due to the extension of the volcanic sequence at Yellowknife to the southwest. The volcanic sequences have three dominant trends – northerly, northwesterly and northeasterly (the late faults through the segment at Yellowknife have been replaced). It is proposed that the basin formed due to the extension with the development of rift valleys or grabens. Such valleys characteristically have zig-zag-trending, fault-bounded margins. Volcanism was concentrated at the basin margins.*

The conventional interpretation of the origin of fault bounded rift basins is that they are due to extension of the crust more or less perpendicular to the trend of the structure (Hobbs et al., 1976). McKenzie (1978) has proposed an extensional model for the formation of sedimentary basins that may have some application to the basin at Yellowknife. According to this model continental lithosphere is rapidly stretched, which results in rifting and block faulting of the crust and, the passive upwelling of hot asthenosphere below the thinned lithosphere. A certain degree of subsidence takes place as a result of the crustal extension but further slower subsidence takes place as the upswelled asthenospheric material cools, becomes more dense, and sinks. The final result is an isostatically compensated deep basin that, not being particularly subject to uplift and erosion, has a good chance of being preserved. The first part of this model is particularly appealing as, in addition to a mechanism for the formation of the rift basin itself, it also provides the thermal anomaly that may help account for the extensive mafic and felsic volcanism that was associated with the formation of the basin and, in addition, some of the early granitoid plutonism that is possibly related to the volcanism (see sections on Amacher Granite and Defeat Plutonic Suite). Bickle and Erikson (1982) have suggested that the main basinal fill at Yellowknife, the greywacke – mudstone turbidites, are a consequence of the second stage of the lithosphere stretching model, the subsidence that follows as a result of the cooling of the upwelled asthenosphere below the lithosphere. However, as the sediments appear to be more or less contemporaneous with the volcanics which they feel are a result of the first stage (locally they are seen to interfinger, and there is good evidence of volcanism taking place as the sediments accumulated), it seems more likely that the greywackes were also related to the extensional stage. If the turbidites are assigned to the extension phase of the model nothing is left to represent the thermal decay stage although it is possible that any such sequence that may have existed as a result of the decay of the thermal anomaly may not have been preserved.

An alternative and, in the case of the Yellowknife basin, preferred method for the formation of grabens is the wedge subsidence mechanism originally proposed by Vening Meinesz (1950) and in a modified form by Bott (1976) and Bott and Mithen (1983). By this mechanism if the crust is under tension, a block between normal fault zones within the brittle upper crust will sink while the bordering terrane rises at the rim of the developing graben. The ductile lower crustal material flows from beneath the down dropped block to beneath the adjacent rising rim area. Bott (1976) calculated that grabens formed in a 10 km thick brittle crust by this mechanism should be between 30 and 60 km wide with subsidence on the order of 5 km. This is more or less the scale of the basin east of Yellowknife assuming the effects of later intrusion and deformation of the supracrustal rocks more or less cancel each other.

As suggested by Bott and Mithen (1983) this mechanism has certain advantages over the lithosphere stretching model. Significantly less extension is required for a given amount of subsidence and subsidence only takes place during active extension of the crust. In the lithosphere stretching model, subsidence continues after rifting due to cooling and sinking of the previously upwelled asthenosphere and can continue to do so for 50 Ma. As previously discussed, the Yellowknife supracrustal rocks probably accumulated in a few million as opposed to tens of millions of years. Thus there is no recognized depositional record for most of this time if the lithosphere stretching model was the mechanism by which the Yellowknife basin formed.

ARCHEAN PLUTONIC ROCKS – SYN- AND POST-YELLOWKNIFE SUPERGROUP PLUTONIC ROCKS

About half the Precambrian terrane in the map area consists of Archean granitoid rocks. Twelve units or plutonic suites have been defined on the basis of their textural and mineralogical characteristics and structural relationships. The various units are either significantly metamorphosed or relatively unaltered. Within these groupings the units are described more or less in order of decreasing concordancy with the adjacent rock units and decreasing age if known.

The oldest granitoid units are the previously discussed Sleepy Dragon and Anton complexes east and west of the main Yellowknife supracrustal basin in the northern part of the area respectively, and are considered to be older than the Yellowknife supracrustal rocks. Two small plutons, the Amacher Granite to the east of the Sleepy Dragon Complex and Detour Granodiorite to the south of the complex are also metamorphosed. The Amacher Granite is considered to be contemporaneous with Yellowknife volcanism and may be a subvolcanic pluton.

The unmetamorphosed Defeat Plutonic Suite, the most extensive plutonic unit in the area is most abundant northwest and southeast of Yellowknife but occurs throughout the area. These plutons are more or less concordant with the supracrustal rocks. The Stagg Plutonic Suite in the northwest of the area varies in composition from tonalite to granite with increasing development of coarse microcline porphyroblasts. It is suggested that this unit may represent a metasomatized and remobilized equivalent of the basement complex. The Awry Plutonic Suite in the northwest and Meander Lake Plutonic Suite along the east border of the area are compositionally heterogeneous units that intrude the basement complex and the supracrustal rocks respectively. Contact relations with the Yellowknife rocks vary from concordant to discordant. The Prosperous Granite in the central part of the area is a two mica granite that commonly has abundant associated pegmatite. The Prosperous Granite bodies are typically sharply discordant to the Yellowknife sediments they intrude. The lithologically similar Redout Granite in the Redout Lake area intrudes the basement complex. The Redout Granite contacts, unlike those of the Prosperous Lake, are migmatitic, suggesting a deeper level of emplacement. The Morose Granite in the northeast and at Hearne Channel, and the Duckfish Granite north of Yellowknife are among the youngest of the Archean granites in the area.

Modal analysis data for most of the granitoid units are summarized in Figure 45.

Geochronology

Geochronological data are available for some of the granites and in general support the temporal ordering of the units based on geological relationships. All geochronological data cited in this report has been recalculated where necessary using the presently accepted decay constants proposed by Steiger and Jaeger (1977) for the potassium-argon, rubidium-strontium and uranium-lead systems.

Minimum zircon ages for the basement granites range between 2585 and 2630 Ma while for the Defeat granitoids, the range of ages from a variety of methods (excluding K-Ar) is between 2535 and 2650 Ma. The geologically younger Prosperous Granite has a Rb-Sr mineral isochron at 2520 Ma. K-Ar data do not appear to have a consistent pattern relative to the apparent geological age of the units (Fig. 46). Twenty-nine biotite dates from 8 granite units range between 1970 and 2595 Ma but for the most part occur in two groups with 4 determinations within 30 Ma of 2200 Ma and 21 within 100 Ma of 2420 Ma. Between the two there is a gap of 100 Ma where there are no dates. Eleven muscovite K-Ar dates from the Prosperous, Defeat, Stagg and Redout granites all occur within 50 Ma of 2515 Ma. Six hornblende K-Ar dates from the Anton Complex and the Defeat, Stagg

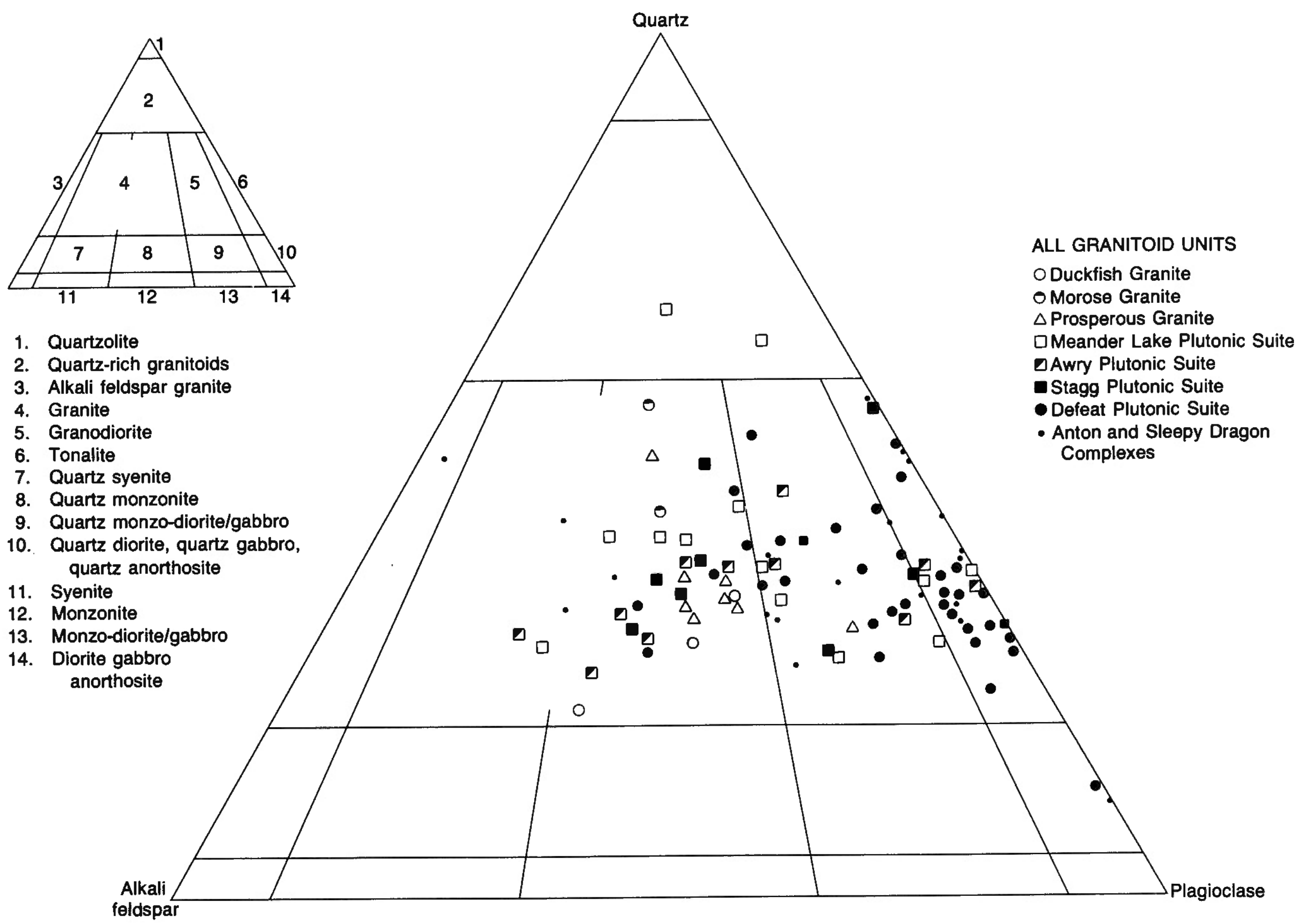


Figure 45. *Plot of all modal analyses of 10 Archean granitoid units in the Yellowknife-Hearne Lake area. The granite nomenclature system is that of Streckeisen (1976).*

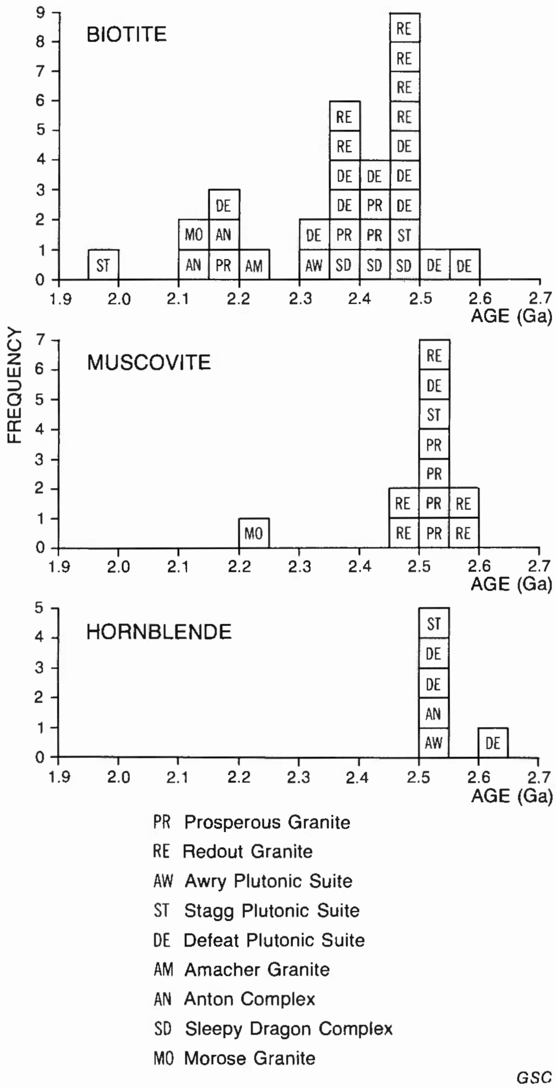


Figure 46. *Histograms of biotite, muscovite and hornblende K-Ar ages for various granitic units in the Yellowknife-Hearne Lake area. These ages reflect the post-intrusion thermal history of the area that is not related to any particular intrusive body. Data from Green and Baadsgaard (1971), Wanless (1970), Wanless (personal communication, 1978), Burwash and Baadsgaard (1962), Stockwell (1962, 1963), Green et al. (1968), and Davidson (1979).*

and Awry plutonic suites fall between 2500 and 2610 Ma and have an average age of 2525 Ma. The K-Ar data are interpreted as reflecting the age of uplift of the region with the present erosion surface passing through the muscovite and hornblende blocking temperature about 2500 Ma ago. The smaller grouping of biotite data suggests a disturbance of the K-Ar system at about 2200 Ma and could be related to a thermal event associated with the emplacement of the Blachford Lake Intrusive Suite that has been dated by both K-Ar and Rb-Sr methods at 2150 Ma (Davidson, 1982; Wanless et al., 1979).

Metamorphosed felsic plutonic rocks

Amacher Granite

The Amacher Granite is located in the northeast part of the map area just west of Sunset Lake and southeast of Amacher Lake. It is a single northerly trending pluton 14 km long and about 3 km wide, and is surrounded by Sunset Lake Basalts.

Contact relations

The granite is generally concordant with the volcanics it intrudes, as it is situated in an anticlinal core of the folded volcanic sequence. In detail the granite is locally discordant, particularly at the northern end. Minor inclusions of mafic material, presumed to be derived from the mafic volcanics, are found in places near the contact. Locally, a few small dykes and veins of the granite intrude the volcanics. Along the northeast margin of the granite Henderson (1939, p. 8) reported the occurrence of a "peculiar fragmental phase of the granite". This clastic unit was described as being 60 cm thick and gradational with the granite. It is composed of rounded to subangular granite clasts in a coarse greenish matrix of granitic detritus. In many instances he found it difficult to differentiate between massive granite and the fragmental unit. As granitic dykes intrude the volcanics in the area, Henderson (1939) suggested the fragmental unit was a brecciated zone rather than a sedimentary conglomerate, although he considered the possibility of more than one age of granite being present.

Lithology

The granite is, for the most part, massive and homogeneous in outcrop. It is a pale pink to light pinkish-grey, medium- to medium coarse-grained, equigranular rock, but typically fractured on a small scale. These fractured surfaces are commonly coated with chlorite or biotite. Larger fracture zones are easily recognized in outcrop by their bleached aspect. Quartz is particularly evident, both from its abundance and relatively coarse grain size and is typically irridescent. At some localities, in particular at the southeast margin of the body, there is a finer grained zone about a metre wide at the contact. No pegmatitic phases were seen.

The granite is mineralogically and texturally varied, ranging from hypidiomorphic-granular aggregates of quartz, plagioclase, microcline and biotite/chlorite to rocks with spectacular porphyritic and intergrowth textures that include perthite and myrmekite (Fig. 47; Table 9). Because of the textural and mineralogical variability, the compositional range is difficult to estimate, although on the basis of a fairly constant mafic mineral content, it may not be excessive.

Plagioclase occurs in euhedral to anhedral grains with rather dull appearing albite twins. The feldspar is only weakly zoned although the distribution of alteration products would suggest the present composition is not primary and that the original crystals were strongly zoned. Indeed, some alteration patterns outline crystal forms that extend across some of the plagioclase aggregates. This suggests the original plagioclase in at least some cases is completely recrystallized. The alteration products, consisting mainly of muscovite, are commonly coarse grained although not uniformly so. The plagioclase contains few or no inclusions. In a few instances quartz intergrowths are present.

Potassium feldspar occurs as both microcline and perthite, with the latter much more abundant. The crystals are generally anhedral with poorer crystal form than the plagioclase. Inclusions of quartz and plagioclase are present in some crystals. In some of the more porphyritic phases, graphic intergrowths of quartz are particularly abundant, with up to half the grain consisting of oriented strained quartz.

Quartz is abundant and occurs as large blocky grains. In many cases the square outline of the beta form is evident, particularly where the quartz occurs as phenocrysts. These grains commonly have long narrow embayments extending inward from the sides of the grain, similar to that commonly seen in volcanic quartz phenocrysts. The quartz may be highly strained, commonly to the point where incipient

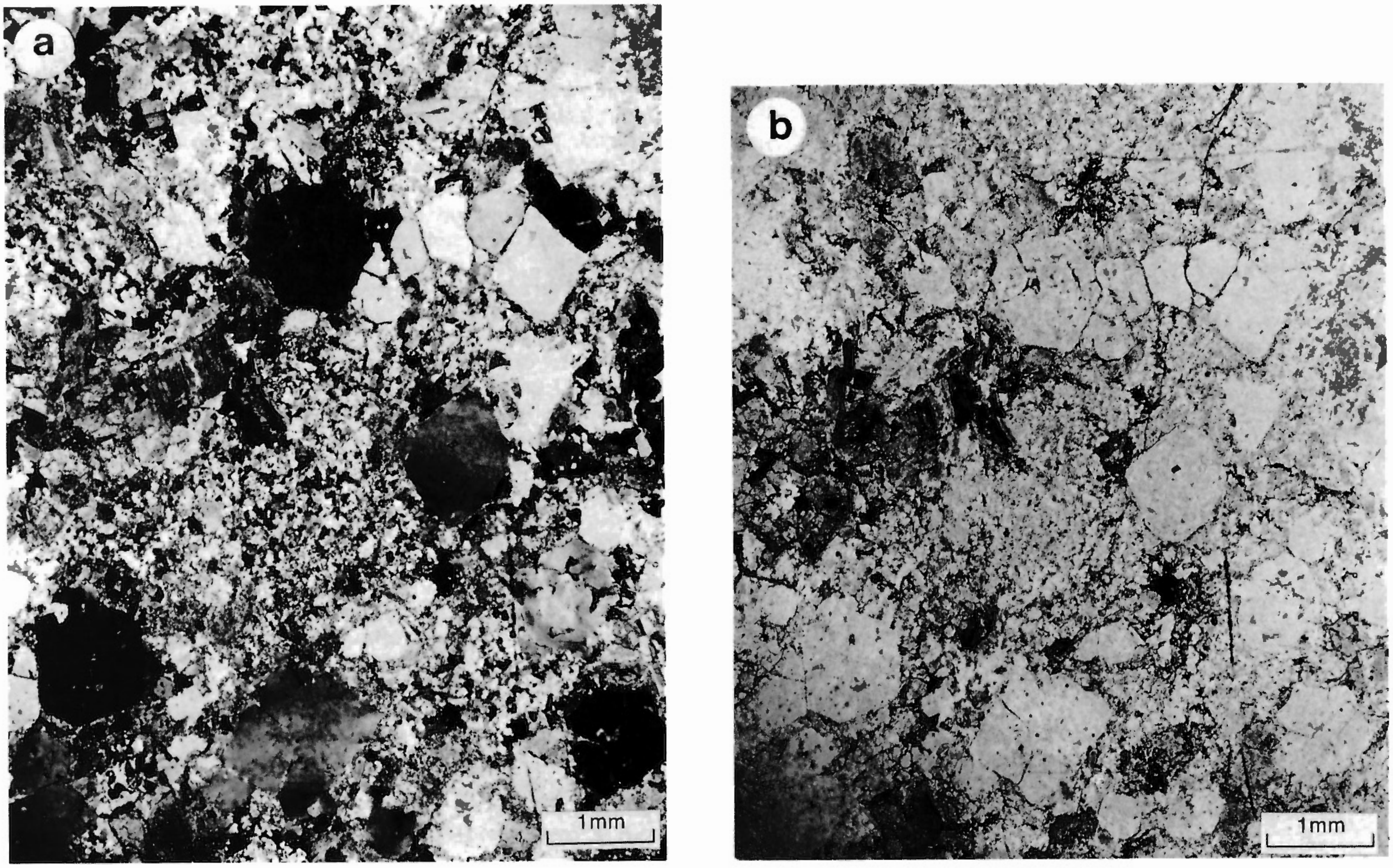


Figure 47. Textures indicating epizonal level of emplacement of the Amacher Granite. Note the porphyritic texture with coarse embayed quartz with square beta outlines. Biotite occurs in fine grained decussate mats indicating that this granite has been metamorphosed. (a) Crossed polarizers. GSC 203660-P. (b) Plane light. GSC 203660-V.

Table 9. Modal analyses, Amacher Granite

	1	2	3	4
Quartz	38	46	41	31
Plagioclase	38	42	44	14
Perthite	30			20
Microcline	1	1	8	
Biotite	2		5	3
Muscovite	1	4	1	
Chlorite	1	6		
Graphic intergrowth				35

fracture planes have formed. Where fracture has taken place, the grain is mantled and the fractures filled with fine granulated quartz. The high degree of strain and fracturing of the quartz could account for its opalescent appearance in hand specimen.

Biotite, where present, typically occurs with a distinctive decussate habit. The randomly oriented fine (0.15 mm) flakes of biotite occur in 2 mm clots interstitial to the quartz and plagioclase. In a few cases, larger flakes of biotite are preserved but invariably are extensively chloritized. Elsewhere, mats of chlorite are present with similar shape to the coarse altered biotite and the decussate biotite clots. A transition from coarse biotite to chloritized biotite to matted chlorite to decussate biotite clots is indicated.

White mica and zircon are the main accessory minerals. Zircon is typically associated with biotite, as are scattered muscovite flakes in some cases. White mica along with carbonate occurs as an alteration product of plagioclase and occurs as flakes up to 0.2 mm in size, but ranges down to very fine material, often in the same grain. Similar white mica occurs in the matrix of some of the more porphyritic varieties. White mica and chlorite are also extensively developed in fractures.

Origin and significance

The porphyritic texture of some phases of the Amacher Granite, in particular with its phenocrysts of beta quartz and abundant exsolution textures in the feldspars, suggests the pluton was emplaced in the volcanic country rock at a relatively shallow level compared with other intrusive bodies in the area. The decussate texture of the biotite that formed when the original biotite flakes were altered to chlorite and, on later heating, were recrystallized into fine grained, randomly oriented biotite, indicates that the pluton was metamorphosed after its emplacement. In addition, the original strongly zoned plagioclase has been homogenized and the alteration products, carbonate and white mica, are commonly coarse grained. In one instance a thin section across the contact between the granite and the mafic volcanic country rock contains fine fractures, extending from the volcanics into the granite, that contain a metamorphic mineral assemblage (clinozoisite but no chlorite) that is compatible with the metavolcanics (hornblende-clinozoisite-plagioclase-biotite). Henderson (1939) originally considered the possibility that this body may have been older than the Yellowknife supracrustal rocks on the basis of the previously mentioned conglomerate-like rocks at the northeast contact of the body. He concluded, however, that since intrusive relationships are well developed elsewhere, the fragmental phase probably represents a brecciated zone, but left open the possibility that more than one age of granite is present. A feature of epizonal plutons is the occurrence of intrusive breccia that forms with the emplacement of the body (Buddington, 1959; Cater, 1969). The fragmental phase in the Amacher pluton may be such a breccia. The fact that the pluton is both emplaced at a shallow level and is also metamorphosed, indicates that the body is old relative to the other intrusive granites. These were intruded at the peak of, or just following the peak of, the associated metamorphism (see Metamorphism). As the Amacher Granite was emplaced presumably at relatively high crustal levels, under low P-T conditions, time would be required for it and the surrounding volcanic rocks to be moved to a deeper level where the present grade of metamorphism was imposed.

When the early age of this pluton and its situation in the centre of a major volcanic pile is considered, it is conceivable that the Amacher Granite is a subvolcanic body, possibly comagmatic with felsic volcanic phases that occur relatively high in the volcanic sequence, south and southeast of the pluton itself.

Conglomerates in the Yellowknife Supergroup, in particular the basal conglomerates of the Jackson Lake Formation at Walsh Lake, contain cobbles derived from epizonal granite plutons. This would suggest that plutons similar to the Amacher Granite and possibly the Amacher Granite itself were unroofed at the time deposition of the Yellowknife supracrustal rocks was taking place.

The only geochronological data available on the Amacher Granite is a single zircon ^{207}Pb-^{206}Pb age of 2550 Ma which is probably a metamorphic age and a K-Ar age of 2222 ± 49 Ma on biotite which probably represents a post-Archean disturbance (R.K. Wanless, personal communication, 1979).

Detour Granodiorite

The Detour Granodiorite is a small elliptical body about 3 km long and 2 km wide at the southwest end of Detour Lake. It has been described in detail by Goetz (1974) and the following account is based mainly on this source. The intrusion is dominantly porphyritic and ranges in composition from tonalite in its core to granodiorite at its margins. It intrudes the Burwash metasediments as dykes from the porphyry extend into the metasediments and unrotated blocks of the sediments occur as inclusions in the immediate contact area. The porphyry is intruded by a white granite on its east side (not differentiated on the map) and dykes from this granite extend completely through the stock and into the sediments on the west side. The porphyry is also intruded by aplitic dykes and pegmatites.

The stock is foliated parallel to its long axis and the regional structure in the area and is metamorphosed.

The rock consists of phenocrysts of quartz and lesser amounts of plagioclase in a finer grained, granoblastic, quartzofeldspathic matrix. The quartz phenocrysts are up to 4 mm and commonly have their original euhedral form preserved. Most quartz grains, however, are polygonized and strongly strained. The plagioclase phenocrysts tend to be smaller, less than 3 mm, but commonly occur in aggregates. In some cases these may represent primary glomeroporphyritic aggregates, but other aggregations are probably due to metamorphic recrystallization of primary phenocrysts. The plagioclase is typically altered with the outline of the original phenocryst being evident only from the distribution of the alteration products. The matrix consists of a granoblastic aggregate of quartz, plagioclase and potassium feldspar with a grain size of about 0.2 mm. Metamorphic biotite occurs as fine grained aggregates to single flakes that are commonly aligned, giving the rock its foliated aspect. Muscovite is a minor component.

The Detour Granodiorite along with the Amacher Granite are the only intrusive bodies that have been metamorphosed subsequent to their emplacement. This implies a relatively old age for the Detour stock. Its porphyritic texture with a fine grained matrix implies shallow emplacement and it may represent a hypabyssal equivalent of the felsic volcanics in the area.

Unmetamorphosed felsic plutonic rocks

Defeat Plutonic Suite

Plutons of the Defeat Plutonic Suite underlie the largest area of any of the granitic units in the map area. Rocks within the various plutons range in composition from granite to tonalite. The largest complexes lie east and west of Yellowknife Bay where they have previously been called the southeastern and western granodiorites (Jolliffe, 1942; Green and Baadsgaard, 1971) although their extent to the west or southeast of Yellowknife Bay was never defined.

The Defeat Plutonic Suite occurs as large plutonic complexes such as that about Defeat Lake, and as scattered isolated individual plutons. The plutonic complexes consist of circular plutonic lobes up to 20 km in diameter to more irregularly shaped bodies. Small satellite plutons about 2 km in diameter occur at the margin of the larger structures as, for example, near the west end of both Harding and Hearne lakes, and on the northeast side of the pluton south of Watta Lake. Individual plutons are commonly defined by narrow screens of inclusions of the supracrustal rocks they intrude, to extensive continuous zones of mixed granitic and supracrustal rocks. Margins of individual plutons are sometimes apparent on airphotos even if no inclusions are present. Some, like the larger plutons northeast and south of Defeat Lake, have prominent aeromagnetic anomalies at their margins (see Geophysics section). The semicircular shape of the terrane northeast of Mirage Islands at the mouth of Yellowknife Bay suggests the presence of a plutonic lobe in this area. The rocks there contain abundant deformed to partially assimilated metasedimentary inclusions, which suggests the main part of the plutonic lobe is not exposed at the present level of erosion. No internal plutonic structure is evident within the large body west of Yellowknife Bay, but if the late faults are restored, two plutonic lobes are evident at the contact with the Kam volcanics (Fig. 13).

Away from the main complexes the intrusions occur as smaller bodies as, for example, between Watta and Blachford lakes. There a series of small plutons on the order of 1 to 4 km in diameter occur as satellites about the 10 km diameter body southeast of Watta Lake, which is itself compound. A similar situation occurs in the vicinity of Tumpline and Desperation lakes. The smallest units are small lenses or sills of granodiorite on the order of a kilometre long and a few hundred metres wide and are generally concordant. Examples of these occur at the south shore of Hearne Lake, southwest of Telluride Lake and south of Cleft Lake.

The Wool Bay Quartz Diorite, a mafic lithodeme (North American Commission on Stratigraphic Nomenclature, 1983) of the Defeat Plutonic Suite consists of several small plutons and stocks southeast of Yellowknife Bay in the vicinity of Ruth Island and Wool Bay, and 5 km west of Defeat Lake. It is discussed briefly in a later section.

The Defeat plutons are massive for the most part, particularly the smaller bodies. West of Yellowknife Bay the granodiorite has a rather weak, generally northeast-trending, foliation that is best developed in the western and northern parts of its outcrop area. Near the contact with the volcanics it is quite massive. The other bodies are massive although locally there is a weak foliation developed in some of the marginal phases.

Contact relations

Contact relations between the Defeat granodiorite and adjacent units are almost always demonstrably intrusive, commonly with sharp, clearly defined contacts. This is generally the case with the supracrustal rocks, although in some areas such as Jennejohn Lake and Yellowknife Bay, and northwest of Desperation Lake, there are large areas of granodiorite with abundant inclusions. The inclusions are generally highly deformed and range from clearly defined blocks to "gneissic" zones where the original nature of the inclusions is completely lost (Fig. 48). Zones of similar inclusions in some cases define the outline of plutonic lobes within a larger complex. In general, the contacts with the supracrustal rocks are concordant, particularly in the case of the larger bodies, such as that about Defeat Lake itself. The only sharply discordant bodies are the smaller ones, such as the satellite plutons in the vicinity of Watta Lake. Even so, many of the smallest bodies are concordant, as for example the elongate body at Drever Lake and the very small, sill to lens-like bodies at the south shore of Hearne Lake, southwest of Telluride Lake and east of the central part of Cleft Lake. Some of the plutons are mantled by mafic volcanics as, for example, the large body west of Yellowknife Bay, which is mantled in part by the Kam Formation. Similar relationships can be seen north of Mason Lake, northeast and southwest of Tumpline Lake and at Doubling Lake. Only Defeat plutons near the margins of the major supracrustal basin east of Yellowknife have a volcanic mantle. Those in the central part of the basin as, for example, in the vicinity of Watta Lake, are not mantled by volcanic units.

The contact relations with other granitic rocks are less clearly understood. The Duckfish Granite appears to be younger than the Defeat pluton north of Yellowknife as it contains inclusions of the granodiorite near its margin, and also large rafts of it in the central part of the body. The contact of the Defeat with the gneisses to the north is poorly known but is probably gradational. It is conceivable that the Defeat may have been derived from the gneisses although

Figure 48. *Migmatitic zone in Defeat pluton southwest of Mason Lake. Dykes and sills of light Defeat granodiorite intrude the darker Burwash Formation metagreywacke-mudstone. The thin light layers are due to incipient melting of the metasediments. GSC 158361*

more work is necessary before this can be established. The contact with the Awry granite to the northwest is also gradational and poorly defined. In the several-kilometre wide contact zone, the Defeat tends to be more gneissic with rather weakly developed compositional layering and an irregular fabric. In some places large blocks of granodiorite to quartz diorite occur in pinkish granitic matrix that is gradational into the pink Awry granite. Narrow dykes of pinkish leucocratic granite, similar to the Awry granite, are common in many places in the western body of the Defeat which would suggest the Awry is younger than the Defeat. The much more massive and homogeneous body of Stagg granitoid northeast of Trout Rock on the North Arm of Great Slave Lake is in sharp contrast with the rather heterogeneous to gneissic marginal phases of the Defeat. Intrusive into the mixed phases of granodiorite and metasediments south of Jennejohn Lake and southeast of Yellowknife Bay are several small plutons of Wool Bay Quartz Diorite to diorite.

A few deformed and metamorphosed mafic dykes occur in the Defeat plutonic complex west of Yellowknife Bay. These dykes are discussed in the section on Archean mafic intrusive rocks.

Metamorphic effects

Most of the Defeat plutons lie within a metamorphic envelope that on the map is outlined by the cordierite isograd in the metasediments. At Yellowknife a metamorphic zonation in the mafic volcanic rocks parallel to the adjacent Defeat granodiorite has been defined by Boyle (1961). The width of the metamorphic envelope to the cordierite isograd is varied, but is normally greater than 1 km and commonly between 2 and 3 km. One exception is south of Drybones Bay on the shore of Great Slave Lake where sediments at greenschist grade on Burnt Island are less than 0.5 km from the granodiorite on the mainland. This could be explained by a fault between the mainland and the island. Low grade metasediments that have been intruded by small Defeat plutons less than 1-1/2 kilometres in diameter commonly do not have amphibolite grade mineral assemblages mappable at the present scale in the contact area (i.e. north of Watta Lake and east of Tibbitt Lake). The width of the metamorphic envelope about the Defeat plutons contrasts with the much wider zone about the Prosperous Granite in the vicinity of Duncan and Prelude lakes where all plutons are contained within the higher grade metamorphic zone outlined by the cordierite isograd. As with the Prosperous intrusions, however, the metamorphic zonation is not thought of as a contact effect due to the emplacement of the hot intrusion into the relatively cold country rock, but more as an additional expression of the same thermal anomaly that was responsible for the generation and emplacement of the intrusion itself. There is evidence that the peak of metamorphism had passed prior to the emplacement of individual plutons. Commonly in the contact area the amphibolite grade metamorphic assemblage becomes progressively retrograded towards the pluton contact and, where preserved, the porphyroblasts are commonly rotated due to the local deformation associated with the emplacement of the pluton (see further discussion on this point in the Metamorphism section).

Lithology

The granodiorite and tonalites are generally homogeneous, equigranular, medium grained, commonly weathering buff to white to pink. Fresh surfaces range from white to dark red, depending on the degree of alteration. Locally highly poikilitic, centimetre scale, megacrysts of late stage potassium feldspar are present. Most of the smaller plutons have a distinctive texture with coarse, almost euhedral, flakes of biotite, in contrast to the larger bodies where the biotite is much finer and more ragged in appearance. In some of the larger bodies the two textural phases occur together. Rare pegmatites are associated with the granodiorite locally. The lithium-bearing muscovite-rich pegmatites in the vicinity of the Defeat pluton at Drever Lake are thought to be coincidental and not related to the granodiorite intrusion.

The intrusions are composed mainly of plagioclase, quartz and biotite. Hornblende is present at a few localities and potassium feldspar, usually microcline, is present, but in highly varied amounts (Table 10, Fig. 49, 50). Plagioclase is the most abundant mineral phase and its habit is a diagnostic feature of the unit. It occurs as stubby subhedral to euhedral crystals, commonly in aggregates, and is typically strongly zoned, often with several cycles of zoning evident within a single crystal. Typically equigranular, it also occurs as somewhat larger phenocrysts. The average composition is about An_{20} but compositions as high as An_{35} occur in the cores of some grains. The cores in some cases are strongly altered, as are certain layers in the zoned cycles. Quartz forms large irregular interstitial masses that are always highly strained and in most cases broken up into polycrystalline aggregates locally with granulated margins. This is particularly the case in the body west of Yellowknife Bay. Biotite is the most common mafic mineral and occurs as large, commonly subhedral, individual crystals particularly in the smaller plutons and also as fine ragged flakes and aggregates. The colour is varied from brown to brownish green. Chloritization is common and the margins of many flakes are irregular and altered. Epidote is commonly associated with biotite aggregates, particularly in the complex west of Yellowknife Bay where alteration and deformation appears to have been more intense. Hornblende occurs locally and where present is interstitial to the plagioclase. Potassium feldspar is invariably interstitial and commonly has abundant inclusions of quartz and plagioclase. In places it forms large, highly poikilitic megacrysts. It is particularly abundant in the core of the pluton northwest of Desperation Lake. Muscovite is present locally, particularly in the more altered rocks and appears to be of secondary origin. Apatite and opaque minerals are the most common accessory minerals.

Some chemical data from the Defeat Plutonic Suite are presented in Table 11.

Wool Bay Quartz Diorite

The Wool Bay Quartz Diorite is considered to be a mafic lithodeme (North American Commission on Stratigraphic Nomenclature, 1983) of the Defeat Plutonic Suite. It was originally recognized by Jolliffe (1942) and remapped by Easton (1982) as several small plutons and stocks southeast of Yellowknife Bay in the vicinity of Ruth Island and Wool Bay. Another 4 km long, elliptical Wool

Table 10. Modal analyses, Defeat Plutonic Suite

	1	2	3	4	5
Quartz	30	31	30	34	36
Plagioclase	52	55	51	54	27
Microcline	10	3	7	1	15
Biotite	7	10	9	9	15
Muscovite			1	1	
Hornblende					1
Chlorite	X		X	X	
Epidote					X

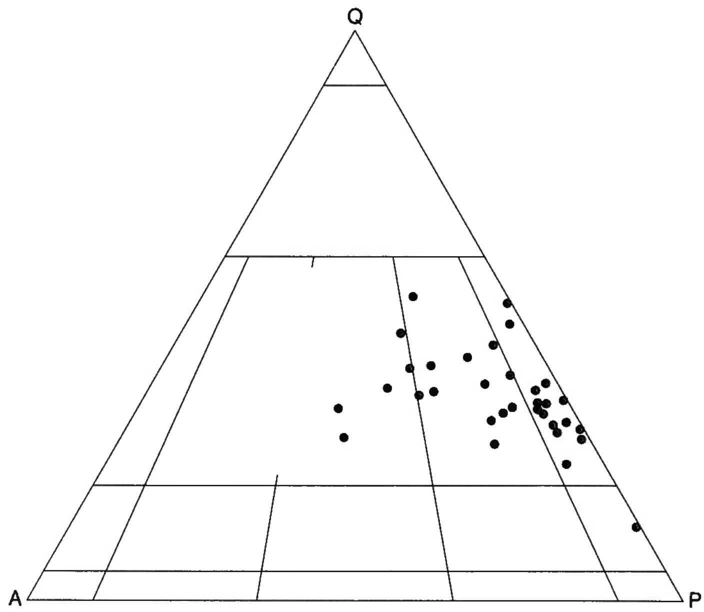


Figure 49. *Plot of modal analyses of the Defeat Plutonic Suite.*

Bay pluton occurs 5 km west of Defeat Lake. In all cases the Wool Bay plutons are in contact with Defeat granodiorites that, on the whole, contain a varied proportion of inclusions of mainly metasedimentary Yellowknife supracrustal rocks. In the Yellowknife Bay area the quartz diorite is older than the adjacent granodiorites as indicated by the local occurrence of veins of the granodiorite in the quartz diorite and blocks of the quartz diorite in the granodiorite. In some places the contact is gradational. On the other hand the Wool Bay intrusion west of Defeat Lake is younger than the adjacent granodiorites as in the border zones it contains inclusions of both Yellowknife metasediments and granodiorite. Both the quartz diorite and the adjacent granodiorite are locally cut by minor pegmatites that are thought to be related to the Defeat Plutonic Suite. The quartz diorite is black to dark green, massive, but varied both texturally and compositionally. The rock consists of hornblende, plagioclase, quartz, biotite and ilmenite in varied proportions. The grain size is varied from fine- to medium-grained and is, in general, even grained but locally contains plagioclase several times the size of the groundmass. Igneous textures are well preserved but the plagioclase and biotite are commonly altered. Several of the plutons have a negative aeromagnetic expression, a feature commonly seen in rock units containing intergrowths of ilmenite-hematite (Rose, 1969, p. 31).

Geochronology

A major program of radiometric dating in the Yellowknife area was carried out by D.C. Green (Green, 1968; Green et al., 1968; Folinsbee et al., 1968, Green and Baadsgaard, 1971). Geochronological data on the Defeat Plutonic Suite from this program and other information subsequent to it are summarized in Table 12. Geochronological work has been restricted to the area southeast and west of Yellowknife Bay in what has previously been termed the western and southeast granodiorites. No data are available for the more remote Defeat plutons. It should be noted that the geochronological study was done before the extent of the Defeat Plutonic Suite west of Yellowknife Bay was defined. Thus some of the data used to make the isochrons and concordia diagrams come from units

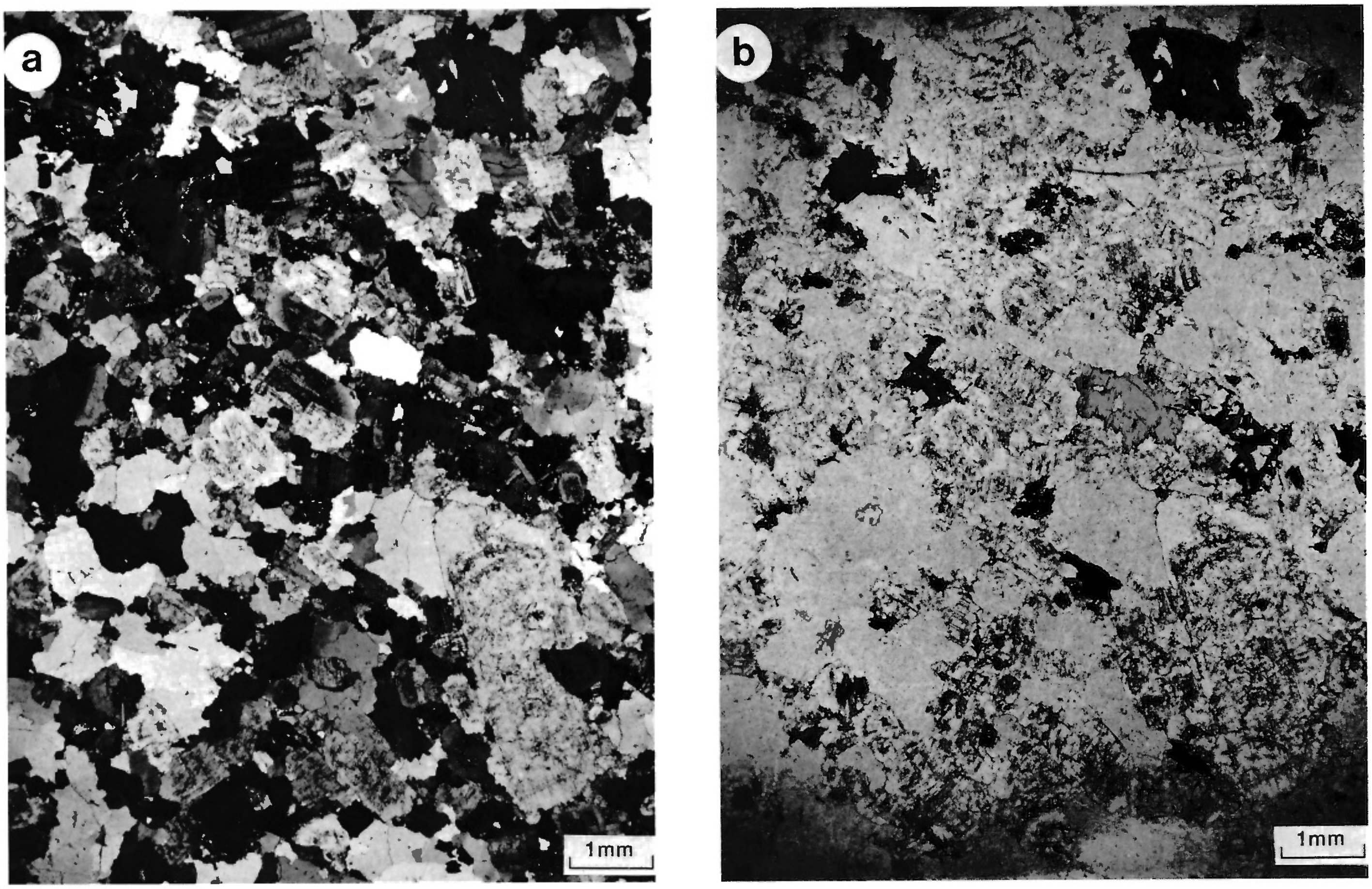


Figure 50. *The Defeat granodiorite is massive to weakly foliated and consists of euhedral to subhedral plagioclase that is commonly strongly zoned. Biotite is the dominant mafic mineral and occurs in large euhedral flakes, particularly in the smaller plutons and border zones of the larger bodies. Microcline is poikiloblastic and can occur discontinuously in optical continuity over a square centimetre or more in area. (a) Crossed polarizers. GSC 203660-W. (b) Plane light. GSC 203660-S.*

Table 11. Chemical analyses, Defeat Plutonic Suite

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	69.57	68.77	69.95	70.67	71.40	72.28	65.90	68.35	68.39	65.52	75.92
TiO_2	.33	.43	.30	.35	0.33	0.32	0.51	0.26	0.33	0.38	0.07
Al_2O_3	16.02	15.18	15.15	15.29	14.71	14.15	16.34	15.49	15.71	16.29	13.46
Fe_2O_3	2.93	3.84	2.63	2.75	2.78	2.04	.44				.11
FeO							4.01	2.36	3.08	3.64	1.51
MnO							.04	0.07	0.06	0.05	.08
MgO	.88	1.21	.75	.91	0.67	.41	1.18	1.34	0.85	1.26	.21
CaO	3.27	3.46	2.61	2.92	2.77	1.94	2.87	1.69	2.36	3.68	1.34
Na_2O	4.28	4.15	4.51	4.79	3.69	3.27	4.00	4.15	4.91	3.63	2.14
K_2O	1.27	1.51	1.61	1.39	1.83	3.63	3.20	4.18	2.41	2.75	4.17
P_2O_5	-	.04	-	.01	-	-	.24	0.10	0.06	0.09	0.10
L.O.I.	.79	1.17	0.62	0.64	0.61	.42	1.14				.77
S											.04

1-6 Defeat granodiorite on road west of Yellowknife: Hill (1980) Appendix III analyses D-256-1(1), D-258(2), D-259(3), D-261(4), D263(5), D-265(6), J.D. Hill, analyst (total Fe as Fe_2O_3)

7 Defeat granodiorite 62°26'N, 114°16'W: Folinsbee et al. (1968), Table 1, analysis 2, O. Van Breemen, analyst.

8-10 Defeat granodoite: Drury (1979) Table 1, analyses 1(8), 2(9), 3(10), G. Hendry, analyst (X-ray fluorescence; total Fe as FeO)

11 Defeat granodiorite, composite sample 3.2 km west of contact with Kam Formation: Boyle (1961) Table 12, R.J.C. Fabry, analyst.

Table 12. Geochronological data, Defeat Plutonic Suite

WEST OF YELLOWKNIFE BAY ("Western granodiorite")

Method	Date	Reference
Lead isochron (zircon)	2650 ± 10*	Thorpe (1971 – data of Green (1968)
Lead isochron (whole-rock)	2635 ± 15	Cumming and Tsong (1975)
Lead-uranium concordia	2557 ± 16*	Green and Baadsgaard (1971)
Rb-Sr isochron (whole-rock)	2555 ± 58	Green et al. (1968)
K-Ar (hornblende)	2500 ± 56	Green and Baadsgaard (1971)
K-Ar (biotite)	2450 ± 28	Green and Baadsgaard (1971)
K-Ar (biotite)	2430 ±	Burwash and Baadsgaard (1962)
K-Ar (biotite)	2660 ± 27	Green and Baadsgaard (1971)
K-Ar (biotite)	2320 ± 27	Green and Baadsgaard (1971)
K-Ar (biotite)	2180 ± 26	Green and Baadsgaard (1971)

EAST OF YELLOWKNIFE BAY ("Southeastern granodiorite")

Method	Date	Reference
Lead temperature isochron	2625 ± 40	Cumming and Tsong (1975)
K-Ar (biotite)	2595	Stockwell (1963)
Rb-Sr isochron (whole-rock)	2585 ± 37	Green and Baadsgaard (1971)
K-Ar (hornblende)	2610 ± 57	Green and Baadsgaard (1971)
Rb-Sr isochron (mineral)	2535 ± 40	Green and Baadsgaard (1971) (from Van Breemen (1965)
K-Ar (muscovite)	2520 ± 28	Green and Baadsgaard (1971)
K-Ar (hornblende)	2510 ± 56	Green and Baadsgaard (1971)
K-Ar (biotite)	2510 ± 78	Green and Baadsgaard (1971)
K-Ar (biotite)	2470 ± 28	Green and Baadsgaard (1971)
K-Ar (biotite)	2360 ± 27	Green and Baadsgaard (1971)

Data recalculated from original sources using the following constants (Steiger and Jaeger, 1977)

Constants
$\lambda(^{238}U) = 1.55125 \times 10^{-10}a^{-1}$
$\lambda(^{235}U) = 9.8485 \times 10^{-10}a^{-1}$
$\lambda(^{87}Rb) = 1.42 \times 10^{-11}a^{-1}$
$\lambda e(^{40}K) = 0.581 \times 10^{-10}a^{-1}$
$\lambda\beta(^{40}K) = 4.962 \times 10^{-10}a^{-1}$
$^{40}K/K = 1.167 \times 10^{-4}$

* When the samples were collected for these two age estimates (the same four data points were used in both cases) the 'Western granodiorite' included all granitoid rocks west of Yellowknife to Rae (about 3 km west of the area). Of the four points, only one comes from the Defeat Plutonic Suite as presently mapped; the others come from the Stagg Plutonic Suite.

other than the Defeat as defined in this report, and so are considered of questionable validity (Table 12). The range of ages presented in Table 12 indicates the times at which the various isotopic systems became closed. The older ages (2625-2650 Ma) are fairly close to the age of the Archean Yellowknife supracrustal rocks they intrude (2610-2640 Ma ^{207}Pb-^{206}Pb ages from zircons in metasediments and metavolcanics at Yellowknife; R.K. Wanless, personal communication, 1969; Green and Baadsgaard, 1971). This would imply the emplacement of the Defeat intrusions may have taken place not long after the extrusion and deposition of the supracrustal rocks. The K-Ar numbers presumably represent cooling ages with perhaps some of the younger values related to later thermal events.

Origin and significance

It would appear that the various Defeat plutons at the present erosion level represent intrusion at a variety of crustal levels. The complex west of Yellowknife Bay, particularly in its westernmost parts, is somewhat gneissic and there is a general northeasterly trending fabric in the western and central parts of the unit. These rocks tend to have a more crushed appearance and on the whole are more altered. The large intrusions about Defeat Lake and south of Watta Lake are more massive and occur in distinct plutonic lobes perhaps indicating the granodiorite is exposed at a somewhat higher level. The plutons in the vicinity of Tumpline Lake on the other hand appear to be exposed at the highest level as the rocks are very massive and commonly porphyritic with strongly zoned euhedral to subhedral plagioclase in a somewhat finer matrix.

Several interpretations have been made as to the origin and significance of the Defeat Lake granodiorite. Boyle (1961, p. 74) suggested that the western complex "was formed principally by granitization of a great thickness of sediments which once lay stratigraphically below the greenstones". This was based mainly on the fact that the granitic mass is flanked by Yellowknife supracrustal rocks, and that locally at the contacts with the supracrustal rocks where inclusions are abundant there is evidence of 'granitization' of the inclusions. On the other hand Folinsbee et al. (1968) and Green and Baadsgaard (1971) suggest a completely magmatic origin for the granodiorite. On the basis of their low strontium initial ratios, they concluded the intrusions were mantle derived and that subsequent fractionation of subsilicic hornblende under moderate pressure (700 MPa), near the base of what they considered to be oceanic crust, contributed to the formation of magmas with calc-alkalic affinities. They further suggested the Defeat plutons were emplaced into the deforming volcanic sequence where they were soon exposed by erosion, and contributed detritus to the accumulating Yellowknife sediments. Drury (1979) concurred with their proposed derivation of the plutons from a mafic source at mantle depths but felt that the hornblende fractionation process was not compatible with his rare-earth element data from the suite, and suggested that the identification of the specific evolutionary path the plutons followed was not possible with the available data.

While this model may explain the strontium data, it also has to accommodate the generation of very large volumes of felsic material from the mantle over a rather short period of time. Not only does the considerable volume of the Defeat plutons themselves have to be accounted for, but the Yellowknife sediments which have a somewhat similar composition and age as the Defeat plutons and were derived from a mixed felsic and granitoid source, represent an even larger volume of felsic material whose ultimate source must be found. If, according to the model of Green and Baadsgaard (1971), the plutons were emplaced in, and the sediments deposited on, oceanic crust without the involvement of pre-existing sialic crust, then the only source available would be derivation from presumably the same mantle source at about the same time as the plutons.

In more recent oceanic environments such as island arcs, the amount of felsic material generated is relatively small; the volcanics consisting mainly of basalt and basaltic andesite and their intrusive equivalents, while the more felsic plutons, where present, are rarely more felsic than quartz diorites and tonalites. Where exposed, these intrusions are rarely more than 15 km in size (for example the Miocene Tanzawa plutonic comples of Japan (Ishizaka and Yanagi, 1977)). The Defeat plutonic complex differs from such oceanic intrusions in that it is very much larger (by a factor of about 7 at present erosion levels) and is an order of magnitude more potassic.

There would appear to be some difficulties here and, particularly in light of other geological constraints, the premise that the plutons are mantle derived perhaps should be re-examined. There is good reason to believe that there was significant sialic crust prior to the deposition of the Yellowknife supracrustal rocks. Examples of sialic basement rocks or what have been interpreted as basement rocks have been recognized or suggested both within the area (Sleepy Dragon and Anton complexes) and elsewhere in the Slave Province: at Point Lake (Stockwell, 1933; Henderson and Easton, 1977); Indin Lake (Frith et al., 1977), and Healey Lake (Henderson et al., 1982; Henderson and Thompson, 1982). Where dated these basement rocks range in age from 2.9 to in excess of 3.1 Ga and so are significantly older than the Defeat Plutonic Suite and the Yellowknife supracrustal rocks. In addition, samples of sialic basement with a similar range of ages occur as clasts in a diatreme within the Kam Formation just south of Yellowknife (Nikic et al., 1980); a location that is almost surrounded by plutons of the Defeat Plutonic Suite. Negative evidence for the existence of sialic crust includes the idea that if, as in the model of Green and Baadsgaard (1971), the Yellowknife mafic volcanics represent remnants of oceanic crust and the Yellowknife sediments were deposited on it, one might expect that the Archean supracrustal basin to the east of Yellowknife would, to some extent, be underlain by remnants of this crust. There appears to be no evidence of this as the mafic volcanics, on the whole, appear to be preserved only where they are presently exposed; that is, rather sporadically along the margins of the area underlain by the sedimentary rocks. The evidence for this limited distribution is based on the distribution of gravity anomalies in the area (Fig. 34; Earth Physics Branch, 1969). If there were a significant thickness of mafic volcanics underlying the sediments one would expect a positive gravity anomaly throughout the basin similar to the anomaly that is only seen at the margins of the basin where thick sequences of volcanic rock are presently exposed. The 'oceanic crust' therefore has been either removed preferentially from the central part of the basin, or was never present and the sedimentary rocks were presumably deposited on sialic crust.

The evidence for sialic basement is hard to ignore. Although compositionally heterogeneous (see sections on the Anton and Sleepy Dragon complexes) there is evidence that at least locally it can be quite potassic as is the case at Point Lake 250 km to the north of the area where basement rocks dated at 3.15 Ga (Krogh and Gibbins, 1978; Henderson et al., 1982) contain up to 3.8% K_2O (Schau and Henderson, 1983). Serious consideration has to be given to the idea that the Defeat Plutonic Suite was formed at least in part from the mobilization of pre-existing sialic crust. Indeed, north of Yellowknife, the apparently gradational nature of the contact between the Defeat Plutonic Suite and the gneisses of the Anton Complex to the north may be an indication that the Defeat was derived from the gneisses.

With regard to the genetic relationship of the Defeat Plutonic Suite to other units and the evolution of the area, the following might be considered as an hypothesis worth further investigation. It is proposed that the first expression of the Defeat plutonic event may have been felsic volcanism that was active during the deposition of the preserved supracrustal rocks and indeed, upon erosion, contributed to the sediments. The volcanic accumulations may have been intruded by comagmatic epizonal plutons that were themselves exposed and eroded, and whose detritus also contributed to the continually accumulating sediments. The large complexes of the Defeat, as exposed at the present erosion level, are mesozonal expressions of the same thermal event that was responsible for the earlier extrusive volcanic rocks and high level plutons and are emplaced in the redeposited, deformed and metamorphosed detritus eroded from these earlier phases of the event. At no single locality is the complete cycle preserved but the various stages of the cycle are represented throughout the region. The earliest part, the felsic volcanic stage, is represented by the major accumulations of felsic volcanics as at Clan Lake and Russell Lake. The Burwash Formation greywacke-mudstone turbidites contain thin felsic tuff beds indicating contemporaneous felsic volcanism as the sediments were deposited. A major component of the greywackes is felsic volcanic rock fragments derived from the volcanic phase of the "Defeat" intrusive-extrusive event. The large volume of the sediments preserved in the area implies a large volume volcanic source and it is suggested that the volcanic precursors of the event (now largely eroded) may have been the source of much of this material. The next stage in the cycle is represented by the Amacher Granite, a high-level epizonal pluton that was emplaced prior to the rise of regional isotherms. The plutons east and south of Tumpline Lake represent a somewhat later stage as they are not metamorphosed but have some relatively high-level characteristics. There are abundant felsic volcanics in the supracrustal rocks about these plutons suggesting a felsic centre nearby. The deepest level and latest stage of the plutonic event is represented by the large plutons about Defeat Lake and west of Yellowknife Bay. The early stage volcanic carapace and high-level intrusions are not preserved, probably having previously been eroded, contributing to the sediments that were metamorphosed and deformed as the final stage mesozonal plutons were emplaced. Although none of the felsic volcanics have been preserved about these mesozonal plutons, some of the coarsest volcanic and plutonic detritus found in the sediments occurs east of Hearne Lake and at Upland Lake close to the complex at Defeat Lake.

Stagg Plutonic Suite

The Stagg Plutonic Suite is a massive to locally weakly foliated, dark, mafic-rich, commonly porphyroblastic granitoid of varied composition that occurs in five separate areas at the present erosion level in the northwestern part of the map area. These include two plutons northwest of Russell Lake, the main part between North Arm, Great Slave Lake to the north border of the area that is divided by a unit of Yellowknife supracrustal rocks at Stagg Lake, and a body on the North Arm at Trout Rock, somewhat removed from the others. The smaller bodies are individually quite varied but all characteristics of the unit can be seen in various parts of the main exposure of the suite. The main body is a more or less parallel sided unit about 15 km wide that, within the map area, is exposed over a length of about 50 km and has the northeasterly trend that is common throughout the region.

Contact relations

At Stagg Lake the northerly trending Yellowknife metasedimentary rocks are intruded by rocks of the Stagg Plutonic Suite. The Stagg, however, is older than, and intruded by, the Awry granitic rocks which occur both east and west of the main body of the Stagg granite. A pluton and several smaller bodies of Defeat granodiorite intrude the Yellowknife supracrustal rocks southwest of Stagg Lake but their relationship to the Stagg granite is equivocal. At one locality, however, blocks of the granodiorite occur in a matrix of Awry type, pink leucocratic granite that also intrudes the Stagg granite.

The nature of the contact between the Stagg Plutonic Suite and the Yellowknife supracrustal rocks appears to be related to the composition of the Stagg. Northeast of Stagg Lake, where the Stagg is more tonalitic in composition, the contact is sharp with few or no dykes or sills of the plutonic rock cutting the metasediments. To the northwest, where the composition is much more granitic and the intrusion is characterized by large porphyroblasts of microcline, the contact is more diffuse extending over several hundred metres from the first sill of granite in the metasediments to the granite itself which, in the contact region, contains abundant inclusions of metasediments. Contact with the Awry granitoids to the east and west is generally gradational ranging from diffuse mixtures of the two types to zones in which blocks of Stagg of varied size occur in the Awry. Southeast of Stagg Lake, for example, the contact is gradational over a distance of 1 km, changing from Stagg granitoid with increasing proportion of dykes of Awry, to Awry granitoid with inclusions of Stagg that become increasingly scattered. Isolated, commonly large inclusions of Stagg granitoid occur within the Awry as much as several kilometres from the contact. To the south, however, in the vicinity of the road, the contact is more continuously gradational with no inclusions or dykes observed in the contact area where the granitic phases tend to have a vague, irregular foliation. On the northwest side of the Stagg the contact is less sharply defined as there the Stagg consists of a lighter, porphyroblastic phase that is somewhat similar to the Awry granitic rocks.

At several localities in the vicinity of Frank Channel, North Arm, metamorphosed diabase dykes in the granite still retain primary textures, such as chilled margins, but are intruded by thin veinlets of granitic material. Somewhat similar but more deformed and recrystallized mafic dykes also occur in the Defeat Plutonic Suite to the east.

Inclusions of Yellowknife metasediments are common within the unit and are particularly abundant in the vicinity of the supracrustal rocks at Stagg Lake. They are typically small and scattered and are most abundant west of Stagg Lake. Northwest and east of Stagg Lake are several large inclusions mappable at 1:250 000 scale. Northwest of Stagg Lake in the porphyroblastic, more granitic phase, the inclusions are highly altered so that all primary structures are lost and their foliation is highly irregular. The metasedimentary inclusions are commonly migmatitic to extensively injected with dykes and sills of granite. To the east where the unit is more mafic the elongate metasedimentary zones are, on the whole, better preserved. They appear to be cut, on a small scale at least, only by the younger Awry granitic phases. At the north border of the map area east of the supracrustal belt at Stagg Lake, a lobe of the more mafic phase of the Stagg granite contains abundant metasedimentary inclusions.

Lithology

The Stagg Plutonic Suite consists of two compositional-textural end members with all gradations between the two (Table 13; Fig. 51). The tonalitic end member occurs mainly adjacent to and east of the metasedimentary belt at Stagg Lake. It also occurs to a much more limited extent southeast of the major felsic volcanic unit at Russell Lake, east of the Stagg River fault, and also as large blocks or inclusions in the Awry granite in the vicinity of Frank Channel. The rock is a massive to locally weakly compositionally layered on a 5 cm scale, medium grained, generally equigranular but locally porphyroblastic, dark, usually grey to reddish-grey, more rarely dark greenish-grey, brownish weathering, very biotite-rich tonalite. It is commonly intruded by the pink to white biotite-poor granites and less commonly by pegmatitic phases of the Awry granites. It is gradational into the porphyroblastic granitic phase which in some cases sharply intrude it. The converse was never noted.

The granitic end member is the dominant phase southwest of the metasedimentary belt at Stagg Lake but also occurs east of the metasedimentary belt, east of the

Table 13. Modal analyses, Stagg Plutonic Suite

	1	2	3	4	5
Quartz	35	54	27	32	31
Plagioclase	21	43	29	26	48
Microcline	14	X	34	28	5
Biotite	30	X	5	9	13
Hornblende		1			
Chlorite	X	3	3	2	1
Epidote				1	
Opaque	X		1	X	1

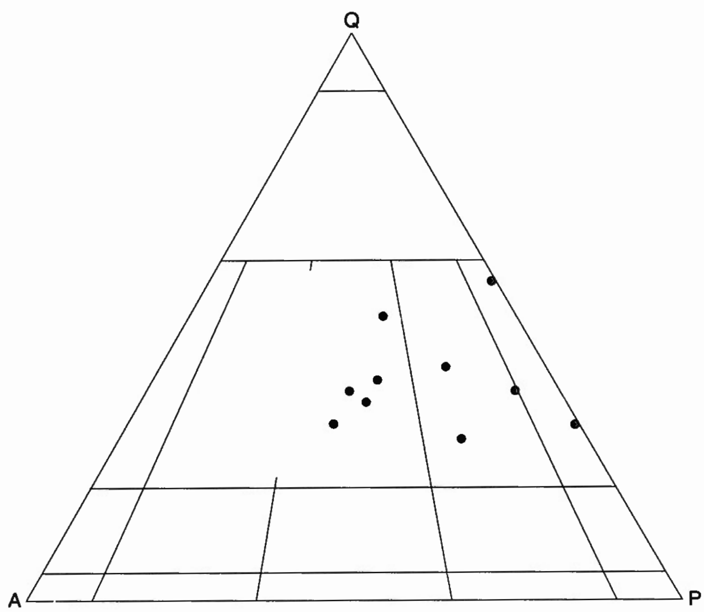


Figure 51. *Plot of modal analyses of the Stagg Plutonic Suite.*

tonalitic phase from which it grades. The rock is a massive to locally weakly foliated, generally homogeneous on an outcrop scale, medium- to coarse-grained, typically porphyroblastic with megacrysts of microcline up to 5 cm in length, dark pink, generally biotite-rich granite. The colour can be quite varied from the most common dark pink, to light pink and even white. The microcline megacrysts are commonly darker pink than the rock itself. As with the tonalitic phase, it is variably intruded by lighter pink, biotite-poor dykes and pegmatitic phases of the Awry granite that locally can make up over half the outcrop area.

The Stagg pluton north of Trout Rock on the North Arm of Great Slave Lake is similar to the more granitic phases of the suite and consists of generally massive, medium- to medium coarse-grained, biotite-rich granite with relatively few dykes of the biotite-poor phase. Contacts with the adjacent granitic units are gradational and to the west consist of diffuse zones of mainly Stagg character alternating with zones of Awry-type granite. Contact relations with the Defeat granodiorite to the east remain equivocal as phases of the Awry granite are present there as well. The main part of the granite is fairly homogeneous and dykes of the Awry granite, although present, are less abundant than elsewhere. The small Stagg granite pluton northwest of Russell Lake at the north border of the map area also consists of the granitic phase and is a homogeneous, medium coarse grained, massive, pink, porphyroblastic granite with opalescent quartz and moderate amounts of biotite. It is sharply intrusive into the Yellowknife metasediments and locally contains small metasedimentary inclusions, particularly near the contact. Muscovite-bearing pegmatites occur locally on the west side of the pluton and on the east side, dykes of granite up to 3 m wide are present.

West of Russell Lake at the west boundary of the map area, the Stagg granite consists of both tonalitic and granitic phases. As in the east it consists of a massive to weakly layered, medium grained, biotite-rich (but commonly also containing hornblende), equigranular tonalite that locally grades to a porphyroblastic granitic phase that at least in one or two localities appears to intrude the tonalitic phase. In much of the unit the tonalite is intruded by dykes of Awry-type granitoid and pegmatite that form a separate unit to the north. The relative proportions of Stagg and Awry-type granitoid are highly variable, ranging from scattered Awry dykes to over half the outcrop area consisting of Stagg granitoid as large angular blocks in an Awry granite matrix. The contact with the adjacent Yellowknife metasediments is sharp and dykes or sills in the metasediments are the younger Awry type. Inclusions of metasediments in this body are very rare. A few exposures along the shore of Russell Lake contain metasediment inclusions suggesting that the contact is probably close to and parallel to the shoreline.

In thin section the most noteworthy feature of the Stagg granite is the relatively high proportion of mafic minerals and the coarse porphyroblastic microcline commonly present in the granitic phases (Fig. 52). The rock is massive, equigranular to porphyroblastic with grain margins irregular to sutured and in some cases granulated. It is medium- to medium coarse-grained with most phases anhedral to subhedral. The tonalite phases tend to be richer in mafic minerals, do not have microcline megacrysts but are similar

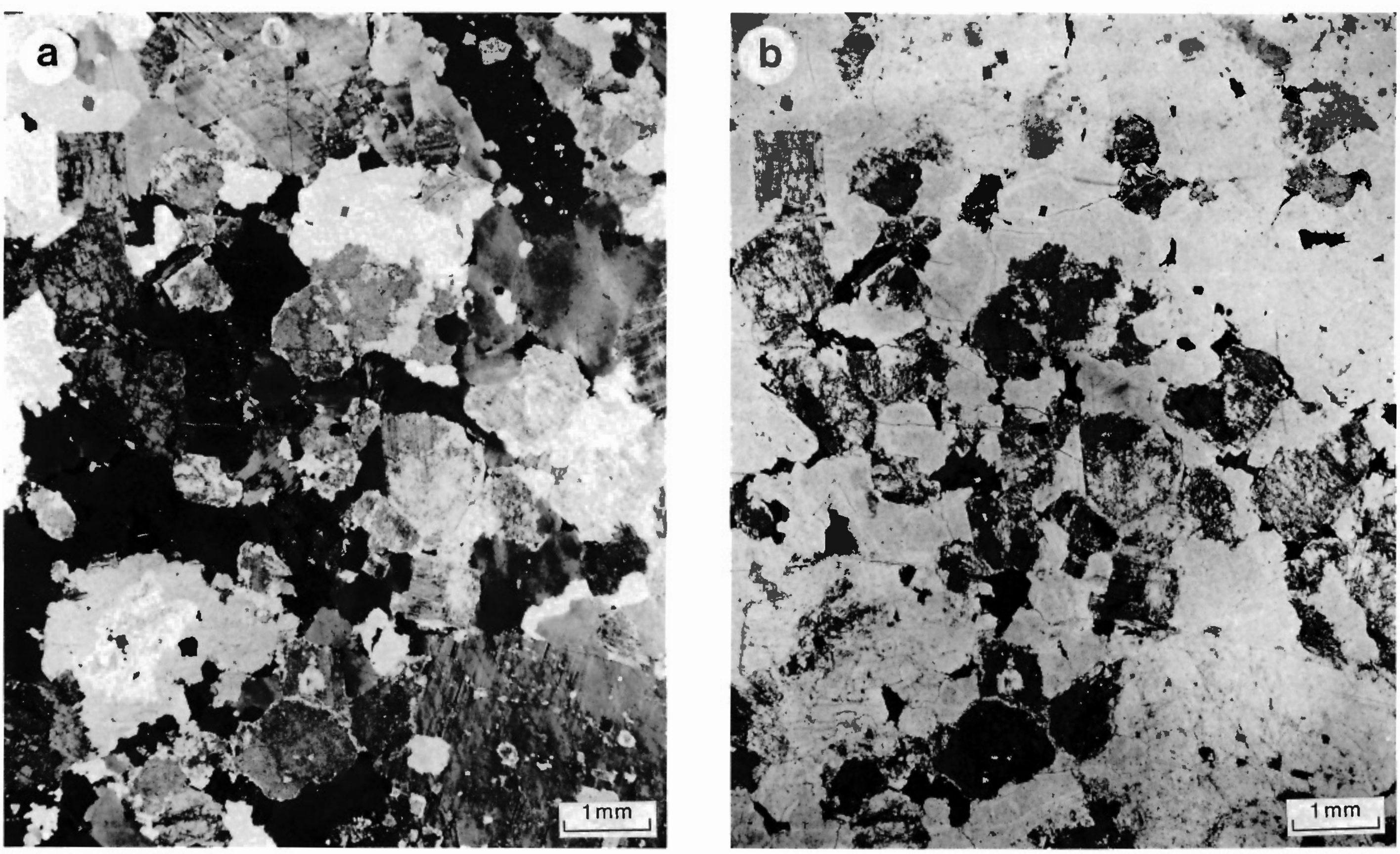


Figure 52. *The Stagg granite is a massive rock consisting of irregular to subhedral plagioclase and coarse megacrysts of microcline with abundant plagioclase inclusions. Quartz occurs in large anastomosing, commonly polycrystalline masses. Both biotite and hornblende can occur as mafic minerals and are commonly altered. (a) Crossed polarizers. GSC 201859-Y. (b) Plane light. GSC 203660-E.*

to the more granitic phases in quartz content and plagioclase texture and alteration. In some cases the plagioclase is somewhat coarser grained than the other minerals in the rock. A weak foliation in some of the tonalite is expressed by poorly oriented aggregates of biotite concentrated between framework grains parallel to the foliation.

The rock consists of plagioclase, quartz, microcline and biotite with altered hornblende in some cases. The plagioclase occurs in stubby, subhedral to anhedral grains that are normally extensively altered to sericite, epidote, and carbonate. No zoning is present, nor is any pre-existing zoning expressed by the distribution of alteration products. Clear discontinuous albitic overgrowths are present in some cases. Fine, rounded inclusions of quartz and, less commonly, biotite are present in many cases. Quartz occurs as irregular, anhedral, interstitial masses that are polycrystalline and highly strained. It commonly has ragged irregular boundaries with adjacent minerals. Some rocks are very quartz-rich, in excess of 50%, in which feldspars and mafic minerals "float" unsupported in a polycrystalline matrix of quartz. Microcline forms a highly varied proportion of the Stagg granitoid. It most commonly occurs as large, unaltered, elongate porphyroblasts up to several centimetres in length that are grossly subhedral but in detail have very irregular margins. Inclusions are abundant and consist mainly of altered plagioclase with a thin clear rim and, less commonly, quartz and biotite. The inclusions can be several millimetres in size. The porphyroblasts are commonly slightly perthitic. Microcline also occurs as fine grained interstitial fillings that are rarely perthitic. Biotite is very abundant in most cases and occurs as scattered very dark brown flakes to aggregates that are commonly altered to some degree, in some cases completely, to chlorite. Hornblende, or at least what is thought to have been originally hornblende but is now aggregates of very fine grained, bright green chlorite, is also associated with the biotite in many cases. Accessory minerals including apatite, opaques, epidote and zircon are also most commonly associated with the mafic aggregates.

Some chemical data from the Stagg Plutonic Suite are presented in Table 14 (Hill, 1980).

Origin and significance

The Stagg granite would appear to have had a more complex history than many of the other granitic units in the area and is deserving of more detailed examination. The oldest phase is the tonalite that is clearly intrusive into the Yellowknife metasediments. It appears to grade into a more granitic phase that is characterized by coarse porphyroblasts of microcline, but in other respects is quite similar to the tonalitic phase. Locally blocks of the tonalite occur in the porphyroblastic granite suggesting that the granitic phase is younger. Both phases are cut by dykes of a younger granite that is similar in many respects to the light pink, leucocratic Awry granite. A period of diabase dyke emplacement in the Stagg took place prior to the Awry intrusion as the metadiabase, still retaining primary features such as chilled margins, is cut by fine pink granitic veinlets, possibly related to the Awry granite intrusion.

The relationship between the tonalitic and granitic phases of the Stagg is not understood. Marmo (1971) suggested that coarse microcline porphyroblasts in some granites are a metasomatic phenomena and that such bodies are a granitized equivalent of more mafic rocks. Perhaps such a mechanism could be invoked to explain the Stagg granitic rocks with the porphyroblastic granitic phase being the metasomatized equivalent of the more tonalitic phase. If such was the case, the 'granitized tonalite' would have to have been mobile enough to subsequently intrude both the supracrustal rocks and the tonalite. The more mafic phases of the Stagg are similar in some respects to the tonalitic gneiss of the Anton Complex. Their contact relations with the Awry are similar as well. Could the Stagg be a partially metasomatized, and to some extent remobilized, equivalent of the Anton? Obviously more detailed work is required before this unit is satisfactorily explained.

Table 14. Chemical analyses, Stagg Plutonic Suite

	1	2	3	4
SiO_2	71.83	66.75	72.95	71.13
TiO_2	0.47	0.65	0.21	0.44
Al_2O_3	13.50	14.10	14.27	14.09
Fe_2O_3	3.56	6.34	1.75	2.62
MgO	1.00	1.43	0.50	0.83
CaO	1.00	2.80	1.19	0.68
Na_2O	2.09	2.66	2.35	2.19
K_2O	5.20	3.92	5.93	7.02
P_2O_5	0.04	0.20	0.04	0.01
L.O.I.	1.12	0.84	0.69	1.10

1-3 Stagg granite Trout Rock area: Hill (1980), Appendix III analyses D268(1), D269(2), D270(3), J.D. Hill, analyst. Total Fe as Fe_2O_3.

4 Stagg granite Frank Channel area: Hill (1980), Appendix III analysis D279-1, J.D. Hill, analyst. Total Fe as Fe_2O_3.

Awry Plutonic Suite

The Awry Plutonic Suite consists of massive, light pink, medium- to medium coarse-grained, biotite-poor granitoid and is located in the central part of the granitic terrane between Yellowknife Bay and Russell Lake. The unit has a generally northeasterly trend between North Arm, Great Slave Lake and Awry Lake. It is also found as inliers in the Paleozoic sediments south of the North Arm and also on islands in the North Arm. The main part of the unit is about 35 km wide and within the map area is exposed over a distance of about 80 km. A second, smaller part of the unit is exposed southeast of Russell Lake where it is separated from the main part of the unit by the Stagg Plutonic Suite. A small body of granite at the northwest corner of the map area is also included as part of this unit.

Contact relations

The Awry Lake granite is younger than the adjacent granitoid units although in many cases the contacts are gradational. The contact between the Awry granite and the Anton Complex in the northern part of the area occurs across a zone about 15 km wide consisting of blocks of Anton quartz dioritic to granodioritic gneiss in a matrix of the Awry granite. The proportion of the two is highly varied ranging from essentially no inclusions to in excess of 80% inclusions. The inclusions of Anton are also varied in size, ranging from a few centimetres to over a hundred metres. The contacts between the blocks and the matrix range from sharp and angular to, less commonly, rounded and rather diffuse. East of the contact within the Anton Complex there are extensive dykes and small bodies of granite in the more or less intact quartz diorite to granodiorite gneiss. To the southeast the contact between the Awry and Defeat plutonic suites is also gradational with the more mafic, locally foliated to gneissic Defeat granodiorite occurring as inclusions in an area of mixed Awry leucocratic granite and generally more mafic granodiorite. The contact of the Stagg pluton northeast of Trout Rock with the Awry granite is also diffuse with a wide zone of dominantly Awry-type granite alternating with

Stagg type. Locally dykes of the light pink, leucocratic Awry type cut the dark, porphyritic, more mafic Stagg granitoid suggesting the Awry is younger. On the northwest side of the Awry outcrop area the contact with the main Stagg granite terrane is gradational although on the whole over a narrower zone than with the Anton. To the east of Stagg Lake there are abundant inclusions of the more mafic Stagg granite in the lighter Awry granite and dykes of Awry cut the adjacent Stagg granitoid to the northwest. To the southwest of the lake the contact is more diffuse with zones of Stagg type granitoid mixed with the Awry type. As elsewhere, blocks of a more mafic phase occur in the Awry Plutonic Suite. At Russell Lake the Awry granite is clearly intrusive into the volcanics with the up to 800 m wide contact zone consisting of dykes of granite in the volcanics to volcanic inclusions in the granite. The contact with the Stagg granite to the southeast, as elsewhere, is gradational.

In addition to inclusions of generally mafic-rich plutonic rock, inclusions of metasediments are locally present. These are particularly abundant between the south end of the metasedimentary belt through Stagg Lake and the road, as blocks, several hundred metres in size, of Yellowknife Supergroup metasediments in which bedding defined by compositional layering is preserved. Metasedimentary inclusions also occur both northwest and east of Awry Lake, mainly as minor or scattered individual small inclusions.

Lithology

The Awry Plutonic Suite consists mainly of heterogeneous, massive, light pink, medium- to medium coarse-grained leucocratic granitic rocks. Locally they are weakly foliated, porphyroblastic and contain moderate amounts of biotite. This heterogeneity is locally evident in an area as small as a few hundred square metres. Pegmatite is common in the unit, usually consisting of only quartz, feldspar and biotite. The northwestern part of the Awry would appear to be more potassic on the basis of anomalously high ^{40}K concentrations indicated on the potassium radioactivity map of the area (Grasty and Richardson, 1972). In the southwestern part of the unit the granite is altered, with the feldspars bleached and grey and the mafic minerals replaced by chlorite. This occurs in distinct narrow zones and is probably related to the extensive fracture system in the area, expressed, in part, by major lineaments parallel to the north shore of North Arm, Great Slave Lake.

A noteworthy feature of the Awry granite is the local occurrence of uranium stain. It typically occurs on flat, glacially polished surfaces as a bright yellow stain occurring both on the granite and pegmatite. It is made particularly evident as lichen will not grow on it. The stain occurs as irregular patches a few centimetres to a metre in diameter that typically occur in groups aligned along fractures in some places but elsewhere apparently randomly distributed. No primary uranium minerals were recognized associated with the secondary uranium minerals which are mainly uranophane. Lang et al. (1962) have suggested that such occurrences are supergene deposits in which the uranium was leached from lean occurrences and then precipitated from long standing ponds of rain or meltwater. The occurrence of this stain on thin dykes of the Awry granite intruding adjacent granitic units which are not stained is not explained by this mechanism. The occurrence of this stain caused considerable interest in the mid-fifties when a large area in the vicinity of Trout Rock was staked. Similar stains also occur on some outcrops of the Prosperous Granite to the east.

In thin section the Awry granite is as heterogeneous on a microscopic scale as it is in outcrop (Fig. 53, 54, Table 15). In general it is a massive granite of medium- to medium

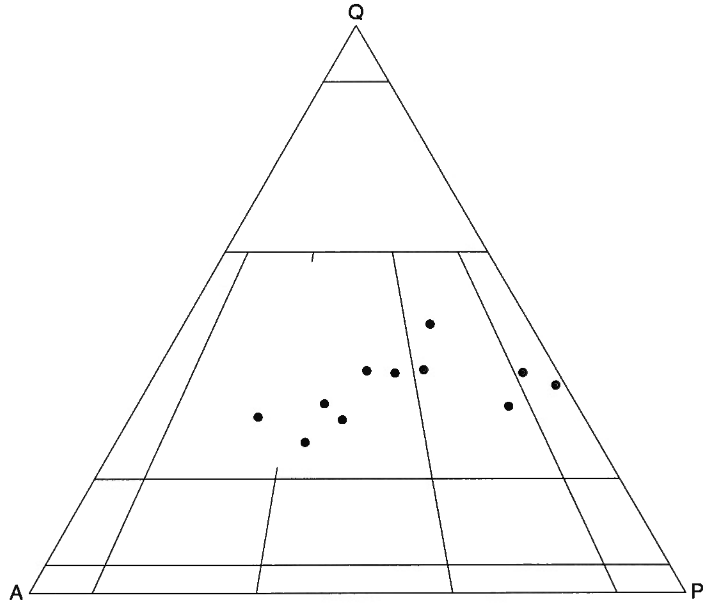


Figure 53. *Plot of modal analyses of the Awry Plutonic Suite.*

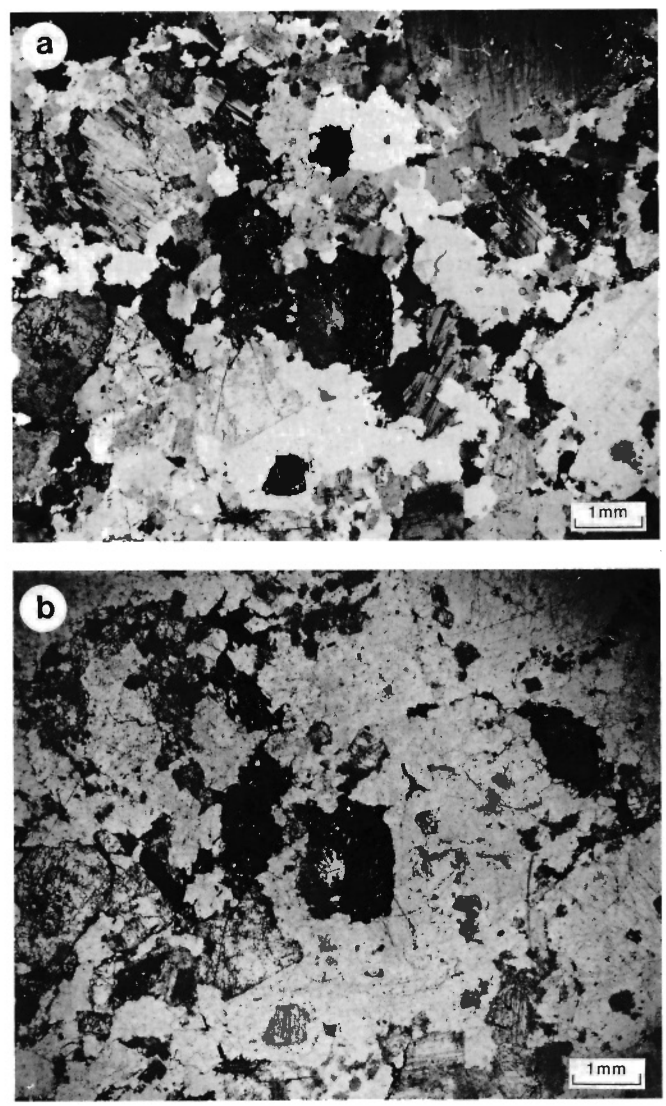


Figure 54. *Awry granite is a massive unit consisting of abundant quartz, subangular, generally altered plagioclase, megacrysts of microcline with abundant inclusions of plagioclase and moderate amounts of biotite. (a) Crossed polarizers. GSC 203660-O. (b) Plane light. GSC 203660-M.*

Table 15. Modal analyses, Awry Plutonic Suite

	1	2	3	4	5
Quartz	30	32	35	29	25
Plagioclase	19	27	37	51	28
Microcline	49	37	16	9	42
Biotite	1	1	8	8	X
Muscovite		X	1	3	2
Hornblende		X			
Chlorite	2	3		1	3
Perthite			2		
Epidote					X
Opaque			1	X	X

Table 16. Chemical analyses, Awry Plutonic Suite

	1	2	3
SiO_2	73.88	71.87	74.40
TiO_2	0.19	0.24	0.10
Al_2O_3	14.29	14.17	13.99
Fe_2O_3	1.58	2.20	1.03
MgO	0.41	.74	0.44
CaO	0.67	1.76	0.26
Na_2O	2.92	2.92	2.34
K_2O	5.50	4.06	6.35
P_2O_5	-	0.06	-
L.O.I.	0.86	0.85	.83

Awry granite from road north of North Arm: Hill (1980), Appendix III, analyses D272(1), D275(2), D277(3). J.D. Hill analyst (Total Fe as Fe_2O_3)

coarse-grain size, equigranular to inequigranular and typically has a small amount of mafic minerals. The texture is granular with most phases subhedral to anhedral. In the megacrystic varieties coarse subhedral laths of microcline are present. Locally the rock has a weak foliation expressed by the alignment of sparse, somewhat shredded, biotite which tends to envelop the framework silicates. Evidence of weak crushing of the rock is shown by narrow zones of annealed finer grained broken quartz and feldspar, and the crudely rounded or milled aspect of the coarser grains in some cases.

The rock is composed mainly of microcline, quartz, plagioclase, biotite and/or chlorite. Microcline, which in some cases is weakly perthitic, is the major component. It occurs as interstitial anhedral grains to coarse, subhedral, commonly Carlsbad twinned megacrysts in excess of 1 cm in length. It is rarely altered. The coarse grains commonly contain inclusions of plagioclase, and to a lesser degree quartz, that can be up to several millimetres in size and indicate that the microcline formed at a relatively late stage. Orthoclase is present in a few samples. Quartz occurs interstitially in large anastomosing areas with small scale cuspate margins against other minerals and locally appears to replace them. In minimally deformed samples, quartz in optical continuity can be traced over 2 cm. In most cases, however, the quartz grains are polycrystalline with irregular margins to polygonized on a fine scale where the rock has been crushed. Stubby to equant subhedral grains of plagiocase 5 mm or less in size are everywhere altered. The alteration products, sericite-epidote, occur with a patchy distribution that does not reflect an earlier compositional zonation. Inclusions of plagioclase in the microcline megacrysts have a clear narrow unaltered rim. Mafic minerals are a minor component and form about 5% of the rock. They consist of fine biotite flakes generally less than 2 mm in size or chlorite that has evidently replaced biotite. Biotite in most cases is chloritized to some degree. Fine secondary muscovite (less than 1 mm) is associated with the biotite locally.

Some chemical data from the Awry Plutonic Suite are presented in Table 16 (Hill, 1980).

Meander Lake Plutonic Suite

The Meander Lake Plutonic Suite consists of heterogeneous, generally massive, medium grained, pink, equigranular to weakly porphyritic, biotite moderate, muscovite-poor, granitic rocks at the eastern margin of the map area. Within the map area it underlies a roughly rectangular area about 60 km long and 20 km wide, but is almost transected by a major belt of Yellowknife supracrustal rocks, dominantly mafic volcanic flows, tuffs, and volcanogenic sediments of the Payne Lake Formation. It is offset by a series of northerly trending left lateral faults and is transected by a major north-northeasterly trending cataclastic to mylonitic zone. Several satellitic plutons that occur away from the main Meander Lake complex have been included in the unit. The largest, a single faulted pluton, occurs in the metasediments south of Sunset Lake while east of Boulding Lake a few small satellite bodies occur in the Burwash Formation near the main complex. The Meander Lake Plutonic Suite was mapped almost entirely by helicopter and air photo interpretation and so coverage is less detailed than most other units.

Contact relations

The granite is younger than Yellowknife supracrustal rocks which it intrudes. The contact with the mafic volcanics and contemporaneous gabbro sills is gradational, consisting of a zone ranging from dominantly mafic rocks with dykes and sills of the granite, to dominantly granite with scattered mafic inclusions. To the south the contact is sharp and straight and coincident with the major cataclastic to mylonitic zone. Although deformation within this zone is commonly extreme, the rocks are derived mainly from a granitic protolith. West of the cataclastic mylonitic zone minor dykes and sills of the granite occur in the volcanic units. Southwest of Payne Lake there is a major lobe of the Meander Lake granite that is not significantly deformed. Southeast of the fault through Meander Lake the granite at the volcanic contact is again minimally deformed or undeformed. In the contact area the dominantly intermediate to felsic volcanic rocks are highly deformed and contain relatively undeformed dykes and sills of the granite, that increase in size and proportion over 1.5 km towards the main granite body. The granite engulfs the southeasterly extension of the volcanic belt at the east margin of the map. The contact with the volcanics is in general concordant although in detail it is discordant.

The southern part of the unit, about Meander Lake itself, is for the most part intrusive into the Burwash metasediments. The northeast margin of this body consists of a poorly defined mixed zone, about 2 km wide, in which there is a high proportion of deformed metasediment inclusions. This contrasts strongly with the sharp contact with the metasediments to the west and south. In contrast to its contact with the volcanics in the north, the granite contact with the easterly trending metasediments is for the most part discordant.

Metamorphic effects

Associated with the Meander Lake granites is a metamorphic envelope in the sediments. It is most clearly evident south of Desperation Lake and southeast of Francois Lake where the aureole, as outlined by the cordierite isograd, is clearly parallel to the intrusive contact of the granite with the sediments. North of Desperation Lake the cordierite isograd swings away to the west and all the metasedimentary terrane to the northwest is in the amphibolite facies of metamorphism. Work in the area has not been sufficiently detailed to distinguish thermal effects related to the Meander Lake granite from metamorphism related to other granites in the immediate area. On the other hand the higher metamorphic grade in this area could also be a regional metamorphic gradient increasing to the north, and unrelated to any specific intrusive event, although this is considered unlikely as the Sunset Lake basalts about Sunset Lake are at greenschist grade. The width of the metamorphic zone apparently related to the Meander Lake granite is varied. South of Desperation Lake it is about 1 km wide, while southwest of Francois Lake it is 2.5 to 5 km wide. The metamorphic aureole is not thought of as a contact effect of the granite with the heat required for metamorphism coming directly from the adjacent granite. Such contact effects are typically very narrow, measurable in tens of metres rather than thousands (see discussion by Turner, 1981). More likely, the metamorphic aureole is an expression of the heat associated with formation or mobilization of the granite, and that accompanied or shortly preceded emplacement of the granite into its present position.

Lithology

Within the Meander Lake Plutonic Suite some aspects of its internal structure are evident. Along the southern margin of the body several small lobes are present that typically have a diameter of only a few kilometres. Deeper within the body along the eastern margin of the map area are at least two much larger plutons that are on the order of 10 km in diameter. Along the southeastern margin of the northernmost of these plutons the granite is gneissic with shallow dips.

Also present within the Meander Lake terrane are large areas of inclusions that represent remnants of Yellowknife supracrustal rocks. These inclusion-rich areas are highly varied and poorly defined, varying from scattered isolated inclusions to dense swarms in which granite matrix is a minor component. The inclusions range from small angular blocks to banded gneisses with dykes and sills of granite. South of Meander Lake, for example, is an area of essentially intact metasediments with preserved primary compositional layering. In general, however, the inclusions are highly deformed and altered. Northeast of the Payne Lake volcanics the inclusions are almost entirely amphibolitic suggesting a mafic volcanic parent. To the southwest the inclusions are quartzofeldspathic and pelitic suggesting derivation from metasediments similar to those that occur to the west. A third, less abundant, inclusion type consists of blocks of granodiorite and tonalite in a matrix of Meander Lake granite. Although their distribution is not defined, the mafic granitoid inclusions appear to be most abundant in the northern part of the body particularly in the west, although examples were noted within the circular pluton at the eastern margin and also northeast of Desperation Lake. These granodioritic to tonalitic inclusions tend to be rounded, commonly have diffuse contacts and in places have a gneissic fabric. They are similar in many respects to the inclusions in the Awry Plutonic Suite that occur in a wide zone adjacent to the older Anton Complex and to a lesser extent adjacent to the Stagg Plutonic Suite, and are believed to be derived from these bodies. A similar origin is proposed for these inclusions in the Meander Lake granite in light of their proximity to the Sleepy Dragon Complex which is thought to be older and probably, at least in part, basement to the Yellowknife Supergroup. Although the Meander Lake Plutonic Suite is separated from the Sleepy Dragon Complex by volcanics along the Beaulieu River, the intermediate, orthogneiss inclusions in the northwest part of the Meander Lake may be evidence of similar granitoid basement east of the volcanics as well.

A striking feature within the Meander Lake granite is the north-northeasterly trending, steeply dipping cataclastic to mylonitic zone in the northeast body of the granite. It is a unique feature within the map area. The zone is up to 2 km wide and ranges from a mildly crushed zone in which the feldspars are reddened and the mafic minerals chloritized, as is the case at the north margin of the area, to a glassy structureless mylonite that occurs locally southwest of Payne Lake. In places there are extensive breccia zones with quartz stockworks that form resistant ridges. Indeed the zone is evident morphologically on the 1:250 000 scale topographic map of the area. Within the zone on a small lake 8 km south of Payne Lake, a small lens of presumed early Proterozoic Great Slave Supergroup conglomeratic arkosic quartzite to quartzite occurs that has been mildly crushed. The zone is displaced by the early Proterozoic faults in the area.

The granite is heterogeneous and may be divisible into several units with more detailed mapping. On the whole, it is generally massive, medium grained and equigranular. Porphyroblastic microcline, however, occurs sporadically, particularly in the north, in megacrysts up to 1 cm in size. The colour is pink but also varied from generally darker in the northwest to white in the south. Biotite is the only mafic mineral and is everywhere present in moderate proportions. Muscovite is more sporadic and tends to be more common to the south. Garnet occurs locally, particularly where the granite contains mafic inclusions. Pegmatite is common but sporadically developed, varying from none to in excess of 30% of the outcrop. For the most part pegmatites are simple quartz-feldspar-(muscovite) assemblages but in a few cases tourmaline is present. In one area east of Boulding Lake, pegmatites containing spodumene occur in the metasediments and are associated with the small satellite plutons. The adjacent granite differs somewhat from the rest of the Meander Lake granite in that it contains abundant muscovite and is white. This part of the Meander Lake granite therefore may be more similar to the Prosperous-type granites which also typically have lithium-bearing pegmatites associated with them. The extent of these Prosperous-type granites within the Meander Lake Plutonic Suite has not been defined and so they have not been mapped as a separate unit.

In thin section the Meander Lake granite is heterogeneous with few if any unifying characteristics. The composition is varied, ranging from granite to tonalite (Table 17, Fig. 55). Texturally the rock is also varied from an even grained granitic texture to porphyroblastic. The rock shows varied degrees of crushing from the homogeneous mylonite in parts of the cataclastic zone to essentially

Table 17. Modal analyses, Meander Lake Plutonic Suite

	1	2	3	4	5
Quartz	35	32	27	26	37
Plagioclase	30	43	53	59	40
Microcline	29	20	18	7	19
Biotite	4	6	2	7	4
Muscovite	2		X	2	1

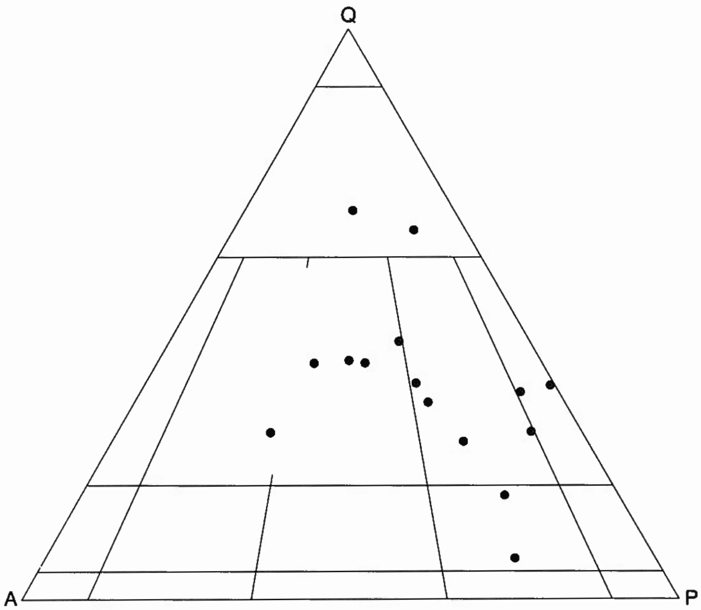


Figure 55. *Plot of modal analyses of the Meander Lake Plutonic Suite.*

undeformed rocks in which primary igneous textures are well preserved. The granites in the southern part of the unit are less deformed than the rocks to the north. Many are between these two extremes exhibiting moderate degrees of cataclasis indicated by the polycrystallinity of the quartz, the somewhat abraided nature of the feldspars, with the finer grained quartzofeldspathic abrasion products occurring between the coarser grains and, in more extreme cases, the fine shredded aspect of the micas and their orientation into a weak fabric.

The rock consists mainly of plagioclase, microcline, quartz, biotite, and muscovite. The plagioclase occurs normally as anhedral equigranular grains, rarely as megacrysts, with very irregular boundaries. In most cases compositional zoning is not preserved except as zones of greater alteration that reflect the crystal outline of the growing grain. Commonly the core is highly altered reflecting its original more calcic composition while the outer part is relatively free of alteration. Plagioclase within microcline porphyroblasts shows a similar clear margin and more altered core. In a few cases myrmekite is developed in some of the crushed plagioclase. Microcline occurs as both coarse porphyroblasts in excess of 1 cm in size and as relatively fine interstitial grains. The coarse porphyroblasts are generally elongate but in detail have irregular margins. They typically contain inclusions of plagioclase and quartz of varied size and shape. They are commonly slightly perthitic. The proportion of microcline is highly varied from none to being the dominant phase in the rock. Quartz also occurs interstitially in anastomosing aggregates that where undeformed can extend in excess of 1 cm in optical continuity. In most cases, however, the quartz reflects the degree of deformation of the rock and is typically polycrystalline, in some cases on a very fine scale. In detail the margins of the quartz are finely irregular to cuspate, suggestive of replacement of the adjacent feldspars. Biotite is the only significant mafic mineral. It occurs in small aggregates to rather fine scattered flakes and, in the more deformed rocks, is altered to chlorite. The biotite is a deep dark reddish brown suggesting a high titanium content that seems to be a characteristic feature of the unit. Pleochroic haloes about zircon(?) within the biotite are strongly developed. Muscovite is a common constituent of the granite and occurs as scattered, relatively coarse flakes, to finer aggregates associated with biotite, to thin very fine oriented flakes within the altered cores of plagioclase. Garnet is locally present in the Meander Lake granitic rocks particularly where associated with supracrustal inclusions.

Prosperous Granite

The Prosperous Granite consists of a group of massive, medium grained, biotite-muscovite granite plutons and associated pegmatite. The main group of plutons occurs between Prosperous, Duncan and Reid lakes. They consist of bodies up to 37 km in length and 14 km wide. Other generally smaller bodies occur on the Yellowknife River at the north border of the map area, at Matonabbee Point, south of Detour Lake, west of Buckham Lake, and north and east of Turnback Lake. An unmapped pluton of unknown extent, that probably is part of the Prosperous Granite, occurs with the Meander Lake Plutonic Suite in the vicinity of Boulding Lake.

Contact relations

The contacts of the Prosperous granites with the adjacent country rock are typically sharp and definite. There are few dykes and sills of granite extending into the metasediments, although pegmatites in some areas are quite common. The pluton on the Yellowknife River west of Clan Lake has a sharp contact with the supracrustal rocks, but the west margin with the granitic gneisses of the Anton Complex is gradational, with extensive sills and dykes of the granite in the gneiss. Between Duncan and Prosperous lakes there is an abundance of small granitic plutons, stocks and pegmatites with a wide range of sizes, although each is sharply defined (Fig. 56). The extent, to say nothing of the contact relations, of the two plutons north of Turnback Lake are poorly known.

Metamorphic effects

The main group of plutons and associated dense swarm of stocks and pegmatites occur within a broad thermal ridge that is outlined by the cordierite isograd on the map. This metamorphism is not directly related to the emplacement of the plutons; that is, it is not considered a contact effect. There is little doubt, however, that the metamorphism and the plutons are ultimately related. Most of the other Prosperous plutons elsewhere in the map area have also been intruded into relatively high metamorphic grade country rocks. Even those plutons not directly associated with higher grade terrane, such as those south of Detour Lake, have a very wide metamorphic zone about them. For example, the pluton north of Consolation Lake, which is only about 2 km in diameter, has an aureole as defined by the cordierite isograd 1.5 to 2 km wide. The metamorphism of part of the main thermal ridge has been described by Ramsay and Kamineni (1977) and is discussed more completely in the section on metamorphism.

Lithology

In many cases the Prosperous intrusions consist of a series of discrete plutonic lobes averaging about 5 km in diameter. These are particularly evident in the largest body southeast of Duncan Lake. The rounded lobes are defined from air photographs as commonly there is little apparent lithological variation between lobes. In some cases, however, distinct changes are evident, as for example in the body north of Consolation Lake where there is an abrupt change in grain size across the nested plutonic boundary. The various lobes within a plutonic complex are commonly intersected at different levels by the present erosion surface. For example, in the large plutonic complex east of Duncan Lake the

Figure 56. *Small Prosperous Granite pluton 3 km south-southwest of Duncan Lake. The pluton and abundant dykes, sills and irregular intrusions of similar granite and pegmatite intrude Burwash Formation metagreywacke-mudstone. The prominent east-northeasterly and northeasterly linear features are Dogrib and Indin diabase dykes respectively. NAPL A8662-108.*

deepest level is represented by the main part of the intrusion, which is very massive, homogeneous and generally free of inclusions. The lobes on the east side represent an intermediate level containing large roof pendants of the intruded sediments. On the west side at the south end of Duncan Lake the present erosion surface intersects just above the top of a plutonic lobe as that area consists of a very high density of small stocks and dykes of granite and pegmatite intruding the Burwash metasediments.

The larger plutons are elongated in a northerly direction. This orientation is generally parallel to the structural trend of the sedimentary rocks they intrude. In detail, however, the plutons are sharply discordant, particularly on the northern and southern margins. This is evident in the vicinity of Sparrow and Hidden lakes. The smaller plutons are even more discordant. Inclusions in the plutons tend to be angular and only minimally deformed. They range in size from very small to large blocks over 2 km in length but are not particularly abundant.

The granite is typically a massive, rather homogeneous rock, medium- to medium coarse-grained and somewhat varied in colour, ranging from white to buff to light pink (Table 18; Fig. 57). Some plutons, for example the body south of Prestige Lake, are weakly porphyritic. A unifying feature of all the granitic bodies of this unit is the presence of two micas. Biotite and muscovite normally are present, although commonly in highly varied proportions. Some bodies, such as those east of Buckham Lake and south of Detour Lake, have large areas of granite with only muscovite. In parts of other bodies biotite is dominant. Locally, the margins of some plutons are weakly foliated as is the case on the northeast margin of the pluton at Matonabbee Point. One major exception is the pluton south of Detour Lake which has a weak and generally variably developed foliation due to the orientation of micas throughout. The Prestige pluton south of Prestige Lake contains northeasterly oriented 3 cm potassium feldspars throughout. This pluton has been described in detail by Kretz et al. (1982) with particular reference to its Na-K-Li geochemistry.

Table 18. Modal analyses, Prosperous Granite

	1	2	3	4	5
Quartz	35	30	30	42	59
Plagioclase	30	34	33	27	27
Microcline	27	28	22	21	6
Biotite	4	2	4	2	5
Muscovite	1	5	11	8	4
Chlorite			X		
Perthite	2.8	X			

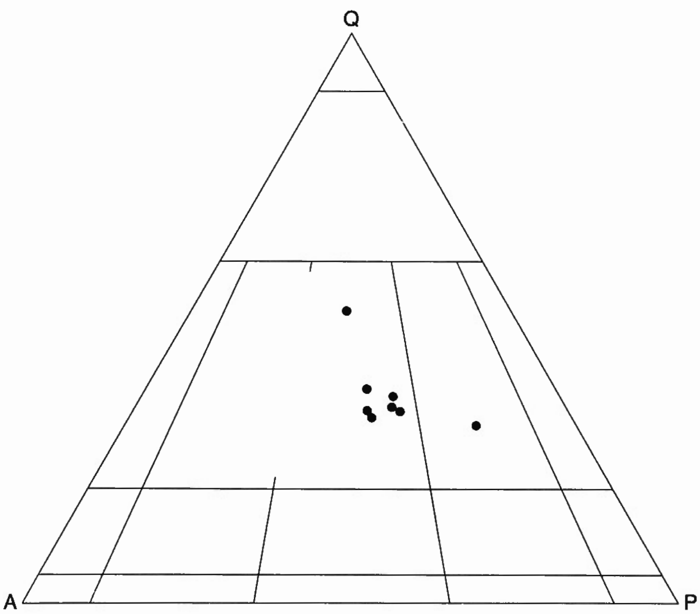


Figure 57. *Plot of modal analyses of the Prosperous Granite.*

A typical Prosperous granite consists of 30% quartz, 35% plagioclase, 25% microcline, 8% muscovite and 2% biotite. The rock is even grained, with a size range of 3 to 5 mm (Fig. 58). In some cases the rock is weakly porphyritic with subhedral Carlsbad twinned microcline crystals up to 8 mm long. Grain margins tend to be embayed on a fine scale. The potassium feldspar is typically anhedral microcline and is usually the coarsest phase. Small inclusions of quartz, plagioclase and mica are common. Plagioclase is also anhedral, generally finer grained, and more altered than the microcline. It is only weakly zoned, although in some cases more calcic cores are outlined by dense alteration products. Inclusions are present but less abundant than in the microcline. Quartz occurs as equant interstitial grains, strained but commonly not polygonized. Muscovite occurs as coarse, evenly distributed, generally unoriented single flakes that in some cases are bent. The very dark reddish-brown biotite, on the other hand, tends to be finer grained (1-2 mm), rather ragged, commonly chloritized, and in some cases has a preferred orientation. Euhedral garnet occurs locally. Apatite, typically associated with biotite, is the most common accessory mineral. Iron oxides are rare to nonexistent.

An interesting feature of this granite is the scattered occurrence of uranophane stain on some outcrops. This can be seen in the main body southeast of Duncan Lake, and also on the pluton east of Matonabbee Point. The stain occurs as scattered bright yellow patches a few centimetres to several tens of centimetres in size, of varied shape, and commonly found on the highest part of an outcrop (Fig. 59). Uranium stain identical to this also occurs in considerably greater abundance on outcrops of the Awry granite north of North Arm, Great Slave Lake and is discussed more completely in the section on that granite.

Pegmatites are commonly associated with the Prosperous granites, particularly in the area between Duncan and Prosperous lakes and occur both in the surrounding sediments and in the plutons themselves. The pegmatites in the vicinity of Prestige Lake, Staple Lake and Sparrow Lake have been studied in detail by Kretz (1968, 1970). Parts of the main body southeast of Duncan Lake can be as much as 40% pegmatite, although most of the plutons are homogeneous granite. The pegmatites within the plutons are typically composed of quartz, plagioclase, potassium feldspar, muscovite and locally, tourmaline. In the surrounding metasediments, on the other hand, the pegmatites can have a very complex mineralogy that includes most importantly, beryl, spodumene, and columbite-tantalite, along with many other minerals in minor amounts. Kretz (1968) reports a mineral zonation of the pegmatites with respect to the adjacent granite bodies with tourmaline, beryl, spodumene and columbite-tantalite occurring at respectively greater distances from the granite. Although not as numerous, lithium pegmatites also occur west of the pluton at Buckham Lake and south of the pluton northeast of Turnback Lake and in the vicinity of Boulding Lake. Boyle (1961) has suggested that the pegmatites in the vicinity of the pluton at Prosperous Lake were derived from the rocks they occur in; that is from the crystallized but still hot granite, or the sediments during their metamorphism. Kretz (1968), on the other hand, suggested that all the pegmatites and the granites had a common source, were separated prior to the emplacement of the granite, with the pegmatitic fluids reintroduced somewhat later along grain boundaries and fine fractures, and eventually accumulated in active dilatent zones in both the granite and country rock.

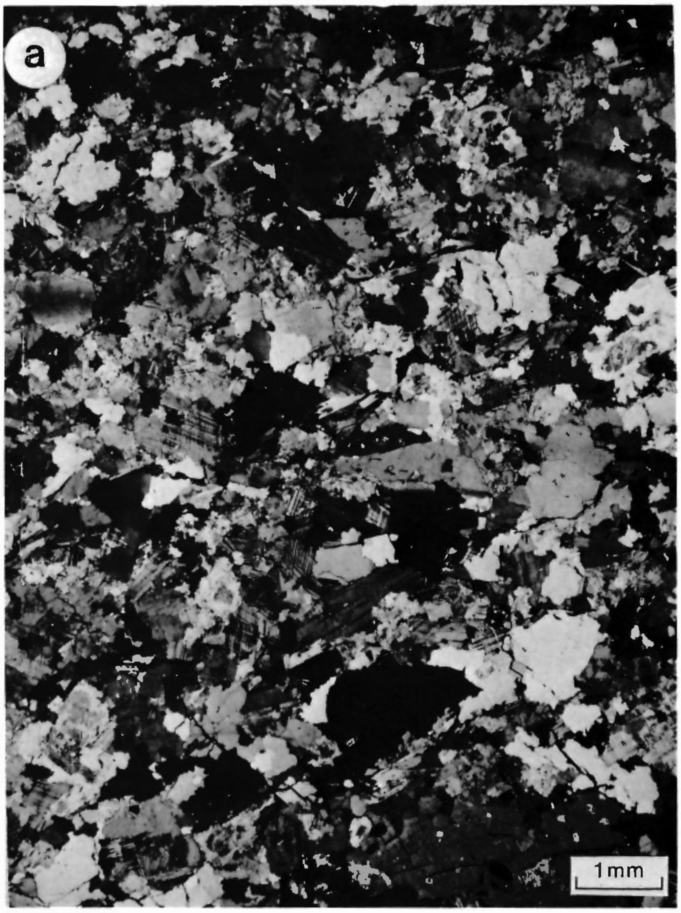
Some chemical data from the Prosperous Granite are presented in Table 19.

Geochronology

The pluton at Prosperous Lake has been dated at 2520 ± 25 Ma (Rb-Sr mineral isochron $\lambda^{87}Rb = 1.42 \times 10^{-11}\ yr^{-1}$ (recalculated from $\lambda = 1.39$), $Sr_i = 0.712 \pm 0.002$; Green et al., 1968) while K-Ar ages (recalculated) from the granite include four muscovite ages at 2520 Ma (Stockwell, 1962), 2520 Ma (Burwash and Baadsgaard, 1962), 2515 Ma (Green et al., 1968) and 2510 Ma (Green and Baadsgaard, 1971) and four biotite ages at 2410 Ma (2; Green et al., 1968), 2390 Ma (Green and Baadsgaard, 1971) and 2190 (Green and Baadsgaard, 1971).

Origin

Both Boyle (1961) and Green et al. (1968) suggest the Prosperous Granite was derived from the Burwash metasediments. Boyle (1961) infers that the mechanism involved the granitization of the sediments in place. Green et al. (1968), on the basis of the high strontium initial ratio of the granite, and the time available for the strontium ratio in the sediments to increase prior to the formation of the granite, suggested the granite formed due to the partial anatexis of the sediments. Drury (1979) suggested the similarity of composition of the granite to that of the minimum melt composition at 300 MPa in the Q-Ab-Or system is compatible with a partial melting origin for the granite, most likely of the metasediments. Both have similar rare-earth element patterns that together with low strontium and europium contents and the high K/Rb ratios of the granites are also consistent with a high degree partial melt



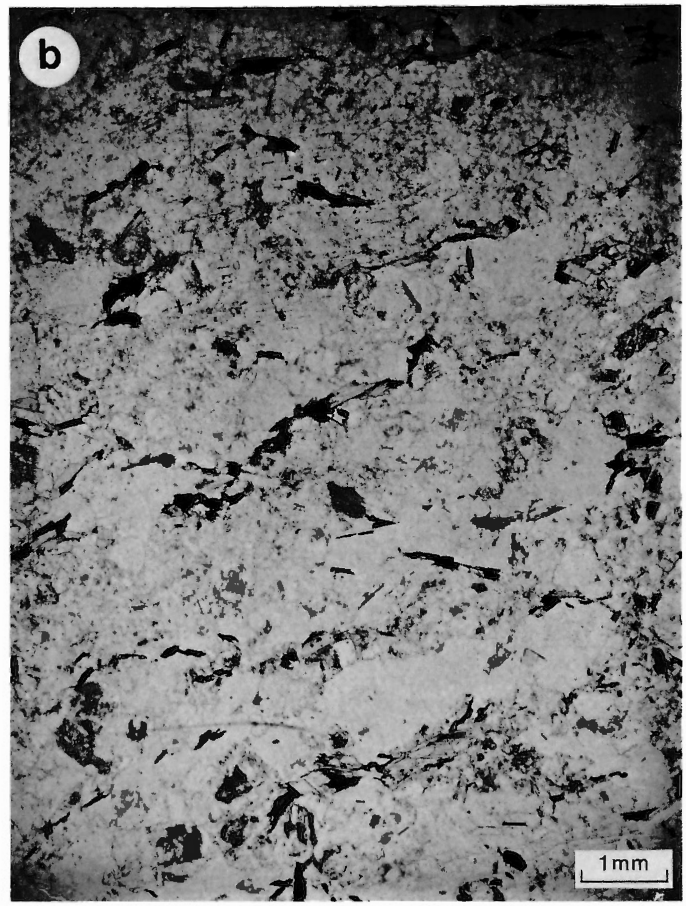


Figure 58. *The Prosperous Granite typically contains both biotite and muscovite and commonly has a weakly crushed aspect with irregular grain boundaries and strained to polycrystalline quartz. Microcline is abundant and occurs in broken aggregates. (a) Crossed polarizers. GSC 203660-D. (b) Plane light. GSC 203660-F.*

Figure 59. *Patches of uranophane stain on the otherwise lichen-covered surface of a Prosperous Granite outcrop east of Duncan Lake. The stain is typically bright yellow, forms on the highest part of the outcrop, and no lichen grows on it. GSC 177599*

removed from a plagioclase-rich residuum. Kretz et al. (1982) tentatively suggested that at least one of the Prosperous plutons (Prestige, south of Prestige Lake) was emplaced diapirically as a totally crystalline but plastic body. This was based on appropriate density requirements, textural features of the granite, and the contrast between apparent P-T metamorphic conditions of the adjacent country rocks and the minimum P-T conditions under which a melt crystallizing muscovite could exist, among others.

Redout Granite

The Redout Granite is a massive to irregularly foliated, fine- to medium-grained, equigranular, heterogeneous, biotite-muscovite granite. Pegmatite is commonly associated with it. It occurs in the southern part of the Sleepy Dragon Complex between Redout Lake and Turnback Lake as five closely spaced, commonly irregularly shaped, plutonic bodies that intrude both the heterogenous rocks of the Sleepy Dragon Complex and the Yellowknife Supergroup. The Redout Granite was first recognized by Henderson (1941b) but was not outlined in its entirety until Davidson (1972) mapped this part of the area. The following account is based mainly on the observations of Davidson.

Contact relations

Although the Redout Granite displays intrusive relationships towards all units with which it is in contact, these contacts are typically diffuse. The sharpest contacts

Table 19. Chemical analyses, Prosperous Granite

	1	2	3	4	5	6	7
SiO_2	75.30	75.32	73.86	74.09	74.28	72.10	71.44
TiO_2	0.02	0.01	.11	0.14	0.07	0.13	0.13
Al_2O_3	14.68	14.52	14.57	14.30	14.38	15.00	14.87
Fe_2O_3	0.82	0.71	.29	0.18	0.32		
FeO			.83	.99	0.77	1.56	1.54
MnO			.04	.02	0.04	0.03	0.05
MgO	0.07	0.01	.30	.31	0.16	1.21	1.23
CaO	0.40	0.35	.66	.82	0.60	0.91	0.88
Na_2O	4.62	4.39	3.50	3.61	4.18	3.93	3.82
K_2O	3.91	3.81	4.88	5.08	4.15	4.95	5.15
P_2O_5	0.16	0.12	.27	0.26	0.11	0.09	0.12
L.O.I.	0.81	0.73	.49	0.27	0.43		
S					tr		
CO_2					0.18		
Cl					0.02		

1,2 Prosperous Granite on road near Prelude Lake: Hill (1980), Appendix III analyses D280(1), D281(2), J.D. Hill analyst (total Fe as Fe_2O_3)

3 Prosperous Granite 62°40'N, 114°10'W: Folinsbee et al (1968) Table 1 analysis 4, D. Thaemlitz, analyst.

4 Prosperous Granite: Green et al. (1968), Table VI, L.R. Campbell, A. Rich, L. Yopik, analysts.

5 Prosperous Granite composite: Boyle (1961), Table 12, J.A. Maxwell analyst.

6,7 Prosperous Granite northeast of Prosperous Lake: Drury (1979) Table 1 Analysis 4(6), 5(7), G. Hendry, analyst (X-ray fluorescence; total Fe as FeO)

occur north and southwest of Redout Lake where the Redout Granite discordantly intrudes the Sleepy Dragon Complex, and also on the eastern arm of Turnback Lake where the granite intrudes Yellowknife supracrustal rocks. More commonly, however, the contact areas are migmatitic with irregular, swirly foliated older phases intimately mixed with highly varied proportions of dykes, sills and lenses to nebulitic masses of Redout Granite and associated pegmatite. Even where the contact is with the Yellowknife rocks, such as between the plutonic lobes in the vicinity of Turnback Lake, the supracrustal rocks become very nebulitic and grade into the Redout Granite. It is commonly difficult to differentiate between highly altered Sleepy Dragon rocks and Yellowknife rocks, as for example southwest of Turnback Lake.

Metamorphic effects

The Redout Granite lies within a metamorphic envelope that has been defined in the metasediments by the first appearance of cordierite in the vicinity of Victory Lake. There the metamorphic isograds are more or less parallel to the contact of the granite bodies and, in the case of cordierite, is about 5 km from it. Within the Sleepy Dragon Complex the granitic rocks were presumably metamorphosed to some extent during emplacement of the Redout Granite, but to date it has not been possible to distinguish between the various metamorphic events that contributed to the complex history of the Sleepy Dragon Complex. To the south and southeast of the Redout Granite the Yellowknife supracrustal rocks have been strongly metamorphosed, but again because of the occurrence of several other granitic units of possibly different ages in the vicinity, it has not yet been possible to relate the metamorphic pattern to particular intrusive events or a regional pattern.

Lithology

The Redout Granite is heterogeneous but in general is a pink to pinkish grey, fine- to medium-grained, equigranular, biotite-muscovite granite. The texture is variable over small areas and, although commonly fairly massive, does show a fabric developed to varied degrees particularly near its contacts with the migmatitic country rock. Inclusions are common, especially in the contact areas, and consist mainly of amphibolites and vague granitoid gneisses presumably derived from the Yellowknife supracrustal rocks and Sleepy Dragon Complex. In general the central parts of the plutonic lobes are free of inclusions and more homogeneous. Pegmatite is common as irregular patchy bodies commonly with diffuse margins and is particularly abundant towards the margin of the plutons.

In addition to the barren patchy pegmatites within the granite, pegmatites related to the Redout Granite, some with rare element concentrations, occur in the Sleepy Dragon Complex between Upper Ross Lake and Redout Lake. These have been described in detail by Rowe (1952) and Hutchinson (1957). The pegmatites show a regional zonation. Abundant large (up to 700 x 200 m) mineralogically simple pegmatites occur as concordant replacement bodies near the granite. Farther away they decrease in size, abundance and morphological complexity tending to be simple discordant vein-like bodies that formed due to injection of a fluid phase. Rare element minerals also show a zonal arrangement with pegmatites containing beryl, columbium-tantalum minerals and lithium minerals occurring at respectively increasing distances from the granite. In 1946 a mill was set up about 2.5 km east of Upper Ross Lake and operated intermittently for 2 years to concentrate columbium-tantalium minerals quarried from nearby pegmatites (Lord, 1951).

In thin section the Redout Granite consists of plagioclase, microcline, quartz, biotite, and muscovite with a granitic, equigranular texture. Individual phases tend to have irregular to somewhat sutured margins but are not granulated or milled. The plagioclase occurs as stubby subhedral grains generally less than 3 mm in size. They are not zoned but are somewhat altered with a rather patchy distribution of alteration products. Minor rounded inclusions of quartz and, less commonly, microcline are present in rare cases. Microcline also occurs as generally anhedral to subhedral grains of about the same size as the plagioclase. It commonly has an irregular shape but is not particularly interstitial to the other phases. It is very slightly perthitic in some cases and commonly contains inclusions of quartz and plagioclase. Quartz occurs as large irregular masses generally less than 5 mm in size. It is generally interstitial but in some cases appears to replace plagioclase. It also occurs as scattered rounded bodies generally about 0.5 mm in size scattered along contacts between coarser grains. The quartz is always strained and commonly fractured but not particularly polycrystalline. Biotite occurs as ragged individual to small aggregates of very dark brown flakes generally less than 1 mm in size. Muscovite is commonly associated with the biotite and has a similar habit although in some cases a few larger flakes are present.

Geochronology

Geochronological data available from the Redout Granite include five K-Ar determinations (recalculated) on muscovite at 2550 ± 28 Ma (2 determinations; Green and Baadsgaard, 1971), 2540 Ma (Stockwell, 1963), 2480 ± 28 Ma and 2460 ± 28 Ma (Green and Baadsgaard, 1971) and six on muscovite from the granite at 2490 Ma (Burwash and Baadsgaard, 1962), 2490 ± 28 Ma, 2460 ± 28 Ma, 2450 ± 27 Ma, 2380 ± 27 Ma and 2370 ± 27 Ma (Green and Baadsgaard, 1971). The average muscovite age is 2515 Ma and for biotite is 2440 Ma. The age of the Redout relative to the other granitic bodies is difficult to determine. It clearly is younger than the Sleepy Dragon Complex that it intrudes. On the other hand, it may be older than the nearby Morose Granite as Redout dykes and pegmatites, which are abundant in the contact areas of the pluton, were not seen to cut the Morose.

Origin

The Redout Granite differs from many of the other intrusive plutonic units in that its generally concordant contact relations are diffuse with migmatite phases developed in the adjacent country rock and diffuse, almost totally assimilated granitoid inclusions. This suggests relatively deeper levels of emplacement as these bodies have many characteristics of catazonal plutons (Buddington, 1959). On the other hand, these plutons have characteristics that they share with the Prosperous granites. These include the presence of both biotite and muscovite, the abundance of pegmatite in the granite and the occurrence of zoned pegmatites with rare element concentrations in the adjacent country rock. Although it has been suggested on the basis of isotopic and trace element data that the Prosperous Granite was derived from partially melting Burwash metasediments, the occurrence of the somewhat similar Redout Granite, mainly within a block of sialic basement to the Yellowknife supracrustal rocks (the Sleepy Dragon Complex), makes a similar origin unlikely.

Morose Granite

The Morose Granite is a massive, coarse grained, generally homogeneous, porphyroblastic, biotite-muscovite granite. Three rather widely spaced plutons have been included in the unit. The largest occurs about, and northeast of, Morose Lake in the core of the Sleepy Dragon Complex. A smaller body occurs to the north at the north boundary of the map area. A similar pluton that occurs on the west margin of the early Proterozoic Blachford Lake Intrusive Suite, north of Hearne Channel, also has been included in this unit. The southwestern half of the main pluton and the southern pluton have been mapped in detail by Davidson (1972). The remaining part of the unit is known only from isolated helicopter landings and air photo interpretation. The following remarks are based mainly on the observations of Davidson.

Contact relations

Contacts between the Morose plutons and adjacent units are, for the most part, sharp. The largest body about Morose Lake is intrusive into the granitoid rocks of the Sleepy Dragon Complex and along the western and northwestern boundaries of the body the contact is sharp with locally a narrow migmatitic zone in the adjacent Sleepy Dragon. In places coarse microcline porphyroblasts in the Morose tend to be aligned parallel to the contact. The contact between a somewhat different phase of the Morose southwest of Languish Lake, along the Ross River, is much less sharply defined, being more or less gradational with the adjacent granitoid gneisses over several kilometres. The relationship between the Morose and the two plutons tentatively assigned to the Prosperous Granite is not known. The contact between the Morose pluton at Hearne Channel and the adjacent Burwash Formation metasediments are sharp with only a few minor dykes at the north end. The Morose is younger than, and intrudes, the small Defeat granodiorite pluton at the southeast corner at Francois Bay. The original eastern margin of the Morose pluton is partly obscured by the intrusion of the early Proterozoic Caribou Lake Gabbro of the Blachford Lake Intrusive Suite, which has metamorphosed (spinel, hypersthene) and locally melted the granite (Davidson, 1978a). A large dyke related to the Caribou Lake Gabbro cuts across the southern part of the granite (Davidson, 1978a).

The metamorphic effects of the Morose plutons have not been defined. The northern two plutons have been emplaced in the Sleepy Dragon Complex which consists mainly of metamorphosed granitoid rocks, and the local metamorphic effects due to the intrusion of the Morose have not been separated from the presumably older regional metamorphism. Similarly to the south, the Morose pluton was emplaced into regional amphibolite grade metasediments, and no attempt has been made to separate the Morose metamorphic effects from any that preceded its emplacement or those that followed.

Lithology

The granite is typically massive except for a weak, local, marginal foliation, fairly homogeneous and light ranging from white to pink to pale brown or buff (Table 20, Fig. 60). Its most obvious characteristic is the occurrence of

Table 20. Modal analyses, Morose Granite

	1	2
Quartz	53	40
Plagioclase	20	35
Microcline	21	26
Biotite	3	6
Muscovite	4	4

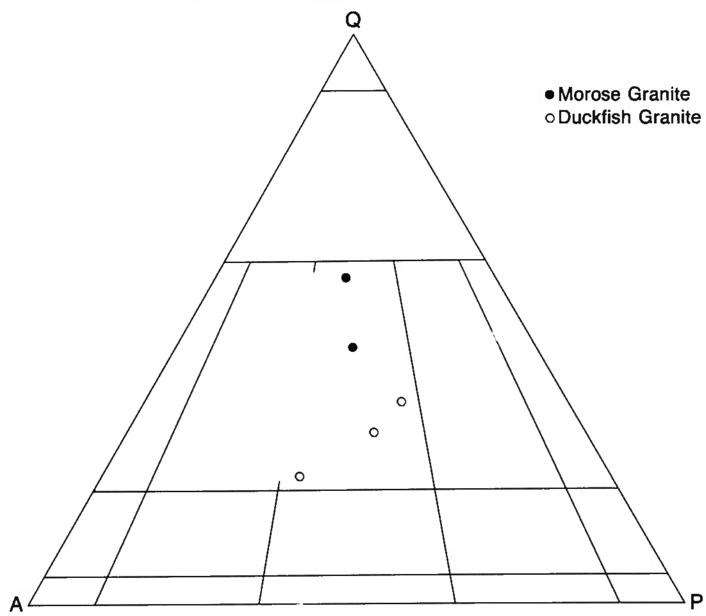


Figure 60. *Plot of modal analyses of Morose and Duckfish granites.*

abundant, elongate microcline megacrysts that are up to 3 cm in length. Both biotite and muscovite are present. Locally, minor pegmatite dykes are present consisting of a lighter coloured equigranular leucocratic granite that is probably related to the main phase. The southernmost lobe of the main body south and east of Languish Lake differs from the main granite. After a fairly abrupt transition the biotite-muscovite granite becomes finer grained, the megacrysts reduced in size and concentration, and the colour, a pinkish grey. Towards its southern and eastern margins it contains abundant, highly altered and deformed, inclusions of amphibolite to granitoid gneiss into which it ultimately grades. The main body of the Morose Granite on the other hand contains large rafts or roof pendants which, in the vicinity of Morose Lake, are up to several kilometres in size and are similar to the granitoid gneisses of the Sleepy Dragon Complex the body intrudes. Abundant large inclusions are also present in the southern pluton although these are generally metasedimentary schists. One large inclusion of Defeat granodiorite occurs near the north end of the pluton.

Two K-Ar ages have been determined from the southern pluton and include a muscovite age of 2201 ± 48 Ma and a biotite age of 2109 ± 47 (Davidson, 1979). Davidson (1979) suggests these ages represent updating by a later thermal event, possibly the emplacement of the adjacent Blachford Intrusive Suite at about 2.15 Ga. A thermal disturbance resulting in K-Ar data of this age may be more widespread as data from the Anton Complex, Defeat Plutonic Suite, Prosperous Granite and Amacher Granite result in similar ages (Fig. 46).

The Morose Granite has a rather weak expression on the Potassium Airborne Radioactivity map (Grasty and Richardson, 1972) and no expression on the aeromagnetic map.

Duckfish Granite

The Duckfish Granite occurs about the lake of that name 13 km north of Yellowknife Bay. The granite occurs as a single, roughly circular pluton 8 km in diameter. As the intrusion contains Defeat granodiorite-type inclusions locally along its western margin, it would appear to be intrusive into the Defeat. Its contact with the Anton Complex to the north is diffuse with extensive dykes and sills of the granite in the easterly trending intermediate composition Anton gneisses. The contact with the Kam volcanics, where seen in the vicinity of Chan Lake, is straight and sharp with no evidence of granite dykes in the volcanics or volcanic inclusions in the granite. The contact may be a fault.

The granite is a massive, medium grained, mauve-pink, leucocratic rock. The rather coarse quartz has an iridescent aspect. Joints are prominent in the southeastern part of the body. Inclusions are not common, but are more abundant at the margins with the adjacent granitic rocks, where they occur as small well-rounded bodies similar to the adjacent granitoid units. Pegmatites and aplitic dykes are rare. The contact with the volcanics is sharp, with a very fine-grained contact phase 15 cm wide noted at one locality. This textural change may be due to shearing rather than chilling. No dykes or small intrusions into the volcanics were seen. The contact to the north with the granitic gneisses is much more diffuse consisting of a zone with extensive sills and dykes of the granite in the gneissic rocks.

The granite consists of about 35% potassium feldspar, 30% plagioclase, 25% quartz, and less than 5% mafic minerals (Table 21, Fig. 60). The potassium feldspar typically occurs as large blocky crystals of euhedral to subhedral perthitic microcline up to 7 mm in size. The amount of albitic intergrowth is varied. The microcline also contains irregular inclusions of plagioclase, commonly zoned, up to

Table 21. Modal analyses, Duckfish Granite

	1	2	3
Quartz	19	29	33
Plagioclase	30	31	37
Microcline	40	2	24
Biotite	2	2	3
Muscovite		1	1
Chlorite	2	X	2
Perthite	6	35	1
Opaque	X	X	

5 mm in size. Subhedral plagioclase grains range in size from 7 mm to less than 0.5 mm. They are typically zoned, usually with what was originally a very calcic core, now almost obliterated by alteration products. Quartz has an interstitial habit and is typically polygonized into irregular aggregates, which causes its iridescent aspect in hand specimen. Very pale brown to greenish-brown biotite is the dominant mafic mineral and is typically chloritized and commonly very ragged in outline. Accessory minerals, including magnetite, apatite, sphene and zircon, are abundant and are typically associated with the biotite aggregates. The magnetite content is reflected on the aeromagnetic map of the area (Geological Survey of Canada, 1969b) which shows a distinct anomaly along the west margin of the pluton. The airborne gamma-ray spectrometry maps of the area (Grasty and Richardson, 1972) show anomalies in potassium, uranium and thorium along the marginal area of the northern half of the pluton, with the largest anomaly on the eastern side. Two samples from the northwestern part of this anomalous region contains sphene crystals up to 1.5 mm long that contain circular to rounded isotropic areas, up to 0.3 mm in diameter, that are presumably the result of radiation damage.

Most of the granite shows evidence of crushing. The coarser feldspars are surrounded by narrow zones of much finer broken aggregates of quartz and feldspar, and in many cases, myrmekite. Some of these coarse feldspars are fractured and have been injected with a similar aggregate of finer broken quartzofeldspathic material. The quartz is almost always extensively polygonized. Many of the alteration products of the plagioclase, in particular, are quite coarse (up to 0.15 mm), and moderately coarse secondary muscovite is present. It is not clear whether this is due to late stage deuteric alteration or perhaps a later thermal event.

The granite is younger than the adjacent granitic units and presumably the Yellowknife supracrustal rocks as well. Its age relative to the younger granitic units such as the Prosperous Granite to the east is not known. However if the alteration and internal deformation are related to a later thermal event, then the one associated with the emplacement of the nearby Prosperous granites is a possible candidate.

ARCHEAN MAFIC INTRUSIVE ROCKS

Mafic rocks of Archean age consist of three groups of dykes or sills and one plutonic unit. The Wool Bay Quartz Diorite, the plutonic unit, is considered part of the Defeat Plutonic Suite which is dominantly of granodioritic to tonalitic composition and has been briefly described in the section on the Defeat. The Archean dykes and sills have been grouped on the basis of their apparent age relative to the Yellowknife Supergroup and are described in the following sections.

Pre-Yellowknife Supergroup mafic intrusions

The oldest mafic bodies occur as highly deformed amphibolite layers and lenses within the granitic to tonalitic gneisses of the Sleepy Dragon Complex east of the Cameron River. These gneisses are thought to predate the accumulation of the Yellowknife Supergroup (Baragar, 1966; Davidson, 1972; this report). The mafic bodies are parallel to the foliation in the surrounding country rocks and Davidson (1972) reported tight intrafolial folds in these bodies in the plane of foliation. All primary textures and structures have been lost so that there is now little evidence as to their original nature. Ancient mafic dykes in a granitic terrane is a reasonable possibility. They are distinct from the clearly younger, much less metamorphosed, dykes and sills in the Sleepy Dragon Complex that are thought to be feeders to the Cameron River volcanics.

Syn-Yellowknife Supergroup mafic intrusions

Mafic sills occur with both the volcanics and sediments of the Yellowknife Supergroup. They are abundant in the volcanic sequences, are probably genetically related to the volcanics, and have been discussed in the sections dealing with the individual volcanic units (see also Lambert, in press). The mafic sills in the sediments, in contrast, are rare and occur in three main areas. Near Clan Lake mafic sills occur both southwest and southeast of the lake as well as in the adjacent volcanics. The sills are concordant with the folded strata, with the sills to the southeast occurring in a dome and basin structural pattern. A zone of mafic sills 10 km long and about 0.5 km wide occurs in the Tibbet Lake area. These sills differ from those in the other areas in that there are no mafic volcanics in the vicinity. On the basis of an interpretation of the structural geometry of the sediments they are contained in (Fig. 61, 62, 63, in pocket), it would appear they are relatively low in the sedimentary section, possibly close to the stratigraphic level of the volcanics 15 km to the east. Again the sills are concordant and range between 1-200 m in thickness. They follow the fold pattern in the sediments very closely and, like the sediments, tend to be thickened at fold noses. The concordancy of the sills and lack of undeformed dykes crosscutting, particularly in fold culminations, indicate the intrusions were intruded prior to folding. In the Consolation Lake area there is a single thick sill that is differentiated with a more felsic zone along the west margin of the body. The adjacent sediments in this area are steeply dipping but face west. This would further indicate that the sill was intruded and differentiated prior to the folding of the sediments.

The sills are compositionally uniform but texturally variable. In most cases the units are massive but where there is poorly defined layering it is due mainly to variation in grain size which can range up to several centimetres. Locally there is shearing parallel to layering. The sills are everywhere metamorphosed so that the original mineralogy and in most cases texture have been altered. The sills are now amphibolite with coarse laths of subhedral hornblende in a matrix of generally finer grained plagioclase. The plagioclase is varied in grain size and is commonly highly altered. The sill at Consolation Lake contains abundant biotite.

In the Sleepy Dragon Complex east of the Cameron River and Upper Ross Lake is a suite of mafic dykes, sills and irregular bodies in a zone several kilometres wide parallel to the volcanic-Sleepy Dragon Complex contact. Most of these bodies are roughly parallel to the contact, changing trend from northwesterly east of Upper Ross Lake to northeasterly east of the Cameron River to the north. They are most abundant in the vicinity of Upper Ross Lake (Fig. 5). Also present are a few dykes that strike directly towards the volcanic contact. This mafic intrusive complex is thought to represent a feeder system to the adjacent Yellowknife volcanics (Baragar, 1966; Lambert, 1982) and so are considered contemporaneous with volcanism and possibly the previously discussed sills in the sediments and volcanics.

The dykes and sills are varied in thickness ranging to in excess of 20 m. All are metamorphosed such that the original mineralogy is replaced by a hornblende-plagioclase assemblage, and in most cases primary microscopic textures are gone. Locally, grosser features such as chilled margins and coarse plagioclase phenocrysts entrained in the central part of the dyke have survived (Fig. 6). The contacts between these intrusions and the country rock are commonly sheared with the margins marked by biotite schist. The sills and irregular bodies conform in shape to the fabric in the country rock and are commonly sheared, particularly along their margins, by later movement parallel to the early foliation. The dykes at a high angle to the gneissic foliation are also locally offset. It was not found possible to trace the dykes across the granitic-volcanic contact due to subsequent movement on the contact, although similar dykes and sills are found within the volcanic sections as well.

Post-Yellowknife Supergroup(?) mafic intrusions

In the Defeat Plutonic Suite west of Yellowknife, and to a lesser extent in the Stagg Plutonic Suite, are metamorphosed mafic rocks that range from narrow zones of amphibolite inclusions to well preserved dykes. The mafic blocks, up to several metres in size, occur in a matrix of granodiorite in sharply defined linear zones. The individual blocks are sharply bounded by curved to rounded surfaces suggesting plastic deformation (Fig. 64). They vary from massive, homogeneous, medium grained amphibolite to laminated to weakly compositionally layered amphibolite. Less abundant, but also present, particularly in the Stagg Plutonic Suite, are tabular mafic bodies that are clearly dykes. They have sharp, straight to only moderately irregular

Figure 64. *Deformed and metamorphosed mafic dyke in the western part of the Defeat Plutonic Suite.*

contacts and in some cases relict chilled margins. As with the narrow zones of amphibolitic inclusions, the original mineralogy and texture is completely altered and now consists of a blocky assemblage of hornblende and plagioclase. These metamorphosed dykes are also cut by narrow veinlets of granitic material.

The occurrence of these dykes in granitic rocks has important implications for the evolution of the granitic rocks, as the deformed and metamorphosed state of such dykes has been interpreted in a variety of ways. As discussed by Watterson (1965) these include (a) the granitization of supracrustal rocks originally containing mafic dykes that resisted and survived the granitization process; (b) the stoping away of the original mafic dyke-bearing rocks by intrusive granite; (c) the remobilization of a pre-existing dyke-bearing granitoid terrane by a subsequent thermal event, and (d) the emplacement of mafic dykes into newly emplaced granite bodies that were still hot and, although solid enough to be capable of fracturing to allow the emplacement of the mafic material, retained enough heat and fluid to plastically deform the dykes and alter their primary mineralogy (i.e. Roddick and Armstrong, 1959; Roddick, 1965). It has already been suggested that both the Defeat and Stagg plutonic suites which contain dykes of this type had a complex history that includes, for example, the suggestion that the Defeat was derived by granitization of pre-existing supracrustal rocks (Boyle, 1961) or by remobilization of gneissic granitoid basement (this report). A detailed study of these dykes may provide in addition to a better understanding of the dykes themselves, an important insight into the evolution of the plutonic rocks they are contained in.

PROTEROZOIC INTRUSIONS

Early Proterozoic Blachford Lake Intrusive Suite

The Blachford Lake Intrusive Suite is a multiphase intrusion of alkaline character situated on the north shore of Hearne Channel. The intrusion was recognized and briefly described as an alkalic complex by Davidson (1972) and Badham (1979) and, after further mapping at 1:50 000 scale in 1977, was more completely described and formally named (Davidson 1978a, 1981). Petrochemical data on the Blachford Lake are available on open file (Davidson, 1981) and have been discussed by Davidson (1982). K-Ar geochronological data from the suite are described and interpreted by Davidson (1979). The following description is abstracted entirely from his work.

The Blachford Lake intrusions consist of six major units that range in composition from the mildly alkaline Caribou Lake Gabbro on the west through relatively younger and increasingly alkaline diorite, Whiteman Lake Syenite and Hearne Channel and Mad Lake granites to highly peralkaline Grace Lake Granite and Thor Lake Syenite bodies that form the easternmost and largest units. The suite intrudes all the Archean units in the area. These include the metamorphosed Burwash greywackes, small satellite plutons of Defeat granodiorite and the southern body of the Morose Granite. Sharp, angular, well preserved inclusions of Archean country rock, particularly the metasediments, are found locally. The intrusions were emplaced at very high crustal levels so that there is only a very narrow contact aureole (Davidson, personal communication, 1978). The suite is intruded by both the north-northwesterly trending Mackenzie diabase dykes, dated elsewhere at about 1200 Ma and a set of east-northeasterly trending diabase, the Hearne dykes, that are particularly common within about 7 km of Hearne Channel between Blachford Lake and the east border of the map area. Preliminary radiometric dating of various units of the suite by K-Ar methods suggest an age of about 2.15 Ga (Davidson, 1979).

Caribou Lake Gabbro

The Caribou Lake Gabbro consists of a narrow body that forms the western and northern margin of the Blachford Lake Intrusive Suite and a narrow dyke-like structure that intrudes the adjacent Archean Morose Granite. Gradational with this marginal phase is an irregularly shaped remnant of leucodiorite southeast of Caribou Lake. The unit was probably originally more extensive but much has been lost due to the intrusion of later units of the suite. Large inclusions of the diorite occur in the Hearne Channel and Grace Lake granites to the east. The gabbro is texturally and compositionally zoned from west to east. The marginal phase, consisting of massive olivine gabbro of irregular grain size with local pegmatitic patches, is chilled against the Archean units to the west and north that now contain a high temperature contact metamorphic mineral assemblage (cordierite-hypersthene-spinel). This is followed at Caribou Lake by a massive to faintly layered, medium- to fine-grained, equigranular olivine gabbro, locally with plagioclase laths aligned parallel to layering and dipping steeply easterly. To the south, a noritic gabbro lacking olivine is present. In both types, interlayered pegmatitic gabbro with coarse magnetite lenses, medium grained gabbro with thin magnetite layers and fine- to medium-grained sulphide-bearing gabbros follow. Finally pegmatitic hornblende-clinopyroxene gabbro and leucogabbro grade into a uniform coarse massive leucodiorite, containing variable proportions of fayalite, ferroaugite, hornblende and biotite, that lies southeast of Caribou Lake and makes up approximately half the unit. Within both gabbro and leucodiorite, particularly south and southeast of Caribou Lake, are rounded to angular inclusions of anorthosite and anorthositic gabbro that Davidson (1978a) suggests represent an early crystallized phase not exposed in place at the present level of erosion. Niccolite-bearing carbonate veins that cut both the gabbro and dykes from the adjacent granitoid units occur in 3 places south of Caribou Lake (Henderson, 1939; Davidson, personal communication, 1981).

In thin section the gabbros consist of pyroxene with brown hornblende overgrowths and plagioclase that ranges in composition from An_{55-40} in the gabbros to An_{40-25} in the leucodiorite where the sodic varieties are commonly antiperthitic. Small amounts of primary deep red-brown biotite indicative of the alkaline nature of the rock, are also present. Forsteritic olivine is restricted to the contact region while fayalitic olivine is present in the leucogabbro and leucodiorite. Intergrown orthopyroxene and clinopyroxene occur in the noritic gabbro.

Whiteman Lake Quartz Syenite

The Whiteman Lake Quartz Syenite occurs south of Whiteman Lake where it intrudes the Archean Defeat granodiorite and the Caribou Lake Gabbro and is intruded by the Grace Lake Granite to the east and the Mad Lake Granite to the south. In the central part of the unit are abundant inclusions of metasediments, granodiorite and Caribou Lake Gabbro. The same compositional and textural zonation seen in the Caribou Lake Gabbro to the southwest can be seen in the distribution of gabbro inclusions in the syenite, suggesting the emplacement of the body was relatively passive. Elsewhere the body is relatively free of inclusions except for minor rounded bodies near the margins.

The unit consists of two phases. The older, which occurs mainly on the southeast, is a dark green syenite with little or no quartz and with fayalite and green clinopyroxene. The larger perthitic grains have antiperthitic oligoclase cores. The younger phase is dominant in the north and west parts of the body and consists of a light green to pinkish-grey, brown and crumbly weathering, massive, uniform, medium grained, hornblende-perthite-quartz syenite that also contains biotite, clinopyroxene and altered fayalite locally. Around Mad Lake both phases are present with the coarse green fayalite-pyroxene syenite, which often is difficult to distinguish from the leucodiorite phase of the Caribou Lake Gabbro, occurring as sparse xenoliths in the medium grained quartz syenite. Dykes of the quartz syenite intrude the Caribou Lake Gabbro to the southwest.

Davidson (1979) reported a hornblende K-Ar age of 2127 ± 79 Ma and a Rb-Sr isochron age of 2080 ± 42 Ma (Wanless, personal communication, 1978).

Hearne Channel Granite

The Hearne Channel Granite occurs on the north shore of Hearne Channel near the southeastern margin of the alkalic complex. It intrudes the Caribou Lake Gabbro that lies to the west and north, and is intruded by both the Mad Lake Granite to the northeast and the Grace Lake Granite to the east. The large body of Caribou Lake-type gabbro in the centre of the intrusion is probably a roof pendant.

The granite is pinkish brown, massive, medium grained and subequigranular. It is compositionally uniform and except for the large roof pendant is generally free of inclusions. The feldspars are commonly perthitic with larger grains containing antiperthitic plagioclase cores and in some there are alternating zones of perthite and antiperthite. The matrix consists of a micrographic quartz-feldspar intergrowth. A diagnostic feature of the unit is scattered xenocrysts of dark andesine that are probably derived from a phase of the Caribou Lake Gabbro. In some cases they are overgrown by perthite or antiperthitic oligoclase. Iron-rich hastingsitic hornblende is the main mafic mineral but biotite (annite) is also commonly present (Davidson, 1981).

Mad Lake Granite

The Mad Lake Granite occurs as two small bodies, one of which is intrusive into the Caribou Lake Gabbro and Whiteman Lake Quartz Syenite, while the other, to the south, is intrusive into the Caribou Lake Gabbro, but seems to grade into the Hearne Channel Granite. Davidson (1978a) has suggested that only the top of the body is exposed at the present erosion level and that the two are one at depth. In both locations they are intruded by the Grace Lake Granite to the east.

The granite is pink, massive and uniform with subporphyritic texture. The phenocrysts are perthite with antiperthitic cores. Quartz also occurs as small partly resorbed phenocrysts. The matrix consists of fine- to medium-grained sodic oligoclase, microcline, quartz, hastingsitic hornblende and biotite.

Davidson (1979) reported a biotite K-Ar age of 2166 ± 47 Ma.

Grace Lake Granite

The Grace Lake Granite, the largest unit in the Blachford Lake Intrusive Suite, forms a circular intrusion 18 km in diameter. Its central part is occupied by the Thor Lake Syenite which is in gradational contact with the granite over at most a few metres. The granite intrudes all the other units with which it is in contact. It contains large inclusions of the Archean metasediments and, in the southwest part of the intrusion, inclusions of Caribou Lake Gabbro. In its west-central part, the granite is intruded by two small stocks of diorite and quartz monzonite, correlated with the Compton laccoliths in the East Arm of Great Slave Lake (i.e. on Blanchet Island). In the northern part of the intrusion are flat lying lenses of pyroxene-bearing syenite. Dykes of the granite and minor pegmatitic phases occur locally, and intrude adjacent older phases.

The fresh granite is light grey to pale greenish grey but with increasing degree of alteration becomes buff to reddish pink. It is a massive coarse grained equigranular rock with euhedral mesoperthitic feldspar as the major constituent. Quartz forms about a quarter of the rock and decreases slightly towards the centre of the body. The major mafic mineral is sodic amphibole ranging from ferrorichterite to riebeckite, forming about 7% of the rock, but increasing slightly towards the centre. Acmite and astrophyllite occur locally and biotite is present only as a fine grained secondary mineral. Fluorite and zircon are the only notable accessories.

K-Ar ages include 2057 ± 56 Ma on riebeckite (Davidson, 1978a) and 2133 ± 112 Ma also on a riebeckite (Davidson, 1979).

Thor Lake Syenite

The Thor Lake Syenite occupies the central part of the Grace Lake Granite with which it has abruptly gradational contacts. It is similar in many respects to the Grace Lake Granite, the main difference being an abrupt decrease in quartz content. Also angular inclusions of the Archean metasediments are much more abundant in the syenite, in particular in the northeast part. Davidson (1978a) has defined 5 types of syenite based on textural and compositional variations. Three form a general concentric zonation while the other two occur on the west side. The narrow outermost zone occurs on the northern and eastern margin of the syenite and is a mafic fayalite-hedenbergite syenite. The inner two zones and two phases on the west side are generally less mafic ferroricherite syenites that differ mainly in the texture of the amphibole. The contacts between the phases are generally gradational over a few metres and appear to be steep. Associated with the Thor Lake Syenite are veins of alkaline pegmatite and related albite-rich rocks that are abundant in the northwestern part of the syenite (Davidson, 1978a, 1982). Patchy mineralization, particularly within a 2 km^2 triangular area within this part of the body contains concentrations of Nb, Be, Li, Th, Y, U and rare-earth elements in an assemblage of fluorite, carbonate, quartz, zircon and rare minerals (see section on Economic Geology).

K-Ar data are from metamorphosed inclusions within the syenite and include a biotite age of 3240 ± 59 Ma (anomalously old due to excess argon) and a whole-rock and biotite age from a second inclusion of 2150 ± 58 and 2064 ± 46 Ma respectively (Davidson, 1979).

On the basis of petrochemical data on the Blachford Lake Suite, Davidson (1982) suggests that each intrusion of the suite represents separate intrusions of magma derived from a fractionating source at some deeper level, so that the youngest, most highly fractionated peralkaline Grace Lake Granite and Thor Lake Syenite spent the longest time lower in the crust prior to being emplaced into their present location. He points out that the Blachford Lake complex is similar in many respects to intrusive complexes elsewhere that are related to thermal doming associated with a proto-rift environment.

Early Proterozoic Compton Intrusive Suite

The Compton Intrusive Suite (Hoffman, 1981) consists of dioritic to monzonitic intrusions that occur along an east-northeast-trending zone over a distance of 250 km through the East Arm of Great Slave Lake. To the northeast beyond the map area the intrusions are more quartz monzonitic while diorite is abundant in the west (Hoffman et al., 1977). Within the present area rocks of the suite occur mainly on Blanchet Island as 4 discrete bodies. In addition two small Compton stocks occur south of the west end of Blachford Lake in the Grace Lake Granite, and a small stock outcrops on the small Island in Hearne Channel south of Narrow Island. This latter and the smaller southwestern stock on the mainland, like the main intrusions on Blanchet Island, are diorites while the larger northern stock in the Grace Lake Granite is of more quartz monzonitic composition. According to Hoffman et al. (1977), the intrusions occur as laccoliths at the contact between the Pethei Group and the overlying Stark Formation. The contact with the underlying Pethei is generally flat but is steep against the adjacent Stark Formation where locally autoclastic blocks of the diorite occur in a breccia in the immediate contact area. Where the top of the intrusion is preserved, it typically contains highly altered blocks of Stark breccia. The diorite is a dark, massive rock consisting of euhedral laths of plagioclase, interstitial hornblende and biotite with varied amounts of quartz. The rock is commonly highly altered with the plagioclase obscured by alteration products and the mafic minerals now consisting of chlorite.

Hoffman et al. (1977) suggest the intrusion of the laccoliths into the Great Slave Supergroup postdates nappe formation in the East Arm which occurred after accumulation of the Great Slave Supergroup was complete. The laccoliths are also younger than the Blachford Lake Intrusive Suite and the east-northeasterly trending Hearne dykes between Blachford Lake and Hearne Channel, as they intrude both. Hoffman et al. (1977) also suggest folding of the Great Slave Supergroup was not complete when the laccoliths were intruded as the steeper dips of the country rocks flatten beneath the laccoliths. The intrusions have a zircon radiometric age of 1865 Ma (Bowering and Van Schmus, 1982), a biotite K-Ar ages of 1795 ± 55 Ma (Hoffman, 1970) and 1864 ± 45 Ma (Davidson, 1979), a hornblende K-Ar age of 1705 ± 58 Ma (Davidson, 1979) and a Rb-Sr isochron of 1811 ± 78 Ma (Wanless, personal communication, 1979).

Proterozoic mafic dykes

Several post-Archean diabase dyke sets occur within the map area. Except for faulting they are all undeformed and primary mineralogy and textures are for the most part well preserved. The dyke groupings have not been distinguished on lithological criteria alone, but have been assigned to the various sets in large part on the basis of their orientation. This presents problems in the case of dykes with an east-northeasterly trend as at least two groups with different ages have been recognized within the area. Many of the Proterozoic dyke sets have been described and geochronological and paleomagnetic data presented by Burwash et al. (1963), Leech (1966), McGlynn and Irving (1975), Gates and Hurley (1973), and Fahrig and Wanless (1963). The dykes are much more abundant in the granite terrane, a reflection of the contrasting mechanical properties of the granitic and metasedimentary rocks.

Dogrib dykes

The east-northeasterly trending Dogrib dykes named by McGlynn and Irving (1975) include the Set 1 dykes of Burwash et al. (1963) and Leech (1966). They consist of rather widely separated individual dykes or small groups of dykes. The dyke through Russell and Stagg lakes is part of the Dogrib set as is the group that passes from west of Yellowknife to east of Gordon Lake. The latter group is most completely mapped, and consists of two discontinuous dykes that pinch out, splay off and in places intersect each other. A minor group of relatively thin dykes occurs in the central part of the granitic terrane west of Yellowknife. The Dogrib dykes are typically the widest dykes in the area with widths up to 180 m in some cases. Some of the very thick dykes, such as the segment east of Ryan Lake, are composite, consisting of as many as four separate intrusions, each with chilled margins against the adjacent earlier diabase. The diabase consists of equal proportions of pyroxene and plagioclase with ophitic texture with minor altered olivine and iron oxides (McGlynn and Irvine, 1975). These dykes have a rather weak aeromagnetic expression relative to their size, and contrast with the similar trending but younger Hearne dykes, north of Hearne Channel, which although generally narrower, have a more pronounced magnetic expression. Considerable effort has gone into determining the age of the Dogrib dykes but, as is commonly the case with mafic dykes the results have not been particularly satisfactory. K-Ar ages range between 910 and 2385 Ma (Fig. 65; Fahrig, 1972; Leech, 1966), while Gates and Hurley (1973) on the basis of an Rb-Sr isochron report an age of 2635 ± 80 Ma ($\lambda^{87}Rb = 1.42 \times 10^{-11}\ yr^{-1}$) although the Prosperous Granite which is intruded by the dyke has a Rb-Sr mineral isochron age of 2520 ± 25 Ma ($\lambda^{87}Rb = 1.42 \times 10^{-11}\ yr^{-1}$) (Green et al., 1968). Paleomagnetic data comparing the Dogrib dykes with the Indin dykes (discussed below), whose K-Ar ages are not really significantly different from those of the Dogrib dykes (Leech, 1966), are not incompatible with significantly different ages for the two sets as the paleopoles are separated longitudinally by about 5500 km (McGlynn and Irving, 1975).

Milt diabase

Gently dipping to horizontal sheets of diabase occur at several localities within the map area. They are most abundant in the vicinity of Detour Lake, Patterson Lake and Sleepy Dragon Lake, and are named the Milt diabase sheets as an extensive, locally well exposed example occurs south of the Milt Lake in the northeastern part of the area. Other occurrences are in the shorter of the two eastern arms of Jennejohn Lake, south of the mouth of the Francois River and between Russell Lake and the 90 degree bend in Stagg River. Like the diabase dykes they are mineralogically and texturally fresh for the most part but differ in that they form sub-horizontal sheets with dips less than 10 degrees. The sheets are generally less than 2 m thick and in most cases occur singly, although this is difficult to tell due to the lack of relief in the area. East of Yellowknife between the Yellowknife River and Duck Lake is a shallow, easterly dipping mafic sheet that has intruded the complexly folded Burwash metasediments. It has been described by Leech (1965, 1966). As can be seen in the road cut across the sheet it is differentiated into an underlying olivine-rich rock (west side of fault) and an overlying gabbro (east side of fault). Eight K-Ar whole-rock and two biotite ages range between 1495 and 2473 Ma (Fig. 65; Fahrig, 1970; Henderson, 1978b; Leech, 1966) and are not considered significant as far as age of intrusion is concerned.

Indin dykes

The Indin dykes as named by McGlynn and Irving (1975) include Sets II and IV of Burwash et al. (1963) and Leech (1966) and are the most abundant in the area. The Indin dykes have been considered to occur in two conjugate sets with northwesterly and north northeasterly trends by Burwash et al. (1963) and McGlynn and Irving (1975). These two directions are evident on the 1 inch:1 mile maps at Yellowknife (Jolliffe, 1942, 1946; Henderson and Brown, 1966) and, in particular, in the Indin Lake region some 200 km north-northeast of Yellowknife (Wright, 1954). Diabase dykes with these trends also occur in the Sleepy Dragon Complex (A. Davidson, unpublished maps 1971; Fortier, 1947). However, in the dominantly granitoid terrane between Yellowknife and Russell Lake, the trends of some 305 dykes (Fig. 66) do not separate out into two distinct populations, but occur within a 90 degree sector that includes the two Indin trends so evident elsewhere. The dykes are well

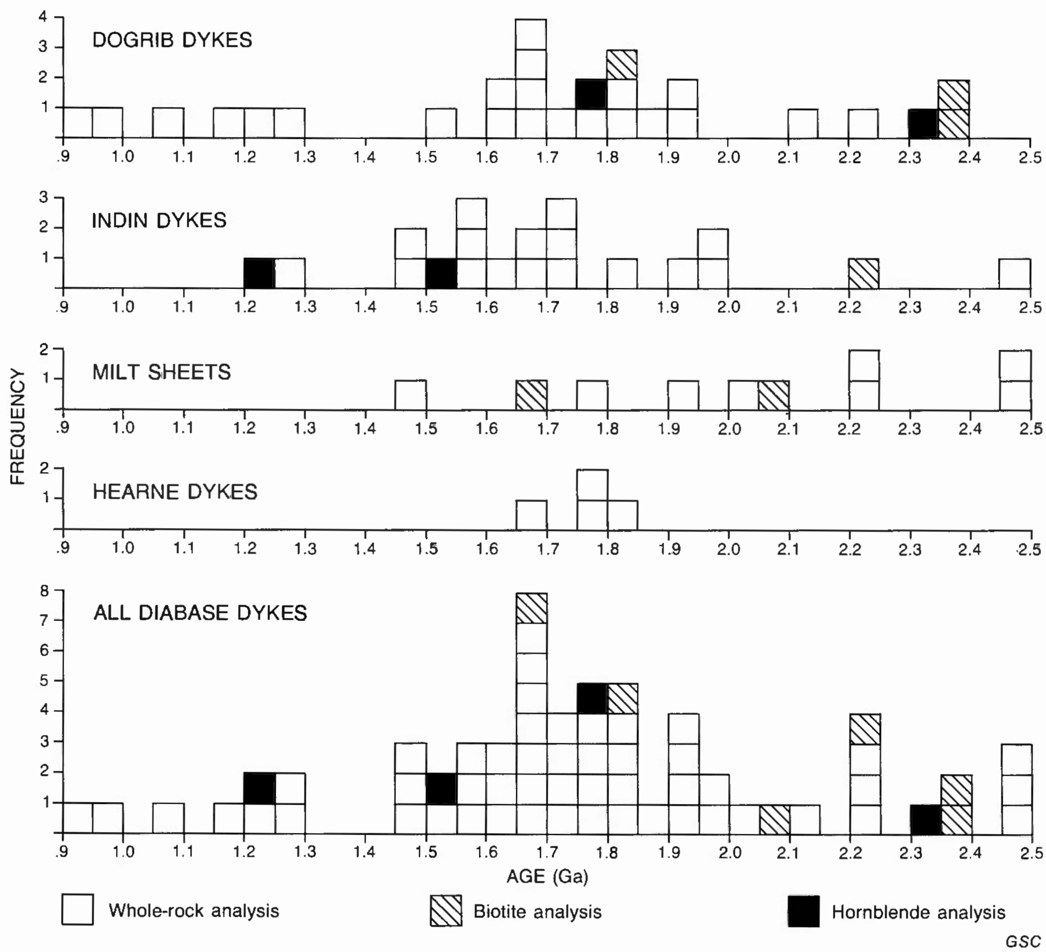


Figure 65. *Histogram of K-Ar whole-rock ages for the Dogrib, Indin, Milt and Hearne diabase dyke sets within the area. Data from Leech (1966), Fahrig (1965, 1972), Henderson (1978b) and Davidson (1979).*

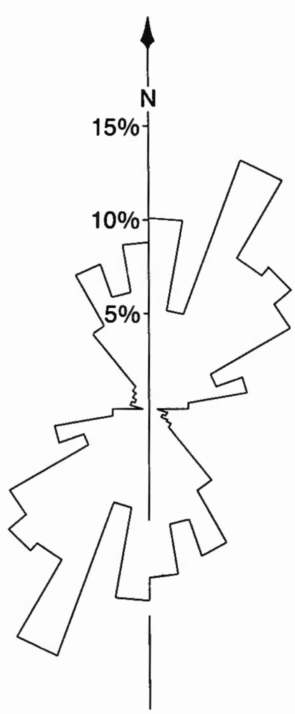


Figure 66. *Rose diagram showing the polymodal trend of 305 mainly Indin dykes of all sizes in the granitoid terrane northwest of Yellowknife.*

preserved with chilled margins, primary textures and original mineralogy. Individual dykes are thinner than the Dogrib dykes, for the most part being less than 50 m, although in number and volume they are much more important. They would also appear to have a much more regional distribution than the Dogrib dykes which, in general, are rather isolated in their occurrence. The Indin dykes do not have an aeromagnetic expression which helps to differentiate them from the younger, somewhat similar trending Mackenzie dykes. Leech (1966) and McGlynn and Irving (1975) found that there was no difference in K-Ar ages and paleomagnetic pole directions between the two Indin dyke trends which, together with conflicting crosscutting relationships between the two trends (Wright, 1954), supports the suggestion that they are conjugate. Henderson and Brown (1966) show that the Indin dykes are younger than the Dogrib dykes and the 5500 km difference in paleomagnetic pole positions for the two (McGlynn and Irving, 1975) suggests the Indin dykes are significantly younger than the Dogrib. K-Ar ages of the Indin dykes range between 1215 and 2495 Ma (Fahrig, 1965; Leech, 1966) while Rb-Sr data range between 2067 and 2174 Ma (Gates and Hurley, 1973).

Hearne dykes

On the north shore of Hearne Channel east of Francois Bay is a set of east-northeast-trending diabase dykes. Although similar in trend to the Dogrib dykes through

Yellowknife and Stagg Lake, they differ from them in that they have a pronounced aeromagnetic expression. They intrude the 2.15 Ga Blachford Lake Intrusive Suite (Davidson, 1981) which suggests they differ in age from the Dogrib dykes as well. They are offset by the left lateral transcurrent faults in the area, and neither the dykes nor the faults are known to involve the rocks of the early Proterozoic Great Slave Supergroup. Davidson (1978a) has suggested that these dykes are correlative with the similarly trending Hearne dykes (Hoffman, 1980, 1981; Hoffman et al., 1977, therein named Simpson dykes) in the pre-Great Slave Supergroup rocks of the western and eastern East Arm of Great Slave Lake. Although most of these dykes do not have an aeromagnetic expression as do those north of Hearne Channel, this may be due to the thinness of many of them and the fact that they are somewhat altered (Hoffman, personal communication, 1982). The emplacement of the Hearne dykes may be related to extension associated with the formation of the Athapuscow Aulacogen (Hoffman, 1981).

Mackenzie dykes

The Mackenzie dykes are the youngest set in the area. To the north they are thought to be related to the Coppermine lavas and the Muskox intrusion all of which were intruded or extruded 1100 to 1200 Ma ago (Fahrig and Jones, 1969). The dykes have a north-northwesterly trend, are particularly abundant through the central part of the Slave Province, but are less abundant to the southwest within the map area. Their trend is similar to that of the Indin dykes (northwest set) but they can be distinguished by their stronger aeromagnetic expression. The Mackenzie dykes are restricted to the easterly part of the map area where only a few examples are present. They occur through the southwest end of Wedge Lake, through Prestige and Pensive lakes and along the southwest shore of Gordon Lake. Others occur through east-central Desperation Lake, a few kilometres west of Payne Lake, and between the north end of Hearne Lake and west end of Blachford Lake.

Geochronology

Determining the age of intrusion of mafic dykes is difficult. Various attempts have been made and are summarized in Figure 65. Most of the data comes from K-Ar whole-rock methods and Leech (1966), who has done the most extensive work on the diabase dykes in the Slave Province, discussed the problems involved in some detail. She found that the older the dyke the more difficult it was to determine its age of intrusion or, as stated in Leech's Law, "Updating and uplifting have the same effect; old dykes and old ladies give young ages" (Leech, 1966, p. 410). With considerable reservations she suggested the oldest ages may represent the age of intrusion of the various sets. However, if the frequency of ages is plotted against time there is really little to distinguish between the resulting pattern for any of the dyke sets (Fig. 65). The K-Ar data appear to have little to do with the age of intrusion of the dykes but are a reflection of argon loss (or gain) of the dykes due to subsequent thermal effects in the crust after they were emplaced. The Milt diabase sheets, for which there are few data available, were presumably exposed to similar conditions. Although no data are available from within the area, Mackenzie K-Ar ages from elsewhere are much more closely grouped and may be more significant with regard to age of intrusion (Leech, 1966).

In addition there are some difficulties in the interpretation of the Rb-Sr data from the mafic dykes. The Dogrib dykes have been dated at 2635 ± 80 Ma (Gates and Hurley, 1973) by this method, which is 115 Ma older than the 2520 ± 25 Ma Rb-Sr age of the Prosperous Granite (Green and Baadsgaard, 1971) that they intrude. Similarly their ages on the Mackenzie dykes are about 140 Ma older than the preferred K-Ar estimates. The apparently older Rb-Sr ages may be due to contamination of the diabase by the country rock as has been suggested in other areas by Patchett et al. (1979) and Baragar and Loveridge (1982).

EARLY PROTEROZOIC GREAT SLAVE SUPERGROUP

Early Proterozoic sediments and volcanics of the Great Slave Supergroup and associated intrusive rocks occur mainly on islands in the East Arm of Great Slave Lake, in the southeast of the map area, separated from the Archean terrane by Hearne Channel. These and other exposures of presumed early Proterozoic rocks are related to Wopmay Orogen (Stockwell, 1970; Fraser et al., 1972; Hoffman, 1980a) which extends south along the west margin of the Slave Province. The Great Slave Supergroup occurs in the preserved remnant of a northeast-trending basin known as the Athapuscow Aulacogen (Hoffman, 1973b; Hoffman et al., 1974), which is at a high angle to the trend of the orogen. The aggregate stratigraphic thickness within the structure is about 12 000 m, while the thickness of the deposits on the adjacent craton is only about 1700 m (Hoffman, 1974). In general, stratigraphic thicknesses increase down the axis of the structure towards the southwest. Sediment transport was strongly controlled by the structure (Hoffman, 1969).

The geology of the Great Slave Supergroup has been described and discussed by Stockwell (1936), Hoffman (1968, 1969, 1973a,b, 1974, 1977, 1981), and Hoffman et al. (1974, 1977). The following account of that part of the map area underlain by Great Slave Supergroup rocks is based almost entirely on these works.

Within the map area the units of the Great Slave Supergroup include both the thinner cratonic sediments and the generally thicker sediments deposited within the aulacogen itself (Fig. 67). For the lower part of the section, this differing structural environment is expressed mainly in the varied thickness of the units. Due to lack of exposure, this is not evident within the map area. For the upper part of the section, however, in the Pethei Group there are profound facies changes at the transition from platform to basin. This transition is well developed on Blanchet Island, which is underlain mainly by formations of the Pethei Group.

Structure

The depositional and structural evolution of the East Arm area during the early Proterozoic are closely interrelated. However, detailed discussion of it is beyond the scope of this report. A general outline for the evolution of the area is provided by Hoffman (1973b, 1981) and Hoffman et al. (1974, 1977).

The Great Slave Supergroup is preserved in a large canoe-shaped asymmetric synclinorium. That part within the map area occurs mainly on the very gently dipping north limb of the structure. The first major synclinal axis occurs just south of the northeasterly part of Blanchet Island. The southern limb dips considerably more steeply than the northern limb.

A major structural event that involved rocks in the Blanchet Island region was the emplacement of large nappes (Hoffman et al., 1977). The nappes, one of which forms the southern part of Blanchet Island, moved in a northwesterly direction and this deformation has moved basinal Pethei and underlying Kahochella rocks towards the platformal Pethei, thus considerably telescoping the original facies distribution.

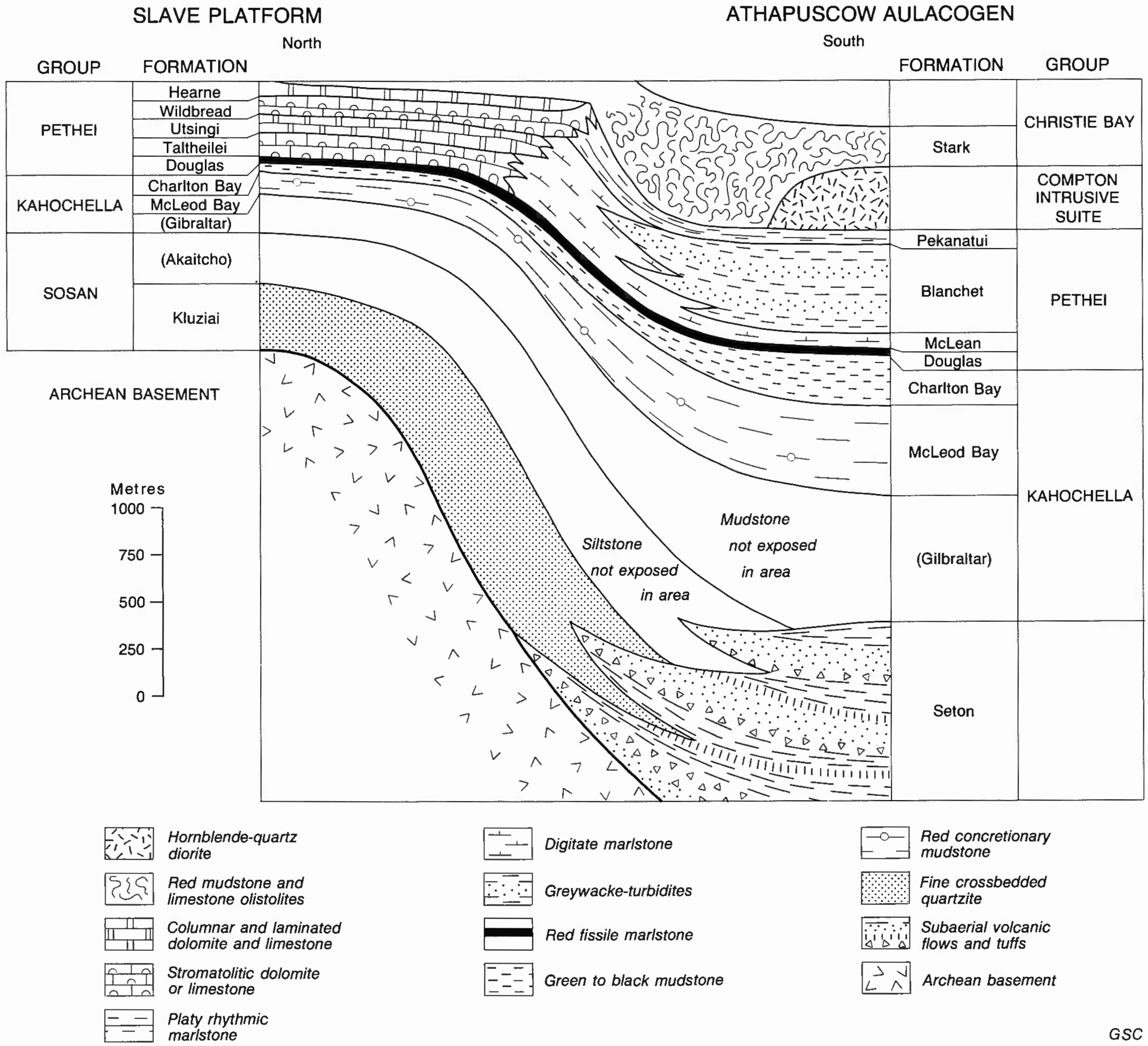


Figure 67. *Section across the Athapuscow Aulacogen showing the stratigraphic relationship of the northern platform and the southern basinal formations of the Great Slave Supergroup exposed in the Hearne Lake map area. After Hoffman et al. (1974).*

Sosan Group

Kluziai Formation

The Kluziai Formation is a sandstone that occurs on the west side of Seton Island where it is conformably overlain by Seton volcanic rocks. The lower part of the formation is not exposed in the map area. The formation consists of a "uniform sequence of cross-bedded, fine grained, even textured, pink to grey sandstones. The beds are lenticular, generally less than 2 feet thick, and are separated by shale partings. The tops of many beds are ripple marked and mud cracked. Petrographically the sandstone is a texturally mature subarkose. Heavy mineral bands and scattered granule-sized quartz grains are typical" (Hoffman, 1968, p. 12). The Kluziai is interpreted as being of fluvial and marginal marine origin (Hoffman, personal communication, 1981).

Seton Formation

The Seton Formation, consisting of volcanics and volcanogenic sediments for the most part, underlies most of Seton Island in the southeast corner of the map area and the smaller islands to the north-northeast. Hoffman (1968, p. 16; personal communication, 1981) reports that: it consists mainly of "volcanogenic rocks with interbedded sandstone, granular hematite and tuffaceous shale. The volcanogenic strata include massive to columnar basalt flows, basalt flow breccias, minor columnar rhyolite flows, and a wide variety of pyroclastic rocks. Ash-fall tuffs and agglomerates are well-bedded rocks containing silt-sized to cobble-sized angular fragments in a matrix of very fine grained ash. Re-worked tuffs are common and contain calcite-cemented, well-sorted and moderately well-rounded volcanic fragments". The volcanics occur as a series of cycles consisting ideally of coarsening upwards sedimentary and volcaniclastic rocks capped by basic subaerial flows (Hoffman, 1973a). The petrography and chemistry of the formation are described by Olade and Morton (1972). The formation represents a complex volcanic centre contemporaneous with the regionally much more extensive Kluziai and Akaitcho River formations of the Sosan Group with which it is interbedded. It is conformably overlain by the Gibraltar Formation siltstones, although this relationship is not exposed in the area.

A Rb-Sr isochron from the Seton Formation suggests an age of 1832 ± 10 Ma ($\lambda^{87}Rb = 1.42 \times 10^{-11}\ a^{-1}$, $Sr_i = 0.7018 \pm 0.0005$, Baadsgaard et al., 1973) although this is somewhat younger than the ^{207}Pb-^{206}Pb zircon age of 1865 Ma for the Compton Intrusive Suite (Bowring and Van Schmus, 1982) which is geologically younger than the Seton.

Kahochella Group

Two formations of the Kahochella Group, the McLeod Bay and Charlton Bay formations, are present only on the south shore of the long island northwest of Seton Island.

McLeod Bay Formation

The McLeod Bay is a red shale with closely spaced calcareous concretions generally less than 10 cm in size that occur in concentrations parallel to bedding. Hoffman (1968) suggests the shales were deposited in a littoral or neritic environment, on the basis of sedimentary structures preserved within the concretions, contemporaneous lag conglomerates of the concretions within the shale and the rare occurrence (outside the map area) of gypsum casts.

Charlton Bay Formation

The Charlton Bay Formation is a dark green shale that conformably overlies the McLeod shales and is overlain by the red marlstones of the Douglas Peninsula Formation of the Pethei Group. The formation consists of "very finely laminated dark green shale or argillite with large widely spaced oblate spheroidal brown weathering calcareous concretions as much as 50 cm across" (Hoffman, 1968, p. 20). The unit contains thin fine grained turbidites and so probably represents a deeper water environment than the underlying McLeod Bay shales.

Pethei Group

The various formations of the Pethei Group are areally the most extensive of the Great Slave Supergroup within the map area. The facies change from shallow water platformal carbonates to deeper water basinal deposits within this group occurs on Blanchet Island.

Douglas Peninsula Formation

The Douglas Peninsula Formation is a thin marlstone that occurs on the south side of the long island northeast of Seton Island. It is the basal unit of the Pethei Group and within the map area is conformably overlain by the basinal McLean marlstones. Elsewhere, it is overlain by the Taltheilei Formation platform carbonates. The formation consists of lithologically homogeneous thin-bedded, fissile red marlstone. The marlstone is made up of "ragged lenticles of pink to light brown argillaceous limestone 1 to 10 cm long and 5 cm thick embedded in dark red brown mudstone" (Hoffman, 1968, p. 21). Elsewhere, crystal casts after gypsum occur. The origin of the formation is uncertain but it changes little in character from where it is overlain by platformal carbonates to where it is overlain by basinal sediments.

Basinal Pethei Group

McLean Formation

The McLean Formation limestones and mudstones are the lowermost unit of the basinal Pethei and conformably overlie the basal Douglas Peninsula marlstones. Within the map area it is conformably overlain in most cases by the basin slope Pekanatui Point Formation or the basinal Blanchet Formation, or the platformal Wildbread Formation on the long island northwest of Seton Island. The lower McLean consists of "medium to thin-bedded grey limestone with closely spaced, one centimeter thick, ragged and crenulated, discontinuous laminations of red brown mudstone... The upper part of the formation consists of brown to green mudstone and flat bottomed, convex upward nodules of argillaceous limestone 3-5 cm across" (Hoffman, 1968, p. 32). Hoffman (1968, 1974) considered the formation to be mainly a basin-floor facies.

Blanchet Formation

The Blanchet Formation greywackes and mudstones conformably overlie the McLean mudstones and limestones and are overlain by the Pekanatui Point limestones. The formation consists mainly of thick-bedded, generally rather massive looking, graded greywacke turbidites separated by thin argillites or interlaminated argillite and limestone. The turbidites were derived from the southwest (Hoffman, 1969). Interbedded with the greywackes in the lower part of the formation are argillite units, up to 25 m thick, containing nodules of argillaceous limestone similar lithologically to the underlying McLean. Similar argillite units, only with thin even-bedded limestone more like the overlying Pekanatui Point Formation, occur in the upper half. The occurrence of presumably deep water turbidites supports the conclusion that this formation and those similar units above and below represent deep water basinal sedimentation.

Pekanatui Point Formation

The Pekanatui Point Formation limestones conformably overlie the basinal Blanchet greywackes and mudstones or the McLean limestones and mudstones and are sharply overlain by the Stark Formation of the Christie Bay Group. The formation consists of thin, even-bedded, light grey, aphanitic limestone separated by siliceous seams or argillite laminations. Convolute bedding and breccia beds with blocks up to 5 m in size are common. In addition to the limestone, units of dark grey argillite to argillaceous dolomite with nodules of limestone and greywacke turbidites are present. Hoffman (1974) has interpreted this unit as representing a basin slope facies.

Platformal Pethei Group

Taltheilei Formation

The Taltheilei Formation is the lowermost unit of the Pethei carbonate platform sequence. It conformably overlies the Douglas Peninsula marlstone and is overlain by the Utsingi carbonates, but laterally passes into the Utsingi Formation towards the basin margin. This transition can be seen at the northeast end of Blanchet Island. The formation consists of light brown weathering thin- to thick-bedded fine- to medium-grained dolomite and minor limestone. Stromatolites in both layers and mounds are abundant and have been discussed by Hoffman (1968, 1974). Excellent exposures of these forms can be seen at the northeast end of Blanchet Island. The formation is of shallow water marine origin and is analogous to more recent carbonate platform reefs (Hoffman, 1968).

Utsingi Formation

The Utsingi Formation lies conformably between the Taltheilei and overlying Wildbread Formation, and where the Taltheilei is not present thickens and takes its position. It consists of "a monotonous succession of thick bedded, blue grey weathering, crystalline limestone with distinctive discontinuous laminations and mottling of brown weathering dolomite" (Hoffman, 1968, p. 27). In the lower part, dolomite laminations occur as concave-up disc-like structures that are vertically superimposed to form vertical to steeply inclined structures a metre or so in length. In the upper part, the dolomite is less prominent and the vertical structure is due to columns of finely laminated limestone. Bedding surfaces form domes elongated parallel to the current and normal to the trend of the unit. Hoffman (1974) favoured a sublittoral origin for this formation on the basis of its relationship to other facies as it grades without a break into basin floor facies.

Wildbread Formation

The Wildbread Formation, a stromatolitic carbonate, conformably overlies the Utsingi dolomites and limestone and is overlain by the Hearne Formation carbonates. The Wildbread is laterally transitional into the less extensive Hearne, particularly to the south towards the edge of the platform. In these areas the Wildbread becomes considerably thicker. The formation consists of thick stromatolitic carbonates; white, recrystallized, rippled, thick-bedded oolitic limestone; stromatolitic laminated, grey limestone and brown dolomite, and thick-bedded, white to grey with a mottled crystalline texture, crenulated, siliceous or dolomitic laminated limestone. Along the shelf margin are stromatolitic mounds or bioherms. Hoffman (1968) suggested the formation represents a back reef lagoonal succession with environmental conditions ranging from shallow marine subtidal to supratidal.

Hearne Formation

The Hearne Formation limestones conformably overlie the Wildbread and are laterally transitional into it along the platform margin. It is overlain by the Stark breccias of the Christie Bay Group. The formation is quite similar to the Utsingi Formation in that it consists dominantly of thick-bedded, white to light grey limestone with laminations and mottling of reddish limestone or dolomite. Bedding surfaces also form ellipsoidal domes oriented parallel to current direction. Well laminated stromatolites are present at the top of the formation. Hoffman (1968) suggested the formation represents a back reef lagoonal environment that was probably permanently submerged.

Christie Bay Group

Stark Formation

The Stark Formation mudstone and carbonate breccias sharply overlie both the platformal and basinal formations of the Pethei Group. The Stark "consists of chaotically dispersed blocks of stromatolitic limestone and dolomite containing abundant shallow water sedimentary structures, in a brecciated matrix of red mudstone" (Hoffman et al., 1977, p. 124). The carbonate blocks are not derived from the underlying Pethei but are indigenous to the Stark Formation. Salt casts are common immediately above and below the formation. Hoffman (1968) and Hoffman et al. (1977) concluded that the formation formed due to solution collapse of a major salt unit overlying the Pethei that was in turn overlain by mudstones and carbonates of the Stark. Solution and collapse took place prior to, and during, the deposition of the lower part of the Tochatwi Formation sandstones (not exposed in the area).

Other early Proterozoic sediments

In addition to the major accumulation of early Proterozoic sediments and volcanics in the East Arm of Great Slave Lake, there are a few other localities where very small exposures of sediments of presumed early Proterozoic age occur.

Eight kilometres south of Payne Lake on a small east-northeast-trending lake is a 1500 by 100 m outlier of conglomeratic sandstone, originally reported by J.F. Henderson (1939). The sediments are quartzites to feldspathic sandstones, with scattered well-rounded cobbles of quartz, quartzite and arkose up to 15 cm in size, although most are 2 to 4 cm. No igneous or volcanic cobbles are present. The sandstone is massive with no bedding or other sedimentary structures preserved. It is extensively fractured, but not sheared. Contact relations with the surrounding Archean granitic rocks were not observed, but the sediments occur in an extensive fault zone and are presumed to have been preserved in a down dropped block in that structure. The relationship of these sediments to the Great Slave Supergroup is not known.

In the Yellowknife area underlying the Paleozoic rocks, a small outcrop of quartzite occurs in the bay west of Spruce Point (R.J.W. Douglas, personal communication, 1976). This quartzite may be related to the early Proterozoic Snare Group that occurs to the north-northwest in the Bear Province.

STRUCTURAL GEOLOGY

Structural events have played an important part in the evolution of the geology of this part of the Slave Province. These events include the original controls on the extrusion and deposition of the Yellowknife supracrustal rocks, their deformation and metamorphism, and finally faulting events, long after the stabilization of the Slave craton as a separate identity after the close of the Archean.

Early regional studies were made by Henderson (1943) and Fortier (1946). Later more detailed studies were concerned mainly with the structure of the supracrustal rocks and, on the whole, were concentrated in two relatively small areas. In the Yellowknife area Campbell (1947) presented the first structural synthesis of the area after the original mapping of the region by Jolliffe (1942, 1946). Henderson and Brown (1952b, 1966) produced a detailed study on the structure of the mafic volcanics at Yellowknife, and Brown (1955) described and interpreted the late faults in that region. Helmstaedt et al., (1981) have discussed aspects of the structure of the southern part of the volcanic belt. The complex structure of the sediments has been discussed and interpreted by Ramsay (1973c), Drury (1977) and Fyson (1978, 1982). The second area of concentration is the Gordon Lake – Ross Lake area that was first described and interpreted by Henderson (1941 a, b). The interpretations made in this area were extrapolated throughout the basin by him (Henderson, 1943). Fortier (1947) produced a 1inch=2mile scale map of the Ross Lake area with abundant structural data. More recently Fyson (1975, 1980), working mainly in the Ross Lake area and Cleft Lake area, 30 km to the southeast, has interpreted the structural data and done a detailed petrofabric study. Thompson (1978), in considering the metamorphism of the province as a whole, developed a structural synthesis incorporating the observations of many workers throughout the province. Fyson (1981) has discussed the tectonics of the southern Slave Province.

Deep crustal structure

The work of Barr (1971) and Clee et al. (1974) has given some insight into the structure of the crust within part of the map area, on the basis of two seismic reflection-refraction experiments along the north shore of Great Slave Lake. Clee et al. (1974), in the more detailed experiment near Yellowknife, found that there is a near surface low velocity layer with a high velocity gradient about 1 km thick across the entire area of the experiment. They suggested this might correspond to a surface zone of fractures and joints that presumably become tighter and fewer in number with depth. East of Yellowknife and south of the mainland along the track of the experiment there is a seismic unit that they interpreted as greenstones (presumably metamorphosed mafic volcanic rocks). They proposed that south of Yellowknife the Kam mafic volcanics extend a considerable distance to the southeast under Great Slave Lake south of the main eastern complex of the Defeat Plutonic Suite. On the basis of the seismic data they suggested the volcanics extend to a depth of about 10 km where their contact with the underlying crustal rocks appears to be very irregular or involves rather heterogenous rocks. The southeastern extension of the volcanic belt is also indicated by the gravity data of the region (Gibb and Thomas, 1980) although the depth of the volcanics on the basis of the seismic model (up to 13 km) is significantly larger than the maximum of 7 km suggested by the gravity model of the belt. Gibb and Thomas (1980) suggested the discrepancy may reflect differing assumptions as to the composition and density of the volcanic and granitoid material in the root zone of the preserved volcanics, and the depth estimates should perhaps be regarded as possible maximum and minimum limits. West of Yellowknife under the granitic terrane is a low velocity layer about 2 km thick at a depth of about 10 km. Clee et al. (1974) suggested this may be due to increased pressure of fluids in pores due to increased pressure and temperature and the sealing of microfractures with depth. The base of the crust which has a rather complex detailed structure occurs at a depth of about 31 km. The earlier, less detailed but more extensive seismic experiment of Barr (1971), indicated that there is an abrupt 4 km difference in the thickness between the Slave crust and the thinner crust under the adjacent Athapuscow Aulacogen of the East Arm of Great Slave Lake.

Regional trends

The dominant structural feature is the northeasterly trend of many of the major rock units. In the Russell Lake area the contact between the Yellowknife supracrustal rocks and the granitic terrane to the southeast has a northeasterly trend that corresponds to a major aeromagnetic break that extends into the Wecho River area (Geological Survey of Canada, 1969c) to the north. The major granitic units within the Yellowknife area also have a northeasterly trend, a trend that is also apparent in potassium anomaly patterns on the radiometric maps of the area (Grasty and Richardson, 1972). The older granitic units on the eastern side of this granitic terrane are weakly foliated (Defeat Plutonic Suite) to strongly foliated (Anton Complex). In both cases the trend of the foliation is northeasterly. In the east the block of pre-Yellowknife basement rocks, the Sleepy Dragon Complex, also has a northeasterly trend, although the structure within it is complex.

The major mafic volcanic sequence at Yellowknife has a similar trend. The apparent northerly trending aspect of the belt as a whole is due to left lateral, north-northwesterly trending faults. The flows, however, have a definite northeasterly trend. On the east side of the supracrustal basin the major volcanic belt along the Cameron River has a northeasterly trend, as does the volcanic belt through Tumpline and Turnback lakes to the volcanic complex at Sunset Lake. The major departures from this trend in the volcanic rocks are the southeasterly trend of the volcanic rocks south of Payne Lake and Victory Lake and the similar trend of the southeast extension of the Kam volcanics under Great Slave Lake (Gibb and Thomas, 1980).

Structure of the granitic units

The granitic units throughout the area are for the most part rather structureless. Even the Amacher Granite, which is thought to be similar in age to the supracrustal rocks and is certainly premetamorphic, is a massive unit. The only exceptions are the previously mentioned weak foliation in the Defeat Plutonic Suite west of Yellowknife and the gneissic units west of the Yellowknife River (Anton Complex) and east of the Cameron River (Sleepy Dragon Complex). Locally in parts of the Defeat Plutonic Suite southeast of Yellowknife there is a foliation at and parallel to plutonic margins. Rocks in the inclusion-rich zones between massive plutonic lobes of the Defeat are typically strongly aligned and foliated. This is also the case with inclusion-rich zones in the Meander Lake Plutonic Suite.

Much of the structure in the Anton and Sleepy Dragon metamorphic complexes may have been acquired prior to the deposition of the Yellowknife rocks. There is a strong structural and metamorphic contrast between these complexes and the adjacent Yellowknife rocks (i.e. east of Yellowknife River and west of Cameron River). The structure within these units has not yet been studied in detail and is poorly understood. In the Sleepy Dragon Complex there are large scale folds within the gneissic rocks (Davidson, 1972) that may represent pre-Yellowknife structures that have been modified by later events. Northwest of Turnback Lake, the post-Yellowknife structures, in which both tectonic slices or inclusions of supracrustal rocks as well as later intrusions are deformed, also dominate the older Sleepy Dragon granitoid rocks (Davidson, 1972).

Structure of the supracrustal rocks

Deformational events in the area are most clearly expressed in the Yellowknife supracrustal rocks. The thick units of mafic and felsic volcanics have behaved in a relatively competent manner compared to the interbedded greywackes and mudstones in which the fold patterns are highly complex (see map, in pocket). What follows is a general description of the supracrustal rocks. Somewhat more detailed descriptions of the structure of specific areas, particularly in the case of the various volcanic belts, are presented in many cases in the sections dealing with the description of the particular units in question.

Volcanic rocks

The major volcanic belts occur either in steeply dipping, homoclinal successions or more complex, but generally steeply dipping, synclinal – anticlinal sequences. In general the homoclinal successions occur where rather thick, dominantly mafic sequences are buttressed on one side against an older basement of granitoid gneisses and plutons and face toward much less competent, typically isoclinally folded basinal metagreywacke – mudstone sediments on the other. The folded metavolcanic structures, on the other hand, are bounded by metasediments or are buttressed on both sides between blocks of basement and/or intrusive granitoid plutons. This pattern of deformation is seen throughout the Slave Province as, for example, at Point Lake where the Point Lake volcanics (Bostock, 1980) south of Point Lake are folded between two basement blocks, while north of the lake the same volcanic belt forms a homoclinal succession with probable basement gneisses on one side and metasediments on the other (Henderson, 1977). Similarly, in the Lac de Gras area in the central Slave Province (Folinsbee, 1949), the northerly trending volcanic belt is in large part a homoclinal succession between granitoid rocks and an extensive terrane of metasedimentary rocks (Dillon-Leitch, 1979; Moore, 1956). In the Indin Lake area in the western part of the province, the metasedimentary-metavolcanic rock units occur as anticlinal structures bounded by metasedimentary rocks (Tremblay, 1954; Tremblay et al., 1953).

Within the map area the Kam mafic volcanic belt at Yellowknife, the dominantly mafic volcanic formations along the Cameron River, the mafic volcanics southwest of Clan Lake and possibly the felsic volcanic rocks east of Russell Lake, occur in homoclinal successions. All are thick sequences bounded by basement blocks or granitoid plutons on one side, face toward metasedimentary rocks on the other, and most are more or less parallel to the previously discussed northeasterly trend seen throughout the region. All are steeply dipping to vertical and locally are overturned towards the sedimentary basin they face. In the case of the Kam Formation, this structural picture should be qualified for that part of the formation in the southernmost part of Yellowknife Bay. Although geological relations are in large part complicated by extensive faulting and are obscured by water, it has been suggested that the Duck Formation mafic volcanic rocks which occur in an anticlinal structure marginal to a major lobe of the Defeat Plutonic Suite on the east side of the bay may be correlative with the Kam Formation on the west. So in this part of the area, between the two Defeat plutonic buttresses, the otherwise homoclinal succession of mafic volcanics may be folded across the major synclinal structure under Yellowknife Bay (Henderson, 1970).

The major thick volcanic sequences in the northeast of the area, on the other hand, do not occur in homoclinal successions but are typically complexly folded. The volcanic complex in the Victory – Tumpline – Turnback lakes area occurs in a series of northeasterly and northwesterly trending structures that are bounded by metasediments. The volcanic structures are both truncated by, and mantle, Prosperous and Defeat granitoid intrusions respectively.

To the northeast, the complexly folded, dominantly mafic, volcanic complex at and southeast of Sunset Lake is confined between the Sleepy Dragon basement complex to the west and the Meander Lake Plutonic Suite to the east and south. The dominant element is a northerly trending synclinal structure through Sunset Lake. The western limb is complicated by the presence of the Amacher Granite in the core of an elongate anticlinal dome. The southeasterly trending limb of the complex consists in large part of steeply dipping, amphibolite grade, volcaniclastic sediments. Although structural data are rather sparse, the anomalous thickening of the unit to about twice its outcrop width, as seen to the north and south, may be explained by the presence of a northerly trending, subhorizontal syncline-anticline pair that transects the southeasterly trending metavolcaniclastic belt at a low angle.

East of Clan Lake is a more or less equidimensional felsic volcanic unit in which primary layering is commonly difficult to see. However, given the highly competent nature of the unit there is a general structural conformity with the northeasterly trending sediments that surround it. From the outline of the unit there appears to be a second direction of folding in a northwesterly direction. This is the same general direction as a prominent regional cleavage in the sediments (discussed later). The volcanics have behaved as a relatively stable buttress about which the less competent sediments and thin-bedded mafic volcanics interbedded with them have been folded in a much tighter style. To the southwest of Clan Lake a similar complex fold pattern can be seen in the sediments, where the outcrop pattern of mafic sills interlayered with the sediments suggests an interference pattern produced by superimposed structural elements.

In some cases some of the Defeat plutons are in part mantled by relatively thin sequences of mainly mafic volcanic rocks. This can be seen at Doubling Lake at the east border of the area and southeast of Yellowknife where the Duck Formation occurs in an anticline between Yellowknife Bay and Preg Lake, parallel to, or striking into, the plutonic lobes. Defeat plutons are also mantled by somewhat thicker and compositionally more diverse volcanics in the vicinity of Tumpline Lake where they are part of the complexly deformed volcanic centre in that region.

One of the best examples of two phases of folding in the volcanics can be seen at Preg Lake, 17 km east of Yellowknife (Fig. 68). There the volcanics that partially mantle a lobe of the intrusive Defeat granodiorite interfinger with the adjacent Burwash greywackes. The main structure is an easterly trending anticline of varied plunge that can be traced from Yellowknife Bay and terminates at the plutonic lobe. In addition there is a series of folds in the volcanics whose trend is normal to the main anticline and parallels the contact of the volcanics with the pluton and the sediments.

Structures within the volcanic units consist mainly of flattening of primary features. Volcaniclastic layers within the volcanic sections have taken up much of the strain and so

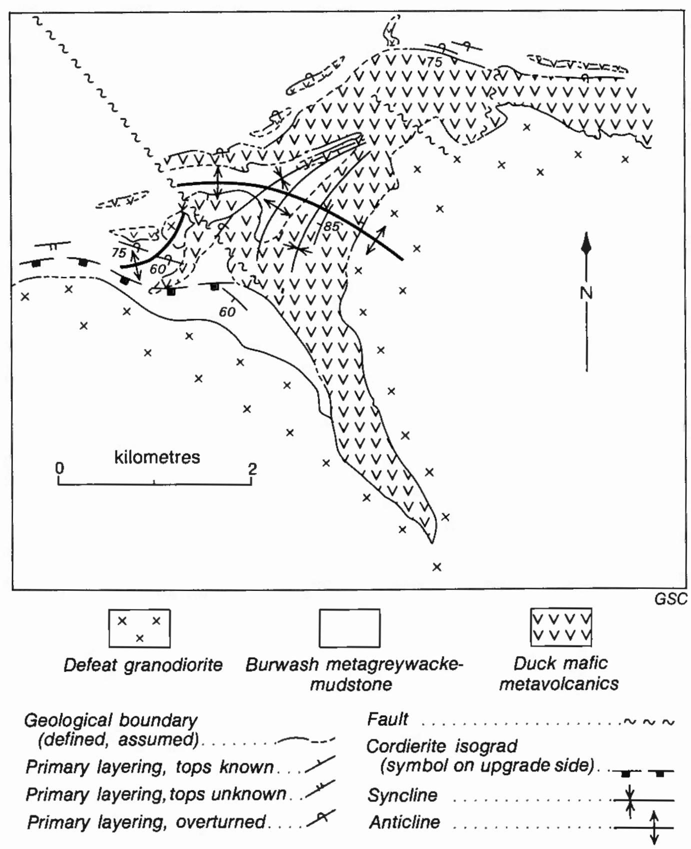


Figure 68. *Superimposed folding in the mafic volcanics at Preg Lake, 17 km east of Yellowknife. An easterly trending anticlinal structure, along which mafic volcanics are discontinuously exposed, can be traced from Yellowknife Bay to this locality. A lobe of the Defeat granodiorite intrudes the core of the structure. A series of cross folds parallel to the intrusive contact, and perhaps related to the expansion of the plutonic lobe, occurs in the volcanics. The interfingering nature of the volcanic and sedimentary rocks is demonstrated in this structure. Modified after Jolliffe (1942).*

tend to be much more deformed than the more competent flows. Even the flows, however, have been deformed to some degree. This is most evident in the pillowed flows. In some cases in horizontal section the pillows appear essentially undistorted, at least to the extent that it is possible to differentiate between weak tectonic distortion and primary deformation that took place as the pillow structures formed. At the other extreme, the pillows are so flattened that the outcrop has a banded aspect and, unless one traces out individual bands, the original pillowed nature is easily missed. In many cases even though pillows can be recognized, their usefulness as top direction indicators is lost, as the originally gently curved margins of the pillows become straightened on the top and bottom surfaces and interdigitated with adjacent pillows on the sides (see Lambert, in press). However, even the apparently most undistorted pillows in horizontal section are considerably elongated in vertical section. The importance of the vertical direction of elongation of the pillows, and other structures as well, may be related to the common proximity of the volcanics to intrusive granitic bodies. The vertical diapiric intrusion of the plutons may be reflected by the vertical extension of elements within the adjacent country rocks. As suggested by Clifford (1972) for similar rocks in the Sioux Lookout area of Ontario, the flattening in a subvertical plane parallel to the contact with the granite may be due to the inflation of the intruding diapir. Such a mechanism might also explain the folding parallel to the contact in the previously mentioned volcanics at Preg Lake.

Sedimentary rocks

The much more complex structure of the metamorphosed greywacke-mudstone turbidites, compared to that of the volcanic sequences, is due mainly to the much less competent nature of the sediments. The greywacke-mudstone sediments are, however, everywhere concordant with adjacent volcanic rocks.

The structural relationships between the sediments and the various granitic units is much more varied. The largest intrusions of the Defeat Plutonic Suite are everywhere concordant with the sediments, although with the smaller plutons there is a greater tendency for discordancy. The Prosperous Granites, on the other hand, tend to be sharply discordant to the sediments. The generally northerly trend of the larger bodies is roughly parallel to the trend of the sediments they have intruded. In detail, however, contacts are commonly discordant at the ends of the intrusions as well as where there is a local variation in the trend of the adjacent sediments. All the smaller intrusions are discordant. Contact relations with the Meander Lake Plutonic Suite on the east side of the basin is more varied. The complex is quite discordant north and east of Francois Lake, but is more concordant to the south. The early Proterozoic Blachford Lake Intrusive Suite is strongly discordant to the trend of the Burwash sediments.

The consistent northeasterly trend seen in some of the granitoid and volcanic units is also evident in the metasediments at Russell Lake. Similar but higher grade metasediments at Stagg Lake, however, are at a rather low angle to the northeasterly trending Stagg Plutonic Suite. This dominant northeasterly trend, on the other hand, is not apparent at all in the Burwash Formation metasediments in the supracrustal basin east of Yellowknife where structural trends are highly varied. In the northern part of the area, the bedding trends form a broad, open, slightly distorted 'N' shape (see bedding trend lines on map, in pocket). From south of Quyta Lake to the north end of Duncan Lake there is a northeasterly trend that swings around to the southeast between Prestige Lake and the south end of Gordon Lake. From there it swings again to the northeast parallel to Cameron River. The southern part of the basin is made up of a broad, rather distorted, 'M' shaped structure. Between Hearne and Campbell lakes, the bedding trend is consistently northerly to about Ross Lake, where it swings around to the southeast through Victory Lake and Consolation Lake to the central part of Cleft Lake, where it once again swings north to the latitude of Tumpline Lake. From there it swings

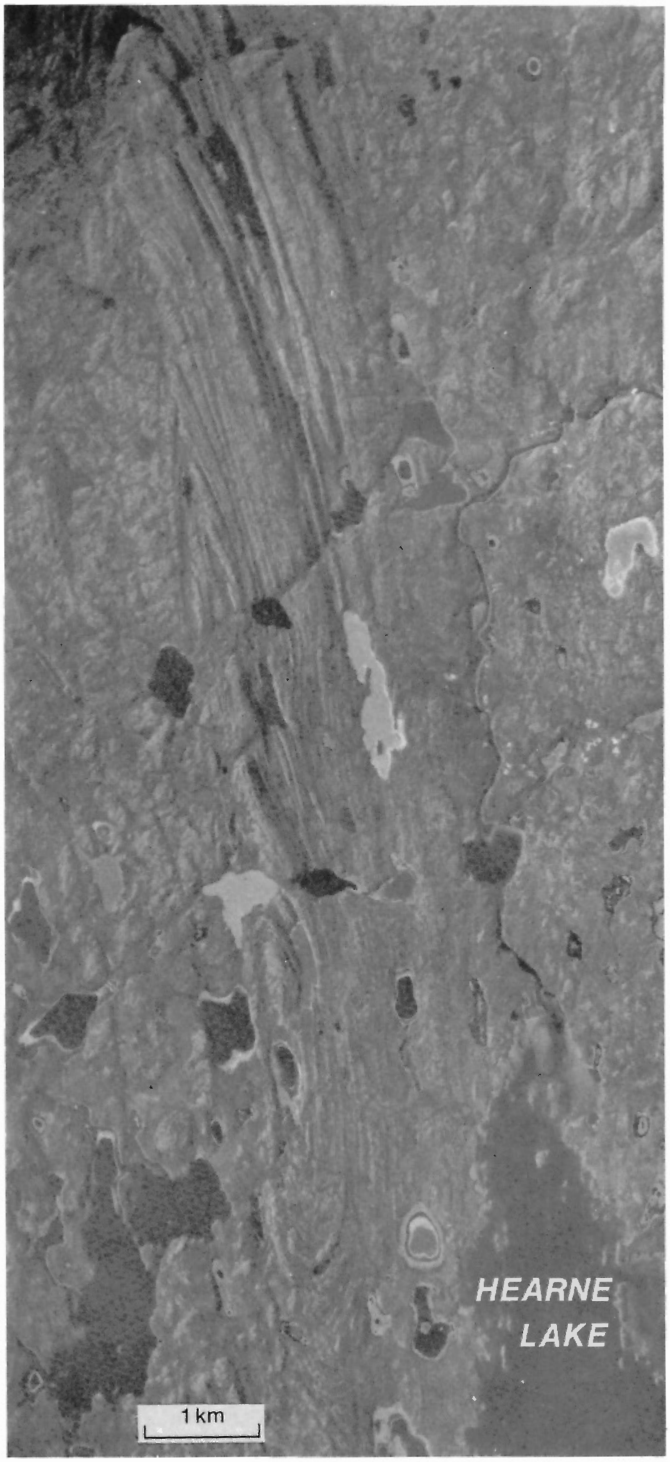


Figure 69. *Part of a large, tight, doubly plunging syncline northwest of Hearne Lake. The layering on the east side of the structure is accentuated by the more or less parallel regional north-northwesterly trending cleavage. NAPL A13747-70*

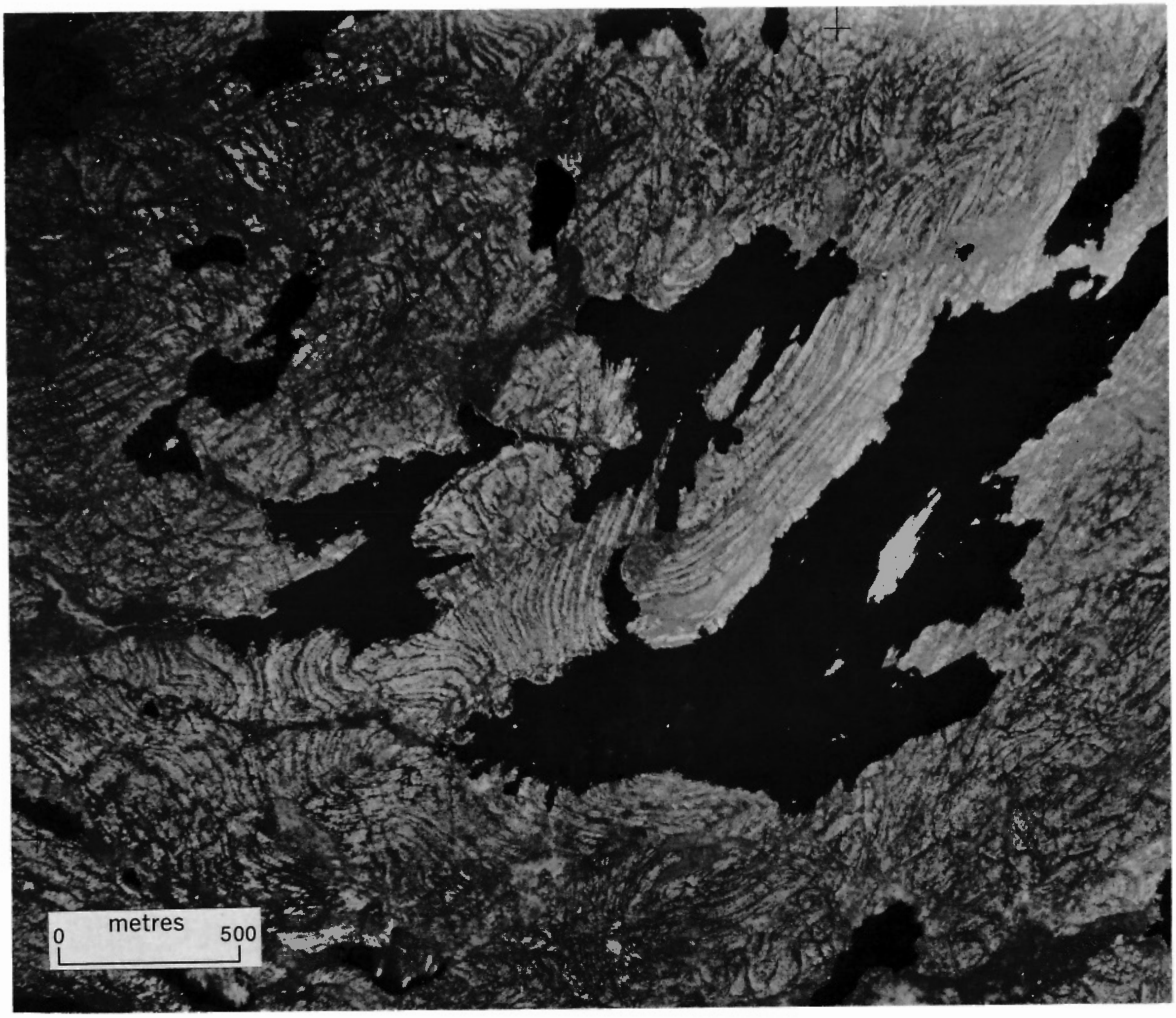


Figure 70. Complex interference fold patterns in the Burwash Formation 10 km west of Francois Lake. NAPL A8566-67

Figure 71. Steeply plunging asymmetric fold in Burwash Formation greywacke-mudstone. GSC 177861

easterly to Desperation Lake. To the southeast in the vicinity of Buckham Lake, the regional bedding trends have essentially the same 'M' shaped map pattern but with a generally decreasing amplitude towards the southeast. In the triangular area between the two plutonic complexes of Defeat granodiorite east and west of Yellowknife Bay and the Prosperous Granite, the bedding trends form a complex tight fold pattern on a smaller scale than the broad-scale patterns elsewhere in the basin.

The local fold patterns within these generalized trends, as outlined on the map, are highly varied. At their simplest they consist of generally upright isoclinal folds with subhorizontal axes that locally plunge gently at angles less than 25°. The scale of these folds is varied, with wavelengths that range from a few tens of metres to over 10 km. These isoclines are the dominant fold style where the bedding trend is consistent over long distances. They are perhaps best documented in the Gordon Lake South area (Henderson, 1941b, 1943) where individual isoclines have been mapped. Another area where these relatively simple isoclines are well developed is between Campbell, Hearne and Ross lakes. One of the largest folds occurs just northwest of Hearne Lake (Fig. 69) where a doubly-plunging synclinal structure can be traced over 20 km along strike. The axial surface traces of adjacent anticlines are 4 and 5 km to the east and west. This particular structure is unusual in that it is one of the few places where the hinge zone of the

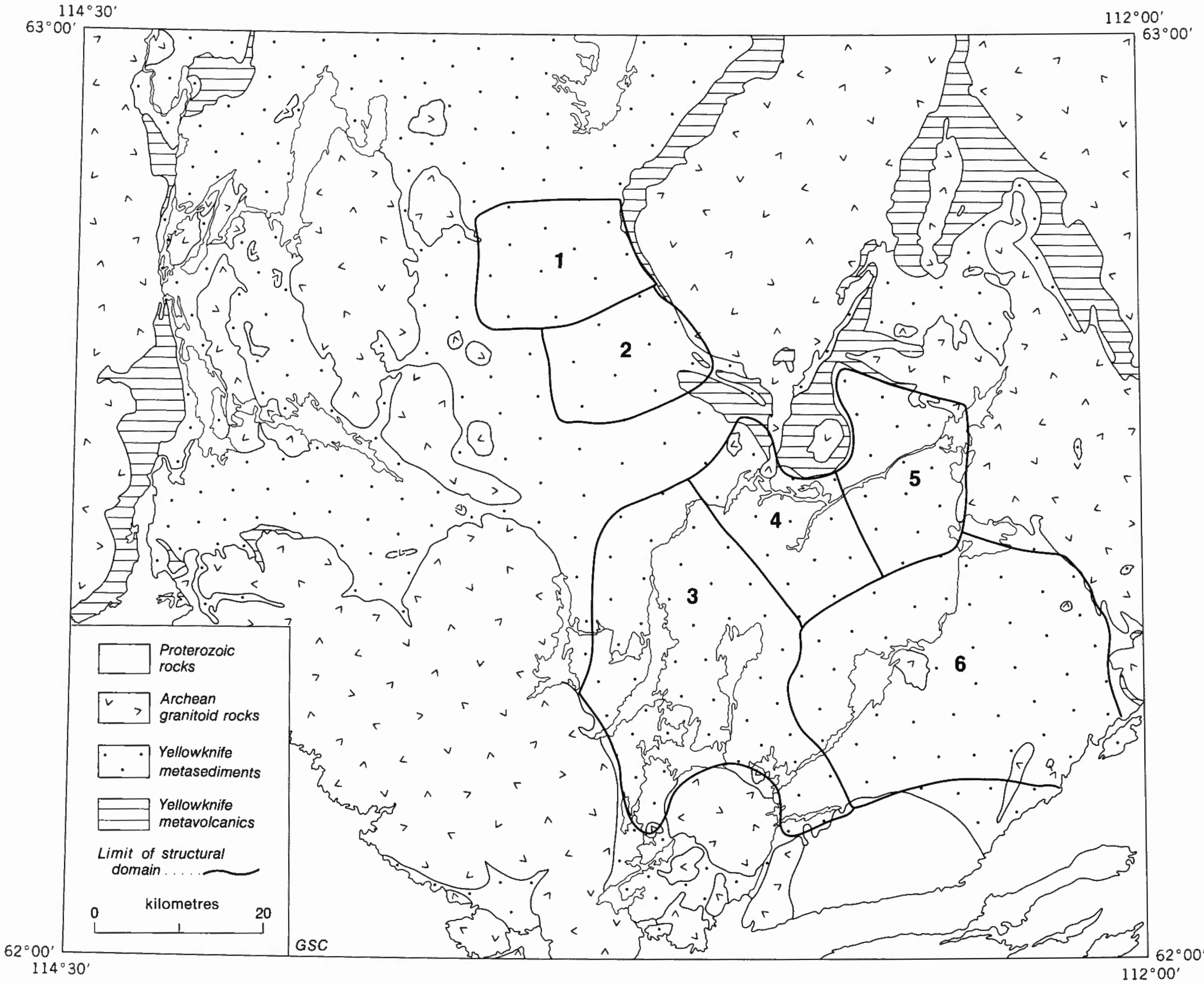


__Figure 72.__ Structural domains based on bedding trends in the Burwash Formation. Structural data for these domains are displayed in Figures 73 and 74.

structure is locally preserved. In most cases, in such structures, the hinge zone is broken and continous layers around the culmination of the fold are not commonly seen. Where the hinge zone is preserved it is evident that there has been great thickening of the beds in the axial zone. Another major synclinal structure occurs at the west margin of the basin parallel to Yellowknife Bay. In this structure the west limb includes the Kam mafic volcanic sequence while the east limb consists mainly of Burwash basinal sediments. A felsic volcanic unit high in the section (the Banting Formation) is considered to occur on both limbs, although Helmstaedt et al. (1980) have suggested the two volcanic units are separate members in a homoclinal succession. Between the axial surface traces of the syncline and the adjacent anticline to the southeast is about 5 km of Burwash turbidites.

Individual isoclines cannot be traced long distances along strike. The longest axial surface trace recognized is about 25 km although most are much less. The amplitude of the folds is difficult to determine, given the apparent subhorizontal orientation of their axes and the lack of topographic relief in the country. A cleavage in the sediments that is thought to be related to the formation of the isoclinal structures, occurs in some outcrops as a planar fabric subparallel to bedding, and is recognized by the flattening of primary structures, in particular mudstone clasts, in the coarser grained parts of beds (Fig. 36). It is not always strongly developed. Folds commonly end by decreasing in apparent amplitude and wave length, and also with the development of an en echelon series of smaller scale folds superimposed on the end of the main fold (Fig. 61, in pocket). Another fold commonly starts nearby, translated laterally from the extended trace of the fold that ended in a manner similar to the models illustrated by Dubey and Cobbold (1977).

Detailed mapping by Fortier (1947) in the vicinity of Ross Lake shows the form and complexity of the structure in the Burwash sediments. Within this area the regional trend north from Hearne Lake swings around at Ross and Pensive lakes to a southeasterly trend through Victory Lake. Figure 61 (in pocket) is taken from this area and has at its approximate corners Redout Lake, Sproule Lake, Reid Lake and Consolation Lake (NTS 85 I/11). Figure 61 is a form surface map of the area, based on the interpretation of Fortier's (1947) structural data and air photos. Form lines indicating bedding have been drawn, and axial plane traces of various folds are interpreted. Although no marker horizons are recognized within the Burwash, four "artificial" marker

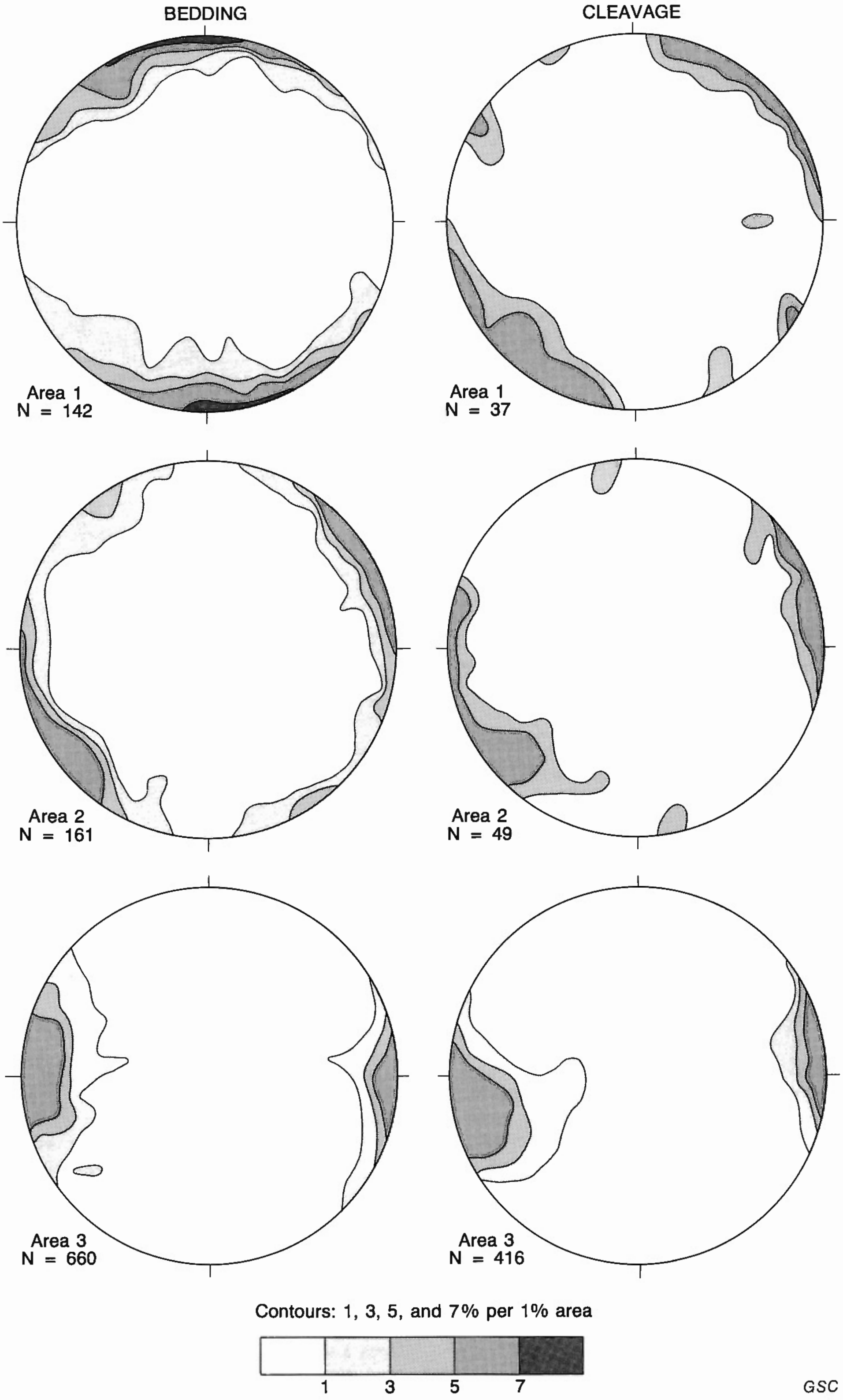


Figure 73. *Stereograms of bedding and cleavage data for six domains defined in Figure 72. Although the bedding trends among the six domains are varied, the cleavage trends in each is consistent among the various domains. This implies the cleavage is not related to the fold styles within the domains. The stereograms are equal area, lower hemisphere plots that are contoured at 1, 3 and 5% per 1% area.*

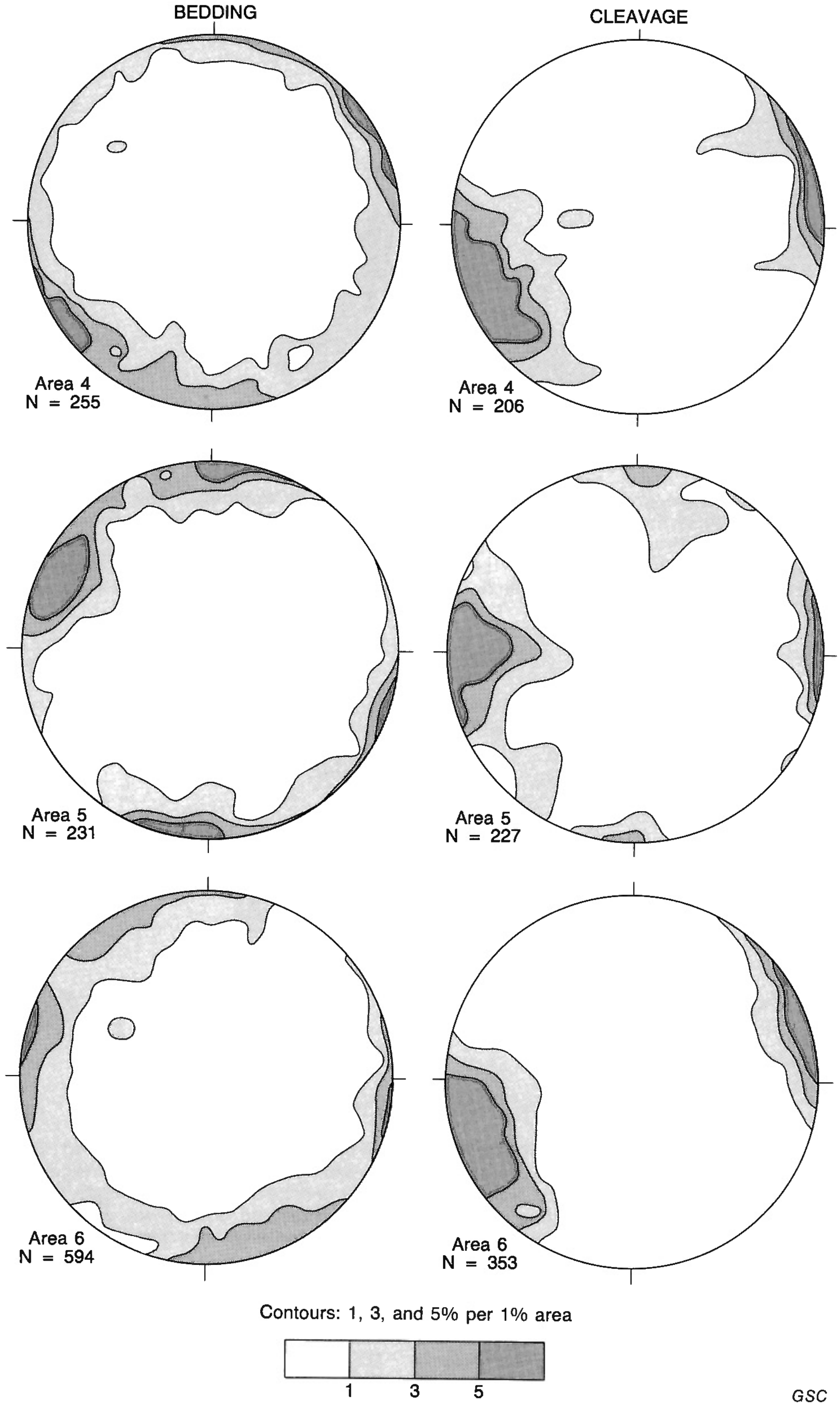


Figure 73 *(cont.)*

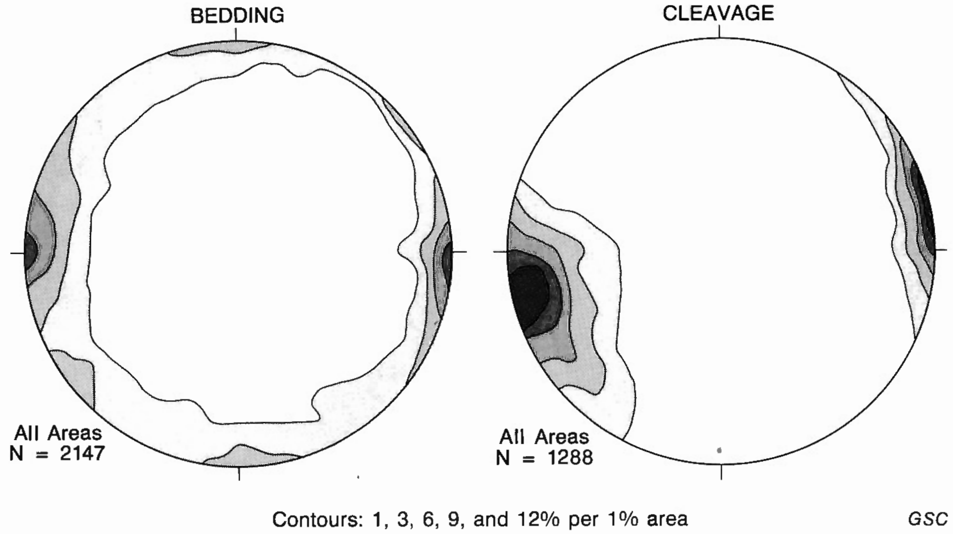


Figure 74. *Stereograms of structural data from all six domains. The bedding planes are generally steeply dipping and have a moderately preferred northerly trend although trends in any direction are well represented. The cleavage, on the other hand, has a consistently north-northwesterly trend and dips steeply to the east-northeast. The stereograms are equal area, lower hemisphere plots that are contoured at 1, 3, 6, 9 and 12% per 1% area.*

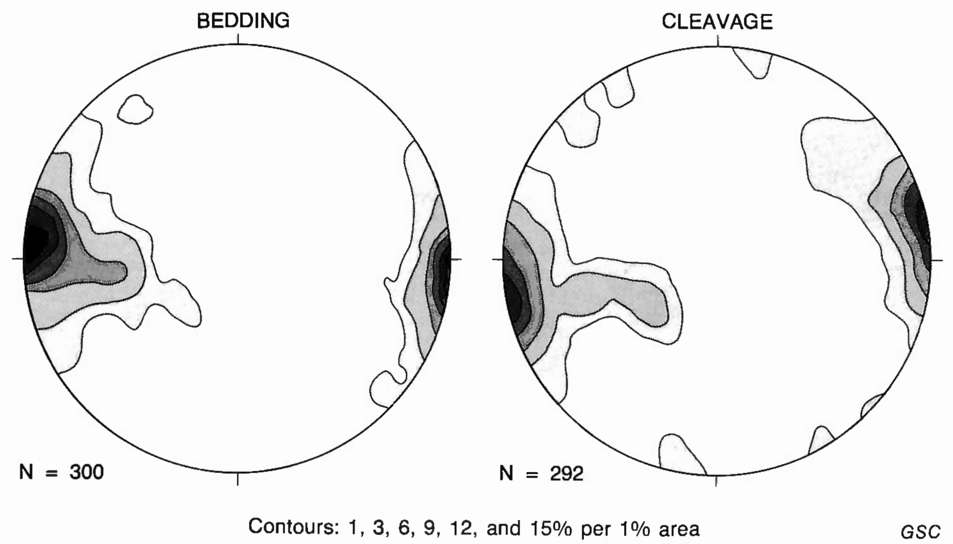


Figure 75. *Stereograms of structural data from the Benjamin Lake area 200 km east-northeast of Yellowknife. As in the Hearne Lake area, the cleavage has a preferred north-northwesterly trend that appears unrelated to the northerly to north-northeasterly trend of bedding in that area. Structural data from Heywood and Davidson (1969). The stereograms are equal area, lower hemisphere plots that are contoured at 1, 3, 6, 9, 12 and 15% per 1% area.*

beds are arbitrarily defined in Figure 62 (in pocket) and traced through the area following the form surface pattern, in order to allow better visualization of the form of the folds. One outcome of this exercise is that it allows an estimate to be made of the stratigraphic position of the mafic sills at Tibbit Lake. On this basis it would appear that they are relatively low in the Burwash Formation section, possibly at the approximate stratigraphic level of the Tumpline Lake Basalt and Sharrie Rhyolite as exposed south of Victory Lake. In Figure 63 (in pocket) a series of hypothetical sections have been drawn, one across the area at a high angle to the trend, a second at least in part parallel to the structure in the southern part of the area considered, and a third along the trough of the major synformal stucture.

Where earlier structures are at a high angle to later directions of folding, spectacular interference patterns result (Fig. 70). For example, east of Yellowknife near Yellowknife Bay the east limb of the major synclinal structure has been refolded into a set of steeply plunging chevron folds (Jolliffe, 1942, 1946). Farther east, near Bighill Lake, a basin and dome pattern has been produced that has some of the lowest dips anywhere in the area (<10°). In the vicinity of Ross Lake where the early isoclines change direction, interference patterns including steeply plunging structures and dome and basin structures (Fig. 62) are developed. Other well developed interference structures occur southwest of Consolation and Cleft lakes where the trend of the isoclines are normal to a later direction of folding. In this regard there appears to be a rough correlation between the occurence of gold and/or scheelite bearing quartz veins and the area where the interference patterns are best developed.

A satisfactory unravelling of deformational phases or events within the sedimentary basin has not proved to be possible at the present scale of mapping. As can be seen in Figure 61 a complex scenario is evident, and may well include local complex deformational styles imposed while the sediments were still unconsolidated, and indeed may have been still accumulating. Subsequent deformation may have been strongly influenced by local constraints such as the presence or emplacement of plutonic bodies or movement of basement blocks, and may well have overlapped in time and space. This complexity becomes even more evident in areas where an attempt has been made to understand the deformational history. For example, east of Yellowknife, Fyson (1982) has suggested a sequence of 8 structural events expressed as a series of folds and cleavages of varied orientation that are thought to overlap in time.

Most of those concerned with the structure in the area have separated a later more regional deformational event from earlier events. For example Henderson (1943) suggested the trends of the earlier isoclines were reoriented into their present broad open form during a later deformation that was expressed locally as steeply plunging folds (Fig. 71). Fortier (1946, 1947) noted that the axial traces of superimposed flexures and cross folds on the earlier structures have a consistent northwesterly trend. He also mapped a regional northwesterly trending cleavage in the Ross Lake area that is parallel to the trend of the superimposed folds. Fyson (1975, 1980) has done petrofabric studies on the effects of this later deformation on earlier structures.

The effect of this late deformation is truly regional. The structures formed by it differ from the earlier isoclines by their consistent orientation throughout the area and beyond. Its most prominent manifestation is the development of a strong cleavage that occurs almost everywhere throughout the basin. Although there are local variations, the cleavage has a consistent north to north-northwesterly trend and dips very steeply to the northeast (Fig. 72-75). Where the cleavage is at a moderate angle to bedding it is commonly refracted as it passes through the graded beds (Fig. 36). It is much more strongly developed than cleavages related to earlier folds. This regional cleavage is locally crenulated (Fyson, 1975; Kamineni et al., 1979).

Although quite consistent in its orientation throughout the area, the development strength of the late cleavage is variable. In places it is quite obscure. This is commonly the case where bedding and cleavage trends are parallel. On the other hand there are zones where the cleavage is particularly strong. For example, north of Consolation Lake and east of Tumpline Lake there are zones a few hundred metres wide where the cleavage is exceptionally well developed, to the extent that the bedding is transposed parallel to it. Between Beaulieu River and Cleft Lake several of these narrow, poorly defined zones of intense cleavage are separated by areas of varied but generally more weakly developed cleavage.

As a result of this later deformation, an additional fold pattern is imposed on the earlier structure. Where the original isoclines are parallel or at a low angle to the second direction of folding the effect has been minimal, or at most, as suggested by Fyson (1975) and Thompson (1978), a tightening of the original fold. For example the large syncline northwest of Hearne Lake appears not to have been affected by later folding. There has, however, been a physiographic effect, for when the bedding and cleavage are parallel, in one limb there has been preferential erosion of the weaker layers producing a pronounced ribbed effect, compared to the relatively flat terrane underlain by the other limb, where the late cleavage is at an angle to bedding (Fig. 69).

The variations in plunge of the early structures may be due, or at least exaggerated, by the later deformation. As previously mentioned, the intensity of the deformation as shown by the development of the cleavage can be quite varied over a small area. Variations in intensity of the late deformation along the trend of an earlier fold could result in the bending of the axis of the earlier fold. On the other hand, the variation of plunge of the early folds has also been interpreted by Fyson (1975) as being due to culminations and depressions formed in an earlier folding event prior to the formation of the isoclines.

As reported by most workers in the region, metamorphic porphyroblasts of cordierite and andalusite are commonly oriented parallel to the cleavage or overgrow it (Henderson, 1943; Fortier, 1946, 1947; Kretz, 1968; Kamineni, 1973b; Ramsay, 1973b; Fyson, 1975; Kamineni and Divi, 1976; Ramsay and Kamineni, 1977; Thompson, 1978; Kamineni et al., 1979). Thus this later deformation predated at least the last stages of metamorphism.

Origin

There is a general consensus among those who have worked on the structure of the metasediments that the deformation pattern is closely related to the activity of the granites. For example Henderson (1943), although he did not propose a mechanism for the formation of the isoclinal folds, assumed that their original trends were more or less consistent over the whole area but were later reoriented by variably directed forces due to the intrusion of later granitic bodies. Fyson (1975, 1978, 1980) on the other hand suggested that the isoclines and 'elliptical depressions and culminations' that preceded them were due to the diapiric rise of granitic basement and granodioritic intrusions, together with marginal subsidence with accompanying gravitational sliding from the uplifted areas. He felt that the curvature of the fold trends was at least in part related to rising diapirs but was subsequently modified by a late regional compressive event. Similarly Drury (1977) proposed that the isoclines formed by gravity sliding of the largely unconsolidated Burwash sediments due to the rise of the Defeat granodiorites as diapirs. The form of the isoclines was subsequently modified by periods of inflation and remobilization of the intrusions.

The model relating the formation and form of the major isoclines to the diapiric uprise of basement and granitic intrusion seems to be generally accepted, although perhaps certain constraints can be applied after consideration of other aspects of the geology reported elsewhere in this report. The basin in which the volcanics were extruded and sediments deposited formed due to partial rifting of a sialic basement, resulting in a terrane of down-dropped grabens and uplifted horsts on the scale of tens of kilometres (McGlynn and Henderson, 1970; Lambert, 1977; Henderson, 1981; this report). Thus the sediments were deposited in a basin contained by buttresses of basement. During the filling of the basin vertical movements continued, both along the margin as well as within the basin with the continued rise (and collapse ?) of basement blocks, most of which are not exposed at the present erosion level. This may have resulted in the contemporaneous deformation of the supracrustal rocks even as they accumulated. The earliest intrusions, such as those of the Defeat Plutonic Suite, were emplaced as diapirs and, together with continued basement block movement, resulted in the formation of major isoclines due to the sliding away of the sediments from the rising diapir (Fyson, 1975; Drury, 1977). These folds would be constrained by the basement buttresses at the basin margins and perhaps by others not exposed at the present erosion level. Later intrusions such as the Prosperous Granite may also have had a similar effect, reinforcing, or perhaps interfering with, earlier fold patterns to produce the complex pattern such as that seen at Ross Lake. Other intrusions such as the Redout and Morose granites in the Sleepy Dragon Complex may have caused expansion of the basement block and may be an explanation of the overturned structure in that vicinity. Thus the major curved isoclines may not have formed as a consequence of a single event but evolved over a period of time as a result of a series of events.

The intruding diapir and/or inflating basement block model may account for the structures between the two Defeat plutonic complexes east and west of Yellowknife (Fyson, 1982) or the curvature of the structure southeast of Ross Lake through Cleft Lake to Desperation Lake. However, the origin of the broad arching of the structure northeast through Duncan Lake and then southeast to Gordon Lake is less evident, but may be related to the early expansion of the terrane to the south prior to the intrusion of the Prosperous granites that are discordant to the structure. The relatively low dips in the area northeast of Duncan Lake may be evidence of this early inflation. The origin of the early folds in the southern part of the area between Campbell Lake and Drever Lake to the southeast is not clear, but may also be related to an early inflationary event in the area outlined by the thermal dome.

It is difficult to do more than speculate on the original form of the earliest folds. It seems unlikely that they were long limbed recumbent isoclines or nappes as nowhere where there are sufficient data is there any evidence of an underlying limb (that would presumably be recognized by reversed facing direction relative to the upper limb) exposed in the cores of anticlines (Fig. 61, in pocket). In Figure 63 (in pocket) a hypothetical section parallel to the trough of a major synclinorium, as drawn, shows little relief in the trough of the syncline although at the southwest end of the section the layering swings past the vertical. As the area of interest occurs at the corner of Fortier's (1947) more detailed map and a large Prosperous granite pluton occurs immediately to the southwest, there are insufficient data to determine if this dip reversal represents the nose of an earlier recumbent isocline or, more likely, later interference.

The curved form of the earlier folds has been intensified by the later, more regional, deformational event that imposed the rather consistent north-northeasterly cleavage throughout the area mapped. While the stresses involved in the earlier deformations appear to be related more to local features or events within the area, the direction of stress that resulted in the later deformation was consistent throughout the region. Indeed, what is thought to be the same structure with the same trend also occurs in the Benjamin Lake area (Heywood and Davidson, 1969) some 50 km to the east, (Fig. 75). The consistent attitude and strong development of regional cleavage, with the accompanying production of folds superimposed on earlier structures implies a regional compression in a west-southwesterly – east-northeasterly direction on a regional scale.

Faults

Several fault sets of varied ages occur within the map area.

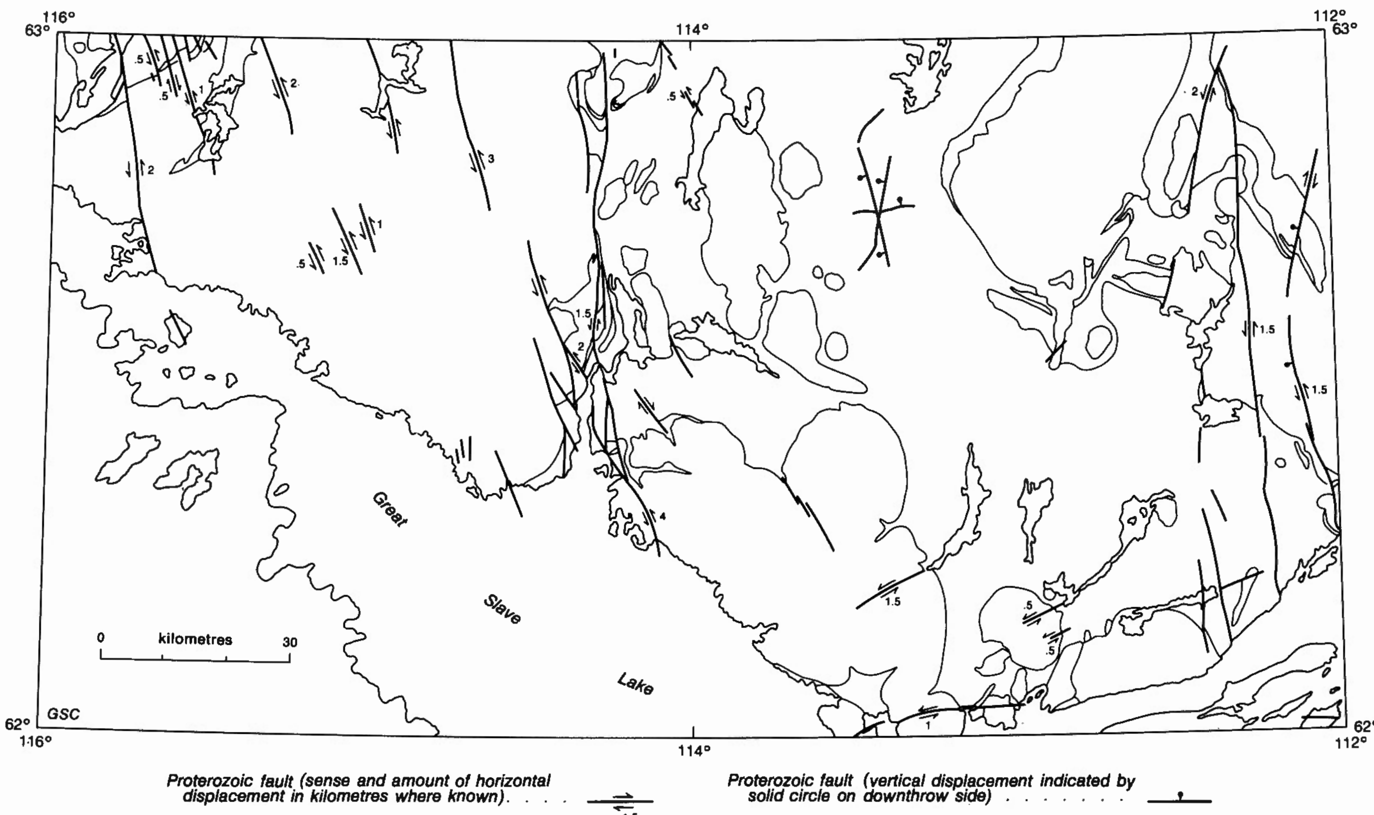


Figure 76. *Most of the prominent post-Archean faults in the area are north-northwesterly striking, left lateral faults with the displacements, where known, indicated in kilometres. A less abundant set has a northeasterly trend and is dominantly right lateral, although several of these faults in the east are left lateral. In the eastern part of the area and in the central part of the basin are faults with a significant vertical component.*

With the exception of the shear zones in the mafic volcanics at Yellowknife (previously described in the section on the Kam Formation) which are believed to be a gravity slide phenomenon contemporaneous with volcanism, faults of Archean age are not easily documented. The supracrustal basin is thought to be a graben-like structure, so presumably there were basin margin faults that, however, have not as yet been identified. Several possible splays from the cryptic basin boundary fault occur in the vicinity of Webb and Patterson lakes in the Sleepy Dragon Complex and have been discussed earlier. Some of the movement on the unconformity between the relatively minimally deformed volcanics and the highly deformed gneissic basement rocks east of the Cameron River may be Archean. Post-Archean faulting is quite common throughout the area. Several fault sets are present but it is commonly difficult to determine absolute or even relative ages of movement on these structures.

The most prominent of the late faults trend northwesterly between 330 and 020° with an average trend of about 345°. In general the trend chages from north-northwesterly in the west to northerly in the east. These are mainly left lateral transcurrent faults with displacements of up to 4.9 km although most range between 0.5 and 3 km (Fig. 76). Faults with displacements of only a few kilometres occur throughout the map area, although they are apparently more abundant in the granitic terrane west of Yellowknife. Across the 200 km width of the area there is an apparent aggregate displacement of about 40 km with the east side north relative to the west. This is a minimum estimate as only the prominent mapped faults are considered. Prominent sets of parallel, closely spaced lineaments are clearly visible on air photographs of the area, particularly in the granitic terranes. Many of these fractures may also be faults, but without markers this is difficult to demonstrate. A few of these have been mapped as faults in the granitic terrane west of Yellowknife where the displacement can be shown by the offset of diabase dykes.

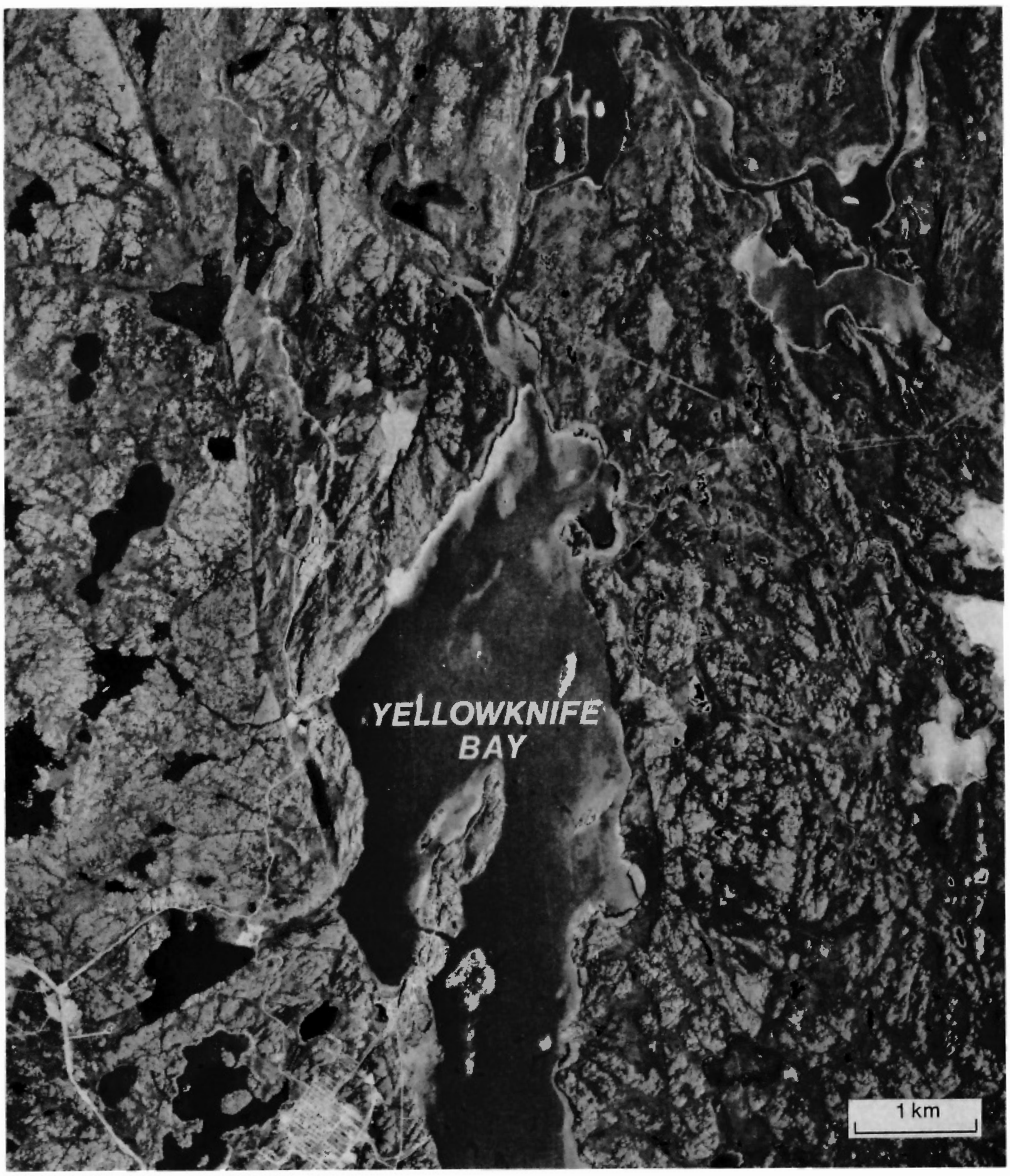


Figure 77. *Part of the prominent north-northwest-trending Proterozoic left lateral West Bay fault along the west side of Yellowknife Bay which has a horizontal offset of 4.9 km and a vertical displacement of 0.45 km at Yellowknife. NAPL A13621-177*

Many of the faults are laterally continuous with some having been traced over 100 km. Most, however, are much shorter. In some cases, as at Tanco Lake near the east margin of the area, the trace of an individual fault shifts abruptly a few hundred metres to the west where it continues with the same northerly trend. The ends of the faults overlap for a kilometre or so, with the severely fractured area between, commonly covered by a lake forming a type of pull-apart basin as discussed by Aydin and Nur (1982); a not uncommon feature of strike slip faults. This happens at all scales. For example, the West Bay fault through Yellowknife disappears 30 km north of Yellowknife in a series of high-angle splays. A second fault, overlapping with the north end of the West Bay fault, begins about 4 km to the west, and continues north for another 100 km. The trend of some faults varies as much as 15 or 20° over its length. In addition at Yellowknife the West Bay fault breaks up into a series of low angle splays that are perhaps a result of the change in mechanical properties of the rock units the fault passes through which change from granite to metavolcanic to metasediment and back across Yellowknife Bay.

Around Yellowknife there are abundant, easily accessible, examples of these faults (Fig. 77). They have been described in detail by Jolliffe (1945), Brown (1955) and Henderson and Brown (1952b, 1966). The faults typically occur as single, sharply defined, more or less vertical planes. Where determined, movement has been dominantly horizontal with a minor vertical component. For example, at Yellowknife the West Bay fault has a horizontal component of 4920 m and a vertical component of 480 m (Campbell, 1947). The fault zones are very narrow, commonly less than 1 m, and in some cases only a few centimetres wide. Displacement has taken place along a single plane. There is no evidence to suggest the faults consist of a series of bifurcating shears as is commonly the case in large transcurrent faults. The initial movement on these faults sharply defined the narrow zone of weakness along which all subsequent movement took place. The fault zones tend to weather low and consist of broken wall rock that ranges from breccia to fine rock powder. The coarser clasts are angular with no suggestion of elongation. Failure was brittle. The contact between the fault zone and the wall rock is generally very sharp. The adjacent wall rock contains fractures related to the faulting and is commonly altered, with the granitic rocks becoming a bright pink. The fault zone itself is commonly silicified and fractures are filled with quartz, hematite and carbonate. In some cases fractures in the country rock are injected with pulverized rock fragments. Pseudotachylyte-filled fractures occur in association with the West Bay Fault in the Defeat granodiorite 9 km north of Yellowknife. At the north end of Duncan Lake a similar occurence of glassy pseudotachylyte-filled fractures within the metasediments is also believed to be related to faulting, although the fault itself has not been identified (Fig. 78). Large quartz veins or areas of brecciated country rock with a quartz matrix, that in some cased are several tens of metres wide occur locally along some of the faults. They are most abundant along the Stagg River Fault and West Bay Fault and tend to occur in granitic rocks where the trend of the fault changes.

In addition to these prominent north-northwesterly to northerly sinistral faults that occur more or less throughout the area, there are other sets that are more restricted in distribution. Between Hearne Lake and Hearne Channel there are several short displacements, generally variably trending, east-northeasterly, left lateral faults. A maximum apparent displacement of about 1 km is seen on the fault between Campbell Bay and Francois Bay. Others through Buckham Lake, Hearne Lake and Blachford Lake have shorter displacements. Farther north, between Gordon lake and Ross Lake, three intersecting faults and a fourth fault 15 km to the north, have a main component of movement that seems to be vertical. These faults have intersected and brought to the surface the cordierite isograd that normally outcrops several kilometres to the west. Similarly, the north-trending fault 8 km west of Desperation Lake has a dominantly vertical component of movement and, like the others to the west, the east side has moved up relative to the west.

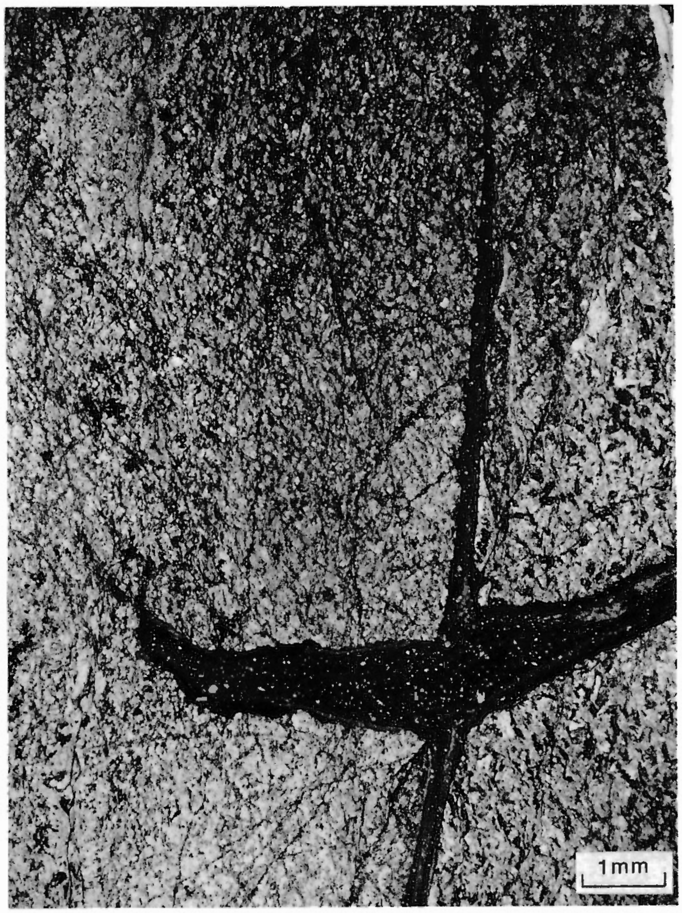


Figure 78. *Pseudotachylyte in Burwash Formation metagreywacke at the north end of Duncan Lake. Pseudotachylyte forms due to the injection of ultra fine grained pulverized rock into fractures formed as a result of high level faulting. The pseudotachylyte is a vaguely banded pale brown material containing scattered clasts of the country rock that grade down into an isotropic, presumably devitrified glassy matrix. Similar structures occur in the Defeat granodiorite near Yellowknife associated with the West Bay fault. No fault has yet been recognized at Duncan Lake. GSC 203660-I*

At the east margin of the area, between Payne Lake and Narrow Island in Hearne Channel are two faults with opposite apparent senses of displacement. The southern fault between Narrow Island and the east side of Meander Lake, like most other faults in the region, has a north-northwesterly trend and an apparent left lateral sense of displacement. The second fault starts on the west side of Meander Lake but has a north-northeasterly trend and a right lateral sense of displacement. The two faults do not appear to intersect in Meander Lake, nor do they continue beyond Meander Lake.

Two sets of short displacement (typically less than a few tens of metres) dextral faults have been described by Henderson and Brown (1966) as a result of their detailed mapping of the mafic metavolcanic rocks at Yellowknife. The two sets have east-northeasterly and east-southeasterly trends and are considered by Henderson and Brown (1966) to be complementary shears and tension faults related to the major north-northwesterly sinistral faults with which they are closely associated.

The ages of the fault sets are difficult to establish. They are clearly post-Archean as they offset all Archean structures and, as Mackenzie dykes are not known to be offset by any of the fault sets, they are presumably pre-1200 Ma, the age of the Mackenzie dykes (Fahrig and Jones, 1969). One of the sinistral, north-northwesterly faults appears to offset the Grace Lake Granite of the Blachford Lake Intrusive Suite, which has an age of about 2.15 Ga, while other faults of this set offset the Hearne dykes which are also younger than the Blachford Lake and predate the Great Slave Supergroup. There is no evidence that these faults affect the nearby early Proterozoic Great Slave Supergroup rocks which have a minimum age of 1865 Ma, the $^{207}Pb/^{206}Pb$ zircon age of the Compton Intrusive Suite laccoliths that intrudes them (Bowring and Van Schmus, 1982). Thus the age of the north-northwesterly fault set would appear to be bracketed between 2.15 and 1.87 Ga. As previously mentioned, the small dextral faults at Yellowknife Bay are considered by Henderson and Brown (1966) to be related to these north-northwesterly sinistral faults. The age of the other faults sets, even their relative ages, are more difficult to determine. In one case the east-northeasterly sinistral faults would appear to postdate the Blachford Lake on the basis of the presence of a prominent linear across the north part of the Grace Lake Granite. It also might be argued that this same linear is offset by the north-northwesterly sinistral set. The group of faults south of Gordon Lake, whose main component of movement appears to be vertical, exist in isolation, and about all that can be said concerning their age is that they are post-Archean.

METAMORPHISM

The supracrustal rocks of the Yellowknife Supergroup have all undergone varied degrees of metamorphism. The metamorphic pattern in the basin is shown in general fashion in the metasediments by the cordierite isograd which delineates areas of greenschist and amphibolite metamorphic grade rocks (Winkler, 1974). Of the granitic rocks, only the Anton and Sleepy Dragon complexes and the Amacher and Detour plutons have undergone significant metamorphism after their emplacement.

Metamorphism of the area on a regional scale was described by Henderson (1941b, 1943), Jolliffe (1939, 1940, 1942, 1946) and Fortier (1947) who recognized a regional metamorphism and a subsequent thermal event related to granitic intrusions. Later more detailed work on the metamorphism of small parts of the area, including detailed chemical studies, includes that of Folinsbee (1942), Boyle (1961), Kretz (1968, 1973), Kamineni (1973a, b, 1974, 1975), Kamineni and Carrara (1973), Kamineni and Divi (1976), Kamineni and Wong (1973), Ramsay (1973a, b, c, 1974a, b), Ramsay and Kamineni (1977) and Kamineni et al. (1979). Thompson (1978) has reviewed the metamorphism of the Slave Province as a whole. The work of Davidson (1967) and Heywood and Davidson (1969) in the geologically similar and nearby Benjamin Lake area is an important study in this regard.

Granitic rocks

Both the Sleepy Dragon and Anton complexes east and west of the main Yellowknife supracrustal basin have been strongly metamorphosed, and indeed it is the contrast between the commonly highly deformed granitoid gneisses of these complexes and the relatively lower grade Yellowknife supracrustal rocks adjacent to them that is an important criterion for the recognition of these units as basement to the Yellowknife. As the major part of the metamorphism of these complexes took place before the deposition of the Yellowknife rocks, their grade and pattern of metamorphism is therefore out of context with that seen in the rest of the area. These rocks have been involved to varied degrees in post-Yellowknife metamorphisms but no attempt has been made to outline the effect of these later events. These metamorphosed units have been described in greater detail in the sections devoted to them.

The only other granitoid units that have been significantly metamorphosed include the Amacher and Detour plutons in the northeastern part of the area. The decussate habit of the biotite aggregates in the Amacher granite suggests that the granite has been metamorphosed despite its well preserved primary igneous textures. It has been interpreted as a possible synvolcanic pluton emplaced and subsequently metamorphosed in the volcanic complex at Sunset Lake. The Detour granodiorite is a quartz-plagioclase porphyry pluton that has been metamorphosed, as indicated by the development of a strong foliation defined by the alignment of fine grained elongate aggregates to single flakes of biotite parallel to the regional foliation. Like the Amacher it is thought to be possibly related to Yellowknife volcanism.

Volcanic rocks

The mafic volcanic rocks are difficult to map from a metamorphic point of view at 1:250 000 scale, as the mineralogical and textural changes with changing grade are often subtle and difficult to detect in the field. The original mineralogy is nowhere preserved although in some cases primary textures are preserved as diffuse outlines. For example, pseudomorphs after originally acicular plagioclase crystals in the lowest grade rocks occur in the least metamorphosed mafic flows. As the grade increases the rocks are increasingly recrystallized with the loss of all traces of primary textures and the growth of chlorite and actinolitic amphiboles. The rocks become increasingly coarser grained and strongly pleochroic hornblende starts to form. In some of the fragmental volcanic units garnet is typically up to 1 cm in size, very irregular in outline, and highly poikilitic. At the highest grades the volcanics are essentially hornblende amphibolites with coarse, well formed, oriented crystals up to 1 cm long. Mineralogical variations are difficult to observe in the field, but with increasing grade the rocks become coarser grained and darker. Primary structures are commonly evident even at the highest grades, provided deformation has not been extreme. Pillow margins in volcanic xenoliths in the intrusive granitic rocks are locally preserved.

Felsic volcanic rocks undergo even less change with increasing metamorphic grade. In general the main effect is a coarsening of grain size while the mainly quartzofeldspathic mineralogy changes little. If highly deformed, the felsic volcanics form quartz-muscovite schists.

Isograds within the mafic volcanic units have not been consistently defined throughout the region. However, in the Yellowknife area Boyle (1961) has defined three metamorphic facies; amphibolite, epidote-amphibolite and greenschist, based on metamorphic mineral assemblages. These facies are parallel to the contact with the Defeat pluton to the west. In the Sunset Lake area, Lambert (in press) has defined a line based on the change from actinolite to hornblende, with the low grade, mainly mafic metavolcanic rocks occupying the core of the main syncline through Sunset Lake. In the sediments to the south, however, the grade is above greenschist. Along the Cameron River the mafic volcanic sequence mainly contains hornblende, but at a few localities at the top of the sequence the rocks are actinolite-bearing (Lambert, in press).

Sedimentary rocks

Jackson Lake Formation

The Jackson Lake Formation lies unconformably above the Kam mafic volcanics at Yellowknife. The sandstones are made up of quartz, felsic volcanic lithic clasts, weathered granitic detritus, and minor varied amounts of plagioclase in a fine matrix of quartzofeldspathic material and muscovite. The metamorphic assemblage consists of randomly oriented to radiating clusters of chloritoid laths, very fine grained white mica throughout the matrix, but more coarsely and abundantly developed in pressure shadows behind framework grains, and irregular patches of chlorite, along with quartz and plagioclase. Carbonate is present in varied amounts but where abundant there is no chloritoid. According to Winkler (1974) and Hoschek (1967) this indicates low grade metamorphism of a rock whose bulk chemistry indicates a high Fe/Mg ratio, a relatively high Al content and low K, Na and Ca. This is supported by the lack of biotite in these rocks. It has been suggested that this somewhat unusual composition (Table 7) is due to weathering of source volcanic and granitoid rocks (see section on Jackson Lake Formation; Schau and Henderson, 1983). This metamorphic assemblage contrasts strongly with that in the metagreywacke-mudstone rocks to the east.

Burwash Formation and other metagreywacke-mudstones

The Burwash Formation and its lithological equivalents at Russell Lake and Stagg Lake are generally of uniform composition and occur over a wide area. The formation is made up of interbedded mudstones and greywackes which, where metamorphosed under varied conditions, contain a variety of mineral assemblages, many of which can be easily identified in the field. This unit then is ideal for reflecting the variation in metamorphism throughout the area.

The general pattern of metamorphism can be seen in the distribution of cordierite (see map, in pocket; Fig. 79, Table 22). The first occurrence of cordierite is an easily mapped feature that has been used to mark the boundary between greenschist and amphibolite grade metamorphosed rocks. The supracrustal rocks in the Russell Lake – Stagg Lake area are all at amphibolite grade, whereas in the terrane east of Yellowknife there are large areas of

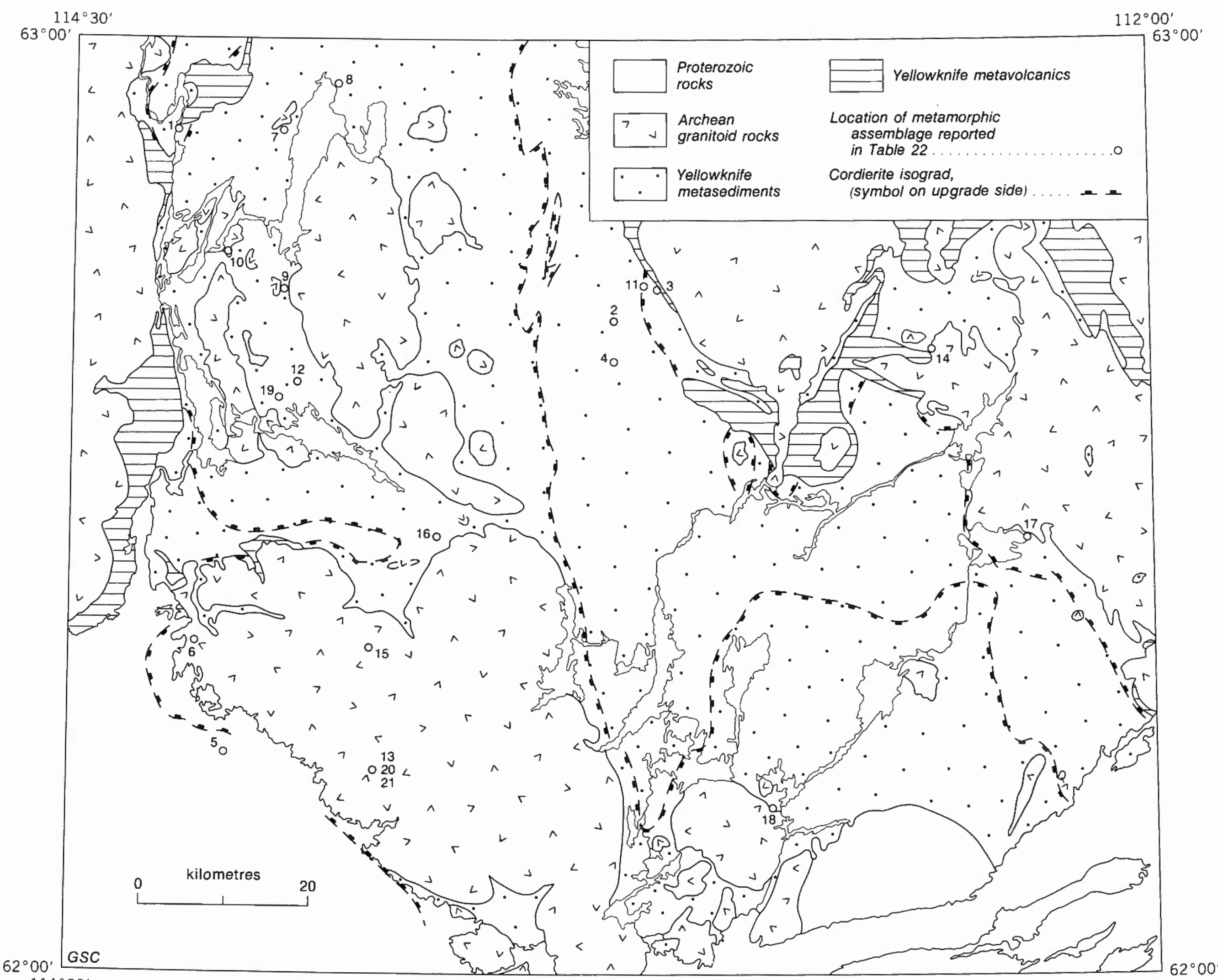


Figure 79. *Location of metamorphic mineral assemblages (1-21) in the Burwash Formation recorded in Table 22.*

Table 22. Metamorphic mineral assemblages in the Burwash Formation

Assemblage	Quartz	Plagioclase	Chlorite	Muscovite	Biotite	Garnet	Cordierite	Staurolite	Andalusite	Sillimanite	Other
1	X	X	C	X							
2	X	X	C		X						
3	X	X			X						
4	X	X	C		X						(E)
5	X	X	C	X	X						
6	X	X	C_2		X	X					
7	X	X	C_2	X	X		X		X		
8	X	X	C_2	2	X		X				
9	X	X			X		X				A
10	X	X	C_2		X	X	X				
11	X	X	C	X	X				X		
12	X	X	C_2		X	X	X	X	X		
13	X	X	C_2		X	X	P		(X)		
14	X	X	C	X	X		X	X			
15	X	X	C,C_2	2	X	X	X				
16	X	X	C	X	X		X				
17	X	X	C,C_2	(2)	X	X	X				
18	X	X	C_2	(X)	X		X		X	X	
19	X	X	C_2	(X)	X		X	(X)	X	(X)	
20	X	X	C_2		X		P			X	
21	X	X		(X)	X		X			X	

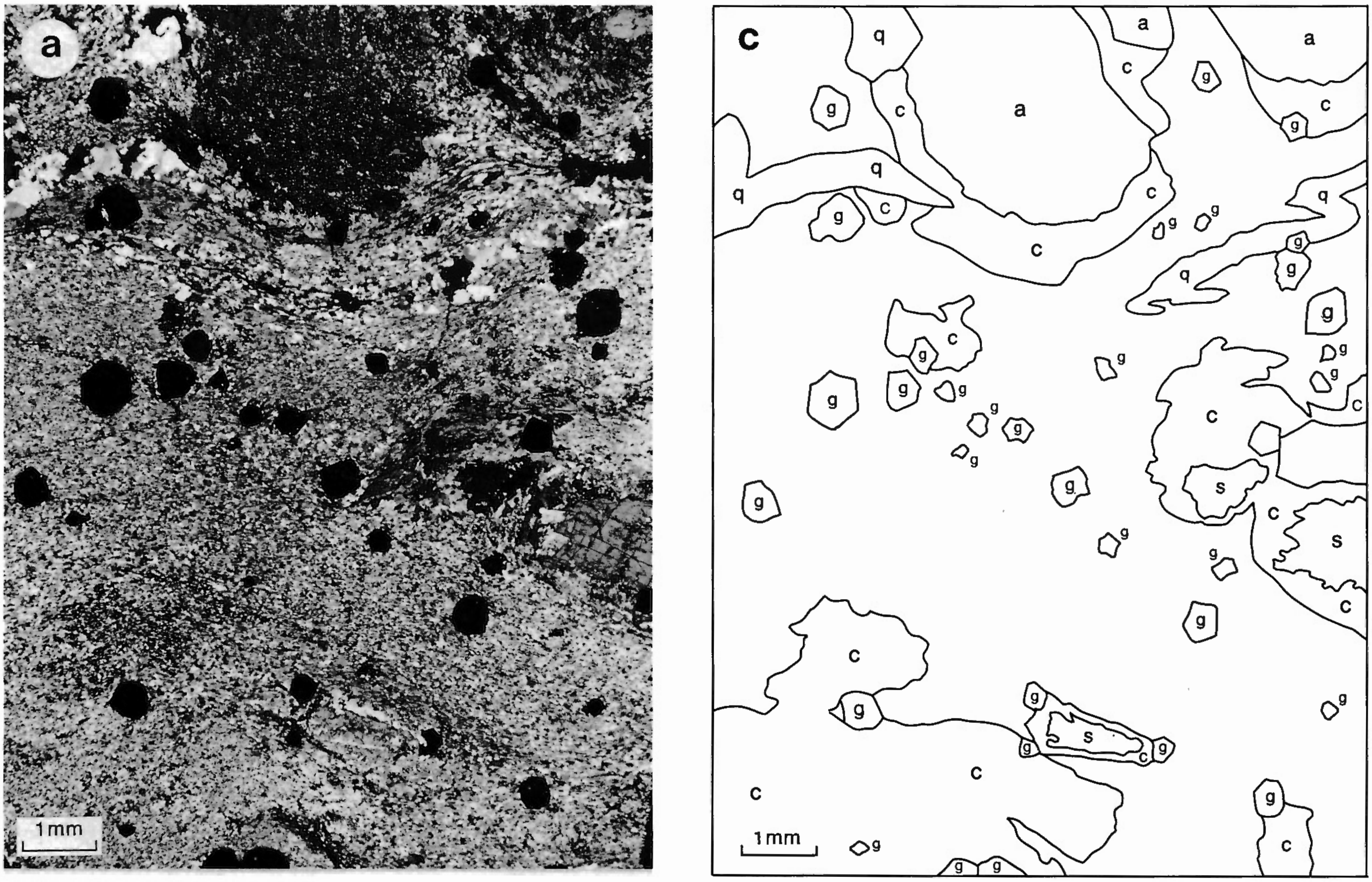
See Figure 79 for location of assemblages
X Mineral present
(X) Trace amount of mineral present
2 Two habits of same mineral present
F Fine grained chlorite
C Coarse grained chlorite
P Pinnite after cordierite
E Epidote
A Anthophyllite

amphibolite grade rocks occurring in the centre of the area and in relatively narrow zones at the margins with greenschist metamorphic grade rocks between.

The rocks of this formation are everywhere metamorphosed to some degree (Fig. 79, Table 22). Nowhere is the original mineralogy preserved. The lowest grade rocks occur in a small area at Yellowknife Bay north of Akaitcho Bay and over larger areas at Gordon Lake (Henderson, 1943) and the north end of Hearne Lake. These rocks consist of the assemblage quartz, plagioclase, chlorite, muscovite, epidote, clinozoisite and carbonate, and are free of metamorphic biotite. The original textures in these rocks are well preserved with quartz, the lithic clasts (dominantly felsic volcanic) and plagioclase having somewhat serrated detrital outlines. The matrix consist of fine quartzofeldspathic material and fine laths of muscovite and chlorite, although the chlorite can occur as larger amoeboid patches and in rare cases as coarse flakes that may represent original detrital biotite, now retrograded. Epidote and clinozoisite occur as scattered equant grains.

At higher grades biotite appears as irregular flakes, commonly with inclusions, and with a wide grain size range. The orientation of this early biotite is strongly controlled by the coarser detrital quartz and feldspar grains of the original sediment. In the very fine pelitic rocks the biotite tends to have a better crystal form and more random orientation than in the coarser greywackes. The origin of biotite in pelitic rocks is not easily understood as indicated by Brown (1974, p. 75) who suggested that 'the reactions proposed for it are almost as numerous as the persons who have written articles on the subject'. Winkler (1974) suggested that phengite and chlorite react to form biotite, a more aluminous chlorite and quartz: a possibility, but not yet recognized. In the Yellowknife area Ramsay (1973a, b, c) has suggested that the biotite in the vicinity of Prosperous Lake formed due to reaction of chlorite, muscovite and ilmenite to produce rutile, K-feldspar, quartz and a muscovite of different composition, along with biotite. This has generated some controversy (Brown, 1974; Ramsay, 1974b). As suggested by Ramsay (1973a) the bulk composition along with the temperature are important controls on the formation of biotite. Commonly, beds of presumably only slightly varied composition on the same outcrop, in the vicinity of the biotite isograd, in some cases will be free of biotite while in adjacent beds it is well developed.

With increasing grade the matrix becomes coarser grained and the finer grained rock fragments recrystallize and commonly lose their identity. Biotite becomes coarser and more dominant in the fabric of the rock. Plagioclase, as at lower grades, is commonly clouded with fine grained



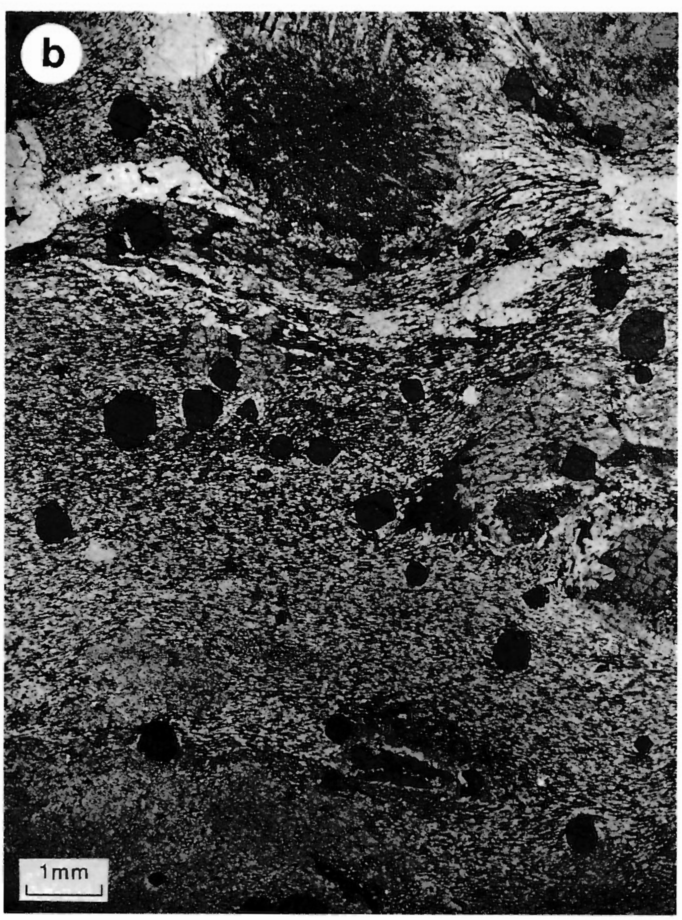


Figure 80. *Burwash Formation metapelite. The mineral assemblage in this pelitic rock from 3 km north of the west end of Prelude Lake is quartz (q)-plagioclase-biotite-andalusite (a)-staurolite (s)-garnet (g)-cordierite (c). The staurolite and andalusite are commonly mantled by cordierite. (a) Crossed polarizers. GSC 201859-W. (b) Plane light. GSC 203660-C.*

alteration products due to the equilibration of composition with metamorphic grade. Eventually the rock becomes totally recrystallized, commonly with the loss of most primary textures. The texture is much more even grained with interlocked grains of quartz, unaltered plagioclase and fine flakes of chlorite and white mica. Biotite can occur as large, commonly poikilitic flakes, or as groups of scattered grains in optical continuity. The mineral assemblage in the metasediments at uppermost greenschist grade includes quartz-plagioclase-biotite-chlorite-white mica and minor epidote and clinozoisite.

Amphibolite metamorphic grade starts with the first appearance of cordierite (Winkler, 1974). The cordierite isograd is easily mapped in the field as the mineral first appears at the isograd as coarse porphyroblasts in rocks of appropriate composition. This isograd has been included on almost all of the geological maps of the southern Slave Province since mapping was begun in the 1930s. At amphibolite grade the metasediments consist of assemblages of the following minerals: quartz, plagioclase, biotite, cordierite, andalusite, garnet, amphibole, staurolite and sillimanite (Table 22). Of these, quartz, plagioclase and biotite are present in almost all of the metasediments while the remainder are more varied in their occurrence. Chlorite is common in most rocks at this grade, but is of secondary origin, reflecting a retrogressive event that postdated the peak of metamorphism.

Cordierite is present in most of the metasediments at this grade. The mineral occurs as elongate rounded porphyroblasts up to 5 cm in length but more typically between 1 and 2 cm. They tend to be highly poikilitic with inclusions of quartz, plagioclase and biotite, which results in a rather diffuse appeareance on the outcrop, although in thin section the mineral is sharply defined. In many cases the inclusions are finer grained than similar minerals in the present matrix of the rock. In rare instances, cordierite mantles other earlier metamorphic minerals such as staurolite and andalusite (Fig. 80). Cordierite is susceptible to alteration, with the margins commonly replaced with a pale yellow pinnite. In some grains, cordierite is entirely altered to chlorite. It has been suggested that cordierite forms due to the reaction of chlorite, muscovite and quartz to form cordierite and biotite (see e.g. Seifert, 1970). For the rocks in the vicinity of the Prosperous Granites, Ramsay (1974a), on the basis of changes in the biotite chemistry, suggested biotite is present on both sides of the reaction. The orientation of the cordierite is highly variable from random to strongly parallel to the cleavage in the host rock (Fig. 81). In most cases the cordierite has overgrown, and is therefore later than, the prominent cleavage in the metasediments. Commonly, strongly oriented to tightly crenulated fine micas occur as inclusions within cordierite porphyroblasts in a matrix containing coarse, randomly oriented biotite plates of presumably later generation than that within the cordierite (Fig. 82). Orientation of some cordierite parallel to the cleavage is thought to be due to mimetic crystallization; in some instances cordierite has grown parallel to two directions of cleavage in the same rock. In the immediate vicinity of some of the Defeat plutons the cordierite porphyroblasts

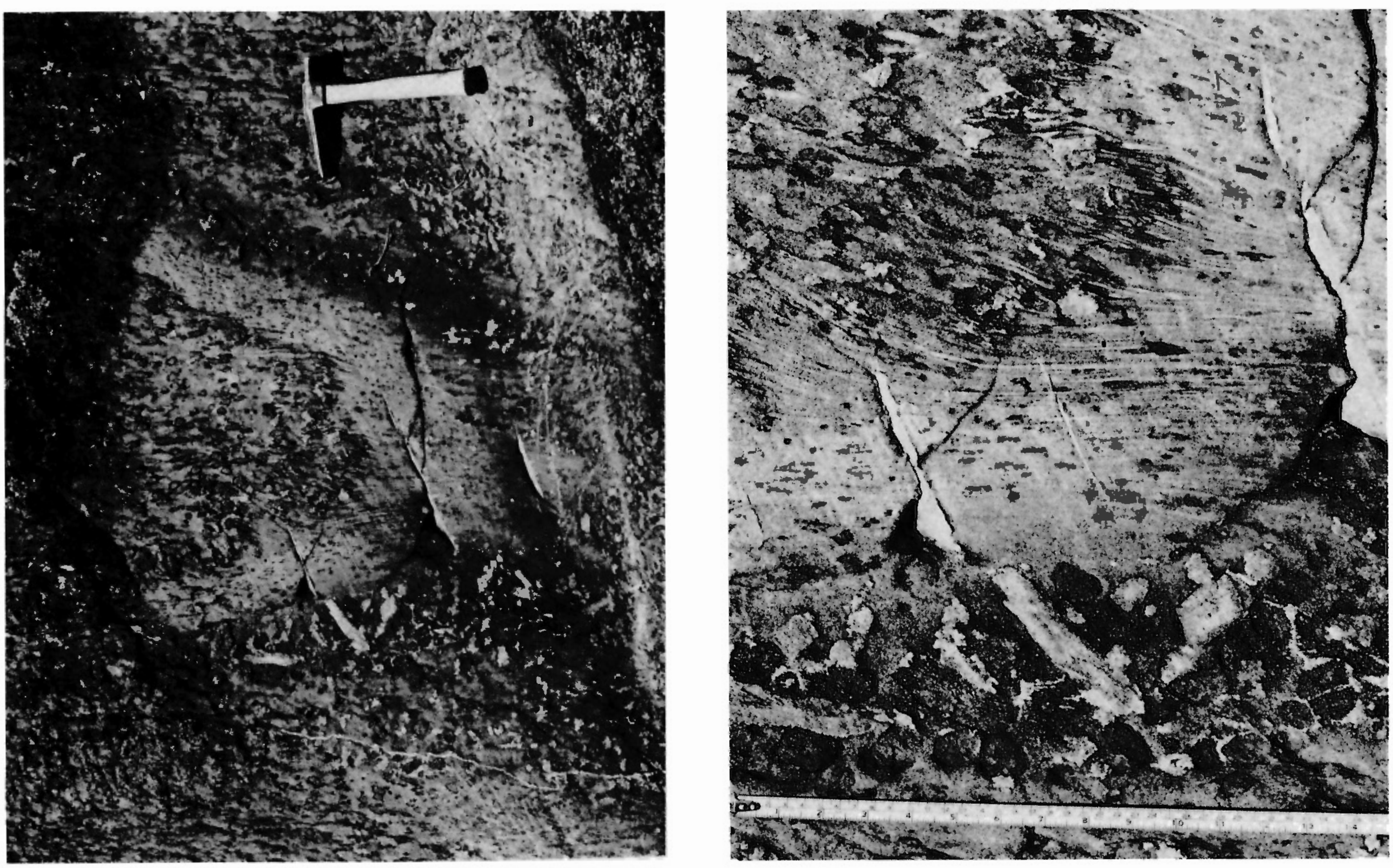


Figure 81. *(a) Tight fold in amphibolite facies Burwash Formation on west side of Duncan Lake. The light layer defining the fold is more psammitic in composition. GSC 177619. (b) Detailed view. The dark megacrysts are cordierite or altered cordierite, and the light metacrysts are andalusite. Note the strongly developed cleavage and the megacrysts flattened in the cleavage near the nose of the fold. GSC 177622.*

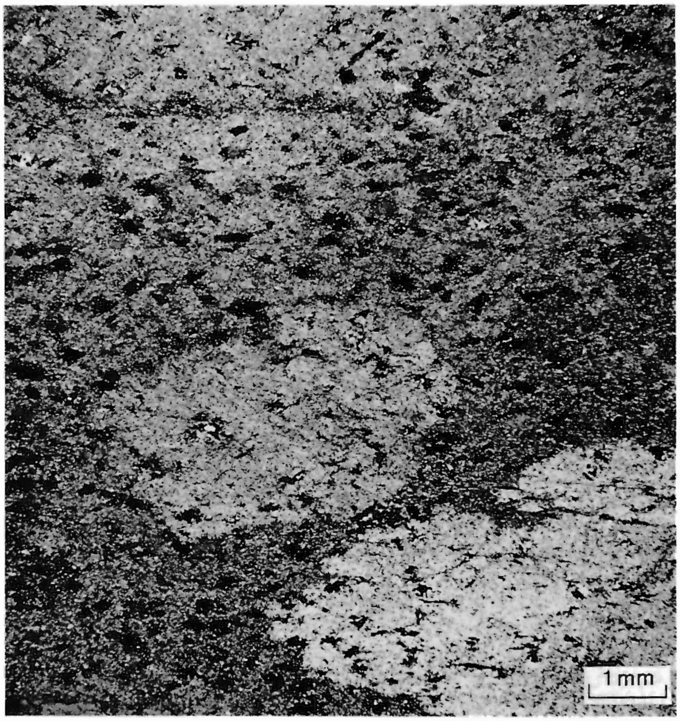


Figure 82. Cordierite in Burwash metagreywacke east of Tumpline Lake. Fabric in the metasediment is defined by the strong orientation of fine muscovite and has been overgrown by both the cordierite and coarse biotite. Compare with Figure 83. (a) Crossed polarizers. GSC 203660-G. (b) Plane light. GSC 203660-H.

are commonly rotated and evidently were involved in deformation after they started forming, possibly related to the emplacement of the nearby pluton (Fig. 83).

Andalusite is the other common coarse porphyroblastic mineral in the metasediments. It is generally smaller, and in many cases tends to have a better crystal form and weather higher, than cordierite. Andalusite is generally less abundant, although in some areas, as near Jennejohn Lake, it is more abundant than cordierite. Coarse pink aggregates of chiastolitic andalusite also occur in segregations with quartz in black carbonaceous metaargillaceous units that are most abundant in the Duncan Lake – Prelude Lake areas. The andalusite porphyroblasts are typically poikilitic and, like the cordierite, the inclusions within the crystals are commonly smaller than the present matrix of the rock. In some andalusite porphyroblasts, sharply defined zones are free of inclusions, while adjacent zones contain more than 50% included material. Throughout most of the area the first occurrence of andalusite is somewhat removed from the cordierite isograd, suggesting it formed under somewhat higher metamorphic grade conditions. In the Benjamin Lake area 60 km to the northeast, Heywood and Davidson (1969) have also noted the coexistence of andalusite and cordierite some distance from the cordierite isograd, and have suggested the andalusite may form due to the reaction of cordierite and muscovite. However, around the Defeat plutons near Jennejohn Lake, andalusite occurs very close to the isograd.

Garnet occurs sporadically in the medium grade rocks. It typically occurs as scattered pink and red anhedral to subhedral crystals. Most are poikilitic to some degree and some are very patchy, consisting of diffuse areas with more inclusions than garnet. They are most abundant in silicate iron formation such as that which occurs on the islands in central Russell Lake. There over half the rock can be formed

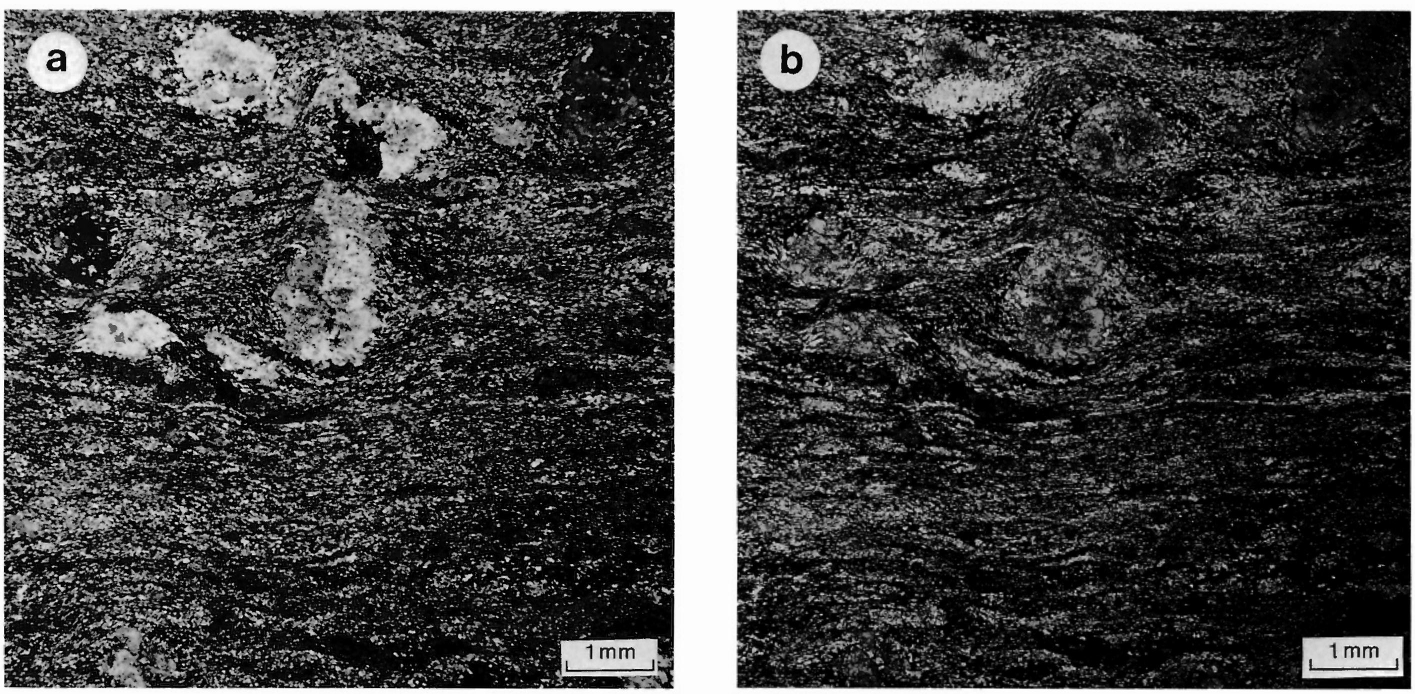


Figure 83. Cordierite in Burwash metagreywacke northeast of Watta Lake. The strong fabric in the metasediment involves all mineral phases with some of the cordierite porphyroblasts appearing to have been rolled in the foliation. This sample comes from near the contact of the Defeat granodiorite pluton at Watta Lake. The deformation of the cordierite in the foliation suggests that the foliation is synchronous with, or postdates, porphyroblast formation. The metamorphism initiating the formation of the cordierite may have preceded, probably by a short period of time, the deformation associated with the emplacement of the pluton. Compare with Figure 82 which is remote from any pluton. (a) Crossed polarizers. GSC 201859-Z. (b) Plane light. GSC 202890-X.

of garnet. Like andalusite, garnet occurs at some distance from the cordierite isograd, generally at higher grades than the first appearance of andalusite. The occurence of garnet is sensitive to bulk compositional constraints (Winkler, 1974). This is evident in the Yellowknife-Hearne Lake region as garnet is common in most parts of the thermal ridge through Prelude and Duncan lakes, although there are also areas where it is not present. In the other large amphibolite grade area about Buckham Lake, on the other hand, garnets are rather sporadically developed. Kretz (1973) in a study on garnets at Staple Lake, south of Duncan Lake, proposed that the garnet formed in an early reaction of chlorite, muscovite and quartz to produce garnet, biotite and cordierite, with the nature and proportion of the products depending on the initial composition of chlorite. This early formation of garnet is supported by the occurrence in at least one locality of an euhedral garnet enclosed in staurolite which in turn is mantled by cordierite (Fig. 80). Jolliffe (1942) reported the occurrence of garnet below the cordierite isograd, but only in "hornblendic" beds that possibly represent tuffs, calcareous concretionary lenses or layers, or silicate iron formation. Garnets are common in greenschist grade silicate iron formations in the Contwoyto Lake area of the central Slave Province (Tremblay, 1976).

Amphiboles occur in the metasediments above the cordierite isograd. Their formation, to an even greater extent than garnet, is very dependent on bulk composition. Gedrite, an aluminous orthoamphibole was described by Kamineni and Wong (1973) from near Sparrow Lake. It occurs as elongate, pale brown crystals with random orientation, particularly in the psammitic parts of generally somewhat rusty weathering beds. Adjacent beds are generally free of the mineral. It occurs throughout the area in metasediments of appropriate composition and metamorphic grade. Kamineni (1973b, 1975) has defined a gedrite isograd between the cordierite and sillimanite isograds at Sparrow Lake. According to Kamineni (1975) gedrite in that area does not occur above the sillimanite isograd. Cummingtonite is also present in the metasedimentary sequence and, like the gedrite, is compositionally restricted, occuring only in calcareous metaconcretions as yellow, elongate, randomly oriented, individual, acicular crystals.

Staurolite is rare in these rocks. It has been reported from several localities in the amphibolite grade area in the vicinity of Prosperous Lake and Hidden Lake (Jolliffe, 1946; Ramsay, 1973c, Kamineni, 1973b, 1975, Ramsay and Kamineni, 1977) and also occur at several localities between Turnback Lake and Desperation Lake. The relatively rare occurrence of staurolite reflects the more iron-rich composition of the metasediments in which it occurs. It forms very poikilitic subhedral to euhedral crystals up to 4 or 5 mm long. In many cases the staurolite is extensively resorbed and is almost everywhere mantled by cordierite.

The highest grade rocks occur as inclusions within the plutons, in mixed zones of metasediments and granitic phases associated with some of the plutons, and in narrow zones immediately adjacent to the plutons. Both the Defeat granodiorite and the Meander Lake granite have abundant inclusions that in some instances are migmatitic and in most cases are highly deformed. Mineral assemblages in the inclusions include quartz, plagioclase, and biotite with varied proportions of garnet, cordierite, sillimanite and muscovite. Inclusions in the Prosperous granites are much less deformed, commonly randomly oriented and with little or no evidence of migmatization. In the Stagg Lake area, where the regional metamorphic grade is higher than elsewhere, the sediments are extensively recrystallized and most primary structures, other than compositional layering, have been lost. There, sillimanite occurs locally as large blocky crystals apparently pseudomorphous after andalusite. High grade rocks occur in narrow aureoles about individual plutons. Jolliffe (1946), Kamineni (1975) and Ramsay and Kamineni (1977) reported fibrolitic sillimanite associated with andalusite in narrow zones adjacent to granitic bodies of the Prosperous type. Sillimanite also occurs with similar habit and association adjacent to some of the larger Defeat plutons. Davidson (1972) reported sillimanite in the highly metamorphosed pelitic rocks west of Turnback Lake.

Pattern of metamorphism

In the most general terms the basin east of Yellowknife represents the lowest grade terrane within the area while that area outside the basin, being dominated by granitoid plutonic rocks, represents a much higher grade. Within the basin the vast majority of amphibolite grade rocks are associated with thermal ridges in the central parts of the basin. The largest of these is the north-trending west-central ridge that continues across the map area and extends as a distinct entity some 30 km to the north of the area. A second thermal high occurs to the east of the first in the southern part of the basin where the basin becomes significantly wider. These thermal highs are surrounded for the most part by basinal rocks at greenschist grade. Much smaller areas of higher grade rocks also occur in narrow zones at the margin of the basin.

The gross pattern of metamorphism, then, is on the scale of the basin. That is to say, it is not the reflection of a broad regional gradient as, for example, appears to be the case in the Healey Lake area in the eastern Slave Province (Henderson et al., 1982; Henderson and Thompson, 1982). There, the southeasterly increasing Archean metamorphic gradient (presumably reflecting an increasingly deeper level of erosion across the map area) appears to be largely independent of local features within the map area. With the metamorphic pattern within the Yellowknife-Hearne Lake area appearing to be closely related to the form of the basin, it is not unreasonable to suggest that the distribution of the thermal anomalies that are responsible for the pattern are perhaps ultimately related to disturbances caused during the original formation of the basin as a graben.

At a more detailed level, in almost all cases the pattern of metamorphism can be related to the emplacement of granitic intrusions. Within a significant part of the metasedimentary terrane east of Yellowknife the supracrustal rocks are at low grade, but the grade rises, commonly quite abruptly, toward the present margins of the basin. For example at Yellowknife there is a relatively steep metamorphic gradient across the volcanic sequence normal to the contact with the intrusive Defeat pluton (Boyle, 1961). Before the contact with the sediments is reached the metamorphic grade is in the greenschist facies. Similarly to the north in the vicinity of Clan Lake the isograds approximately parallel the contact with the granitic rocks, at least some of which are younger than the supracrustal rocks. On the east side of the basin, in the vicinity of Upper Ross and Victory lakes, the metamorphic gradient as outlined by the cordierite isograd is not normal to the supracrustal-granitic rock contact. There the isograd passes from the metasediments, towards the Cameron River mafic metavolcanics and the older Sleepy Dragon Complex. The cordierite isograd within the metasediments is approximately parallel to the margin of the Redout Granite which is intrusive into the basement complex (Davidson, 1972; this report). Lambert (in press) reports that the actinolite-hornblende transition in the mafic volcanic rocks along the Cameron River is near the top of the volcanic section, which would suggest the metamorphic conditions indicated by the cordierite isograd in the metasediments probably continued northeasterly within the volcanic sequence. The far eastern side of the basin north and south of Desperation Lake is also at higher metamorphic grade although the width of the zone as indicated by the cordierite isograd is quite varied.

Within the basin, as previously mentioned are several large, higher grade, metamorphic areas. In the vicinity of Duncan and Prelude lakes the broad thermal ridge as defined by the cordierite isograd is, within the map area, 40 km wide and 60 km long, but also extends some 30 km north beyond the area. In the greenschist area between the Gordon Lake and Pensive Lake a lens of amphibolite grade rocks has been raised to the present erosion level. Although the amount of movement on the fault is not known, the occurrence of these higher grade rocks a kilometre or so east of the main cordierite isograd may suggest a relatively moderate dip of the isograd. The main thermal ridge contains numerous plutons, stocks and pegmatites of the Prosperous type. To the south, east of Yellowknife and separated from the Prosperous thermal ridge by a narrow tongue of greenschist grade metasediments, is a higher grade area occupied mainly by a large complex of Defeat plutons. The plutonic complex is mantled by a relatively narrow zone, commonly only 2 or 3 km wide, of amphibolite grade metasediments. This narrow zone continues around the plutonic complex to its southwest side where the cordierite isograd can be traced between the islands in Great Slave Lake, coming as close as 1 km from the intrusion. A third major high grade area occurs north of Hearne Channel, centred about Buckham Lake. This thermal dome, as defined by the cordierite isograd, contains within it only a few, relatively small, plutons of the Defeat granodiorite in the south and a single small Prosperous type pluton in the central northeast. The much younger early Proterozoic Blachford Lake Intrusive Suite, which occurs in the southern part of this high grade area, has a contact aureole of its own superimposed on the Burwash sediments (Davidson, personal communication, 1978). Other Archean metamorphic zones include higher grade zones about the small Prosperous type plutons in the vicinity of Consolation Lake. Although these plutons are only a few kilometres in size, the distance between the pluton and the cordierite isograd that surrounds them is about the same as that about the very much larger Defeat plutonic complex to the southwest.

Although all the intrusions are mantled by amphibolite grade metasediments that, particularly in the case of the Defeat plutons, may be only a few kilometres wide and closely conform to the shape of the complex, the aureoles are not thought of as contact metamorphism in the strict sence. Contact metamorphism due only to the heat contained by the intrusive body rarely extends laterally beyond 1 km from the intrusive contact (Turner, 1968). Even the aureoles about the Defeat plutons that closely parallel the contact are too broad to be due to a purely contact effect. Perhaps they should be considered more a result of heat evolved from the same source that formed the intrusion, that rose faster than the pluton itself, metamorphosing the country rocks shortly before the arrival of the pluton. In some cases a true contact effect due to the presence of the intrusion is superimposed on the thermal aureole in the metasediments surrounding it. This appears to be an apparent retrogressive event, for as the contact with the intrusion is approached, the normal large cordierite porphyroblasts in the aureole become smaller and eventually disppear over a distance of one or two hundred metres.

In the case of the thermal ridge in the vicinity of Prelude and Duncan lakes, the various plutons, stocks and pegmatites occupy a relatively small proportion of the area outlined by the cordierite isograd. At this level of erosion the thermal event that presumably resulted in the formation of the granite bodies is expressed more by the extent of the thermal ridge than by the amount of plutonic material. The high grade area to the southeast about Buckham Lake is perhaps an even more extreme example of this situation, as only a few small plutons and pegmatites are exposed within it at the present erosion level.

The ages of the various metamorphic high grade areas may differ one from another as there is evidence that some of the plutons associated with the thermal highs differ in age. For example, the Defeat plutons are generally quite concordant with the supracrustal rocks they intrude, suggesting they may be older than the Prosperous granites to the north which are more discordant. Radiometric age data supports this as a Rb-Sr whole-rock isochron on the Defeat granodiorite (southeast granodiorite) indicates an event at 2585 ± 37 Ma compared to 2520 ± 25 Ma for a Rb-Sr mineral isochron on the Prosperous granite (Green and Baadsgaard, 1971; Green et al., 1968). The thermal aureoles associated with these intrusions presumably also differ in age, so that it can be argued that between Jennejohn Lake and Reid Lake the cordierite isograds representing metamorphic gradients associated with the two temporally separated events may intersect (Fig. 84). Similarly, the high grade area about Buckham Lake contains granitic plutons of both the Defeat and Prosperous types among others. If these plutons are of similar age to the main bodies of the respective suites there are presumably overlapping thermal events within these areas as well.

The relative ages of the various thermal events at the margin of the present day basin are less well understood. Geochronological data from the Defeat Plutonic Suite west and southeast of Yellowknife are more similar (Table 12) and

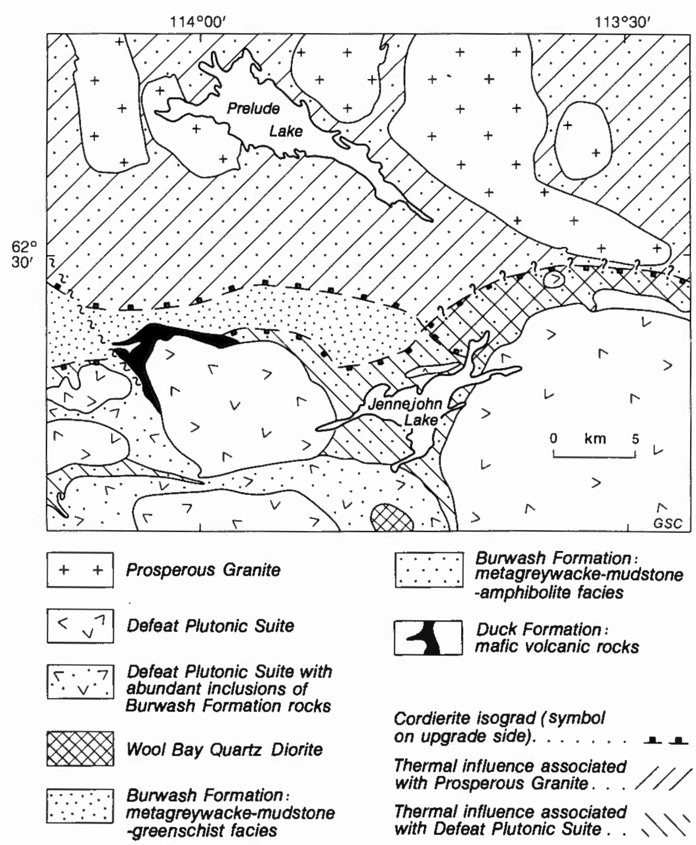


Figure 84. *Zone of possible superimposed metamorphism. The metamorphic zonation brought about by the broad thermal ridge around the Prosperous granites on the basis of isograd patterns may intersect the relatively narrow metamorphic zonation about the Defeat granodiorite, as defined in both cases by the cordierite isograd northeast of Jennejohn Lake.*

so the associated metamorphic events are probably of similar age. To the north, near Clan Lake, the intrusive events are more complex and not well understood. One pluton of the Prosperous suite is involved but the intrusive bodies within the Anton complex may represent several ages, so the metamorphic history may well be very complex. The thermal zonation at Victory Lake seems to be related to the Redout Granite but the age of this granite relative to the other intrusive events is not known. The metamorphism about the Prosperous type granites south of Detour Lake may be similar to that of the main Prosperous Granite. The relative age of the Meander Lake granite along the east margin of the map area to the other granitoid units is not known. The metamorphic picture in the vicinity of the Beaulieu River east of Turnback Lake is probably complex, as four suites of plutons of possibly more than one age are present in the region.

The conditions of metamorphism are difficult to define due to "... insufficient and contradictory experimental data, the variety of petrogenetic grids in the literature, and the correspondence of isograds in the Slave Province with mineral reactions that have not been calibrated experimentally" (Thompson, 1978, p. 93). At the simplest level, the metamorphic assemblages within the map area represent relatively low pressures, as indicated by the lack of kyanite among the aluminosilicates. Some detailed work has been done to investigate the conditions of metamorphism for small areas. In the higher grade region between Yellowknife and Ross Lake, Ramsay and Kamineni (1977), and Kamineni et al. (1979) have suggested, on the basis of paragenetic mineral relationships, that several phases of metamorphism were involved in the formation of the thermal aureole, although they stress that these phases are part of a single event in which temperature and pressure conditions varied. Ramsay and Kamineni (1977) suggested that for this area pressures during the first phase ranged between 300 and 450 MPa with temperatures up to 500° based on the initial occurrence of staurolite before andalusite and formation of garnet rather than cordierite which comes in later. This first stage was followed by a drop in pressure to between 250 and 350 MPa with the maximum temperature rising close to 700°C near the contact with the plutons when cordierite, gedrite, andalusite and sillimanite assemblages formed. Thompson (1978) however has presented arguments by which these same mineral parageneses can develop entirely at lower pressures without an initially higher pressure. As similar mineral assemblages occur throughout the area, metamorphic conditions, though possibly separated to some extent in time, were everywhere at relatively low pressures.

Throughout the region there has been a localized retrograde event that has resulted in the alteration of biotite to chlorite although the original biotite textures are preserved. More commonly the replacement of the mineral is only partial. East of Yellowknife, Ramsay and Kamineni (1977) have related these retrogressive events to the late transcurrent faults which is supported by the locally bleached aspect of the rock adjacent to these faults. In addition, the range of K-Ar dates in the region (Fig. 46, 65) suggest that during the early Proterozoic there may have been several thermal events possibly associated with the movement of water that may have resulted in the retrogression of the Archean rocks.

PALEOZOIC GEOLOGY

The area southwest of the North Arm of Great Slave Lake is underlain by a very gently southwesterly dipping homocline of Paleozoic shales, sandstones, carbonates and evaporites. A preliminary account of these rocks and their distribution over a much larger area was given by Douglas and Norris (1960) and a more complete description was presented by Norris (1965) and Richmond (1965). Law (1971) has reviewed the subsurface Paleozoic geology of the southern Northwest Territories (Fig. 85). In the course of mapping the area the Paleozoic rocks were not re-examined

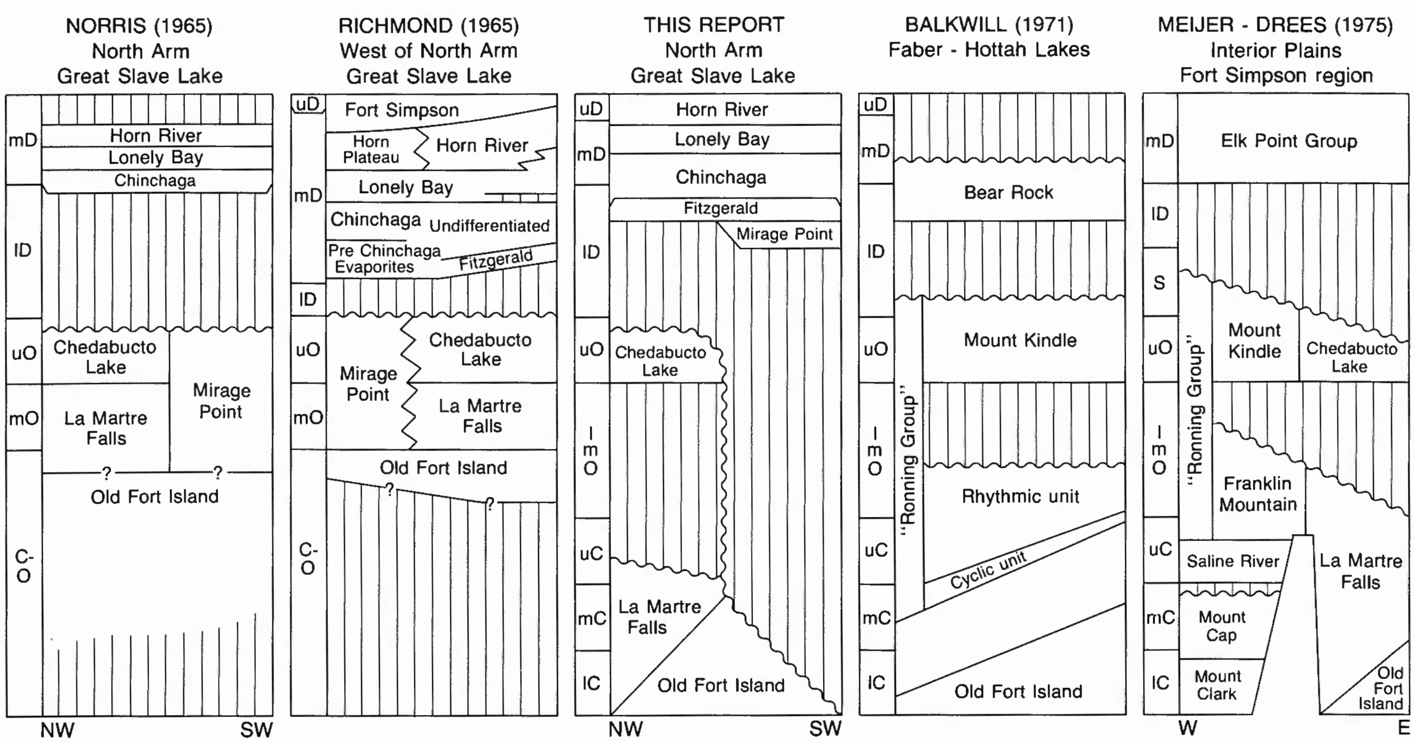


Figure 85. Correlation chart for Paleozoic stratigraphy. The stratigraphy used in this report is compared with the stratigraphy used by previous workers in the area and with two regions to the west and northwest. Vertical pattern represents no record due to nondeposition or erosion and the wavy contact lines represent unconformities.

and the following descriptions are abstracted mainly from Douglas and Norris (1960), Norris (1965) and Richmond (1965). Norris (1965) and Richmond (1965) in particular have presented detailed section descriptions and lists of fossil assemblages from various units in the region.

The Paleozoic rocks are restricted to the area southwest of North Arm, Great Slave Lake, except for a small outlier at Rae Point. No other outliers of Paleozoic rocks were found in the course of mapping the Archean terrane along the north shore. The Paleozoic section consists of eight formations. The Old Fort Island, La Martre Falls, and Chedabucto Lake formations, are of Ordovician age or older. They are unconformably overlain by the Mirage Point, Fitzgerald, Chinchaga, Lonely Bay and Horn River formations of Devonian age.

Within the map area the southwesterly dipping homocline has a dip of about 5 m/km. Topographic irregularities on the Precambrian erosion surface strongly influenced the distribution and thickness of the older formations. The maximum thickness of the Ordovician and older units occurs to the northwest of Great Slave Lake. They become thinner to the south and southwest. A southwesterly trending high, termed the Fort Rae arch (Douglas and Norris, 1960), over which the units are significantly thinner, extends from the vicinity of Rae Point for some 150 km. The area was subjected to periods of erosion during Silurian and Early Devonian time and also for a period during Middle Ordovician time (Fig. 85). Lineaments on the Paleozoic terrane fall within the orientation range of the northeasterly and north-northwesterly faults in the Precambrian terrane to the north and east. There is no evidence of Paleozoic movement on these features (Norris, 1965).

Old Fort Island Formation

The Old Fort Island Formation is a sandstone that lies unconformably above the Archean basement. It is overlain by shales, sandstones and dolomites of La Martre Falls Formation north of Alexander Point, while to the south it is overlain by carbonates of the Mirage Point Formation. It is poorly exposed but can be seen on the south side of Wrigley Point, on Old Fort Island (the type locality) and on the Louise Islands. It consists of thin- to thick-bedded, fine- to coarse-grained, varicoloured but mainly white, friable quartzose sandstone with some thin beds of greenish-grey and dusky red siltstone and minor laminae or partings of green shale. Where exposed on Old Fort Island the lowermost beds are crossbedded. No complete section is exposed. The formation is unfossiliferous but near Hottah Lake, 300 km to the north-northwest, it is overlain conformably by Middle Cambrian strata (Balkwill, 1971). As Balkwill (1971, p. 13) suggests, a Lower to Middle Cambrian age for the formation is indicated "although as a basal, transgressive clastic deposit its age probably differs from place to place along the margin of the Canadian Shield".

La Martre Falls Formation

La Martre Falls Formation is dominantly shale and lies between the overlying Chedabucto Lake dolomites and the underlying Old Fort Island sandstones, although in the vicinity of Pointe du Lac it directly overlies Archean granitic rocks. Within the map area it is poorly exposed but can best be seen between Redrock Point and Wrigley Point. It is about 30 m thick within the area but increases to in excess of 110 m towards the northwest. To the southeast it is truncated by the Mirage Point Formation argillaceous dolomite in the vicinity of Alexander Point. The formation consists of soft, olive-green, rarely black, fissile, noncalcareous shale with minor thin beds of friable, whitish greenish-grey, quartzose sands and minor sandy to silty, dark brownish-grey dolomite. There is some primary and secondary gypsum in the lower part. As originally defined by Norris (1965), the uppermost 10 m of the formation at the type locality northwest of the map area contain Middle or Upper Ordovician fossils. No other diagnostic fossils have been recovered from the lower part. However, outside the map area to the north-northwest of Hottah Lake, fossils tentatively assigned to the Middle Cambrian occur in strata approximately correlative with the middle part of La Martre Falls (Balkwill, 1971). Williams (1974) has suggested the upper La Martre Falls is correlative with the lower part of the Upper Ordovician Mount Kindle Formation of the Franklin Mountains (equivalent to the Chedabucto Lake Formation – see below) and that the unconformity underlying the Mount Kindle, that is clearly present in the west and north is also present within La Martre Falls Formation as originally defined (Fig. 85). Williams (1974) also suggested that the name La Martre Falls be abandoned in favour of earlier terminology for units defined to the west (La Martre Falls – Mt. Cap, Saline River and Franklin Mountain formations). On the other hand, Meijer-Drees (1975) recommended the retention of La Martre Falls Formation, with Williams revised definition, in the vicinity of Great Slave Lake, as it is difficult to separate the Mount Cap, Saline River and Franklin Mountain formation equivalents in this area.

Chedabucto Lake Formation

The Chedabucto Lake Formation is a dolomite that overlies La Martre Falls shales and is unconformably overlain by the Chinchaga evaporites and carbonates in the central and northwestern part of the Paleozoic terrane. The unit has a variable thickness ranging from a maximum of about 80 m in the northwest to where it thins and disappears in the vicinity of Alexander Point. The formation is a relatively resistant scarp-forming unit with many exposures, although no complete sections through the formation are known. It consists of thick-bedded to massive, fine grained, granular, in places minutely vuggy, medium brown dolomite that weathers pale orange to orange brown. Purplish-red mottling of the dolomite is prevalent. Chert, in the form of blebs and nodules to very fine anastomosing veinlets, is present in some beds in the lower part of the formation. Fossils collected from the formation northwest of the map area indicate an Upper Ordovician (Richmondian) age. Within the map area fossils are scarce and not very diagnostic. Williams (1974) correlated the Chedabucto Lake Formation with the lower part of the Mount Kindle Formation that occurs to the west in the Franklin Mountains, suggesting that the term Chedabucto Lake was unnecessary and should be dropped. Meijer-Drees (1975) however, considered that the term was still useful in the Great Slave Lake area. Williams (1974) suggested that the Ordovician unconformity underlying the Mount Kindle Formation in the west extended as far east as Great Slave Lake. Thus the Chedabucto Lake Formation is contained both above and below and is truncated to the south by unconformities.

Mirage Point Formation

The Mirage Point Formation is a dolomite that occurs in the southeastern part of the Paleozoic terrane south of Alexander Point. It overlies the basal Old Fort Island Formation, where present, and is overlain by the Chinchaga evaporites and carbonates. Only the upper part of the formation is exposed in the map area, with the best exposure being a 10 m section at Baker Bay. In its outcrop area the formation is estimated to be about 60 m thick, although where known in well cores to the south it is up to 200 m thick. Where exposed the formation consists of thinly interbedded dark dusky red, purple and orange-red, in some cases green mottled dolomite, argillaceous dolomite, variable sandy dolomite, gypsiferous dolomite, dolomitic mudstone, very fine grained dolomitic sandstone, green and red shale, gypsiferous shale, gypsum and satin spar. Mud cracks occur on some surfaces. The formation is not known to contain any fossils. The Mirage Point was originally interpreted as the lateral stratigraphic equivalent of La Martre Falls and Chedabucto Lake formations (Norris, 1965; Richmond, 1965). Williams (1974) has suggested the Mirage Point may be much younger. He suggested that where the pre-Middle Devonian unconformity cut through the resistant Mount Kindle Formation (Chedabucto Lake Formation) an escarpment was formed in the recessive La Martre Falls shale that was held up by the more resistant Chedabucto Lake Formation. The carbonates, red beds and evaporites that make up the Mirage Point Formation filled the erosional depression in Lower Devonian time and were succeeded by Lower to Middle Devonian units.

Fitzgerald Formation

The Fitzgerald Formation is a dolomite unit that was recognized in the area by Richmond (1965). It conformably overlies the Devonian Mirage Point Formation and probably unconformably overlies the Chedabucto Lake Formation where present. As described by Richmond (1965) it consists of a thick bedded, medium to light brown, calcareous, conglomeratic dolomite. It contains pebble size coral fragments and spindle-shaped clasts in a slightly silty, fine-grained matrix. The rock weathers massive to rubbly and on a flat bedded surface has a mottled patchy appearance. He reported that an 'organic hash' unit occurs at the top of the formation both west and south of Great Slave Lake. The formation is well developed in the vicinity of Chedabucto and Bras D'Or lakes. The Fitzgerald Formation is almost certainly correlative with the Ernestina Lake Formation of the Elk Point basin of Alberta and is Lower Devonian (G.K. Williams, personal communication, 1981).

Chinchaga Formation

The Chinchaga Formation, as the term is used in the area, is dominantly an evaporitic unit with some carbonates. Within the map area it conformably overlies the Fitzgerald Formation and is overlain by the Lonely Bay Formation carbonates. Because of its evaporitic composition, the formation is very poorly exposed and contact relations with adjacent units are not seen. The thickness of the formation is between 90 and 110 m. Gypsum, which forms most of the unit, occurs as recessive, white to banded, light to dark grey beds that are commonly contorted and brecciated. More resistant carbonate units, in the form of massive to thin-bedded, brown to pale brown, gypsiferous limestone to dolomite, occur at and near the base, in the middle, and in upper parts of the formation. These units are not laterally continuous and are commonly brecciated. As might be expected by its evaporitic nature, the formation is only sparsely fossiliferous. The formation is lithologically similar to and homotaxial with the Bear Rock Formation south of Great Bear Lake (Balkwill, 1971) and is considered to be Lower to Middle Devonian (G.K. Williams, personal communication, 1981).

Lonely Bay Formation

The Lonely Bay Formation is a limestone that conformably overlies the Chinchaga evaporites and is overlain by the Horn River shales and limestones. It is best exposed along its basal northeast margin where it forms a resistant scarp against the very recessive Chinchaga evaporites. The contact with the Horn River shales is not exposed and the area underlain by the Lonely Bay formation is defined on the basis of its somewhat higher topographic expression. The estimated thickness of the formation ranges from 40 m in the southeast to 90 m to the northwest beyond the map area. The lower part of the formation consists of limestones ranging in character from massive, dark brown, aphanitic and in part stylolitic; to thinly bedded, light grey, orange-brown weathering and fine grained to aphanitic; to irregularly thinly bedded, light olive grey to medium grey and fine grained; to medium bedded, grey, aphanitic and dolomitic; to thin bedded, pale brown and slightly argillaceous rocks. The upper part of the formation, where exposed 20 km west of the map area, consists of massive, dark to medium brown, fine grained, fetid limestone and interbedded nodular and irregular thin bedded, medium brown, fine grained to aphanitic limestone. The Lonely Bay Formation contains abundant Middle Devonian fossils. It is correlative with the Keg River platform of Alberta and the upper Hume Formation of Norman Wells, N.W.T. (Law, 1971). The unit is considered to be Middle Devonian, possibly Lower Givetian (Norris, 1965).

Horn River Formation

The Horn River Formation southwest of the map area consists mainly of bituminous shales. No outcrops of the formation have been recognized within the area. Norris (1965) on the basis of the recognition of clasts of shale and argillaceous limestone from this unit on the shore of Lonely Bay, 25 km to the south of the area, extrapolated the lower contact through the southwest corner of the area. Richmond (1965), by projecting the lower contact east from exposures 100 km to the west of the area, concluded the lower contact is about 5 km southwest of the area. The Horn River Formation is time transgressive but in this vicinity is Frasnian in age (Fuller and Pollock, 1972).

QUATERNARY GEOLOGY

The area is everywhere glaciated although thick glacial deposits are of limited extent, with the exception of the northeast corner of the area and off the Shield southwest of Great Slave Lake. Glacial striae indicate a consistent ice flow direction towards 240°. This corresponds to the general direction of flow interpreted by Craig (1965) to the southwest of the area in a down ice direction. There, Craig reported a second ice movement direction based on striae trending northwest. Craig felt there was insufficient evidence to determine whether these different striae trends could be interpreted as representing two distinct glaciations or merely changes in the direction of flow near the margin of the last ice sheet. No striae corresponding to the northwesterly set were observed in the area although striae were not systematically measured.

Most of the area is exposed bedrock and the Quaternary deposits that are preserved are mainly thin sands, silts and clay. Thicker deposits of glacial drift are more abundant on the plateau above 400 m that forms most of the southern Slave Province. There is comparatively little glacial cover on the slope between the plateau and the boundary of the Shield to the southwest. Only in the north-central part of the area, which reaches the level of this plateau, is there any significant accumulation of glacial and postglacial deposits. The terrain southwest of the Shield area underlain by flat lying Paleozoic rocks has a generally thicker mantle of glacial and postglacial deposits. Within the map area this terrain rises gently from the level of Great Slave Lake towards the southwest.

In the east there is a series of discontinuous deposits of sand and gravel, the largest of which is up to 10 km long and up to one km wide (Fig. 86). They form a discontinuous branching system that is more or less parallel to the glacial

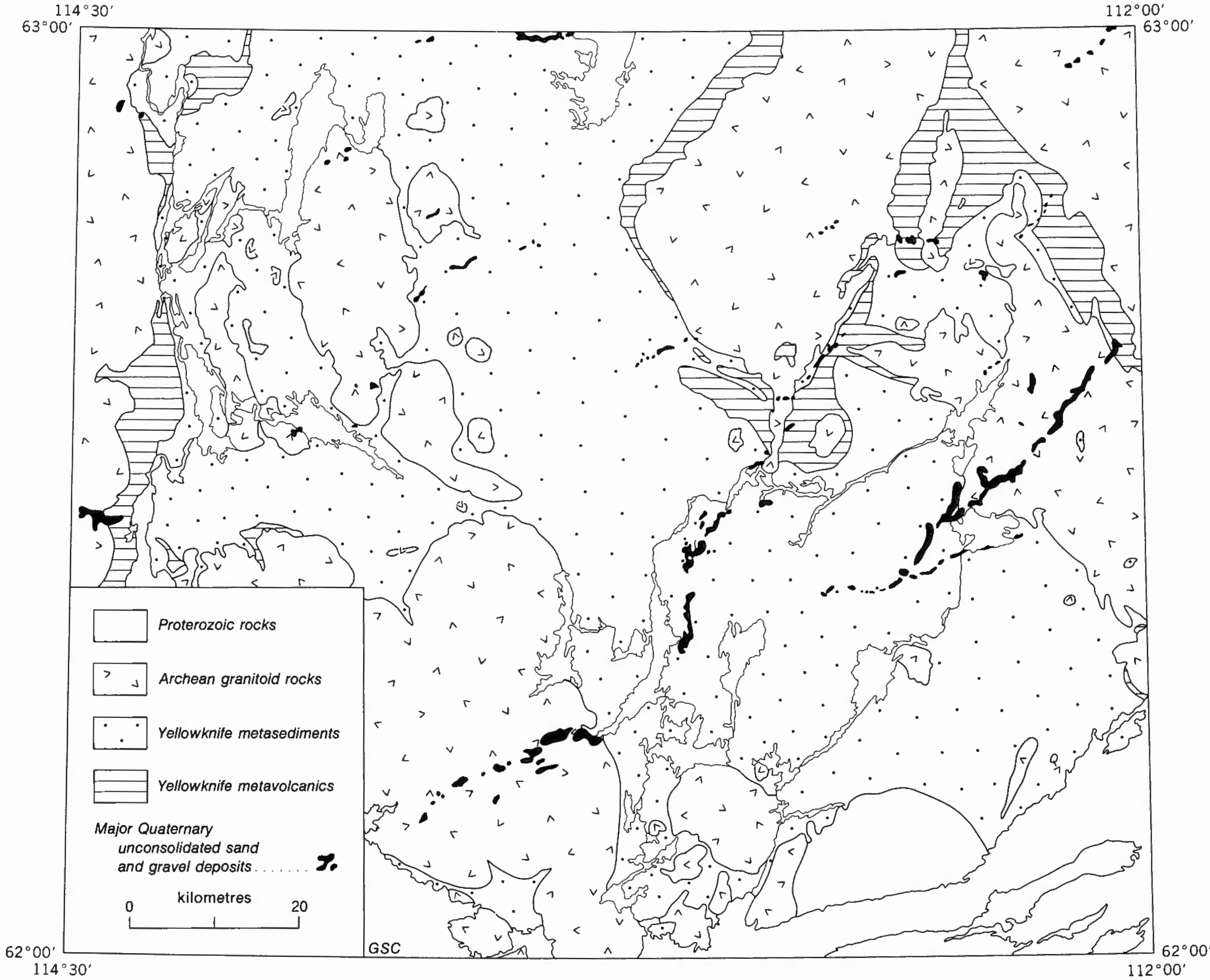


Figure 86. *Distribution of major Quaternary unconsolidated sand deposits. The deposits are mainly eskers with locally developed esker fan or delta complexes, and represent the main drainage pattern during the deglaciation of the area. Southwest of the granitic terrane in the northeast part of the area the esker deposits have been largely reworked by glacial Lake McConnell, and raised beaches are locally developed on them (Fig. 87).*

flow direction and to the present topographic gradient, and are interpreted as parts of an esker system with two main tributaries that join in the east-central part of the area. The northernmost branch extends from the northeast corner of the area through Turnback, Tumpline and Consolation lakes to Hearne Lake, more or less following the course of the present Beaulieu River. The southern branch can be followed through Meander Lake, between Desperation and Francois lakes, where it is joined by a second order tributary near Hearne Lake. The esker continues to the southwest through Hearne Lake to Great Slave Lake. Another smaller, less well preserved esker can be followed as a series of discontinuous sand ridges from Blaisdell Lake, southwest of Gordon Lake, through Bliss Lake and Prelude Lake to the mouth of the Yellowknife River. The sand and gravel deposits at Yellowknife may be related to this system. In the northeast part of the area the sands form sharp crested ridges that can be seen northeast of Sunset Lake and Meander Lake. However, throughout most of the area the ridges are less well developed or preserved towards the southwest, where they tend to form low flat hills. At several localities, notably west of Meander Lake, north of Francois Lake, and between Hearne and Woyna lakes, the sand complexes are elongated in a north-south direction and have an impressive series of narrow terraces developed on them. These may represent esker fan or delta complexes with superimposed beach ridges (Fig. 87). Southwest of Hearne Lake the sand deposits have been completely reworked into low terraced hills, possibly by the action of the glacial lake that formed with the decay of the ice. The fact that the sand deposits in the southwest part of the area tend to form low flat hills suggests that the esker formed at the edge of the ice where the sub-ice river suddenly lost velocity on emerging from the confined channel under the ice (Bannerjee and McDonald, 1975). Bannerjee and McDonald suggest the steep-sided esker ridges, seen in the northeast part of the complex, may be due to the sand being deposited under the ice, the steep ridge being due to the confinement of the deposit by the ice walls of the channel.

Glacial Lake McConnell (Craig, 1965) formed at the Shield margin between Great Bear Lake and Lake Athabaska on the retreat of the ice from the area about 10,000 years

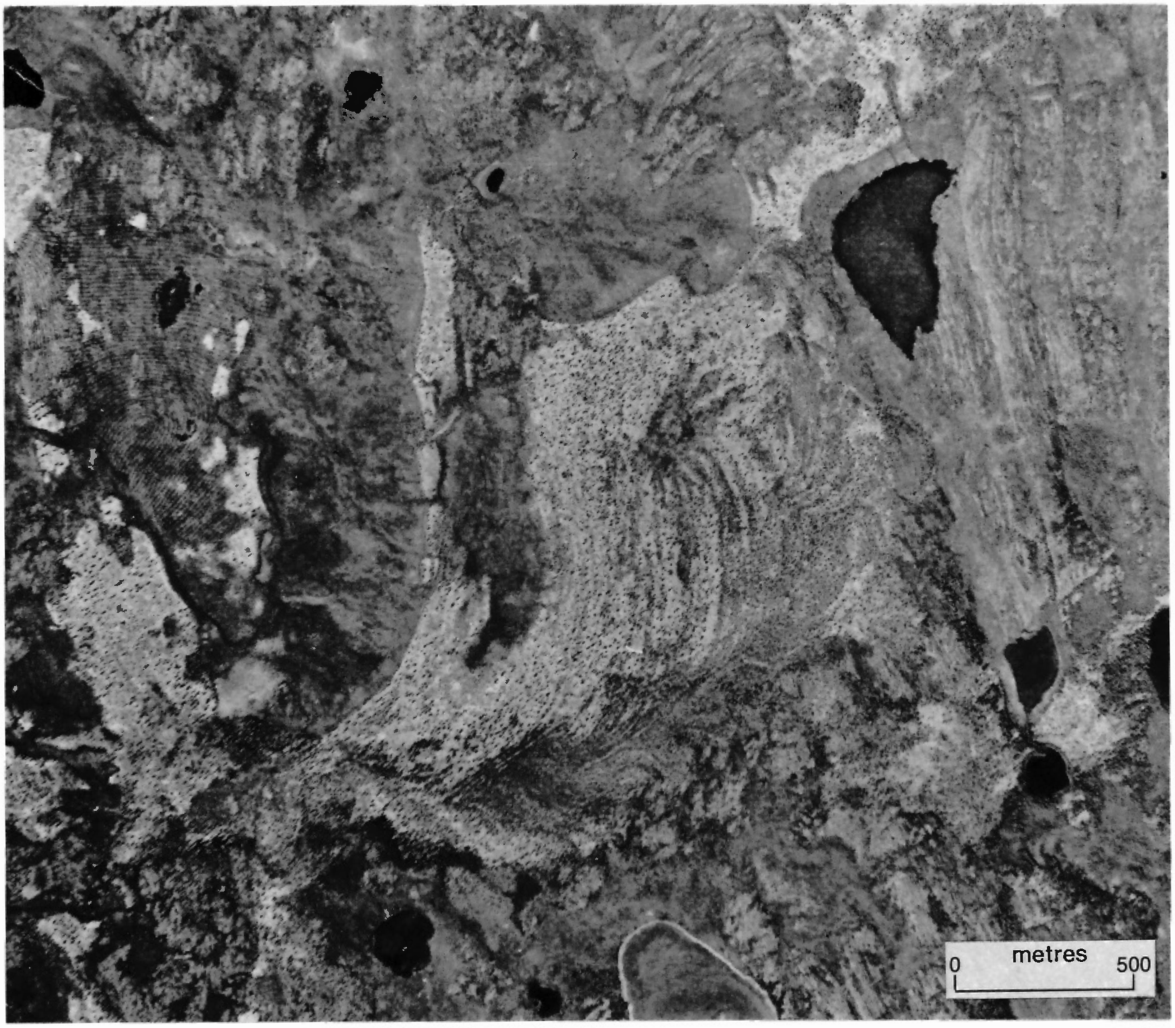


Figure 87. *Reworked Quaternary esker fan or delta deposit 8 km northeast of Hearne Lake. Prominent ribs are a series of raised beaches. NAPL A8566-123*

ago (Prest, 1969). The maximum level of the lake was at about 280 m above present sea level along the Shield margin. If the reworking of the esker sands to form beaches was due to glacial Lake McConnell then the lake may have extended onto the Shield as far as Meander Lake at a present-day elevation of about 335 m.

GEOPHYSICS

Aeromagnetic pattern

Many of the features on the aeromagnetic map of the area (Geological Survey of Canada, 1969 a,b) can be related to geological features exposed at the surface. In some cases their magnetic characteristics are particularly useful for tracing the extent of such units. The magnetic intensity in the area ranges between a minimum of 59 800 nanoteslas (gammas) and a maximum of 67 000 nanoteslas, although over most of the terrain it is between 60 200 and 60 900 nanoteslas; a range of 700 nanoteslas. The extreme high values are over magnetite iron formation or other magnetic rock units, while the lowest values in the southeast part of Yellowknife Bay occur over some of the quartz dioritic intrusions in that area.

The most prominent linear magnetic features are due to three diabase dyke sets. Only the widest of the east-northeasterly trending Dogrib dykes (McGlynn and Irving, 1975; this report) have a moderate magnetic response. These include the single thick dyke through Russell and Stagg lakes and the pair of thick dykes between Yellowknife and Gordon Lake. Other narrower Dogrib dykes that occur in the region have no aeromagnetic expression. The Hearne dykes which have a similar east-northeasterly trend have a much more pronounced aeromagnetic expression. They occur in a zone 12 km wide immediately north of Hearne Channel. The north-northeast-trending Mackenzie dykes that occur only in the eastern half of the area have a strong magnetic expression. The Indin dykes in the area have no aeromagnetic expression.

In general there is little in the way of contrast in magnetic response between the Defeat Plutonic Suite and the Yellowknife supracrustal rocks they intrude, except in one case between Jennejohn and Harding lakes (Fig. 88). There, a prominent circular anomaly and adjacent elliptical anomaly to the southwest, both with steep magnetic gradients, correspond closely to plutonic lobes in the Defeat plutonic complex. Magnetite is more abundant toward the margin of these plutonic lobes, which in other respects are similar to

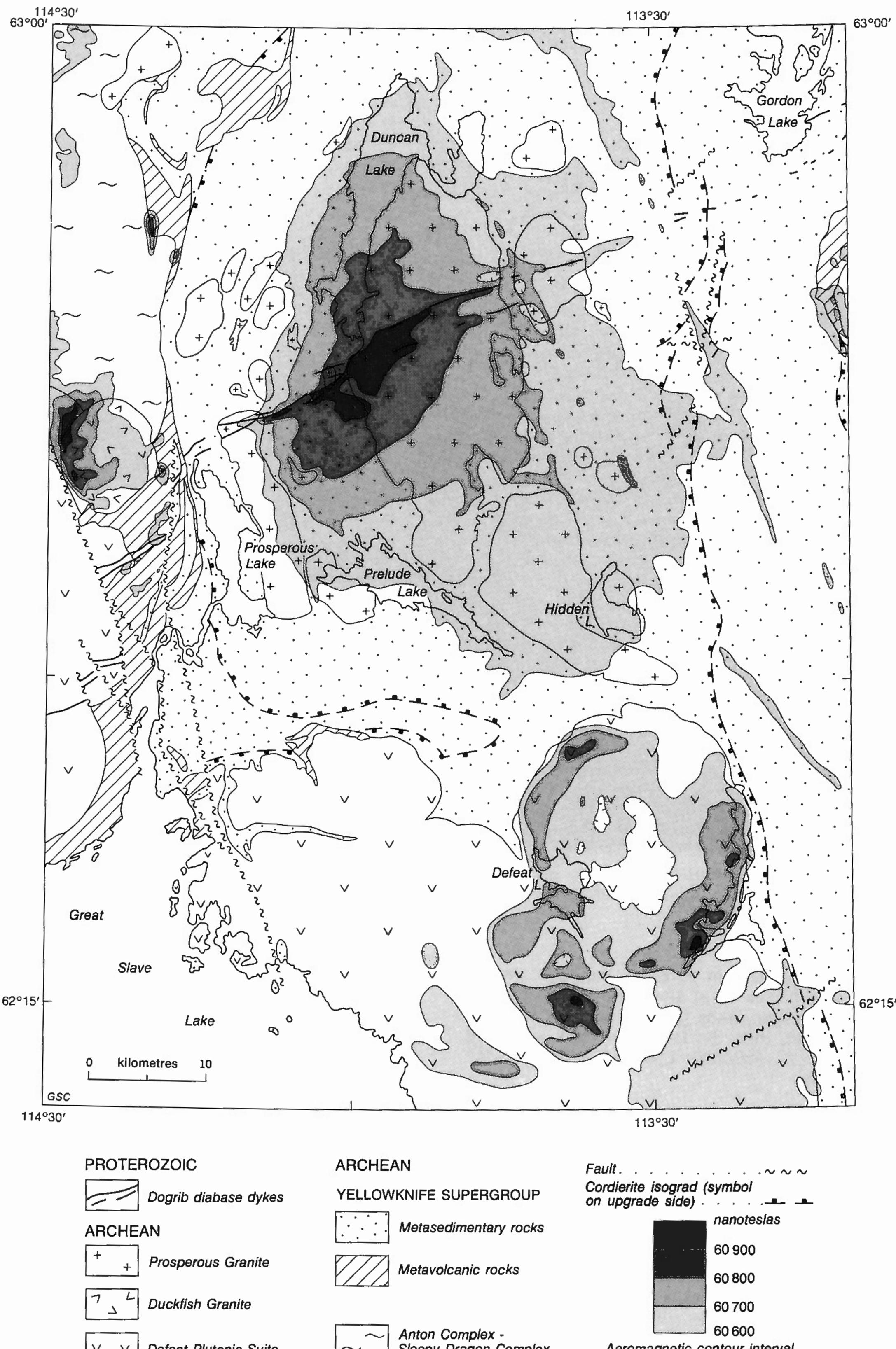


Figure 88. *Contrasting magnetic anomalies east of Yellowknife Bay. The northern broad anomaly in the vicinity of the Prosperous granites, and presumably related to them, reflects a relatively deep source as at the surface there is no magnetic contrast between individual granite plutons and the metasediments they intrude. In contrast the prominant circular anomaly, that closely corresponds to the margin of the large pluton of the Defeat Plutonic Suite, represents a surface or near surface source related to the marginal phases of the pluton.*

the rest of the complex (E. Lisle, personal communication, 1977). On the other hand, the broad anomaly in the vicinity of the Prosperous granites between Prelude and Duncan lakes does not correspond particularly well at the present erosion level to the outcrop of the plutons or the pattern of metamorphism, at least as expressed by the cordierite isograd (Fig. 88). This, together with the broad shape of the anomaly and the gentle magnetic gradients, may suggest a relatively deep source for the magnetism that is presumably related to the same thermal event that resulted in the emplacement of the Prosperous granites and the metamorphic pattern at the present erosion level. An even more diffuse and lower amplitude magnetic dome as in the vicinity of Buckham Lake similarly may be related to a relatively deep source, that is also related to the development of the broad thermal dome in the area.

The Duckfish Granite, 25 km north of Yellowknife, has a distinctive magnetic response that contrasts strongly with the volcanic and other granitic rocks it intrudes. The magnetite content of certain members of the Blachford Lake Intrusive Suite, such as the Caribou Lake Gabbro (Davidson, 1978a) and possibly the Thor Lake Syenite, is reflected in the prominent positive anomalies between Blachford Lake and Hearne Channel. The most sharply defined and intense anomalies are related to magnetite iron formations, the largest of which occurs near Amacher Lake. Other smaller anomalies due to iron formation occur 45 km north of Yellowknife near Bell Lake.

The granitic terrane northwest of Yellowknife has a patchy, irregular magnetic signature that roughly corresponds to the trend of the granitic units within it. The anomalies are offset by the series of north-northwesterly trending, dominantly sinistral, transcurrent faults that are particularly abundant in this area. An abrupt drop in the magnetic intensity to the northwest of the granitic terrane corresponds to the contact between the granitic rocks and the Yellowknife supracrustal rocks. This magnetic break extends north well beyond the border of the map area. The Anton Complex gneissic terrane north of Yellowknife that may, at least in part, be older than the supracrustal rocks, has a generally higher and more irregular magnetic topography. The anomalies become larger in size and intensity to the north-northeast beyond the map area (Geological Survey of Canada, 1969c) suggesting this gneissic terrane may continue some distance to the north. Although not as strongly developed, the magnetic signature of the Sleepy Dragon Complex between the Cameron and Beaulieu rivers, also thought to be older than the Yellowknife rocks, has a somewhat higher, more irregular response. This higher, irregular magnetic topography is noted on other possible basement blocks in the Slave Province, such as west of Healey Lake (Henderson and Thompson, 1982).

Other anomalies that are not satisfactorily explained include the large strong positive anomaly that is mainly underwater in the bay north of Mattonabbee Point on Great Slave Lake. Islands in the bay are of mafic-rich granitic rocks quite distinct from the two-mica granite immediately to the south. The horseshoe-shaped anomaly in the Paleozoic terrain southwest of Waite Island corresponds to a series of inliers of granitic rocks. The anomaly may represent a plutonic lobe of the Stagg Plutonic Suite similar to that in the Defeat granodiorite at Defeat Lake. Four kilometers from the northwest corner of the map area is a very intense, small, circular anomaly of unknown origin. It has a similar shape and intensity, although is considerably smaller in area compared to the anomaly over the Big Spruce alkalic complex 65 km to the north (Lord, 1942; Irving and McGlynn, 1976).

Gravity pattern

The Bouguer anomaly gravity map of the region that includes the Yellowknife-Hearne Lake area (Earth Physics Branch, 1969) shows several large scale regional features within the area of interest. North of Hearne Channel a positive gravity gradient that increases to the southeast is part of the regional gravity high that underlies the East Arm of Great Slave Lake. Hoffman et al. (1977) suggested this anomaly may be due to extensive mafic or perhaps ultramafic intrusions at depth, of which the large funnel-shaped intrusions that reach the surface in the East Arm may be part. There is also a regional positive gravity gradient that increases from the Slave Province to the Bear Province and which occurs over the western part of the area. The gravity pattern over the area underlain by Paleozoic rocks is in general significantly more positive than that over the adjacent Archean terranes.

Since the gravity stations are spaced at about 10 km intervals, many features on the geological map are not reflected on the gravity map. Thus it is possible for a major mafic volcanic belt to be hidden if few or no gravity stations are situated on it. This appears to be the case for the volcanics in the Cameron-Beaulieu area. However at Yellowknife where the stations are more closely spaced, the volcanic rocks are expressed by a positive anomaly.

Gibb and Thomas (1980), on the basis of a much more detailed gravity survey and three-dimensional gravity modeling south and southeast of Yellowknife Bay, have suggested that the Kam mafic volcanics extend south under Great Slave Lake and then swing abruptly southeast through the Mirage Islands and continue for another 60 km. According to their model the southeasterly continuation of the Kam volcanics is up to 15 km wide and up to 3 km thick, which would make it the largest volcanic belt in the Slave Province. Between the volcanics offshore and the Defeat granodiorite onshore to the northeast are sediments.

Using the regional gravity data available an attempt has been made to model the geological situation along a profile normal to the margin of the supracrustal basin at Yellowknife, in order to determine whether or not a significant layer of mafic volcanic rock occurs in the basin below the metasediments (McGrath and Henderson, 1983). Using average densities of units, where exposed at the surface, the model calculated closely approximates the observed gravity signature and suggests there is little evidence for the existence of such a layer (Fig. 34). This supports the idea, expressed elsewhere in this report, that the volcanic sequences occur mainly at the margins of the basins. However, it should be kept in mind that the gravity model is a model of the present day situation, and does not take into account the possibility that such a layer could have existed as part of the Yellowknife Supergroup, but has since disappeared during the course of subsequent deformation and intrusion.

The areas underlain by gneissic rocks east and west of Gordon Lake, that are thought to be at least in part older than the supracrustal rocks, tend to have the most negative gravity expression in the area.

Airborne gamma-ray spectrometry pattern

Uranium, thorium and potassium gamma radiation over the Yellowknife-Hearne Lake area have been measured in an aerial survey flown at an average 2.5 km line spacing (Grasty and Richardson, 1972). Certain features on the resulting map can be related to defined geological units, while others reflect variations of the various radioactive elements within a given unit.

The radiation measured from all the elements is generally higher in the granitic rocks west of Yellowknife. This may in part be due to generally better and larger outcrop exposures, as well as to the composition of the rock. West of the Stagg River fault the radiation due to uranium is significantly higher than in the rocks east of the fault. Within the Awry granite there is a broad ridge-like anomaly in all three elements that extends from Rae Point to Awry Lake. This trend parallels the trend of the granitic units in the area. The prominent anomaly in all three elements over the Duckfish Granite may be due to its high potassium composition and the abundance of zircon and sphene as accessory minerals. The two-mica granites of the Prosperous type typically have positive potassium and uranium anomalies. The uranium pattern has local peaks over some of these granitic bodies. An increase in U/Th ratio with uranium is evident in several plutons of the Prosperous Granite, in particular, south of Prelude Lake, south of Hidden Lake, at Mattonabbee Point, Buckham Lake and near Consolation Lake. Charbonneau (1982) in a radiometric study of several granitic bodies elsewhere on the Shield has suggested that an increase in U/Th ratio with U but not Th may be indicative of postmagmatic redistribution of uranium, and so may be an economic criterion. Thorium has no expression over these granites except for the pluton south of Wedge Lake. The Defeat granodiorites do not have a significant expression and are indistinguishable from the adjacent sediments. The mafic volcanic belts in general have a negative expression. The granitic rocks along the eastern margin or the area have a moderate, rather patchy expression that may in part be due to the poorer exposure in that area. The gneissic units north of Yellowknife and between Cameron and Beaulieu rivers have an irregular patchy expression that may reflect the variable composition of the unit. The Blachford Lake Intrusive Suite has a moderately positive expression with local peaks, at least one of which has been shown to be of some economic interest (Davidson 1978a; this report).

The significance of the radioactivity pattern over the west half of the Hearne Lake area (NTS 85 I) has been discussed by Newton and Slaney (1978).

ECONOMIC GEOLOGY

The Yellowknife-Hearne Lake region has been the most productive in the Northwest Territories due almost entirely to the production of the Con and Giant Yellowknife gold mines at Yellowknife. Other commodities that have attracted interest, although to date are largely uneconomic, include lithium, beryllium, columbium and tantalum in pegmatites, tungsten in scheelite-bearing quartz veins, base metals and nickel (Fig. 89).

There is a large literature on the economic geology of the area and what follows is only a very brief summary of the main deposit types. Detailed descriptions of the individual deposits are available in various government reports. Deposits discovered prior to 1950 are summarized by Lord (1941, 1951). McGlynn (1971) has described properties active during the fifties and Baragar (1961, 1962), Baragar and Hornbrooke (1963), Schiller and Hornbrook (1964), Schiller (1965), and Thorpe (1966, 1972) properties active in the sixties. Properties active during the seventies have been described in the Department of Indian and Northern Affairs Mineral Industry Reports (Padgham et al., 1975; Gibbins et al., 1977; LaPorte et al., 1978; Lord et al., 1978, 1981; Padgham et al., 1978): a continuing series of reports.

Gold

Gold was first discovered in the Yellowknife area in 1898 by E.A. Blakenay on claims within 16 km of the mouth of the Yellowknife River, but it was not until 1934 that serious gold exploration began with the discovery of a high grade gold-bearing quartz vein in the low grade metasediments of the Burwash Formation on the east side of Yellowknife Bay, opposite the present site of Yellowknife (Lord, 1951). The following year gold was discovered in the volcanic Kam Formation on the west side of the bay, which lead to the establishment of the Con mine in 1938 and adjacent Negus mine in 1939. During this period intensive prospecting resulted in the discovery of many of the gold properties in the region (Table 23) of which the Ptarmigan, 10 km northeast of Yellowknife, and Thompson-Lundmark at Thompson Lake 45 km northeast of Yellowknife, developed into small, rather short-lived but profitable mines. The largest gold mine in the Northwest Territories, the Giant Yellowknife mine was brought into production in 1948. Both the Con and Giant Yellowknife mines have remained in continuous production since their start, except for a short period at the end of World War II for the Con mine.

Gold in shear zones in mafic volcanic rocks

The major producing gold mines in the area are in parts of a large complex shear zone system in the Kam mafic volcanics described in detail by Boyle (1961, 1979), Henderson and Brown (1966), Brown and Dadson (1953) and Brown et al. (1959), and summarized in this report in the section on the Kam Formation. The geochemistry of the deposits has been discussed by Boyle (1961). The gold deposits occur in large discordant zones, commonly as narrow discontinuous bodies localized near the ends of unsheared blocks within the shear zones where deformation has been particularly intense and complex and dilatent zones were available for quartz vein formation. Concordant shear zones, commonly developed in interflow tuff beds, are generally barren. Gold occurs both in chlorite-sericite-carbonate-sulphide schists, where it is associated with quartz, carbonates, pyrite, arsenopyrite, chalcopyrite, stibnite, sulphosalts, sphalerite and galena in decreasing order of abundance, and in quartz-carbonate-filled fractures, where it is associated with quartz, carbonate and small amounts of pyrite, arsenopyrite, chalcopyrite and tetrahedrite (Boyle, 1961).

Most authors have concluded the shear zones formed due to faulting associated with the folding of the volcanic sequence and intrusion of the Defeat granodiorite. Henderson (1978a, and this report), on the other hand, suggested the initial movement on what was to become the shear zones took place in the form of large scale gravity slides while volcanism and sedimentation of the Yellowknife Supergroup was still taking place. The original slide surfaces could have provided a conduit for meteoric waters downward and, as volcanism was still active, for fumarolic materials upward, either of which could have facilitated early alteration of the adjacent volcanics and provided a zone of weakness along which subsequent movement during deformation, intrusion of granitoid plutons and metamorphism of the volcanic sequence could take place. Boyle (1961) suggested the gold was derived from the altered volcanic rocks in the shear zones. Carbon dioxide, water, sulphur and some other elements were liberated from the mafic volcanic sequence during metamorphism. These were concentrated along shear zones that formed towards the end of the deformation of the supracrustal rocks, resulting in their alteration and the liberation of silica, potassium, calcium and iron among other elements, but in particular gold. The gold along with quartz, carbonate and a variety of sulphides were subsequently precipitated as vein fillings in dilatent zones in the shears. Kerrich et al. (1977), noted the reduced state of iron in the shear zones, and suggested that if the system is buffered by the rock, a large water/rock ratio (> 3), and a large temperature gradient is required in order for sufficient reduction of iron to take place. They suggested this requires

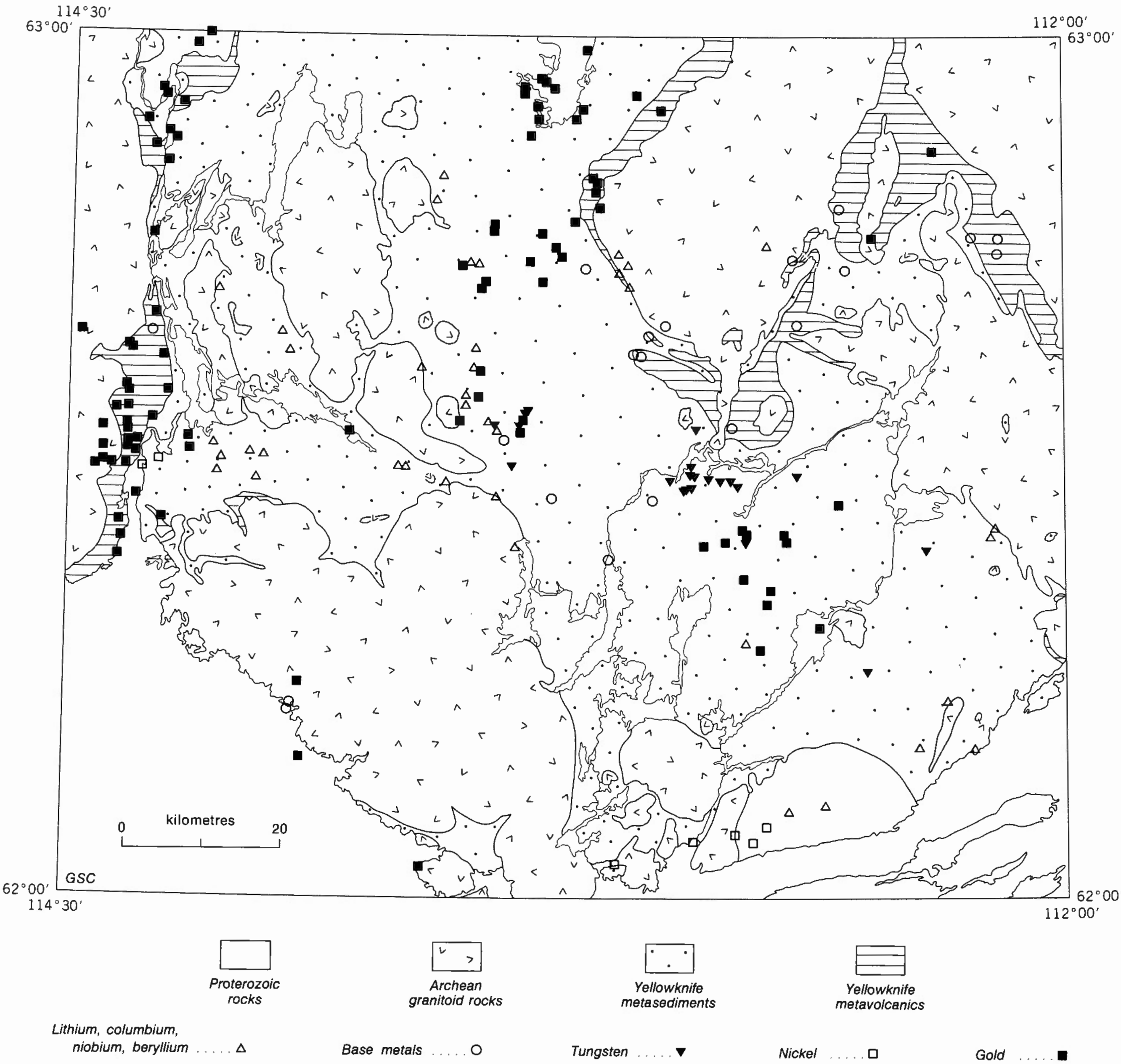


Figure 89. *Distribution of economic mineral prospects in the central and eastern part of the map area. Gold occurs mainly in shear zones and quartz veins in the mafic volcanic sequences and in quartz veins in the metasediments, particularly where there is a change in direction of major regional bedding trends. Similarly, tungsten, as scheelite in quartz veins, occurs in the metasediments in two main areas in the central part of the area. Lithium, columbium, niobium and beryllium minerals occur in pegmatites mainly associated with Prosperous and Redout granites, although other important pegmatites occur in the southeast part of the area and have not as yet been related to any granitic body. Base metal prospects occur in the Yellowknife volcanics, particularly where felsic volcanics and carbonate units are present. A linear array of nickel prospects are found in the southern part of the area parallel to Hearne Channel.*

a hydrothermal system, with the water and gold derived from the dehydration of the volcanic rocks during metamorphism (Kerrich, 1981), driven by energy from the adjacent granitoid batholith, moving along highly permeable conduits opened in the shear zones by hydraulic fracturing (Allison and Kerrich, 1981).

Gold in quartz veins

Gold also occurs in quartz veins at many localities in the Burwash Formation metagreywackes and mudstones (Fig. 89). Most are small, high grade prospects that have produced at most less than 16 kg (500 oz.) of gold (Table 23) although some, such as the Ptarmigan mine 10 km northeast of Yellowknife, the Thompson-Lundmark mine 45 km northeast of Yellowknife, and Camlaren at Gordon Lake, have

Table 23. Gold production in the Yellowknife-Hearne Lake area[1]

Property	Location	Type	Production period	Tonnes (Tons) milled	Kilograms (Ounces) produced	Reference
Giant Yellowknife (Lolar and Supercrest)	62°30'N 114°22'W	Shear zones in Kam mafic volcanics	1948-1981[2]	10 132 200 (11 132 200)	175 875 (5 654 500)	Franklin and Thorpe (1982), CMH[3]
Con and Rycon	62°26'30"N 114°21'30"W	Shear zones and quartz veins in shear zones in Kam mafic volcanics	1938-1981[2]	5 946 200 (6 554 500)	101 338 (3 258 100)	Franklin and Thorpe (1982), CMH[3]
Negus	62°25'45"N 114°20'44"W	Quartz veins in shear zones in Kam mafic volcanics	1939-1952	445 352 (490 908)	7955.60 (255 807)	Lord (1951)
Thompson-Lundmark	62°36'45"N 113°28'15"W	Quartz veins in Burwash metagreywacke	1941-1949	121 537 (133 969)	2187.54 (70 339)	Lord (1951)
Camlaren	62°59'05"N 113°12'15"W	Quartz veins in Burwash metagreywacke	1962-1963	11 044 (12 174)	431.82 (13 885)	Lord (1951)
Ptarmigan	62°31'10"N 114°11'46"W	Quartz veins in Burwash metagreywacke	1941-1942	31 234 (34 429)	370.74 (11 921)	Lord (1951)
Nose Group	62°54'50"N 114°14'00"W	Quartz veins in felsic volcanics	1967	1034 (1140)	16.02 (515)	Schiller (1965)
Pensive Yellowknife	62°43'N 113°19'W	Quartz veins in Burwash metasediments	1939 ?	?	10.89 (350)	Lord (1951)
Joon Group	62°26'N 112°52'W	Quartz veins in Burwash metasediments	1977-1978	?	9.35 (300)	Lord et al. (1981) Padgham (1981b)
DAF Group	62°54'21"N 113°13'58"W	Quartz veins in Burwash metagreywacke	1947-1948	279 (307)	8.43 (271)	Lord (1951)
Old Paar	62°44'N 113°31'W	Quartz veins in Burwash metagreywacke	1963-1965 1966-1968		7.25 (233)	Lord (1951) Baragar (1961)
Rich Group	62°28'01"N 114°19'16"W	Quartz veins in Burwash metagreywacke	1935	137 (151)	6.00 (193)	Lord (1951)
Ruth mine	62°27'45"N 112°34'15"W	Quartz veins in Burwash metagreywacke	1942	169 (186)	4.73 (152)	Lord (1951) McGlynn (1971)
Norma Group	62°24'46"N 112°54'22"W	Quartz veins in Burwash metagreywacke	1947-1948	419 (462)	1.49 (48)	Lord (1951)
Tin (Star)	62°32'37"N 114°10'25"W	Quartz veins in Burwash metagreywacke	1950	27 (30)	.72 (23)	Lord (1951)
Rod Group		Shear zone in granodiorite	1978-1979	11 (12)	.45-.60 (15-20)	Padgham (1981b)

[1]Modified after Padgham (1981b).
[2]Production continued after 1982.
[3]Canadian Mines Handbook, Northern Miner Press Ltd., Toronto.

produced in total almost 3000 kg (100 000 oz.). These deposits have been described by Lord (1951), Boyle (1961), Henderson and Jolliffe (1939), Coleman (1957), Wiwchar (1957), Wanless et al. (1960) and Ramsay (1973c). The quartz veins, ranging in size up to many tens of metres long and a metre or so wide, are discordant to bedding, parallel to bedding or occur in the broken axial zones of isoclinal folds. The veins contain, in addition to quartz, less than one or two per cent other minerals including pyrite, sphalerite, galena, chalcopyrite, arsenopyrite, tourmaline, feldspar, and carbonate. Veins with sulphides are more likely to contain gold. Most authors agree that the veins formed due to the formation of dilatant zones during the deformation of the country rocks, although it is possible some of the veins parallel to bedding may be related to chert layers interbedded with the sediments. Ramsay (1973c), in his study of several quartz veins south of Prosperous Lake, concluded they formed in dilatent zones after the relaxation of compression caused by the emplacement of the large granite plutons. This would imply more than one age of veining. On the basis of the regional distribution of the vein deposits (Fig. 89) they appear to be concentrated in areas where there is a change in large scale regional structural trends. Thus there is a concentration of deposits at Gordon Lake where the northwesterly trend west of the lake swings around to a northeasterly trend east of the lake. Similarly, north of Ross Lake there is a concentration of deposits where the northerly trend to the west changes to southwesterly through Victory Lake. A third concentration northeast of Campbell Lake marks the change from the northerly trend through Campbell Lake to a more easterly trend north of Buckham Lake.

Boyle (1961) has suggested the quartz and metal content of the veins is derived from the sediments with the mobilization of fluids during metamorphism, although from the distribution of the veins (Fig. 89) there is no apparent grade at which this process is favoured. Ramsay (1973c), on the basis of a fluid inclusion study of the quartz in some of the veins, concurred with Boyle's interpretation.

Gold in iron formation

In the Point Lake-Contwoyto Lake area, 350 km north-northeast of Yellowknife, there are important gold deposits associated with sulphide-bearing iron formations (Tremblay, 1976; Bostock, 1980). Although such iron formations do occur in the Yellowknife-Hearne Lake area they are rare. One example of this type occurs on some small islands in Russell Lake, 4 km south of the north boundary of the map area, where a sulphide-bearing sample of the silicate iron formation contained up to 235 ppb gold (Boyle, 1976). The Andrew Yellowknife Mines property a few kilometres north of Russell Lake, and outside the area, is a similar deposit (Lord, 1951). Northeast of Yellowknife, Gibbins et al. (1977) report iron-rich bands in the sediments of the WT claim group northwest of Upper Ross Lake. Baragar (1962) mentions actinolite-rich country rock associated with the gold quartz veins at Hidden Lake. These deposits may be of the Contwoyto Lake type.

Archean rare element pegmatites

Archean rare element pegmatites that occur in many localities east of Yellowknife have attracted attention due to the local occurrence of lithium minerals. Also present, although much less abundant, are beryllium, niobium, tantalum, and tourmaline. Between 1946 and 1948 some attempts were made to recover columbite-tantalite from three properties in the area, but according to Lord (1951) the combined total of concentrates was probably less than 5 tonnes (5 tons). Between 1955 and 1960, and again in 1975, the pegmatites of the region were intensely prospected for lithium minerals, although to date none have been mined for lithium.

The pegmatites of the region have been described in detail by Jolliffe (1944), Fortier (1947), Rowe (1952), Hutchinson (1955), Mulligan (1965) and Kretz (1968). The pegmatites are associated with the Archean Prosperous Granite plutons, the Redout Granite and apparently part of the Meander Lake Plutonic Suite southwest of Francois Lake. Similar pegmatites occur over a wide area between Campbell Lake and Hearne Channel to the southwest where the only granitic bodies exposed are the small Prosperous pluton at Buckham Lake and the elongate Defeat granodiorite pluton at Drever Lake, although elsewhere the Defeat intrusions have only minor, simple pegmatite associated with them. The occurrences of pegmatites in this area together with the broad zone of amphibolite grade metamorphism, suggests that more Prosperous-type granite bodies may exist below the present erosion level.

The pegmatites are varied in size and shape. Most are tabular and typically 20 to 100 m long and less than 2 m wide, although some are up to 1 km in length and 20 m wide. Some have a more bulbous shape. They are grey, coarse grained, and commonly internally zoned where rare element minerals are present. Almost all contain quartz, plagioclase, potassium feldspar and muscovite, with the most abundant rare element minerals being coarse green spodumene, amblygonite, columbite-tantalite and beryl. Kretz (1968) reported a total of 41 minerals that occur in the pegmatites. The pegmatites only occur in country rock (either Burwash metasediments or Sleepy Dragon meta- granitoid rocks) that are metamorphosed to amphibolite facies. In the vicinity of Upper Ross Lake and Sparrow Lake the pegmatites are radially zoned with respect to the plutons with which they are associated. Typically, although not universally, the rare element minerals are absent from pegmatites within or immediately adjacent to the granite. Beryl, followed by columbite-tantalite, and finally by lithium minerals occur with increasing distance from the granite, although there is some degree of overlap of the zones. No mineralogical zonation has been noted in the Buckham Lake region. Hutchinson (1955) suggested the pegmatites in the Upper Ross Lake area were derived from the associated Redout granite through a process of crystallization of a pegmatitic fluid and replacement of the country rock. Kretz (1968), in the Prosperous Lake – Duncan Lake region, suggested a common source for both the granite and the pegmatites, but felt some material in the pegmatites was also contributed from the country rock. Boyle (1961) considered it likely that the pegmatites were derived from their immediate sedimentary or granitic host rocks during metamorphism and immediately following crystallization of the granite.

To date the pegmatites have proved to be of little economic value. As previously mentioned, approximately 5 tonnes (5 tons) of columbite-tantalite concentrates were recovered between 1946 and 1948. Lasmanis (1978) estimated the lithium potential for the region at about 44 million tonnes (49 000 000 tons) of pegmatite containing 1.40% LiO_2.

Early Proterozoic rare element deposits

An important tantalum-columbium deposit occurs in the Thor Lake Syenite of the early Proterozoic Blachford Lake Intrusive Suite, and has been described in some detail by Davidson (1978a). The main zone consists of a roughly triangular area, with sides approximately 2 km long, below and south of 'Thor Lake', the large lake in the western part of the syenite body. The zone consists of dark, extensively altered syenite of varied grain size, composed mainly of albite and iron oxides but also containing rare element minerals. To the north of the main zone are smaller zones of black rock in sharp contact with slightly altered syenite with pink to buff fluorite veins and masses. All zones have above background radiation counts with local patches that are particularly high due to concentrations of uranium or thorium or both. A broad positive magnetic anomaly is centred over the alteration zones. Davidson (1978a) suggested the alteration zones formed under the influence of a late-stage magmatic vapour phase, probably related to the emplacement of veins of alkaline pegmatite and related albite-rich rocks that occur thoroughout the syenite, but are particularly abundant in a northwesterly trending zone east of 'Thor Lake'. Other albitic veins and pegmatitic veins with rare minerals also occur, but much less abundantly, in the surrounding Grace Lake granite and in one case across into Yellowknife Supergroup rocks. Highwood Resources Ltd. reports a 64 million tonne (70 million ton) drill-indicated and inferred rare element deposit whose grade is approximately 0.25 kg/tonne (0.6 lbs/ton) tantalum oxide, 3.3 kg/tonne (8.0 lbs/ton) niobium oxide, 0.8 kg/tonne (2 lbs/ton) samarium oxide, 8.2 kg/tonne (20 lbs/ton) cerium oxide, 4.9 kg/tonne (12 lbs/ton) lanthanum oxide and 29 kg/tonne (70 lbs/ton) zirconium oxide (Thomas, 1982).

Tungsten

Scheelite occurring at several localities in the region (Fig. 89) was first recognized in the Gilmour Lake area in 1941, and for a few years the region was closely prospected. Only a few tens of tonnes of material was ever mined from the more promising prospects. Most of the individual properties have been described by Lord (1951) and Little (1959).

Most scheelite deposits occur either in the Gilmour Lake area about 15 km north of Campbell Lake or near Tibbit Lake, 25 km to the west. The Gilmour Lake deposits occur in the greenschist grade Burwash metasediments, typically in quartz veins similar to the much more abundant gold-quartz deposits. Indeed, some of the gold-quartz prospects contain minor scheelite. In some cases the scheelite veins contain, in addition to quartz, actinolite, garnet, and epidote in considerable amounts. This may indicate the sedimentary units in which the deposits occur are more calcic than is normal for the Burwash Formation as a whole. In this regard Boyle (1961) has reported that scheelite occurs in 'small skarn lenses' (probably metaconcretions) in the Burwash Formation at either greenschist or amphibolite metamorphic grade.

The other important locality for scheelite is in the vicinity of Tibbit Lake where Lord (1951) reported the occurrence of about 150 scheelite-bearing quartz veins. As in the Gilmour Lake area, these deposits occur in greenschist grade metasediments, but in this locality extensive thick metagabbro sills intrude the sediments. The most important scheelite deposits are in the gabbro sills, where scheelite occurs in quartz veins, in pegmatitic quartz veins associated with coarse garnet, clinozoisite, feldspar and carbonate, and in altered fractures of shear zones where the scheelite-clinozoisite-feldspar-garnet-chlorite-quartz-carbonate assemblage is gradational with the gabbro.

Scheelite also occurs in some gold-bearing quartz lenses in the shear zones in the Con and Negus mines in amphibiotite grade mafic volcanics of the Kam Formation (Boyle, 1961; Little, 1959).

Base metals

Base metal prospects in the Yellowknife-Hearne Lake area are concentrated in the volcanic units in the north-central part of the area. A few minor veins with base metal sulphides also occur locally in the metasediments or more rarely in the granitic units. Most individual properties in the area are described by Lord (1951), McGlynn (1971), and Thorpe (1972).

The largest deposits (Lord, 1951· Baragar, 1961) occur on Turnback Lake (XL, XLX Groups). The zinc-copper-lead mineralization occurs as fine grained sphalerite and pyrrhotite with irregularly distributed coarse grained chalcopyrite, galena and pyrite through a distance of several kilometres in highly metamorphosed Yellowknife supracrustal rocks, notably an amphibolite gneiss (Padgham et al., 1975). Shelgelski and Thorpe (1972) suggested the mineralizaiton is confined to one stratigraphic horizon, and is associated with a metamorphosed limestone unit. As the most massive mineralization is immediately adjacent to, and cut by, pegmatite, they suggested that heat from the pegmatite mobilized and concentrated the originally much more disseminated stratiform mineralization into its present form. To the west, south of Victory Lake, the base metal mineralization (sphalerite, galena, pyrite, arsenopyrite, pyrrhotite and minor chalcopyrite) occurs in an anticlinal structure at the sheared contact between felsic volcanics and Burwash slates (Thorpe, 1972). Minor sphalerite-galena mineralization occurs to the north in the mafic volcanics, and also at one locality at the unconformity between the brecciated pre-Yellowknife basement and the overlying Raquette Lake Formation conglomerates (see Raquette Lake Formation). Minor pyrrhotite, pyrite, sphalerite, galena, chalcopyrite mineralizaiton also occurs in the metamorphosed metasedimentary rocks southwest of Payne Lake (Thorpe, 1972).

Base metal deposits are not present to any extent in the Kam mafic volcanics. One unusual exception occurs at Homer Lake in the northern part of the belt where a 30 m wide quartz porphyry dyke contains lead-zinc-silver mineralization in lenses along its margin, and a quartz-feldspar porphyry dyke at the north end of nearby Likely Lake contains pyrite, pyrrhotite, chalcopyrite, molybdenite and fluorite (Lord, 1951; Shegelski and Thorpe, 1972).

Nickel

A number of narrow veins containing niccolite, other arsenides and carbonates occur along the north shore of Hearne Channel from Gros Cap to south of Mad Lake. Although most are associated with the Caribou Lake Gabbro of the early Proterozoic Blachford Lake Intrusive Suite, some do occur in the Archean rocks at the mouth of the Francois River, the mouth of the Beaulieu River and as far west as Gros Cap. The linear array may suggest the veins are related to post-Archean faulting parallel to the north shore of Hearne Channel.

Nickel is also found in an ultramafic differentiate of a diabase sill 8 km northeast of Yellowknife.

Uranium-thorium

A few uranium prospects occur in the vicinity of Russell Lake and Stagg Lake. Mineralization appears to be related to the north-northwesterly trending faults in the area. One such prospect occurs on the Stagg River fault 16 km north of Great Slave Lake, and consists of low grade secondary mineralization within the weathered zone. (Thorpe, 1972). A second occurrence is found in a fault zone 20 km north-northeast, 3 km from the north boundary of the map area (McGlynn, 1971).

Spectacular uranium staining occurs in the vicinity of Trout Rock on the North Arm of Great Slave Lake (McGlynn, 1971). No evidence of primary mineralization was found. Such staining of the surface of outcrops of the Awry granite is fairly common and has been described in the section on that granite. Similar bright yellow stains also occur locally on outcrop surfaces of the Prosperous granites.

REFERENCES

Allison, I and Kerrich R.
1981: History of deformation and fluid transport in shear zones at Yellowknife; in Proceedings of the Gold Workshop, ed. R.D. Morton; Yellowknife Gold Workshop Committee (1980), Yellowknife N.W.T., p. 202-218.

Aydin, A. and Nur, A.
1982: Evolution of pull-apart basins and their scale independence; Tectonics, v. 1, p. 91-106.

Baadsgaard, H., Morton, R.D., and Olade, M.A.D.
1973: Rb-Sr isotopic age for the Precambrian lavas of the Seton Formation, East Arm of Great Slave Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 10, p. 1579-1582.

Baldham, J.P.M.
1979: Geology and petrochemistry of lower Aphebian (2.4-2.0 Ga) alkaline plutonic and hypabyssal rocks in the East Arm of Great Slave Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 16, p. 66-72.

Balkwill, H.R.
1971: Reconnaissance geology, southern Great Bear Plain, District of Mackenzie; Geological Survey of Canada, Paper 71-11, 47 p.

Bannerjee, I. and McDonald, B.C.
1975: Nature of esker sedimentation in glaciofluvial and glaciolacustrine sedimentation; in Glaciofluvial and Glaciolacustrine Sedimentation; ed. A.V. Joplin and B.C. McDonald; Society of Economic Paleontologists and Mineralogists, Special Publication no. 23, p. 132-154.

Baragar, W.R.A.
1961: The mineral industry of the District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 61-3, 29 p.

1962: Mineral industry of District of Mackenzie and part of District of Keewatin, 1961; Geological Survey of Canada, Paper 62-1, 41 p.

1966: Geochemistry of the Yellowknife volcanic rocks; Canadian Journal of Earth Sciences, v. 3, p. 9-30.

1975: Miscellaneous data from volcanic belts at Yellowknife, Wolverine Lake and James River, NWT; in Report of Activities Part A, Geological Survey of Canada, Paper 75-1A, p. 281-286.

Baragar, W.R.A. and Hornbrook, E.H.
1963: Mineral industry of District of Mackenzie, 1962; Geological Survey of Canada, Paper 63-9, 44 p.

Baragar, W.R.A. and Loveridge, W.D.
1982: A Rb-Sr study of the Natkusiak Basalts, Victoria Island, District of Franklin; in Rb-Sr and U-Pb Isotopic Age Studies, Report 5, in Current Research, Part C, Geological Survey of Canada, Paper 82-1C, p. 167-168.

Baragar, W.R.A. and McGlynn, J.C.
1976: Early Archean basement in the Canadian Shield: A review of the evidence; Geological Survey of Canada, Paper 76-14, 20 p.

Baragar, W.R.A., Plant, A.G., Pringle, G.J., and Schau, M.
1979: Diagenetic and postdiagenetic changes in the composition of an Archean pillow; Canadian Journal of Earth Sciences, v. 16, p. 2102-2121.

Barr, K.G.
1971: Crustal refraction experiment: Yellowknife 1966; Journal of Geophysical Research, v. 76, p. 1929-1947.

Bickle, M.J. and Erikson, K.A.
1982: Evolution and subsidence of early Precambrian sedimentary basins; in The evolution of Sedimentary Basins, ed. P. Kent, M.H.P. Bott, D.P. McKenzie, and C.A. Williams; The Royal Society of London, Philosophical Transactions, v. A305, p. 225-247.

Bostock, H.H.
1968: Gold-arsenopyrite-loellingite-pyrrhotite deposits in amphibolite, Itchen Lake-Contwoyto Lake area, District of Mackenzie; in Report of Activities, Part B; Geological Survey of Canada, Paper 68-1B, p. 72-76.

1977: The composition of hornblende, grunerite, and garnet in Archean iron formation of the Itchen Lake area, District of Mackenzie, Canada; Canadian Journal of Earth Sciences, v. 14, p. 1740-1752.

1980: Geology of the Itchen Lake area, District of Mackenzie; Geological Survey of Canada, Memoir 391, 101 p.

Bott, M.H.P.
1976: Formation of sedimentary basins of graben type by extension of the continental crust; in Sedimentary Basins of Continental Margins and Cratons, ed. M.H.P. Bott; Tectonophysics, v. 36, p. 77-86.

Bott, M.H.P. and Mithen, D.P.
1983: Mechanisms of graben formation – the wedge subsidence hypothesis; in Processes of Continental Rifting, ed. P. Morgan and B.H. Baker; Tectonophysics, v. 94, p. 11-22.

Bouma, A.H.
1962: Sedimentology of Some Flysch Deposits; Elsevier, Amsterdam, 168 p.

Bowering, S.A. and Van Schmus, W.R.
1982: Age and duration of igneous events, Wopmay Orogen, NWT, Canada; Geological Society of America, Abstracts with Programs, v. 14, p. 449.

Boyle, R.W.
1961: The geology, geochemistry and origin of the gold deposits of the Yellowknife District; Geological Survey of Canada, Memoir 310, 193 p.

1976: Mineralization processes in Archean greenstone and sedimentary belts; Geological Survey of Canada Paper 75-15, 45 p.

1979: The geochemistry of gold and its deposits; Geological Survey of Canada, Bulletin 280, 584 p.

Brown, C.E.G. and Dadson, A.S.
1953: Geology of the Giant Yellowknife Mine; Canadian Institute of Mining and Metallurgy, Transactions, v. 56, p. 69-86.

Brown, C.E.G., Dadson, A.S., and Wrigglesworth, L.A.
1959: On the ore-bearing structures of the Giant Yellowknife gold mine; Canadian Institute of Mining and Metallurgy, Transactions, v. 62, p. 107-116.

Brown, E.H.
1974: The origin of biotite in some Archean meta-sediments near Yellowknife, NWT, Canada: A Discussion; Contributions to Mineralogy and Petrology, v. 46, p. 75-76.

Brown, I.C.
1950: Christie Bay, Northwest Territories; Geological Survey of Canada, Paper 50-21, 8 p., 1 map.

Brown, I.C. (cont.)
1955: Late faults in the Yellowknife area; Geological Association of Canada, Proceedings, v. 7, p. 123-138.

Buddington, A.F.
1959: Granite emplacement with special reference to North America; Geological Society of America, Bulletin, v. 70, p. 671-747.

Burwash, R.A. and Baadsgaard, H.
1962: Yellowknife-Nonacho age and structural relations; in The Tectonics of the Canadian Shield, ed. J.S. Stevenson; Royal Society of Canada, Special Publications no. 4, p. 22-29.

Burwash, R.A., Baadsgaard, H., Campbell, F.A., Cumming, G.L., and Folinsbee, R.E.
1963: Potassium-argon dates of diabase dyke systems, District of Mackenzie, N.W.T; Canadian Institute of Mining and Metallurgy, Transactions, v. 66, p. 303-307.

Campbell, N.
1947: Regional structural features of the Yellowknife area; Economic Geology, v. 42, p. 687-698.

1949: The Con-Rycon Mine, Yellowknife, NWT; Canadian Institute of Mining and Metallurgy, Bulletin, v. 42, no. 446, p. 288-292.

Cater, F
1969: The Cloudy Pass epizonal batholith; Geological Society of America, Special Paper 116, 52 p.

Charbonneau, B.W.
1982: Radiometric study of three radioactive granites in the Canadian Shield: Elliot Lake, Ontario, Fort Smith, and Fury and Hecla, N.W.T.; in Uranium in Granites, ed. Y.T. Maurice; Geological Survey of Canada, Paper 81-23, p. 91-99.

Clee, T.E., Barr, K.G., and Berry, M.J.
1974: Fine structure of the crust near Yellowknife; Canadian Journal of Earth Sciences, v. 11, p. 1534-1549.

Clifford, P.M.
1972: Behavior of an Archean granitic batholith; Canadian Journal of Earth Sciences, v. 9, p. 71-77.

Coleman, L.C.
1957: Mineralogy of the Giant Yellowknife gold mine, Yellowknife, NWT; Economic Geology, v. 52, p. 400-425.

Condie, K.C. and Baragar, W.R.A.
1974: Rare earth element distributions in volcanic rocks from Archean greenstone belts; Contributions to Mineralogy and Petrology, v. 45, p. 237-246.

Craig, B.G.
1965: Glacial Lake McConnell and the surficial geology of parts of Slave River and Redstone River map areas, District of Mackenzie; Geological Survey of Canada, Bulletin 122, 33 p.

Cumming, G.L. and Tsong, F.
1975: Variations in the isotopic composition of volatilized lead and the age of the western granodiorite, Yellowknife, Northwest Territories; Canadian Journal of Earth Sciences, v. 12, p. 558-573.

Davidson, A.
1967: Metamorphism and intrusion in the Benjamin Lake map-area, Northwest Territories; unpublished Ph.D. thesis, University of British Columbia, Vancouver, 204 p.

Davidson, A. (cont.)
1972: Granite studies in the Slave Province (Parts of 85 I); in Report of Activities, Part A; Geological Survey of Canada, Paper 72-1A, p. 109-115.

1978a: The Blachford Lake Intrusive Suite: An Aphebian alkaline plutonic complex in the Slave Province, Northwest Territories; in Current Research, Part A; Geological Survey of Canada, Paper 78-1A, p. 119-127.

1978b: Interpretation of potassium-argon isotopic age GSC 76-188; in R.K. Wanless, R.D. Stevens, G.R. Lachance and R.N. Delabio, Age determinations and geological studies, K-Ar isotopic ages, Report 13; Geological Survey of Canada, Paper 77-2, p. 39-40.

1979: Interpretation of potassium-argon isotopic ages GSC 78-124 to GSC 78-136; in Age Determinations and Geological Studies, K-Ar Isotopic Ages, Report 14, ed. R.K. Wanless, R.D. Stevens, G.R. Lachance and R.N. Delabio; Geological Survey of Canada, Paper 79-2, p. 34-38.

1981: Petrochemistry of the Blachford Lake complex, District of Mackenzie (85 I/1, 2); Geological Survey of Canada, Open File 764.

1982: Petrochemistry of the Blachford Lake complex near Yellowknife, Northwest Territories; in Uranium in Granites, ed. Y.T. Maurice; Geological Survey of Canada, Paper 81-23, p. 71-79.

Dillon-Leitch, H.
1979: Preliminary geology map of 76 D/3, Courageous Lake, District of Mackenzie, N.W.T.; Department of Indian and Northern Affairs, Canada, Exploration and Geological Services Unit, EGS 1979-7.

Douglas, R.J.W. and Norris, A.W.
1960: Horn River map-area, Northwest Territories; Geological Survey of Canada, Paper 59-11, 23 p.

Douglas, R.J.W., Norris, A.W., and Norris, D.K.
1974: Horn River, District of Mackenzie; Geological Survey of Canada, Map 1372A.

Drury, S.A.
1977: Structures induced by granite diapirs in the Archean greenstone belt at Yellowknife, Canada: Implications for Archean geotectonics; Journal of Geology, v. 85, p. 345-358.

1979: Rare-earth and other trace element data bearing on the origin of Archean granitic rocks from Yellowknife, Northwest Territories; Canadian Journal of Earth Sciences, v. 16, p. 809-815.

Dubey, A.K. and Cobbold, P.R.
1977: Noncylindrical flexural slip folds in nature and experiment; Tectonophysics, v. 38, p. 223-239.

Duffield, W.A., Stieltjes, L., and Varet, J.
1982: Huge landslide blocks in the growth of Pitou de la Fournaise, La Réunion, and Kilauea Volcano, Hawaii; Journal of Volcanology and Geothermal Research, v. 12, p. 147-160.

Eade, K.E. and Fahrig, W.F.
1971: Geochemical evolutionary trends of continental plates – a preliminary study of the Canadian Shield; Geological Survey of Canada, Bulletin 179, 51 p.

Earth Physics Branch
1969: Rae, Northwest Territories; Earth Physics Branch, Department of Energy, Mines and Resources, Canada; Gravity Map Series no. 90.

Easton, R.M.
1982: Geology of the east side of Yellowknife Bay; Department of Indian and Northern Affairs, Canada, Exploration and Geological Service Unit, EGS 1982-5, 3 maps.

Easton, R.M. and Jackson, V.
1981: Geology of the Walsh Lake map area; Department of Indian and Northern Affairs, Canada, Exploration and Geological Services Unit; EGS 1981-3.

Easton, R.M., Boodle, R.L., and Zalvsky, L.
1982: Evidence for gneissic basement to the Archean Yellowknife Supergroup in the Point Lake area, Slave Structural Province, District of Mackenzie; in Current Research, Part B, Geological Survey of Canada, Paper 82-1B, p. 33-41.

Fahrig, W.F.
1965: Interpretation of potassium-argon isotopic age GSC 63-55; in Age Determinations and Geological Studies, Part 1 – Isotopic Ages, Report 5, ed. R.K. Wanless, R.D. Stevens, G.R. Lachance and R.Y.H. Rimsaite; Geological Survey of Canada, Paper 64-17 (Part 1), p. 47.

1970: Interpretation of potassium-argon isotopic age GSC 67-82; in Age Determinations and Geological Studies, K-Ar Isotopic Ages, Report 9, ed. R.K. Wanless, R.D. Stevens, G.R. Lachance and R.N. Delabio; Geological Survey of Canada, Paper 69-2A, p. 45.

1972: Interpretation of potassium-argon isotopic age GSC 70-69; in Age Determinations and Geological Studies, K-Ar Isotopic Ages, Report 10, ed. R.K. Wanless, R.D. Stevens, G.R. Lachance and R.N. Delabio; Geological Survey of Canada, Paper 71-2, p. 42.

Fahrig, W.F. and Jones, D.L.
1969: Paleomagnetic evidence for the extent of Mackenzie igneous events; Canadian Journal of Earth Sciences, v. 8, p. 455-467.

Fahrig, W.F. and Wanless, R.K.
1963: Age and significance of diabase dyke swarms of the Canadian Shield; Nature, v. 200, p. 234-937.

Folinsbee, R.E.
1942: Zone facies of metamorphism in relation to the ore deposits of the Yellowknife-Beaulieu Region, Northwest Territories; unpublished Ph.D. thesis, University of Minnesota, Minneapolis, 180 p.

1949: Lac de Gras, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 977A.

Folinsbee, R.E., Baadsgaard, H., Cumming, G.L., and Green, D.C.
1968: A very ancient island arc; American Geophysical Union, Geophysical Monograph 12, p. 441-448.

Fortier, Y.O.
1946: Yellowknife-Beaulieu Region, Northwest Territories; Geological Survey of Canada, Paper 46-23, 1 map with marginal notes.

1947: Ross Lake, Northwest Territories; Geological Survey of Canada, Paper 47-16, 4 p., 1 map.

Franklin, J.M. and Thorpe, R.I.
1982: Comparative Metallogeny of the Superior, Slave and Churchill Provinces; in Precambrian Sulphide Deposits, ed. R.W. Hutchinson, C.D. Spence and J.M. Franklin; Geological Association of Canada, Special Paper 25, p. 3-90.

Fraser, J.A.
1964: Geological notes on northeastern District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 63-40, 20 p.

Fraser, J.A., Hoffman, P.F., Irvine, T.N., and Mursky, G.
1972: The Bear Province: in Variations in Tectonic Styles in Canada, ed. R.A. Price and R.J.W. Douglas; Geological Association of Canada, Special Paper 11, p. 453-504.

Fremlin, G. (ed.)
1974: The National Atlas of Canada, 4th ed.; The Macmillan Co. of Canada Ltd., Toronto.

French, B.M.
1973: Mineral assemblages in diagenetic and low-grade metamorphic iron formation; Economic Geology, v. 68, p. 1063-1074.

Frith, R.A.
1981: Geology of the Mara River (E/2) and Beechey Lake (W/2) map area; Geological Survey of Canada, Open File 785.

Frith, R.A. and Hill, J.D.
1975: The geology of the Hackett-Back River greenstone belt-preliminary account; in Report of Activities, Part C; Geological Survey of Canada, Paper 75-1C, p. 367-370.

Frith, R.A. and Percival, J.A.
1978: Stratigraphy of the Yellowknife Supergroup in the Mara-Back rivers area, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 78-1C, p. 89-98.

Frith, R., Frith, R.A., and Doig, R.
1977: The geochronology of the granitic rocks along the Bear-Slave Structural Province boundary, northwest Canadian Shield; Canadian Journal of Earth Sciences, v. 14, p. 1356-1373.

Fuller, J.G.C.M. and Pollock, C.A.
1972: Early exposure of Middle Devonian reefs, southern Northwest Territories, Canada; 24th International Geological Congress, Section 6, Stratigraphy and Sedimentology, p. 144-155.

Fyson, W.K.
1975: Fabrics and deformation of Archean metasedimentary rocks, Ross Lake – Gordon Lake area, Slave Province, Northwest Territories; Canadian Journal of Earth Sciences, v. 12, p. 765-776.

1978: Structures intruded by granite diapirs in the Archean greenstone belt at Yellowknife, Canada: implications for Archean geotectonics: a discussion; Journal of Geology, v. 86, p. 767-769.

1980: Fold fabrics and emplacement of a granitoid pluton, Cleft Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 17, p. 325-332.

1981: Divergent fold overturning and regional tectonics, southern Slave Province, Northwest Territories; Precambrian Research, v. 14, p. 107-118.

1982: Complex evolution of folds and cleavages in Archean rocks, Yellowknife, N.W.T.; Canadian Journal of Earth Sciences, v. 19, p. 878-893.

Gates, T.M. and Hurley, P.M.
1973: Evaluation of Rb-Sr dating methods applied to the Matachewan, Abitibi, Mackenzie and Sudbury dyke swarms in Canada; Canadian Journal of Earth Sciences, v. 10, p. 900-919.

Geological Survey of Canada
1969a: Hearne Lake, District of Mackenzie; Map 7190G, Aeromagnetic Series, Geological Survey of Canada, Geophysics Paper 7190.

1969b: Yellowknife, District of Mackenzie, Map 7191G, Aeromagnetic Series, Geological Survey of Canada, Geophysics Paper 7191.

1969c: Wecho River, District of Mackenzie, Northwest Territories, Map 7196 G, Aeromagnetic Series, Geological Survey of Canada, Geophysics Paper 7196.

Gibb, R.A. and Thomas, M.D.
1980: Correlation of gravity anomalies with Yellowknife Supergroup rocks, North Arm, Great Slave Lake; Canadian Journal of Earth Sciences, v. 17, p. 1506-1516.

Gibbins, W.A., Seaton, J.B., Laporte, P.J., Murphy, J.D., Hurdle, E.J., and Padgham, W.A.
1977: Mineral industry report, 1974, Northwest Territories; Department of Indian and Northern Affairs, Canada, Exploration and Geological Service Unit, EGS 1977-5.

Goetz, P.A.
1974: Petrology and emplacement of the Detour Lake tonalite-granodiorite porphyry; unpublished B.Sc. thesis, University of Waterloo, 60 p.

Goodwin, A.M.
1977: Archean volcanism in Superior Province, Canadian Shield; Geological Association of Canada, Special Paper 16, p. 205-241.

Grasty, R.L. and Richardson, K.A.
1972: Airborne radioactivity map, Yellowknife, Northwest Territories; Geological Survey of Canada, Open File 124.

Green, D.C.
1968: Precambrian geology and geochronology of the Yellowknife area, N.W.T.; unpublished Ph.D. thesis, Department of Geology, The University of Alberta, 166 p.

Green, D.C. and Baadsgaard, H.
1971: Temporal evolution and petrogenesis of an Archean crustal segment at Yellowknife, N.W.T., Canada; Journal of Petrology, v. 12, p. 177-217.

Green, D.C., Baadsgaard, H., and Cumming, G.L.
1968: Geochronology of the Yellowknife area, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 5, p. 725-735.

Griffiths, P.S.
1980: Box-fault systems and ramps: atypical associations of structures from the eastern shoulder of the Kenya Rift; Geological Magazine, v. 117, p. 579-586.

Hauer, L.M.
1979: Preliminary geological map, northern end of the Yellowknife greenstone belt; Department of Indian and Northern Affairs, Canada, Exploration and Geological Services Unit, EGS 1979-10.

Helmstaedt, H., King. J., and Boodle, R.
1980: Geology of the Banting and Walsh Lakes map area NTS 85J/9; Department of Indian and Northern Affairs, Canada, Exploration and Geological Services Unit, EGS 1980-5.

Helmstaedt, H., Goodwin, J.A., Patterson, J.G., and King, J.
1979: Preliminary geological map-southern end of the Yellowknife greenstone belt; Department of Indian and Northern Affairs, Canada, Exploration and Geological Services Unit, EGS 1979-9.

Helmstaedt, H., King, J., Goodwin, J.A., and Patterson, J.G.
1981: Geology of the southwest end of the Yellowknife greenstone belt; in Proceedings of the Gold Workshop, ed. R.D. Morton; Yellowknife Gold Workshop Committee (1980), Yellowknife N.W.T., p. 232-239.

Henderson, J.B.
1970: Stratigraphy of the Yellowknife Supergroup, Yellowknife Bay – Prosperous Lake area, District of Mackenzie; Geological Survey of Canada, Paper 70-26, 12 p.

1972: Sedimentology of Archean turbidites at Yellowknife, Northwest Territories; Canadian Journal of Earth Sciences, v. 9, p. 882-902.

1973: A contribution to the structural geology at Yellowknife, District of Mackenzie; in Report of Activites, Part B; Geological Survey of Canada, Paper 73-1B, p. 106-109.

1975a: Sedimentological studies of the Yellowknife Supergroup in the Slave Structural Province; in Report of Activities, Part A; Geological Survey of Canada, Paper 75-1A, p. 325-330.

1975b: Sedimentology of the Archean Yellowknife Supergroup at Yellowknife, District of Mackenzie; Geological Survey of Canada, Bulletin 246, p. 62.

1975c: Archean stromatolites in the northern Slave Province, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 12, p. 1619-1630.

1977: Keskarrah Bay, Northwest Territories; Geological Survey of Canada, Open File 447.

1978a: Age and origin of the gold bearing shear zones at Yellowknife, Northwest Territories; in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, p. 259-262.

1978b: Interpretation of potassium-argon isotopic ages GSC 76-184 – GSC 76 187; in Age Determinations and Geological Studies, K-Ar Isotopic Ages, Report 13, ed. R.K. Wanless, R.D. Stevens, G.R. Lachance and R.N. Delabio; Geological Survey of Canada, Paper 77-2, p. 38, 39.

1981: Archean basin evolution in the Slave Province, Canada; in Precambrian Plate Tectonics, ed. A. Kroner; Elsevier, Amsterdam, p. 213-235.

Henderson, J.B. and Easton, R.M.
1977: Archean supracrustal-basement rock relationships in the Keskarrah Bay map-area, Slave Structural Province, District of Mackenzie; in Report of Activities, Part A; Geological Survey of Canada, Paper 77-1A, p. 217-221.

Henderson, J.B. and Thompson, P.H.
1980: The Healey Lake map area (Northern Part) and the enigmatic Thelon Front, District of Mackenzie; in Current Research, Part A; Geological Survey of Canada, Paper 80-1A, p. 165-169.

1982: Geology, Healey Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 860.

Henderson, J.B., Loveridge, W.D., and Sullivan, R.W.
1982: A U-Pb study of zircon from granitic basement beneath the Yellowknife Supergroup, Point Lake, District of Mackenzie; in Rb-Sr and U-Pb Isotopic Age Studies, Report 5, in Current Research, Part C, Geological Survey of Canada, Paper 82-1C, p. 173-178.

Henderson, J.B., Thompson, P.H., and James, D.T.
1982: The Healey Lake map area and the Thelon Front problem, District of Mackenzie; in Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p. 191-195.

Henderson, J.F.
1938: Beaulieu River area, Northwest Territories; Geological Survey of Canada, Paper 38-1, 20 p., 1 map.

1939: Beaulieu River area, Northwest Territories; Geological Survey of Canada, Paper 39-1, 16 p., 1 map.

1941a: Gordon Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 644A.

1941b: Gordon Lake South, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 645A.

1943: Structure and metamorphism of early Precambrian rocks between Gordon and Great Slave Lakes, N.W.T.; American Journal of Science, v. 241, p. 430-446.

Henderson, J.F. and Brown, I.C.
1948: Yellowknife, Northwest Territories (Sheets 1 and 2); Geological Survey of Canada, Paper 48-17, 6 p., 2 maps.

1949: Yellowknife, Northwest Territories (Sheet 3); Geological Survey of Canada, Paper 49-26, 1 map.

1950: Yellowknife, Northwest Territories (Sheet 4); Geological Survey of Canada, Paper 50-34, 1 map.

1952a: Yellowknife, Northwest Territories (Sheet 5); Geological Survey of Canada, Paper 52-24, 1 map.

1952b: The Yellowknife Greenstone Belt, Northwest Territories; Geological Survey of Canada, Paper 52-28, 41 p., 1 map.

1966: Geology and structure of the Yellowknife Greenstone Belt, District of Mackenzie; Geological Survey of Canada, Bulletin 141, 87 p.

Henderson, J.F. and Jolliffe, A.W.
1939: Relation of gold deposits to structure, Yellowknife and Gordon Lake areas, Northwest Territories; Canadian Institute of Mining and Metallurgy, Transactions, v. XLII, p. 314-336.

1941: Beaulieu River, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 581A.

Heywood, W.W. and Davidson, A.
1969: Geology of the Benjamin Lake map-area, District of Mackenzie; Geological Survey of Canada, Memoir 361, 35 p.

Hill, J.D.
1980: The structural development and crystallization of the Kenoran granitoid plutons in the Nose Lake – Back River area, Northwest Territories; unpublished Ph.D. thesis, The University of Western Ontario, London, Ontario, 239 p.

Hobbs, B.E., Means, W.D., and Williams, P.F.
1976: An Outline of Structural Geology; John Wiley & Sons, Inc., 571 p.

Hoffman, P.F.
1968: Stratigraphy of the Great Slave Supergroup (Aphebian), east arm of Great Slave Lake, District of Mackenzie; Geological Survey of Canada, Paper 68-42, 93 p.

Hoffman, P.F. (cont.)
1969: Proterozoic paleocurrents and depositional history of the east arm fold belt, Great Slave Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 6, p. 441-462.

1973a: Aphebian supracrustal rocks of the Athapuscow Aulacogen, east arm of Great Slave Lake, District of Mackenzie; in Report of Activities, Part A, Geological Survey of Canada, Paper 73-1A, p. 151-156.

1973b: Evolution of an early Proterozoic continental margin: the Coronation Geosyncline and associated aulacogens, northwest Canadian Shield, in Evolution of the Precambrian Crust, ed. J. Sutton and B.F. Windley; The Royal Society of London, Philosophical Transactions, Series A, v. 273, p. 547-581.

1974: Shallow and deep water stromatolites in a lower Proterozoic platform to basin facies change, Great Slave Lake, Canada; American Association of Petroleum Geologists, Bulletin v. 58, p. 856-867.

1977: Preliminary geology of Proterozoic formations in the east arm of Great Slave Lake, District of Mackenzie; Geological Survey of Canada, Open File 475.

1980a: Wopmay Orogen: A Wilson cycle of early Proterozoic age in the northwest of the Canadian Shield; in The Continental Crust and its Mineral Deposits, ed. D.W. Strangway; Geological Association of Canada, Special Paper 20, p. 522-549.

1980b: Conjugate transcurrent faults in north-central Wopmay Orogen (early Proterozoic) and their dip slip reactivation during post-orogenic extension, Hepburn Lake map-area, District of Mackenzie; in Current Research, Part A, Geological Survey of Canada, Paper 80-1A, p. 183-185.

1981: Autopsy of Athapuscow Aulacogen: A failed arm affected by three collisions; in Proterozoic Basins of Canada, ed, F.H.A. Campbell; Geological Survey of Canada, Paper 81-10, p. 97-102.

Hoffman, P.F., Bell, I.R., Hildebrand, R.S., and Thorstad, L.
1977: Geology of the Athapuscow Aulacogen, east arm of Great Slave Lake, District of Mackenzie; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 117-129.

Hoffman, P.F., Dewey, J.F., and Burke, K.
1974: Aulacogens and their genetic relation to geosynclines, with a Proterozoic example from Great Slave Lake, Canada; in Modern and Ancient Geosynclinal Sedimentation, ed. R.H. Dott and R.H. Shaver; Society of Economic Paleontologists and Mineralogists, Special Publication no. 19, p. 38-55.

Hoschek, G.
1967: Untersuchungen zum Stabilitatsbereich von Chloritoid and Staurolith; Contributions to Mineralogy and Petrology, v. 14, p. 123-162.

Hurdle, E.
1983: Geology of a volcanic pile at Clan Lake, N.W.T.; Department of Indian and Northern Affairs, Canada, Exploration and Geological Service Unit, EGS 1983-5, 3 maps.

Hutchinson, R.W.
1955: Regional zonation of pegmatites near Ross Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Bulletin 34, 50 p.

Hyde, R.S. and Walker, R.G.
1977: Sedimentary environments and the evolution of the Archean greenstone belt in the Kirkland Lake area, Ontario; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 185-190.

Illies, J.H.
1981: Mechanism of graben formation; in Mechanism of Graben Formation, ed. J.H. Illies; Tectonophysics, v. 73, p. 249-266.

Irving, E. and McGlynn, J.C.
1976: Polyphase magnetization of the Big Spruce Complex, Northwest Territories; Canadian Journal of Earth Sciences, v. 13, p. 476-489.

Ishizaka, K. and Yanagi, T.
1977: K, Rb, and Sr abundances and Sr isotopic composition of the Tanzawa granitic and associated gabbroic rocks, Japan: low potash island arc plutonic complex; Earth and Planetary Science Letters, v. 33, p. 345-352.

Jenner, G.A., Fryer, B.J., and McLennan, S.M.
1981: Geochemistry of the Archean Yellowknife Supergroup; Geochimica et Cosmochimica Acta, v. 45, p. 1111-1129.

Jolliffe, A.W.
1939: Quyta Lake and parts of Fishing Lake and Prosperous Lake area, Northwest Territories; Geological Survey of Canada, Paper 39-6, 3 maps.

1940: Quyta lake and parts of Fishing Lake and Prosperous Lake areas, Northwest Territories; Geological Survey of Canada, Paper 40-14, 9 p.

1942: Yellowknife Bay, Northwest Territories; Geological Survey of Canada Map 709A.

1944: Rare element minerals in pegmatites, Yellowknife-Beaulieu area, Northwest Territories; Geological Survey of Canada, Paper 44-12, 23 p.

1945: Yellowknife faults and orebodies; The Precambrian, May 1945.

1946: Prosperous Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 868A.

Jolliffe, F.
1936: Yellowknife River area, Northwest Territories; Geological Survey of Canada, Paper 36-5, 10 p.

Kamineni, D.C.
1975: Chemical mineralogy of some cordierite bearing rocks near Yellowknife, Northwest Territories, Canada; Contributions to Mineralogy and Petrology, v. 53, p. 293-310.

1974: Variation in the distortion index of cordierite east of the Sparrow Lake granite pluton, District of Mackenzie; Canadian Mineralogist, v. 12, p. 419-421.

1973a: X-ray and Mossbauer characteristics of a cummingtonite from Yellowknife, District of Mackenzie; Canadian Mineralogist, v. 12, p. 230-232.

1973b: Petrology and geochemistry of some Archean metamorphic rocks near Yellowknife, District of Mackenzie; unpublished Ph.D. thesis, University of Ottawa, Ottawa, 228 p.

Kamineni, D.C. and Carrara, A.
1973: Comparison of the composition of porphyroblastic and fabric – forming biotite in two metamorphic rocks; Canadian Journal of Earth Sciences, v. 10, p. 948-953.

Kamineni, D.C. and Divi, S.R.
1976: Growth of cordierite in the Archean metasediments near Yellowknife, District of Mackenzie, Canada; Geologische Rundschau, v. 66, p. 255-263.

Kamineni, D.C. and Wong, A.S.
1973: Gedrite from Yellowknife – Beaulieu region, District of Mackenzie, N.W.T.; Canadian Mineralogist, v. 11, p. 1012-5.

Kamineni, D.C., Divi, S.R., and Tella, S.
1979: Time relations of metamorphism and deformation in Archean metasedimentary rocks near Yellowknife, Canada; Neues Jahrbuch fur Mineralogie Monatshefte, Jahrgang 1979, p. 34-48.

Kelley, V.C.
1978: Tectonics, middle Rio Grande Rift, New Mexico; in Rio Grande Rift: Tectonics and Magmatism, ed. R.E. Riecker; American Geophysical Union, Washington, p. 57-70.

Kerrich, R.
1981: Archean lode gold deposits: a synthesis of data on metal distribution, rare earth elements and stable isotopes, with special reference to Yellowknife; in Proceedings of the Gold Workshop, ed. R.D. Morton; Yellowknife Gold Workshop Committee (1980), Yellowknife, N.W.T., p. 95-171.

Kerrich, R., Fyfe, W.S., and Allison, I.
1977: Iron reduction around quartz veins, Yellowknife district, Northwest Territories, Canada; Economic Geology, v. 72, p. 657-663.

Kretz, R.
1968: Study of pegmatite bodies and enclosing rocks, Yellowknife-Beaulieu region; Geological Survey of Canada, Bulletin 159, 109 p.

1970: Variation in the composition of muscovite and albite in a pegmatite dyke near Yellowknife; Canadian Journal of Earth Sciences, v. 7, p. 1219-1235.

1973: Kinetics of the crystallization of garnet at two localities near Yellowknife; Canadian Mineralogist, v. 12, p. 1-20.

Kretz, R., Garrett, D., and Garrett, R.G.
1982: Na-K-Li geochemistry of the Prestige pluton in the Slave Province of the Canadian Shield; Canadian Journal of Earth Sciences, v. 19, p. 540-554.

Krogh, T.E. and Gibbins, W.
1978: U-Pb isotopic ages of basement and supracrustal rocks in the Point Lake area of the Slave Structural Province, Canada; in Abstracts with Programs, Geological Association of Canada – Mineralogical Association of Canada, v. 3, p. 438.

Lambert, M.B.
1974: Archean volcanic studies in the Slave-Bear province; in Report of Activities, Part A, Geological Survey of Canada, Paper 74-1A, p. 177-179.

1976: The Back River volcanic complex, District of Mackenzie; in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, p. 363-365.

1977: Anatomy of a Greenstone Belt, Slave Province, Northwest Territories; in Volcanic Regimes in Canada, ed. W.R.A. Baragar, L.C. Coleman and J.M. Hall; Geological Association of Canada, Special Paper 16, p. 331-340.

Lambert, M.B. (cont.)
1978: The Back River volcanic complex – A cauldron subsidence structure of Archean age, in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, p. 153-157.

1982: Synvolcanic intrusions in the Cameron River volcanic belt, District of Mackenzie; in Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p. 165-167.

– The Cameron and Beaulieu River volcanic belts, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Bulletin 349. (in press)

Lambert, M.B. and Henderson, J.B.
1980: A uranium-lead age of zircons from volcanics and metasediments of the Back River Volcanic Complex, eastern Slave Province, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 80-1C, p. 239-242.

Lang, A.H., Griffith, J.W., and Steacy, H.R.
1962: Canadian deposits of uranium and thorium; Geological Survey of Canada, Economic Geology Series, no. 16, 2nd edition, 173 p.

Laporte, P.J., Gibbins, W.A., Hurdle, E.J., Lord, C., Padgham, W.A., and Seaton, J.B.
1978: Mineral industry report 1975, Northwest Territories; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 1978-5.

Lasmanis, R.
1978: Lithium resources in the Yellowknife Area, Northwest Territories, Canada; Energy, v. 3, p. 399-407.

Law, J.
1971: Regional Devonian geology and oil and gas possibilities, upper Mackenzie River area; Canadian Petroleum Geology, Bulletin, v. 19, p. 437-486.

Leech, A.P.
1965: A reconnaissance: basic intrusive rocks of the Precambrian Shield, Canada; unpublished M. Sc. thesis, University of Alberta, Edmonton, p.

1966: Potassium-argon dates of basic intrusive rocks of the District of Mackenzie, NWT.; Canadian Journal of Earth Sciences, v. 3, p. 389-412.

LeMaitre, R.W.
1976: The chemical variability of some common igneous rocks; Journal of Petrology, v. 17, p. 589-637.

Little, H.W.
1959: Tungsten deposits of Canada; Geological Survey of Canada, Economic Geology Series no. 17, 251 p.

Lord, C., Laporte, P.J., Gibbins, W.A., Hurdle, E.J., Seaton, J.B., and Padgham, W.A.
1978: Mineral industry report 1976, Northwest Territories; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 1978-11.

Lord, C., Laporte, P.J., Gibbons, W.A., Seaton, J.B., Goodwin, J.A., and Padgham, W.A.
1981: Mineral industry report 1977, Northwest Territories; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 1981-11, 202 p.

Lord, C.S.
1941: Mineral industry of the Northwest Territories; Geological Survey of Canada, Memoir 230, 136 p.

1942: Snare River and Ingray Lake map-areas; Geological Survey of Canada, Memoir 235, 55 p.

1951: Mineral industry of District of Mackenzie, Northwest Territories; Geological Survey of Canada, Memoir 261, 336 p.

Marmo, V.
1971: Granite Petrology and the Granite Problem; Elsevier, Amsterdam.

McGlynn, J.C.
1971: Metallic mineral industry, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 70-17, 194 p.

1977: Geology of Bear-Slave structural provinces, District of Mackenzie; Geological Survey of Canada, Open File 445.

McGlynn, J.C. and Henderson, J.B.
1970: Archean volcanism and sedimentation in the Slave Structural Province; in Symposium on Basins and Geosynclines of the Canadian Shield, ed, A. Baer; Geological Survey of Canada, Paper 70-40, p. 31-44.

McGlynn, J.C. and Irving, E.
1975: Paleomagnetism of early Aphebian diabase dykes from the Slave Structural Province, Canada; Tectonophysics, v. 26, p. 23-38.

McGrath, P.H., Henderson, J.B., and Lindia, F.M.
1983: Interpretation of a gravity profile over a contact zone between an Archean granodiorite and the Yellowknife Supergroup using an interactive computer program with partial automatic optimization; in Current Research, Part B, Geological Survey of Canada, Paper 83-1B, p. 189-194.

McKenzie, D.P.
1978: Some remarks on the development of sedimentary basins; Earth and Planetary Science Letters, v. 40, p. 25-32.

Meijer-Dress, N.C.
1975: Geology of the lower Paleozoic formations in the subsurface of the Fort Simpson area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 74-40, 65 p.

Middleton, G.V. and Hampton, M.A.
1973: Sediment gravity flows: mechanics of flow and deposition; in Turbidites and Deep-water Sedimentation, ed. G.V. Middleton and A.H. Bouma; Society of Economic Paleontologists and Mineralogists, short course lecture notes.

Moore, J.C.G.
1956: Courageous-Matthews Lakes area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Memoir 283, 52 p.

Moore, J.C.G., Miller, M.L., and Barnes, R.W.
1951: Carp Lakes, Northwest Territories; Geological Survey of Canada, Paper 51-8, 1 map with marginal notes.

Moore, J.G.
1966: Gravity slide origin of rift zones of some Hawaiian volcanoes; Bulletin Volcanologique, v. 29, p. 719-720.

Mulligan, R.
1965: Geology of Canadian lithium deposits; Geological Survey of Canada, Economic Geology Report 21, 131 p.

Newton, A.R. and Slaney, V.R.
1978: Geological interpretation of an airborne gamma-ray spectrometer survey of the Hearne Lake area, Northwest Territories; Geological Survey of Canada, Paper 77-32, 14 p.

Nikic, Z., Baadsgaard, H., Folinsbee, R.E., Krupicka, J., Leech, A.P., and Sasaki, A.
1980: Boulders from the basement, the trace of an ancient crust?; in Selected Studies of Archean Gneisses and Lower Proterozoic Rocks, Southern Canadian Shield, ed. G.B. Morey and G.N. Hanson; Geological Society of America, Special Paper 182, p. 169-175.

Norris, A.W.
1965: Stratigraphy of middle Devonian and older Paleozoic rocks of the Great Slave Lake region, Northwest Territories; Geological Survey of Canada, Memoir 322, 180 p.

North American Commission on Stratigraphic Nomenclature
1983: North American Stratigraphic Code; American Association of Petroleum Geologists, Bulletin, v. 67, p. 841-875.

Olade, M.A.D. and Morton, R.D.
1972: Observations on the Proterozoic Seton Formation, East Arm of Great Slave Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 9, p. 1110-1120.

Padgham, T., Caine, T.W., Hughes, D.R., Jefferson, C.W., Kennedy, M.W., and Murphy, J.D.
1978: Mineral industry report 1969 and 1970, Northwest Territories, west of 104 W longitude; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 78-6.

Padgham, W.A.
1974: Preliminary geology map of High Lake, NWT, 76M/7; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, Geological Survey of Canada, Open File 208.

1980: An Archean ignimbrite at Yellowknife and its relationship to the Kam Formation basalts; Precambrian Research, v. 12, p. 99-113.

1981a: Geology of the Yellowknife volcanic belt; in Proceedings of the Gold Workshop, ed. R.D. Morton; Yellowknife Gold Workshop Committee (1980), Yellowknife, N.W.T., p. 288-322.

1981b: Gold deposits in the Northwest Territories; in Proceedings of the Gold Workshop, ed. R.D. Morton; Yellowknife Gold Workshop Committee (1980), Yellowknife, N.W.T., p. 174-200.

Padgham, W.A. and Bryan, M.P.D.
1975: Geology Agricola Lake 76G/12, District of Mackenzie, Northwest Territories; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 1975-2.

Padgham, W.A. and Ronayne, E.A.
1976: Geology Hackett River 76F/16, District of Mackenzie, Northwest Territories; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS 1976-8.

Padgham, W.A., Kennedy, M.W., Jefferson, C.W., Hughes, D.R., and Murphy, J.D.
1975: Mineral industry report 1971 and 1972, Northwest Territories, west of 104° west longitude; Department of Indian and Northern Affairs, Exploration and Geological Services Unit, EGS-1975-8.

Patchett, P.J., van Breemen, O., and Martin, R.F.
1979: Sr isotopes and the structural state of feldspars as indicators of post magmatic hydrothermal activity in continental dolerites; Contributions to Mineralogy and Petrology, v. 69, p. 65-73.

Payne, A.V., Baadsgaard, H., Burwash, R.A., Cumming, G.L., Evans, C.R., and Folinsbee, R.E.
1965: A line of evidence supporting continental drift; in The Upper Mantle Symposium, New Delhi, ed. C.H. Smith and T. Sorgenfrei; International Union of Geological Sciences, Copenhagen, p. 83.

Pettijohn, F.J.
1957: Sedimentary Rocks; Harper & Bros., New York, 718 p.

Prest, V.K.
1969: Retreat of Wisconsin and Recent ice in North America; Geological Survey of Canada, Map 1257A.

Ramsay, C.R.
1973a: The origin of biotite in Archean metasediments near Yellowknife N.W.T., Canada; Contributions to Mineralogy and Petrology, v. 42, p. 43-54.

1973b: Controls of biotite zone mineral chemistry in Archean meta-sediments near Yellowknife, Northwest Territories, Canada; Journal of Petrology, v. 14, p. 467-488.

1973c: Metamorphism and gold mineralization of Archean metasediments near Yellowknife, N.W.T., Canada; unpublished Ph.D. thesis, University of Alberta, Edmonton.

1974a: The cordierite isograd in Archean metasediments near Yellowknife, NWT., Canada – variations on an experimentally established reaction; Contributions to Mineralogy and Petrology, v. 47, p. 27-40.

1974b: The origin of biotite in Archean metasediments near Yellowknife, N.W.T., Canada: a reply to E.H. Brown; Contributions to Mineralogy and Petrology, v. 46, p. 77-79.

Ramsay, C.R. and Kamineni, D.C.
1977: Petrology and evolution of an Archean metamorphic aureole in the Slave craton, Canada; Journal of Petrology, v. 18, p. 460-486.

Ramsay, J.G.
1967: Folding and Fracturing of Rocks; McGraw-Hill, New York, 568 p.

Reches, Z.
1978: Analysis of faulting in three dimensional strain field; Tectonophysics, v. 47, p. 109-129.

Richmond, W.O.
1965: Paleozoic stratigraphy and sedimentation of the Slave Point Formation, southern Northwest Territories and northern Alberta; unpublished Ph.D. thesis, Department of Geology, Stanford University, U.S.A.

Roddick, J.A.
1965: Vancouver, North Coquitlam and Pitt Lake map-areas, British Columbia; Geological Survey of Canada, Memoir 335, 276 p.

Roddick, J.A. and Armstrong, J.E.
1959: Relict dykes in the coast mountains near Vancouver, British Columbia; Journal of Geology, v. 67, p. 603-613.

Rose, E.R.
1969: Geology of titanium and titaniferous deposits of Canada; Geological Survey of Canada, Economic Geology Report no. 25, 177 p.

Rowe, R.B.
1952: Pegmatite mineral deposits of the Yellowknife-Beaulieu region, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 52-8, 36 p.

Schärer, U. and Allègre, C.J.
1982: Investigation of the Archean crust by single grain dating of detrital zircon: a greywacke of the Slave Province, Canada; Canadian Journal of Earth Sciences, v. 19, p. 1910-1918.

Schau, M. and Henderson, J.B.
1983: Archean chemical weathering at three localities on the Canadian Shield; Precambrian Research, v. 20, p. 189-224.

Schiller, E.A.
1965: Mineral industry of the Northwest Territories, 1964; Geological Survey of Canada, Paper 65-11, 62 p.

Schiller, E.A. and Hornbrook, E.H.
1964: Mineral industry of District of Mackenzie, 1963; Geological Survey of Canada, Paper 64-22, 43 p.

Seifert, F.
1970: Low-temperature compatibility relations of cordierite in haplo-pelites of the system K_2O-MgO-Al_2O-SiO_2-H_2O; Journal of Petrology, v. 11, p. 73-100.

Shegelski, R.J. and Thorpe, R.I.
1972: Study of selected mineral deposits in the Bear and Slave provinces; in Report of Activities, Part A, Geological Survey of Canada, Paper 72-1A, p. 93-96.

Steiger, R.H. and Jaeger, E.
1977: Subcommission on geochronology: Convention on the use of decay constants in geo- and cosmo-chronology, Earth and Planetary Sciences Letters, v. 36, p. 359-362.

Stockwell, C.H.
1933: Great Slave Lake – Coppermine River area, Northwest Territories; Geological Survey of Canada, Summary Report 1932, Part C, p. 37-63.

1936: Eastern portion of Great Slave Lake (west half); Geological Survey of Canada, Map 337A.

1962: Interpretation of potassium-argon isotopic age GSC 60-49; in Age Determinations by the Geological Survey of Canada, Report 2, Isotopic Ages, ed. J.A. Lowdon; Geological Survey of Canada, Paper 61-17, p. 29.

1963: Interpretation of potassium-argon isotopic age GSC 61-64, GSC 61-66 and GSC 61-68; in Age Determinations and Geological Studies (including Isotopic Ages – Report 3), ed. J.A. Lowdon, C.H. Stockwell, H.W. Tipper and R.K. Wanless; Geological Survey of Canada, Paper 62-17, p. 39-42.

1970: Geology of the Canadian Shield; in Geology and Economic Minerals of Canada, ed. R.J.W. Douglas; Geological Survey of Canada, Economic Geology Report no. 1, p. 44-54.

Streckeisen, A.
1976: To each plutonic rock its proper name; Earth Science Reviews, v. 12, p. 1-33.

Swanson, D.A., Duffield, W.A., and Riske, R.S.
1976: Displacement of the south flank of Kilauea volcano: the result of forceful intrusion of magma into the rift zones; United States Geological Survey, Professional Paper 963, p. 39.

Teal, P.R. and Walker, R.G.
1977: Stratigraphy and sedimentology of the Archean Manitou Group, northwestern Ontario; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 181-184.

Thomas, D.G.
1982: Thor Lake Tantalum-Columbium-Rare Earths Property; Highwood Resources Ltd. Annual Report, 1981, p. 2-3.

Thompson, P.H.
1978: Archean regional metamorphism in the Slave Structural Province – a new perspective of some old rocks; in Metamorphism in the Canadian Shield, ed. J.A. Fraser and W.W. Heywood; Geological Survey of Canada, Paper 78-10, p. 85-102.

Thorpe, R.I.
1966: Mineral industry of the Northwest Territories, 1965; Geological Survey of Canada, Paper 66-52, 66 p.

1971: Comments on rock ages in the Yellowknife area, District of Mackenzie; Report of Activities, Part B, Geological Survey of Canada, Paper 71-1B, p. 76-79.

1972: Mineral exploration and mining activities, mainland, Northwest Territories, 1966 to 1968 (excluding the Coppermine River area); Geological Survey of Canada, Paper 70-70, 204 p.

Tirrul, R. and Bell, I.
1980: Geology of the Anialik River greenstone belt, Hepburn Island map-area, District of Mackenzie; in Current Research, Part A, Geological Survey of Canada, Paper 80-1A, p. 157-164.

Traill, R.J.
1950: Dacites of the Lower Yellowknife Group; unpublished B.Sc. thesis, Department of Geology, Queen's University, Kingston.

Tremblay, L.P.
1954: Chalco Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1023A.

1976: Geology of northern Contwoyto Lake area, District of Mackenzie; Geological Survey of Canada, Memoir 381, 56 p.

Trembly, L.P., Wright, G.M., and Miller, M.L.
1953: Ranji Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1022A.

Turner, C.C. and Walker, R.G.
1973: Sedimentology, stratigraphy and crustal evolution of the Archean greenstone belt near Sioux Lookout, Ontario; Canadian Journal of Earth Sciences, v. 10, p. 817-845.

Turner, F.J.
1968: Metamorphic Petrology; McGraw-Hill, New York, 403 p.

Turner, F.J. (cont.)
1981: Metamorphic Petrology: Mineralogical, Field and Tectonic Aspects; 2nd edition, McGraw-Hill Book Co., New York, 524 p.

Van Bemmelen, R.W.
1954: Mountain Building; Martinus Nyhoff, The Hague, 177 p.

Van Breemen, O.
1965: Thermally induced relocation of strontium and rubidium in a granodiorite; unpublished M.Sc. thesis, University of Alberta, Edmonton, Alberta, 34 p.

Vening Meinesz, F.A.
1950: Les graben africains, résultat de compression ou de tension dans la croûte terrestre?; Bulletin des Seances, Institute Royal Colonial Belge, v. 21, p. 539-552.

Walker, R.G.
1978: A critical appraisal of Archean basin-craton complexes; Canadian Journal of Earth Sciences, v. 15, p. 1213-1218.

Wanless, R.K.
1970: Isotopic age map of Canada; Geological Survey of Canada, Map 1256A.

Wanless, R.K., Boyle, R.W., and Lowdon, J.A.
1960: Sulphur isotope investigation of the gold quartz deposits of the Yellowknife district; Economic Geology, v. 55, p. 1591-1621.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Delabio, R.N.
1979: Age determinations and geological studies; K-Ar isotope ages, Report 14; Geological Survey of Canada, Paper 79-2, p. 34-38.

Watterson, J.
1965: Plutonic development of the Ilordleq area, south Greenland, Part 1: Chronology, and the occurrence and recognition of metamorphosed basic dykes; Grønlands Geologiske Undersøgelse, Bulletin 51, p. 147.

Williams, G.K.
1974: Lower Paleozoic, Slave River map-area, District of Mackenzie (NTS 85); in Report of Activities, Part B; Geological Survey of Canada, Paper 74-1B, p. 287-290.

Winkler, H.G.F.
1974: Petrogenesis of Metamorphic Rocks; Springer-Verlag, New York, 320 p.

Wiwchar, M.B.
1957: Consolidated Discovery Yellowknife Mine; in Structural Geology of Canadian Ore Deposits, v. 2, ed. G. Gilbert; Canadian Institute of Mining and Metallurgy, p. 201-210.

Wright, G.M.
1954: Ghost Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1021A.

INDEX