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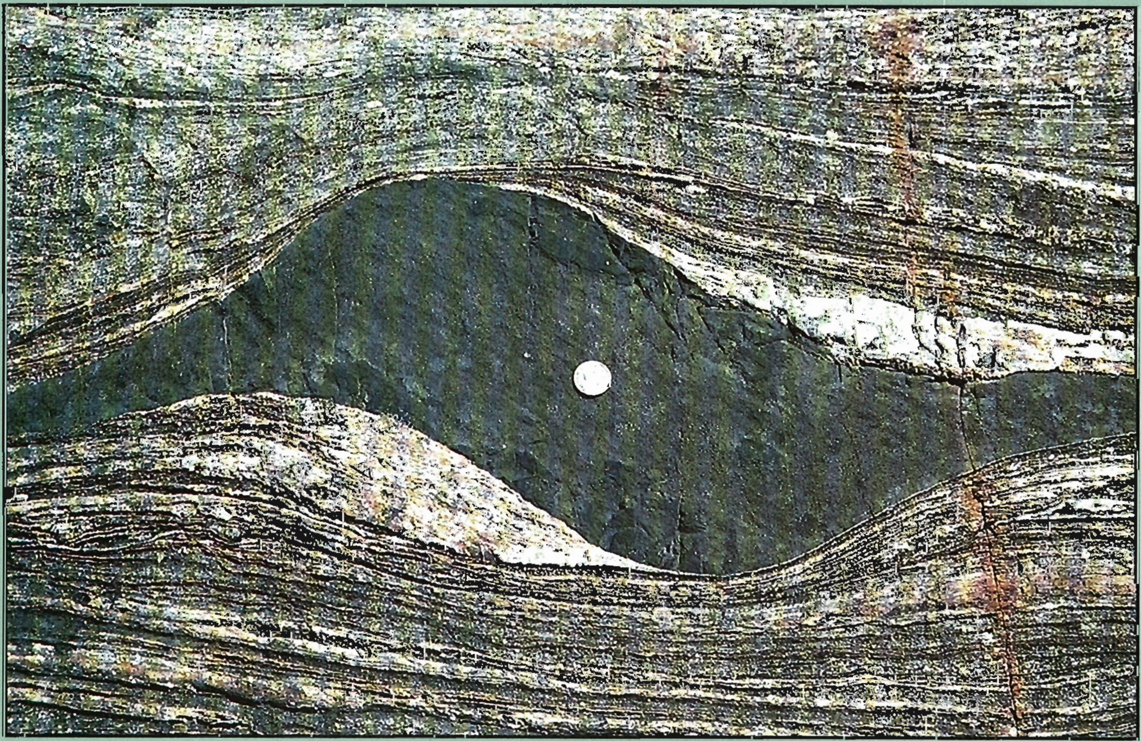

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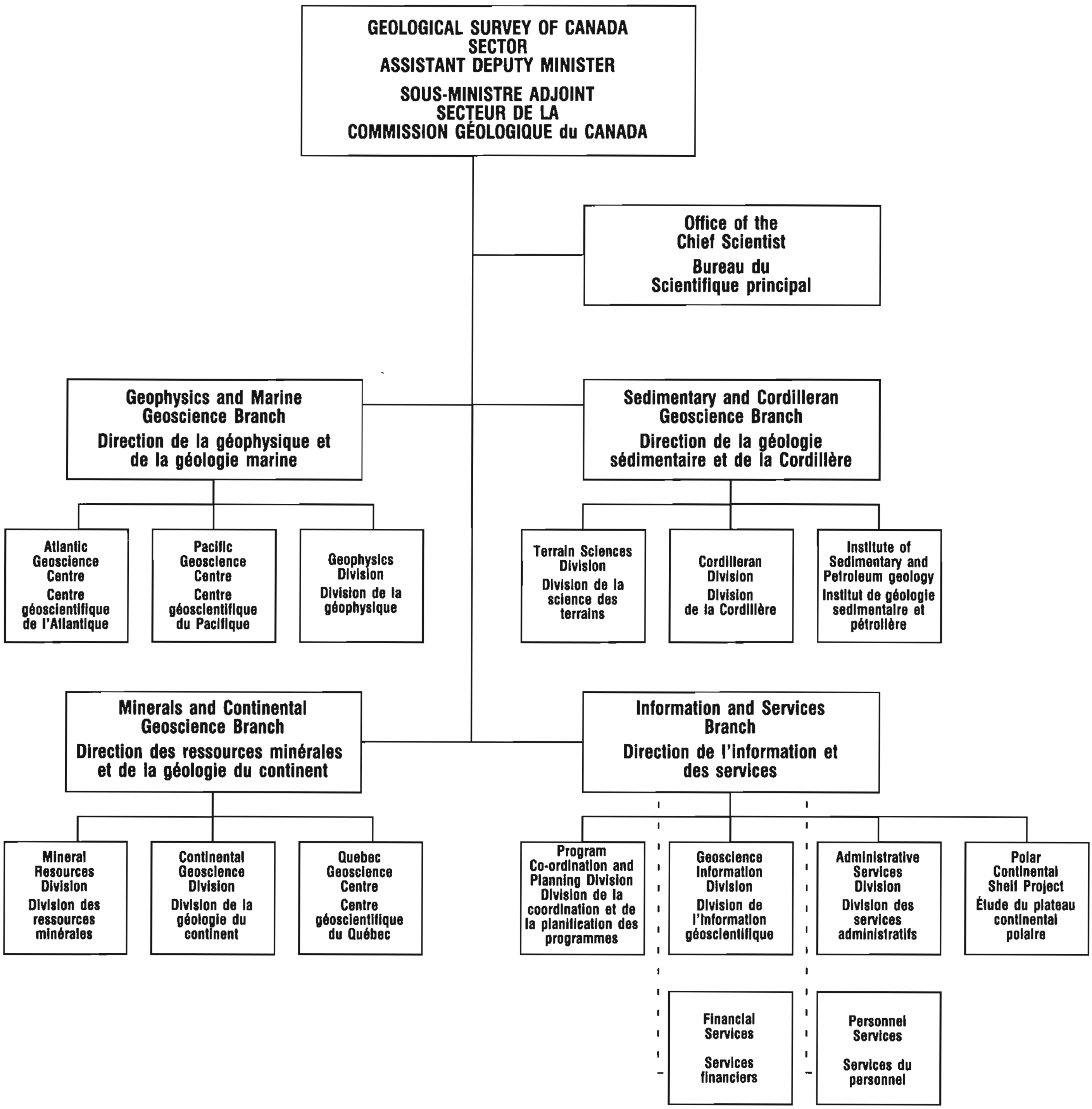
Rotated 'swell' in a heterogeneously extended mafic dyke in mylonite flanked by pegmatitic pressure shadows, observed in the plane perpendicular to the mylonitic foliation and parallel to the extension lineation. The orientations of the 'swell' and the pressure shadows indicate dextral shear. Great Slave Lake shear zone, N.W.T. Photo by Simon Hanmer.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The Grenville Province to the east of Val d'Or, Quebec: a geological reconnaissance and a possible extension of the Abitibi greenstone belt in the Grenville parautochthonous belt

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Abstract

East of Val d'Or, gneisses with a dominant mafic and a widespread, minor ultramafic component can be followed into the Grenville Province some 80 km eastward from the Grenville Front. These gneisses are oriented east-west. This orientation may be inherited from Archean structures. Other large areas are underlain by gneisses of intermediate to felsic composition with widespread subordinate portions of complexly deformed and highly metamorphosed mafic and supracrustal rocks. The Grenville Front can be observed in outcrop to juxtapose rocks of profoundly different metamorphic and structural history across a braided mylonite zone.

The mafic gneisses may be the highly metamorphosed extension of the Abitibi greenstone belt within the Grenville paratochthonous belt. If further research supports this hypothesis, a large area can be considered of interest for mineral exploration.

Résumé

À l'Est de Val d'Or, des gneiss à composantes mafiques et ultramafiques ont été observés dans la province géologique de Grenville sur une distance de 80 km en direction est, à partir du front de Grenville. Ces gneiss sont orientés est-ouest, signe qu'il puisse s'agir d'un héritage des éléments structuraux archéens. D'autres secteurs importants sont occupés par des gneiss de composition intermédiaire à felsique. On retrouve également dans ces secteurs, en moindres proportions, des gneiss mafiques et des roches supracrustales à hauts degrés de métamorphisme qui présentent des déformations complexes. Les roches du front de Grenville forment une zone de mylonite anastomosée, juxtaposant des roches caractérisées par des empreintes métamorphiques et structurales distinctes.

Les gneiss mafiques représentent possiblement l'extension métamorphisée de la ceinture de l'Abitibi dans la ceinture parautochtone de Grenville. Dans la mesure où d'autres études pourraient confirmer cette hypothèse, cette région présente un intérêt particulier au niveau de l'exploration minérale.

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INTRODUCTION

Within the last few years, ideas concerning the Grenville Geological Province have evolved from the position that the rocks of the Grenville are exotic with respect to the neighboring Archean rocks, to the position of, for example, Wynne-Edwards et al. (1966) that portions of the Grenville consist of re-worked Archean or other older rocks (e.g. summary by Moore, 1986). The present study was conducted based on two ideas. First, the Archean Abitibi greenstone belt, which is truncated by the Grenville Front, hosts productive mineral deposits along its entire east-west length. Second, the rocks near the Grenville Front have been shown in general to be parautochthonous. The area east of the Abitbi belt, however, had not been previously mapped. So the question arises "Can we identify portions of the former extension of the Abitibi belt within the Grenville geological province?" If portions of the Abitibi belt can be identified within the Grenville Province, how can we evaluate the mineral potential of such areas?

This paper offers a brief summary of recent field investigations in the previously unmapped area to the east of Val d'Or, Quebec.

LOCATION AND ACCESS

The location and boundaries of the study area are illustrated in Figure 1. The western side of the study area is located some 50 km to the east of the city of Val d'Or, Quebec. Road access to the region is adequate for very broad reconnaissance-style field investigations, but would not presently support more detailed work except in areas of active or recent logging. In the northeast, access is still too limited for vehicle-supported field studies.

PREVIOUS WORK

Most of the area under study has no record of previous geological mapping, mineral exploration, regional geochemical or geophysical study. Figure 2 illustrates the available geological maps in and adjacent to the study area. The exceptions to the general observations on the lack of mineral exploation are the limited work carried out in the Echouani area by Camchib Resources Inc. and reported in summary form by Sethuraman (1984), and the geological reconnaissance carried out in part of the area by Birkett (1981) for Shell Canada Resources Ltd. Compilation by Avramtchev and LeBel-Drolet (1981) indicated two occurrences of copper mineralization in the southern part of the study area (NTS 31N/16). Ground traverses in the areas of these reported occurrences were unsuccessful in locating outcrop, let alone mineralization. In the east, regions mapped by Wynne-Edwards et al. (1966) and Laurin (1965) have been re-examined to allow some basis for extending from known into unknown areas. The area near the Grenville Front has been examined by Bell (1932) and by Lacoste et al. (1987). Many of the outcrops in the study area were created by relatively recent road construction to support logging activities, and thus were not available for examination by earlier workers.

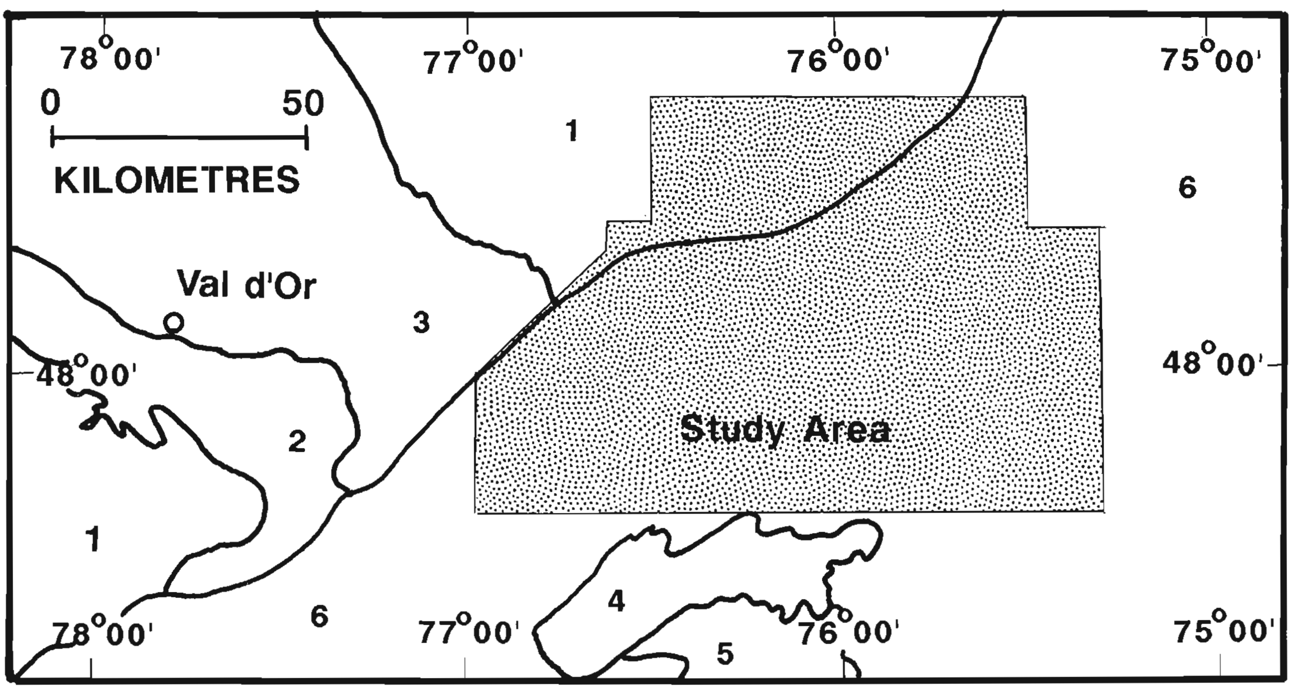


Figure 1. Location map of the study area with the city of Val d'Or indicated as a nearby geographical reference. **1** Granitoid and other rocks (undivided) of the Superior Province; **2** Metasedimentary rocks of the Pontiac Subprovince; **3** Metavolcanic and associated rocks of the Abitibi greenstone belt; **4** Bouchette anorthosite; **5** Réservoir Cabonga Terrane of Indares and Martignole (1990); **6** Rocks of the Grenville Province (undivided).

PRESENT STUDY

Limitations — caveat emptor

The results of this work and of future prospecting are limited by the lack of access in some areas, and by the general lack of outcrop in the study area as a whole. Aside from sporadic shoreline exposures, almost all bedrock examined in the course of this study was uncovered by logging or road-building activites. Away from these areas, a continuous, if generally thin layer of sand and gravel covers the bedrock surface. Locally eskers provide thicker deposits of overburden. In general, outcrop is considerably less than 1% in the study area. The outcrops found are generally flat or of low relief, with local remnants of glacial polish. Acquisition of both structural data and samples was generally difficult.

Methodology

Data from the field investigations were compiled at the scale of 1:50 000 for eventual publication at 1:250 000. The reconnaissance-scale geological map presented as Figure 3 has been drawn using the field observations with much reliance on available aeromagnetic maps. Minor structures such as small folds and mineral lineations have been used to project the units across areas of limited outcrop, with the realization that in a region of superposed folding and without a stratigraphic context such exercises are imprecise.

Results

An interpretive map of the large-scale features of the area is given in Figure 4. Several features are noteworthy.

The rocks of the Grenville Province in this area were metamorphosed under conditions of the amphibolite or granulite facies and are strongly deformed.

The Archean autochthon to the northwest of the Grenville Front contains areas of amphibolite to granulite facies metamorphic rock, as well as regions of supracrustal rocks of medium metamorphic grade. Northwest of the Grenville Front, within a distance of less than 1 km from the Front, the Archean metamorphic rocks locally show the effects of retrograde metamorphism, as well as late structural events marked by small shear zones and associated potassic alteration.

The Grenville Front is exposed in the west-central portion of the study area as a profound tectonic structure juxtaposing greenschist and granulite facies metamorphic rocks across a braided mylonite zone as much as 2 km wide. Progressive development of mylonite from various protoliths can be observed. The last event interpreted in the area is the local development of pseudotachylite in the mylonites. Structural studies indicate a complex history of movement within this mylonite zone, dominantly uplift of the Grenvillian side of the structure.

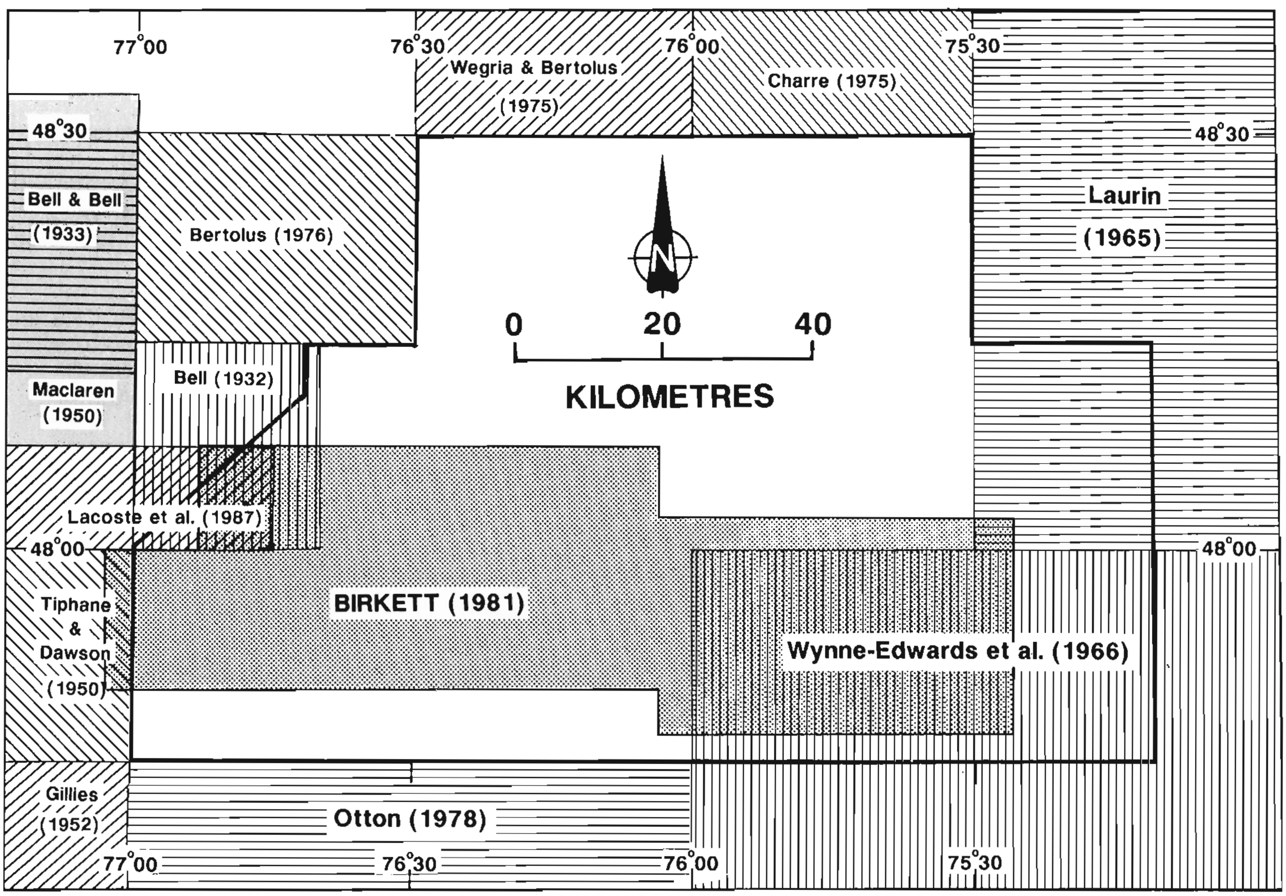
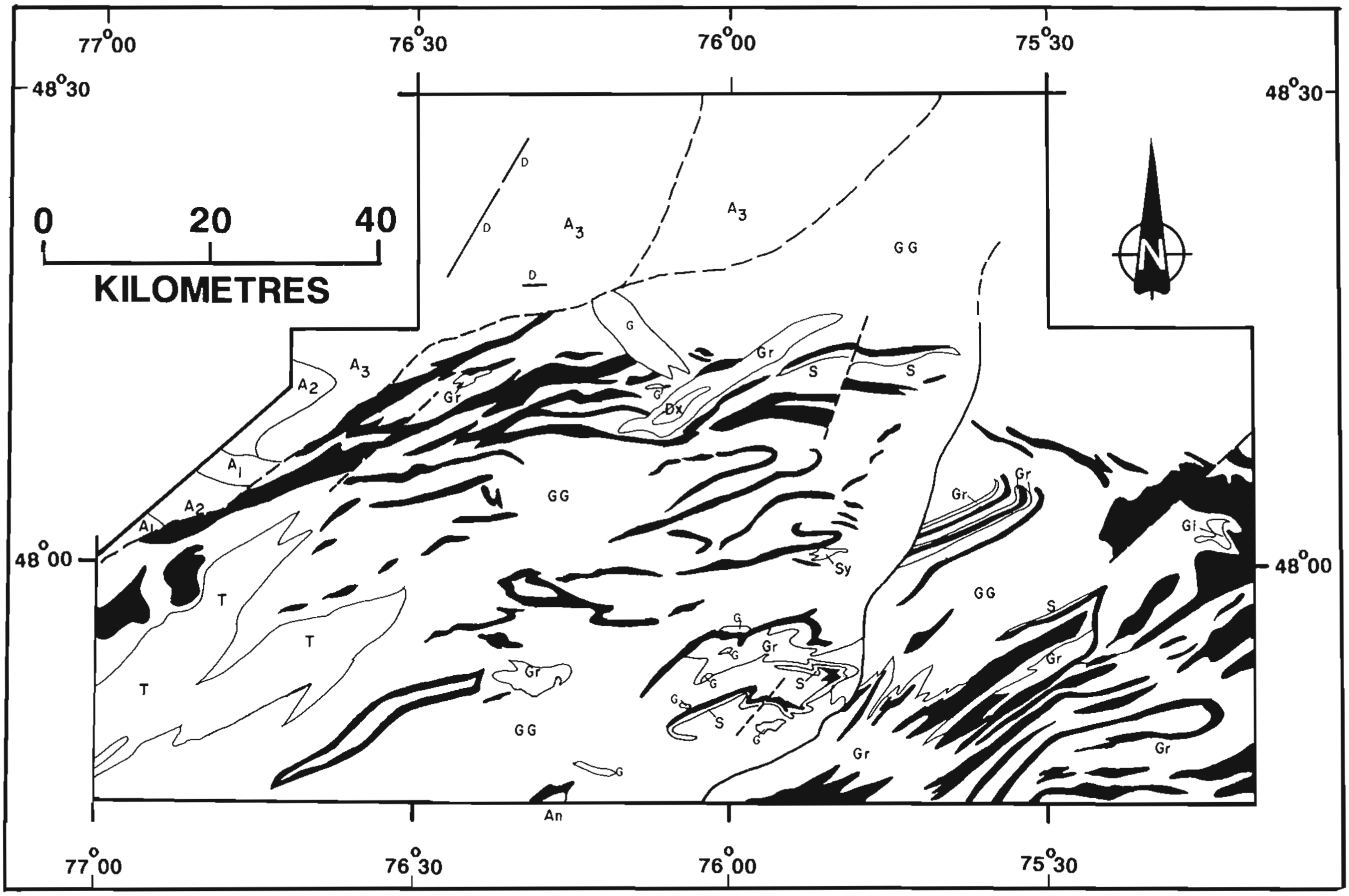


Figure 2. Previous geological mapping in and adjacent to the study area (heavy outline).



LEGEND

G	GABBRO syntectonic, massive to foliated, coronitic	S	SUPRACRUSTAL ROCKS siliceous and aluminous gneiss, iron formation
An	ANORTHOSITE	GG	GREY GNEISS of intermediate composition — feldspar, biotite, quartz ± garnet ± hornblende; includes a common minor component of mafic gneiss
Gi	GRANITE		
T	TONALITIC INTRUSIVE intermediate to felsic, deformed		
Dx	DIATEXITE massive to slightly foliated or gneissic granitic rock	D	DIABASE
Sy	SYENITE, metamorphosed and deformed	A_3	HIGH-GRADE METAMORPHIC ROCKS and granitic rocks of various compositions
■	GNEISS OF MAFIC COMPOSITION hornblende, plagioclase ± garnet ± diopside ± hypersthene; includes small areas of ultramafic rock	A_2	MEDIUM-GRADE METAMORPHOSED VOLCANIC ROCKS part of the Abitibi greenstone belt
Gr	GRANITIC GNEISS	A_1	MEDIUM GRADE METAMORPHOSED SEDIMENTARY ROCKS part of the Abitibi greenstone belt

Figure 3. A simplified lithological map of the study area.

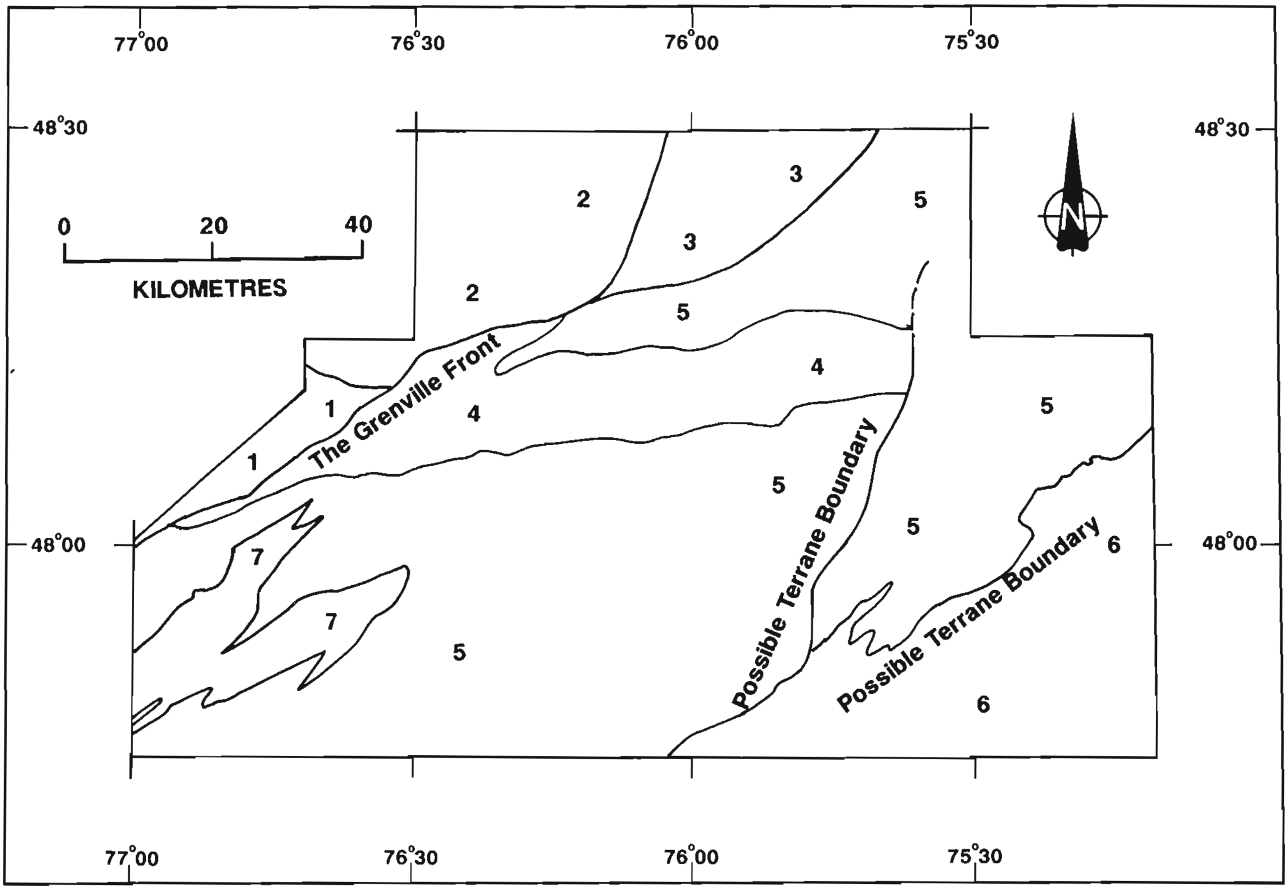


Figure 4. An interpretive map of the study area, indicating the major regions recognized to date. **1** medium-grade metasedimentary and metavolcanic rocks of the Abitibi greenstone belt; **2** intrusive and high-grade metamorphic rocks of generally intermediate to felsic composition and of Archean age; **3** high-grade metamorphic rocks with a major component of rocks of mafic composition, age of last metamorphism unknown (Archean or Grenvillian); **4** high-grade metamorphic rocks with a major component of rocks of mafic composition showing the effects of Grenvillian metamorphism and deformation; **5** high-grade metamorphic rocks dominated by gneisses of intermediate to felsic composition with a minor but widespread component of complexly deformed mafic rocks and supracrustal rocks, polymetamorphic, last metamorphosed in the Grenvillian event; **6** high-grade metamorphic rocks with subequal amounts of mafic and intermediate compositions, minor supracrustal rocks, possibly monocyclic, metamorphosed in the Grenvillian event; **7** intermediate to felsic intrusive rocks, deformed and metamorphosed in the granulite facies.

In the Archean autochthon, granitic and tonalitic rocks are exposed in the central portions of the area, and high-grade metavolcanic and associated supracrustal rocks in the east. It is assumed that these rocks are Archean in age, based on the prominent north-trending structural grain, although geochronological support for this hypothesis is presently lacking.

Near the Grenville Front, within the Grenville Province, gneisses of mafic composition are an important part of the lithological assemblage. These rocks display a general east-west orientation of lithologies and of the gneissosity. This general rock package can be followed some 80 km in an east-west direction. No primary structures allowing the assignment of a volcanic or sedimentary origin to the protoliths have been recognized. A small but significant proportion of the rocks in this area are of ultramafic composition. These are recognized in the field by the common presence of abundant orthopyroxene porphyroblasts. Chemically, the hypothesis of an igneous origin for these ultramafic rocks is supported by their tenors of Cr and Ni, which are in the range of hundreds to thousands of ppm.

In the eastern part of the study area, a zone of complex, poorly exposed geology separates two major rock packages, the east-west trending lithologies referred to above, and northeast-directed lithologies to the east. The zone is marked by distinct pattern changes in the airborne magnetic survey maps, and by overprinting of one gneissosity by another. This zone may be a terrane boundary. The area requires more study.

Rocks of intrusive aspect are not rare in the region under study, although areally subordinate. Most outcrops within the Grenville Province contain a mobilizate of local derivation now crystallized as coarse grained quartz-feldspar rich rock. These pegmatites form a few per cent of the rocks in the amphibolite facies, 15 per cent or more in the granulite facies. In the south and the north of the study area, small bodies of gabbro have been mapped. These rocks are syn- to late tectonic, locally massive and locally foliated (within single continuous outcrops) and commonly display corona textures indicative of reaction between plagioclase and olivine or pyroxene. A metamorphosed syenite of small extent has been identified in the central part of the area. Alkali granite in the eastern part of the region reported by Wynne-Edwards et al. (1966) has not yet been seen in the field. Diatexites have been seen in the north-central part of the study area. In the western portion of the study area, a substantial body of tonalitic rock is exposed. Its limits are not yet known adequately. This body may be of Grenvillian age, as it is less deformed than the large majority of rocks in this region. Within the Superior Province, granitic and tonalitic rocks dominate in the northwestern part of the study area, to the north of the Abitibi greenstone belt.

DISCUSSION AND CONCLUSIONS

Geology

The assignment of some of the mafic rocks of the study area to the metamorphosed extension of the Abitibi greenstone belt is based on the juxtaposition of the western portion of these rocks with the eastern portion of the Abitibi belt within the Superior Province. Unless this juxtaposition is fortuitous, there is thus no evidence for major lateral offset across the Grenville Front in the study area. Instead, the Grenville Front marks a zone of pronounced uplift of rocks which were probably tectonically buried earlier in the Grenvillian orogeny.

The general description of the relations across the Grenville Front parallels that of Ciesielski (1988) from the Chibougamau area.

The Grenville Front, in the western part of the study area, has juxtaposed rocks of similar composition but differing metamorphic histories. The main effect of the Grenvillian events on the rocks to the west of the Front has been limited retrogressive metamorphism and brittle strurctural development. Major domains can be outlined within the Grenville Province, based on proportions of rock types, structural trends and metamorphic grade and history. In particular, major packages of rock with an important component of mafic composition and a persistent component of ultramafic composition can be followed some 80 km into the Grenville Province.

Metallogeny and implications for mineral exploration

The only previous significant mineral exploration reported work in this area was by Sethuraman (1984). After area selection, an exploration program of airborne surveys with ground follow-up and diamond drilling was carried out. Minor base metal (Cu) and precious metal (Au) mineralization was encountered over narrow widths. The interpretation concerning the geological context of the mineralization provided by Sethuraman (1984) was that of a mafic-felsic contact in a setting of metamorphosed volcanic rocks, with intercalated chemical sediments (sulphide and oxide facies iron formation).

Although mineral deposit models appropriate to Archean greenstone belts were assumed in the area Sethuraman (1984) reported on, the terrane analysis of Indares and Martignole (1990), extended to the northeast here, suggests that the work discussed by Sethuraman (1984) might have been mostly or entirely in rocks of the monocyclic metamorphic terranes, possibly of Proterozoic age. Indares and Martignole (1990) assigned the Bouchette anorthosite and the Réservoir Cabonga Terrane to the monocyclic belt (Fig. 1). It seems that the rocks possibly of original Archean age, in the west of the present study area, have not yet been explored.

The rocks of mafic composition indicated as unit 4 in Figure 4 underlie some 1200 km^2 of the study area. These rocks may be of original Archean age, may be a metamorphosed extension of the Abitibi belt, and are, from available data, unexplored. Other areas of supracrustal rocks may also be prospective for base or precious metals.

The lack of outcrop renders traditional prospecting methods of little value in this area. The generally transported nature of the post-glacial deposits and the high proportion of sand in the overburden lead to the suggestion that direct regional geochemical study may not be an appropriate tool here for mineral exploration. The unconsolidated sediments show a general reverse grading from sand in the lower areas to gravels on the tops of the lower hills in the region. This sequence suggests that most of these materials are reworked shore deposits, with little immediate application for mineral exploration. On the higher hills, above the level of post-glacial lakes, sandy, blocky tills may still be locally present. The overburden, although nearly continuous and of varying thickness, is, however, generally poor in clay and geophysically transparent. Given the almost complete lack of outcrop, airborne geophysical methods may be the most appropriate tool for mineral exploration once intitial area selection has been carried out.

FUTURE STUDIES

Geological investigations in this area will now be directed to testing the hypothesis that the Abitibi belt can be identified and followed in the Grenville Province. The methods envisaged include (1) improving the lithological and structural map, (2) studies of the metamorphic histories of the different regions to constrain theories of various terranes, (3) isotopic age dating better to elucidate the development of the region, (4) isotopic tracer and chemical studies to test correlations across the Grenville Front, and (5) structural analysis to allow more confidence in the interpretation of local and regional geological patterns.

Nature hints at secrets here, but scarcely reveals them. In the area we have examined, the topography reflects poorly the geology, and the sands of postglacial Lake Ojibway-Barlow blanket an area as complex as it is unknown. Yet there is hope, not only of scientific knowledge, but of practical reward from further study.

ACKNOWLEDGMENTS

We are indebted to Pierre Brouillette, Michel Dion, Sylvie Lévesque, Dan Richardson and Maxime Tellier for good-natured assistance in the field at various times in the course of this work. Denis Racicot (MERQ, Val d'Or) helped with searching the database for previous geoscientific studies in the area. Serge Paradis and Michel Parent (GSC) provided informed discussion of the Quaternary geology and its potential for applications in mineral exploration. James Moorhead and Jean-François Couture (MERQ) offered stimulating discussion of structural problems in the field. Léopold Nadeau and Louise Corriveau (GSC) and Alain Simard and James Moorhead (MERQ) have provided constructive reviews of earlier versions of this work. Their comments have improved the presentation of ideas.

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The first surface rupture from an earthquake in eastern North America

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Adams, J., Wetmiller, R.J., Drysdale, J., and Hasegawa, H.S., The first surface rupture from an earthquake in eastern North America; in Current Research, Part C, Geological Survey of Canada, Paper 9-1C, p. 9-15, 1991.

Abstract

For the first time, an earthquake in eastern North America is confirmed to have produced surface faulting. The rupture was found as a result of locating aftershocks during a field survey (in July 1990) of the magnitude (M_S) 6.3 earthquake of 25 December 1989 in northern Quebec. It is centred at 60.12°N, 73.60°W and extends 8.5 km along an average trend of 038° (concave to the northwest). The fault appears to dip steeply to the southeast and the sense of faulting is reverse with the southeast side upthrown. The maximum throw is 1.8 m and tapers to less than 0.3 m at each end. Surface effects include: deformed lake shorelines, discontinuous fault scarps, a left-lateral strike-slip fault, torn muskeg above some traces, sand volcanoes, freshly cracked boulders, and a partly-drained lake. Two lakes were discoloured by silt.

Résumé

Pour la première fois, dans l'est de l'Amérique du Nord, il a été confirmé qu'un séisme avait généré des failles en surface. On a découvert cette rupture du sol durant une étude sur le terrain (juillet 1990) entreprise pour situer les répliques sismiques du séisme de magnitude (M_S) 6,3 survenu le 25 décembre 1989 dans le nord du Québec. La rupture est centrée à 60,12°N et 73,60°O et s'étend sur 8,5 km selon une direction moyenne de 038° (concave au nord-ouest). La faille semble s'incliner fortement vers le sud-est, et les failles sont de type inverse, le compartiment sud-est étant soulevé. Le rejet vertical maximum est de 1,8 m et s'amincit à moins de 0,3 m à chaque extrémité. Les effets observés en surface sont en particulier: les rivages lacustres déformés, les escarpements de faille discontinus, un décrochement latéral senestre, le déchirement du muskeg au-dessus de quelques lignes de faille, des injections de sable, des blocs fraîchement fissurés, et un lac partiellement drainé. Deux lacs étaient décolorés par le silt en suspension.

INTRODUCTION

The Geological Survey of Canada (GSC) originally located the epicentre of the Ungava Earthquake at 60.02°N 73.63°W ± 20 km in the middle of the unpopulated Ungava Peninsula of northern Quebec (Fig. 1A), and concluded that the focal depth was shallow. From the position of the midpoint of the fault rupture, the epicentre should be revised to 60.12°N 73.60°W ± 5 km with the uncertainty representing whether the rupture initiated at the middle or at one end of the fault.

The mainshock was preceded by a $m_{bLg} = 5.1$ (GSC) foreshock 10 hours earlier (04:25 UT). There were 5 earthquakes of $M > 3$ between the foreshock and the mainshock, and at least 8 $M > 3$ aftershocks in the two weeks after the mainshock, the largest being $m_{bLg} = 4.4$. There have been very few subsequent aftershocks detected by the Canadian Seismograph Network.

The mainshock was felt quite strongly in Kuujjuaq (Fort Chimo) at 360 km, but not felt at Iqaluit (Frobisher Bay) at 500 km. Although not well established, it is likely that the felt area was considerably smaller than for the $m_b = 5.9$, $m_{bLg} = 6.5$ Saguenay Earthquake of 1988 (North et al., 1989; Drysdale, 1990). Other magnitudes for the Ungava Earthquake are: $m_b = 6.2$ (from the U.S. Geological Survey National Earthquake Information Service), and $m_{bLg} = 6.1$ (GSC). The seismic moment of

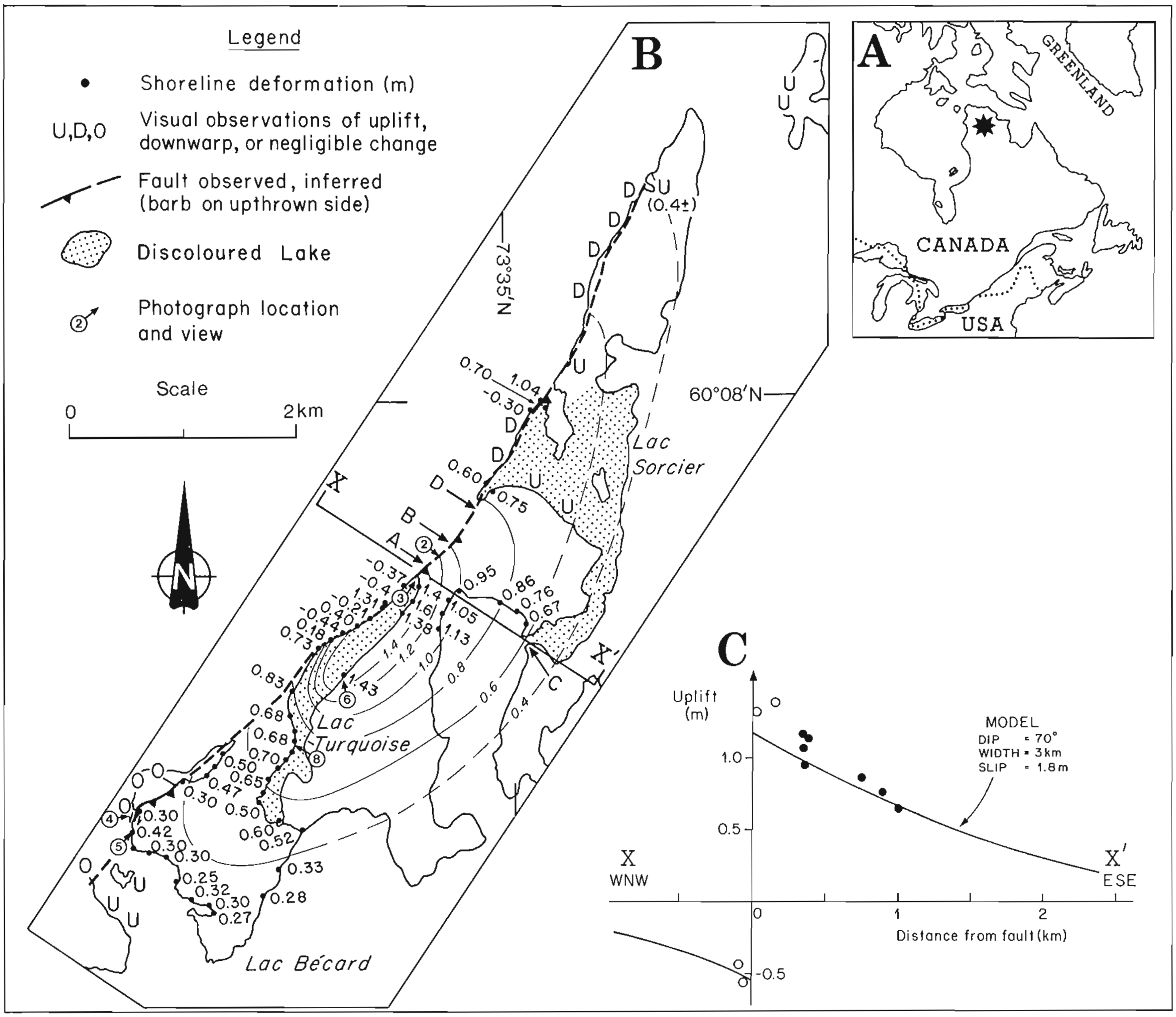

Figure 1. A) Top right: Location map. B) Map showing location of the surface rupture, site of shoreline deformation measurements (not all dots are labelled), sites for which visual estimates were made from helicopter (U, D, 0), and the approximate contours of regional deformation inferred (fine and dashed lines). The two discoloured lakes (stipple) and the location and view direction of figures 2 — 6 and 8 are also shown. C) Profile X — X' from Figure 1B showing the deformation expected for a reverse fault with a 70° dip, 3 km width, and 1.8 m throw (solid line), and the observed shoreline deformation on lakes Bécard (solid dots) and Turquoise (open circles).

1.1×10^{18} N-m (National Earthquake Information Service) gives a moment magnitude of 6.0. The felt area of the Ungava relative to the Saguenay earthquake is consistent with their relative m_{bLg}'s, but not with their relative teleseismic magnitudes.

The epicentral area is typical of the ~2 Ga old Superior Province of the Canadian Shield. It lies within the area of continuous permafrost in northern Canada and is north of the tree-line. The gneiss bedrock has been heavily glaciated and attains a maximum elevation of 250 m with local relief of only 70 m. The ground surface generally consists of muskeg, boulders, till, or outcrop. About 50% of the region is covered by water, with a large lake, Lac Bécard, originally surrounding the epicentre on three sides; the presence of these lakes is fortuitous as it makes possible much of our analysis.

Temperatures of −35°C, snow cover, short daylight (<3 hr/day), and remoteness precluded an immediate field study to record aftershocks. Instead, the GSC carried out an aftershock survey from 7-25 July 1990. During this survey the aftershock zone was located and then the surface faulting described below was discovered. At the time of discovery, the unnamed lake paralleling the fault scarp was coloured turquoise, distinct from all other lakes in the region. The colouration was due to a large amount of silt suspended in the water and was an indirect effect of the earthquake. Thus the fault scarp has been named "The Lac Turquoise Fault Scarp". Lac Turquoise and Lac Sorcier are informal geographic names. Lac Sorcier ('Wizard') is named for the shape of the wizard's hat.

The observations reported in this paper were made during four visits by helicopter during routine servicing of the field seismographs, with a total of only 17 hours on the ground. The ground examination was supplemented by direct observation from the helicopter and by subsequent examination of extensive videotape.

SURFACE FAULTING

Unlike many modern fault scarps, the Lac Turquoise Fault Scarp did not often present itself as a vertical free face, most probably because the bouldery muskeg and till hindered direct propagation of the fault to the surface. The line of the fault was easily seen, however, by the torn and buckled muskeg, which appeared to have failed in a brittle manner (it would have been frozen at the time of the earthquake). Some of the failure modes resemble those described by Gordon (1971). However we observed several places (e.g. A and B in Fig. 1B) where a scarp could be measured (Fig. 2), and many places where large differences in shoreline elevation occurred within a short distance (about 10 m; Fig. 3). In all cases the upthrown block was found to be on the southeast side of the fault.

Left-lateral faulting

Towards the southern end of the main rupture, a surface crack striking 174° through the bouldery muskeg was observed to have systematic left-lateral offset of 110 mm (4 observations, e.g. Fig. 4). This was the only place we saw faulting across bedrock. At one point along this rupture, thrusting on a 70° plane dipping east produced a throw of 65 mm. We consider the crack to be a rupture associated with the main fault because: i) The displacements were systematic along the crack and represented predominantly lateral movement, while superficial ground cracking due to subsidence would probably have produced uneven widening. ii) At the time of the earthquake (25 December) the muskeg was probably frozen solid, and so could be expected to fail in a brittle manner despite its peaty appearance. iii) The left-lateral fault makes an angle of 106° with the reverse fault at their intersection. After including the small amount of convergence across the left-lateral fault, its slip direction is within 4° of being normal to the reverse fault, suggesting a clear structural relationship. iv) We observed clear uplift of the shoreline on video images of the south side of the bay (as shown in Fig. 1B). The observed left-lateral fault has the correct sense to transfer reverse motion to this segment.

The left-lateral slip of 110 mm combined with differential uplift of 300 mm (Fig. 1B) on the reverse fault yields

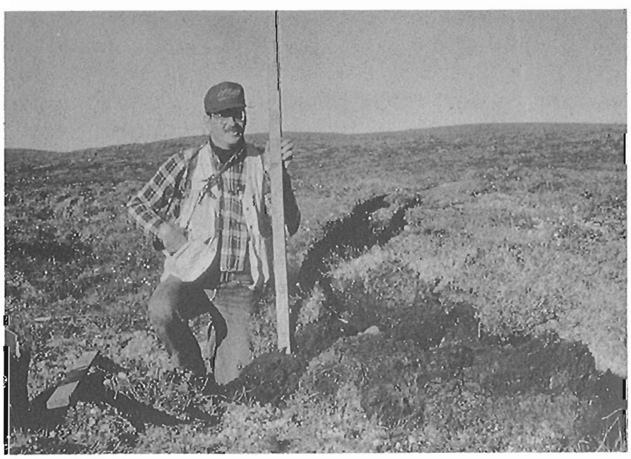

Figure 2. Fault scarp in peat, about 0.6 m high, in the saddle between Lac Turquoise and Lac Sorcier. Scarp trends into the distance behind and left of the man. Some collapse of the scarp has apparently occurred.

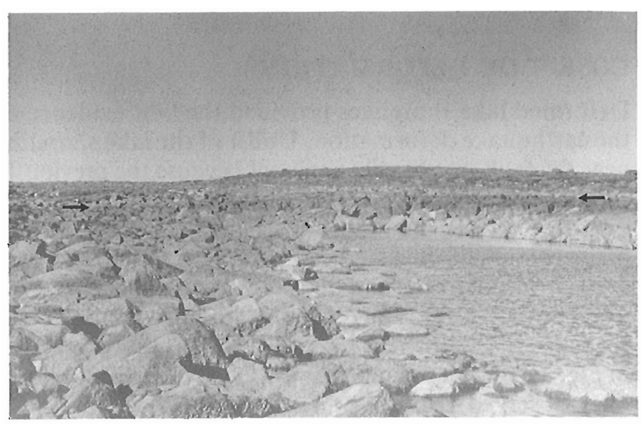

Figure 3. Deformation of old lake shoreline at the north end of Lac Turquoise by the fault scarp. Note the downwarping of the dark, stained band, which marks the rock just below the old lake level (arrows). The unstained rock below it is visible only to the right (east) of the fault.

Figure 4. Fault offset of 110 mm between a boulder and formerly-adjacent peat along the left-lateral fault near the south end of the rupture.

a dip of 70° for the main fault. While this value need not apply to the remainder of the fault, it suggests that the main rupture dips steeply.

SHAPE OF THE FAULT

The primary feature of the Lac Turquoise scarp is its concave plan; concave to the northwest away from the upthrown block. The sense of curvature in relation to the upthrown block is the opposite to that found for the Australian surface faults at Meckering (Gordon, 1971; Gordon and Wellman, 1971) and Tennant Creek (Choy and Bowman, 1990). In detail, the trace of the fault appears to follow the shorelines of lakes Bécard, Turquoise and Sorcier quite closely, at places deviating through angles of up to 25° from the smooth concave curve. We infer from this that the rupture was controlled by a pre-existing fault in the bedrock structure, the fault that also controlled the glacial erosion which produced the present topography and hence lake shores. At this time we have no evidence whether or not the Lac Turquoise scarp represents the only surface rupture at this site since the deglaciation.

COSEISMIC DEFORMATION

Deformed lake shorelines provided the best evidence of the earthquake deformation. Uplift of the lake shoreline near the fault was readily apparent because the previously submerged rocks were stained black and coated with reddish algae rather than moss and lichen. The reddish and black colours of the emerged rocks were very distinctive, even from the air, and allowed a quick estimate of uplifted shorelines. The position and throw of the northern 2 km and southern 1 km of the fault rupture (mapped in Fig. 1B) are based on clear evidence of uplifted shorelines observed from the air, as time did not permit a ground examination.

On single large boulders, such as are very common along the shorelines of all the lakes, the colour contrast allowed the level of the old lake to be determined to within a few centimetres. Figure 5 shows a typical boulder on the uplifted shoreline of lac Bécard. When making the shoreline measurements relative to the current lake level, it was seldom possible to make a direct measurement as in Figure 5. Instead, short leveling surveys were performed (Fig. 6).

The method involved an Abney hand level, resting on a 1.63 m staff, and sighting to a 1.8 m rod. Where the shoreline was uplifted more than 1.63 m, a second levelling was made from an intermediate point. For the larger shoreline displacements, the sight lengths were sometimes as much as 10 m, but normally a careful choice of profile limited the sight lengths to less than 5 m. The Abney level was checked after the field experiment and found to have essentially negligible bias, when set at zero. Reading precision under office conditions (including setting the zero) amounted to 16 mm (1σ) over a sight length of 15 m.

Figure 5. Shoreline uplift of 0.42 m on Lac Bécard near the southern end of the Lac Turquoise fault scarp. Pre-earthquake water level on this large boulder is indicated by black arrow, July 1990 water level by the white arrow.

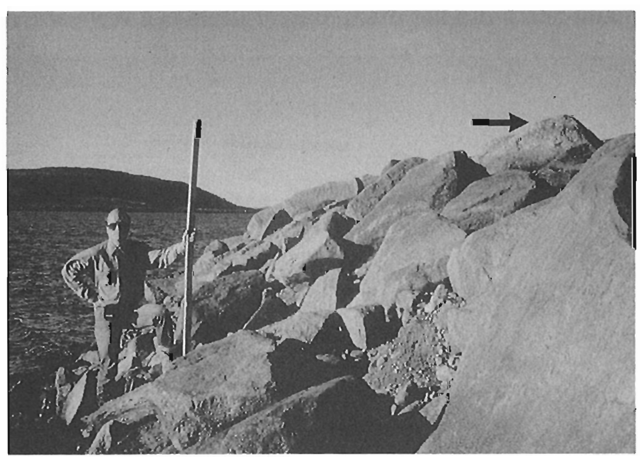

Figure 6. Shoreline change of −2.6 m on Lac Turquoise. Pre-earthquake water level is shown by the horizontal arrow. Of the 2.6 m, about 1.2 m represents drainage of the lake (see text) and 1.4 m is coseismic uplift. The fault rupture follows the far (northwest) side of Lac Turquoise.

Boulder movement by ice push in winter was recognized and avoided. Special care was taken to choose large boulders with clear colour banding, and as close horizontally to the lake as possible, so as to minimise the errors involved in longer sight lengths. All measurements are relative to water levels on 20-22 July. Repeated measurements at a few sites, and consistency of levels along a few kilometres of shoreline, suggest that the measurements are precise to about 0.1 m for the larger shoreline displacements (> 1 m) and 0.05 m for the smaller ones (< 0.5 m).

On a single lake, such elevation measurements provide a direct measure of relative uplift. Because Lac Bécard extends for 20 km to the south and west of the epicentre, it is a large lake relative to the extent of the coseismic deformation and so the shoreline uplift near the epicentre represents very closely the absolute elevation change relative to the rest of the earth.

The deformed region involved three lakes (Bécard, Sorcier, and Turquoise), and so their shoreline measurements must be tied together. From aerial photographs flown in July 1955 which show shallow water in the strait between Bécard and Sorcier (Fig 1B, arrow C) and the elevation of the old shoreline on Lac Bécard we believe Lac Sorcier was sometimes a bay of Lac Bécard. We have little information on the seasonal variation of lake levels in this region, though a second set of aerial photographs flown in August 1957 shows the strait to be dry. This suggests variation of >0.4 m and that the separation of the two lakes may have been a seasonal phenomenon. The 1989 Ungava Earthquake uplifted the strait 0.67 m (Fig. 1B), perhaps sufficient that Lac Sorcier will never again become a bay of Lac Bécard. Irrespective of the seasonal variation, the clear colour banding of the raised shoreline suggests that the former high water level was a stable reference level, and in any event the relative shoreline change measurements used here are unaffected by seasonal level variations.

On 22 July, the water level on Lac Sorcier was 0.25 m higher than on Lac Bécard as measured by Abney level over a 65 m sight length. Thus for Lac Sorcier the new water level is 0.25 m higher than it would have been if still connected to Lac Bécard. Therefore measured shorelines elevations on Lac Sorcier need to have their displacements increased by 0.25 m to convert them to elevations relative to Lac Bécard. The water level increase also accounts for the unusual amount of apparent subsidence (Fig. 1B) along the northwestern shore of Lac Sorcier (downthrown side of the fault) and the apparent absence of uplift along the eastern shore (upthrown side).

Lac Turquoise was always higher than Lac Bécard. In July the water level of Lac Turquoise was 1.8 m below its outlet into Lac Bécard. This was partly due to coseismic uplift and partly to drainage. Because the earthquake occurred in winter, we do not know the lake level immediately after the earthquake. We suspect that the lake drained through surficial deposits at the north end of the lake, and emerged in an area of extensive sand volcanoes (see below) near Lac Sorcier. From a mean shoreline elevation of 2.2 m above current water level of Lac Turquoise and a lake area of about 5×10^5 m^2, the volume of water that drained away was at least 1×10^6 m^3.

The lowest shoreline measured on Lac Turquoise was on the downthrown side of the fault and was 0.65 m above the July water level. This represented the minimum amount of lake drainage and supposes that the uplifted block was uplifted relative to a stable "downthrown" block. We estimate the amount of net drainage in July in two ways. Firstly, if, as is shown later by our ground deformation modelling program, one block was uplifted by about two-thirds the fault offset and the other downthrown by one-third, then from the highest (2.6 m) and lowest (0.65 m) shoreline measurements the mean level of the lake before drainage was 0.65 + (2.6 — 0.65)/3 m or 1.3 m above the current lake level. Secondly the shoreline elevation changes of 1.8 m on Lac Turquoise at the outlet and 0.5 m on nearby Lac Bécard suggest about 1.2 m, after allowing for 0.1 m of extra uplift for the outlet which is 200 m closer to the fault.

With a correction of − 1.2 m for Lac Turquoise and + 0.25 m for Lac Sorcier, the shoreline changes on all three lakes can be analyzed together.

The pattern that was derived is rather simple, an elongate half dome of uplift on the upthrown side (Fig. 1B), and localized observations of subsidence on the downthrown side (where little time was spent in the field). From the contours, the maximum uplift exceeds 1.4 m and the maximum subsidence is 0.55 m. The largest elevation changes are restricted to the central 1 km of fault and the uplift tapers to 0.3 m over 3 km to the south and 0.4 m over 4 km to the north.

We made a few direct measurements of scarp height and these and the fault throw inferred from the shoreline elevation near the fault are shown as a profile along the fault in Figure 7. As is evident, the maximum throw is restricted to a short portion of the fault. The mean uplift averaged along the length is 0.8 m.

Deformation modelling results

We used the Mansinha and Smylie (1971) formulation to compute the expected vertical deformation from the earthquake. In plan view, the concave surface fault was modelled by three 3-km fault segments of strike 048°, 038°, and 028° and throws of 0.7, 1.8, and 0.7 m. Fault dip was taken to be 70° to the southeast, and fault width to vary over 3, 4, 5, and 6 km. For a 3-km fault width

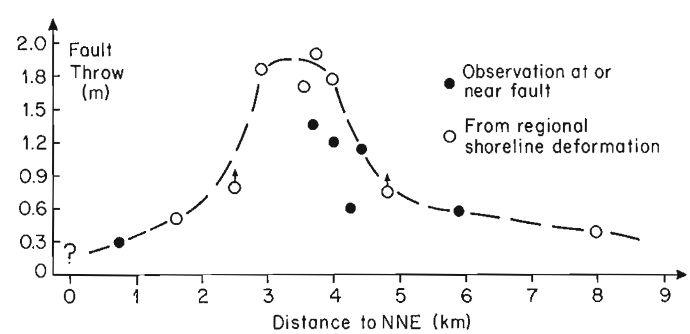

Figure 7. Profile along the Lac Turquoise fault scarp, showing measured or inferred throw across the fault and the throw implied by the maximum difference in shore elevation across Lac Turquoise. Dashed envelope line suggests the trend.

we also tried fault dips of 60°, 70°, 80°, and 90°. We note from the models (which are not shown) that the ratio of uplift to subsidence is very sensitive to fault dip and that the far-field uplift (for which we have no measurements yet) is sensitive to fault width. For dips less than 80° there is a preponderance of uplift, as was observed.

While the fault parameters were derived from the observed shoreline offsets at the fault, the regional deformation can be matched to the independent observations of shoreline uplift away from the fault. A NW-SE profile (X-X' in Fig. 1B) from our fault model, with a 3-km fault width and a 70° dip, best matches shoreline tilt on Lac Bécard between Lac Turquoise and the outlet to Lac Sorcier (Fig. 1C).

OTHER EFFECTS

Liquefaction Features

Liquefaction features were evident but not numerous. We infer that the frozen ground in December, and the presence of continuous permafrost, severely limited the amount of water available for expulsion. We did not observe many fine grained surficial sediments likely to liquefy.

Sand volcanoes, the best defined liquefaction features, were found along the shores of Lac Turquoise, below the pre-earthquake lake level (Fig. 8). They were likely vented through newly exposed, unfrozen lake bottom soon after the earthquake uplifted and partly drained the lake. A few had ejected medium-fine sand, but the majority comprised very fine sand and silt. They ranged from 350 mm to 2 m in diameter.

Many liquefaction-like features were found in a restricted area between Lac Turquoise and Lac Sorcier (Fig. 1B, Arrow D). Some of these features resembled conventional sand volcanoes, having clogged themselves, but others had open centres suggesting passage of a large volume of water, although they were dry in July. Further evidence for a large flow was a "tidemark" of silt on boulders about 0.5 m above the general level of the ground. We hypothesize that Lac Turquoise drained underground to the north through the bouldery muskeg and that the silty water emerged in this area and flowed into Lac Sorcier, discolouring it. The large volume of silt required could have come from the bottom sediments of Lac Turquoise and the fine fraction of the sediment between lakes Turquoise and Sorcier. Both Lac Turquoise and the southern half of Lac Sorcier were discoloured at the time of our visit.

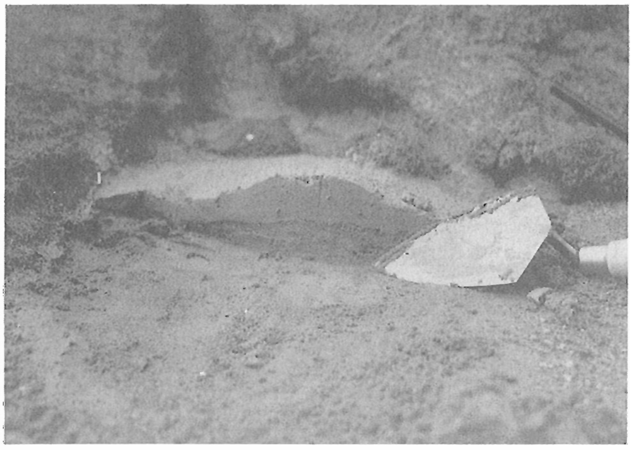

Figure 8. Cross-section through 350-mm diameter sand volcano on the shore of Lac Turquoise showing the clear contrast between the clean, grey ejected silty-sand and the oxidised, brown sand of the lake bottom.

In addition to the above, we observed a dried-up spring at the south end of Lac Turquoise a new, replacement spring about 3 m lower on the slope, and a second new spring just west of Lac Turquoise where copious water was flowing.

Slope failures

We saw only a few fallen rocks (i.e. rocks without lichen cover), even beneath steep slopes of shattered bedrock close to the fault where many would have been expected. We also did not see any evidence of rock-hopping or other evidence of vertical accelerations exceeding 1 g; both phenomena would have been inhibited by the ground being frozen at the time of the earthquake and by the snow cover. Although we saw isolated, freshly cracked boulders on the newly-exposed lakeshore (where they were very visible because of their fresh colour), we do not yet have any explanation for them. Some surficial sliding on moderate slopes near the fault trace may have been the result of the earthquake shaking, or may be a regular spring thaw phenomenon.

FAULT AND EARTHQUAKE PARAMETERS

From the above surface observations we conclude that the Ungava Earthquake ruptured a curved fault about 8.5 km long, that the sense of motion was reverse-slip with the southeast side up, that the surface throw was a maximum of 1.8 m and tapered towards each end, and that the fault is likely to dip steeply to the southeast but not likely to extend below 4 km depth. From a mean throw of 0.8 m (slip $\approx$ throw for dips $<60°$), a fault length of 8.5 km, a fault width of 3 — 5 km and a crustal rigidity of 3.3×10^{10} N-m^{-2}, we compute a seismic moment of $0.7 - 1.1 \times 10^{18}$ N-m. The published moment of 1.1×10^{18} N-m (from the Earthquake Data Reports of the USGS, and derived independently by USGS and Harvard from teleseismic wave analysis) is very similar, the good agreement indicating that the observed surface throw is sufficient to explain the seismic energy release.

The short-period P-nodal focal mechanism solution (Fig. 9) we determined for the mainshock is well-constrained by the data shown and represents reverse faulting in response to NNW compression. The two possible orientations for the fault plane at the start of the rupture are indicated by the solution's nodal planes, which strike E-W and dip south or NE-SW and dip northwest. Our near-field ground deformation analysis rules out the northwest-dipping plane, as thrust motion on this

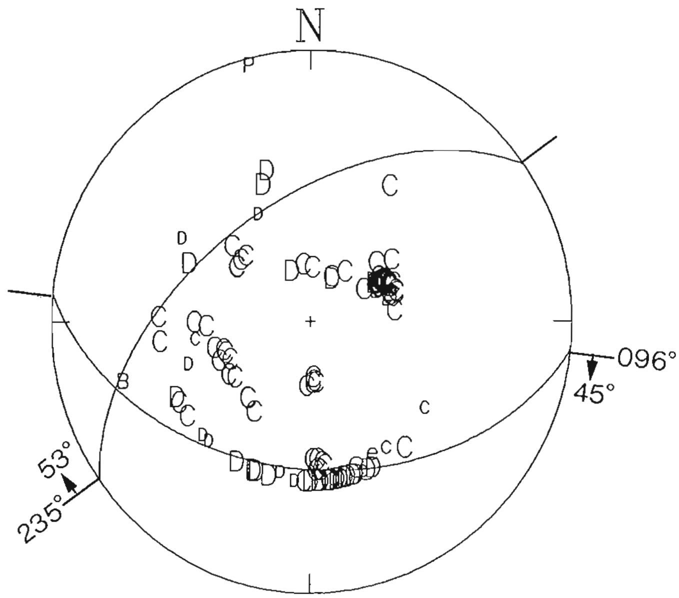

Figure 9. Short-period P-nodal mechanism for the Ungava Earthquake. The mechanism is a lower hemisphere projection. C and D are compression and dilatation; P is the pressure axis.

plane would produce subsidence to the southeast and uplift to the northwest of the fault, opposite to that observed. Moreover the complementary plane that dips to the south strikes about 40° clockwise of the surface rupture. The fact that neither our mechanism, nor the USGS or Harvard moment tensor solutions (not shown), agrees with the field observations of the strike and dip of the surface break is a problem that remains to be resolved.

CONCLUSIONS

It is too soon yet to assess how the first surface faulting from an earthquake in eastern North America will help our assessment of seismic hazard for the Canadian Shield. Although the present evidence of surface faulting and deformation from the Ungava Earthquake is plentiful, its preservation potential is poor. Thus prehistoric surface ruptures elsewhere in the shield might easily be overlooked. We suspect that, despite the indirect nature of the evidence, the silt deposited from suspension onto the lake bottoms (Doig, 1986) may provide the best long-term record of the 1989 Ungava or other large shield earthquakes.

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Duality of magmatism along the Kirkland Lake-Larder Lake fault zone, Ontario

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Abstract

Magmatism in the Kirkland Lake-Larder Lake fault zone (KLF) is characterized by compositional duality. Two suites, syenitic (quartz-free) and granitic (quartz-normative or quartz-bearing) rocks, are present within three magmatic domains. Domain 1, the fault zone, contains both suites. In Domain 2, immediately north of the KLF, only granitic rocks are observed. Domain 3, south of the KLF, contains both syenitic and granitic rocks. These occur either together as phases of composite intrusions or separately as discrete plutons. The Larder Lake fault (LLF), forming the south margin of the KLF, is the boundary between two levels of emplacement: rocks to the south are plutonic, whereas those to the north are hypabyssal or extrusive. This supports previous interpretations for relative upward movement of the crust south of the fault zone. The syenitic and granitic rocks in Domain 1, which host gold, share mineralogical and compositional similarities with the magmatic rocks in other domains.

Résumé

Le magmatisme à l'intérieur et près de la zone faillée de Kirkland Lake-Larder Lake (FKL) se caractérise par une dualité compositionnelle. Deux séries, une syénitique (sans quartz) et l'autre granitique (avec quartz modal ou normatif) se retrouvent dans trois domaines magmatiques. Dans le domaine 1, qu'est la zone faillée, on retrouve les deux séries. Le domaine 2, immédiatement au nord de la FKL, est composé seulement de roches granitiques. Enfin, le domaine 3 au sud de la FKL possède les deux séries. Dans ce dernier domaine, les deux séries se retrouvent sous forme d'intrusions composées ou distinctes. La faille de Larder Lake, bordure méridionale de la FKL, agit comme frontière démarquant deux niveaux d'emplacement distincts: au sud les roches sont plutoniques, alors qu'au nord elles sont hypabyssales ou effusives. Ceci vient appuyer les interprétations antérieures d'un soulèvement relatif de la croûte au sud de la zone faillée. Les roches syénitiques et granitiques de la zone faillée (domaine 1), qui renferment la minéralisation aurifère, partagent des traits minéralogiques et compositionnels communs avec les roches magmatiques des deux autres domaines.

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INTRODUCTION

The Kirkland Lake-Larder Lake fault zone (KLF), at the western termination of the major east-trending Cadillac-Larder Lake fault, is host to one of North America's major gold camps. Magmatism, which is closely associated with gold, shows a compositional duality, in that both syenitic (quartz-free) and granitic types are present. These are variably distributed between three domains. Domain 1 comprises the KLF; Domains 2 and 3 are, respectively, the country north and south of the KLF (Fig. 1).

Domain 1 is largely composed of sedimentary rocks of the Timiskaming Group (Fig. 1), which lie unconformably on metamorphosed greenstone basement. Alkaline volcanic units (trachytes) are interbedded with the Timiskaming sedimentary rocks. The Timiskaming is cut by elongate intrusions of alkali-feldspar syenite and lamprophyre and by similarly elongate quartz-monzonite (feldspar porphyry) bodies. Domain 3, south of the KLF, also contains both magmatic types, which occur as different phases of large plutons, such as the Otto and Murdock Creek stocks. Other plutons within this domain appear to be exclusively syenitic (Lebel) or exclusively granitic (McElroy). In Domain 2, north of the KLF, only quartz-bearing rocks occur, these as elongated plutons in and around McVittie Township.

Rocks from the two magmatic suites are exposed at different levels of emplacement along the KLF. Rocks within and immediately north of the KLF are almost exclusively hypabyssal or extrusive, while those to the south are distinctly plutonic. These different styles of emplacement are clearly distinguished on the map (Fig. 1): hypabyssal and extrusive units are thin and elongated, whereas plutonic rocks are usually elliptical.

The spatial relationship between gold mineralization and felsic/alkaline intrusions in the Superior Province has long been recognized (Hodgson and MacGeehan, 1982; Cherry, 1983; Colvine et al., 1988; Burrows and Spooner, 1989). Only recently have attempts been made to establish the temporal relationship between gold-bearing veins and magmatism. The results are controversial: some data show a small time gap between gold mineralization and intrusion (Marmont and Corfu, 1989; Claoué-Long et al., 1990), while others show the gold to be significantly younger. For example, Bell et al. (1989) assign a Proterozoic age to the mineralizing event.

At Kirkland Lake there is a distinct time difference between the syenitic intrusions that are the principal host to the gold and the mineralizing event. The elongate intrusions of Domain 1 were formed during an extensional phase of tectonism, while the gold was introduced during a subsequent phase of transpression (Cameron, 1990). Thus the ore-bearing fluids could not have been derived from the batch of magma that formed these intrusions. Archean gold has frequently been ascribed to fluids expelled during prograde regional metamorphism. However, as Jolly (1978) showed, regional metamorphism at Kirkland Lake was complete prior to the deposition of the Timiskaming Group and thus well before alkaline magmatism.

There remains an enigma of an apparently close spatial connection between magmatism and gold mineralization at Kirkland Lake, but not an exact temporal overlap between the two. This enigma is present in other Archean gold camps. It has prompted a series of studies on the nature of alkaline magmatism at Kirkland Lake (Levesque, 1989; S.M. Rowins, pers. comm., 1990; Rowins et al., in press). For the Murdock Creek stock (Fig. 1), Rowins et al. (in press) found that the magma was intrinsically oxidized, which matches the unusually oxidized nature of the gold-bearing fluids at Kirkland Lake (Cameron and Hattori, 1987). Rowins et al. (in press) proposed that the magma was derived from oxid-

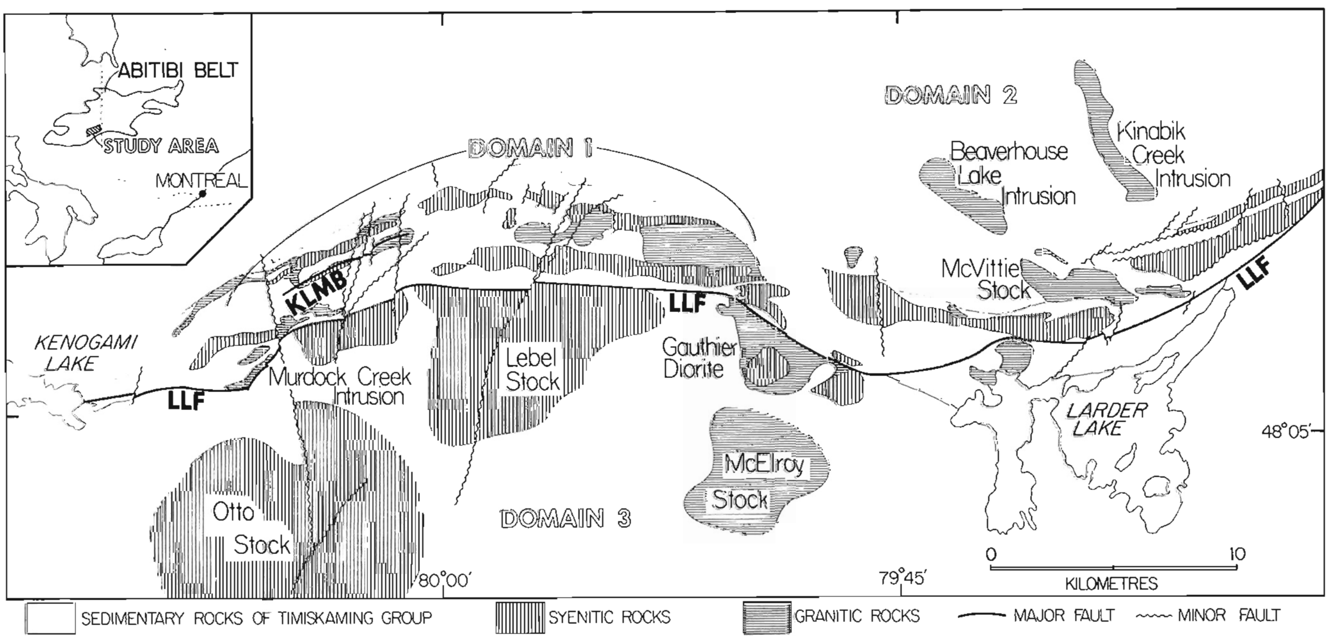

Figure 1. Geology and distribution of magmatic rocks along the Kirkland Lake-Larder Lake fault zone (KLF). The KLF is bounded to the south by the Larder Lake fault (LLF), and to the north by the basal Timiskaming unconformity (dashed line). (KLMB = Kirkland Lake Main Break).

ized and metasomatized upper mantle; they suggest a common genetic link between magma generation, CO_2 (represented by carbonatization), and gold-bearing fluids. The objective of the present study, which commenced in 1990, is to examine the relationship between the quartz-bearing and the alkaline igneous rocks. This paper presents some preliminary observations on this topic.

DESCRIPTION AND PETROGRAPHY OF ROCK UNITS

All syenitic plutonic rocks described are classified according to the IUGS nomenclature (Streckeisen, 1976). Some of the granitic rocks cannot, at present, be similarly classified, given the difficulty in identifying the primary constituents of their microcrystalline groundmass. Petrographic examination was of least-altered specimens for each rock type.

Units within the Kirkland Lake-Larder Lake fault zone (Domain 1)

Six igneous rock units occur within the KLF: alkali-feldspar melasyenite, alkali-feldspar syenite, trachyte, lamprophyre, feldspar porphyry, and quartz-feldspar porphyry. Several descriptive names have been used in past literature; a short review of these names, and a proposed classification is given.

Alkali-feldspar melasyenite

This is the largest unit of the intrusive syenitic complexes within the fault zone. Traditionally named augite syenite, several other terms have been used to describe this unit; these include basic syenite, mafic syenite, lamprophyre, and augite porphyrite. However, most primary pyroxene compositions (Levesque, 1989) fall in or above the diopside field of the IMA pyroxene nomenclature (Morimoto, 1989).

Least-altered alkali-feldspar melasyenite is dark olive-grey, medium grained, and hypidiomorphic to idiomorphic. This unit is texturally slightly heterogeneous, essentially composed of two phases: a porphyritic phase containing large pyroxene and biotite phenocrysts set in a fine grained groundmass of alkali feldspar (Fig. 2c), and an equigranular phase consisting of interlocking pyroxene, biotite, and alkali-feldspar crystals.

Alkali-feldspar melasyenite (Fig. 2d) (Colour Index(CI) = 40) is distinctive by the abundance (30 to 40 vol.%) of large (0.2 to 1.0 cm) euhedral pyroxenes that are commonly zoned. Other mafic minerals include biotite, magnetite, and apatite (total 10 vol.%). Subhedral perthitic feldspar, composed of roughly equal amounts of alkali feldspar and albite, constitutes 50 to 60 vol.% of this rock type.

Alkali-feldspar syenite

This is the least abundant intrusive rock type within the KLF. It has been called syenite, felsic syenite, acid syenite, massive syenite, and red syenite. It is the felsic equivalent of the alkali-feldspar melasyenite unit. Predominance of alkali feldspar over pyroxene imparts a pinkish-brown colour to hand specimens. Contact relationships between this rock type and the alkali-feldspar melasyenite are varied: in areas, alkali-feldspar syenite grades into alkali-feldspar melasyenite, while in others a sharp intrusive contact is observed. This indicates that the rock types are consanguinous and were emplaced penecontemporaneously. The alkali-feldspar syenite contains turbid, reddish microperthitic alkali feldspar (85 to 95 vol.%). Sparse pyroxene and biotite phenocrysts, along with fine grained magnetite, are common (less than 15 vol.%).

Trachyte

Trachyte, in the terminology of Cooke and Moorehouse (1969), constitute the volumetrically most important, and only extrusive member of the syenitic suite within the KLF. Unfortunately, these are the least well-preserved rocks within the fault zone. Nonetheless, three distinct trachytic phases can be identified on the basis of characteristic relict textures: felsic trachyte, mafic trachyte, and pseudoleucite-bearing trachyte.

Felsic trachyte is extensively altered, cataclastically deformed, and, for the most part, identifiable only by association to mafic and/or leucite-bearing trachytes. This rock type is typically fine grained due to complete recrystallization, displays mottled pink (hematite) and green (chlorite + carbonate) patches, and is schistose. A few outcrops display large, centimetre-sized fresh alkali feldspars in a fine grained reddish-orange groundmass composed of fine acicular alkali feldspar, chlorite, and carbonate.

Mafic trachyte (pyroxene trachyte) (Fig. 2a) is generally altered and cataclastically deformed. The mafic phenocryst is commonly weathered out, or completely altered to a yellowish-brown colour. Textures are somewhat better preserved in this unit, and primary mineral alignment, defined by altered pyroxenes, is often recognized (Fig. 2a). The recrystallized groundmass is composed of alkali feldspar and carbonate with minor iron oxides, epidote, and biotite. Relatively unaltered pyroxenes can be observed in a few locations (Fig. 2b), including an excellent exposure on the eastern shore of Bear Lake, in the western part of McVittie Township.

Pseudoleucite trachyte, which originally contained primary leucite (Cooke and Moorhouse, 1969), is characterized by equant, euhedral patches of fine grained orthoclase, carbonate, and minor sericite and iron oxide. These patches represent polygonal pseudoleucite crystals. The groundmass is usually composed of alkali feldspar, albite, carbonate, and iron oxide. Locally, this rock type contains pyroxene as both a phenocryst and groundmass phase; this rock was named leucite tephrite by Cooke and Moorhouse (1969). More rarely, the pseudoleucite crystals are completely altered to a fine interlocking mesh of sericite, carbonate, and alkali feldspar. This gives the rock a distinct green-spotted appearance, hence the name green-spotted trachyte (MacLean, 1956).

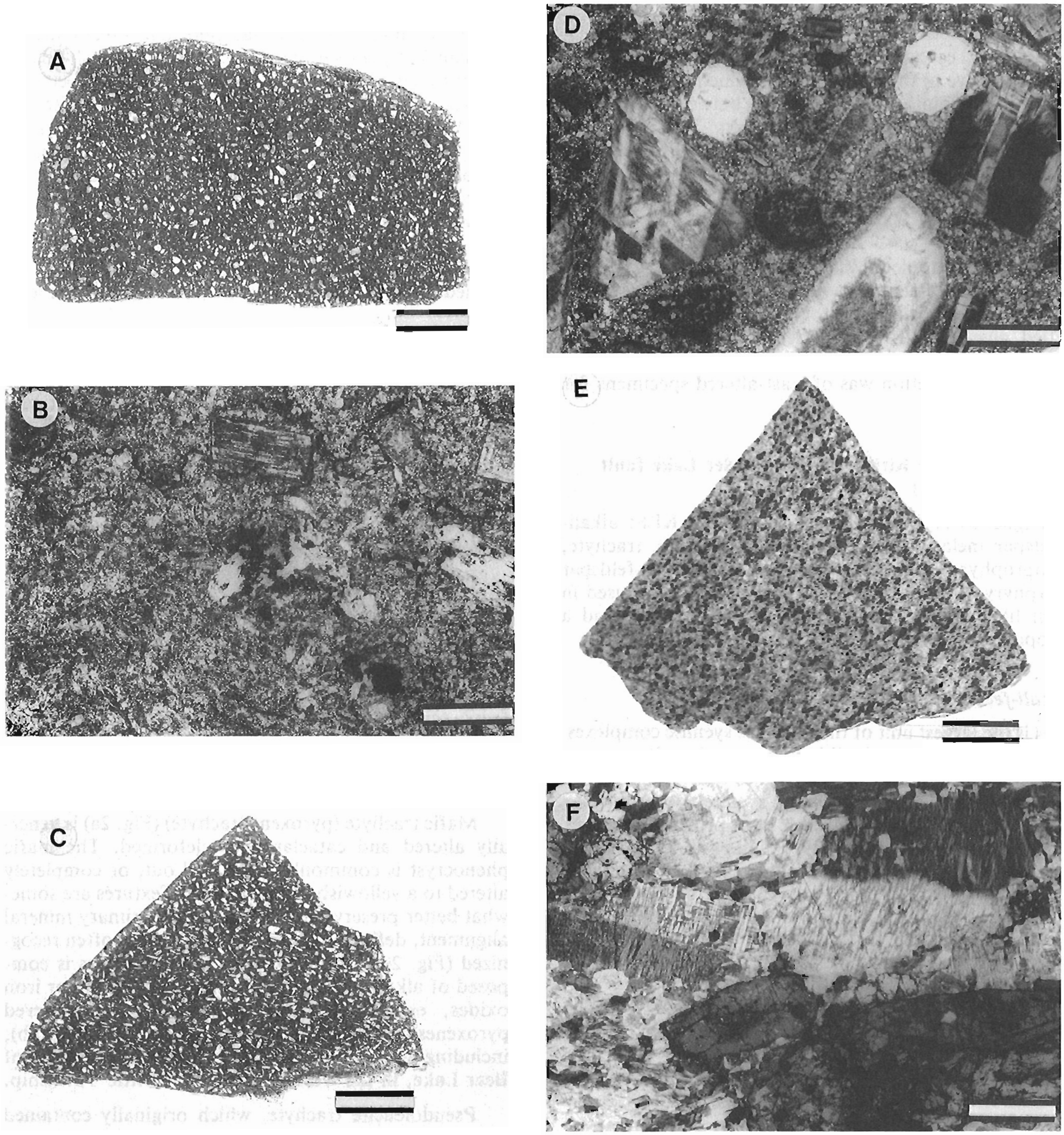

Figure 2. Spectrum of syenitic rocks observed along the KLF; scale bar for A, C, and E measures 2 cm; scale bar for B, D, and F is 2 mm.
A. extrusive mafic trachyte. Note primary pyroxene alignment; B. photomicrograph of mafic trachyte, showing subparallel feldspar and pyroxene phenocrysts; C. hypabyssal alkali-feldspar melasyenite; D. photomicrograph of alkali-feldspar melasyenite, displaying large euhedral pyroxenes; E. plutonic alkali-feldspar melasyenite (Otto stock); F. photomicrograph of Otto stock melasyenite displaying large perthitic alkali feldspar.

Lamprophyre

Two different types of calc-alkaline lamprophyre, following the IUGS nomenclature (Streckeisen, 1979), are observed within the fault zone: minette and vogesite. Minette dykes are most common; they contain large Mg-rich biotite phenocrysts set in a finer groundmass of alkali feldspar, plagioclase, and magnetite, with minor carbonate and chlorite. The second type of lamprophyre, vogesite, is characterized by the abundance of commonly zoned clinopyroxene phenocrysts enclosed in a groundmass of biotite, alkali feldspar, magnetite, chlorite, plagioclase, and carbonate.

Feldspar porphyry

Feldspar porphyry is the common name given to the most abundant hypabyssal granitic rock found within the KLF. It has also been called syenite porphyry (Thompson, 1950). A detailed description of this unit is given by Hicks and Hattori (1988), which is the basis for the description below.

Feldspar porphyry (Fig. 3a) is typically hypidiomorphic and dark olive-grey, containing abundant phenocrysts set in a fine groundmass composed of varying amounts of plagioclase and orthoclase, with minor quartz, magnetite, apatite, and titanite (Fig. 3b). Three phases can be distinguished based on phenocryst assemblages: plagioclase-biotite porphyry, plagioclase-orthoclase-biotite porphyry, and plagioclase-hornblende-biotite porphyry. As with the two alkali-feldspar syenite units, intrusive contacts vary from gradational to sharp, indicating the consanguinity of these three phases.

The name feldspar porphyry has been kept at present because of the difficulty in estimating modal percentages of primary alkali feldspar and plagioclase in the groundmass. However, these rocks have been classified as quartz monzonites and quartz monzodiorites by Hicks and Hattori (1988).

Quartz-feldspar porphyry

This forms a minor part of the granitic rocks within the KLF. It is practically indistinguishable from feldspar porphyry in hand specimen. Both units are mineralogically identical; the presence of large, rounded quartz phenocrysts distinguishes this unit. The subhedral nature, roughly hexagonal outlines, and embayed crystal edges of quartz phenocrysts are indicative of rapidly cooled magmatic rocks. This is commonly referred to as amoeba quartz. The strong similarities between feldspar porphyry and this unit indicate that both are genetically related; this unit would probably fall in or near the granodiorite field of the QAP diagram of Streckeisen (1976).

Rock units north of the KLF (Domain 2)

Three intrusive bodies lie immediately north of the KLF, in and adjacent to McVittie Township; they are the McVittie stock, which transects the northern limit of the fault zone, the Beaverhouse Lake intrusion, and the Kinabik Creek intrusion (Fig. 1). These rocks have many features in common with the granitic magmatic rocks found within the KLF; they occur as thin, elongated bodies and display porphyrytic textures.

The McVittie stock and the Beaverhouse Lake intrusion are composed of euhedral plagioclase, hornblende, and rare, large, subhedral alkali-feldspar phenocrysts set in a fine grained groundmass of plagioclase, alkali feldspar, and minor magnetite, apatite, titanite, and quartz. These two intrusions are identical to the feldspar porphyry bodies observed within the fault zone. The Kinabik Creek intrusion (Fig. 3c) differs from the other two intrusions only by the presence of large, subhedral, often embayed quartz phenocrysts, which occur along with the plagioclase, hornblende, and alkali-feldspar phenocrysts (Fig. 3d). Accordingly, the Kinabik Creek intrusion is a quartz-feldspar porphyry.

Rock units south of the KLF (Domain 3)

The large plutons located to the south of the KLF include the predominantly syenitic Otto, Murdock Creek, and Lebel stocks, the granitic McElroy stock, and a diorite intrusion in Gauthier Township. These plutons are usually rounded, although those in fault contact with the southern limit of the KLF are truncated and appear more elongate.

Recent investigations on the Murdock Creek intrusion (Rowins et al., 1989; Rowins et al., in press) and the Otto stock (Smith and Sutcliffe, 1988) have revealed that these largely syenitic plutons contain quartz-bearing phases. The Murdock Creek stock is composed of an extensive alkali-feldspar syenite and melasyenite core enclosed by a thin mafic margin composed of clinopyroxenite, meladiorite, melamonzonite, and hornblendite (Rowins et al., 1989). The Otto stock (Fig. 2e) is composed of five lithological units: syenite, quartz syenite, mafic syenite (Fig. 2f), porphyritic syenite, and hornblendite/diorite (Smith and Sutcliffe, 1988). Leucocratic monzonite also appears as a very minor unit.

The Lebel stock is composed of at least two transitional syenitic units: a foliated, marginal alkali-feldspar melasyenite (CI = 30), and alkali-feldspar syenite (CI = 20). Large perthitic alkali-feldspar laths define the foliation in the marginal unit. Clinopyroxene, which is commonly zoned, biotite, magnetite, minor apatite and titanite account for up to 35 vol.% of this phase. Traces of primary quartz and zircon are also present. In the alkali-feldspar syenite, hornblende replaces clinopyroxene as the main mafic mineral phase, and alkali feldspar is present as large, equant, sometimes porphyritic, subhedral crystals. Biotite, magnetite, titanite, apatite, and zircon are accessory phases.

The McElroy stock (Fig. 3e) is a large, homogeneous quartz monzonite (CI = 10) intrusion. This unit consists principally of microcline, albite, and quartz; hornblende, magnetite, pyroxene, titanite (up to 2 vol.%) and apatite are also present (Fig. 3f). Myrmekitic textures are commonly observed at alkali feldspar-plagioclase contacts.

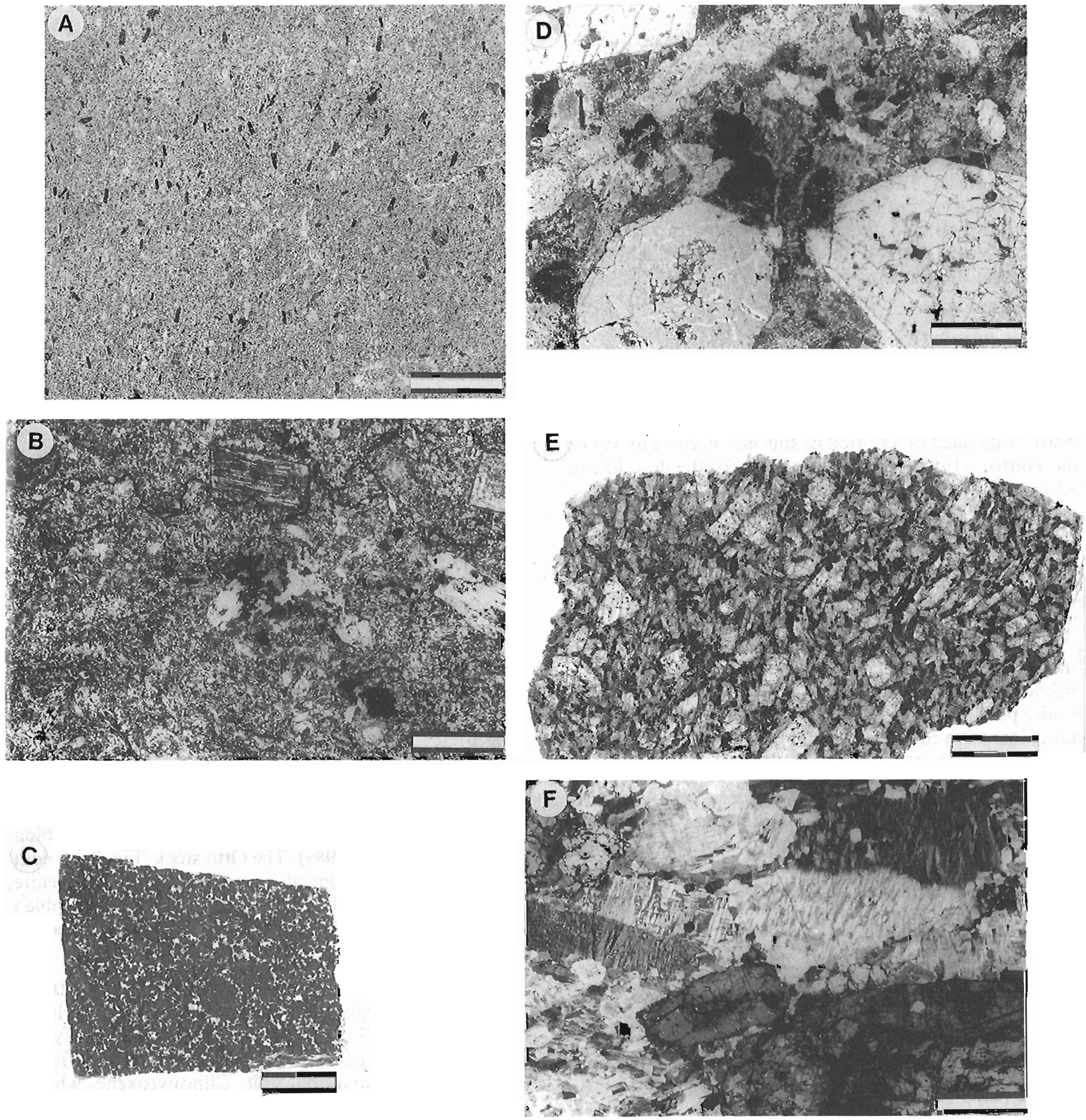

Figure 3. Spectrum of granitic rocks observed along the KLF; scale bar for A, C, and E measures 2 cm; scale bar for B, D, and F is 2 mm.
A. hypabyssal quartz monzonite (feldspar porphyry); B. photomicrograph of feldspar porphyry, showing large alkali-feldspar phenocrysts (lower right corner), and smaller abundant plagioclase phenocrysts; C. hypabyssal quartz-feldspar porphyry (Kinabik Creek intrusion); D. photomicrograph of quartz-feldspar porphyry. Note euhedral pseudohexagonal quartz phenocrysts; E. plutonic quartz monzonite (McElroy stock); F. photomicrograph of McElroy stock quartz monzonite, displaying subhedral amphibole.

A large dioritic intrusion, located north of the McElroy stock, is in fault contact with the southern limit of the KLF. Only a few occurrences of diorite have been noted south of the KLF. The rock is generally equigranular, hypidiomorphic, and medium grained. Common to rocks within the fault zone, diorite is also slightly heterogeneous; it ranges from leucodiorite to diorite to meladiorite. Leucodiorite (CI = 15) and diorite (CI = 40) are composed of plagioclase, hornblende, quartz, magnetite, and titanite. The elongate plagioclase laths commonly enclose hornblende and magnetite. The meladiorite contains large euhedral clinopyroxene crystals (up to 80 vol.%) with interstitial plagioclase and titanite. Hornblende alteration rims around pyroxene are commonly observed.

Rocks found to the south of the KLF are distinctive in the textures they display. Large, rounded plutons displaying medium- to coarse-grained hypidiomorphic, equigranular textures are not observed anywhere else along the fault zone. These rocks were emplaced at depth, and in most cases, contain both granitic and syenitic units. This indicates that the compositional duality of magmatism is not restricted to rocks within the KLF.

DISCUSSION AND SUMMARY

Two suites of magmatic rocks are present within and near the Kirkland Lake-Larder Lake fault zone. Syenitic (quartz-free) rocks and granitic (quartz-bearing or quartz-normative) rocks occur in three well-defined magmatic domains (Fig. 4). Within the fault zone (Domain 1), both syenitic and granitic rocks outcrop as thin, elongated units. Magmatism immediately north of the KLF (Domain 2) is characterized by elongate granitic bodies. South of the KLF (Domain 3), three types of large, rounded intrusions are encountered. Predominantly syenitic composite intrusions with minor mafic margins (Otto, Murdock Creek, Lebel stocks) are most common. One homogeneous granitic pluton (McElroy stock) outcrops south of the fault zone, and a diorite intrusion borders the fault zone in Gauthier Township.

The Larder Lake fault (LLF), which marks the southern limit of the KLF, defines two zones of magmatic emplacement: rocks to the south are distinctly plutonic, while rocks north of the LLF are either hypabyssal or extrusive (Fig. 5). The rounded shape and textures of the intrusions lying south of the KLF are strong evidence for forceful emplacement by diapiric rise, while the porphyritic and elongated nature of the syenitic and granitic bodies within and north of the fault zone indicate high-level passive emplacement, or extrusion in the case of the trachytes.

The syenitic phases of the large intrusions south of the KLF (Domain 3) exhibit similarities to the syenitic bodies within the fault zone (Domain 1). Both fall in the alkali-feldspar syenite field of the QAP diagram (Streckeisen, 1976), with diopsidic clinopyroxene and Mg-rich biotite as the dominant mafic mineral phases. Compositionally, the clinopyroxenes from these two domains are identical (Levesque, 1989). Alkali feldspar in both domains is perthitic, containing roughly equal amounts of orthoclase and albite.

The granitic rocks (feldspar porphyry and quartz-feldspar porphyry) north of the fault zone are mineralogically and texturally identical to the granitic rocks found within the fault zone. They also share a common style of intrusion and level of emplacement, occurring as thin, elongated hypabyssal bodies. Furthermore, these rocks share common mineralogical characteristics with the diorite intrusion (hornblende and plagioclase assemblage in the leucodiorite and diorite) and the monzonitic phases (quartz-bearing) of the composite intrusions south of the KLF. The majority of granitic rocks within the

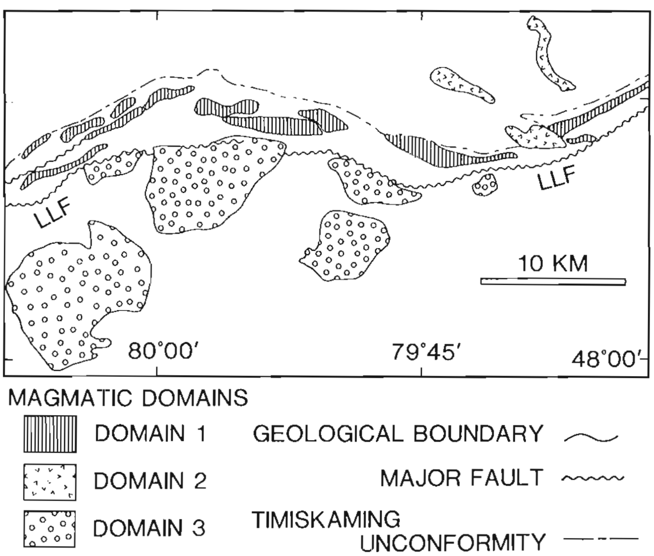

Figure 4. Magmatic domains along the KLF. The Larder Lake fault (LLF) separates Domain 3 from Domain 1, while the northern limit of the KLF (dashed line) marks the boundary between Domain 1 and Domain 2.

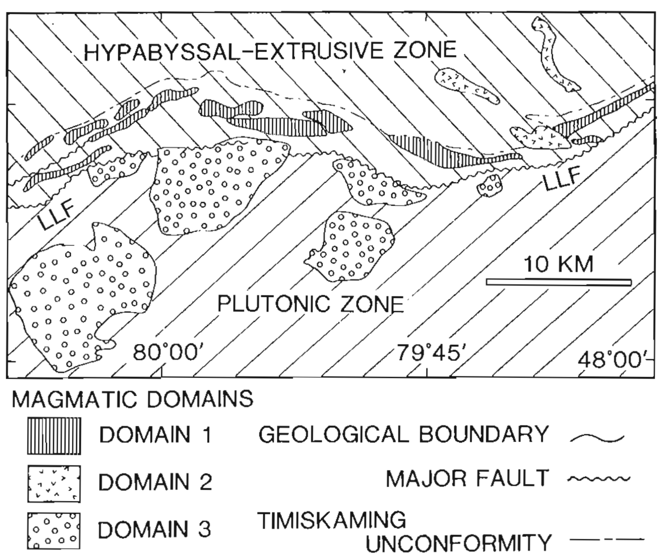

Figure 5. Emplacement levels of magmatic rocks along the KLF. The Larder Lake fault (LLF) defines the boundary between the plutonic zone and the hypabyssal-extrusive zone.

KLF are cataclastically deformed and intensely altered; those north of the fault zone are less altered and undeformed. These three intrusions (McVittie, Beaverhouse Lake, and Kinabik Creek) provide an opportunity to examine unaltered rocks from the granitic suite, which are a host to gold within the KLF.

Textural differences between the magmatic rocks occurring within and to the north of the KLF (Domains 1 and 2), and those found to the south (Domain 3) are striking (Fig. 5). Rocks to the south exhibit coarse grained, holocrystalline, hypidiomorphic, equigranular to inequigranular textures, indicative of emplacement at depth. Rocks found within and to the north of the KLF feature hiatal-porphyritic and trachytic textures, with idiomorphic phenocrysts (including amoeba quartz), and a microgranular groundmass. These textures are indicative of rapid cooling and high-level emplacement, or extrusion.

Movements along the Kirkland Lake and Larder Lake faults have been interpreted as south side up (Thompson, 1950). The textural evidence given above supports this; subsequent to intrusion, deeper levels to the south have been uplifted. Our working hypothesis is that the hypabyssal syenitic bodies of Domain 1 and the syenitic plutons of Domain 3 were generated from the same parental magma, differing only in their level of emplacement. This scenario may possibly be extended to the rocks of the granitic suite.

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Preliminary assessment of wallrock alteration in uranium open-pit mines, Athabasca basin, Saskatchewan: mineralogical and chemical characterization

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Abstract

A suite of wallrock samples from open-pit uranium mines was collected during two reconnaissance trips to the Athabasca basin. The samples were collected to assess the extent of recent wallrock alteration and the potential for long-term contamination of surface and groundwaters following decommissioning of the open pits. The samples from Cluff Lake and Rabbit Lake mine camps have similar mineralogical and chemical characteristics. Trace elements of environmental concern are slightly enriched relative to regional sandstones. Water samples have elevated levels of As, Cr, Ni, and U.

Résumé

Une série d'échantillons de roche encaissante, provenant de mines d'uranium exploitées à ciel ouvert, ont été recueillis durant deux visites de reconnaissance dans le bassin d'Athabasca. Ces échantillons serviront à évaluer le degré de l'altération récente subie par la roche encaissante, et les possibilités de contamination à long terme des eaux souterraines et des eaux superficielles après la fermeture de ces mines. Les échantillons provenant des campements miniers du lac Cluff et du lac Rabbit présentent des caractéristiques minéralogiques et chimiques similaires. Les grès régionaux sont légèrement enrichis en éléments à l'état de traces qui pourraient causer des problèmes écologiques. Les échantillons d'eau contiennent des taux élevés d'As, Cr, Ni et U.

INTRODUCTION

Of major environmental concern is the impact of mining on the natural environment. It is estimated that by the year 2000 there will be over 172 000 hectares of land despoiled through mining activities in Canada (Brooks et al., 1989). In addition to the physical disturbance of the land, acidic conditions may be generated in sulphide tailings, waste-rock piles, and on mine walls. Acid effluents can be detrimental to the neighbouring terrestrial and aquatic environments through *in situ* leaching of heavy metals. Therefore containment is a key step in effective environmental management of active and abandoned mine sites.

The purpose of this study is to assess whether alteration of the wallrock in mined-out uranium open pits in the Athabasca basin will release environmentally hazardous elements into the surface and groundwater systems. Although radionuclides such as ^{226}Ra, U, and Th are of prime importance, other associated heavy metals such as As, Co, Cu, Ni, Mo, Pb, Se, V, and Zn are also of concern.

This report describes preliminary results of reconnaissance trips to the Athabasca basin to collect wallrock and water samples from the open pits. The objectives of the study are: (1) to characterize the materials in the wallrock mineralogically and chemically; (2) to determine the extent of wallrock alteration/weathering; and (3) to evaluate/predict site-specific conditions for leachability of environmentally toxic heavy metals and radionuclides into surface and groundwaters.

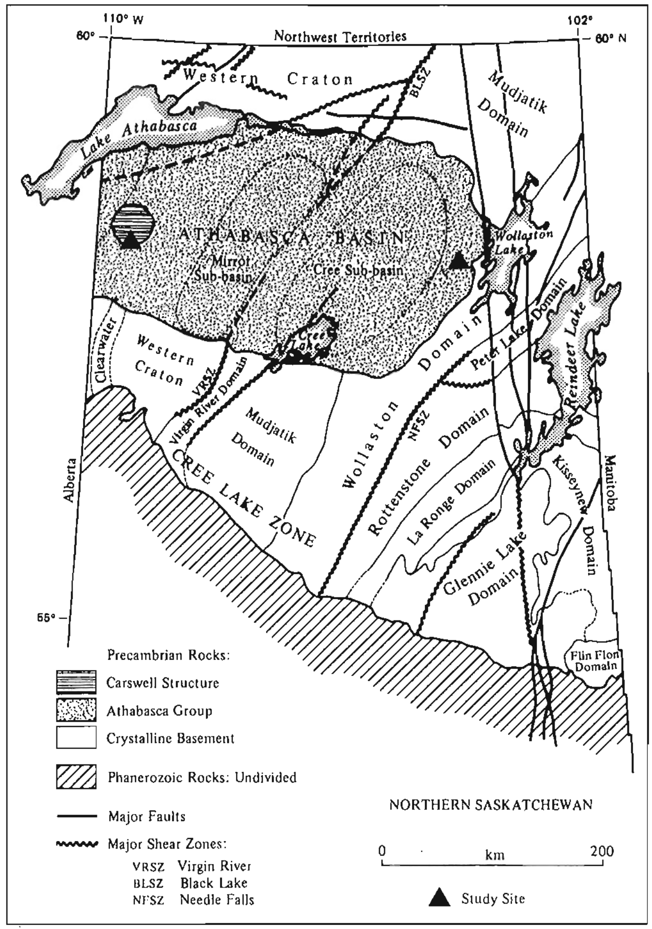

Figure 1. Regional geology and location of study area in northern Saskatchewan (after Macdonald and Broughton, 1980).

GEOLOGICAL SETTING

Regional geology

The crystalline basement of northern Saskatchewan is composed of Archean and Aphebian sedimentary, volcanic, and plutonic rocks of the western Churchill Structural Province and forms part of the Trans-Hudson Orogen (Hoffman, 1981) (Fig. 1). The basement rocks can be subdivided into three fault-bounded tectonic zones which have been variably affected by thermotectonic events during the Hudsonian Orogeny (ca. 2000-1700 Ma; Lewry et al., 1985).

Intense weathering of the crystalline basement followed the Hudsonian Orogeny. Remnants of the regolith, about 50 m thick, are preserved under the Athabasca Group. The Paleohelikian Athabasca Group sedimentary succession (Fig. 2), consisting of fluvial quartz sandstone and conglomerate with minor marine sediments, covers an area in excess of 100 000 km^2, and reaches a thickness of about 1400 m (Sibbald, 1985).

Athabasca sandstones had a complex diagenetic history, involving intensive oxidation and postdepositional leaching (Hoeve et al., 1980). The sandstones are now composed of orthoquartzites with a clay-rich matrix of illite, kaolinite, and hematite (Ramaekers, 1983).

The U deposits in the Athabasca basin are associated with the unconformity between the Athabasca Group sandstones and the Hudsonian basement (Fig. 3). The unconformity-type deposits occur west of the Needle Falls Shear Zone and appear to be related to a belt of Aphebian supracrustal rocks (Fogwill, 1985). These epigenetic deposits are polymetallic and are composed of pitchblende and coffinite associated with accessory minerals containing Ni, Co, Cu, Pb, Zn, Mn, Fe, V, Ag, Au, PGEs, S, As, Bi, and Se (Hoeve, 1978). The first stage of mineralization occurred between 1350 and 1250 Ma ago and is represented by pitchblende, Ni-Co arsenides, and sulpharsenides. A second phase, 1100 to 1050 Ma ago, is characterized by the presence of dull, massive pitchblende, base metal sulphides, siderite, Fe-chlorite, tourmaline veins, and euhedral quartz. Later remobilization between 300 to 250 Ma ago has also been recorded (Hoeve and Quirt, 1984).

Detailed geology

Cluff Lake Camp

The Cluff Lake deposits occur along the southern margin of the Carswell Structure, located in the western part of the Athabasca basin (Fig. 3). The uplifted basement core of the structure consists of feldspathic and mafic gneisses

(Earl River Complex) overlain by aluminous gneisses (Peter River Gneiss). The basement is uncomformably overlain and surrounded by deformed Athabasca Group clastic sedimentary rocks and Carswell stromatolitic dolomites which form a series of rings about 4 to 5 km wide (Tona et al., 1985; Sibbald and Quirt, 1987). The Ordovician Carswell event produced vein and dyke-like intrusives of Cluff breccia and overprinted the earlier tectonic fabric (Tona et al., 1985).

The D ore deposit occurs along the overturned unconformity between Aphebian basement and Athabasca Group sandstones. The basement, consisting of garnet-rich aluminous gneisses, overlies interbedded siltstones, conglomerates, and sandstones. Both units are crosscut by an altered mylonitic zone (Ey et al., 1985; Sibbald and Quirt, 1987). The ore occurs as discontinuous lenses in the sheared and chloritized siltstones. The ore is composed of pitchblende, uraninite, and coffinite in association with native Au and Au tellurides, native Se and Pb, Bi, Ni, and Co arsenides, jordisite, pyrite, galena, and chalcopyrite (Tona et al., 1985; Sibbald and Quirt, 1987).

The D ore body was mined by open-pit methods between 1979 and 1981, producing 110 000 tonnes of ore at an average grade of 3.79% (Tona et al., 1985). The pit is about 90 m wide, 200 m long, and 23 m deep. The pit was flooded in the spring of 1983 when Boulder Creek overtopped its banks. Monitoring of the surface and groundwater quality commenced in 1983 (Saskatchewan Environment and Public Safety, 1990).

The Claude deposit occurs entirely within basement gneisses (Peter River gneisses). The ore, consisting of uraninite with U-Ti minerals, clausthalite, galena, pyrite, chalocopyrite, sphalerite, and jordisite, is associated with subvertical shear zones containing rotated fault blocks (Tona et al., 1985; Sibbald and Quirt, 1987). An experimental open pit was opened in 1982 and mining was completed in 1989 after removing 2097 tonnes of U at 0.36% grade. Currently, the pit remains open with flooding occurring naturally at a slow rate.

The Dominique-Janine deposit also occurs within the basement gneisses. The fault-controlled contact between the Peter River and Earl River gneisses is crosscut by the subvertical north-trending mineralized zone. The deposit contains 878 tonnes of 3.8% U (Sibbald et al., 1990) and is currently being mined by open-pit methods.

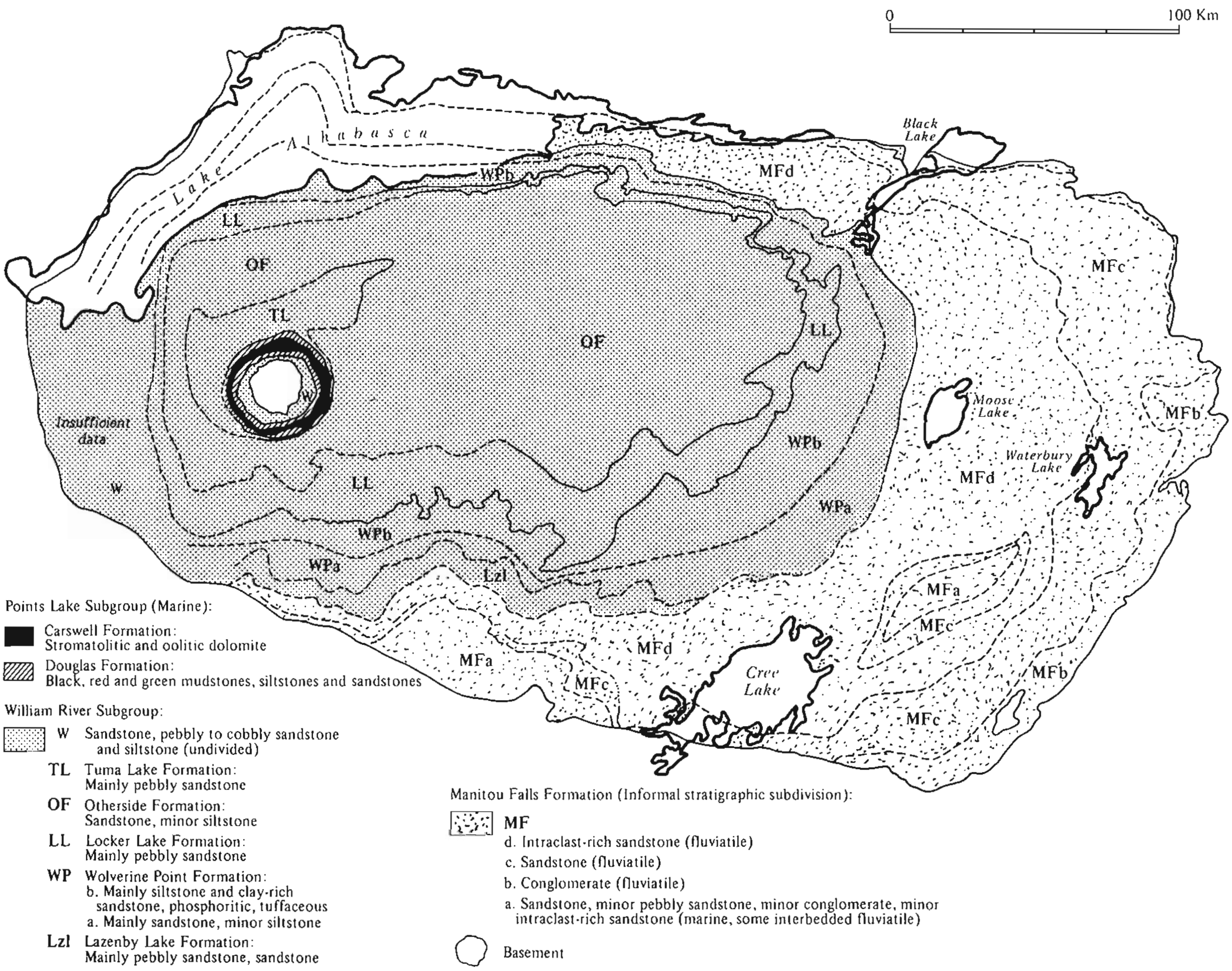

Figure 2. Detailed lithology of the Athabasca basin (after Macdonald and Broughton, 1980).

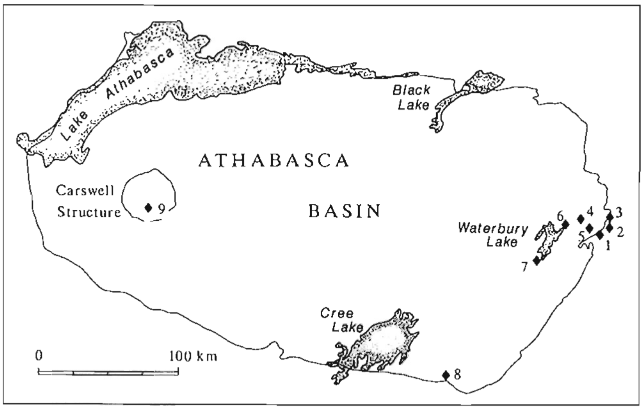

Figure 3. Location of uranium deposits in the Athabasca basin (1 = Rabbit Lake; 2 = Collins Bay; 3 = Eagle Point; 4 = Dawn Lake; 5 = McClean Lake; 6 = Midwest Lake; 7 = Cigar Lake; 8 = Key Lake; 9 = Cluff Lake).

Rabbit Lake Camp

The Rabbit Lake mine camp which includes the Rabbit Lake, Collins Bay A-Zone, B-Zone, D-Zone, and Eagle Point deposits occurs along the eastern margin of the Athabasca basin (Fig. 3). In the Collins Bay area, mineralization occurs within Aphebian Wollaston Group metasedimentary rocks and the Athabasca Group sandstones and conglomerates. These two units unconformably overlie Archean granitoid gneisses. Mineralized zones occur near the Paleohelikian-Aphebian and Paleohelikian-Archean unconformities (Ward, 1989).

The Collins Bay B-Zone deposit is located in the footwall of the Collins Bay Fault, within the Athabasca Group sandstones and conglomerates. The ore consists of pitchblende in association with coffinite, nicolite, rammelsbergite, gersdorffite, maucherite, bravoite, violarite, chalcopyrite, and other Cu sulphides and occurs as lenses that parallel the strike of the fault (Sibbald and Quirt, 1987). The deposit is being mined as an open pit.

SAMPLING AND ANALYTICAL METHODS

Four uranium mines in the Athabasca basin were visited in September 1989: Key Lake, Cigar Lake, Rabbit Lake, and Cluff Lake (Fig. 3). The trip served two purposes: (1) to collect samples for preliminary analyses; and (2) to gain an overview of the environmental problems associated with decommissioning of the U open-pit mines. Eleven samples were collected from the N wall of the D pit at Cluff Lake and a suite of ten samples was provided from Collins Bay B-Zone pit by mine officials at Rabbit Lake. One surface water sample from the D pit and three water samples from seeps proximal to sample locations from the Collins Bay B-Zone pit were also collected. Results from these samples are described in this report.

During detailed sampling in June 1990, 53 wallrock samples were collected from the D, Claude, and Dominique-Janine open pits at Cluff Lake mine and Collins Bay B-Zone open pit at the Rabbit Lake mine camp. Samples included rocks representative of the different lithologies, material from fractures and fault zones, zones of radioactivity as determined by scintillometer, and zones with distinct colouration produced by the occurrence and possible leaching of heavy metals. Water samples were collected from appropriate nearby sites, especially from seeps and fracture zones along the open-pit walls.

Mineral compositions of samples were determined on bulk and clay-size separates using X-ray diffraction analysis. This will be complemented by petrographic analysis, scanning electron microscopy, and microprobe analysis of selected samples.

Major elements, Ba, Nb, Rb, Sr, and Zr concentrations in bulk powdered samples were determined by conventional X-ray fluorescence (XRF) analysis. Concentrations of FeO, H_2O, CO_2T, CO_2, C, and S were determined by rapid chemical methods. The trace elements Be, Co, Cr, Cu, La, Mo, Ni, V, Y, Yb, and Zn were analyzed by inductively-coupled plasma emission spectrometry (ICP-ES). Arsenic, Bi, and Sb were analyzed by quartz tube atomic absorption spectrophotometry (QT-AAS) by hydride generation. Uranium was analyzed by activation analysis-delayed neutron counting at X-Ray Assay Laboratories, Toronto. For water samples, the major elements, except Na and K, and the trace elements Ba, Co, Cr, Cu, Ni, Sr, V, and Zn were analyzed by ICP-ES. Na and K were analyzed by AAS. Boron, Hf, Ta, Th, U, and Zr were analyzed by inductively-coupled plasma mass spectrometry (ICP-MS) whereas As, Bi, and Sb were analyzed by QT-AAS (hydride method). The radionuclides ^{226}Ra, ^{210}Pb, and ^{210}Po were analyzed at the Saskatchewan Research Council. A sequential extraction procedure (Percival, 1989) has been applied to selected wallrock samples to determine the nature of partitioning of the metals (e.g. As, Co, Cu, Mo, Ni, Pb, U, Se, Th, V, and Zn) and thereby enable prediction of their relative mobility. Analyses of the extractant solutions are in progress.

RESULTS AND DISCUSSION

Wallrock samples collected from the D pit are aluminous gneisses (Peter River Gneiss) composed of quartz remnants and hematized garnets surrounded by a clay-rich matrix. Analyses of the clay-size fraction show that the samples are dominated by either illite or chlorite (sudoite variety) with subordinate hematite and quartz (Table 1).

Wallrock samples taken from the Collins Bay B-Zone open pit show more lithological variation. Samples CB-1, -2, -3, -6, and -8 are fine- to medium-grained Athabasca Group sandstones. Many of these samples are partially or completely bleached. Hematite and limonite occur as patches or as liesegang rings. Samples CB-4, -5, and -10 are graphitic gneisses and samples CB-7 and -9 are quartzofeldspathic gneisses. Sample CB-7 is a chloritic, brecciated pegmatite whereas CB-9 is representative of the red paleoweathering zone (regolith). Clay mineral analyses show that four of the five sandstone samples are dominated by illite and quartz with minor to trace amounts of kaolinite and hematite (Table 1). The excep-

Table 1. Mineral composition of clay-size fraction of samples from D pit, Cluff Lake Mine (PNA-89) and Collins Bay B Zone pit, Rabbit Lake Mine (PNA-CB). Chlorite is the sudoite variety (Al-chlorite); I/S = illite/smectite mixed-layer clay mineral; minerals listed in order of abundance.

SAMPLE NO.	COMPOSITION
PNA-89-1	Chlorite, Illite, Quartz, Hematite
PNA-89-2	Chlorite, Illite, Quartz, Hematite, I/S mixed-layer
PNA-89-3	Illite, Chlorite, Quartz, Hematite
PNA-89-4	Illite, Chlorite, Hematite, Quartz
PNA-89-5	Chlorite, Illite, Quartz, Hematite
PNA-89-6	Illite, Chlorite, Hematite, Quartz
PNA-89-7	Illite, Chlorite, Quartz, Hematite
PNA-89-8	Illite, Chlorite, Hematite, Quartz
PNA-89-9	Quartz, Hematite, Illite, Chlorite, Smectite, Siderite
PNA-89-10	Illite, Chlorite, Hematite, Quartz, Kaolinite(?)
PNA-89-11	Illite, Quartz, Chlorite
PNA-CB-1	Quartz, Illite, Kaolinite
PNA-CB-2	Illite, Quartz, Kaolinite, Hematite
PNA-CB-3	Illite, Quartz, Kaolinite
PNA-CB-4	Chlorite, Illite, Plagioclase Feldspar, Quartz, Graphite, Kaolinite(?)
PNA-CB-5	Chlorite, Illite, Kaolinite(?), Graphite
PNA-CB-6A	Illite, Goethite, Quartz, I/S mixed-layer
PNA-CB-6B	Illite, Quartz, Goethite, I/S mixed-layer
PNA-CB-7A	Illite, Chlorite, Quartz, Diaspore
PNA-CB-7B	Chlorite, Illite, Quartz, Diaspore
PNA-CB-8	Kaolinite, Quartz, Illite
PNA-CB-9	Illite, Chlorite, Quartz, Hematite, Smectite
PNA-CB-10	Illite, Chlorite, Quartz, Graphite

Table 2. Chemical analyses of bulk, powdered rock samples from D pit, Cluff Lake Mine (PNA-89).

	PNA-89-01	PNA-89-02	PNA-89-03	PNA-89-04	PNA-89-05	PNA-89-06	PNA-89-07	PNA-89-08	PNA-89-09	PNA-89-10	PNA-89-11
SiO2 (wt%)	71.3	74.6	66.2	54.1	79.8	59.9	54.2	59.7	90.3	67.3	56.2
TiO_2	0.68	0.64	0.16	1.13	0.33	0.99	0.90	0.88	0.06	0.78	0.81
Al_2O_3	15.5	13.4	17.5	26.9	11.2	22.2	19.4	20.9	4.1	18.6	26.3
Cr_2O_3	0.01	0.01	0.02	0.01	0.02	0.03	0.01	0.02	0.00	0.01	0.04
Fe_2O_3	2.2	1.8	4.6	3.1	0.7	3.6	8.5	5.7	2.5	3.1	1.1
FeO	0.2	0.2	0.5	0.3	0.3	0.5	3.1	0.7	0.3	0.3	0.3
MnO	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.05	<0.01	<0.01	0.00	0.00
MgO	2.93	3.01	2.54	2.38	2.82	2.25	1.89	1.74	0.39	1.07	1.51
CaO	0.14	0.02	0.04	0.03	<0.02	0.02	0.24	0.16	0.26	0.04	0.10
Na_2O	<0.1	<0.1	<0.1	<0.1	<0.1	0.1	0.1	0.1	<0.1	<0.1	0.1
K_2O	2.68	2.01	3.69	6.75	1.47	5.55	5.38	5.76	1.01	5.11	7.96
H_2OT	4.0	3.7	4.0	5.1	3.3	4.4	4.5	4.2	0.8	3.1	4.3
CO_2T	<0.1	<0.1	0.1	<0.1	0.1	0.2	1.8	0.2	0.1	0.1	0.1
C	-	-	-	-	-	-	-	-	-	-	-
P_2O_5	0.15	0.05	0.15	0.11	0.02	0.05	0.11	0.13	0.21	0.14	0.11
S	0.01	0.01	0.01	0.01	<0.01	0.01	0.01	0.11	<0.01	0.01	0.01
Total	99.9	99.5	99.6	100.1	100.1	99.8	100.2	100.3	100.0	99.8	99.2
As (ppm)	1.9	2.3	3.6	2.2	1.4	5.2	12.5	10.0	22.4	3.4	1.3
Ba	85	97	104	185	31	265	386	393	77	298	669
Be	1.9	1.8	2.1	2.9	1.6	2.4	3.2	3.3	0.5	2.0	6.5
Co	6	6	6	9	5	13	14	12	2	6	18
Cr	64	69	140	120	120	190	100	110	52	93	300
Cu	2	1	4	4	2	8	6	8	8	2	3
La	64	40	150	100	3	38	58	49	54	150	18
Mo	<2	<2	<2	<2	3	<2	<2	<2	<2	<2	<2
Ni	59	59	54	50	52	53	30	33	14	19	22
Nb	21	22	2	24	23	23	24	27	8	15	17
Rb	44	36	55	96	18	112	114	129	15	81	234
Sr	187	93	390	304	<0.1	114	124	126	31	318	91
U	1.6	1.6	3.0	1.9	1.8	2.4	2.4	2.3	2.1	2.0	4.7
V	88	81	220	170	52	180	160	170	54	110	200
Y	19	13	120	48	33	43	35	35	9	28	150
Yb	2.0	1.4	10	5.0	3.7	4.1	3.8	3.9	0.7	3.0	16
Zn	<0.5	<0.5	<0.5	2	<0.5	10	8	7	<0.5	1	27
Zr	276	381	137	287	278	313	348	316	20	404	287

Table 3. Chemical analyses of bulk, powdered rock samples from Collins Bay B Zone pit, Rabbit Lake Mine (PNA-CB).

	PNA-CB-1	PNA-CB-2	PNA-CB-3	PNA-CB-4	PNA-CB-5	PNA-CB-6A	PNA-CB-6B	PNA-CB-7A	PNA-CB-7B	PNA-CB-8	PNA-CB-9	PNA-CB-10
SiO2 (wt%)	93.1	90.3	90.1	57.4	68.1	93.2	93.1	70.3	67.1	91.5	76.1	74.1
TiO_2	0.09	0.26	0.15	0.90	0.77	0.17	0.26	0.35	0.13	0.19	0.34	0.41
Al_2O_3	4.8	6.0	6.7	19.9	17.8	2.5	2.7	17.3	18.7	5.4	11.8	14.3
Cr_2O_3	<0.01	<0.01	<0.01	0.01	0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
Fe_2O_3	0.1	0.5	<0.1	1.4	<0.1	2.2	1.6	0.5	0.7	<0.1	3.4	0.6
FeO	0.1	<0.1	0.1	3.6	0.3	<0.1	0.1	0.4	0.5	<0.1	0.1	0.1
MnO	<0.01	<0.01	<0.01	0.02	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
MgO	0.18	0.26	0.23	6.89	2.59	0.15	0.15	2.23	3.77	0.11	2.13	2.45
CaO	<0.02	<0.02	0.04	0.15	0.04	<0.02	<0.02	0.18	0.61	<0.02	0.07	0.08
Na_2O	<0.1	<0.1	<0.1	<0.1	0.1	<0.1	<0.1	0.1	0.1	<0.1	<0.1	<0.1
K_2O	0.99	1.67	1.07	3.15	2.14	0.61	0.70	4.20	3.35	0.54	2.75	2.91
H_2OT	1.0	1.1	1.0	6.2	5.0	0.7	0.7	3.4	4.6	1.5	2.8	3.4
CO_2T	<0.1	<0.1	<0.1	0.1	<0.1	<0.1	<0.1	0.1	<0.1	<0.1	<0.1	0.9
C	-	-	-	-	1.6	-	-	-	-	-		
P_2O_5	0.03	0.04	0.03	0.14	0.03	0.03	0.04	0.17	0.59	0.03	0.08	0.11
S	<0.01	<0.01	0.15	<0.01	<0.01	<0.01	<0.01	0.11	0.02	<0.01	0.01	0.01
Total	100.4	100.2	99.5	100.1	98.7	99.6	99.5	99.4	100.2	99.3	99.6	99.4
As (ppm)	124	528	36	15	35	17	14	2.6	0.7	2.0	1.2	0.7
Ba	20	47	151	270	298	40	64	171	80	32	142	146
Be	0.3	1.0	0.5	3.6	2.4	0.5	0.6	2.2	2.9	0.3	2.5	2.6
Co	5	3	<0.3	42	5	<0.3	<0.3	4	6	<0.3	6	5
Cr	6	11	11	110	75	14	13	21	20	10	22	37
Cu	4	4	9	2	2	7	6	4	4	1	2	2
La	15	27	13	54	30	14	18	100	190	15	36	67
Mo	3	<2	6	<2	2	4	3	<2	2	<2	<2	<2
Ni	48	80	20	360	420	4	7	27	36	3	23	31
Nb	12	15	13	30	28	18	18	23	2	15	20	19
Rb	16	20	17	161	62	17	13	98	66	6	148	66
Sr	59	252	74	98	71	53	71	208	580	70	81	218
U	10.2	56.5	64.1	11.8	21.3	100	127	8.7	6.2	9.1	4.0	2.7
V	4	130	61	130	230	110	90	55	61	18	48	55
Y	4	9	8	51	18	11	13	42	24	9	21	24
Yb	0.4	1.2	1.2	5.3	2.4	1.2	1.5	4.0	1.8	1.1	1.9	2.6
Zn	<0.5	<0.5	<0.5	40	6	<0.5	<0.5	3	<0.5	<0.5	20	1
Zr	215	479	328	220	214	393	467	343	39	476	131	184

Table 4. Chemical analyses of water samples taken from D pit, Cluff Lake Mine (PNA-89W) and Collins Bay B Zone pit, Rabbit Lake Mine (PNA-CBW). Water quality objectives for Saskatchewan listed in last column.

	PNA-89W-1	PNA-CBW-1	PNA-CBW-3	PNA-CBW-10	WATER[a] OBJECTIVES
Si (ppb)	2991	7730	8224	15210	
Ti	54	51	49	55	
Al	23	112	116	256	
Fe	894	8027	2666	6181	1000
Mn	54	133	65	845	
Mg	9934	3020	2983	9364	
Ca	14820	6830	6496	22390	
Na	3070	3140	3510	3170	1×10^5
K	1044	1445	1192	3202	
P	321	1065	0	0	
S	3749	2265	1581	2118	
B	37	38	16	22	500
Ba	11	15	4	32	1000
Co	4	31	13	12	50
Cr	78	1575	536	160	20
Cu	0	15	0	0	10
Hf	<0.2	<0.2	<0.2	<0.2	
Mo	3.2	11.0	3.7	1.8	
Ni	139	1015	434	195	25/100[b]
Sr	75	33	21	195	
Ta	<0.2	<0.2	<0.2	<0.2	
Th	<0.2	<0.2	<0.2	<0.2	
U	11.5	0.4	2.4	8.9	10
V	<3	<3	<3	<3	100
Zn	3	106	8	8	50
Zr	<0.2	<0.2	<0.2	<0.2	
$^{11}B/^{10}B$	4.14	3.97	3.97	3.99	

[a] Data from Saskatchewan Environment and Public Safety (1988). Water quality objectives for B, Co, U and V based on levels for irrigation purposes (Table 4.6, p. 16-17); other data for protection of aquatic life and wildlife (Table 4.2, p. 10-11).

[b] 25 ppb where hardness ≤ 100 ppm $CaCO_3$; 100 ppb where hardness > 100 ppm $CaCO_3$.

tion, sample CB-8, contains major kaolinite and minor quartz and illite. The remaining samples are composed of major illite and chlorite (sudoite variety) with minor to trace amounts of quartz, plagioclase feldspar, kaolinite, graphite, goethite, diaspore, and illite/smectite mixed-layer minerals.

The results of major and trace element analyses for these samples are shown in Tables 2 and 3. For the aluminous gneisses from the D pit (Table 2), the SiO_2 content ranges from about 54 to 90%. Increases in alteration intensity are marked by decreases in SiO_2 content and concomitant increases in Al_2O_3 and K_2O. Fe_2O_3 and MgO contents are variable but low. Trace element concentrations for all samples are variable. Concentrations greater than 100 ppm are exhibited by Ba, Cr, La, Rb, Sr, V, and Zr only. Of the elements of environmental concern (e.g. As, Co, Cu, Ni, Mo, U, V, and Zn), only As, Ni, and V occur in amounts greater than 20 ppm. The trace element concentrations, in general, are low with respect to mineralized zones but are slightly enriched relative to regional Athabasca sandstones (see Quirt, 1985).

The sandstone samples from the B-Zone pit (Table 3) contain more than 90% SiO_2 and the gneissic samples contain from 57 to 76% SiO_2. Al_2O_3 content is less than 7% for the sandstones and greater than 11% for the gneisses. The K_2O content varies with Al_2O_3. Fe_2O_3 contents are low, with all samples containing less than about 2%. MgO content is variable with one sample, CB-4, containing 6.9%.

Trace element concentration ranges are similar to the D pit samples in Table 2. Arsenic is enriched in two samples, CB-1 and CB-2, relative to the others and Ni concentrations are highest for the graphitic gniess samples CB-4 and CB-5. Uranium contents are enriched relative to the basin background (1-2 ppm) with a few samples containing greater than 50 ppm.

The rock samples collected from the two pits show similar mineralogical and chemical characteristics within each lithological group. Concentrations of the trace elements of environmental concern are slightly enriched relative to background, regional Athabasca Group sandstones and could become a long-term source of surface water contamination. However, their concentrations are very low with respect to mineralized zones. To assess the potential leachability of the trace elements, a suite of samples was selected for detailed study. The sample suite includes both bulk powdered and clay-size material from samples PNA-89-3, -5, and -9, and PNA-CB-1, -2, -5, -6A, and -6B. These samples have been subjected to a series of partial extractions to leach metals associated with exchangeable sites, carbonates, uraninite, sulphides and organics, amorphous Fe-oxides, crystalline Fe-oxides, and residual minerals (Percival et al., 1990). Analyses of the extractant solutions are in progress.

Chemical analyses of water samples, expressed in ppb, are shown in Table 4. Saskatchewan surface water quality objectives for some elements of concern are also listed. The concentrations listed are based on permissible limits for the protection of aquatic life and wildlife or irrigation standards (Saskatchewan Environment and Public Safety, 1988). The water quality standards listed for irrigation purposes appear to be less restrictive than for aquatic life and wildlife.

For the D pit water sample, elements of concern are well below permissible levels with the exception of Cr, Ni, and U. The U concentration is slightly higher than the allowable level for irrigation.

One of the main concerns with regard to the D pit is the presence of gersdorffite in the F1 fault zone. This could be a source of As and Ni in the surface water as a result of spring and storm-event discharge. Levels above permissible limits were detected in the fault water in the spring of 1987 (Saskatchewan Environment and Public Safety, 1990).

ACKNOWLEDGMENTS

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Sphalerite-rich breccias in the footwall to the Ansil copper deposit, Noranda, Quebec

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Abstract

The flanks of the copper-rich Ansil volcanogenic massive sulphide deposit in Noranda mining camp are underlain by a series of sphalerite-rich breccias. The location of the breccia zones is related to offsets in the footwall contact believed to represent synvolcanic faults. One component of the breccia zones crosscuts the rhyolite footwall rocks at a high angle, and is overlain by a second component of stratiform breccia. The breccias are restricted in their distribution to the intersections between two sets of synvolcanic faults, which also control the morphology and distribution of the overlying massive sulphide lense.

Sphalerite-rich breccias in the footwall to the Ansil deposit formed: a) from phreato-magmatic explosions caused by the mixing of seawater and gaseous, cooling rhyolite, with metal-rich hydrothermal fluids using the breccias as conduits, or b) by overpressuring of a shallow hydrothermal aquifer.

Résumé

Les flancs du gisement Ansil, gisement de sulfures massifs d'origine volcanique riche en cuivre du camp minier de Noranda, se trouvent au-dessus d'une série de brèches riches en sphalérite. L'emplacement des zones bréchifiées est lié à des décalages dans le contact du mur qui correspondraient à des failles synvolcaniques. L'une des composantes des zones bréchifiées traverse les roches du mur rhyolitique selon un fort pendage et est recouverte par une seconde composante de brèche stratiforme. La distribution des brèches est limitée aux intersections de deux ensembles de failles synvolcaniques qui exercent un contrôle aussi sur la morphologie et la distribution de la lentille de sulfures massifs sus-jacente.

Les brèches riches en sphalérite du mur du gisement Ansil ont été formées par : a) des explosions phréatomagmatiques causées par le mélange d'eau de mer et de rhyolite gazeux se solidifiant avec des fluides hydrothermaux riches en métaux empruntant les brèches pour se déplacer, ou b) la surpression d'un aquifère hydrothermal peu profond.

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INTRODUCTION

This is a preliminary report on the nature of synvolcanic, sphalerite-rich breccias stratigraphically underlying the Ansil volcanogenic massive sulphide deposit, and their importance in understanding the formation of this deposit. This aspect is part of a larger study being carried out by the principal author to examine the paleo-hydrothermal system associated with this deposit, and to determine the nature of the fluids involved.

The Ansil Mine is in the Noranda mining camp of northwestern Quebec, 14 km north of the city of Rouyn-Noranda (Fig. 1). The deposit was first discovered in 1981 and brought into production in 1989. Pre-production ore reserves were 1.58 million tonnes grading 7.2% Cu, 0.9% Zn, 26 g/t Ag and 1.6 g/t Au (Riverin et al., 1990).

The Ansil deposit is hosted by the Upper Blake River Group in the south-central portion of the Archean Abitibi Belt. In the Noranda area the Upper Blake River Group includes the Noranda Volcanic Complex (Gibson and Watkinson, 1990) which is composed of a thick sequence of subaqueous, bimodal felsic and mafic volcanic formations that have been divided into five cycles (Spence, 1967). The Ansil deposit is within the third cycle, which hosts eleven of the fourteen massive sulphide deposits in the Noranda area. The third cycle, or Mine Sequence, is bordered to the west by the trondhjemitic, syn-volcanic Flavrian and Powell plutons and to the east by the post-volcanic Lac Dufault granodiorite.

The volcanic formations between the three plutons dip 20 to 50° to the east and are transected by several large east-northeast striking faults that were originally syn-volcanic (de Rosen-Spence, 1976). Several of these faults and their subsidiaries served as controls for volcanic eruptions and contemporaneous hydrothermal activity (Gibson, 1989; Gibson and Watkinson, 1990). Rocks have been affected by sub-greenschist grade regional metamorphism.

MINE GEOLOGY

The stratigraphy hosting the Ansil deposit has been described by Gibson (1989) and the geology of the deposit itself by Riverin et al., (1990). The deposit is at the contact between a thick sequence of massive to flow-banded rhyolite flows and minor flow breccia (Northwest Formation) and the overlying Rusty Ridge Formation massive to pillowed, basaltic-andesite flows. Also present

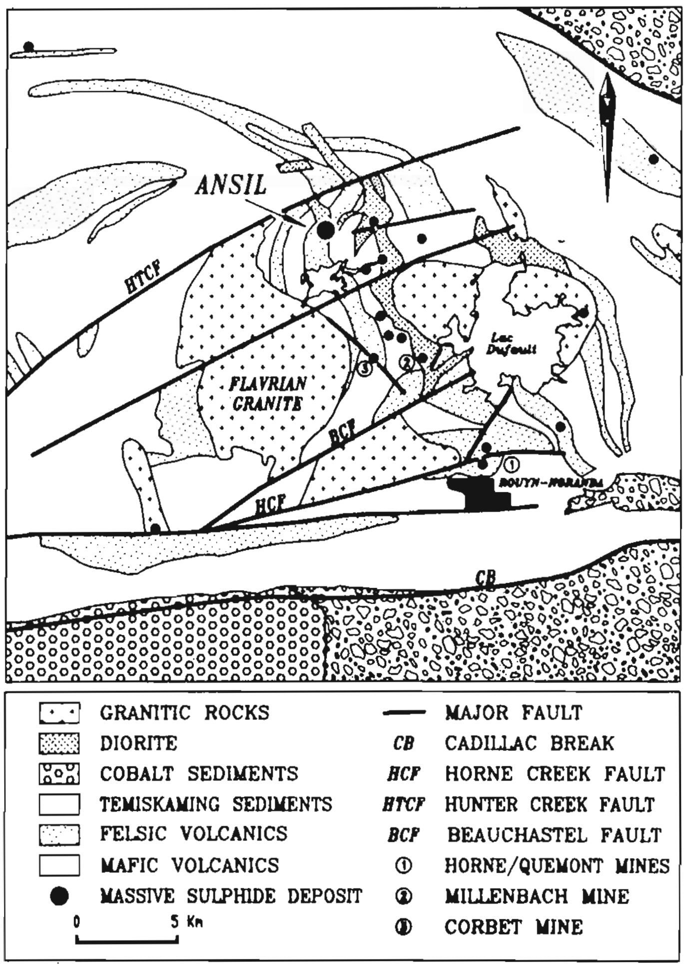


Figure 1. General geology of the Rouyn-Noranda area (from Riverin et al., 1990).

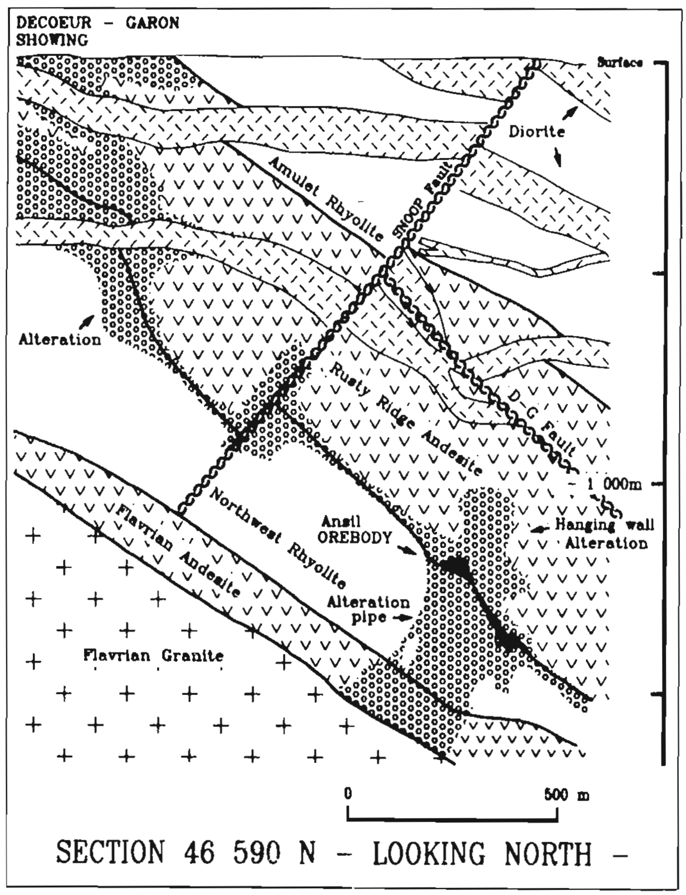


Figure 2. Section looking north through the Mine Sequence hosting the Ansil deposit (from Riverin et al., 1990).

along the mine horizon are thin, discontinuous units of interlayered chert-sulphide (Lewis Tuff) and felsic quartz-phyric pyroclastic tuff (QP Tuff).

The volcanic stratigraphy dips east at 45 to 50° and is transected by the northwest-trending D-G Fault and the east-northeast-trending Snoop Fault. The step-like configuration of the footwall to the Ansil massive sulphide may be due to north-westerly trending synvolcanic faults (Riverin et al., 1990). These two faults systems are believed to be part of a series of northwest and east-northeast synvolcanic faults recognized in the region that control the location of the volcanogenic massive sulphide deposits in the Noranda area (Watkins, 1980; Gibson, 1989). Syn-kinematic shears and faults are present, one of which controls the Northwest Formation rhyolite-Rusty Ridge Formation andesite contact, and commonly defines the upper contact of the orebody. Preliminary analysis from this study indicates that the dominant motion along this fault was reverse, with a variable component of oblique, sinistral movement.

The Ansil deposit consists of a 350 by 120 m single lense of massive sulphide that strikes north and dips 50° to the east. It is thickened at both ends to a maximum of 35 m with a thin central portion. The lense is composed of massive to coarsely-banded chalcopyrite and pyrrhotite, with thin chert layers present in the lower parts of the deposit. Both domes are underlain by up to 15 m of massive magnetite that in some mine sections also forms a core to the sulphide domes. The upper contact of the massive sulphide is overlain by a 10 to 50 cm thick, discontinuous lense of massive magnetite.

Along the upper contact of the Northwest Formation rhyolite are a number of zinc-rich breccias that underlie the north and south flanks of the massive sulphide lense. It is the description of these breccias that is the object of this paper.

SPHALERITE-RICH BRECCIAS

A zinc-rich breccia zone on the north flank of the lower massive sulphide dome has been mapped on three levels (over 180 vertical metres) at 1:150 scale (Fig. 3, 4). The location of this zone corresponds to a 40 m offset in the upper contact of the footwall rhyolite. A wedge of QP Tuff is restricted to the south side of the offset, which also defines the northern limit of the massive sulphide lense. The offset is not apparent along the overlying massive sulphide-andesite contact, suggesting it formed prior to massive sulphide deposition.

The footwall rhyolite in the vicinity of the offset is composed of massive and brecciated rock containing vein-controlled and disseminated sphalerite and pyrite (Fig. 3, 4). Massive rhyolite defines a sharp right-angle at the offset contact. Rhyolite breccias within and overlying the massive flows are divided into disconformable and conformable components (Fig. 3, 4).

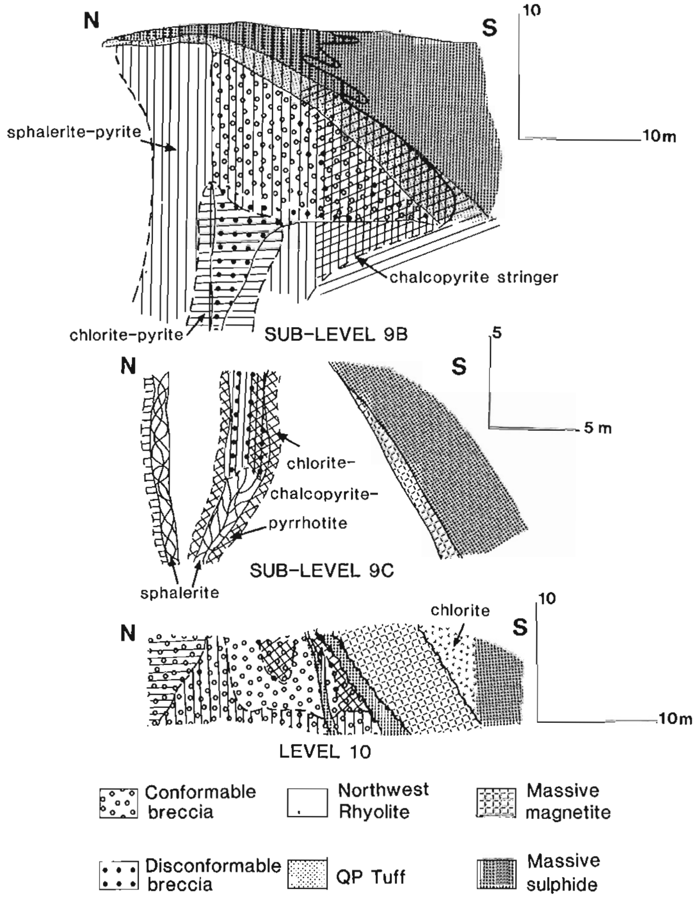


Figure 3. Plan view of the geology at the north end of the massive sulphide lense as observed on Sub-Level 9B (modified from company data).

Disconformable breccia

The disconformable breccia is principally restricted to a zone that extends into the massive rhyolite along strike of the offset contact, where it is exposed on Level 9 and Sub-level 9B (Fig. 3). The breccia is 12 m wide at its upper contact narrowing to the west for at least 20 m, forming a funnel-shaped structure. The narrow end of the breccia zone is composed of strongly chloritized, angular to rounded fragments averaging 5 to 10 mm in a fine grained quartz-chlorite matrix with 2 to 3% fine grained, disseminated pyrite (Fig. 5a).

Nearer the wide mouth of the structure, chloritized fragments become coarser, with 15 to 20% coarse pyrite cubes in the quartz-chlorite groundmass (Fig. 5b). Peripheral to the chlorite-pyrite core of the structure, the breccia also contains sericite and 5 to 10% disseminated sphalerite in both the groundmass and within breccia fragments. With the presence of sphalerite and sericite, the breccia fragments and groundmass become increasingly silicified towards the margin of the breccia. The fragments commonly have purplish, sphalerite-rich cores and white, quartz-rich margins (Fig. 5c). The transition

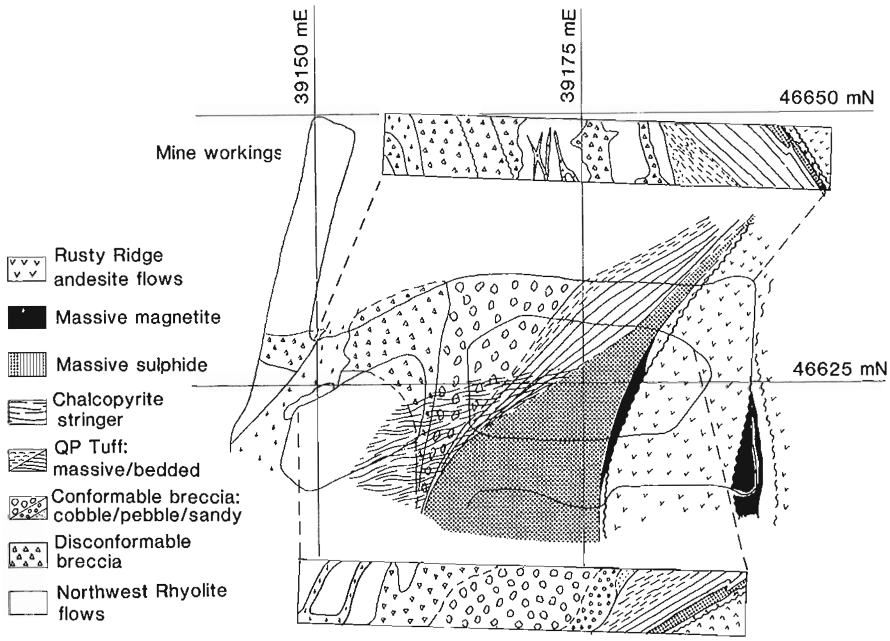


Figure 4. Schematic sections through various levels of the North Zinc Zone.

between the chlorite-pyrite and quartz-sphalerite-sericite breccias is marked by sphalerite-rich fragments in a chlorite-rich groundmass.

The massive rhyolite surrounding the disconformable breccia is strongly chloritized with irregular patches of fine grained silica giving the rock a mottled appearance. Along the north margin of the disconformable breccia the massive rhyolite is quartz-sericite-sphalerite-pyrite-rich with 10 to 15% disseminated pyrite and 5 to 7 mm patches of fine grained sphalerite in a strongly bleached, creamy white rock.

Along the massive rhyolite contact that defines the footwall offset there are several disconformable breccia zones 50 to 100 cm wide containing strongly bleached, sphalerite-pyrite rich fragments in a groundmass of semi-massive sphalerite-pyrite with sericite and quartz (Fig. 3, 5c). Many of the smaller fragments within these breccia zones are strongly angular, with convex faces (Fig. 5d). The larger fragments are strongly silicified and display a variety of textures from massive with perlitic fractures to banded. The banding is formed by discontinous, quartz-rich streaks. The two rock types are randomly oriented and intermixed.

On Sub-level 9B (SL 9B) the sphalerite-rich, disconformable breccias were internally disrupted, with fragments clearly displaced and shape of their margins bearing no physical relation to surrounding fragments. On Sub-level 9C (SL 9C) there are two disconformable breccia zones (Fig. 4). One is similar in texture to those described on SL 9B, whereas the other consists of the in situ fragmentation of massive rhyolite by a series of sphalerite-quartz-pyrite-rich veins (Fig. 6a). The veins resemble chalcedony with their cherty, fine grained texture and fine banding of quartz- and sphalerite-rich layers. The massive rhyolite within the vein system is purple due to a large concentration of very fine grained sphalerite. This sphalerite-rich stringer zone has a 10 to 15 cm wide margin rich in chlorite and pyrite.

The other breccia zone on SL 9C consists of both in situ and disrupted breccia. Purple sphalerite-rich fragments with their distinctive white rims are cut by thin chalcopyrite-pyrrhotite-chlorite veins; the margin of the breccia has a halo of chlorite, pyrrhotite and chalcopyrite (Fig. 4).

Conformable breccia

Overlying the funnel-shaped disconformable breccia on SL 9B is a 15 by 15 m, wedge-shaped stratiform breccia that is bounded to the north by the offset footwall contact, and thins to the east and south (Fig. 3). The breccia is crudely graded, with a 10 to 12 m wide base of poorly sorted, coarse breccia, overlain by up to 3 m of moderately- to well-sorted pebble breccia (Fig. 3). Overlying the pebble breccia is a small wedge of coarse, sandy material.

The conformable breccia is composed of strongly silicified and sericitized fragments with purplish, sphalerite-rich cores and white rims (Fig. 6b). The groundmass is quartz, sphalerite, pyrite and sericite, with minor chlorite. The southern edge of the breccia is overprinted by a chalcopyrite-pyrrhotite vein system that is part of the disconformable sulphide stringer zone below the lower suphide dome. The overprinting results in a distinctive breccia containing purple, sphalerite-rich fragments in a chalcopyrite-pyrrhotite-rich groundmass (Fig. 6c). The same phenomenon is observed to a lesser extent within the conformable breccia on Level 10 (Fig. 4).

Hangingwall relationships

The North Zone sphalerite-rich breccia and massive rhyolite are overlain by a southward-thinning wedge of massive to well-layered QP Tuff. The well-layered part of the tuff contains 10 to 20% disseminated sphalerite. The QP Tuff is in turn overlain by the massive sulphide lense, and there is interlayering of the tuff's finely layered, cherty

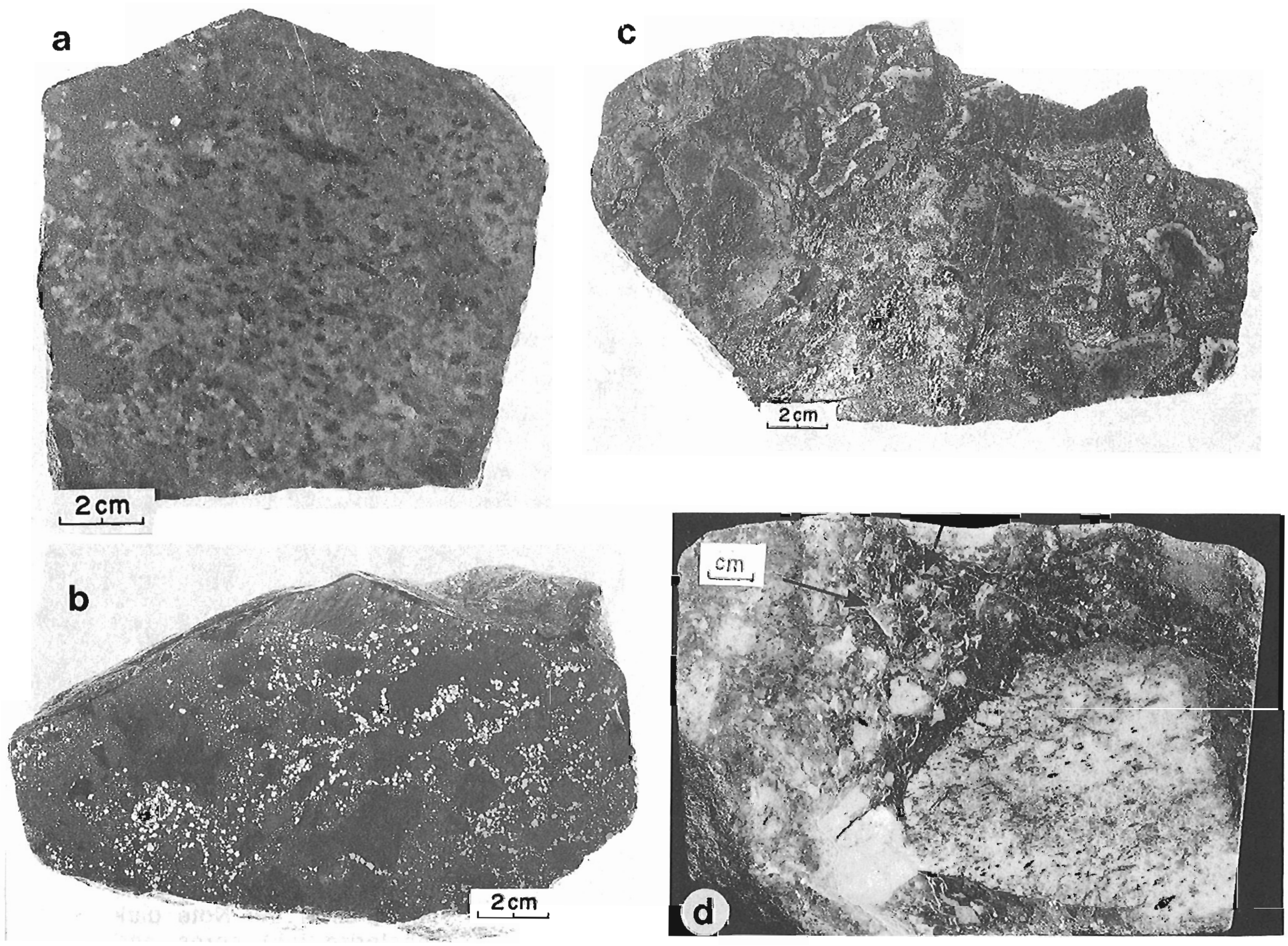


Figure 5. a) Chlorite-pyrite rich microbreccia from west end of the disconformable breccia zone on Sub-Level 9B. b) Chlorite-pyrite rich breccia from the top of the disconformable breccia zone on Sub-Level 9B. c) Sphalerite-quartz-sericite-pyrite rich breccia from small disconformable breccia zone on Sub-Level 9B. Note dark cores and white rims to fragments. d) Sphalerite-quartz-sericite-pyrite rich breccia from small disconformable breccia zone on Sub-Level 9B. Note angular fragments with convex faces in groundmass (arrows).

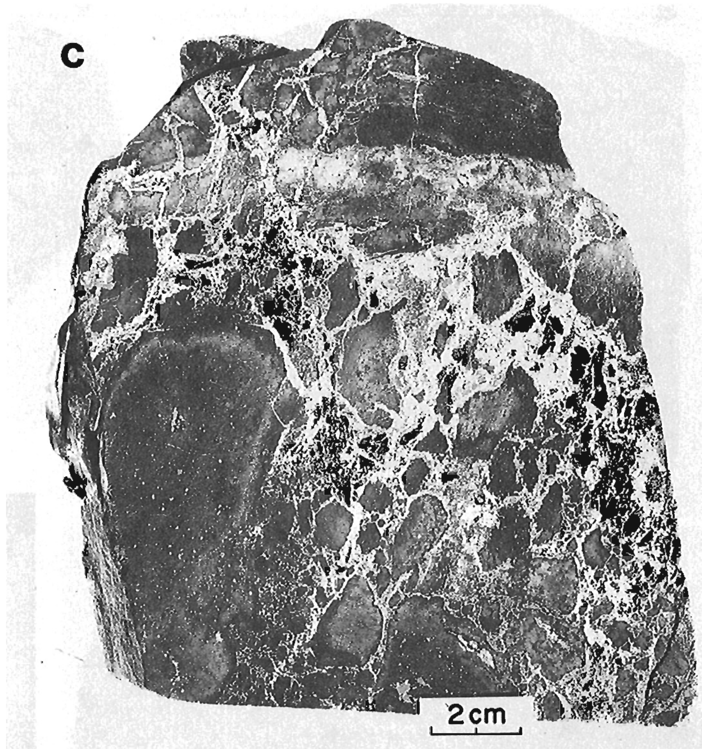
top and massive sulphide. Directly above the sphalerite mineralized breccias, the massive sulphide lense contains interlayered sphalerite and chalcopyrite-pyrrhotite.

Below Level 10 the sphalerite content of the well-layered massive sulphide lense increases significantly. Diamond drilling has revealed an extensive, sphalerite-rich breccia directly below the northeast flank of the lense, indicating the presence of another zinc-rich footwall zone.

DISCUSSION

The relationship of disconformable and conformable, zinc-rich breccias with offsets in the upper contact of the Northwest Formation rhyolite suggests that they are related to synvolcanic faulting that was coeval with massive sulphide deposition. The breccias have many of the characteristics of phreato-magmatic explosion structures. These characteristics include the funnel shape of the disconformable zones, and the change towards the upper contact of the rhyolite from in situ to disrupted breccias in which fragments of differing textures are intermixed, suggesting transport. The funnel shape of the breccia zone is typical of breccias formed from a rapidly expanding medium, such as a boiling fluid, approaching the volcanic surface. The presence of small, shard-shaped fragments and perlitic textured larger fragments is also characteristic of phreato-magmatic breccia zones (Gibson and Watkinson, 1990). The distinctive white rims on the sphalerite-rich fragments within the disrupted breccias are similar to the vein margins observed in the vein systems forming the lower part of the breccia zones.

The conformable breccia overlying the disconformable breccia zone may have formed in several ways. If the disconformable breccia formed from a rapidly expanding fluid or from a shallow, overpressured aquifer, it would explosively breach the rock-seawater interface resulting in a pile of debris being deposited directly above the breccia pipe on the rhyolite surface. This is a similar model to that proposed for the formation of the Buttter-cup Hill breccias (Gibson and Watkinson, 1990), which are located several kilometres southwest of Ansil.



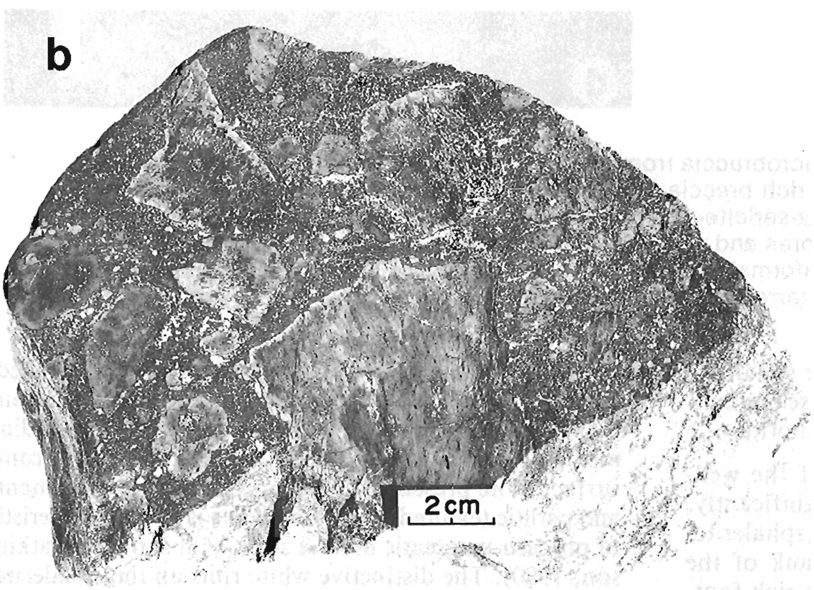


Figure 6. a) Coarse breccia from conformable breccia zone on Sub-Level 9B. Note dark (sphalerite-rich) cores and white rims to fragments. b) Sphalerite-rich vein system in Zinc North Zone on Sub-Level 9C. Further north of two breccia zones exposed on this sub-level. c) Chalcopyrite-pyrrhotite veins in conformable breccia zone on Sub-Level 9B. Darker fragments have sphalerite-rich cores.

Alternatively the conformable breccia may have formed from slope instability along the faulted rhyolite surface. This latter model would explain the grain-size grading within the conformable breccia, a phenomenon more common in debris flows than phreato-magmatic breccias.

Examination of 1:250 scale mine plans and sections has revealed that the distribution of the zinc-rich breccias is localized near the intersection of the northerly trending step-faults (Riverin et al., 1990) and a series of east-west trending offsets along the upper contact of the footwall rhyolite (Fig. 7). The east-west offsets are only present where the upper contact of the rhyolite has a relatively shallow dip. It is proposed that these offsets represent synvolcanic faults, and the restriction of their offset to the shallow-dipping rhyolite contacts is due to their normal, vertical movement (Fig. 7). The proposed east-west trending faults parallel the down-plunge axis of the massive sulphide lense.

The rhyolite contact indicates that paleo-topography drops off to the east and south, which is compatible with normal displacement along both the north-south step faults and the proposed east-west faults. This is in agreement with Gibson's (1989) model showing the Ansil deposit situated along the southeast flank of the Northwest Rhyolite.

There is a direct spatial relationship between the distribution of sphalerite in the massive sulphide lense and the zinc-rich breccia zones. The exception to this is the Zinc South Zone which lies along the south flank of the upper sulphide dome (Fig. 7). Its relationship with the massive sulphide lense is obscured by a syn-kinematic, oblique-reverse fault that truncates the upper contact of the zinc-rich breccia. If the breccias were channelways for fluids depositing zinc contemporaneously with copper from the larger stringer zone, this indicates massive sulphide formation from two fluids. This is contrary to the present models developed from the study of active seafloor sulphide mounds which attempt to explain metal zonation in individual sulphide chimneys by the fluctuation of fluid temperature within the hydrothermal vent (Paradis et al., 1988). In agreement with the modern seafloor model is the observation at Ansil that the quartz-sphalerite-sericite-pyrite mineralogy within the breccia zones is overprinted by chlorite-pyrite, chlorite-pyrrhotite-chalcopyrite assemblages, suggesting an evolving fluid system in which early, low temperature zinc mineralization is replaced by a copper-rich mineral assemblage. Systematic, detailed petrography, whole-rock and isotope geochemistry is presently underway to study this relationship.

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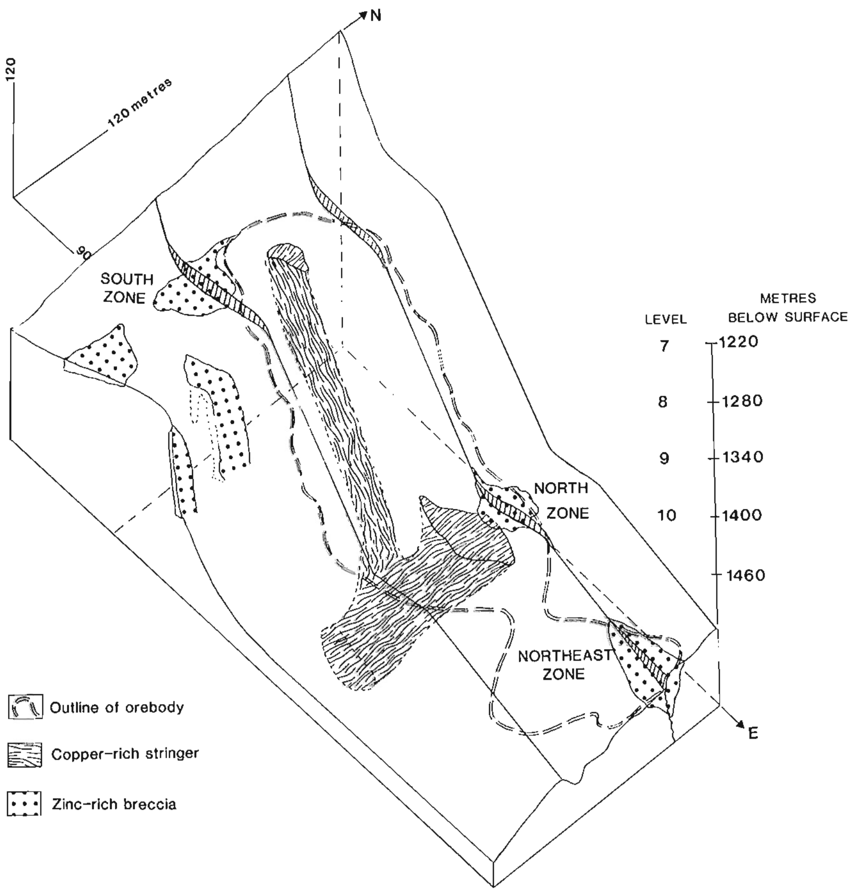

Figure 7. Isometric diagram showing the surface of the footwall rhyolite and distribution of zinc zones.

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Profiles of altered zones at ca 2.45 Ga unconformities beneath Huronian strata, Elliot Lake, Ontario: evidence for early Aphebian weathering under anoxic conditions

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Abstract

Petrographic studies and chemical analyses of samples representing complete drillcore sections through 4 altered zones on basalt beneath ca 2.45 Ga subarkose of the Matinenda Formation, Huronian Supergroup, at Quirke II uranium mine support previous conclusions that these ubiquitous, unconformity-related, altered zones were developed through subareal weathering under anoxic conditions. Pedological features (varieties of cutans) can be discerned in upper parts of the profiles. There is evidence of erosion of the top surface and of compaction. Analyses are given for many more elements than were considered in previous studies. Carbonate enrichment accompanying Fe, Mn, Mg and K depletion is characteristic of basal sections of the zones. In one case extreme carbonate enrichment is overlain by local restoration of Fe and Mg. Topmost sections of the paleosols show the most extreme variations, notably: losses of Na, Ca, Sr, Mg, Mn, Fe; gains of K, Rb, Ba, Be, S, U, and perhaps Au.

Résumé

Des études pétrographiques et des analyses chimiques d'échantillons, représentant des sections complètes de carottes de forages traversant quatre zones altérées sur le basalte sous le grès arkosique datant d'environ 2,45 Ga de la formation de Matinenda du supergroupe d'Huronien à la mine d'uranium Quirke II, appuient des conclusions antérieures à l'effet que ces zones altérées, présentes partout et reliées à une discordance se sont formées par météorisation subsuperficielle dans des conditions anoxiques. Des entités pédologiques (variétés de cutanes) peuvent être discernées dans les parties supérieures des profils. Il y a des indications d'érosion de la surface supérieure et de compaction. Des résultats d'analyses sont présentés pour un nombre beaucoup plus élevé d'éléments que celui traité par les études antérieures. Un enrichissement en carbonate accompagnant l'appauvrissement en Fe, en Mn, en Mg et en K est caractéristique dans les sections basales des zones. Dans un cas extrême, l'enrichissement en carbonate est sous-jacent à une restitution par endroits du Fe et du Mg. Les sections supérieures des paléosols présentent les variations les plus considérables, notamment : pertes en Na, Ca, Sr, Mg, Mn, Fe et gains en K, Rb, Ba, Be, S, U et peut-être en Au.

INTRODUCTION

Coarse subarkose and quartz-pebble conglomerate beds of the Matinenda Formation, host of the Elliot Lake area uranium deposits, overlie altered zones developed on ca 2.7 Ga metavolcanic, metasedimentary, and granitic rocks and also on ca 2.45 Ga Huronian basalt flows. The alterations exhibit upward gradations to argillite and have been studied by several workers all of whom considered they represent paleoweathering profiles (Roscoe, 1957, 1969, 1973, 1981; Roscoe and Steacy, 1958; Pienaar, 1963; Robertson, 1964; Frarey and Roscoe, 1970; Fryer, 1977; Gay and Grandstaff, 1980; Kimberley et al. 1984; Goddard, 1987; G-Farrow and Mossman, 1988). Most of the profiles studied show chemical changes similar to those found in younger soil profiles with some exceptions — most notably, an upward loss of iron and a lack of significant enrichment of ferric iron relative to ferrous iron. This has been considered to indicate that the weathering occurred under reducing conditions (Pienaar, 1963 and later workers cited above). Robertson (1964) suggested that "such a reduction may have been due to exclusion from the atmosphere by the overlying material or to an atmosphere deficient in oxygen". Since the paleosols are directly overlain nearly everywhere by fluvial sediments, it is unlikely that they could have formed in environments effectively protected from interactions with atmospheric gases. Oxygen deficiency in the early Aphebian atmosphere is also indicated by lack of ferric iron oxide minerals and presence of authigenic pyrite and detrital pyrite and uraninite in clastic sediments in the lower Huronian Elliot Lake and Hough Lake groups. This contrasts with the Cobalt Group (uppermost Huronian), which contains redbeds and black sand heavy mineral concentrations (Roscoe, 1969, 1973, 1981) and which has been found overlying granite weathered under oxidizing conditions (Rainbird et al., 1990).

Problems have arisen in interpretations of available data on the early Huronian paleosols. Gay and Grandstaff (1980) and Kimberley et al. (1984), for example, reported increases in iron and/or ferric : ferrous iron ratios in some granitic paleosols, whereas earliest analyses of granitic paleosols by Pienaar (1963) and Robertson (1964) indicated the opposite. Variations have also been found in paleoweathering profiles of mafic rocks. These differences could be due to many factors, such as inhomogeneities in the protoliths, differing rates of weathering and erosion, post-weathering diagenetic, metamorphic or metasomatic alterations, post-Huronian weathering, insufficient samples, inadequate control of stratigraphic position of samples, the presence of corestones, and analytical errors. Three profiles are very well described by G-Farrow and Mossman (1988), and variations of components are illustrated graphically, but the raw analytical data are not given, nor is the source of these data (Goddard, 1987) referenced. Other papers provide minimal information on minerals and textures in the altered zones. Analyses given by Kimberley et al. (1984) for two profiles include data only on Al, K, Na, Fe, Mn, Co, and Eu contents. Other published reports provide only data on major elements but nothing on contents of CO_2, S, U, Th, REE, Au, base metals, other minor and trace elements, or isotopic variations.

The ancient weathering profiles should provide important information on sources of elements in suprajacent pyritic uraniferous paleoplacers and on evolution of the atmosphere and attendant changes in geochemical processes relevant to the formation of a number of types of ore deposits. A much more extensive database is required, however, if this potential is to be realized. Imminent closures of several Elliot Lake uranium mines spurred the writers to examine underground exposures of unconformities and obtain samples through paleoweathered protoliths. In this interim report we present data on drillcore samples through 4 paleosol sections from underground drillholes in the Rio Algom Limited Quirke II Mine.

GEOLOGICAL SETTING

Quirke II mine is at the north end of the extensive south-dipping conglomeratic uranium ore zone at Quirke Lake. Quirke II mine and the Elliot Lake ore zones are at north and south sides of Quirke Lake Syncline. General geology of the area and sources of detailed information have been reviewed most recently by Robertson (1986, 1987).

The several conglomeratic beds of the Quirke Lake ore zone are within the Manfred Member (Roscoe, 1981) of the Matinenda Formation. This unit of the Huronian Supergroup overlies pre-Huronian granitic rocks and metavolcanic rocks to the east and basalt flows of the Huronian Dollyberry Formation (Robertson, 1971) to the west. The Dollyberry lavas and the Matinenda Formation are considered to be approximately coeval with rhyolite of the Coppercliff Formation which has been dated (U-Pb zircon) at 2450 +25/−10 Ma (Krogh et al., 1984). The youngest dated pre-Huronian (Archean) rock near the area is syenite at East Bull Lake which has yielded a U-Pb zircon date of 2665 +1.6/−1.4 Ma (Krogh et al. 1984). The oldest dated post Huronian rock is Nipissing gabbro dated (U-Pb baddeleyite) at 2217.5 +/−1.6 Ma (Andrews et al., 1986). Huronian strata were folded prior to intrusion of Nipissing gabbro and were probably deposited during a pulsed uplift-downwarp (Frarey and Roscoe, 1970) event that could have occurred over several tens (rather than hundreds) of million of years.

Archean rocks were metamorphosed to greenschist facies prior to deposition of Huronian strata. The latter are mainly at sub-greenschist facies in Quirke Lake Syncline but were altered locally at contacts with gabbro intrusions and contain greenschist minerals in areas where they are most deformed — presumably during folding related to Penokean Orogeny at ca 1.85 Ga. Whole rock and mica K-Ar dates are dominantly about 1.6 Ga.

SAMPLES AND ANALYSES

Data presented herein are from four underground drillholes from Quirke II Mine that intersected basal Matinenda subarkose and radioactive pyritic quartz pebble conglomerate, the basal unconformity and underlying basement rocks. Drill sections from least altered basement rocks through increasingly altered zones to the base of Matinenda strata were sampled systematically for chemical and petrographic studies. Cores were sawn lon-

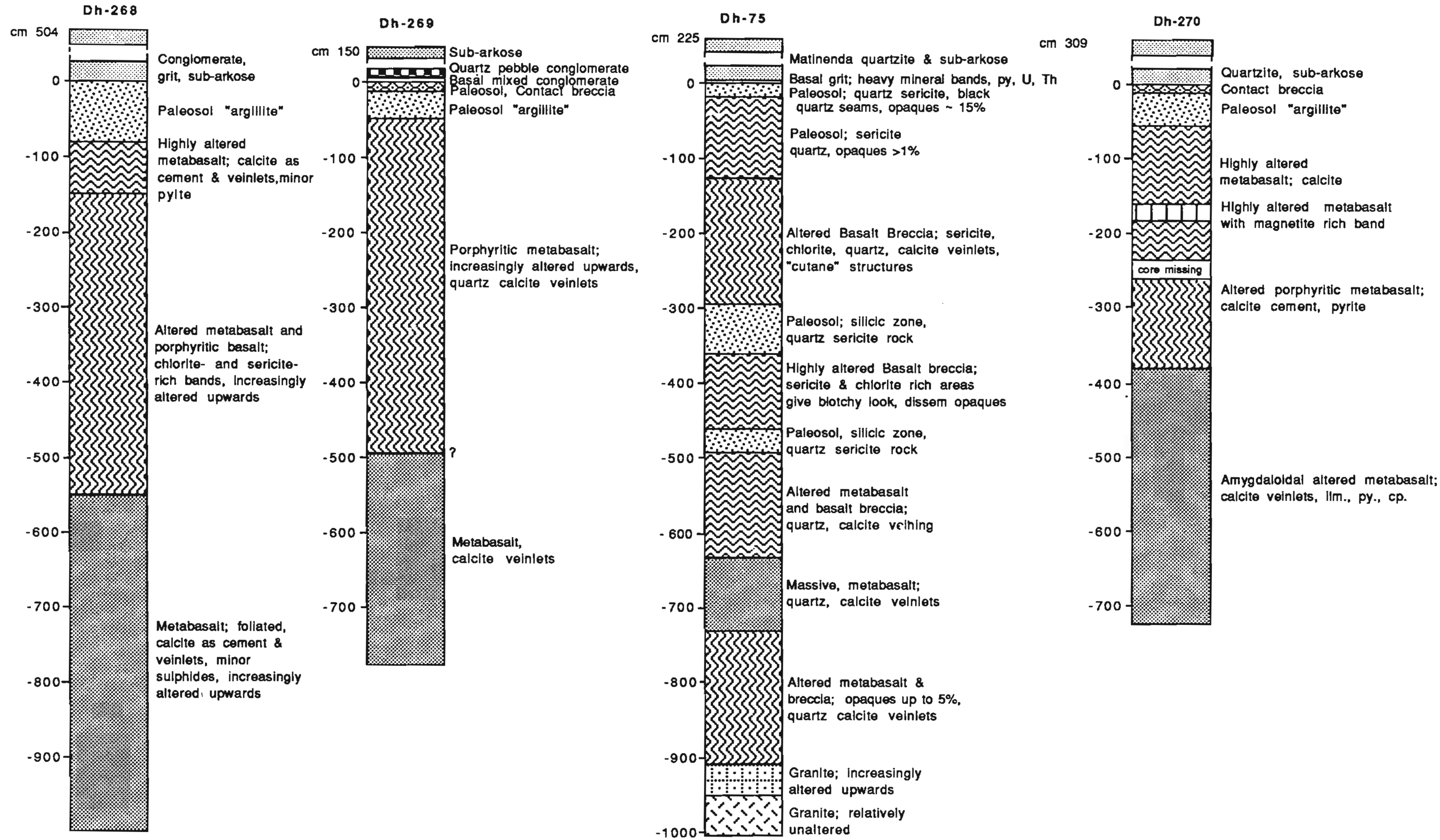

Figure 1. Lithologs of Quirke II drillholes. 0 cm is the unconformity surface between basement and the Huronian sediments.

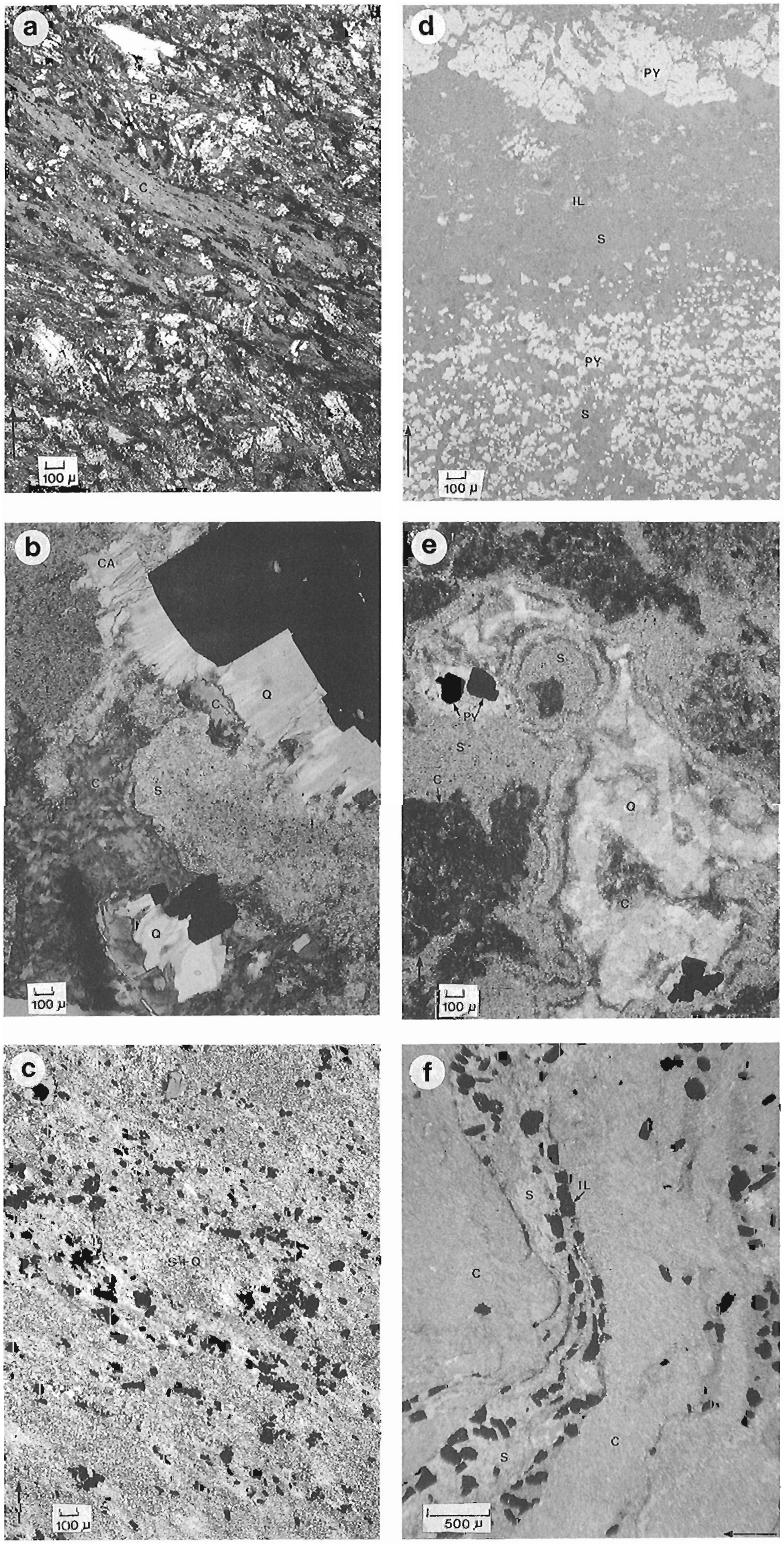

Figure 2. Paleosols showing various degree of alteration and fabric. All photos are under polarized light (except 2d). The arrow indicates stratigraphically upwards direction; Q-quartz, Qa-quartz aggregate, P-plagioclase, C-chlorite, CA-calcite, S-sericite, PY-pyrite, IL-ilmenite.

(a): DH-268-33 (−945cm), Metabasalt, foliated volcanic texture preserved, mafics altered to chlorite but relatively unaltered.

(b): DH-75-21 (−513cm), secondary quartz and minor calcite growths on pyrite cubes as cavity filling in stress shadows.

(c): DH-268-49 (−7cm), paleosol, extreme alteration, laminations are all destroyed, very fine grained sericite rich rock ("argillite"), opaques are mostly ilmenite, few pyrite grains.

(d): DH-75-31(0 cm), top of the paleosol in contact with Matinenda subarkose (not in the photo) with pyrite clusters (concretions ?) at the very contact. Fine grained disseminated ilmenite in lower 1/3 of the photo tightly packed pyrite crytals in sieve-like texture (in reflected light.)

(e): DH-75-23 (−378 cm), cutane structure in highly altered metabasalt.

(f): DH-75-27 (−12 cm), paleosol, intensely altered, showing pedogenic fabric, ilmenite in sericite rich bands.

Table 1. Chemical analyses of Quirke II drill core samples.

a — DH-269

SAMPLE	1	2	3	4	5	6	7	8	9	10	11
DEPTH *(cm)*											
TO	-715	-583	-463	-325	-140	-43	-15	-10	-5	5	10
FROM	-775	-715	-583	-463	-325	-140	-43	-15	-10	-5	5
Percent											
SiO_2	44.8	43.6	45.9	46.0	44.1	43.2	42.7	43.5	45.1	60.8	56.6
TiO_2	2.37	2.48	2.20	2.39	2.11	4.05	3.31	4.63	4.60	1.92	2.71
Al_2O_3	17.1	16.8	15.9	15.7	14.6	27.3	25.6	28.1	26.4	19.5	20.1
Fe_2O_3	1.6	6.5	1.3	0.9	1.1	2.5	2.4	2.8	2.9	1.9	1.9
FeO	16.2	12.0	11.9	9.0	9.0	5.4	7.5	4.0	1.8	1.3	2.4
MnO	0.24	0.25	0.23	0.22	0.24	0.20	0.15	0.18	0.03	0.02	0.09
MgO	2.35	2.33	1.61	1.45	1.59	1.17	1.69	0.97	0.98	0.63	0.65
CaO	2.18	2.40	5.14	8.23	10.96	1.24	2.15	1.22	3.30	1.55	1.74
Na_2O	2.9	1.5	2.7	4.1	4.3	0.1	0.1	0.2	0.1	0.1	0.2
K_2O	2.92	4.11	4.47	2.84	1.48	9.63	8.1	10.07	9.62	9.00	9.26
H_2O^t	4.7	4.8	3.3	2.8	3.0	4.1	4.7	3.9	4.0	2.4	3.0
CO_2	0.4	0.9	3.3	5.0	7.1	0.0	0.0	0.0	<0.1	<0.1.	<0.1
C	0.1	0.1	0.1	0.0	0.0	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
P_2O_5	0.51	0.48	0.43	0.44	0.39	0.49	0.63	0.17	0.16	0.29	0.68
S	0.15	0.13	0.13	0.07	0.02	0.02	0.01	0.0	0.07	0.21	0.18
Total	98.7	98.7	98.9	99.4	100.1	100.0	99.6	100.3	99.8	100.3	100.2
ppb											
Au	<1	2	<2	<2	<5	<1	<1	<7	<10	n.a.	n.a.
ppm											
Ba	2241	2812	2507	1469	1012	4827	4005	4969	4636	3613	3520
Be	1.7	2.1	1.4	1.4	1.5	10	8.5	10	15	6.4	6.5
Co	50	55	48	44	41	24	32	14	70	27	24
Cr	260	290	270	290	270	520	440	580	620	120	280
Cu	150	170	170	130	80	110	10	1	57	43	25
La	41	38	47	46	46	150	180	170	410	190	450
Ni	200	190	150	140	120	62	79	23	130	27	45
Pb	0	0	0	3	1	2	1	1	42	35	76
Rb	71	121	132	76	44	448	401	488	471	366	379
Sr	101	84	140	191	289	80	72	51	73	40	64
Th	4.8	4.6	5.1	5.6	4.7	8.5	6.2	10.0	26.0	n.a.	n.a.
U	0.9	1.6	0.9	1.6	2.3	2.6	3.9	7.3	16.4	n.a.	n.a.
V	250	290	230	170	180	490	450	490	650	240	220
Y	69	72	74	74	34	47	57	61	118	64	103
Zn	130	130	110	90	95	51	71	24	48	24	54
Zr	312	337	293	331	305	554	477	642	718	1010	1585

n.a. - not analysed

b — DH-75

SAMPLE	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
Height *(cm)*																		
TO	-951	-920	-912	-807	-732	-632	-527	-485	-463	-364	-299	-184	-147	-120	-72	-17	0	4
FROM	-1007	-951	-920	-912	-807	-732	-632	-527	-485	-463	-361	-299	-184	-147	-120	-72	-17	0
Percent																		
SiO_2	67.1	69.7	71.5	56.4	51.1	44.6	42.4	61.7	79.4	48.7	73.2	54.1	50.9	49.1	46.8	44.5	44.3	73.8
TiO_2	0.32	0.25	0.26	1.69	2.18	2.79	2.64	1.52	0.74	2.11	0.3	1.57	2.38	2.59	3.68	4.12	4.59	1.04
Al_2O_3	15.0	12.3	9.7	12.7	13.9	17.2	16.6	10.9	7.6	14.0	6.9	12.4	14.7	18.6	28.2	31.4	31.5	12.0
Fe_2O_3	0.5	0.9	(9.8)	1.5	(17.5)	1.2	2.0	(13.8)	0.5	(18.3)	1.1	1.5	(18.9)	(14.8)	1.7	1.6	1.2	(4.2)
FeO	4.6	5.8		10.9		16.2	17.9		4.5		7.4	15.3			1.5	0.9	2.0	
MnO	0.04	0.03	0.04	0.09	0.09	0.12	0.12	0.08	0.04	0.12	0.06	0.11	0.09	0.04	0.00	0.0	0.00	0.00
MgO	1.79	2.22	2.59	3.08	2.89	3.36	3.53	1.98	0.92	2.81	1.55	3.25	3.38	2.15	0.94	0.84	0.83	0.28
CaO	0.99	0.30	0.54	3.27	1.89	1.63	1.72	2.57	1.40	3.39	2.61	1.95	0.85	0.72	1.45	0.44	0.20	0.09
Na_2O	4.3	1.5	0.4	0.1	0.2	0.4	0.4	0.1	0.1	0.5	0.1	0.1	0.1	0.1	0.1	0.2	0.2	0.1
K_2O	3.05	4.19	3.16	3.10	3.99	4.38	3.34	2.51	2.37	3.42	1.07	1.43	2.35	5.24	9.68	10.74	10.60	6.20
H_2O^t	2.1	2.5		4.0		5.1	5.6		1.8		2.6	5.2			3.8	4.1	4.0	
CO_2	0.2		0.1	0.9	0.7	0.5	0.6	1.4	0.8	1.9	1.9	0.9						
C	0.1		0.1	0.1	0.1	0.0	0.1	0.0	0.1	0.1	0.1	0.1						
P_2O_5	0.12	0.10	0.08	1.31	0.50	0.56	0.54	0.38	0.13	0.44	0.04	0.40	0.47	0.37	0.95	0.30	0.13	0.05
S	0.03	0.13	1.33	0.85	2.36	0.35	0.82	1.68	0.25	1.00	0.70	0.27	1.29	2.98	0.25	0.09	0.82	2.82
Total	100.4	100.2	99.7	100.2	97.6	98.7	98.4	98.8	100.8	96.9	99.7	98.6	95.7	97.1	99.7	99.6	100.7	101.0
ppm																		
Ba	983	1509	1283	1139	1541	1618	1215	897	764	1326	418	515	773	1544	2589	2661	2405	1327
Be	0.9	1.3	0.8	3.1	2.2	3.3	2.7	2.1	1.8	1.6	1.1	1.9	4.1	6.9	8.7	9.5	10	3.0
Co	17	26	49	56	70	42	67	77	24	95	59	62	100	210	21	6	16	38
Cr	150	140	120	71	67	79	76	61	54	70	49	150	360	190	510	68	72	38
Cu	11	5	11	30	82	160	67	100	24	60	53	150	440	1000	300	0	0	74
La	21	10	11	39	54	80	78	57	24	72	42	25	70	63	150	32	160	58
Ni	52	53	64	70	54	81	86	75	31	100	65	230	360	500	150	48	18	17
Pb	3	4	12	6	9	4	4	1	1	0	2	0	4	4	0	0	7	23
Rb	81	140	90	144	164	207	181	138	134	141	55	66	135	315	467	461	457	214
Sr	195	71	51	68	63	56	41	43	57	68	33	73	49	53	32	67	57	32
V	59	78	85	180	230	280	270	180	110	220	55	180	260	310	420	420	440	81
Y	23	16	32	38	38	44	39	75	66	37	33	60	48	41	57	53	63	30
Zn	20	21	30	50	130	98	150	120	140	140	140	130	100	100	65	38	130	42
Zr	133	113	122	310	320	419	363	257	203	333	241	282	346	428	577	668	699	789

Table 1. Chemical analyses of Quirke II drill core samples. (cont'd.)

c — DH-268

SAMPLE	33	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50
DEPTH *(cm)*																		
TO	-896	-796	-696	-596	-496	-396	-296	-196	-156	-121	-101	-76	-65	-30	-19	-7	6	25
FROM	-996	-896	-796	-696	-596	-496	-396	-296	-196	-156	-121	-101	-76	-65	-30	-19	-7	6
Percent																		
SiO_2	48.3	49.1	47.9	50.2	46.5	48.3	48.1	46.6	43.8	36.7	38.7	51.6	40.0	42.2	38.3	42.9	50.5	57.4
TiO_2	2.01	1.91	1.90	1.89	1.92	1.89	1.89	1.92	2.26	2.46	2.35	2.72	4.07	4.34	4.05	4.52	3.15	3.00
Al_2O_3	13.5	13.2	12.7	12.9	13.0	12.6	12.6	13.0	13.6	17.8	17.3	17.7	26.7	27.9	25.7	29.6	25.4	22.0
Fe_2O_3	2.1	2.0	2.1	1.9	2.4	1.4	1.3	1.5	1.6	1.7	(24.9)	2.9	3.0	2.9	(13.1)	2.6	2.4	2.3
FeO	10.4	10.3	10.2	10.5	10.0	10.5	11.0	10.7	12.2	16.6		11.8	8.1	5.4		4.8	3.2	2.5
MnO	0.28	0.29	0.26	0.24	0.28	0.37	0.39	0.37	0.31	0.33	0.25	0.20	0.21	0.17	0.16	0.18	0.12	0.10
MgO	5.11	4.91	4.82	4.26	2.92	4.74	4.11	2.53	1.31	1.64	1.77	1.31	1.02	0.76	0.65	0.69	0.61	0.55
CaO	5.57	6.07	6.26	6.07	8.29	6.99	7.44	9.01	8.88	7.21	1.95	0.79	1.00	1.26	0.83	0.21	0.37	0.63
Na_2O	3.6	3.4	3.1	3.1	3.8	2.8	2.8	3.7	4.0	3.4	0.7	0.2	0.2	0.2	0.2	0.2	0.2	0.2
K_2O	1.41	1.49	1.66	0.76	0.59	0.89	0.39	0.35	0.62	1.83	3.05	4.51	9.06	9.95	9.02	10.18	9.38	8.04
H_2O^t	3.8	3.8	3.9	4.1	3.5	4.2	4.3	3.5	3.4	4.6	<0.1	4.5	4.5	3.9	<0.1	4.2	3.6	2.9
CO_2	2.4	3.1	3.5	3.0	4.7	4.0	4.2	5.9	6.0	4.8	0.6	<0.1	>0.1	<0.1	<0.1	<0.1	<0.1	<0.1
C	0.1	0.1	0.1	0.1	0.1	0.0	0.0	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
P_2O_5	0.42	0.34	0.32	0.32	0.48	0.33	0.32	0.33	0.36	0.42	0.45	0.48	0.68	0.79	0.50	0.12	0.19	0.13
S	0.02	0.04	0.06	0.05	0.12	0.04	0.03	0.06	0.08	0.39	3.81	0.41	0.09	0.19	5.30	0.01	0.02	0.01
Total	99.1	100.0	98.8	99.4	98.7	99.1	99.0	99.5	98.5	100.1	96.1	99.6	99.3	100.5	98.3	100.7	99.8	100.2
ppb																		
Au	17	10	15	11	6	23	5	<2	<4	4	180	28	40	11	190	25	7	n.a.
ppm																		
Ba	345	354	263	264	355	234	225	280	451	1188	1718	2371	4424	5325	4842	5038	4336	3655
Be	1.8	1.9	2.1	1.8	1.6	2.0	1.5	1.7	1.4	2.4	3.2	4.5	8.2	8.4	7.8	8.6	7.4	6.5
Co	52	55	56	52	60	54	52	52	68	120	200	82	48	42	1800	23	25	16
Cr	260	260	260	250	260	240	240	240	260	300	260	330	500	510	410	550	160	180
Cu	240	240	230	220	220	200	220	210	180	410	13000	1000	15	130	27	6	20	21
La	39	38	41	39	37	37	37	32	42	56	36	40	62	96	54	98	200	180
Ni	140	150	150	150	160	150	150	140	150	200	400	150	130	89	310	55	76	35
Pb	27	33	52	42	15	51	13	33	11	9	12	25	11	10	16	10	14	17
Rb	134	139	180	73	66	90	59	28	39	81	145	219	460	473	409	451	403	323
Sr	169	171	173	155	197	183	178	202	198	163	69	35	60	39	52	38	65	43
Th	3.9	4.1	4.1	3.6	3.5	3.6	3.9	4.7	4.4	5.3	4.4	5.1	8.7	9.2	9.1	8.3	50.0	n.a.
U	0.8	1.0	1.0	0.7	1.1	0.9	1.1	0.7	0.9	1.2	1.3	1.2	2.1	2.7	3.3	4.5	18.6	n.a.
V	220	220	220	210	220	210	210	210	210	270	270	300	450	450	420	470	380	340
Y	70	74	55	66	64	69	61	32	38	40	50	40	45	65	53	35	44	47
Zn	290	250	270	240	180	250	200	300	130	150	220	120	42	20	28	18	13	29
Zr	280	256	258	257	261	263	263	255	278	319	291	336	532	593	559	578	630	567

d — DH-270

SAMPLE	51	52	53	54	55	56	57	58	59	60	61	62	63	64	65	66	67	68	69	70
Height *(cm)*																				
TO	-705	-680	-654	-631	-591	-531	-455	-405	-376	-331	-261	-219	-179	-161	-111	-66	-45	-28	-11	26
FROM	-731	-705	-680	-654	-631	-591	-531	-455	-399	-371	-331	-231	-219	-179	-161	-111	-66	-45	-28	0
Percent																				
SiO_2	56.9	53.1	46.6	43.7	49.0	42.7	42.9	45.9	44.4	43.1	40.9	43.8	43.6	40.1	43.3	40.5	46.4	43.7	62.8	77.2
TiO_2	1.75	2.06	2.76	2.32	2.00	2.08	2.39	1.94	1.94	1.81	1.97	2.39	2.23	2.04	2.15	2.43	2.94	3.45	2.36	0.41
Al_2O_3	14.8	14.3	18.0	16.8	14.1	16.0	17.0	14.6	13.8	12.5	13.5	15.5	15.4	16.3	18.2	19.8	24.4	27.7	17.9	10.7
Fe_2O_3	2.9	3.5	3.9	4.6	3.2	(23.5)	4.5	1.8	0.8	0.8	1.7	5.1	3.8	4.8	3.7	4.6	3.6	3.4	2.3	0.0
FeO	10.2	12.2	13.7	16.4	12.7		16.9	5.47	3.4	3.4	5.4	16.6	17.5	14.9	16.8	16.2	5.5	4.7	3.1	1.7
MnO	0.15	0.15	0.23	0.25	0.23	0.23	0.24	0.21	0.23	0.25	0.29	0.27	0.29	0.30	0.26	0.27	0.15	0.17	0.11	0.0
MgO	1.88	2.24	2.33	2.87	2.06	2.64	2.75	0.83	0.44	0.46	0.74	2.48	2.56	2.33	2.47	2.71	1.27	1.12	0.76	0.29
CaO	1.87	2.62	1.31	1.34	4.06	2.26	1.04	11.56	14.20	16.95	14.99	1.60	2.22	5.27	1.41	1.44	0.87	0.65	0.37	0.40
Na_2O	0.4	0.7	2.0	0.9	0.4	0.1	0.4	6.0	6.3	6.0	5.9	0.2	0.2	0.3	0.3	0.1	0.0	0.12	0.0	0.1
K_2O	4.75	3.35	4.15	4.24	4.82	4.91	5.22	1.52	1.29	0.87	0.76	5.66	4.59	4.40	4.73	4.84	8.75	9.96	6.54	6.79
H_2O^t	3.6	4.2	4.5	5.3	3.9	<0.1	5.5	1.8	1.2	1.0	1.9	5.0	5.1	4.9	5.5	6.1	4.1	4.2	2.9	1.2
CO_2	<0.1	0.1	<0.1	<0.1	2.3	0.7	0.1	8.3	10.5	12.7	11.3	0.4	0.6	3.3	0.3	0.2	0.0	<0.1	<0.1	<0.1
C	<0.1	<0.1	<0.1	<0.1	0.1	0.1	0.1	0.0	0.1	0.0	0.0	0.1	0.1	0.1	0.0	0.1	0.1	<0.0	<0.1	<0.1
P_2O_5	1.18	1.26	0.65	0.51	0.51	0.51	0.44	0.35	0.34	0.33	0.35	0.42	0.68	0.57	0.44	0.53	0.63	0.44	0.12	0.05
S	0.05	0.20	0.07	0.09	0.34	1.11	0.16	0.12	0.06	0.06	0.22	0.45	0.48	0.47	0.15	0.18	0.45	0.00	0.02	0.41
Total	100.9	100.2	100.7	99.8	100.1	97.2	100.0	100.5	99.1	100.5	100.1	100.3	99.8	100.4	100.1	100.3	99.7	100.3	99.8	101.1
ppb																				
Au	3	6	2	2	4	7	6	2	3	2	6	9	11	3	2	5	4	<1	<1	<1
ppm																				
Ba	3754	2417	3125	3208	3369	3379	3336	639	556	364	370	3603	3324	2994	2358	2299	4030	4957	4053	6572
Be	4.3	4.0	4.6	2.4	2.1	1.7	1.8	1.2	1.5	1.3	1.3	1.8	2.0	3.9	4.5	5.4	10.0	11.0	6.8	2.6
Co	39	60	44	55	60	110	58	28	12	12	29	62	79	84	70	83	68	22	16	11
Cr	96	170	200	240	220	140	120	210	200	220	220	170	180	200	160	250	320	410	310	42
Cu	90	490	78	280	120	220	100	210	140	100	200	350	340	610	190	190	630	44	21	21
La	29	41	19	18	21	44	66	28	33	28	26	64	27	58	40	37	70	150	150	89
Ni	110	130	120	150	110	110	170	81	53	49	120	120	100	120	180	200	62	80	38	0
Pb	0	5	3	2	1	3	1	0	6	5	4	6	5	9	7	7	5	5	7	34
Rb	145	106	133	103	121	113	119	41	44	24	21	173	124	199	219	260	418	446	274	191
Sr	66	70	84	65	98	62	69	235	302	332	317	87	91	147	60	38	37	30	40	78
Th*	7.2	7.8	7.3	6.2	5.5	5.1	6.2	5.1	4.7	4.7	5.2	6.3	5.4	5.2	44.0	5.2	7.9	7.7	6.4	146.0
U*	0.9	1.7	1.7	1.4	1.3	1.2	1.3	0.9	0.8	0.9	1.0	1.0	1.0	0.9	11.4	0.8	1.5	2.8	4.5	40.0
V	180	190	270	210	170	180	220	120	140	140	180	200	200	200	230	250	320	370	250	62
Y	24	40	24	26	21	25	28	24	24	27	29	41	33	37	36	38	48	48	38	27
Zn	82	100	120	150	100	150	150	46	23	18	43	160	160	140	150	160	44	35	67	1
Zr	214	281	305	251	225	217	247	204	200	200	213	235	236	231	345	286	368	434	302	324

* U and Th analyses by ICP-MS method by the G.S.C. Labs.

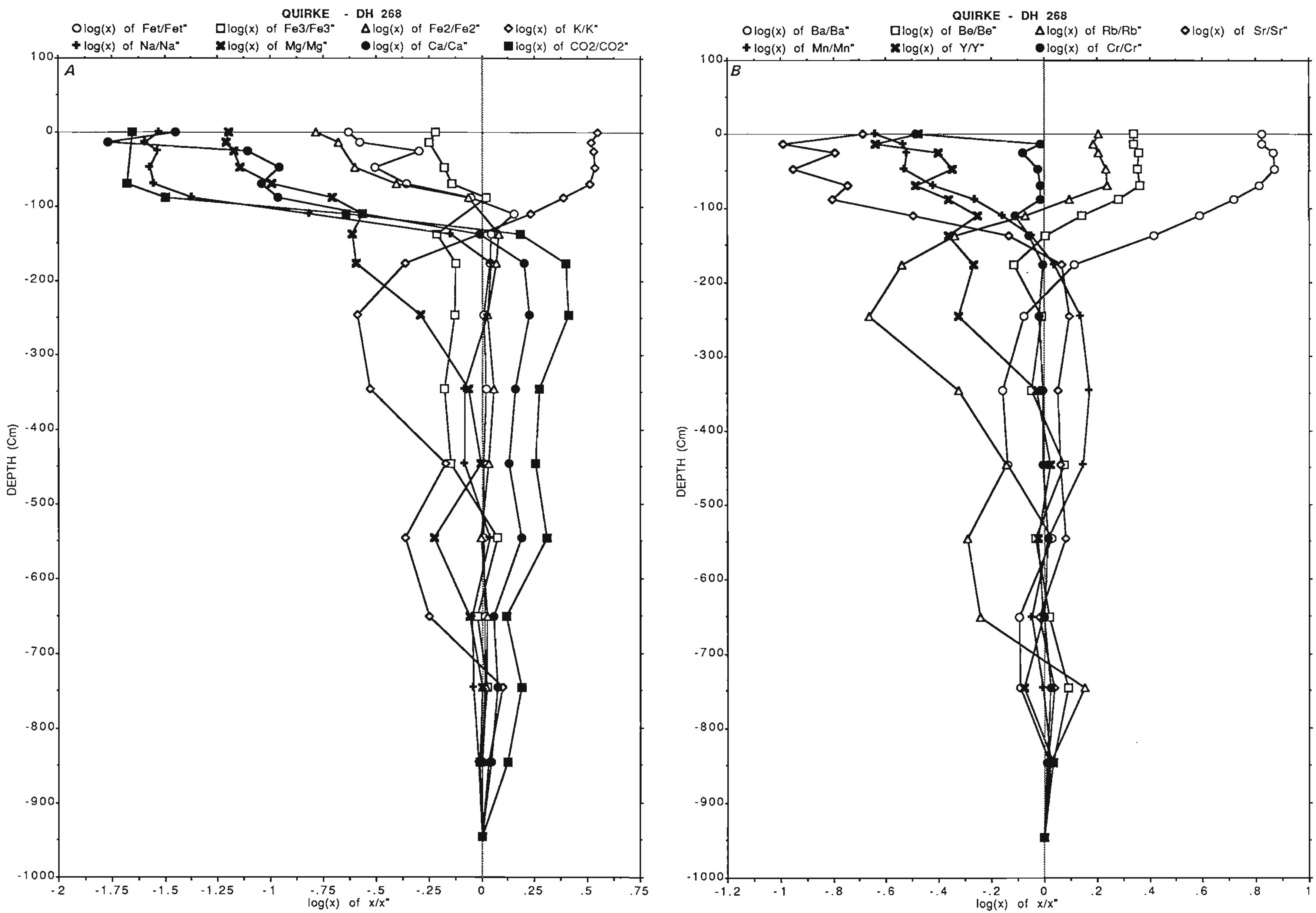

Figure 3. Chemical profiles of Quirke II drill cores.

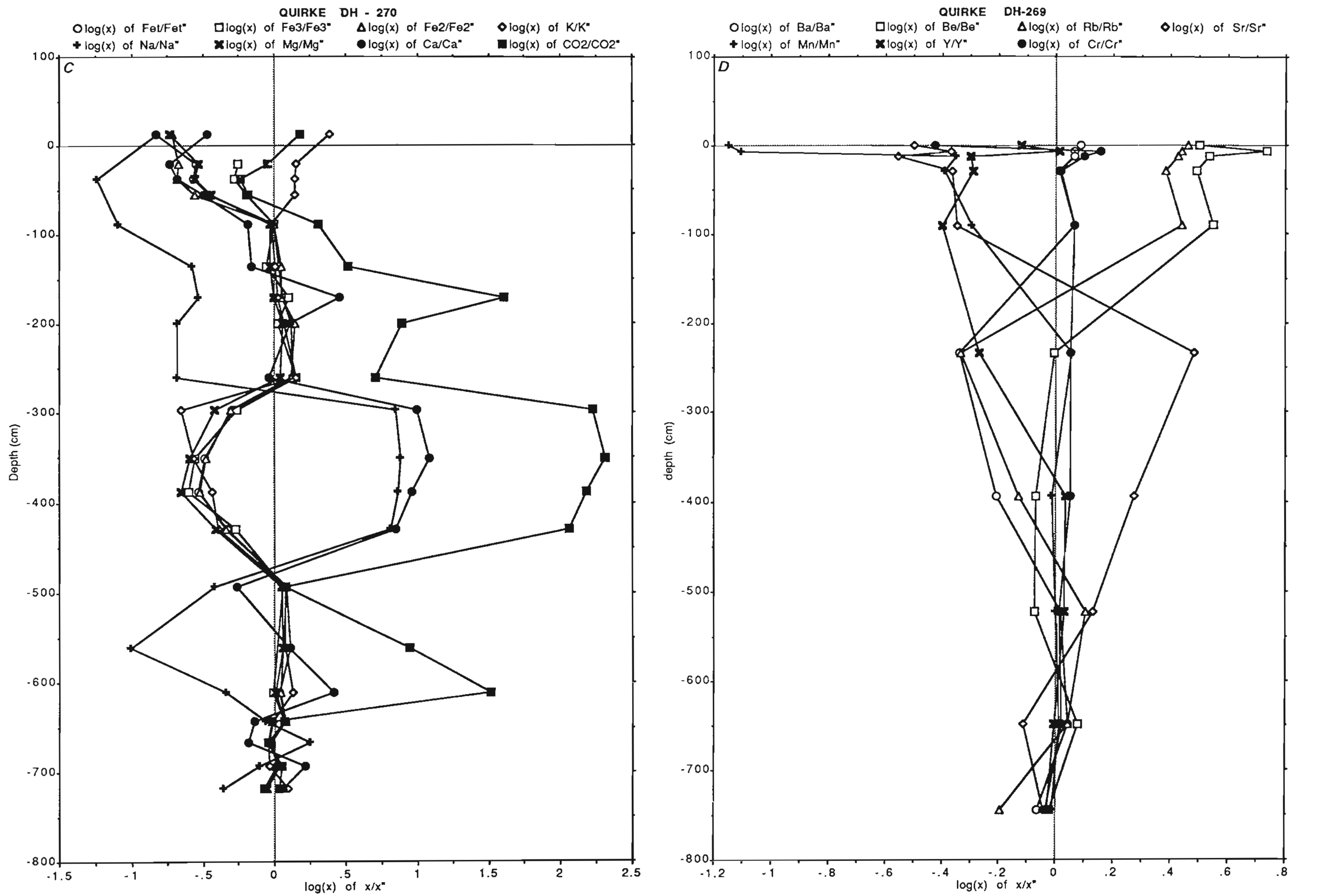

Figure 3. (cont'd.)

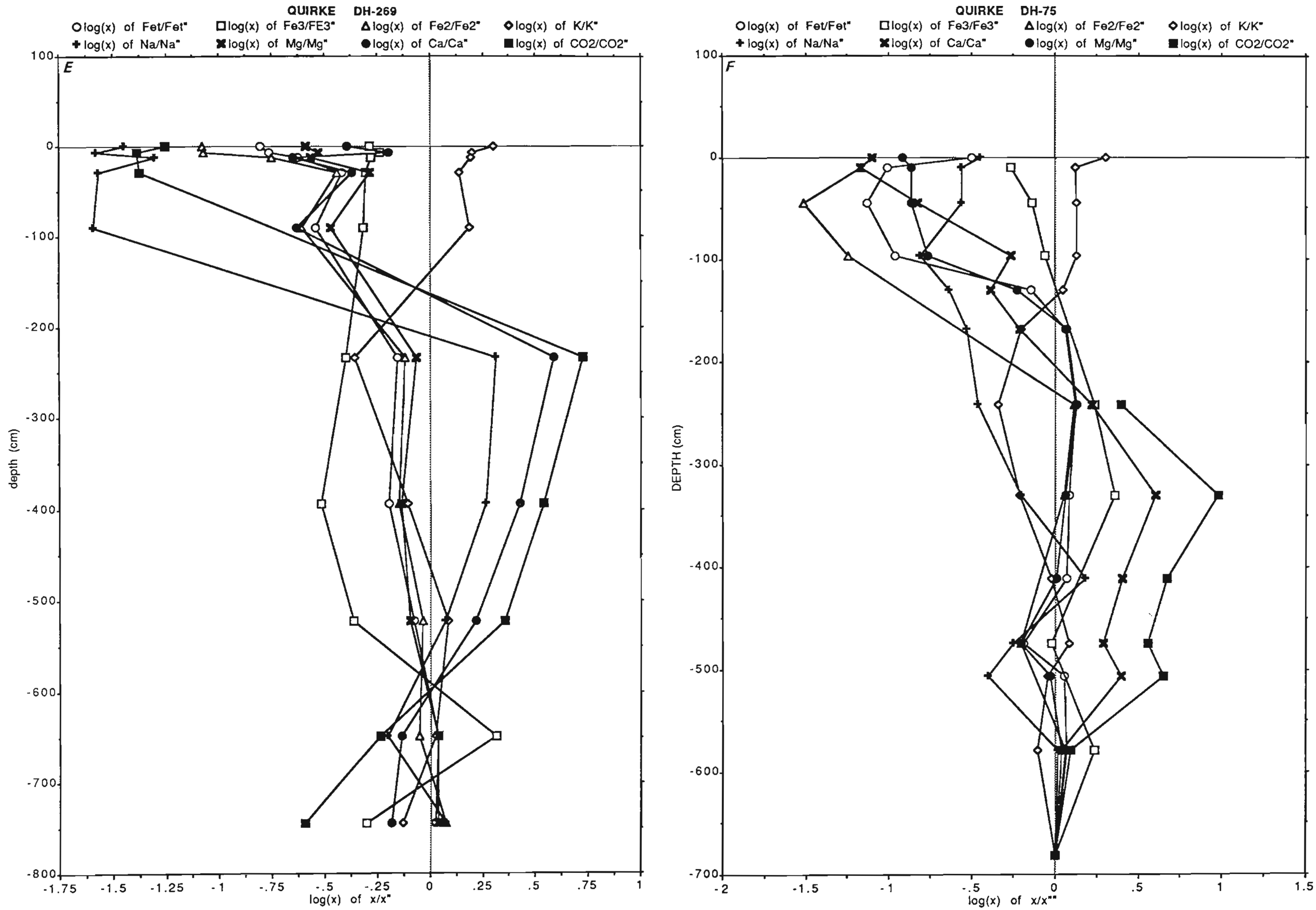

Figure 3. (cont'd.)

gitudinally and one half was analyzed in contiguous sections for major, minor and trace element compositions. The sampled sections are designated by their mid-section core distances stratigraphically below bases of Matinenda strata in Figure 3. Sample intervals and lengths are normally about 100 cm in basal, least altered, sections and as little as 5 cm at the top of the altered zones. At least one specimen in each chemically analyzed section was studied in thin section.

Except for U, Th and Au which were analyzed by X-Ray Assay Laboratories, Toronto, all analyses were done by Analytical Chemistry Section, Mineral Resources Division, Geological Survey of Canada. U was done by delayed neutron activation counting, Th by x-ray fluoresence and Au by fire assay and atomic absorption. The major oxides were done by X-ray fluorescence method. Pb was done by atomic absorption. The trace elements were determined by I.C.P.-ES. The FeO, H_2O^t, $CO_2{}^t$ and S were analyzed by rapid chemical methods. Fe_2O_3 is calculated using the formula: $Fe_2O_3 = Fe_2O_3$ (XRF) — 1.11134 × FeO (volumetric). FeO analyses are not given on samples having more than 0.5% S as these are unreliable, because S-bearing minerals would reduce Fe^3 to Fe^2. Actual Fe_2O_3 contents then cannot be calculated in these cases, hence total Fe (XRF) is given as Fe_2O_3 with the values bracketed (xxx) to distinguish them from actual calculated Fe_2O_3.

PETROLOGY

Lithological units of the 4 drillcore sections are shown graphically in Figure 1. Core lengths as presented are only slightly greater than true stratigraphic thicknesses. It is difficult to distinguish between Archean and Huronian basalts. The least altered of those in each of the drillcores have similar high Ti, and K, and other chemical and structural and textural similarities. Dh 75 appears to have repeated highly altered zones interpretable as interflow weathered zones in Dollyberry basalts. All are tentatively considered Huronian but a magnetite-rich band in highly altered metabasalt in Dh 270 suggests that it might be Archean. A highly altered metabasalt associated with altered granitic rock at the base of the Dh 75 profile may be Archean.

Dh 268, Dh 269 and Dh 270 show single cycles of progressive upward alteration of basalt through 4 or 5 m culminating in a cap of residual argillite 50 to 90 cm thick. Granitic rock at the base of Dh 75 is presumably Archean as no post-Archean granitoid rocks are known near Quirke Lake. The altered granite, between unaltered granite and the altered basalt at the base of this hole, is interpreted as paleosol that was subsequently covered by a basalt flow, which in turn was subsequently altered throughout. Successive thin flows show alterations that may represent 3 weathering cycles. Possibly the granite is a boulder in interflow or basal conglomerate. Another possible interpretation is that a hybrid zone separates Archean metabasalt from intrusive Archean granite, and that the Archean rocks were intensely weathered prior to eruption and weathering of unconformably overlying Dollyberry basalt.

The original products of weathering were modified during early and late diagenesis, early post Huronian mild metamorphism and late (Penokean) mild metamorphism. The resulting products are mixtures of chlorite, sericite, quartz, and calcite in varied proportions. Chlorite is dominant in lower, less altered sections of basalt. Progressive upward destruction of plagioclase laths results in increases in sericite, quartz and calcite. The final product, a cap of 'argillite', is dominantly sericite with little or no chlorite. It is identical to matrices of Matinenda subarkose, quartz pebble conglomerate, argillite partings in Huronian arenites and the pelitic fraction of Huronian siltstone-argillite formations, notably the McKim and Pecors formations. There is no evidence that the sericite precursor was other than a K-rich illite formed prior to burial of the paleosol beneath Matinenda sands.

Contacts between paleosol "argillite" and overlying sorted quartzose clastic sediments of Manfred Member are sharp and 'tight' in Dh 268 and Dh 75. In Dh 269, the contact is less obvious due to evident erosion at the top of the paleosol and mixing through a few centimetres of flakes and clasts of argillite and variably 'argillitized' basalt with quartz granules and pebbles. In Dh 270, residual argillite and Matinenda subarkose are separated by 11 cm of vein-breccia material containing quartz, carbonate, chlorite, and pink felspar. which was not sampled.

The abundance of carbonate in the altered sub-Matinenda rocks is noteworthy as this feature has not been emphasized by previous investigators. The calcite is present as fine grained intergranular "cement", as thin veinlets and also as secondary sparry calcite growths in cavity fillings. Affects of diagenesis and/or burial are exemplified by quartz crystal growths in the pressure shadows of pyrite cubes (Fig. 2b) and recrystallized compressed quartz-rich layers that impart a laminated appearance.

Certain elongated arcuate textures identified as cutans are present. Cutans or "clay skins" are common soil features widely distributed in Phanerozoic formations and are considered as plasma concentrations or separations connected with surfaces of grains or voids (Teruggi and Andreiss, 1971). There are several types of cutan fabric recognized in these paleosols. Some are arcuate where sericite is lined with thin wisps of chlorite. Others are channel cutans (Fig. 2e) (Ettensohn et al., 1988, Figure 17D) which are thin irregular channels with alternating sericite rich bands lined with coatings of dark brown mineral ending in a sac shaped cavity at the core of which there is a large grain of chlorite and oxide mineral. In this section ilmenite appears to be concentrated in the sericite-rich bands (Fig. 2f). In addition to the cutans there are some oval structures, 1.5 mm long 1 mm wide, with fine grained interlocking quartz grains in elongate arcuate shapes surrounded by chlorite in a groundmass of very fine grained sericite. The formation of these structures is not clear but probably related to the pedogenic processes. Another characteristic feature of the paleoweathering of basalt is its laminated appearance. The pale white laminae are sericite-quartz-rich and dark laminae are light to medium green and chlorite-rich. However in more intensely weathered parts of the section the laminae are obliterated (Fig. 2c) and the rock exhibits a blotchy, mottled appearance. Highly altered parts of the profile also have brecciated structure, which resulted from chemical and physical breakdown.

Ilmenite, rutile and pyrite are main opaque minerals. Chalcopyrite, pyrrhotite, sphalerite and magnetite have also been identified. In some core sections titanite is the predominant accessory mineral occurring as wedge shaped crystals. Pyrite occurs in several morphological forms: as discrete euhedral to subhedral cubes, as accumulations of fine grained subhedral to anhedral cystals that in places form a sieve texture (Fig. 2d), as euhedral (>100 μm size) cubes filling voids and cavities (Fig. 2b) and as thin veinlets a few millimetres wide that cut across the foliation.

PETROCHEMISTRY

Analyses of the samples from 4 drillhole sections are given in Table 1. The progressive gains and losses of major and and some minor components relative to Al_2O_3 from relatively unweathered parts to increasingly altered parts (towards the unconformity surface) are shown graphically (as logs of ratios) in Figures 3A to 3F. This and other methods of graphic representation of chemical weathering have been described by Reiche (1943). It is unlikely that aluminum is totally immobile but it is the least mobile of major elements. The normalization helps clarify changes and it also reduces variations due to inhomogeneities in the original unweathered rock. Composition of the least altered sample or an average of several samples is taken as the closest available approximation to 'parent' rock composition. These are vertically farthest away from the unconformity except in Dh 75. The formula used in calculation is :

$$M_{value} = \text{Log } [(MO_2^{sample}/MO_2^{parent})/(Al_2O_3^{sample}/Al_2O_3^{parent})]$$

Dh 268 shows the least complex chemical profiles. The profiles from other drillholes are complicated by abundance of intergranular calcite, translocated paleosols or more than one paleosol. Two main alteration zones can be distinguished, a lower section of initial alteration 3 to 6 m thick extending to about 1 m beneath the unconformity and a topmost intensely altered section about 1 m thick. In Dh 270, the lower zone can be subdivided into a 2 m low Fe subzone containing up to 29% calcite and a suprajacent 1.5 m subzone where Fe was restored to original levels (Fig. 3C).

Initial alterations are marked by increases in CO_2, Ca, and Na, and losses in Fe^3, Fe^2, Mg, K and Rb, except in Dh 270 where the initial abundance of latter elements are restored above a carbonate zone. Fe^3/Fe^2 ratios remain constant, unaffected by variations in carbonate content although total iron is antipathetic to CO_2.

In the uppermost (argillite) altered zone, K, Ba, Be, S, Rb, increase upward whereas CO_2, Ca, Na, Mg, Fe (other than that in pyrite), Mn, Cu, Zn, Ni, P, Y and Sr decrease. Si is slightly depleted in the argillite. Th and U are slightly enriched, with U and U/Th ratios increasing progressively to the top of the paleosol. Au is evidently contained in pyrite as its concentrations tend to correlate with those of S. Both elements are enriched in residual argillite in Dh 268 and Dh 269. Concentrations of Zn, Ni and Pb, on the other hand are antipathetic to S. Consistencies in ratios of Ti, P, V, Zr, La and Y to Al support the supposition that absolute increases of these elements are the result of compaction of residual clay. Fe^2 decreases more than Fe^3 in the argillite cap, resulting in increase in Fe^3/Fe^2 ratio. Some of the Fe^2 dissolved, presumably from silicates, is recaptured in pyrite which is the only iron-bearing mineral identified in the argillite. The extreme separation of Rb and Sr — increase of Rb and loss of Sr — at the top of the paleosol could be exploitable for isotopic Rb-Sr dating of the paleoweathering.

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A geological transect of the Leaf River area, northeastern Superior Province, Ungava Peninsula, Quebec

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Abstract

The Minto subprovince in the Lake Minto-Leaf River area is made up of four northwest-trending lithotectonic terranes, all dominantly plutonic in character. From west to east these include: 1) Tikkerutuk terrane, of homogeneous hornblende-biotite granodiorite and granite, a possible continuation of the Bienville subprovince to the south; 2) Lake Minto terrane, comprising orthopyroxene-bearing diatexite, granodiorite and granite with abundant relics of paragneiss and rare metavolcanic rocks; 3) Goudalie terrane, containing biotite ± hornblende tonalite and tonalite gneiss, with paragneiss and probable metavolcanic remnants, cut by mafic dykes in turn cut by late granites; and 4) Utsalik terrane, made up of granodiorite and granite with distinct hornblende- and orthopyroxene-bearing phases.

Résumé

La sous-province de Minto dans la région du lac Minto et de la rivière Leaf est composée de quatre terranes lithotectoniques à direction nord-ouest, de caractère principalement plutonique. Ce sont, de l'ouest à l'est : 1) le terrane de Tikkerutuk, de granodiorite et de granite à hornblende et biotite homogènes, prolongement possible de la sous-province de Bienville au sud; 2) le terrane de Minto Lake, comprenant de la diatexite, de la granodiorite et du granite à orthopyroxène avec des restes abondants de paragneiss et de rares roches métavolcaniques; 3) le terraine de Goudalie, contenant de la tonalite et du gneiss tonalitique à biotite avec plus ou moins de hornblende accompagnés de restes de paragneiss et de roches probablement métavolcaniques, recoupés de dykes mafiques à leur tour recoupés de granites tardifs; et 4) le terrane d'Utsalik, composé de granodiorite et de granite avec des phases distinctes à hornblende et à orthopyroxène.

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INTRODUCTION

This report summarizes results of the second season of a three year field project aimed at mapping a 50-150 km wide transect across Ungava Peninsula (Fig. 1). The latitude of Lake Minto- Leaf River was chosen as it provides water access at a high angle to the regional structural grain. The eastern part of the transect, described here, was mapped in 1990, continuing from the western region examined in 1989 (Percival et al., 1990). Some preliminary regional subdivisions have been defined on the basis of the reconnaissance investigations.

The project constitutes a transect through the Minto subprovince (Card and Ciesielski, 1986) of the Superior Province. Based on Stevenson's (1968) regional helicopter reconnaissance, Card and Ciesielski distinguished the Minto subprovince with northerly structural trends and high metamorphic grade, from lower grade, easterly-striking subprovinces to the south. Work in 1989 and 1990 (Percival et al., 1990) permits subdivision of the region into several lithotectonic terranes.

LITHOTECTONIC TERRANES

The NNW-trending terranes, defined on the basis of their lithological components and structural character, are from west to east (Fig. 2a,b): 1) Tikkerutuk terrane; 2) Lake Minto terrane; 3) Goudalie terrane; and 4) Utsalik terrane. All consist dominantly of plutonic rocks and have intrusive contacts with their neighbours.

Tikkerutuk terrane

The well-exposed region between Hudson Bay and western Lake Minto is underlain by homogeneous hornblende-biotite granodiorite and granite. Sparse metre-scale inclusions of diorite, gabbro and pyroxenite are ubiquitous. Inclusions of supracrustal rock were not observed in the present area but Stevenson (1968) reported iron-formation from the Inukjuak region, 175 km to the northwest. Granodiorite and granite form discrete NW-striking belts 5-50 km wide except in broad contact zones where granite dykes intrude granodiorite forming kilometre-scale gneissic units.

Granodiorite is medium grained, with 15-25% combined hornblende and biotite, and ranges in texture from foliated to massive. Clinopyroxene occurs locally as cores in hornblende. Granite is coarse grained to pegmatitic and generally massive, with 5-10% biotite and less hornblende. Foliation strikes consistently NNW, with steep dips (75-90°). Local fracture, gouge and chlorite-epidote alteration zones occur in narrow, WNW-trending zones, suggesting late brittle faults.

Rocks of similar lithological and structural character occur south of Lake Minto, where the dominant rock type is granite. Granodiorite occurs as kilometre-scale bodies, presumably inclusions. Granite of at least two ages occurs in this region: foliated to gneissic biotite granite occurs as inclusions in massive leucogranite.

Several small (5-500 m scale) bodies of pyroxenite breccia occur at wide intervals within the terrane. These are metre-scale blocks of orthopyroxene — clinopyroxene — hornblende ± biotite rock, separated by centimetre-wide dykes of granite. The breccia locally has a pyroxenitic matrix. Gabbro and diorite, with hornblende-biotite mafic mineral assemblages, occur as metre-scale inclusions.

The eastern boundary of the Tikkerutuk terrane is defined lithologically by the appearance of supracrustal units and diatexite characteristic of the Lake Minto terrane. Because hornblende-biotite granodiorite of similar character occurs in both terranes, the boundary is gradational. Structural features trend parallel to the boundary.

Lake Minto terrane

The Lake Minto terrane extends from western Lake Minto in the west to the western end of Leaf River, approximately 130 km across strike. It terminates in the south in central Lake Minto along a WNW-trending ductile deformation zone; the northern boundary was not observed. The region is characterized by high relief and rugged topography.

Lithological diversity characterizes the Lake Minto terrane, both in supracrustal and intrusive rock types. Supracrustal rocks occur in minor proportions as kilometre-scale units and are widespread as inclusions in plutonic rock types. The most abundant lithology of inferred supracrustal origin is probably sedimentary, migmatitic paragneiss, occurring in belts up to 3 km wide. Migmatitic paragneiss consists of a medium grained, schistose paleosome of biotite — plagioclase — quartz ± garnet ± orthopyroxene ± cordierite and a medium- to coarse-grained white granite leucosome with some

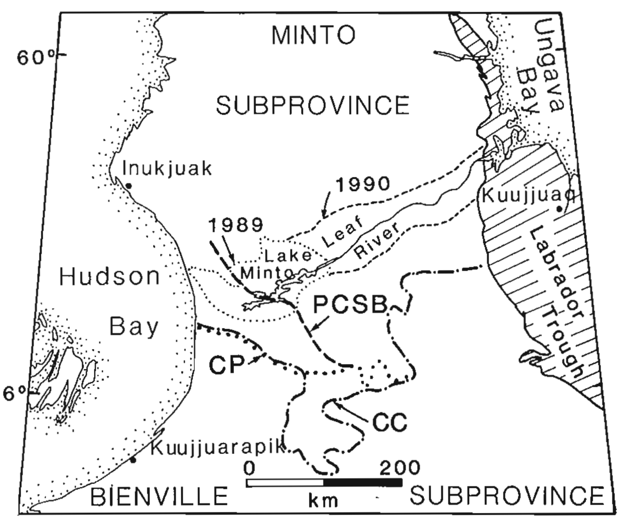


Figure 1. Location map showing the position of the Lake Minto-Leaf River transect and the Bienville-Minto subprovince boundary as defined by Card and Ciesielski (1986) (CC), Ciesielski and Plante (1990) (CP) and this paper (PCSB).

garnet, biotite and orthopyroxene. Paragneiss commonly occurs as inclusions within diatexite and is associated with white leucogranite. Metamorphic conditions, estimated through microprobe analyses of garnet-orthopyroxene-plagioclase-quartz assemblages are in the range 4.5-6 kb, 750-825°C. Thinly-layered quartz — magnetite iron-formation occurs in association with paragneiss, and as metre-scale blocks in diatexite; arkosic units are rare. Rocks of probable volcanic origin occur as thin (< 1 km wide) units of dominantly mafic composition (see Percival et al., 1990) within a 20 km-wide region of central Lake Minto. Rare associated thinly-layered felsic rocks contain zircon of 2695 ± 5 Ma age (J.K. Mortensen, pers. comm., 1989).

Intrusive rock types make up over 95% of the terrane. Four major units are recognized and appear to have the following chronological order: 1) dioritic gneiss, with biotite — clinopyroxene — orthopyroxene assemblages; 2) hornblende — biotite granodiorite and associated rocks; 3) orthopyroxene — biotite ± garnet ± cordierite diatexite; and 4) granite.

Dioritic gneiss (1) occurs in several poorly-preserved, kilometre-scale oval plutons and as ubiquitous inclusions in later intrusions, particularly diatexite. In a few locations diorite contains paragneiss inclusions. It has a distinctive texture, preserved in the larger bodies, comprising a paleosome of centimetre-scale oval domains of fine grained orthopyroxene-biotite-plagioclase and clinopyroxene-biotite-plagioclase, and coarse grained tonalitic leucosome with pyroxenes up to 1 cm. In small inclusions, schistose to migmatitic textures are defined by biotite alignment. A suite of hornblende-bearing rocks (2) varies in composition from granite to pyroxenite but granodiorite is the most common, followed in decreasing abundance by granite, diorite, gabbro and pyroxenite. The rocks are homogeneous and massive-to-foliated with local migmatitic and gneissic phases. Both ortho- and clinopyroxene form major constituents of granodiorite, diorite and gabbro; clinopyroxene is common as cores to hornblende in granodiorite. Mafic mineral assemblages define igneous facies: hornblende-biotite; hornblende-clinopyroxene-biotite and orthopyroxene-clinopyroxene-hornblende-biotite. Granite contains similar mafic mineral assemblages and grades to granodiorite through K-feldspar megacrystic varieties. Inclusions of paragneiss, iron-formation, diorite and gabbro are common in granodiorite, and granodiorite inclusions occur in granite. In some areas granodiorite is interlayered with diatexite on a decametre to kilometre scale.

Diatexite (4) is a coarse grained rock of granodioritic composition, with heterogeneous textures showing affinities to migmatites. According to Brown's (1973) definition, diatexites are migmatites containing more than 50% neosome component. Units range in thickness from metres to 20 km and contain varied amounts of inclusions of paragneiss and diorite. Diorite and gabbro inclusions commonly have dehydrated margins in which biotite and hornblende are depleted with respect to the inclusion interior. Mineral assemblages in the coarse grained granodioritic component most commonly include orthopyroxene and biotite, with local garnet and rare cordierite. Coarse grained orthopyroxene-biotite leucogranite occurs as layers on the centimetre to kilometre scale.

Foliation is defined by biotite ± orthopyroxene alignment in granodiorite and by the orientation of inclusion pods and layers. Structural trends are generally WNW and steep, but abrupt strike variations are common on the outcrop scale.

Several varieties of granite (4), in addition to the hornblende-bearing granite described above, form mappable units. Coarse grained, homogeneous, massive orthopyroxene-biotite granite forms discrete plutonic bodies, whereas coarse grained to pegmatitic, homogeneous leucocratic biotite granite occurs both as batholithic masses and as sub-outcrop scale dykes, sills and pods. Gneissic biotite granite occurs locally as inclusions in massive granite. Leucogranite contains inclusions of all other rock types and is thus inferred to be the youngest Archean unit in the region.

The eastern margin of the Lake Minto terrane is marked by a decrease in relief and amount of exposure and the boundary with the Goudalie terrane to the east is therefore not well defined. However, similar lithological units occur in both terranes and it is probable that the boundary is intrusive.

Goudalie terrane

This unit represents the narrowest of the four terranes, extending about 70 km across poorly-exposed ground from western Leaf River to the Vizien River. It consists dominantly of foliated to gneissic tonalitic rocks with sporadic supracrustal relics, cut by granodiorite, several distinctive suites of granite and a small swarm of mafic dykes.

Supracrustal rocks include paragneiss, calc-silicate gneiss and mafic gneiss. Paragneiss, in units up to 0.5 km thick, is a garnet-biotite-plagioclase-quartz ± cordierite ± sillimanite rock with up to 20% leucosome of white granite. Rare calc-silicate pods up to several metres thick consist of calcite-quartz-garnet-clinopyroxene assemblages. Mafic gneiss units up to several metres thick are layered on a millimetre to centimetre scale, with assemblages of hornblende-plagioclase ± clinopyroxene ± epidote (Fig. 3a). Irregular zones and pods rich in clinopyroxene and/or epidote resemble calcic alteration zones in altered mafic volcanic sequences.

Tonalite also contains enclaves of mafic and intermediate plutonic rocks. Mafic inclusions are pods or layers of homogeneous medium- to fine-grained hornblende-plagioclase ± clinopyroxene amphibolite, locally with plagioclase phenocrysts up to 2 cm. Diorite is similar, with more biotite and stronger foliation.

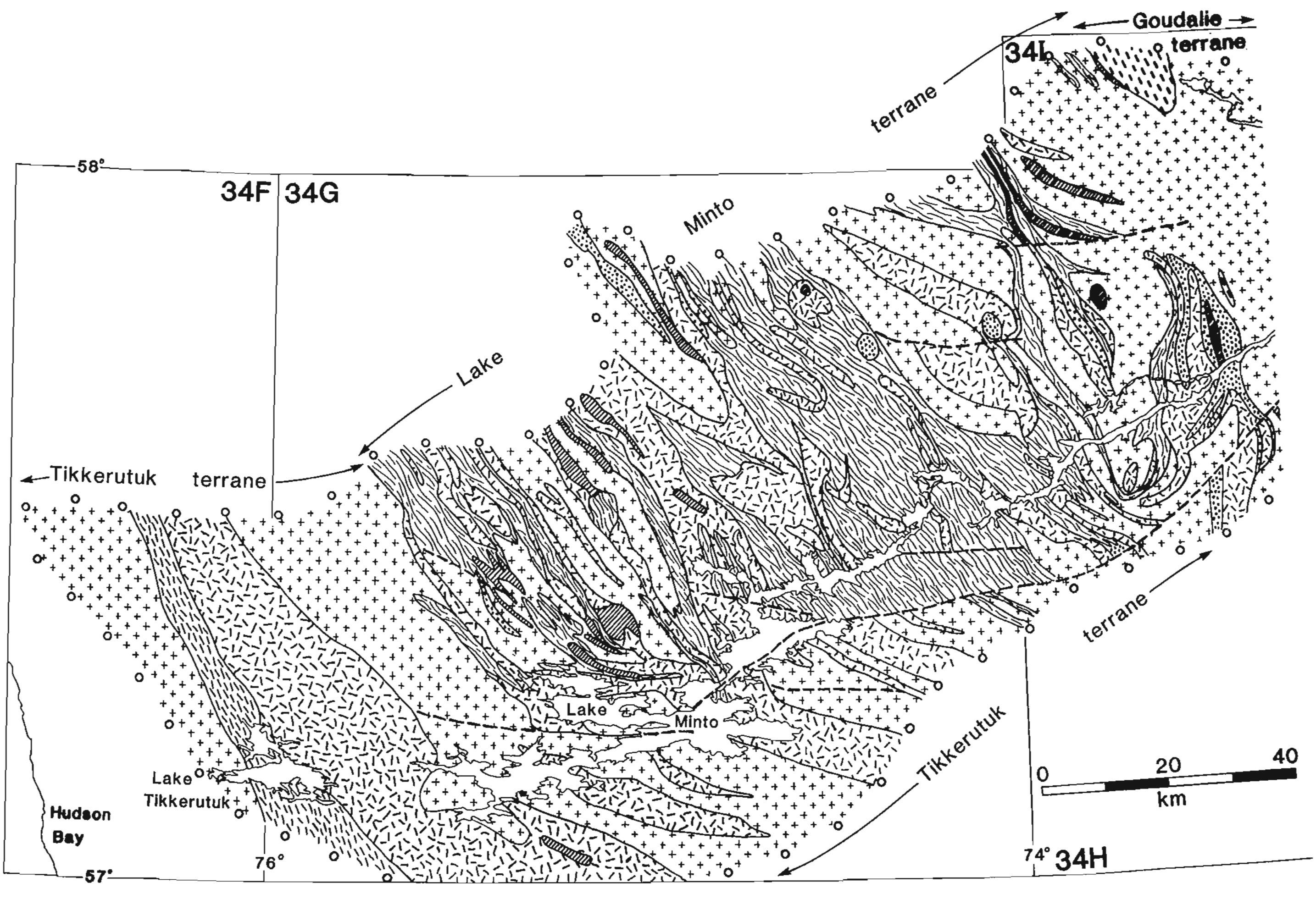
Tonalite, which makes up most of the terrane, comprises several discrete mappable units. In the south, the rocks are homogeneous, massive to foliated biotite ± hornblende tonalite with rare inclusions and very minor late granite dykes and veins. In the north, hornblende-biotite assemblages are more common in well foliated to gneissic rocks. Mafic to intermediate inclusions are common and layers and dykes of late granite are widespread. Gneissic units (Fig. 3b) are made up of centimetre-scale layers of tonalite with up to 10% mafic gneiss enclaves and 10-40% concordant granite veins.

Several suites of homogeneous plutonic rocks cut tonalite. Hornblende-biotite granodiorite occurs in elongate plutons up to 2 km wide. At least three types of granite were recognized: 1) ubiquitous biotite leucogranite that occurs as dykes and pods in tonalitic gneiss and as small plutons and is foliated in WNW-trending ductile deformation zones; 2) white orthopyroxene-biotite granite that occurs as metre-scale pods; and 3) biotite-apatite leucogranite that occurs as concordant layers cutting the WNW-trending deformation zones.

A narrow (30 km wide) swarm of mafic dykes, with individual members 0.5-20 m wide, cuts tonalites and some granites along the eastern margin of the terrane. The dykes are disrupted (Fig. 3c) by later granite veins and are stretched and boudinaged in WNW-striking deformation zones.

Metamorphic grade recorded by assemblages in mafic rocks and paragneiss is generally upper amphibolite facies. Paragneiss is migmatitic and contains garnet-cordierite-sillimanite-biotite assemblages. Mafic rocks are generally hornblende-plagioclase-biotite. Gneissic tonalite in the central part of the terrane locally displays "blebby" texture, defined by coarse hornblende in leucosome (Fig. 3d). Such textures have been interpreted elsewhere (e.g. West Greenland; McGregor et al., 1986) as being characteristic of retrograded granulite. In the eastern few kilometres of the terrane, the grade increases to granulite facies. Mafic rocks have ortho- and clinopyroxene and tonalites are locally migmatitic, with garnet-orthopyroxene-biotite assemblages. In this region, hornblende-bearing mafic enclaves have centimetre-thick dehydrated rims consisting of orthopyroxene and plagioclase (Fig. 3e); selvages of similar character border centimetre-scale granite dykelets cutting mafic rocks. Reactions and metamorphic conditions attending formation of these dehydration zones are under investigation as part of a B.Sc. project by A. Ross.

Figure 2a. Geological map outlining major lithological units, lithotectonic terranes and structural zones within the western part of the Minto subprovince.



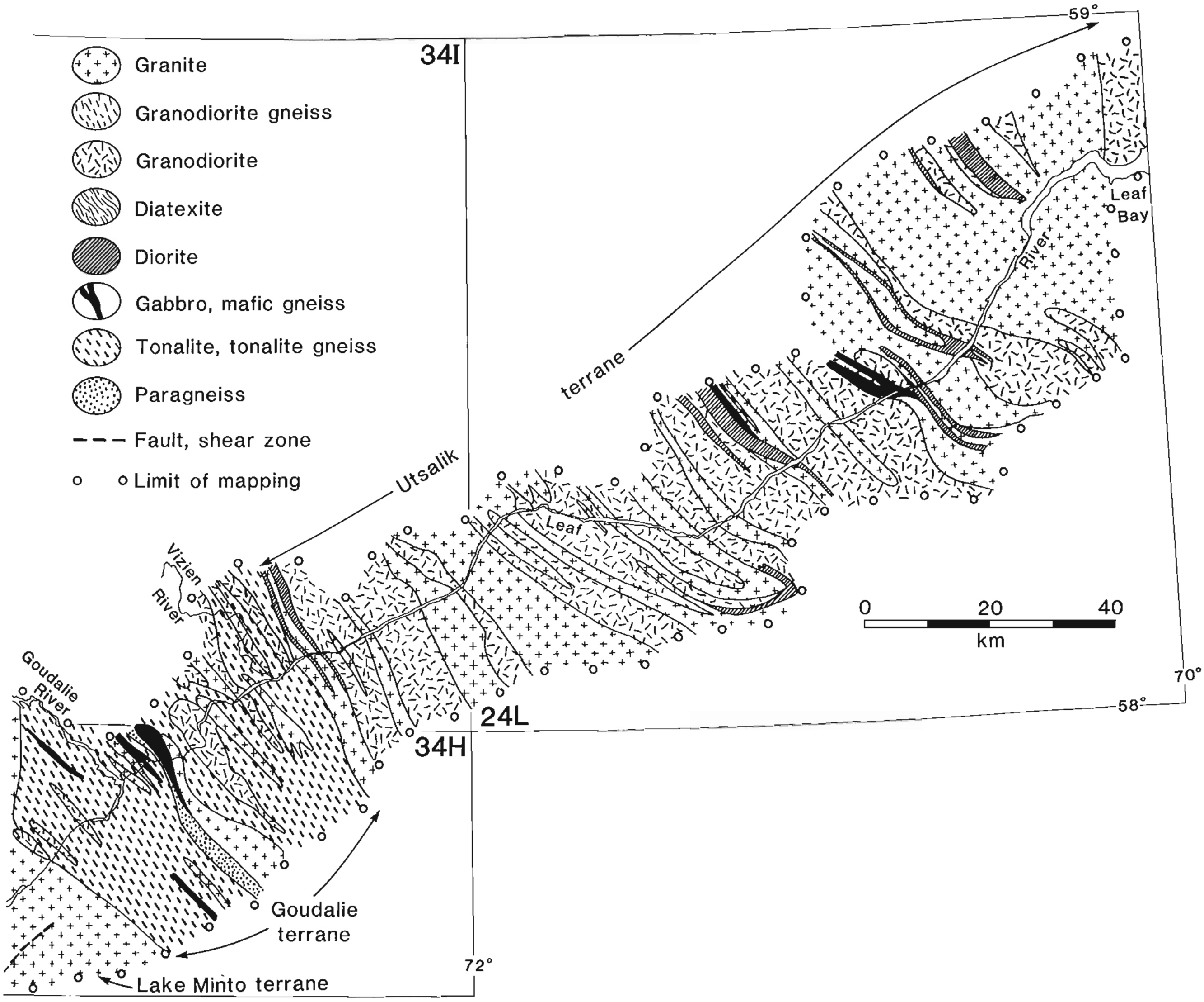


Figure 2b. Geological map outlining major lithological units, lithotectonic terranes and structural zones within the eastern part of the Minto subprovince.

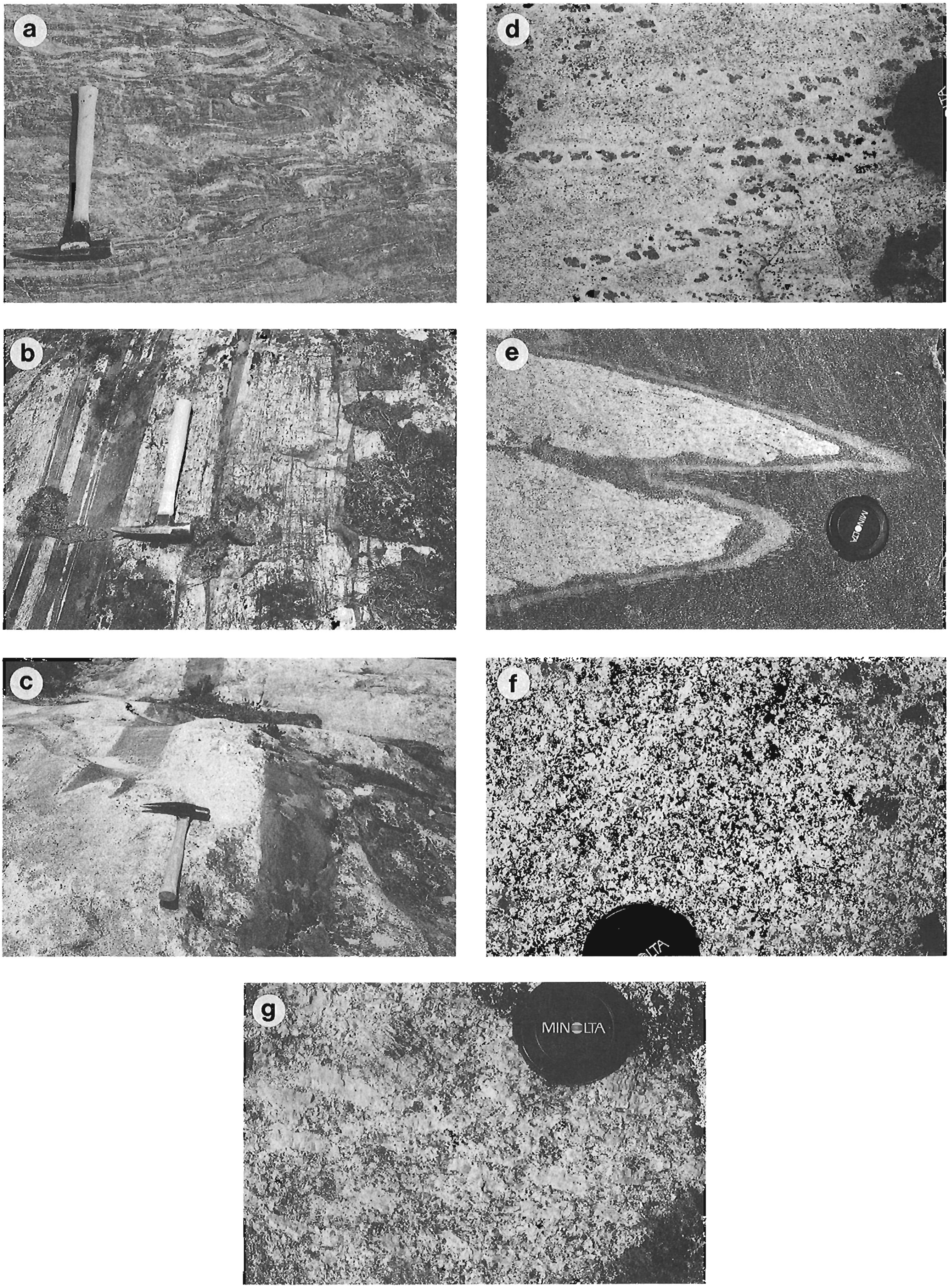
The orientation of foliation and gneissosity are generally NNW and steep. A zone of superimposed ductile deformation, in which structures including mafic dykes are reoriented to WNW orientations, cuts the Goudalie terrane. In the 15-25 km-wide zone, strong foliation, gneissosity, protomylonite zones, large-scale shear bands and local ultramylonite are variably developed. Kinematic indicators are mainly sinistral although dextral indicators were noted in some parts. A major sinistral ductile swing also occurs in the Lake Minto area (Percival et al., 1990).

The boundary between the Goudalie and the Utsalik terrane is dominantly lithological, separating tonalitic rocks with abundant mafic dykes from homogeneous granodioritic rocks. The granodiorites adjacent to the boundary have primary igneous textures and two-pyroxene mineral assemblages, in contrast to the hornblende-biotite assemblages and foliated to gneissic textures in tonalites of the Goudalie terrane.

Utsalik terrane

This region extends from the Vizien River in the west to the Proterozoic unconformity in the east, a width of about 150 km. It is characterized by moderate relief and varied, generally good, exposure. Intrusions of granodiorite and granite form elongate bodies up to 10s of kilometres in width.

The oldest units in the terrane are sparse inclusions of plutonic rock ranging in composition from gabbro to diorite, with rare pyroxenite. Diorite inclusions occur as isolated angular to equant blocks on a metre scale and as trains of angular, tabular blocks resembling disrupted dykes. Some coarse grained gabbroic bodies have medium grained dioritic margins up to 30 cm thick. Other gabbros are net-veined by centimetre-scale granite dykes. The nature and origin of inclusions in the western part of the terrane are under study as part of G. Shore's B.Sc. project.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MINOLTA

The terrane consists dominantly of homogeneous, massive to foliated granodiorite with several mineral facies. Medium- to coarse-grained hornblende-biotite rocks are dominant; these grade through facies with common clinopyroxene cores in hornblende into massive, homogeneous rocks containing orthopyroxene, clinopyroxene and biotite, all of apparent igneous origin (Fig. 3f).

Granodiorite is cut locally by granite with similar hornblende-bearing assemblages. Granite occurs also as large plutonic masses, with mafic mineral assemblages which include hornblende-biotite, biotite, clinopyroxene-hornblende-biotite, orthopyroxene-biotite and orthopyroxene-clinopyroxene-biotite. Coarse allanite is a common accessory mineral. Both granodiorite and granite are cut by late leucogranite (Fig. 3g).

Structural trends in the Utsalik terrane, defined by mineral foliation in granodiorite, are dominantly steep NW (315 ± 20°). Deformation zones, which also affect the late granites, strike westerly (270 ± 20°). The southeastern part of the terrane is more complex, with highly varied structural trends and lithologies on the metre scale. Here several varieties of granodiorite and granite are mixed on a metre to kilometre scale.

Figure 3 a. Mafic gneiss showing irregular patches and discontinuous layers rich in clinopyroxene, with some garnet, in a matrix of hornblende — plagioclase — clinopyroxene. Hammer handle = 30 cm long.

b. Tonalitic gneiss consisting of thinly-layered hornblende — biotite tonalite, mafic gneiss lenses and layers, and concordant granitic pegmatite. Hammer handle = 30 cm long.

c. Discontinuous mafic dyke in tonalite. Note wispy dyke termination and en echelon continuation. Hammer handle = 30 cm long.

d. "Blebby texture" defined by coarse ragged hornblende in leucosome in tonalitic gneiss. The hornblende may represent retrogressed orthopyroxene that formed in granulite-facies leucosome. Lens cap = 5 cm.

e. Dehydration rim on mafic inclusion in tonalite. Hornblende — plagioclase assemblages, characteristic of the inclusion, give way to orthopyroxene — plagioclase adjacent to tonalite, which contains assemblages of orthopyroxene — biotite — plagioclase — quartz ± garnet. Lens cap = 5 cm.

f. Coarse grained, massive granodiorite containing orthopyroxene, clinopyroxene and biotite. Lens cap = 5 cm.

g. K-feldspar megacrystic granite with weak foliation. Lens cap = 5 cm.

DYKES

Proterozoic dykes cut all terranes but are more abundant in the eastern part of the region. The most abundant are 2-30m, fresh diabase dykes striking 110 ± 10°. Both olivine-bearing and olivine-absent dykes were noted. An older set of NNW-striking clinopyroxene-plagioclase dykes occurs locally.

BOUNDARIES AND LINKAGES

Boundaries between terranes are defined on the basis of changes in lithological character, involving both plutonic and supracrustal assemblages. The amalgamation of the terranes occurred prior to, or during, intrusion of the hornblende-bearing granodiorite suite, which occurs in all terranes, provided that this igneous suite can be considered a time marker. Preliminary U-Pb zircon geochronology on hornblende granodiorite from Lake Minto (J.K. Mortensen, pers. comm., 1990) indicates an age of 2720 ± 10 Ma, similar to the 2721 ± 3 Ma age obtained on hornblende granodiorite from Leaf Bay, 300 km to the east by Machado et al. (1989).

A suite of diorites also appears to link the Lake Minto, Goudalie and Utsalik terranes. The fine grained diorite, with orthopyroxene-clinopyroxene-biotite-plagioclase assemblages and coarse grained pyroxene-bearing leucosome, forms ubiquitous inclusions in diatexite of the Lake Minto terrane and occurs also in larger oval plutons. In the Goudalie terrane, diorite of similar texture and mineralogy occurs as elongate enclaves at a high angle to foliation in tonalite. These may be deformed dykes. A few large (10 m) enclaves of diorite of similar texture and mineralogy occur in granodiorite of western Utsalik terrane. Dykes of massive granodiorite cleanly cut foliated to gneissic diorite. Thus the diorites, presumably equivalent in age, predate granodiorites of the Lake Minto and Utsalik terranes but may postdate tonalite of the Goudalie terrane. If the diorites are the same age, the field relations would imply an older age for the Goudalie terrane.

Rocks of the Goudalie terrane range in grade from amphibolite facies (hornblende-biotite-plagioclase assemblages) in the western and central parts to local granulite facies (two pyroxene and garnet-orthopyroxene assemblages) in the east. In the Utsalik terrane adjacent to the occurrences of granulite facies, granodiorites have igneous textures and assemblages of ortho- and clinopyroxene. The relationships imply a possible contact metamorphic effect of the pyroxene-bearing granodiorite on lower grade tonalite.

SUPERIOR PROVINCE SUBPROVINCE STRUCTURE

Card and Ciesielski (1986) defined the Minto subprovince on the basis of its high metamorphic grade, contrasting with plutonic rocks in the amphibolite facies in the Bienville subprovince (Fig. 1) to the south, and its northerly structural grain. Results of the present work suggest that NNW-striking terranes, similar in scale (75-150 km wide) to that of the southern, east-striking subprovinces, are present *within* the Minto subprovince. Following the Card and Ciesielski (1986) classification, the four terranes of the Minto block are plutonic in character, but vary systematically in mineralogy. The Tikkerutuk and Goudalie terranes have dominantly hornblende-biotite-bearing plutonic rocks and inclusions in the amphibolite facies. The Lake Minto terrane has uniformly high grade inclusions and pyroxene-bearing plutons. The Utsalik terrane is characterized by patchy pyroxene-bearing plutonic phases.

The Tikkerutuk terrane closely resembles the Bienville subprovince in terms of major rock types, mineralogy (Ciesielski and Plante, 1990) and aeromagnetic character. Although rocks from the Tikkerutuk terrane have not been dated, if granodiorites are similar in age to those of the Lake Minto and Utsalik terranes, they are also similar in age to granodiorites from the Bienville subprovince (2712 +3/−2 Ma; Mortensen and Ciesielski, 1987). Aeromagnetic patterns (Geological Survey of Canada, 1988) extend across the subprovince boundary defined by Card and Ciesielski (1986). It is therefore possible that the Tikkerutuk terrane represents a continuation of the Bienville subprovince.

The Lake Minto terrane is unique among the four terranes in terms of its high proportion of supracrustal remnants and in the presence of diatexite. Similar characteristics, including crystallization pressures in the 4.5-6 kb range, occur in the Ashuanipi complex (Percival, 1987; Percival, 1990), along strike about 350 km to the southeast and separated by the Bienville subprovince. Preliminary geochronology suggests that diatexites of the Lake Minto terrane may be about 20 Ma older than those of the Ashuanipi complex and hence direct correlation may be untenable.

The Goudalie terrane has many characteristics of granite-greenstone terranes of the southern Superior Province, including dominantly tonalitic composition and remnants of probable volcanic and sedimentary origin. It appears to be an isolated unit, surrounded by later plutonic bodies. Geochronology in progress will attempt to date its components to aid in correlation and tectonic reconstruction.

The Utsalik terrane resembles the Tikkerutuk terrane lithologically, although the scale and style of granodiorite and granite units differ. The patchy development of pyroxene-bearing phases also sets it apart from both the Tikkerutuk terrane and the Bienville subprovince.

TECTONIC FRAMEWORK

An outstanding question in Ungava geology is the relative age of NNW-striking structures, characteristic of the Minto subprovince, and WNW-trending structures which dominate farther south. Within the Minto subprovince, older NNW-trending structures are locally deflected into westerly (WNW) orientations in superimposed deformation zones. A similar bimodal distribution of structural trends was noted by Stevenson (1968) in plutonic rocks of the Bienville subprovince, with westerly trends becoming more dominant to the south. Although detailed geochronology is required to date units which bracket the different fabric-forming events, it is probable that the westerly trends are late features superimposed on NNW-trending structures or on massive units. Steep westerly structures in lower grade terranes of the southern Superior Province have been explained by a structural model involving dextral transpressive deformation (e.g., Corfu and Stott, 1986; Borradaile et al., 1988; Hudleston et al., 1988). To date, tectonic interpretations of Superior Province have been constrained only by the geology of the southern subprovinces (e.g., Percival and Williams, 1989; Card, 1990). The addition of information from the Minto subprovince, particularly geochronological data linking events to the southern belts, may require a reassessment of the tectonic system controlling the evolution of the Superior Province.

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Preliminary report on mineral resource assessment of a proposed national park, Bluenose Lake area, District of Mackenzie[1]

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Jones, T.A., Insinna, A., and Jefferson, C.W., Preliminary report on mineral resource assessment of a proposed national park, Bluenose Lake area, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 65-70, 1991.

Abstract

Geological and geochemical fieldwork were conducted as part of a mineral resource assessment of the Bluenose Lake area, N.W.T., a proposed national park. Typical outcrops of all geological units outcropping within an area of interest surrounding the proposed park were investigated and sampled in order to establish their mineral resource potential. Minor amounts of detailed and fill-in mapping were completed between 118°30' and 120° longitude, and north of latitude 68°. Two outcrops of Mesozoic age were located in an area shown on current maps as Paleozoic. Little evidence of mineralization was noted. Bulk stream sediment samples, collected in each of 25 watersheds draining the area, have been submitted for analysis. Single gold flakes (300 by 300 microns) were separated from two samples during preparation of heavy mineral concentrates. The samples are from short streams draining north into the Arctic Ocean.

Résumé

Des travaux géologiques et géochimiques sur le terrain ont été réalisés dans le cadre d'une évaluation des ressources minérales de la région du lac Bluenose (T.N.-O.), emplacement proposé pour l'aménagement d'un parc national. Les affleurements typiques de toutes les unités géologiques comprises dans une zone d'intérêt entourant le parc proposé ont été analysés et échantillonnés afin d'établir leur potentiel en ressources minérales. On a cartographié de façon détaillée et complète de petites zones situées entre 118°30' et 120° de longitude et au nord de la latitude 68°. Deux affleurements d'âge mésozoïque ont été localisés dans une zone dite plaéozoïque sur les cartes actuelles. On a relevé peu d'indices de minéralisation. Des échantillons en vrac de sédiments fluviatiles ont été recueillis dans chacun des 25 bassins versants drainant la région et ont été analysés. Des paillettes d'or (300 sur 300 microns) ont été extraites de deux échantillons au cours de la préparation de concentrés de minéraux lourds. Les échantillons proviennent de petits cours d'eau se déversant dans l'océan Arctique au nord.

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[1] Contribution to assessment of various resources of the Bluenose Lake area by Environment Canada, Parks. Project carried by the Geological Survey of Canada, and contracted to the first author.
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INTRODUCTION

The work outlined in this report was performed as part of a study of the non-renewable resources of the Bluenose Lake area, N.W.T., for Parks Canada planning purposes. The area has been proposed as an addition to the national parks system. It is the policy of Indian and Northern Affairs Canada to ensure that an inventory of the non-renewable natural resource potential of areas in the Yukon and Northwest Territories be compiled prior to their formal establishment as new national parks. The Geological Survey of Canada has traditionally conducted the required mineral and energy resource assessments (MERAs) at the request of Parks Canada (Jefferson et al., 1988, p. 7). The information acquired may be used to modify the boundaries of the proposed park, or, in areas of particularly high mineral potential, may suggest that establishment of a park is inappropriate if development of mineral resources is desired.

This summer's fieldwork was part of a two year MERA of the Bluenose Lake area following the format used by Jefferson et al. (1988). Fieldwork included: 1) evaluating the lithological characteristics and potential for economic mineralization of each of the major rock units outcropping within the map area; 2) geochemical sampling of river sediments; and 3) limited detailed geological mapping in areas of particular interest, or where previous work was lacking in detail, or incorrect.

DESCRIPTION OF CURRENT SURVEY

The Bluenose Lake area is located on the Arctic coastline, some 250 km. east of the Mackenzie (Dehcho) River delta (Fig. 1), and immediately east of the community of Paulatuk. Parks Canada has already proposed provisional boundaries for the park, but the study will evaluate the non-renewable resource potential of a somewhat larger area, here termed the area of interest (see boundaries, Fig. 1), which is approximately 62 000 km^2 in size.

In order to evaluate the mineral resource potential of the area, we were required to identify the mineral deposit models applicable to the stratigraphic package present in the area of interest. Those models identified to date are listed in Table 1.

The first field component involved describing and sampling at least one outcrop of each known rock unit within the area of interest, in order to evaluate its economic mineral potential by comparison with an appropriate mineral deposit model. Some 55 locations were investigated during one month of fieldwork.

The second field component involved dividing the area of interest into 25 watersheds, ranging in size from (approx.) 300 — 3000 km^2. One 8 — 10 kg. sample of − 1 cm sieved stream sediment material was taken in each watershed. From each bulk sample a silt fraction and a heavy mineral concentrate (HMC) were produced, as described later, for geochemical analysis.

In addition to the foregoing, several other tasks were completed. Limited bedrock mapping was done in several areas where current maps are essentially blank. One such area is centred on Bluenose Lake, and extends from 68° N latitude to just south of the Arctic coast. An oriented bulk sample of a prominent diabase sill was collected for A.N. LeCheminant (Continental Geoscience Division) to determine its U-Pb age as part of an ongoing geochronology program on the Mackenzie and Franklin igneous events.

Logistical support was shared with A. Pedder (Institute of Sedimentary and Petroleum Geology) in order to facilitate sampling and description of an isolated outcrop of postulated middle Devonian age, dated by him some 20 years ago on the basis of a fossil collection, but never visited until now due to its isolation.

It was also requested that we note and identify to Parks Canada personnel any sites which might form part of a geoscience interpretive program, should a park eventually be established.

Finally, the first author was charged with keeping the community of Paulatuk advised of his activities in the area. The proposed park abuts and may eventually include some areas which form part of the Western Arctic land claims settlement. Public discussions are ongoing.

PREVIOUS WORK

Only reconnaissance level bedrock and surficial geological maps have been completed for most of the area. The coastline geology was originally mapped by John Richardson in 1826 (Franklin, 1828), except for the southern portion of Darnley Bay, which was mapped by the Canadian Arctic Expedition in 1915 (O'Neill, 1924). The inland portion of the area of interest was not mapped until the staging of helicopter-supported Operations Coppermine and Norman in the late 1950s and 1960s. All of the area of interest east of longitude 124° was mapped in 1958 at a scale of 1:500 000 (Fraser et al., 1960); that portion of the area lying west of longitude 120° and north of latitude 68° was re-mapped in 1968 at 1:250 000 (Balkwill and Yorath, 1970, 1971; Cook and Aitken, 1969; Yorath et al., 1974). The area south of latitude 68° and west of longitude 119° was also re-mapped in 1968 at 1:500 000 (Cook and Aitken, 1971).

Table 1. Deposit models used in evaluation of mineral resource potential, Bluenose Lake area

Evaporites (1)*
Stratiform Phosphate (4)
Mississippi Valley Type Pb, Zn (6.1)
Kupferschiefer Type Cu (6.3a)
Continental Red Bed Cu (6.3b)
Sandstone Type U (6.4)
Carlin Type Carbonate Hosted Au (8.1)
Shale — hosted Pb, Zn (9.2)
Gabbroid — associated Ni, Cu, Platinum group elements (12.2)
Unconformity — associated U (21)
Arsenide Vein Ag, U (22)
ALSO: Coal, Carving Stone

* numbers in brackets are those used in Eckstrand (1984)

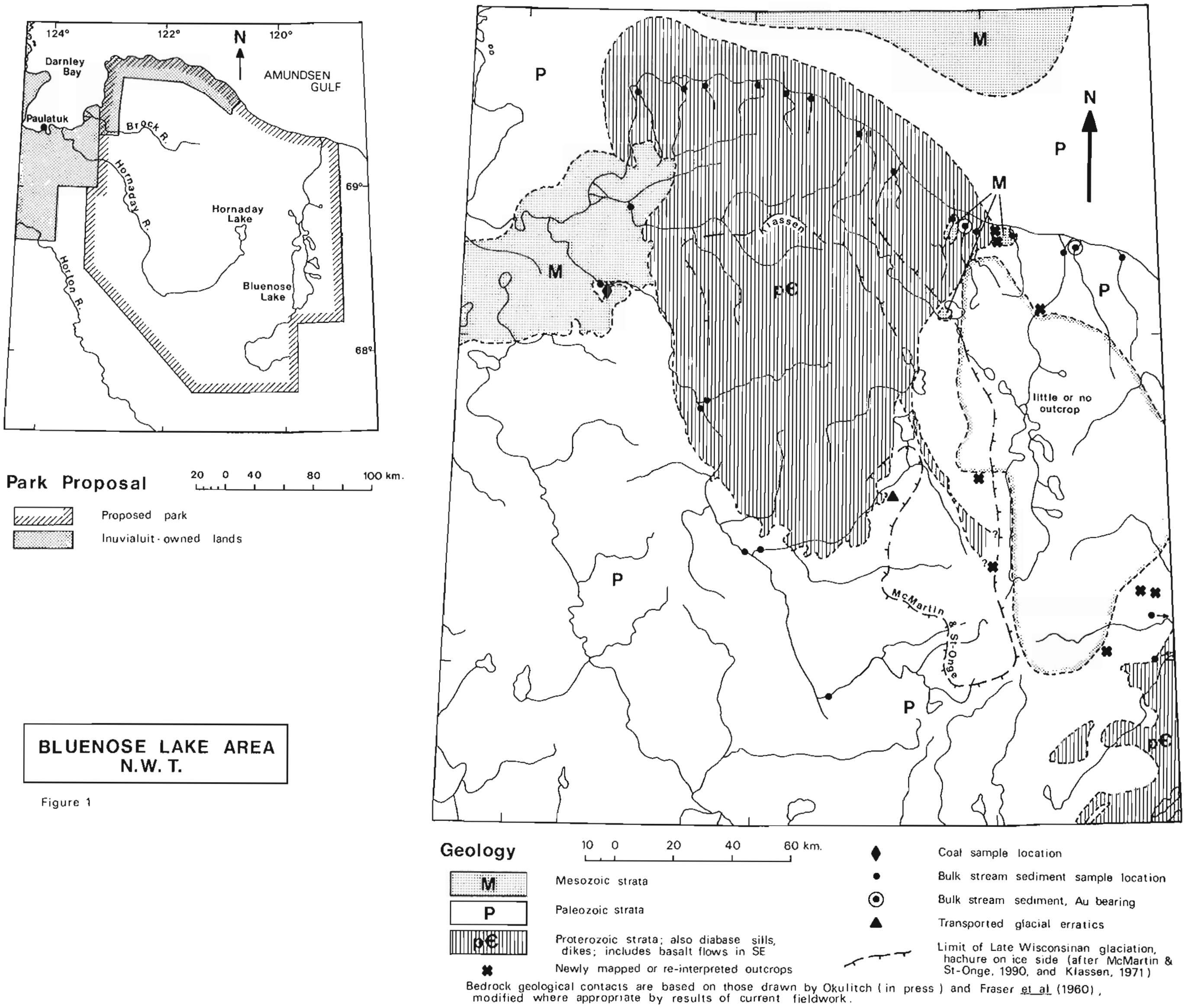

Figure 1. Map of Bluenose Lake area showing boundaries of proposed park and key sample localities mentioned in text.

The surficial geology of the region was mapped by Craig (1960), and that of most of the area of interest by Klassen (1971). St-Onge and McMartin (1987) and McMartin and St-Onge (1990) have recently mapped and sampled the Quaternary geology of the area to the east of longitude 120°, but this work is mostly outside the area of interest.

The most recent bedrock map of the area is a compilation of the entire Horton River Map sheet (NR-9/10/11/12) at scale of 1:1 000 000 (Okulitch, in press); this includes the entire area of interest north of latitude 68°.

Much reliance was placed on interpretation of air photographs in previous bedrock and surficial geological mapping. Therefore extending bedrock and surficial geological mapping and sampling across the area of interest is one of the important goals of the project.

GENERAL GEOLOGY

The bedrock geology of the area has been studied within the framework proposed by Okulitch (in press, sheet 2), and is summarized in the following based on the above-cited references and with reference to Figure 1.

The central portion of both the proposed park and the area of interest is underlain by Proterozoic sediments of the Shaler Group which form an uplifted topographic high known geologically as the Brock Inlier. These sediments have been intruded by late Proterozoic diabase dykes and sills, and are flanked on the west, south and east by onlapped Paleozoic sediments of Cambrian to Devonian age. A relatively thin, discontinuous veneer of Cretaceous sediments unconformably overlies the older rocks, mostly near the present day coastline, both west and east of the Brock Inlier. Balkwill and Yorath (1971) noted that Cretaceous coal seams are generally lignitic and of limited extent.

Quaternary glacial deposits include 1) spectacular terminal moraines and relict glacial ice around Bluenose Lake; 2) drumlin fields, and 3) glacio-fluvial outwash deposits in the northern part of the area. Surficial deposits in the central part of Brock Inlier do not to appear to have been modified by Late Wisconsinan glacial processes.

GEOLOGICAL FIELDWORK — DESCRIPTION AND PRELIMINARY RESULTS

On the basis of the above geology, 55 specific sites or outcrops were identified to be investigated in 1990. Sites were chosen to:

1) offer a view of each of the important rock units known in the area, so that each unit might be examined for economic mineral potential;

2) allow investigation of specific stratigraphic sequences and/or structures thought to be prospective for particular mineral deposit types (ie. carbonates near thrust faults for Carlin-type Au deposits)

3) map areas with little or no previously recorded outcrop (due to sparse field observations and/or mantling by Quaternary deposits).

Little or no evidence of mineralization of potential economic interest was recognized during the course of this summer's survey.

A report of possible copper mineralization in Shaler Group sediments by Young (1977) was investigated, but no copper minerals or oxidation fronts were located. Intensely hematitized Cambrian sandstone unconformably overlying Proterozoic carbonates is similar to the lower portion of the Athabasca Group in northern Saskatchewan. No secondary uranium minerals were noted, but minor thorium or uranium mineralization of sandstone type is possible.

Exposed Mesozoic coal seams were examined and sampled at the junction of George Creek and Hornaday River. Cretaceous coal beds there constitute approximately 5 m out of a 6.5 m measured thickness of friable, marcasitic sandstones and red ashy muds (evidence of partial combustion). Minor amounts of asphaltic material were noted immediately across the river from the coal, in coarse vugs of a Paleozoic carbonate unit which is tentatively assigned to the Devonian Bear Rock Formation. Samples of the coal and asphaltic material have been submitted to F. Goodarzi of ISPG for analysis.

At each outcrop visited, a sample was taken for geochemical analysis of each principal rock type, as well as any veins, contact sweats or other lithological anomalies considered to be prospective for economic mineralization. All samples have been submitted for geochemical analysis.

Fill-in bedrock mapping was undertaken in areas where previously recorded bedrock data were sparse or non-existent. The principal area in question is located between 118° 30′ and 120° longitude, and between latitude 68° and the coastline. Several previously unidentified outcrops were located, and re-examination of other previously located outcrops allows them to be assigned to specific stratigraphic units (Fig. 1, 2). All outcrops located within the above area are Paleozoic carbonates, except for two outcrops located on the coastline immediately west of Tinney Point (approx. 69° 20′ N, 119° 55′ W; see Fig. 1). Although Okulitch (in press) has identified these outcrops as Ordovician/Silurian, field evidence suggests a Mesozoic age on the basis of woody fossil remains (Fig.3). This interpretation is consistent with previous mapping (O'Neill, 1924; Fraser et al., 1960).

West of Bluenose Lake, the width of the zone which was unglaciated during the last glaciation may be quite narrow. Obviously transported erratics there (e.g. diabase, basal Hornby Bay Formation conglomerate?) overlie Paleozoic bedrock in only one location, about 40 km west of the interpreted terminal moraine (Fig. 1; see also Fig. 1 of McMartin and St-Onge, 1990). No definitive erratics could be identified north, west, and in the immediate vicinity of Hornaday Lake, confirming the lack of direct glacial effects on this area. Finally, large fragments of wood, including entire conifer stumps, were found on the shores of Hornaday Lake (elevation 517 m), at least 125 km northeast of the current treeline. The wood is believed to be eroding out of Quaternary shoreline sediments (Stephen Zoltai, pers. comm., 1990). Samples have been submitted for radiocarbon dating.

GEOCHEMICAL SAMPLING — DESCRIPTION AND PRELIMINARY RESULTS

Two types of geochemical samples were collected in summer 1990.

Rock samples, often multiple, were collected at each of the bedrock sites visited. All samples have been submitted for multi-element inductively coupled plasma analysis. In addition, specific samples considered pros-

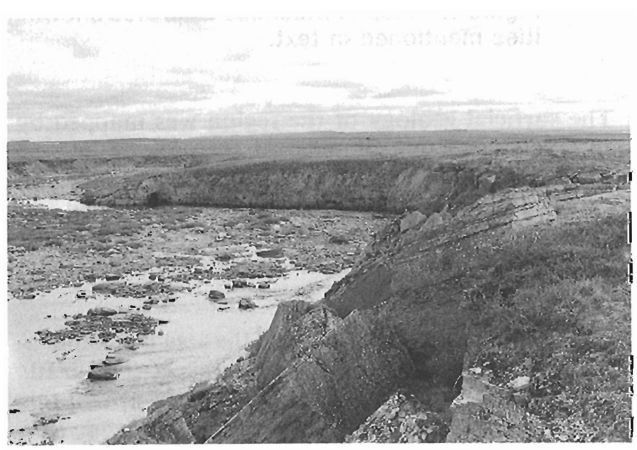

Figure 2. Typical outcrop of Paleozoic carbonate south and east of Bluenose Lake, exposed by stream erosion of a thin but continuous veneer of glacial till. Coarse vuggy horizons with drusy quartz suggest that this unit is the upper Franklin Mountain Formation, of Ordovician age (Okulitch, in press; Balkwill and Yorath, 1971, p16). Vertical exposure 4 — 7 m high.

Figure 3. Bedding plane view, Mesozoic chert pebble conglomerate. Arrows indicate fossil woody trunk and branch fragments. Hammer handle approximately 33 cm. long.

pective for gold, platinum group elements, or uranium are being analyzed by neutron activation and/or fire assay followed by direct current plasma emission spectroscopy.

Twenty eight bulk stream sediment samples were also collected from 25 first-order watersheds (Fig. 1). Samples average 8 to 10 kg and consist of field-seived (– 1cm) stream sediment material taken at the downstream apex of each watershed. Considerable difficulty was experienced in locating sufficient fine material in many of the streams. All samples were taken from active or recently active sediment, avoiding bank slump material.

All samples have been wet-sieved, and divided into + 850 micron, 850 – 177 micron, and – 177 micron fractions. A 3 gm. aliquot of the – 177 micron fraction has been submitted as a standard "silt" sample for geochemical analysis (multi-element inductively coupled plasma emission spectroscopy after extraction by nitric aqua regia). The remainder of the – 177 micron fraction was recombined with the – 850 micron fraction and passed twice on a Deister concentrating table to produce a heavy mineral concentrate (HMC) without the use of heavy liquids. The non-magnetic parts of these HMCs are being analyzed commercially for gold plus a standard multi-element package by neutron activation.

Two samples yielded single rectangular flakes of gold, measuring approximately 300 by 200 by 30 microns and 300 by 200 by 70 microns respectively. Both grains are from short rivers draining carbonate bedrock watersheds into the Arctic Ocean (Fig. 1). The source of the gold is unknown.

Evaluation of the significance of the geochemical results for both rock and stream sediment samples will be done by comparison with similar sample sets from other similar areas (particularly Jefferson et al., 1988; Spirito et al., 1988), as insufficient samples of any one type were collected to allow for statistical analysis.

SUMMARY OF RESULTS AND IMPLICATIONS FOR FURTHER WORK

Minor evidence of economic potential was noted as follows:

1) single gold flakes approximately 300 by 300 microns were separated from each of two stream sediment samples while heavy mineral concentrates were being prepared. The beaten flakes are of unknown provenance;

2) minor hematite gossan, possibly indicative of U mineralization, was found in one outcrop;

3) poor quality Mesozoic coal seams are limited in extent; and

4) minor amounts of asphaltic material are in vugs in Devonian carbonate.

90 rock samples and 25 bulk stream sediment samples have been sent for geochemical analysis; no results are yet available. Limited detailed mapping was completed between 118° 31′ and 120° longitude, in an area largely underlain by Paleozoic carbonates. Several new outcrops were located and described, several outcrops previously located were assigned to specific stratigraphic horizons, and two outcrops near the coast were tentatively reassigned a Mesozoic age.

Follow-up work (prospecting, mapping, further geochemical sampling) is being planned for those two sites already noted to contain gold. Geochemical analysis of stream sediment samples may also suggest further follow-up work when results are available. Such work would be conducted in 1991.

After all follow-up work and analyses are completed, an assessment of the mineral resource potential will be combined with an analysis of the oil and gas resource potential (to be completed by the Canada Oil and Gas Lands Administration). The combined report will be published as a GSC Open File and submitted to Parks Canada as part of the data base for public consultation.

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Crustal structures from the southwestern Grenville Province: a progress report on the processing and interpretation of reflection data in lakes Ontario, Erie and Huron

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Forsyth, D.A., Milkereit, B., Hinze, W.J., Mereu, R.F., and Asmis, H., Crustal structures from the southwestern Grenville Province: a progress report on the processing and interpretation of reflection data in lakes Ontario, Erie and Huron; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 71-76, 1991.

Abstract

Newly licensed reflection data in the Great Lakes provide first images of basement structures from the southwestern Grenville province. This study includes the first use of extended correlation during the reprocessing of marine Vibroseis data to resolve midcrustal structure. Seismic images from central Lake Ontario reveal a cross-section of the Central Metasedimentary Belt (CMB) dominated by a remarkable series of easterly dipping, apparently folded thrust sheets extending to the middle crust. Eastern Lake Erie and southern Lake Huron data image structural features near the Grenville Front. When the reprocessing is completed, the seismic data will enable an almost continuous structural transect of Paleozoic-covered Grenville basement to be constructed beneath the Great Lakes.

Résumé

Des données de réflexion nouvellement autorisées recueillies dans les Grands Lacs donnent les premières images des structures du socle dans le sud-ouest de la province de Grenville. La présente étude consiste, entre autres, à recourir pour la première fois à une corrélation élargie au cours d'une reprise du traitement des données marines obtenues par Vibroseis pour établir la structure de la croûte moyenne. Des images sismiques du centre du lac Ontario révèlent une coupe de la zone métasédimentaire centrale où prédomine une série remarquable de nappes de charriage à pendage vers l'est, apparemment plissées, atteignant la croûte moyenne. Les données provenant de l'est du lac Érié et du sud du lac Huron indiquent les caractéristiques structurales de la zone voisine du front de Grenville. Lorsque la reprise du traitement des données sera terminée, les données sismiques permettront de construire sous les Grands Lacs un transect structural quasi continu du socle de Grenville recouvert de roches paléozoïques.

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INTRODUCTION

More than 1300 km of Great Lakes seismic data have been licensed recently by the Geological Survey of Canada (GSC) from Grant Norpac of Houston, Texas. The seismic data resulted from marine Vibroseis surveys conducted in 1971 and airgun surveys completed in 1985. Part of the interpretation presented here is based on the first use of extended correlation of marine Vibroseis data in the study of deep crustal structure. These data, from Lakes Ontario, Erie and Huron (Fig. 1), complement GLIMPCE profiles in northern Lake Huron (Green et al., 1988), the COCORP reflection profiles in Ohio (Pratt et al., 1989) and Michigan (Zhu and Brown, 1986) and the Columbia Gas Transmission data in Ohio and northwest Pennsylvania (Beardsley and Cable, 1983). The seismic data beneath the Great Lakes will provide (a) an important calibration of the magnetic and gravity anomaly patterns over the region, (b) a major link between the Paleozoic-covered Grenville basement and the mapped geology of the exposed Precambrian terranes in Canada (Davidson, 1986; Rivers et al, 1989), and (c) new constraints on proposed intra-Grenville suture zones in the northern United States (Culotta et al., 1990).

The original seismic data, which were processed to only 3 seconds, showed clear indications of reflections from the deeper parts of the sections. Reprocessing tests showed that significant enhancement of many of the structural features was possible and suggested that custom processing was warranted to optimize both shallow and deeper crustal structures. For example, extended correlation of marine Vibroseis data, application of pre- and post-stack deconvolution techniques and detailed stacking

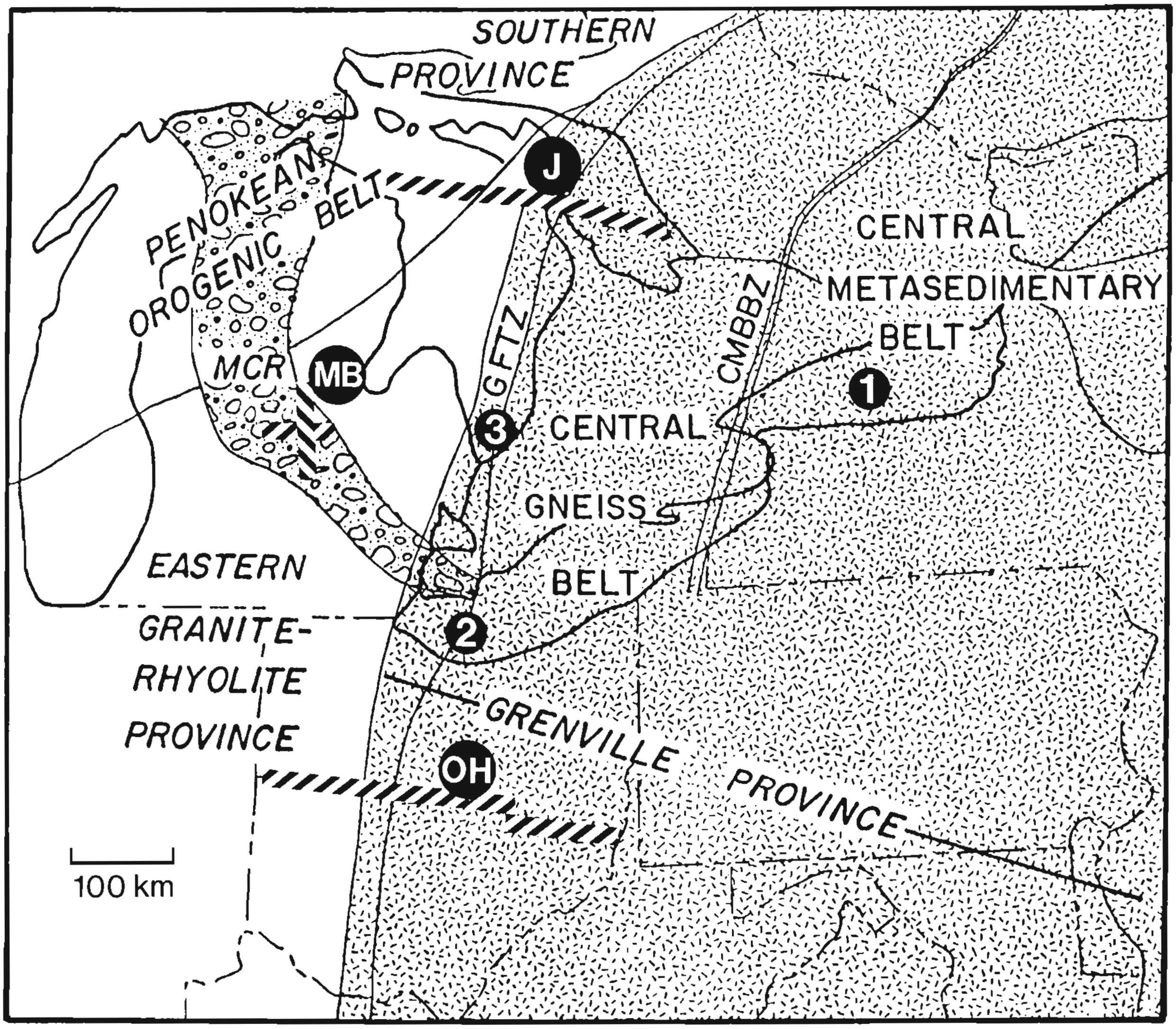

Figure 1. Locations of crustal seismic data relative to the principal Precambrian subdivisions of southern Ontario and northern United States (based on Carter and Easton, 1990). Abbreviations: MCR — Midcontinent Rift, GFTZ — Grenville Front Tectonic Zone, CMBBZ — Central Metasedimentary Belt Boundary Zone; J — GLIMPCE reflection profile (Green et al., 1988), MB — Michigan Basin profiles (Zhu and Brown, 1986), OH — Ohio profiles (Pratt et al., 1989). Numbers 1 to 3 refer to key seismic data examples described in the text.

velocity analyses were helpful in optimizing structural information to two-way times of 8 seconds or depths of approximately 25 km. As efforts are still under way to reprocess many of the data to clarify mid-crustal structures, this report describes only a selection of features evident in the "test" portions of the data. The description that follows is based on images derived from both the final stack and migrated seismic plots. Examples of data from three key areas are shown in figures 2, 3, and 4.

REGIONAL SETTING

The 1300 km of newly-licensed seismic data will enable an almost continuous structural transect to be constructed to midcrustal depths from the Grenville Province beneath eastern Lake Ontario to the Grenville Front Tectonic Zone (GFTZ) beneath Lake Huron (Fig. 1). In particular, the lines in Lake Ontario will provide a cross section from the eastern Central Metasedimentary Belt (CMB) to the possible extension of the Central Metasedimentary Belt Boundary Zone (CMBBZ) beneath western Lake Ontario (Fig. 1). The lines in Lake Erie cross the CMBBZ and the Central Gneiss Belt (CGB). Although of slightly poorer quality, the lines in Lake Huron will provide the first images of the structure near the Grenville Front and its junction with the CAN-AM structure (CAS), a major circular feature outlined from potential field data in southern Lake Huron (Forsyth et al., 1990).

CENTRAL LAKE ONTARIO SECTION

The Lake Ontario data provide the first seismic images of structural relationships between the terrane subdivisions of the CMB. The correlation of the structures with the mapped geology will provide strong constraints on interpretation of the tectonic development of the CMB.

Seismic data from central Lake Ontario (area 1 in Figure 1) show a 40-50 km wide zone of strong, easterly dipping reflections typified by the style shown in Figure 2. The structure appears to be a sequence of repeated folds successively truncated along easterly dipping fault planes that suggest thrusting toward the west. The axial regions of these folds have typical lateral dimensions of 10 km and occur at times between 1 and 7 seconds (Fig. 2). The easterly dips vary from 20 to 30 degrees and events extend to the middle crust. To the west, the repeated thrust pattern terminates against a strongly reflective region, similar to the fabric seen at the lower left in Figure 2. To the east, the crust is significantly less reflective with general indications of shallow to subhorizontal dips. Thus, the zone in central Lake Ontario is distinct from the areas on either side.

Magnetic anomalies associated with the Sharbot Lake Terrane, a subdivision of the Elzevir Terrane mapped on the exposed Frontenac Axis (Moore, 1982; Easton, 1988), can be traced southwards to the zone typified by the structures shown in Figure 2. To the southwest in New York state, the magnetic anomalies suggest continuity with a zone that contains faulted Middle Ordovician through Silurian rocks called the Clarendon-Linden Fault System (Van Tyne, 1975; Hutchinson et al., 1979). The Clarendon-Linden Paleozoic system, which is characterized by offsets of up to 30 m, is narrower than the magnetic and seismic expression of the Precambrian Sharbot Lake Terrane. The dips, dimensions and structural juxtaposition of the interpreted thrust sheets within what is probably the extension of the Sharbot Lake Terrane beneath Lake Ontario are similar to those interpreted for thrust sheets in the Haliburton area to the northwest (Hanmer et al., 1985). The reflection fabric also resembles that of the mid-crustal section from the eastern portion of GLIMPCE Line J in Georgian Bay

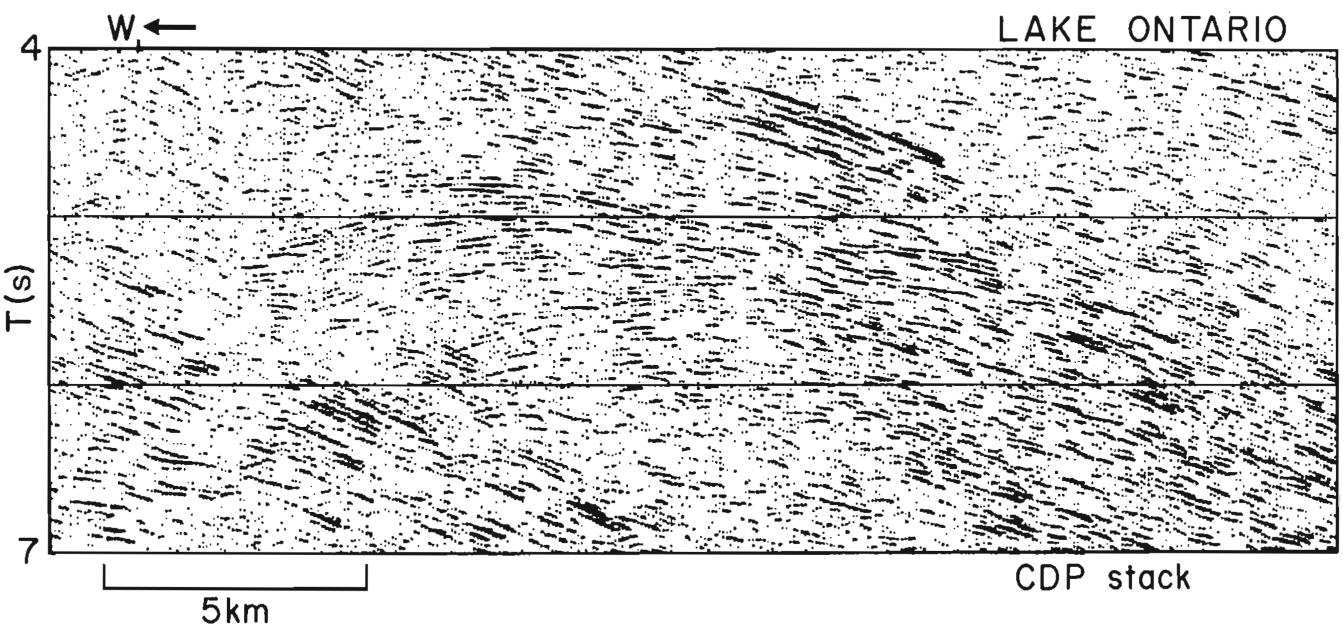

Figure 2. Section 1. An example of mid-crustal structure from migrated marine Vibroseis data continuation of the Sharbot Lake Terrane beneath central Lake Ontario.

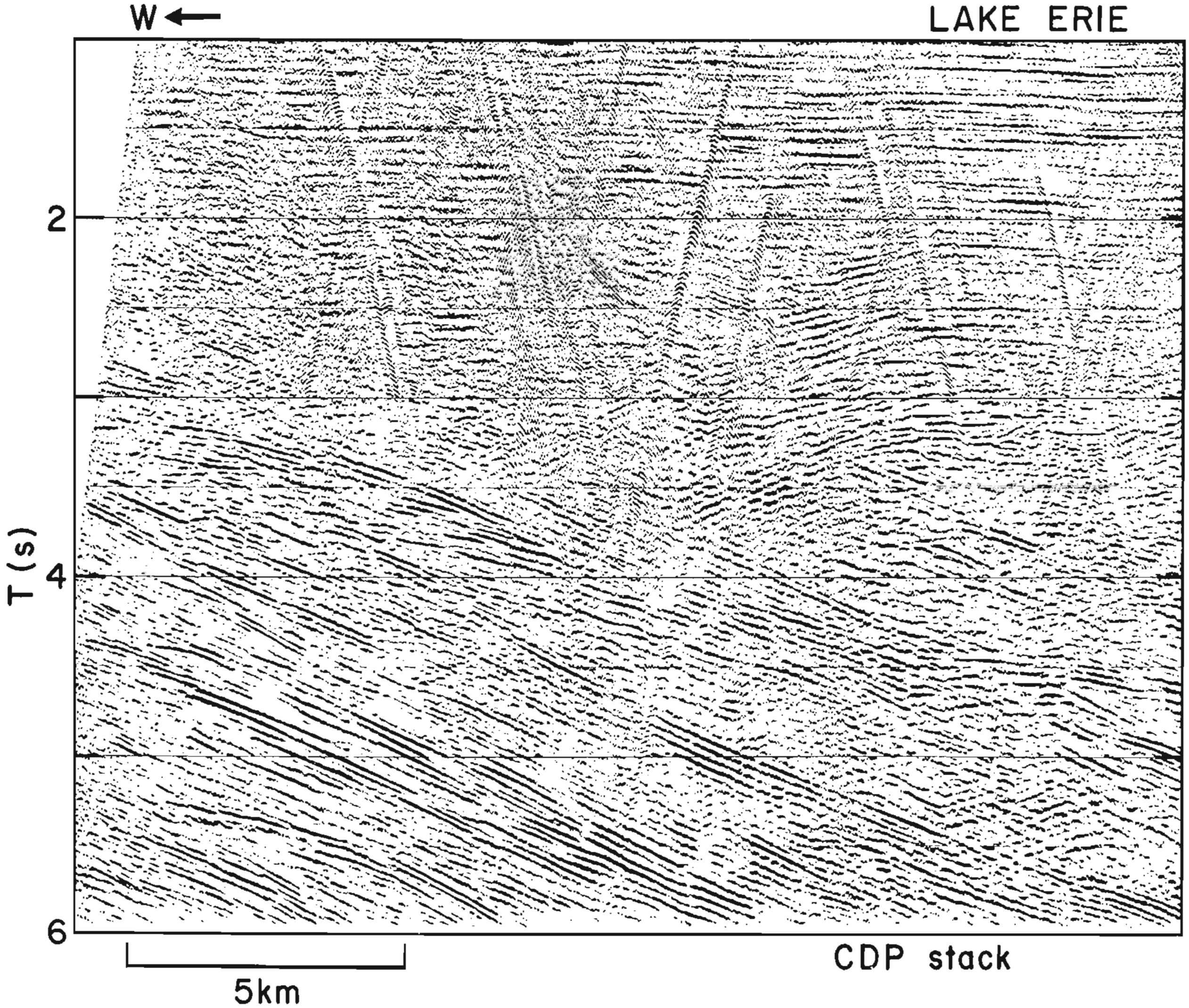

Figure 3. Section 2. An example of mid-crustal structure beneath western Lake Erie. Note the truncation of an upper complex zone that includes moderate westerly dips by a lower strongly reflective zone of easterly dipping structure resembling structures imaged on GLIMPCE Line J in northern Lake Huron.

(Green et al., 1988; Milkereit et al., 1990) and that of the shallow seismic sections in the northern Unites States shown by Beardsley and Cable (1983). Such similarity from widely separate areas suggests that an easterly dipping structural fabric is the norm for the southwest Grenville province.

LAKE ERIE SECTION

Data example 2 is from an area beneath western Lake Erie (see Figure 1). The seismic section (Fig. 3) shows a mid-crustal structure at between 2 and 3 s with westerly dips, terminating abruptly against a sequence of very strong easterly dipping reflections that begins at about 3 s (Fig. 3) and continues to the middle crust. The reflections have easterly dips of 30 degrees and resemble those of the GFTZ seen on GLIMPCE line J in northern Lake Huron. However, the Lake Erie data are significantly east of the southern extension of the Grenville Front suggested by Lidiak and Hinze (1990) on the basis of magnetic trends in Lake Huron. Pratt et al., (1989) have identified a 40-50 km wide area of east-dipping reflectors as the GFTZ in central Ohio. From the south, Lucius and Von Frese (1988) summarized core data that suggest the Grenville Front and associated tectonic zone continue north through Ohio toward the western end of Lake Erie. From the north, Carter and Easton (1990) summarized core data that indicate the eastern edge of the GFTZ lies beneath western Lake Erie close to the position suggested by the seismic data. Additional data are still required to define the Grenville Front and GFTZ between Lake Erie and GLIMPCE Line J in Lake Huron.

LAKE HURON SECTION

Compared to the Lakes Erie and Ontario examples the southern Lake Huron data (area 3 in Figure 1) are located beneath a thicker sedimentary section on the eastern

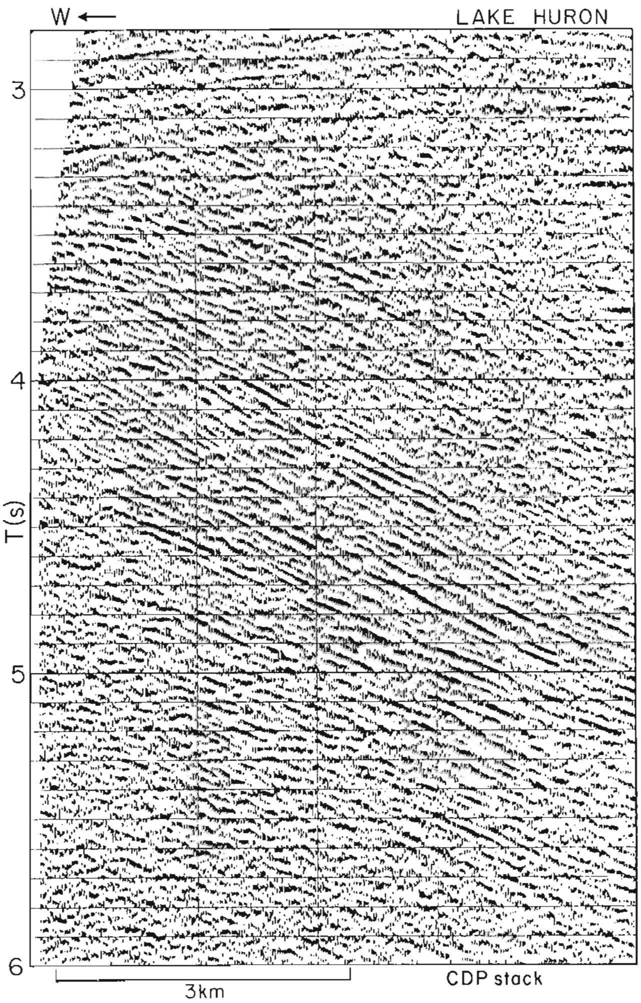

Figure 4. Section 3. Part of an east-west section in southern Lake Huron. The east-dipping reflectors correlate with south-dipping reflectors observed on a north-south line and indicate a true dip to the southeast. Although probably part of the GFTZ, the structure may also reflect effects of the CAN-AM Structure.

margin of the Michigan Basin. The seismic signal from the Precambrian basement therefore suffers more attenuation, and structural resolution is generally less than beneath the other lakes. In area 3, near the northern edge of the CAS, the midcrustal section below 3.0 seconds (Fig. 4) is characterized by a sequence of easterly dipping reflections. The similarity with the structure within the GFTZ in northern Lake Huron suggests that area 3 is within the GFTZ in southern Lake Huron. However, along an intersecting north-south line these reflections correlate closely with south-dipping events, thus indicating a true dip to the southeast. The structure therefore also dips towards the centre of the CAS (Forsyth et al., 1990). Distinguishing between east to southeast dipping Grenville Front structures and the CAS anomaly requires additional data. Furthermore, to understand the complications associated with the potential superposition of Grenville Front Structures, effects of the Mid-Continent rift system immediately to the west, features associated with subdivisions within the western CGB and the effects of the CAS will require further work.

ACKNOWLEDGMENTS

The reflection data were provided to the Geological Survey of Canada (GSC) from Grant Norpac of Houston, Texas under the terms of a license agreement between the GSC and Grant Norpac. The data processing group at Grant Norpac has been very helpful in optimizing processing parameters. Support for the purchase of the data was provided under the terms of a memorandum of understanding between the GSC and Ontario Hydro. Constructive comments on the manuscript by A.G. Green, D. White and R. Kurtz are greatly appreciated.

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Lithotectonic studies in the Central Metasedimentary Belt of the southern Grenville Province: lithology and structure of the Saint Jovite map area, Quebec

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Abstract

The Saint Jovite map area includes the highly strained gneisses of the Labelle ductile shear belt between the Central Metasedimentary Belt (CMB) in the west and the Morin terrane (MT) in the east. In the Morin terrane, charnockitic rocks from the Morin anorthosite complex aureole dominate. In the Central Metasedimentary Belt, metaplutonites consist of porphyritic monzonite complexes and are subordinate to metasediments and hornblende- and biotite-bearing quartzofeldspathic orthogneisses, interlayered with amphibolite masses. Granulite facies metamorphic assemblages occur in the east whereas lower grade assemblages occur in the west. A kilometre-scale S-shaped fold with a moderately plunging SSW axis and a complex dome characterize the Labelle shear belt in this area. This structure appears as a late deformation, later than the establishment of the orthopyroxene isograd and the juxtaposition of terranes on both sides of the Labelle shear belt.

Résumé

La carte de la région de Saint-Jovite comprend les gneiss très déformés de la zone de cisaillement ductile de Labelle entre la zone métasédimentaire centrale à l'ouest et le terrane de Morin à l'est. Des charnockites de l'auréole du complexe anorthositique de Morin dominent dans la partie est. Par contre, à l'ouest, les plutonites consistent de monzonite porphyritique et sont subordonnées aux métasédiments et gneiss quartzofeldspathiques à biotite et hornblende intercalés avec des masses d'amphibolites. On trouve, à l'est, des assemblages métamorphiques du faciès des granulites tandis que ceux à l'ouest sont de plus faible degré métamorphique. Un pli en S à plongée sud-sud-ouest et un dôme complexe caractérisent la zone de cisaillement. Cette structure apparaît comme une déformation tardive qui succède à la formation de l'isograde de l'orthopyroxène et à la juxtaposition des terranes de part et d'autre de la zone de cisaillement de Labelle.

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INTRODUCTION

Understanding the relationship between the Central Metasedimentary Belt and the Morin terrane hinges on an improved understanding of the Labelle shear belt, which separates these two regions, and the history of igneous and metamorphic activities before and after the terrrane boundary was established.

Geological field investigations summarized in this report are based on 1:50 000 mapping of the Saint Jovite map area (NTS 31J/2; 75°00'-74°30'N, 46°00'-46°15'W; Fig. 1) by J. Martignole (1988 to 1990), L. Corriveau (1990), G. Engelbretch and L. Corriveau (1981), and J. Lévesque and J. Marchand (1989) (unpublished reports). Data from all preceeding sources will be integrated within a 1:50 000 scale geological map sheet (31J/2).

The Saint Jovite map area (Fig. 2) is underlain by Precambrian rocks, of Mid- to Late-Proterozoic age, except for a few diabase dykes. The rocks fall into two main groups forming the Central Metasedimentary Belt to the west and the Morin terrane to the east, separated by a major terrane boundary, the Labelle shear belt (LSB: Indares and Martignole, 1985; Rivers et al., 1989). East of the Labelle shear belt, the rocks are of dominantly plutonic origin, most of them being charnockitic and within the aureole of the Morin plutonic complex (Martignole and Schrijver, 1970). West of the Labelle shear belt, the rocks are dominantly of metasedimentary origin (and some may be metavolcanic) with two 15 km^2-size plutons and a few mappable gabbro sills and masses. The dominant tectonic grain in the area is north-south. In the Labelle shear belt, many rocks display evidence of high strain.

Previous work

The Saint Jovite map sheet was first mapped by Osborne (1936; east-half) and by Pollock (1955; west-half). The area was later included in the Mont Laurier — Kempt

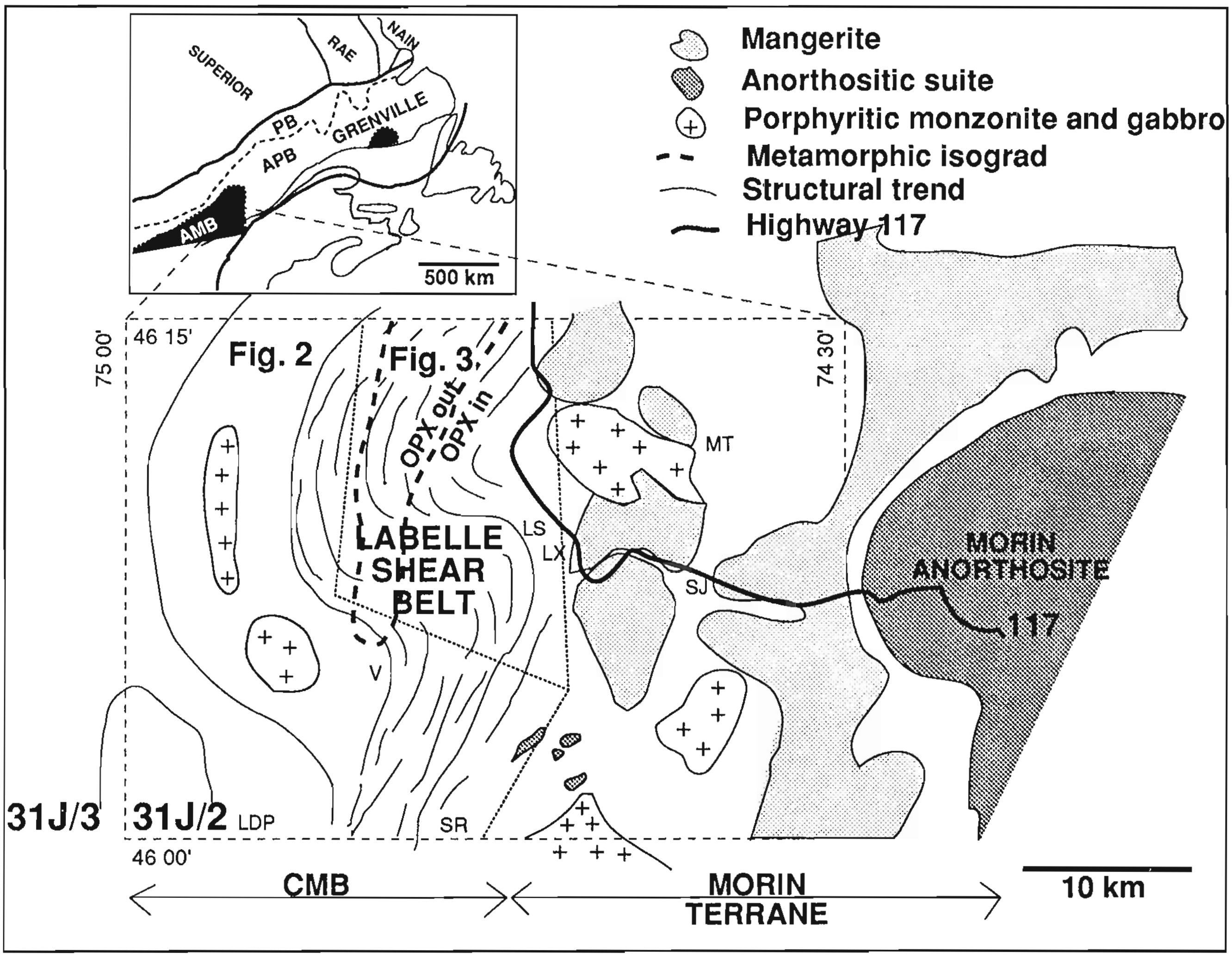

Figure 1. Divisions of the Grenville Province (inset; Rivers et al., 1989) and details of the Central Metasedimentary Belt in southern Québec showing plutonic suites of the Mont-Laurier area and Morin terrane. Abbreviations: AMB — allochthonous monocyclic belt; APB — allochthonous polycyclic belt; PB — parautochthonous belt; LS — Lake Simon; LDP — Lac des Plages; LX — Lake Xavier; OPX — orthopyroxene; SJ — Saint-Jovite; SR — Saint Rémi; V — Vendée.

Lake map of Wynne-Edwards et al. (1966). In his subdivision of the Grenville Province, Wynne-Edwards (1972) recognized the highly strained rocks (cataclasites of Wynne-Edwards et al., 1966) in the Saint Jovite area as part of a major boundary between the Central Granulite terrane to the east and the Central Metasedimentary Belt to the west. The southeastern part of the present study area was also included in a study of the Morin anorthosite complex by Martignole and Schrijver (1970) and more recently in a metamorphic study by Indares and Martignole (1990).

GNEISSES AND MARBLE

Lithological units in the Saint Jovite area consist of a series of supracrustal rocks including marble (calcareous or dolomitic), calc-silicate rocks, quartzite and impure quartzite, amphibolite, graphite and sulphide-bearing quartzofeldspathic (rusty) gneisses and metapelitic (sil-grt-bt) and semipelitic (bt-grt) gneisses with rare kyanite (Fig. 4,5; mineral abbreviations, Kretz 1983). Rocks of orthogneissic origin consist of massive to foliated amphibolite interstratified with gneissic rocks and homogeneous granitic gneisses and hornblende biotite andesine gneisses. These rocks are crosscut by rare mafic dykes that were deformed and metamorphosed to amphibolite facies.

Amphibolite

Amphibolites form units, metres to tens of metres thick, of alternating, centimetre to decimetre scale, melanocratic and mesocratic layers. This layering may have been enhanced by deformation. Clinopyroxene, hornblende and biotite are the main mafic minerals whereas orthopyroxene, rare in granitic veins, is common in crosscutting noritic veins (Fig. 5B). Garnet is locally present in the amphibolite and commonly rimmed by plagioclase. This rock type alternates with pink quartzofeldspathic gneisses (e.g., central part of the map east of Lake Xavier and north of Lake Simon; Fig. 1,3,4A-D,5B). These sequences may correspond to a basalt-rhyolite bimodal volcanic pile as they have good compositional layering with garnet and occur within supracrustal sequences.

Graphite and sulphide bearing quartzofeldspathic (rusty) gneisses

Quartzofeldspathic rusty-weathering gneisses are common in the map area (Fig. 2). Their accessory minerals consist of disseminated graphite and biotite flakes, locally garnet, almost invariably pyrrhotite and more rarely pyrite and chalcopyrite. A similar rock type with a coarse to pegmatitic grain size often occurs at the interface between marble and quartzofeldspathic gneiss. Rusty-weathering gneiss is intercalated with quartzite, semipelitic and pelitic gneisses, impure marble and calc-silicate rock often at the outcrop scale.

Marble (calcareous or dolomitic), calc-silicate rocks, skarns

Marble, which forms units metres to hundreds of metres thick throughout the map area, is white on fresh surface, medium- to coarse-grained and granoblastic except in mylonitic zones where it is very fine grained and dark to pale grey. The purest marbles have dominant calcite with about 5% diopside, local sparse titanite and graphite. Accessory minerals are mostly dolomite, plagioclase, tremolite, vesuvianite and wollastonite. Marbles, except where present as thin units, display a chaotic outcrop pattern with numerous dislocated fold hinges and contorted layers of calc-silicate rocks, amphibolites or pegmatites which were apparently less ductile than the surrounding marbles at some stage of deformation. These intense deformation features indicate that the measurable thickness of marble units is not representative of the original thickness.

Calc-silicate rocks and quartzofeldspathic diopside- and graphite-bearing gneisses occur as erratic pods and decimetre-scale layers within the marble units. These calc-silicate rocks are rich in diopside and locally contain grossularite, wollastonite, titanite, scapolite, quartz, K-feldspar and plagioclase. Calcite-bearing amphibolite are also locally present (Fig. 4F). Regular layers of calc-silicate rocks within marble gives the marble a more coherent outcrop pattern and suggests that these calc-silicate rocks were probably derived from the metamorphism of marly sediments.

Skarns occur as massive, medium- to coarse-grained, diopside- and tremolite-bearing rocks, with minor grossularite, titanite, wollastonite and quartz (Fig. 5A). These skarns are most abundant in the eastern part of the map where they occur as inclusions of various size within charnockitic gneisses and plutonic rocks.

Metapelitic (sil-grt-bt), semipelitic (bt-grt) gneisses and quartzite

Aluminous metasedimentary rocks occur throughout the entire map area as medium-grained, garnet- and biotite-bearing gneisses locally with sillimanite. Kyanite has been observed in one outcrop at Lake Xavier. These rocks are rusty on weathered surfaces and black to bluish on fresh surfaces. They form centimetre-metre-scale layers, intercalated with quartzite and quartzofeldspathic gneisses and are invariably invaded by a network of quartzofeldspathic veins. This rock assemblage may locally reach a few hundreds of metres in thickness. Sillimanite has been observed to form 20% of some layers up to 1m thick. Black tourmaline may be an abundant accessory mineral but no tourmalinite layers have been observed. Garnet (1-10 mm in diameter) is common. It has a typical lilac colour when associated with sillimanite and biotite, otherwise it is red. Garnetite (< 10 cm thick) is rare. Locally graphite and sulphide (usually pyrrhotite) are present.

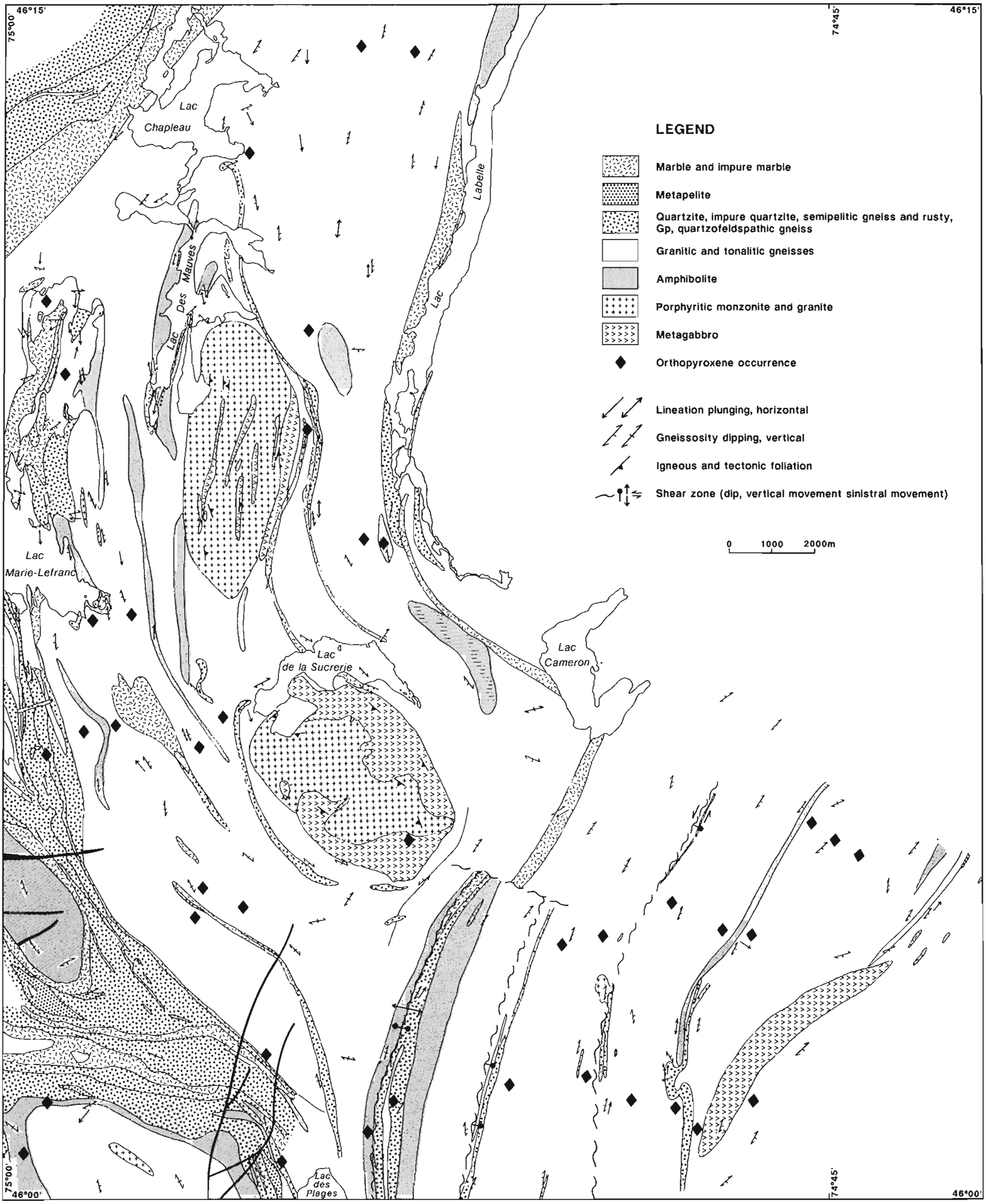

Figure 2. Geological map of the western portion of the 31J/2 map sheet based on geological data collected by L. Corriveau, Rio Algom (Map G4757 by G.H. Engelbrecht, pers. comm. to L. Corriveau, 1990) and SOQUEM (Lévesque and Marchand, pers. comm. to L. Corriveau, 1990).

Quartzite (qtz >90%) and impure quartzite form units up to a few hundreds of metres thick. Impurities consist of feldspar, garnet and biotite. Layers of fine grained amphibolite and hornblende- and biotite-bearing quartzofeldspathic gneisses are locally intercalated with quartzites. The thickest and most regular quartzite layers occur in the western part of the map area.

Rocks of mainly orthogneissic origin

Amphibolite masses within the supracrustal rocks tend to be elongate, massive, granoblastic and medium grained, with a few small remnants of an ophitic texture with orthopyroxene among laths of plagioclase. Compositional layering and a well developped metamorphic mineral foliation occur locally. The mineral paragenesis

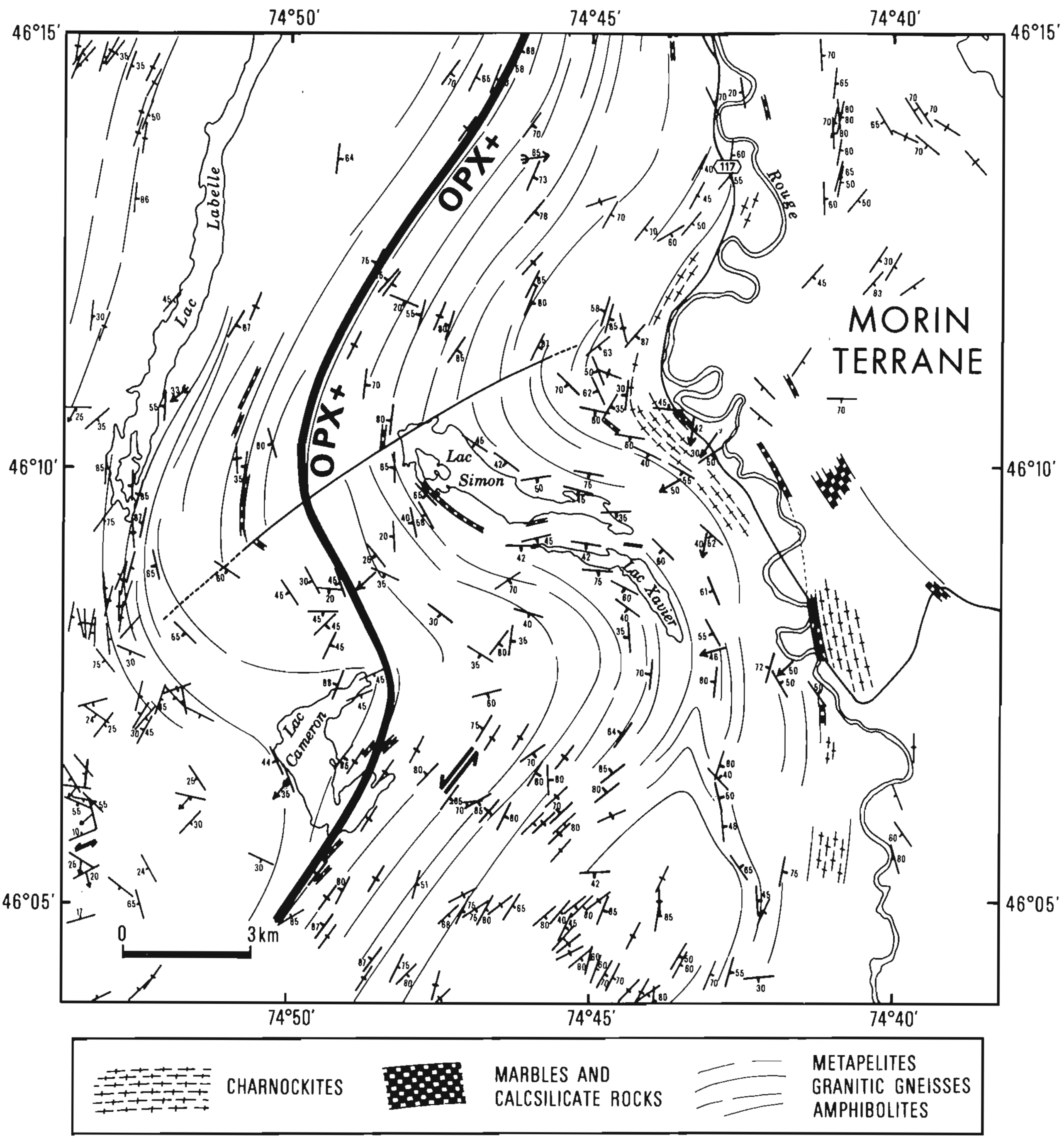

Figure 3. Structural map of the Lac Simon S-shape fold in eastern portion of the 31J/2 map sheet based on geological data collected by J. Martignole.

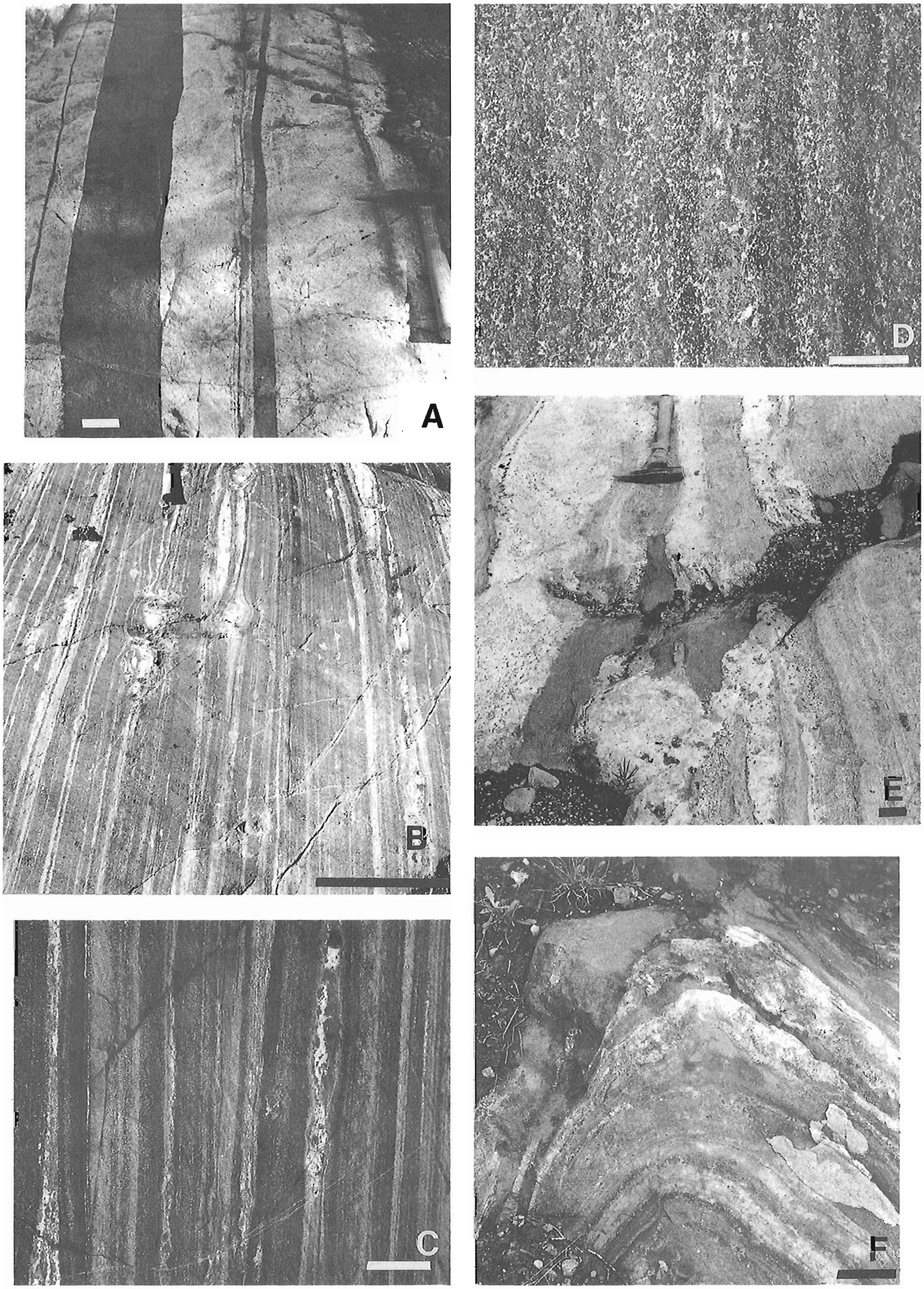
A
B
C
D
E
F

Figure 4. A) Pink, fine-grained quartzofeldspathic gneisses intercalated with amphibolite layers near Lake Xavier (scale bar: 10 cm).
B) Layered amphibolite interpreted to be metavolcanic with transposed leucosomes near Lake Xavier (scale bar: 1 m).
C) Details of layered amphibolite showing compositional layering (scale bar: 10 cm).
D) Detail of layered amphibolite showing garnet- and clinopyroxene-bearing layers intercalated with garnet- and hornblende-bearing layers (scale bar: 1 cm).
E) Hornblende- and biotite-bearing grey quartzofeldspathic gneisses with leucosomes along the metamorphic foliation crosscut by fine-grained mafic amphibolite dykes (scale bar: 10 cm).
F) Folded layered amphibolite and calc-silicate rocks crosscut by a fine-grained amphibolite dyke (scale bar: 10 cm).

(Color index [CI]: 20-30) consists of hornblende and plagioclase with accessory quartz, magnetite, apatite and titanite and indicates that recrystallization took place at amphibolite facies. Folded and foliated dykes of mesocratic amphibolite (CI: 40) and monzogranitic augen gneiss (CI: 15) occur locally. Granitic veins are also common. In the less deformed part of the amphibolite masses, fine grained lenticular mafic enclaves are aligned and define a relict igneous foliation. Where solid-state deformation occurs, these enclaves are folded and realigned along a NNW-SSE tectonic foliation and a SSW lineation also defined by the alignment of hornblende and biotite and aggregates of amphibole and plagioclase. The SSW-lineation commonly undulates around the horizontal indicating a regional E-W warping. Incipient melting is common in this unit as shown by the presence of abundant, ameboid-shaped hornblende- and biotite-bearing granitic or tonalitic leucosomes, up to 30 cm in diameter distributed homogeneously throughout the outcrops. In some locations, the leucosomes are aligned perpendicular to the tectonic foliation, in others, they parallel it. These observations suggest that some leucosomes emplaced along the structural extension direction were later rotated into the foliation plane. The paragenesis of quartz, feldspar, hornblende and biotite in the leucosomes and their orientation indicate that anatexis was later than or contemporaneous with the amphibolite facies metamorphism and NNW-SSE foliation. The sporadic occurrence of orthopyroxene in ameboid-shaped leucosomes throughout most of the western portion of the map area indicates that granulite facies conditions were locally met during partial melting. Variation in the fluid composition may have permitted orthopyroxene crystallization (Wendlant, 1981).

Grey quartzofeldspathic gneisses with hornblende and biotite and granitic gneisses dominate in the western portion of the 31J/2 map area (Fig. 2). They form discrete corridors bordered by late mylonite zones, some of which have down-dip lineations. Most of these gneisses have a paragenesis diagnostic of amphibolite facies metamorphic conditions. East of Vendée, where granulite facies is pervasive, orthopyroxene is present in these rocks and in their leucosomes. Metamorphic layering is well developped and the leucosomes occur as irregular veins parallel or oblique to the foliation. Amphibolite layers (10 cm-3 m thick) are commonly intercalated in these units and are interpreted as transposed mafic dykes. The dykes are fine grained and locally have centimetre-sized laths of plagioclase phenocrysts internally recrystallized to a white and fine-grained aggregates.

Fine-grained and granoblastic amphibolite dykes (20-50cm thick; Fig. 4E, F) crosscut the paragneissic and orthogneissic units and are themselves folded. In contrast to other amphibolite units, these dykes are generally devoid of leucosomes and appear to be among the youngest Grenvillian mafic magmatism in the area.

PLUTONIC ROCKS

Plutonic rocks dominate the eastern part of the map sheet (Fig. 1). The charnockitic rocks include charnockites, mangerites, jotunites and leuconorites associated with some granitic rocks which may be transitional into charnockitic rocks. These plutons occur only in the Morin terrane (Fig. 1). Monzonitic and gabbroic complexes with porphyritic monzonite, quartz monzonite and granite, diorite, gabbro and net-veined dykes were observed throughout the map area (Fig. 1). Pink aplitic granite and pegmatite are the latest Grenvillian magmatic activities observed in the area. Diabase dykes are most prominent in the southwestern part of the map (Fig. 2).

Noritic gabbro and anorthosite near Saint Rémi

Three masses of anorthositic and noritic rocks have been mapped at the eastern margin of the Labelle shear belt, northwest of Saint Rémi (Fig. 1,2); the gabbro (1 km by 5 km; Fig. 2) consists of coarse grained orthopyroxene and plagioclase with an ophitic texture. Two masses of anorthosite have been found southeast of the gabbro mass. They consist of white, recrystallized plagioclase and coarser grey plagioclase porphyroclasts with mafic aggregates of fine grained amphibole and pyroxene surrounded by garnet coronas. These three occurrences appear on both sides of a late northeast-trending shear zone, and are themselves highly deformed at least at their borders. It is thus possible that these gabbroic and anorthositic masses represent slices of an originally larger anorthositic massif of the Morin terrane incorporated within the Labelle shear belt.

Charnockitic rocks

Charnockites (sensu lato) occur along the eastern margin of the map area and are part of the complex plutonic aureole around the Morin anorthosite (Fig. 1). Mangerites, jotunites and leuconorites associated with some granitic rocks which may be transitional into charnockitic rocks are also included in the following description. These rocks range in composition from orthopyroxene (pigeonite)-plagioclase to mesoperthite-orthopyroxene-quartz (true charnockites). Some may constitute mappable units;

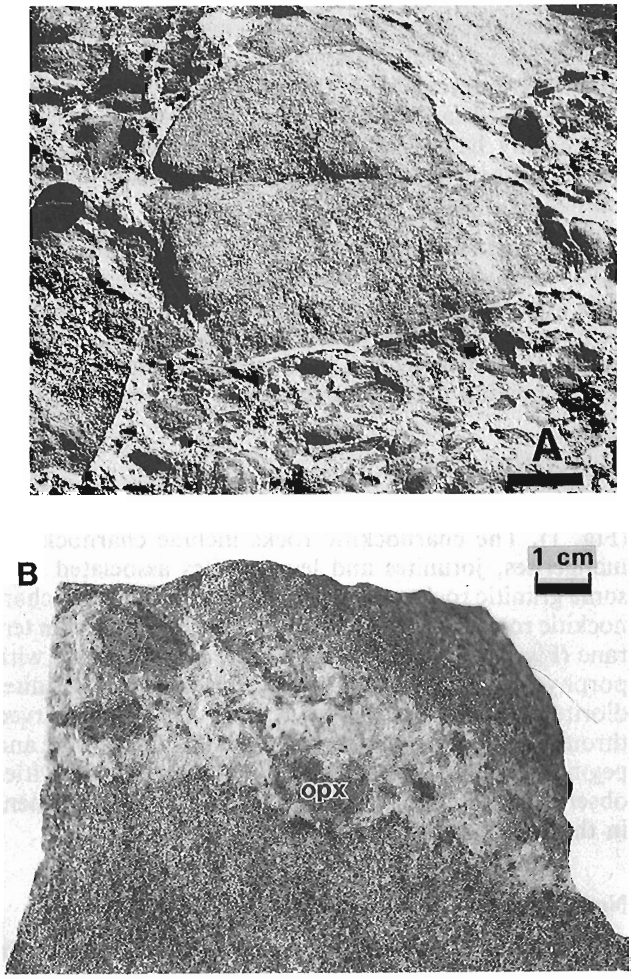

Figure 5. A) Skarns. B) Charnockitic leucosomes in mafic granulite (UTM:521900E 5122000N)

others are of highly variable composition even at an outcrop scale. Grain size of these rocks varies from fine to coarse or even porphyritic and many carry plagioclase megacrysts with Carlsbad-albite twins. Garnet is commonly present in true charnockitic rocks but rare in granitic rocks. Most outcrops include several types of xenoliths and autoliths (centimetre to 10 m thick). Detailed mapping would no doubt reveal the presence of much larger scale metasedimentary inclusions. The xenoliths, in order of decreasing abundance are quartzite, skarn, amphibolite and metaluminous gneisses. Locally (Mont Tremblant area; Fig. 1), disaggregated quartzite inclusions occur as bluish to purplish quartz seams accounting for about 30% of the rock composition. The texture of charnockitic rocks (sensu lato) varies from hypidiomorphic granular with an igneous foliation to gneissic and mylonitic. Strain gradients are sometimes high enough to see all these textural types within a few tens of metres.

Charnockitic rocks (sensu stricto) away from the Morin anorthosite may grade into coarser-grained, hornblende- and biotite-bearing monzonitic to quartz monzonitic rocks, many of which are porphyritic.

Porphyritic monzonite, quartz monzonite and granite, metagabbro

Two monzonitic complexes (15 km^2) occur west of the Labelle shear belt. They consists of: 1) foliated porphyritic monzonite (CI: 15) with 10 to 20% coarse laths of K-feldspar in a fine grained matrix of plagioclase, hornblende and biotite, 2) foliated equigranular biotite-granite or porphyritic biotite-and hornblende-bearing granite, and 3) granoblastic amphibolite or medium grained noritic metagabbro with relict ophitic texture. Foliation in the monzonite is defined by the alignment of K-feldspar laths, biotite, hornblende and cogenetic enclaves. The laths of K-feldspar are euhedral and their long axis locally defines a lineation. In zones intensely deformed and recrystallized, the monzonite is an augen gneiss with lenticular K-feldspar surrounded by a fine-grained aggregate similar to that of the matrix of plagioclase, biotite and hornblende. In the granite, the foliation is defined by elongate quartz grains or by K-feldspar laths. In the metagabbro, relict plagioclase laths, with an internal granoblastic texture, are aligned parallel to the post-consolidation foliation defined by biotite and hornblende.

The monzonitic and metagabbroic rocks of the Lac des Mauves complex (Fig. 2) occur as sheet-like bodies parallel to the north-striking, subvertical structural trend of the country rocks. The monzonite was emplaced as homogeneous sheets or is interdigitated with dioritic material, having numerous elongate microdioritic inclusions (4 cm-2 m width) with aspect ratios between 5:1 to 20:1, or large sheets of medium-grained diorite and gabbro, now recrystallized to an amphibolite. Hybrid zones with a net-veined texture are common, the microdiorite inclusions are granular and elongated parallel to the foliation in the monzonite, they may enclose some porphyritic K-feldspar. Leucosomes are rare in the monzonite but are present in the metagabbro. The net-veined zones and the monzonite appear largely to have retained their igneous texture and foliation. The parallellism between the regional foliation, the intrusive sheets and the igneous and post-consolidation foliation and the near absence of leucosomes suggest that this body was emplaced after peak metamorphism and partially recrystallized at amphibolite grade during the last episode of deformation.

The Lac des Sucreries complex (Fig. 2) occurs within a complex domal structure located west of the major S fold of the Labelle shear belt. The base of the dome is paragneissic with layers of marble, quartzite and calc-silicate rocks. At the contact with the intrusion, quartzo-feldspathic gneisses intercalated with layers of quartzite 10-20 cm thick are most common. Structurally overlying these paragneisses is an undulating sheet of porphyritic leucocratic monzonite, overlain by intercalated grey hornblende and biotite-bearing tonalitic gneisses and medium grained amphibolite and metagabbro with relict ophitic texture. Hornblende- and biotite-bearing quartz monzonite and granite also occur. A few zones of an alkaline olivine and biotite-bearing melagabbro (CI: 85) occur in the centre of the gabbro sheet. The biotite forms thin randomly oriented plates up to 2 cm in length that are characteristic of olivine-bearing alkaline mafic rocks. All units show a range in the extent of their recrystallization, from massive (especially the gabbro) to having a poorly defined subhorizontal foliation and lineation, to a layered gneiss with hornblende and orthopyroxene-bearing leucosomes and irregular veins in both the monzonite and the amphibolite rocks. East-striking shear zones are common. Some have a moderate dip to the south and a down-dip lineation parallel to the Labelle shear belt stretching lineation. The monzonitic pluton at Lac des Sucreries appears to be more deformed than the Lac des Mauves complex. Whether the extent of deformation reflects a variation in age of emplacement of the complexes or is a function of their respective locations with respect to the major S fold of the Labelle shear belt cannot be established at present. The later interpretation is more likely considering the similarities in rock types which suggests a magmatic association between these two monzonitic complexes.

Granites

Pink aplitic leucocratic biotite granites form discrete masses in the southeast corner of the map and are probably late compared to the porphyritic monzonites in which they occur as metre-scale dykes. White medium grained granites occur as small masses and dykes in the western part of the map area.

Diabase dykes

A swarm of E-W- or N-S-striking diabase dykes (10 cm-10 m thick; Fig. 2) occur in the western part of the map area. The dykes have a strong aeromagnetic signature and are visible on air photographs. The diabase are fine to medium grained, orthopyroxene is common.

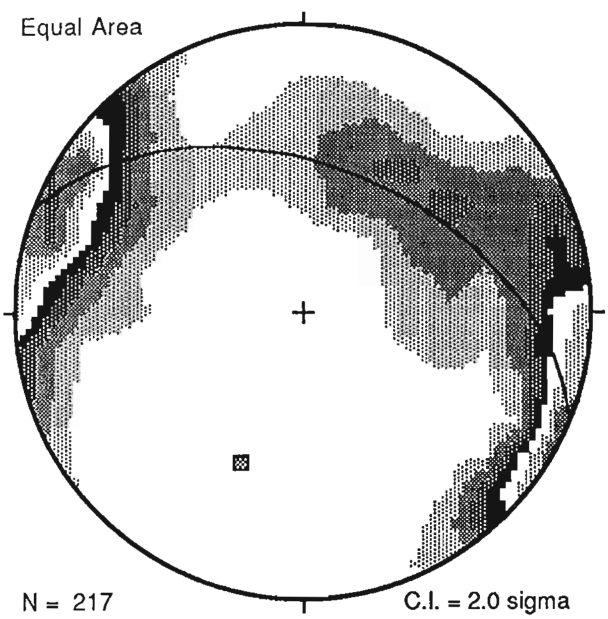

Figure 6. Stereographic projection of foliation planes from the Lac Simon fold of the Labelle shear belt.

STRUCTURE

Folds

The Labelle shear belt (LSB: Indares and Martignole, 1985, 1989, 1990) intersects the Saint Jovite map area around latitude 74°45′. Instead of running N-S as in the area just north of the Ottawa river, it is folded into an S-shaped, plurikilometric reclined fold. Foliation planes (Fig. 3,4) define a girdle with an axis dipping 43° to the SSW. The axial plane of the fold dips steeply towards the SE. Along the limbs of this fold, as well as in its hinge, minor folds have their axial planes subparallel to the regional foliation and their axes tend to be oriented toward that of the main fold (Fig. 3,4). This geometry is interpreted as the result of uncompleted reorientation of preexisting folds during the formation of the S-shaped fold.

Ductile shear zones

Rocks of plutonic origin are not readily deformable into folds because of the lack of a ductility contrast. Deformation of plutonic rocks in the Saint Jovite map area thus results in the development of gneissic or mylonitic fabrics. Ductile shear zones intersect charnockitic as well as monzonitic rocks. Some of them in charnockites probably developed in early stages of the subsolidus history of the rocks as attested by the totally recrystallized, polygonal texture of the associated mylonites. The majority of mylonite zones, however, seem to have been formed under amphibolite grade conditions, as attested by the disappearance of orthopyroxene and its replacement by hornblende or biotite in the mylonites. The orientation of high strain zones in the plutonites is variable, dominantly NE-SW in the southeast corner of the map, and NS in the northeast corner.

Evidence of high strain is also found in metasedimentary rocks and in orthogneisses, many of which can be described as straight gneisses or as mylonites. This type of deformation is visible throughout a roughly 10 km-thick strip of subvertical gneissic rocks, limited to the east by highway 117 and to the west by Lake Labelle and the village of Vendée (Fig. 1,2,3). Several quartzofeldspathic mylonites have been observed along this strip, with kinematic indicators in agreement with a transcurrent sinistral sense of shear (Fig. 2,3). A few ductile mylonite zones at amphibolite facies with down-dip lineation occur in the south central part of the map area; the best exposed example occurs northeast of Lac des Plages (Fig. 2).

Brittle fracturing

Evidence for brittle deformation is not widespread, but brecciated rocks were found near Lake Simon; they are compatible with the existence of a NE-SW-trending fault visible on air photos and satellite images. This fault, being subparallel to the axial plane of the S-shaped fold, could well originate, in a late stage, through the same deformation event which produced the fold.

METAMORPHISM

The largest part of the area is in the granulite facies, as attested by the presence of orthopyroxene in most mafic and intermediate quartzofeldspathic rocks. Several observations attest for a slight decrease in the metamorphic conditions from east to west, namely away from the anorthositic complex. In the charnockites, orthopyroxene is ubiquitous, both as a primary, igneous phase, or as recrystallized grains. At several localities, charnockites are transitional with hornblende-biotite monzonitic and granodioritic porphyritic rocks, an amphibolite facies mineralogy. Garnet coronas are common in charnockitic rocks, between plagioclase and pyroxene. They have been interpreted as resulting from subsolidus reactions of the type opx + pl = grt + qtz which could be achieved during a cooling event at high pressure (Martignole and Schrijver, 1973) or during cooling under increasing load pressure (Martignole, 1986). Granulite facies mobilisates are found at several localities in amphibolites and quartzofeldspathic (mafic-rich) gneisses but are more abundant in the eastern half of the map area and along its western border. They consist of noritic to charnockitic veins and pods, a few centimetres-thick, which in some cases give the outcrop a migmatitic aspect. These mobilizates are interpreted as resulting from partial melting under conditions of low water activity. Incongruent melting of phlogopite has been shown experimentally (Wendlandt, 1981) to produce such assemblages under temperature conditions in the vicinity of 740°C. The presence of orthopyroxene in mafic and intermediate assemblages as well as the presence of charnockitic veins has been used to delineate the orthopyroxene isograd (Fig. 1,3).

In calcareous and dolomitic marbles, high-grade assemblages are calcite-diopside-grossular-(graphite), whereas the assemblage plagioclase-diopside-grossular occurs in calc-silicate layers and skarns. The assemblage calcite-quartz is stable in the western part of the area, but it is replaced by wollastonite where marbles occur in contact with charnockites or metagabbro intrusions.

In the metapelites, the most common assemblage is garnet-biotite-sillimanite-(K-feldspar-quartz), but kyanite has been found at one locality (between lakes Xavier and Simon). The presence of both aluminium silicates in an area where the temperature has reached temperatures of about 740°C constrains the metamorphic pressure around 800 MPa.

ECONOMIC GEOLOGY

Several occurrences of sulphides have been noted in garnet-sillimanite gneisses, in rusty paragneisses and in some shear zones. An analysis of a quartzofeldspathic mylonite containing pyrite-chalcopyrite veinlets gave: 0.15% Co and 793 ppm Cu (UTM: 528000E, 5109300N), whereas 2-3 mm thick veins of sulphides along an outcrop scale zone gave 0.12% Cu and 221 ppm Co. A vein of pyrite with 20 ppb of Au occur in a shear zone (UTM: 530000E, 5113400N). The industrial minerals wollastonite, graphite, diopside, dolomite and tourmaline occur at several localities, but none are presently worked. Wollastonite is usually found at the contact between marbles and charnockitic rocks. Over thirty occurrences of wollastonite have been reported in the Saint Jovite area (Simandl, 1989a) but none is presently being exploited. Graphite, also, is widespread in calc-silicate rocks, quartzofeldspathic gneisses and metapelites (Simandl, 1989b; this work). It was mined in the past near Saint Rémi where it occurs with wollastonite. Several showings of graphite are presently under investigation by SOQUEM in the western and central parts of the map area.

CONCLUSION

The Saint Jovite map area includes part of the Morin terrane and the Central Metasedimentary Belt, separated by the Labelle shear belt, a zone of high strain affecting both metasedimentary and metaplutonic rocks. The orthopyroxene isograd of Wynne-Edwards et al. (1966) has been revised. Granulite facies occurs throughout most of the 31J/2 area with the exception of a 10km wide strip along the central part of the Labelle shear belt. Marbles, quartzites and metapelites are found on both sides of the Labelle shear belt, their distribution has been defined. Rocks of the anorthosite and gabbro suite typical of the Morin terrane occur only as far west as the eastern boundary of the Labelle shear belt. This suite appears to have been disrupted and incorporated within the Labelle shear belt as lenses a few kilometres long. Whether this terrane boundary is one along which major displacement took place is still unknown. It does display evidence for transcurrent, incremental strain and late subvertical movement.

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Lithotectonic studies in the Central Metasedimentary Belt of the southwestern Grenville Province: plutonic assemblages as indicators of tectonic setting

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Abstract

A subdivision of the Central Metasedimentary Belt of Québec into two lithotectonic terranes extending southward into Ontario has been proposed based on the distribution of potassic alkaline plutons. Inter-terrane contrasts among pre-amalgamation plutonic suites and similiarities among later ones, and the existence of major shear zones along proposed terrane boundaries would thus be expected. Sediment-hosted Zn-mineralization in the Adirondack and Mont Laurier areas occurs in different terranes and would not belong to the same basin. In order to alleviate the shortage of modern mapping and test this model, regional mapping has been initiated in the Central Metasedimentary Belt. Five plutonic associations are recognized as potential lithotectonic assemblages: porphyritic monzonite and metagabbro, charnockite and anorthosite, noritic gabbro and syenite, potassic alkaline plutons, and granite. A major, possibly inter-terrane, shear zone has been recognized, but its regional extent is presently unknown. Zn and graphite mineralization, wollastonite occurrences and potential sites for building stones have also been located.

Résumé

La subdivision de la Ceinture métasédimentaire centrale du Québec (CMB) en deux terrains lithotectoniques qui se prolongent en Ontario est principalement basée sur la répartition de plutons alcalins potassiques. Ce modèle implique des contrastes interterrains dans les suites plutoniques qui précèdent la juxtaposition de ces terrains et des similarités dans celles qui lui succèdent, et l'existence de zones de cisaillement majeures le long de ces terrains. Les minéralisations zincifères logées dans les sédiments des Adirondack et de la région de Mont-Laurier se trouvent dans des terrains différents et ne se seraient donc pas formées dans le même bassin. Afin de pallier au manque de données cartographiques récentes et pour vérifier ce modèle, un programme de cartographie régionale a été instauré dans la CMB. Cinq associations plutoniques sont reconnues comme assemblages lithotectoniques possibles: monzonite porphyrique et métagabbro, charnockite et anorthosite, gabbro noritique et syénite, plutons alcalins potassiques, et granite. Une zone de cisaillement majeure, possiblement de nature interterrain, a été reconnue mais son étendue régionale est présentement inconnue. Des minéralisations zincifères, des indices de graphite et de wollastonite et des sites potentiels pour des pierres à bâtir ont aussi été localisés.

INTRODUCTION

The purpose of this study is to examine the magmatic and tectonic evolution of the Central Metasedimentary Belt of Québec (CMB; Fig. 1) by carrying out systematic regional mapping in the Mont Laurier area and petrological studies of its plutonic suites. The basis for using plutonic suites as indicators of tectonic settings is described below, followed by a description of the plutonic suites and a discussion of their potential as lithotectonic assemblages. Sites of potential economic interest are located precisely by their UTM co-ordinates.

Geological setting and statement of problem

The Central Metasedimentary Belt (Fig. 1) is part of the allochthonous monocyclic belt (AMB; Rivers et al., 1989) and consists of supracrustal rocks (ca. 1.3 to 1.2 Ga) and plutonic suites emplaced during the Grenvillian orogenic cycle (ca. 1.3-1.0 Ga; Moore and Thompson, 1980). The Central Metasedimentary Belt has been thrust northwestward over the allochthonous polycyclic belt (APB; Rivers et al., 1989) along the Central Metasedimentary Belt boundary zone (CMBBZ; Fig. 1); a stage of this thrusting has been dated at 1060 Ma (van Breemen and Hanmer, 1986). The Central Metasedimentary Belt is separated to the east from the adjacent Morin and Adirondack terranes of the allochthonous monocyclic belt by the Labelle shear zone and the Carthage-Colton mylonite zone respectively (Fig. 1; Rivers et al., 1989).

Despite being easily accessible and hosting significant Zn and graphite mineralization, the Central Metasedimentary Belt of Québec lacks reliable geological maps; only a few of its 25 1:50 000 map sheets are up to date (Dupuis et al., 1989; Gauthier, 1982; Hébert, 1988; Langlais, 1988). In addition to Zn and graphite, rare earth and platinum group (?) elements, P, U, building stones, pure marble, dolomitic marble, micas, silica and wollastonite, locally occur in quantities and quality worthy of further exploration in this belt. Furthermore, the Central Metasedimentary Belt is considered a key area for

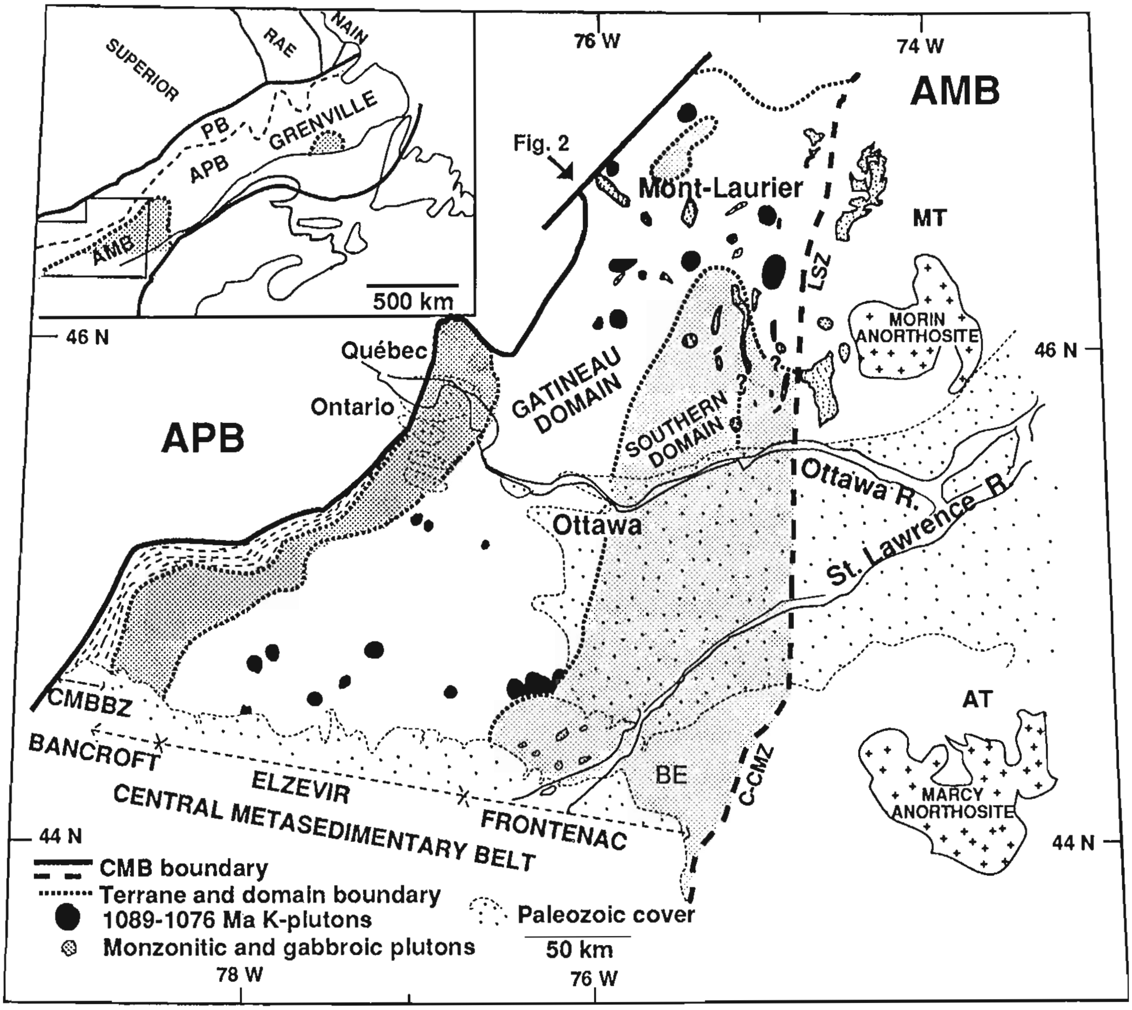


Figure 1. Divisions of the Grenville Province (inset) and the Central Metasedimentary Belt (CMB) in eastern Ontario and southern Quebec (terminology from Rivers et al., 1989, modified from Corriveau 1990). Abbreviations: AMB — allochthonous monocyclic belt; APB — allochthonous polycyclic belt; AT — Adirondack terrane; MT — Morin terrane; PB — parautochthonous belt; C-CMZ — Carthage-Colton mylonite zone; CMBBZ — Central Metasedimentary Belt boundary zone; LSZ — Labelle shear zone; BE — Balmat-Edwards Zn- deposit.

understanding the tectonic evolution of the Grenville Province and will be the host of future Lithoprobe transects. In order to alleviate the shortage of modern mapping, which has previously been an impediment to mineral exploration and tectonic investigations, a program of regional mapping in the Central Metasedimentary Belt of Québec was initiated during the 1990 summer. This work is taking place in conjunction with others with the aim of jointly gaining a regional perspective on the economic potential and tectonic evolution of the southwestern Grenville Province.

During the summer of 1990, six weeks were spent carrying out 1:50 000 mapping to complete the western portion of the Saint Jovite sheet (NTS 31J/2; Martignole and Corriveau, 1991). This mapping was followed by a field investigation of plutonic suites and their country rock in the Mont Laurier and eastern Gatineau areas.

Method and model

The study of the tectonothermal evolution of deep crustal Precambrian orogens, such as the Grenville Province, cannot be based on well established stratigraphic columns, fossil records, volcanic sequences or ophiolite complexes as any of these features are rare in such orogens. However, plutonic suites are common and, if their magmatic lineages and geographical extent are assumed to reflect their original tectonic setting, they can be used to establish paleo-environments and to delineate terranes and their timing of amalgamation and accretion. To this end, it is necessary to establish mineral paragenesis, geochemical evolution, magmatic lineages, timing of emplacement and geographical extent of the suites. For igneous rocks that have been metamorphosed, it is also necessary to establish if metamorphism was isochemical before geochemical signatures are used to interpret magmatic lineages. This approach was applied to syenitic plutons in the Mont Laurier area of Québec (Corriveau, 1989; GSC project 760061 to A. Davidson). It included two summers' field mapping, a petrological study of eight plutons, and U-Pb geochronology on three plutons. The U-Pb ages (Corriveau et al., 1990) not only provided the age of emplacement, but also highlighted an eastward migration of this magmatism and demonstrated that the plutons were part of a 450 km long plutonic province extending into Ontario. The integration of field, petrological and geochronological investigations led to (1) the first identification of a Proterozoic potassic alkaline plutonic belt with arc-related geochemistry, (2) further delineation of terranes in the Central Metasedimentary Belt, and (3) a new tectonic model describing the late stage of the Grenvillian orogenic cycle (Corriveau, 1990).

The tectonic model suggests that (1) the Mont Laurier area is not a single terrane, but consists of two suspect terranes which extend southward into Ontario: the Gatineau domain, a northward extension of the Elzevir terrane and the southern domain, a northward extension of the Frontenac terrane (Fig. 1), (2) between 1089 and 1076 Ma, the Elzevir terrane was situated above a southeast dipping subduction zone, and (3) that amalgamation of the Elzevir and Frontenac terranes and the allochthonous polycyclic belt took place after the emplacement of 1089 to 1076 Ma plutons, likely during the Ottawan orogeny. A post-1076 Ma accretion of the Central Metasedimentary Belt to the allochthonous polycyclic belt is consistent with the 1060 Ma age of a major episode of northwest-directed thrusting along the Central Metasedimentary Belt boundary zone (van Breemen and Hanmer, 1986) (Fig. 1). In the model, the Labelle shear zone represents a major boundary between two terranes amalgamated after emplacement of the potassic plutons which postdate the regional metamorphism. The Labelle shear zone is an up to 10km wide, north-striking transcurrent sinistral ductile shear belt at amphibolite to granulite facies which comprises discrete corridors bordered and overprinted by late mylonite zones, some with down-dip lineations (thrusting or extension?) (Martignole and Corriveau, 1991). Supracrustal rocks appear compositionaly similar on both sides of the Labelle shear zone, but there are differences in metamorphic grade and plutonic assemblages across this tectonic boundary.

The proposed tectonic model also has key implications for mineral exploration in the Central Metasedimentary Belt. Zn exploration in the Mont Laurier area is based, in part, on the hypothesis that equivalents of supracrustal rocks hosting the Balmat-Edwards Zn deposits in the Frontenac-Adirondack Lowlands terrane (Fig. 1) occur in the Mont Laurier area. However, according to the present model, the sediment-hosted Zn-mineralizations in the Adirondack and Mont Laurier areas occur in different terranes, hence would not belong to the same basin.

As a consequence of the proposed subdivision of the Mont Laurier area into suspect terranes, inter-terrane contrasts among pre-amalgamation plutonic suites and similiarities among later ones should be expected, and major shear zones should occur along proposed terrane boundaries. This model can thus be tested by regional mapping of the Central Metasedimentary Belt and by petrological studies of plutonic rocks within the belt and across the Labelle shear zone. This report presents the first phase of this testing, the second phase will involve geochemical and geochronological studies, which are needed before the validity of the tectonic model can be assessed.

PLUTONIC SUITES IN THE CENTRAL METASEDIMENTARY BELT OF QUEBEC

This belt in Québec is underlain by supracrustal rocks interlayered with felsic and mafic orthogneisses. These rocks were metamorphosed to amphibolite or granulite facies and intruded by five groups of plutons (Fig. 2,3,4): 1) gabbro and metagabbro (unit 15 in Wynne-Edwards et al., 1966; unit G10 in Avramtchev and Piché, 1981), 2) mangerite (unit 17/G12), 3) porphyritic monzonite and quartz monzonite associated with minor diorite (unit 18/G17), 4) biotite-bearing syenite and associated rocks (unit 21/G18), and 5) pink granite (unit 20/G19). Reliable and precise ages are not available for any lithologies or thermotectonic events in the Central Metasedimentary Belt in Québec, except for the 1089 to 1076 Ma U-Pb zircon and baddeleyite ages from biotite-bearing syenites

(Corriveau et al., 1990). Reported occurrences of units 15, 17, 18 and 20 (Wynne-Edwards et al., 1966) were examined in the Mont Laurier area and in the eastern part of the Gatineau area.

Gabbroic rocks

Several of the gabbroic masses previously reported are very poorly exposed, and gabbro, metagabbro, amphibolite, syenite and surrounding quartzite country rock were all included in the masses (unit 15). Foliated and/or lineated amphibolite with relict ophitic texture occurs in the masses elongated parallel to the regional structural trend. Gabbros were found in two plutons and were associated with syenites.

The Montagnes Noires pluton occurs in the southern domain of the Mont Laurier area (Fig. 2,3I,J). It is a circular body of 30 km^2 with noritic gabbro in the south and syenite and quartz syenite in the north. The gabbro is coarse grained and consists of plagioclase laths with interstitial hornblende, magnetite and minor orthopyroxene (Fig. 3J). Plagioclase locally defines an igneous foliation. The syenite is massive and consists of coarse blocky K-feldspar, some of which are zoned with a centimetre-size core of grey K-feldspar surrounded by a fine inclusion-rich rim and an outer rim of pink K-feldspar (Fig. 3I). Accessory minerals include hornblende, magnetite and quartz with minor apatite, baddeleyite, biotite, titanite and zircon. The contact zone between the pluton and the country rock is heterogeneous with fine grained mafic enclaves; deformation and recrystallization is restricted to small shear zones, otherwise the syenite is undeformed at the contact. This syenite and gabbro pluton thus postdates regional metamorphism and deformation. Based on mineral assemblages (hypersthene

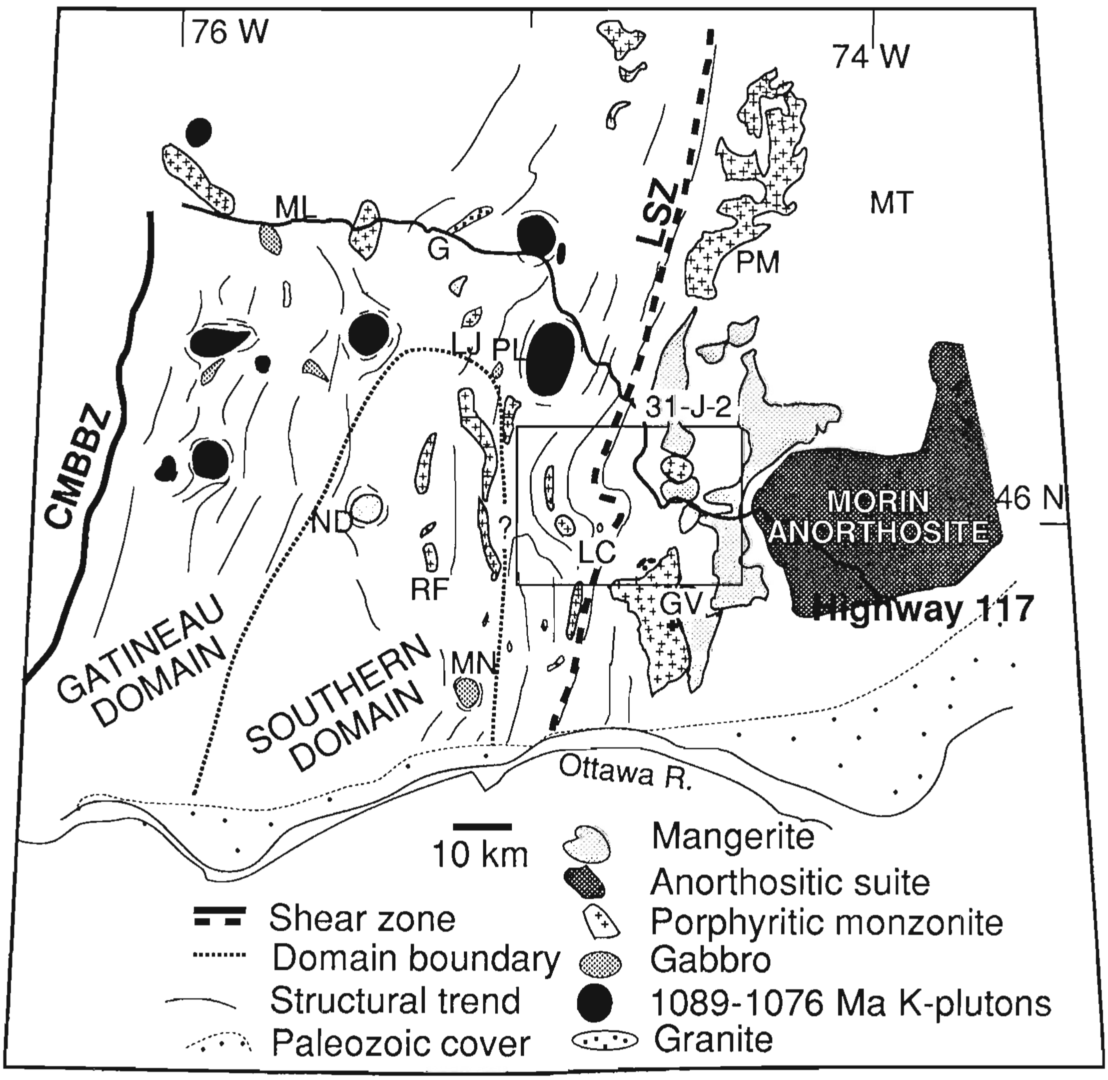


Figure 2. Details of the northern part of the Central Metasedimentary Belt showing plutonic suites and subdivisions of the Mont Laurier area. Abbreviations: G — Guenette; GV — Grey Valley; LC - Lake Chevreuil; LJ — Lake Joinville; ML — Mont Laurier; MT — Morin terrane; PL — Petit Lac Nominingue; PM — Parc du Mont-Tremblant; MN — Montagnes Noires; ND — Notre-Dame du Laus; RF — Réserve Faunique Papineau-Labelle. Other abbreviations accompany Figure 1.

and minor biotite in the gabbro, abundance of hornblende and minor amounts of titanite and biotite in the syenite), it is suggested that the pluton is not part of the potassic suite but represents a different type of late Grenvillian magmatism.

The gabbroic pluton west of Mont Laurier (Fig. 2) occurs in the Gatineau domain and comprises foliated grey biotite and hornblende-bearing syenites along highway 117 and massive grey noritic gabbros.

These two gabbro-syenite plutons are relatively fresh, display significant lithological diversities and occur in different terranes. They may thus be useful markers in the unravelling of the tectonic evolution of the Central Metasedimentary Belt and will be the object of future detailed mapping and petrological studies.

Mangerite

Two mangeritic bodies, one east of Lake Chevreuil, the other southeast of Mont Laurier (Fig. 2), have been reported in the Central Metasedimentary Belt of Québec whereas no anorthosites were found (Wynne-Edwards et al., 1966; Avramtchev and Piché, 1981). This contrasts markedly with the prominent mangerite, charnockite, gabbro and anorthosite suite in the adjacent Morin terrane (Fig. 2). Investigation of mangeritic bodies in the Central Metasedimentary Belt is aimed at establishing whether they belong to the same suite as those of the Morin terrane. A few outcrops of a coarse grained biotite- and garnet-bearing monzonite with minor hornblende and quartz and orthopyroxene-bearing monzodiorite were found north of Lake Chevreuil. This body ($< 0.5 \times 1$ km) occurs within granulite facies country rocks. Its mineral parageneses and textures are unlike those of porphyritic mangerite and monzonite in the Morin terrane. The body south of Guenette (Fig. 2) has not been investigated yet. Anorthosite and noritic gabbros have been observed as lenses a few kilometres long only within the eastern part of the Labelle shear zone (Fig. 2) in the 31J/2 map sheet (Martignole and Corriveau, 1991) and probably represent part of the Morin terrane. A pluton of mangerite and monzonite was discovered in charnockitic gneisses near Notre-Dame du Laus (Fig. 2). The brown weathering and greenish colour on fresh surface found in a few outcrops suggest that the pluton is partly mangeritic, however, no orthopyroxene was observed in the field. A coarse hornblende- and biotite-bearing monzonite is predominant and is associated with granitic rocks in the Notre-Dame du Laus pluton. The texture ranges from hypidiomorphic with subhedral K-feldspar and interstitial hornblende and biotite to an augen gneiss. The presence of coarse grained mangeritic rocks suggest that this pluton is slightly different from the porphyritic monzonite suite in the Central Metasedimentary Belt described below.

Porphyritic monzonite and quartz monzonite suite

Porphyritic monzonite and quartz monzonite rocks are common throughout the Central Metasedimentary Belt (Fig. 2). These rock types form spectacular and homogeneous outcrops with widely spaced joints in plutons at Grey Valley and Parc du Mont Tremblant (potential sites for building stones: 523900E, 5086155N; 522855E, 5087950N; 522050E, 5089150N; 537050E, 5151175N). Only a few sites south of Mont Laurier are presently quarried for local use. In the Central Metasedimentary Belt, the monzonitic plutons are most abundant in the vicinity of the Central Metasedimentary Belt boundaries. They occur mainly as large sheet-like complexes, subparallel to the north-striking regional structural grain. The plutons predominantly consist of porphyritic monzonite with coarse K-feldspar laths in a fine- to medium-grained matrix of plagioclase, K-feldspar, biotite and hornblende with minor magnetite and titanite (Fig. 3G). Biotite and hornblende-bearing quartz monzonite and granite, granoblastic amphibolite, metadiorite (Fig. 3D), noritic metagabbro with relict ophitic texture and net-veined dykes (Fig. 3E) are ubiquitous in these plutons. Fine grained mafic dykes locally crosscut foliated monzonites. A few plugs of alkaline olivine- and biotite-bearing melagabbro (CI:85; $<$ 10m in size) are associated with the gabbroic rocks. They appear less deformed, hence younger, than their host metagabbro.

The monzonitic plutons typically contain a weakly to moderately developed foliation defined by alignment of euhedral to subhedral K-feldspar megacrysts, ferromagnesian minerals and elongate enclaves. The proportion of phenocrysts and the grain size of the matrix are highly variable within the monzonite units, but the mineral paragenesis of K-feldspar, plagioclase, biotite, hornblende and magnetite is constant throughout. Diorites have mineral assemblages and grain size similar to the matrix of the porphyritic monzonite; plagioclase appears recrystallized, however it is not possible to establish from field observations whether the paragenesis is mainly igneous or metamorphic. Metagabbros have preserved their ophitic texture, but plagioclase laths are white to grey and internally recrystallized to fine- to medium-grained granoblastic aggregates. Hornblende, biotite and orthopyroxene with accessory magnetite, apatite and rare titanite and pyrite occur between laths. Most of the igneous orthopyroxene is replaced by amphibole. In situ leucosomes are generally absent in the monzonitic units but may occur in the gabbroic ones. Metamorphic mineral assemblages in the pluton and along its contact zones indicate that recrystallization took place at amphibolite facies. This contrasts with the occurrence of incipient melting features and granulite facies mineral assemblages in the country rocks and indicates that the plutons were emplaced after peak metamorphism.

Dioritic enclaves are common in the monzonitic and dioritic phases of these plutons (Fig. 3D). They are lenticular, up to 1m long, fine-to medium-grained and equigranular, however, rare K-feldspar phenocrysts occur. Aspect ratios between 3:1 and 5:1 are common. The enclaves are not equivalent to the gneissic country-rocks and are best explained as cogenetic inclusions. The net-veined dykes (Fig. 3E) consist of microgabbroic or dioritic enclaves in a hybrid, fine- to medium-grained felsic matrix. The composition of the net-veined dykes varies and may represent the range of magma-types associated

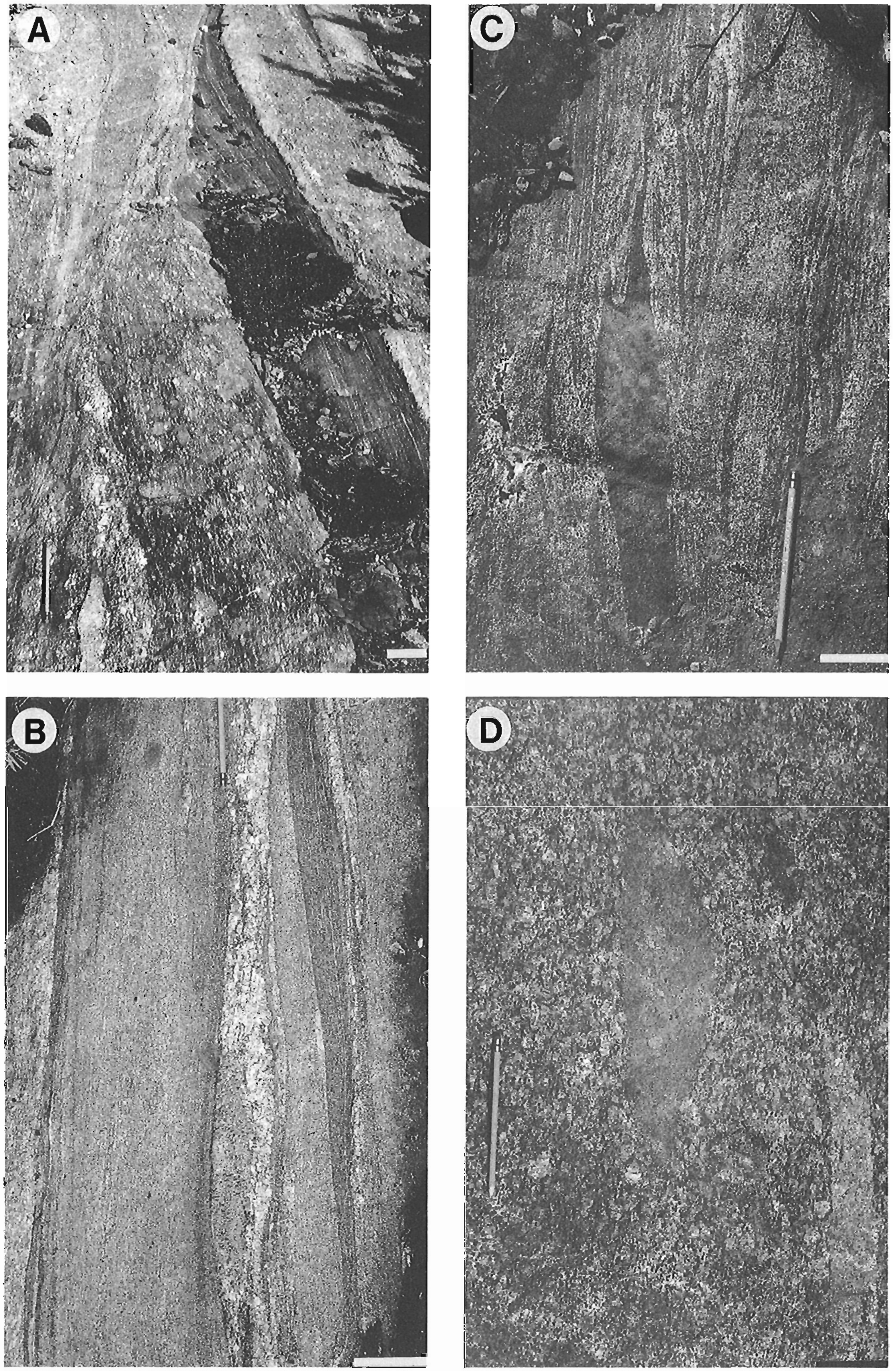


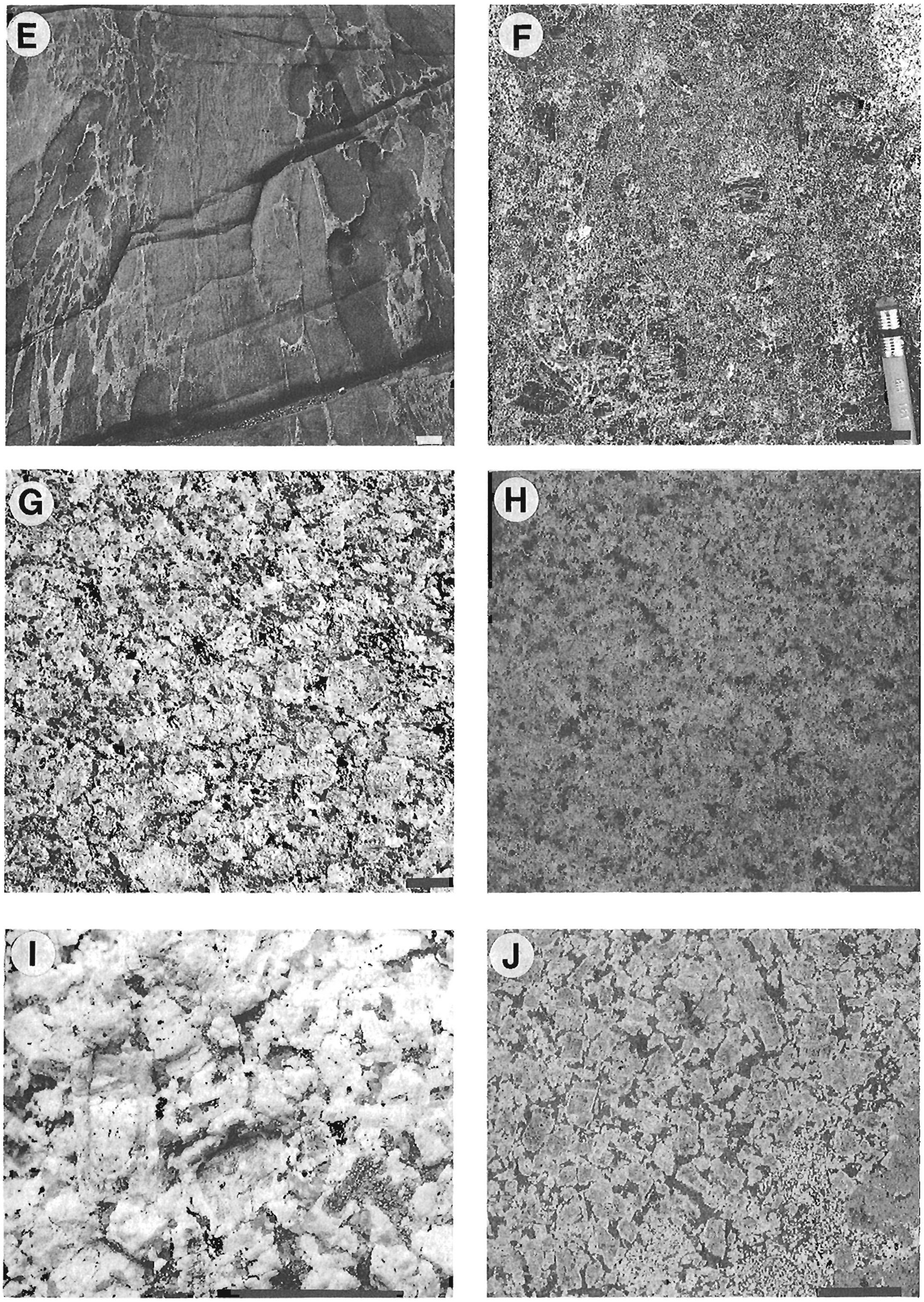
Figure 3. A to G: Solid-state deformation along contact zone of a porphyritic monzonite and diorite pluton with net-veined dykes (UTM: 484500E, 5099185N; scale bars for A-E are 10 cm).

A) Mylonitized contact of monzonitic pluton within ten metres of the quartzite country rock.

B) Mylonitized net-veined dykes forming a compositionally layered orthogneiss.

C) Deformed net-veined dykes with mafic enclaves with aspect ratios greater than 7:1.

D) Diorite with weak foliation defined by the alignment of feldspar laths, biotite, hornblende and microdioritic enclaves with an aspect ratio of 4:1.



E) Net-veined dykes with gabbroic enclaves having an aspect ratio around 4:1 and showing cuspate boundaries. The igneous foliation is defined by the elongation of the enclaves and by laths of plagioclase that were apparently unaffected by recrystallization (Fig. 3F).

F) The preferred alignment of pristine laths of plagioclase defines an igneous foliation (scale bar: 2 cm).

G) Monzonite with K-feldspar phenocrysts in a matrix of plagioclase, K-feldspar, biotite and hornblende (scale bar: 2cm).

H) Guenette granite with biotite and muscovite (UTM: 481200E, 5154700N; scale bar: 2cm).

I) Montagnes Noires syenite with zoned K-feldspar (UTM: 487350E, 5071575N; scale bar: 1cm).

J) Montagnes Noires gabbro (UTM: 486770E, 5066700N; scale bar: 1cm).

with the monzonitic magmatism. The enclaves and matrix of the dykes have a foliation defined by preferred shape orientation of plagioclase, hornblende and biotite. Dark grey laths of plagioclase, apparently unaffected by recrystallization (Fig. 3F), occur locally within these enclaves. This foliation is thus interpreted to be igneous. It is parallel to the elongation of the mafic enclaves in the net-veined dykes and the contacts of these dykes. In turn, these features are parallel to the elongation of the enclaves and K-feldspar phenocrysts in the host plutonic units, the plutonic contacts and the metamorphic foliation in the country rocks. The enclaves and the abundance of net-veined dykes suggest formation by mingling of mafic and felsic magmas at the site of emplacement. The alignment of enclaves and phenocrysts is interpreted to be related to magmatic flow whereas the weak biotite and hornblende foliation is likely tectonic in origin. Solid-state deformation is most apparent along a highly mylonitized, 100m wide zone at the contact between a monzonitic pluton and its quartzite country rocks; the monzonite becomes an augen gneiss (Fig. 3A) and the inclusions in the net-veined dykes and dioritic plutons (Fig. 3D) are transposed into compositionally layered gneisses (Figs. 3B,C). Tectonic foliation in the plutons is thus parallel to regional structures and to the preserved igneous foliation. In summary, the observations of pluton emplacement as sheets and the presence of abundant zones of magma mingling suggest that emplacement was synkinematic with respect to the last episode of regional deformation.

In the Morin terrane (Fig. 2), the monzonitic plutons are large, irregular and associated with a charnockitic suite (Martignole and Corriveau, 1991). The monzonite and mangerite are coarse grained with porphyritic K-feldspar in a matrix of medium- to coarse-grained plagioclase, biotite, hornblende and magnetite. The sporadic occurrence of garnet and orthopyroxene is characteristic of monzonites in this terrane. Rapakivi texture was observed in the pluton near Grey Valley (UTM: 522550E, 5088100N).

Pink granites

The Guenette granite (Fig. 2,3H,4A) is an elongate body of fine- to medium-grained, leucocratic, hypidiomorphic granular and massive granite. Biotite and muscovite are the main mafic minerals. Magnetite, titanite, garnet and fluorite occur in trace amounts. Along highway 117 (Fig. 2), the granite is pink, very homogeneous, with widely spaced joints and has been quarried for several decades. In the northern part of this pluton exposure is poor, however, a few outcrops occur along the power line (UTM: 485900E, 5156100N). The granite is white on weathered surface, pale grey on fresh surface and contains very few joints. It is massive, very homogeneous and is not magnetite-bearing. The overburden in this area appears to be only a few metres thick and the granite represents a potential source of pale grey granite for building stones.

Dykes of pink granite similar in texture and mineral assemblage crosscut the porphyritic monzonite plutons and the country rocks. The massive character of the Guenette granite and the occurrence of the pink granitic dykes indicate that this magmatism postdates the episode of monzonitic plutonism. The elongation of the granitic bodies parallel to the regional structural trend suggest that they were emplaced during the waning stages of regional deformation and metamorphism. The presence of biotite and muscovite suggests that the pluton is peraluminous and derived by fusion of the crust during granulite facies metamorphism. The presence of angular enclaves of country rocks in this pluton and cogenetic dykes crosscutting the country rocks throughout the area around Mont Laurier suggests that this magmatism is not the product of local anatexis.

Fine grained biotite-bearing mafic lamprophyre dykes with magnetite and biotite phenocrysts crosscut the Guenette granite (Fig. 4A,B). Dyke contacts are straight or show cuspate boundaries, especially in the vicinity of dyke terminations, which are very irregular. This suggests that the dykes were emplaced prior to complete solidification of the granite and may be part of the same magmatic event. Geochemical and mineralogical characteristics of lamprophyre rocks are generally unique indicators of their tectonic setting (e.g., Rock, 1987), and the association of lamprophyre with this granitic magmatism may thus prove important for the tectonic studies.

Only one outcrop of granite was found within the reported granite pluton west of Mont Laurier (Wynne-Edwards et al., 1966). Outcrop exposure is poor and granitic rocks may only occur as masses, few metres in size, within the country rocks.

TECTONIC ZONES

Reconnaissance mapping in the Réserve Faunique de Papineau Labelle (Fig. 2) has established a series of 10-50 cm wide subvertical ductile mylonite zones at amphibolite facies along major lineaments. Sinistral north- or west-striking shear zone with subhorizontal stretching lineations appear to be most common, however, down-dip lineations have been observed in a few zones (e.g., southwest-striking, steeply-dipping mylonite zones southeast of Lake Joinville; Fig. 2).

A major (> 10 m), possibly inter-terrane, northwest-striking ductile shear zone with moderate southwesterly dip and down-dip stretching lineation has been recognized south of the Petit Lac Nominingue (Fig. 2). The lineation is defined by alignment of quartz, hornblende or sillimanite. This is a fine grained porphyroclastic mylonite zone, with coarse fragments of undeformed pegmatite, porphyroclasts of K-feldspar probably derived from the pegmatite and mylonitized blocks of gabbro some of which have relict ophitic texture in undeformed zones. In some exposures, kinematic indicators (C, C' and S fabrics) show northeast-directed thrusting. Sillimanite is present along shear plane in a quartzofeldspathic gneiss and may represent leached zones formed during mylonitization. The regional extent of this shear zone is presently unknown, but it is located along the northeastern part of the proposed terrane boundary between the southern and Gatineau domains. Shear zones with similar characteristics were not found elsewhere during the study.

ECONOMIC GEOLOGY

A Zn-mineralization (sphalerite) was found in a rusty biotite-bearing quartzofeldspathic gneiss in the Réserve Faunique de Papineau Labelle (Fig. 2; UTM: 482050E, 5122400N). Disseminated fine grained sphalerite with a metallic grey to honey brown colour constitutes up to 2% of the mineralized zone. This mineralization has not been assayed or characterized by geochemistry and petrography. As discussed above, graphite mineralizations and wollastonite occurrences occur in the 31J/2 sheet (Martignole and Corriveau, 1991) and potential sites for building stones are common in the porphyritic monzonite plutons and the Guenette granite. Graphite showings near Lake Carmin (SW of 31J/2) occur in a sequence of quartzofeldspathic gneisses, quartzites and metapelites occupying the eastern flank of a large fold; the western flank extend into the southeastern part of the 31J/3 map sheet and is a potential site of additional graphite mineralization. A spectacular outcrop of layered amphibolite (metavolcanic?) was also observed by the author in the southeastern part of 31J/3 in 1981. The southeastern part of 31J/3 and southwestern part of 31J/2 are thus of sufficient potential economic interest to warrant detailed mapping at 1:20 000 and metallogenic studies by provincial or federal surveys in the future.

DISCUSSION

Reconnaissance mapping indicates that the gabbroic rocks are far less widespread than inferred from the existing 1:250 000 map. This will limit their used as lithotectonic indicators. Biotite and hornblende-bearing grey syenite with K-feldspar laths have locally been mapped as gabbro. These syenites are deformed and do not appear similar to the potassic syenites, but represent a different period of syenite magmatism. The presence of a porphyritic mangerite and monzonite pluton at Notre-Dame du Laus in what was previously considered charnockitic gneisses and intrusives too deformed to be included in the charnockite intrusive series, is another example of the urgent need for modern mapping in the Central Metasedimentary Belt of Québec. Porphyritic monzonite

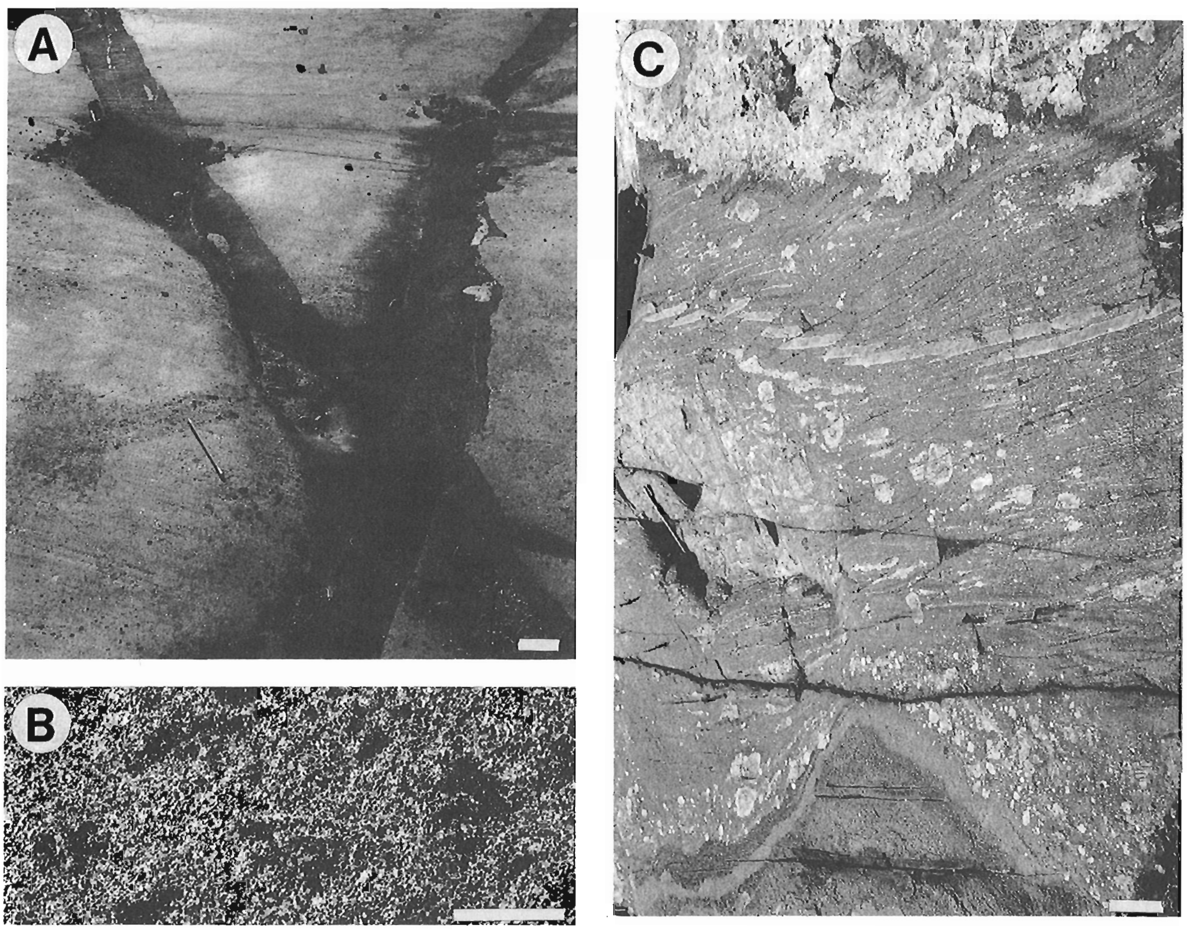


Figure 4. A) Guenette granite crosscut by two lamprophyre dykes (UTM: 482700E, 5155800N; scale bar: 10cm).

B) Close-up of the lamprophyre dykes with biotite and magnetite phenocrysts (scale bar: 1cm).

C) Porphyroclastic mylonite with fragments of pegmatite unaffected by deformation. Porphyroclasts of K-feldspar are probably derived from pegmatite and the compositional layering presumably formed by transposition and mylonitization of an unknown protolith (UTM: 495050E, 5130630N; scale bar: 10cm).

bodies are as abundant and homogeneous as reported, but the position of pluton boundaries needs revision. The abundance of features suggesting magma mingling is charcteristic of these plutons and needs further petrological investigation.

Five plutonic associations were recognized in the Central Metasedimentary Belt and the adjacent Morin terrane: porphyritic monzonite and metagabbro, charnockite and anorthosite, noritic gabbro and syenite, potassic alkaline plutons, and granite. The charnockite and anorthosite suite is restricted to the Morin terrane and disrupted members have been incorporated along the eastern margin of the Labelle shear zone (Fig. 2). Porphyritic monzonites occur in both the Central Metasedimentary Belt and the Morin terrane (Fig. 2). Based on field observations alone, it is not possible to establish whether these monzonitic rocks belong to the same suite or if there are significant differences between them across the Labelle shear zone. Their plutonic association appears slightly different, with mangerites and charnockites in the Morin terrane, and net-veined dykes and metagabbro or diorite in the Central Metasedimentary Belt. The noritic gabbro and syenite rocks occur in the Central Metasedimentary Belt in two domains that have been proposed to be distinct (Fig. 2). The major difference in the field is that gabbro and syenite west of Mont Laurier are more deformed and recrystallized than the Montagnes Noires pluton. Furthermore, the syenite associated with this gabbro is very similar to syenites associated with a large porphyritic monzonite pluton located northwest of Mont Laurier. None of the plutons investigated in the southern domain and Morin terrane have characteristics of the potassic alkaline suite. The proposed extent of this suite (Corriveau, 1990) is thus corroborated by the present investigation. The pink granites appear restricted to the Gatineau domain with a geographical extent similar to that of the potassic alkaline suite. These two suites never occur together in a single magmatic centre and are not considered part of the same event, however, the significance of their similar geographical extent warrants further investigation. During the course of this field work, a major, possibly inter-terrane, shear zone has been recognized, but its regional extent is presently unknown. Zn and graphite mineralizations, wollastonite occurrences and potential sites for building stones have also been located.

Integration of field mapping with future petrological, structural and geochronological studies of plutonic suites in the Central Metasedimentary Belt and of para- and orthogneissic country rocks is needed to further our understanding of the tectonic evolution of this key area in the Grenville Province.

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Final field report on the Contwoyto-Nose Lakes map area, central Slave Province, District of Mackenzie, N.W.T.[1]

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King, J.E., Van Nostrand, T., Bethune, K, Wingate, M.T., and Relf, C., Final field report on the Contwoyto-Nose Lakes map area, central Slave Province, District of Mackenzie, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 99-108, 1991.

Abstract

The northeast quadrant of the Contwoyto-Nose Lakes map area is underlain predominantly by granitoids with minor screens and inclusions of iron-formation-bearing metaturbidites and metavolcanic rocks. Re-evaluation of the contact between the iron-formation-bearing Contwoyto Formation and the iron-formation-barren Itchen Formation east of Contwoyto Lake indicates that the Itchen Formation does not extend east of the Wishbone monzogranite.

Résumé

Le quadrant nord-est de la carte illustrant les lacs Contwoyto et Nose est caractérisé par un sous-sol composé surtout de granitoïdes avec des écrans et des inclusions moins importantes de métaturbidites et de roches métavolcaniques contenant des formations ferrifères. La réévaluation du contact entre la formation ferrifère de Contwoyto et la formation non ferrifère d'Itchen à l'est du lac Contwoyto révèle que la formation d'Itchen ne s'étend pas à l'est du monzogranite de Wishbone.

[1] Includes work carried out under Canada-N.W.T. Mineral Development Agreement 1987-1991. Project carried by the Geological Survey of Canada.
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INTRODUCTION

This report presents the results of the final field season of a four-year mapping project in the central Slave Province during which the Contwoyto Lake (76E) and the west half of the Nose Lake (76F) sheets were mapped at a scale of 1:100 000 (Figs. 1, 2). Most of the 1990 mapping was focussed in the northeast quadrant of the map area (compare Fig. 2 with fig. 2 of King et al., 1990), although localities from across the area were re-examined.

The map area is underlain by metaturbidites (Contwoyto and Itchen Formations, Bostock, 1980), metavolcanic rocks, and a variety of plutons that were emplaced between about 2660 to 2580 Ma. The area has experienced at least three phases of Archean deformation and preserves high-temperature — low pressure metamorphic assemblages. Previous work, general geology and previously mapped units in the area are described by King et al. (1988, 1989, 1990) and Relf (1989, 1990). Geochronological data for the area is summarized by Mortensen et al. (1988) and van Breemen et al. (1989).

New results from the 1990 field season include: 1) the re-evaluation of the contact between the Contwoyto and Itchen formations west of Contwoyto Lake, 2) identification of a belt of metabasic (possibly metavolcaniclastic) rocks north of Nose Lake, 3) recognition of significant volumes of biotite-hornblende monzogranite of the syn-main-deformation calc-alkaline plutonic suite; 4) recognition of a distinctive suite of porphyritic biotite-hornblende monzogranites intermediate in age between the syn-main-deformation, calc-alkaline plutonic suite and the post-main-deformation peraluminous plutonic suite; and 5) subdivision of plutonic units and map-scale xenoliths in a third large monzogranite complex north of Nose Lake.

CONTWOYTO-ITCHEN FORMATION CONTACT

The two metaturbiditic formations in the map area, the Contwoyto and Itchen Formations, are distinguished by the restriction of iron-formation to the Contwoyto Formation (Bostock, 1980) although previous work has also recognized that bedding within the Itchen Formation is generally thicker than in the Contwoyto Formation and that Ca-rich concretions appeared to be unique to the Itchen Formation (Bostock, 1980; King et al., 1988). All of these characteristics were utilized to define the Contwoyto-Itchen Formation contact in the map area (e.g. King et al., 1990). King et al. (1988, 1989, 1990) and Relf (1989, 1990) interpreted the contact between the Contwoyto and Itchen formations west of Contwoyto Lake to trend northwest from the west shore of Contwoyto Lake to the north edge of the Wishbone monzogranite (see fig. 2 in King et al., 1990). This contact could not, however, be extended to the east side of Contwoyto Lake because rare iron-formations there occur along strike from rocks interpreted as Itchen Formation on the west side of the lake, and because Ca-rich concretions occur in metaturbidites along strike from rocks thought to be Contwoyto Formation. This problem led King et al. (1990) to suggest that either the Contwoyto and Itchen formations were complexly interfolded within the migmatite terrane or that the distinction between the two formations breaks down east of Contwoyto Lake.

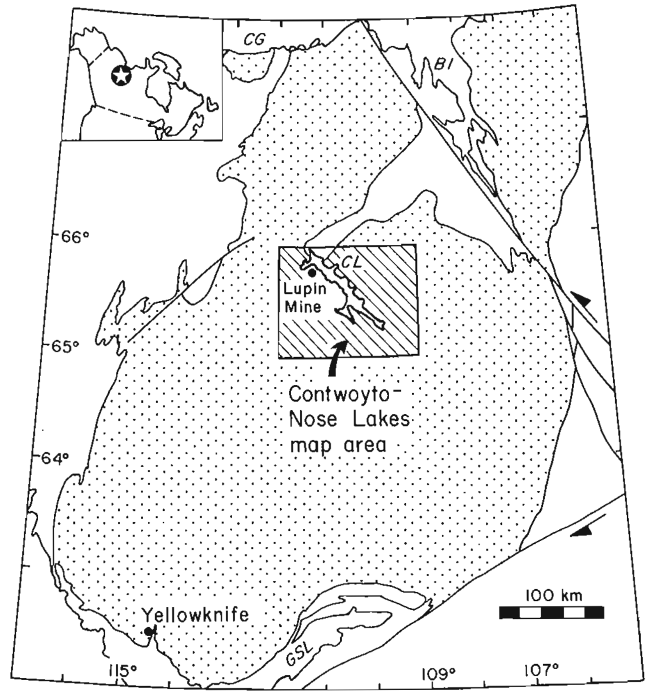


Figure 1. Location of the Contwoyto-Nose Lakes project area. Stippled area is the Slave Structural Province. Inset shows the position of the Slave Province in Canada. BI — Bathurst Inlet; CG — Coronation Gulf; CL — Contwoyto Lake; GSL — Great Slave Lake

Two sets of observations this summer provide a new perspective on this problem. Ca-rich concretions were observed in known Contwoyto Formation at, and northwest of Nose Lake and northeast of Contwoyto Lake (Fig. 2). The concretions are therefore not unique to the Itchen Formation and cannot be used to distinguish the two formations. Also, detailed mapping located several occurrences of iron-formation northeast of the Siege tonalite (location A, Fig. 2). On the basis of the definition above, metaturbidites in this area are therefore correlated with the Contwoyto Formation rather than with the Itchen Formation. These observations suggest that the Itchen Formation does not extend east of the Wishbone monzogranite and that the metaturbidites east of Contwoyto Lake could be Contwoyto Formation. However, the possibility that the distinction between the two formations breaks down east of the lake, or that they are complexly interfolded, remains valid.

IRON-FORMATION IN THE MIGMATIZED METATURBIDITES

Numerous occurrences of silicate-facies iron-formation and related rocks were observed within migmatized metaturbidites northwest of Nose Lake (locations denoted with circled crosses in Fig. 2). The composition of these units ranges from grunerite-garnet-quartz-hornblende iron-formation to biotite-garnet-quartz schists. Pyrite, pyrrhotite and magnetite are common constituents. Layering is typically tightly to isoclinally folded and boudinaged. The iron-formation is greatly disrupted by the boudinage and by granitic intrusion and typically occurs as fragments 1-3 m in size although they locally are up to 30 m across (strike length is not known due to discontinuous exposure, but is probably less than 100 m).

CANYON METABASITE BELT

A 1 km-wide belt of dominantly metabasic rocks, here referred to informally as the Canyon metabasite belt, was recognized in the northeastern corner of the map area (locality B, Fig. 2). The main rock type in the belt is hornblende ± biotite ± plagioclase ± quartz gneiss that varies from basic to intermediate in bulk composition. Thin (10-30 cm) layers of sillimanite-bearing pelitic schists also occur throughout the belt. The gneisses are characterized by well defined composition layering and foliation that strike north and dip near vertically. Four elements constitute the compositional layering: (1) a principal banding composed of diffuse to sharply to diffusely bounded, very continuous laminations 0.2-5 cm in width that are defined by variations in relative abundance and grain size of component minerals (Fig. 3a,b); (2) discontinuous elongate pods and segregations of plagioclase ± quartz interpreted to be metamorphic in origin (Fig. 3b); (3) 0.5-5 cm wide *lit-part-lit* veins of leucocratic biotite-hornblende tonalite to quartz diorite (Fig. 3c); (4) relatively late, but deformed, pink granitic veins that both cross-cut and parallel other layering (Fig. 3c). Collectively, these elements impart a gneissic aspect to the rocks. A well developed foliation defined by the alignment of ferromagnesian minerals and flattened feldspar-quartz aggregates parallels compositional layering. Lineations are not well developed, but where present are defined by the linear alignment of hornblende and plunge moderately to steeply on the foliation. The plagioclase-quartz segregations and granitic veins are commonly isoclinally folded and boudinaged. The granitic veins are locally greatly attenuated and mylonitized.

It is difficult to ascertain whether the principal layering in the belt is dominantly tectonic or primary in origin. The compositional variation and scale of the principal layering is similar to that of beds in volcaniclastic units observed at lower grade. Also, sillimanite-grade metaturbidites are intercalated with the metabasites at the scale of the layering. These two features suggest that this layering is at least in part primary and that the belt is composed predominantly of clastic rocks. However, the bedding component has clearly been tectonically transposed and metamorphically enhanced. It is not known whether the mylonitic strain recorded by the granitic veins represents the general strain in the belt (not easily recognized in the metabasic rocks), or a localized, inhomogeneous strain.

Although apparently isolated within the present map area (Fig. 2) the Canyon metabasite belt is located 3 km west of remnant metavolcanics interpreted as part of the volcanogenic Hackett River Group (see GSC Map 1619A in Frith, 1987). The rocks of the belt may therefore be part of the Hackett River Group.

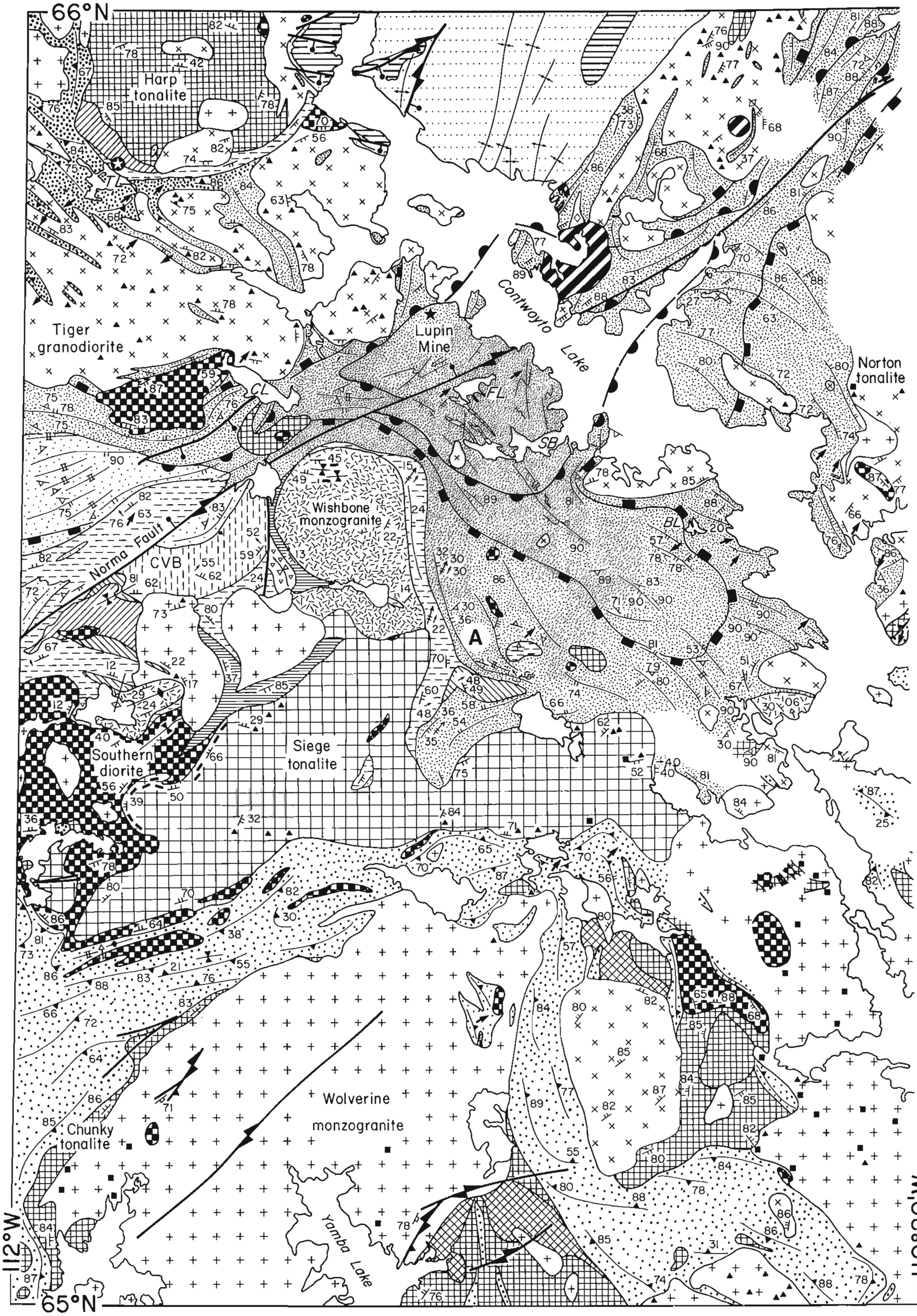
PLUTONIC UNITS

Plutonic units in the area were previously classified (King et al., 1988, 1989, 1990) by relative age of emplacement into six suites referred to as C1, C2 ... C6 ('C' denotes geographic location near Contwoyto Lake). Mapping this summer extended the subdivision of the granitic units to the northeastern corner of the map area (Fig. 2), provided new information on the compositional range and scale of intrusion of the C4 diorite-tonalite suite, identified a new group of plutons intermediate in age to the C4 and C6 suites, and outlined a third large composite C6b body. Nomenclature of rock types follows Streckeisen (1976).

C4 Hornblende ± biotite gabbro, diorite, quartz diorite, tonalite, granodiorite and monzogranite

As previously described (King et al., 1988, 1989, 1990; Davis, 1989; Davis et al., 1990), the C4 suite of intrusions includes dominantly calc-alkaline biotite ± hornblende gabbro to granodiorite, with minor hornblendite and monzodiorite, that are characterized by poorly- to well-developed foliations. This plutonic suite is interpreted to have been emplaced during the main phase of deformation (D_2) (King et al., 1988, 1989, 1990). The lithologies, map pattern and strain state of the C4 group of plutons are very similar to those of the dioritic to granodioritic intrusions included in the Regan Intrusive Suite described immediately east of the present map area (Hill and Frith, 1982; Frith, 1987; Frith and Fryer, 1985). Two C4 plutons have yielded U-Pb zircon ages of about 2608 Ma (van Breemen et al., 1989). Interpretation of geochemical data suggests that this group of rocks is juvenile and may have affinities to modern continental margin-type magmatism (Davis, 1989; Davis et al., 1990).

This summer's field work identified numerous bodies interpreted to be correlative with the C4 suite (Fig. 2). Of these, two plutons are of particular interest in that their composition ranges from biotite-hornblende diorite to biotite-hornblende monzogranite (Fig. 2). Previously monzogranite was identified as only a minor component of the C4 suite in the Contwoyto-Nose lakes area (King et al., 1990) (cf. Frith and Fryer, (1985) regarding the Regan Intrusive Suite immediately east of the present map area). The Canyon granodiorite (Fig. 2) comprises predominantly biotite ± hornblende granodiorite but also includes diorite, quartz diorite, tonalite, quartz monzodiorite, granodiorite and monzogranite. Textures vary between rock types, suggesting that they are different units formed by multiple intrusions (although lichen cover and discontinuous outcrop preclude tracing internal contacts in the field). The Bar monzogranite (Fig. 2), varies from hornblende-biotite quartz diorite in its southern half to hornblende-biotite monzogranite in the north. The change in composition is continuous, with quartz content gradually decreasing as K-feldspar content increases.


66°N
65°N
112°W
110°20'W
Harp tonalite
Tiger granodiorite
Lupin Mine
Contwoyto Lake
Norton tonalite
CL
FL
SB
BL
Wishbone monzogranite
Norma Fault
CVB
A
Siege tonalite
Southern diorite
Wolverine monzogranite
Chunky tonalite
Yamba Lake

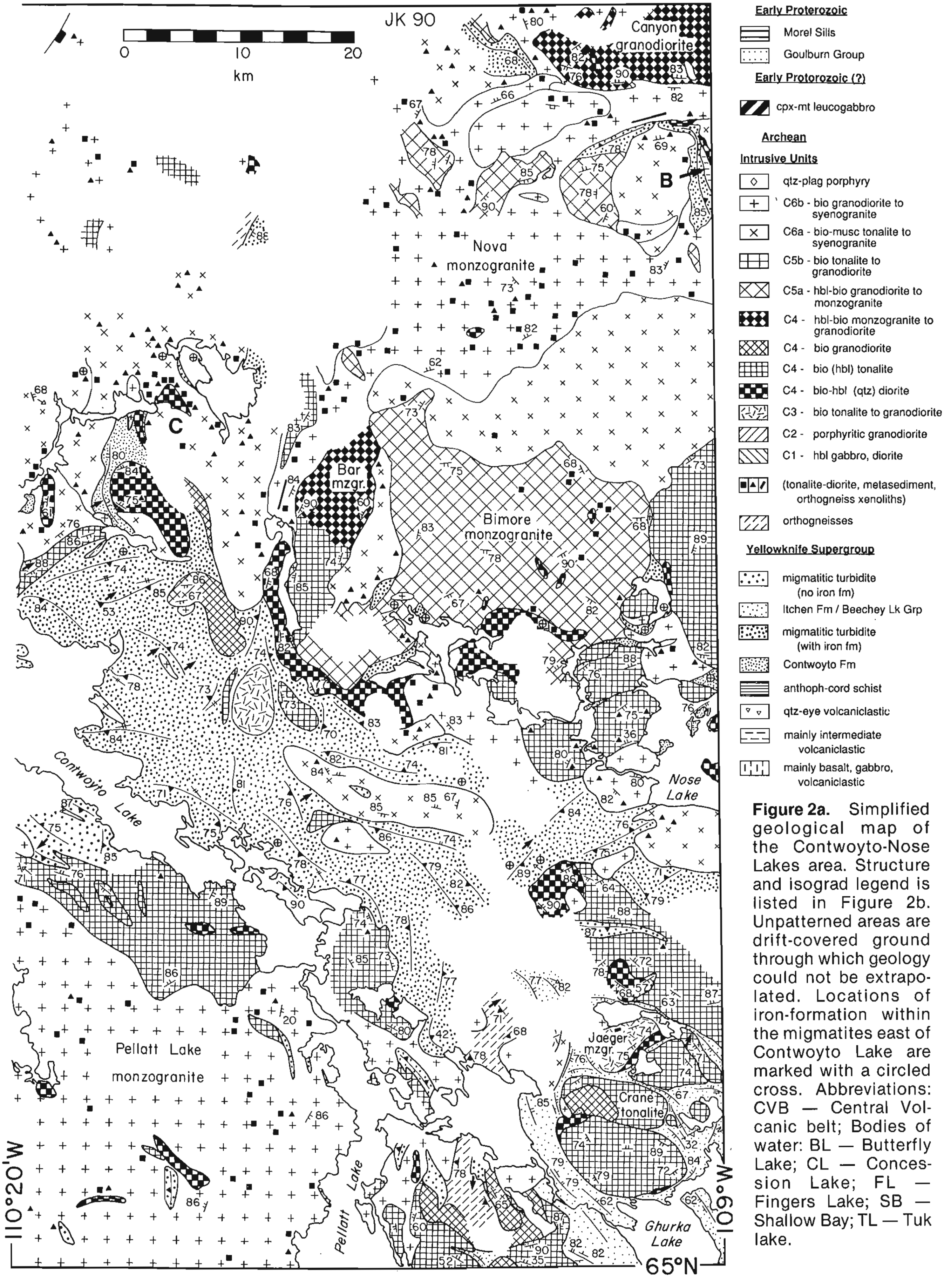


Figure 2a. Simplified geological map of the Contwoyto-Nose Lakes area. Structure and isograd legend is listed in Figure 2b. Unpatterned areas are drift-covered ground through which geology could not be extrapolated. Locations of iron-formation within the migmatites east of Contwoyto Lake are marked with a circled cross. Abbreviations: CVB — Central Volcanic belt; Bodies of water: BL — Butterfly Lake; CL — Concession Lake; FL — Fingers Lake; SB — Shallow Bay; TL — Tuk lake.

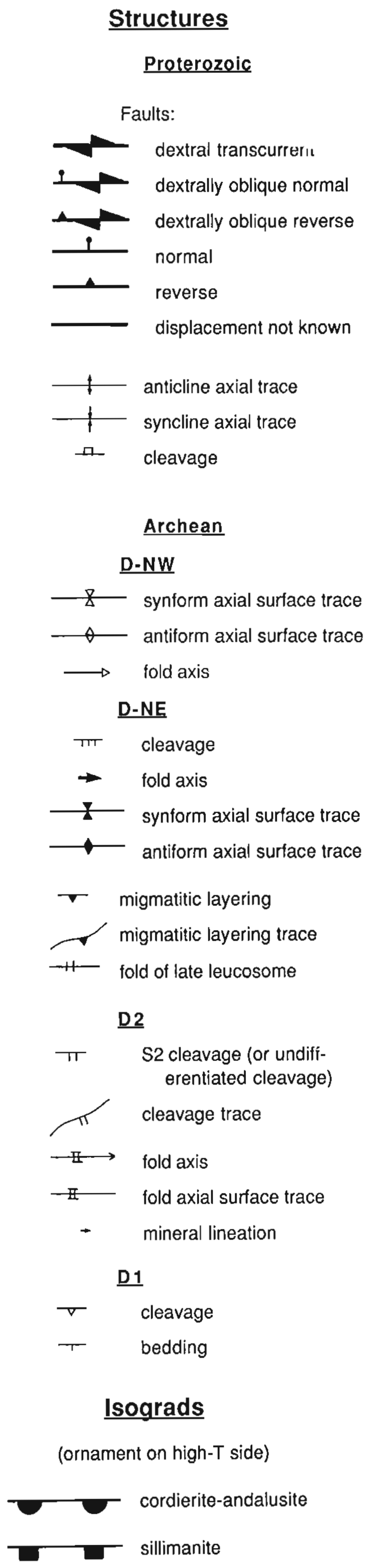


Figure 2b. List of structures and isograds for geological map of the Contwoyto-Nose Lakes area shown in Figure 2a.

Figure 3a. Principal banding in the Canyon metabasite belt defined by variations in mineral abundance and grain size variation. Scale is in centimetres. GSC 205300D.

Across the change in composition, the texture of the body is uniformly equigranular and medium grained and hornblende maintains a distinctive euhedral form. The continuous nature of the compositional change, together with the textural homogeneity suggests that the body is zoned rather than composite. Ferromagnesian minerals constitute 15-20% of the rock; biotite forms 10-20% and hornblende 0-5%. Magnetite and titanite are common accessory phases. Both plutons are typically medium grained and equigranular, with a local tendency to be K-feldspar- or plagioclase-porphyritic. A moderately well developed tectonic foliation is defined by the alignment of ferromagnesian phases and, phenocrysts K-feldspar. Where these bodies are monzogranitic they are difficult to distinguish from some C6b units (see description below). Although surrounded by C6b monzogranite at the map scale (Fig. 2) the margin of the Canyon granodiorite is typically rimmed by 2-10 m of paragneisses, suggesting that its original contact is generally preserved and that the C6b monzogranite preferentially intruded the metasediments in this area.

C4 bodies typically occur as 4-20 km-wide plutons and can be either a single unit or composite (Fig. 2) (King et al., 1988, 1989, 1990). Less commonly they form sill-like bodies, tens to hundreds of metres in length, that may also be either simple or composite. These smaller bodies generally occur in isolation or in groups of two to five spatially distinct bodies. Between the Norton tonalite and the Bar monzogranite a swarm of numerous small bodies, ranging in composition through hornblendite, biotite ± hornblende ± clinopyroxene gabbro, biotite ± hornblende diorite and hornblende ± biotite tonalite, form a 4 km-wide composite body (location C, Fig. 2). This abundance and variety of small bodies continues as a swarm of closely packed inclusions, together with intervening metaturbidites, for 2 km into the intruding C6 monzogranite. A similar swarm of compositionally mixed inclusions occurs 7 km to the west (Fig. 2).

Figure 3b. Primary layering in the Canyon metabasite belt accentuated by pods and segregations of plagioclase + quartz. Arrows point to attenuated granitic vein. Scale is in centimetres. GSC 205300S.

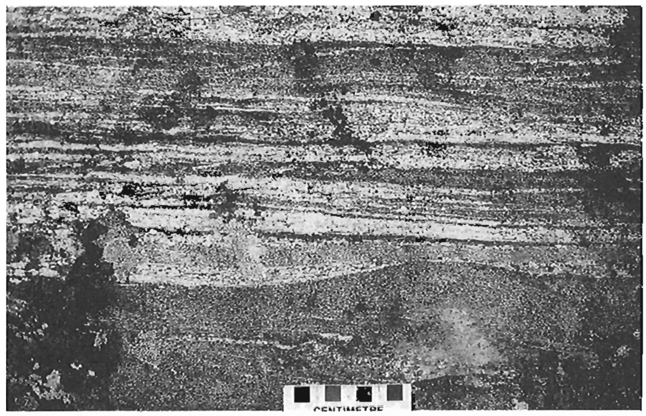


Figure 3c. Quartz diorite to tonalite veins parallel to principal layering in the Canyon metabasite belt impart a strongly gneissic aspect. Scale is in centimetres. GSC 205300M.

C5 K-feldspar-porphyritic hornblende ± biotite monzogranite

This summer a distinctive group of plutons was observed to intrude the C4 suite and to be intruded by, and occur as inclusions in, the C6 Nova monzogranite (Fig. 2). This group therefore correlates with the relative timing denoted as C5 although its age relative to the previously documented C5 Siege tonalite (King et al., 1988, 1989, 1990) is not known. On the map the units are distinguished as C5a (porphyritic hornblende ± biotite monzogranite) and C5b (Siege). The composition of the C5a plutons is typically monzogranite although they range from tonalite to syenogranite. Their most distinctive features are a pronounced K-feldspar porphyritic texture and relative abundance of ferromagnesian minerals (15-20%) (Fig. 4). The largest body is the Bimore monzogranite northwest of Nose Lake; in the northeastern corner of the map area this group of plutons occurs as map-scale xenoliths in the C6b Nova monzogranite (Fig. 2).

Two major rock types are recognized in this unit, an unrecrystallized biotite monzogranite to syenogranite and a recrystallized hornblende ± biotite granodiorite to monzogranite. The unrecrystallized biotite granite is restricted to the northern half of the Bimore monzogranite. It is typically medium grained, non-foliated and the K-feldspar phenocrysts, 7-30 mm in diameter, are unrecrystallized to partially recrystallized about their rims. Biotite forms 10-20% of the rock. Magnetite is a common accessory phase. Contacts with C4 diorite to quartz diorite inclusions are sharp. The hornblende ± biotite granodiorite to monzogranite occurs in the southern half of the Bimore monzogranite and as map-scale inclusions in younger plutons (C6b Nova monzogranite) in the northeastern part of the map area. It also occurs in small zones in the north part of the Bimore monzogranite where it is intruded by the non-recrystallized unit. The hornblende ± biotite monzogranite is medium- to coarse-grained and moderately to very well foliated. The K-feldspar phenocrysts, up to 40 mm in diameter, are partially to completely recrystallized; the degree of phenocryst recrystallization correlates directly with the intensity of foliation. Biotite forms 10-30% and hornblende 0-15% of the rock. Magnetite is a common accessory phase. Although the two rock types in this unit are texturally and mineralogically distinct, no sharp contacts were observed and contacts appeared to be gradational.

Although comparable to the Siege tonalite (C5) in relative time of emplacement, these two units are compositionally and texturally different — the Siege tonalite is leucocratic, equigranular, fine- to medium-grained and contains accessory apatite, magnetite and muscovite (King et al., 1988). In addition the Siege contains only poorly developed foliations (although this is partly a function of its leucocratic character). The intensity of foliation and presence of hornblende suggest that the porphyritic monzogranite is more closely related to the C4 suite than to the Siege tonalite, but the occurrence of map-scale, sharply bounded, inclusions of C4 bodies requires a post-C4 correlation.

C6b Biotite monzogranite to syenogranite

C6b plutons are peraluminous, medium- to coarse-grained biotite granodiorites to syenogranites that are commonly K-feldspar porphyritic and have only minor associated pegmatitic veins and pods (King et al., 1988, 1989, 1990; Davis et al., 1990). They are interpreted to have been emplaced after the main deformation (D_2) and syn- to post-cross folding (D_3) (King et al., 1988, 1989, 1990). The plutons typically form large complexes (e.g. Wolverine and Pellatt monzogranites) (Fig. 2). Ages derived from three different C6b bodies range from 2585-2582 Ma (van Breemen et al., 1989).

A third large (30 km diameter) C6b complex, the Nova monzogranite, underlies much of the northeastern part of the map area (Fig. 2). The composite Nova monzogranite comprises a variety of units ranging in composition from granodiorite to syenogranite. Their texture is typically medium grained with a tendency to be K-feldspar porphyritic (Fig. 5a), but varies to equigranu-

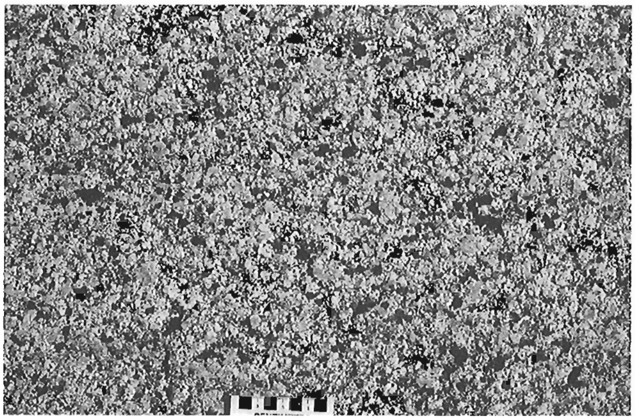


Figure 4a. Non-foliated, K-feldspar-porphyritic, biotite monzogranite. typical of the northern part of the Bimore monzogranite. This unit contains inclusions of C4 diorite and is intruded by and occurs as xenoliths in C6 monzogranites. Scale is in centimetres. GSC 205300I.

Figure 4b. Well foliated K-feldspar porphyritic, hornblende-biotite monzogranite typical of the southern part of the Bimore monzogranite. Scale is in centimetres. GSC 205300F.

lar to pegmatitic. Internal contacts could generally not be mapped because of lichen cover and discontinuous exposure, but one very distinctive body of medium- to coarse-grained, K-feldspar-megacrystic monzogranite could be distinguished (Figs. 2, 5b). The southern part of the Nova monzogranite contains abundant angular to rounded inclusions of paragneisses, C4 units, and C5a porphyritic biotite monzogranite that range in size from centimetres to tens of metres. The abundance of inclusions decreases northward.

C6b bodies are typically not foliated although they are folded by F_3 (King et al., 1988, 1990) and contain local foliations defined by the alignment of biotite and flattened quartz and feldspar (Fig. 2) (King et al., 1990). The age of these foliations is not constrained and in Figure 2 they are drawn as undifferentiated foliations. King et al. (1990) suggested that their zonal distribution and orientation in the Pellatt monzogranite indicate that they are shear zone fabrics, possibly related to a regionally developed set of brittle-ductile faults (King et al., 1989). Observations this summer indicate that at least some of these fabrics are Archean. The C6b Nova monzogranite contains narrow, discontinuous zones of moderately to well developed foliations defined by the alignment of biotite, K-feldspar phenocrysts and flattened quartz. The boundary of the foliation zones, observed in two localities, are abrupt strain gradients in which the the foliation disappears across 1 cm. The foliation zones are therefore interpreted to be heterogeneous shear zones. In one locality, the foliation is cut by 1-2 cm wide dykes of biotite-muscovite monzogranite, correlated with the C6a suite, that contain biotite foliations parallel to their margins and at a high angle to the foliation in the host. The foliation in the Nova monzogranite at this locality must therefore have formed during or after its emplacement but before the intrusion of the C6a vein. Available geochronology constrains this interval to ca 2585-2580 Ma (van Breemen et al., 1989).

C6a Muscovite — biotite tonalite to syenogranite

Plutons included within the C6a group are peraluminous, fine- to coarse-grained, muscovite-biotite tonalite to monzogranite with accessory apatite, tourmaline and garnet and abundant associated pegmatitic dykes and pods (King et al., 1988, 1989, 1990, Davis et al., 1990). C6a plutons postdate main deformation (D_1, D_2) and are syn- to post-D_3. One of the C6a plutons, near Lupin Mine, has been dated at about 2581 Ma (van Breemen et al., 1989).

Two distinct types of C6a bodies were recognized this summer. The two large C6a plutons at the east map boundary (Fig. 2) comprise homogeneous, medium-grained, equigranular monzogranite with 2-4% muscovite. They have few or no inclusions and display no foliations. In contrast, the C6a bodies between the Bar monzogranite and the Norton tonalite (Fig. 2) are texturally very inhomogeneous monzogranites that range in grain size from fine to pegmatitic. They have <1 to 4% muscovite and contain abundant inclusions of metaturbidites and granitoid units (C4, C5). A distinctive feature is the blue-grey colour of the K-feldspar on weathered surfaces. Where it intrudes migmatitic metaturbidites, the blue K-feldspar-bearing monzogranite contributes significantly to the migmatitic character. It forms textures ranging from distinct veins that crosscut the schistose to gneissic, metasedimentary melanosome (that contains earlier leucosome phases (King et al., 1990)), to a pervasive vein network, to almost complete assimilation of the metasediments. In the latter case it is essentially a schlieric monzogranite. Where occurring as distinct veins the blue-K-feldspar-bearing monzogranite is deformed together with foliation in the paragneisses about tight to isoclinal folds (Fig. 6) (cf. King et al., 1990).

The relative timing of the C6a and C6b suites of plutons was previously not well known because they are rarely in mutual contact and because their U-Pb mona-

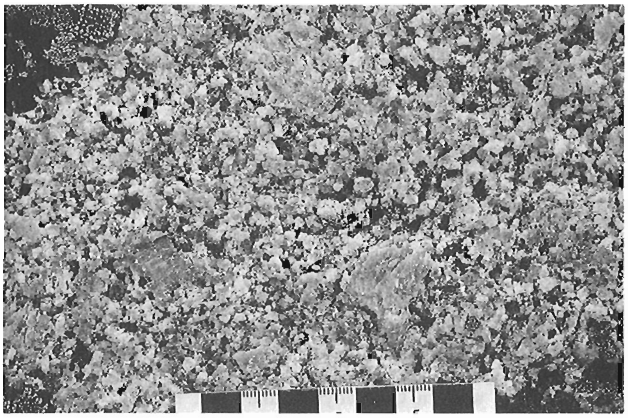


Figure 5a. K-feldspar-porphyritic biotite monzogranite typical of the Nova monzogranite. Scale is in centimetres. GSC 205300N.

Figure 5b. K-feldspar-megacrystic biotite monzogranite distinguished within the Nova monzogranite (Fig. 2). Alignment of K-feldspar megacrysts is interpreted to be of magmatic origin. Scale is in centimetres. GSC 205300B.

zite ages overlap within error (van Breemen et al., 1989). However, lengthy contacts between the two units are exposed in the northeast corner of the map area and relative time relations there can be determined. The contacts are variable in character, changing along strike from sharp to mixed zones up to 2 km wide. Along each of these contacts veins of the C6a biotite-muscovite monzogranite intrude C6b biotite monzogranites and inclusions of C6b rocks occur in C6a plutons. In this area, therefore, C6a plutons postdate C6b plutons.

The blue-K-feldspar-bearing C6a monzogranite also intrudes and contains inclusions of the Norton tonalite. The Norton tonalite is correlated with the C6a suite on the basis of its composition (King et al., 1990; Relf, 1990). However it contains the D_2 foliation, suggesting that it is older than most of the post-D_2 C6a bodies in the map area (ibid).

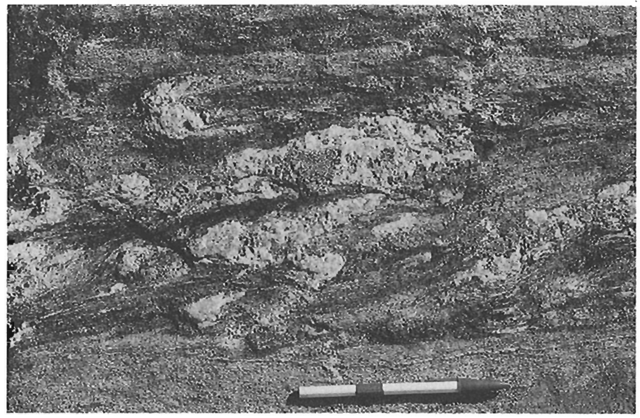


Figure 6. Veins of blue-K-feldspar-bearing monzogranite folded together with foliation in migmatitic paragneisses. Pen is 13.5 cm. GSC 205392.

SUMMARY

Two new units, the Canyon metabasites and C5 porphyritic hornblende-biotite monzogranite, were recognized in the northeast quadrant of the Contwoyto-Nose Lakes map area during the 1990 field season. In addition, although previously monzogranite was identified as only a minor component of the C4 suite in the map area, significant volumes of hornblende ± biotite monzogranite were recognized in the Canyon and Bar plutons. Other plutonic units mapped out in 1990 are correlative with previously recognized units of King et al. (1988, 1989, 1990).

Identification of iron-formation in the turbidites northeast of the Siege tonalite requires that these rocks be correlated with the Contwoyto Formation; the contact between the Itchen and Contwoyto formations is therefore thought to terminate at the Wishbone granodiorite.

A small belt of metabasic gneisses was recognized in the northwest corner of the map area. Primary layering in these rocks is interpreted to have been derived from bedding and the belt is thought to comprise predominantly metavolcaniclastic rocks.

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Janien Schwarz not only provided her usual outstanding cuisine, but also acted as geological assistant for three weeks. We would like to thank our expeditor, Rod Stone, for his capable and efficient services and Dave Ingibergsson of Latham Island Airways for getting us where we needed to go. The Polar Continental Shelf Project provided helicopter support. Critical reading by Simon Hanmer improved the text.

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Evolution of Archean and early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec

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Abstract

Tectonostratigraphic units within the internal zone of the Ungava Orogen belong to three tectonic domains: (1) (par)-autochthonous Superior Province basement; (2) Cape Smith Belt allochthons; and (3) the allochthonous Narsajuaq terrane. The Superior Province comprises voluminous tonalite and monzogranite plutons emplaced in mafic to ultramafic intrusive units and clastic metasedimentary rocks. The Narsajuaq terrane is characterized by high-grade metasedimentary rocks and a layered tonalite-quartz diorite unit intruded by tonalite and monzogranite plutons. The plutonic suites in domains (1) and (3) show a temporal calcic to potassic evolution and may represent the plutonic foundations of magmatic arcs. Cape Smith Belt rocks occur as klippen (outliers) and thrust imbricates separating domains (1) and (3). Preaccretion structures and granulite-facies assemblages are recognized in domains (1) and (3), whereas synaccretion and postaccretion deformation and amphibolite-facies histories are common to all three domains. The regional distribution of tectonic domains provides a first-order constraint for mineral exploration in northern Quebec.

Résumé

Les éléments tectono-stratigraphiques dans l'arrière-pays de l'orogène de l'Ungava se regroupent en trois domaines tectoniques: (1) le socle parautochthone de la province du lac Supérieur, (2) les nappes de charriage de la zone de Cape Smith, et (3) le terrane allochthone de Narsajuaq. La province du lac Supérieur compte d'importants plutons tonalitiques et monzogranitiques mis en place dans des unités plutoniques mafiques et ultramafiques ainsi que dans des roches métasédimentaires d'origine détritique. Le terrane de Narsajuaq est caractérisé par des roches métasédimentaires de degré métamorphique élevé et par une unité litée de tonalite et diorite à quartz recoupée par des plutons de tonalite et monzogranite. Les séries plutoniques dans les domaines (1) et (3) font état d'une évolution magmatique qui passe de calcique à potassique. Ces roches constituent peut-être les racines plutoniques d'arcs magmatiques. Les roches de la zone de Cape Smith forment des lambeaux tectoniques et des nappes de charriage qui séparent les domaines (1) et (3). Certains éléments structuraux et des cortèges de minéraux du faciès des granulites sont identifiés comme précédant l'accrétion dans les domaines (1) et (3). Par contre une déformation et une évolution métamorphique au faciès des amphibolites se reproduisent dans les trois domaines tectoniques. La répartition régionale des domaines tectoniques constitue une indication de premier ordre pour guider les travaux d'exploration minière dans le nord du Québec.

INTRODUCTION

Field work by the Geological Survey of Canada in the Ungava Peninsula of Québec continued during the second field season of a three-year multidisciplinary project initiated in 1989. Field aspects of the project included bedrock geological mapping at 1:50 000 scale (Continental Geoscience Division), mapping of surficial deposits at 1:250 000 scale (Centre géoscientifique de Québec), and detailed gravity surveys (Continental Geoscience and Geophysics divisions). This report focuses on the preliminary field results of the bedrock mapping program in parts of the following map areas: Cap de Nouvelle-France (NTS 35I), Cratère du Nouveau-Québec (NTS 35H), Lacs Nuvilik (NTS 35G), Sugluk (NTS 35J), and Wakeham Bay (NTS 25E).

The tectonostratigraphic units of the northeastern Ungava Peninsula can be grouped into three tectonic domains (Fig. 1): (1) (par)-autochthonous Archean Superior Province basement; (2) early Proterozoic Cape Smith Belt allochthons (Povungnituk, Chukotat, and Watts groups); and (3) the allochthonous early Proterozoic Narsajuaq terrane (Lucas and St-Onge, 1990; St-Onge and Lucas, 1990a,c). Units from all three domains are recognized in the internal zone of the Ungava Orogen (St-Onge and Lucas, 1990a) (Fig. 1) where significant structural relief (> 18 km; Lucas, 1989) exposes a deep crustal cross-section. This report will: (1) describe the principal tectonostratigraphic units from the current map area (Fig. 1); (2) outline the structural history of the three domains; (3) document their polychronic metamorphic histories; and (4) summarize the implications of the 1990 field work for future research and mineral exploration in the area.

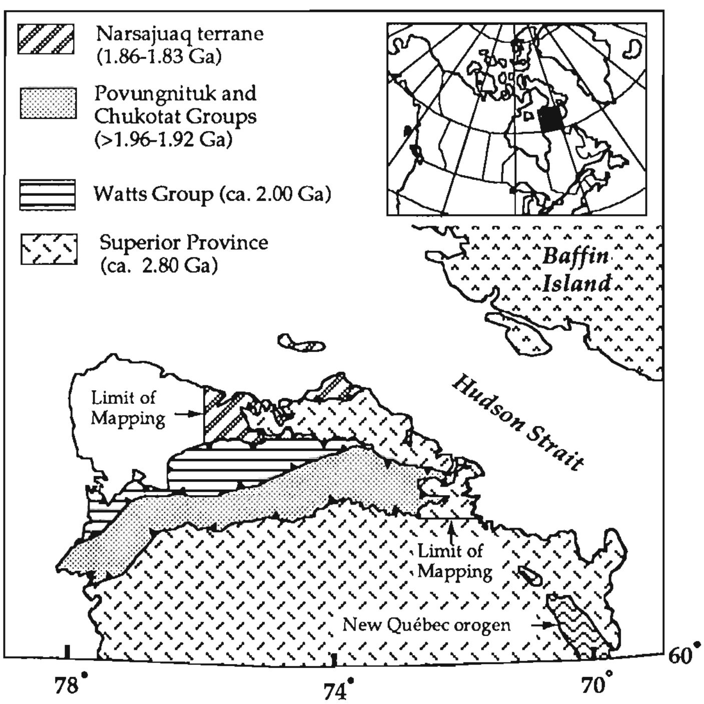

Figure 1. Map illustrating the location of the 1990 field area and the tectonic elements in the Ungava Orogen. Ages are from Parrish (1989; pers. comm., 1990).

TECTONOSTRATIGRAPHIC ASSEMBLAGES

Five principal tectonostratigraphic assemblages have been documented in the map area (Fig. 2): (1) Superior Province basement units; (2) the Spartan Group (of the Cape Smith Belt); (3) the Povungnituk Group (of the Cape Smith Belt); (4) the Akimmialuq plutonic suite (new name) of the Narsajuaq terrane; and (5) the Sugluk Group (new name) also of the Narsajuaq terrane. All units within these assemblages are deformed and metamorphosed. However, modifiers describing their deformation state (e.g. gneiss) or metamorphic character (e.g. meta-) are dropped for simplicity where appropriate. The metamorphic mineral assemblages for the various plutonic and supracrustal rocks are described in the metamorphic history section of this report.

Superior Province basement

Within the map area (Fig. 2), the Superior Province basement is characterized by distinct intrusive units which range in composition from quartz diorite to syenogranite. The principal tectonostratigraphic units are described below in order of decreasing age as inferred by field relationships.

Mafic-ultramafic units

In the Wakeham Bay area and east of Deception Bay (Fig. 2), mafic and ultramafic intrusive units are found as map-scale screens between, and rafts (Fig. 3) within, voluminous tonalite plutons. The mafic units include both quartz diorites and amphibolites. The quartz diorites are generally medium grained, equigranular, and foliated. They contain hornblende and biotite. The quartz diorites are commonly marked by an outcrop-scale compositional layering defined by the modal distribution of plagioclase and ferromagnesian minerals. The hornblende-biotite amphibolites are fine- to medium-grained, commonly homogeneous, and well foliated. Both units are crosscut by tonalite and monzogranite veins. The mafic units are most commonly observed as centimetre- to tens of metre-scale enclaves within tonalite and granite plutons. They may in part represent boudined dykes or sills, and in part correspond to fragments of older mafic plutonic bodies which were entrained in the tonalite plutons (Fig. 4). The ultramafic units consist of peridotite and pyroxenite. They occur as enclaves within tonalite and granite plutons, or within the mafic units as both enclaves and interlayered intrusive bodies.

Sedimentary rocks

Fine grained clastic sedimentary rocks are preserved as large screens within tonalite and monzogranite plutons, and are principally found east of Deception Bay (Fig. 2). Rusty semipelite is volumetrically the most important unit, while locally, less than 1 m thick bands of pelite are interlayered with the semipelite. A well developed foliation is characteristic of both the semipelite and pelite. Small pods (0.5 m in thickness) of calcite marble occur within tonalite and monzogranite bodies of the Wakeham Bay area (Fig. 2).

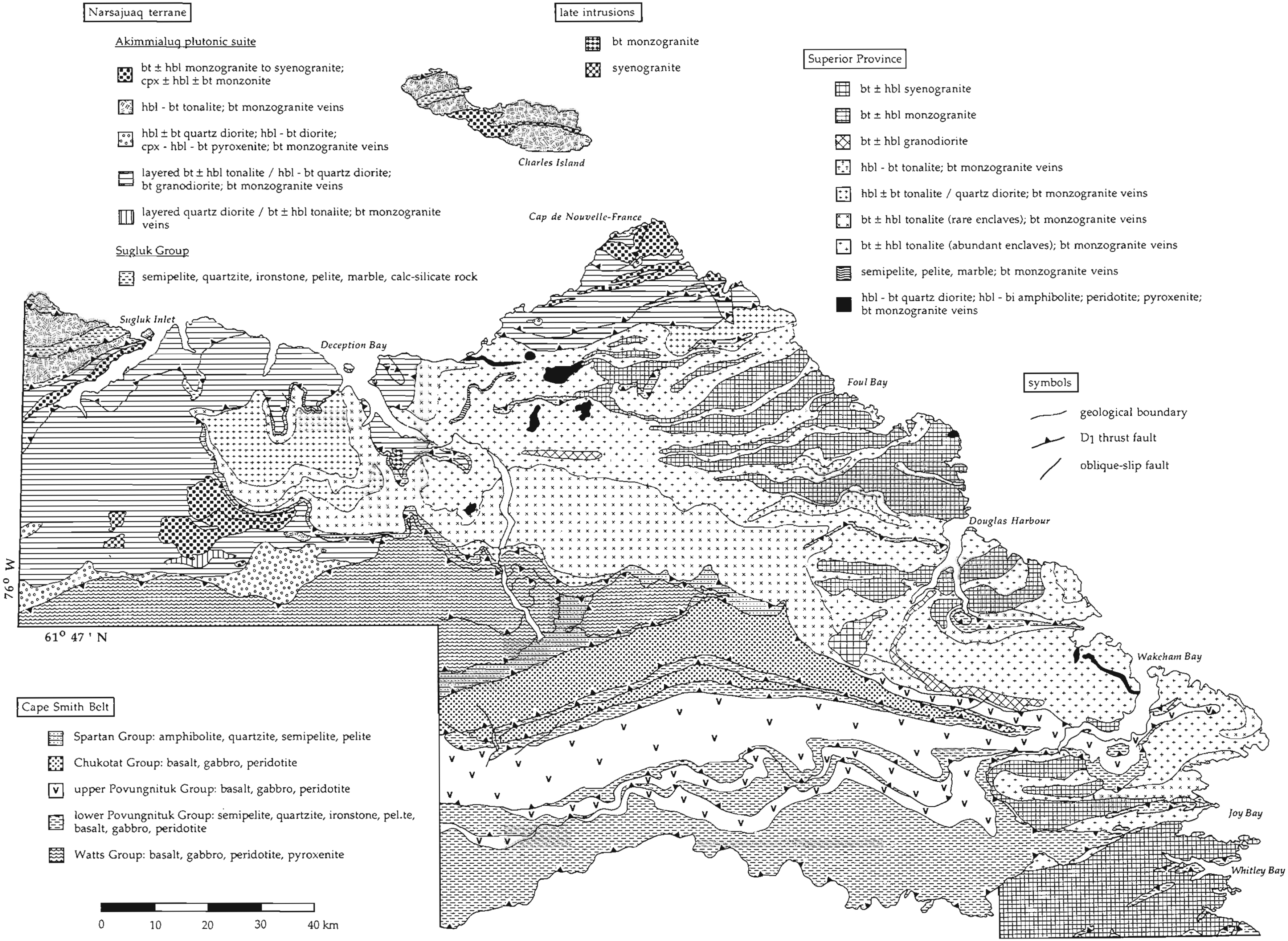

Figure 2. Geological compilation map for the Sugluk Inlet-Wakeham Bay area.

Biotite ± hornblende tonalite

Biotite ± hornblende tonalite is the dominant lithology in the mapped Superior Province basement (Fig. 2). The tonalite is grey, medium grained, and equigranular. It displays a well developed gneissic foliation and is locally plagioclase megacrystic. Two distinct generations of tonalite plutons are recognized on the basis that they largely predate and postdate granite veining. The older tonalites have been grouped into three principal map units (Fig. 2): (1) biotite ± hornblende tonalite with abundant mafic to ultramafic enclaves (Fig. 4); (2) biotite ± hornblende tonalite (rare enclaves); and (3) hornblende-biotite tonalite with interlayered quartz diorite. The veins and sheets of monzogranite and granodiorite cutting the older tonalites are deformed, and are generally subparallel to the tonalite foliation. Two samples from the tonalite with rare mafic enclaves unit (collected southeast of Deception Bay, Fig. 2) have U-Pb (zircon) igneous crystallization ages of 2882 Ma and 2780 Ma (Parrish, 1989).

Hornblende-biotite tonalite

Southwest of Foul Bay (Fig. 2) two distinct plutons of hornblende-biotite tonalite are emplaced in the older tonalites. These younger tonalites are medium grained, equigranular, and foliated. The two plutons contain abundant enclaves of amphibolite and quartz diorite. A relatively minor proportion of monzogranite veins (compared to the older tonalites) suggests that the younger tonalites may have been intruded concurrently with the monzogranite plutons.

Biotite ± hornblende granite

Biotite ± hornblende granite plutons intrude the previously described units west of Douglas Harbour-Foul Bay and in the Whitley Bay area (Fig. 2). The tabular, up to kilometre-scale bodies are oriented approximately east-west, and their contacts parallel the gneissosity in the country rocks. The granites vary in composition from granodiorite to syenogranite, although monzogranite is overwhelmingly the more common type (Fig. 2). Based on pluton geometry and deformation state, it appears that the oldest units were granodiorites and monzogranites, and that the youngest pluton is a syenogranite (Fig. 2). However, individual monzogranite plutons can display a range in composition from granodiorite at the margins to syenogranite near the centres. The granites are medium- to coarse-grained, and vary from equigranular to K-feldspar megacrystic. Accessory magnetite and allanite are common. The monzogranite and granodiorite plutons are in general foliated, but some have massive cores. The plutons commonly contain centimetre-scale to kilometre-scale inclusions of tonalite, quartz diorite, amphibolite, and pyroxenite. The deformed granitic veins observed in all older map units are interpreted as being related to this episode of voluminous, syndeformation granite plutonism.

Clinopyroxene monzonite occurs as small (tens of metres in diameter) plugs and dykes emplaced in the tonalites, and as a border phase to some of the larger monzogranite plutons. The monzonites are fine- to medium-grained and foliated along contact margins with the tonalites. The cores of the plugs are massive.

Deformed gabbro dykes

A swarm of north-northwest-trending gabbro dykes crosscuts all units described above. The vertical dykes are medium- to coarse-grained hornblende-biotite gabbro, and generally less than 10 m wide. They are metamorphosed to the ambient grade of the host Superior Province rocks (see below) and contain a foliation approximately parallel to that in the adjacent units. In total, it appears that the dykes were emplaced before or during the regional deformation-metamorphism episode in the Superior Province.

Little-deformed gabbro dykes

In the Joy Bay area (Fig. 2), a second set of vertical, north-trending dykes occurs in the Superior Province units. The dykes are medium grained hornblende-biotite gabbro with distinct chilled margins. Where the dykes are

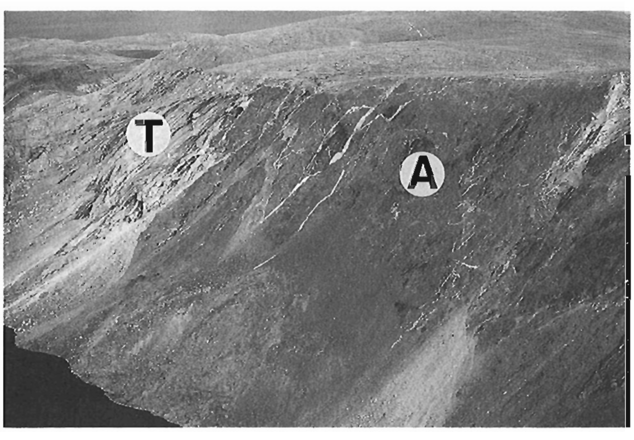

Figure 3. Kilometre-scale raft of hornblende-biotite amphibolite (A) in tonalite (T) from the Superior Province north of Wakeham Bay (Fig. 2). Height of cliff is 500 m. (GSC 205312 H)

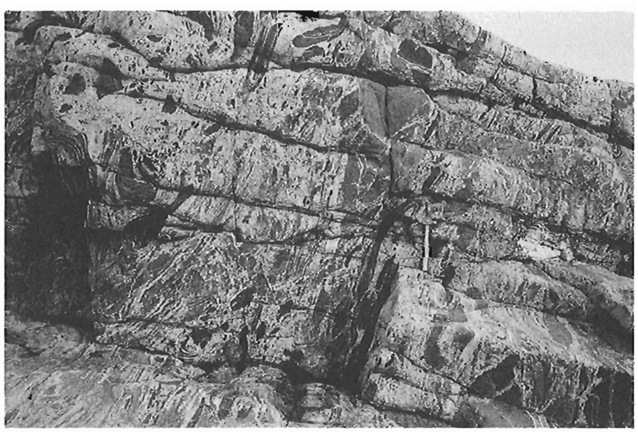

Figure 4. Biotite-hornblende tonalite (light) with abundant mafic enclaves (dark) from the Superior Province north of Wakeham Bay (Fig. 2). Hammer for scale is 39 cm long. (GSC 205313-N)

in contact with Superior Province units, they are essentially undeformed. However, the dykes are truncated by the basal décollement which separates the Superior Province basement from the allochthons (Povungnituk Group) of the Cape Smith Belt (Fig. 5). Within a few metres of the basal fault, a strain gradient from massive gabbro to hornblende-actinolite-biotite schist is developed in the dykes. The metamorphic mineral assemblages in the schist are consistent with assemblages in the overlying Povungnituk Group units (Bégin, 1989). Two origins can be considered for the late dykes: (1) they are late Archean in age and were emplaced following the Archean deformation and metamorphism recorded in the Superior Province country rocks; or (2) the dykes are early Proterozoic in age and related to the mafic volcanism which gave rise to the Povungnituk Group rift basalts. Dating of the dykes by R. Parrish (Continental Geoscience Division) should test one of the two hypotheses.

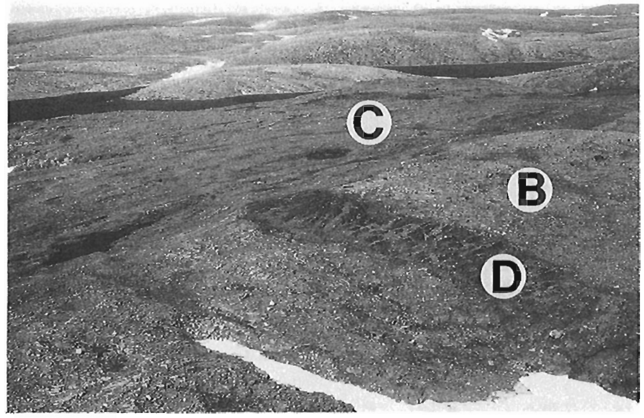

Figure 5. Little-deformed gabbro dyke (D) crosscutting the gneissic fabric in the Archean basement (B) and truncated by the basal décollement of the Cape Smith Belt allochthons. The dyke and the imbricated early Proterozoic cover (C) are metamorphosed to greenschist facies. Photo is from west of Joy Bay (Fig. 2). Width of dyke is 10 m. (GSC 205312-N)

Spartan and Povungnituk groups

The early Proterozoic Spartan Group (Fig. 2) is characterized by two principal units in the map area: clastic sedimentary rocks and hornblende-biotite mafic rocks. The clastic sedimentary rocks comprise quartzite, semipelite, and pelite (locally graphite-bearing) which are interlayered on a metre-scale. The layering in the sedimentary rocks is assumed to be transposed bedding, although no sedimentary structures are preserved. In general, the mafic rocks are fine grained, homogeneous amphibolites interpreted as mafic flows. These are locally interlayered with bands of medium grained amphibolite which display internal compositional layering, and are interpreted as layered mafic sills. To date, the age and tectonic significance of the Spartan Group is not known. These questions are being addressed by analysis of detrital zircon populations in clastic rocks and by geochemical analysis of the amphibolites.

The Povungnituk Group (Fig. 2) is characterized by clastic sedimentary rocks (semipelite, micaceous sandstone, arkosic sandstone and minor pelite) overlain by massive and pillowed basalt flows (Hynes and Francis, 1982; St-Onge and Lucas, 1990b). All lithologies are intruded by gabbro and layered peridotite-gabbro sills. A rhyolite from near the top of the Povungnituk Group has been dated by Parrish (1989) at 1959 Ma.

Narsajuaq terrane

The Narsajuaq terrane was recognized by St-Onge and Lucas (1990a) as an assemblage of plutonic rocks distinct from both the Superior Province units and the early Proterozoic units of the Cape Smith Belt (see also Lucas and St-Onge, 1990; St-Onge and Lucas, 1990c). Recent geochronological work (R.R. Parrish, pers. comm., 1990) has confirmed this field-based assertion. U-Pb (zircon) studies have yielded igneous crystallization ages for Narsajuaq terrane units that range from 1863 to 1834 Ma. This contrasts with the 2882-2780 Ma range for tonalites of the Superior Province and the 1998-1918 Ma range for igneous units of the eastern Cape Smith Belt (Parrish, 1989). St-Onge and Lucas (1990a) also distinguished the Narsajuaq terrane from the Sugluk terrane, found north of Sugluk Inlet (Hoffman, 1985) based on the apparent restriction of sedimentary rocks and granulite-facies metamorphism in the Sugluk terrane. However, given (1) the 1835-1830 Ma igneous ages from the Sugluk terrane (Parrish, 1989) and (2) recently recognized relict granulite-facies mineral assemblages in the eastern Narsajuaq terrane, it is no longer necessary to distinguish the two allochthonous terranes. Rather we consider that the former Narsajuaq and Sugluk terranes compose a single composite terrane, named Narsajuaq terrane. Within the Narsajuaq terrane we will informally describe the sedimentary rocks as the Sugluk group and refer to the plutonic rocks as the Akimmialuq plutonic suite.

Akimmialuq plutonic suite

Layered tonalite-quartz diorite

The principal unit in the Akimmialuq plutonic suite is a well layered sequence of tonalite and quartz diorite (Fig. 6). The tonalite is biotite ± hornblende-bearing, medium grained and equigranular, and volumetrically composes 70 to 80% of the layered unit. Locally biotite granodiorite occurs as 0.5-1.0 m thick bands within the layered sequence. The layers of hornblende-biotite-bearing quartz diorite range in thickness from 2-3 cm to 1 m, and are often disrupted through boudinage within the enveloping tonalite (Fig. 6). Preserved crosscutting relationships indicate that tonalite commonly intrudes quartz diorite, although quartz diorite veins intruding and interlayered with tonalite have been documented. Metre-scale amphibolite and pyroxenite enclaves occur locally within the tonalite sheets. All components of the layered tonalite-quartz diorite unit are crosscut by biotite ± hornblende monzogranite to syenogranite veins (Fig. 7). The granite veins vary from relatively massive (Fig. 6)

Figure 6. Layered tonalite-quartz diorite unit from the Narsajuaq terrane west of Cap de Nouvelle-France (Fig. 2). Gneissic layering is defined by tonalite (grey), quartz diorite (dark), and monzogranite (light) bands. Hammer for scale is 39 cm long. (GSC 205313-S)

to well foliated (Fig. 7) and range in thickness from several millimetres to over ten metres. The oldest, deformed veins are generally transposed into parallelism with the host rock layering, although they can be responsible for significant disruption of the pre-existing layering (Fig. 7). U/Pb dating (zircon) of the layered tonalite-quartz diorite unit south of Sugluk Inlet (Fig. 2) has yielded the following igneous crystallization ages: (1) 1863 Ma for a tonalite layer; (2) 1861 Ma for a granodiorite layer; and (3) 1844 Ma for a quartz diorite layer (R.R. Parrish, pers. comm., 1990).

Mafic intrusive complex

A small composite mafic intrusive complex emplaced in the layered tonalite-quartz diorite unit was mapped on the northeastern side of Deception Bay (Fig. 2). The complex is dominated by a hornblende-biotite quartz diorite which grades into hornblende-biotite tonalite toward the interior of the pluton. The quartz diorite contains inclusions of pyroxenite, diorite, and tonalite. Associated with the quartz diorite is a mafic diorite phase dominated by large ovoid plagioclase phenocrysts up to 3 cm in length. The plagioclase phenocrysts are cored by hornblende, and garnet inclusions are locally present. Clinopyroxene-hornblende ± biotite pyroxenite forms a second distinct intrusive phase associated with, and marginal to the quartz diorite. All units of the intrusive complex are crosscut by monzogranite veins. Foliated margins and a massive core dissected by discrete shear zones characterize the complex.

Hornblende-biotite tonalite

Large bodies of hornblende-biotite tonalite are emplaced in the Narsajuaq terrane north of Sugluk Inlet and on Charles Island (Fig. 2). The kilometre-scale tabular bodies are foliated and lie parallel to the layering in the older tonalite-quartz diorite unit. The tonalites are generally medium grained and equigranular. The intrusions are

Figure 7. Layered tonalite-quartz diorite unit from the Narsajuaq terrane west of Cap de Nouvelle-France (Fig. 2). Gneissic layering is disrupted by anastomosing, deformed monzogranite veins (light). Hammer for scale is 39 cm long. (GSC 205313-B)

characterized by abundant, centimetre- to tens of metre-scale enclaves of (1) sedimentary rocks (Sugluk Group), (2) tonalite-quartz diorite gneiss, (3) pyroxenite, (4) amphibolite, and (5) anorthosite. Layer-parallel to crosscutting veins of biotite monzogranite intrude the tonalite bodies. One of the tonalite bodies north of Sugluk Inlet has 1830 Ma igneous zircons (Parrish, 1989).

Biotite ± hornblende granite

Metre- to kilometre-scale sheets of granite intrude the layered tonalite-quartz diorite unit in the Cap de Nouvelle-France area and the young tonalite unit on Charles Island (Fig. 2). The pink-weathering granites vary in composition from biotite ± hornblende monzogranite to syenogranite. They are equigranular and medium grained at their margins and commonly K-feldspar megacrystic toward their cores. A fine grained, leucocratic aplitic phase is associated with the larger body south of Cap de Nouvelle-France. The granites are in general deformed and locally display gneissic layering. Their margins are marked by enclaves of tonalite, quartz diorite (Fig. 8), and/or sedimentary rocks. U/Pb (zircon) dating of two foliated monzogranite plutons south of Sugluk Inlet (Fig. 2) has yielded igneous ages of 1834 Ma and

1836 Ma (R.R. Parrish, pers. comm., 1990). Another monzogranite body north of Sugluk Inlet has yielded an igneous crystallization age of 1835 Ma (Parrish, 1989).

Small, intermediate intrusive complexes are spatially associated with the granites in the Cap de Nouvelle-France area (Fig. 2). The bodies are hundreds of metres in scale and are dominantly clinopyroxene ± hornblende ± biotite monzonite. They vary locally to hornblende-biotite monzodiorite and hornblende-biotite diorite. The mafic plutonic bodies are emplaced in the granites and are not crosscut by granitic dykes and veins. These complexes are generally massive although their margins contain a weak foliation.

Sugluk group

The Sugluk group comprises highly deformed quartzites, semipelites, pelites, ironstones, marbles, and calc-silicate rocks which outcrop in the northern portion of the Narsajuaq terrane (Fig. 2). Semipelite is the most abundant rock type, while pelite, marble, and calc-silicate rocks are relatively rare. Relatively homogeneous quartzite can be interlayered with semipelite and mafic bands, or with thin (millimetre-scale) pelite bands (e.g. west of Cap de Nouvelle-France; Fig. 2). On Charles Island and north of Sugluk Inlet (Fig. 2), the quartzite layers can contain abundant magnetite and grade locally into ironstone. No primary structures are preserved in the deformed sedimentary rocks.

Two structural settings have been documented for the sedimentary rocks. The lower (southern) contact of several of the sedimentary rock packages is interpreted as a thrust fault (Fig. 2), whereas the upper contact is intrusive, defined as the last occurrence of significant sedimentary rock within the engulfing tonalite bodies. Other bands of Sugluk Group rocks display only intrusive contacts and therefore are interpreted as large map-scale screens between hornblende-biotite tonalite plutons. U-Pb studies of the Sugluk Group (Parrish, 1989) document that: (1) a semipelite north of Sugluk Inlet (Fig. 2) has zircon cores recording ages of >2230 Ma; while (2) a sandstone (also north of Sugluk Inlet) has a population of young detrital igneous grains (1830-1863 Ma) and a population of older detrital grains (>2525 Ma). The two detrital zircon populations in the quartzite suggest a mixed provenance from both an older source and a younger source, possibly recording unroofing of Akimmialuq plutonic suite units or their eruptive equivalents.

Figure 8. Biotite ± hornblende monzogranite (light) with quartz diorite enclaves (dark) from the Narsajuaq terrane on Charles Island (Fig. 2). Pen for scale is 15 cm long (GSC 205313-V)

STRUCTURAL HISTORY

The structural history of the Ungava Orogen can be viewed from the perspective of deformation before, during, and after the collisional event (termed accretion event) responsible for the juxtaposition of three tectonic domains. Distinct, early (preaccretion) structures are recognized in both the Superior Province basement and the Narsajuaq terrane. Late, synaccretion (assembly) and postaccretion deformation and metamorphic histories are common to all rocks in the internal zone of the orogen. The well documented D_1 crustal thickening event (Lucas, 1989; St-Onge and Lucas, 1990b) recorded in the Cape Smith Belt units (Fig. 2) corresponds broadly to the accretion event. In detail, the early piggyback-sequence thrusting recorded in the Povungnituk and Chukotat groups (Fig. 2; Lucas, 1989) precedes the accretion event, which is coeval with out-of-sequence thrusting in the belt.

Preaccretion structures in Superior Province units

A penetrative deformation event predating the accretion event produced a foliation in all units of the Superior Province basement except the little-deformed dykes. As a consequence of the early deformation, all units are characterized by a variably developed fabric. The foliation within individual lithologies is generally defined by a mineral growth fabric and/or a shape fabric of deformed quartz and feldspar. A gneissosity is produced in some units by the interlayering of tonalite ± diorite with monzogranite veins. The planar fabric appears to have developed at granulite-grade conditions, as indicated by aligned orthopyroxene crystals in the foliation. The presence of high-grade foliations adjacent to significantly lower grade reworked basement rocks (Lucas, 1989; St-Onge and Lucas, 1990a) or Cape Smith Belt units (Bégin, 1989) constrains the relative age of the foliation to predate the early Proterozoic accretion event.

The regional distribution of strain related to the preaccretion deformation is highly variable, although the older tonalites and their enclaves appear to record a higher bulk strain than the younger tonalites and granites. Even within the older tonalites, enclaves with pre-existing foliations rarely show evidence of significant straightening during tonalite deformation (Fig. 4). Local high-strain zones (metre- to kilometre-scale) occur in the older tonalite unit, characterized by a well developed, often flaggy, foliation and a stretching lineation. While evidence for bulk noncoaxial deformation exists along these zones, no regionally consistent pattern has been recognized as yet. In summary, the preaccretion deformation event appears to be: (1) synmagmatic, its onset predating the emplacement of the older tonalites and its termination postdating intrusion of the older set of gabbro dykes; and (2) synmetamorphic, characterized by the development of high-grade fabrics. As such, its presumed Archean age can be tested by isotopic analysis of both igneous and metamorphic zircons from the high-grade plutonic rocks.

Preaccretion structures in the Narsajuaq terrane

The outcrop-scale structures in Narsajuaq terrane rocks are, in general, the result of a deformation event which was broadly coeval with magmatism and granulite-grade metamorphism (see below), and which predated accretion to the Superior Province margin (Lucas and St-Onge, 1990; St-Onge and Lucas, 1990a,c). The principal planar fabric observed throughout the Narsajuaq terrane is a gneissic foliation defined by compositional layering (e.g. alternating bands of tonalite, quartz diorite and/or monzogranite; Fig. 6). Mineral growth and deformation-induced shape fabrics within individual layers augment the larger scale gneissic foliation. An east-west-trending, subhorizontal stretching lineation, locally defined by aligned orthopyroxene crystals, is associated with the planar fabric. St-Onge and Lucas (1990a) suggested that the preaccretion deformation reflected transcurrent bulk shear within Narsajuaq terrane, and inferred a dextral sense of shear for the event from kinematic indicators observed north of Sugluk Inlet.

The most significant new observations concerning preaccretion deformation in Narsajuaq terrane pertain to the heterogeneous distribution of penetrative strain during that event. Deformation appears to have been underway before emplacement of the granites and younger tonalite bodies, and outlasted the emplacement of the youngest monzonite plutons. The highest penetrative strains, as shown by the development of "straight gneisses" and boudinage of more competent layers (Fig. 6), are recorded in the layered tonalite-quartz diorite unit and in the sedimentary rocks. Bulk strain within all younger units is significantly lower, as indicated by the absence of straightening deformation in enclaves (Fig. 8). In the youngest granite, quartz diorite, and monzonite plutons, penetrative deformation is generally restricted to the margins, with the cores deforming along small-scale shear zones. These observations suggest that the transcurrent deformation may have localized into narrow corridor(s), or was waning, during the emplacement of these younger (ca. 1835-1830 Ma) plutonic bodies.

Accretion event (D_1)

Narsajuaq terrane was wedged in at the base of the Cape Smith Thrust Belt during the accretion (D_1) event (Lucas and St-Onge, 1990). Only a relatively thin (5-25 m) thrust sheet of Spartan Group amphibolites and/or sedimentary rocks separates the terrane from the underlying Superior Province basement (Fig. 9). The Spartan Group rocks are strongly deformed, while accretion-related strain in Narsajuaq terrane units appears to be restricted to areas directly adjacent (< 10 m) to thrusts. The thrust faults bounding the Spartan Group thrust sheets (Fig. 2) are synthermal peak with respect to the Spartan Group rocks. In contrast, the present roof thrust to the Narsajuaq terrane wedge, found south of Sugluk Inlet (Fig. 2), is a postthermal peak, south-verging thrust which has truncated the original structure. A number of composite Spartan Group-Narsajuaq terrane imbricates were mapped south of Cap de Nouvelle-France, and appear to represent horses in a large-scale duplex structure.

Mapping of the Superior Province basement from Joy Bay to Deception Bay has not revealed any faults of significant displacement that may have been intracontinental décollements during the D_1 event. As such, the assertion that the Superior Province basement is autochthonous with respect to the structurally overlying Cape Smith Belt units and the Narsajuaq terrane (Lucas, 1989; St-Onge and Lucas, 1990a,b) appears to be justified. Parautochthonous Superior Province rocks, generally limited to within a kilometre below the basal décollement, include those rocks involved in: (1) basement-cover imbrication (see Lucas, 1989); (2) penetrative reworking, reorientation, and retrograde metamorphism of pre-existing (Archean) fabrics (Lucas, 1989; see below); and (3) small-scale shear zones with distinct "metasomatic" mineral assemblages in equilibrium with thrust belt assemblages (e.g. kyanite-garnet-muscovite assemblages in biotite tonalites).

Postaccretion crustal-scale folding events

All pre-existing structures were redeformed during two postaccretion folding events. A series of east-west-trending, regional-scale folds extend from Hudson Strait to the western limit of mapping (Fig. 1) and correspond to the D_2 structures described in the Cape Smith Belt (Lucas, 1989; St-Onge and Lucas, 1990b). The D_2 structures are deformed by north-trending D_3 folds of basement and cover, resulting in a map-scale dome-and-basin interference pattern (Lucas, 1989; St-Onge and Lucas, 1990b). This interference pattern is responsible for the

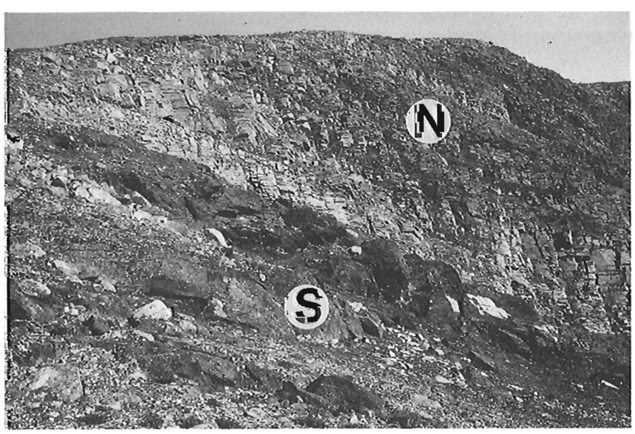

Figure 9. Layered tonalite-quartz diorite unit of the Narsajuaq terrane (N) above a thrust imbricate of Spartan Group amphibolites (S) southeast of Cap de Nouvelle-France. Cliff section is 150 m in height. (GSC 205312-D).

occurrence of a series of tight, doubly-plunging outliers of Povungnituk Group and Spartan Group units both north (Fig. 2, 10) and south of the main Cape Smith Thrust Belt (Fig. 2). The outliers to the north of the belt occur on the flanks of the regional, predominantly west-plunging antiform (Kovik antiform; Hoffman, 1985).

Two observations indicate that Superior Province basement units deformed by folding during the D_2 event. First, fold axes determined by poles to basement (preaccretion) foliation are statistically equivalent to those determined from D_2 minor folds in adjacent Cape Smith Belt units (Fig. 11). Second, mesoscopic folds of layered basement units correspond in attitude and geometry to the map-scale D_2 folds of basement and cover. D_2 folding appears to have been accommodated by millimetre- to centimetre-scale, anastomosing shear zones in the granitic and tonalitic gneisses. The shear zones contain biotite ± chlorite ± muscovite ± garnet assemblages.

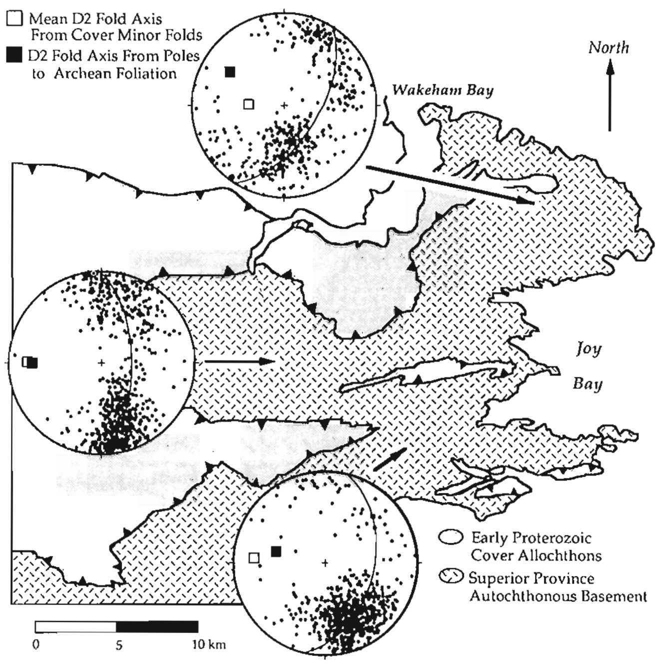

Figure 11. Outline of the eastern end of the Cape Smith Belt with equal area stereonet plots of poles to foliation in the Superior Province basement. Best fit great circle and eigenvectors to the poles are shown, and the corresponding mean D_2 fold axis is indicated. For comparison, the mean D_2 fold axis for adjacent Cape Smith Belt units is shown, representing the vector mean of small-scale D_2 fold axes.

METAMORPHIC HISTORY

The regional metamorphic zonation in the internal part of the Ungava Orogen is interpreted as the result of three distinct thermal events: (1) an older (preaccretion) event associated with the plutonism in the Superior Province; (2) a younger (preaccretion) event related to plutonism in the Narsajuaq terrane; and (3) a final event coeval with tectonic thickening in the Cape Smith Belt and accretion of the Narsajuaq terrane. Mineral assemblages associated with the three metamorphic events are described in the following three sections.

Metamorphic history of the Superior Province units

Mineral assemblages in the Superior Province document conditions ranging from amphibolite to granulite facies. However, all map units contain high-grade assemblages, either in local pockets within a regional lower grade zone, or in regionally high-grade zones delineated by the orthopyroxene isograd (Fig. 12). The older quartz diorite and amphibolite units and the late, deformed gabbro dykes contain subsets of the maximum phase assemblage garnet-clinopyroxene ± orthopyroxene-hornblende-biotite-plagioclase ± quartz. Pelitic units within the clastic sedimentary rock screens east of Deception Bay (Fig. 2) contain the assemblage garnet-sillimanite-biotite-K-feldspar-plagioclase-quartz-granitic pods whereas the predominant semipelite unit contains garnet-biotite-plagioclase-quartz ± orthopyroxene. The older, voluminous tonalite plutons, the granitic bodies, and the younger tonalite intrusions all contain subsets of the maximum phase assemblage orthopyroxene-hornblende-biotite-plagioclase-quartz ± garnet ± clinopyroxene (plus K-feldspar in the granites). As discussed above, high-temperature assemblages are interpreted as preaccretion (i.e. Archean) in age. Geochronological studies of metamorphic and igneous zircons from the Archean units will permit an assessment of the role of advective heating by plutonism in generating the regional, granulite-grade metamorphism.

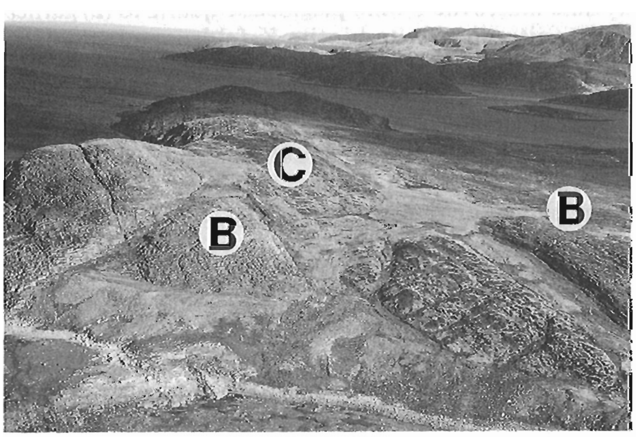

Figure 10. Tightly folded outlier of Povungnituk Group sedimentary rocks and amphibolites south of Douglas Harbour (Fig. 2). Abbreviations are; B — Superior Province basement; C — Povungnituk Group cover. Width of outlier is approximately 125 m. (GSC 205312-A)

The distribution of amphibolite-facies mineral zones in the Archean units appears to be spatially related to early Proterozoic synaccretion thrust structures (Fig. 12). Field observations of retrograde reaction rims in the amphibolite-facies rocks (orthopyroxene rimmed by clinopyroxene; clinopyroxene by hornblende or biotite; garnet rimmed by plagioclase, etc.) suggests that most

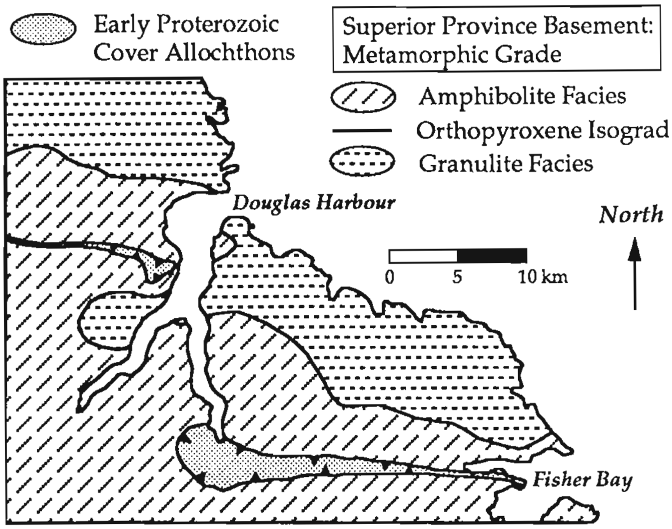

Figure 12. Metamorphic map for the Douglas Harbour area showing the distribution of granulite and amphibolite-facies rocks in the Archean basement, and the location of the allochthonous early Proterozoic rocks in the outliers.

amphibolite-facies assemblages are retrograde. Finally, Archean units with demonstrable early Proterozoic fabric (Lucas, 1990) are characterized by lower amphibolite-facies assemblages (muscovite-biotite ± epidote). These observations suggest that the distribution of subgranulite-facies assemblages in the Superior Province is strictly a retrograde phenomena, presumably reflecting a re-equilibration driven by hydration of footwall basement units. This hypothesis will be further tested with petrological and isotopic work on the amphibolite-facies assemblages.

Pre-accretion metamorphism in the Narsajuaq terrane

The early Proterozoic units of Narsajuaq terrane contain mineral assemblages that range from middle amphibolite to granulite facies. Mafic bands in the layered tonalite-quartz diorite unit contain subsets of the maximum phase assemblage garnet-orthopyroxene-clinopyroxene-hornblende-biotite-plagioclase ± quartz. Tonalites and granites contain subsets of the maximum phase assemblage orthopyroxene-hornblende-biotite-plagioclase-quartz ± garnet ± clinopyroxene (plus K-feldspar in the granites). Pyroxenite enclaves commonly contain orthopyroxene-clinopyroxene ± hornblende. Finally, the sedimentary rocks of the Sugluk group contain the following maximum phase high-temperature mineral assemblages: garnet-sillimanite-biotite-K-feldspar-plagioclase-quartz-granitic pods and garnet-sillimanite-cordierite-hypersthene-biotite-K-feldspar-plagioclase-quartz-granitic pods in pelitic layers; garnet-biotite-quartz-granitic pods in the semipelites; and garnet-sillimanite-quartz in the quartzites. The geochronological studies of Parrish (1989) show significant overlap in the ages of igneous and metamorphic zircons from various granulite-facies units north of Sugluk Inlet. These results strongly suggest that the granulite-grade regional metamorphism in the Narsajuaq terrane is a related to plutonism in the early Proterozoic magmatic arc.

While the granulite-facies mineral assemblages can be related to Narsajuaq terrane arc magmatism, three field observations suggest that the amphibolite-facies mineral assemblages may be largely retrograde and thus postdate the arc magmatism. First, patchy occurrences of granulite-grade mineral assemblages are preserved (often in quartz diorite layers) in the southern, dominantly amphibolite-grade thrust imbricates of the Narsajuaq terrane (Fig. 2). Second, retrograde reaction textures are present in a number of amphibolite-facies assemblages including hornblende rimming clinopyroxene and biotite replacing hornblende. Third, the highly deformed zones adjacent to the faults which imbricate the Narsajuaq terrane (Fig. 2) are characterized by amphibolite-facies mineral assemblages even when the thrust sheets contain granulite-facies assemblages.

Synaccretion metamorphism

The greenschist- to amphibolite-grade regional metamorphism documented in the Cape Smith Belt units (Bégin, 1989) has been interpreted as a consequence of tectonic thickening within the thrust belt (St-Onge and Lucas, in press) and accretion of the Narsajuaq terrane (St-Onge and Lucas, 1990a). In the present map area, the thrust imbricates and outliers of Spartan Group (Fig. 2) provide an essential monitor on the evolving metamorphic conditions in internal zone units during accretion of the Narsajuaq terrane. At the longitude of Cap de Nouvelle-France (Fig. 2), maximum phase assemblages in Spartan Group pelites display a south to north transition from (1) garnet-kyanite-muscovite-biotite-plagioclase-quartz to (2) garnet-kyanite-biotite-K-feldspar-plagioclase-quartz-granitic pods to finally (3) garnet-sillimanite-kyanite-biotite-K-feldspar-plagioclase-quartz-granitic pods. At the longitude of Deception Bay (Fig. 2), pelitic assemblages show a south to north change from (1) garnet-kyanite-muscovite-biotite-plagioclase-quartz to (2) garnet-sillimanite-muscovite-biotite-K-feldspar-plagioclase-quartz-granitic pods to (3) garnet-sillimanite-biotite-K-feldspar-plagioclase-quartz-granitic pods. These assemblages document a northward increase in synaccretion temperature within the internal zone, and an apparent eastward increase in pressure, consistent with the westward plunge of the structures (Fig. 2).

IMPLICATIONS FOR RESEARCH AND MINERAL EXPLORATION

Mapping within the internal zone of the Ungava Orogen has clearly delineated three distinct crustal domains, a result which provides first-order constraints for any mineral exploration strategy in northern Quebec. As an example, ultramafic sills occur in the outlier of Povungnituk Group rocks mapped southeast of Douglas Harbour (Fig. 2). The sills could be related to the mafic-ultramafic magmatism which gave rise to the lac Cross, Katinik, and Raglan (Donaldson) orebodies in the southern thrust belt (see Giovenazzo et al., 1989).

Mapping of the Superior Province within the Ungava Orogen enables the study of Archean crustal evolution in the context of a crustal cross-section with greater than 18 km of structural relief (Lucas, 1989). The predominant westward plunge of D_2 structures in the map area (Fig. 12) allows the geological map (Fig. 2) to be viewed as an oblique cross-section through the Superior Province crust. The tonalite with abundant mafic enclaves in the east (lowest structural levels) passes up structural section to tonalite with rare mafic enclaves in the west. Mafic-ultramafic screens are found at lower structural levels (east) whereas sedimentary screens characterize higher structural levels to the west. Finally, the locus of the voluminous monzogranite sheets is clearly the lower structural level in the east. In total, the mapped units of the Superior Province are interpreted as forming the plutonic edifice of a magmatic arc which experienced a temporal evolution from calcic to more potassic magmatism. The youngest plutons were emplaced at relatively deep crustal levels and may have contributed to the granulite-grade regional metamorphism concurrent with deformation.

The Narsajuaq is similarly exposed in an oblique crustal section, and as such presents a significant opportunity to study the growth of new continental crust in an early Proterozoic magmatic arc. The Narsajuaq terrane also offers the advantage of having different lateral and vertical sections exposed as a consequence of its imbrication during the accretion event. These oblique sections not only provide an excellent opportunity for geochemical and isotopic studies of crustal evolution, but also enable examination of both the synmagmatic, granulite-grade deformation and metamorphism, and the subsequent collisional accretion event.

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Tantato domain, northern Saskatchewan: a segment of the Snowbird tectonic zone

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Hanmer, S., Ji, S., Darrach, M., and Kopf, C., Tantato domain, northern Saskatchewan: a segment of the Snowbird tectonic zone; in *Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 121-133, 1991.*

Abstract

The Snowbird tectonic zone in the Stony Rapids area comprises a triangle of mylonites; the Tantato domain. A footwall of steeply dipping granulite and upper amphibolite mylonites is overlain by a hanging wall of granulite mylonites. The footwall comprises two kinematic sectors, defined on the basis of syn- and post-thermal peak structures: dextral in the west and sinistral in the east. The hanging wall was initially emplaced as a thrust sheet. Post-granulite facies extensional(?) movements occurred at the hanging wall — footwall interface. The Tantato domain may lie at the northeastern end of a 300 by 100 km crustal scale "boudin" structure subjected to bulk pure shear.

Résumé

La zone tectonique Snowbird, dans la région de Stony Rapids est constituée d'un triangle mylonitique: le domaine de Tantato. Un compartiment inférieur de mylonites formées au faciès des amphibolites et des granulites est chevauché par un compartiment supérieur de mylonites formées au faciès des granulites. Le compartiment inférieur est divisé en deux secteurs cinématiques, définis grâce à des structures syn- et post-apogée thermique: dextre à l'ouest et senestre à l'est. Le compartiment supérieur a été mis en place pendant une période de chevauchement. Par la suite, des déplacements additifs(?) ont eu lieu à l'interface entre les deux compartiments. Le domaine de Tantato pourrait représenter la terminaison nord-est d'un "boudin" crustal qui mesurerait 300 km sur 100 km.

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INTRODUCTION

A high amplitude linear anomaly in the horizontal gravity gradient map of the Canadian Shield extends from the Canadian Rocky Mountains to the coast of Hudson Bay, near Baker Lake and beyond (Goodacre et al., 1987). Low amplitude anastomosing linear elements in the magnetic field also occur along the trend of the gravity anomaly and define an elliptical area (300 by 100 km) to the southwest of Stony Rapids (Geological Survey of Canada, 1987). The linear gravity anomaly, corresponding to the Snowbird tectonic zone (Hoffman, 1988), has been considered as a possible Early Proterozoic suture (Walcott and Boyd, 1971; Gibb and Halliday, 1974; Thomas et al., 1988; Hoffman, 1988). Others considered that it reflects a fundamental intra-plate boundary, the limit of "Hudsonian" deformation (Lewry et al., 1978; Lewry and Sibbald, 1980; Thomas and Gibb, 1985).

The Snowbird tectonic zone (Fig. 1) is well exposed in the Tantato domain, Stony Rapids area, northeast of Lake Athabasca (Fig. 2). Earlier workers identified the Tantato domain (Fig. 1) as a triangular area (75 by 80 by 125 km) underlain by granulite to upper amphibolite facies metamorphosed sediments, volcanic rocks and granitoids, folded into a broad, southwest-plunging synform. The contacts with the flanking Dodge and Mudjatik domains (Fig. 3) were identified as narrow, low temperature mylonite zones (see Gilboy, 1980; Gilboy and Ramaekers, 1981 and Hanmer, 1987 for review and references).

The Tantato domain lies at the northeastern end of the magnetically defined elliptical area, along the trend of the Snowbird tectonic zone (Fig. 1). The Tantato domain has been interpreted as part of what is now termed the Rae Province, flanked on its southeastern margin by the reworked Archean rocks of the Cree Lake zone (Hearne Province), the continental hinterland to the Trans-Hudson Orogen (Lewry and Sibbald, 1977; Lewry et al., 1978; Hoffman, 1988). Because trends in the potential fields which are associated with the Snowbird tectonic zone appear to truncate those associated with the 2.0-1.9 Ga Taltson Magmatic Zone, Hoffman (1988) suggested that activity in the Snowbird tectonic zone is, at least in part, post 1.9 Ga.

A brief field reconnaissance in 1986 (Hanmer, 1987) indicated that much of the material underlying the Tantato domain comprises penetratively developed granulite to upper amphibolite facies mylonites and that its northwestern and eastern sides are characterized by dextral and sinistral transcurrent shear, respectively. Building on this preliminary study, we have conducted systematic fieldwork during 1990 in the areas around Bompas Lake, Chipman River, the Clut Lakes and Reeve Lake, supported by boat and fixed wing aircraft.

PRINCIPAL ELEMENTS

Wall rocks

West of Reeve Lake, the Dodge domain wall rocks are coarse muscovite-biotite schists and gneisses with abundant coarse granitic sweats or veins which appear to represent a folded package of coarse muscovite-biotite migmatites. To the east, in the Bompas Lake area, the Mudjatik domain wall rocks are sillimanite-garnet metapelites, semipelites and garnet quartzites which pass eastward into cordierite-garnet metapelites with abundant cordierite-bearing quartz sweats. The metapelites are intruded by concordant sheets of foliated leucogranite, several metres wide, which are probably related to a large body of isotropic to very poorly foliated granite, east of the metasediments.

The Tantato domain can be divided into 2 principal parts (Fig. 3); a footwall of steeply dipping granulite to amphibolite facies mylonites, overlain to the south by an hanging wall panel of shallowly to steeply dipping granulite facies mylonites. Further subdivision of the Tantato domain is outlined below. The relationship between map units is indicated where known.

Chipman batholith

The eastern part of the footwall was formerly described as a "hybrid gneiss" (e.g. Gilboy and Ramaekers, 1981). Northeast of Woolheather Lake it is a clinopyroxene-amphibole tonalitic batholith (Fig. 4, 6), which contains abundant 10 cm to 1 km wide anorthosite, pyroxenite, anthophyllitite, hornblendite and banded amphibolite inclusions, carries variably developed and preserved mylonitic fabrics and is intruded by biotite-hornblende granites and a voluminous swarm of mafic dykes. The tonalite

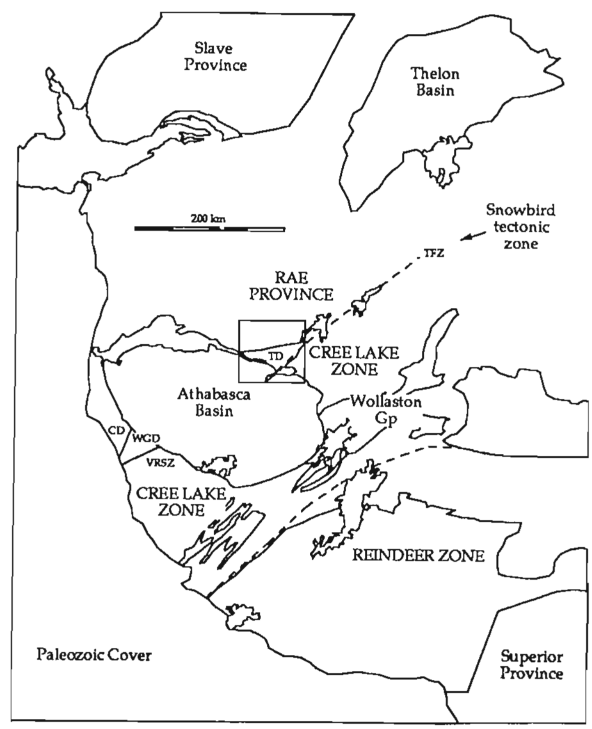

Figure 1. Location of Tantato domain (TD) with respect to the Snowbird tectonic zone and other selected tectonic elements of the western Canadian Shield. Clearwater domain (CD), Tulemalu Fault Zone (TFZ), Virgin River Shear Zone (VRSZ) and Western Granulite domain (WGD). Box locates Figures 2, 3 and 6.

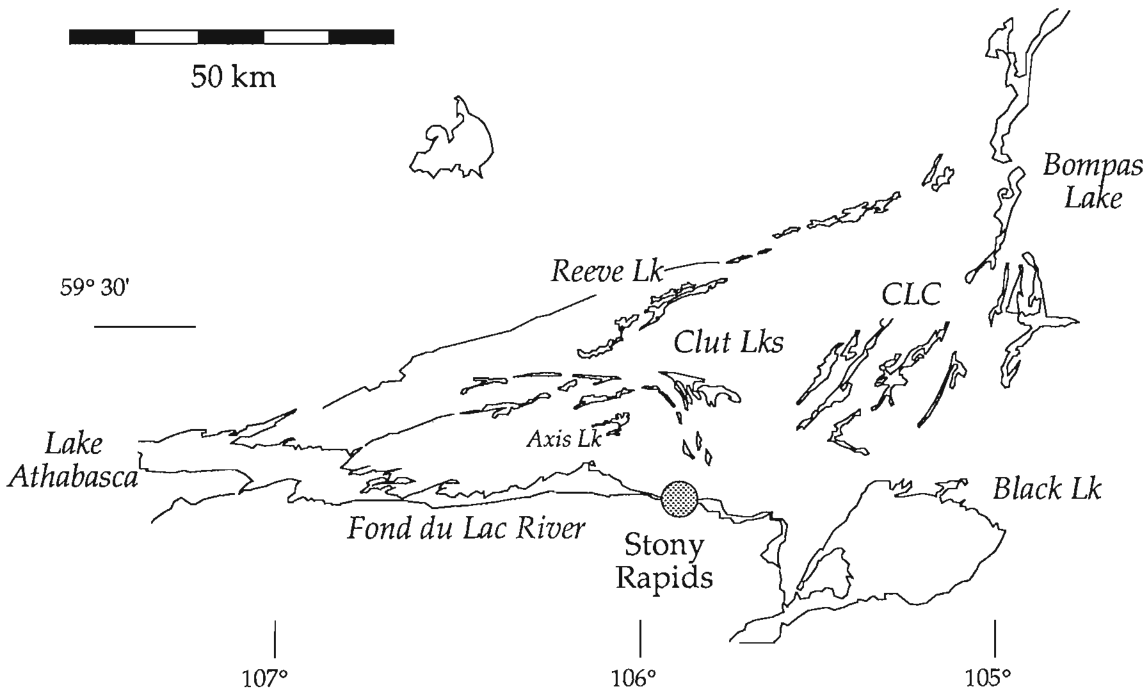

Figure 2. Locations in the Stony Rapids area around and within the triangular Tantato domain. CLC = Cora-Lytle-Chipman Lakes.

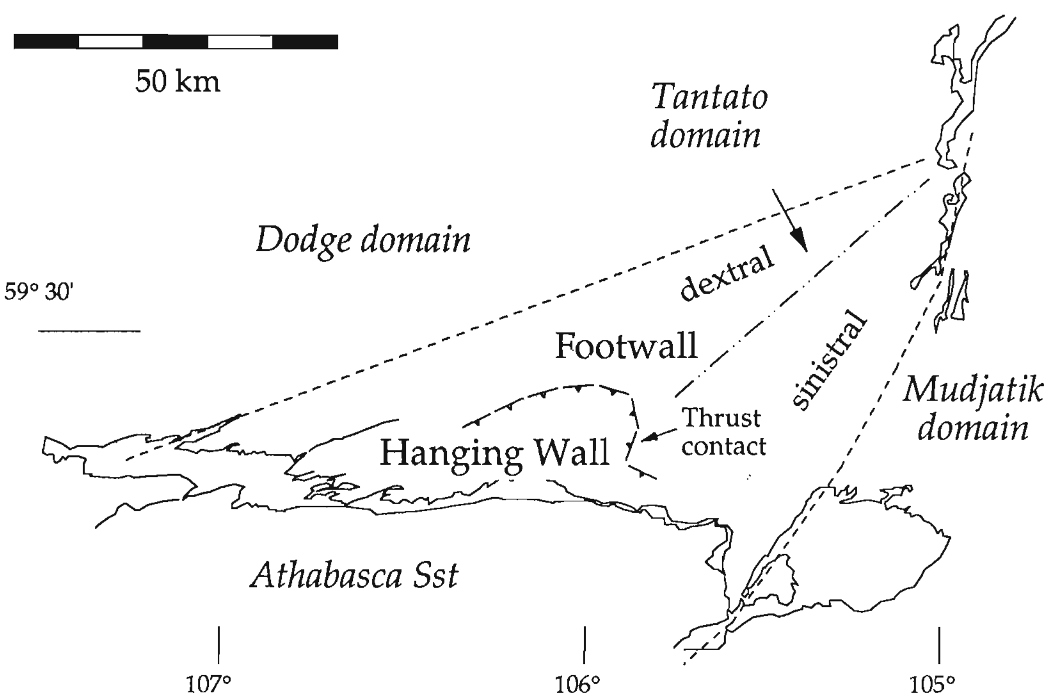

Figure 3. Principal structural elements of the Tantato domain. Dextral and sinistral kinematic sectors are schematically indicated, separated by dot-dash line.

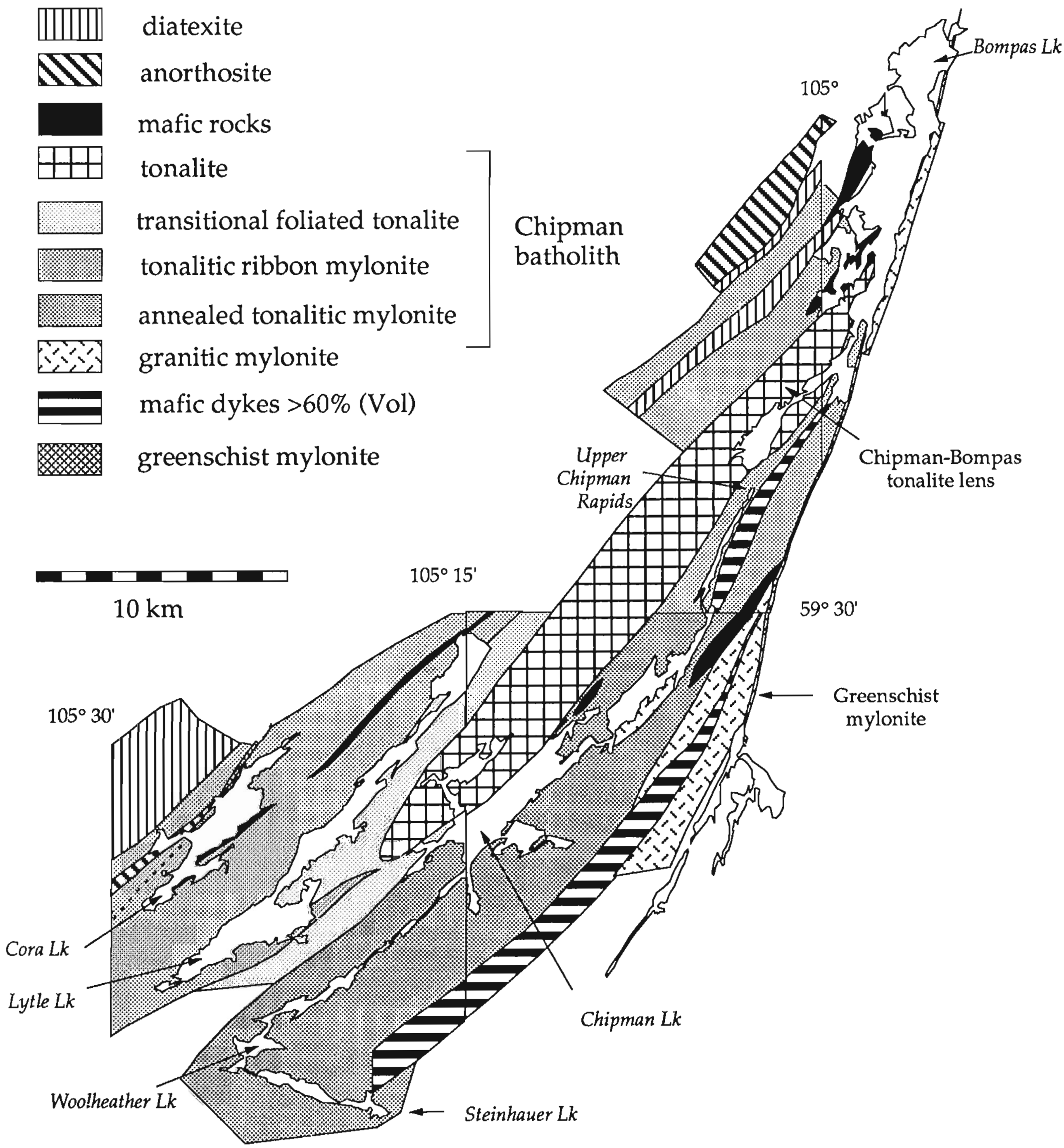

Figure 4. Schematic geology of the Chipman Lake-Bompas Lake area.

is compositionally homogeneous, except for a higher proportion of mafic inclusions in its eastern half. A single xenolithic raft of garnetiferous paragneiss was observed which superficially resembles the diatexite described below.

Tectonic fabrics are mylonitic, with the exception of a lenticular area (5 by 30 km) between Chipman and Bompas lakes where the tonalite is moderately foliated and deformed (Fig. 4). There, the tonalite is a coarse (2 cm +) plutonic rock, charged with refractory material derived either from its source, or picked up en route. In either case, the mafic inclusions appear to represent components of a magmatically dismembered and dispersed layered mafic complex. Compositional banding, due to thin (< 25 cm) lenses or concordant layers of the included mafic material, reflects flow during magmatic emplacement. The banding is folded with an upright, northeast-trending first foliation developed in the fold axial planes. Fold axes are parallel to a strong, generally southwest-plunging extension lineation, irrespective of fold tightness. The foliation, accompanied by a general tightening and transposition of the associated folds, passes progressively into quartz and feldspar ribbon tonalitic mylonite. This is well exposed in large outcrops west of south Bompas Lake, where 50-100 m wide lenses of moderately deformed tonalite are wrapped around by anastomosing belts of quartz and feldspar ribbon mylonite, even at the scale of a single outcrop.

These strain and fabric gradients are small-scale analogues of the relationship of the Chipman-Bompas lens and tonalitic quartz and feldspar ribbon mylonites to ultramylonites to the northwest. Biotite-hornblende-bearing quartz and feldspar ribbon mylonites on the northwest shore of Lytle Lake pass progressively northwestward into spectacular, fine grained garnet-clinopyroxene tonalitic ultramylonites at Cora Lake. There, the tonalitic mylonites are associated with thick (10-100 m), concordant layers of clinopyroxenite, orthopyroxenite, magnetite-garnet clinopyroxenite, amphibolite, anorthosite and sillimanite-garnet metapelite.

East of Chipman Lake, the tonalite is a strongly annealed, well banded straight gneiss, with a generally sugary texture and medium grain size (< 1mm). The straight layering is derived by the transposition of locally preserved metre-scale folds which deform an earlier banding, similar to that described above for the Chipman-Bompas lens (Fig. 4). Locally, plagioclase occurs as polycrystalline streaks and ribbons, or as 1-5 cm porphyroclasts disposed as isolated grains or arranged in trains along the plane of the layering. These observations indicate the initially coarse grain size of parts of the tonalitic straight gneiss and the strong deformation and grain size reduction that it has undergone. Therefore, the straight gneiss is an annealed tonalitic mylonite in which the quartz shape fabric has apparently been destroyed by diffusive dispersion (Hanmer, 1984). Thin, concordant amphibolite, hornblendite, clinopyroxenite and anorthosite layers are an integral part of the annealed mylonite banding. However, after the straight mylonitic layering had formed, thicker mafic layers (> 1 m) behaved more competently than the bounding tonalitic mylonite and extended heterogeneously, probably in response to changing deformation conditions (e.g. decreasing temperature). Coarse tonalite and granitic pegmatite have crystallized in the gaps between competent blocks. Locally, heterogeneous flow around blocks of competent material led to the initiation and amplification of irregular folds of the once straight layering. The result is a disrupted straight gneiss or annealed mylonite of highly irregular aspect.

The presence of the assemblage of anorthosite, pyroxenite, anthophyllitite and banded amphibolite inclusions in the annealed and disrupted tonalitic mylonite to the southeast, and the moderately foliated tonalite and quartz ribbon tonalitic mylonite to the northwest, suggests that the tonalitic units are components of the same body, and that the principal differences between them are reflections of spatial variation in (i) total strain, (ii) the ratio of strain rate vs. recrystallisation (annealing) rate and (iii) the volume of distributed mafic material and of later syntectonic granite intrusions.

Biotite-hornblende granites have intruded the tonalite along its eastern margin and to the southwest of Woolheather Lake (Fig. 4, 6). Except for locally very biotite-rich garnet-hornblende megacrystic variants, they are pinkish-red, coarsely equigranular leucogranites. Although some large, discontinuous bodies outcrop between Chipman Lake and Upper Chipman Rapids, most of the granite within the tonalite batholith occurs as thin (0.1-1 m) veins and is under-represented in Figure 4. The largest granite bodies occur along the eastern margin of the batholith and even intrude the contact between the tonalite and the paragneisses of the immediately adjacent Mudjatik domain wall rocks (Fig. 4). The granites are variably deformed. Smaller veins crosscut the straight banding of the annealed tonalitic mylonites. Isotropic to variably foliated granite is associated with disrupted parts of the annealed tonalitic mylonites and with heterogeneous extension of the competent mafic inclusions. Thicker granite sheets (metres to kilometres) are concordant to the regional foliation in the batholith. They may be poorly foliated and occur immediately adjacent to strongly banded annealed tonalitic mylonite, or they may themselves be penetratively mylonitized and even annealed.

The map units of the Chipman batholith are truncated by the strike of the eastern boundary of the Tantato domain, which changes from northeast-north-northeast-trending at the latitude of Upper Chipman Rapids (Fig. 5). At Bompas Lake they abut against a 750 m wide belt of quartz and feldspar ribbon tonalitic mylonite to ultramylonite derived from the same tonalite precursor.

Mary Lake batholith

Much of the western part of the footwall is occupied by a compositionally monotonous, strongly garnetiferous granite to granodiorite batholith which contains few inclusions (Fig. 5, 6). In the field, two principal sub-units were distinguished which alternate on the scale of metres to kilometres.

The dominant sub-unit is a garnet-hornblende granodiorite. Volumetrically important (20% +) 2-5 mm blood red garnets occur in a very strongly foliated, fine grained, grey, granular matrix of quartz and feldspar. While quartz ribbons may be absent, highly attenuated polycrystalline streaks and ribbons of feldspar are common fabric forming elements. The volumes of the individual feldspar ribbons indicate that the original feldspars were up to several centimetres in size and that the initially coarse grained rock has undergone very strong deformation and severe grain size reduction. Ignoring the garnets, the fabric of the quartzofeldspathic component is mylonitic to ultramylonitic. Much of the garnet is rimmed by fine grained, polycrystalline hornblende which extends as tails, or ribbons from the garnet cores into the quartzofeldspathic matrix. In places, especially in the vicinity of syntectonic granite sheets, hornblende may be accompanied (replaced?) by biotite.

The second sub-unit is clearly a leucocratic buff to reddish pink granite mylonite, at least locally derived at the expense of an original megacrystic texture. Throughout the sub-unit, garnets occur as small (< 1 mm), clear, pinkish-red grains arranged in rims, one grain thick, around aggregates of fine hornblende with which they appear to be in equilibrium. Locally, relict green clinopyroxene occurs with the hornblende. However, the garnet rims clearly occur at the contacts between fresh clinopyroxene and ribbons of polycrystalline plagioclase. These observations suggest that the rim-type garnets in this sub-unit are quite distinct from the core-type described above and that they are the product of metamorphic reactions rather than a relict phase.

Many relatively minor granitic intrusions were emplaced into the Mary Lake batholith, immediately below the hanging wall-footwall interface, north of the Clut Lakes (Fig. 5). North of west Clut Lake, a set of discontinuous leucogranite lenses, up to 1 km thick, lie concordant to the regional foliation in the Mary Lake batholith (granitic mylonite of Fig. 5). These granites are now porphyroclastic quartz ribbon mylonites, remarkable for the very fine grain size and dark colour of their matrix. Other medium- to medium-coarse-grained biotite leucogranites occur as 1-2 km thick concordant map units within the Mary Lake batholith north of east Clut Lake

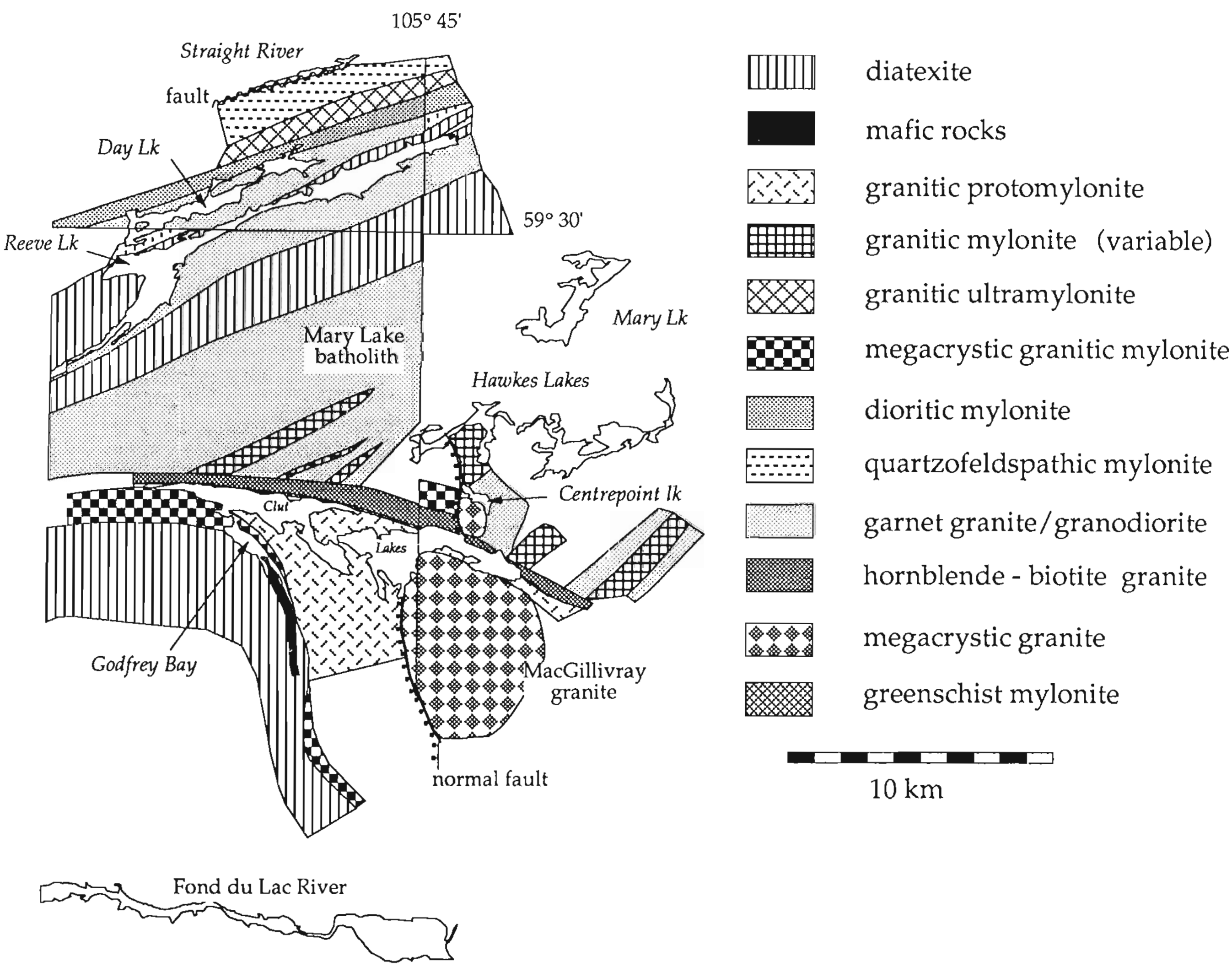

Figure 5. Schematic geology of the Clut Lakes-Reeve Lake area.

(also granitic mylonite of Fig. 5). They are variably deformed and locally form excellent protomylonites to mylonites and are spatially associated with other moderately to poorly foliated garnet-hornblende leucogranitic units of similar dimensions. In contrast to these concordant northeast-striking granitic bodies, a 500 m thick hornblende-biotite granite sheet trends east-west, along the north shore of the Clut Lakes (Fig. 5). The granite is variable in grain size and is locally protomylonitic. Clearly, its east-west strike truncates the northeast grain of the rest of the footwall. The relationships between the various granitic bodies mentioned here are not yet known. However, an elliptical body of poorly foliated megacrystic hornblende-biotite granite outcropping on the south side of Centrepoint lake is flanked along its west margin by a 50 m thick, shallowly to moderately west dipping, diplineated, coarsely porphyroclastic hornblende bearing mylonite (Fig. 5). The similarity to the situation of the MacGillivray granite is obvious.

Axis Lake (hanging wall) panel

The hanging wall panel is essentially composed of diatexite and mafic granulites (Fig. 5, 6). The former predominates in the lower part of the hanging wall panel while the latter are more important in the upper part. Identical diatexite is widespread in the footwall.

In general, the diatexite is white to tan and comprises a banded, leucocratic, quartzofeldspathic matrix with a highly variable content of lilac to delicate pink garnets, local orthopyroxene, graphite and sprays of coarse sillimanite. Clear yellow monazite crystals, up to 1 mm in size, are visible in outcrop. The granitic matrix is a fine grained, quartz and feldspar ribbon mylonite to ultramylonite, banded on a 5-10 cm scale due to variations in grain size, garnet content, composition and colour. Coarse, strongly foliated, garnetiferous pegmatitic layers with attenuated monomineralic plates of polycrystalline quartz and feldspar occur in sharp contact with bands of fine grained, quartz and feldspar ribbon ultramylonite, indicative of leucosome formation during deformation. Some layers may be essentially garnet-free while adjacent ones may have up to 80% + garnet. Garnet grain size is generally constant within a given layer, but may vary from 1-10 mm between bands; rarely garnets attain diameters in excess of 5 cm. The most garnetiferous layers are up to a metre in thickness and are essentially composed of garnet +/− magnetite with a matrix of interstitial quartz, or garnet-clinopyroxene +/− magnetite (iron-formation of Slimmon and Macdonald, 1987). Other layers are quartz-sillimanite in composition, sometimes with spectacular sprays of very coarse aluminosilicate.

Garnet-orthopyroxene-clinopyroxene +/− quartz mafic granulites (Fig. 6) occur as a series of discrete sheets, 100 m to several kilometres thick, separated one from the other by intervals of diatexitic quartz and feldspar ribbon mylonite. Compositions vary from gabbroic to leucogabbroic. The fabrics of the mafic granulites are highly variable, ranging from extremely fine grained and isotropic to pyroxene ribbon ultramylonites. Locally,

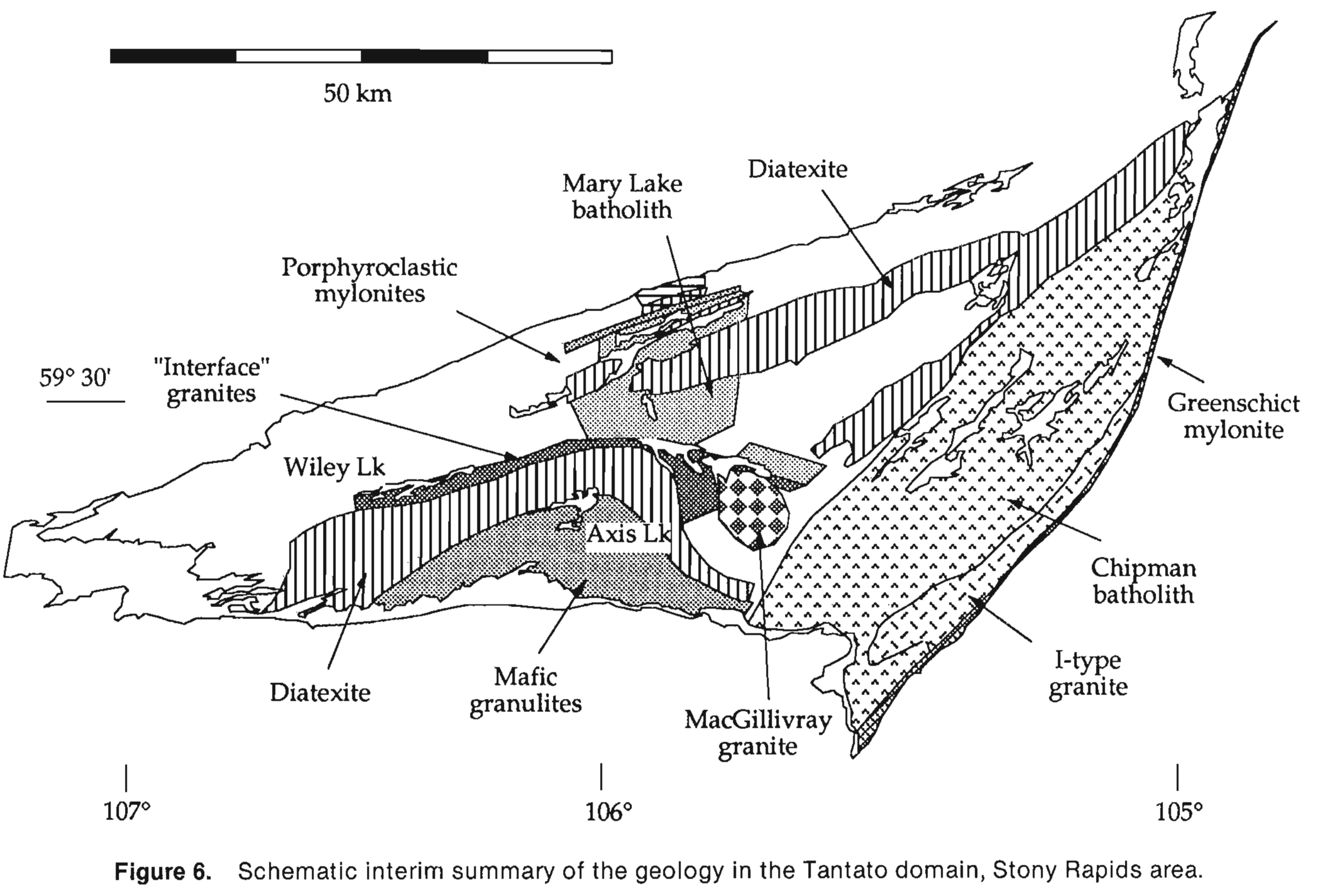

Figure 6. Schematic interim summary of the geology in the Tantato domain, Stony Rapids area.

relict igneous textures (microgabbro to gabbro) are preserved, especially in YZ sections of the finite strain ellipsoid. In coarser grained examples, coronitic structures are common (clinopyroxene rimmed by orthopyroxene, or by garnet and magnetite). Quartz ribbons derived from the deformation of syntectonically introduced quartz veinlets are widespread in a strike-parallel band stretching from south Axis Lake to Fond du Lac River. Subtle compositional banding is only locally visible, although orthopyroxenite layers, 10s of metres thick, occur at Axis Lake. Contacts with the diatexitic ribbon mylonites are either abrupt or sheeted. At the contacts, the mafic granulite is either a pyroxene ribbon ultramylonite or may preserve moderate angles between layering and foliation or between layering/foliation and garnet-orthopyroxene quartzofeldspathic pegmatites. In the Axis Lake area, the thin diatexitic mylonite intervals separate mafic granulite sheets of very different composition and fabric. This suggests that the present disposition of the mafic granulite sheets is, in part, the result of tectonic juxtaposition (stacking). At present we cannot unequivocally determine the relative timing of the plutonic emplacement of the mafic rocks with respect to the metamorphic history of the diatexite.

Clut Lakes (interface) granites

The hanging wall-footwall interface is occupied by a zone of coarse equigranular to megacrystic biotite-hornblende granites (Fig. 5, 6). Among these there are three distinct units (Fig. 5). (i) A 1 km wide garnet-clinopyroxene granitic quartz and feldspar ribbon mylonite extends from west Clut Lake, via Godfrey Bay, to the south (megacrystic granite mylonite of Fig. 5). The original megacrystic texture is locally preserved. Small (< 1 mm), clear, pinkish-red garnets form rims, one crystal thick, around centimetre size clinopyroxene crystals. In the mylonite, few feldspar porphyroclasts are preserved, reflecting the high metamorphic temperatures of the deformation. (ii) A 3 km wide pink to pinkish-red, coarse grained (1-2 cm), equigranular, biotite +/− hornblende granitic protomylonite extends from Wiley Lake to south and east of the Clut Lakes (granitic protomylonite of Fig. 5). Local xenoliths of fine grained homogeneous amphibolite are now transposed into concordant layers. Matrix-rich, quartz and feldspar ribbon protomylonites pass into subordinate mylonites to ultramylonites. (iii) The pinkish-red MacGillivray granite (Fig. 5), south of east Clut Lake, is an extremely coarse grained (< 10 cm) hornblende-biotite megacrystic pluton. It is approximately circular and, although it carries a primary flow fabric, it is essentially non-foliated. It is important to stress that the western margin of the granite contains xenolithic rafts of, and is therefore clearly intrusive into, the protomylonitic wallrocks. The megacrystic phase of the granite is cut by 1-100 m wide vertical sheets of very light coloured, medium- to coarse-grained leucogranite, itself cut by 1m thick veins of very coarse grained (sub-pegmatitic), salmon pink granite. The western margin is heterogeneously deformed. Over a distance of several hundreds of metres from the sharp contact, a foliation is developed in 10 m wide zones which anastomose about large elliptical pods of non-foliated granite. At the contact itself, all three phases of the granite are deformed in a 50-100 m thick, steeply west dipping mylonite zone. Whereas the equigranular granite phase developed an ultramylonite fabric, a very coarsely porphyroclastic mylonite developed at the expense of the megacrystic granite, where 5 cm + feldspar and 2 cm hornblende porphyroclasts are set in a fine grained, feldspar ribbon mylonite matrix.

Western margin

In contrast to the broadly homogeneous Chipman and Mary Lake batholiths, the diatexite and the mafic granulites of the hanging wall panel, the Reeve Lakes area comprises an intercalation of 1-2 km thick, laterally continuous bands of diatexite, grey garnet granite-granodiorite (identical to the Mary Lake batholith), pink garnet leucogranite and hornblende diorite, as well as two-pyroxene pyroxenites, hornblendites and amphibolites (Fig. 5). With the exception of the diatexite, plus some garnetiferous megacrystic granite, these lithologies are mostly fine grained, very straight banded mylonites to ultramylonites. Northwest of Reeve Lake, feldspar and quartz ribbon mylonites alternate irregularly with sugary, annealed mylonites whose ribbon fabrics have been partially to completely destroyed (dispersed). Feldspar porphyroclasts with attenuated polycrystalline tails, etched feldspar ribbons on weathered surfaces and progressive transitions into splendid ribbon mylonites, demonstrate that these annealed rocks are mylonites and ultramylonites. While some diatexite shows excellent quartz ribbon mylonite fabrics, the 2 km thick elliptical volume of diatexite separating the North and South Arms of Reeve Lake is coarse grained, homogeneous and near isotropic (Fig. 5). Mylonites envelope the ellipse as anastomosing belts. This anastomosing structure is also present within the enclosing mylonites at all scales from 1 m to 100s of metres. Although hornblende is a common porphyroclastic phase and biotite is a common foliation forming mineral, widespread garnet-clinopyroxene and orthopyroxene-sillimanite-biotite-K-feldspar mineral assemblages indicate that mylonitization initially occurred under granulite facies metamorphic conditions. Whereas polycrystalline feldspar ribbons are common, one of the most remarkable features in these mylonites, is the common occurrence of 2 cm feldspar porphyroclasts. In other words, the spatial distribution of abundant porphyroclasts in high temperature mylonites coincides with that of heterogeneous, flow in the Tantato domain.

The abrupt contact between the high grade mylonites and the retrogressed migmatites of the Dodge domain wall rock northwest of Reeve and Day Lakes is a clean fault break of unknown displacement (Fig. 5). While minor cataclasite (1 m wide breccia) and some bull-quartz veins (1 m +) occur at Day Lake, they lie well within the Tantato domain.

Eastern margin

A 300m wide belt of biotite +/− chlorite mylonites forms the eastern margin of the Tantato domain between Bompas Lake and the latitude of north Chipman Lake (Fig. 4). They are derived at the expense of various protoliths from both the Tantato and Mudjatik domains. The sillimanite-garnet metapelites of the latter are reduced to a strongly porphyroclastic muscovite-biotite schist with hornblende bearing amphibolite inclusions. At Bompas Lake, the schist is intruded by a medium- to coarse-grained leucogranite (see *Chipman batholith* above) which is itself reduced to a protomylonite. The westernmost 50 m of the biotite schist is further reduced to a chlorite mylonite with concordant sheets of leucogranitic ultramylonite. The point to retain here is that the lowest grade mylonites occur within, rather than at the margin of, the Tantato domain. As in the case of the western margin, except for some minor quartz veining, there is very little evidence of cataclasis or macroscopic distributed brittle failure.

MYLONITES

With few exceptions, mylonites are penetratively developed throughout the footwall and the hanging wall of the Tantato domain. The mylonites within both the footwall and the hanging wall panel are remarkable for the plasticity of their constituent feldspars and generally contain very few, if any, feldspar porphyroclasts. Porphyroclasts described from the annealed tonalitic mylonites were probably derived at the expense of syntectonic pegmatites. Monomineralic, polycrystalline ribbons of clinopyroxene or hornblende, 10s of centimetres in length by less than a millimetre in thickness, in tonalitic, anorthositic and granitic mylonites, as well as orthopyroxene ribbons of similar dimensions in mafic granulites, are common throughout the Tantato domain. The extreme plasticity of the silicate phases is a reflection of the high metamorphic temperatures at which mylonitization occurred.

Mylonites of the hanging wall panel contain syntectonic sillimanite-orthopyroxene-biotite-garnet-K-feldspar or garnet-orthopyroxene-clinopyroxene-plagioclase assemblages. Although garnet porphyroclasts in the footwall mylonites are pseudomorphed by non-distorted hornblende and/or biotite aggregates and represent static post-mylonite retrogression, much of the mylonite in the footwall contains similar granulite facies syntectonic assemblages to those of the hanging wall panel. However, the footwall also contains mylonites with hornblende and biotite as fabric forming elements. The mylonites of the Chipman and Mary Lake batholiths have already been described; other hornblende-biotite mylonites form a kilometre wide belt northwest of Cora Lake where they are clearly derived by the reworking of earlier garnet-clinopyroxene bearing mylonites and represent a dynamic retrogression.

DYKES

A swarm of mafic dykes (Chipman Sill (sic) Swarm of Macdonald, 1980) occurs throughout the Chipman batholith. Individual dykes are up to several 10s of metres wide, but very close spacing results in zones of dyke material 100s to 1000s of metres thick, especially to the southeast of Chipman Lake (Fig. 4). The dyke swarm is spatially coincident with a positive gravity anomaly (unpublished 1:50 000 scale Bouguer anomaly compilation, Geophysics Division, Geological Survey of Canada), which might suggest the presence of more voluminous dykes or a large mafic intrusion at depth. Chilled margins and concentrations of plagioclase phenocrysts near the centres of the dykes are common. In a few notable cases, the dykes are densely charged with euhedral to rounded plagioclase phenocrysts up to 5 cm in size. The dykes are variably deformed. Some are well foliated and subsequently isoclinally and sheath folded, whereas others preserve misoriented wall-rock inclusions, delicate joint-controlled irregularities decorating their contacts and apophyses projecting into the wall rocks.

The dykes contain several distinctive garnet-associated transitional granulite-upper amphibolite facies metamorphic assemblages, although garnet-free assemblages also occur throughout the dyke swarm, particularly at Bompas Lake. The commonest metamorphic assemblage is clinopyroxene-garnet-plagioclase +/− hornblende, with a uniform 1-2 mm grain size. Within initially fine grained hornblende-plagioclase amphibolites, the clinopyroxene-garnet assemblage developed on the concave side of commonly preserved, centimetre scale, curved, bud-like 'fronts' of garnet ("cauliflower structure"). Coalescence of the individual buds results in a homogeneous granulite facies mineral assemblage throughout the dyke. Dykes with the clinopyroxene-garnet assemblage are found in sharp contact with dykes containing a coarse plagioclase-hornblende-garnet assemblage characterized by spectacular euhedral garnet crystals of uniform grain size up to 10 cm, set in a coarse, hornblende dominated matrix. Regular variation in garnet grain size within a given dyke may impart a banded or lenticular structure parallel to the dyke walls. In many examples, elongate tails of quartz-plagioclase develop at the garnet/matrix interface and constitute the principal tectonic fabric forming elements within the dyke. Locally, the very coarse garnets are partially to completely replaced by hornblende-plagioclase aggregates which are strongly deformed in the internal fabric of the dyke. The coarsely plagioclase-phyric dykes are also foliated to non-deformed and may contain clinopyroxene phenocrysts set in a plagioclase-hornblende matrix. Locally, the plagioclase phenocrysts have very thin rims of fine, clear garnets.

Many of the dykes are clearly intrusive. However, there is a complete range of mafic sheets and lenses within the tonalite batholith, from dykes with clear crosscutting contacts and other primary intrusive features mentioned above to disrupted, lenticular inclusions of amphibolite with spectacular euhedral garnets identical to those in many of the crosscutting dykes. As a general rule, the mafic dykes are numerically and volumetrically more important, more commonly crosscutting and less deformed in the eastern part of the Chipman batholith, compared with the west. Whereas some earlier, disrupted dykes are intruded by minor granite veins, the discrete, crosscutting mafic dykes systematically post-date even the major granite sheets which occur near the eastern margin of the batholith. Thin mafic dykes also occur within the biotite and chlorite mylonites at the edge of the batholith (east shore of Bompas Lake) where they are clearly both crosscutting and folded with the mylonite foliation. Taken together, these observations suggest that the western limit of mafic dyke emplacement migrated eastward during progressive deformation, toward the margin of the Tantato domain.

STRUCTURE

Foliations and lineations

The footwall mylonites are generally north-northeast- to east-northeast-trending and steeply dipping. Extension lineations porpoise about the horizontal with a distinct bias toward southward plunges. However, along the eastern margin of the footwall, in the Reeve Lake area and the Bompas Lake area plunge values are highly variable, from strike to dip-parallel. Elsewhere in the footwall plunge values are shallow to locally moderate. In the Reeve Lake and Bompas Lake areas, the variation in plunge appears to reflect the flow required to maintain strain compatibility in the presence of complex flow plane geometries (anastomosing and triangular apex, respectively; see above).

Diatexitic and mafic mylonites at the base of the hanging wall panel and granitic mylonites at the hanging wall-footwall interface dip shallowly under the panel, the direction of dip varying with the curvature of the base of the hanging wall. Within the hanging wall panel, dips become progressively steeper towards the interior. Locally along the Fond du Lac River foliation and layering dip to the northwest. The extension lineation trends approximately east-west in the diatexitic and mafic mylonites, as well as the underlying garnet-clinopyroxene granitic mylonites (Fig. 6). Where the mylonites strike north-south, the extension lineation is close to dip-parallel. Where the mylonites trend east-west to east-northeast, the lineation porpoises about the horizontal. In other words, the pitch of the lineation varies from low to high. However, in the granitic protomylonites at the hanging wall-footwall interface (Fig. 6) the pitch of the extension lineation is everywhere moderate to high; in other words it is discordant to the lineation pattern in all the overlying mylonites. The coarsely porphyroclastic mylonites along the west margin of the MacGillivray granite dip steeply to the west and the extension lineation is dip-parallel. The steeply south-dipping actinolite mylonites along the north shore of Clut Lakes are also dip-lineated.

Folds

Folds within the Tantato domain generally have amplitudes and wavelengths of less than a few metres at most. Fold axes are everywhere coaxial with the extension lineation, irrespective of their tightness and irrespective of whether they deform the mylonite fabric or carry it as an axial planar foliation. The folds either occur sporadically, or form fold trains (packets) without associated regional scale folds. Nevertheless, rare sheath folds which deform the mylonite fabrics within the Tantato domain do occur spatially associated with open to isoclinal folds which also fold the mylonite foliation. Moreover, these occur at all metamorphic grades. We do not envisage discrete polyphase events with radical episodic rotations of the kinematic framework of the flow to account for these observations. Rather, we suggest that lineation-parallel folds initiated in a regime of general noncoaxial flow associated with a component of shortening within the shear plane oriented at an high angle to the flow direction, whereas the sheath folds initiated in the same flow at an high angle with respect to the flow direction in response to local strain rate variations within the shear plane.

Kinematics

Flow in the hanging wall mylonites appears to have been remarkably homogeneous because very few perturbations of the flow plane have either amplified or been preserved. Folds of the mylonite fabric are exceedingly rare and macroscopic shear indicators even more so, despite the presence of abundant rigid garnet porphyroclasts. On the other hand, asymmetrical extensional shears, winged inclusions, asymmetrical pull-aparts and oblique S fabrics in concordant veins, although not abundant, are widespread in the footwall and in the granitic mylonites of the hanging wall-footwall interface. A remarkable, if as yet incomplete picture has emerged. All observed shear-sense criteria east of a line from Centrepoint lake to the apex of the Tantato domain are sinistral, whereas all those to the west are dextral (Fig. 3). Except for the Centrepoint lake area itself (Fig. 5), we have not yet mapped in the critical central area and could only speculate as to the history of interaction between the sinistral and dextral sectors. However, we note that (i) the volumes of dextral and sinistral mylonites are roughly equivalent, (ii) dextral shear-sense indicators have not been observed in the sinistral sector and vice versa, even in areas of relatively low deformation where earlier finite strain increments might be preserved.

The garnet-clinopyroxene granitic mylonites and biotite granitic protomylonites of the hanging wall-footwall interface, as well as the coarsely porphyroclastic mylonites on the west sides of the MacGillivray granite and its Centrepoint lake homologue, contain asymmetrical extensional shears which consistently indicate top-side-down (extensional?) displacements of the hanging wall panel, along the direction of the extension lineations. From the metamorphic mineral assemblages, the relative plasticity of the feldspars, the thickness of the mylonite zones and the isotropic nature of the protolith prior to mylonitization in each case, the displacement-emplacement history is complex, but tractable. The garnet-clinopyroxene megacrystic granite was emplaced at the base of the hanging wall and accommodated top-side-down to the west displacement at upper amphibolite facies. The hornblende — biotite equigranular granite was emplaced beneath the mylonitized megacrystic granite and accommodated top-side-down to the southwest displacements at lower(?) amphibolite facies. The MacGillivray and Centrepoint lake megacrystic granites were emplaced later (see *Clut Lakes granite* above) and focussed top-side-down to the west displacements at upper greenschist facies.

DISCUSSION

The protoliths to the mylonites of the Tantato domain represent a tonalite (Chipman) batholith intruded into a dismembered layered mafic complex and a pile of monotonous pelitic metasediments, with widespread minor iron-rich impure psammites and possible mafic volcanics. The pelitic metasediments may also have been intruded by a granite-granodiorite (Mary Lake) batholith and (Axis Lake-Fond du Lac River) gabbros, according to one possible interpretation of their sheeted, intercalated contacts. During mylonitisation, significant volumes of granite and mafic dykes were emplaced into the eastern part of the tonalite batholith. Combined with the granulite to upper amphibolite facies metamorphic conditions, the lithological association of the Tantato domain is suggestive of the deep roots of a magmatic arc (e.g. Coward et al., 1987).

The large volume of high grade mylonites, plus the fact that so little low grade mylonite, cataclasite and brittle fault products (pseudotachylyte, quartz veining) are developed, indicate that most of the deformation history of the Tantato domain occurred at significant crustal depths. Later uplift was accompanied by relatively minor ductile deformation. Granulite facies diatexitic mylonites in the hanging wall panel are identical to those present throughout much of the footwall. The latter are variably reworked and retrogressed to amphibolite facies mylonites in the central and western parts of the footwall and either formed at, or were retrogressed to, upper amphibolite facies in the eastern part. The important point here is that, assuming (quite reasonably) that the Tantato domain has not been subsequently rotated through 90° +, if the granulite mylonites of the hanging wall are contemporaneous with the granulite and amphibolite facies mylonites of the footwall, then the hanging wall mylonites are thrust related. If the extension lineation is approximately parallel to the transport direction, then thrusting was from either the east or the west.

The qualitative geometrical balance between sinistral and dextral sectors within the high grade, strike-slip footwall mylonites suggests that, if they are contemporaneous, they formed in response to pure shear boundary conditions, at the scale of the Tantato domain. The truncation of the mylonitized Chipman batholith by tonalitic ribbon mylonites adjacent to the eastern margin of the Tantato domain indicates that there were local, but significant, adjustments to those boundary conditions during progressive deformation. Given that the rocks of the Dodge and Mudjatik domains have not significantly participated in this progressive deformation, tectonic flow within the footwall must have led to a relative southwestward displacement of the footwall rocks with respect to the wall rocks on either side. Clearly, the relationship between flow in the footwall and thrust-related progressive deformation in the hanging wall is not simple.

Post-granulite facies, top-side-down mylonites developed in syntectonic granites at the base of the hanging wall panel. Magmatic intrusion along the trailing edge of the upper plate of a possibly extensional shear zone strongly suggests a causal relationship between the locus of emplacement and the locus of progressive deformation. The deformation would create a zone of potential dilation and the soft, recently emplaced, warm granites would localize continued tectonic displacement. Further extensional movements post-date the emplacement of the MacGillivray and Centrepoint lake granites which, if related, might represent two stocks of a 'stitching' granite sealing the east-west branch of the top-side-down protomylonites in east Clut Lake.

In the eastern part of the Tantato domain, there is a coincidence between the distribution of extensive strongly annealed mylonites, absence of anhydrous mafic metamorphic minerals, a gravity high and the presence of a relatively densely packed mafic dyke swarm. If there is a genetic relationship between these various elements, it is unlikely to be due to a thermal anomaly related to the mafic intrusions, given the existence of dry granulite facies mineral assemblages to the west where the dykes are absent. Rather, we might speculate on the possibility that the mafic magmas could have played a role in the introduction of hydrous fluids into hot tonalitic and granitic mylonites, leading to enhanced grain boundary mobility, diffusive dispersion of quartz ribbons (Hanmer, 1984) and hydration of mafic mineral phases.

The abrupt, faulted nature of the western boundary, coupled with the absence of mafic dykes in the Mudjatik domain wall rocks, despite their presence in the greenschist facies mylonites of the eastern boundary, highlights the importance of major lateral displacements along discrete brittle faults at both margins of the Tantato domain. Could major post-mylonite discrete faulting account for the truncation of the magnetic anomaly of the 2.0-1.9 Ga Taltson Magmatic Zone by that of the Tantato domain, effectively leaving the timing of mylonitization unconstrained as of this writing?

Regional context

As has been suggested by other workers, there appear to be a number of lithological and structural similarities between the northern rocks described here and the region south of the Athabasca sandstone. The lithological assemblage of the Chipman batholith resembles the eastern part of the Western Granulite domain (Lewry and Sibbald, 1977; Crocker and Collerson, 1988). The western margin of the Tantato domain has been equated with the Clearwater domain (Wallis, 1970; Lewry et al., 1978). The eastern margin has been correlated with the Virgin River shear zone to the south (Wallis, 1970a, 1970b; Lewry and Sibbald, 1977; Lewry et al., 1978). Descriptions of the Virgin River domain so closely resemble the western Mudjatik domain to the north (see also Gilboy, 1978) as to suggest extension of the former into the northern region. Possible correlation of megacrystic granites in the eastern part of the Chipman batholith with the Junction Granite in the south was pointed out by Gilboy (1978). Mafic dykes in the Virgin River shear zone (Carolan and Collerson, 1988, 1989) were equated with the dykes in the Chipman batholith by Macdonald (1980). Note also that Carolan and Collerson found that mafic dykes are absent in the wall rocks immediately east of the Virgin River shear zone.

To the northeast of the Tantato domain, Gilboy and Ramaekers (1981) identified a narrow low grade "cataclastic" zone, the Gaste Lake shear zone. Our observation that the eastern margin of the Tantato domain truncates both the Chipman batholith and the trend of the western margin of the tectonic zone, suggests that the Gaste Lake structure is probably the extension of the eastern margin.

In light of the foregoing, it is interesting to note that, whereas the eastern margin of the Tantato domain is associated with sinistral shear along a southward-plunging extension lineation in the Stony Rapids area, the upper greenschist to lower amphibolite facies Virgin River shear zone shows dextral shear along a northward-plunging lineation (Carolan and Collerson, 1988, 1989). Furthermore, the sinistral and dextral mylonites of the Tantato domain, the Virgin River shear zone and the Clearwater domain occupy the four quadrants located at the ends of the large elliptical area (Fig. 1), apparent in the magnetic field of Northern Saskatchewan (Geological Survey of Canada, 1987). While noting that the Virgin River shear zone is a relatively minor, low grade structure and that the internal magnetic structure of the elliptical area is abruptly truncated by the trend of the Clearwater domain, we speculate that the geophysically defined ellipse represents the lower part of a continental scale "island" of stiff material, subjected to a major component of regional scale pure shear.

Age Constraints

There are no direct age constraints on the geological history of the Tantato domain. However, in light of the above noted similarities on either side of the Athabasca sandstone, the 1.82 Ga magmatic age of the Junction Granite and the circa 2.0-2.3 Ga metamorphic ages of Mudjatik domain gneisses (U-Pb in zircon; Bickford et al. 1986, 1987, 1988) are indicative of the importance of Early Proterozoic events in the tectono-thermal history of the region.

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Distribution of distinctive Hudson Bay erratics and the problem of the Cochrane limit in Abitibi, Quebec

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Abstract

The distribution and elevation of ice-rafted Paleozoic and Proterozoic erratics first brought in by Cochrane readvance(s) in Lake Ojibway, in northern Abitibi, and redistributed by icebergs and small ice rafts, indicate that the level of the lake was at that time about 100 m lower than its maximum. A Cochrane ice grounding line farther south and farther east than previously proposed is suggested, based on the elevation and on the distribution of the Hudson Bay erratics, on the presence of a widespread compact and massive pebbly clay, and on its geotechnical properties.

Résumé

L'altitude et la répartition de blocs et galets glaciels provenant des roches paléozoïques et protérozoïques du bassin de la baie d'Hudson, transportées d'abord par la (les) récurrence(s) du Cochrane dans le lac Ojibway et ensuite par les icebergs et de petits radeaux de glace, indiquent que le niveau du lac était alors environ 100 m sous son niveau maximal dans le nord de l'Abitibi. La présence d'une argile caillouteuse et compacte, ses propriétés géotechniques de même que l'altitude et la répartition des erratiques provenant de la baie d'Hudson, indiquent que l'aire du glacier Cochrane en contact avec le substratum s'étend plus loin vers l'est et vers le sud que ce qui a été avancé jusqu'ici.

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INTRODUCTION

The mapping of selected Hudson Bay erratics discussed in this report is part of a surficial geology mapping program of the Abitibi mineralized belt undertaken in 1988 by Terrain Sciences Division (Fig. 1). The area lies at the heart of the eastern sector of the Laurentide Ice Sheet (Fig. 2). It has been subjected to major westward and southwestward ice flows in Late Wisconsinan time, and converging southeast and southwest ice flows during the final opening of the ice mass along the Harricana-Lake McConnell interlobate glaciofluvial system at deglaciation (Veillette, 1989). Except for a few high points, most of the study area was flooded by proglacial Lake Ojibway (Vincent and Hardy, 1979). The Hudson Bay erratics under discussion were found in the northern part of the study area (north of 49°N), which is underlain primarily by Archean granite, syenite, massive monzonite, granodiorite, and gneiss. Minor amounts of metasediments consisting of wacke, siltstone, argillite, and felsic metavolcanic rocks occur as sinuous east-west elongated bands. Ultramafic and mafic intrusions occur locally throughout the area (MERQ-OGS, 1983). The discovery of distinctive Hudson Bay erratics farther south and farther east than had been known or anticipated, their stratigraphic position, and their association with glaciolacustrine sediments postdating the southernmost Cochrane readvance (Hardy, 1976) suggest that they were deposited by icebergs or smaller ice rafts detached from larger icebergs. The distribution and elevation of the erratics are used to infer the position of the southern margin of Lake Ojibway shortly before final drainage of the lake into Hudson Bay about 8000 years BP (Hardy, 1976).

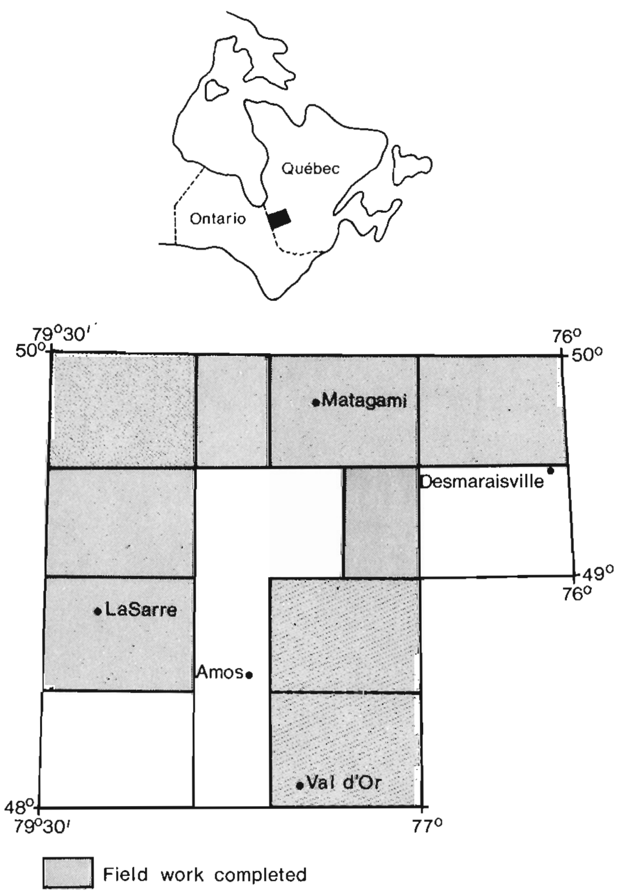

Figure 1. Area covered by the Abitibi surficial geology mapping program.

The use of ice-rafted erratics to delineate a former water plane is not a common technique in the glaciolacustrine environment. Examples are more numerous in the marine environment. Dionne (1972) showed that the relative abundance of ice-rafted Precambrian clasts along the shores of the Lower St. Lawrence can be used to infer the maximum level reached by the Goldthwait Sea. In the Cloridorme area of northeastern Gaspésie, the distribution of ice-rafted Precambrian erratics above the present shoreline and along a rugged, rocky coast that prevented the development of strandlines, also provides a useful minimum marine limit marker since the interior of the peninsula is devoid of Precambrian rocks both in the bedrock and in the glacial sediments overlying it (Veillette, 1988). Similarly, in a model that may have involved both ice-rafting and direct floating, "pumice levels" dated using radiocarbon ages of nearby driftwood and whalebone pieces embedded in beach deposits have been used to correlate raised beaches of different elevations in the Canadian Arctic and in Kong Karls Land in the Spitsbergen area (Blake, 1970; Salvigsen, 1981).

METHOD

Nearly all the zone of occurrence of Hudson Bay erratics is under forest cover. This constraint is compounded by the presence of a widespread, 20 to 30 cm thick, organic-rich topsoil that masks dropstones lying at the surface in undisturbed forested areas. Thus the search is restricted to cleared areas in the vicinity of logging roads, borrow pits, and river and lake banks. Since all lakes and rivers drain into James Bay, that is, northward or northwestward, postglacial fluvial transport of the Hudson Bay erratics by seasonal ice, south of their depositional limit by icebergs of Cochrane age, is impossible. Nearly all accessible roads and borrow pits were examined and several hundred kilometres of lake and river banks were

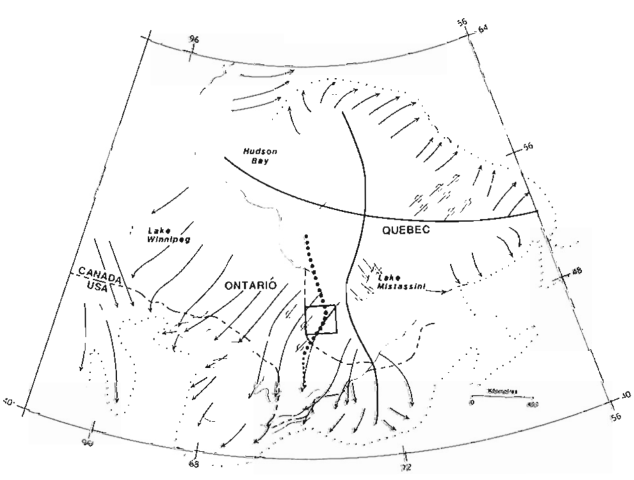

Figure 2. Flow lines during the last glacial maximum through Abitibi-Témiscamingue area (shaded); the dotted line represents the Harricana Moraine (modified from Dyke and Prest, 1987).

inspected by foot traverse along selected shoreline sections. Presence or absence and the relative abundance of the erratics at a particular site were systematically noted. Because the erratics are commonly of cobble and boulder size, they can often be seen from a slow moving vehicle, in cleared areas along logging roads, or from a slow moving motorboat, along the riverbanks.

Vane tests and Atterberg limits were used to determine the shear strength of undisturbed varved clays and massive pebbly clays containing Paleozoic rocks from the Hudson Bay and/or James Bay areas, to distinguish between loaded and unloaded areas by glacier ice. All Atterberg limit test results are not yet available.

DESCRIPTION OF HUDSON BAY ERRATICS

The exotic erratics identified in this survey are classified in two broad groups. The first includes fossiliferous limestone, dolomite, and a yellowish sandstone with a calcareous cement, the latter found so far only in western Abitibi (Fig. 3). The source rocks are presumed to be Paleozoic carbonates of Devonian, Silurian or Ordovician age of the Hudson and James Bay basins although a small outlier is present in Lake Waswanipi, Québec (Fig. 4). Some Ordovician limestone fragments from the Lake Waswanipi outlier (Norris and Sanford, 1968) have been reported in the sandy, predominantly granitic till of northeast provenance in the immediate vicinity of Lake Waswanipi. Although detailed granule counts and till matrix carbonate contents are not yet available for the areas to the southwest of Lake Waswanipi, regional till surveys for mineral exploration purposes in the same area have revealed Paleozoic carbonates in the till overlying the Precambrian substrate (J. Brunet, personal communication, 1990). But the close association between the Paleozoic carbonates and the distinctive low-grade metasedimentary "omars" (see below), their consistent stratigraphic position, and the associated deformed clays (see next section), suggest that the Paleozoic carbonates surveyed are all of Hudson Bay provenance.

Fine grained, dark grey Proterozoic greywackes and siltstone, commonly referred to as greystones since the early descriptions of Bell (1886), form the second group of erratics. These commonly display calcareous, more rarely ferruginous or chloritic concretions of variable sizes which, when weathered out, give the rock a distinctive pitted surface (Fig. 5). The known source area is the Belcher Islands of southeastern Hudson Bay but it may underlie most of the area between the islands and the east coast of Hudson Bay (see Fig. 4; Prest, 1990). Prest has proposed the word "omar" to refer to the erratics derived from the Omarolluk Formation of the Belcher Islands and mapped their distribution (mostly for those erratics showing concretions or concretion pits) in Ontario and western Canada. The omars are widespread in northern Ontario and have been found right up to the Rocky Mountains in western Canada. The reader is referred to Prest and Nielsen (1987) and to Prest (1990) for a detailed description of these erratics, their distribution in Canada, and their special significance in the glacial history of the Laurentide Ice Sheet.

Figure 3. Large dropstone (Paleozoic sandstone) in post-Cochrane sands in northwestern Abitibi. (GSC 205354E

The Proterozoic erratics of Abitibi found on higher grounds are invariably striated, commonly elongated, up to 1 m in diameter (Fig. 6), but most commonly of cobble or small boulder size; some show a whitish ring when chemically altered. Their resistance to breakdown by mechanical forces or weathering makes omars good survivors of glacial transport over long distances and may explain their common occurrence as isolated clasts along wave-washed lake shorelines and riverbanks. On the other hand, subaerial exposure of the Paleozoic carbonates often results in breakdown of the erratics along bedding planes (Fig. 7). Because of this, the carbonates most commonly occur as a litter of fragments of various shapes. Their common association with a blocky, calcareous, pebbly clay suggests that those clasts that are found well preserved at the surface result from erosion of the enclosing clay in late postglacial time.

In the predominantly granitic Precambrian terrain of northwestern Québec, both groups of erratics are readily and positively identified from a brief visual inspection.

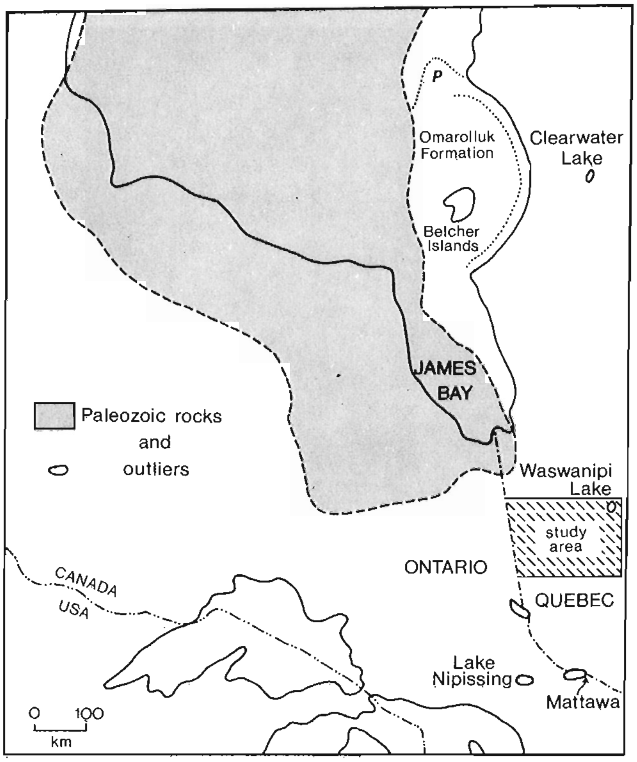

Figure 4. Paleozoic and Proterozoic (P) rocks of Hudson Bay Lowlands and related Paleozoic outliers in Québec and Ontario (modified from Norris and Sanford, 1968).

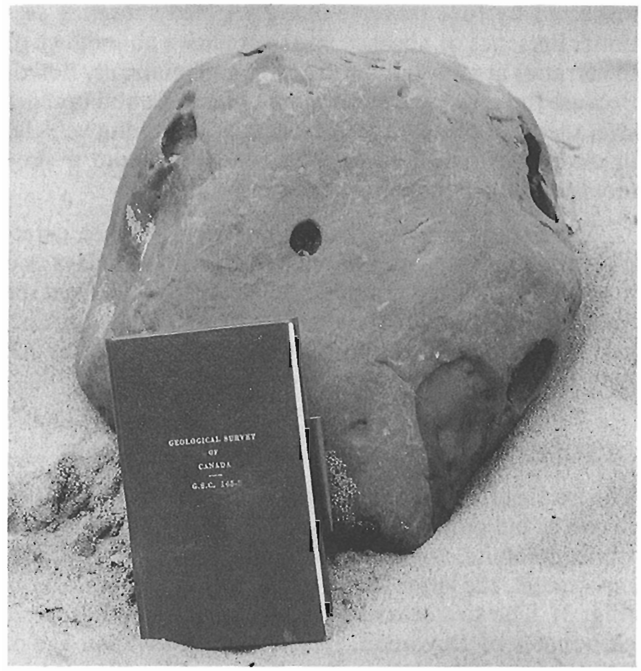

Figure 5. Proterozoic ice-rafted greywacke (omar) in Lake Waswanipi area. (GSC 205353F)

ASSOCIATED SEDIMENTS AND STRATIGRAPHIC POSITION OF ERRATICS

The erratics occur as isolated dropstones on the surface of the former Ojibway lake bed, on glaciolacustrine deltas and terraces, within a brown, stiff, blocky, gritty calcareous clay in the northern part of the area, or within a brown or olive, only slightly gritty non calcareous (or of low carbonate content) blocky clay veneer in the southern part of the area. Pockets of the stiff, brown, calcareous, pebbly clay occur in several places in association with deformation structures, on the flanks of eskers that probably acted as barriers to the movement of icebergs carrying debris in the Ojibway basin (Fig. 8). Iceberg "wallows" are especially well preserved on the northwest flanks of NE-SW trending eskers in northeastern Abitibi. Both the Paleozoic carbonates and the Proterozoic greywackes are always present at the surface of deglaciation deposits and not within the deposits.

DISTRIBUTION OF HUDSON BAY ERRATICS AND PROBABLE ASSOCIATION WITH COCHRANE SURGES

Figure 9 shows the limits of the two Cochrane and the Rupert readvances defined by Hardy (1976), the southern limit of Cochrane ice proposed by Chauvin (1977) south of 50°N, and the revised limit proposed here (with gaps for areas not yet investigated in the field) of the southernmost Cochrane readvance. All of the above limits delineate grounded glacier ice and, except for the revised limit introduced here, were drawn primarily on the basis of large flutings inscribed on the surface of the Ojibway clay plain. Within the revised southernmost limit, the uppermost sediment is a pebbly clay, strongly calcareous, and overconsolidated (with a natural water content in the lower range of the plasticity index) on the crest and flanks of higher points where pits dug with a backhoe show the overconsolidated clay. Vane tests gave a shear strength exceeding 26 t/m^2 (the limit of the instrument used) at several locations. In the north of the study area, lows between flutings are covered with post-Cochrane till varves or fine grained glaciolacustrine sediments rarely exceeding 1 m in thickness. These suggest a relatively short period of deep water following the deposition of the pebbly clay (the southernmost Cochrane readvance and its retreat) and final drainage of Lake Ojibway. Iceberg furrows in the clay are abundant in this zone (Dionne, 1977) and reach lengths of several kilometres, with a maximum width of 200 m and depths of 6 m in the clay.

The second broad zone toward the periphery is characterized by the brown, blocky and discontinuous clay veneer previously mentioned that locally shows a sharp contact with the underlying Ojibway varves but also a gradual one in many locations. Iceberg furrows are also present but are much smaller and not as well developed as in the zone to the north. The predominantly massive clay of variable texture, containing deformation structures and small dropstones, is found in association with patches of clay that appear relatively undisturbed. This clay is extensive in the upper portion of the Bell River drainage basin south of Matagami. Its southeastern extent has not yet been mapped.

Figure 6. Large omar with pea-size concretions in the Lac au Goéland area. Pen for scale. (GSC 205353A)

Figure 7. Shattered Paleozoic limestone clast exposed on crest of esker. (GSC 205353C)

The eastern and southern limits of erratics mapped in this survey and presumed to have been ice-rafted are shown in Figure 9. Beyond this limit there is only one unconfirmed report of a limestone erratic found at the surface of the clay plain, in the Barraute area. South of the southernmost Cochrane ice grounding line, the elevation of the erratics was controlled by the lake level and elevations cluster in a narrow range in the vicinity of 300 m, with some higher ones in the 320-330 m range to the northeast. The highest values thus provide the best estimate, albeit a minimum one, for the former Ojibway water plane, at the time of the Cochrane invasion. North of the grounding line, Hudson Bay erratics are found at much higher elevations, suggesting that glacier ice was responsible for their transport or that a higher glaciolacustrine level existed earlier in Cochrane time. This last interpretation is unlikely since erratics are absent above 330 m south of the grounding line. Extensive surveys over large areas in the elevation range of 330-450 m in the Lake Waswanipi-Lac au Goéland region failed to produce a single Hudson Bay erratic, although Lake Ojibway varves commonly occur at 350 m in this region. For this reason, and because all ice-rafted erratics occur in a narrow elevation range, it is probable that Lake Ojibway, in Abitibi, was about 100 m below its maximum level at the time of the southernmost Cochrane readvance.

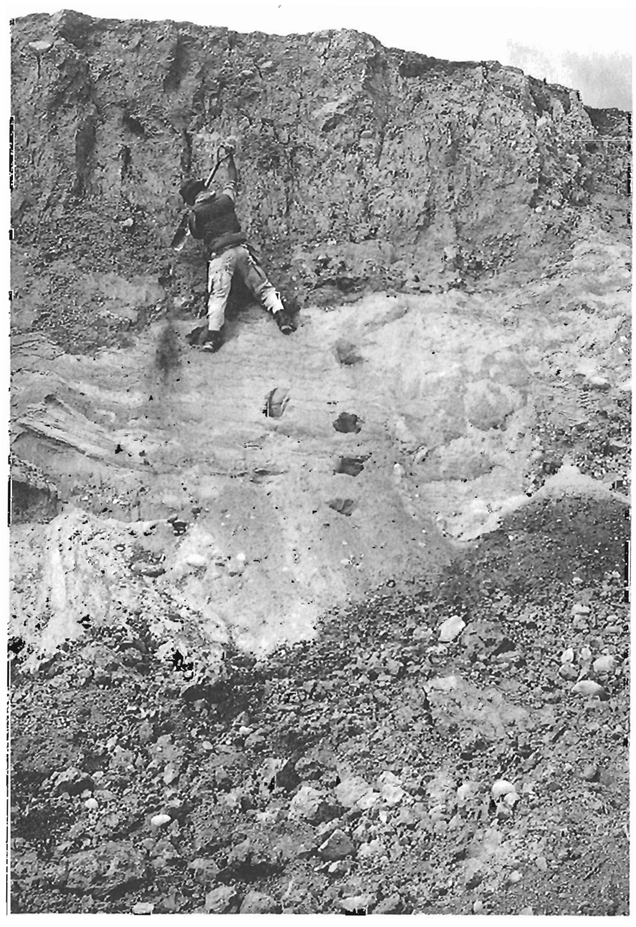

Figure 8. Pocket of massive calcareous pebbly clay (top) deposited by an iceberg on the crest of an esker in the Lake Waswanipi area, 150 km east of the revised (see Fig. 9) Cochrane limit. (GSC 205354G)

DISCUSSION AND IMPLICATIONS OF RESULTS

Unlike northeastern Ontario, where sediments of Cochrane age overlie glacial deposits also containing abundant Paleozoic and Proterozoic erratics from the Hudson Bay basin, Cochrane sediments in northwestern Québec rest on glacial deposits largely devoid of these rocks (Veillette, 1989). This characteristic, the consistent stratigraphic position of the distinctive erratics used, and the distance of transport in excess of 200 km (in some cases) from the classical Cochrane limit provide unequivocal evidence that they were transported by icebergs, probably following the southernmost Cochrane readvance into Lake Ojibway.

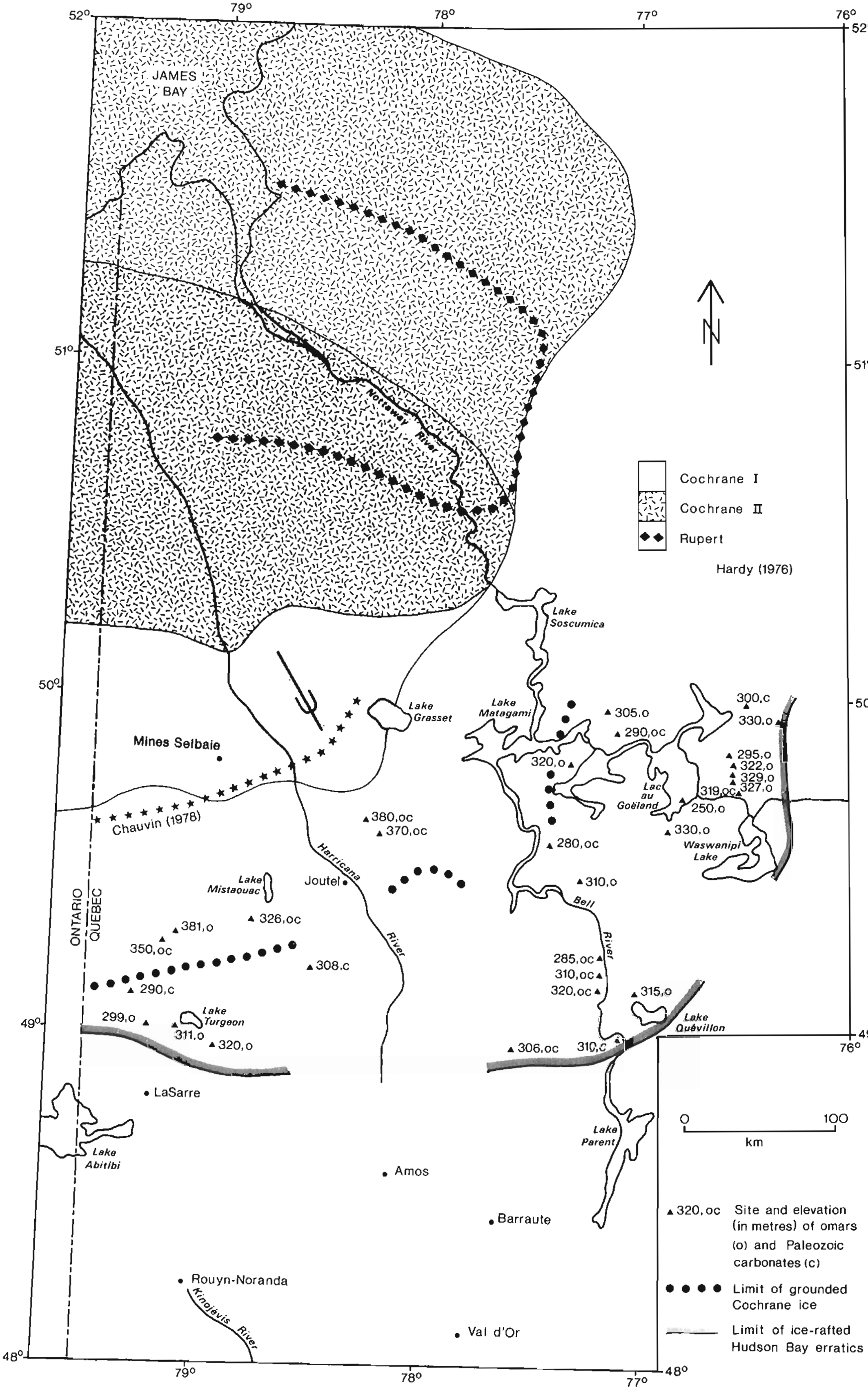

Figure 9. Location and limit of ice-rafted erratics used, revised limit of grounded Cochrane ice, and limits proposed by earlier investigators. The arrow indicates ice flow direction of the Cochrane ice.

The distribution of distinctive ice-rafted omars and Paleozoic carbonates from the James Bay and Hudson Bay lowlands, scattered in a narrow elevation range in eastern Abitibi over the predominantly granitoid glacial sediments of northeast provenance, provides an estimate of the minimum elevation of the southern margin of Lake Ojibway during or after the southernmost Cochrane readvance shortly before 8000 years BP, when Lake Ojibway drained into Hudson Bay. It is probable that the lake was then draining southward through the Kinojénis outlet, as proposed by Vincent and Hardy (1979, Fig. 3H and 3I) but the level of Lake Ojibway must have been considerably lower than the 375-400 m they assigned for the Abitibi region, otherwise Hudson Bay erratics would be widespread below that elevation in Abitibi. Lake Ojibway in central Abitibi (south of 49°N) was then too shallow to allow unrestricted movement of large icebergs. Iceberg furrows do not occur south of the limit of ice-rafted erratics and attain their maximum development only well within this limit (i.e. in the vicinity of the Cochrane grounded line). Iceberg tracks show a preferred NW-SE orientation, which was attributed by Dionne (1977) to prevailing winds from the northwest. This may explain the pronounced southeastward extension of ice-rafted erratics in eastern Abitibi, especially in the large, low-elevation drainage basin of the Bell River system.

The discovery of Hudson Bay erratics that far south and east in Abitibi — a minimum distance of 700 km from the presumed distal contact of the Omarolluk Formation in southeastern Hudson Bay — has direct implications for the paleogeography of the later phases of Lake Ojibway and may guide future research. The erratics offer potential to positively identify the lake outlet(s) farther south in Abitibi that was(were) active at the time of the Cochrane readvance. Seasonal ice or small icebergs may have carried erratics along the outlet channels. The distribution of the erratics analyzed in conjunction with the orientation of iceberg tracks may allow the reconstruction of iceberg "routes" dictated by prevailing winds and/or the position of outlets. It is unlikely that the icebergs responsible for this transport in eastern Abitibi carried Hudson Bay erratics from the Matheson till of northeastern Ontario because of the large, north-south barrier created by the Harricana Moraine (Fig. 2). The flanks of the moraine show abundant scour marks and deflection of iceberg paths coming from both sides of the moraine south of 50°N. Cochrane ice east of the Harricana Moraine is a more likely source for the erratics dispersed to the southeast and east of the area.

From an applied perspective, the widespread occurrence of ice-rafted erratics so far removed from their source rocks should serve to caution those involved with surface boulder tracing studies for mineral exploration purposes in Abitibi. Icebergs efficiently transported glacial debris in Lake Ojibway over several hundred kilometres. While the data presented here apply to icebergs of Cochrane age, it is probable that iceberg activity was also prevalent in earlier stages of Lake Ojibway. Boulder tracing or long distance provenance studies should then preferably be limited to areas that are known to be well above the maximum water plane reached by the lake or, if below this level, should be restricted to clasts found in till and not at the surface of it. This is not to say that boulder tracing at the surface is totally unreliable in a glaciolacustrine environment, but that a constant awareness of the possibility of ice-rafted clasts is necessary for an intelligent interpretation of glacial transport data. Abundant mineralized boulders found within a small area are significant indicators of mineralized sources.

CONCLUSIONS

The distribution of dropstones and ice-rafted erratics consisting of Paleozoic carbonates and sandstone from the Hudson Bay basin and of distinctive Proterozoic greystones from southeastern Hudson Bay (Prest and Nielsen, 1988; Prest, 1990) mapped in Abitibi, Québec, since 1988, provides an estimate of the elevation of the southern margin of proglacial Lake Ojibway shortly before its final drainage into the Tyrrell Sea about 8000 BP (Hardy, 1976). The erratics were first transported by southeastward Cochrane ice and then, ice-rafted to their present position a total minimum distance of 700 km for the greystones from the distal contact of the Proterozoic rocks in Hudson Bay. All dropstones reported here were found at the surface of eskers, moraines, glaciolacustrine deltas, and other topographic high points that formed shoals or islands on the Ojibway lake bed. The highest level for a Hudson Bay dropstone was 330 m but most occur in the 300-320 m range. Lake Ojibway left shorelines and washing limits at elevations of up to 430 m in the area where dropstones are most abundant. Thus the highest elevation of 330 m for dropstones suggests that the elevation of the lake was considerably lower in Abitibi during the Cochrane readvance. The absence of large iceberg furrows near the limit of ice-rafted Hudson Bay erratics indicates that the erratics were probably carried into place by small bergs or ice rafts detached from larger icebergs. The frequency of erratics decreases toward the limit.

Data and field observations obtained to date suggest, especially for northwestern Abitibi, that the traditional interpretation of Hughes (1959), that is, that the Cochrane event is characterized primarily by a massive pebbly clay, locally moulded into large-scale flutings that truncate Ojibway varves and other deglaciation features, still applies. The influence of icebergs, however, on the sedimentology and structure of the underlying sediments remains poorly known and requires further research for a better understanding of the late deglaciation history of northern Abitibi.

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Older ice flows in the Matagami-Chapais area, Quebec

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Abstract

Bedrock-inscribed ice flow measurements, mainly cross-striations from the Matagami-Chapais area of Québec, suggest the presence of a NE-SW oriented ice divide that probably existed in Late Wisconsinan time, between the two regions. This ice mass configuration preceded that of the last deglaciation, probably at or some time after the last glacial maximum. The model is in accord with previous ice flow data from Abitibi-Témiscamingue and with lithological indicators of glacial transport. Striated planes indicating former westward and northwest flows in the Matagami area and a southeast flow in the Chapais area are in general agreement with flow lines obtained from glaciological models of the ice sheet during the last glacial maximum.

Résumé

Des marques d'écoulement glaciaire, pour la plupart des recoupements de stries sur le socle de la région de Matagami-Chapais, au Québec, suggèrent l'existence à une époque antérieure à la dernière déglaciation, d'une ligne de partage des glaces entre ces deux régions. Un tel modèle correspond aux données d'écoulement glaciaire déjà recueillies en Abitibi-Témiscamingue et aux indicateurs lithologiques de transport glaciaire. Ces nouvelles mesures de terrain, témoignant d'un mouvement antérieur vers l'ouest et le nord-ouest dans la région de Matagami, et vers le sud-est dans la région de Chapais, appuient les grandes lignes d'écoulement glaciaire déduites de modèles glaciologiques de l'inlandsis au dernier pléniglaciaire.

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INTRODUCTION

A major objective in the reconstruction of the glacial history of a region is to determine the general configuration of the former ice mass that covered it. The positions of ice divides are commonly estimated for the ice mass during the last deglaciation phase using the orientation of glacial landforms and striations inscribed on bedrock and/or by measuring the tilt of strandlines in former glaciolacustrine and glaciomarine environments. Reconstruction of an ice mass configuration that predates that of the last deglaciation, however, can also be attempted using ice flow indicators providing that (1) the former position(s) of the ice divide(s) was (were) substantially different from those associated with the final deglaciation, (2) an adequate number of ice flow features have survived the eroding action of the ice sheet during the last deglaciation, and (3) these former ice flow features can be reliably identified and correlated. To be meaningful, results must be based on a large number of field measurements to compensate for the absence of ice flow macrofeatures that are generally destroyed during the last deglaciation. The satisfaction of all the above provisos has allowed the establishment of a relative ice flow chronology from cross-striated outcrops for that part of Abitibi-Témiscamingue west of the Harricana Moraine in northeastern Ontario and northwestern Québec (Veillette, 1986, 1989). This "stratigraphy" of erosional surfaces evolved from an ordained and predictable sequence: striated surfaces from an oldest west-southwest flow are crossed by striations of an intermediate southwest flow, and finally by those of a last southeast flow at final deglaciation, for the area between Lake Nipissing and James Bay (Fig. 1).

This report presents new, preliminary, bedrock-inscribed ice flow measurements for the Matagami area east of the Harricana Moraine, in the northeastern Abitibi region (Fig. 1), mapped in 1986 and during the summer of 1990. The presence of indicators of westward and northwest (toward James Bay) ice flows that predate the widespread southwest flow of final deglaciation is evaluated in conjunction with evidence of an older southeast flow, immediately to the east in the Chapais and Chibougamau areas (Bouchard and Martineau, 1985; Prichonnet and Beaudry, 1990). These recent data suggest the presence of a NE-SW ice divide in this sector, preceding the ice mass configuration that prevailed during final deglaciation, with a dispersion zone to the northeast responsible for the widespread southwest (190°-220°) flow in the study area. The ice mass configuration reconstructed using these field measurements is compared with configurations derived from glaciological models of the eastern sector of the Laurentide Ice Sheet by Boulton et al. (1985), Fisher et al. (1985), and by an ice flow model for the last glacial maximum by Dyke and Prest (1987).

Most of the study area, except for a few high points, has been flooded by proglacial Lake Ojibway (Vincent and Hardy, 1979) The substrate consists primarily of Archean granitic and gneissic rocks, with minor amounts of felsic metavolcanics, wackes, siltstone, and argillite (MERQ-OGS, 1983). A small outlier of Ordovician limestone is present in Lake Waswanipi.

CROSS-STRIATED SITES

The location of 61 cross-striated sites involving striated surfaces of two or more distinct ice movements along with unidirectional striations is shown in Figure 2. An extensive network of logging roads built in recent years in the dense black spruce forest between Matagami and Desmaraisville, an area covering some 8000 km² (Fig. 2), along with a chain of large navigable lakes (Matagami to Waswanipi) and rivers, provides reasonable access to the area. The scarcity of glaciofluvial sand and gravel of suitable quality forced logging operators to use till as a substitute material for road construction. The sandy till that constitutes the tail of crag and tails deposited by southwest flowing ice at deglaciation, became the favourite target of road builders since several of these streamlined forms outcrop above the clay plain, with their uppermost portions located above the water table. This situation led to the uncovering of numerous down-ice surfaces of rock knobs (crags) — a portion of the bedrock surface that is rarely observed under normal field conditions. Many of the surfaces were inscribed with striations and ice flow marks of older ice movements, which had been preserved in these sheltered positions from the erosive action of the last southwest flow. In addition, the excavations provided numerous spectacular longitudinal sections in the streamlined till.

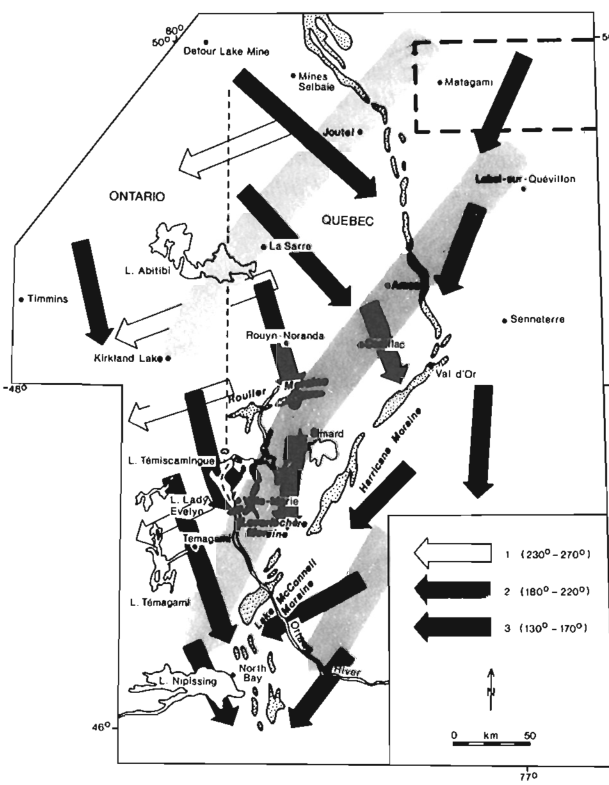

Figure 1. Chronology of ice flows of Abitibi-Témiscamingue, 1 is oldest. Azimuths 130° to 170° refer to the area west of the Harricana Moraine; the study area is in the northeast corner. Adapted from Veillette (1989)

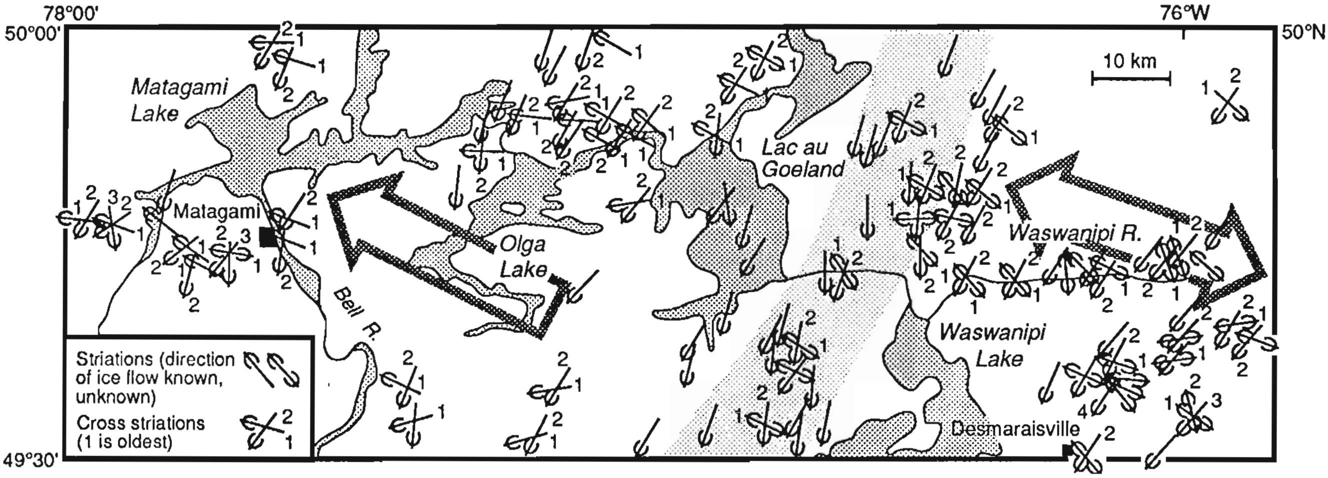

Figure 2. Location of 61 cross-striated sites in northeastern Abitibi. West of the shaded zone between Lake Waswanipi and Lac au Goéland, striations indicate older westward and northwestward ice movements. The large arrow represents the direction of an older northwest regional ice flow. Double-headed arrow indicates uncertain direction.

Older westward or northwestward ice flow

West of a line between Lac au Goéland and Lake Waswanipi, definite westward (260°-270°) and northwestward (270°-310°) striated surfaces with stoss-and-lee topography have been preserved on the lee side of crags and beneath till deposited by the last southwestward (190°-220°) flowing ice (Fig. 3). Sites showing similar ice movements, not reported here, were observed by the senior author in the Lake Soscumica area about 40 km north of the study area and along the Matagami-James Bay road north of 50°N.

Several outcrops showing preferred NW-SE or W-E striated planes, in sheltered positions in relation to the last southwest flow at deglaciation, were also found in the area east of Lake Waswanipi (Fig. 2), confirming earlier reports for north of Lac au Goéland by R.D. Thomas and I. Haga (personal communication, 1988). Some doubt remains, however, as to the prevalent older sense of ice movement in this area. Marks indicative of both northwestward and southeastward flows and others without distinctive ice flow directions were observed. For the present only the axis of ice movement (large double-headed arrow, Fig. 2) is shown for these older ice flows.

In the region south of Lake Olga a preferred orientation to the southwest (240°-250°) for the older ice movement has been observed at some locations (Fig. 2).

Older eastward and southeastward flows

While the correct sequence of ice flows remains to be clarified in the Lake Waswanipi area, marks of older eastward and southeastward flows are abundant in the Chapais-Chibougamau area (Fig. 4; Prichonnet and Beaudry, 1990). Earlier reports by B. Lauriol (personal communication to the senior author, 1987) of former striated planes with a 145° direction of ice movement crossed by those of the dominant regional flow at 235°, and with intermediate striations toward 190°, to the west of Chapais were confirmed and documented by subsequent detailed investigations of ice flow and glacial transport studies in 1989 by Prichonnet and Beaudry (1990) and Beaudry and Prichonnet (1990).

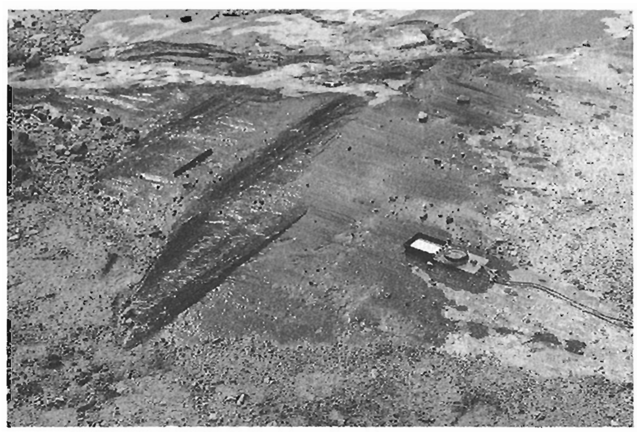

Figure 3. Outcrop on lee side of crag northeast of Lake Matagami showing striated plane toward 300° (toward upper right hand corner of photo), crossed by striations of the last (190°) ice movement at deglaciation (compass). (GSC 205355)

A NORTHEAST-SOUTHWEST ICE DIVIDE

The 250 km east-west transect shown in Figure 4 summarizes our present knowledge of older ice flows in this area. Providing the basic assumption that older southeastward striations in the Chapais region are roughly synchronous with older northwestward striations in the Matagami area is correct, then at some time, prior to final deglaciation, an ice divide must have existed between the two areas. This model provides a valid explanation for the absence (preliminary observations during summer 1990) of Proterozoic lithological indicators of glacial transport — either from Hudson Bay or from the Lake Mistassini area (see Fig. 5 for location) — in the regional till of the Matagami area, if this condition prevailed for most of the Late Wisconsinan. During late deglaciation, the gradual opening of the Labrador Sector of the Laurentide Ice Sheet along the Lake McConnell-Harricana Moraine glaciofluvial complex (Fig. 1) initiated the drawdown that shifted the original northwest flow of northeastern Abitibi (Matagami) to a westward flow and finally to the southwest flow of final deglaciation. During the same deglaciation period, the ice margin configuration responsible for the original southeast flow, farther east in the Chapais-Chibougamau area, shifted first to a southward flow and then to the prevalent southwest flow of final deglaciation as reported by Prichonnet and Beaudry (1990).

The flow lines derived from those new data (Fig. 2 and 4) and that from Abitibi-Témiscamingue data have been compared to flow lines from three recent glaciological models of the ice sheet during the last glacial maximum (Fig. 5). Assuming that the former northwest and southeast flows (Fig. 4) are representative of ice flow conditions at or near the last glacial maximum, the models of Fisher et al. (1985) and Boulton et al. (1985) with deformable beds under Hudson Bay (Fig. 5A and 5B), with some modifications, show the best agreement with the data on ice flows acquired during 1990. Both models show an elongated NE-SW ice divide to the southeast of James Bay and Hudson Bay. The northwest oriented striations and stoss-and-lee topography, investigated during summer 1990, could then represent the northwest flank of the NE-SW-oriented divide proposed by these authors, while the former southwestward striations of western Abitibi (Veillette, 1989) would mark the southwest extremity of the divide. While the model of Boulton et al. (Fig. 5A) shows good agreement with the Chapais-Chibougamau southeast older striations, the one of Fisher et al. (Fig. 5B) does not account for this flow, but could explain, with minor modifications, the northwest older striations of the Matagami area. The principal merit of the two models (Fig. 5A and 5B) when compared to striation measurements is that they imply synchronous ice movements toward Hudson Bay (northwest) and toward the St. Lawrence Valley (southeast). The preliminary provenance data in the area, that is, absence of Hudson Bay Proterozoic erratics in the Matagami area (except in the Cochrane till, see Veillette et al., 1991, this volume), and evidence for former glacial transport to the southeast in the Chapais-Chibougamau-Lake Mistassini area (Dionne, 1986; Beaudry and Prichonnet, 1990) support a similar configuration of the ice mass in pre-late deglaciation time. The model with a rigid substrate under Hudson Bay (Fig. 5C) results in different ice flow conditions at the last glacial maximum but, like the previous models, shows a dispersion centre to the southeast of James Bay and Hudson Bay. The model presented by Dyke and Prest (1987), which incorporates results of striations available at the time, also favours the presence of an ice divide (the Mistassini ice divide) between Lake Mistassini and the Abitibi area during the last glacial maximum (Fig. 6).

The persistent absence of differential weathering on the two or more facets of hundreds of cross-striated planes in the Abitibi-Chapais-Chibougamau area

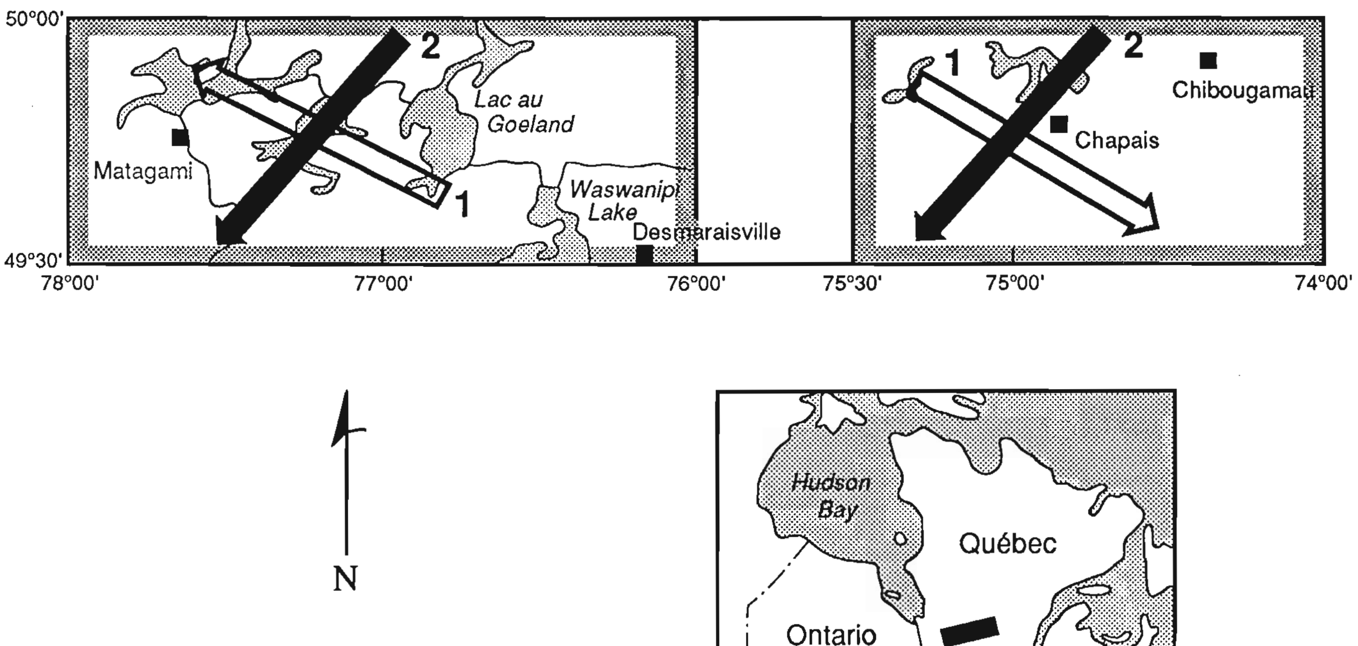

Figure 4. Transect between Matagami and Chapais, Québec, showing the generalized sequence of ice flow (1 and 2). Measurements west of 76°W are from this study; those east of 75°30'W are from Prichonnet and Beaudry (1990)

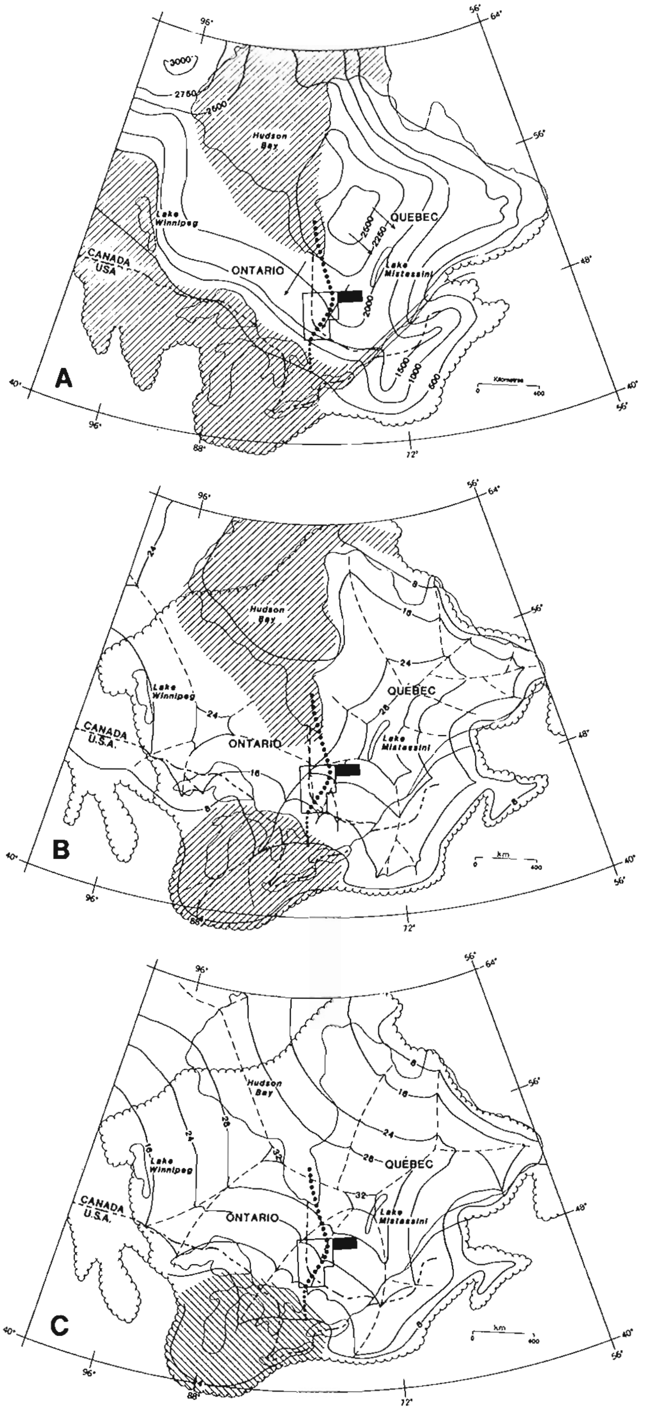

Figure 5. Configuration of the Labrador sector of the Laurentide Ice Sheet at the last glacial maximum. Deformable beds indicated by hatching.
A. Modified from Boulton et al. (1985); ice thickness is in metres.
B and C. Modified from Fisher et al. (1985); ice thickness is in hundreds of metres.

observed since 1985, is interpreted as supporting evidence that the cross-striated planes discussed have formed under an ice sheet subjected to changes in ice flow directions without intervening ice-free periods.

CONCLUSIONS

It is becoming more evident, with the accumulation of bedrock-inscribed ice flow indicators in recent years, that the hard, resistant crystalline bedrock of the Canadian Shield has preserved a reliable record of the various movements of the Laurentide Ice Sheet.

In the Matagami-Chapais area, the distribution and direction of striated surfaces associated with older ice flows and protected in sheltered positions from the erosive action of the last ice flows, suggest the existence of an ice divide between the two regions in Late Wisconsinan time. At final deglaciation, the configuration of the ice sheet shifted and generated the widespread southwest flow uniformly covering the area and visible in ice flow macrofeatures on air photographs. This interpretation is in agreement with our present knowledge on distribution and provenance of lithological indicators of glacial transport in this region. The high mobility of the ice sheet in Late Wisconsinan time as revealed by the older striations, gives some credibility to the patterns of crossing megalineations revealed by satellite images (Boulton and Clark, 1990) and attributed to movements of the Wisconsinan ice mass.

Further field work will be necessary to establish the extent of the former regional flows detected by striations in the Matagami-Chibougamau area, and their potential associations with former directions of glacial transport.

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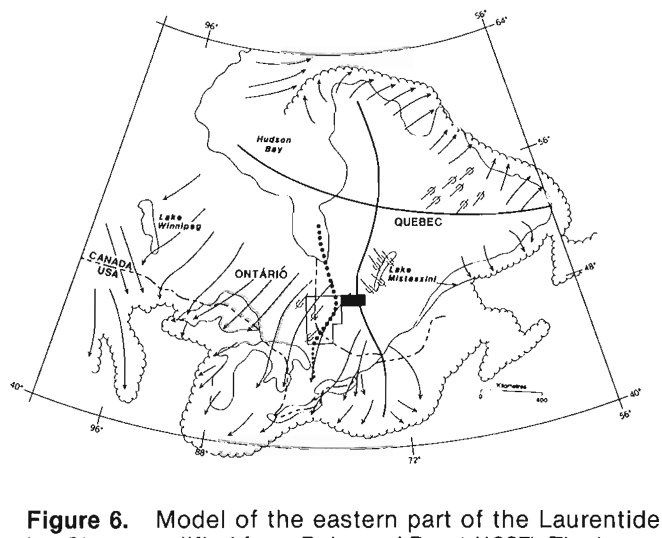

Figure 6. Model of the eastern part of the Laurentide Ice Sheet modified from Dyke and Prest (1987). The heavy lines represent the ice divides.

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Geology of the Whitehills-Tehek area, District of Keewatin: an Archean supracrustal belt with iron-formation-hosted gold mineralization in the central Churchill Province[1]

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Henderson, J.R., Henderson, M.N., Pryer, L.L., and Cresswell, R.G., Geology of the Whitehills-Tehek area, District of Keewatin: an Archean supracrustal belt with iron-formation-hosted gold mineralization in the central Churchill Province; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 149-156, 1991.

Abstract

Bedrock mapping in the Whitehills-Tehek area placed iron formation-hosted gold mineralization into a regional tectonic setting. In this part of the Churchill Structural Province the deformation is almost entirely late Archean in age. Complex structural development of the area includes early stratal disruption, possibly by low-angle faulting, that obscures the original stratigraphy. Gold mineralization occurs in iron sulphide-magnetite-quartz bearing iron formation in contact with chemically altered and foliated ultramafic rocks. Diamond drilling by an exploration company showed that highest gold values occur in nearly vertical beds of iron formation which form the hinge of a recumbent fold. Regional mapping aided by aeromagnetic vertical-gradient maps showed that the mineralized structure is a parasitic fold on the overturned limb of a regional north-vergent F_2 antiform.

Résumé

La cartographie géologique de la région de Whitehills-Tehek situe la minéralisation aurifère en formation ferrifère dans un contexte tectonique régional. Dans cette partie de la province structurale de Churchill, la déformation date presque entièrement de la fin de l'Archéen. L'histoire structurale complexe de cette région comprend tout d'abord la dislocation des strates, peut-être par faille subhorizontale, qui masque la stratigraphie primaire. La minéralisation aurifère se trouve dans une formation ferrifère à sulfure de fer, magnétite et quartz en contact avec des roches ultramafiques affectées par l'altération chimique et la foliation. Les forages au diamant effectués par une compagnie d'exploration indiquent que les valeurs en or les plus élevées s'observent dans des couches de la formation ferrifère presque verticales formant la charnière d'un pli couché. La cartographie régionale, réalisée à l'aide de cartes aéromagnétiques du gradient vertical, indique que la structure minéralisée consiste en un pli parasitique sur le flanc inverse d'un antiforme régional P_2 orienté vers le nord.

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INTRODUCTION

The Whitehills-Tehek lakes area is located in the District of Keewatin, north of the hamlet of Baker Lake (Fig. 1). The study area includes four adjoining portions of NTS 1:250 000 scale maps 56D (Baker Lake), 66A (Schultz Lake), 66H (Amer Lake), and 56E (Woodburn Lake), and was chosen for mapping in order to place gold mineralization, associated with iron formation in the Third Portage Lake area (Barham & Mudry, 1990), into a regional structural and stratigraphic context. Recent high resolution aeromagnetic total field and vertical magnetic gradient maps (Energy, Mines and Resources Canada, 1990a, b, c, d) provide some constraints on the location of formational contacts in areas of covered bedrock.

Acknowledgments

We thank Phil Mudry, Toby Hughes, Sandy Barham, Mark Balog, and other members of Comaplex Minerals' Meadowbank River exploration camp for sharing their knowledge of the geology of the Third Portage Lake diamond drilling area with us. Comaplex also are thanked for logistical support, including casual charter of their helicopter. Susan Fazakas served as a proficient assistant in the field and office. John Broome and Dennis Teskey provided coloured images of the aeromagnetic flight-line data. Subas Tella, Tom Wright, Tony LeCheminant, and Fred Chandler are gratefully acknowledged for reviewing the manuscript and suggesting numerous significant improvements to the text and figures.

Previous geological mapping

The Whitehills-Tehek area is included in the much larger region mapped by Wright (1967) and published at 1:1 000 000 scale. Geological mapping at 1:250 000 scale in the Whitehills-Tehek region was done by Donaldson (1966, Schultz Lake), Tella and Heywood (1983, Amer Lake), and Fraser (1988, Woodburn Lake). Geological maps at 1:50 000 scale that overlap with the mapping done in this study were completed by Taylor (1985, 66A/9), and Ashton (1988, 66H/1). In addition Nadeau (1981) mapped 56D/12, and Heywood and Schau (1981) and Schau et al. (1982) published sketches showing the geology of the Baker Lake area (56D).

Geological summary

Bedrock in the region northwest of Hudson Bay (Fig. 1) consists of Archean and Proterozoic supracrustal assemblages that are distinguished on the basis of cross-cutting relationships of deformation and plutonism. For example the early Proterozoic Penrhyn, Amer and Hurwitz groups rest unconformably on late Archean (Kenoran) orogenic granites, whereas the Archean Prince Albert and Woodburn groups are intruded by Kenoran granites.

Although the region lies within the Churchill Structural Province defined by early Proterozoic (Hudsonian) K-Ar dates obtained from various minerals, it is generally understood that much of the terrane consists of Archean rocks that have been reworked during the Hudsonian orogeny. However, details of the regional extent of Hudsonian deformation of Archean crust are unknown. LeCheminant et al. (1987) reported late Hudsonian U-Pb ages for an extensive suite of syndeformational calc-alkaline granites north of Wager shear zone, thus documenting widespread early Proterozoic tectonism north of the Wager and Amer shear zones. Other workers (Patterson, 1986) assumed that extensive Hudsonian tectonism had affected the region to the south beyond the Whitehills-Tehek area. Ashton (1988), though, obtained a late Kenoran age for granite intruding supracrustal rocks west of Tehek Lake. Our mapping determined that this granite and satellite dykes are not deformed, indicating that the Hudsonian orogeny had little effect on rocks of the Whitehills-Tehek area. The maximum extent of this pre-Hudsonian craton is defined by the deformed and metamorphosed early Proterozoic Amer Group to the north, Hurwitz Group to the south, and by the early Proterozoic emplacement of the Daly Bay Complex (Gordon, 1988) to the east. To the west 1.7-1.8 Ga igneous and sedimentary rocks occupy the Thelon and Baker Lake basins (Fig. 1) forming an overlap sequence on older tectonized Precambrian rocks. Separation of Hudsonian and Kenoran structural and metamorphic fabrics in future should better define the extent of un-reworked Archean crust in the Churchill Structural Province.

Geology of the Whitehills-Tehek Area

Based on previous mapping and geochronology (Ashton, 1988) it seems certain that the deformed supracrustal rocks in the Whitehills-Tehek area (Fig. 2) are Archean.

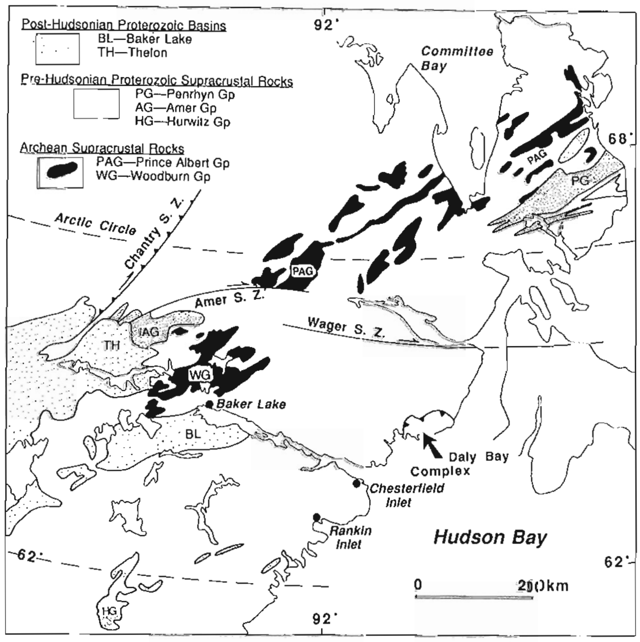

Figure 1. Geological map of the region northwest of Hudson Bay showing the distribution of Precambrian supracrustal rocks and some major shear zones. The Whitehills-Tehek area is centred about 50 km north of Baker Lake village (beneath the letters 'WG').

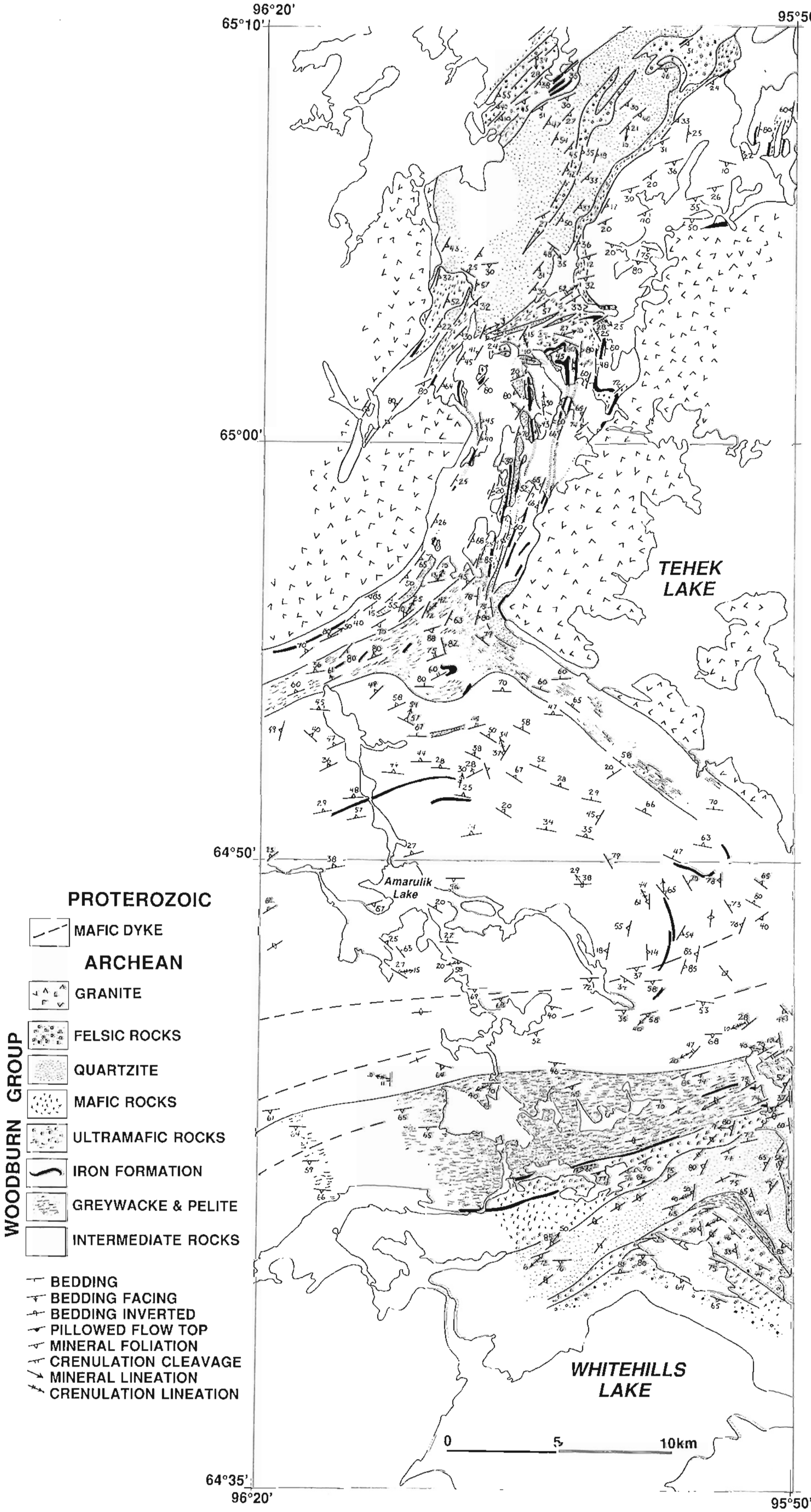

Figure 2. Geological map of the Whitehills-Tehek area.

Ashton (1988) determined a 2621 Ma U-Pb zircon age for unfoliated granite that intrudes the Woodburn group (informally named Woodburn Lake group by Ashton, 1981, 1982) near the lake shown in the northwest corner of the area (Fig. 2). Preliminary results of U-Pb dating of igneous zircons from dacite and detrital zircons from adjacent greywacke collected south of Amarulik Lake (Fig. 2) produced ages of about 2786 and 2770 Ma (H. Chapman personal communication, 1990), respectively, which correlate with a U-Pb zircon age of 2798 +24/-21 Ma from Woodburn Group dacite porphyry in the Amer Lake map area (Tella et al., 1985), thus confirming the Archean age of the Woodburn group.

Stratigraphy

Stratigraphy of the Woodburn group is uncertain because very few stratigraphic facing directions were found, few bedding measurements were obtained, and mappable units are disconformable and discontinuous, suggesting that the sequence was disrupted early in the structural history. Thrusting was invoked by Taylor (1985) to explain inconsistent stratigraphic relationships in the area north of Whitehills Lake, and by Ashton (1988) for the area northwest of Tehek Lake. Fraser (1988) concluded that in the Woodburn Lake map area, the Woodburn Lake group comprises a conformable succession consisting of a 'lower sequence' of metavolcanic rocks and volcanogenic metasedimentary rocks, and an 'upper sequence' of shallow-water-platform type metasedimentary rocks. A similar stratigraphic sequence was proposed by Ashton (1988), but with an unconformity between the metavolcanic-dominated lower sequence and the quartzite-dominated upper sequence.

In the present study it was not possible to confirm either the existence of an unconformity between the two sequences composing the Woodburn group, or the shallow-water depositional environment of the quartzite-dominated sequence. The south-dipping structural sequence defined by S_1 that is exposed between Amarulik Lake and Whitehills Lake (Fig. 2) represents our best estimate of the regional stratigraphy in the Whitehills-Tehek area. The sequence is listed in apparent stratigraphic order in the legend of Figure 3, and brief descriptions of the nine map units follow.

Intermediate rocks

This basal unit of the Woodburn group consists of massive, fine-grained, grey meta-andesite, dacite and interbedded iron formation occurring mainly around Amarulik Lake. Its base has not been recognized. Similar-looking intermediate volcaniclastic rocks are present within this map unit, but it is not clear to what extent these rocks correlate with the intermediate volcanic rocks.

Greywacke and pelite

Metagreywacke overlies intermediate rocks south of Amarulik Lake with apparent conformity. Metagreywacke is psammitic and thick bedded; in a few localities graded beds were recognized. Pelitic beds are rare. Black meta-argillite or slate overlies metagreywacke south of Amarulik Lake. Northwest of Tehek Lake metagreywacke appears to be absent from the succession, and iron formation occurs between intermediate rocks and ultramafic rocks.

Iron formation

Iron formation is found discontinuously above the greywacke and pelite unit, and beneath the unit of ultramafic rocks. Iron formation consists mainly of thin magnetite-rich and quartz-rich interbeds. In places iron-silicate-rich beds composed of chlorite, garnet and amphibole are found. Rarely, rusty-weathering magnetite-bearing beds containing pyrite and pyrrhotite were observed. Iron formation occurs in the Woodburn group within the underlying sequences of intermediate rocks and within the greywacke and pelite unit, and the overlying unit of ultramafic rocks. Stratigraphically, however, the association with ultramafic rocks is the most important. Notably, iron formation was never observed to occur interbedded with quartzite.

Ultramafic rocks

A thirty metre wide layer of spinifex-textured komatiite occurs above the unit of iron formation and below the mafic rocks located about five kilometres north of Whitehills Lake. This remarkably undeformed exposure was described by Taylor (1985, p. 8). Ultramafic rocks consisting mainly of chlorite-talc-serpentine schist and dolomite are most abundant in the region west and north of Tehek Lake. Northwest of the study area spinifex-textured komatiite occurrences have been well-documented by Ashton (1981, 1982, 1988) and Annesley (1981a, 1981b, 1989).

Mafic rocks

Pillowed metabasalts extensively overlie komatiite, iron formation and greywacke north of Whitehills Lake. This unit is now mainly a chlorite-actinolite-plagioclase schist. Similar mafic schist overlies quartzite north of Whitehills Lake, but it is not known whether these mafic rocks correlate with the unit of mafic rocks underlying quartzite (see below). Mafic schist forms an important component of the unit mapped as ultramafic rocks in the region northwest of Tehek Lake, but in this part of the map area (Fig. 2) the structural geometry is too complex to separate mafic and ultramafic rocks without additional large scale mapping.

Quartzite

North of Whitehills Lake, quartzite and quartz-muscovite schist occur between two units of mafic rocks, and north of Tehek Lake ultramafic schist appears to surround a large area mapped as quartzite. No evidence of unconformity was observed at the base of the quartzite formation, except for the varied lithologies it overlies in the area. The most likely explanation is that repetitions and omissions of quartzite in the apparent stratigraphic sequence are due to low-angle faulting during the earliest deformation.

Previous workers have noted the purity of quartzite in the Woodburn group, and have assigned it to a shallow-water platformal environment of deposition. However bedding was rarely identified in the Whitehills-Tehek area, and quartzite was seen to grade gradually into well-foliated quartz-muscovite schist and quartz-feldspar "porphyry", and in places quartzite was seen to contain more than fifty per cent vein quartz. The origin of quartzite is uncertain; some of it apparently was quartz-rich sediment, but the scarcity of primary fabrics discourages a unique assignment to its origin.

Felsic rocks

Felsic rocks occur as discontinuous lenses at several apparent stratigraphic positions throughout the study area. The largest masses occur north of Whitehills Lake where they overlie quartzite and mafic rocks; several smaller bodies were mapped in the area west of Tehek Lake. Felsic rocks include quartz-feldspar porphyry and "quartz-eye" schist; no distinction could be made regarding their origin as shallow intrusions or rhyo-dacite flows.

Granite

Two large masses of coarse-grained granular, biotite granite intrude the Woodburn group west of Tehek Lake. The granites are not penetratively foliated, but contain discrete foliation zones in places. The granites are similar in texture and composition, and may connect in the subsurface. Undeformed granite dykes locally intrude the Woodburn group near the granite contacts. The granites were intruded late in the deformational history of the Whitehills-Tehek area.

Mafic dykes

Two east-trending and undeformed mafic dykes up to several metres wide were observed south of Amarulik Lake. The dykes show a positive aeromagnetic signature which was used to extrapolate them along strike. The dykes are porphyritic and contain centimetre-wide white plagioclase phenocrysts in an aphanitic black matrix.

Structure

Structural fabric elements recognized in the field are bedding and pillowed flow tops (S_0), mineral foliation (S_1), mineral lineation (L_1), crenulation cleavage (S_2), and crenulation lineation (L_2). Axial planes and axes of mesoscale F_2 folds also were measured. No mesoscale F_1 folds were recognized.

The largest structure observed in the Whitehills-Tehek area is an elliptical dome defined by S_1 foliation dip-reversals around Amarulik Lake (Fig. 3). No mesoscale fabrics are associated with this fold. In the Third Portage Lake area, north-trending upright folds form a tight W-shaped synform between the late Archean batholiths cropping out to the east and west. North and south of Third Portage Lake the tight upright folds in the S_1 foliation disappear. Although the dome and the W-shaped synform are nearly orthogonal it is not possible to determine their relative sequence of formation because

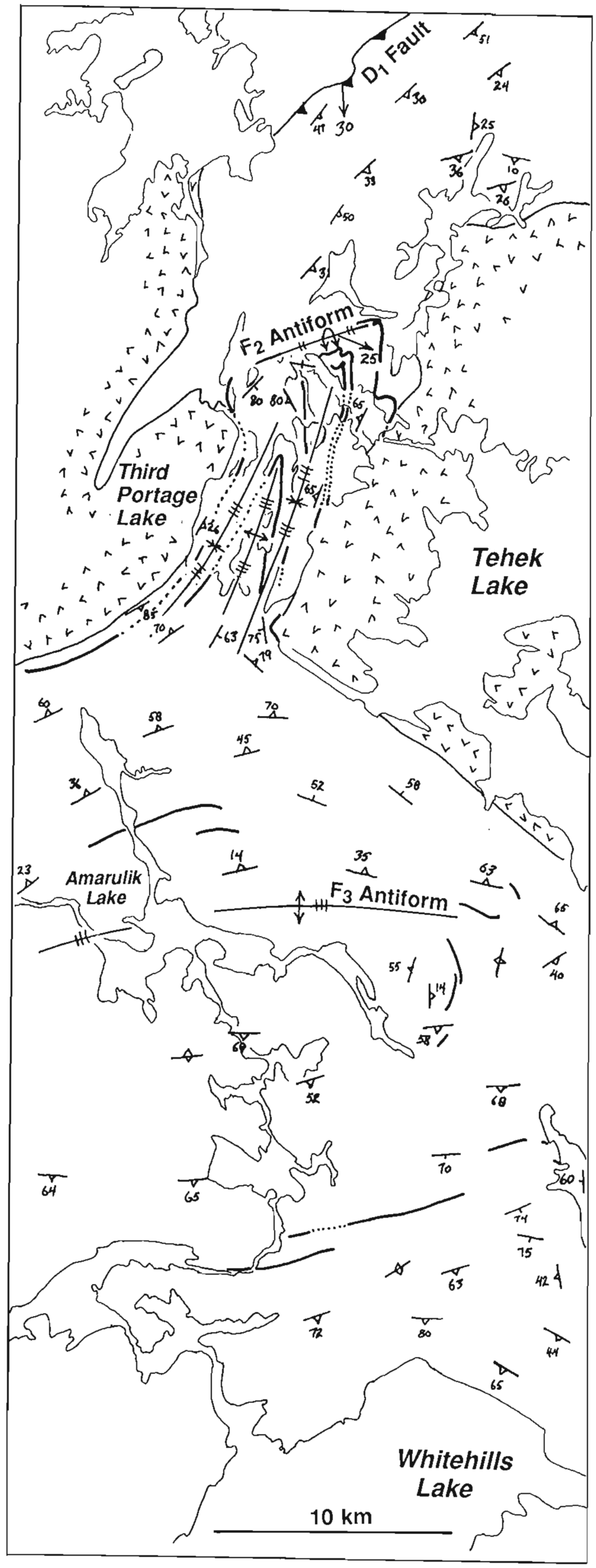

Figure 3. Structural map of the Whitehills-Tehek area showing F_2 and F_3 fold axial-surface traces, and an inferred D_1 low-angle fault (barbs on upper plate). Dotted lines in Third Portage Lake link observed Iron formation units coincident with aeromagnetic anomalies.

they lack associated mesoscopic fabrics that would show cross-cutting relations. For this reason they are all labelled F_3 on the structural map (Fig. 3).

Older structures recognized are reclined to recumbent mesoscale folds in the main fabric, S_1 (Fig. 4), and related microscale subhorizontal S_2 crenulations of S_1 (Fig. 5). Axes of F_2 folds and crenulations plunge 5-30 degrees northerly from the Third Portage drilling area (Fig. 6) to the north end of Third Portage Lake. East of Third Portage Lake, axes of F_2 folds plunge east at 5-20 degrees and the F_2 folds are consistently south-vergent. The mesoscale F_2 folds are north-vergent in the easternmost north-trending belt of iron formation (Figs. 3, 6); this vergence change indicates the approximate location of the axial-surface trace of the kilometre-scale, east-southeast plunging, reclined F_2 fold shown in Figure 3.

Figure 4. Field photograph (view east) showing east-plunging, south-vergent, reclined F_2 folds in the quartzite. Located on the north shore of Third Portage Lake (Fig. 4)

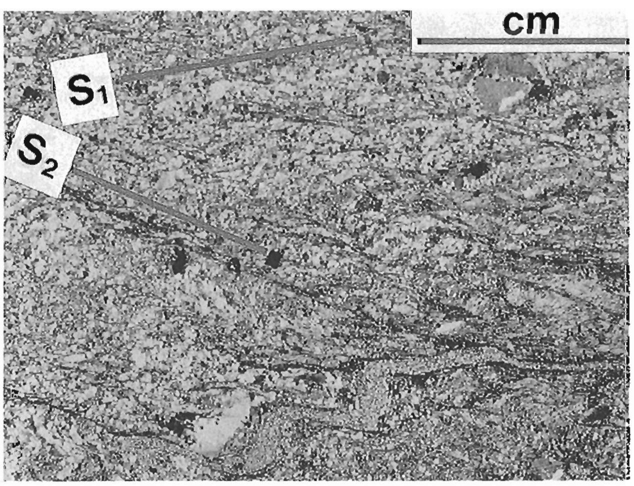

Figure 5. Photomicrograph showing S_1 foliation and S_2 crenulation cleavage developed in quartz-muscovite schist composing the quartzite unit.

The oldest structures observed in the Whitehills-Tehek area are related to formation of the pervasive S_1 cleavage in the supracrustal rocks. North of Third Portage Lake, the effects of younger deformations seem to be minimal, and S_1 dips southeast at less than 40 degrees (Fig. 3). Sparse bedding observations indicate that S_0 dips in the same direction as S_1, but more steeply; this suggests northwest subhorizontal transport of the supracrustal rocks during D_1, and possible stratigraphic inversion (see also Ashton, 1982, p. 155).

Metamorphism

Widespread carbonate alteration of ultramafic rocks, the presence of chlorite, biotite and muscovite in pelitic and intermediate igneous rocks, as well as rare occurrence of garnet in silicate facies iron formation indicate that large fluxes of H_2O-CO_2 fluids accompanied greenschist facies regional metamorphism. In zones of gold mineralization more than 10 per cent sulphide replacement of magnetite has taken place, suggesting that sulphur and small quantities of gold also were present in the metamorphic fluids. No contact metamorphism related to granite emplacement was seen, and the regional metamorphic grade appears to be the same throughout the Whitehills-Tehek area; it should be noted, however, that kyanite and muscovite occur in quartzite north of Third Portage Lake (Ashton, 1988; S. Tella, personal communication, 1990), although adjacent ultramafic rocks are chlorite-talc schists, and garnet is not present in iron silicate-rich iron formation.

Economic geology

Gold occurrences in the area of Third Portage Lake (Fig. 6) are spatially related to sulphide-bearing magnetic iron formation, especially where iron formation is adjacent to chlorite-talc schist. Highest concentration of gold occurs in the vertical hinge region of an F_2 fold, where vuggy quartz-pyrite-gold mineralization overprints earlier quartz-pyrrhotite-gold mineralization in iron formation (Barham and Mudry, 1990). Introduction of gold-bearing fluids apparently occurred during both F_1 and F_2 by hydraulic fracturing of brittle magnetite-quartz iron formation. Most favourable areas for prospecting appear to be where F_2 fold hinges occur in iron formation adjacent to ultramafic schist, and where there has been more than 10 per cent sulphide replacement of magnetite.

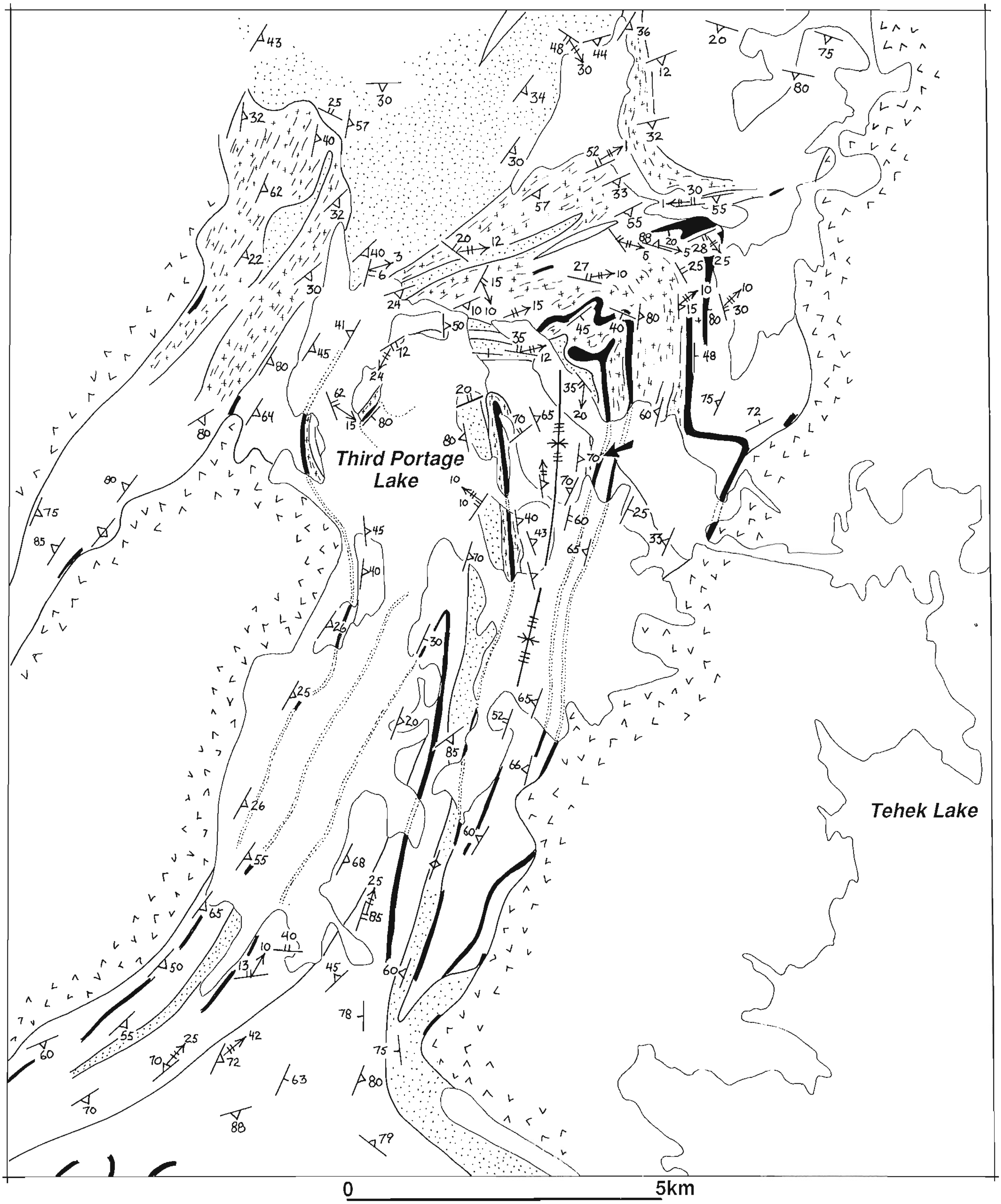

Figure 6. Geological map of the Third Portage Lake area. The bold arrow east of Third Portage Lake points to the gold-bearing band of iron formation (located west of the F_3 synform axial-surface trace) described as the Third Portage Prospect (Barham & Mudry, 1990). Other symbols are the same as in Figure 3.

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New geological developments in the internal zone of Wopmay orogen, District of Mackenzie

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Hildebrand, R.S., Paul, D., Pietikäinen, P., Hoffman, P.F., Bowring, S.A., and Housh, T., New geological developments in the internal zone of Wopmay orogen, District of Mackenzie; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 157-164, 1991.

Abstract

The internal zone of Wopmay orogen comprises a large klippe of 2.0-2.55 Ga gneissic rocks, termed Bent gneiss, and 1.90 Ga volcanic and sedimentary cover of the Grant-Akaitcho groups. Rocks of the Grant-Akaitcho groups probably formed near the leading edge of Hottah terrane just prior to the initiation of the Calderian orogeny. The gneissic basement and its cover were tectonically interleaved along west-vergent thrusts during the earliest stages of the Calderian orogeny. They were subsequently emplaced upon Slave craton. Lineations in the klippe and the autochthon indicate that the emplacement of the klippe was oblique dextral. Final Calderian compression resulted in the formation of large northerly-trending folds.

Résumé

La zone interne de l'orogène de Wopmay comprend une large klippe de roches gneissiques de 2,0-2,55 Ga, appelée gneiss de Bent, et une couverture volcanique et sédimentaire de 1,90 Ga des groupes de Grant-Akaitcho. Les roches des groupes de Grant-Akaitcho se sont probablement formées près de la bordure frontale du terrane de Hottah juste avant que s'amorçe l'orogenèse caldérienne. Le socle gneissique et les roches sus-jacentes ont été tectoniquement interfoliés le long de chevauchements à vergence ouest durant les toutes permières étapes de l'orogenèse caldérienne. Ils ont par la suite été mis en place sur le craton des Esclaves. Selon les linéations observées dans la klippe et les roches autochtones, la mise en place de la klippe découle d'un mouvement dextre oblique. La compression caldérienne finale a provoqué la formation d'immenses plis à direction nord.

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INTRODUCTION

During the summer of 1990, two months were spent mapping rocks in the Calder River (86F) map area. The map area includes: (1) Archean rocks of Slave province; (2) rocks of Hottah terrane, interpreted as an exotic terrane that collided with the western margin of Slave craton during the Calderian orogeny (Hildebrand et. al., 1983); (3) rocks of the Akaitcho and Grant groups, formerly interpreted as part of a rift succession deposited on the western margin of Slave craton prior to the Calderian orogeny (Easton, 1980; 1981; Hoffman and Bowring, 1984; but see Bowring and Grotzinger, 1988); (4) plutons of the syn-collisional Hepburn batholith (Lalonde, 1986); and (5) rocks of Great Bear magmatic zone, a post-collisional magmatic arc (Hildebrand et. al., 1986). Previous investigations within the Calder River area are listed in Hildebrand et. al. (1987). The major aims of the project are to unravel the geological evolution of the central Great Bear magmatic zone, the medial zone (Hildebrand et. al., 1990), and the western part of the internal zone of Wopmay orogen. The purpose of this report is to summarize some of the geological results of the year's fieldwork, with particular emphasis on the stratigraphy and structure of the internal zone where nearly all of the work was concentrated. The principal results of the fieldwork are summarized below.

1. Within the medial zone, on the east limb of Eyston anticline (Fig. 1), Archean gneisses and unconformably overlying metasedimentary rocks of the Odjick Formation are exposed beneath an easterly-vergent thrust fault. We have dated rocks above the thrust in three places. In one case orthogneisses yield a U-Pb zircon age of about 2.0 Ga (Hildebrand, et. al., 1990) and the other two samples are about 2.5 Ga. The assemblage of gneisses above the thrust is termed Bent gneiss. It forms the trailing edge of a klippe (Turmoil klippe) that cores a regional syncline, known as Robb River syncline (Fig. 1), between the western limb of Exmouth anticline and Eyston anticline.

2. Within Turmoil klippe, volcanic and sedimentary rocks of the Akaitcho and Grant groups sit unconformably upon Bent gneiss. Sequences of rocks identical to those of the type Grant Group (Easton, 1981) within the medial zone were traced laterally into rocks of the Akaitcho Group, indicating that the two groups were deposited within the same basin. The basin was apparently located on the leading edge of Hottah terrane.

3. Although the internal structure of Turmoil klippe is complex and incompletely resolved, we have identified a number of westerly-vergent thrust faults along which the basement and cover are interleaved. The faults were folded about northwesterly-trending axes (F_2) during emplacement of Turmoil klippe; subsequently refolded about northerly-trending axes (F_3) during terminal stages of Calderian shortening; and still later, broadly-folded about an east-northeast trend (F_4) by an apparently unrelated deformational event.

4. Removal of the effects of F_3 and F_4 folds reveals subhorizontal, northeasterly-trending stretching lineations within Proterozoic rocks of the autochthon of Exmouth anticline and rocks of the Turmoil klippe. The lineation was likely generated as Turmoil klippe was transported northeastward over Slave craton and suggests a dextral transpressive emplacement.

5. Some possible Hepburn plutons within the thrust stack have S_1 Calderian fabrics, and others, such as the Robb, are little-deformed. This, coupled with isotopic evidence which indicates that most Hepburn plutons were derived from west of Slave craton (Bowring and Podosek, 1989; Housh et. al., 1989), suggest that the westward-vergent thrusts, F_1, and S_1 formed prior to the emplacement of Turmoil klippe on Slave craton.

EYSTON ANTICLINE: POSSIBLE AUTOCHTHON WITHIN THE MEDIAL ZONE

St-Onge et. al. (1984) recognized that within the internal zone, Archean rocks and their autochthonous Proterozoic sedimentary cover are exposed beneath a basal décollement in a series of large-scale northerly-trending folds (Exmouth anticline, Acasta syncline, Scotstoun anticline, and Carousel anticline-Figure 1). They also argued that movement on the décollement predated the folding because the fault is folded. King (1986) and St-Onge and King (1987) argued that because the fold axes of the large-scale folds are coaxial with Calderian folds, and because the traces of mineral isograds both transect and are folded by the folds, that the folding was a late result of progressive deformation during the Calderian orogeny. Consequently, they divided the Calderian orogeny into two stages: D_1, represented by northerly-trending thrusts and related folds; and D_2, represented by the relatively large-scale basement-involved folds.

In the medial zone, on the western limb of a large anticline, Hildebrand et. al. (1990) discovered relationships similar to those exposed on the large-scale folds of western Slave craton, in that an easterly-vergent thrust fault structurally superposes high-grade gneiss on Proterozoic metasedimentary rocks which themselves sit unconformably on Archean gneiss. During the 1990 season we traced the triad another 15 km northward (Fig. 1).

The section of Proterozoic metasedimentary rocks exposed beneath the thrust fault contains quartzite, garnet amphibolite, carbonate, conglomerate, and semipelite. The presence of a 10 m interval of white quartzite suggests that the section might be correlative with the lower Odjick Formation. Overall, the lithologies are similar to those present beneath the décollement on the western side of Exmouth anticline. The metasedimentary rocks are folded about axial planes that dip easterly at 50°-70°, the same inclination as bedding, the overlying thrust fault and gneissic layering in the rocks above it. Additionally, the folds have an asymmetry suggesting easterly-directed topside-over-bottom movement.

Whereas the rocks of the structurally overlying plate are continuous eastward to Exmouth anticline (Fig. 1), the Archean and its cover may be continuous with Slave craton. If they are structurally continuous, beneath Robb River syncline, with the autochthonous core of Exmouth anticline, then Eyston anticline represents the known western limit of Slave craton.

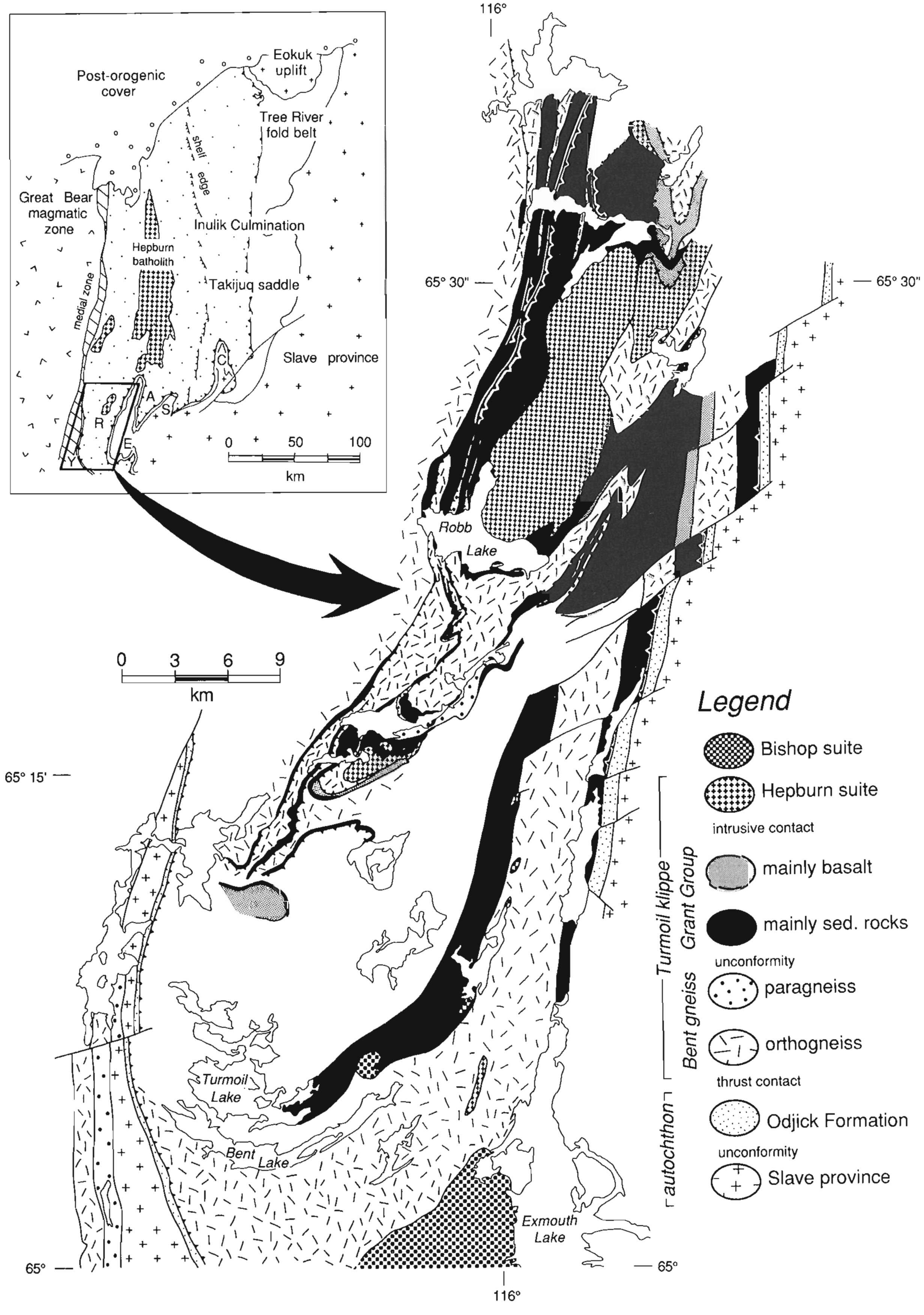

Figure 1. Geological map of the southeastern Calder River map area. Inset map after Hoffman et. al. (1988) showing location of map and regional geology. A = Acasta syncline; E = Exmouth anticline; C = Carousel anticline; S = Scotstoun anticline; Y = Eyston anticline; R = Robb River syncline.

TURMOIL KLIPPE

Structurally overlying the Proterozoic metasedimentary rocks of Eyston anticline are gneissic rocks previously interpreted, on the basis of 2.0 Ga U-Pb zircon ages, to be part of Hottah terrane (Hildebrand et. al., 1990); however, preliminary geochronological results on samples collected in 1990 indicate that parts of the complex have U-Pb zircon ages around 2.5 Ga. The rocks occupy the core of Robb River syncline (Fig. 1) and, as they are structurally isolated from the main part of Hottah terrane, constitute a klippe, here termed Turmoil klippe after outcrops in the Turmoil Lake area (Fig. 1). The gneisses within Turmoil klippe are named Bent gneiss after superb outcrops in the Bent Lake area (Fig. 1). Overall, basement and cover units as mapped during 1990 do not correspond to those of St-Onge et. al. (in press)

Bent gneiss comprises mainly tonalitic orthogneiss with subordinate amounts of granitic to dioritic orthogneiss, paragneiss, and amphibolite. The tonalitic gneisses are equigranular to plagioclase-porphyroclastic, biotite-bearing rocks. They appear to be intruded by a variety of porphyritic and even-grained, but gneissic, biotite and hornblende-biotite granitoid rocks ranging from diorite to granite. In general, the granitic rocks are porphyritic and the dioritic rocks more even-grained. Metasedimentary rocks occur mainly as metre to ten-metre-wide enclaves in the tonalitic gneiss. Amphibolite occurs as lozenges within the gneisses and probably represents torn-apart mafic dykes, although in some cases there are so many lozenges and blobs of amphibolite that magma mingling is a possibility. Additionally, there are large coarse grained meta-gabbroic sills, some of which are unconformably overlain by sedimentary rocks of Grant Group. The sills are heterogeneously strained with 0.5-2.0 cm wide zones of high strain.

Supracrustal rocks

Easton (personal communication, 1982) thought that within the Hepburn Lake map area, rocks of Akaitcho Group sat unconformably on granitic basement. The granites were subsequently dated (Bowring, 1984) to be about 2.51 Ga and King et. al. (1987) subsequently confirmed the unconformable relationship. During the 1990 field season we discovered that rocks of Akaitcho and Grant groups sit unconformably on Bent gneiss (Fig. 2). This is a major result of the summer's work because it further defines the type and age of basement to these interesting, but problematic, rocks. It is clear from Nd and Pb isotopes (Bowring and Podosek, 1989; Housh et. al., 1989) and dating of ash-beds within Kilohigok basin (Bowring and Grotzinger, 1988) that rocks of the two groups are exotic with respect to Coronation margin.

Rocks of Akaitcho Group were mapped in the northern and eastern part of the area, where they are continuous with those of the type area (Easton, 1980) and a distinctive sequence of rocks, termed Grant Group (Easton, 1981), was mapped in the southern and western part of the internal zone. However, rocks of the two groups are present within the same thrust slices and are lateral facies equivalents. For simplicity we use the term Grant Group to refer to the 1.90 Ga (Bowring, 1984) cover sequence that sits unconformably on Bent gneiss.

Grant Group

The unconformity with Bent gneiss is well-exposed in numerous places. In the southern part of the area a few metres of pyritic semipelite typically sit directly on Bent gneiss or are separated from it by 10-15 cm of grus. Relief on the unconformity is typically less than 1 m. Deformation precludes any accurate estimate of thicknesses for stratigraphic units and top determinations are very sparse. However, abundant outcrops of the basal unconformity allowed us to determine local facing directions. Overlying the semipelite are 3-20 m of grey to buff weathering carbonate. The carbonate preserves tight reclined folds which are difficult to see in other lithologies. In general, the folds are asymmetric with a sense of topside-over-bottom movement to the west. Above the carbonate are pillow basalts or medium-grained psammites. The metabasalts are mostly aphyric and typically strongly deformed, although nearly pristine pillows occur in a few places. The metabasalts are intruded by abundant metagabbroic sills.

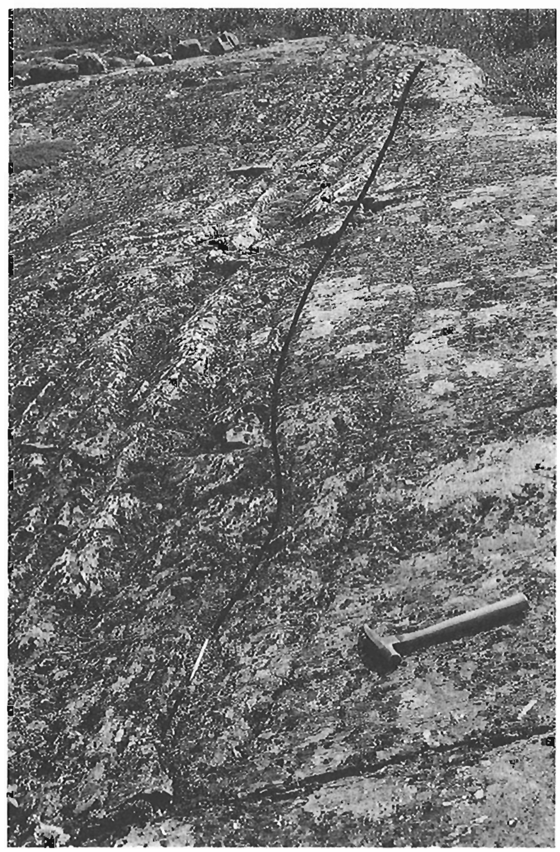

Figure 2. Photograph showing well-bedded sedimentary rocks of the Grant Group sitting unconformably on Bent gneiss south of Robb Lake. Hammer handle lies in the plane of foliation within Bent gneiss. Trace of unconformity shown by black line.

The psammites range from quartzite to arkose and are exposed in only one thrust slice. They are well-bedded rocks with bed thicknesses ranging from 15 cm to 1 m.

Farther north, the sections are different in that they contain greater percentages of sedimentary and siliceous volcanic rocks. Immediately above an unconformity of little relief is a thick fining-upward section of sandstone. Typically the sandstone is overlain by rhyolitic or basaltic lava flows. The entire sequence is intruded by gabbroic sills and siliceous porphyries. The gabbroic sills are coarse grained rocks that typically form high-standing ridges.

Locally, such as within a thin strip along the western side of Exmouth anticline (Fig. 1), possible rocks of Grant Group are more metamorphosed than elsewhere. Sillimanite porphyroblasts and/or granitic melt pods are common. As the contact with Bent gneiss was not exposed we were unable to be sure that rocks within the strip are cover but northeast of Robb Lake metasedimentary rocks of the Grant Group can be traced to higher grades toward the northeast. Accordingly, we tentatively include rocks of the strip within Grant Group.

STRUCTURE

The oldest post-depositional structures seen in the internal zone are brittle-ductile thrusts within Turmoil klippe. The thrust-sheets are thin, very continuous along strike, and contain both crystalline basement and cover. Locally, ramps (Fig. 3) with higher cut-off angles occur. The thrust faults truncate penetrative fabrics in the basement (Fig. 2) and have no apparent associated mineral lineation. Reclined, tight to isoclinal folds (F_1) of bedding and a sub-horizontal axial-planar foliation (S_1) are particularly well-developed in the supracrustal rocks. Many of the F_1 folds are outcrop-scale and are distinctly asymmetrical, with an asymmetry suggesting topside-over-bottom movement to the west. Thus, we infer a westward vergence for the thrust faults. The occurrence of thin thrust sheets containing both crystalline basement and cover is interesting but by no means unique, for such structures are known from the Scandinavian Caledonides (Björklund, 1985) and the Cape Smith belt (Lucas, 1989).

A southwest-trending mineral lineation (L_2) is well-developed in allochthonous and autochthonous cover rocks along the west side of Exmouth anticline. It plunges 30°-60° on steeply-dipping to vertical bedding and cleavage planes. Where, bedding and/or cleavage dip eastward, such as on the western limb of Robb River syncline, the lineation trends and plunges southeast. The brittle thrusts and S_1 are tightly to isoclinally folded (F_2) about axes that trend northeast and plunge moderately, perpendicular to L_2 (Fig. 3). A well-developed crenulation cleavage (S_2), axial planar to F_2, is present in mica-rich units.

The temporal relationship between L_2 and F_2 was not satisfactorily resolved in the field but as the two linear elements are at a high angle to one another it is likely that the two were generated by the same deformational event. Furthermore, as L_2 are clearly folded by F_3 folds and plot on a great circle (Fig. 4), they must have been dispersed from a discrete cluster by the F_3 folds. This indicates that they were not refolded by F_2, supporting the argument that F_2 and L_2 were generated during the same deformational event. When the effects of F_3 folds are removed the lineation is subhorizontal and northeasterly-trending on the autochthon of Exmouth anticline; therefore, the original orientation of the cluster was probably the same. This suggests that emplacement of Turmoil klippe onto Slave province was likely dextral oblique.

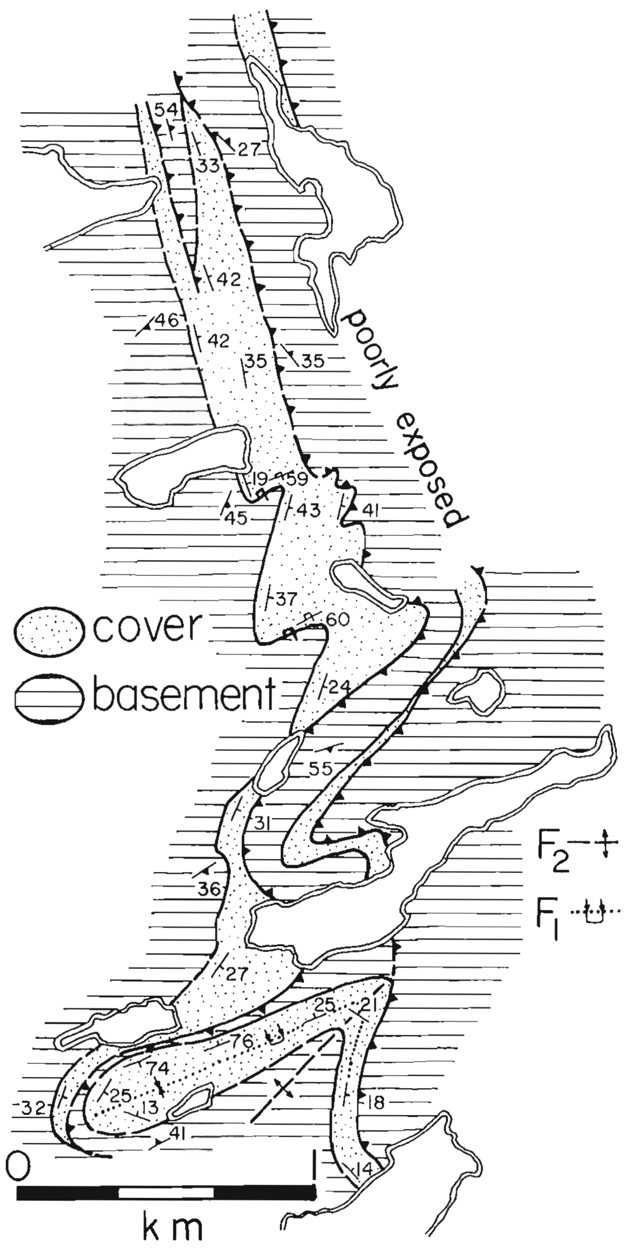

Figure 3. Detailed geological map showing geology south of west end of Robb Lake. Note the thin imbricate thrust sheets involving Bent gneiss and Grant Group cover. One F_1 and one F_2 fold axis are indicated.

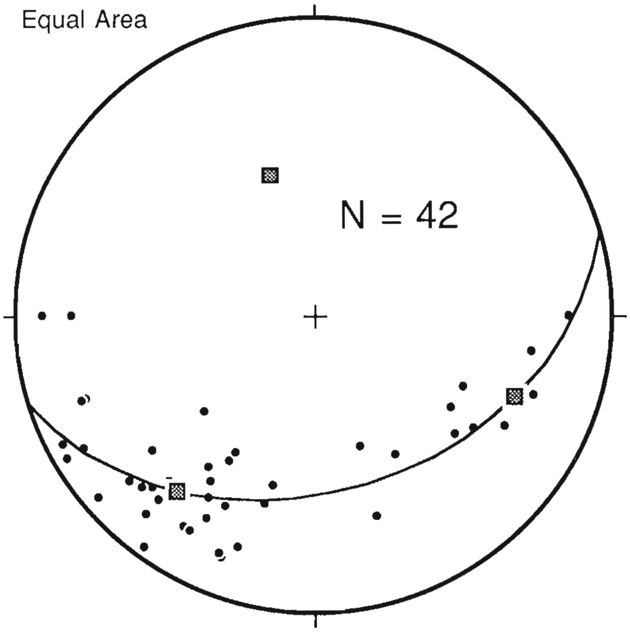

Figure 4. Equal area lower hemisphere stereonet plot of L_2 lineations collected from Robb River syncline.

All of the above structures are folded by northerly-trending folds (F_3) that involve Archean basement of the autochthon. They are thick-skinned folds interpreted to have been generated during the latest stages of the Calderian orogeny (King, 1986; Hoffman et. al., 1988). Examples include Exmouth anticline and Robb River syncline. They have wavelengths of 10-30 km and amplitudes of 10-15 km. The lower hemisphere plot of L_2 (Figure 4), collected from Robb River syncline north of the west-southwest trending axis of even younger F_4 folds, clearly plot on a great circle, indicating that F_3 folds are similar in their geometrical form (Ramsay, 1967). This makes any estimate of the amount of shortening impossible as constant bed length cannot be assumed. The indicated axis for the F_3 fold is steeper and more westerly than expected from the trend of contacts. This is probably because the lineations used for the plot were collected on the northern limb of an F_4 anticline that plunges westerly, thus, slightly rotating and steepening the axis. Counterclockwise rotation by right-lateral faults of the area could also account for the rotation, but would not cause the steepening of the axis.

HEPBURN INTRUSIVE SUITE

We mapped several plutons that are possibly members of Hepburn intrusive suite (Lalonde, 1986) and they range from strongly deformed to non-deformed. The most deformed, and probably the oldest, plutons are protomylonitic megacrystic granites containing potassium feldspar phenocrysts 5 to 10 cm long. Biotite is the sole ferromagnesian mineral. At least one of the deformed granitic plutons appears to intrude Bent gneiss, but the pluton is not dated.

A large pluton located on the north side of Robb Lake is a non-deformed to slightly-foliated biotite granite. It also contains 5-10 cm long potassium feldspar phenocrysts and is notable because it intrudes deformed Hepburn plutons, Bent gneiss with a folded Calderian S_1 fabric, and rocks of Grant Group. Because isotopic data indicate that the magmas for Hepburn intrusive suite were generated through interaction of mantle-derived basalt with crustal melts derived from isotopically homogeneous crust in the age range 2.0-2.4 Ga, not the isotopically heterogeneous Archean of Slave craton (Bowring and Podosek, 1989; Housh et. al., 1989), and because no member of the suite is known to intrude autochthonous rocks of Slave craton despite careful mapping (St-Onge et. al., 1984), it is generally assumed that the plutons were emplaced outboard of the western edge of Slave craton and transported on top of it. Accordingly, the westward-vergent thrusts, F_1, and S_1 formed prior to the intrusion of the Robb pluton, and were subsequently transported onto Slave craton, as was the Robb pluton, within Turmoil allochthon.

DISCUSSION

A first order problem that arises from the field work and preliminary U-Pb zircon dating is the significance of the circa 2.5 Ga rocks within Bent gneiss. There are now a large number of U-Pb zircon ages in the internal zone that are between 2.5 and 2.58 Ga (Bowring, 1984; unpublished data). They form a discrete cluster of ages that is younger than most known rocks of Slave province (Hoffman, 1989). As they are overlain by rocks dated at 1.90 Ga they were unlikely to have been in contact with Slave craton at that time for passive margin rocks of Coronation Supergroup were being deposited then. It is interesting that west of the medial zone, Hottah terrane comprises mostly rocks in the age range 2.0-1.9 Ga with no rocks as old as 2.5 Ga, whereas Turmoil klippe may be dominated by rocks about 2.5 Ga and contain few rocks in the age range 2.0-1.9 Ga. The obvious question is whether rocks of Turmoil klippe are part of Hottah terrane, were once part of Slave craton, or represent something entirely different such as a composite terrane. Whatever their ultimate origin, it is likely that rocks of Turmoil klippe were not in contact with Coronation margin between 1.967-1.90 Ga and probably were part of Hottah terrane during the Calderian orogeny.

Given that rocks of Grant Group are not products of initial rifting on Slave craton, what is their tectonic setting? We know that the upper parts of the sequence erupted just prior to the Calderian orogeny at about 1.90 Ga, but do not know the age of initiation of magmatism or sedimentation. Since the sequence constitutes the easternmost parts of Hottah plate, and is not compositionally like an arc sequence, it is reasonable to assume that it was deposited trenchward of any arc on that plate. If so, rocks of Grant Group were deposited near the leading edge of Hottah terrane. Since forearc regions are generally considered to be regions of low heat flow, a mechanism is needed to account for the magmatism, not only of Grant Group, but also of Hepburn intrusive suite (Hoffman et. al., 1988), which starts as Grant magmatism wanes (Bowring, 1984). Two modern analogues exist. Ridge subduction (DeLong and Fox, 1977), particularly in a small basin such as in the Woodlark Basin, where subduction of the Woodlark spreading ridge beneath the New Georgian forearc causes abundant magmatism, yet just precedes collision of the Australia-India plate with the Solomon arc (Perfit et. al., 1987; Johnson et. al., 1987). In fact, many cases of forearc magmatism generated by ridge subduction are known and include cases in Japan (Miyake, 1985; Hibbard and Karig, 1990), the Aleutians (Hill et. al., 1981), and Chile (Forsyth and Nelson, 1985).

The second analogue generates its magmatism by asthenospheric upwelling concomitant with roll-back of the lower plate such as in the Marianas, Japan and New Zealand. Asthenospheric upwelling may drive the roll-back (Tatsumi et. al., 1989) or be a passive response to roll-back induced by subduction of progressively older, hence denser, oceanic lithosphere. Whatever mechanism is invoked to generate the Grant-Hepburn magmatism it apparently occurred near the leading edge of Hottah terrane just prior to the start of the Calderian orogeny.

EVOLUTION

The overall evolution of the internal zone of Wopmay orogen is as follows.

1. Eruption and deposition of Grant Group on Bent gneiss. Based on distribution and comparison of chemical analyses the most likely tectonic setting was a former forearc region. The intense magmatism and subsidence in a normally cold forearc region may have been caused by ridge subduction as in the Woodlark Basin and/or aesthenospheric upwelling due to slab roll-back as in the Marianas.

2. Intrusion of earliest plutons of Hepburn intrusive suite, coeval with westerly-vergent thrusting involving Bent gneiss and rocks of Grant Group. Formation of F_1 and S_1. The thrusts may have formed as back thrusts related to ramping of Turmoil allochthon over Slave cratonic margin.

4. Northeastward transport of Turmoil klippe onto Slave craton. Formation of F_2 folds and L_2 lineation.

5. Formation of F_3 folds during terminal phase of Calderian orogeny.

ACKNOWLEDGMENTS

We owe the results of the summer to Aletha Buschman and John Vezina who were enthusiastic and knowledgable field assistants. We also thank Rod Stone for expediting, Latham Island Airways and Canadian Helicopters for safe and efficient transportation, and the Polar Continental Shelf Project for providing some helicopter support. Marc St-Onge critically reviewed the manuscript.

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Prehnite-pumpellyite to amphibolite facies metamorphism near Flin Flon, Manitoba

Scott Digel[1] and T.M. Gordon[1]

Digel, S. and Gordon, T.M., Prehnite-pumpellyite to amphibolite facies metamorphism near Flin Flon, Manitoba; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 165-172. 1991.

Abstract

The Flin Flon area is located within the Flin Flon domain of the Trans-Hudson orogen. The southern Flin Flon domain consists largely of volcanic and sedimentary rocks metamorphosed at prehnite-pumpellyite facies in the south to lower amphibolite facies in the north. The northern Flin Flon domain is dominated by sediments of greywacke composition metamorphosed at lower amphibolite facies and higher.

Deformation in the lower-grade southern Flin Flon domain is dominantly brittle. The Flin Flon area is subdivided into distinct lithologic blocks, each bounded by postmetamorphic faults. The most complete metamorphic sequence is preserved in the Hook Lake block which is relatively free of large intrusions. Changes in regional metamorphic mineral assemblages are obscured by: earlier contact metamorphism associated with mafic dykes and felsic plugs and dykes; later brittle faulting; and lithologies containing non-diagnostic mineral assemblages.

Résumé

La région de Flin lon est situé au sein du domaine de Flin Flon dans l'orogène trans-hudsonien. Le sud du domaine de Flin Flon est composé en grande partie de roches volcaniques et sédimentaires métamorphisées du faciès des prehnites-pumpellyites dans le sud au faciès inférieur des amphibolites dans le nord. Le nord du domaine de Flin Flon est caractérisé par des sédiments de grauwacke métamorphisés du faciès inférieur des amphibolites à des degrés de métamorphisme plus élévés.

La déformation dans le sud du domaine de Flin Flon où le degré de métamorphisme a été plus faible est surtout de type cassant. La région de Flin Flon est subdivisée en blocs lithologiques distincts, chacun limité par des failles post-métamorphiques. La séquence métamorphique la plus complète se trouve dans le bloc de Hook Lake dans lequel on ne trouve à peu près pas de grandes intrusions. Les modifications dans les associations minérales métamorphiques régionales sont masquées par : un métamorphisme de contact précoce associé à des dykes mafiques et des bouchons et dykes felsiques, la formation de failles cassantes tardives, et des lithologies contenant des associations minérales non révélatrices.

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INTRODUCTION

The Trans-Hudson orogen is a 500 km wide collage of arc-related volcanic, sedimentary and plutonic rocks which formed and coalesced in the Early Proterozoic. In Saskatchewan and Manitoba, it has been subdivided into several distinct lithotectonic domains (Bickford et al., 1990). The northern boundary of the Flin Flon domain (Fig. 1) has been extended to include amphibolite facies rocks thought to be metamorphic equivalents of rocks of the Flin Flon domain (K.E. Ashton, pers.comm., 1990). The Flin Flon area is located within the Flin Flon domain and is approximately centred on the town of Flin Flon, Manitoba at 54°45′N latitude 100°45′W longitude (Fig.1).

The Flin Flon region has a long history of mineral production and geological exploration. Much of the Flin Flon area was mapped at a scale of 1:20 000 by Bailes and Syme (1989) and Syme (1986, 1987, 1988) and they provide a discussion of the history of geological investigations in the area. They have also established the stratigraphic and structural framework in which the geology of this region is discussed.

Digel and Gordon (1990) reported that mineral assemblages diagnostic of regional metamorphism from prehnite-pumpellyite to lower amphibolite facies exist in the study area. Field work was conducted in the summer of 1990 with three objectives, to:

a) examine in detail the transition from prehnite-pumpellyite to greenschist facies
b) investigate the extent and effects of contact metamorphism and
c) document the metamorphic and structural transition from the southern to the northern part of the Flin Flon domain.

Preliminary analysis of data and samples collected in 1989 and 1990 has addressed each of these objectives. The prehnite-pumpellyite to greenschist transition is complicated by earlier contact metamorphism, postmetamorphic faulting, and the presence of rocks containing nondiagnostic mineral assemblages. Contact metamorphism associated with mafic dykes and felsic intrusions is fairly widespread and predates the regional metamorphism. The increase in metamorphic grade and the intensity of ductile deformation from south to north is relatively uniform. There are no fundamental structural or metamorphic breaks in this area.

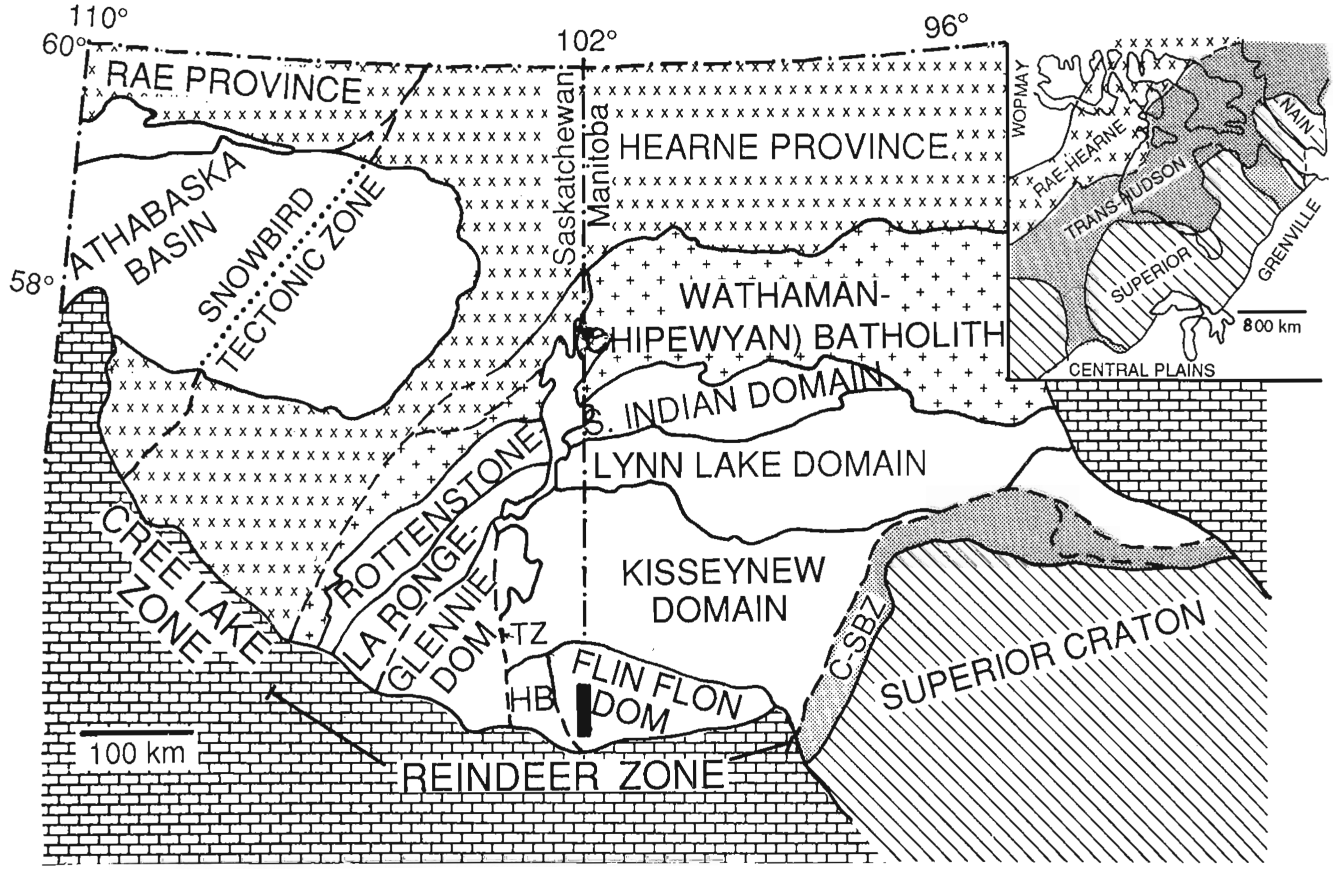

Figure 1. Domains of the Trans-Hudson orogen in Saskatchewan and Manitoba. The Flin Flon area is marked as a small black rectangle. Brick pattern is Paleozoic cover. Modified from Bickford et al. (1990). Inset shows the general extent of the Trans-Hudson orogen.
TZ: Tabbernor fault zone
HB: Hanson Lake block
C-SBZ: Churchill-Superior boundary zone

GEOLOGY OF THE FLIN FLON AREA

Figure 2 is a simplified geological map of the Flin Flon area. Volcanic rocks of the ca. 1886 Ma (Syme et al., 1987) Amisk Group underlie most of the study area. The Amisk Group is unconformably overlain by conglomerates, arenites, and wackes of the 1832 ± 2 Ma (Gordon et al., in press) Missi Group. In subsequent rock descriptions, mineral abbreviations, given in Table 1, will be used.

Several groups of pre-Missi felsic and mafic intrusions are recognized in the Flin Flon area. The 1840 ± 7 Ma (K. Ansdell, pers. comm. 1990) Phantom Lake Granodiorite contains approximately 10% euhedral 1-8 cm Kfs, 40-50% 1-5 mm euhedral Pl, and 10% 1-4 mm anhedral Qtz phenocrysts. The groundmass is composed of fine-grained feldspar and Qtz with accessory Bt, Hbl, Chl, and Ep. The Phantom Lake Granodiorite is typically fresh-looking, unfoliated and cuts all other lithologies. No younger rocks have been observed to cut the Phantom Lake Granodiorite, however it is deformed by brittle faults of P5 age. The relative timing between deposition of the Missi Group and intrusion of the Phantom Lake Granodiorite is somewhat uncertain. There are no clear contact relations between the two units in the map area. Bailes and Syme (1989) stated that dykes of porphyritic granodiorite similar to the Phantom Lake Granodiorite cut the Missi but do not refer to a specific location. Within the stated errors, there is overlap between the ages given for the Missi Group and Phantom Lake Granodiorite in this area. It is also possible that the Missi Group is older than 1832 Ma here at the west end of the Flin Flon domain where it is undated.

In some areas aphanitic and porphyritic mafic dykes and plugs may constitute up to 80% of the outcrop. Porphyries contain up to 60% Pl and/or Cpx phenocrysts. The aphanitic dykes are dark brown to black and scattered and have Qtz amygdules. The dykes are up to 2 m in width and cut all other rocks except the Phantom Lake Granodiorite. Porphyritic and aphanitic dykes cut each other suggesting that they were emplaced contemporaneously. The petrography and field relations of these dykes are similar to those of some rocks of the Boundary suite of intrusions (Syme and Forester, 1977; Bailes and Syme, 1989). Further work is planned to test this hypothesis. Fine- to coarse-grained, generally unfoliated, plug-like bodies of quartz diorite to gabbro are locally abundant in the Flin Flon area. Most of these bodies are less than 1 km^2 in areal extent.

Five deformational events (P1 — P5) have been identified by Bailes and Syme (1989) in the Flin Flon area. Wilcox (1990) discusses the deformational history of the

Table 1. Mineral abbreviations (from Kretz, 1983).

Act: Actinolite	Bt : Biotite	Cal: Calcite
Chl: Chlorite	Cpx: Clinopyroxene	Ep : Epidote
Grt: Garnet	Kfs: K Feldspar	Hbl: Hornblende
Ms: White Mica	Pl : Plagioclase	Pmp: Pumpellyite
Prh: Prehnite	Qtz: Quartz	Sil: Sillimanite

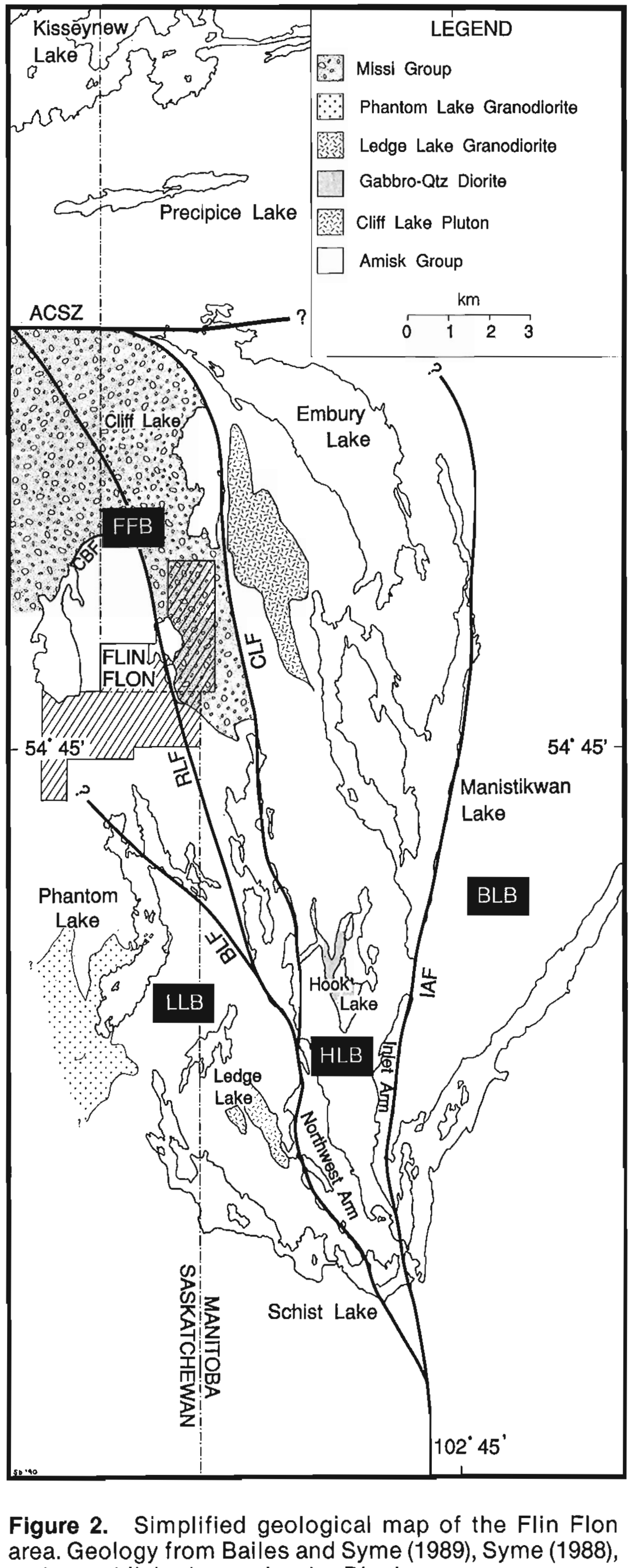

Figure 2. Simplified geological map of the Flin Flon area. Geology from Bailes and Syme (1989), Syme (1988), and unpublished mapping by Digel.

LLB: Ledge Lake block
HLB: Hook Lake block
BLB: Bear Lake block
FFB: Flin Flon block
BLF: Burley Lake fault
ACSZ: Annabel Creek shear zone
CLF: Cliff Lake fault
RLF: Ross Lake fault
CBF: Club Lake fault
IAF: Inlet Arm fault

Flin Flon area in the context of a larger scale tectonic picture. P1 folds deformed the Amisk Group prior to deposition of the Missi Group. P2 to P4 folds and their associated fabrics are not widely developed in the region. They are open to tight folds, with or without an associated axial planar fabric, which tend not to be equally well developed within the Missi Group and Amisk Group rocks. P2 to P4 folds are pre- to syn-metamorphic and do not deform isograds nor do the associated foliations overprint metamorphic mineral assemblages.

P5 is a widespread brittle to brittle-ductile faulting event which dominates the structural fabric over most of the area. Foliations are generally well developed only within several metres of P5 faults. Zones affected by the faults are generally up to a few tens of metres in width. Outside these zones, the rocks are essentially undeformed and primary volcanic and sedimentary structures are well preserved. Fault-related fabrics overprint metamorphic mineral assemblages, hence P5 is postmetamorphic.

Bailes and Syme (1989) have subdivided the geology of the Flin Flon area into stratigraphically distinct blocks (Fig. 2) separated by block-bounding faults of P5 age. They have determined that each block contains a distinct internal volcanic stratigraphy which can not be correlated across the block-bounding faults.

The regional metamorphism in the Flin Flon area increases northward from prehnite-pumpellyite facies (Prh-Pmp in metabasites) in the southern Hook Lake block to middle amphibolite facies (Sil-Grt-Bt in metapelites) on the south shore of Kisseynew Lake (Fig. 3). The highest grade metavolcanic rocks are pillows containing Hbl-Pl-Qtz north of Precipice Lake. More detailed textural descriptions of these rocks and minerals are given by Digel and Gordon (1990).

Ledge Lake Block

The Ledge Lake block is underlain by abundant pillowed flows and less abundant mafic breccia and scoria tuff of the Amisk Group (Fig. 2). Outcrops of Phantom Lake Granodiorite are most abundant in this block, particularly near Phantom Lake. Dykes and plugs of Phantom Lake Granodiorite have intruded and metamorphosed the volcanic rocks throughout much of the block. The undated Ledge Lake Granodiorite is exposed over approximately 4 km^2 in the centre of the Ledge Lake block. It cuts the Amisk Group rocks but, unlike the Phantom Lake Granodiorite, is commonly strongly foliated. In one location a Phantom Lake Granodiorite dyke was observed to cut the Ledge Lake Granodiorite. No rocks containing Prh or Pmp were observed within several hundred metres of this unit and it is likely that it has also had some contact metamorphic effect on the country rocks. Abundant intrusions of fine-grained gabbro to diorite occur in the southern half of the Ledge Lake block. All of these bodies are less than 1 km^2 in area and have no obvious thermal aureoles.

Brittle deformation is widespread over much of the Ledge Lake block, especially in the north half where faulting is most pervasive. Foliations are present only in the immediate vicinity of these P5 faults. Centimetre-scale

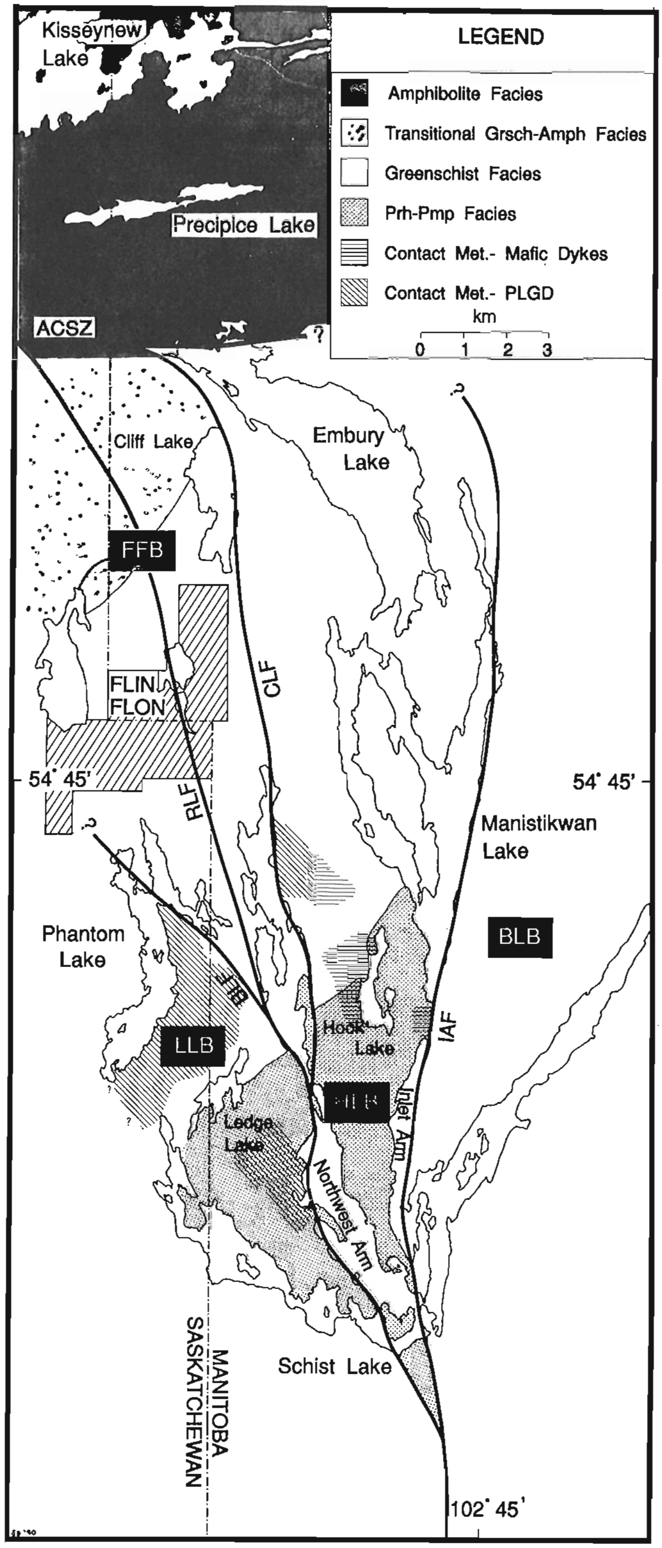

Figure 3. Metamorphic map of the Flin Flon area. Facies are primarily determined from assemblages in basaltic metavolcanic rocks. Abbreviations as in Figure 2.

brittle fractures and joints cut hard, hornfelsed volcanic rocks near P5 faults. These P5 structures deform all rocks in the Ledge Lake block.

Rocks bearing Prh and Pmp are abundant in the southern Ledge Lake block; however, thermal metamorphism associated with the Phantom Lake Granodiorite and Ledge Lake Granodiorite has resulted in an irregular distribution of mineral assemblages (Fig. 3). Volcanic rocks unaffected by the Phantom Lake Granodiorite contain Chl-Qtz-Ep-Pl $\pm$ Prh $\pm$ Pmp. Basaltic rocks which have been contact metamorphosed to amphibolite facies are shiny black and are cut by abundant Ep veins. In thin section they contain fine-grained Hbl-Pl-Qtz-Bt $\pm$ Ep $\pm$ Chl. Contact metamorphism is especially extensive in the north and west parts of the Ledge Lake block (Fig. 3).

Gabbro intrusions are cut by networks of randomly oriented millimetre-scale veins. Vein assemblages are Prh-Qtz $\pm$ Pmp and Ep-Qtz suggesting that intrusion predated the regional metamorphism. Most outcrops are cut by one type of vein although both types occur at a few locations. Mineral assemblages in these veins have an irregular spatial distribution.

It appears that a large portion of the Ledge Lake block was regionally metamorphosed at Prh-Pmp facies conditions. Large areas experienced contact metamorphism during intrusive activity before the regional metamorphism. Prh-Pmp facies assemblages are developed only in rocks isolated from the contact effects of the Ledge Lake Granodiorite and Phantom Lake Granodiorite. This resulted in the irregular distribution of mineral assemblages over this block (see discussion for Hook Lake block).

Flin Flon Block

Volcanic rocks of the Amisk Group underlie most of the Flin Flon block south of the Club Lake fault (Fig. 2). These rocks are mostly pillow basalts with lesser amounts of pillow breccia and scoria tuff. Missi Group arkosic sandstones with centimetre-scale conglomerate layers cover most of the area north of the Club Lake fault and south of the Annabel Creek shear zone (Fig. 2). The basal conglomerate is well exposed just south of Annabel Creek and northeast of the Club Lake fault.

The Flin Flon block contains structural elements transitional to the largely brittle deformation in the southern Flin Flon domain and ductile deformation of the northern part of the Flin Flon domain. In the southern part of the Flin Flon block, foliations exist only near P5 faults, and deformation style is similar to that of the Ledge Lake block . The onset of penetrative flattening (ductile deformation) is seen in volcanic rocks just north of Flin Flon. This fabric becomes increasingly well developed in Missi Group rocks north of the Club Lake fault and swings to the northwest near Annabel Creek.

Three major faults occur in the Flin Flon Block. The Club Lake fault, a pre-P3 (P2?) fault which has thrust the Amisk Group over the Missi Group (Bailes and Syme, 1989), was subsequently folded by the P3 Hidden Lake synform. The fault plane is well exposed north of Flin Flon (Fig. 2); it is an intensely foliated zone about a metre thick. Pillows are strongly flattened within 10 m of the fault.

The Ross Lake fault, within the Flin Flon block, is a brittle fault which apparently cuts the P5 Annabel Creek shear zone at the north end of the area (Fig. 2). In the south it appears to merge with the Burley Lake fault and probably the Cliff Lake fault (Fig. 2). There must be some sort of transfer of motion between the termination of the Cliff Lake fault and continuation of the Ross Lake fault but there is insufficient exposure to characterize the mechanism. Motion on the Ross Lake fault must be moderate because:

a) the Phantom Lake Granodiorite occurs on either side of it
b) Prh-Pmp facies rocks exist on either side of the Ross Lake fault with no large metamorphic offset and
c) similar Amisk Group stratigraphy exists on both sides of the fault.

The Burley Lake fault is a poorly exposed, dominantly brittle fault, which separates the Ledge Lake block and Flin Flon block. It juxtaposes volcanic rocks of the Ledge Lake block, which have been metamorphosed by abundant intrusions of Phantom Lake Granodiorite, against volcanic rocks of the Flin Flon block which are free of Phantom Lake Granodiorite dykes and their thermal effects. Volcanic rocks of the Flin Flon block are of greenschist facies at the south end of the block and reach upper greenschist near the Club Lake fault. Assemblages in the south end are Act-Ep-Chl-Pl-Qtz $\pm$ Bt (assemblage A). Near the Club Lake fault the volcanic rocks contain actinolitic Hbl-Pl-Bt $\pm$ Ep $\pm$ Chl (assemblage B).

The transition from assemblage A to assemblage B was mapped in volcanic rocks across a zone less than 30 m wide (Fig. 3). Rocks with assemblage A are green, relatively soft, and amphibole is barely visible with a hand lens, although it is abundant in thin section. Rocks with assemblage B however, are dark green to black, hard, and contain prismatic amphiboles several mm in size. Accompanying the increase in amphibole size is a sharp decrease in the amount of Ep and Chl in these rocks. The area over which rocks with assemblage B can be mapped is limited by the absence of volcanic rocks north of the Club Lake fault. Metamorphism of the Missi Group rocks north of the Club Lake fault is discussed in the section concerning the Annabel Creek shear zone.

Hook Lake Block

The Hook Lake block is dominantly underlain by Amisk Group volcanic rocks (Fig. 2). Outcrop constitutes as much as 70% the area. A distinctive mafic tuff is composed of rounded lapilli to bomb sized clasts of mafic scoria in a fine-grained matrix. Clasts may comprise from 0-70% of the unit. Bedding, defined by different proportions of clasts, is well preserved. Pillows and pillow breccias are also widespread.

An elongate body of Phantom Lake Granodiorite, 500-600 m^2 in size, on the northeast shore of the Northwest Arm of Schist Lake is the largest body of granodiorite away from Phantom Lake. Metre-scale dykes of Phantom Lake Granodiorite are scattered across the central part of the block. The dykes are the youngest rocks in the block and locally cut the mafic dyke complexes.

Mafic dyke complexes are found only in the Hook Lake block (Fig. 3). Within the dyke complexes the Amisk Group volcanic rocks have been thermally metamorphosed and fresh surfaces are shiny black in appearance. There is no penetrative fabric. Rocks which are thermally metamorphosed grade into those showing no contact effects. Areas were mapped as belonging to the mafic dyke complex if dykes constituted more than 30% of the outcrop.

Massive gabbro to quartz diorite intrusions are common over the central part of the Hook Lake block. The larger intrusions are shown in Fig 2. Where they occur near the dyke complexes, they are cut by the mafic dykes. Bailes and Syme (1989) described the petrology and field relations of these rocks in detail and suggested that they are closely related in composition and time to the volcanic rocks that they intrude.

The Cliff Lake fault is a brittle fault separating the Flin Flon block and Ledge Lake block from the Hook Lake block. The Hook Lake block contains considerably more mafic breccia and scoria tuff than the Flin Flon block and Ledge Lake block . There are considerable similarities in terms of lithology and metamorphic grade between these three blocks, however, suggesting that movement on the Cliff Lake fault is moderate.

Postmetamorphic (P5) faults are common in the Hook Lake block. Exposure in the fault zones is usually quite poor, however, and they are usually occupied by lakes or swamps. Sense and amount of movement on these faults is difficult to establish. The Hook Lake block is not penetratively deformed, and foliations are restricted to P5 fault zones.

The Hook Lake block contains the largest and best preserved exposures of Prh-Pmp facies rocks in the Flin Flon volcanic belt and the Prh-Pmp to greenschist facies transition is best exposed here. The transition is not, however, as simple as suggested by Bailes and Syme (1989) and Digel and Gordon (1990) and is complicated by:

a) the occurrence of massive lithologies (gabbros, diorites) containing non-diagnostic mineral assemblages,
b) the effects of P5 faults and
c) contact metamorphism associated with mafic dyke complexes.

The Prh-Pmp to greenschist facies transition zone in the Hook Lake block is occupied by extensive intrusions of gabbro to quartz diorite (Fig. 2). There is no contact metamorphism associated with these intrusions. The ability of metamorphic fluids to permeate the rocks has a critical bearing on the development of very low-grade index minerals such as Prh and Pmp. These minerals are best developed in the most amygdaloidal (permeable) rocks and tend not to occur in the massive lithologies except in veins. Prh + Pmp veins (southern Hook Lake block) and Qtz-Ep veins (northern Hook Lake block) cut these gabbros. Probably these veins were filled during the regional metamorphism. The distribution of veins containing these index minerals is structurally controlled, making it difficult to constrain accurately mineralogical changes (isograds) across the intrusions.

P5 fault zones have a strong fabric and a non-diagnostic mineral assemblage of Cal-Chl-Ep. It is possible to trace undeformed units with abundant Prh and Pmp into fault zones where these same units develop a strong fabric and the Cal-Chl-Ep assemblage (no Prh or Pmp). Movement along these faults must offset the metamorphic zones, however mineralogical offsets are difficult to detect here since Prh and Pmp are rare in this area and have been obliterated in the fault zones.

The existence of widespread mafic dyke complexes further obscures the transition. Volcanic rocks in the complexes are thermally metamorphosed at middle to upper greenschist facies (Act-Ep-Chl-Pl-Bt). In the southern part of the dyke complexes (Fig.3), veins of Prh $\pm$ Pmp cut the thermally metamorphosed rocks and the mafic dykes. In the northern part of the complexes, veins with identical field relations to the Prh-Pmp veins contain Ep-Qtz. These veins are related to those which cut the gabbros. Presumably they reflect a regional metamorphic gradient superimposed on rocks which were previously contact metamorphosed by the mafic dykes (Table 2).

Bear Lake Block

The stratigraphy of the western Bear Lake block is dominated by pillows with little of the mafic breccia and scoria tuff characteristic of the Hook Lake block. The Phantom Lake Granodiorite does not occur in the Bear Lake block.

The Inlet Arm fault is a ductile north-south fault separating the Hook Lake block and Bear Lake block. It is a dominant structure in the map area. Kinematic indicators in the Inlet Arm fault suggest that the Bear Lake block has moved up and to the north relative to the Hook Lake block, consistent with the juxtaposition of Prh-Pmp and greenschist facies rocks across southern part of the fault. Within the Bear Lake block, the rocks are penetratively deformed. Pillows are typically flattened to ratios in excess of 10:1 and primary volcanic structures have been largely destroyed. A consistent, steeply-

Table 2. Timing of geological events in the Hook Lake block

1886 Ma	Amisk Group volcanism Intrusion of massive gabbros/diorites Mafic dyke intrusion → contact metamorphism
1840 Ma	Phantom Lake Granodiorite stock and dykes → contact metamorphism
1815 Ma	Regional metamorphism P5 faulting and associated deformation

dipping, north-south flattening fabric overprints the metamorphic mineral assemblage throughout the area. Rocks throughout the Bear Lake block are metamorphosed to the greenschist facies. The usual mineral assemblage is Act-Chl- Qtz-Ep-Pl.

Annabel Creek Shear Zone

This shear zone is an east-west trending zone of left lateral ductile deformation (Fig. 2). It is the northern boundary of the tectonic block system of Bailes and Syme (1989). The north-south trending block faults and their associated fabrics are dragged into the Annabel Creek shear zone, forming a distinctive kilometre-scale sinistral drag fold at Embury Lake (Fig. 2). Outcrop-scale kinematic indicators in the study area and to the west (Wilcox, 1990) are consistently sinistral. The brittle Ross Lake fault is the only structure which cuts the Annabel Creek shear zone.

Missi Group sediments develop a progressively stronger foliation northward from the Club Lake fault to the Annabel Creek shear zone. This fabric gradually intensifies toward the Annabel Creek shear zone and is rotated into the Annabel Creek shear zone. Conglomerate horizons in Missi Group sandstones contain strongly rodded cobbles. Near the Club Lake fault the lineation plunges moderately to the southeast but is gradually rotated to plunge gently to the east just south of Annabel Creek where the basal conglomerate is exposed. North of the creek, strongly foliated chlorite schists are interpreted to be highly deformed Amisk Group volcanic rocks. The Annabel Creek shear zone is not a single fault but a zone of penetrative ductile deformation containing numerous higher strain zones. These observations are consistent with those made by Wilcox (1990) and Ashton et al. (1987).

Near the Club Lake fault, Missi Group sandstones contain Qtz-Pl-Ms-Bt. The first appearance of Grt about 1 km south of the Annabel Creek shear zone is irregular and apparently strongly influenced by bulk composition. The metamorphic assemblage in similar rocks near Annabel Creek is Qtz-Pl-Bt-Ms Grt ± Hbl ± Chl. Assemblages in metavolcanic rocks are Hbl-Pl-Qtz ± Bt (amphibolite facies). Volcanic rocks within the high strain zones have been retrograded to greenschist facies (Chl-Qtz-Pl-Ep-Act).

Wilcox (1990) proposed that the Annabel Creek shear zone is a P5 structure at the transition from the high-level, low-grade, brittle rocks in the south to the deeper-level, high-grade, ductile rocks to the north. Our structural and metamorphic observations in the Annabel Creek shear zone west of Embury Lake are consistent with his view that the Annabel Creek shear zone is part of a metamorphic and structural continuum. Progressively deeper levels of the crust are exposed from the Hook Lake area north to Kisseynew Lake. There is no evidence in the study area that the Annabel Creek shear zone is a fundamental structural or metamorphic break.

Precipice Lake Area

The Precipice Lake area encompasses the deepest level and highest-grade rocks observed in the study area. It is largely underlain by coarse gabbros and Amisk Group metasedimentary rocks of greywacke composition. Volcanic rocks are interlayered with the sediments but comprise less than 30% of the outcrop. Pillows and pillow breccias, flattened up to 15:1, were recognized in a few areas. Massive to well foliated gabbros occur as sill-like bodies up to several hundred metres thick.

A penetrative flattening fabric exists and is parallel to the foliation in the Annabel Creek shear zone (280°/65°). Contacts between lithologic units are transposed parallel to this foliation. Centimetre- to metre-scale isoclinal folds of the flattening fabric are abundant. Larger scale folds probably exist but are difficult to recognize.

Rocks in the Precipice Lake area have been metamorphosed to amphibolite facies. Volcanic rocks contain Hbl-Pl-Qtz-Bt. Metagreywackes near the Manitoba-Saskatchewan border on the south shore of Kisseynew Lake contain Bt-Qtz-Pl-Sil. Enclaves of chlorite schist and other greenschist facies rocks are probably the result of local retrogression in high strain zones.

CONCLUSION

A prograde sequence of metamorphic rocks from prehnite-pumpellyite to amphibolite facies is present in the Flin Flon area. Brittle deformation is predominant in the low-grade southern part of the area. Concomitant with increasing metamorphic grade (deeper structural levels) is the development of a penetrative flattening fabric and other ductile deformational features. The Annabel Creek shear zone is a kilometre-scale ductile high strain zone near the transition to the amphibolite facies. It does not appear to be a fundamental structural or metamorphic break in this area.

Contact metamorphism associated with mafic dyke complexes and felsic intrusions is widespread in the low-grade area. Contact metamorphic assemblages reach amphibolite facies and are overprinted by the regional metamorphism.

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Reconnaissance geology of the Precambrian Shield of the Boothia Uplift, northwestern Somerset Island and eastern Prince of Wales Island, District of Franklin

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Abstract

The northern end of the crystalline core of the Boothia Uplift on northwestern Somerset Island represents a continuation of the major lithologies and structural trends present to the south but the degree of deformation is significantly greater. Granitoid gneisses and aluminous paragneisses, all in granulite facies, are deformed into north-south trending, tight, generally upright folds. Flattening and mylonitization are commonly intense.

On easternmost Prince of Wales Island, the gneisses are similar in metamorphic grade and structural style but are almost entirely granitoid. They are heavily retrograded along the high-angle reverse faults separating them from the Paleozoic at the margin of the Boothia Uplift.

A major post-tectonic K-feldspar-megacrystic granite (1703 ± 2 Ma old) has intruded in a linear zone along the west coast of northern Somerset Island. K-feldspar-quartz porphyry dykes, locally fluorite-bearing, in the bordering gneisses are doubtless related to the granite.

Résumé

L'extrémité nord du noyau cristallin du soulèvement de Boothia dans le nord-ouest de l'île Somerset représente un prolongement des principales lithologies et directions structurales présentes au sud mais la déformation y est beaucoup plus importante. Les gneiss granitoïdes et les paragneiss alumineux, dans les deux cas du faciès des granulites, sont déformés en plis généralement droits, serrés, à direction nord-sud. L'aplatissement et la mylonitisation ont été généralement intenses.

À l'extrême est de l'île Prince of Wales, les gneiss présentent un degré de métamorphisme et un style structural semblables mais ils sont presque entièrement granitoïdes. Ils ont subi un métamorphisme régressif important le long de failles inverses à fort pendage les séparant des roches paléozoïques en bordure du soulèvement de Boothia.

Un important granite mégacristallin à feldspath potassique post-tectonique (1703 ± 2 Ma) a pénétré par intrusion dans une zone linéaire longeant la rive ouest du nord de l'île Somerset. Les dykes de porphyre à quartz et à feldspath potassique, contenant par endroits de la fluorite et qui se sont formés dans les gneiss de bordure, sont sans aucun doute liés au granite.

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INTRODUCTION

The Boothia Uplift is a major, north-plunging structural feature extending 800 km from southern Boothia Peninsula to northwestern Devon Island. It is cored by high grade metamorphic rocks of the Churchill Structural Province of the Canadian Shield and flanked by Helikian to Devonian sediments more than 4000 m thick. Uplift occurred largely in Siluro-Devonian time, by either vertical block-faulting (Kerr, 1977) or west-directed compression (Okulitch et al., 1986).

The crystalline core of the Boothia Uplift, consisting of granulite facies gneisses and supracrustals and intrusive rocks and exposed as far north as northwestern Somerset Island, is the subject of the present study, which is based on 1:250 000 scale geological mapping north of 71°N. Fieldwork in the region between 71 and 73°N was begun in 1986 (Frisch et al., 1987) and continued in 1987. In 1990, during 5 weeks of work, mapping was extended northward to the northern limit of the exposed Shield and westward to the Shield exposures in easternmost Prince of Wales Island at the margin of the Boothia Uplift.

This report summarizes the results of the 1990 fieldwork and presents a simplified geological map of the Precambrian Shield terrane of the Boothia Uplift north of 71°N.

LITHOLOGY

Gneisses and supracrustal rocks

The coastal regions of western Somerset Island between 73°N and M'Clure Bay (Fig. 1B) and much of easternmost Prince of Wales Island (Fig. 1A) are underlain mainly by the dark weathering (hornblende-)biotite-orthopyroxene gneisses of tonalitic to granodioritic composition that form the bulk of the crystalline terrane to the south (Frisch et al., 1987). Pink, K-feldspar-rich gneisses, locally augen textured, form concordant sheets, centimetres to metres thick, in the pyroxene gneiss and impart a migmatitic aspect to many outcrops. Rusty weathering, garnetiferous schistose and felsic paragneisses occur as intercalations, up to a few hundred metres thick, in the pyroxene gneisses on Somerset Island but are rare on Prince of Wales Island.

Retrogression of the dark pyroxene gneisses, typically marked by reddening due to increased K-feldspar (microcline) content, is common throughout the Boothia-Somerset Shield terrane but is particularly marked in a broad zone (up to several hundred metres wide) along the Shield-Proterozoic/Paleozoic contact on the west shore of Peel Sound. This contact, especially with the Paleozoic, is commonly along a high-angle reverse fault and much of the adjoining gneiss is dusky red, chloritic and shot through with quartz veins and lenses.

The pyroxene and granitic gneisses and paragneisses have all been deformed together into tight folds, with north-striking, vertical to slightly overturned axial planes and shallow to moderate plunges (15-40°) north and south. With few exceptions, the gneisses show considerable strain and high grade mylonites, with quartz ribbons, recrystallized orthopyroxene, and garnets strung out along foliation planes, or highly attenuated mafic lenses, are common. Some shear zones are pervaded by K-feldspathization, which has transformed the brown or grey tonalitic gneisses into pink granitic ones.

The central part of the Shield terrane of northern Somerset Island is occupied largely by supracrustal rocks, mainly migmatitic (graphite-cordierite-)garnet-sillimanite-biotite gneiss and garnet-biotite gneiss, commonly rusty weathering and schistose. Associated with these rocks are white, garnetiferous, biotite-free leucogneiss and subordinate quartzite and mafic to ultramafic layers, lenses and pods of garnet-free pyroxene-amphibole rock. However, green- or brown-weathering metaperidotite bodies, such as those occurring in southern Somerset Island and northern Boothia Peninsula (Frisch et al., 1987), have not been found. In this context, it should be pointed out that the ultramafic body on Arcedeckne Island (Fig. 1D) tentatively identified as garnet peridotite by Frisch et al. (1987) actually consists of layered garnet + fayalite + apatite + quartz ± Fe-rich orthopyroxene rock and probably represents metamorphosed shaly-phosphatic iron formation.

The major supracrustal gneiss belt of northern Somerset Island is broken up by narrow, north-trending grabens filled by poorly-consolidated sandstones of the Tertiary Eureka Sound Group.

Neither marble nor calc-silicate rock has been found in the supracrustal belt. Indeed, the only substantial metacarbonate outcrops known north of 73°N occur east of M'Clure Bay (Fig. 1B), associated with a thin supracrustal intercalation of pyroxene amphibolite and quartzite in pyroxene gneiss. The metacarbonate is a pale grey weathering (forsterite-)diopside-calcite marble, locally altered to ophicalcite, forming isolated outcrops that are the remnants of a tectonically disrupted layer — the typical outcrop pattern of carbonate rock in the Shield terrane of the Boothia Uplift. The scarcity of marble north of 73°N relative to the areas south is probably a consequence of the higher intensity of deformation in the north.

The eastern half of the Shield terrane on northern Somerset Island (the area east of the dotted line in Fig. 1B and C) is an upland of low relief covered by felsenmeer, mud and till. Widely scattered outcrops indicate that this area is underlain by crystalline rock similar to that to the west but, although the felsenmeer is largely in situ and in places faithfully reflects the underlying bedrock (e.g. dykes, fold structures), rock units cannot be traced far. No contact between gneiss and Paleozoic rock was seen but there is no evidence of any paleo-weathering in the gneisses closest to the Paleozoic.

K-feldspar-megacrystic granite and related rocks

Aside from diabase dykes, the only intrusive rock of significance in the Shield terrane of northern Somerset Island is a massive, pink granite rich in K-feldspar megacrysts. Its main outcrop borders M'Clure Bay, where it underlies an area 22 km long and up to 10 km wide (Fig. 1B); hence the rock is informally named the M'Clure Bay granite.

At M'Clure Bay, the granite is entirely massive with abundant, randomly oriented, pink K-feldspar phenocrysts commonly 3-4 cm long in a pink, fine- to medium-grained, biotite granite matrix. The appearance of the granite changes little near the contact with the country rock gneisses but phenocryst size is slightly reduced (to 2-3 cm). Fine grained porphyritic and aphyric granite veins, doubtless offshoots from the main granite body, crosscut gneisses on small islets at the southern end of M'Clure Bay.

Outcrops of M'Clure Bay granite are found on the Peel Sound coast as far as 25 km south of M'Clure Bay. Excellent wave-washed exposures display a well-defined contact zone, wherein gneiss inclusions, up to 0.5 m across, have been rotated but have not reacted with the granite melt. Granite at the main contact with the gneisses shows ghost-like mafic-felsic layering parallel to the N-S foliation in the basement.

Clearly, the M'Clure Bay granite is a posttectonic intrusion emplaced in a linear zone paralleling the structural trend of the gneisses. Uranium-lead dating confirms that the granite is relatively young (see below).

Probably related to the M'Clure Bay granite are K-feldspar-quartz porphyry dykes that occur sporadically in the Shield terrane of northern Somerset Island. The dykes are up to 10 m thick and consist of several varieties of porphyry, differing mainly in grain size. One variety has pink K-feldspar phenocrysts (1 cm) and smoky quartz phenocrysts (0.5 cm) in a very fine grained quartzofeldspathic matrix. In one such dyke, cutting the M'Clure Bay granite near the south end of the bay, possible rapakivi texture is evidenced by white (plagioclase?) feldspar rimming the pink K-feldspar phenocrysts. Another variety of dyke is miarolitic and has abundant small feldspar and quartz phenocrysts and accessory fluorite, readily visible in hand specimen. All the dykes have a distinctly high crustal level aspect and show no preferred orientation. Although not abundant, they appear to be most common in the general vicinity of M'Clure Bay but one dyke, trending north, was found on the large island off the Somerset Island coast just north of 73°N. Felsic porphyry dykes are unknown in the Shield terrane to the south.

Diabase dykes

Diabase dykes, mostly belonging to the Mackenzie and Franklin swarms, are considerably more abundant north of 73°N than to the south. Major trends are southwest and northwest but a number of dykes strike westerly and northerly. At least two lithological varieties exist, one weathering black, the other (less common) brown (olivine-bearing?). The brown-weathering dykes consistently trend northwest and cut westerly-trending black dykes but are themselves cut by southwest-trending black dykes.

Unmetamorphosed Proterozoic rocks, west coast of Peel Sound

The Proterozoic sediments and basalt overlying the gneisses at the western margin of the Boothia Uplift were briefly examined on, and south of, Prescott Island (Fig. 1A).

In the central part of the Precambrian terrane on Prescott Island, sediments of the Proterozoic Aston Formation nonconformably overlie gneiss. A basal conglomerate, consisting of well-rounded gneissic and granitic pebbles (averaging 2-3 cm but reaching 10 cm in diameter) in a crossbedded calcareous sandstone matrix, rests on steeply-dipping, migmatitic pyroxene gneisses that are little altered. The conglomerate is at most 2 m thick and little of it is preserved. It is overlain by a gently west-dipping, thick succession of mainly dusky red and pink quartz sandstones, ripple marked and crossbedded, with minor red siltstone interbeds and, near the base, red silty dolomite. The distinctive red, stromatolitic dolomite marker unit near the base of the Aston Formation on Somerset Island (Stewart, 1987) occurs several hundreds of metres higher in the formation on Prescott Island, just below a thick basalt sill. It can be readily followed south across Pandora Island to the southernmost coastal outcrop of the Proterozoic succession on Prince of Wales Island. In these southern outcrops, however, as noted by Dixon et al. (1971), the marker unit is a pink, cherty-silty dolomite without stromatolites. In this area (Area 1 of Dixon et al. (1971)), a thin conglomerate-breccia is locally preserved at the base of the Aston Formation, resting on pink, mylonitic granitic gneiss.

The southern of the two westerly-trending diabase dykes on Prescott Island (Fig. 1A) runs into the thick basalt sill with which the stromatolitic-cherty dolomite unit is associated. The dyke could not be traced through the sill and thus appears to be a feeder to it.

STRUCTURAL GEOLOGY

As already noted, the degree of deformation of the crystalline rocks on Somerset Island north of 73°N and on Prince of Wales Island is higher than in the south. Mylonitization and flattening in the N-S plane are intense, major folds are tight and at most only slightly overturned, and divergence of structural trends from north-south is minimal. The dome and basin interference structures that characterize southern Somerset Island and northern Boothia Peninsula are absent. Thus the F_3 folding event so pervasive throughout the crystalline core of the Boothia Uplift (Frisch et al., 1987) is here even more dominant.

North-south faulting, some of it probably as young as Tertiary, appears to be particularly significant north of 73°N, as evidenced by the Tertiary-filled grabens and offset westerly-trending dykes.

Further examination in 1990 of the circular structure southwest of Stanwell-Fletcher Lake (Fig. 1C) has thrown some light on basement-cover relationships in the Boothia-Somerset Shield terrane. The structure is marked by a ring of outward-dipping marble and calc-silicate rock, concordantly mantled by paragneiss, that forms the central part of a remarkably large (ca. 10 km diameter) dome, which is believed to have formed as a consequence of interference between east-west F_2 folding and north-south F_3 folding (Frisch et al., 1987). The rocks within the carbonate ring, in the very core of the structure, are orthopyroxene-bearing granitoid gneisses and show a north-south, moderate to steep foliation similar to the

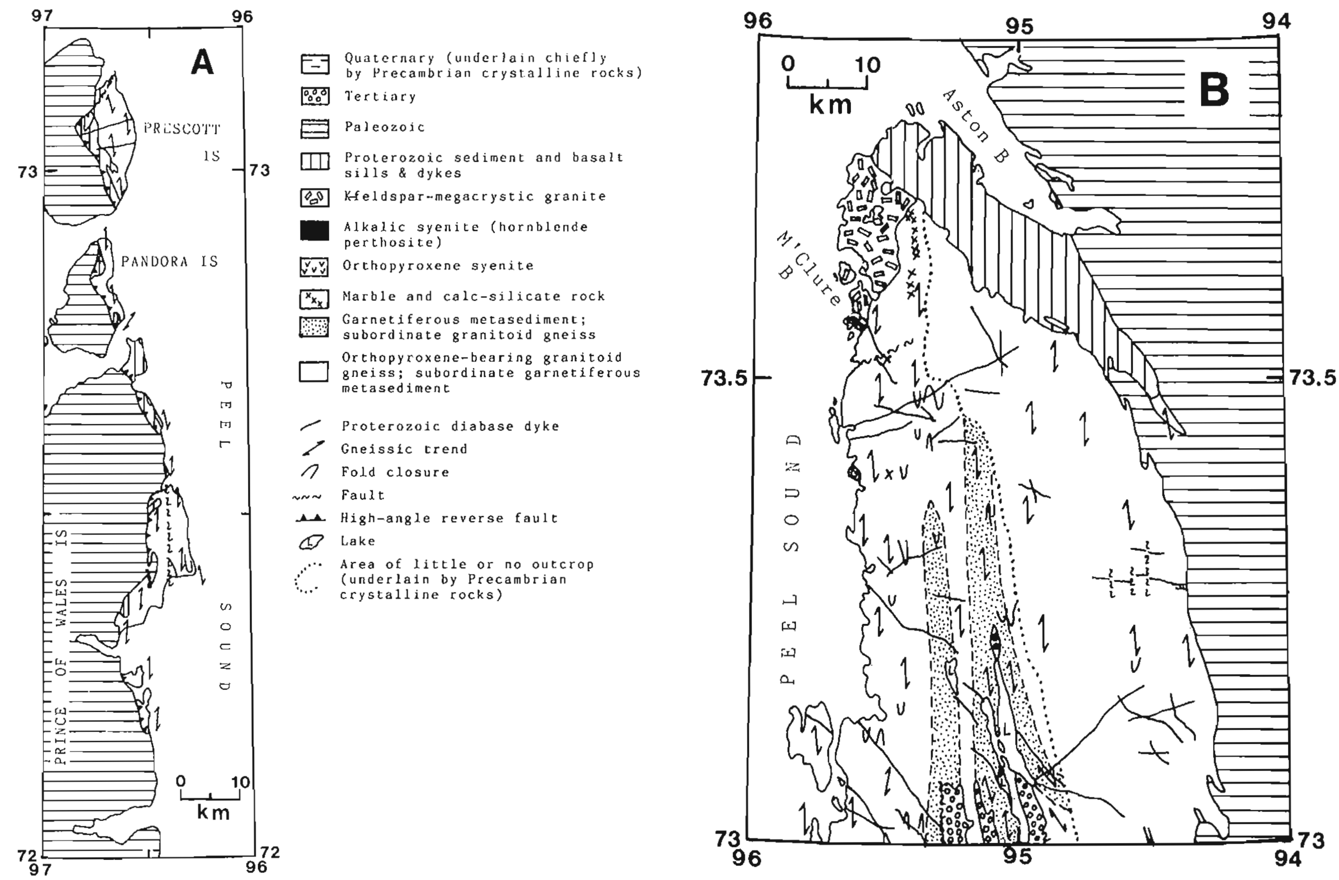

Figure 1. Sketch map of the geology of eastern Prince of Wales Island (A), northwestern Somerset Island (B), southern Somerset Island (C) and northern Boothia Peninsula (D). Distribution of unmetamorphosed Proterozoic and Phanerozoic rocks slightly modified after Stewart (1987) and Christie et al. (1971).

regional one outside the dome. As gneisses and supracrustals apparently were not folded together, a detachment between them may be inferred. Radiometric dating of the core gneisses should provide an age for basement in the Boothia Uplift.

GEOCHRONOLOGY OF THE CRYSTALLINE ROCKS

Little is known of the age of the metamorphic rocks of the Boothia Uplift. Reconnaissance whole-rock Sm-Nd dating (E. Hegner, personal communication, 1989) of four samples of granitoid gneiss from three localities in northern Boothia Peninsula has yielded T_{DM} ages ("crustal separation" ages based on the "depleted mantle" model) in the range 2.2-3.0 Ga. Sm-Nd ages of intrusive rocks are Proterozoic: 2.4 Ga for gneissic orthopyroxene syenite at Cape Bird (Fig. 1C) and 2.4 Ga for syntectonic alkalic syenite from northern Boothia Peninsula. The Cape Bird body has given a preliminary zircon U-Pb age of ca. 1.93 Ga (R.R. Parrish, personal communication, 1989), which probably approximates the age of the main granulite facies metamorphism in the area. Xenocrystic zircons, presumably derived from basement at depth, from kimberlite emplaced in the Paleozoic of northeastern Somerset Island have given several ages around 1935 ± 5 Ma and one of 2500 Ma (Reichenbach & Parrish, 1988).

A single fraction of zircon from a sample of M'Clure Bay granite collected north of the bay by R.D. Stevens in 1976 has given a virtually concordant age of 1703 ± 2 Ma (R.R. Parrish, personal communication, 1989). This is an unusually young age for granitic rock in the northern Canadian Shield and only slightly older than the K-Ar mineral ages of 1660 and 1670 Ma from the Boothia-Somerset Shield terrane reported by Blackadar (1967).

Thus, while it appears that much of the Boothia-Somerset Shield terrane was consolidated by mid-Proterozoic time, indications of a geological record extending back to 3.0 Ga suggest much work remains to be done to resolve the earlier history of the crystalline core of the Boothia Uplift.

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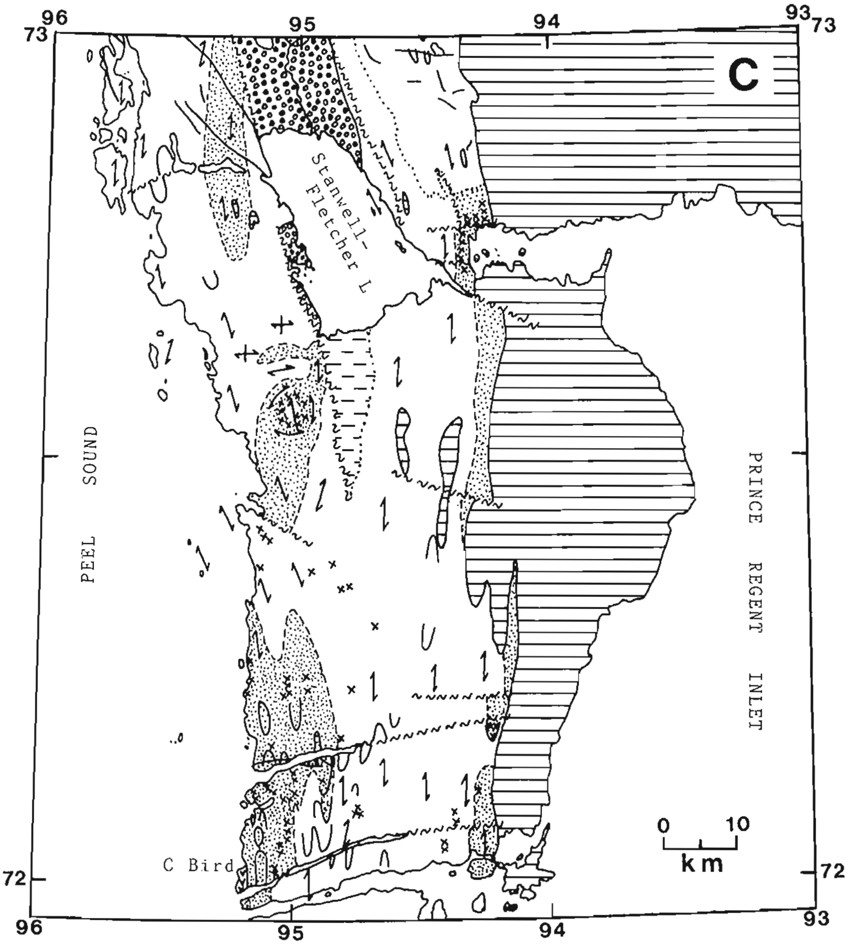
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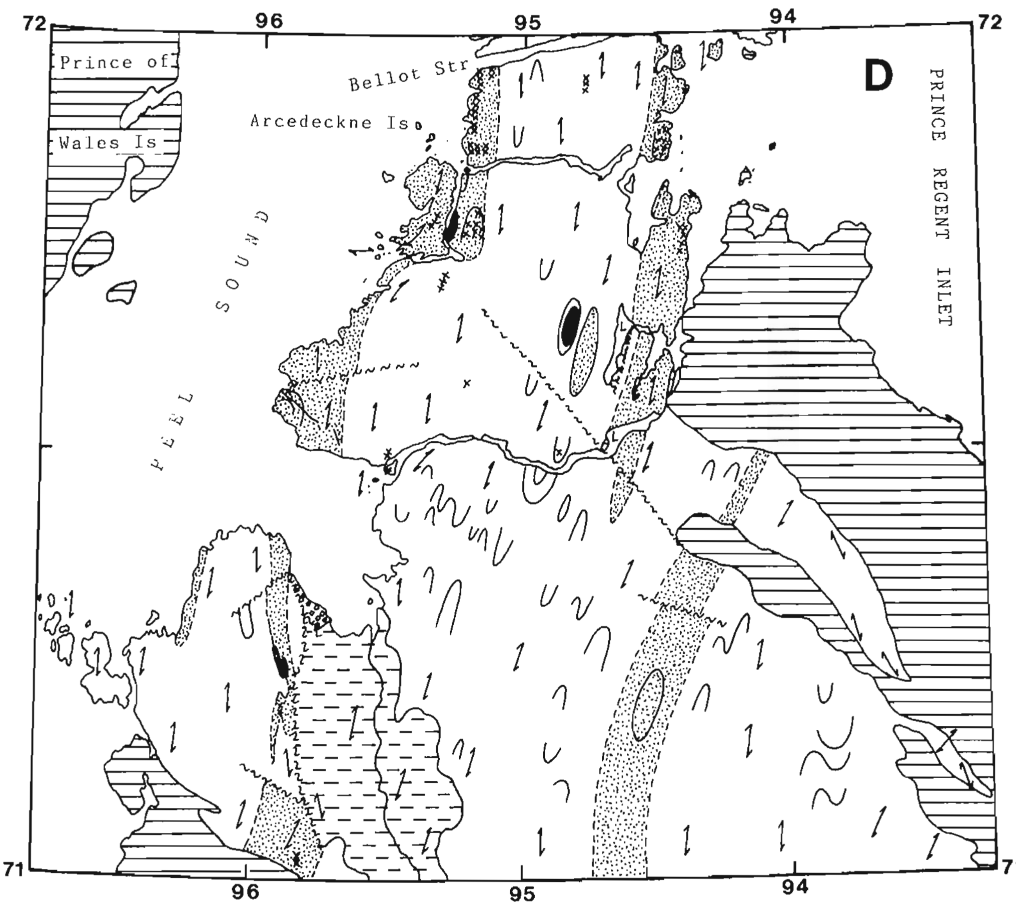
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Field relationships in the early Proterozoic Purtuniq Ophiolite, Lac Watts and Purtuniq map areas, Quebec

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Abstract

The meta-igneous rocks of the Watts Group comprise pillowed mafic volcanic rocks, sheeted mafic dykes, rare plagiogranite intrusions, and cumulate textured mafic and ultramafic rocks. Detailed mapping in the eastern portion of the Watts Group supports previous interpretations that these rocks represent the crustal portion of a dismembered early Proterozoic ophiolite. The sheeted dyke to pillowed basalt transition was mapped at 1:5 000-scale at a well-exposed locality 5 km east of lac Watts. Primary relationships between dykes and pillowed basalts are locally well preserved, although in most localities the contacts are deformed. A suite of late, fine grained mafic dykes crosscuts the sheeted dykes, pillowed basalts and the underlying cumulate mafic rocks. Future studies of primary igneous minerals collected in two detailed sampling traverses will help to constrain more fully the tectonomagmatic evolution of the ophiolite.

Résumé

Les roches méta-ignées du Groupe de Watts consistent de laves mafiques en coussins, de dykes stratifiés, d'intrusions plagiogranitiques, et de cumulats mafiques et ultramafiques. La cartographie en détail dans le secteur oriental du Groupe de Watts corrobore les interprétations précédentes c'est-à-dire que ces roches représentent la partie crustale d'une ophiolite du Protérozoïque inférieur. La transition entre les dykes stratifiés et les laves mafiques en coussins a été cartographiée à l'échelle de 1/5 000 dans une région bien à découvert, environ 5 km à l'est du lac Watts. Les relations primaires entre les dykes stratifiés et les laves en coussins sont localement très bien conservées, mais dans la plupart des cas, les contacts sont déformés. Un ensemble de dykes mafiques à grain fin recoupe les dykes stratifiés, les laves mafiques en coussins, et les cumulats sous-jacents. Des études futures sur les minéraux ignés primaires conservés dans une série d'échantillons prélevés le long de deux traverses détaillées faciliteront l'interprétation de l'évolution tectonomagmatique de l'ophiolite.

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INTRODUCTION

The Cape Smith Belt is a southward-verging thrust belt preserved in a regional-scale infold (Lucas, 1989; St-Onge and Lucas, 1990) and is related to the early Proterozoic Trans-Hudson orogen (Hoffman, 1989) (Fig. 1). The Watts Group is the structurally highest tectonic unit in the belt, and comprises mostly ultramafic and mafic cumulate rocks and strongly foliated mafic rocks (St-Onge and Lucas, 1990). In one less deformed locality, pillowed volcanic rocks, sheeted mafic dykes and a plagiogranite intrusion have been recognized. The Watts Group contains all of the igneous crustal components of an ophiolite suite, albeit rather disrupted and deformed. It has been interpreted as obducted oceanic crust (the Purtuniq Ophiolite, Scott et al., 1989) despite the apparent lack of a tectonized mantle component.

The positive identification of ophiolitic rocks derived from oceanic crust in the ca. 2.0 Ga Cape Smith Belt would have considerable geotectonic significance. Ophiolites extend the record of oceanic lithosphere well beyond that in extant ocean basins, and provide the prime evidence for the operation of plate-tectonic processes in the past. Ophiolites should also constrain mantle evolution, as oceanic crustal structure and composition can be directly related to mantle temperature and composition (Klein and Langmuir, 1987; McKenzie and Bickle, 1988). However, ophiolite preservation appears to decrease exponentially with increasing age, such that ophiolites older than the Phanerozoic are rare. The oldest well established ophiolites are ca. 850 Ma (Arabian Shield, Pallister et al., 1988) with the exception of the ca. 1.96 Ga Jormua complex, Finland (Kontinen, 1987) and the Purtuniq ophiolite. The Purtuniq ophiolite, 1998 ± 2 Ma (Parrish, 1989) in age, presents a valuable opportunity to examine tectonic and magmatic processes which were active in the early Proterozoic. This report summarizes the results of a two week excursion in 1990 during which a detailed examination was made of localities which display relationships critical to the interpretation of the ophiolite. The objectives included detailed mapping of intrusive relationships in the sheeted dyke outcrops, study of mafic dykes cutting gabbroic cumulates, and detailed sampling profiles in the ultramafic cumulate rocks.

FIELD RELATIONSHIPS

The geology of the lac Watts and Purtuniq 1:50 000 scale map areas (St-Onge and Lucas, 1989a, b) is summarized in Figure 2. The meta-igneous components of the ophiolite comprise the Watts Group, which structurally overlies the deep-water facies sedimentary rocks of the Spartan Group, which in turn structurally overlies the pillowed basalts of the Chukotat Group.

Observations were made in the meta-volcanic sequence along the west side of lac Watts, and on a hilltop 5 km east of lac Watts on which the sheeted dyke complex is exposed (outlined with Locality A, Fig. 2). Plutonic rocks examined in detail include layered gabbros along the shores of lac Watts, a cumulate ultramafic body northeast of lac Watts, and the cumulate ultramafic and mafic rocks in the vicinity of rivière Déception. Specific

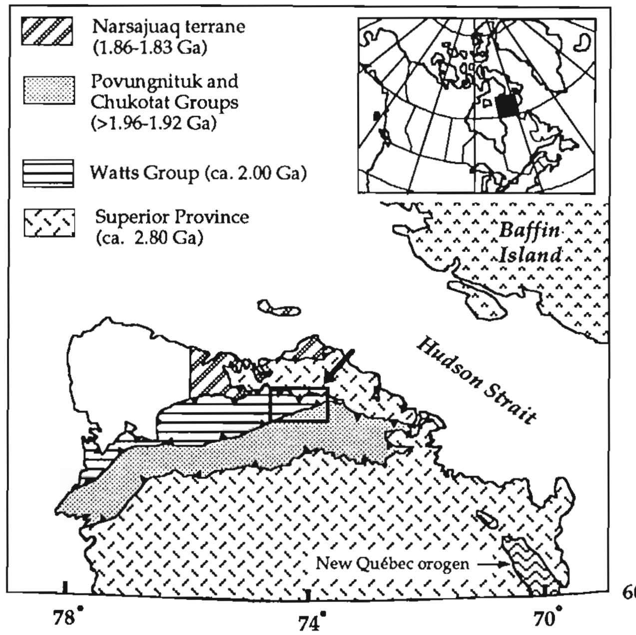

Figure 1. Map showing the location of the Cape Smith Belt, northern Quebec, Canada, with simplified tectono-stratigraphic units shown. Inset (upper right-hand corner) locates the main figure in northern Canada. The eastern portion of the ophiolite, as shown in Figure 2 and described in the text, is outlined in the main figure. From St-Onge and Lucas (1991).

locations subsequently referred to in the text are indicated on Figure 2. Field observations on the sheeted mafic dykes, volcanic rocks, and the layered cumulate rocks are discussed in the following sections.

Sheeted mafic dykes and pillowed metabasalts

Foliated mafic rocks form a large component of the Watts Group, and original volcanic textures are relatively rarely observed. However primary relationships between sheeted dykes and pillowed mafic volcanic rocks are reasonably well exposed on a prominent hill 5 km east of lac Watts (Fig. 2, locality A) as previously described by Scott et al. (1989). Due to the importance of 100% sheeted dyke complexes in providing compelling evidence for an origin in mafic oceanic crust, a section through part of this outcrop has been mapped in detail at 1:5000-scale (Fig. 3, 4).

Extensive outcrop occurs on the top and down the gently north-west sloping side of the hill, with the best exposures of the sheeted dykes occurring along the southeast margin of the hill (unit 1, Fig. 3). The steeper southeastern and northeastern unexposed slopes are scree-covered, obscuring the inferred location of the 'out-of-sequence' retrograde thrust which dips approximately 35° to the northwest under this tectonic unit (Lucas, 1989; St-Onge and Lucas, 1989a). Figure 3 illustrates the transition from 100% sheeted dyke outcrop with screens of massive finer-grained amphibolite and coarser-grained amphibolite, through a zone of alternating dykes, massive amphibolites and pillowed metabasalt upward into more continuous outcrop of deformed pillowed basalts.

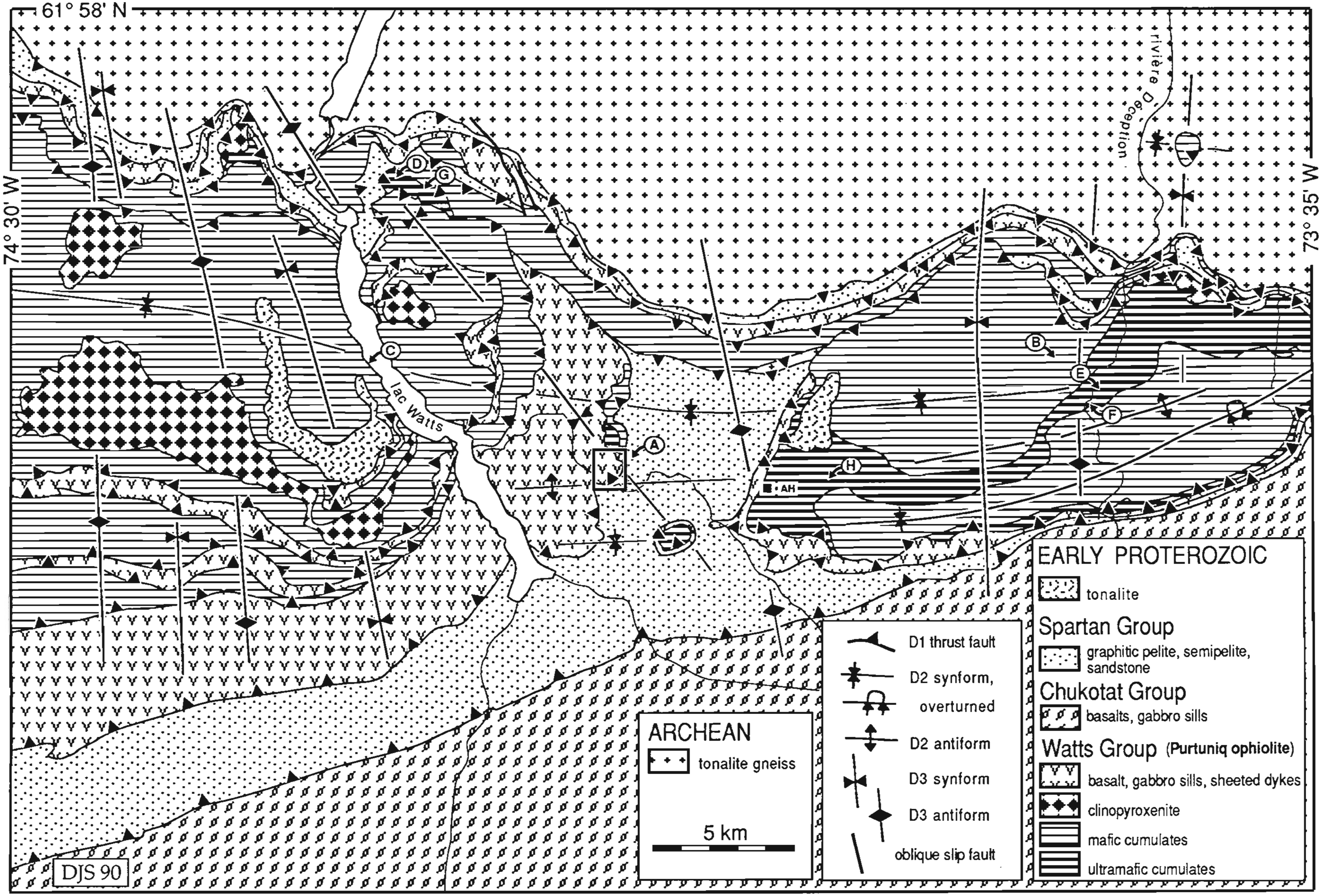

Figure 2. Generalized geology of the eastern portion of the Purtuniq ophiolite, Cape Smith Belt, northern Quebec. Location of this figure is outlined in Figure 1. Circled letters A-H identify locations referred to in text. AH refers to the former Asbestos Hill mine at Purtuniq. Modified from St-Onge and Lucas, 1989a, b.

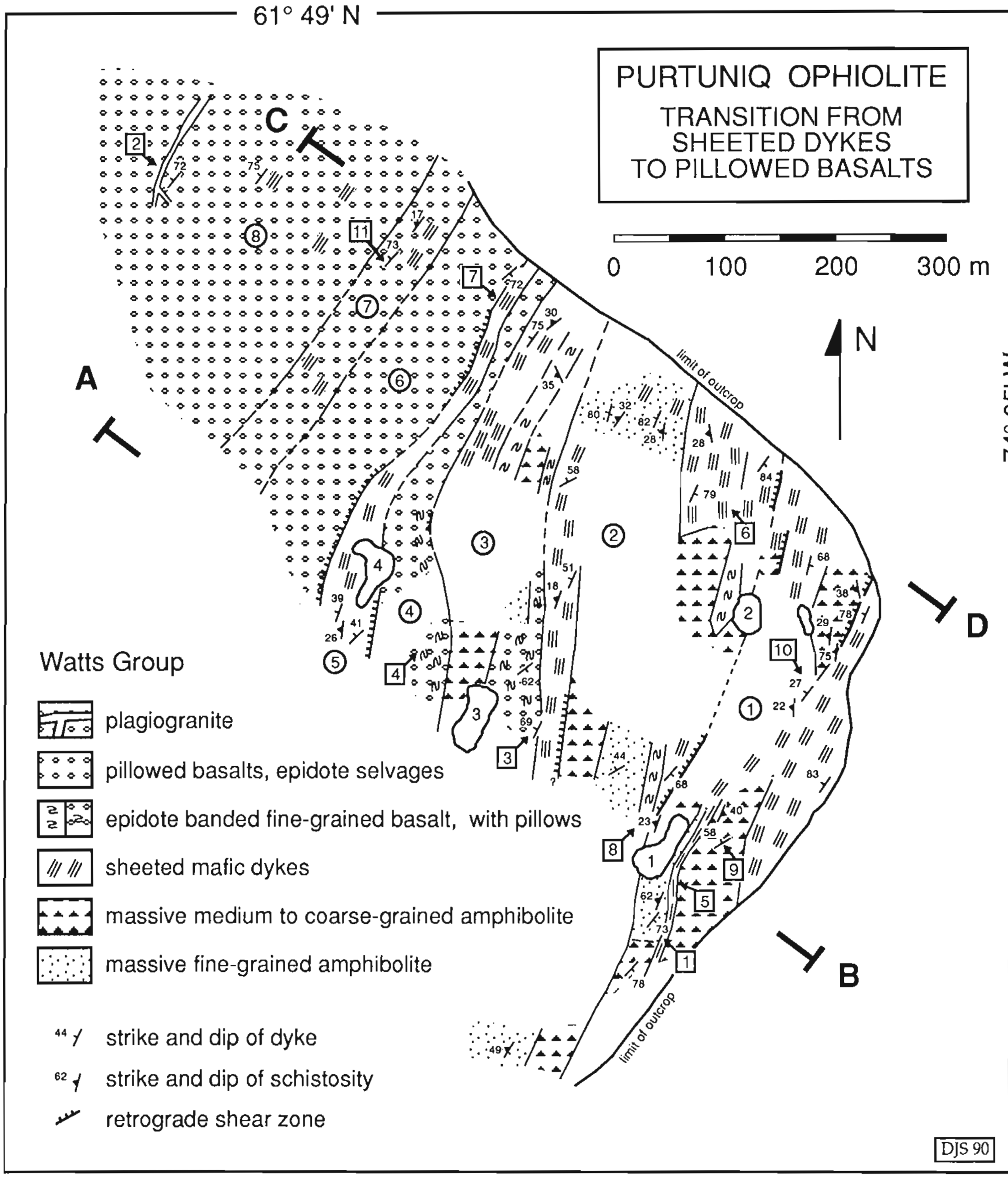

Figure 3. Detailed geological map of a Locality A (Fig. 2), a hill-top about 5 km east of lac Watts. Circled numbers refer to map units described in text. Numbers enclosed by squares are locations referred to in text. The four irregularly-shaped bodies numbered 1-4 are small lakes.

The sheeted dykes and massive amphibolites

The sheeted dykes are best exposed along the southeast margin of the hill (unit 1, Fig. 3) and crosscutting relationships at one locality are described in detail below. The northeast-striking and steeply dipping, 20 to 40 cm thick dykes comprise about 1 mm grained amphibolites in association with both massive amphibolites of similar appearance and coarser grained amphibolites. Lichen cover commonly obscures details on outcrops, and discrimination of dykes in the finer grained amphibolites often requires detailed, cm by cm searches for contacts delineated by grain size changes. An unknown amount of the fine grained amphibolite shown in Figure 3 may be sheeted dykes. The coarser grained amphibolite, which contains up to 10 × 5 mm, strained, darker-green amphibole porphyroclasts partly replaced by a finer grained, foliated matrix of pale amphibole, chlorite, epidote, plagioclase and calcite, may have been derived from a relatively coarse gabbro; these coarser amphibolites are interpreted as screens between the sheeted dykes.

Figure 5 illustrates crosscutting relationships in dykes over a 3.5 m section of previously cleaned, lichen-free outcrop (Fig. 5a, b: Locality 1 in Fig. 3). Dyke contacts are delineated by grain-size and petrographic differences. Chilling directions were inferred in the field from grain size fining in the metamorphic amphibolite assemblages, and are marked in Figure 5b where confirmed by observations made on thin sections. Two possible dyke contacts mapped in the field were revealed in thin section as a 1 cm wide chlorite- and actinolite-rich dykes cutting uniform amphibolites. The wider dykes typically show one-way chilling; whereas most of these observed "half-dykes" preserve only their northern margins, the population is far too small to imply a statistically valid paleospreading direction. The intrusive relationships allow determination of a relative sequence of dyke intrusion ages, but lithological differences are insufficient to allow correlation with other outcrops, with the exception of the relatively rare, uniformly late, more mafic dyke phase. The most common dyke lithology is an approximate

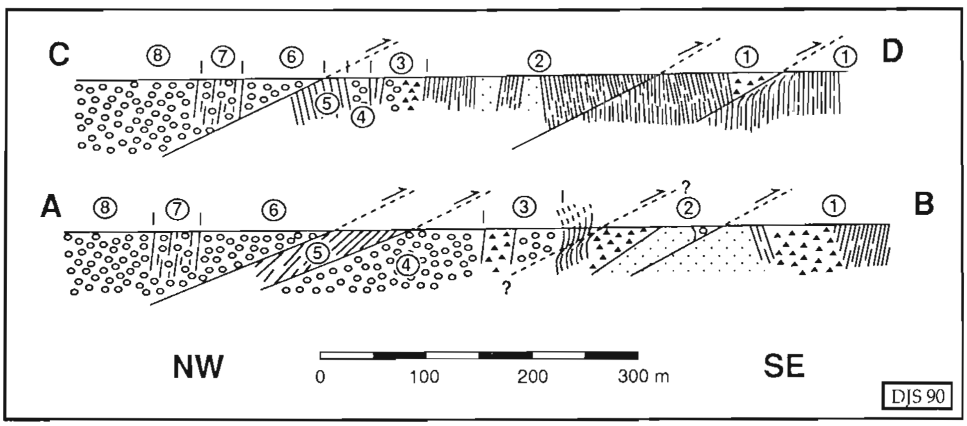

Figure 4. Schematic cross-sections drawn along lines AB and CD on Figure 3, described in text. Patterns as in Figure 3.

Figure 5a. Outcrop photograph and sketch of sheeted dykes with lichen removed. View looking NE. Locality 1 on Figure 3. Hammer is 35 cm long.

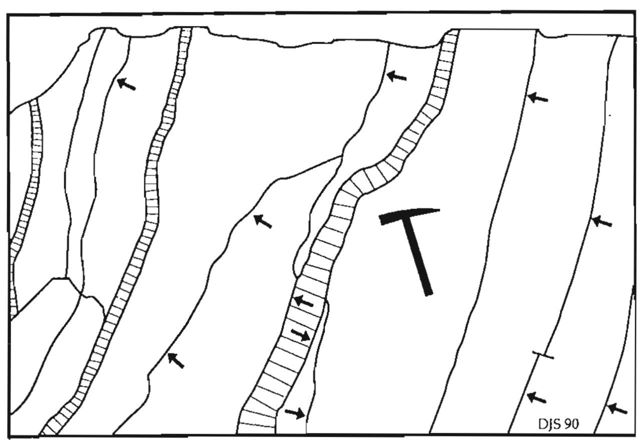

Figure 5b. Sketch of area shown in Figure 5a. Chilled margins are indicated with arrows in the direction of decreasing grain size. Crosscutting 'chloritic' dykes are indicated with striped pattern.

1 mm grain-size amphibolite in which green amphiboles (hornblende) are variably rimmed by colourless amphibole (actinolite) in a matrix of colourless amphibole, plagioclase, epidote, chlorite, and minor amounts of calcite and sphene. These dykes are subsequently referred to as 'amphibolite' dykes. This metamorphic assemblage is typical of the Watts Group in which a regional, thermal peak, amphibolite facies assemblages are partly retrogressed to greenschist facies (Bégin, 1989). The main petrographic differences between dykes are grain size variations and the presence or absence of coarser amphibole or plagioclase crystals, presumably pseudomorphing phenocrysts.

The younger, more mafic 'chloritic' dyke phase has a darker green colour and is characterized by a more chlorite-actinolite rich composition than the other dykes. These minerals often define a strong foliation in the late dykes. This phase is frequently intruded as thin, about 10-15 cm dykes at low but variable angles to the older sheeted dykes. Some of the less deformed examples contain about 5 mm chlorite-filled pseudomorphs, possibly after olivine. At two localities chlorite-rich dykes exhibited variolitic textures. Petrographically similar dykes were found widely dispersed across the entire map area (Fig. 3), and, significantly, cut the plagiogranite observed in the northwestern corner.

Pillowed basalts

Epidote-banded mafic schists represent the most abundant lithological unit in the map area of Figure 3. Where less deformed and well exposed, generally on sections parallel to schistosity, the epidote bands are seen to form the selvedges of pillows (Fig. 6). Preferential epidote growth in oxidized pillow selvedges has been documented in other metabasalt terranes (Holland and Norris, 1979). The more foliated or less-well exposed, strongly epidote banded mafic rocks are interpreted as derived from pillowed basalts. The pillows, 1 to 2 m long, are variably elongate in the S_1 foliation plane. The few facing directions determined from pillow shapes and very rare lava withdrawal shelves are always up to the northwest, consistent with the stratigraphic location of the sheeted dykes to the southeast. The attitude of the pillow units is poorly defined, but is inferred as approximately parallel to the S_1 cleavage, and perpendicular to the less foliated dykes.

The transition from sheeted dykes to extrusive mafic rocks

The mapped transition from predominantly sheeted dykes to pillowed metabasalts has been subdivided into units 1 to 8 in Figure 3 on the basis of informal lithological associations. Some of the contacts between units are recognized as tectonic (see below).

Unit 1 is defined as outcrops composed entirely of sheeted dykes with screens of fine- or coarser-grained amphibolite. The contact between units 1 and 2 is taken as a well-exposed retrograde shear zone approximately 25 m north of lake 1 (Fig. 3, photograph Fig. 7). Unit 2 contains well-exposed sheeted dykes on the northeastern margin of the hill and much undifferentiated massive amphibolite, particularly in the less-clean outcrop in the central parts. Rare metre-scale screens of foliated mafic schist occur in this unit and contain anastomosing, cm scale epidote-rich bands reminiscent of the better-preserved pillowed metabasalts described previously (units 4, 6, 7 and 8).

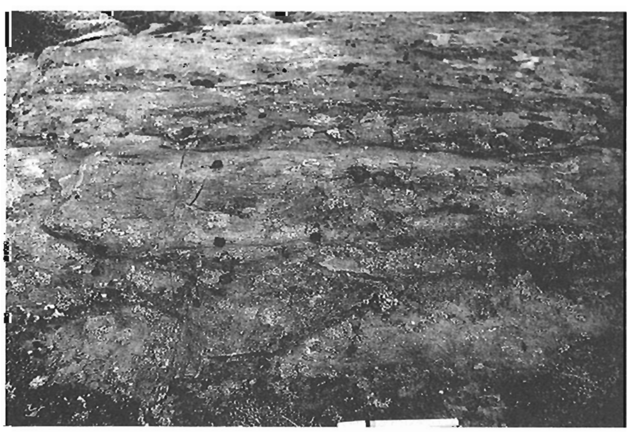

Figure 6a. Watts Group pillowed basalts, with selvedges outlined by epidote. Pen at bottom of photo is 15 cm long.

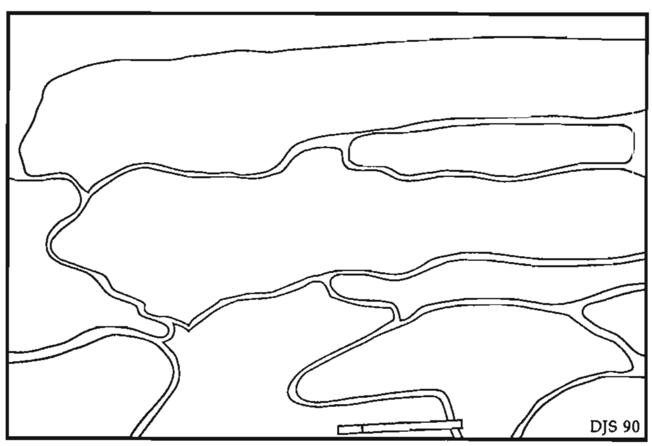

Figure 6b. Sketch of area shown in Figure 6a.

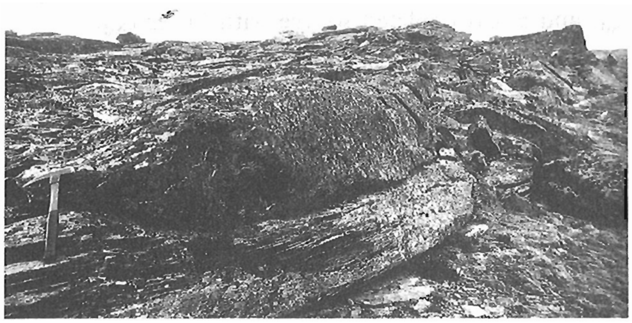

Figure 7. Massive amphibolite dyke (centre of photo) truncated by strongly foliated mafic rocks (foreground) with retrograde mineralogy, dipping 23° to the northwest (left side of photo). The approximate 50 cm wide dyke can be followed along strike (background, right) to an upright position, dipping 68° toward the southeast. Locality 8 on Figure 3. Hammer is 35 cm long.

The third unit, up to 100 m in width (Fig. 3), is dominated by epidote-banded, fine grained mafic rock, thought to have been derived from pillowed basalt. Sheeted dykes also occur in this unit, together with fine- and coarser-grained amphibolites which are intercalated with these epidote-banded mafic rocks. Most contacts are obscured, but a well-exposed locality 40 m east of lake 3 (locality 3) preserves primary intrusive relationships between a set of sheeted amphibolite dykes and epidote-banded, foliated mafic rocks cropping out to the northwest. The dykes occur on the northwest margin of a 25-50 m unit of sheeted dykes. The dykes truncate anastomosing epidote bands in the mafic schists at high angles. Similar alternations between sheeted dykes and screens of pillowed basalt are a characteristic component of the transitional zone at the top of sheeted dyke complexes in Phanerozoic ophiolites (eg. Coleman, 1977).

Unit 4 contains predominantly epidote-banded foliated, mafic rock. Outcrops 80 m northwest of lake 3 (Locality 4 on Fig. 3) exhibit recognizable pillow shapes outlined by epidote-rich selvedges facing up and to the northwest. Unit 5 comprises up to 50 m of sheeted dykes and is overlain by a dominantly pillowed metabasalt sequence. Units 6 and 8 comprise variably deformed mafic rocks in which about 5 cm epidote-rich selvedges outline recognizable pillow shapes. These are cut by rare, narrow crenulated chloritic mafic dykes. The deformed pillowed mafic rocks continue up-section for several kilometres (St-Onge and Lucas, 1989a). Unit 7 contains epidote-rimmed pillowed basalts cut by amphibolitic dykes which are locally sheeted in aspect. The pillowed volcanic rocks of Unit 8 are intruded by the plagiogranite described below.

Plagiogranite

One 60 m long and <1 m thick plagiogranite sill was mapped in the northern part of the area (Locality 2 in Fig. 3). The sill is fed by a plagiogranite dyke which crosscuts the pillowed basalts. In turn, the plagiogranite was crosscut by narrow dykes from the late 'chloritic' suite. The sill contains a foliation defined by fine grained plagioclase, quartz, and minor actinolite and muscovite.

Structure

Deformation fabric intensity is highly variable. The 'amphibolite' dykes often have only a weakly developed schistosity, whereas pillowed metabasalts display a well-developed schistosity parallel to regional S_1 (Fig. 3). The schistosity is probably a composite fabric, in that retrograde chlorite and actinolite growth along the fabric is thought to be associated with the development of post-thermal peak(post-S1) out-of-sequence thrusting (Lucas, 1989; St-Onge and Lucas, 1989a, b). In the more mafic, chlorite-rich dykes, the gently northwest-dipping penetrative fabric is often prominently crenulated about northwest-plunging axes with axial planes dipping steeply southwest. This crenulation has a similar orientation to D_3 structures mapped regionally by St-Onge and Lucas (1989a, b).

Sheeted dyke outcrops show a correlation between dyke dips and the intensity of foliation development, consistent with the bulk shear regime associated with the thrusting deformation. The least deformed dykes, such as those cropping out at Localities 1, 5, 6, and 7 (Fig. 3) are poorly foliated and dip between 70° and 90° to the southeast. With increasing foliation intensity, dykes rotate to dip at progressively shallower angles toward the northwest. The most strongly foliated and strained dykes dip between 27 and 40° toward the northwest, and have been rotated through angles of up to 90° (Fig. 8). In these localities (9, 10, and 11 on Fig. 3), the dykes remain clearly visible on foliation surfaces, but become very difficult to discriminate on surfaces perpendicular to the foliation. The association of dyke orientation and internal strain, the consistent rotation of the more deformed

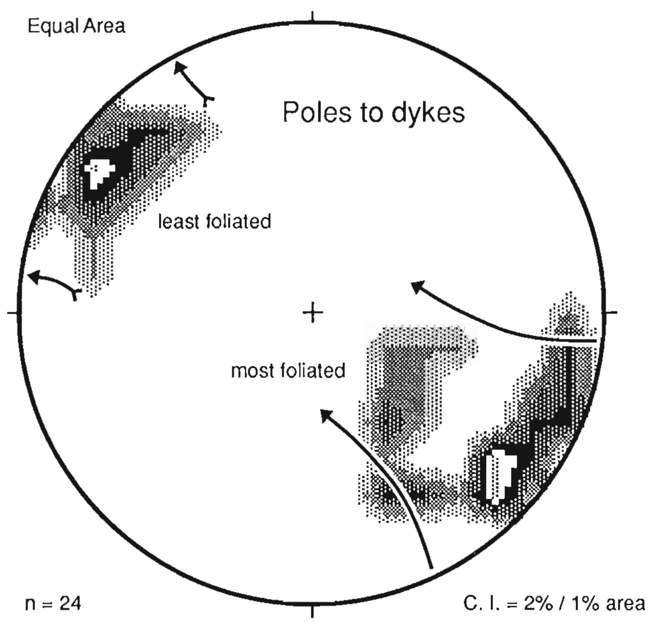

Figure 8. Equal area, contoured stereoplot of poles to dyke orientations.

dykes, and the maximum dyke rotations of 90° are all consistent with the observed higher strain zones being shear zones. The sense of rotation of the more deformed dykes implies a southeastward displacement of the hanging-wall. Such displacement is consistent with the regional direction of thrusting (St-Onge and Lucas, 1990). Several narrow and discrete shear zones were mapped (Fig. 3). The best exposed, at locality 8 (Fig. 3) truncates a dyke in the hanging-wall (Fig. 7) and contains about 0.5 m of strongly retrograded mafic schist with chlorite and actinolite-rich assemblages. These shear zones are inferred to have developed during the post-thermal peak out-of-sequence thrusting responsible for the major thrust which underlies the Watts Group in the area (Fig. 2). The relationship between these discrete retrograde shear zones and the larger-scale shear strains inferred from the dyke rotations is uncertain. The presence of retrograde mineral growth along the main penetrative fabrics associated with both dyke rotation and the retrograde shear zones suggests that dyke rotation and shear zone development are coeval. Further petrographic study is required to accurately constrain the relative timing of the shearing events.

Two interpretive sections through the map area (Fig. 3) are shown in Figure 4. Shear zones have been located on the basis of several criteria with varied degrees of confidence. The contact between Units 1 and 2 is taken as the well-exposed (Fig. 7) retrograde shear zone at locality 8 (Fig. 3) in the southwest, and may be tentatively joined to a small shear zone along strike from Lake 2 (Fig. 3). Foliated and rotated dykes at locality 10 (Fig. 3) within Unit 1 are interpreted to reflect further shearing (not shown in Figure 4) within this unit. A shear zone exposed at the base of the sheeted dykes in Unit 2 may explain variable rotations within the dyke complex. However at locality 3 (Fig. 3), an intrusive contact is preserved between sheeted dykes of Unit 2 and epidote-banded mafic schists of Unit 3. Similarly, no evidence was seen of a tectonic contact between Units 3 and 4. Shear zones are shown bounding Unit 5 because dykes are strongly deformed and rotated in the southwest. Dyke rotation occurs to a lesser extent along strike to the north. Further, the sinuous outcrop trace of the contact between Units 5 and 6 is consistent with the relatively shallow dip characteristic of the exposed tectonic contacts.

Due to the excellent exposure in the area covered by Figure 3, we are confident that all of the significant shear zones have been recognized. Consequently, the partly tectonized transition from sheeted dykes with screens of metagabbro, up through sheeted dykes and pillowed metabasalts, to the upper, most extensive unit of dominantly pillowed metabasalts (Fig. 4), is inferred to be largely primary in origin.

Coarse grained cumulate rocks

Coarse-grained cumulate-textured rocks are volumetrically the largest component of the Watts Group in the eastern part of the Cape Smith Belt. Foliated coarse metagabbroic rocks (Fig. 9, at Locality B on Fig. 2) form much of the cumulate sequence and are intercalated with distinctive bodies of layered dunitic, wehrlitic, and clinopyroxenitic ultramafic rocks. Scott et al. (in press) and Scott and Hegner (1990) show that the cumulate rocks are derived from source regions with two distinct Nd isotopic compositions, a more depleted source with higher ε_{Nd}, and a less depleted source with lower ε_{Nd}.

Well-exposed and well-foliated metagabbroic rocks with about 1 cm grain size crop out along the west shore of lac Watts. Their primary igneous mineralogy has been completely replaced by sodic plagioclase, calcic amphibole, and epidote (Bégin, 1989). Rare outcrops preserve pseudomorphed igneous layering which ranges from clinopyroxenite to anorthosite. However most outcrops are strongly foliated, and primary compositional layering is not apparent. Variably deformed, fine-grained amphibolite dykes form a small but distinctive component of the outcrops (Fig. 10, at Locality C on Fig. 2). If the metagabbroic rocks represent a deeper structural level than the sheeted dykes in the proposed ophiolite complex, then mafic dykes in the gabbroic rocks should correlate with the sheeted dyke complex. This assertion will be tested by analyzing chemical and Nd isotopic compositions.

The layered ultramafic rocks preserve rare relict igneous olivine and chromium-rich spinel, and abundant unaltered igneous clinopyroxene (Scott et al., 1989; in press). Igneous layering comprises alternations from dunite through wehrlite to clinopyroxenite on scales of centimetres to tens of metres (Fig. 11, at Locality D on Fig. 2). Two detailed sampling traverses were made through the layered ultramafic rocks at locations D and E (Fig. 2) to supplement preliminary examinations of primary igneous phases by Scott et al. (in press; see also Scott, in prep.). At location D northeast of lac Watts, 14 m were sampled at 0.8 m intervals, with 3.5 m sampled continuously. A 75 m section was sampled at less than 1 m intervals at the rivière Déception locality (Fig. 2, Locality E).

At the rivière Déception locality, the layered ultramafic rocks are overlain by a thin sequence of compositionally layered gabbros, and a sequence of fine grained

Figure 9. Medium grained layered gabbro; dark layers are metapyroxenitic composition, light layers are anorthositic. Locality B on Figure 2. Pen is 15 cm long.

(about 1 mm) amphibolites. The amphibolites are finer grained than most of the metagabbroic rocks in the area, and are at best only weakly compositionally layered at the outcrop scale. These rocks have been interpreted by Picard et al. (1990) as metavolcanic in origin, whereas St-Onge and Lucas (1989b, 1990), Scott et al. (in press) and Scott (1991) interpret them as fine grained metagabbros. Ultramafic layers centimetres to metres in thickness were observed in the layered gabbroic rocks and in the fine grained amphibolites, just above the contact with the underlying layered ultramafic cumulate rocks. This suggests that the layered gabbroic rocks and the fine grained amphibolites are are all part of a single, layered mafic-ultramafic body. The chemical and Nd isotopic composition of the fine grained amphibolites are indistinguishable from the underlying gabbroic and ultramafic rocks with which they appear to be in stratigraphic contact, and distinct from Watts Group metavolcanic rocks in other thrust sheets (Scott et al., in press; Scott, 1991). However, the significance of their uniformly fine grain size and the lack of any igneous structures in these rocks is still puzzling, and requires further study.

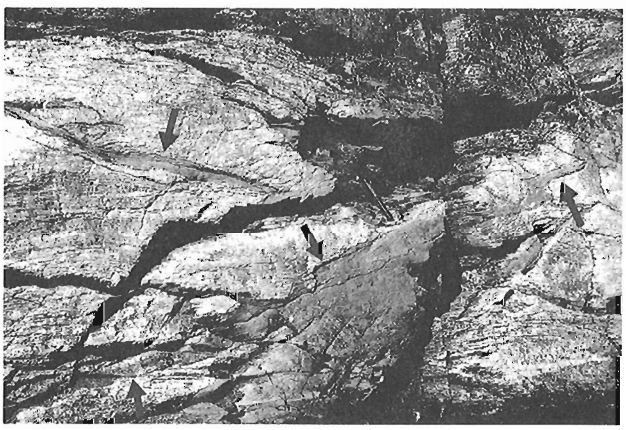

Figure 10. Fine grained mafic dykes (indicated by arrows) cutting layered gabbro, west shore of lac Watts. Hammer in centre of photo is 35 cm long.

A prominent but widely spaced set of mafic dykes striking north-south and northeast-southwest was mapped cutting the rivière Déception layered ultramafic rocks and overlying gabbroic and fine-grained amphibolitic rocks. These dykes are lithologically similar to the late chlorite-rich set which intrude the sheeted dyke complex.

ECONOMIC GEOLOGY

The ophiolitic nature of the rocks of the Watts Group suggests two types of mineralization of possible economic interest. Stratabound, massive base-metal sulphide deposits are common in the pillowed volcanic sequences of numerous Phanerozoic ophiolites (eg Troodos, Cyprus; Samail, Oman; Bay of Islands, Newfoundland; see Coleman, 1977 for a review). Podiform chromite lenses are associated with tectonized dunitic rocks in the preserved mantle sections of numerous younger ophiolites (Coleman, 1977).

Evidence of base-metal sulphide mineralization, such as development of gossan zones, was not observed in the mafic volcanic section of the Watts Group in the study area during 1:50 000 scale mapping (eg. St-Onge et al., 1987; 1988, and references to previous workers within). Detailed aeromagnetic surveying of the volcanic sequence may be a useful exploration technique for locating sulphide bodies not exposed at surface.

The absence of preserved mantle material in the Purtuniq ophiolite suggests it is unlikely that classical podiform chromite deposits are preserved. However, stratiform layers of chromite have been observed in the cumulate ultramafic rocks at several localities (F, G, and H, Fig. 2). The layers are locally up to 20 cm thick, and although disrupted, can be traced along strike for tens of metres (Fig. 12, at locality F in Fig. 2). Despite the relatively small volumes of the chromite concentrations in the cumulate ultramafic rocks, electron microprobe analysis indicates that the grains are chromium-rich (Scott et al., in press; Scott, 1991).

The deposit of asbestos at Asbestos Hill is no longer considered an economic commodity.

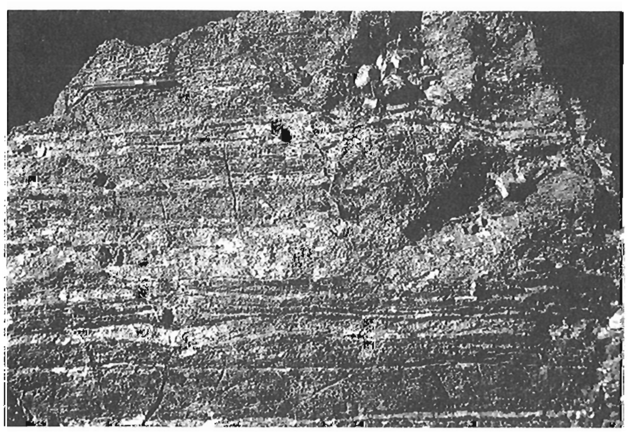

Figure 11. Layered cumulate ultramafic rock. Darker bands are clinopyroxene-rich, lighter bands are dunitic. Locality D on Figure 2. Pen in upper left-hand corner is 20 cm long.

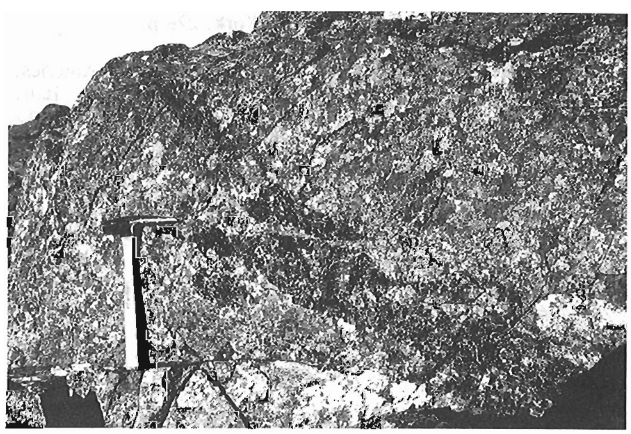

Figure 12. Dark black bands to the right of the hammer are monomineralic chromite seams. Locality F on Figure 2. Hammer is 35 cm long.

DISCUSSION

The association of pillowed metabasalts, rare plagiogranite, sheeted dykes, the transition from sheeted dykes into pillowed basalts, coarse gabbroic and ultramafic cumulate rocks, and mafic dykes cutting the gabbroic rocks is consistent with previous assertions that the Watts Group represents the crustal portion of a dismembered early Proterozoic ophiolite (St-Onge et al., 1987, 1988; Scott et al., 1989; St-Onge and Lucas, 1990). The structural setting of the Watts Group in the internal part of an orogenic belt emplaced on the northern margin of the older Superior province continental crust (Hynes and Francis, 1982) is consistent with this interpretation. The only lithological unit missing from a complete ophiolite sequence is a tectonized peridotite which has undergone high-temperature mantle deformation.

The chemical complexities revealed by Scott et al. (in press) suggest a complex magmatic evolution, but such complexity is characteristic of many ophiolite bodies (eg. Troodos and Samail, see summary by Coleman, 1977). Geochemical studies to constrain further the relationships of units within the Watts Group and their magmatic evolution are in progress.

ACKNOWLEDGMENTS

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Detailed gravity traverses across the early Proterozoic Ungava orogen, northern Quebec

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Abstract

Approximately 320 km of detailed gravity profiling (stations 1 to 3.5 km apart) has been completed within the eastern Ungava orogen in support of ongoing geological studies. The new gravity data define more precisely a large positive gravity anomaly associated with the Cape Smith Belt (CSB), outlined by older regional surveys. A principal objective of the gravity program is to refine earlier gravity models of the CSB, in the light of current geological knowledge. The detailed gravity data will enable the crustal structure of the CSB predicted from down-plunge constrained, structural cross-sections to be tested. In addition, the new data will be used to examine structure within the flanking Superior Province and Narsajuag terrane. A northward increasing gravity gradient within the Narsajuaq terrane is attributed to a northward increase in metamorphic grade, culminating in granulite grade rocks exposed north of Sugluk Inlet.

Résumé

Pour appuyer les études géologiques en cours, on a réalisé dans l'est de l'orogène d'Ungava des profils gravimétriques détaillés sur environ 320 km (1 à 3,5 km entre les stations). Les nouvelles données gravimétriques ont permis de définir avec plus de précision une grande anomalie gravimétrique positive associée à la zone de Cape Smith délimitée par des levés régionaux antérieurs. Le programme gravimétrique a pour principal objectif de perfectionner les modèles gravimétriques de cette zone à la lumière des connaissances géologiques actuelles. Les données gravimétriques détaillées permettront de vérifier la structure crustale de cette même zone établie à partir de coupes structurales restreintes suivant le plongement. De plus, les nouvelles données serviront à analyser la structure au sein de la province du lac Supérieur et du terrane de Narsajuaq adjacent. On attribue l'accroissement vers le nord du gradient gravimétrique au sein du terrane de Narsajuaq à une augmentation vers le nord du métamorphisme, culminant en roches du faciès des granulites exposées au nord de l'inlet Sugluk.

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INTRODUCTION

The early Proterozoic Ungava orogen, which is an integral part of the Trans-Hudson orogen (Hoffman, 1981, 1988), has been defined recently by St-Onge and Lucas (1990a). It comprises 3 fundamental tectonic elements: (1) Superior Province (Archean) basement, (2) the Cape Smith Belt, and (3) the Narsajuaq terrane (Fig. 1). The Cape Smith Belt forms part of the Circum-Superior Belt of earlier authors (e.g. Baragar and Scoates, 1981).

Prominent gravity anomalies (Fig. 2) associated with the Cape Smith Belt and its hinterland, the Churchill Province, were interpreted by Thomas and Gibb (1977) to signify a continental collision zone. The collisional suture was presented in a gravity model in part as a major density discontinuity extending to the base of the crust, and in part as the Cape Smith Belt (Fig. 3). Near the eastern extremity of the belt, the suture was described as cryptic. Hoffman (1985) modified this model by suggesting that the suture lay not at the northern margin of the Cape Smith Belt, as suggested by Thomas and Gibb (1977), but farther north within the Churchill Province. In recent years, a mapping program carried out within the Cape Smith Belt and its northern hinterland has greatly expanded knowledge on the structure and tectonic history of the Ungava orogen (e.g., St-Onge et al., 1988; Lucas, 1989; St-Onge and Lucas, 1990a, b). The collisional model is now well established; St-Onge et al. (1988) having documented rift-fill, transitional-crust and ophiolite suites, which record a history of ocean opening and closure.

Downplunge-constrained, strike-perpendicular cross-sections of the eastern part of the Cape Smith Belt have been constructed by Lucas (1989). The downplunge sections are made possible by at least 18 km of structural relief present at the east end of the belt. The validity of the derived sections has been supported by comparison with crustal models derived from detailed gravity profiles (Feininger et al., 1985), which show relatively close agreement. Independant insights into the third dimension of the crust offered by the gravity method were considered a useful adjunct to ongoing geological mapping north of the Cape Smith Belt, and in summer 1990, gravity surveys were incorporated into the program. Three traverses, totalling 320 km, were completed. Preliminary data resulting from the surveys are presented in this report.

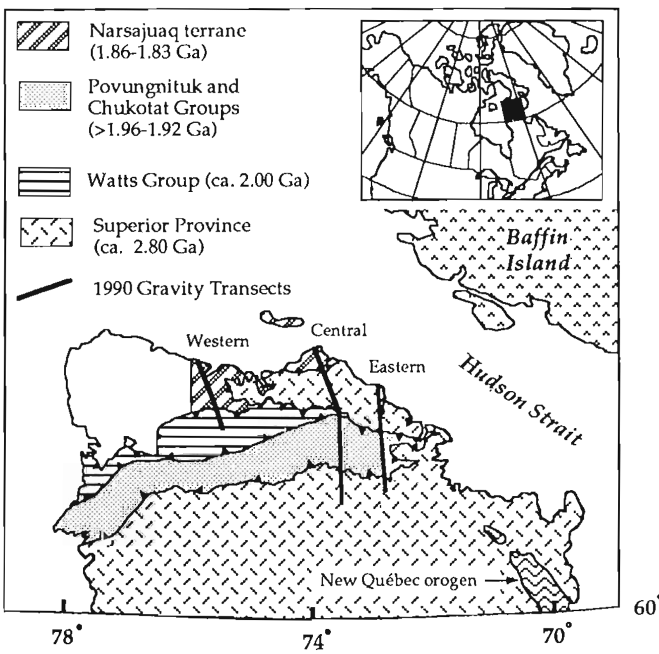

Figure 1. Simplified geological map illustrating the main tectonic elements of the Ungava orogen (modified from Lucas and St-Onge, 1991). Positions of the three detailed gravity traverses are indicated.

GRAVITY TRAVERSES

Along all three traverses, gravity stations were spaced at relatively close intervals in anticipation of defining details of the gravity field that could be directly related to smaller scale aspects of the geology. Intervals ranged generally from about 1 to 3.5 km, with the nominal spacing being about 2 km. The traverses are referred to as western, central and eastern (Figs. 1, 2).

The western traverse, 75 km long, trends south-southeast across the central part of the orogen from a point near Cap Tiblemont on the coast of Hudson Strait (Figs. 1, 2). It crosses both the Narsajuaq terrane and the northern margin of the Cape Smith Belt. The central traverse, 150 km long, is located near the eastern end of the orogen (Figs. 1, 2). It extends from Cap Buteux southwards across the Narsajuaq terrane, Superior Province basement rocks, the Cape Smith Belt, and finally ends on the Superior Province. The 95 km long eastern traverse commences in the north on Promontoire de Martigny and extends southward, crossing from the Superior Province basement into the Cape Smith Belt, and back again into the Superior Province basement south of the belt (Figs. 1, 2).

Transport for the gravity surveys was provided by helicopter. A Lacoste and Romberg Geodetic gravity meter, No. 172, was used for the measurements. All observations were tied to local base stations, which had been established by control ties to base stations of the National Gravity Network. A second Lacoste and Romberg gravity meter, No. 932, was employed for the ties. Vertical control was provided by altimetry tied to Topographical Survey monuments or lake elevations established by airborne profiling radar surveys. Horizontal positions were determined in most cases using visual fixes on 1:50 000 scale topographic maps. A global positioning system was used in a few cases where map determinations were not possible. Bouguer gravity anomalies were computed using a uniform crustal density of 2.67 g/cm^3 and sea level datum. Terrain corrections for gravity stations have not yet been computed, but are estimated to be generally less than 2 mGal. Not including terrain corrections, Bouguer anomalies are estimated to be accurate to within ± 2 mGal.

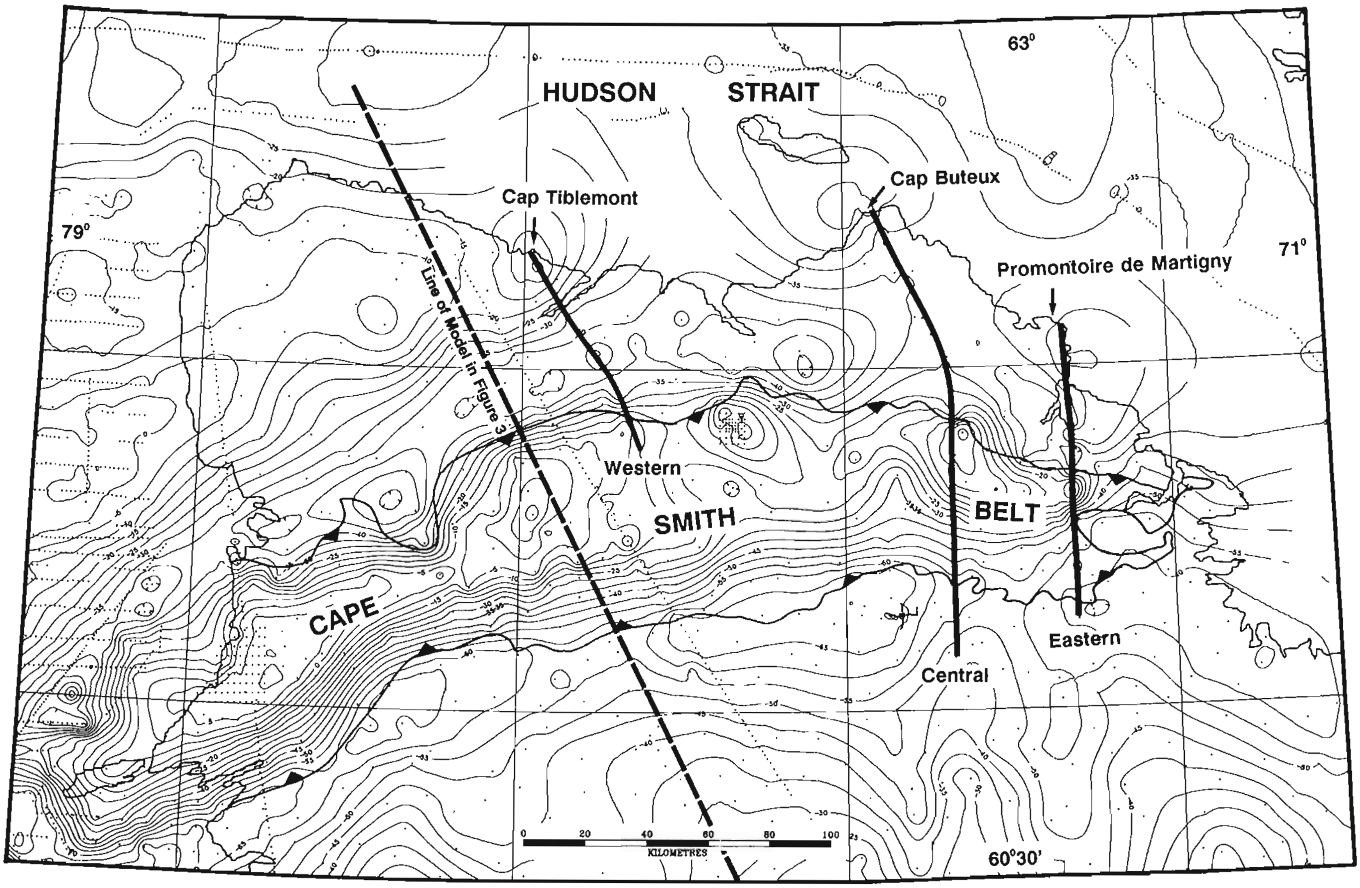

Figure 2. Bouguer gravity field over the Ungava orogen; contour interval is 5 mGal. Positions of new detailed gravity traverses are indicated by heavy solid lines. Line of gravity model in Figure 3 is indicated by a dashed line.

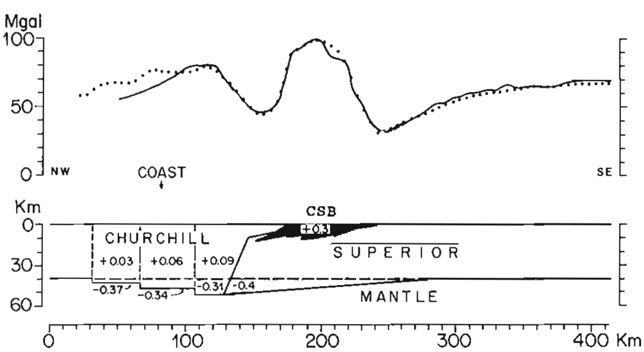

Figure 3. Gravity model across the Ungava orogen after Thomas and Gibb (1977). CSB, Cape Smith Belt (solid black unit). Density contrasts are in units of g/cm^3.

PRELIMINARY RESULTS

The preliminary results of the gravity surveys are presented simply as Bouguer gravity profiles accompanied by the principal surface tectonic domains along the traverses. The eastern (Fig. 4) and central (Fig. 5) traverses, which cross the entire exposed width of the Ungava orogen in the Ungava Peninsula display a broadly similar pattern of anomalies. Both are characterized by strong positive anomalies over the Cape Smith Belt, reflecting the denser nature of the constituent rocks, which are largely mafic and ultramafic, and by generally flat and featureless gravity fields over the Superior Province and Narsajuaq terrane. A mean density calculated for rock samples from within the belt is 2.92 g/cm^3, which compares with mean densities of 2.69 g/cm^3 and 2.77 g/cm^3, for felsic plutonic rocks to the south and north of the belt, respectively (T. Feininger, pers. comm., 1990). The samples of plutonic rocks used for density determinations collected north of the Cape Smith Belt are exclusively from the Narsajuaq terrane.

The difference in densities of the basement to either side of the belt is also manifested in the gravity profiles, the gravity field to the north being at a higher level than that to the south. Thomas and Gibb (1977) attributed the change in level to differences in mean crustal density and crustal thickness, crust north of the Cape Smith Belt being modelled as thicker and denser than crust to the south (Fig. 3).

The eastern profile (Fig. 4) contains a prominent positive anomaly in the position of a narrow outlier of the Cape Smith Belt. The outlier is located at the southern extremity of Southeast Arm of Douglas Harbour (Lucas and St-Onge, 1991), and comprises metasedimentary

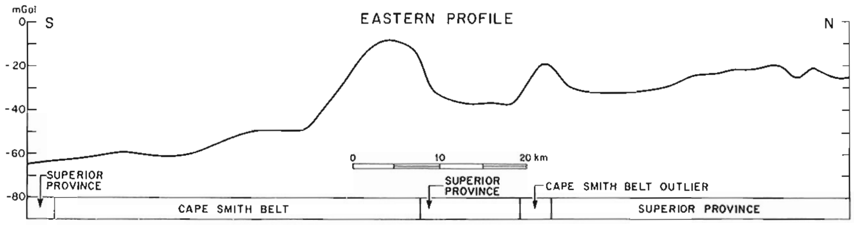

Figure 4. Gravity profile and principal tectonic elements along the eastern traverse.

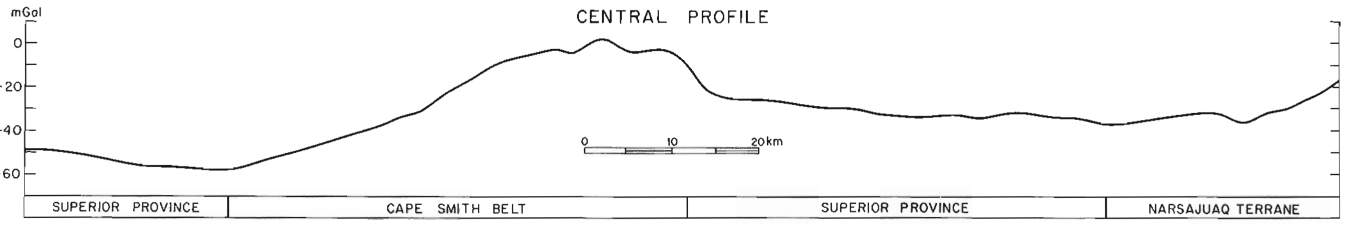

Figure 5. Gravity profile and principal tectonic elements along the central traverse.

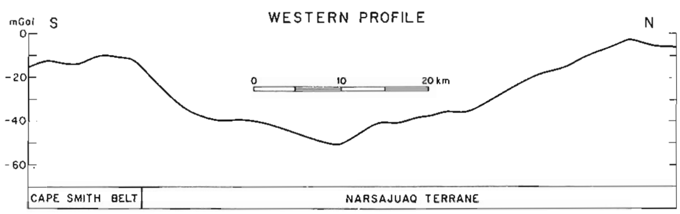

Figure 6. Gravity profile and principal tectonic elements along the western traverse.

rocks, gabbro, basalt and peridotite. A preliminary estimate of its depth, based on the amplitude of the gravity anomaly (15 mGal) and assumed densities, is between 1 and 2 km. The association of this unit with a distinct gravity anomaly provides an opportunity to model its third dimension.

It is noted that the northern flank of the positive anomaly associated with the Cape Smith Belt extends northward just beyond the boundary of the belt in all profiles. This feature is consistent with the northern contact of the belt being overturned to the south and dipping steeply northward, as documented geologically by St-Onge et al. (1988), Lucas (1989) and St-Onge and Lucas (1990a). A similar relationship between the positive anomaly and the Cape Smith outlier (Fig. 4) implies that the structure of the northern margin of this unit mimics that of the main belt.

The western profile (Fig. 6) contrasts with its central and eastern counterparts in that the gravity field north of the Cape Smith Belt is not as flat, and rises to levels higher than that attained in the belt itself. From a point near the centre of the Narsajuaq terrane, the field rises gently at first, both southward and northward. It becomes steeper southward as it approaches the Cape Smith Belt, and also becomes steeper northward from a point near an internal fault within the Narsajuaq terrane, and continues to rise to a peak near the coast. The northward increase in the gravity field and its gradient are probably related in part to a concomitant northward increase in metamorphic grade, which culminates with the development of granulite grade rocks. The change in slope of the gravity profile at the fault offers the possibility of modelling the subsurface extension of the fault.

FUTURE PLANS

Densities will be measured on rock samples collected from the region to provide a constraint for modelling studies. The latter will proceed as a collaborative venture between the project geologists and geophysicists. The principal objective is to derive quantitative gravity models of the crust across the entire orogen. Such models will in turn provide additional guidelines for structural analysis and establishing tectonic models.

ACKNOWLEDGMENTS

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Orthogneisses, paragneisses and high strain zones on northeast Coats Island, District of Keewatin, N.W.T.

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Schau, M., Orthogneisses, paragneisses and high strain zones on northeast Coats Island, District of Keewatin, N.W.T.; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 195-203, 1991.

Abstract

Northeast Coats Island is at the apex of an uptilted block formed along the southern edge of a Cretaceous graben. It is underlain by gneiss that is basement to nearby Palaeozoic carbonates. The gneiss complex includes early, shallow dipping, granitic to granodioritic orthogneisses and pelite, marble, and quartzite bearing paragneisses, which were intruded by a metagabbroic suite and later subjected to intrusion of granite and local development of migmatite. Subsequently, a middle to lower amphibolite grade regime of steeply dipping, northeast-trending, high-strain zones was imposed on the region, rotating a large proportion of the previous shallowly dipping gneissosity into steep, north east trending orientations. A later, east-west trending, steeply dipping, kilometre wide, high strain zone has been superimposed on the complex. The mineral potential of the gneiss complex is low to very low.

Résumé

Le nord-est de l'île Coats se trouve au sommet d'un bloc incliné vers le haut formé le long de la bordure méridionale d'un graben crétacé. Cette zone repose sur un gneiss qui est également le socle des carbonates paléozoïques voisins. Le complexe gneissique comporte des orthogneiss granitiques à granodioritiques précoces à pendage peu profond ainsi que de la pélite, du marbre et des quartzites contenant des paragneiss qui ont été recoupés par une suite métagabbroïque et ont été, ultérieurement, affectés par des intrusions de granite et par la formation locale de migmatite. Par la suite, un régime métamorphique du faciès d'intermédiaire à inférieur des amphibolites dans des zones très déformées à direction nord-est et à fort pendage a affecté la région, faisant subir une rotation à une grande partie de la structure gneissique antérieure à pendage peu profond de telle sorte que les orientations ont une direction nord-est et sont abruptes. Une zone fortement déformée d'environ un kilomètre de largeur à pendage abrupt et à direction est s'est par la suite superposée au complexe. Le potentiel minéral du complexe gneissique varie de faible à très faible.

INTRODUCTION

A reexamination of the Precambrian geology of Coats Island in the light of new regional geological mapping and geophysical surveys provides new constraints for models of the complex geological history of northern Hudson Bay. While previous workers generally agreed that major east trending structures with pronounced geophysical expression underlay the region, much debate centred on their geological nature. Recent examinations of subaqueous bedrock distributions in Hudson Bay (Sanford and Grant, 1990) have shown the presence of grabens with Cretaceous fill coincident with geophysical breaks. The extent to which the geophysical features reflect Precambrian rock distributions and structures as opposed to Phanerozoic distributions requires re-evaluation. Data from this study will provide some geologic constraints on models of the development of northern Hudson Bay region. Concurrently, a pilot study was successfully completed showing the feasibility of generating digital working and compilation maps during the field component of a project.

GEOLOGY

Regional

East trending second-derivative gravity anomalies in the NE Arctic are prominent (Goodacre et al, 1988). One, traversing northern Hudson Bay, is roughly coincident with the "sapphirine line" (Schau 1982), and possibly, in part, related to a Chesterfield Fault Zone (Heywood and Schau 1978) and/or a Snowbird Tectonic Zone (Hoffman 1988), and is near anomalous or thinned continental crust (Barr,1969).

Anorthositic gabbros are exposed along this same east trending zone from the Kramanituar complex near Baker Lake (Schau and Hulbert, 1977, Schau and Ashton, 1979, Schau et al 1982), through the Hanbury Complex (Tella and Annesley, 1988), and Daly Bay complex (Gordon, 1988) near the coast, to Southhampton, Coats and Walrus Islands in Hudson Bay (Heywood and Sanford, 1976) (Fig. 1). On land this zone is manifested by an important positive gravity anomaly (Gibb and Halliday,1976).

In Hudson Bay, south of Southampton Island, a steep gradient in the gravitational field is situated over water (Goodacre et al.,1988), lying subparallel with the Bell Arch (Sanford and Grant, 1990) and the gravity low is coincident, in part, with a Phanerozoic graben, on the southern flank of which Coats Island, Bencas Island, and Walrus Island (and the matching positive gravity anomaly) are situated (ibid). Cretaceous sediments occur in this graben sited in Evans Strait, bounded to the south by faults, adjacent to and subparallel with the west, north west striking coast line of the island (ibid).

Although the general trends of geology from the mainland project easterly to Coats Island, structures mapped by Heywood and Sanford (1976) suggested northeasterly trends, and they considered the gneiss units to be most like those to the north, on Southhampton Island.

Coats Island

The northeast corner of Coats Island is part of a tilted fault block of exposed Precambrian gneisses (Heywood and Sanford, 1976), rising several kilometres above similar, subsurface basement. From the apex at the northeast extremity of the island, the top of the exposed gneiss complex slopes evenly to the west and south, a remnant of a Cambrian peneplain, to the Palaeozoic carbonate sediments that cover the majority of the island (fig. 2). The crystalline part of the island is cut by steep valleys which are largely glacial incisions and modifications of fault-related joint sets.

In their memoir, Heywood and Sanford (1976) distinguished rocks of their unit 3 (layered quartzofeldspathic gneisses) from their unit 5 (migmatites) on Coats Island and suggested that future work would show regional structures on the island.

In this report, their unit 3 (ibid) is subdivided into paragneiss and orthogneiss; their unit 7 (ibid), (ultramafic and anorthositic gabbros) are lumped as metagabbroic suite; and their unit 5 (ibid), (migmatite) is discussed as migmatites and granites. Their unit 6a (ibid), (quartz monzonite) may be equivalent with some of the unit called granite in this report. The present study located distinctive marble, quartzite and graphite pelite remnants in the paragneisses, and tectonic slivers of a widely distributed metagabbroic suite; both assemblages are recognized in the gneisses on Southhampton Island as unit 2 and unit 7 respectively.

Gneissic Complex

Paragneisses

In a large area near the unconformity (see Fig 2) paragneisses with abundant biotite, and cordierite-garnet schists, garnet granofels, as well as more common hornblende-bearing granodioritic gneisses have varied, commonly shallow dipping orientations. Where layering is preserved, it is steeper than a cross-cutting shallow-dipping cleavage fabric. These fabrics are preserved in cross cutting metagabbro contact zones and thus predate the intrusion of gabbro. Near granites, graphitic sillimanite-bearing, garnet free, biotite schists , are locally abundant.

In two superposed high strain zones (see below) the planar fabric orientations are steep, and the paragneisses occur mainly as lenses aligned parallel with the general trend shown by the country rock. In a large lens, shown on figure 2, hornblende and biotite rich quartzofeldspathic gneiss are intercalated with thin quartzite, marble, and biotite-sillimanite schists as well as thicker garnetiferous granofels. A several metre thick marble horizon, containing many complex and detached folds developed on interlayered calcsilicate gneiss layers, has been traced for about 3 kilometres along strike within the "earlier" high strain zone. A horizon of planar layered quartzite, a few metres thick overlaps in part with the marble and was traced intermittently for an additional kilometre along the same zone. Under the microscope the quartz-

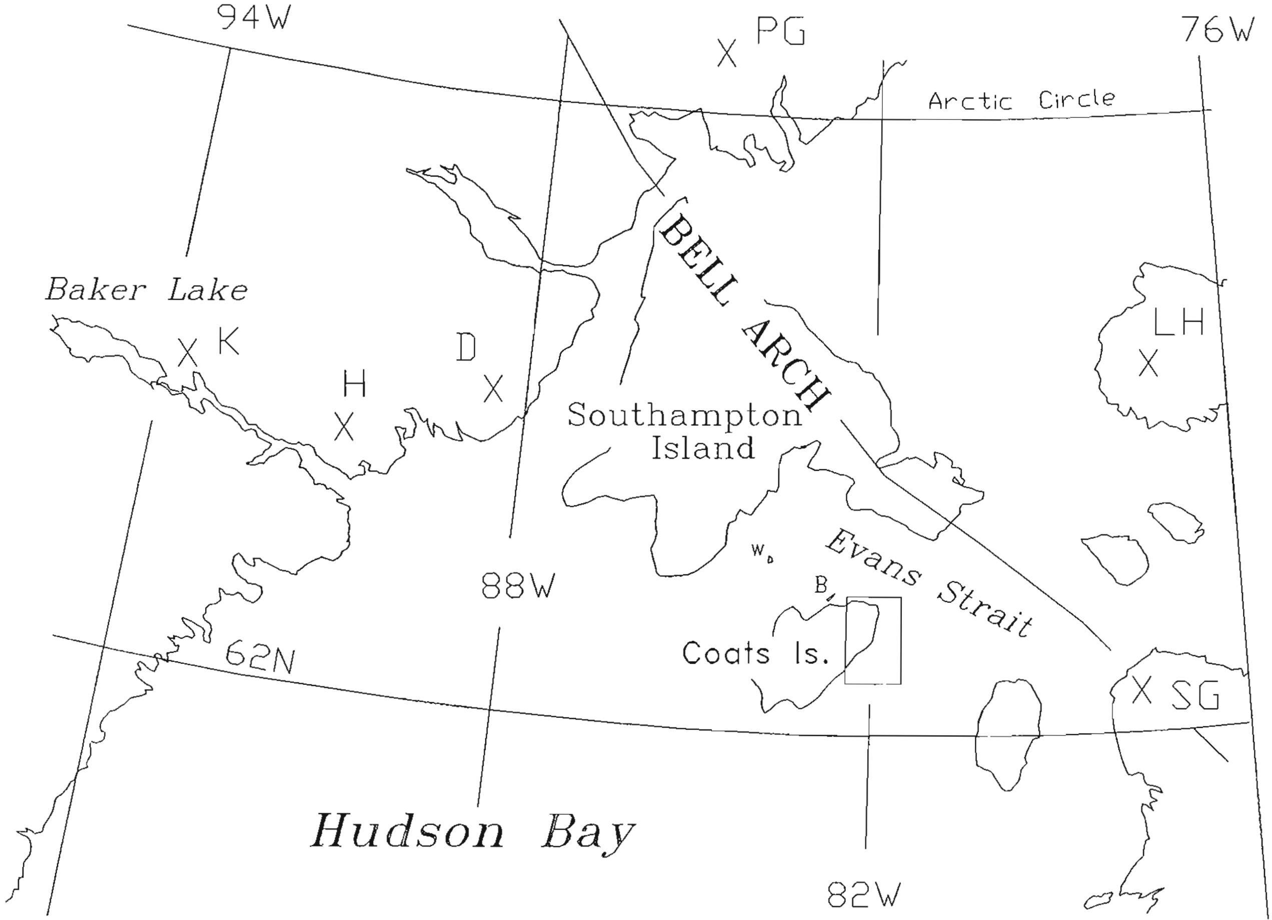


Figure 1. Location Map. X marks localities of rock units mentioned in text: K Kramanituar Complex, H Hanbury Island Shear Zone, D Daly Bay Complex, PG area of Penrhyn Group, LH area of Lake Harbour Group, SG area of Sugluk Group. The Bell Arch is a broad tectonic element and the trace on the map is generalized. The small capitals without x indicate the location of two small islands; W Walrus Island, and B Bencas Island. Rectangle outlines Figure 2.

ite can be seen to include small amounts of microcline and muscovite with minor plagioclase and biotite and a few scattered garnets. The granofels is seen in thin section to consist of garnet and quartz with minor amounts of plagioclase, magnetite and accessory biotite and apatite, zircon, and sphene. Many smaller lenses of paragneiss and associated metagabbro are found in the high strain zones. Within the later high strain zone the lenses are metre sized, and only extreme compositions are readily assigned to the paragneiss; such as the calc-silicate, biotite-sillimanite schist, or quartzite boudins.

Although the paragneisses locally contain dense rocks, with specific gravities in excess of 3.0 being common among the garnetites and calcsilicate marbles, the volume of these exposed rocks is considered insignificant, and is unlikely to account for regional gravity anomalies.

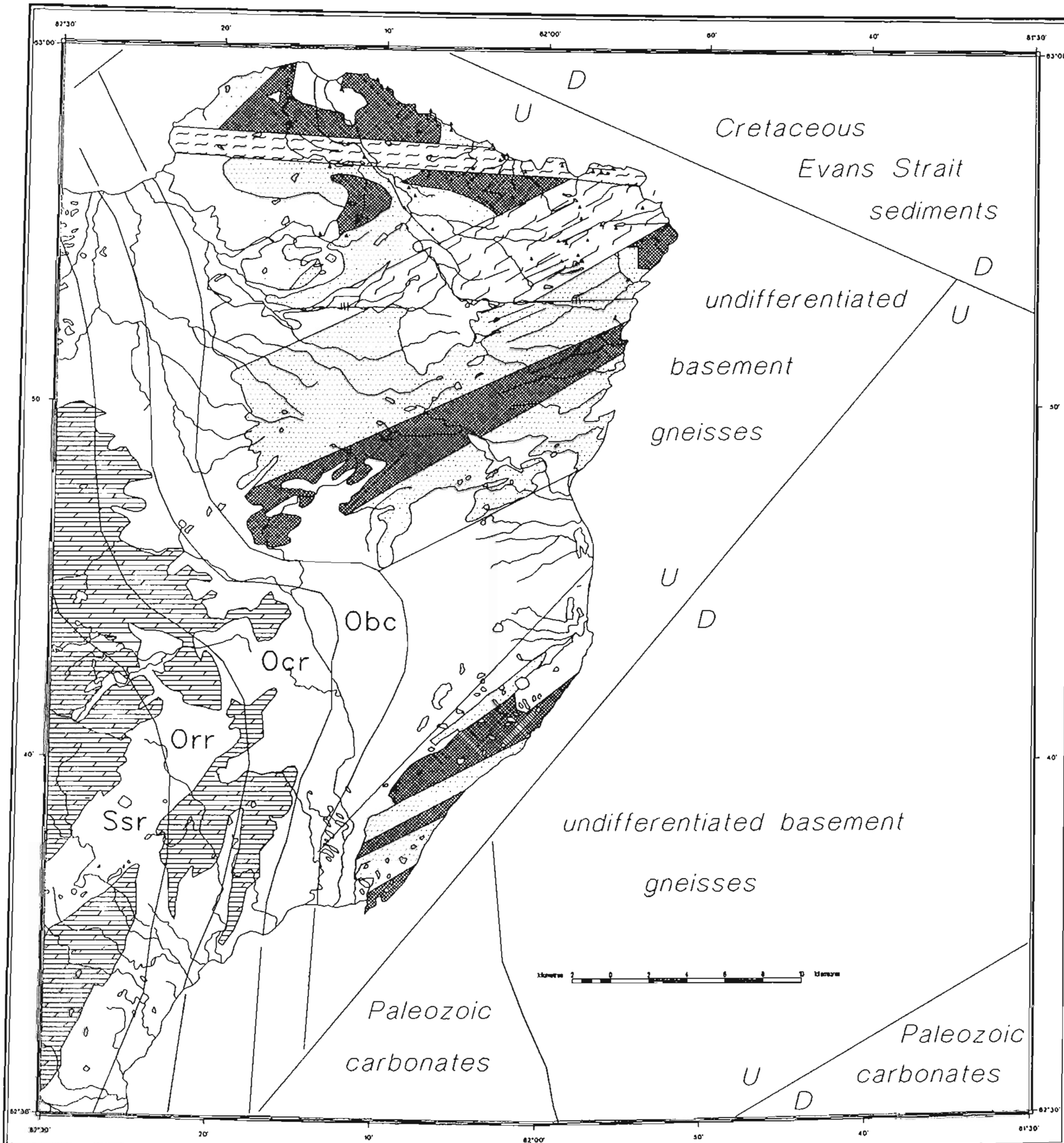
Orthogneisses

The orthogneisses are grey, foliated, biotite bearing quartzofeldspathic rocks. Variation in grain size, abundance of biotite, and presence of small amounts of hornblende constitute the principal variations in the unit. The rock is more foliated than layered, sometimes with a well defined lineation.

Optically it is medium to fine-grained, equigranular, foliated with aligned biotite, and vaguely layered with quartz, oligoclase, and microcline in varying proportions. It lies within granite to granodiorite compositional fields. Orthogneiss in the high strain zones is locally recognizable, but it is generally difficult to distinguish early orthogneiss from gneissose, sparingly garnetiferous granite formed in the high strain zones. The foliated and lineated fabric of the orthogneiss predates the metagabbro suite since it cuts across the fabric in the less deformed regions. In migmatites of the lower strain regions, massive granite encloses fragments of foliated orthogneiss.

Metagabbro suite

Northwest trending, hornblende rich metagabbro dykes, tens of metres wide, transect shallowly dipping gneisses west of the northeast trending shear zone. Within high strain zones, metagabbro and paragneiss lenses contain



LEGEND

Undifferentiated Quaternary

PALEOZOIC

Carbonate Formations, Obc, Ocr, Orr, Ssr

PRECAMBRIAN

L *Later High Strain Zone*

E *Earlier High Strain Zone*

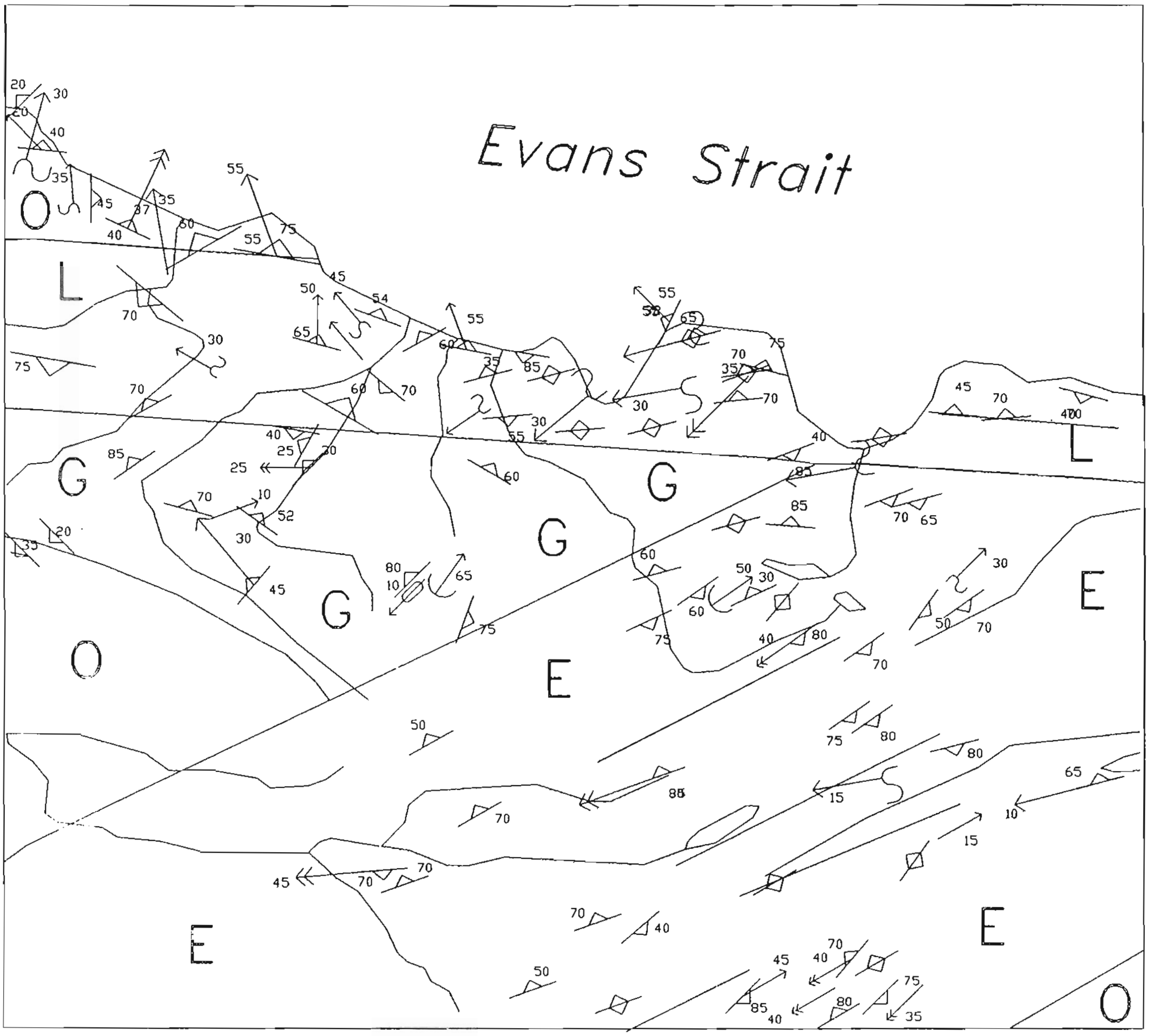
G *Granite-Granodiorite*

Metagabbro, inclusions

P *Paragneiss, rare quartzite and marble*

O *Orthogneiss, granodioritic*

Figure 2. Geological map of the Precambrian Gneisses on Coats Island. Modified from Heywood and Sanford (1976), and Sanford and Grant (1990). Lines offshore are fault traces and the relative movement of blocks are marked by D (down) and U (up).



LEGEND

Gneissosity, inclined, vertical

Minor folds with plunge

Lineations, 1st, 2nd, elongation

Figure 3. Detail of geology from northern coast of Coats Island showing intersecting high strain zones.

discordant structures. Primary layering in the metagabbro is commonly present in the larger lenses, and compositions of metagabbro fragments vary from lens to lens. The metagabbro suite appears to consist of a tectonically dispersed, once more extensive, layered mafic complex. Massive anorthosite occurs as an outlier in the south of the mapped area as a hill in the Palaeozoic outcrop area (Heywood and Sanford, 1976). Gabbroic anorthosite blocks are found in the late high strain zone. Layered gabbro lenses are found in both high strain zones and hornblendites and other ultramafic lenses occur sparingly. The least deformed part of the gabbro is a wide crosscutting body, with areal differences in development of layering. A thin section of an amphibolite, shows decussate hornblende aggregates pseudomorphing prismatic minerals, relict amphibole speckled by opaques, and partially replaced and rimmed by amphibole, zoned plagioclase, and regions consisting of intergrown opaques, amphibole, and plagioclase. Some phases are garnetiferous and a thin section show very complicated textures such as garnets with symplectic rims intergrown with amphibole, decussate amphiboles, relict and replaced amphiboles and clinopyroxene, and plagioclase with symplectic cores possibly of spinel. The metagabbros have locally intruded deformed paragneisses and orthogneisses, and elsewhere occur as local inclusions in granite. The high strain zones which contain metagabbro lenses postdate the granite. The microscopic textures described above indicate that static heating, rather than recrystallization in a dynamic environment, was the latest major metamorphic event to affect the rock. This could result from

i/ a late (post-high-strain-zone) static heating of the whole region, which is thought unlikely because of lower grade muscovite bearing assemblages in the surrounding strained rocks, or more likely from

ii/ an older thermal event possibly associated with granite emplacement. The subsequent high strain and lower metamorphic grade in the strain zones has been localized in the borders of the lenses and thus has not affected the interiors of these structural relics.

Migmatites and granites

These rocks form most of the northeast coast of the island. Typical outcrops comprise equal thirds of orthogneiss inclusions, later crosscutting granite and several generations of pegmatites, the latest of which is planar. Locally the proportion of granite increases and relict inclusions and their structures become nebulous. In thin section the granite is seen to be composed of about equal proportions of microcline, quartz, and plagioclase, with minor amounts of biotite and scattered red garnets. Inclusions of metagabbro and paragneiss dot the granite. Outside the high strain zones, the granites are massive and medium-grained. Within the high strain zones granite still shows a few garnets but most minerals are strained, hydrated or polygonitized. Biotite and elongate quartz grains define the foliation. Late pegmatite dyke sets that trend north-northeast with moderate to steep northwest dips, postdate the granite but probably predate the high strain zones.

High strain zones

Two steeply dipping high strain zones have been recognized on a map scale (fig. 3).

a) Earlier zone

An earlier east-northeast trending straight zone, several kilometres across is developed in orthogneiss and granite and contains decametre to kilometre long lenses of paragneiss and metagabbro. The rocks are well foliated, and show a planar fabric. Sometimes folia are closely spaced such that the rock is a phyllonite. Foliations typically strike south of west and dip steeply to the north-northwest and contain mineral lineations, elongations and minor folds with both shallow to moderate plunges both to northeast and southwest. Ellipsoidal lenses of paragneiss and gabbro are common, and plunge southwesterly. At one locality, adjacent layers showed mineral lineations with opposed moderate plunges. The shallow northeast plunging structures are apparently the latest, but more work is needed to substantiate this. For example, a northeast plunging Z-fold developed on a flattened metagabbro fragment which, within its relict fabric, and detached from the country rock, contains southwest plunging Z-folds.

b) Later zone

This east-trending high strain zone is narrower but is continuous along strike for at least 24 kilometres. It consists of subparallel zones of straight gneisses developed in the local country rock. Tectonic clasts tend to be metre to decametre sized, but trains of metagabbro clasts extend for up to a kilometre along strike of the strain zone. Deformed and rotated pegmatites are also present in the high strain zone. Other east-trending lineaments are also narrow high strain zones, but their full extent has yet to be mapped.

Phanerozoic units

Since the first systematic mapping of the Palaeozoic of Coats Island (Heywood and Sanford, 1976), new information obtained from marine and on shore investigations during 1985 to 1988 has resulted in an improved lithostratigraphic and stratigraphic classification of the Ordovician succession (Sanford and Grant 1990). On Fig. 1, four formations outlined by Grant and Sanford (1990) are indicated. Only a few outcrops of the Ordovician Bad Cache Rapids Formation are known, and the area marked as Obc on the map includes these, and the overlying Boas River Formation, for which no outcrops are known on the island. The Ordovician Churchill River (Ocr), and Redhead Rapids Formations (Orr) and Silurian Severn River Formations (Ssr) are better exposed (Grant and Sanford, 1990).

Off-shore, similar, submerged Palaeozoic sediments dip southwestward, as they do on land. Northeast trending normal faults bound the island and abut against west-northwest trending faults (ibid).

Cretaceous sediments occur northeast of a fault which parallels the west-northwest striking shore of the island. These Evans Strait sediments, described by Sanford and Grant (1990), reach thicknesses of many hundreds of metres near the island. The magnitude of throw on the bounding fault can be estimated from matching Precambrian surfaces on either side of the fault, and is in the order of a kilometre near the island. The fault appears to be part of the Bell Arch System and demonstrates Cenozoic displacement. The northern coast is very steep (Grant,1969), and may represent a fault plane related to this fault system.

Lineaments, obvious on air photos and mapped by Aylsworth and Shilts (1987), are debris filled valleys that run subparallel to the dominant trend in the earlier ductile shear zone. Specimens from the walls of one such lineament are cut by narrow veinlets of cataclasite and carbonate veins. They show feldspars that are heterogeneously transformed into lenses of fine-grained carbonate aggregates, chloritized biotite, and strained quartz. The lineaments are interpreted as Phanerozoic faults that have been excised by glacial scouring.

Aylsworth and Shilts (1987) record the Pleistocene emergence of the island from beneath an ice sheet. Contrary to Craig (1969), the highest Holocene marine sediments occur at about 124 m as marked on Aylsworth and Shilts' (1987) map, some tens of metres lower in altitude than this author is used to seeing either to the north on Melville Peninsula or to the west near Baker Lake. In addition to island derived carbonate fragments, Dubawnt Group fragments (from near Baker Lake) are found in the glacial sediments (cf Shilts, 1980). and so are fragments of ripple-marked, well indurated, orthoquartzite, and hematite iron formation, grey slate, greenstone-grade mafic volcanics (termed 'dark erratics' by Shilts 1980) and a characteristic type of spotted greywacke. These latter rock types are not indigenous to the island, and likely come from the southeast (Shilts, 1980) maybe from as far south as the Belcher Islands. Most glacial striations trend east-northeastward, but a few late striae trend west of north. On the north coast, raised carbonate beach deposits at an elevation of about 8 metres encorporate a large number of thick-billed murre bones, among which, fragments of chert tools are found.

Speculative history of the gneiss complex

The Coats Island gneisses are very similar to those exposed on Southhampton Island to the north. Metasedimentary gneisses hint at a quartzite, carbonate, and shale protolith association, although the preserved volumes are very small and are tectonically interleaved with orthogneisses. The early Proterozoic Penrhyn Group (Henderson, 1983) and late Archean? sedimentary rocks at Daly Bay (Gordon,1988) are the most likely candidates for correlation with the paragneisses to the north and west. Gordon (1988, p.20) postulated that . . . the Daly Bay Complex is only part of a much larger region composed of imbricated slices of middle and lower crustal material . . . He concludes they were emplaced from the south upward into the gneisses to the north circa 1850 Ma (ibid). The Lake Harbour Group (Jackson and Taylor, 1972) to the north east, and the Sugluk Group (Taylor, 1982, Lucas and St-Onge, in press) to the southeast are also possible correlatives. The latter is considered allochthonous and active circa 1863-1834 Ma.(Lucas and St-Onge, in press).

Coats Island gneisses could well represent part of this 'larger region'. The seaward projection of mainland geology mentioned above suggest some sort of structural continuity. The separation of fabrics into those of early shallow dipping structures possibly of widespread occurrence, and later, more local, steeply dipping, high strain zones as well as the recognition of a metagabbro suite of intermediate age fits well with the mainland geology. A possible scenario is outlined below:

1/ Early, shallow dipping foliated orthogneisses and interleaved paragneisses, showing minor structures with a relative north over south sense of movement suggest that lateral transport was an important early component. Gabbro suite was emplaced after much of this transport. Granite intrusion and migmatite formation followed. Perhaps elevation and associated decompression, subsequent to depression of the gneisses under a thrust mass from the north, led to localized partial melting of gneisses.

2/ Later, steeply dipping, high strain zones: an earlier, several kilometre wide, northeast trending zone, and a later, kilometre wide, east-trending zone with lenticular inclusions of previous units, cut the above units. These zones carry muscovite developed in the foliation so the zones are formed under middle or lower amphibolite conditions, perhaps imposed as the total, cooling gneiss assemblage was approaching the surface.

3/ Several sets of brittle faults, associated with the Bell Arch transect the island, and deform a pre-Ordovician peneplain, and form the edge of a graben immediately off shore, which contains Cretaceous sediments.

Thus similar Precambrian units and structures probably occur on either side of the graben on Coats Island and Southhampton Island, and the geophysical parameters principally reflect the effects of Phanerozoic deformation.

MINERAL POTENTIAL

Gneissic complex

The crystalline bedrock is predominantly composed of orthogneisses of moderate to high metamorphic grade which have generally been considered poor prospects for mineral exploration. Paragneisses carrying quartzite, marble, and sillimanite schists, may be marginally more prospective, but the low volumes, and extremely sheared, and lenticular nature of these gneisses, along with scarce pyrite in garnet rich rocks, suggest the mineral potential for these rocks is extremely low as well. Graphite has been noted in some of the marble and biotite schists, creating possible interference with electric prospecting methods. No copper mineralization was noticed associated with metagabbro suite. Several generations of simple type pegmatite, with large books of biotite, and devoid of exotic mineral assemblages are also low in mineral potential.

Current models of gold deposit development (cf Cameron, 1989) involve processes that leach sheared rocks such as those exposed on Coats Island, to concentrate deposits elsewhere in more favoured sites. Should these models prove applicable, then the mineral potential of these rocks is lessened significantly. As the rocks have been sheared extensively, the mineral potential is considered to be very low.

Phanerozoic units

The debris from many units have been seen in the carbonate till cover and along the carbonate beaches. Two slabs, one from a creek bed draining marine till deposits, and the other from an elevated beach deposit, consist of 3 cm thick slabs some several tens of centimetres in diameter, of pure layered pyrite. It is possible that carbonate hosted base metal deposits are present nearby. Current patterns and ice rafting (Pelletier, 1969) suggest that the source might be to the north on or near eastern Southhampton Island. However, if slabs were derived from reworked entrained glacial debris, then, since the fragments coexist with fragments from as far away as Baker Lake, Richmond Gulf, and Belcher Islands, no particular location for their potential source can be deduced.

Sanford and Grant (1990) note that Cretaceous units in which oil and gas could accumulate, occur on the bottom of Evans Strait. Pelletier (1969) noted that the content of organic carbon in bottom sediments of the strait is high (in excess of 1.5%) and attributed this to the shallow depth and rapid sedimentation rates which deter oxidation in the strait. However, the total Quaternary sections do not exceed thicknesses of 5 to 10 metres in this region (Josenhans et al.,1988). An alternate explanation is required for elevated organic carbon in the bottom sediments of Evans Strait. Pitch balls or other signs of hydrocarbon pollution were not seen in traverses along the beaches facing the Strait.

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