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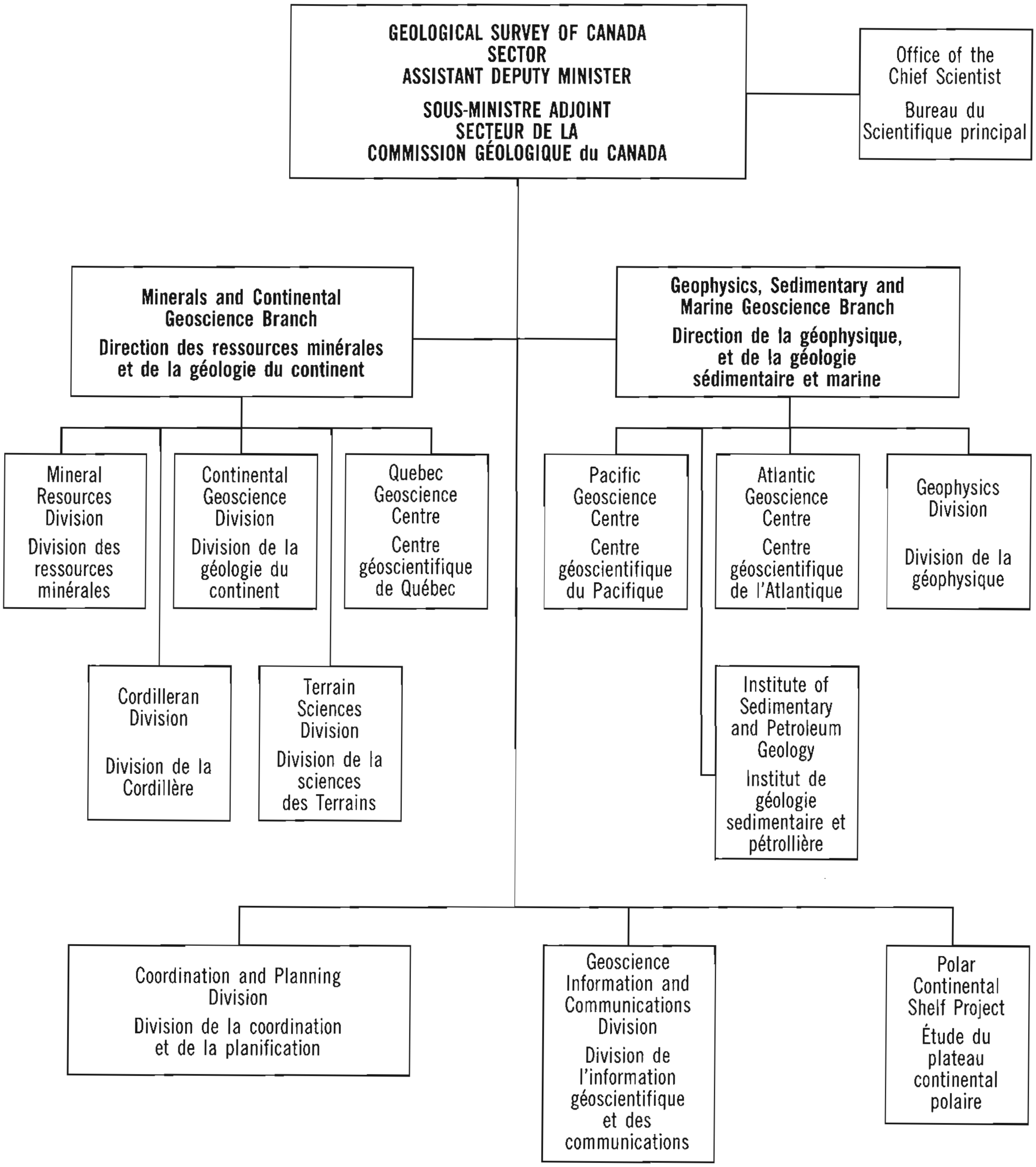
Cover description

Ductilely dismembered mafic dyke in heterogeneously deformed anorthositic gabbro showing an apparent dextral sense of ductile shear; Uvauk complex, Gibson Lake map area (55N), Northwest Territories. See report by Tella et al., p. 197-208. Photo by S. Tella.

Description de la photo couverture

Dyke mafique démembré par déformation ductile dans du gabbro anorthositique à déformation hétérogène, montrant un cisaillement ductile vers la droite; complexe d'Uvauk dans le secteur cartographique du lac Gibson (55N), Territoires du Nord-Ouest. Voir rapport de Tella et al., p. 197-208. Photo prise par S. Tella.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Geological studies in the Precambrian of the western margin of Boothia Uplift, Prince of Wales and Somerset islands, District of Franklin, Northwest Territories

Thomas Frisch
Continental Geoscience Division

Frisch, T., 1993: Geological studies in the Precambrian of the western margin of Boothia Uplift, Prince of Wales and Somerset islands, District of Franklin, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 1-6.

Abstract: Anorthosite has been discovered at the margin of the M'Clure Bay granite, a large rapakivi granite intrusion (1703 Ma old) in northwestern Somerset Island. It appears to form a sheet, 20 m thick, and is probably cogenetic with the granite.

Substantial sequences of the Proterozoic Aston Formation lie nonconformably on basement adjacent to the Paleozoic terranes on eastern Prince of Wales Island and Prescott and Pandora islands. A representative section measured on Prince of Wales Island comprises 700 m of shallow-marine rocks: a thin breccia-conglomerate resting on gneiss, 200 m of reddish arkose, and a 500 m-thick succession mainly of more mature quartz sandstone. A 200 m-thick gabbro sill (1268 Ma) has intruded the section. The section terminates at the high-angle reverse fault juxtaposing Proterozoic and Paleozoic strata at the margin of the Boothia Uplift. The intense deformation evident in rocks on both sides of the fault is consistent with a compressive tectonic regime.

Résumé : On a découvert une anorthosite sur la marge du granite de M'Clure Bay, vaste intrusion de granite rapakivique (âgée de 1 703 Ma) située dans le nord-ouest de l'île Somerset. Cette anorthosite, probablement cogénétique du granite, constitue apparemment une nappe de 20 m d'épaisseur.

Des séquences substantielles de la Formation d'Aston (Protérozoïque) reposent en discordance sur un socle adjacent aux terranes paléozoïques dans l'est de l'île du Prince-de-Galles et dans les îles Prescott et Pandora. Un profil stratigraphique représentatif mesuré dans l'île du Prince-de-Galles englobe 700 m de roches marines épicontinentales: un mince conglomérat bréchiforme reposant sur un gneiss, 200 m d'arkose rougeâtre, et une succession de 500 m d'épaisseur principalement composée de grès quartzeux plus mature. Un filon-couche gabbroïque de 200 m d'épaisseur (âgé de 1 268 Ma) a pénétré les roches du profil. Ce dernier se termine contre une faille inverse très inclinée juxtaposant des strates protérozoïques et paléozoïques sur la marge du Soulèvement de Boothia. La déformation intense visible dans les roches des deux côtés de la faille est un phénomène compatible avec un régime tectonique de compression.

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 Province of the Canadian Shield and flanked by Helikian to Devonian sedimentary rocks 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).

Previous work by the author on the crystalline core of the Boothia Uplift in 1986, 1987 and 1990 was confined essentially to the exposures on Boothia Peninsula north of 71°N and on Somerset Island (Frisch et al., 1987; Frisch and Sandeman, 1991). Operational difficulties restricted examination of the Precambrian rocks of the western margin of the Boothia Uplift, exposed in a narrow belt along the west shore of Peel Sound, to no more than a few person-days. Fieldwork in 1992 focussed, therefore, on the shield terranes of easternmost Prince of Wales and Prescott islands and was carried out from July 18 to August 8. A further 9 days were spent in the area about M'Clure Bay on Somerset Island. Results of the 1992 fieldwork are summarized in this report.

METAMORPHIC AND IGNEOUS ROCKS OF THE BASEMENT

Crystalline rocks of the Canadian Shield were investigated in the area between Cape Brodie and Savage Point peninsula on Prince of Wales Island, on eastern Prescott Island, and in the region about M'Clure Bay on western Somerset Island (Fig. 1).

Foot traverses across the gneiss terranes on the west shore of Peel Sound confirmed earlier observations (Frisch and Sandeman, 1991) that dark-weathering, orthopyroxene-bearing granitoid gneisses predominate and garnetiferous supracrustal rocks are rare. The gneisses are intensely flattened, commonly mylonitic, and steeply to vertically dipping. Mafic layers and lenses, many of which are highly attenuated, and migmatitic layering are common. Folding is most easily recognized in the migmatitic rocks and its style is similar to that in the basement on Somerset Island and Boothia Peninsula: steeply-dipping axial planes and moderately north- or south-plunging axes.

Garnetiferous gneiss was found at three localities. Migmatitic, rusty, garnet-biotite paragneiss in belts 30-40 m across and 200-300 m long occur north of Cape Brodie and on eastern Prescott Island. Mafic garnet-hornblende-pyroxene gneiss outcrops in a small fault-bounded basement block on Prescott Island (Fig. 1). These garnetiferous rocks should yield useful information on pressure-temperature conditions of metamorphism in the westernmost crystalline core of the Boothia Uplift.

The basement rocks immediately below the Proterozoic sedimentary cover on Prescott Island are little altered and no change in their appearance is discernible as the nonconformity is approached. The gneisses below the measured Proterozoic section on Prince of Wales Island (see below) are highly shearedand are consequently altered and weathered along the shear zones. It is emphasized, however, that no regolith is preserved in the basement.

Shearing, K-feldspathization, quartz and K-feldspar veining, and chloritization are widespread in the basement rocks adjacent to faulted contacts with Proterozoic and Paleozoic sedimentary rocks. Such contacts are invariably near the main high-angle reverse fault juxtaposing Precambrian and Paleozoic rocks at the margin of the Boothia Uplift. As the sedimentary rocks are generally little affected, the alteration of the basement rocks must have occurred prior to deposition of the Proterozoic strata. Shear zones in the basement were apparently reactivated as zones of movement in the Paleozoic tectonism that created the Boothia Uplift.

It may be pointed out here that the body of ultramafic rock south of Flexure Bay shown on Map 3-1967 of Blackadar (1967) actually consists of chloritized, sheared granitoid

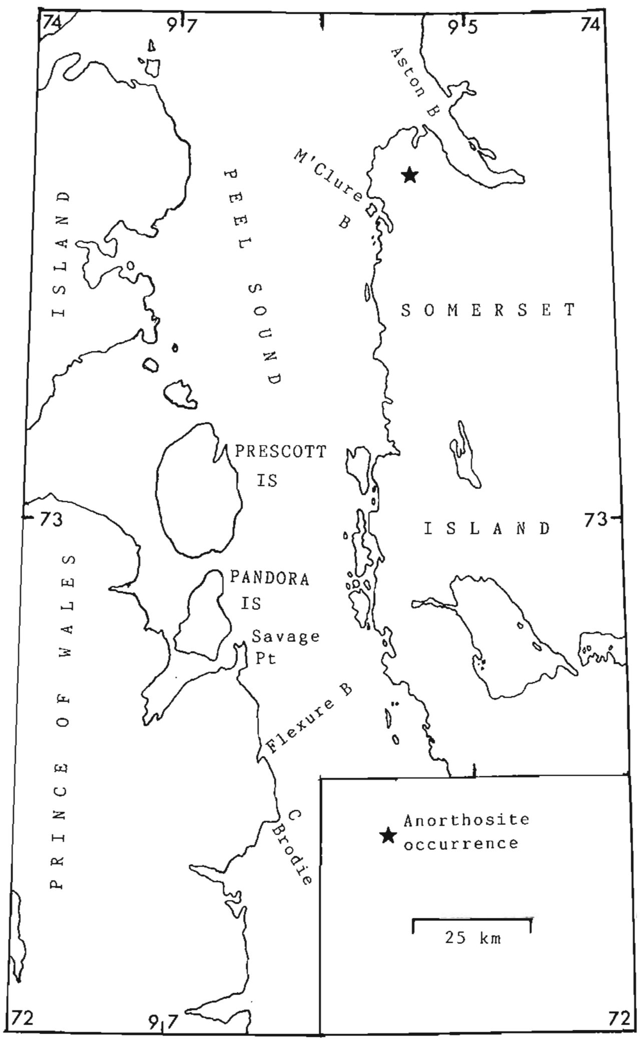

Figure 1. Location map of the study area.

gneiss and amphibolite deformed in the tectonic zone of the Precambrian-Paleozoic contact, which lies a few hundred metres away.

On Somerset island, anorthosite, previously unknown in the Boothia Uplift, was discovered at the margin of the large rapakivi granite intrusion at M'Clure Bay (Fig. 1). The M'Clure Bay granite is a posttectonic (U-Pb zircon age 1703 ± 2 Ma) intrusion of massive, pink, coarse grained, K-feldspar- megacrystic rock, which underlies virtually the whole coastal shield terrane from M'Clure Bay north to Aston Bay (Frisch and Sandeman, 1991). Ten kilometres northeast of M'Clure Bay, the contact between the granite and gneiss runs northeastward through a 200 m-high hill, which is capped by anorthosite. The anorthosite is coarse grained (grain size 5-13 mm) and grey or greenish on the fresh, and white on the weathered, surface. Minerals interstitial to plagioclase, which comprises about 90% of the rock, are pyroxene and Fe-Ti oxide (ilmenite?). The anorthosite occurs largely as rubble but appears to form a sheet, ca. 600 m long, up to 300 m wide and 20 m thick, underlain by granite and bordered to the east by gneiss. No contacts were seen but the anorthosite presumably is intrusive into the gneiss. The adjacent granite is a finer grained variety of the M'Clure Bay granite, similar to the rock found elsewhere in the margins of this body. Neither the granite nor the anorthosite is deformed. The age relation between the two rock types is unknown but if, as seems likely, they are genetically related, as in the well-known Proterozoic anorogenic anorthosite-rapakivi granite association, the anorthosite is probably the older rock (Emslie, 1978).

UNMETAMORPHOSED PROTEROZOIC STRATA

Proterozoic sedimentary rocks and intrusive gabbro were examined on the east shore of Prince of Wales Island between Savage Point and Flexure Bay and on Prescott Island (Fig. 2). On both islands, substantial sequences of the Aston Formation (Dixon et al., 1971) rest nonconformably on

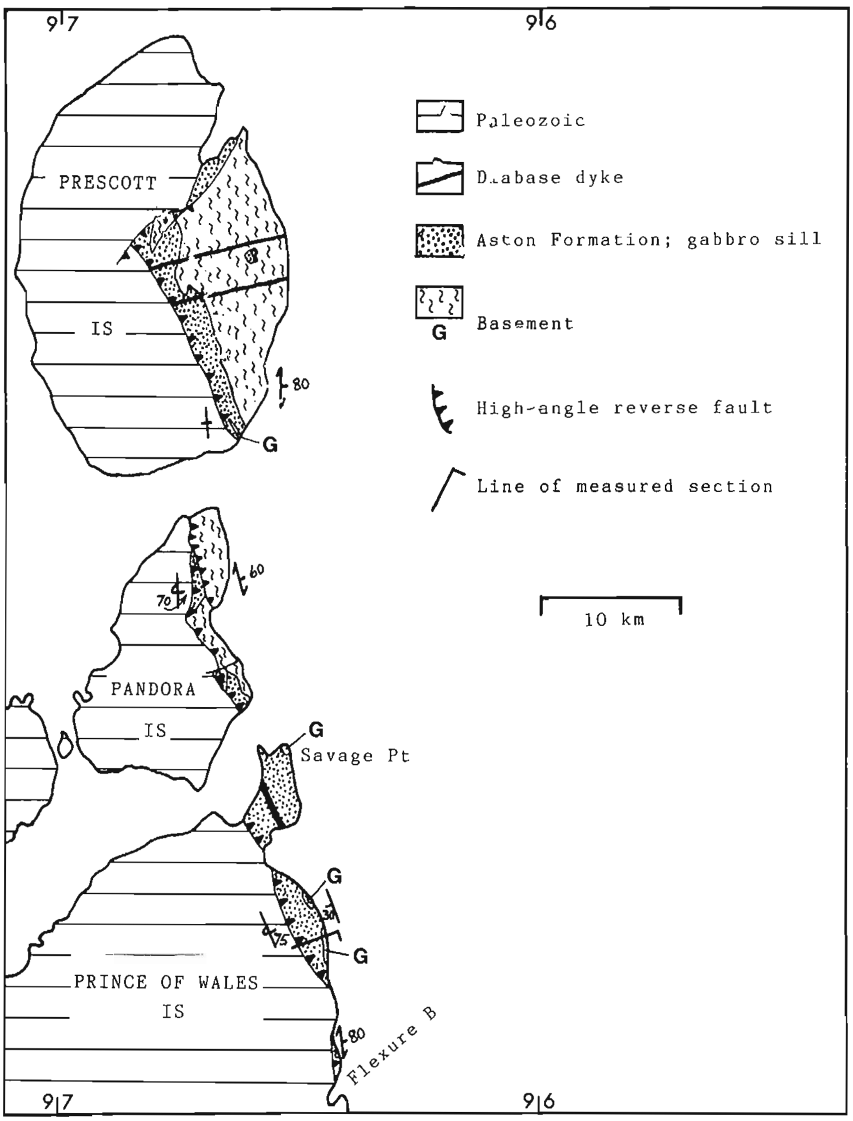

Figure 2.

Geological sketch map of Prescott Island, Pandora Island and part of Prince of Wales Island. Geology of Pandora Island after Christie et al. (1971).

crystalline basement adjacent to the Paleozoic terrane but the sedimentary section on Prescott Island is severely disturbed by tectonism. Thin outliers of the Aston were found on eastern Prescott Island, well to the east of the main Proterozoic section (Fig. 2). The outcrop of Hunting Formation - the carbonate unit overlying the Aston on Somerset Island - reported by Miall (1969, p. 260) as lying 13 km northwest of Cape Brodie on Prince of Wales Island was found to belong to the Paleozoic section. No other occurrences of Hunting Formation have been reported from west of Peel Sound and it appears that the Hunting is restricted to Somerset Island.

The best exposed and structurally simplest Proterozoic sedimentary succession is that between the Savage Point peninsula and Flexure Bay. Some 700 m of reddish, yellow and grey arkosic and quartzose sandstones and minor dolostone, siltstone and conglomerate-breccia, dipping 25-30° southwest, rest on gneiss. A ca. 200 m-thick gabbro sill (U-Pb baddeleyite age 1268 Ma; LeCheminant and Heaman, 1991) has intruded the section, roughly dividing it into upper and lower halves. A stratigraphic column is shown in Figure 3 and a descriptive summary of the lithology follows.

The base of the succession consists of a one metre-thick layer of breccia-conglomerate, which lies on highly sheared and consequently strongly weathered gneiss. The clasts are of gneiss and quartz and mostly angular; most of them are 5-10 cm long and the largest seen measured 20 cm. The matrix of the conglomerate is a brick-red, silty arkose with abundant millimetre-sized rock and quartz grains. Overlying the breccia-conglomerate is a 200 m-thick sequence of dark pink, flaggy arkose with minor, thin silty interbeds and, near the top, several thin beds of more quartzose sandstone. There follow lighter-coloured, grey, yellow to buff, and pale pink quartz sandstones and orthoquartzite, totalling ca. 90 m in thickness. Like the underlying arkosic rocks, the quartzose sandstones tend to be thin bedded to flaggy. Although ripple marks and planar, low-angle crossbeds in sandstone and shrinkage cracks in siltstone are common, they are not abundantly developed. Furthermore, due to the intense frost shatter in these steep coastal exposures, sedimentary structures are seen practically only in dislodged blocks of rock and orientation measurements are rarely meaningful. From limited data, a westerly sediment transport direction is inferred, in agreement with the observations of Dixon et al. (1971).

At a height of about 300 m in the section, the lithology changes abruptly with the appearance of a resistant unit, 45-50 m thick, composed largely of dolomite and chert. Although lithologically variable along strike, the unit may be divided into a lower part, consisting of thinly interbedded pink to white dolomite and red chert, and an upper part, made up of slightly recessive red, silty chert overlain by ca. 14 m of pink to white, cherty, stromatolitic dolomite. The stromatolites are of the columnar type, forming both discrete and branching columns with circular to elliptical cross-sections and diameters typically 3-7 cm, enclosed in silty dolomite. Whereas earlier work (Dixon et al., 1971, p. 738; Frisch and Sandeman, 1991, p. 175) indicated that

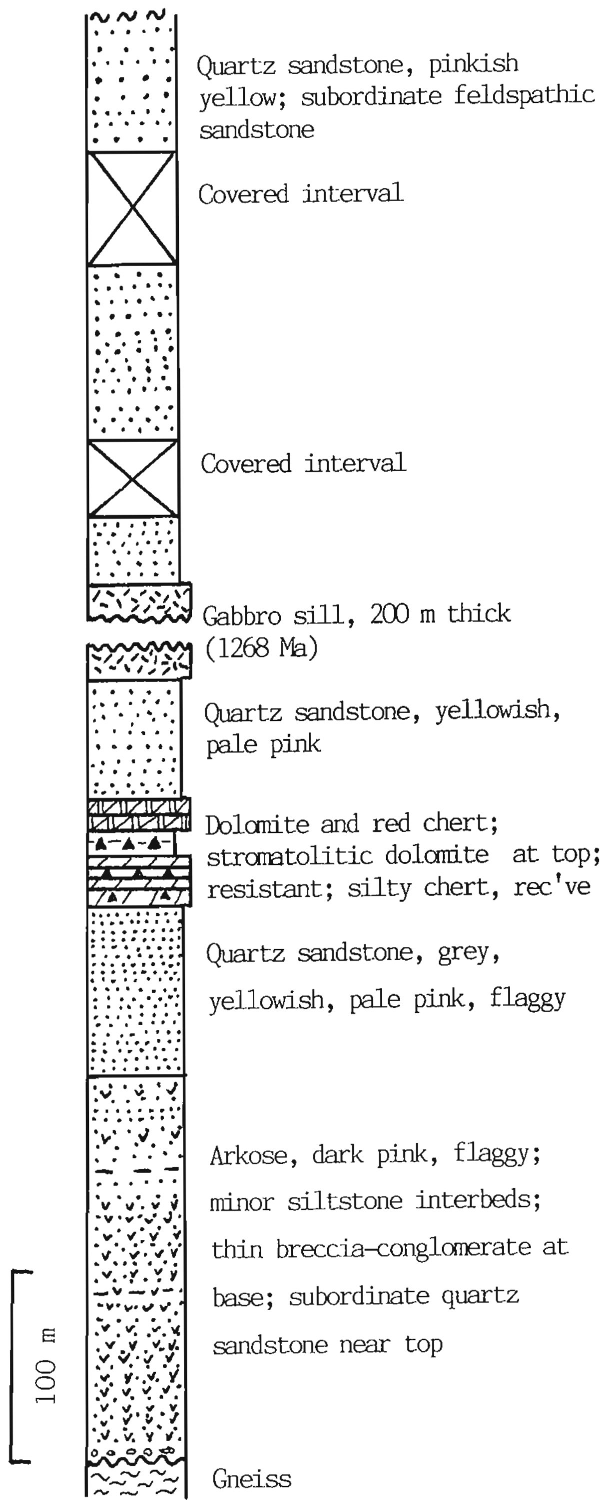

Figure 3. Stratigraphic column for the measured section of the Aston Formation on Prince of Wales Island (see Fig. 2).

columnar stromatolites are absent from the dolomite-chert unit in the southern part of the Proterozoic succession on Prince of Wales Island, fieldwork in 1992 showed that they do indeed occur at the top of the unit along its entire length, even if locally their presence is marked only by loose rock in rubble. Vague internal structures and wavy laminations, in part at least, probably of algal origin, are conspicuous throughout much of the dolomite-chert unit.

The succession overlying the dolomite-chert unit is about 360 m thick and comprises thin-bedded, pale-coloured, predominantly pinkish-yellow, quartz sandstone and subordinate more immature, feldspathic sandstone. The rocks are not particularly well exposed, some 100 m of section being drift-covered, and the section terminates in a deep stream valley along the faulted contact with the Paleozoic.

In the main Proterozoic sedimentary succession on Prescott Island, the basal beds, up to 200 m thick, of the Aston Formation weather a distinctive brick red, a colour seen in the Proterozoic on Prince of Wales Island only in individual beds or thin sequences. From aerial observations, however, similarly brick red-weathering rocks form the bulk of the Proterozoic sedimentary succession on Pandora Island. The rocks on Prescott Island, both those at the base of the main succession adjacent to the Paleozoic and in the outliers in the basement terrane to the east, are mainly arkose and silty arkose, subordinate immature quartz sandstone, and minor siltstone. In the main succession, the lowermost rock is pink conglomerate made up of 1-3 cm-long rounded pebbles of felsic gneiss and quartz in a matrix of pink arkose; maximum measured thickness of conglomerate is 6 m. The underlying basement gneiss is little altered. No conglomerate was seen in the heavily weathered outliers of Aston Formation to the east but these exposures consist largely of a thin veneer of rubble on poorly exposed basement.

The sill, which has intruded the sedimentary succession about half-way up, is a generally homogeneous body of medium grained gabbro, in which the chief minerals are pyroxene and plagioclase. It is slightly finer grained near the margins but an actual contact with country rock was never seen. Some outcrops of the basal part of the sill on Prince of Wales Island are made up of black gabbro with ca. 5% biotite. This rock is characterized by a pitted surface and closely spaced flat joints parallel to the intrusive contact. The biotite gabbro probably crystallized from magma in which volatiles were concentrated near the margin of the intrusion.

Although concordant overall, the sill is locally transgressive. It normally lies above the dolomite-chert unit but cuts it at several places on Prescott and Prince of Wales islands. At the intrusive contact of the northernmost exposure of the sill on Prescott Island, ophicalcite is extensively developed as a result of contact metamorphism of the dolomite-chert rock.

The predominance of uniformly sand-sized material and absence, except for the basal layer, of conglomerate in the Proterozoic sedimentary succession on the west side of Peel Sound indicates a marine origin for the sediments. The linearly extensive, if thin, stromatolite unit, the occurrence of shrinkage cracks at many levels in the succession, and the basal conglomerate all point to shallow-water sedimentation. The scarcity of shale implies a moderately high-energy environment of deposition. Limited data on orientation of crossbeds indicate westward transport, presumably off a granitic terrane in the east, now exposed as the cratonic core of the Boothia Uplift. The abundance of hematite-stained arkosic rocks and the thicker basal conglomerate on Prescott Island suggest that these rocks were deposited closer to their source, or were deposited more rapidly, than were those on Prince of Wales Island.

The top of the Proterozoic sedimentary succession is bounded by a major, steeply-dipping reverse fault that separates the moderately westward-dipping Proterozoic from the steeply to vertically dipping Paleozoic strata. On Prince of Wales Island, the fault is marked by a deep stream valley and the Proterozoic-Paleozoic contact is unexposed. On Prescott Island, the contact lies in or adjacent to a shallower stream valley and the rocks are highly shattered. Aston Formation sandstone in stream-bed exposures near the fault is a dense, flinty rock showing a high degree of flattening. At the mouth of the valley, on the south shore of the island, a thin fault slice of basal Aston arkose and underlying granitoid basement has been so strongly sheared that they are not easily distinguished.

The evidence of a high degree of deformation everywhere along the Proterozoic-Paleozoic contact favours an origin of the Boothia Uplift by compression, rather than vertical block movements.

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Thanks are extended to B. Paquin for assistance and good companionship in the field. The Polar Continental Shelf Project generously provided Twin Otter and helicopter support, accommodation at Resolute Bay and other assistance and courtesies. G.D. Jackson read the manuscript and made useful suggestions for its improvement.

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Geological Survey of Canada 860001

Sedimentology and sequence stratigraphy of the Neoproterozoic Reynolds Point Formation, Minto Inlier, Victoria Island, Northwest Territories[1]

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Morin, J. and Rainbird, R.H., 1993: Sedimentology and sequence stratigraphy of the Neoproterozoic Reynolds Point Formation, Minto Inlier, Victoria Island, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 7-18.

Abstract: The Reynolds Point Formation of the Neoproterozoic Shaler Group represents part of an extensive shallow-water carbonate platform deposited within the intracratonic Amundsen Basin. It comprises four members.

The Lower Clastic Member is a mudstone to quartzarenite succession representing a prograde river-fed delta reworked by storms. The Lower Carbonate Member comprises rhythmitic, stromatolitic and oolitic dolostone magnafacies recording mid to outer shelf, mid shelf, and inner shelf deposition on a cyclically prograding storm-dominated carbonate ramp. The Upper Clastic Member is composed of crossbedded quartzarenite and sandstone-carbonate rhythmite representing sand-shoal to barrier-beach and back-barrier lagoon environments. The Upper Carbonate Member comprises stromatolitic and thin-laminated dolostone deposited in a lagoon that evolved to a subaerial setting.

The Reynolds Point Formation defines five, third-order(?) sequences. These are interpreted, with the exception of the first sequence, which could be tectonic, as global eustatic sea level variations within a passive subsidence regime.

Résumé : La Formation de Reynolds Point du Groupe néoprotérozoique de Shaler représente une partie d'une immense plate-forme de roches carbonatées mise en place à l'intérieur du bassin intracratonique d'Amundsen. Cette formation, constituée principalement de dolostone et d'une petite quantité de quartzarénite, se compose de quatre membres.

Le membre clastique inférieur représente la progradation d'un delta ultérieurement influencé par des processus marins et particulièrement par des tempêtes. Le membre carbonaté inférieur est défini par l'alternance de trois magnafaciès (rythmitique, stromatolitique et oolitique) qui représentent respectivement des sédiments déposés dans la zone de plate-forme moyenne à externe, plate-forme moyenne et plate-forme interne. Le membre carbonaté inférieur représente la progradation en quatre étapes d'une rampe de roches carbonatées dominée par l'effet des tempêtes. Le membre clastique supérieur représente un milieu de sédimentation variant d'un haut-fond sablonneux silicoclastique à un cordon littoral et lagon interne subissant une possible influence fluviale. Le membre carbonaté supérieur s'est accumulé à l'intérieur d'un lagon peu profond qui a évolué en zone subaérienne.

La Formation de Reynolds Point est caractérisée par cinq séquences de troisième-ordre (?). À l'exception de la première séquence qui peut être d'origine tectonique, les séquences sont interprétées comme étant des fluctuations eustatiques d'échelle planétaire du niveau marin, superposées à un régime de subsidence passive.

[1] Contribution to Canada-Northwest Territories Mineral Initiatives 1991-1996, a subsidiary agreement under the Canada-Northwest Territories Economic Development Agreement. Project funded by the Geological Survey of Canada.

INTRODUCTION

In this paper, we present the results of ongoing field studies in the Wynniatt Bay area of Victoria Island, in the northeast domain of Minto Inlier (Fig. 1). Data were collected as part of a Canada-Northwest Territories Minerals Initiative (MDA) study designed to focus on basin analysis and 1:50 000 scale mapping of potential base metal and hydrocarbon-hosting sedimentary strata of the Neoproterozoic Shaler Group. The first phase of the study will establish the detailed stratigraphic framework of the lower Shaler Group for mapping and intrabasinal correlation. In addition, sedimentological investigations will be employed to explain the regional paleogeography and dynamic evolution of the Amundsen Basin.

Fieldwork in 1992 focused on key stratigraphic sections and regional mapping of the Reynolds Point Formation, an extremely well preserved sequence of platformal carbonates with subordinate siliciclastic rocks. This paper presents a refined lithostratigraphic subdivision of the Reynolds Point Formation, along with a preliminary interpretation of depositional environments using contemporary facies models and sequence stratigraphic analysis.

GENERAL GEOLOGY

Neoproterozoic sedimentary and subordinate volcanic rocks of the Shaler Group outcrop in Minto Inlier on northern Victoria Island in the Canadian Arctic (Thorsteinsson and Tozer, 1962; Fig. 1). This sequence forms an arcuate northeast-striking belt cored by the shallow-limbed Holman Island Syncline. The Shaler Group unconformably overlies Paleoproterozoic sedimentary rocks of the Goulburn Supergroup, which in turn overlie granite of possible late Archean age on the southwest side of Hadley Bay (Fig. 1). The Shaler Group is unconformably overlain by a homoclinal succession of Cambro-Ordovician carbonate and quartzarenite.

STRATIGRAPHY OF THE SHALER GROUP

The Shaler Group was deposited on an extensive, tectonically stable, low gradient, shallow marine platform within the intracratonic Amundsen Basin (Rainbird, 1991). The stratigraphy of the Shaler Group in Minto Inlier includes, in stratigraphic order, the Glenelg, Reynolds Point, Minto Inlet, Wynniatt, Kilian, and Kuujjua formations (Thorsteinsson and Tozer, 1962; Young, 1981; Jefferson, 1985; Rainbird, 1991; Fig. 1). The Shaler Group was amended to include the Natkusiak Formation, an unconformably to conformably overlying sequence of basalt flows and minor volcaniclastic rocks (Rainbird, 1991). The flows were fed by 723 Ma diabase dykes and sills of the Franklin magmatic episode (Heaman et al., 1992). A lower age limit for the Shaler Group of 1110 Ma has been established by U-Pb dating of detrital zircons from the lower Reynolds Point Formation (Rainbird et al., 1992b). The Shaler Group is about 5000 m-thick in the northeast domain of Minto Inlier.

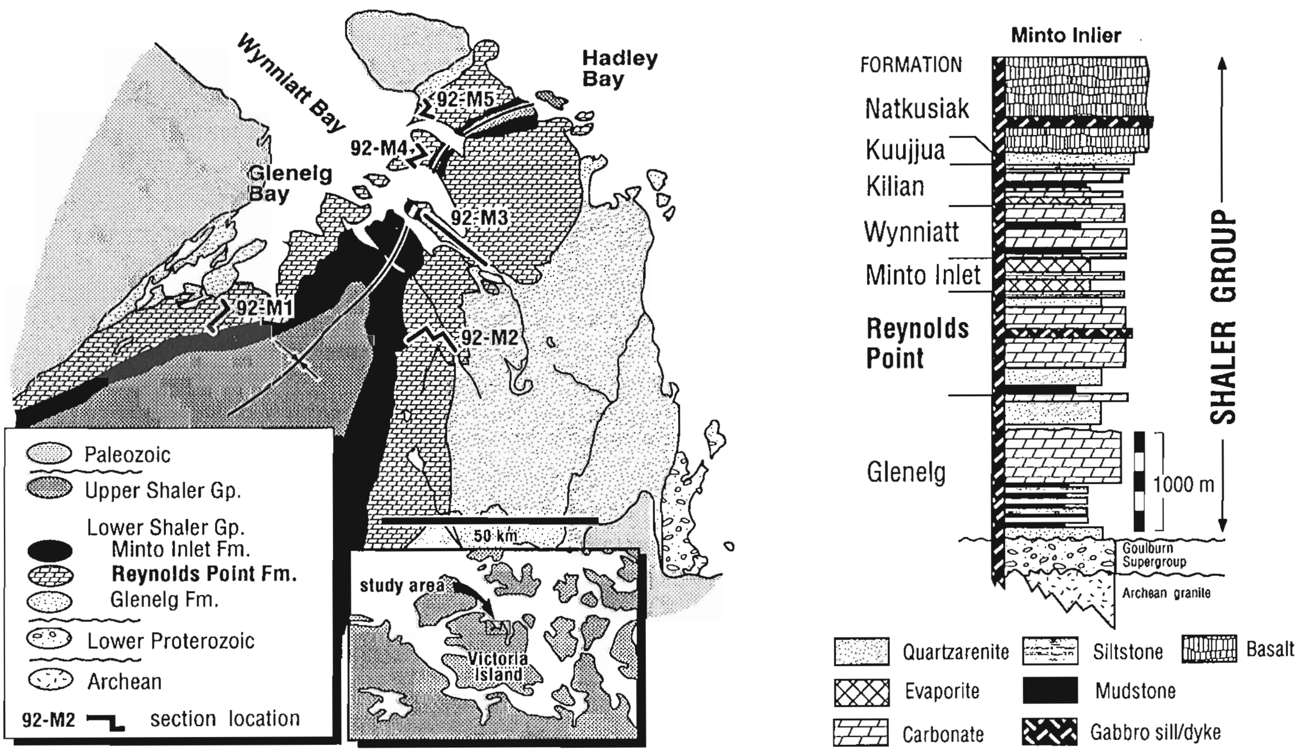

Figure 1. Generalized stratigraphy of the Shaler Group, section locations, and geological map of the northeastern end of Minto Inlier, Wynniatt Bay area, Victoria Island.

REYNOLDS POINT FORMATION

The Reynolds Point Formation was named by Thorsteinsson and Tozer (1962) after a promontory on the south side of Wynniatt Bay. This carbonate-dominated formation conformably overlies the orange-weathering stromatolite member of the Glenelg Formation and is overlain by gypsiferous redbeds of the Minto Inlet Formation (Fig. 1 and 2). The most complete sections of the Reynolds Point Formation were measured at sections 92-M3 and 92-M2 where it attains a thickness of about 550 m (Fig. 3). It increases in thickness northwestward toward sections 92-M1 and 92-M5, although these sections are incomplete. Young (1981) measured 1100 m of Reynolds Point near section 92-M1 of the present study (Fig. 3).

The Reynolds Point Formation has been subdivided lithostratigraphically into four members: lower clastic, lower carbonate, upper clastic and upper carbonate (Young and Long, 1977a; Fig. 2). Parts of the stratigraphy and sedimentology have been described by Young (1974), Young and Jefferson (1975), Young and Long (1977a), Young and Long (1977b), and Jefferson (1977). These members and newly proposed subdivisions of the relatively thick lower carbonate member are discussed below.

Lower Clastic Member

Description

The Lower Clastic Member conformably overlies the regionally extensive orange-weathering stromatolite member of the Glenelg Formation (Jefferson and Young, 1989; Rainbird et al., 1992a). The contact is rarely exposed but at a single locality, east of section 92-M3, it is gradational over 10 to 20 cm with mudstone of the Lower Clastic Member. The contact with the overlying Lower Carbonate Member is gradational over several metres and is placed where carbonate dominates over quartzarenite. Thickness of the Lower Clastic Member ranges from 60 m at sections 92-M3 and 92-M2 to 125 m at section 92-M5 (Fig. 3). Young and Long (1977a) noted that it is at least 215 m thick near our section 92-M1.

The base of the Lower Clastic Member is characterized by an abrupt change from a stromatolitic dolostone to thin parallel-laminated red mudstone, which coarsens abruptly upward through flaser- and lenticular-bedded siltstone to trough- and planar-tabular crossbedded quartzarenite. The crossbedded quartzarenite grades up into parallel-laminated red mudstone with common mudcracks. Possible pseudomorphs of gypsum nodules are present at section 92-M4. Similar fining upward cycles were described from the Lower Clastic Member by Young and Long (1977a). The red mudstone is overlain by small- to medium-scale, tabular-planar crossbedded quartzarenite displaying unimodal northwesterly paleocurrents. An erosional unconformity marked by a sporadic mudstone-pebble conglomerate separates the quartzarenite from an overlying heterogeneous unit composed of hummocky crossbedded, fine grained quartzarenite interbedded with parallel-laminated dolosiltite and dololutite. Basal contacts of the smaller quartzarenite layers are conformable and flat, but the larger ones are scoured into the underlying fine-grained carbonate. Load structures are commonly developed where sandstone overlies siltstone and mudstone. The top of the heterogeneous unit is composed of brown dolomitic mudstone containing small starved ripples of quartzarenite. Mud content increases upward to almost pure mudstone, which in turn grades into wavy bedded and small-scale hummocky crossbedded sandy dolostone with rare gutters. Polygonal ripples are preserved on the tops of the hummocky layers. Quartz sand content diminishes upward with sandy dolostone being replaced by low-angle planar crossbedded dolarenite with interbedded guttered rhythmite (see lithofacies description below).

Interpretation

The Lower Clastic Member of the Reynolds Point Formation shares many characteristics with the Upper Clastic Member of the Glenelg Formation, which is interpreted as the deposit of a prograding river-dominated delta (Rainbird et al., 1992a). Young and Long (1977a) interpreted the Lower Clastic Member as a fluvio-marine delta complex that was reworked by tidal processes. Our work suggests that storms may have played a more significant part in reworking the delta top. Metre-scale fining upward cycles and unimodal north-westerly paleocurrents in the lower quartzarenites of the Lower Clastic Member and in other fluvial quartzarenites of the Shaler Group, suggest that these deposits probably are part of the same extensive fluvial system that traversed across the North American continent in Neoproterozoic time (Rainbird et al., 1992b).

Deposition of thin, parallel-laminated mudstone directly over the orange-weathering stromatolite at the top of the Glenelg Formation is interpreted to indicate rapid basin deepening. Consequently, the coarsening upward sequence of fluvial quartzarenite indicates relatively rapid sedimentation and basin infill.

The heterogeneous unit in the upper part of the member is interpreted as a storm-wave influenced inner shelf deposit that reworked the fluvial-delta front deposits represented by the underlying planar crossbedded quartzarenites. Features such as gutters, broad scours, loading, and various scales of hummocky crossbedding in the sandstone layers all are indicative of storm deposition (Brenchley, 1985). Thin carbonate interlayers represent sedimentation in between storm events. Gradual marine transgression is indicated by the upsection increase in fines and decrease in siliciclastic input.

Lower Carbonate Member

Description

The Lower Carbonate Member is the thickest member of the Reynolds Point Formation ranging from 350 m at section 92-M3 to more than 500 m at section 92-M1, where the base of the member is missing. It is mainly composed of ooid grainstones, stromatolites of diverse morphology, and a

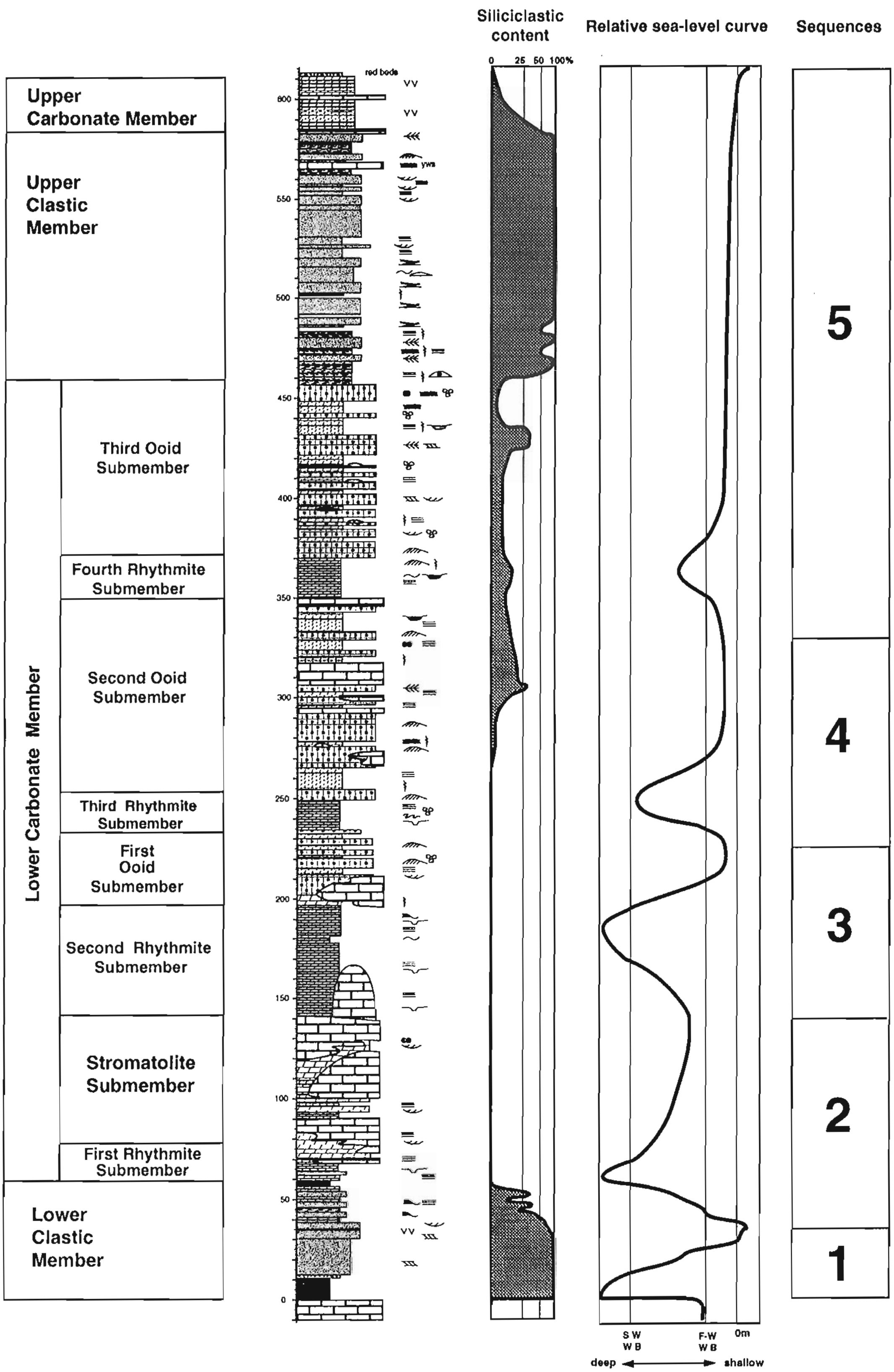

Figure 2. Generalized stratigraphy of the Reynolds Point Formation in the Wynniatt Bay area in relation to siliciclastic content and relative sea level variations.

Table 1. Summary of characteristics for the eight submembers identified within the Lower Carbonate Member of the Reynolds Point Formation. Stratigraphic sections (e.g. 92-M5) are located on Figure 3

Submember	Lithology	Basal Contact	Thickness
Third Ooid	Oolitic magnafacies with increased component of quartz relative to underlying submembers. Quartz is disseminated within carbonate but occurs more significantly as sandstone rhythmite and as herringbone cross-bedded, medium-grained quartzarenite. It is more abundant toward the top of the submember. Stromatolites are common but occur mainly as very small patch reefs.	Conformable under the lowermost ooid bed and conformably overlain by the Upper Clastic Member.	From 70 m at 92-M3 to 170 m at 92-M1.
Fourth Rhythmite	Quartzose, wavy to thin parallel-bedded scoured rhythmite interbedded with dololutite. Hummocky cross-stratification present at 92-M3 and 92-M1 and gutters are common at 92-M1.	Conformable over the uppermost ooid bed.	From 15 m at 92-M3 to 25 m at 92-M1.
Second Ooid	Oolitic magnafacies and stromatolite magnafacies, including patch reefs and up to 15 m tabular reefs. Hosts the lowermost significant occurrence of quartz sand in the Lower Carbonate Member. Quartzarenite beds, sandstone rhythmite and sandy ooid grainstone are common in the upper part.	Conformable under the lowermost ooid bed.	From 75 m at 92-M3 to 120 m at 92-M1.
Third Rhythmite	Guttered rhythmite with rare oolitic packstone.	Conformable over the uppermost ooid bed.	About 15 m thick at all sections.
First Ooid	Oolitic magnafacies, rare dolostone lithofacies. Quartz sand present in minor amounts within coarse dolostone lithofacies in the upper part of the unit at 92-M4.	Conformable under the lowermost ooid bed.	From 15 m at 92-M3 to 80 m at 92-M1.
Second Rhythmite	Rhythmitic magnafacies with guttered rhythmite, non-guttered rhythmite and nodular dolosiltite lithofacies. Ooid packstone beds with erosive bases and interbedded with non-guttered rhythmite are abundant in the upper part of the submember at 92-M4 but are a minor constituent at 92-M3 and 92-M1 and are absent from 92-M2 and 92-M5. A 2 m thick thinly bedded to parallel-laminated purple siltstone was observed at 92-M5, 92-M4, and 92-M3.	Conformable over the uppermost stromatolite.	From 60 m at 92-M2 to 1.50 m at 92-M5.
Stromatolite	Stromatolitic magnafacies with stromatolites and subordinate coarse dolostone, scoured rhythmite and guttered rhythmite lithofacies. Oolitic to intraclast packstone also is present in the upper part at 92-M2. Stromatolite reef morphology varies from large patch reefs to tabular reefs and to pinnacle reefs on the top of the submember. At 92-M2 and 92-M3, three tabular reefs separated by coarse dolostone lithofacies but at 92-M5, it forms a single 75 m thick tabular reef.	Conformable under the lowermost stromatolite or at the level coarse dolostone dominates over the rhythmite.	From 55 m at 92-M3 to 75 m at 92-M5.
First Rhythmite	Guttered rhythmite with minor amount of thin-laminated quartzarenite at the base. At 92-M5 and 92-M4, the guttered rhythmite passes upward into non-guttered rhythmite, nodular dolosiltite and back into guttered rhythmite. The upper part contains thin to medium beds of coarse dolostone lithofacies interbedded with guttered rhythmite lithofacies.	Conformable, located where the guttered rhythmite dominates over quartzarenite of the Lower Clastic Member.	From 15 m at 92-M3 up to 40 m at 92-M5.

complete spectrum of dolosiltite-rich rhythmites, with minor amounts of quartzarenite. Quartzarenite is absent from the base but becomes gradually more abundant toward the top (Fig. 2). The contact is gradational between the Lower Carbonate Member and the Lower Clastic Member, at the base, and with the Upper Clastic Member at the top.

The Lower Carbonate Member is composed of three magnafacies: oolitic, stromatolitic, and rhythmitic magnafacies. The rhythmitic magnafacies alternates with either stromatolitic or oolitic magnafacies such that eight locally mappable submembers can be identified (see Table 1). These are, in ascending stratigraphic order: first rhythmite, stromatolite, second rhythmite, first ooid, third rhythmite, second ooid, fourth rhythmite and third ooid submembers (Fig. 2 and 3).

Magnafacies [1]

Rhythmitic magnafacies

Description. The rhythmitic magnafacies is composed of interbedded dololutite dolosiltite and dolarenite, forming repetitive graded beds or rhythmites. This assemblage is typified by guttered rhythmite, nodular dolosiltite, oolitic packstone, and scoured rhythmite lithofacies.

The guttered rhythmite is the most common lithofacies and is composed of 5 to 15 cm thick beds of parallel-laminated dolosiltite to fine dolarenite with common basal gutter casts overlain by hummocky crosslaminated to crossbedded dolosiltite with 1 to 4 cm thick dololutite interbeds. Polygonal ripples having wavelengths averaging 20 cm and amplitudes averaging 3 cm commonly are preserved on the tops of the dolosiltite layers. Rare syneresis cracks are present in the dololutite interbeds.

[1] *Magnafacies* is used in place of the prase lithostratigraphic assemblage

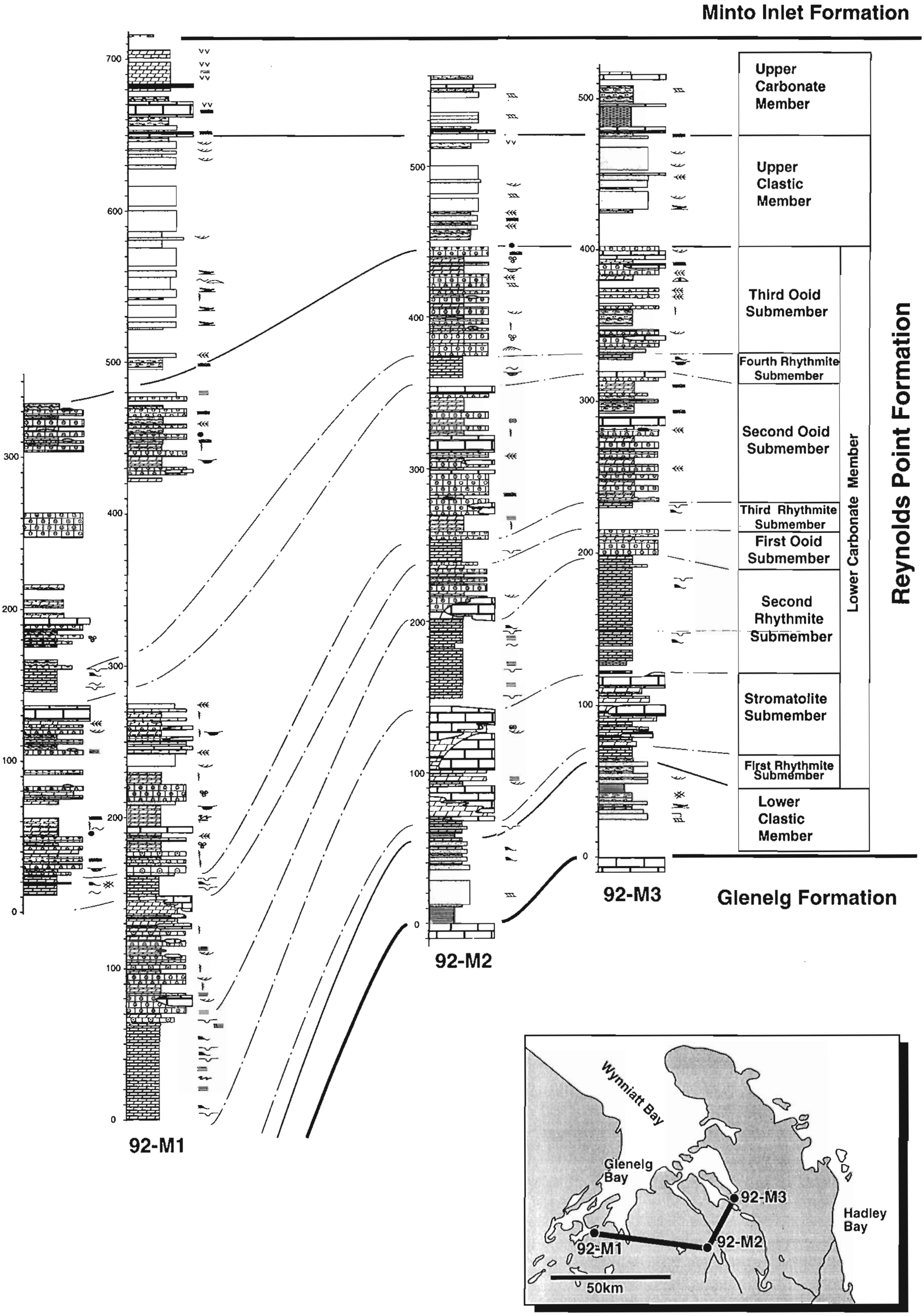

Figure 3. Cross-section of the Reynolds Point Formation in the Wynniatt Bay area. Thick solid lines represents formation boundaries, thin solid lines represent member boundaries, and dotted lines represent submember boundaries.

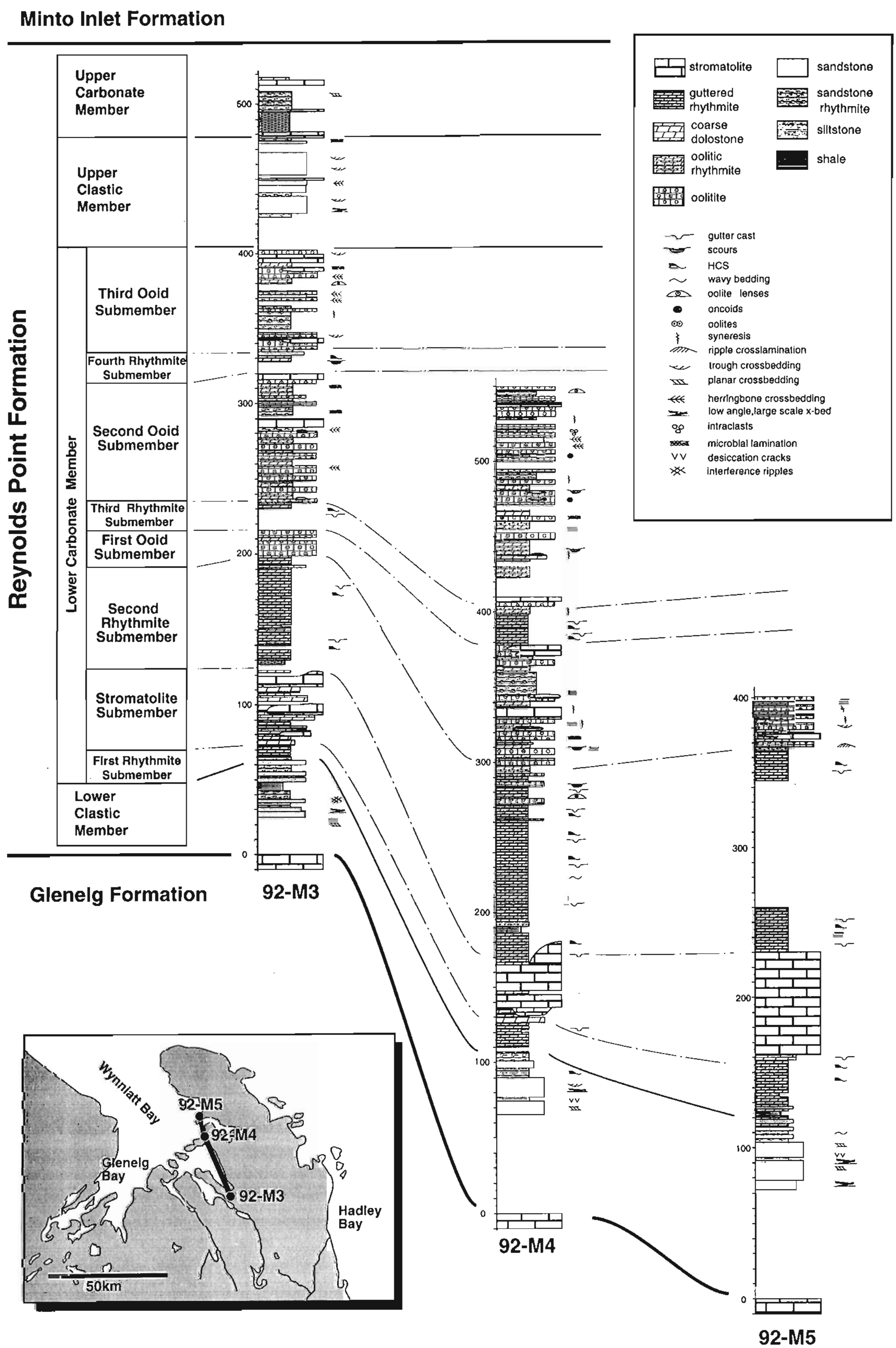

Figure 3. cont.

The nodular dolosiltite lithofacies is composed of nodular dololutite to dolosiltite with faint parallel lamination. This facies occurs in units 1 to 4 m thick and generally is bounded by thin parallel to ripple crosslaminated nonguttered rhythmites.

The oolitic packstone lithofacies is composed of sand to granule size ooids with poorly sorted, granule- to pebble-size, subrounded intraclasts of ooid grainstone and laminated dolosiltite with packstone and rare grainstone texture. It occurs as normally graded and erosive-based lenticular beds that range in thickness from 5 cm to 1 m. The oolitic packstone lithofacies is interbedded with scoured rhythmite lithofacies, a dolosiltite rhythmite with basal scours exceeding 1.5 m width and 8 cm depth.

Interpretation. The rhythmitic magnafacies represents sediments that have been deposited in water depth ranging from below to within the storm wave action zone (Fig. 4). The occurrence of graded rhythmites with gutter casts suggests episodic high-energy event deposition within a relatively quiet environment. A storm origin for the rhythmites is strongly inferred from the co-occurrence of hummocky cross-stratification and polygonal ripples (e.g. Brenchley, 1985; Duke et al., 1991).

The nodular dolosiltite and associated nonguttered rhythmite lithofacies are characterized by normal grading and by a lack of erosional sedimentary structures suggesting distal shelf deposition at or below normal wave base and affected only by infrequent storm deposition.

The oolitic packstone lithofacies exhibits textures and structures suggestive of high-energy transport and final deposition within a relatively quiet environment. It is therefore interpreted as a storm-ebb deposit where sediment was reworked on the shallower part of the shelf and transported into deeper water.

Stromatolitic magnafacies

Description. The stromatolitic magnafacies is composed of large stromatolites interbedded with coarse dolostone lithofacies and rhythmites.

The stromatolite reefs vary from tabular to large patch reefs ranging from 50 m to tens of kilometres wide and 2 to 70 m thick. Most reefs are composed of erect digitate stromatolites with branching columns 1 to 3 cm wide. Stromatolites at patch reef margins are characterized by more strongly radiating columns.

Stromatolite reefs pass laterally and vertically to coarse dolostone lithofacies, composed of stromatolitic fragments. Rock fragments vary from well sorted and well rounded dolorudite and dolarenite grainstone to poorly sorted, subangular dolorudite to dolarenite packstone. This lithofacies occurs in beds ranging from a few centimetres to 2 m thick with some tabular crossbeds 10 to 30 cm thick. Normal grading from coarse dolostone to massive dolosiltite lithofacies is common. The coarse grained dolostone is interbedded with guttered rhythmite, scoured rhythmite, and massive dolosiltite. Massive dolosiltite occurs in thick beds adjacent to some large reefs at section 92-M2 but is medium-bedded at other sections.

Interpretation. The stromatolitic magnafacies is interpreted to have been deposited within the storm wave action zone but below fair-weather wave base (Fig. 4). This is supported by the dominance of stromatolites forming tabular to large patch reefs and their association with guttered and scoured rhythmite lithofacies. The stromatolitic composition of intraclasts of the coarse dolostone lithofacies and its occurrence adjacent to the reefs indicates local erosion and deposition. Variations in grain size, degree of sorting, rounding, and mud content of the coarse dolostone lithofacies suggests reworking by periodic high-energy events such as storms. The thick layers of massive dolosiltite/dololutite associated with the coarse dolostone likely represent fine material derived from the reef and deposited on the reef flanks during slack water periods between storm events.

Oolitic magnafacies

Description. The ooilitic magnafacies comprises ooid grainstone, stromatolite patch reef, and oolitic rhythmite lithofacies that pass laterally and vertically into each other. Sandstone rhythmite lithofacies is also present but only in the upper part of the Lower Carbonate Member.

The ooid grainstone lithofacies occurs in medium to thick beds, up to 1.5 m, with abundant small- to medium-scale bidirectional trough crossbedding. Small-scale dunes with wavelength ranging from 50 to 90 cm are preserved on some bed tops. The ooid grainstone is composed of ooids ranging in size from 1 to 3 mm and common rounded intraclasts ranging in size from 0.5 to 40 cm. Some intraclasts are composed of ooid grainstone and even some contain compound intraclasts of ooid grainstone. In the upper part of the Lower Carbonate Member, oncoidal coating is commonly developed around large intraclasts.

Stromatolite patch reefs commonly range from 1 to 50 m wide and from 50 cm to 20 m thick and pass laterally into ooid grainstone. The patch reefs are composed of branching erect to highly radiating stromatolites with column width ranging from 1 to 4 cm. Tabular reefs 1 to 15 m-thick and more than 100 m wide also are present, but are less abundant than in the stromatolitic magnafacies.

Oolitic rhythmite lithofacies is composed of 3 to 8 cm thick rhythmites that are normally graded from lenticular ooid packstone at the base (0.5 to 3 cm), through dolosiltite (2 to 5 cm), and finally to dololutite (1 to 5 cm). The bases of the graded beds commonly display scours up to a few metres wide and less than 5 cm deep. The dolosiltite typically is parallel-laminated and overlain by rare 1 to 3 cm sets of ripple crosslamination. The dolosiltite contains abundant syneresis cracks and microbial lamination.

Sandstone rhythmite is the siliciclastic equivalent of the oolitic rhythmite and occurs only in the upper part of the formation. It is characterized by laminated to thinly bedded, fine-grained quartzarenite with rare crosslamination and rare oolitic lenses, interbedded with thinly laminated shaly dololutite with abundant microbial lamination and rare syneresis cracks.

Interpretation. The oolitic magnafacies is interpreted to have been deposited in water depths within or close to the fair-weather wave action zone (Fig. 4). It probably represents sediments deposited on inner-shelf shoals composed of ooid banks and stromatolite patch reefs that formed a barrier to offshore waves and restricted water circulation to a back-barrier lagoon where sediments of the ooilitic rhythmite and the sandstone rhythmite were deposited.

The presence of crossbedded ooid grainstone implies a relatively high-energy depositional environment within the fair weather wave action zone. Ooids are today forming in only a few metres of water as sand shoals adjacent to beachs, barrier islands, or shelf breaks (Tucker and Wright, 1990). Patch reefs commonly are interspersed with the ooid grainstone suggesting a co-existing depositional environment.

The oolitic rhythmite lithofacies is characterized by repetitive erosive and normally graded strata suggesting episodic high-energy conditions such as sediment reworking by storm wave washover into a quiet depositional environment. Abundant microbial laminites in the upper part of the oolitic rhythmite suggests a restricted depositional environment, such as a lagoon, which should be expected along a barred coastline setting. Features such as normal grading indicate that the sandstone rhythmite lithofacies was deposited in a similar depositional setting to the ooid rhythmite, however lower energy conditions are assumed because evidence for syndepositional erosion is lacking. The sandstone rhythmite is interpreted as reworked river-borne detritus deposited in the coastal side of the lagoon.

Interpretation of the Lower Carbonate Member

The Lower Carbonate Member is characterized by a cyclic alternation of relatively deep outer shelf deposits (rhythmite submembers) and relatively shallow inner shelf deposits (stromatolite and ooid submembers, Fig. 2). The repetition is caused by a four-stage progradation of a storm-dominated carbonate ramp that was composed of several depositonal belts represented by the three magnafacies described above (Fig. 4). A ramp model is preferred because coarse gravity flow deposits that characterize the basinal portions of rimmed-shelf platforms are absent from the Reynolds Point Formation. The ooilitic magnafacies represents formation of ooid-patch reef shoals and back-shoal lagoons in an inner-shelf setting. The stromatolitic and rhythmitic magnafacies were deposited in mid-shelf to possibly outer-shelf settings mainly within the storm wave action zone. Storm influence is indicated by a complete spectrum of rhythmite lithofacies containing features like hummocky crossbedding, scouring, and graded bedding that occur on all parts of the shelf, from the lagoon to beneath storm wave base (Fig. 4).

Overall progradation of inner ramp facies is suggested by the gradual up-section increase in the ratio of shallower water (ooid and stromatolite) with respect to deeper water (rhythmitic) magnafacies. Specifically, there is an up-section decrease in size and tabularity of the stromatolite reefs and there is more abundant crossbedded ooid grainstone and quartz sand. A general shallowing upward trend is also suggested by more common lagoonal facies such as oolitic rhythmite and microbial lamination in the upper part of the formation. A large lagoon is also inferred by the presence of abundant bidirectional crossbedding indicating increased influence of tidal currents. Nodular dolosiltite, interpreted as

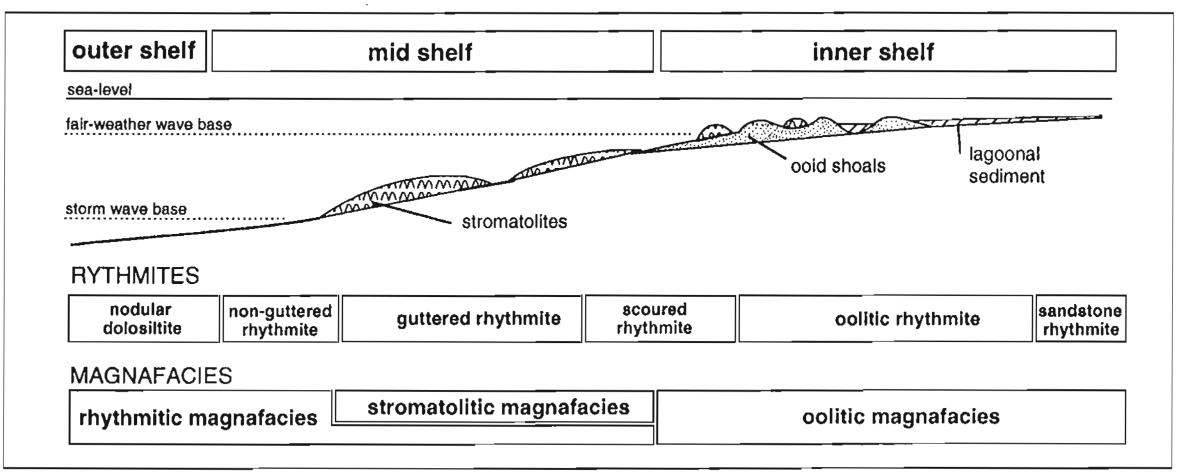

Figure 4. Idealized cross-section of the proposed carbonate ramp setting for the Lower Carbonate Member. The oolitic magnafacies was deposited in an inner shelf position and the stromatolitic and rhythmitic magnafacies were deposited in mid to possibly outer shelf position. The diagram also shows the relative water depths for the rhythmite lithofacies present in the different magnafacies.

the deepest water facies, is common in the first rhythmite submember, rare in the second rhythmite submember and absent from the upper part of the member.

Sedimentary structures that reflect the shallowing upward pattern include gutter casts and hummocky crossbedding, which are abundant within the first rhythmite submember, rare in the second and third rhythmite submembers, and absent from the fourth rhythmite submember. In the fourth rhythmite submember gutters are absent but scoured rhythmite is abundant, reflecting a shallower water setting more strongly influenced by storm events.

The ramp was probably oriented southwest-northeast as suggested by increasing sediment thickness toward the northwest. This is corroborated by the southeast-northwest paleocurrents from gutter casts in the rhythmitic magnafacies and frombidirectional crossbedding of the oolitic magnafacies.

Upper Clastic Member

Description

The basal contact of the Upper Clastic Member is drawn at the top of the uppermost ooid grainstone bed, where there is a gradational change over 50 m between sandy ooid grainstone upward into medium grained quartzarenite and sandstone rhythmite. This member ranges in thickness from 65 m at section 92-M3 to 170 m at section 92-M1 (Fig. 3).

The lower part of the Upper Clastic Member is composed of sandstone rhythmite interbedded with medium bedded, medium grained quartzarenite with bidirectional cross-stratification. This is overlain by subhorizontal planar stratified to low-angle crossbedded quartzarenite with interbedded sandstone rhythmite. At section 92-M1 this unit is overlain by massive to thick bedded but otherwise massive, medium grained quartzarenite.

The upper part of the Upper Clastic Member is composed of medium bedded, medium grained quartzarenite with abundant trough to planar, mainly tabular crossbeds. Paleocurrents from planar crossbeds in the upper part of the member indicate unidirectional flow toward the northeast. The crossbedded unit is overlain by parallel laminated to thinly bedded, fine- to medium-grained quartzarenite interbedded with very thin mudstone interbeds. Desiccation cracks were observed in this unit at section 92-M2. The last few metres of the member are either composed of dolomitic sandstone rhythmite, with microbial laminite and dololutite interbeds or by medium-grained quartzarenite with herring bone crossbedding.

Interpretation

The depositional setting of the Upper Clastic Member is interpreted to be similar to that of the Lower Carbonate Member but with a much higher proportion of siliciclastic detritus in the former. Tidal current influence is clearly recorded by an abundance of bidirectional crossbeds, which are most commonly preserved in tidal channel deposits (see summary in Reinson, 1992). The presence of low-angle crossbedding is common in barrier-beach environments. The sandstone rhythmite facies may therefore represent lagoonal or abandoned tidal channel deposits.

Crossbedding in quartzarenites of the Upper Clastic Member yield bimodal-bipolar, unimodal and polymodal paleocurrent patterns suggesting a complex interaction of marine currents (see data in Young and Jefferson, 1975). A strong unimodal northeasterly trend has been observed in the upper part of the member at sections 92-M1 and 92-M2 suggesting possible fluvial influence, however, regionally and stratigraphically consistent northwest transport has been recorded in underlying fluvial quartzarenites.

Upper Carbonate Member

Description

The Upper Carbonate Member conformably overlies the Upper Clastic Member and is conformably overlain by evaporitic redbeds of the Minto Inlet Formation. The Upper Carbonate Member was present only in three sections (92-M2, 92-M1, and 92-M3) where its thickness is consistently about 65 m; but apparently it thickens to over 200 m in the Minto Inlet area (Young and Long, 1977b). The base of the Upper Carbonate Member is marked by a yellow-weathering stromatolite unit that can been traced throughout Minto Inlier (Young and Jefferson, 1975; Jefferson, 1977; Young and Long, 1977b). Yellow colouration probably is related to the presence of ferrugenous carbonate or fine grained oxidized pyrite. At most sections in the Wynniatt Bay area, the stromatolite is composed of two tabular units, the lower being composed of columnar, laterally linked forms with various amount of intercolumnar siliciclastic material, and the upper being composed of stratiform laterally linked forms with common 30 to 60 cm wide mudcracks. The stromatolite units are separated by parallel-laminated to thinly bedded quartzose dololutite and dolosiltite at section 92-M3 and at section 92-M1. At section 92-M2, the inter-stromatolite unit is composed of medium bedded, medium grained quartzarenite with tabular trough crossbeds. The stromatolite unit is overlain by sandstone rhythmite and by a highly dolomitized tabular stromatolite. The upper part of the member is composed of white to pale grey parallel laminated and mudcracked carbonate (possibly magnesitic) siltite.

Interpretation

Strata of the Upper Carbonate Member are interpreted as shallow marine to lagoonal deposits that gradually evolved into restricted intertidal and possibly to subaerial conditions at the contact with the Minto Inlet Formation. The lower yellow stromatolite is underlain by trough crossbedded sandstone and overlain by sandstone rhythmite suggesting a transition from shoal to lagoonal setting. The upper yellow stromatolite is composed of laterally linked columns with abundant large mudcracks indicating periodic subaerial exposure, perhaps in the intertidal zone. The occurrence of parallel-laminated magnesitic carbonate(?) with abundant mud cracks in the upper part of the submember suggests a restricted peritidal environment. Subaerial deposition may be recorded in redbeds of the overlying Minto Inlet Formation.

SEQUENCES

The submember subdivision of the Reynolds Point Formation defines cyclical alternations of deep water and shallow water facies, which are interpreted as five transgressive-regressive sequences (Fig. 2). The period of this cyclicity is indeterminate because the ages of the strata are unknown; however, the scale of the sequences is comparable to transgressive-regressive sequences in the overlying Minto Inlet and Kilian formations that have been compared with third-order cycles from Phanerozoic sedimentary successions (Rainbird, 1991).

With the exception of the basal sequence, the sequences of the Reynolds Point Formation display no apparent unconformity or abrupt change in sedimentation at their sequence boundaries; transitions from stromatolitic or oolitic magnafacies to the rhythmitic magnafacies are gradational over a few metres. Unconformity-bounded sequence stratigraphy (Vail et al., 1977), however, is applicable to the first sequence composed exclusively of siliciclastic strata, where a prograding fluvio-deltaic high-stand systems tract is separated by an unconformity from overlying shallow marine deposits of a transgressive systems tract (Fig. 2).

The second sequence comprises mixed quartzarenite and dolostone transgressive systems tract and an overlying high-stand system tract composed of rhythmitic to stromatolitic magnafacies. The upper boundary is located at the regressive surface where ooids are present.

The third and fourth sequences are characterized by a transgressive systems tracts of rhythmitic magnafacies overlain by high-stand systems tracts of oolitic magnafacies. The upper boundaries of both sequences are difficult to place because of the gradual transition with the overlying sequence.

The fifth sequence is composed of a relatively thin transgressive system tract of rhythmitic magnafacies and thick high-stand system track composed of ooilitic magnafacies overlain by crossbedded and rhythmitic quartzarenite and stromatolitic dolostone interpreted as shallow marine to restricted lagoonal deposits. The sequence is limited at the top by a possible unconformity beneath supratidal to possibly subaerial redbeds of the Minto Inlet Formation.

The origin of third-order fluctuations are interpreted as global eustatic variations within a passive subsidence regime, because of the symmetric shape of the cycles and the gradual transition between transgressive and regressive deposits.

CONCLUSIONS

The Reynolds Point Formation is composed of four distinct members. The Lower Clastic Member is composed of a mudstone to quartzarenite coarsening upward succession representing the rapid progradation of a river-dominated delta that was later reworked by storms in an inner shelf setting. The Lower Carbonate Member is composed of dolostone that passes gradually to quartzarenite-rich dolostone. It is characterized by an alternation of three magnafacies: rhythmitic, stromatolitic, and oolitic, that represent sediment deposited in mid to outer shelf, mid shelf, and inner shelf settings, respectively. The alternation defines eight lithostratigraphic units that are, in ascending order, first rhythmite, stromatolite, second rhythmite, first ooid, third rhythmite, second ooid, fourth rythmite, and third ooid submembers. The Lower Carbonate Member is interpreted as a four-stage progradation of a storm-dominated carbonate ramp. The Upper Clastic Member comprises crossbedded and rhythmicly bedded quartzarenite, and minor dolostone representing a sand-shoal to barrier-beach and back-barrier lagoon with possible fluvial influence. The Upper Carbonate Member is composed of stromatolitic dolostone, quartzarenite and mudcracked carbonate (magnesitic?) that probably was deposited in a shallow lagoon that evolved to more restricted intertidal and finally to a subaerial setting.

The Reynolds Point Formation is characterized of five third-order(?) sequences bounded by unconformities or by equivalent conformities. With the exception of the first sequence, which could be tectonic, the relative sea-level fluctuations are interpreted as global eustatic variations within a passive subsidence regime.

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Geological Survey of Canada Project 890011 KD

Preliminary Precambrian geology in the vicinity of Eqe Bay, Baffin Island, Northwest Territories[1]

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Abstract: The area under investigation lies between the Archean Committee Fold Belt and the Proterozoic Foxe Fold Belt. It contains quarzofeldspathic gneisses and two greenstone belts of probable Archean age. These belts contain metavolcanic rocks, metasedimentary rocks, and iron-formation that may be correlative with the Mary River Group. A sporadic assemblage of marble, calcsilicates, and metapelite in the northwest part of the area may be correlative with the Proterozoic Piling Group. Gneisses and supracrustal rocks across the map area are intruded by a number of granitic plutons of probable Proterozoic age. Metamorphic grade is mainly greenschist in the southeast part of the area. To the northwest, the Isortoq Fault Zone marks a transition from amphibolite to granulite facies across several moderately to steeply southeast-dipping ductile-brittle faults. Notable economic mineralization in the greenstone belts is associated with mafic volcanic rocks, iron-formation, ultramafic rocks, and quartz-carbonate veins.

Résumé : Le secteur actuellement à l'étude se situe entre la zone de plissement de Committee, d'âge archéen, et la zone de plissement de Foxe, d'âge protérozoïque. Il contient des gneiss quartzo-feldspathiques et deux zones de roches vertes probablement d'âge archéen. Ces zones contiennent des roches métavolcaniques, des roches métasédimentaires et une formation ferrifère qui peuvent peut-être être mises en corrélation avec le Groupe de Mary River. Un assemblage sporadique de marbre, de calco-silicates et de métapélite apparaît dans la partie nord-ouest du secteur et pourrait être corrélatif du Groupe de Piling d'âge protérozoïque. Les gneiss et les roches supracrustales de toute cette région cartographiique sont traversés par plusieurs plutons granitiques probablement d'âge protérozoïque. Le degré de métamorphisme est principalement celui du faciès des schistes verts dans la partie sud-est du secteur. Au nord-ouest, la zone de failles d'Isortoq marque la transition entre le faciès des amphibolites et celui des granulites de part et d'autre de pluseurs failles ductiles et cassantes de pendage sud-est moyen à prononcé. Une minéralisation appréciable des zones de roches vertes, d'intérêt économique, est associée à la présence de roches volcaniques mafiques, d'une formation ferrifère, de roches ultramafiques et de filons carbonatés quartzeux.

[1] Contribution to Canada-Northwest Territories Mineral Initiatives 1991-1996, a subsidiary agreement under the Canada-Northwest Territories Economic Development Agreement. Project funded by the Geological Survey of Canada.

INTRODUCTION

This report presents preliminary results of 1:50 000 scale mapping of Precambrian rocks exposed in the vicinity of Eqe Bay, northwestern Baffin Island (Fig. 1). The map area is within the northeast quarter of NTS 37 C (Koch Island). Work initiated during the 1992 season is part of a three year mapping program under the Mineral Development Agreement between the Geological Survey of Canada and the Government of the Northwest Territories.

The earliest mapping in the area was undertaken by Blackadar (1958, 1963). A more comprehensive coverage was obtained during Operation Bylot (Jackson, 1968, 1969), which mapped north-central Baffin Island. At the same time, an assessment was made of the geological setting and resource potential of iron-formation in the vicinity of Eqe Bay by Patino Mining Corporation in 1969, and also by Crawford (1973). Further mapping was carried out in 1974 and 1975, resulting in a 1:250 000 scale geological map for the area (Morgan, 1982). The aim of the present project is to produce four preliminary geoscientific maps (open file) of the area at 1:50 000 scale and a final A-series compilation map at 1:100 000 scale, coupled with an assessment of the mineral resource potential.

GEOLOGICAL SETTING

The work of Jackson (1966, 1991, in press), Jackson and Taylor (1972), Jackson et al., (unpub. data, 1990; 1990), Morgan (1982) and Henderson et al. (1989) has established a geological framework for the map area. Two northeast-trending orogenic belts of Precambrian age are exposed on northwestern Baffin Island and adjacent Melville Peninsula (Fig. 1). The Committee Fold Belt comprises Archean greenstone belts (Mary River and Prince Albert groups) plus granitic and gneissic rocks. The Foxe Fold Belt consists of early Proterozoic supracrustal rocks (Piling and Penrhyn groups) which unconformably overlie Archean basement rocks of the Committee Fold Belt. Rocks of the Foxe Fold Belt were thrust northwestward and were subsequently intruded by voluminous plutons related to the 1.89-1.85 Ga Cumberland batholith. The northwest limit of thrusting associated with the Foxe Fold Belt is interpreted to coincide with the Isortoq Fault Zone which strikes northeast across Baffin Island (Jackson, in press) and is inferred to extend into the map area along Isortoq Fiord (Fig. 2).

Compilation and mapping by Morgan (1982) delineated two northeast-trending greenstone belts in the Eqe Bay area (Fig. 2). Two U-Pb bulk zircon ages indicated an Archean age for these rocks. Morgan (1982) tentatively correlated them with rocks of the Mary River Group (Jackson, 1966) of the Committee Fold Belt. Protoliths of both belts are mafic and felsic volcanic rocks, semipelitic to pelitic sedimentary rocks, and banded iron-formation. Morgan also identified a different association of supracrustal rocks characterized by quartzite, pelite, marble, and calcsilicate rock in the northwest part of the map area. These rocks were correlated with the Proterozoic Piling Group on the basis of common marble and calc-silicate rocks (W.C. Morgan, pers. comm., 1992). Morgan also interpreted several granitic plutons in the map area to have a Proterozoic age.

In 1992, fieldwork concentrated on rocks northwest of Isortoq Fiord and in the northeast-trending belts of Archean supracrustal rocks: the 'Isortoq Belt', located southeast of Isortoq Fiord and the 'Eqe Bay Belt', passing through Eqe Bay (Fig. 2).

LITHOLOGY

Quartzofeldspathic and mafic gneisses

The dominant map unit northwest of Isortoq Fiord comprises quartzofeldspathic to mafic gneiss. These gneisses are typically well foliated and layered from centimetre- to decimetre-scale. Layering is defined by variation in mafic mineral content of adjacent layers. The gneisses are commonly migmatitic, with up to 25% layer-parallel, granitic leucosome. White weathering, medium grained, quartzo-feldspathic gneiss makes up a large part of this assemblage. This rock varies from tonalite to granite in composition but is most commonly granodiorite. It has a variable foliation, normally defined by thin biotitic seams. Locally, the seams are more concentrated and appear to define relict xenoliths with gradational contacts. Intrusive relations are inferred where metre-scale blocks of mafic gneiss, amphibolite, or pelitic gneiss are bounded by sharp contacts within the quartzofeldspathic gneiss. These inclusions generally increase in abundance toward the greenstone belts.

K-feldspar megacrystic granite gneiss is a distinctive unit associated with the Isortoq greenstone belt. The gneiss is medium grained, and typically contains scattered, 2-3 cm long K-feldspar megacrysts; a subsidiary component of this gneiss is pink, fine grained granite that forms centimetre- to decimetre-scale sheets or dykes that cut the foliation at a low angle.

Isortoq Belt

The Isortoq greenstone belt (Fig. 3) is divided into two parts by a steeply southeast-dipping fault, designated the 'Main Fault' on Figure 3, and which is described in more detail below. A crude tectonostratigraphy can be recognized in part of the footwall (FW, Fig. 3). From northwest to southeast, metapelitic gneiss with iron-formation is the first rock unit encountered structurally above quartzofeldspathic gneisses. This is overlain by a thin layer of amphibolite, in places interlayered at decimetre-scale with metapelite. The amphibolite is fine- to medium-grained, strongly foliated and layered at centimetre- to decimetre-scale. Layering is defined in part by subtle variations in quantity of mafic minerals and in part by grain size variation. Layering is also enhanced by elongate aggregates and stringers of coarse green clinopyroxene. Primary structures in these rocks are generally obscured by strain; however, the bulk composition of the amphibolite and its close association with meta-

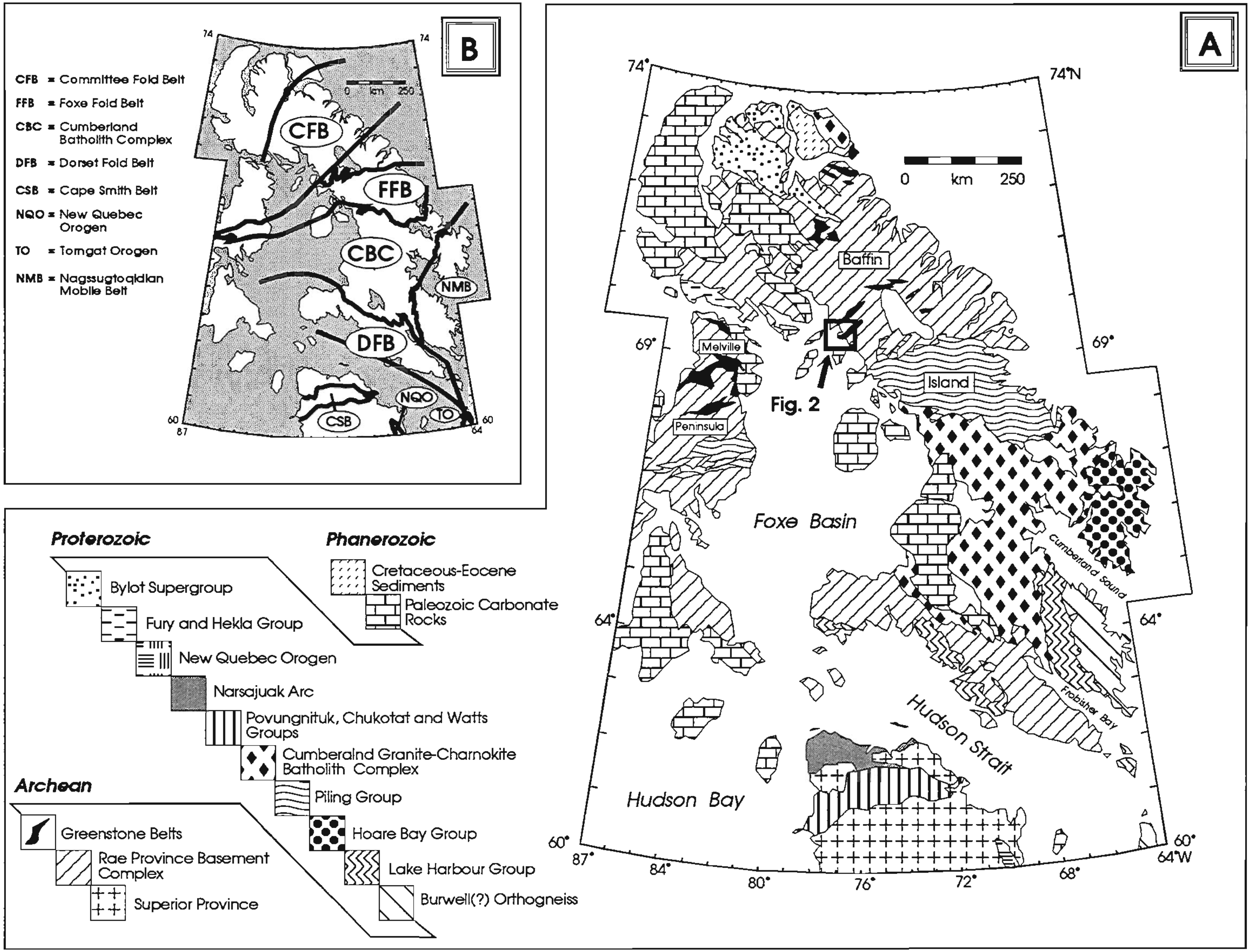

Figure 1. General geology and tectonic subdivisions of Baffin Island (after Hoffman, 1988; Jackson et al., unpub. data, 1990; 1990).

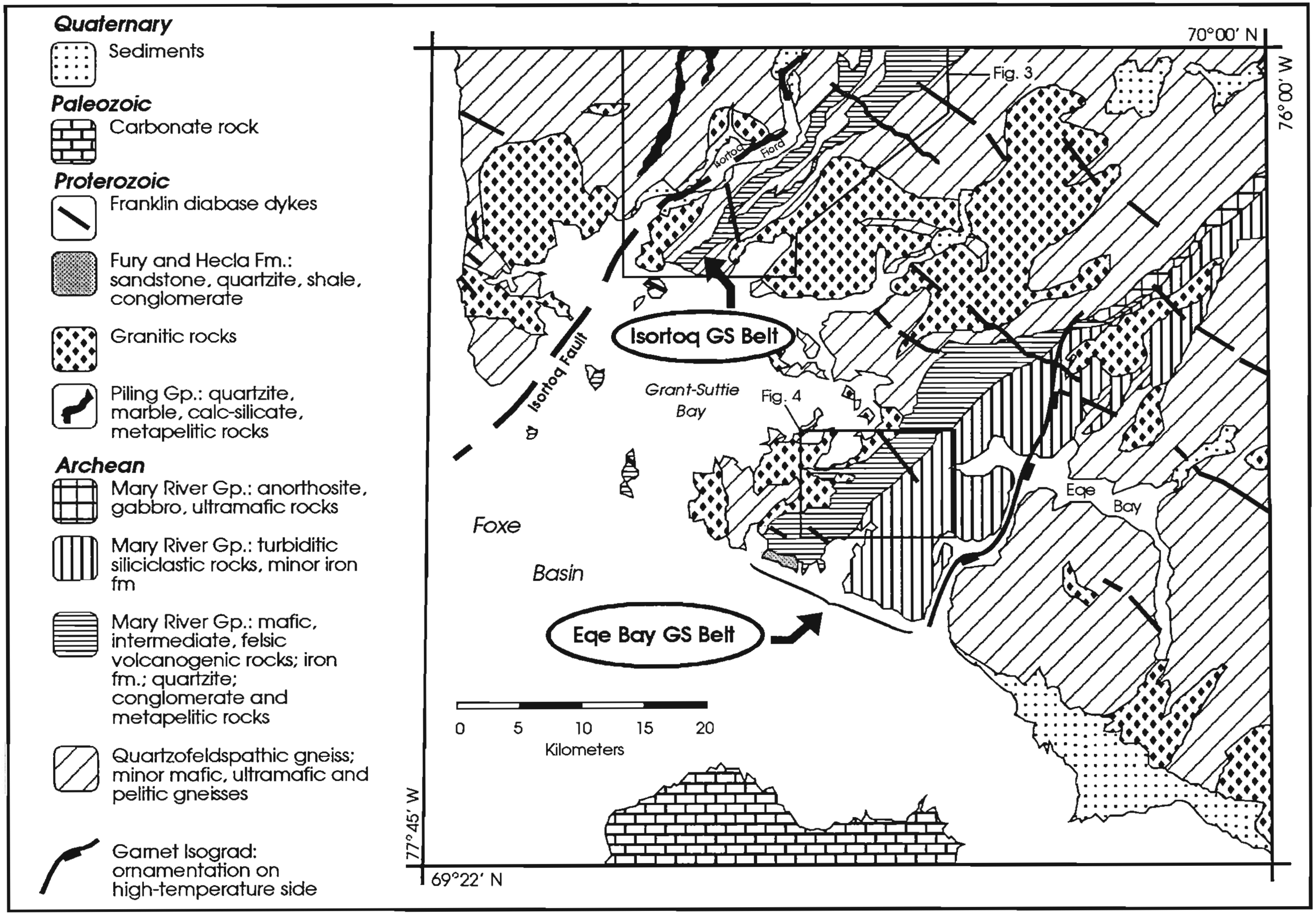

Figure 2. Geology in the vicinity of Eqe Bay (mainly after Morgan, 1982).

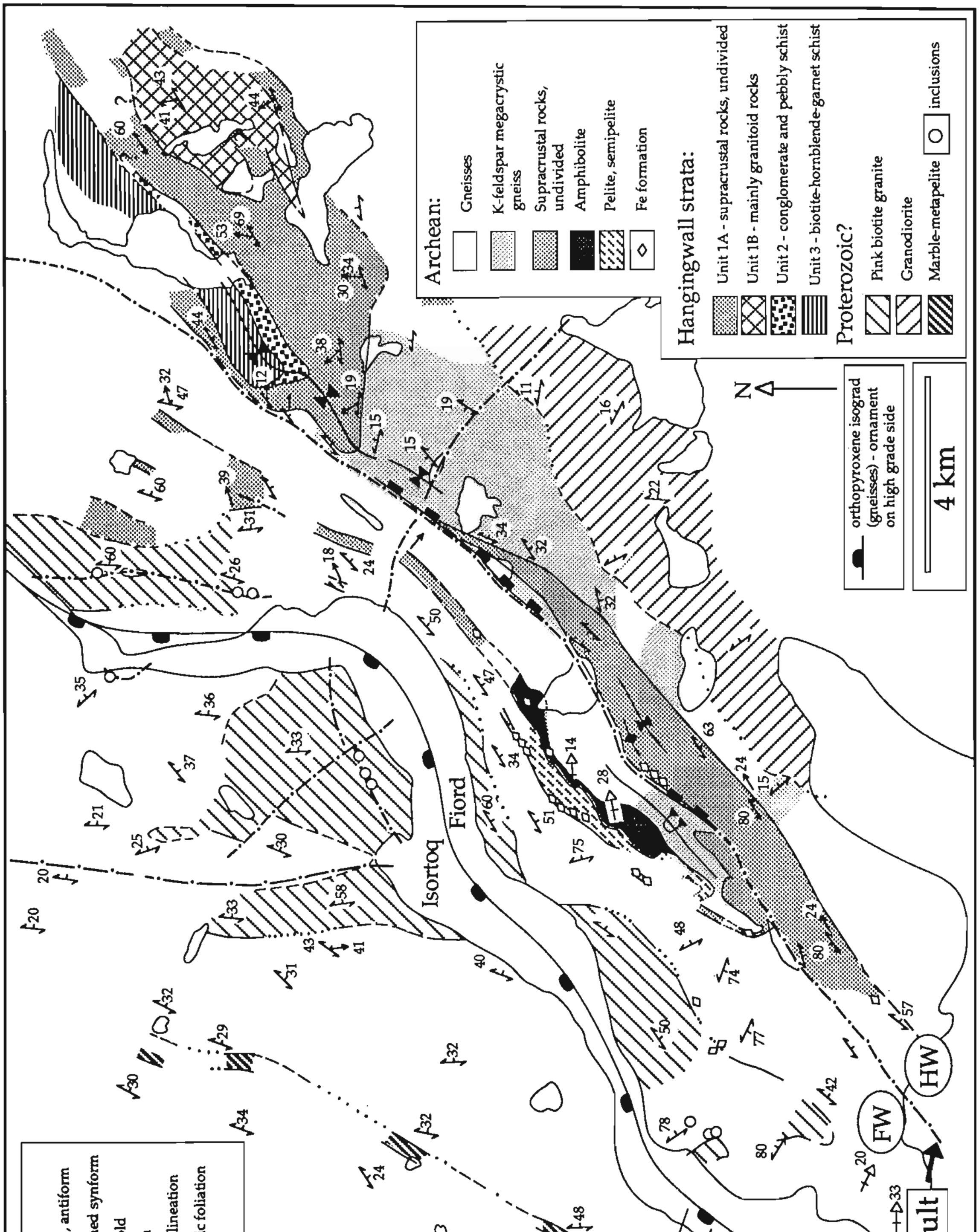

Figure 3. Preliminary geology in the vicinity of Isortoq Fiord.

sedimentary rocks strongly suggests a mafic volcanic protolith. Some of the more finely layered amphibolite may represent former volcaniclastic sedimentary, or epiclastic rocks.

In the southeast part of the belt, the supracrustal rocks of the hanging wall (HW, Fig. 3) comprise interlayered amphibolite, pelite, and semipelite. The stratigraphic order of these rock types is not known but structural observations and interlayering of different rock types indicates probable repetition by folding (see section on structure below). In contrast, a predictable stratigraphy has been identified in supracrustal rocks in the northeast part of the hanging wall. Three composite units were mapped and are described below in order of ascending structural level.

The first unit (unit 1A; Fig. 3) comprises a heterogeneous succession of dominantly mafic to intermediate metavolcanic rocks and associated volcaniclastic metasedimentary rocks. Subordinate rock types include felsic metavolcanic rocks, semipelite, pelite, and banded iron-formation. The unit grades from mainly metabasaltic rocks at the base to intermediate volcanic and volcaniclastic rocks at the top, with an accompanying upward increase in amount of semipelitic material.

The second unit (unit 2) is a discontinuous volcaniclastic orthoconglomerate that reaches a maximum thickness of about 80 m. The base of this conglomerate is characterized by elongate, pebble- and cobble-sized clasts of volcanic rocks (andesite, dacite, rhyolite) and quartzite. This rock grades upward into a paraconglomerate with similar types of clasts in a schistose matrix. The paraconglomerate is gradationally overlain by a biotite-hornblende-garnet schist with thin interlayers of pebble-granule conglomerate.

The third unit (unit 3) is a distinctive biotite-amphibole-garnet schist. In places it contains large sprays of actinolite and porphyroblastic garnet up to 2 cm in diameter.

Iron-formation

Iron-formation in the footwall is primarily oxide facies with alternating millimetre- to centimetre-scale layers that are magnetite-hematite- or silica-rich. Oxide facies iron-formation locally grades into, or is interlayered with silicate facies iron-formation. The latter contains grunerite, with or without garnet. The iron-formation occurs as discontinuous, foliation-parallel layers and lenses, ranging from metres to tens of metres wide, within both gneisses and supracrustal rocks. Various 'segments' appear to be large-scale boudins, pulled apart along strike. Some are also clearly disrupted by folding. In contrast to the footwall, iron-formation in the hanging wall is less abundant and the silicate facies seems to predominate over oxide facies.

Eqe Bay Belt

In contrast to middle to upper amphibolite facies rocks of the Isortoq belt, supracrustal rocks of the Eqe Bay Belt are dominantly at greenschist facies. Three principal rock units are recognized within this belt (Fig. 2). From northwest to southeast, these are: 1) a thick succession of mafic to intermediate volcanic rocks with minor felsic volcanic rocks, quartzite, conglomerate, and iron-formation; 2) a thin unit of intermediate to felsic volcanic rocks, and 3) a thick succession of siliciclastic rocks with subordinate iron-formation. Units 2 and 3 appear to be everywhere separated by a conglomerate containing sedimentary, volcanic, and plutonic clasts. The contact between unit 1 and 2 is a gradational zone wherein intermediate volcanic rocks become increasingly abundant within the succession. A transect between Grant-Suttie Bay and Eqe Bay (Fig. 4), in which most rock types of this belt are encountered, is described below.

Mafic volcanic rocks rocks are green, fine grained and vary from massive to highly schistose or slaty. Relict plagioclase phenocrysts are rare. Altered amygdules, vesicular texture, and pillow structures are locally preserved. Fine grained material is commonly in gradational or sharp contact with medium grained metagabbro that probably constitutes former sills or thick flows. One layer of quartz-rich metasandstone within the volcanic succession is fine- to medium-grained, partly recrystallized, and greater than 10 m thick.

Discontinuous layers of iron-formation, up to tens of metres thick and several kilometres long, occur within mafic volcanic rocks. Oxide and silicate facies are intergradational; minor carbonate facies is also present. The oxide facies consists of alternating, centimetre-scale beds of hematite-magnetite and red chert (jasper). In some locations, chert has been totally converted to crystalline quartz. The silicate facies is characterized by bedding of similar scale containing combinations of quartz-grunerite-magnetite (± hematite). Grunerite occurs as very fine stellate rosettes that overgrow compositional layering. Both silicate- and oxide-facies iron-formation are commonly interlayered with magnetite-bearing mafic volcanic rocks. One prominent layer of iron formation is structurally overlain by a 1 to 5-metre-thick, discontinuous conglomerate characterized by attenuated, pebble-sized clasts of predominantly mafic volcanics with subordinate quartzite and iron-formation, in a fine grained argillaceous matrix.

The succession of mafic volcanic rocks with banded iron-formation is structurally overlain by an approximately 10 m thick layer of brecciated iron-formation. The breccia contains angular, cobble- to boulder-sized clasts of highly deformed iron-formation in a fine grained, green matrix. Stucturally above it is a unit of felsic to intermediate volcanic rocks represented at this location by a fine grained, light grey plagioclase-phyric dacite (Crawford, 1973). Along strike, the same unit consists of a grey to pink quartz-feldspar porphyry that intrudes the mafic metavolcanic rocks. Locally, this porphyry grades into a rock that contains rhyolite lapilli and quartz and feldspar phenocrysts in a fine grained matrix; the latter rock may have a volcaniclastic origin. Intermediate to felsic crystal and vitric tuffs are also present within the intermediate to felsic volcanic unit.

Intermediate to felsic volcanic rocks are gradationally overlain to the southeast by a roughly 10 m thick layer of volcanic agglomerate (or breccia) consisting of attenuated lapilli to bomb-sized fragments of plagioclase-phyric dacite and rhyolite in a fine grained, schistose matrix. Southeastward, this volcaniclastic rock grades up structural section into a paraconglomerate of similar thickness. The conglomerate contains highly elongate pebble- to cobble-sized clasts of siliciclastic rocks, mafic and intermediate volcanic rocks, and granitoid rocks in a strongly foliated silty to arkosic matrix.

The paraconglomerate has a relatively sharp contact with predominantly shaly rocks at the base of the overlying metasedimentary unit. A discontinuity between volcanic rocks and the overlying clastic sedimentary rocks is suggested by the following: 1) the contact is marked by a thick conglomerate that indicates a significant erosional episode between volcanism and depostion of clastic sedimentary rocks; 2) in the southern part of the map area there is significant discordance between the strike of the mafic volcanic-iron-formation unit, and the strike of overlying siliciclastic rocks; 3) no evidence exists for a major dislocation (i.e. thrust, shear zone) along the contact; 4) the contact does not appear to be deformed to the same degree as underlying units. These observations strongly suggest that the discontinuity represents an angular unconformity, but does not rule out the possibility that it is a fault. Note that these relationships, if stratigraphic, also imply that rock units are broadly southeastward-facing. The breccia on the northwest side of the intermediate to felsic volcanic unit may also indicate a period of disruption (of uncertain origin) between mafic and felsic volcanism.

The thick metasedimentary rock unit of the Eqe Bay Belt is typified by alternating sandy to shaly, and rarer pebbly beds that are laterally continuous (over tens of metres) and have sharp bases (so-called 'turbidites'). Bedding varies between millimetre and decimetre scale. Individual beds generally have an even grain size distribution; however, some are graded and have associated ripple cross-laminations. Much of the succession is composed of fine- to medium-grained, moderately well-sorted arkose or lithic-arenite. Beds of coarser arkose and pebble-granule conglomerate may be two or more metres thick. East-northeast of Eqe Bay, iron-formation occurs in association with these clastic rocks.

Marble-metapelite Assemblage near Isortoq Fiord

Interlayered pelitic to semipelitic schist, marble, calc-silicate rock, and minor mafic gneiss and quartzite comprise this assemblage although not all of these rock types are present at all locations. Northwest of Isortoq Fiord (Fig. 3), the unit occurs as a discontinuous layer several kilometres long and less than 50 m wide. Elsewhere, it occurs as metre- to ten-metre-scale

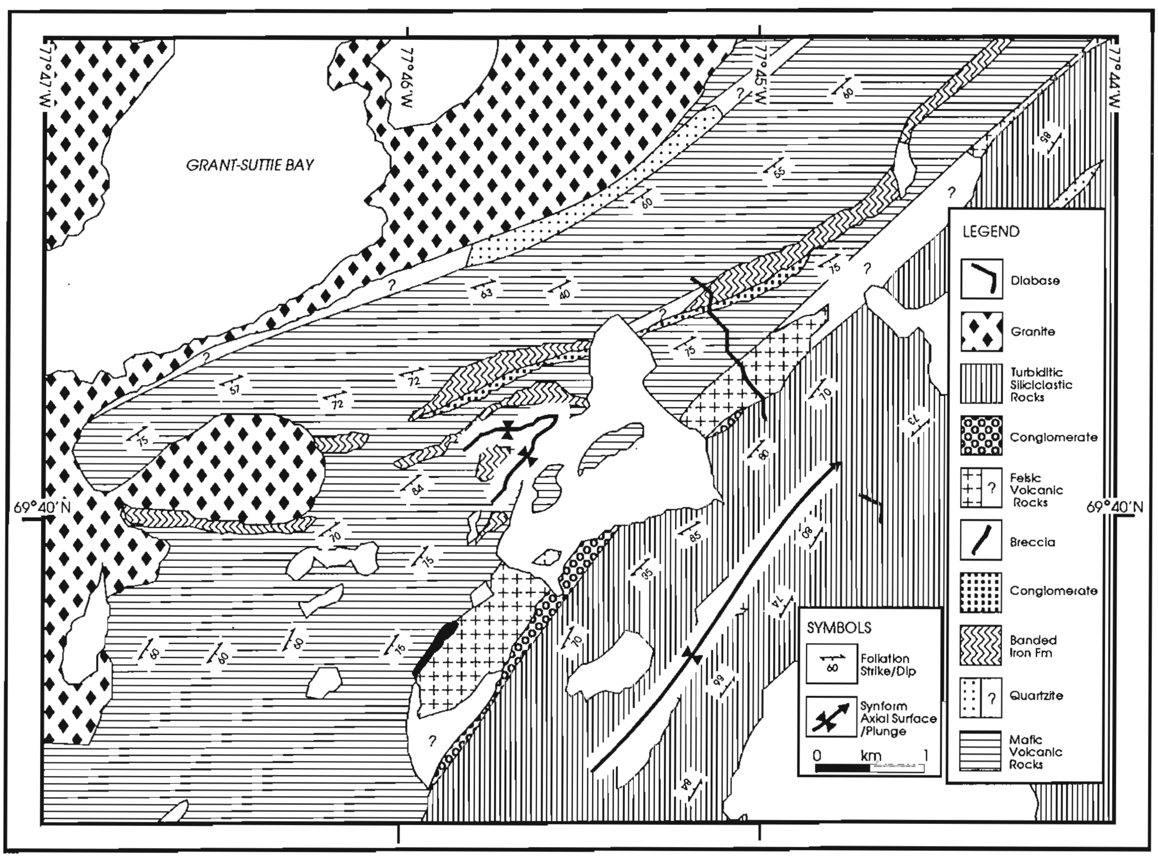

Figure 4. Geology between Grant-Suttie and Eqe Bay (in part after Crawford, 1973).

marble-bearing xenoliths within younger granitic rocks. It is also possible that some metapelite within quartzofeldspathic gneisses represents part of this assemblage.

Intrusive Rocks

Granitoid plutons

Several granitic plutons intrude gneisses throughout the map area (Fig. 2). Northwest of Isortoq Fiord (Fig. 3), two of these comprise uniform, pink-weathering biotite granite which is medium grained, equigranular, and is generally weakly foliated; locally, it contains patches of syenogranite pegmatite. Southeast of the Isortoq belt, gneisses are in gradational contact with a weakly foliated, white-weathering, biotite-hornblende granodiorite. Near Grant-Suttie Bay, equigranular to K-feldspar megacrystic biotite-hornblende granodiorite intrudes supracrustal rocks of the Eqe Bay Belt. In contrast to plutonic units in the vicinity of Isortoq Fiord, this pluton is massive.

Anorthosite and ultramafic rocks

A narrow kilometre-scale intrusion of gabbroic anorthosite structurally underlies mafic volcanic rocks in the northeast part of the Eqe Bay Belt; similar anorthosite also occurs as minor inclusions within granites and gneisses along the southeastern margin of the Isortoq Belt. The gabbroic anorthosite consists of rounded plagioclase crystals, up to 3-4 cm in diameter, set in a matrix of green hornblende and actinolite. Throughout the map area, minor ultramafic rocks occur as metre- to tens of metre-scale boudins or lenses in quartzofeldspathic gneisses, and are also locally associated with metapelitic, anorthositic and mafic metavolcanic rocks. Metamorphism has obliterated primary textures in these rocks.

Franklin diabase dykes

Diabase occurs as vertical, northwest-striking dykes, 1-100 m in width. They cut all Precambrian map units and are considered members of the late Proterozoic Franklin swarm (Morgan, 1982).

STRUCTURE

Quartzofeldspathic and mafic gneisses

Where they have been most thoroughly examined along Isortoq Fiord, quartzofeldspathic and mafic gneisses exhibit significant ductile strain in the form of rare tight to isoclinal or intrafolial folds, boudinage of metre-scale layers of mafic gneiss or metapelite, and variable transposition of leucosomes of different generations into the main foliation plane. It is notable that mineral lineations are not well developed in these gneisses except in close proximity to supracrustal rocks. Rocks of the marble-pelite assemblage are locally tightly folded with neighbouring gneisses about subhorizontal axes that are coaxial with high-grade mineral lineations.

Isortoq Belt

The Isortoq belt is divided into two parts by a prominent, steeply southeast-dipping brittle fault ('Main Fault'; Fig. 3). In the footwall of this fault, southeast-dipping supracrustal rocks and their bounding gneisses are folded about east-northeast-plunging axes, forming a regional-scale synform that closes to the southwest. The southeast limb of this fold is overturned and is truncated by the fault. Supracrustal rocks of the immediate hanging wall are predominantly southeast-dipping. To the northeast, supracrustal rocks in this part of the hanging wall taper out but farther along strike reappear at the same structural level, where they are exposed in an east-northeast-plunging synform. The previously described K-feldspar megacrystic gneiss occupies most of the core of the regional synform in the footwall and also structurally underlies supracrustal rocks of the hanging wall. In the vicinity of the main fault, this gneiss is moderately southeast-dipping and strongly sheared; C-S fabric, shear bands, and rotated porphyroclasts indicate a normal sense of displacement. At outcrop-scale, supracrustal rocks and gneisses of the Isortoq Belt have a strong foliation and lineation. The foliation is defined by compositional layering and coplanar metamorphic minerals. The lineation is defined by aligned, elongate metamorphic minerals and mineral aggregates. It generally plunges moderately to steeply east-northeast and is coaxial with centimetre- to metre-scale, open to isoclinal folds that fold the metamorphic foliation.

Rocks in the vicinity of Isortoq Fiord are cut by numerous cataclastic and brittle faults. Several north-northeast-trending faults cut the pink foliated granite. Most occur along 'weak' zones of inclusions of supracrustal rocks, notably marble. Two of the faults have narrow mylonite zones that are overprinted by discrete, epidote-chlorite grade faults with slickenslides. C-S fabrics and slickenstriae indicate a ductile to brittle, extensional kinematic history for one of these faults.

Eqe Bay Belt

Rocks in the northwest part of the Grant-Suttie Bay-Eqe bay transect form a steeply southeast-dipping panel. Just southeast of their contact with volcanic rocks, siliciclastic rocks show a dip reversal that marks the trace of a tight regional synform; a few top determinations suggest that the southeast limb of this fold is overturned. At outcrop scale most rocks have a near bedding-parallel, schistose to slaty to spaced cleavage and an almost dip-parallel lineation. The exact nature of this lineation is uncertain due to the fine grain size of these rocks. Conglomerate clasts and relict pillows are typically highly elongate, with elongation axes coaxial with the regional lineation. Three phases of coaxial folding were identified within the iron-formation at outcrop scale. Folding was also recognized within adjacent volcanic rocks, but delineation of chronological fold phases is difficult. At least two phases of co-axial folding of iron-formation are also present in the siliciclastic succession near Eqe Bay.

METAMORPHISM

Northwest of Isortoq Fiord gneisses normally have the greasy green appearance typical of granulite-facies rocks; mafic layers commonly contain hornblende, orthopyroxene, and clinopyroxene, locally with garnet. Granitic veins and sweats may also carry orthopyroxene or garnet. Layers and inclusions of metasedimentary gneiss contain biotite and garnet, with or without sillimanite or cordierite. Southeast of the fiord, the main brittle fault of the Isortoq Belt separates upper-amphibolite-facies rocks in the footwall from amphibolite- or lower-amphibolite-facies rocks in the hanging wall. In the footwall migmatitic pelites generally contain garnet and sillimanite; amphibolites contain hornblende, diopside, garnet and biotite. In contrast, sillimanite has not been identified in the hanging wall and hornblende is locally accompanied by actinolite, epidote, or chlorite. A lower grade is therefore inferred for the hanging wall. In the vicinity of Isortoq Fiord, marble contains diopside, scapolite, tremolite, or phlogopite.

Most supracrustal rocks between Grant-Suttie and Eqe bay are metamorphosed to lower greenschist facies. Metabasites contain biotite, chlorite, and actinolite. Metapelites contain biotite and muscovite. Approaching Eqe Bay from the west, a regional isograd (Fig. 2) marks the appearance of garnet in siliciclastic rocks and iron-formation. Along the northwest arm of Eqe Bay, garnet in metapelites is accompanied by porphyroblasts of andalusite, staurolite, and cordierite which overgrow a strong micaceous foliation.

ECONOMIC GEOLOGY

Economic mineralization is associated with mafic volcanic rocks, iron-formation, ultramafic rocks, and late veins in both greenstone belts. Gossanous rocks invariably contain sulphide minerals, including pyrite, chalcopyrite, and pyrrhotite. Some iron-formation is composed of greater than 60% magnetite or hematite, or both. Quartz and quartz-carbonate veins cutting mafic volcanic rocks at Eqe Bay commonly contain pyrite, chalcopyrite, or malachite. Malachite and sulphide minerals are also associated with ultramafic rocks and with late brittle faults. Along Isortoq Fiord, inclusions of the marble-pelite assemblage along fault zones within the pink granite also show minor sulphide mineralization.

DISCUSSION

Within the map area, the Isortoq and Eqe Bay belts have supracrustal rocks at different metamorphic grade making detailed correlation between rock types of the two belts difficult. However, the following lithological similarities indicate a broad correlation: 1) both belts are characterized by a thick succession of volcanogenic rocks and related volcaniclastic deposits, plus iron-formation, and 2) protoliths for high grade mafic rocks in the Isortoq Belt can readily be envisaged to have once been similar to mafic rocks that are better preserved in the Eqe Bay Belt. Despite these similarities, the belts have one striking difference - the Isortoq Belt lacks a thick succession of turbiditic clastic rocks. Whether this difference primarily reflects contrasting depositional environments or is the result of later structural or metamorphic effects is uncertain. Regional correlation of supracrustal rocks of the Isortoq and Eqe Bay belts with the Mary River Group (Morgan, 1982; Crawford, 1973; Jackson, in press) seems reasonable considering overall lithological similarities (see Jackson, 1966). However, two bulk U-Pb zircon ages from the same unit of intermediate to felsic volcanic rocks at Eqe Bay (2888 ± 35 and 2752 ± 30 Ma; Morgan, 1982) and recent, more precise U-Pb zircon ages from dacite within the Mary River Group (2718 +5/-3 Ma; Jackson et al., 1990) barely overlap within error, indicating that this correlation requires further testing.

A subject of concern to Jackson and co-workers in the Mary River area is the relationship between the Mary River Group and surrounding gneisses. Conglomerates close to the base of the Mary River Group imply that subjacent gneisses represent an older basement. This was later confirmed by an age difference between basement gneisses and supracrustal rocks (Jackson et al., 1990). In the Eqe Bay Belt, granitic cobbles within conglomerate overlying volcanic rocks similarly imply erosion of a granitic source terrain that might represent basement. However, where they have been closely examined to date, contact relationships between quartzo-feldspathic gneisses and greenstone belt supracrustal rocks are ambiguous owing to high levels of strain. It is therefore impossible to ascertain whether the gneisses represent basement or are intrusive. Note that it is possible that these gneisses have both older and younger, remobilized components.

The Isortoq Fault Zone was inferred to extend through the map area along Isortoq Fiord (Jackson, in press). This zone was originally delineated across Baffin Island on the basis of a sharp break in regional structural trends, metamorphic grade, and aeromagnetic and gravity signatures (Jackson et al., 1978; unpub. data, 1990; 1990; Jackson, in press). Its most obvious expression in the field area is a change from amphibolite to granulite facies across the fiord. Tight folds within the Isortoq Belt are evidence for large-scale contractional strain, some of which may have been penecontemporaneous with granulite facies metamorphism northwest of the fiord. Several faults, including the main brittle fault of the Isortoq Belt and a mylonitic fault on the northwest side of the fiord, postdate these folds but show no evidence for appreciable displacements. Preliminary mapping indicates that some of these faults are normal, consistent with the southeastward decrease in metamorphic grade across the belt. The faults appear to be superimposed on, and to have possibly enhanced, a pre-existing structural and metamorphic transition. If the high-grade rocks of the marble-metapelite association northwest of the fiord are part of the Piling Group, as inferred by Morgan (1982), some of the folding and granulite facies metamorphism along Isortoq Fiord must be Proterozoic.

As deduced by Jackson and Morgan (1978), the northwestward increase in metamorphic grade near Isortoq Fiord is regional in nature and is not due to granitic magmatism. In contrast, the generally lower grade rocks of the Eqe Bay Belt increase in grade eastward, where they exhibit a garnet isograd of regional extent. It is not yet certain whether this isograd is a product of regional or contact metamorphism. However, the latter possibility is strongly supported by the presence of porphyroblasts of andalusite, cordierite, and staurolite that overgrow the foliation, and the presence of granitoid bodies in the eastern part of the belt.

SUMMARY

The following points briefly summarize results of preliminary mapping in 1992:

1) The field area contains quarzofeldspathic gneisses and two greenstone belts characterized by mafic to felsic volcanic rocks, banded iron-formation, and silisiclastic, semipelitic, and pelitic sedimentary rocks that may be correlative with the Archean Mary River Group. An assemblage of supracrustal rocks characterized by marble and metapelite occurs in the northwest part of the area and may be correlative with the early Proterozoic Piling Group. All of the above rock types are intruded by plutons of inferred Proterozoic age.

2) Supracrustal rocks of both greenstone belts have tight regional folds indicative of large-scale contractional strain. Gneisses bordering the belts also exhibit evidence of significant ductile strain. The marble-metapelite assemblage is tightly infolded with granulite facies gneisses northwest of Isortoq Fiord.

3) In the field area the Isortoq Fault Zone (Jackson, in press) coincides with: a transition from amphibolite to granulite facies; tight regional folds that trend northeast; and a set of moderately to steeply southeast-dipping ductile-brittle faults whose latest displacements are normal.

4) A systematic northwestward increase in metamorphic grade from lower amphibolite facies to granulite facies occurs in the northwest part of the area toward and across Isortoq Fiord. In contrast, supracrustal rocks at Eqe Bay are at lower metamorphic grade (greenschist-amphibolite facies) and grade increases in the opposite direction. Whereas metamorphism at Isortoq Fiord appears to be due to regional burial processes, that in the vicinity of Eqe Bay appears to be related to regional contact effects.

5) Banded iron-formation, mafic volcanic rocks, quartz-carbonate veins and some faults are the prime targets for economic mineralization in the map area.

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Geological Survey of Canada Project 690061-KG

Regional metallogenic significance of Cu, Mo, and U occurrences at DeVries Lake, southern Great Bear magmatic zone, Northwest Territories[1]

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Gandhi, S.S. and Prasad, N., 1993: Regional metallogenic significance of Cu, Mo, and U occurrences at DeVries Lake, southern Great Bear magmatic zone, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 29-39.

Abstract: DeVries Lake area is underlain by Aphebian metasedimentary rocks of the Snare Group, in part migmatized, and younger deformed felsic volcanic rocks of the 1865 Ma old Great Bear magmatic zone. Both assemblages have been intruded by granitoid plutons.

Chalcopyrite-pyrite concentrations occur along magnetite-rich beds in metasiltstones at the 'Kol' showing. The sulphides have been redistributed locally by deformation and cut by granitic veins. Granite-related hydrothermal veins containing molybdenite, pitchblende, tourmaline, pyrite, copper sulphides, and rare-earth elements occur in metasiltstones at the 'Nori' showings.

The felsic volcanic assemblage includes uraniferous magnetite-rich units at the UGI and FXO prospects. This volcanogenic mineralization is a regional metallogenic feature of the Great Bear magmatic zone. The uranium-magnetite association represents a stage in metallogenic evolution of the magmatic zone, that resulted in the monometallic and polymetallic magnetite-rich breccias and veins that occur in the region.

Résumé : La région du lac DeVries contient des roches métasédimentaires aphébiennes du Groupe de Snare, partiellement migmatisées, et des roches déformées plus récentes, de nature volcanique et de teinte claire, de la zone magmatique de Great Bear, datée de 1 865 Ma. Les deux assemblages sont traversés par des plutons granitoïdes.

Des concentrations de chalcopyrite et pyrite apparaissent le long de couches riches en magnétite contenues dans des métasiltstones observés à l'endroit de l'indice de «Kol». Les sulfures ont été localement redistribués par la déformation et recoupés par des filons granitiques. Des filons hydrothermaux associés aux granites et contenant de la molybdénite, de la pechblende, de la tourmaline, de la pyrite, des sulfures de cuivre et des terres rares se manifestent dans des métasiltstones observés à l'endroit des indices de «Nori».

L'assemblage volcanique felsique comprend des unités uranifères riches en magnétite à l'endroit des gîtes possibles UGI et FXO. Cette minéralisation volcanogénique est une caractéristique métallogénique régionale de la zone magmatique de Great Bear. L'association uranium-magnétite représente une étape de l'évolution métallogénique de la zone magmatique, qui a permis la formation des brèches et filons monométalliques et polymétalliques riches en magnétite rencontrés dans la région.

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INTRODUCTION

A subcircular remnant of metasedimentary rocks centred on DeVries Lake was noted in early regional mapping (Wilson and Lord, 1942). This feature attracted the attention of explorationists, and intermittent intensive exploration since then has led to the discovery of several occurrences of U, Mo, and Cu (Fig. 1). The present study was undertaken to evaluate the regional metallogenic significance of these occurrences. Results of five weeks of fieldwork in 1991 by the writers, and related laboratory studies are summarized here.

EXPLORATION HISTORY

Prospecting by Radiore Uranium Mines Limited led to the discovery in 1954 of uranium occurrences near the north shore of DeVries Lake, which was then known as Culbert Lake. The showings were staked as the 'Nori' claims. Trenching revealed high grade pitchblende veins in tightly folded biotite-magnetite-rich metasiltstones (Byrne and McMorland, 1955; McGlynn, 1971, p. 110). A regional aeromagnetic survey by the Geological Survey of Canada in 1963 recorded strong magnetic responses over metasedimentary and volcanic rocks of the area (Map 2928G, NTS sheet 86C/7). In 1967, Shield Resources and Angus Petroleum Consultants Limited conducted a helicopter-supported gamma-ray spectrometer survey that led to new occurrences in the vicinity of 'Nori' showings, which were restaked as the 'RA' claims, and on the southeast side of DeVries Lake that were staked as the 'DV' claims (Curry, 1968). Surface exploration revealed molybdenite in old trenches on the 'RA' claims, and diamond drilling was undertaken in 1968. A total of 918 m was drilled in 17 holes, but results were not encouraging enough to continue the work (Curry, 1968; Thorpe, 1972, p. 68-72). Westrim Mining Corporation Limited also explored the region in 1968 with a helicopter-borne gamma-ray spectrometer survey, and drilling on the Tyke uranium showing (Coxhall, 1968; Thorpe, 1972, p. 59, 67). Five vertical drill holes, totalling 254 m, intersected arkosic quartzite and low uranium values (< 0.01% U).

The 'DV' claims were optioned by Giant Yellowknife Mines Limited, and were explored by geological mapping, a ground magnetic survey, and trenching (Legagneur, 1969). The work revealed weak radioactivity over a large zone of magnetite-rich felsic volcanic rocks, and a few pitchblende stringers. In 1974, Uranerz Exploration and Mining Company Limited initiated a four-year program to further explore the uranium potential of the region (Paterson, 1974, 1975a, b). The 'DV' claims were restaked as the 'UGI' claims. The program included an airborne radiometric survey that led to the discovery of the FXO uranium showings (Fig. 1). The exploration culminated with testing of radioactive zones of the UGI claims by eight vertical diamond drill holes totaling 449 m in length (Male, 1978). Uranium values intersected were too low to be of economic interest.

Noranda Exploration Company examined the Nori/RA area in 1982, and discovered and trenched the 'Kol' copper showing. In 1990, Aber Resources Limited staked the northern part of the DeVries Lake area, including the Nori, Kol, and FXO occurrences. The company carried out prospecting and mapping of the area, and detailed resampling of the Kol trenches in 1991.

GEOLOGY OF THE DEVRIES LAKE AREA

The DeVries Lake area is located in the southern Great Bear magmatic zone (GBmz) near its eastern boundary which is marked in this region by the north-trending Wopmay fault zone (Fig. 1, inset). The oldest rocks in the area are metasedimentary rocks regarded as equivalents of the Snare Group in the east (Wilson and Lord, 1942; Frith et al., 1977). These are polydeformed, and have been intruded by granite-granodiorite gneiss. They are in part migmatized, as seen in the western part of DeVries Lake. A younger assemblage of feldspar porphyry and quartz feldspar porphyry on the east side of the lake was noted by Wilson and Lord (1942). The present study revealed volcanic features in parts of this assemblage. It is similar to other assemblages of felsic extrusives and subvolcanic porphyries in the southern Great Bear magmatic zone (Gandhi, 1989, 1992b). Its relationship with the metasedimentary rocks in the study area is obscured by folds, faults and younger intrusions. An unconformable relationship between two lithologically similar assemblages is seen, however, in the Lou Lake area 80 km south of DeVries Lake (Gandhi and Lentz, 1990). The present study also revealed a nearly horizontal, north-northwest-trending lineation in the felsic assemblage of the DeVries Lake area, which is not observed in the felsic volcanic aasemblages elsewhere in southern Great Bear magmatic zone. This trend is parallel to lineation/foliation in the associated rocks, and is attributed to regional deformation that affected a broad zone along the Wopmay fault zone.

Metasedimentary rocks (Snare Group)

The metasedimetary rocks occur in a broad synclinorium plunging gently to the northwest (Fig. 1). They are divided into three lithological subunits: (i) quartzite-arkosic quartzite, (ii) interbedded quartzite and metasiltstones, and (iii) metasiltstone-argillite.

The first subunit occurs mainly along the east side of DeVries Lake. It is mostly medium- to coarse-grained, sugary white quartzite that ranges from massive to lineated rock containing well developed streaks of coarse quartz 1 to 10 cm long. Their contact with intrusive gneissic granodiorite-quartz monzonite is sharp and is folded in some places (Fig. 2). The arkosic variety is difficult to distinguish in the field from some of the rocks of the felsic volcanic-intrusive assemblage that resemble it closely in colour and texture. Hence, in many places the distinction is tentative or arbitrary.

The second metasedimentary subunit is thick to thin bedded, fine- to medium-grained quartzite-arkosic quartzite intercalated with zones and lenses of thin bedded metasiltstones. Rocks of this subunit underlie central part of the study area. Bedding is well preserved and commonly dips gently or moderately to the northwest or east. Siltstone beds

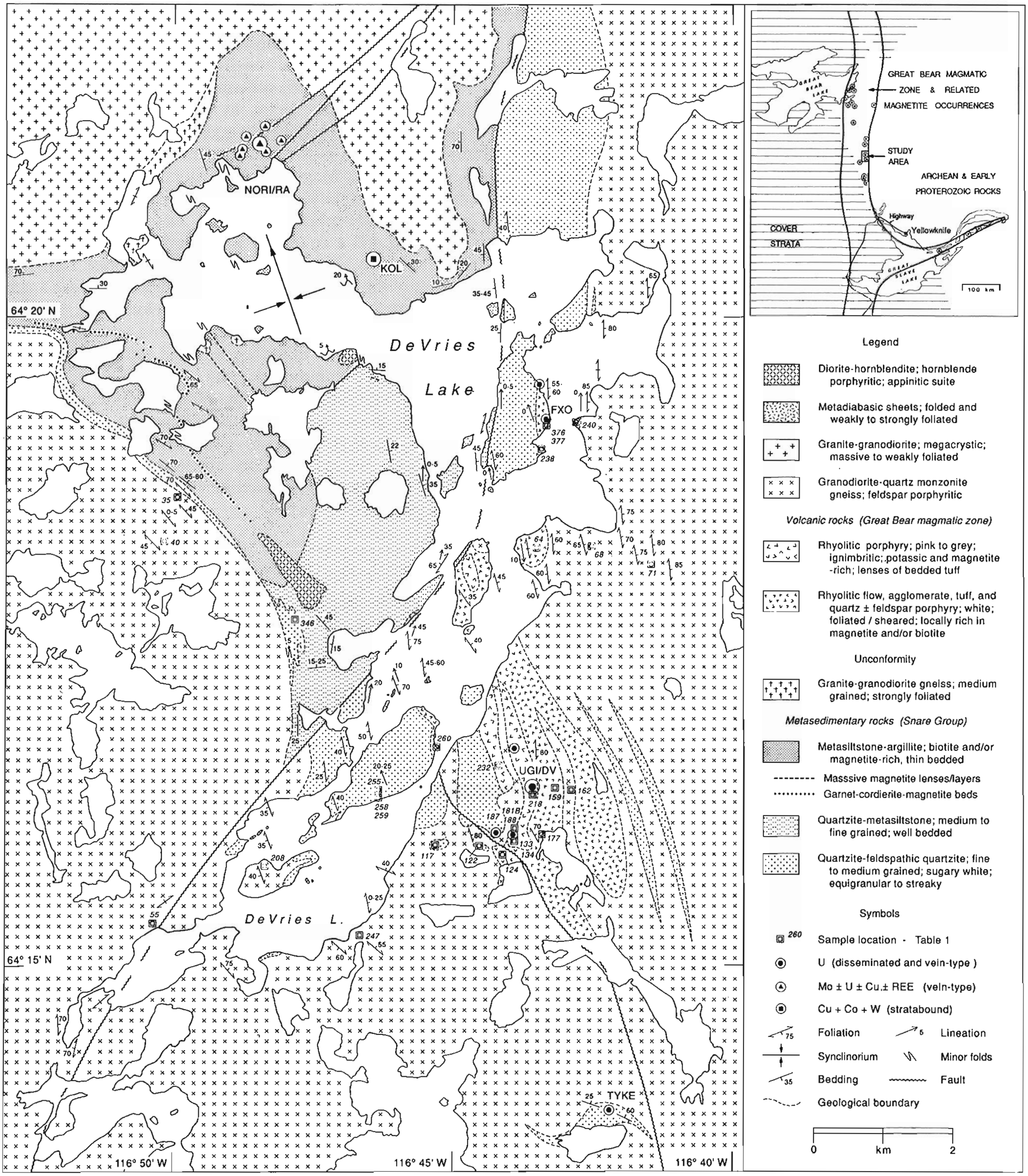


Figure 1. Geology and mineral occurrences of the DeVries Lake area, Northwest Territories.

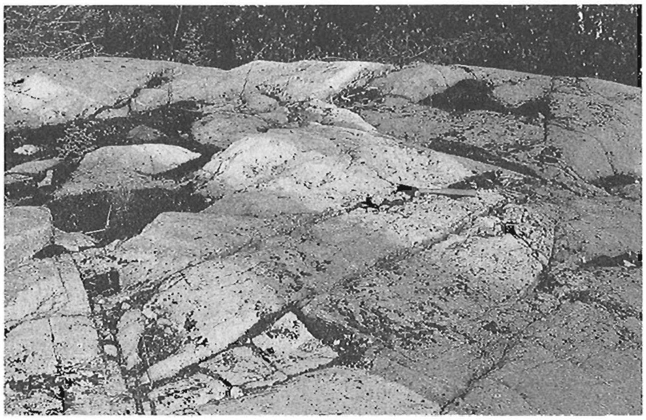


Figure 2. Folded intrusive contact between quartzite (white) and gneissic granodiorite-granite (grey); 4.5 km south-southwest of Kol prospect (Fig. 1). (GSC 1992-116G)

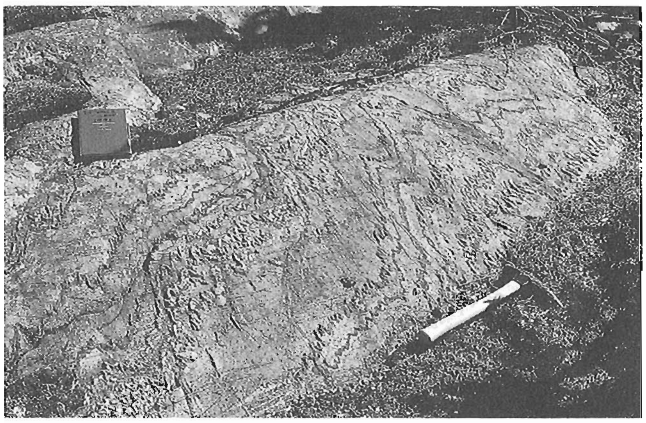


Figure 3. Folded metasiltstone of the Snare Group, 2 km southwest of Kol prospect, near shore (Fig. 1). Looking northwest. (GSC 1992-116B)

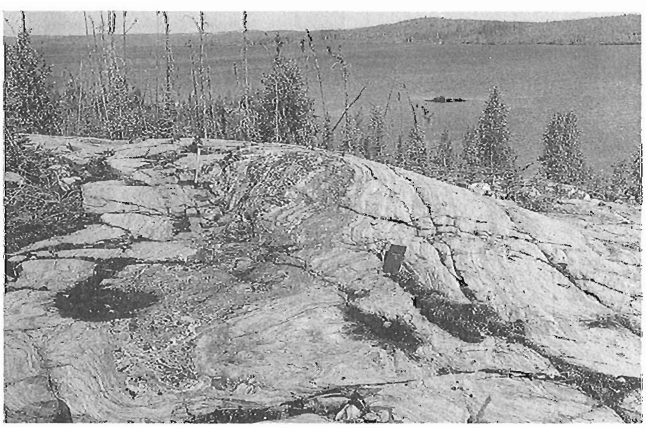


Figure 4. A magnetite-rich boudin-like lens in metasiltstone, 1.5 km south of Kol prospect (Fig. 1). Looking north-northwest. (GSC 1992-116O)

and lenses are more common towards the northwest. Along the eastern fault-boundary with the quartzite-arkosic quartzite subunit, these rocks are cut by a steeply dipping, north-trending set of hornblende veins.

The third subunit is comprised of thinly bedded siltstones and interbedded biotitic argillaceous beds of variable thickness. The beds of this subunit are characterized by minor tight folds on the centimetre- to metre-scale (Fig. 3). They include some magnetite-rich quartzo-feldspathic siltstone beds, cordierite-garnet-magnetite beds, and massive magnetite lenses. Magnetite in siltstones and garnetiferous beds occurs as finely disseminated grains and aggregates. Garnet-cordierite-bearing beds and lenses are as much as 10 m thick, and a couple of them persist for more than a kilometre (Fig. 1). These reflect the overall iron-rich character of the subunit. The massive magnetite lenses are as much as a metre thick and several metres long (Fig. 4). They are cut by quartz-albite±tourmaline veins and aggregates. Rocks of this subunit in the western and southwestern parts of the area are commonly crenulated biotite-rich meta-siltstones that contain abundant granitic material as thin concordant lenses and veins, which impart a migmatitic character. Such migmatitic paragneiss is only locally developed elsewhere, and does not form a continuous ring around the subcircular metasedimentary area as indicated by Wilson and Lord (1942).

Granite-granodiorite gneiss

Moderately to strongly foliated or streaky, medium grained granite-granodiorite gneiss occurs in small area on the northwest shore of DeVries Lake. It is regarded here as a probable equivalent of the Hepburn intrusive suite that intruded the Snare Group east of the Wopmay fault zone (Frith et al., 1977; Gandhi, 1992a).

Felsic volcanic assemblage

Rhyolitic tuffs, flows, and ignimbrites, and quartz-feldspar porphyritic rocks form a north-trending zone on the east side of DeVries Lake. They also occur as lenses and patches along the south and southwest margins of the metasedimentary rocks. They are characteristically white on weathered and fresh surfaces, except for a distinctive pink to grey potassic quartz-feldspar porphyry unit (Fig. 1).

The white felsic rocks include several textural and compositional varieties, reflecting different flows, tuffs, fragmental units, and intrusive porphyries. Their boundaries are, however, poorly defined and the stratigraphic sequence is uncertain. Most of them are lineated and/or foliated, and have been partially or wholly recrystallized. Their boundaries with the white weathering feldspathic quartzites and quartzite are obscured due to deformation. The picture is further complicated by bands and lenses of intrusive granodiorite-quartz monzonite.

The white felsic rocks commonly contain feldspar phenocrysts in a fine grained granoblastic matrix, as well as lenticular quartz eyes or streaks of quartz aggregates that are a few millimetres to a few centimetres long. An agglomeratic character was seen at several localities (Fig. 5). The proportions of mafic minerals, mainly biotite and magnetite, vary greatly. Some varieties are almost completely devoid of mafic minerals, and closely resemble fine grained quartzite in the field, but their felsic volcanic nature has been confirmed by thin section study and chemical analyses (Table 1). Other varieties contain varying proportions of biotite and/or magnetite, which occur as disseminated grains or as aggregates, lenticles, or folia.

The pink to grey quartz-feldspar porphyry is well exposed on the ridge top at the UGI showing. It is characterized by coarse feldspar phenocrysts, abundant disseminated grains of magnetite in the matrix, and high contents of potassium and uranium (Tables 1 and 2). It is in part ignimbritic (Fig. 6). It contains thinly layered tuffaceous magnetite-rich beds near its south end. Its contact with the white felsic volcanic rocks is sharp and in some places sheared, but the unit as a whole appears to be relatively less deformed and sheared than these rocks.

Younger granitic plutons

Granodiorite-quartz monzonite and granite-granodiorite plutons surround the metasedimentary and volcanic rocks and are regionally extensive (Wilson and Lord, 1942). The two phases distinguished are geographically separated; the former underlies a large area on the east, south, and southwest sides of DeVries Lake, and the latter is situated to the north. The granodiorite-quartz monzonite is characterized by well developed, steeply dipping foliation due to biotite-rich folia, and feldspar crystals commonly 0.5 to 2 cm long. In some places it has an augen-like texture. Its quartz content varies from 10 to 25%. It occurs as north-trending bands and lenses with the quartzite-feldspathic quartzite and felsic volcanic assemblages. Attitudes of foliation, xenoliths, and contacts suggest intrusion of the granodioritic material as domes and concordant sheets. The coarse porphyritic granite-granodiorite pluton to the north also forms a large, subconcordant, thick sheet-like body above the gently dipping siltstones of the Snare Group. It is characterized by K-feldspar crystals 2 to 5 cm long, and ranges from massive to weakly foliated.

Metadiabasic sheets

A number of thin sheets of metadiabasic rock occur in the region of felsic volcanic rocks (see Fig. 1 and 11). The sheets are amphibolitic in texture, and have sharp boundaries with the granitic and felsic volcanic rocks. Chilled margin is recognizable in some of them. The attitude of the sheets is variable due to folding, but a gentle dip to the north is common. Lineation is well developed in some places. It is north-trending, and horizontal or plunging gently to the north.

Diorite-hornblendite

Texturally distinctive, hornblende porphyritic mafic diorite intrusions form hills in the central part of the area (Fig. 1), and also occur as dykes and irregular bodies elsewhere in the area. This mafic diorite is characterized by euhedral phenocrysts of hornblende approximately 1 cm long, in fine- to medium-grained dioritic matrix. Traces of pyrite and chalcopyrite occur locally. A very coarse hornblendite phase is abundant in large black hills in the centre of the area. Some layering is also developed in these hills. The intrusions are appinitic in character, and are undeformed and unaltered. They have sharp intrusive contacts with the metasedimentary rocks, and contain a number of xenoliths of these rocks.

PETROCHEMISTRY

Chemical analyses of thirty felsic volcanic, feldspathic quartzite, and intrusive rocks are given in Table 1. The volcanic rocks are rhyolitic, and in the case of magnetite-rich pink varieties the silica contents are low because of their very high iron contents. On the other hand, some of the white

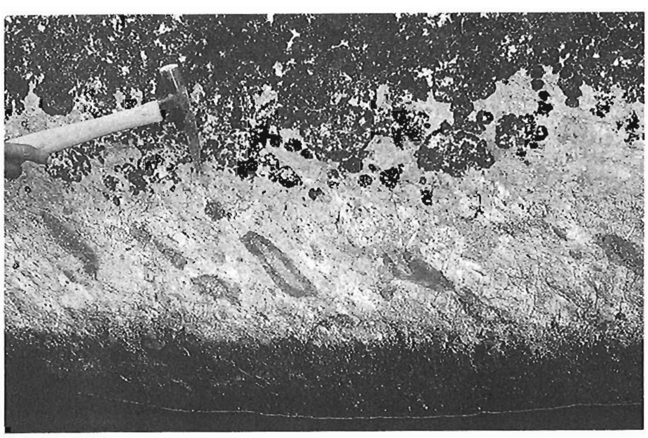


Figure 5. Photograph of rhyolitic agglomerate on shore at the south end of an island between the UGI and FXO showings (Fig. 1). Note porphyritic character of the fragments. (GSC 1992-116 J)

Figure 6. Ignimbrite texture in magnetite-rich pink feldspar porphyry, between trench 3 and 4, UGI prospect (Fig. 9). Looking north. (GSC 1992-116N)

Table 1. Chemical analyses of selected rocks from DeVries Lake area, Northwest Territories

Sample #	177	159	188	124	260	181B	376	238	240	64
Rock unit	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol	Rhy Vol
SiO_2 %	69.90	70.10	70.80	71.30	71.40	72.00	70.00	71.50	69.40	72.70
TiO_2	0.36	0.46	0.26	0.38	0.55	0.28	0.24	0.26	0.35	0.43
Al_2O_3	14.00	16.80	13.10	14.20	12.90	13.70	13.20	14.00	14.60	14.90
Fe_2O_3	0.00	0.00	1.30	0.00	0.50	0.40	0.90	0.50	0.20	0.00
FeO	1.00	0.60	1.50	0.70	1.30	0.80	1.70	1.20	1.10	0.40
MnO	0.00	0.00	0.01	0.00	0.01	0.00	0.01	0.02	0.00	0.01
MgO	2.10	0.47	0.80	0.53	4.78	2.19	1.33	1.45	1.17	0.20
CaO	3.97	4.60	1.62	4.23	2.41	3.51	4.04	2.62	2.65	2.41
Na_2O	5.00	5.80	2.40	5.50	3.10	4.60	5.00	4.60	4.30	3.80
K_2O	1.77	0.53	6.42	0.95	2.50	1.30	1.26	1.54	4.36	4.68
H_2Ot	0.50	0.20	0.50	0.30	0.70	0.60	0.40	0.80	0.40	0.20
CO_2t	0.10	0.10	0.50	0.20	0.10	0.20	0.00	0.10	0.10	0.20
P_2O_5	0.30	0.53	0.14	0.26	0.13	0.35	0.42	0.17	0.12	0.02
S	0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	0.02	< 0.02	< 0.02	< 0.02
Total	99.02	100.19	99.35	98.55	100.38	99.93	98.52	98.76	98.75	99.95
Ba ppm	35	66	1364	10	129	44	20	96	713	742
Be	2.7	3.5	1.4	5	2.9	5.1	4.5	9.3	3.9	5
Co	< 5	< 5	< 5	< 5	7	< 5	< 5	6	5	< 5
Cr	14	< 10	< 10	< 10	27	< 10	11	15	18	12
Cu	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10
La	< 10	26	< 10	< 10	< 10	14	< 10	< 10	11	11
Nb	30	35	44	21	12	17	33	15	23	21
Ni	< 10	< 10	10	< 10	< 10	11	27	11	< 10	< 10
Rb	110	22	210	43	280	100	95	90	180	130
Sc	12	3.7	2.2	3.6	29	12	5.9	11	7.8	2.7
Sr	130	230	110	130	140	110	140	150	120	140
V	15	12	37	< 5	20	35	8	< 5	11	< 5
Y	41	92	64	69	7	27	64	26	44	73
Yb	3.7	7.8	5.9	5.7	0.7	2.3	6.1	2.7	4.1	7.1
Zn	< 5	< 5	< 5	< 5	5	< 5	< 5	10	< 5	< 5
Zr	210	240	170	250	250	180	140	160	180	260

Sample #	133	134	218	232	187	377	346	259	258	255	122	117
Rock unit	Rhy Por	Rhy Por	Rhy Por	Rhy Por	Rhy Por	Rhy Por	Fel Qtz	Fel Qtz	Fel Qtz	Fel Qtz	Fel Qtz	Fel Qtz
SiO_2 %	51.00	52.70	60.60	63.80	69.00	68.10	73.40	75.40	75.60	78.70	74.90	75.20
TiO_2	0.27	0.33	0.38	0.22	0.23	0.23	0.42	0.44	0.34	0.58	0.35	0.30
Al_2O_3	9.50	10.70	11.00	11.40	13.10	14.20	14.60	15.10	14.20	12.70	15.00	14.30
Fe_2O_3	20.60	17.00	10.80	9.00	2.30	2.50	0.00	0.00	0.00	0.00	0.00	0.00
FeO	9.40	8.00	6.00	4.60	1.60	1.80	0.60	0.30	0.40	0.18	0.40	0.30
MnO	0.03	0.02	0.02	0.01	0.01	0.00	0.01	0.01	0.01	0.01	0.00	0.00
MgO	2.09	1.75	2.50	1.11	0.47	1.66	0.47	0.14	0.24	0.15	0.13	0.12
CaO	1.87	1.28	2.68	0.57	0.63	3.12	4.26	3.37	2.73	2.22	2.70	2.55
Na_2O	2.10	1.40	2.40	1.10	1.30	5.30	4.20	4.80	5.10	4.60	5.30	5.40
K_2O	2.36	6.26	3.21	7.98	8.95	1.57	0.60	0.56	0.71	0.37	0.75	0.48
H_2Ot	1.20	0.50	0.70	0.50	0.40	0.50	0.40	0.20	0.30	0.20	0.30	0.20
CO_2t	0.10	0.20	0.00	0.20	0.10	0.10	0.10	0.20	0.10	0.20	0.20	0.10
P_2O_5	0.14	0.21	0.14	0.20	0.13	0.13	0.17	0.04	0.03	0.02	0.02	0.02
S	< 0.02	< 0.02	< 0.02	< 0.02	0.02	0.02	0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02
Total	100.66	100.35	100.43	100.69	98.24	99.23	99.25	100.56	99.76	99.93	100.05	98.97
Ba ppm	88	888	455	1415	1756	31	76	15	47	61	84	81
Be	3.4	1.4	2.1	1.2	0.8	7.6	3.4	6	6.7	6.1	5.3	4.5
Co	9	10	8	6	< 5	7	< 5	< 5	< 5	< 5	< 5	< 5
Cr	< 10	< 10	< 10	< 10	11	19	12	10	< 10	15	23	< 10
Cu	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10
La	< 10	< 10	110	35	13	< 10	20	< 10	< 10	< 10	< 10	10
Nb	24	21	23	< 10	21	34	11	36	33	31	15	30
Ni	52	57	33	15	11	47	< 10	< 10	< 10	< 10	< 10	< 10
Rb	48	200	130	250	290	120	23	12	15	12	15	18
Sc	7.4	6.7	10	8.5	4	4.6	13	1.2	0.7	1.8	3.5	1
Sr	51	65	87	60	98	130	210	200	150	170	210	190
V	28	36	33	5	< 5	7	6	< 5	< 5	< 5	< 5	< 5
Y	31	27	39	13	56	62	63	8	< 5	11	35	7
Yb	2.7	2.6	8.7	1.2	5.1	7.6	6.3	1.1	0.5	1.4	3.7	1
Zn	23	17	8	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5
Zr	130	140	210	140	160	130	290	240	230	150	160	180

Sample #	247	71	68	162	208	35	40	55
Rock unit	Grd Gn	Grd Gn	Grd Gn	Grd Gn	Grd Gn	Grd Gn	Grd Gn	Gr dyke
SiO_2 %	68.40	69.30	71.20	72.00	73.10	73.30	76.80	74.30
TiO_2	0.67	0.48	0.32	0.26	0.39	0.31	0.35	0.17
Al_2O_3	13.90	14.70	13.90	14.00	14.20	14.30	13.20	13.70
Fe_2O_3	1.00	0.90	0.40	1.60	0.30	1.20	< 0.2	0.40
FeO	4.30	2.90	2.00	1.00	0.70	1.10	0.30	0.80
MnO	0.05	0.05	0.04	0.01	0.01	0.01	0.01	0.01
MgO	1.06	0.78	0.50	0.63	0.76	0.44	0.43	0.60
CaO	1.95	1.90	1.18	0.82	1.24	2.90	3.63	0.55
Na_2O	2.50	2.40	2.60	2.20	2.60	4.70	4.00	3.00
K_2O	4.75	5.37	6.06	7.20	6.05	0.82	0.48	5.47
H_2Ot	0.90	0.60	0.50	0.60	0.80	0.30	0.10	0.70
CO_2t	0.20	0.40	0.10	0.10	0.00	0.10	0.20	0.10
P_2O_5	0.23	0.17	0.15	0.14	0.18	0.15	0.10	0.05
S	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02
Total	99.91	99.95	98.95	100.56	100.33	99.63	99.60	99.85
Ba ppm	741	978	759	1219	943	80	< 30	1500
Be	3.1	2.8	2.8	1.8	2.3	5.1	3.4	2.2
Co	12	6	7	5	< 5	< 5	< 5	< 5
Cr	16	11	10	< 10	< 10	<10	11	< 10
Cu	13	11	68	< 10	< 10	10	12	< 10
La	150	64	55	19	25	15	10	53
Nb	20	12	24	17	22	31	24	< 10
Ni	13	< 10	< 10	10	< 10	< 10	< 10	< 10
Rb	240	240	300	240	180	54	22	200
Sc	14	8.3	5.9	5.7	8.3	4.8	8.3	3
Sr	100	110	74	100	150	170	180	120
V	25	15	7	15	10	10	11	< 5
Y	80	47	52	41	57	64	56	14
Yb	7.5	4.6	5	4.2	5.6	6.6	5.7	1.2
Zn	26	41	30	6	< 5	< 5.0	< 5	< 5
Zr	470	300	220	160	310	230	200	110

Notes :

1) Analyses by the Analytical Chemistry Section, Mineral Resources Division, Geological Survey of Canada, Ottawa; All analyses by x-ray fluorescence method except FeO, H2OT, CO2T and S by rapid chemical methods; Fe2O3 is calculated using formula : Fe2O3 + Fe2O3 (xrf) - 1.11134 X FeO (volumetric).

2) Location of samples in Figure 1; Sample numbers are of GFA-'91-1 to 400 series; Following samples are from UGI/DV trenches shown in Figure 11, # 181B, 187 and 188 from Tr-A1, # 133 and 134 from Tr-A2, and # 218 from Tr-3; Samples 255, 258 and 259 from a steep, 50 m high escarpment facing southeast, and represent its middle, upper and basal parts respectively; Samples 376 and 377 from the vicinity of main FXO uranium showing near shore.

3) Rock unit abbreviations: 'Rhy Vol' - Rhyolitic volcanic rocks, mostly flows except for samples 238 and 260 from agglomearte-tuff horizons, and possible intrusive porphyry sample 346; 'Rhy Por' - Rhyolitic porphyry, pink to grey, ignimbritic, potassic and magnetite-rich; 'Fel Qtz' - Feldspathic quartzite; 'Grd Gn' - Granodioritic gneiss; 'Gr dyke' - Granite dyke, medium grained, cutting the granodioritic gneiss.

Table 2. Spectrometer readings at selected mineralized localities in DeVries Lake area, Northwest Territories

Location (metres)	eK %	eU ppm	eTh ppm	Mag* SI
NORI / RA Prospect				
Zone A, TR.-2 (From W to E)				
2 S	9.4	11	21.4	50
2 E	16.6	54	32.9	87
3.5	13.8	211	36.4	350
4.2 SSW	14.4	458	64.5	439
4.2 SSE	22.3	783	140	400
7 E	13.9	45.4	29.5	125
8.8 E	13.9	115	34.8	127
9.8 E	13.1	28.2	32.9	60
Tr-2A				
6 SE	11.8	244	33.9	159
7 SE	10.6	34	21.4	150
8 SE	O.S.	O.S.	2854	484
Tr-1 (From W to E)				
0.5 E	10.7	10	25.1	60
3.5 E	13.0	9.9	28.9	64
5.5 E	12.3	19.5	30.6	85
Tr-3 (From W to E)				
8.5 E	13.0	37	23	69
6.5 E	15.5	86	28.6	28
3 E	13.3	74	25.1	215
3.7 E	21.5	861	33.5	363
2 E	14.5	50	28.3	65
0.5 E	15.7	22.7	29.5	83
Tr-2B (From W to E)				
0	11.3	151	25.3	146
1.5 E	O.S.	O.S.	O.S.	520
2 NE	13.9	242	37.2	746
2.5 E	23.1	708	105	O.S.
Tr-4 (From W to E)				
8.5 E	13.3	8.6	33	61
6.5 E	14.6	66	27.9	122
4.5 E	15.8	31.1	29.7	38
3 E	15.3	32.8	31.7	150
0.8 E	13.6	10.4	28.7	85
TR.-5				
W wall	16.3	114	33	576
Tr-67 m NW of TR.-105(E to W)				
1 W	16.1	37.4	26.6	127
1.8 W	65.9	570	395	0 52
2.9 W	11.9	50	23.4	237
4 W	12.1	38.2	27.7	150
4.8 W	16.9	461	61.3	0 13
TR.-128 m NW of TR.-105 (E to W)				
1 W	13.2	161	39.5	223
1.8 W	19.4	529	93	258
3.2 W	13.0	77	37.9	137
5 W	13.6	55	26.9	275
Tr-105 (From E To W)				
2.7 W	16.2	321	35.7	33
2.7 WSW	O.S.	911	815	31
1.5 W	O.S.	937	841	60
4 SW	11.2	33.5	14.5	62
Tr-104				
E wall	9.8	182	27.2	44
Tr-103 (From NE to SW)				
2 SW	12.5	143	25.4	39
3.5 SW	12.2	118	26.7	107
7 SW	14.3	45.8	26.9	160
11.5 SW	10.1	12.3	18.5	51
Tr-102 (From E to W)				
0.1 W	10.3	69	24	101
2.5 W	O.S.	644	203	0 8
3.3 W	10.7	93	21.3	127
5 W	13.6	47.9	25.6	99
6.6 w	25.9	46.9	125.8	55
8 W	8.4	20	18.3	29
Tr-101; N wall (From W to E)				
1	15.2	21.1	31.7	137
3.8	O.S.	795	576	3.3
5	11.8	233	25.2	95
7	10.1	54	21.9	277
7.7	O.S.	1080	1064	2
8.1	16.8	563	64.1	0 9
8.8	O.S.	1708	1866	12
9.5	13.6	300	33.4	0 10
Tr- 42 m E of Tr-102 (From W to E)				
0	12.1	108	27.9	43
2.5 N	9.8	24.7	24.3	72
2 E	12.6	160	26.7	178
1.5 E	O.S.	1009	1107	27
Tr-106				
N-wall	18.4	591	73	4

Location (metres)	eK %	eU ppm	eTh ppm	Mag* SI
Moly Trench				
0 (N-wall)	11.5	69	20.7	55
0.7 SE	34.0	643	188	9
TR.-55 m ESE of TR. 102 (From W to E)				
3 E	10.2	252	18.5	129
4 E	14.6	143	28.1	0 9
1.8 E	13.9	358	31.8	46
UGI / DV Prospect				
Tr-2 N to S (from N-end)				
3	6.4	11.9	17.4	297
1	0.7	8.4	6.4	507
1.5	1	11.7	8.8	273
2	0.9	13.7	9.4	422
3	1	12.7	7.1	471
3.5	1.7	63.6	20.5	343
Tr-A1 NW to SE (from NW-end)				
1.5	8.6	15.2	28.7	0.5
1.5	10.3	31.6	40	48
3	O.S.	640	565	0.2
4.5	12	59	46	43
6	6.9	223	91	0.4
8	7.8	28	27.8	13
9	10.1	452	102	0.8
TR-A2				
At N-end	13.9	548	123	0.2
Tr-3 (From W to E)				
22 SE	0.3	7.4	21	290
17 SSE	3.1	183	28.5	389
0.5 E	4	178	68	191
2 E	2.3	127	40.3	290
4.7 E	2.2	129	40	193
7 E	2.8	144	33.5	218
2.0 N	1.5	99	23.1	290
6 ENE	1.3	48	28.7	209
52 NE	1.3	66	19.2	232
Hot Spots (near Tr-3)				
0	3.5	271	41.8	100
50.5 S	3	51.8	14.8	250
47 S	2.4	68.4	23.6	80
50 N	4.3	246	61	175
Tr-4 (W to E)				
1.5	2.5	69	26.3	276
3	2.7	26.3	28.1	378
4.5	8.1	46	24.9	115
8 ENE	3.7	65	16.9	137
16 ENE	1.9	83	16	278
23 NE	2.8	80	25.7	111
40 S (pit)	11.7	566	91	360
40 S 1 W	3.4	77	18.6	398
40 S 2 W	1.6	49.7	14.3	415
Near ddh DV-4				
3 SE	1.3	83	18.9	264
7 SE	1.5	58	15.2	216
7.5 E	1.8	118	17	242
Near ddh DV-12				
1 W	0.6	3.4	14.5	289
2 S	0.7	2.6	17.7	293
4 SW	0.5	4.4	19.2	207
5 SE	0.6	6.5	27	0.2
Section W of Tr-4 (From W-end of Tr)				
58	7.8	6.4	22	27
99	6.4	5.5	23.9	6
137	6.4	5.3	25.4	15
183	7.8	1.6	21	20
183E 23 SW	8	1.1	10.9	300
FXO Main Showing				
From East shore				
0	6.1	95	11.9	309
2 W 2 S	6.6	152	8.4	243
2 W 5 N	5.5	62	13.4	242
9 W 2 S	4.8	32	17.1	135
3.5 W 2S	21.9	925	153	234
Showing (appx. 50 W from shore)				
50 W 15 N	3.8	225	16.7	14
43 W 15 N	3.7	244	17.9	17
42 W 16 N	6.0	329	24.3	25
42 W 19 N	3.7	204	16.4	275
Section W-E				
38 W 15 N	4.1	13	16.5	236
58 W 15 N	1.3	14.2	25.3	175
98 W 15 N	0.6	4.7	20.3	< 1
108 W 15 N	0.4	2.3	22.6	0.1

Notes: Tr - Trench; Mag* - magnetic susceptibility readings in 10E-3 SI units using Exploranium KT-5 magnetic susceptibility meter; O.S. - Off Scale; eK, eU, eTh - radiometric equivalent K, U, Th contents obtained from Exploranium GRS-256 gamma-ray spectrometer readings.

varieties are poor in iron, have a very high soda/potash ratio, and are chemically indistinguishable from the rocks regarded here as feldspathic quartzite. These rocks are deformed, recrystallized, and have undergone alkali metasomatism. Iron may also have been mobile to some extent during these events.

MINERAL OCCURRENCES

Three main types of mineral occurrences found in the DeVries Lake area are described below: (i) stratabound chalcopyrite-pyrite concentrations, (ii) hydrothermal Mo-U veins, and (iii) felsic volcanic-associated U occurrences. They are too small and/or low grade to be economically significant, but they represent styles of mineralization that are important for regional metallogeny of the Great Bear magmatic zone. In addition, magnetite-rich lenses and boudins are present in metasedimentary rocks, and they frequently contain uranium concentrations along fractures and in finely disseminated grains. Quartz-pyrite± chalcopyrite, quartz-tourmaline, and quartz veins are also common in the area.

Stratabound chalcopyrite-pyrite occurrence

Chalcopyrite and pyrite are concentrated along beds in metasiltstones at the 'Kol' showing (Fig. 7). The sulphide layers are as much as 0.5 cm thick, and are coarse grained. They alternate with micaceous siltstone beds as much as 1 cm thick. Most of the sulphides are along conformable layers, but some discordant veins are also present. The beds, including the sulphide-rich layers, occur in a small synclinorium, which has associated asymmetric west-verging minor folds and plunges gently to the north. Magnetite is abundant in the sulphide-rich zone, and is coarser grained than magnetite in the siltstone layers. Sparsely distributed granitic veinlets, some of them ptygmatitic, postdate the sulphides. Suphide-rich zones exposed in trenches (Fig. 7) were sampled by Covello, Bryan and Associates for Aber Resources Limited in 1991. Assays for chip and grab samples, and drill core samples obtained using a GSC portable drill machine, ranged from 0.5 to 4.5% Cu over widths of 0.5 to 2 m (G. Thomas, Aber Resources Limited, pers. comm., 1992). The section exposed in Trench E west averaged 2.17% Cu over 8.5 m. The samples contained 70 to 650 ppm Co, and 310 to 3200 ppm W. A spectrometer survey by the writers showed negligible contents of uranium and thorium (Table 2).

Hydrothermal Mo-U veins

Veins containing molybdenite, pitchblende, tourmaline, and traces of sulphides are seen in a number of trenches on the 'Nori/RA' claims (Fig. 8). The veins pinch and swell, and range in width from a few centimetres to 0.5 m. The larger ones are as much as 30 m long. They have variable strike, and are generally steeply dipping. They are discordant to the bedding in the folded biotite-rich metasiltstone host. Aggregates of coarse biotite are developed along margins of

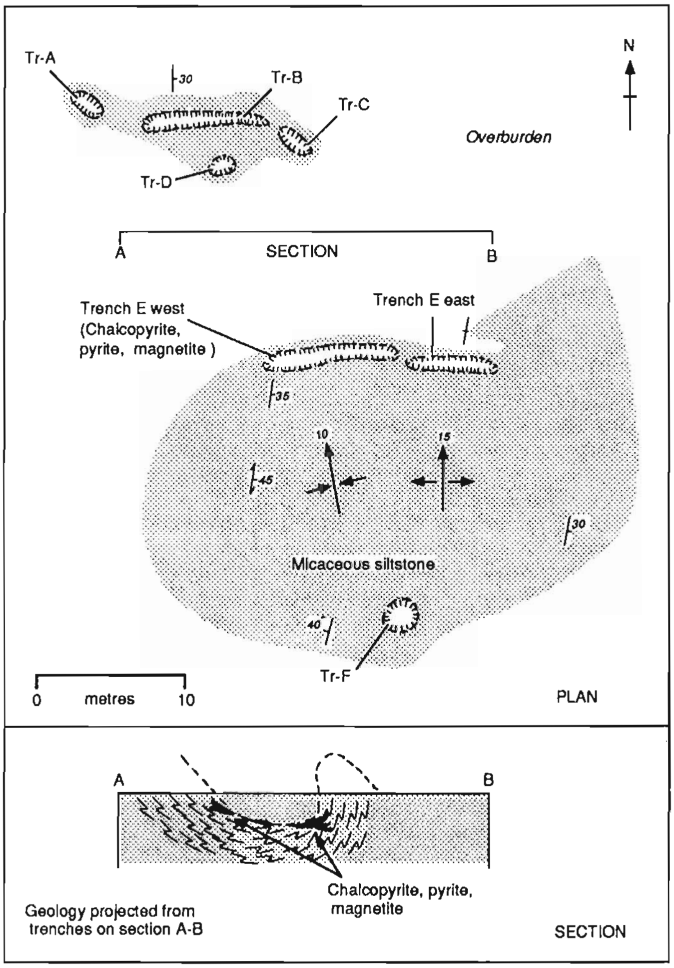


Figure 7. Geology and trenches of the Kol prospect, DeVries Lake.

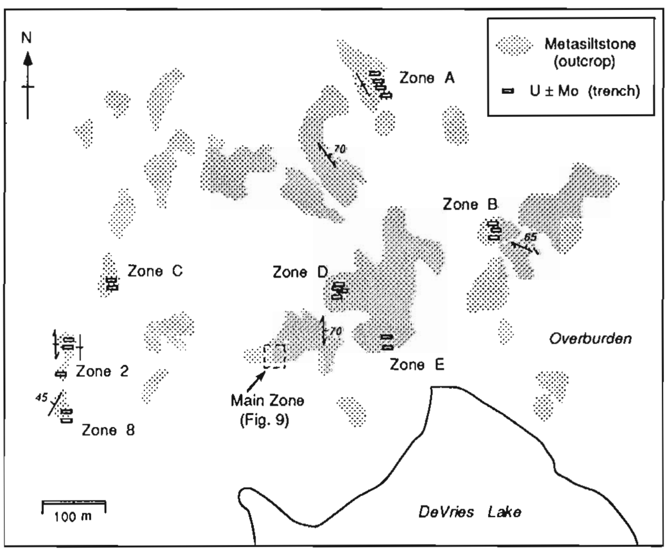


Figure 8. Mo-U showings of the Nori/RA claims, DeVries Lake.

the veins. The presence of tourmaline, apatite, and Ce-La allanite were reported by Miller (1982a, b). Variable amounts of magnetite occur in the veins. Molybdenite and pitchblende are irregularly distributed in them, and are commonly concentrated along coarse biotite-rich material. These minerals are not readily visible on the surface. Miller (1982a) reported that uraninite occurs as zoned euhedral crystals in some veins and as spherical grains in others. Some of the euhedral uraninite crystals occur as inclusions in biotite, and

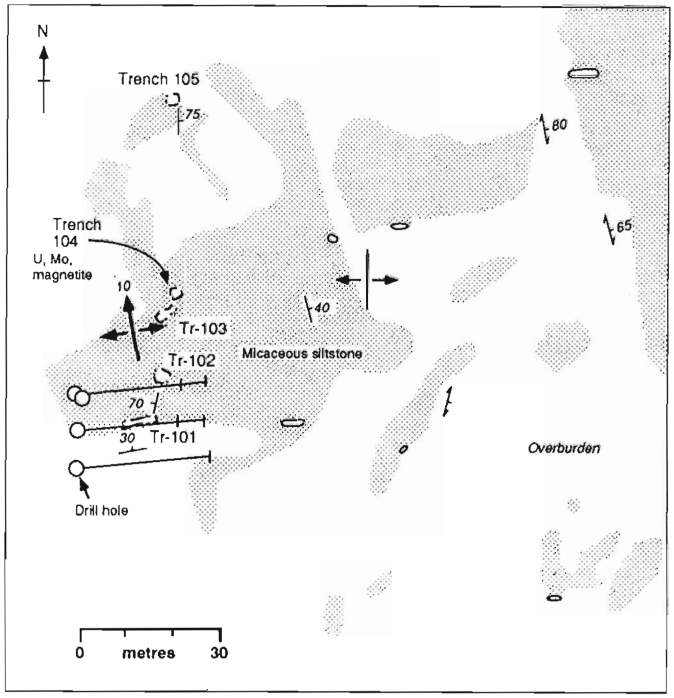
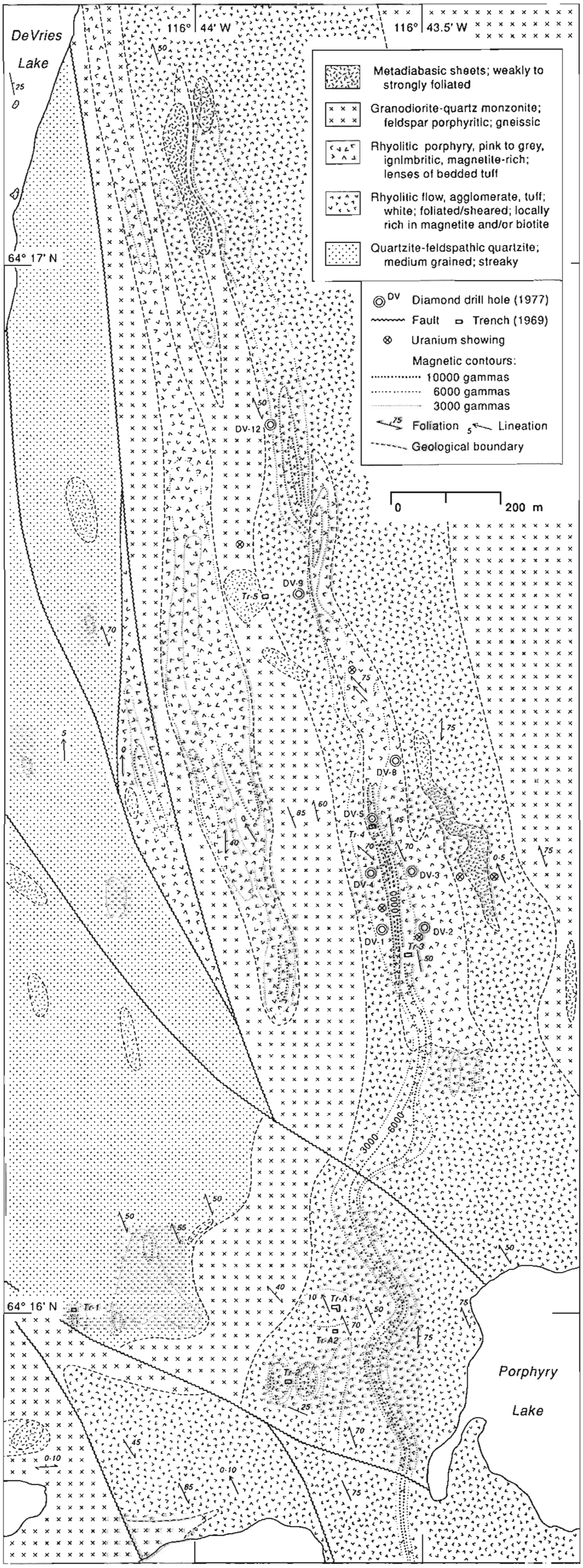


Figure 9. Plan of trenches and diamond-drill holes of the Main zone, Nori/RA claims, DeVries Lake.

Figure 10. Mo-U bearing biotite-tourmaline veins, trench 101, Main zone, Nori/RA prospect (Fig. 8 and 9). Looking north. Note molybdenite-rich zone dips west from pencil to notebook; scintillometer at uranium-rich spot. (GSC 1992-116A)



some others have overgrowths of younger uraninite. Isotopic analyses of uraninites and pyrite reported by Miller (1982a, b) show dates of 1853 ± 19 Ma and 1836 ± 32 Ma, respectively.

Assays of samples from the trenches as well as from drill core are summarized by Thorpe (1972, p. 68-72). The best mineralization is seen in the north-trending Main zone explored by five trenches and five drill holes (Fig. 9). The best assay values reported are 2.4% MoS_2 and 0.42% U over a width of 1 m in trench 101. The mineralized zone is 3 m wide at this trench (Fig. 10), and narrows to the north. The northernmost trench , however, has 0.07% MoS_2 and 0.12% U over a width of 1 m. A sample from a massive pitchblende vein from this trench contains abundant rare-earth elements, Nb and Sc (percentage of Ce : 3.5, Dy : 0.23, Er : 0.12, Eu : 0.026, Gd : 0.19, Ho : 0.051, La : 2.0, Nd : 1.1, Sm : 0.2, Tm : 0.027, Y : 1.21, Yb : 0.18, Nb : 0.13 and Sc : 0.035). Thorium contents are also high in many of the uranium-rich localities, and the Th/U ratio approaches 1 in some cases (Table 2).

Some uraniferous veins contain little or no molybdenite and chalcopyrite, as in the case of Zone A. On the other hand, zones 2, 8, 10 (RA fault zone), and D are pyrite-rich, contain little or no biotite and magnetite, and have only traces of uranium. Pyrite forms coarse aggregates with quartz, and comprises as much as 35% of the vein material in these zones. Traces of chalcopyrite, molybdenite, and silver are present in the sulphide-rich parts.

Felsic volcanic-associated U occurrences

Anomalously high concentrations of uranium, as much as 100 ppm, and a few pitchblende stringers and aggregates occur in the felsic volcanic assemblage at the UGI and FXO showings (Fig. 1 and 11). High radioactivity coincides with strongly magnetic zones in the assemblage. Most of these zones are in the pink to grey, potash-rich porphyry. Spectrometer readings show that it commonly contains 20 to 100 ppm U (Table 2). Other zones are in the white felsic volcanic units that contain disseminated magnetite and/or streaky aggregates of biotite, hornblende, and magnetite, and rarely magnetite streaks as much as 10 cm long.

Uranium is preferentially concentrated in these mafic aggregates. It occurs mainly as euhedral crystals of uraninite, but pitchblende is also present as grains and overgrowth on uraninite (Paterson, 1975b). Miller (1982a, b) noted an association of epidote, apatite, and fluorite, and some scapolitization in the host rocks. He obtained a date of 1864 ± 14 Ma for the uraninites in relatively more mafic rocks from the showing.

Zones of uranium concentration are discontinuous, and high grade parts are only a few centimetres wide and a few metres long, as seen from trenches and drill core (Legagneur, 1969; Paterson, 1974, 1975a, b; Male, 1978). They trend

Figure 11. Geology, trenches, and diamond-drill holes of the UGI uranium prospect, DeVries Lake.

northerly and dip moderately to steeply to the east. The best intersection obtained was from drill hole DV-2 at about 10 m depth, and it averaged 0.07% U over 1.5 m (Male, 1978).

DISCUSSION

Iron oxide-rich metasiltstones and argillite are favourable host rocks for synsedimentary and epigenetic mineralization. Miller (1982a, b) advocated a metaplacer origin for the magnetite and associated uraninite at DeVries Lake and at a few other localities in the Great Bear magmatic zone. Gandhi (1992a), however, favoured chemical precipitation of iron during deposition of the sediments, and a later introduction of uranium. This is supported by: i) scarcity or absence of typical detrital minerals such as zircon, ii) lack of lithologies and structures expected in high energy environments of placer-hosting sequences, and iii) the abundance of argillaceous beds in the siltstones. Chemical deposition of iron is common in early Proterozoic platform-shelf sequences, and the Snare Group is not an exception. Its metamorphism has led to the formation of magnetite-rich siltstone and argillite, and development of iron-rich silicates in appropriate lithologies. The presence of boudin-like massive magnetite lenses in the metasiltstones is, however, intriguing (Fig. 4). These could be either due to local high concentration of iron during deposition or later segregation during deformation and metamorphism. These relatively competent magnetite lenses were fractured during deformation, and were the preferred sites for emplacement of pegmatitic veins and pods. Anomalously high radioactivity is common in these magnetite lenses due to concentration of uranium, mainly along fractures and partly as disseminations. Oxidation of magnetite to hematite creates favourable conditions for reduction and precipitation of uranium from circulating solutions, which may be magmatogenic or meteoric.

The concentration of sulphides along thin beds in metasiltstone at the Kol prospect is considered here as syngenetic, modified by deformation. Such sulphide occurrences are, however, not common in the iron oxide-dominated metasedimentary rocks of the area. Cobalt is found in some hydrothermal veins in the southern Great Bear magmatic zone (Gandhi and Lentz, 1990). A later hydrothermal event may have introduced Co and W into a chemically favorable zone at the Kol prospect.

The U-Mo association seen in veins at the 'Nori/RA' claims is typical of granite-related mineralization in many parts of the world. The presence of flourite, tourmaline, apatite, and allanite in them further supports this interpretation. Development of coarse biotite in association with the veins is clearly a result of hydrothermal alteration of the host siltstones, which contain variable amounts of finer grained biotite. Magnetite is also present in the host rocks and forms coarse aggregates in the Mo-U occurrences. This indicates local reconstitution during mineralization. Isotopic analyses of uraninites from these veins reported by Miller (1982a, b) result in an upper intercept at 1853 ± 19 Ma and a lower intercept at 70 ± 84 Ma. The time of formation of the uraninites is close to that of the volcano-plutonic activity in the southern Great Bear magmatic zone, as indicated by U-Pb zircon dates in the range 1870 to 1860 Ma (Gandhi and Mortensen, 1992). It is thus very likely that the Mo-U mineralization is related to a granitic pluton of this age at depth, and the veins are part of a large hydrothermal system.

The uranium concentrations at the UGI and FXO prospects are restricted to the felsic volcanic assemblage, and occur along zones that are broadly conformable with the stratigraphy. These features and the occurrence of much of the uraninite as disseminated grains indicate its close genetic link with the host felsic volcanic rocks. This linkage is confirmed by close association of uraninite with primary magnetite and/or mafic silicates seen in these rocks to a depth of as much as 55 m (Male, 1978). Supporting evidence is provided by an isotopic date of 1864 ± 14 for the uraninites from the UGI prospect (Miller, 1982a, b). The host rocks have not yet been dated, but they are similar to the felsic volcanic rocks elsewhere in the southern Great Bear magmatic zone that are 1870-1860 Ma old (Gandhi and Mortensen, 1992). The high grade uraninite/pitchblende veins and fracture fillings resulted from migration of this primary uranium into openings formed during deformation and local reconcentration at the margin of mafic intrusions and pegmatitic veins. Weak spotty radioactivity found in arkosic quartzite at the Tyke showing may be due to a more distant secondary concentration of uranium.

The Great Bear magmatic zone is characterized by an abundance of felsic volcanic rocks of the type seen at DeVries Lake. Hence potential for larger volcanogenic uranium deposits is regarded as very good. Furthermore the iron-uranium enrichment in these rocks represents a stage in metallogenic evolution of the magmatic zone, leading eventually to magnetite-rich breccias and veins that occur in the region, for example the polymetallic Sue-Dianne and Mar deposits (Gandhi, 1989; 1992b). These deposits have similarities with monometallic Kiruna deposits in northern Sweden and polymetallic Olympic Dam-type deposits in South Australia (Gandhi and Bell, in press).

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 750010

The Snowbird tectonic zone in District of Mackenzie, Northwest Territories

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Hanmer, S. and Kopf, C., 1993: The Snowbird tectonic zone in District of Mackenzie, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 41-52.

Abstract: Snowbird tectonic zone has been traced from northern Saskatchewan to 61° 32'N. It includes a 5-10 km thick belt of granulite to upper amphibolite facies ribbon mylonites derived from a protolith of diatexite, anorthosite, garnet pyroxenite and granite. It is a dextral strike-slip shear zone whose highly sinuous trace is controlled by the existence of crustal-scale, pre-existing stiff mafic to intermediate lozenges of possible arc affinity. Although the mylonite fabrics are spectacular, the curvilinear shear zone cannot have accommodated very large strike-slip displacements.

Résumé : La zone tectonique de Snowbird s'étend depuis le nord de la Saskatchewan jusqu'à 61° 32' de latitude nord. Elle comprend une zone de mylonites, 5 à 10 km de large, parvenues au faciès des granulites et des amphibolites suite à la déformation des diatexites, des anorthosites, des pyroxénites à grenat et des granites. Elle représente un décrochement dextre dont la trace serpentine à été influencée par l'existence de masses importantes de roches mafiques ou dioritiques, provenant vraisemblablement d'un arc magmatique. Bien que les fabriques mylonitiques soient spectaculaires, la zone de cisaillement serpentine ne pourrait pas être le site de déplacements tectoniques importants.

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INTRODUCTION

Snowbird tectonic zone, clearly delimited as a pronounced linear anomaly in the horizontal gravity gradient map of the Canadian Shield (Goodacre et al. 1987), is well exposed as a triangle of ca. 2.6 Ga, high grade mylonitic rocks (East Athabasca mylonite zone) in the Stony Rapids area, northern Saskatchewan (Hanmer 1987; Hanmer et al. 1991, 1992). The granulite facies mylonites lie at the northeast end of the largest of a chain of magnetically defined elliptical domains (Geological Survey of Canada, 1987; Athabasca, Selwyn and Three Esker lozenges), which may represent relatively stiff, crustal-scale 'boudins' (Fig. 1).

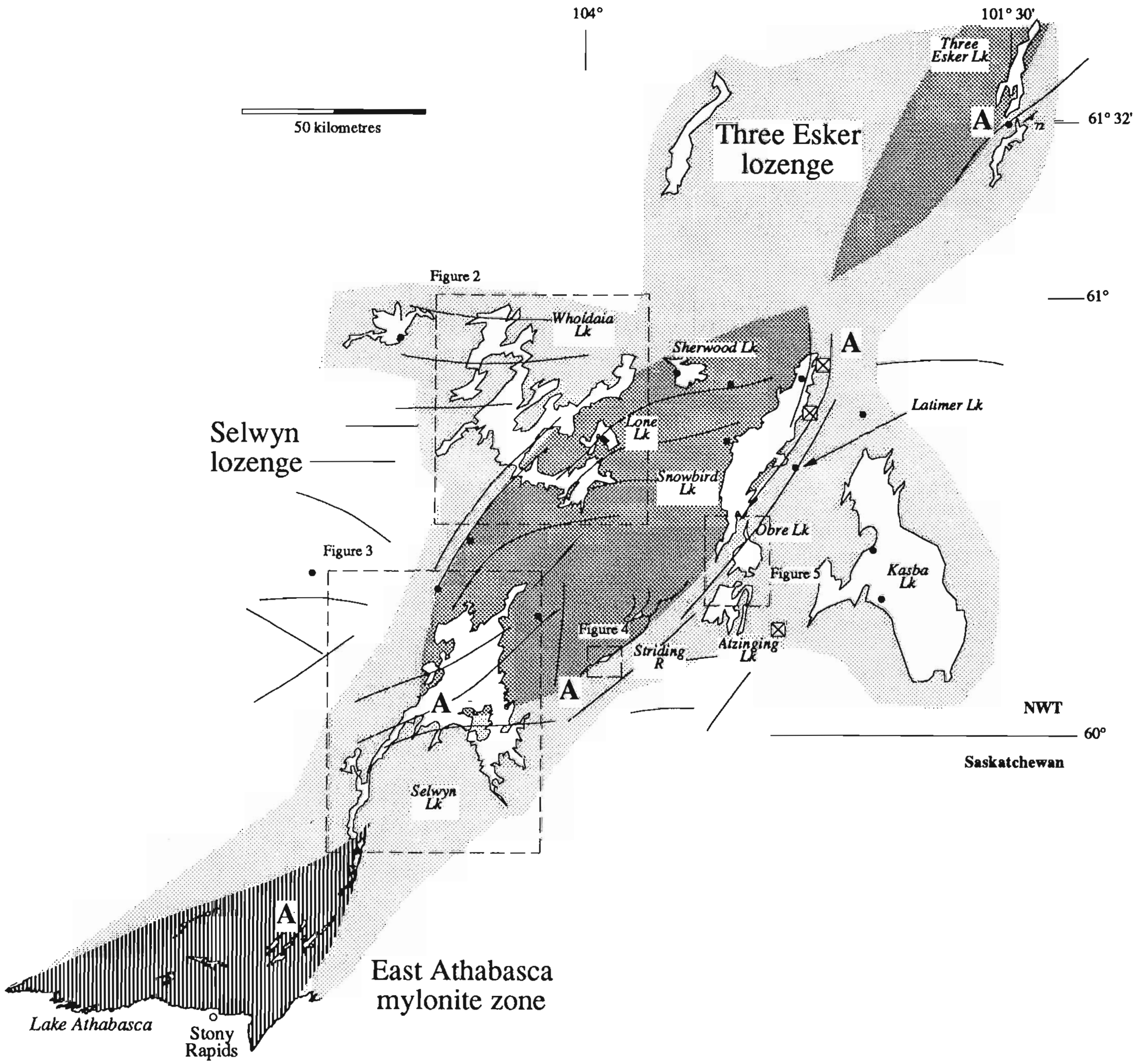

Figure 1. Location of principal geological and geophysical elements of the Snowbird tectonic zone. Medium grey shading indicates the magnetically defined Selwyn and Three Esker lozenges. Ruled shading indicates the magnetic expression of the East Athabasca mylonite zone (Hanmer et al. 1992), at the northeast end of the magnetically defined Athabasca lozenge (not shown). Solid lines are foliation trajectories taken from Taylor (1963, 1970). Boxes show Figure locations. Crossed squares indicate traverses or fly camps as opposed to fly-in stations indicated by black dots. A = anorthosite.

The expression of Snowbird tectonic zone in the potential field patterns extends across Northwest Territories to Hudson Bay, east of Baker Lake. However, high grade mylonites have not been reported from the vicinity of Snowbird tectonic zone to the southwest of Baker Lake (Eade, 1985; Tella and Eade, 1985). Furthermore, low grade ca. 2.6-2.7 Ga tuffs have been identified within 20 km of the trace of Snowbird tectonic zone at Angikuni Lake (Eade, 1986; Loveridge et al., 1988).

We have spent six weeks examining the geology which underlies the northeastward extension of the geophysical expression of Snowbird tectonic zone, principally in the Wholdaia Lake and Snowbird Lake map areas (Figs. 1-5), previously mapped at 1:250 000 scale by Taylor (1963, 1970). The area is blanketed by Quaternary deposits and poorly exposed. Our objectives were to determine:

- To what extent the geological signature of East Athabasca mylonite zone can be extrapolated along the geophysical trace of Snowbird tectonic zone to the northeast;
- The nature and origin of the magnetically defined elliptical lozenges along the trace of Snowbird tectonic zone; and
- The existence, or otherwise, of a metamorphic field gradient along the trace of Snowbird tectonic zone, possibly reflecting shallower structural levels towards the northeast.

Guided by the lenticular pattern in the regional magnetic field, and constrained by the outcrop distribution as shown on Taylor's maps, we focused our attention on the principal elements of the geology both inside and outside the magnetically defined lozenges, but more particularly at their margins.

Our principal conclusions are:

(1) Upper amphibolite facies ribbon mylonites form a remarkably sinuous belt, 5-10 km wide, apparently traceable along strike for at least 300 km into the District of Mackenzie. The upright mylonite belt, associated with a subhorizontal extension lineation and formed by dextral transcurrent shear, is the extension of East Athabasca mylonite zone.

(2) The aeromagnetically defined lozenges are, at least in part, underlain by mafic to intermediate rocks, whereas their wallrocks are composed of diverse, multicomponent granodioritic and granitic gneisses intruded by granite plutons.

(3) Although the mylonite belt runs along the southeast margins of the lozenges, only discontinuous strands of high grade mylonite are present on their northwest side. Accordingly, the geological origin of the lozenges is independent of, and older than, the shearing which produced the mylonites.

(4) Metamorphic grade associated with the Snowbird tectonic zone does not decrease toward the northeast. Therefore,the distribution of Archean rocks of both high and low metamorphic grade along Snowbird tectonic zone must reflect later faulting.

Furthermore

(5) The mylonite belt is spatially associated with a diatexite - anorthosite protolith assemblage. This protolith assemblage is a fundamental characteristic of the Snowbird tectonic zone.

(6) Spatial restriction of anorthosite to the southeast side of a large volume of mafic to intermediate rocks in Selwyn lozenge is similar to the relationship between anorthosite and Bohica mafic complex in the East Athabasca mylonite zone and appears to represent a systematic regional relationship.

Finally

(7) It should be clearly understood that "Snowbird tectonic zone" refers to a geophysically defined feature which, in the study area, is geologically manifested as two components: mylonites and the pre-existing crustal-scale lozenges which they bound.

SELWYN LOZENGE

Three elliptical domains, or lozenges, are discernible in the regional aeromagnetic field (GSC 1987; Fig. 1). The Athabasca lozenge is the largest, 300 x 100 km, and extends southwest from eastern Lake Athabasca. It is largely obscured by the Athabasca Sandstone and the Phanerozoic cover. The Three Esker lozenge is the smallest and is extremely poorly exposed; we have only found one appropriately located exposure (61° 32'N by 101° 30'W). Between them lies the Selwyn lozenge (125 x 50 km).

The Selwyn lozenge is lithologically diverse. Its western and northern part is principally composed of homogeneous to banded amphibolite and abundant hornblendite, all intruded by a vein network of leucodiorite to tonalite (e.g. Wholdaia Lake; Fig. 2 and 6A), or metagabbro intruded by folded granitic veins and cross-cutting lamprophyre dykes (e.g. Lone Lake, Fig. 1). Its eastern margin, at Striding River, is composed of an extensive, monotonous, fine grained, annealed mafic gneiss of uncertain origin. Relict granulite facies assemblages (garnet - clino/orthopyroxene) are locally preserved in the mafic rocks at both Wholdaia Lake and Striding River. West of Snowbird Lake, the Selwyn lozenge is predominantly composed of irregularly folded migmatitic granitoid orthogneiss, or granodioritic gneiss with multiple phases of variably (poorly) deformed granitoid veins and intrusions. North Selwyn Lake is underlain by variably deformed hornblende diorites to granodiorites. Coarse garnet-sillimanite diatexite and foliated gneissic granitoid rocks occur throughout the lozenge.

For the most part, rocks within the Selwyn lozenge are not strongly deformed (e.g. Fig. 6B). They preserve cross-cutting relationships, open fold profiles, igneous textures in foliated plutonic rocks etc. In the case of the foliated plutonic rocks, it is uncertain whether their deformation state is a reflection of the magnitude of the total regional strain, or the timing of their intrusion with respect to the deformation history. The principal exception to the foregoing concerns the monotonous mafic rocks northwest of Striding River (Fig. 4) which texturally and compositionally resemble the annealed granulite facies ultramylonites derived from a norite protolith in the East Athabasca mylonite zone (Hanmer et al. 1992). However, no protolith was identified in the Striding River case and the structural significance of the textures there remains unresolved. Foliation trends within the lozenge are curvilinear with a significant east-west component, clearly discordant to the northeast trending lozenge boundaries (Fig. 1).

WALLROCKS

The wallrocks on the northwest side of the Selwyn lozenge comprise a variety of folded and injected granitoid gneisses, associated with poorly foliated granitoid plutons of unknown age. Foliation trends are highly variable, but are predominantly east-west (Fig. 1). Southeast of the Selwyn lozenge, the poorly exposed wallrock is apparently dominated by the voluminous mass of isotropic biotite granite around Kasba Lake.

Between the granite and the Selwyn lozenge is a 10 km wide belt of amphibolites and finely laminated pelitic and volcaniclastic(?) sedimentary rocks which we examined along a river section at the latitude of Atzinging Lake (Fig. 1). The enveloping surfaces of the stratification (S_0) and a penetratively developed bedding-parallel cleavage (S_1) are generally flat-lying (Fig. 6C), and a mineral extension lineation (L_1) is oriented ca. 030°. The cleavage is locally deformed about 1 cm to 10m scale lineation-parallel folds (F_2). The fold axial planes are shallowly to moderately east

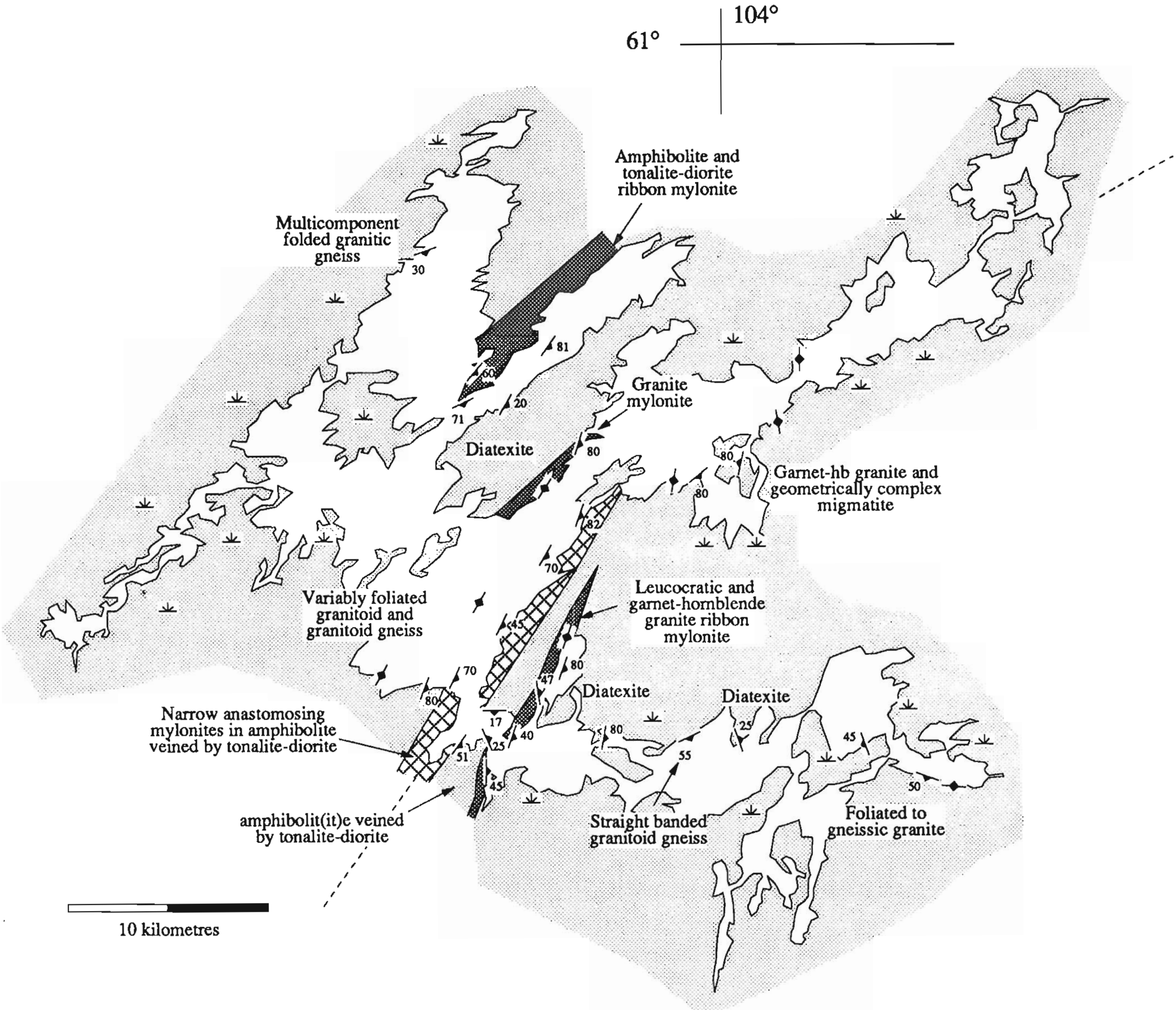

Figure 2. Generalized geology of Wholdaia Lake. Dashed line is the magnetically defined limit of the Selwyn lozenge. Details are discussed in the text.

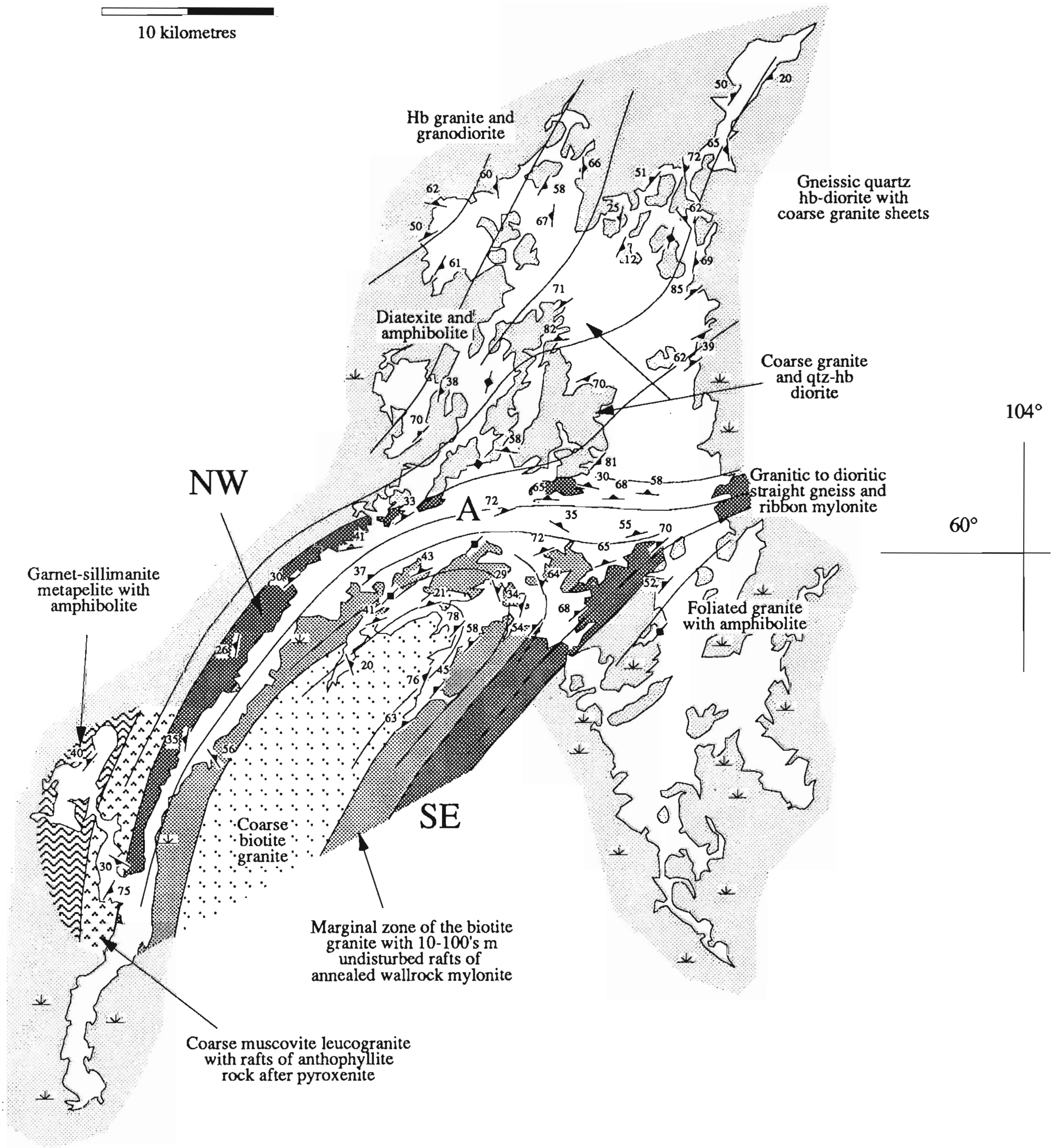

Figure 3. Generalized geology of Selwyn Lake. Solid lines are foliation trajectories. A = anorthosite. Detailed are discussed in the text.

or west dipping, and the folds verge in both directions. No S_2 cleavage is present. The principal metamorphic mineral assemblage (hornblende-chlorite) is indicative of lower amphibolite facies, although 1m thick S_1-parallel zones of chlorite schist, crenulated by F_2, suggest that retrogression to greenschist facies occurred between D_1 and D_2. In brief, these are typical supercrustal rocks which lie at the southwestern end of the Rankin-Ennadai greenstone belt (e.g. Wright 1967).

STRIDING MYLONITE BELT

A belt of upper amphibolite facies mylonites occurs along the southeastern side of the Selwyn lozenge from Selwyn Lake, via Striding River, to the north end of Snowbird Lake: Striding mylonite belt (new name). The trace of the mylonite belt is remarkably sinuous and closely follows the aeromagnetically defined trace of the Selwyn lozenge. The mylonite belt trends northeast along the southwest arm of Selwyn Lake, swings eastward across the middle of the lake, then returns to a northeast trend along Striding River and southern Snowbird Lake, turning to a northward strike as one approaches northern part of the lake (Fig. 1). The isolated outcrop at Three Esker Lake, in combination with regional trends (see Wright, 1967), suggests that the sinuous character of the trace of the mylonite belt persists to, at least, 61° 30′ latitude.

The Striding mylonite belt is apparently between 5-10 km thick, except at Selwyn Lake where it bifurcates into two thinner strands (see below; Fig. 3). It is associated with an upright mylonitic foliation, subhorizontal extension lineation and dextral shear sense (e.g. Hanmer and Passchier, 1991). Although compositionally diverse along the southwest arm of Selwyn Lake (see below), the mylonites are principally derived from diatexite, anorthosite and mafic rocks, and granites.

At Striding River (Fig. 4), the river section is composed of pink leucogranitic ribbon ultramylonite with map-scale bands of ribbon ultramylonite derived from garnet anorthosite and garnet hornblendite (ex-pyroxenite?) protoliths. Southeast of the river, the leucogranitic ribbon mylonites are banded due to the presence of amphibolites, metres to hundreds of metres thick (Fig. 7A, B). Locally, banded garnet - clinopyroxene mafic granulites are preserved, suggesting that the early part of the mylonitization may have witnessed granulite facies.

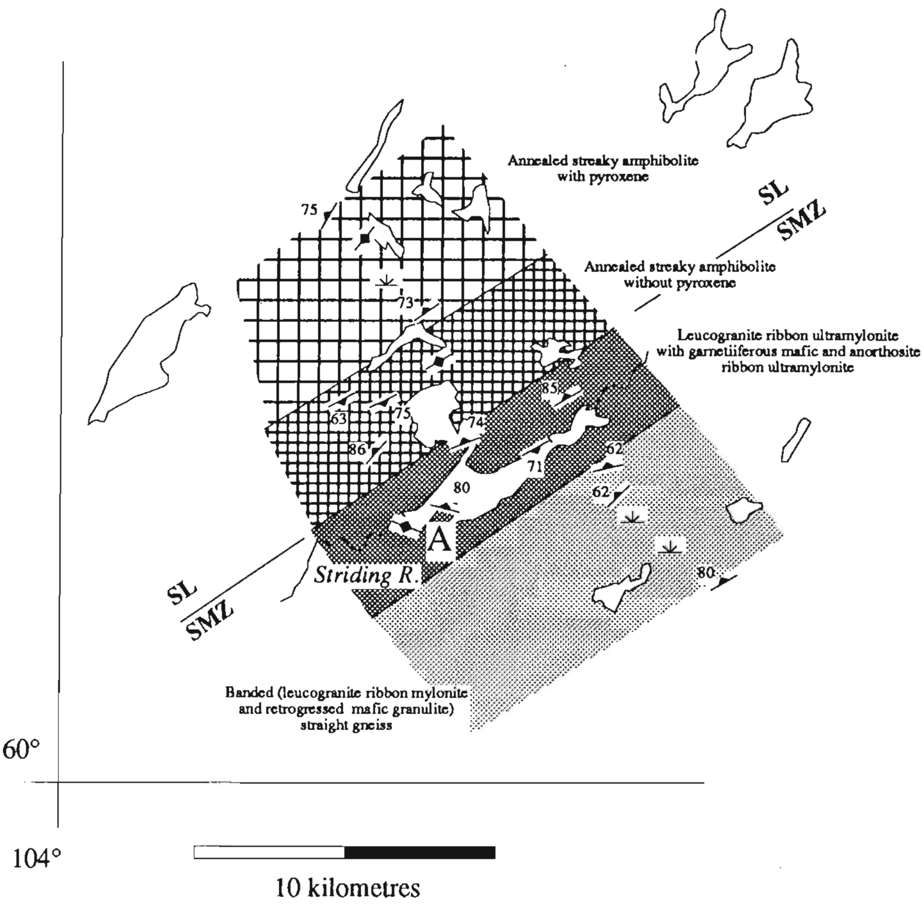

Figure 4. Sketch of geological observations in the vicinity of part of the Striding River. SL = Selwyn lozenge, SMZ = Striding mylonite zone, A = anorthosite. Details are discussed in the text.

In south Snowbird Lake (Fig. 5), ribbon mylonites are derived from diatexites with intrusive sheets of white leucogranite. At nearby Obre Lake, coarse granitoids are deformed by discontinuous narrow mylonite zones, whereas at Latimer Lake, just east of mid-Snowbird Lake (Fig. 1), non-descript granitic and amphibolitic gneisses have suffered cataclastic shearing.

The best exposures of the Striding mylonite belt occur east of northern Snowbird Lake (Fig. 1). From the lakeside to the east, we have found (ultra)mylonites derived from a very coarse granite, banded garnet anorthosite with garnet hornblendite (meta-pyroxenite?) and garnet - sillimanite diatexite (Fig. 8A-C). The mylonite belt appears to be ca. 8 km wide, on the assumption that the lower strain fabrics observed at the eastern end of our traverses represents wallrock, as opposed to low strain pods within the shear zone.

The large exposure at Three Esker Lake is predominantly composed of diatexite ribbon ultramylonite (Fig. 8D) with veins of mylonitised pink leucogranite. However, the most

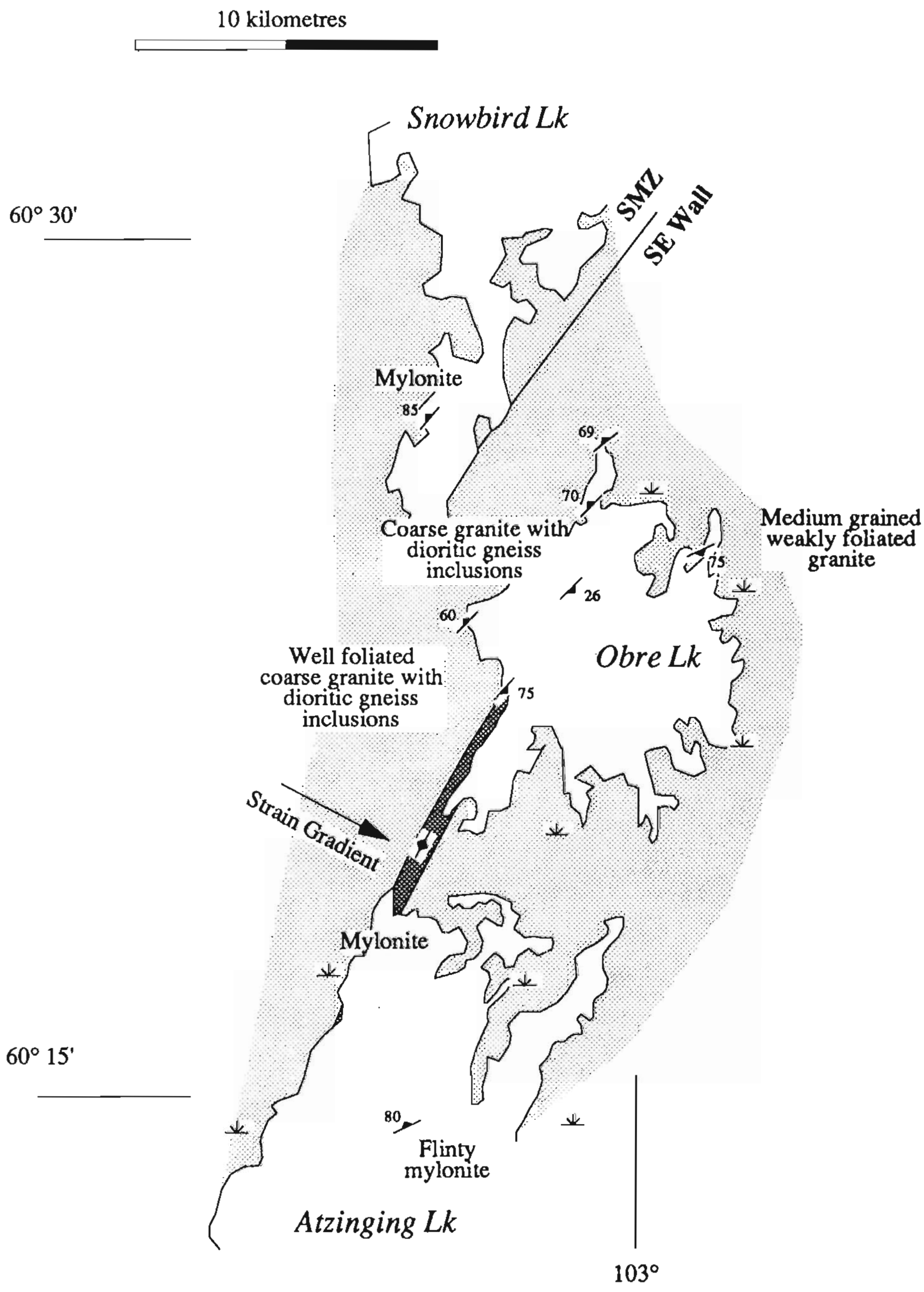

Figure 5. Sketch of geological observations in the Atzinging - Obre - south Snowbird Lakes area. SMZ = Striding mylonite zone. Details are discussed in the text.

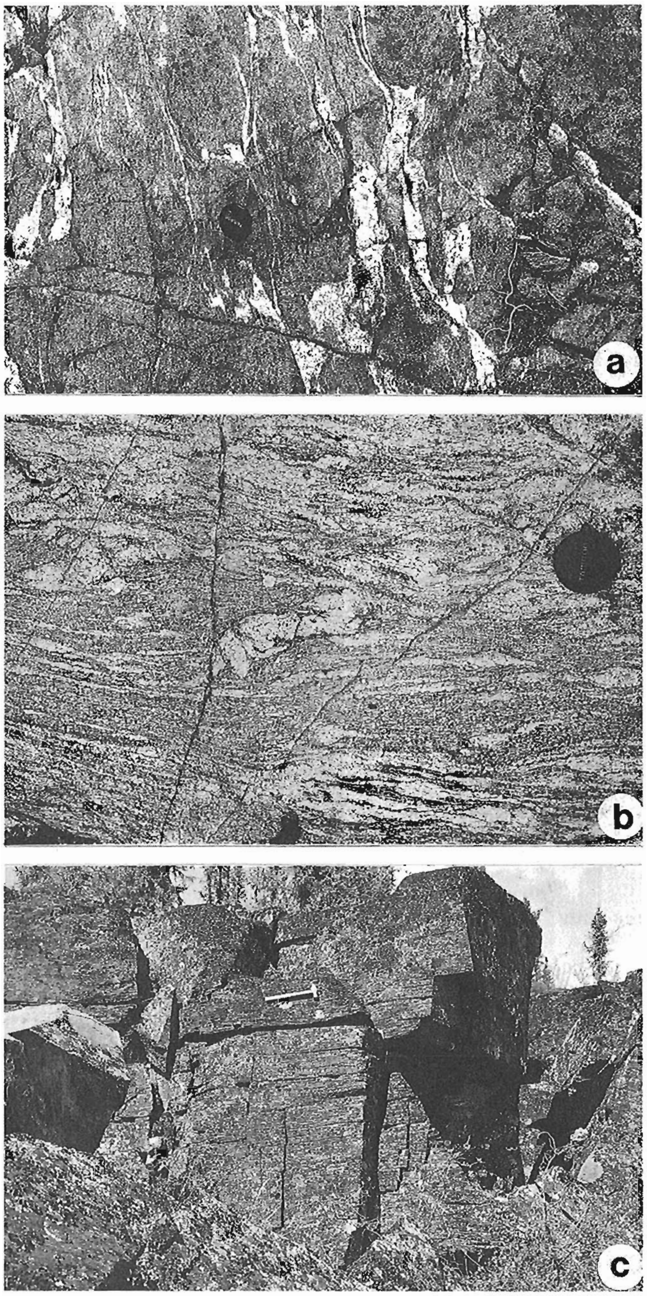

Figure 6.

a. Amphibolite intruded by branching quartz diorite or tonalite veins, Wholdaia Lake (GSC 1992-241R).

b. Well preserved migmatitic structure, Wholdaia Lake (GSC 1992-241J).

c. Flat lying bedding-parallel cleavage in fine grained mafic volcaniclastic(?) material, part of the greenstone belt between Atzinging and Kasba Lakes (GSC 1992-241A).

striking feature of this outcrop is the presence of ribbon mylonites after banded garnet anorthosite and garnet hornblendite, identical to those along strike to the southwest.

WHOLDAIA LAKE

Wholdaia Lake is well situated to permit examination of the northwestern margin of the Selwyn lozenge (Fig. 1 and 2). Although the magnetic expression of the boundary is discrete and trends northeast through the middle of the lake, it does not coincide with any simple geological feature. The western shore of the lake exposes a granitic gneiss, with folded banding and folded, cross-cutting granitic veins. The overall strike here is discordant to the general trends of the mylonites in the middle of the lake. Similarly, in the southeast arm of

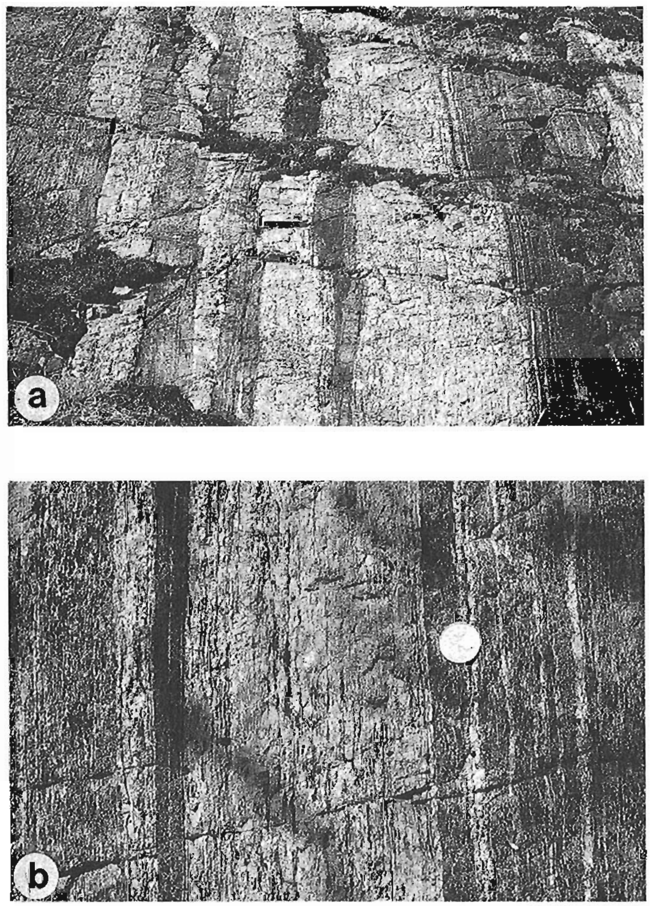

Figure 7.

a. Banded granite and amphibolite ribbon mylonites, southeast of Striding River (GSC 1992-241I).

b. Detail of A (GSC 1992-241H).

the lake, poorly foliated diatexite and foliated to gneissic granite are also structurally discordant (Fig. 2). The mylonites are everywhere associated with an upright foliation, subhorizontal extension lineation and dextral sense of shear.

At first glance it would appear that the mylonites should define the geological limit of the Selwyn lozenge, separating wallrock gneisses on the western lake shore from the lozenge interior on the eastern side of the lake. To our surprise, the mylonites themselves, although very well developed at the outcrop scale, do *not* define the lozenge boundary. Although their thickness may exceed hundreds of metres, the mylonites are laterally discontinuous and do not extend beyond Wholdaia Lake. Despite a directed search, we were unable to find any trace of the mylonites along the lozenge margin between Wholdaia and Selwyn Lakes. Although it is possible that the foliated granitoids of north Selwyn Lake (Fig. 3) could conceivably represent late intrusions which mask the structure of the original lozenge margin, there is no evidence to support such speculation.

SELWYN LAKE

The Striding mylonite belt bifurcates in the southern half of Selwyn Lake (NW and SE in Fig. 3). In the northwest strand, northeast-trending mylonites run up the southwest arm and swing progressively to strike east-west across the centre of the lake. The southeast strand strikes ca. 045° across the main body of the lake. Most of the mylonite is present as annealed, slabby quartz leucodioritic – amphibolite straight gneiss (Hanmer, 1988), locally with excellent development/preservation of feldspar porphyroclasts and quartz or feldspar ribbons. However, the east-west striking mylonites in the central part of the lake are principally ribbon ultramylonites (Fig. 9A, B) derived from garnet-sillimanite diatexite and garnet anorthosite protoliths. The annealed mylonites are cut by an array of 1-10 m thick sheets of generally concordant pink biotite granite in various states of deformation, even within a given outcrop. The most strongly deformed granite sheets are garnetiferous ribbon ultramylonites which may

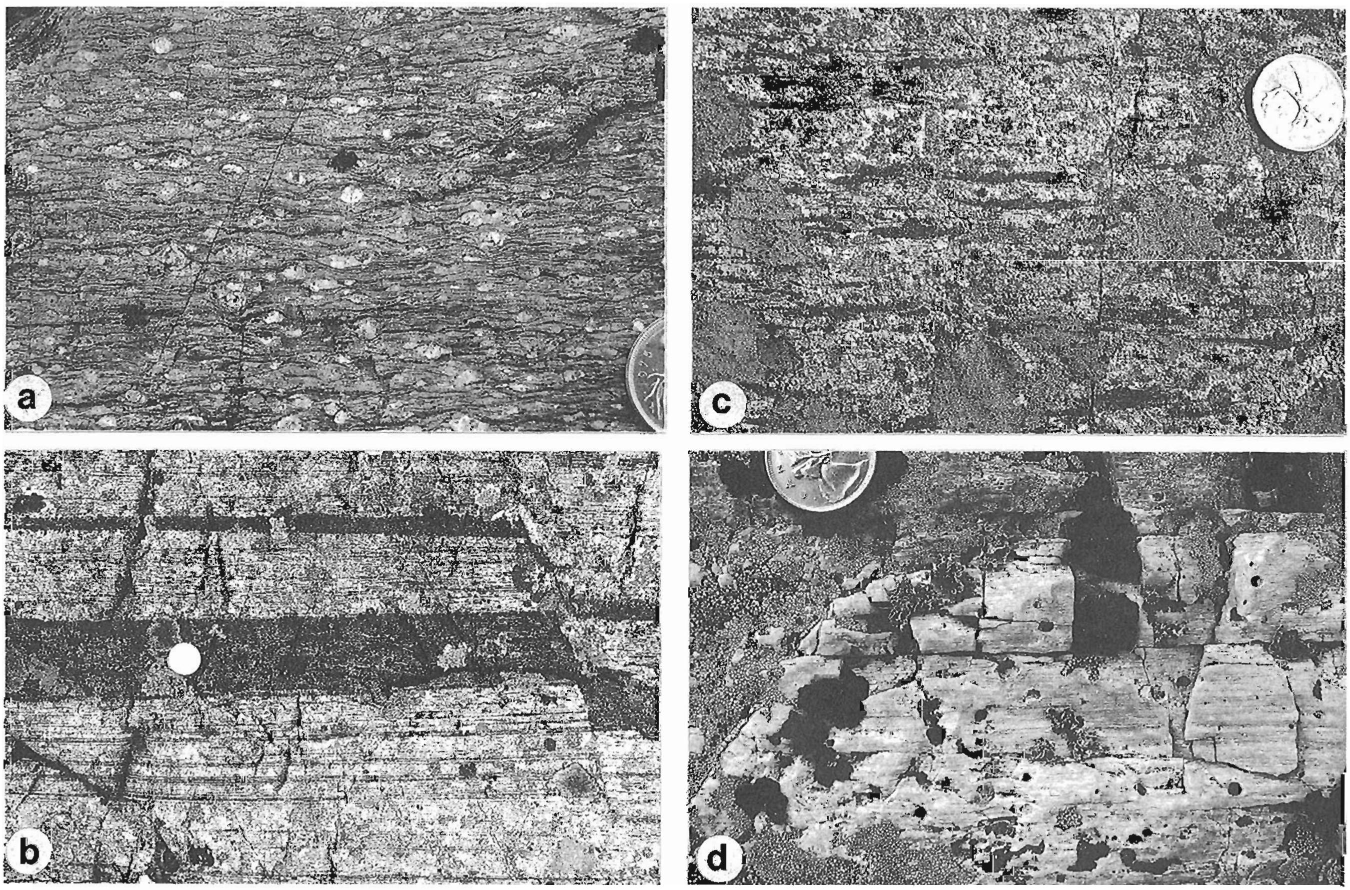

Figure 8.

a. Granitic ribbon mylonite (GSC 1992-241V).

b. Garnet anorthosite and garnet hornblendite (meta-pyroxenite) ribbon ultramylonite (GSC 1992-241S).

c. Garnet anorthosite ribbon ultramylonite with hornblende ribbons after pyroxene (GSC 1992-241W).

d. Detail of garnet granite ribbon ultramylonite from diatexite (GSC 1992-241Z).

A-C are from northern Snowbird Lake, D is from Three Esker Lake.

themselves be cut by less deformed sheets of similar granite. The mylonites in both strands dip moderately to the west and north, carry a subhorizontal extension lineation and were formed by dextral transcurrent shear.

The two strands of mylonite are separated by a body of coarse grained, isotropic biotite granite (Fig. 3). The marginal zone between the granite and the mylonites is composed of the isotropic to poorly foliated granite with large (hundreds of metres long by tens of metres thick) rafts of annealed mylonite, somewhat coarser grained than the adjacent mylonites outside of the granite, but still recognizable as the same material. The planar fabrics within the rafts and, consequently, the shapes of the rafts themselves, lie in the same orientation as the external mylonites, except at the northeast termination of the granite where they have been re-oriented into parallelism with the granite contact.

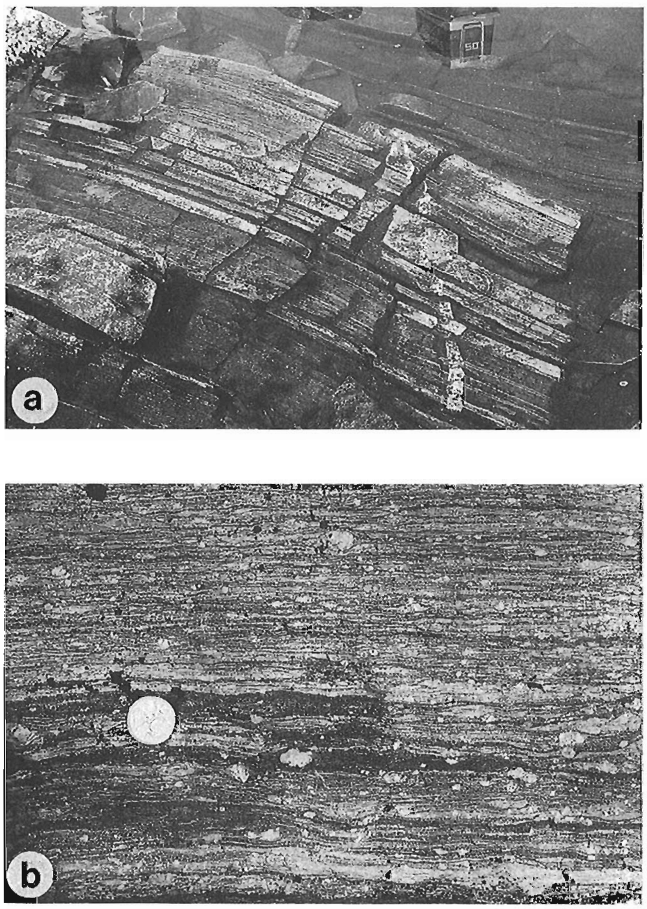

Figure 9.

a. Slabby ribbon ultramylonites in garnet diatexite, northwestern mylonite strand, Central Selwyn Lake (GSC 1992-241G).

b. Granitic ribbon mylonite, southeastern mylonite strand, Central Selwyn Lake (GSC 1992-241C).

It appears that the biotite granite has intruded forcefully into and along a dextral strike-slip shear zone. However, the presence of mylonitized to non-mylonitized granite veins in the wallrock suggest that the final emplacement of the granite was preceded by its own vein cortege and that the granite is in fact late syntectonic with respect to the Striding mylonite belt. We note that, far from "lubricating" the shear zone into which it was intruded (e.g. Hollister and Crawford, 1986), the emplacement of the granite appears to have provoked bifurcation of the shear zone and to have witnessed the cessation of deformation.

SUMMARY

Mylonites

Mylonites along what is now called the Snowbird tectonic zone were first referred to by Taylor (1963). He drew a discrete fault along the Striding River and through Snowbird Lake, to the east of which he mentioned the presence of mylonitization. High grade mylonites of the Snowbird tectonic zone were first mapped and described from the East Athabasca mylonite zone in northern Saskatchewan (Hanmer, 1987; Hanmer et al. 1991, 1992; however, see plate vi in Taylor 1963, p 17).

Prior to this summer's fieldwork, the East Athabasca mylonites appeared to end at the apex of the triangular mylonite zone, just southeast of Selwyn Lake (Hanmer et al., 1992). Our new data suggest that the mylonite strands of Selwyn Lake project southwards to merge with the East Athabasca mylonite zone. Accordingly, the Striding mylonite belt is the extension of the East Athabasca mylonite zone, and together they constitute a significant component of the Snowbird tectonic zone.

The sinuous course of the Striding mylonite belt imposes an important constraint on the displacement it has accommodated. Between Selwyn Lake and the north end of Snowbird Lake, the azimuth of the mylonites swings through 90°. This is, to say the least, an extremely inefficient geometry for a transcurrent fault (e.g. Saucier et al., 1992). However, the kinematical configuration of the triangular East Athabasca mylonite zone to the southeast (Hanmer et al., 1992), essentially composed of two conjugate shear zones, is equally inefficient in terms of the amount of regional wallrock displacement it can have accommodated. Accordingly, it would appear that the spectacular nature of the Snowbird tectonic zone mylonite fabrics is more a reflection of the nature of their development at vcry high metamorphic grade, rather than an indication of extremely large displacements (see Hanmer, 1987).

Lozenges, lithology, and rheology

The Selwyn and Three Esker lozenges have clearly controlled the sinuous trace of the Striding mylonite belt. Furthermore, we can infer that the Athabasca lozenge has played a pivotal role in the location and complex kinematic development of the East Athabasca mylonite zone (see Hanmer et al., 1992).

Presumably all three lozenges have exerted their influence as a result of a rheological contrast between themselves and the softer wallrocks. Only the Selwyn lozenge is readily accessible. In general, it appears to be more mafic to intermediate in composition, compared with the granodioritic to granitic composition of the wallrocks. Mafic to intermediate compositions are also reported from the southwestern part of the Athabasca lozenge (Crocker and Collerson 1988). Nevertheless, our limited observations, while tending to support the notion of a stiffer rheology within the Selwyn lozenge, can only be tentatively extrapolated to explain the existence and rheology of the other lozenges, especially in light of the abundant mafic mylonites in the East Athabasca mylonite zone (Hanmer et al., 1992).

The Striding mylonite belt is associated with a lithological assemblage of semipelitic, anorthositic, pyroxenitic and probably gabbroic/noritic origin, indistinguishable from the equivalent rocks in the East Athabasca mylonite zone. Our field observations allow us to correlate the trace of the mylonite belt with a discrete, sinuous, positive anomaly in the horizontal gravity gradient map (Goodacre et al., 1987). The amplitude of the anomaly is strongest along the east side of the Selwyn lozenge between the Striding River and the north end of Snowbird Lake, precisely where we have observed the largest exposed volumes of anorthosite-garnet hornblendite (meta-pyroxenite) mylonite. However, the volumes of observed mafic mylonite would appear to grossly under-represent the volumes which must be present at depth, and/or covered by drift, in order to account for the intensity of the gravity anomaly. In brief, the anorthosite-pyroxenite-diatexite association appears to have played an important role in the localisation of the Snowbird tectonic zone mylonites.

The symmetrical relationship between the East Athabasca mylonite zone and the magnetically defined Athabasca lozenge is indicative of a genetic relationship (Hanmer et al., 1992). However, the asymmetry of the distribution of Striding mylonite belt along only the southeastern side of the Selwyn lozenge indicates that the origin of the lozenge geometry is independent of, and probably pre-dates, the mylonites of the Snowbird tectonic zone. If both the lozenges and the anorthosite-pyroxenite-diatexite association have influenced the development of the Snowbird tectonic zone, what is the relationship between them? What, indeed, is the origin of the lozenges? For now, we can only speculate, but we note the following:

- The anorthosite-pyroxenite systematically lies to the southeast of the mafic and intermediate rocks of the Selwyn lozenge and the Bohica complex (East Athabasca mylonite zone).
- Similarly, the single outcrop on the southeast side of the Three Esker lozenge contains anorthosite-pyroxenite. The southeast side of the Athabasca lozenge is obscured by younger rocks.
- The anorthosite in the East Athabasca mylonite zone was intruded and partly dismembered by the tonalitic Chipman batholith prior to ca. 3.0 Ga, possibly before 3.2 Ga (R. Parrish, unpublished data in Hanmer et al., 1992).
- The Bohica mafic complex was intruded into the semipelitic protolith to the Reeve diatexite (Hanmer et al., 1992).
- The granulite facies ribbon mylonites in the East Athabasca mylonite zone were formed at ca. 2.6 Ga (R. Parrish, unpublished data in Hanmer et al., 1992).
- Initial Nd model ages for the protoliths to the East Athabasca mylonite zone fall in the range 3.5-2.9 Ga (Kopf, unpublished data), in contrast to the younger (2.9-2.5 Ga, Dudas et al. 1991) or mixed ages (3-3-2.4 Ga, Collerson et al. 1989) obtained elsewhere in the western Churchill Province.

Accordingly, we are now undertaking research to address the following speculations: Are the anorthosites genetically associated with the adjacent mafic rocks to the northwest? If so, do they, together with the diatexites, represent the basal part of a pre-ca. 3.2 Ga, or older, arc intruded into an accretionary wedge? Were the anorthosites and associated rocks structurally transported to their present relative structural level, perhaps by northwest-side-up, dip-slip faulting, long before being reworked by (intra-continental?) strike-slip mylonitization (pre-3.0 Ga *vs* 2.6 Ga)? If so, was dip-slip faulting the crustal-scale response to a collisional event during the early accretionary history of the Churchill Province? Answers to such speculative questions will shed light on the Mid-Archean crustal evolution of this part of the Churchill Province. In particular, they should address the possibility of the existence and scale of a ca. 3.2 Ga, or older, arc and of an ultimately collisional suture of similar vintage.

Faulting (Early Proterozoic?)

The mylonites at Three Esker Lake formed at upper amphibolite facies. Clearly, there is no indication of a significant northeastward progression to shallower structural levels along the trace of the Snowbird tectonic zone. Therefore, the absence of extensive high grade mylonites (Eade, 1985; Tella and Eade, 1985), and the proximity of low grade ca. 2.6-2.7 Ga supracrustal rocks to the trace of the Snowbird tectonic zone at Angikuni Lake (Loveridge et al. 1988), 100 km to the northeast of the present study area, are perhaps most readily ascribed to the action of later brittle faulting.

Similarly, if the Kasba-Atzinging greenstones are of the same ca. 2.7 Ga vintage as other parts of the Rankin - Ennadai belt (Chiarenzelli and Macdonald, 1986;Mortensen and Thorpe, 1987; Tella et al., 1992) and the mylonite belt is of the same vintage as the East Athabasca mylonite zone (ca. 2.6 Ga, R. Parrish, unpublished data in Hanmer et al. 1992), then the former cannot represent a cover sequence with respect to the latter. Accordingly, their present close juxtaposition is strongly suggestive of important post-2.6 Ga fault movements. Such faults, if they exist, are cryptic and unadorned by cataclasites, breccias, gouge or quartz veining (see also Hanmer et al. 1991, 1992). Because they would follow the sinuous trace of the pre-existing mylonites, they are as unlikely as the mylonites to have accommodated important strike-slip displacements. However, the geometry of their surface trace would not have impeded them from accommodating significant dip-slip movement.

A similar problem is present in the East Athabasca mylonite zone where the high pressure (10+ kbars, M.L. Williams, pers. comm.) strike-lineated mylonites are bounded by cordierite and sillimanite-andalusite bearing wallrocks. In all of these cases, we are obliged to invoke post-ca. 2.6 Ga discrete, presumably dip-slip, brittle faulting, to account for the present structural and metamorphic configurations.

ACKNOWLEDGMENTS

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Cu-Ni-Mo-PGE-Au-rich mafic inclusions in the Fort Smith (Konth) granite, Northwest Territories

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Charbonneau, B.W. and Harris, D.C., 1993: Cu-Ni-Mo-PGE-Au-rich mafic inclusions in the Fort Smith (Konth) granite, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 53-59.

Abstract: Unusual mafic inclusions have been found in the Konth granite, a component of the Taltson magmatic zone extending south from the Great Slave Lake Shear Zone. These inclusions, up to several metres long, have high Cu (1826 ppm), Ni (1321 ppm), Mo (310 ppm), PGEs (Pt+Pd) (0.12 ppm) and a trace of Au (40 ppb). Major elements are Fe_{tot} (30%), Al_2O_3 (19%), SiO_2 (39%) and up to a few per cent each of CaO, Na_2O, K_2O, MgO. The inclusions are composed mainly of nickel-rich chlorite (chamosite) with some albite, quartz, biotite, muscovite, epidote, potassium feldspar, zircon, rutile, titanite and a range of sulphide minerals (pyrite, chalcopyrite, molybdenite and a trace of bismuth, bismuthinite, violarite and galena).

The mineralization in these inclusions is similar to that found at Rutledge Lake, 75 km to the northeast, where it has attracted considerable economic interest. The extent of the mafic inclusion swarm is not known.

Résumé : Des inclusions mafiques inhabituelles ont été découvertes dans le granite de Konth, composante de la zone magmatique de Taltson qui se prolonge au sud de la zone de cisaillement de Great Slave. Ces inclusions, qui peuvent atteindre plusieurs mètres de longueur, ont des taux élevés de Cu (1 826 ppm), de Ni (1 321 ppm), de Mo (310 ppm), d'ÉGP (Pt+Pd) (0,12 ppm) et contiennent quelques traces d'Au (40 ppb). Les principaux éléments sont Fe_{tot} (30 %), Al_2O_3 (19 %), SiO_2 (39 %) et les éléments accessoires, représentés par quelques points de pourcentage, sont CaO, Na_2O, K_2O, MgO. Les inclusions se composent principalement de chlorite nickélifère (chamosite) accompagnée d'un peu d'albite, de quartz, de biotite, de muscovite, d'épidote, de feldspath potassique, de zircon, de rutile, de titanite et de divers minéraux sulfurés (pyrite, chalcopyrite, molybdénite et des traces de bismuth, bismuthinite, violarite et galène).

Dans ces inclusions, la minéralisation est semblable à celle observée à Rutledge Lake, à 75 km au nord-est, qui a éveillé un intérêt considérable en raison de son potentiel exploitable. L'étendue de l'essaim d'inclusions mafiques est inconnue.

INTRODUCTION

During the course of field investigation of airborne gamma spectrometric surveys (Geol. Surv. Can., 1972) (Fig. 1) of the Fort Smith (Konth) megacrystic granite south of the East Arm of Great Slave Lake (Charbonneau, 1980, 1991) numerous inclusions were observed in the granite. Although interesting these inclusions were not focussed on in detail with the bulk of the field effort being put into evaluation of the airborne anomaly patterns pertaining to the major rock units.

Most of the inclusions within the Konth granite were readily recognizable as representative of lithologies mapped in the area i.e. Tsu Lake gneiss, Slave granite, etc. (Bostock, 1982). However there are two types of unusual and interesting inclusions:

1. Those with a uranium-thorium-rare earth-rich, biotite-ilmenite-magnetite-monazite-zircon-apatite assemblage were reported on principally by Burwash and Cape (1981), as well as Charbonneau (1980, 1991). These inclusions are of uncertain origin (heavy mineral placers,

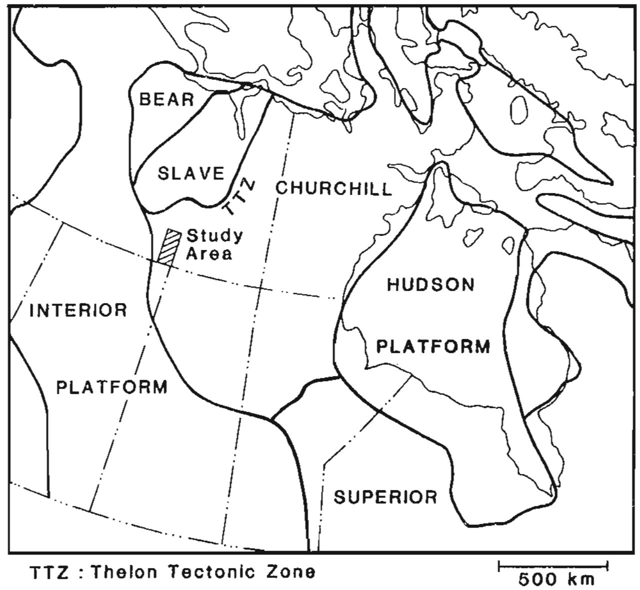

Figure 1. Location of study area.

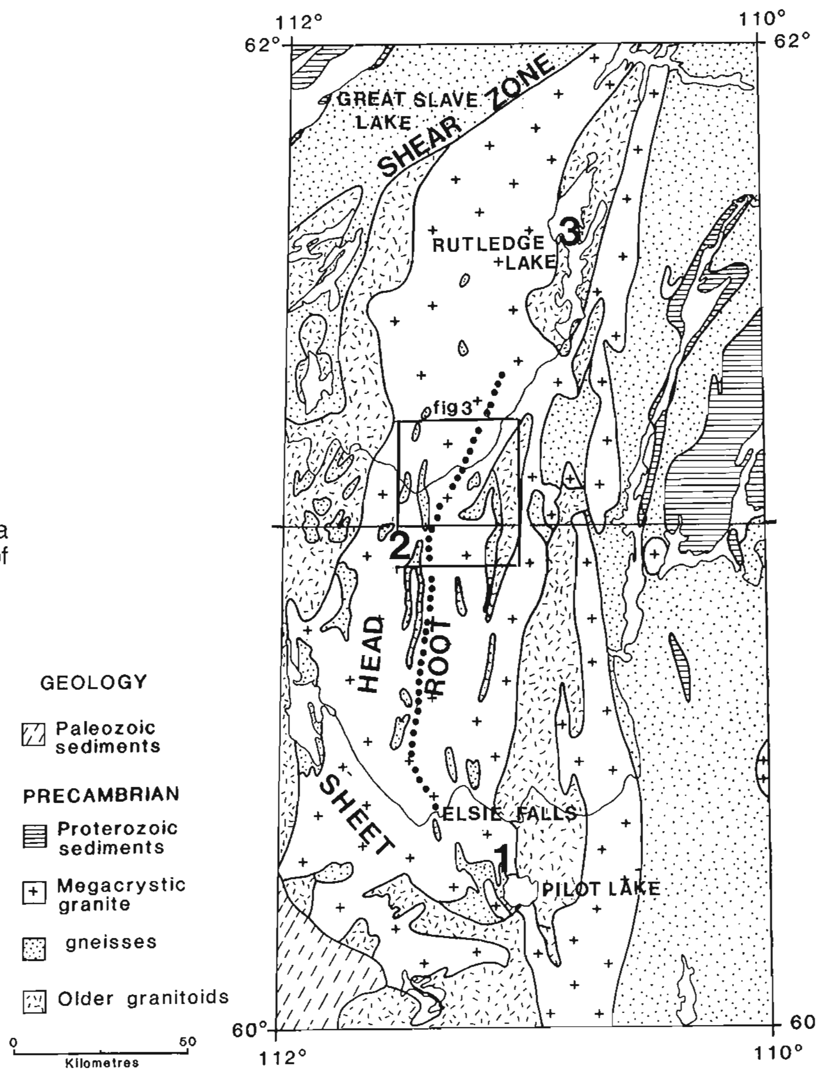

Figure 2.

Generalized geology, detailed study area (black box), mineralized localities, trend of magnetic anomaly (dots).

restites or dykes) and are best manifested around Pilot Lake although they are found elsewhere in the Konth granite.

2. The Cu-Ni-Mo-PGE-Au-rich mafic inclusions northeast of Nelson Lake, composed mainly of chlorite with some albite, quartz, biotite, muscovite, epidote, potassium feldspar, zircon, rutile, titanite and a wide range of sulphide minerals.

Representative localities for these two types of inclusion are shown in Figure 2. Type 1 is best represented at Pilot Lake (locality 1)(Burwash and Cape, 1981). Type 2, which is the focus of this paper, occur northeast of Nelson Lake (locality 2), at 60°59'N, 111°30'W. At locality 3, near Rutledge Lake, the lithology is clearly of mafic or ultramafic igneous origin, comprising pods of a Cu-Ni-rich mafic lithology in gneiss (Culshaw, 1984). The Cu-Ni metal content of locality 2 inclusions, discovered in recent analyses, indicates the unusual nature of these inclusions and their possible economic significance.

GEOLOGICAL SETTING OF THE NELSON LAKE MAFIC INCLUSIONS

The megacrystic Konth granite has been subdivided into three structural components by Bostock (1981). These subdivisions are the "Root Zone", the "Collapsed Head" and the "Sheeted Area". These domains, indicated in Figure 2, have distinct gamma ray spectrometric signatures. The "Root Zone" is highly thoriferous whereas the Head and Sheeted subdivisions typically have higher U/Th ratio (Charbonneau, 1991).

The location of the mafic inclusions described in this paper is near the boundary between the "Root" and "Head". This boundary is accompanied by a zone of aeromagnetic anomalies that can be traced southward to Elsie Falls on Taltson River and has a total length of some 100 km as shown in Figure 2 by the line of dots which follows the trend of the anomaly. Figure 3 (Geol. Surv. Can., 1964) shows the aeromagnetic pattern over part of this belt including the area where the mafic inclusions described in this paper were discovered. The airborne geophysical patterns (magnetic, gamma ray spectrometric and VLF-EM) are more sharply defined in Geological Survey of Canada (1990). The mafic inclusions have distinctly anomalous magnetic susceptibility and it is suggested that the long linear magnetic anomaly described above may in part be due to enclaves of this material which would then be preserved as screens between the "Root" and "Head" of the granite.

The mafic inclusions in the field are illustrated in Figure 4. They range up to several metres in size. The mafic inclusions are highly anomalous with respect to average values of some major and trace elements in ultrabasic rocks (Turekian and Wedepohl, 1961).The inclusions contain approximately 30% Fe_{tot}, 40% SiO_2 and nearly 20% Al_2O_3 and minor amounts (up to a few per cent each) of CaO, MgO, Na_2O and K_2O. The trace elements are most interesting, with levels of 1826 ppm Cu, 1321 ppm Ni, 310 ppm Mo, 0.12 ppm PGE (Pt + Pd) and a trace of Au (40 ppb) found in the inclusion shown in Figure 4. The Pt:Pd ratio is slightly greater than one.

MINERALOGY

Major minerals present in the mafic inclusions are chlorite (var chamosite) with some albite, quartz, biotite, muscovite, epidote, potassium feldspar, zircon, rutile, titanite and a wide range of sulphide minerals. The sulphide minerals are pyrite, chalcopyrite, molybdenite and a trace of bismuth, bismuthinite, violarite and galena. The total sulphide content is typically about 3 or 4 per cent in the enclaves observed. The nickel is principally in the chlorite with trace amounts in violarite. Average nickel values in the chlorite are about 0.5%. The PGE values are distinctly anomalous (L. Hulbert, pers. comm.) although not as high as values reported for some ultramafic bodies in the Northwest Territories where samples containing tens of thousands of ppb have been reported at the Muskox intrusion (Hulbert, 1992).

Typical mineral textures are shown in Figure 5. Molybdenite and chalcopyrite are usually not in contact. The chalcopyrite usually cuts the pyrite. In one section violarite was seen to transect chalcopyrite.

Geochemical analyses of the granite and a representative mafic inclusion are shown in Table 1. An electron microprobe analysis of a nickel-rich, high iron chlorite is given in Table 2. The granite is low in all of the economically significant elements present in the mineralized mafic inclusions. The major elements and the nickel values are consistent with nickel-rich, high iron chlorite. The combination of high nickel, copper, molybdenum and PGEs with gold is unusual.

ORIGIN

Didier (1973) has outlined a variety of theories that can explain the origin of enclaves in granite.

1. Complete transformation of sedimentary or metamorphic xenoliths caught up by the granite magma during emplacement.
2. External igneous origin.
3. Restites i.e. refractory residues of anatexis.
4. Segregations of basic minerals that appeared in the magma at the beginning of crystallization.
5. Congenetic igneous origin involving the appearance of a basic magma in the genesis of certain granites.

Theory 1 or 2, above, would appear to be the most likely explanation for these enclaves. There are many possibilities for the origin of this mineralization. The theories for origin must explain the unusual geochemical signature, i.e. high Cu-Ni-Mo-PGEs + Au and high Al_2O_3 as well as low Cr, MgO.

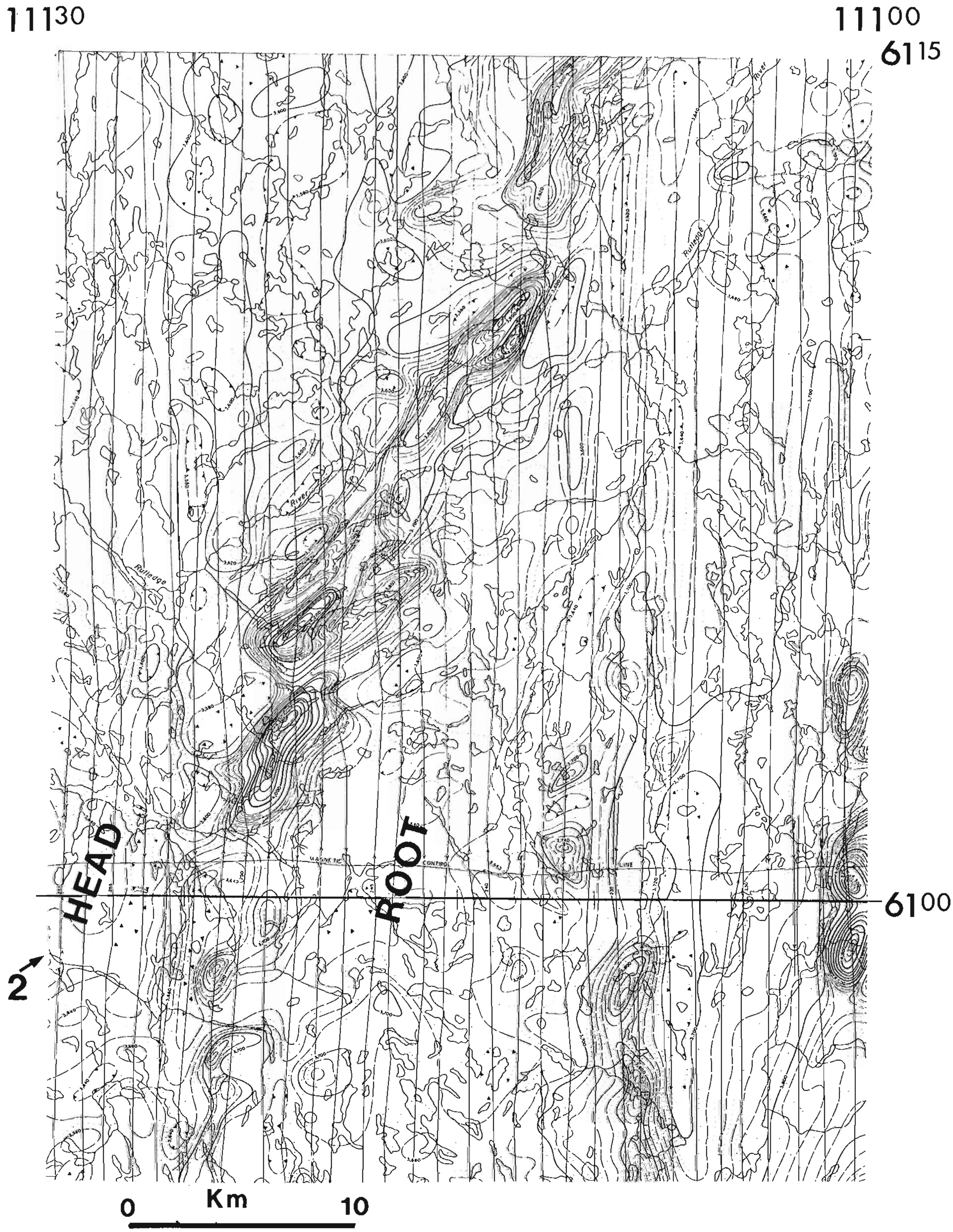

Figure 3. Standard aeromagnetic map showing magnetic anomaly which lies on boundary of "Root" and "Head" granite subdivisions.

The genetic nature of the mineralization may be 1) ultramafic, 2) metasedimentary (iron-formation, laterite), or 3) volcanogenic massive sulphide. The nearest "similar" material is found at Rutledge Lake some 75 km away. The Rutledge Lake mineralization has been described by several workers and groups who have offered a range of geological models for the Rutledge Lake area (Culshaw, 1984; Bostock, 1987, 1988; Enexco Int. Ltd., 1983, 1986, 1987). Culshaw mapped the area as dominantly paragneiss with mafic bands and pods (some volcanics?) and granitic rocks. It is within these rocks that the mineralization occurs. The mineralization is variable but is consistently enriched in Cu-Ni-Mo-PGEs (Pt + Pd) Au as indicated by the Enexco reports. The mineralization is in part related to ultramafic pods, but is also in sulphide-rich layers which may be brecciated and usually occur near bands of pyroxene-bearing or chloritic mafic rocks. These bands resemble material found in the inclusions reported on in this study. The chloritic layers have been postulated as volcanic related (Enexco, 1983) or metasedimentary (Enexco, 1987). Because of the presence of the mafic bands and pods it is likely that the mineralization is related to these rocks which occur in limited amounts at the surface but may be more extensive at depth

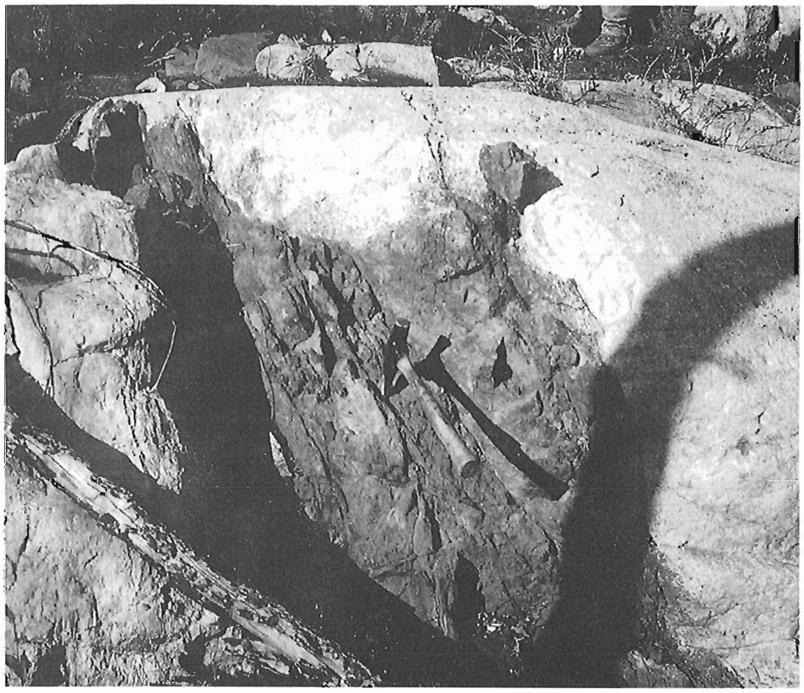

Figure 4.

Field photograph of typical mafic inclusion.

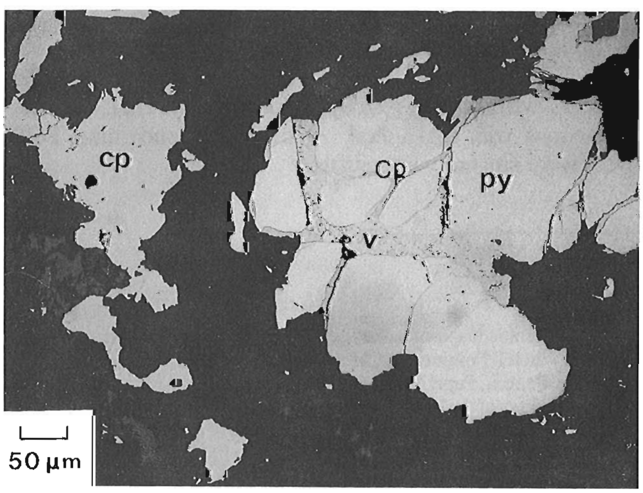

Figure 5a. Photomicrograph of anhedral pyrite (PY) enclosed and partly replaced by chalcopyrite (CP) with a younger violarite veinlet (V) oil immersion.

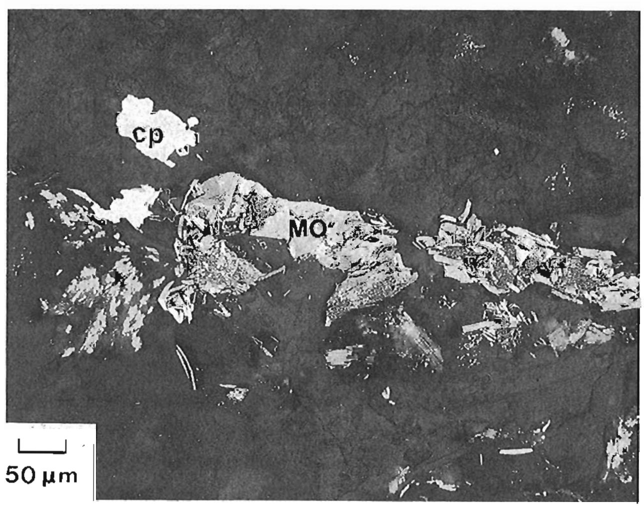

Figure 5b. Photomicrograph of molybdenite flakes (Mo) with chalcopyrite (CP); oil immersion.

Table 1. Geochemical analyses of enclosing granite and mafic inclusion

CZ-19-1-85 (Granite)								
SiO_2	TiO_2	Al_2O_3	Fe_{tot}	MgO	CaO	Na_2O	K_2O	
69.70	0.12	14.90	1.10	0.23	0.64	3.44	7.05	
Cu^1	Ni^1	Mo^2	$Pt + Pd^3$	Au^2	Cr^1	Ce^4	U^2	Th^2
4	<2	<2	<.01	<.005	7	<10	16.0	24.0
CZ-19-10-85 (Mafic inclusion)								
SiO_2	TiO_2	Al_2O_3	Fe_{tot}	MgO	CaO	Na_2O	K_2O	
39.30	1.17	19.00	30.70	2.93	0.59	0.10	2.12	
Cu^1	Ni^1	Mo^2	$Pt + Pd^3$	Au^2	Cr^1	Ce^4	U^2	Th^2
1826	1321	310	0.12	<040	233	348	3.4	4.8

Major elements (%) C.D. Plasma
Trace elements (ppm) AA 1
NA 2
Fire Assay 3
XRF 4

Table 2. Electron microprobe analyses of chlorite (var) chamosite

Chlorite (Ave 5) Var Chamosite	
MgO	4.0%
FeO	38.3%
SiO_2	23.4%
MnO	0.2%
Al_2O_3	19.5%
NiO	0.56%

Enexco (1987). Most of the mineralization at Rutledge Lake is pyrrhotite-rich but the chloritic inclusions observed in this study contain little if any pyrrhotite.

An ultramafic origin for the chloritic inclusions themselves would seem unlikely solely on geochemical grounds, i.e. low MgO, Cr and high Al_2O_3 and Mo, but the common spatial relation between ultramafics and anomalous mineralization seen at Rutledge Lake suggests an association of these phenomena. Iron-formation would seem a possibility but the high Al_2O_3 would not be characteristic of grunerite although it would be characteristic of chlorite (chamosite) iron-formation. An unusual lateritic type of sediment might be a speculative possibility. The multi-metal signature could be consistent with volcanogenic sulphide mineralization and the chloritic material could be an alteration.

ECONOMIC CONSIDERATIONS

This work has draws attention to an isolated occurrence of Cu-Ni-Mo-PGE-Au mineralization similar to that which has attracted substantial exploration effort at Rutledge Lake. The extent of the mafic inclusion swarm carrying this mineralization is not known and the surrounding area has only received reconnaissance investigation. Nevertheless the inclusions occur along an extensive zone of aeromagnetic anomalies which follows a structure of regional scale. These relationships merit further investigation and may prove economically rewarding.

ACKNOWLEDGEMENTS

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Geological Survey of Canada Project 880029

Regional geology in the Winter Lake – Lac de Gras area, central Slave Province, District of Mackenzie, Northwest Territories[1]

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Thompson, P.H., Ross, D., Froese, E., Kerswill, J.A., and Peshko, M., 1993: Regional geology in the Winter Lake – Lac de Gras area, central Slave Province, District of Mackenzie, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 61-70.

Abstract: The Winter Lake-Contwoyto Lake and Courageous Lake-Lac de Gras supracrustal domains with high potential for gold and base metals are connected by metasedimentary and metavolcanic rocks of the Yellowknife Supergroup in an area previously mapped as granitoid rocks.

A sequence dominated by metavolcanic rocks commonly occurs between a heterogeneous metagranitoid complex in the central part of the area and metasedimentary rocks and younger, homogeneous granitoids to the north and east. Metamorphosed mafic dykes cross-cutting structures in the complex are similar to dykes that intruded Yellowknife Supergroup prior to low-P/high-T regional metamorphism and three phases of deformation. Isograds cross older granitoids and are cut by late granites.

Extensive potential basement rimmed by volcanic rocks, low-P/high-T metamorphism, and abundant granitoids are consistent with shortening and overthickening of previously-thinned sialic crust near the margin of a major supracrustal basin.

Résumé : Les domaines supracrustaux de Winter Lake-Contwoyto Lake et de Courageous Lake-Lac de Gras, qui prhsentent tous deux un potentiel élevé du point de vue des minéralisations en or et en métaux communs, sont reliés par des roches métasedimentaires et métavolcaniques du Supergroupe de Yellowknife dans une région dont les roches avaient auparavant été identifiées comme étant de nature granitoïde.

Une séquence principalement composée de roches métavolcaniques se manifeste souvent entre un complexe métagranitoïde hétérogène dans la partie centrale de la région, et des roches métasédimentaires et des granitoïdes homogènes plus jeunes, au nord et à l'est. Des dykes mafiques métamorphisés recoupant des structures du complexe sont semblables à des dykes qui ont pénétré le Supergroupe de Yellowknife avant le métamorphisme régional à faible P et T élevée et trois phases de déformation. Les isogrades traversent des granitoïdes plus anciens et sont recoupés par des granites tardifs.

La présence possible d'un vaste socle bordé de roches volcaniques, l'influence du métamorphisme à faible P et T élevée et ondance des granitoïdes, sont des phénomènes compatibles avec le raccourcissement et l'épaississement considérable d'une croûte sialique auparavant amincie, près de la marge d'un grand bassin supracrustal.

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INTRODUCTION

The Winter Lake – Lac de Gras map area (east half NTS 86A, west half 76D) contains extensive areas of granitoid rocks of unknown origin and parts of three major supracrustal domains with high gold and base metal potential: Winter Lake – Contwoyto Lake, Courageous Lake – Lac de Gras, and Yellowknife – Beaulieu River (Fig. 1). The purpose of the current project is to provide new geological knowledge, link recent work north and south of the area, and contribute

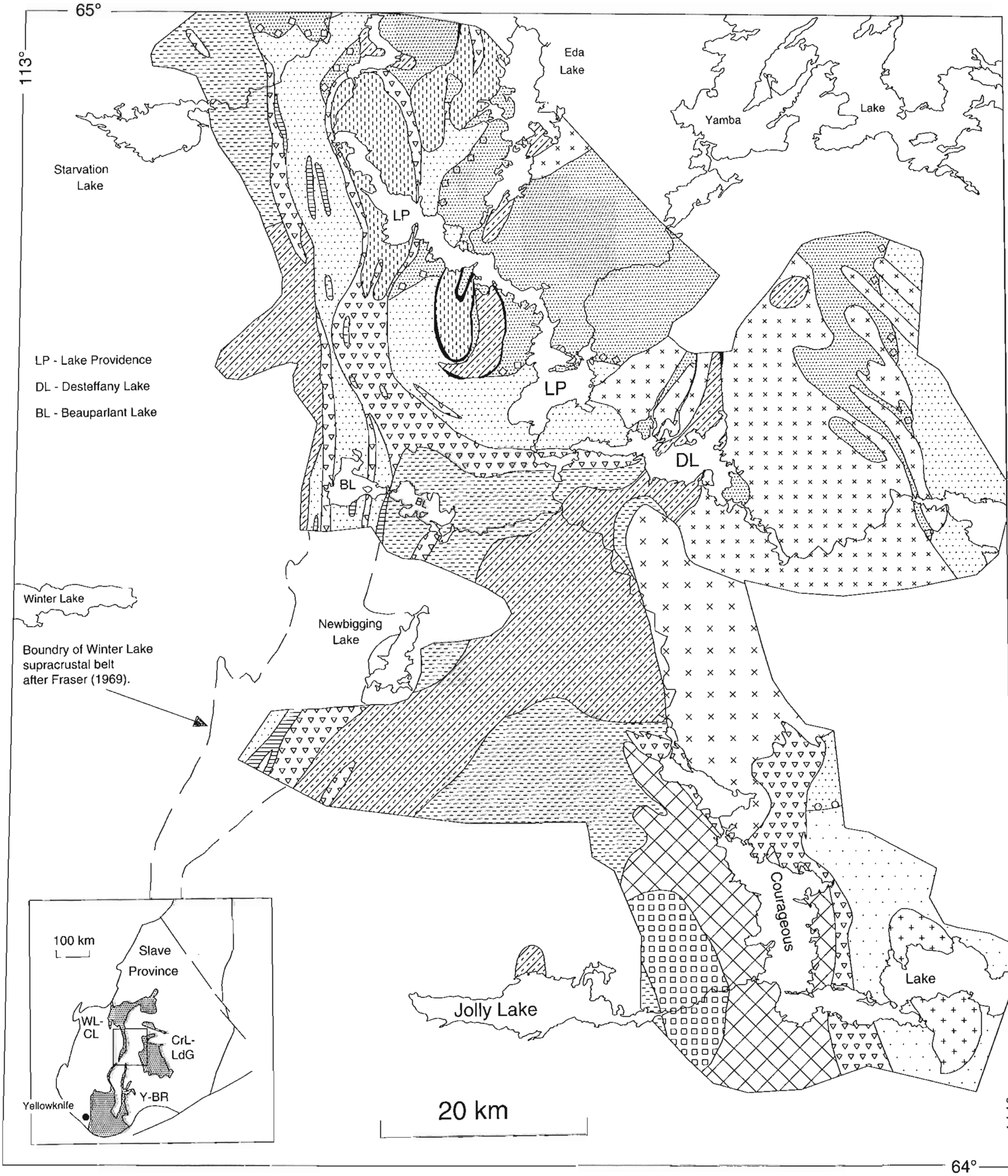

Figure 1. Geological map of the portion of the Winter Lake – Lac de Gras area (86A east half, 76D west half) mapped during the 1991 and 1992 field seasons. Supracrustal domains: WL-CL – Winter Lake – Contwoyto Lake; CL-LdG – Courageous Lake – Lac de Gras; Y-BR – Yellowknife-Beaulieu River.

to a broader regional synthesis. Of particular interest is the evolution of supracrustal and granitoid domains and their relation to each other. Mapping an area 100 km x 110 km at a scale of 1:250 000 will elucidate aspects of stratigraphy, deformation, magmatism, and metamorphism, and provide a framework for more detailed studies. Mineral deposits will be assessed within this larger geological context.

In addition to regional mapping and metallogenic investigations, the project involves special studies in several fields. R. Dilabio (Terrain Sciences) sampled till for kimberlite indicator minerals. Samples were collected by M. Villeneuve (Continental Geoscience Division (CGD)) for geochronology and by mapping crews for gravity/ aeromagnetic studies (P. McGrath, CGD), measurement of thermal properties (A. Judge, Terrain Sciences; T. Lewis, Pacific Geoscience Centre), and study of diabase dykes (A. LeCheminant, K.Buchan, W. Baragar, CGD). The utility of maps integrating LANDSAT TM, aeromagnetic, and reconnaissance geological data as a planning tool was established (A. Rencz, Mineral Resources Division; D. Baril CGD). A. Lalonde (University of Ottawa) began petrological studies of granitoid rocks. Field work was completed for two B.Sc. theses and a fourth year research project.

The geological framework of the east half of the area (Lac de Gras) was outlined by Folinsbee (1949) and that of the west half (Winter Lake) by Fraser (1969). Only the Courageous Lake volcanic belt (Fig. 2) has been mapped in detail (Moore, 1956; Dillon-Leitch, 1981, 1984). Areas adjacent to the north were mapped on a regional scale by Bostock (1980) and King et al. (1991), and to the south by Moore et al. (1951) and Henderson (1944). Major structures and rock units described by Stubley (1991) and Stubley and Irwin (1992) continue across the southern border of the area.

The Slave Structural Province is a segment of the northwestern Canadian Shield that has remained relatively stable since a major orogenic event 2.6 Ga ago (Henderson, 1981; Hoffman, 1989; Thompson, 1989a; Padgham, 1990). It is distinguished from granite-greenstone belts of similar age by a high proportion of greywacke-mudstone relative to mafic and felsic volcanic rocks in the supracrustal component

YELLOWKNIFE SUPERGROUP

Sedimentary Rocks

Phyllite, Slate, Greywacke (chlorite, biotite)

knotted schist isograd

Knotted schist/psammite (cordierite, andalusite, sillimanite)

migmatite isograd
> 5-10 % leucosome

Layered migmatite, diatexite (cordierite, sillimanite, +/- garnet)

Conglomerate

Volcanic and Related Rocks (VARR)

Amphibolite, gabbro, mafic/felsic volcanics, volcaniclastics, carbonate schist, calcsilicate and quartzo-feldspathic gneisses quartz arenite, ultramafic rocks, and gossan.
< 200m thick = solid black

GRANITOID ROCKS

Younger Granitoid Rocks

White Granite

Pink Biotite Granite

Globular Granite

Courageous Lake Pluton

Older Granitoid Rocks

Courageous Lake Metaplutonic Suite

Foliated meta-granite/tonalite

Foliated hornblende meta-granite

Granitoid Migmatite/Gneiss

Legend to Figure 1

(Yellowknife Supergroup, Henderson, 1970). Older granitoid rocks in various parts of the Slave Province represent basement to the supracrustal rocks. Syn- to post-tectonic intrusions underlie extensive areas. Crustal shortening and overthickening was accompanied by low-pressure/high temperature regional metamorphism.

ROCK UNITS

Older granitoid rocks

Metamorphosed and deformed rocks of the granitoid complex in the central and western parts of the area, together with similar rocks near Lake Providence, have been subdivided into granitoid migmatite/gneiss, two foliated metagranitoid units and the Courageous Lake metaplutonic suite (Fig. 1).

Granitoid migmatite/gneiss

This major unit is characterized by textural heterogeneity at outcrop scale. Pinkish-grey- to white-weathering rocks vary from foliated granite/granodiorite to migmatitic foliated granitoid to biotite-quartzofeldspathic gneiss. Even in the relatively homogeneous, foliated granitoid phases, ghostly relicts of gneissic or migmatitic textures are common. Quartz, K-feldspar, plagioclase, and biotite are essential minerals; the occurrence of hornblende is sporadic. Tonalitic compositions are also present in subordinate amounts. Inclusions or boudins of mafic gneiss and metagabbro are less common in the occurrences near Lake Providence than they are to the south and west. Younger-looking, massive, fine grained pink granite dykes and veins constitute 40-50 per cent of the outcrop at some localities. Granitoid migmatite in this unit is distinguished from metasedimentary migmatite by: 1) absence of aluminosilicates and cordierite; 2) relatively low mica content; 3) pinkish-grey rather than rusty-grey weathered surface; and 4) granitoid aspect of protolith (mesosome), its textural and mineralogical homogeneity. Deformed metamorphosed mafic dykes cut across the gneissosity at several localities south of Desteffany Lake and south of Newbigging Lake, indicating the unit has been metamorphosed twice. Gneissosity in the complex is perpendicular to that in overlying supracrustal rocks near the southeast end of Lake Providence. Foliations and gneissosity are locally oblique to the Winter Lake supracrustal belt (Thompson, 1992; Hrabi et al, 1993), but bend around and converge with the trends in the Desteffany Lake volcanic belt as the units are folded into a north-northeast orientation north of the lake.

Foliated metagranite/tonalite

A large part of the granitoid complex in the central and western part of the map area is underlain by variably-foliated, biotite-bearing rocks ranging from granite to tonalite. Hornblende is present in some of the granodioritic to tonalitic compositions. Although the unit is relatively homogeneous, in some areas, nebulitic traces of layering and migmatization make definition of the contact with granitoid migmatite/gneiss difficult. It is distinguished from massive younger homogeneous granitoids by the presence of a fabric and evidence of a pervasive reduction of grain size associated with metamorphism and deformation. Inclusions, if present, are metagabbro or quartzofeldspathic gneiss similar to those in adjacent granitoid migmatite/gneiss. South of Desteffany Lake, some of the metagabbro bodies in metagranite/tonalite are boudinaged dykes. West of the Winter Lake supracrustal belt and north of Beauparlant Lake, the hills of amphibolite/metagabbro alternate with low areas of foliated granitoid. There are no metasedimentary inclusions. Commonly, veins or dykes of massive, pink-weathering fine grained granite or pegmatite crosscut the foliation in the metamorphosed plutonic rocks. Elsewhere, the crosscutting dykes are strongly deformed. Foliation trends are consistent with the regional structure.

Foliated hornblende metagranite

West and north of Lake Providence, pink-grey weathering, fine- to medium-grained, foliated, hornblende-biotite metagranite forms the core of several elongated domal structures. Pink feldspar, plagioclase, quartz, biotite, and hornblende are essential minerals. Scattered occurrences of fluorite are a distinctive feature. Foliations, and locally, lineations, are defined by preferred orientation of lenticular aggregates of quartz and feldspar (rimmed by mafic minerals) and by clots of mafic minerals. The medium- to coarse-grained lenticular or augen structure is preserved in spite of extensive grain size reduction that has transformed augen of quartz and feldspar into fine grained aggregates. Foliations in the granitoid are concordant with planar structures in thin zones of volcanic and related rocks that, along most of the contact, separate the granitoid from sillimanite schist and metasedimentary migmatite. At several localities (e.g., near Lake Providence), a narrow zone (tens of metres) of granitoid migmatite/gneiss, locally containing boudins of metagabbro, separates the granite from adjacent volcanic and related rocks.

Courageous Lake metaplutonic suite

West of Courageous Lake, variably-foliated biotite/hornblende tonalite to granodiorite is the main element in outcrops containing up to six different lithologies. Granitoid gneiss and amphibolite derived from volcaniclastic and massive mafic volcanic rocks and gabbro, together with intrusive phases ranging from gabbro to aplite and pink feldspar pegmatite, are present. On the east shore of Courageous Lake, metamorphosed gabbro dykes which intrude foliated metatonalite and metagranite have been mapped as continuous with mafic sills within the overlying volcanic sequence (Moore, 1956; Dillon-Leitch, 1981).

Yellowknife Supergroup

Volcanic and related rocks (VARR)

This unit commonly separates older granitoids from overlying greywacke/mudstone (Fig. 1). It includes volcanic rocks mapped by previous workers and referred to by Thompson (1992) as the Courageous Lake, Desteffany Lake and "Noname" lake (south of Newbigging Lake) volcanic belts. As Hrabi et al. (1993) referred to "Noname" lake as Sherpa lake, Sherpa lake volcanic belt is used here (Fig. 2).

Pillowed and massive mafic lavas, gabbro, and volcaniclastic rocks are predominant. Moore (1956) and Dillon-Leitch (1981) related metagabbro dykes and sills to mafic volcanism. White-weathering, felsic volcanic rocks, both massive and with clastic textures, are prominent in the Courageous Lake volcanic belt and smaller volumes occur north and west of Desteffany Lake. North of Beauparlant Lake (Fig. 2), they are up to several hundred metres thick. Intrusive quartz-feldspar porphyries have been interpreted by previous workers as being coeval with felsic volcanism.

Other distinctive lithologies, present in subordinate amounts, are characteristic of the Volcanic and related rock unit (VARR). Fine- to very fine-grained, leucocratic, biotite-quartz-feldspar gneisses (layers <1 cm) are associated with the volcanic rocks west and north of Desteffany Lake.

Thin-layered amphibolite and calc-silicate gneiss in the northern half of the area may be high grade (clinopyroxene-garnet) equivalents of carbonate-chlorite-muscovite schist and micaceous marble that occur within the Courageous Lake volcanic belt.

Gossans in sulphide-bearing, felsic and amphibolitic rocks occur sporadically throughout the volcanic and related rocks unit. Usually, they are associated with siliceous and/or carbonate-bearing rocks.

Massive and thin-layered ultramafic rocks, locally exhibiting a heterogeneous fragmental texture, are interlayered with mafic volcanic rocks along the western edge of the Winter Lake supracrustal belt and in the northern part of the Courageous Lake volcanic belt. Another body occurs between the granitoid and metavolcanic rocks near Desteffany Lake.

Quartz-arenite separates mafic volcanics from sillimanite schist or migmatite on the west shore of Lake Providence and, 13 km to the southwest, near the western contact of the Winter Lake belt. Similar rocks occur near the contact with granitoid rocks south and west of Desteffany Lake. Thompson (1992) described quartz-arenite (see also Hrabi et al., 1993) interlayered with metavolcanic rocks and metagabbro, southwest of Newbigging Lake.

At several localities near Lake Providence, K feldspar-quartz-biotite pegmatite has been intruded at a high angle to the layering and foliation. At Desteffany Lake, garnet-bearing aplite and pegmatite both cut across, and intrude parallel to layering.

Phyllite/schist/migmatite

Metamorphosed greywacke-mudstone typical of the Yellowknife Supergroup (Henderson, 1970) is an important element in the western, northern, and eastern parts of the map area. Current mapping has documented the transition from grey-green phyllite/metagreywacke through grey-brown cordierite/andalusite schist/psammite and reddish-brown sillimanite schist/psammite to layered migmatite and diatexite. Bedding is well-preserved at low and medium grade and it persists well into the migmatite zone. Leucosome first appears in pelitic layers of an outcrop. As the proportion of leucosome increases, more psammitic layers are affected, producing, ultimately, a rock made up of remnants of psammitic layers and foliated lenses of biotite-sillimanite melanosome in a relatively homogeneous matrix of quartz, feldspar, cordierite and biotite. Cordierite is prominent in leucosome throughout the migmatite zone. Garnet is present only locally at high and medium grade. In migmatitic rocks, sillimanite is not as common as it is in the highest grade part of the knotted schist zone. Typically, leucosome and melanosome are complexly folded. Compositional layering commonly parallels that in adjacent Volcanic and related rocks unit and, locally, the two rock units are interlayered on the scale of metres.

Conglomerate

Lenses of polymictic conglomerate several hundred metres thick and up to 10 km long (Fraser, 1969; Rice et al., 1990) occur in the Winter Lake supracrustal belt. Variably-deformed clasts of leucocratic granitoid and mafic to felsic volcanic rocks occur in a mica-rich, quartzofeldspathic matrix. Conglomerate occurs between volcanic rocks and pelitic schist/psammite, between metasediments and foliated meta-granite/tonalite, and interlayered with sillimanite-biotite schist and psammite. Foliations are parallel and metamorphic grade is the same in the conglomerate and adjacent rock units. Primary contact relations were not observed.

Younger granitoid rocks

All but one of four massive homogeneous granitoids intrude metasedimentary rocks that occur north and east of the variably-oriented distribution of Volcanic and related rocks unit. Contacts cross-cutting metamorphic grade and structures in the supracrustal rocks indicate that younger plutons intruded relatively late in the tectonic evolution. There is structural evidence that deformation continued after intrusion.

White granite

North and east of Desteffany Lake, there is a mainly white-weathering, massive, medium- to coarse-grained, biotite granite to granodiorite. Characteristic subhedral to euhedral feldspar (0.5-1.5 cm) is evenly distributed within a finer and more even-grained matrix of quartz, feldspar and fresh black

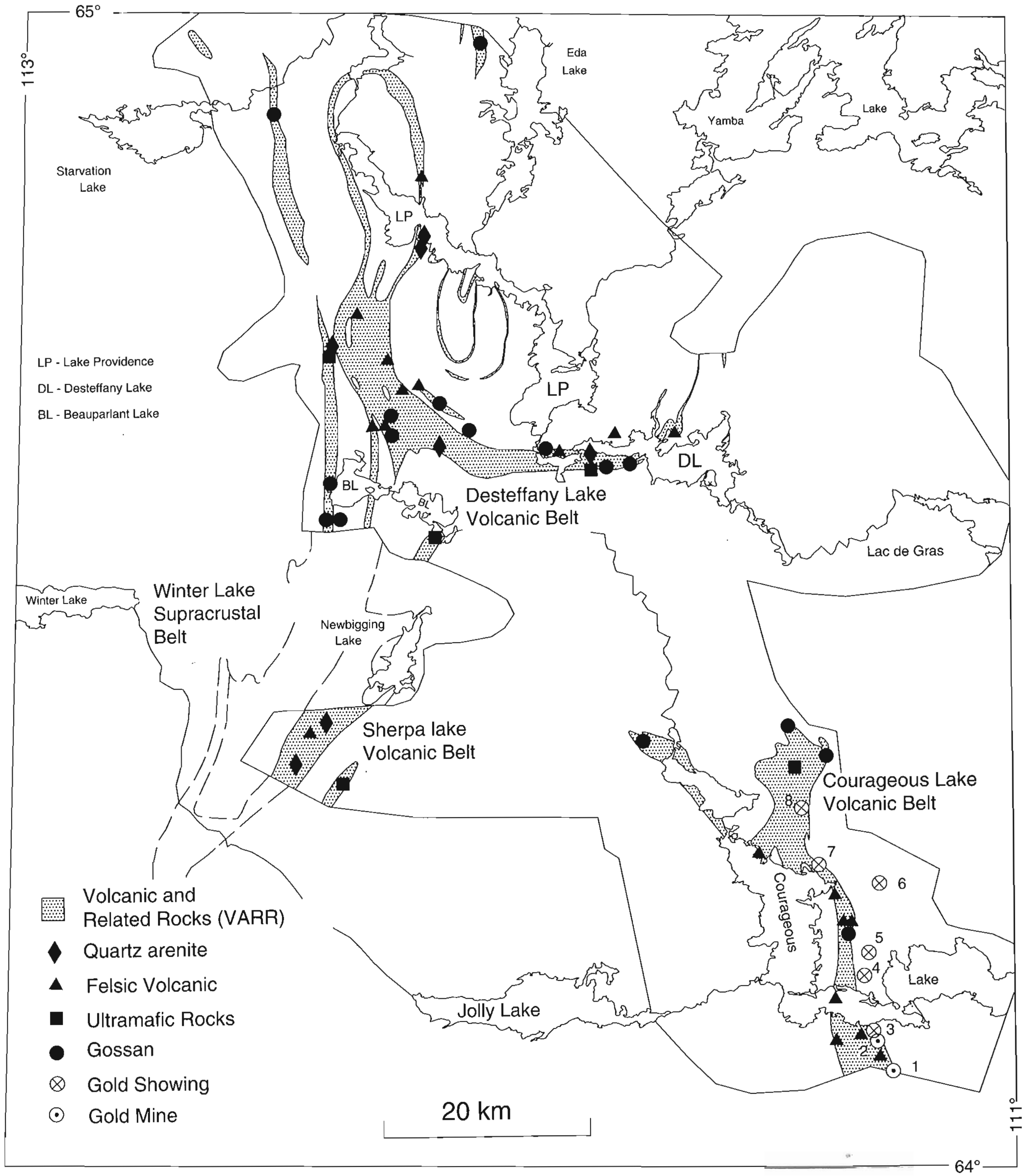

Figure 2. Distribution of volcanic and related rocks (VARR) unit. Gold mines and occurrences: 1 – Salmita; 2 – Red 24 open pit; 3 – Tundra joint venture; 4 – 1725; 5 – Payne; 6 – Kennedy; 7 – TMK; 8 – Jax Lake.

biotite. Discrimination of two feldspars is difficult. Inclusions of metasedimentary migmatite, sillimanite schist/ psammite, granitoid migmatite, amphibolite gneiss and metagabbro are present. The nebulitic aspect of some large metasedimentary and granitoid migmatite inclusions suggests they may be samples of the source area of the granite. At outcrop scale, dykes of white granite and pegmatite have been boudinaged and folded without developing a foliation. Evidently, deformation outlasted intrusion without developing a fabric in the granite.

Pink biotite granite

North of Courageous Lake is a massive, medium- to coarse-grainedd biotite granite. K-feldspar and plagioclase are easily distinguished. Biotite is the main mafic mineral but the proportion is generally less than 10-15 per cent. Garnet (<2 mm) is present in several localities. Inclusions are absent and pegmatite is rare. This granite is discordant with respect to structures to the north and west.

Globular granite

West of Courageous Lake, a massive granite contains distinctive, anhedral to subhedral grains (0.5-1.0 cm) of quartz. On the weathered surface, yellowish-grey quartz stands out against chalky-white feldspar. On a fresh surface, small grains of black biotite appear to be sprinkled through the rock. Inclusions are absent. Here and there, the granitoid is cut by straight-sided dykes (one to several tens of centimetres thick) of pink aplite.

Courageous Lake pluton

This is a zoned pluton surrounded by greenschist facies metasedimentary rocks. A strong aeromagnetic signature and absence of a mappable contact aureole (Folinsbee, 1949) characterize this unit. North of the lake, diorite and quartz diorite mantle a granitic core. Distinctive mafic and ultramafic inclusions in the mafic zone range in size from a few millimetres up to several centimetres. South of the lake, several hundred metres of coarse grained mafic syenite separate diorite/monzodiorite from the country rocks. Inclusions of deformed metasedimentary rocks occur in the syenite. Contacts are concordant with bedding and cleavage in country rock.

STRUCTURAL GEOLOGY

In the north half of the area (Fig. 1, 3), older granitoids, volcanic and related rocks, and metasedimentary rocks, together with traces of the principal foliation outline a complex fold interference pattern that can be related to three main phases of ductile deformation. Foliations and gneissosity commonly dip steeply. Stretching lineations, where present, plunge down the dip of planar structures. The youngest phase of deformation, D_3, produced large-scale, relatively open, steeply-plunging folds with steep, northerly-trending axial surfaces spaced tens of kilometres apart. D_3 deforms isoclinal folds (D_2) of lithological contacts and the concordant principal foliation and gneissosity. Also steeply dipping, D_2 axial surface traces are between 5 and 20 km apart. Present geometry suggests doubly-plunging synforms

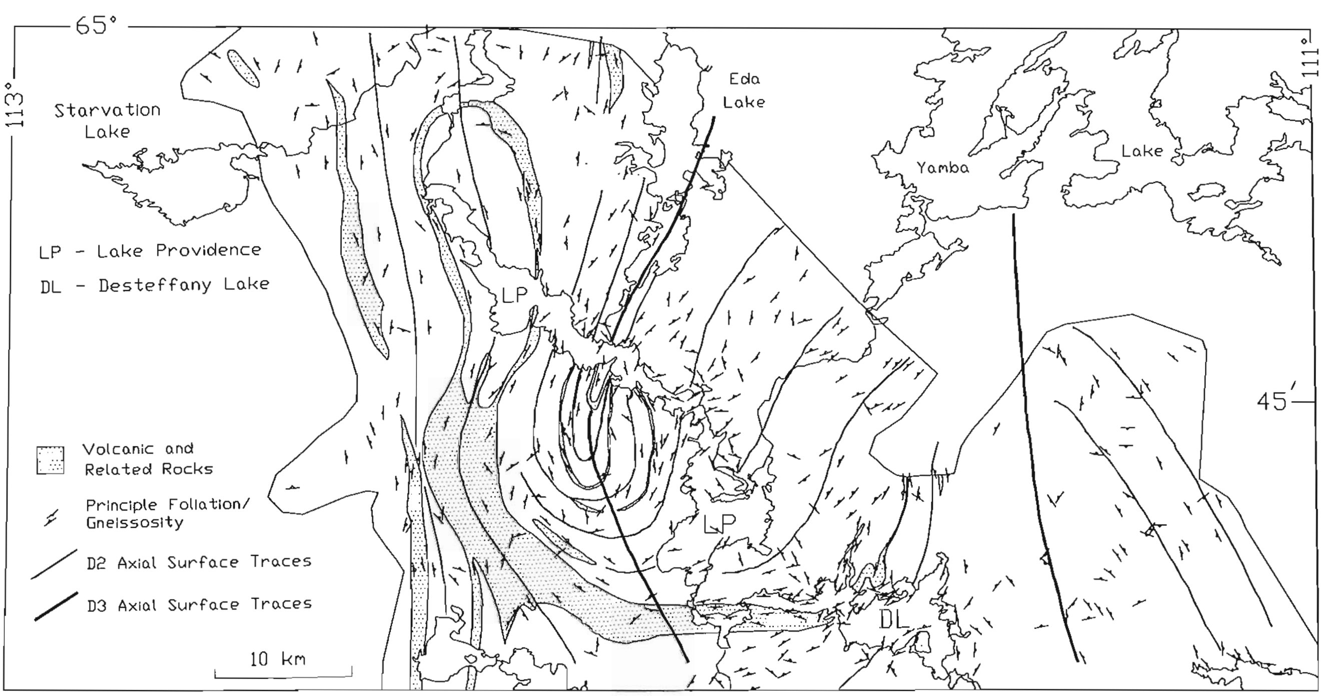

Figure 3. Structural geometry in the north half of the Winter Lake - Lac de Gras map area. To simplify the diagram, VARR is the only rock unit represented.

and antiforms were produced during D_2, possibly reflecting variable strain along the axial surface. Pre-D_3 orientation of D_2 axial surfaces is inferred from enveloping surfaces to be easterly to southeasterly. Unfolding of D_2 structures leaves the principal foliation/gneissosity conformable with a lithological sequence comprising older granitoids, volcanic and related rocks, and metasedimentary rocks. Map scale folds related to D_1 were not recognized in the north half of the area. However, Dillon-Leitch (1981) described structures in low grade supracrustal rocks in the Courageous Lake area that correspond to such an early phase of deformation. Although thrust-related intercalation of granitoid migmatite/gneiss and supracrustal rocks (Thompson, 1992) may have occurred on a relatively small scale during D_1 or D_2, the regional distribution of lithological units can be described in terms of a fold interference pattern.

Foliations in the metagranite/tonalite unit and gneissic layering and isoclinal folds in granitoid migmatite/gneiss are cut by mafic dykes that are themselves deformed and metamorphosed. Similar dykes and sills intruded supracrustal rocks prior to deformation and metamorphism. It is inferred, therefore, that a major deformational event occurred in older granitoid rocks before deposition of the Yellowknife Supergroup. Such an event is consistent with the angular discordance between planar structures in granitoid migmatite/gneiss south of Newbigging Lake and east of Lake Providence (Thompson, 1992) and that of the adjacent supracrustal rocks. A similar history is described by James and Mortenson (in press) in the basement rocks of the Sleepy Dragon Complex.

METAMORPHISM

The isograds in pelitic rocks (Fig. 1) are common throughout the Slave Province; they are based on the first appearance of cordierite and/or andalusite and leucosome. In mafic rocks, cordierite/andalusite isograd corresponds approximately to the transition from chlorite-actinolite schist to amphibolite. Mineral assemblages indicate low pressures (3-5 kbars) and medium to high temperatures (400-700°C). The indicated uplift and erosion of 10-15 km since peak conditions were attained, implies that the crust was overthickened to that extent at that time (45-50 km total thickness). Such high temperatures at low pressures imply high geothermal gradients with the potential for extensive melting in the overthickened crust.

Isograds cut across D_2 structures and are weakly folded by D_3 (compare Fig. 1 and 3). They intersect the boundaries between older granitoid units and overlying supracrustal rocks. Andalusite and cordierite porphyroblasts contain an early fabric (S_1). The porphyroblasts themselves commonly define a preferred orientation parallel to the main schistosity (S_2). In the migmatite zone, complex folding of leucosomes and foliated sillimanite-biotite aggregates occurred during D_2 and D_3. Metamorphic grade is inferred to have peaked during D_2 or soon after D_2 ended, and the absence of retrogression indicates that it remained high during D_3.

ECONOMIC GEOLOGY

The Courageous Lake supracrustal domain contains three past gold producers (Tundra, Salmita, and Red 24), significant prospects including the Main and Carbonate zones on the Tundra joint venture property, numerous other gold occurrences, as well as several polymetallic volcanogenic massive sulphide occurrences (Ransom and Robb, 1986; Comeau, 1990; Hearn, 1990).

The mines and most gold occurrences straddle or are near the contact between the volcanic-dominated supracrustal succession and phyllite/metagreywacke (Fig. 2). Gold mineralization occurs in and/or adjacent to veins and/or fractures in a variety of variably deformed and altered rocks. However, the polymetallic massive sulphide occurrences are associated with felsic volcanic rocks lower in the succession.

Investigation of a gossan discovered in 1991 at the sedimentary/volcanic contact 5 km west of Desteffany Lake confirmed the presence of sulphide iron-formation (hornblende- and pyrrhotite-rich). Sulphide-poor silicate iron-formation (garnet- and grunerite-rich), possibly more extensive than the sulphide-rich variety, is also present. The sulphide iron-formation is interlayered with sulphidic, tuffaceous to cherty, felsic volcanic rocks and several stratiform accumulations of massive pyrrhotite and pyrite. A single sample of sulphide iron-formation collected in 1991 assayed 220 ppb gold.

Iron formation was also identified near the contact between dominantly mafic volcanic rocks and phyllite/metagreywacke about 1 km north of the TMK gold occurrence as well as in trenches on the Jax Lake property (Fig. 2). Visits to pyritic chert locations (Dillon-Leitch, 1981, 1984) along the volcanic/sedimentary contact revealed the presence of additional iron-formation just south of the map area (6 km south-southeast of Nodinka Narrows, 75M/14). The presence of iron-formation in the contact zone further supports Dillon-Leitch's (1981) proposal that exhalative activity was associated with the close of volcanism.

Work to date (Kerswill, 1992) suggests a complex history of metal concentration in the Courageous Lake area. Although late shear zones appear to have controlled gold distribution at the Salmita and Tundra mines and several other occurrences near the volcanic-sedimentary contact, synvolcanic structures and processes may have contributed to gold concentration on the Tundra joint venture property and at Red 24. Syn- volcanic/synsedimentary processes controlled base metal concentration in the massive sulphide deposits and may have contributed to sulphide concentration in the pyritic, tuffaceous to cherty rocks and iron-formation.

DISCUSSION

Supracrustal rocks

Extensions of the Desteffany Lake volcanic belt to the northwest (40 km) and northeast (15 km) provide new targets for mineral exploration and link the economically important

Winter Lake – Contwoyto Lake and Courageous Lake - Lac de Gras supracrustal domains. There is no apparent change in the metamorphosed greywacke-mudstone sequence from one domain to the other. The lack of iron-formation in the metasedimentary sequence suggests a correlation with the Itchen Formation (Bostock, 1980; King et al., 1991). However, iron-formation does occur at the contact between volcanic and sedimentary rocks west of Desteffany Lake and near Jax Lake (Fig. 2). Mafic volcanic rocks and gabbro are similar throughout the area but, in the Winter Lake supracrustal belt and Desteffany Lake volcanic belt, felsic volcanics are present in much smaller quantities than in the Courageous Lake volcanic belt. Except for one conglomerate occurrence south of Courageous Lake (Dillon-Leitch, 1981) volcanic/granitoid pebble conglomerate and quartz arenite are restricted to the northern and western parts of the map area. Calc-silicate gneiss at Desteffany Lake and Lake Providence is probably the higher grade equivalent of the carbonate schist and phyllite in the Courageous Lake belt. Rare orthoamphibole schist and ultramafic rocks occur in all three belts. In the map area, there is no need to separate the quartz-rich rocks that occur above, below, and within the volcanic-dominated supracrustal sequence from the Yellowknife Supergroup as proposed by Padgham (1990).

Granitoid rocks

Subdivision of Fraser's (1969) and Folinsbee's (1949) granitoid unit 4 into eight units (Fig. 1) resolved major discrepancies along the boundary between the Winter Lake and Lac de Gras map areas. Younger granitoid units are discordant to other rock units, structure and regional metamorphism.

Four lines of evidence are consistent with the older granitoid rocks being basement to the supracrustal rocks. First, there is a stratigraphy, comprising older granitoids separated from overlying sedimentary rocks by a thin volcanic sequence, that can be traced along strike and around major folds for tens of kilometres. Although volcanic rocks are not continuous everywhere in the northern part of the Winter Lake supracrustal belt, the complete sequence is present in the southern part (Hrabi et al., 1993). In general, at map scale, the bodies of older granitoid are concordant with stratigraphy and structure in overlying supracrustal sequences; but, at several localities, there is a marked discordance between planar structures in the two units. Isograds defined in metasedimentary rocks cut across the granitoid complex indicating it is older than metamorphism and not a source of heat. Finally, mafic dykes that have intruded foliated metagranitoid rocks west of Courageous Lake (Moore, 1956) and granitoid migmatite/gneiss south of Desteffany and Newbigging lakes, are themselves deformed and metamorphosed. Similar dykes and sills intruded the volcanic belts and associated sedimentary rocks before they were deformed and metamorphosed. The granitoid complex is interpreted as part of the pre-Yellowknife Supergroup crust documented for instance in the Sleepy Dragon Complex 150 km to the southwest (James and Mortensen, in press) and in the Point Lake area 75 km to the northwest (Easton, 1985).

Regional metamorphism

Folinsbee (1949), Moore (1956), and Dillon-Leitch (1981) suggested that granitoid rocks were a likely source of heat for metamorphism. Intersection of isograds with the contact between older granitoids and supracrustal rocks, absence of a mappable contact aureole around the Courageous Lake pluton, and intrusion of White granite across the migmatite isograd (Fig. 1) indicate that granitoid rocks are either older or younger than the metamorphism and therefore not the main source of heat.

Thompson (1989b) suggested that the anomalously high geothermal gradients associated with low-pressure/high-temperature regional metamporphism are inherited in part from a period of crustal thinning and sedimentation immediately preceding the orogeny (crustal shortening and overthickening) that ends with exhumation of metamorphic rocks to the earth's surface. If the model is correct, in addition to providing estimates of peak geothermal gradients and of the magnitude of overthickening during metamorphism, metamorphic mineral assemblages in the Winter Lake – Lac de Gras are evidence of a period of ensialic basin formation.

Tectonic setting

Northward and eastward, there is a major lithological transition from an area dominated by migmatitic, gneissic, and foliated granitoid rocks that are considered to be basement to supracrustal rocks, to one where massive, discordant granitic plutons and metasedimentary rocks are predominant. Two mica granites occur only northeast of 1992 mapping (Folinsbee, 1949). In general, the two main rock packages are separated by thin belts of volcanic and related rocks. The asymmetrical distribution of rock units and of their relative ages are thought to indicate the margin of a deformed supracrustal basin. Further regional mapping and related studies will determine whether an exhumed basin margin is present in the map area and provide important constraints on tectonic models of the Slave Structural Province.

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Geological Survey of Canada Project 870008PT

Geology of the Winter Lake supracrustal belt, central Slave Province, District of Mackenzie, Northwest Territories[1]

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Abstract: Supracrustal rocks in the Winter Lake belt have been tentatively subdivided into three lithological associations. Stratigraphically at the base are felsic volcanic rocks, orthoquartzite, oligomictic conglomerate, and basaltic komatiite that are overlain by mafic volcanic rocks and greywacke-mudstone turbidites, typical of the Yellowknife Supergroup rocks elsewhere. The youngest association consists of compositionally immature crossbedded sandstones and polymictic conglomerate, similar to Jackson Lake conglomerates in the Slave Province and thought to be analogous to Timiskaming-type sedimentary rocks of the Superior Province. A tonalite gneiss complex with abundant mafic dykes flanks the western margin of the belt, but significant strain at the contact between this complex and the supracrustal sequence was not observed. In contrast, the eastern margin of the belt is a zone of mylonitization, interrupted by numerous late plutons, which brings the supracrustal rocks into contact with a granite-gneiss complex to the east.

Résumé : On a provisoirement subdivisé les roches supracrustales de la zone de Winter Lake en trois associations lithologiques. Sont situées stratigraphiquement à la base de cette zone, des roches volcaniques felsiques, un orthoquartzite, un conglomérat oligomictique et une komatiite basaltique qui sont recouverts par des roches volcaniques mafiques et des turbidites à grauwacke et mudstone, typiques des roches du Supergroupe de Yellowknife présentes aileurs. L'association la plus récente est constituée de grès à stratification obliique, de composition immature, et de conglomérat polymictique semblable aux conglomérats de Jackson Lake dans la Province des Esclaves, et peut-être analogue aux roches sédimentaires du type de Timiskaming dans la Province du lac Supérieur. Un complexe à gneiss tonalitique contenant d'abondants dykes mafiques borde la marge occidentale de la zone, mais l'on n'observe aucune déformation significative au contact entre ce complexe et la séquence supracrustale. Par contre, la marge orientale de la zone est une zone de mylonitisation, interrompue par de nombreux plutons tardifs; cette zone sert à mettre les roches supracrustales en contact avec un complexe granitique et gneissique à l'est.

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INTRODUCTION

The purpose of the Winter Lake Mineral Initiatives Project is an updating of the present geological understanding, including better defining of the geological context of mineral showings, of the Winter Lake supracrustal belt by 1:50 000 scale mapping. The belt is located in the east half of the Winter Lake sheet (NTS 86A), central Slave Province (Fig. 1). Project funding is provided predominantly by the Canada-Northwest Territories Mineral Initiatives (Project C4. 120). M. Villeneuve (GSC) is undertaking geochronological support studies in the project area.

Previous work in the Winter Lake area includes 1:250 000 scale regional mapping by Fraser (1969), detailed mapping of the sedimentary rocks southwest of Newbigging Lake (Rice et al., 1990), and a brief reconnaissance by Thompson (1992). Mapping this summer concentrated on the south half of the supracrustal belt, in order to evaluate the overall synformal geometry of this part of the belt described by Fraser (1969) and to examine the extension of an approximately north-striking fault system mapped by Stubley (a,b) in the Carp Lakes sheet (NTS 85P) to the south. A simplified map of the results is presented in Figure 2.

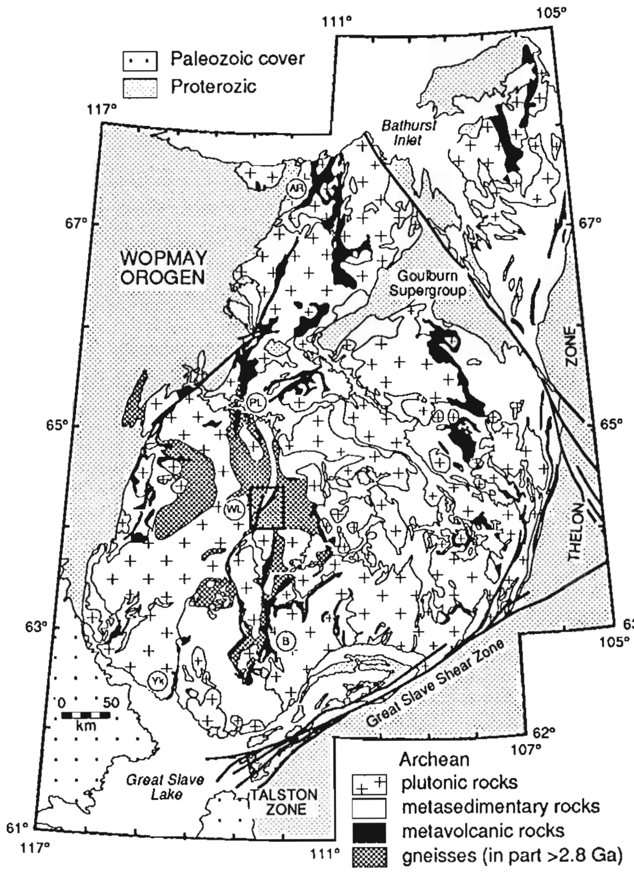

Figure 1. Location of the Winter Lake supracrustal belt in the Slave Province (outlined). WL - Winter Lake, AR - Anialik River, B - Beaulieu River, PL - Point Lake, Yk - Yellowknife. Modified after Hoffman (1989).

This preliminary report concentrates on: 1) the supracrustal rocks, particularly a unit of pillowed basaltic komatiite, discovered in the map area this field season; 2) the contact relationships between the supracrustal rocks and the bounding granitoid and gneissic rocks; and 3) a preliminary interpretation of the deformation history. All rock types have been metamorphosed under greenschist or amphibolite facies conditions, but the prefix "meta" has been omitted in this paper.

SUPRACRUSTAL ROCKS

The Winter Lake supracrustal belt contains a great variety of rock types rarely found together within other individual supracrustal belts of the Slave Province (Fig. 2). These are tentatively subdivided into three lithological associations: 1) an association of felsic volcanic rocks, orthoquartzite, oligomictic conglomerate, basaltic komatiite found at the base of the stratigraphic sequence; 2) mafic volcanic rocks and greywacke-mudstone turbidites; and 3) a group of compositionally immature crossbedded sandstones and polymictic conglomerate.

Sequences such as the Dwyer Lake Formation north of Yellowknife, and the Beniah Formation in the Beaulieu River supracrustal belt contain some of the rock types found in the first association (Helmstaedt and Padgham, 1986; Covello et al., 1988). However, to date no supracrustal belt has been found to contain all of these rock types together. The second association is common to all supracrustal belts in the Slave Province and correspond to the rocks previously described as the Yellowknife Supergroup (Henderson, 1970). Lithological units similar to the third association include the Jackson Lake Formation in the Yellowknife supracrustal belt (Helmstaedt and Padgham, 1986), the Beaulieu Rapids Formation in the Beaulieu River supracrustal belt (Roscoe et al., 1989; Rice et al., 1990), and the String Lake conglomerate in the Anialik supracrustal belt (C. Relf, pers. comm., 1992) and are considered to postdate the Yellowknife Supergroup.

Pre-Yellowknife Supergroup

Felsic volcanic rocks (unit 1)

Felsic volcanic rocks, found south and west of Newbigging Lake (Fig. 2), comprise the stratigraphically lowest unit informally referred to as the Newbigging formation. The unit is found predominantly on the east side of the belt, with minor occurrences mapped on the southwestern margin of the belt. The unit has a maximum thickness of about 3800 m which includes approximately 1000 m of mafic intrusive rocks along the southwest shore of Newbigging Lake .

The felsic volcanic unit is composed of white- to pink-weathering, massive, extremely siliceous rhyolites and white- to buff-weathering, tuff and lapilli tuff. Bedding in the volcaniclastic rocks is generally poorly defined, but where observed, beds ranged in thickness from 3-50 cm. Some horizons in the tuffs have well-preserved fiammé (Fig. 3). Facing indicators are scarce, consisting of graded bedding in the volcaniclastic rocks, and indicate that tops are to the west

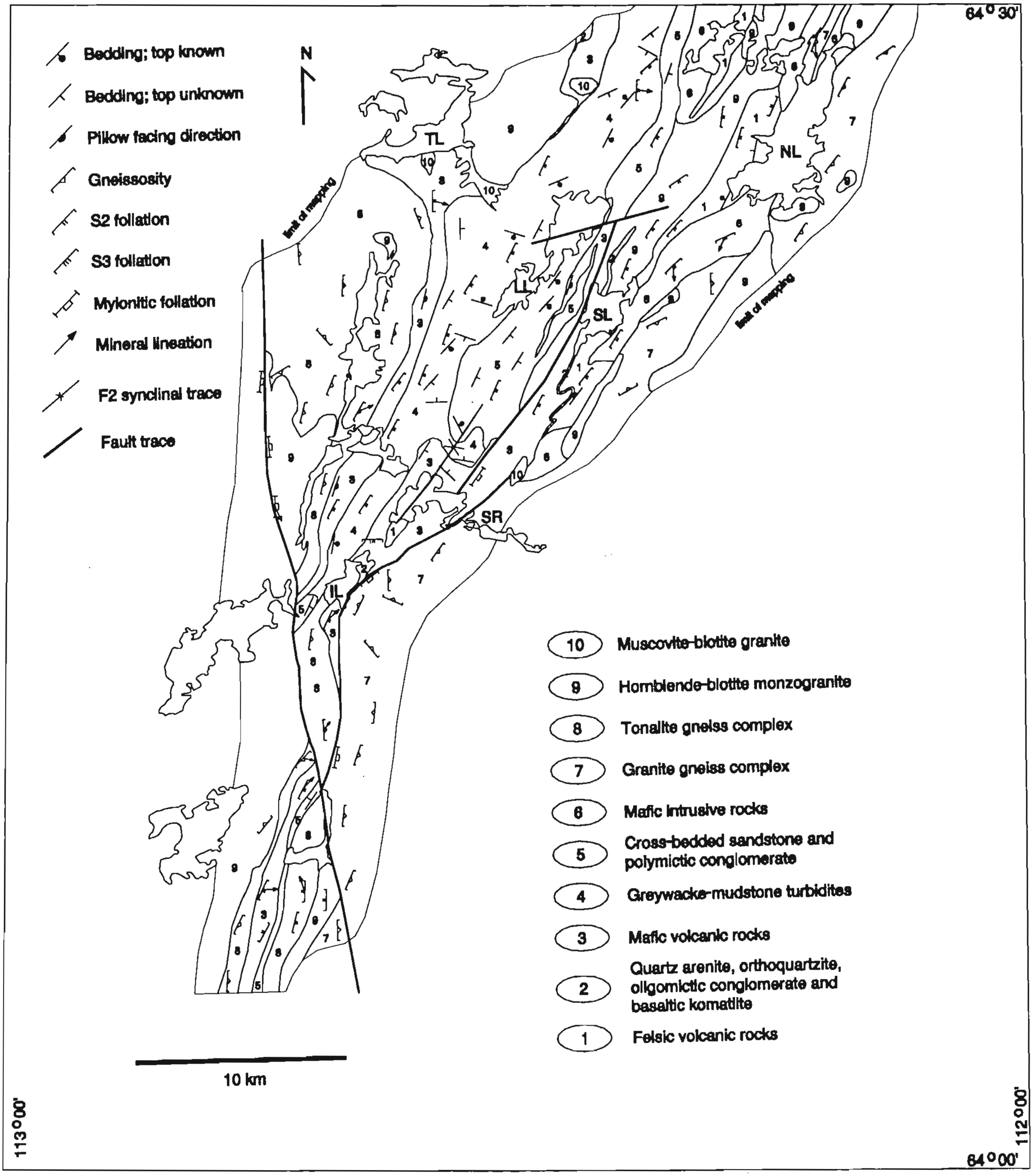

Figure 2. Simplified geological map of the Winter Lake supracrustal belt. IL - Izabeau lake, Left Lung lake, NL - Newbigging Lake, SL - Sherpa lake, SR - Snare River, TL - Terminus lake.

in this unit. Except for parasitic folds south of Sherpa lake, the sequence on the east margin thus appears to be homoclinal. The felsic volcanic rocks have been intruded by a large number of gabbro and fewer ultramafic dykes and sills, that represent up to 50% of the exposed surface area of the unit.

The contact with the adjacent gneissic rocks on the eastern margin is not well exposed. However, at the contact along the shore of Newbigging Lake, the felsic rocks show a progressively stronger foliation approaching gneissic layering and tight isoclinal folds over a distance of 50 m. This is interpreted to be a structural contact.

Quartz arenite, orthoquartzite and oligomictic conglomerate (unit 2)

Thompson (1992) mentioned two sequences of quartz-rich and pelitic sedimentary rocks in the vicinity of Sherpa lake and drew attention to their similarity with the Beniah Formation (Roscoe et al., 1989). These sedimentary rocks are found both stratigraphically above the felsic rocks and interbedded with them. White-weathering orthoquartzites are present north and south of Sherpa lake. The orthoquartzites have very thin, grey laminae, thought to represent mud layers (Fig. 4) which define subtle crossbeds in the orthoquartzite, indicating tops to the west on the east side of the belt. In the core of a fold south of the Snare River valley, rusty brown-weathering, medium grained quartz arenites and oligomictic conglomerate are found in the same relative stratigraphic position as the orthoquartzite and are interpreted to be part of the same lithological association. The oligomictic conglomerate is clast-supported with well-rounded clasts and moderate sorting. The clasts are dominantly felsic in composition and appear to include both clasts of massive felsic volcanic rock and orthoquartzite.

Minor units of the quartz-rich sedimentary rocks are interlayered with the felsic volcanic rocks but the base of the main part of this unit was not observed in contact with the

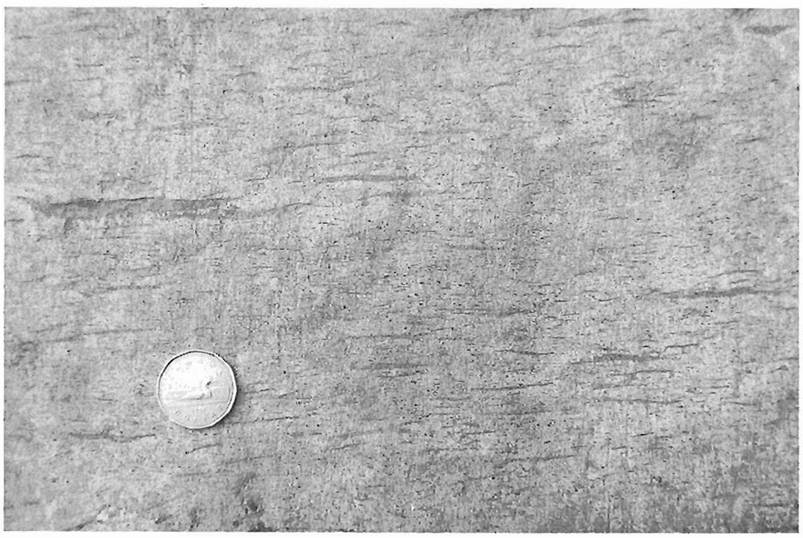

Figure 3.

Flattened fiammé in felsic volcani- clastic rock south of Newbigging Lake.

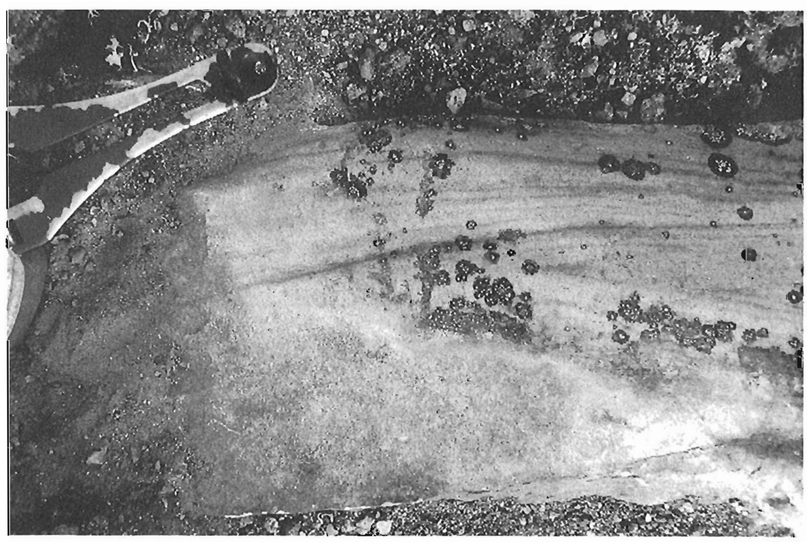

Figure 4.

Crossbedding in orthoquartzite defined by thin lamellae of fine grey sediment. Outcrop is located on the north shore of Sherpa lake.

felsic volcanic rocks. On the map-scale, however, the unit is stratigraphically above the felsic volcanic rocks and in general are structurally conformable with them.

Basaltic komatiite (unit 2)

One of the most intriguing units mapped this summer is a thin but extensive unit of basaltic komatiite, representing the first recorded occurrence of such rocks in the Slave Province. This unit, estimated to be approximately 20 m thick can be traced along most of the east margin of the belt and into the hinge zone of the synform south of the Snare River valley (Fig. 2). It has also been found locally along the west margin of the belt. The basaltic komatiite occurs stratigraphically above the felsic unit along the east margin and stratigraphically above a sequence of quartz-rich sandstone and conglomerate in the hinge zone of the synform. The contact is conformable and bedding is subparallel in both the felsic volcanic rocks and the basaltic komatiite. As no evidence for tectonic reactivation has been found, contact relationships are interpreted as conformable or disconformable.

The basaltic komatiite weathers light- to medium-green and is medium green on a fresh surface. The rock is very soft, and rock powder has a slippery-talc consistency. Most of the unit consists of pillowed rocks (Fig. 5), but other related rock types are tremolite-talc schist, and locally reworked epiclastic sedimentary rocks.

Three samples from the cores of pillows in the basaltic komatiite were analyzed by XRF both at Queen's University and at XRAL Laboratories, Toronto. Results are shown in Table 1 and plotted as a Jensen Cation Plot in Figure 6. The 12-18 wt.% MgO and the position of the samples on the Jensen Plot substantiate the field classification of these rocks as basaltic komatiite.

Yellowknife Supergroup

Mafic volcanic rocks (unit 3)

Mafic volcanic rocks are found on both east and west limbs of the supracrustal belt and in the synformal hinge zone. The unit ranges in thickness at its widest point from approximately 1100 m on the west limb to a minimum of approximately 1500 m on the east limb. The mafic volcanic rocks are green- to dark green-weathering and dark green on fresh surfaces. The unit is composed predominantly of pillowed flows but also contains massive flows. Coarse grained, massive units are interpreted as subvolcanic gabbro intrusions.

Figure 5.

Pillowed basaltic komatiite flow facing northwest. Outcrop is HLB-92-B0207, located southeast of Sherpa lake and is the location of one of the whole-rock geochemistry samples.

Table 1. Table of whole-rock geochemistry results of pillowed basaltic komatiite samples. Sample sets -1 and -2 analysed by XRF at Department of Geological Sciences, Queen's University. Sample set -3 analysed by XRF at XRAL Laboratories, Toronto.

SAMPLE	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	LOI	TOTAL
HLB-92-B0207A-1	53.27	0.65	11.92	11.02	0.19	12.43	8.43	2.20	0.07	0.06	1.35	101.59
HLB-92-B0207A-2	53.31	0.65	11.89	11.02	0.19	12.40	8.43	2.20	0.07	0.06	1.35	101.57
HLB-92-B0207A-3	52.40	0.66	11.70	10.90	0.21	12.10	8.59	2.23	0.08	0.06	1.15	100.08
Standard Deviation	0.51	0.01	0.12	0.07	0.01	0.18	0.09	0.02	0.01	0.00	0.12	
HLB-92-J0684A-1	50.83	0.58	11.46	12.25	0.19	17.70	5.62	0.57	0.38	0.05	2.33	101.96
HLB-92-J0684A-2	50.85	0.57	11.43	12.28	0.19	17.58	5.60	0.53	0.38	0.05	2.33	101.79
HLB-92-J0684A-3	50.00	0.57	11.20	11.60	0.19	17.10	5.65	0.65	0.37	0.06	2.40	99.79
Standard Deviation	0.49	0.01	0.14	0.38	0.00	0.32	0.03	0.06	0.01	0.01	0.04	
HLB-92-P0356B-1	49.60	0.63	10.78	11.41	0.18	15.90	8.56	1.15	0.24	0.06	2.73	101.24
HLB-92-P0356B-2	49.60	0.62	10.76	11.40	0.18	15.90	8.56	1.25	0.24	0.06	2.73	101.30
HLB-92-P0356B-3	48.90	0.64	10.60	11.30	0.18	15.70	8.57	1.31	0.24	0.06	2.40	99.90
Standard Deviation	0.40	0.01	0.10	0.06	0.00	0.12	0.01	0.08	0.00	0.00	0.19	

The lower contact of the mafic volcanic rocks with other supracrustal rocks is conformable. At the western side of the belt and in the core of the synform, the mafic flows overlie either the komatiite or thin units of silicate or oxide facies ironstone, or are in contact with foliated to gneissic tonalite which contain a high number of crosscutting mafic dykes. On the east side, mafic volcanic rocks overlie the komatiite unit (Fig. 2), or are in contact with the granite-gneiss complex or younger felsic intrusions.

Greywacke-mudstone (unit 4)

The centre of the Winter Lake belt is occupied by greywacke and mudstone turbidites without obvious ironstone or volcanic intercalations, that are typical of the Yellowknife Supergroup turbidites. Metamorphic mineral assemblages indicate the turbidites are in the biotite, cordierite, and sillimanite zones, with the metamorphic grade increasing from the centre of the belt toward the west margin and to the south of the Snare River valley. Greywacke beds range in thickness from 10 to 100 cm and the 3 to 5 cm mudstone beds are commonly defined by layers of resistant andalusite. Graded bedding is the only way-up indicator in these rocks. The maximum outcrop width of the unit is 5400 m but poorly defined fold repetition does not allow an accurate estimate of the true unit thickness.

Crossbedded sandstone and polymictic conglomerate (unit 5)

Fraser (1969) showed an area of conglomerate west of Sherpa lake with some interlayering of greywacke which Rice et al. (1990) mapped in detail. These rocks are here distinguished as a separate association. Beige-weathering, fine- to coarse-grained feldspathic arenite to feldspathic wacke sandstone grade upward to granule conglomerates and then to pebble and cobble conglomerate. The sandstones are commonly crossbedded (Fig. 7) or rippled. The conglomerates have minor intercalated beds of sandstone and one thin unit of intercalated fine grained mudstone. The largest clast measured in the conglomerate is 60 by 120 cm. The conglomerate is extremely heterolithic with clasts of foliated to gneissic tonalite, unfoliated tonalite, unfoliated granite, mafic volcanic, mafic intrusive, quartz vein, and probable orthoquartzite found in it (Fig. 8). The unfoliated tonalite is the most common clast type, although near the contact with mafic volcanic rocks the percentage of mafic volcanic clasts significantly increases. Mapping confirmed that the top directions from crossbedding and channel structures consistently show the top direction to the east (Rice et al., 1990).

The lower contact of this unit with the underlying pelites is not well exposed, but the sandstone lacks the early folding of the underlying greywacke-mudstone. North of the Snare River valley, the sandstone-conglomerate is in contact with

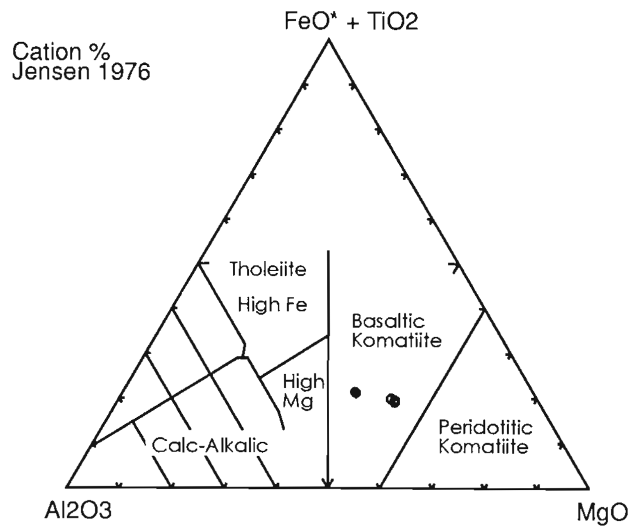

Figure 6. Jensen Cation Plot of pillowed basaltic komatiite samples. Plot produced using a recently updated version of Newpet software.

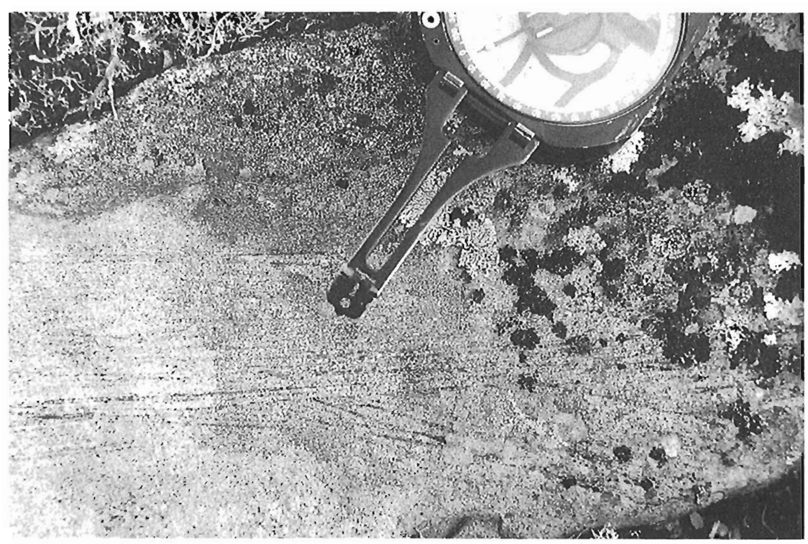

Figure 7.

Crossbedding in feldspathic wacke, part of the upper sedimentary rock association.

Figure 8.

Representative photo of rounded, poorly sorted polymictic conglomerate of the upper sedimentary rock association.

the mafic volcanic rocks and is interpreted to have erosively cut down through the turbidites. These relationships indicate that the sandstone and conglomerates are younger than the greywacke-mudstones and mafic volcanic rocks and unconformably overlie them. West of Sherpa lake an east-facing wedge of mafic volcanic rocks separates a thick section of conglomerate from the rest of this group of sedimentary rocks. This may represent an infolded and fault repeated section of mafic volcanic rocks and conglomerate or an interlayered mafic sequence as was interpreted by Rice et al. (1990). Units of polymictic conglomerate are also recognized in the south end of the belt, but are transposed and intercalated with other units including turbidites. The high state of strain in this part of the belt precludes any stratigraphic interpretations here.

INTRUSIVE ROCKS

Although mapping of the Winter Lake belt is concentrating on the supracrustal rocks, traverses were extended several kilometres into the bounding granites and gneisses, in order to examine the contact relationships along the belt margins and to provide the required continuity of this mapping with the 1:250 000 scale mapping of Thompson et al. (1993).

Eastern margin granite-gneiss complex (unit 7)

Supracrustal rocks of the Winter Lake belt are bounded along much of their eastern margin by a felsic gneiss complex (Fig. 2) containing large blocks of amphibolite and intruded by numerous unfoliated to weakly foliated granite intrusions. Most outcrops consist of varying proportions of both gneiss and the later intrusive granite. Fabric in the gneiss varies from a distinct gneissosity with well defined compositional layering 1-10 cm thick, to a much more subtle compositional variation and then to a foliation based on allignment of micas and flattening of quartz grains.

Away from the belt, gneissosity is swirly and irregular; with proximity to the belt contact, layering is abruptly transposed into a relatively straight-layered gneiss. This transposition culminates in many places along the belt margin as a 3 to 20 m wide mylonite zone. Locally, as for example immediately northwest of Izabeau lake, the straight gneiss is missing from the sequence and swirly gneiss is in direct contact with the mylonite which separates it from the supracrustal rocks.

Western margin tonalite gneiss complex (unit 8)

Along most of the western margin of the belt, as well as in isolated areas on the eastern margin, a schistose to gneissic tonalite complex is present. Within the tonalite complex, distinct dykes of relatively unfoliated tonalite intrude the deformed tonalites. One notable characteristic of this complex is the large percentage of amhibolite in the foliated to gneissic tonalite near the contact with the overlying mafic volcanic rocks. Some of this amphibolite may represent mafic xenoliths, but many mafic dykes with chilled contacts were observed. Individual dykes could not be followed through the tonalite complex and into the mafic volcanic sequence. Similar examples of tonalite complexes containing significant amphibolites and mafic dykes, such as the Anton complex, were originally interpreted to be basement to overlying supracrustal rocks. However, when the 'type' Anton complex was dated it proved to be a post-volcanic ca. 2.64 Ga intrusion (Dudas et al., 1990).

At a distance from the belt margin, gneissic layering varies from relatively straight to distinctly swirly and irregular. Adjacent to the belt margin, foliation in the schistose to gneissic tonalites is parallel to the contact, but no obvious change in strain intensity was observed in the mafic volcanic rocks close to the margin.

Late felsic intrusions (units 9 and 10)

Biotite to hornblende monzo- to syenogranites intrude both gneissic complexes and all the supracrustal rocks except the crossbedded sandstone and conglomerate unit. These rocks are pink- to red-weathering, medium- to coarse-grained and generally equigranular. Generally they have only a weak foliation, although a "ghostly" compositional banding was locally observed which is interpreted to represent almost completely digested xenoliths. One large pluton intrudes the tonalite gneiss complex, occupying a large portion of the southwest margin of the supracrustal belt (Fig. 2). Along the eastern margin, a number of smaller bodies of similar composition intruded preferentially along the supracrustal margin. North of Terminus lake, near the edge of present mapping, an intrusion of K-feldspar megacrystic granodiorite intrudes the supracrustal rocks. With additional mapping, this type of intrusion may warrant a separate subdivision. Small bodies of muscovite-biotite granite to granodiorite intrude the gneissic rocks and all the supracrustal rocks except the post-Yellowknife Group sedimentary rocks.

DEFORMATION HISTORY

The earliest deformation recognized to date in the supracrustal rocks is early folding in the greywacke-mudstone unit which predates the formation of a dominant northeast-striking foliation. No axial planar foliation is present in these early tight folds which have upright, east-striking axial surfaces. Figure 9 illustrates the incompatibility in orientation of the early folds formed in the greywacke-mudstone with the dominant foliation. Local facing reversals south of Terminus lake suggest that isoclinal folds are present in this area. Such folds were not recognized in the hinge zone of the F2 folds and it is therefore impossible to determine yet whether these are early pre-D2 isoclinal folds or minor parasitic folds on the limbs of larger F2 folds. It is important to note that similar early folds were not recognized in either the mafic volcanic rocks underlying the greywacke-mudstone or in the overlying sandstone and conglomerates. Bedding in the mafic volcanic rocks is not well defined and these may be internally folded as well or the rheologically different greywacke-mudstone may have delaminated from the mafic volcanic rocks and folded during D1. Bedding is well defined in the overlying sandstone and the more likely explanation for the lack of folding in this unit is that the sandstone was deposited unconformably on previously folded greywacke-mudstone.

The dominant foliation (S2) throughout the belt is oriented approximately parallel to the belt margins at 010-030° and generally dips steeply to the east. Folds correlated with the D2 event are bisected by S2 and occur at various scales from folds with metre-scale wavelengths to a belt-scale synform. Small-scale, upright, tight folds in the synform hinge zone plunge moderately to north (Fig. 10). The belt has an overall synformal geometry as observed by Fraser (1969) which is attributed to compression during the D2 event. However this is not a simple synclinal structure. There is a distinct asymmetry in the distribution of rock units on opposite sides of the belt which remains to be explained and faults complicate the structure.

Where the belt narrows south of the Snare River valley, complex folding patterns indicate that F2 folds are refolded. The geometry of the refolding pattern is not well defined and this is interpreted as a progressive deformation during D2. In the southern, narrowest part of the belt, all units are are transposed parallel to the belt margin. The foliation which is oriented parallel to these transposed units is interpreted to be a higher strain effect of D2 deformation.

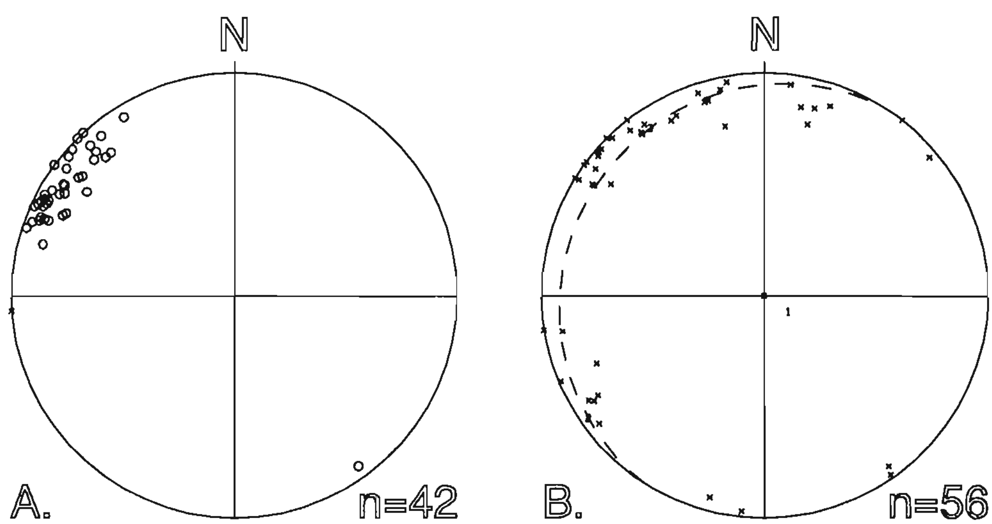

Figure 9. Stereonet plot of poles to dominant foliation (A) and bedding (B) in greywacke-mudstone west and northwest of Left Lung lake. Folds (F1) in bedding are incompatible with synchronous development of the dominant foliation (S2) and are interpreted to be pre-cleavage folds.

Figure 10.

Folds in layered quartz arenite sandstone of the lower sedimentary rock association found west of Izabeau Lake. Folds have weak axial planar cleavage not visible in the photo which is consistent in orientation with the dominant S2 foliation.

Figure 11.

Minor folds (F3) with axial planar foliation (S3) deforming bedding and S2 foliation.

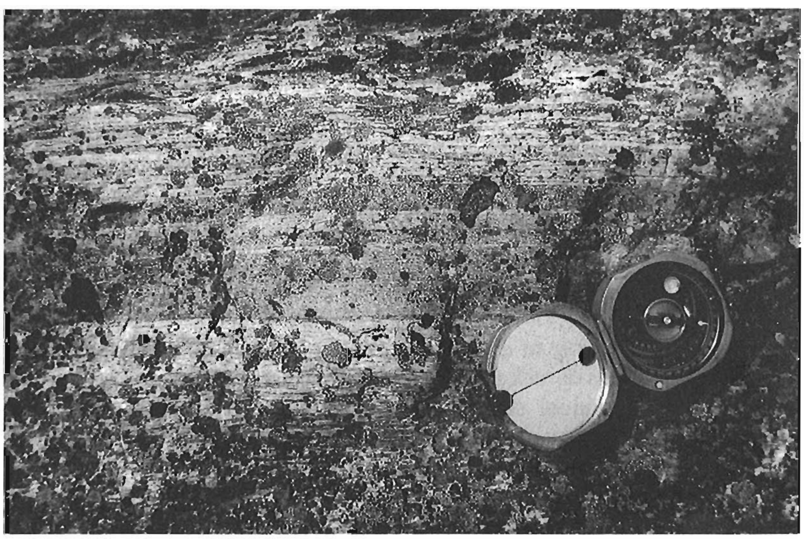

Figure 12.

Mylonitic foliation in mylonite zone separating supracrustal belt from external swirly gneiss complex on the east margin of belt. Outcrop located west of Izabeau lake.

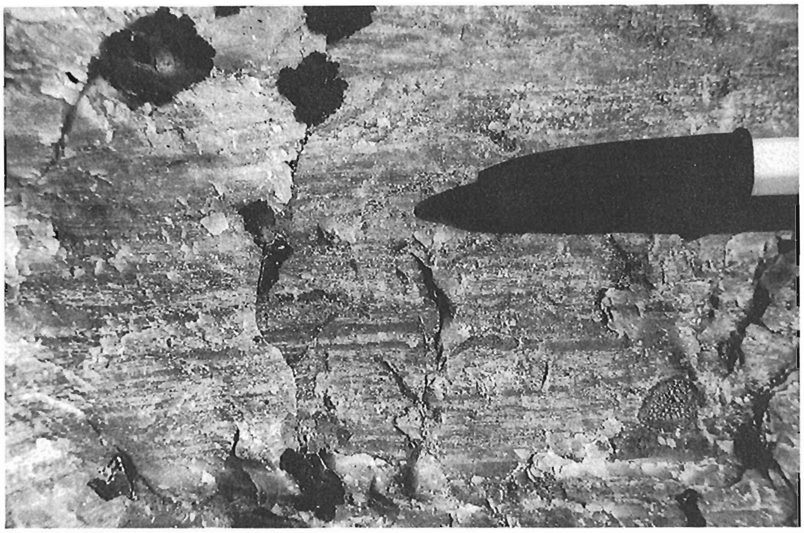

Figure 13.

Close-up of mylonitic foliation in the late mylonite cutting the southern end of the supracrustal belt. Outcrop located southwest of Izabeau lake.

Bedding and the S2 foliation are deformed by small, upright, steeply-plunging, east-striking folds which are bisected by a weakly developed, spaced cleavage (Fig. 11) striking between 80-110°. These features do not significantly affect the geometry of the units but an interpretation of their overall significance cannot be determined without a more regional perspective.

Boundary mylonite

A 3 to 20 m wide mylonite is found along part of the eastern boundary of the supracrustal belt (Fig. 12). Near Izabeau lake, the mylonite separates supracrustal rocks near the base of the stratigraphic sequence from the granite-gneiss complex which has an irregular planar fabric that strikes at high angles to the belt margin. Both steep and shallow lineations are observed in this mylonite, suggesting that movement along the zone may have been complex, although the relative timing of such motion has not been determined. The occurrence of this mylonite indicates that the Winter Lake belt is not strictly autochonous with respect to the bounding granite-gneiss complex. Further study is required to understand the timing and kinematics of this mylonite.

Post-monzogranite mylonite

The southern end of the supracrustal belt is truncated and sinistrally offset by a second, distinct mylonite zone which then continues to the north on the west side of the belt. The mylonite is up to 100 m wide, strikes at 350-360° and dips steeply to the east. Northwest of the belt, the mylonite cuts a hornblende-biotite monzogranite, and a well-developed symmetrical progression from unfoliated and weakly foliated monzogranite to a protomylonite with flattened quartz grains and feldspar porphyroclasts to an ultramylonite can be observed (Fig. 13). Lineations in this zone are well developed and consistently plunge gently to moderately to the southeast. These observations suggest that the fault has at least a final component of sinistrally-oblique reverse slip.

The mylonite zone is the northerly extension of a regional fault mapped through most of Beaulieu River supracrustal belt (Stubley, 1989, 1990a, b) and which follows a topographic lineament that possibly continues as far north as the western Point Lake area (M. Stubley, pers. comm., 1992). The sense of sinistral offset of the southern end of the supracrustal belt is consistent with the 15 km of sinistral offset along the same fault in the Carp Lakes sheet (M. Stubley, pers. comm., 1992) although the magnitude of strike separation is different. Movement along this fault post dates intrusion of the biotite-hornblende monzogranite which is the youngest Archean igneous suite mapped in the area and as yet there are no contraints on the minimum age of the mylonite.

ECONOMIC GEOLOGY

Of the numerous gossans within the mafic volcanic unit, some are silica-enriched ferrous mudstone or alteration zones with high silica and sulphide contents whereas others are essentially massive sulphide. A massive unit located north of the Snare River valley on the east side of the belt and was recognized by Fraser (1969). Another massive sulphide unit was found at the southwest shore of Terminus lake. The former is a 1 m thick unit of massive pyrite-pyrrhotite and the latter is a 15 m thick unit of siliceous chemical metasediment with various contents of disseminated sphalerite, chalcopyrite, pyrite, and pyrrhotite and includes a section of massive sulphide.

DISCUSSION

Mapping to date in the Winter Lake belt has recognized a series of rock associations which rarely occur together within individual supracrustal belts in the Slave Province. The discovery of basaltic komatiite flows is to date unique in the Slave, and their association with felsic volcanic rocks and mature quartz arenites produces an assemblage that is similar to those found in many greenstone belts of other Archean cratons such as the Prince Albert Group in the Rae Province (Schau, 1977), and most of the greenstone belts in the Sachigo Subprovince, Superior Province in Ontario (Thurston and Chivers, 1990). Where these sequences are recognized, they form the oldest preserved parts of the greenstone belt. Quartz arenite sequences without recognized komatiite but locally including ultramafic intrusive rocks, are found in other supracrustal belts in the Slave Province such as the Dwyer and Octopus formations in the Yellowknife belt (Helmstaedt and Padgham, 1986), and the Beniah Lake formation in the Beaulieu River belt (Covello et al., 1988). A preliminary date of >2.8 Ga from a rhyolite in the Dwyer Formation (Isachsen et al., 1991) is preliminary evidence for a similar relationship in the Slave Province as this is considerably older than the Yellowknife Supergroup volcanic rocks.

The rhyolite-orthoquartzite-komatiite sequences have been interpreted to have formed on shallow, stable platforms of passive margins (Schau, 1977; Covello et al., 1988; Rice et al., 1990; Thurston and Chivers, 1990). Problems associated with such an interpretation in the Winter Lake belt are: 1) the granite-gneiss complex which is the likely basement to the sequence is in structural contact with the supracrustal belt, and; 2) it is uncertain whether the felsic volcanic rocks are analogues for the alkalic volcanism typical of early continental rifting (Hoffman, 1989). Observations suggest that submarine volcanism, represented by the pillowed mafic volcanic rocks was conformably or disconformably deposited on the lower sequence.

The polymictic conglomerates and associated crossbedded sandstones may be distinctly younger than the mafic volcanic and greywacke-mudstone associations. They are interpreted to lie unconformably on both of these units, appear to have been deposited after the D1 deformation and contain granitic clasts which are likely post-volcanic in age. In all these respects, this association is similar to Timiskaming-type sequences in the Superior Province.

Significant work remains to be done to fully understand the timing, kinematics, and significance of the mylonite zones, particularly the mylonite along the eastern margin of the belt. Although it is known that this mylonite separates the supracrustal rocks from a granite-gneiss complex which is a candidate for basement-age rocks, the significance of the mylonite in the tectonic history of the supracrustal belt has yet to be determined.

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Caroline West and Adrienne Jones were both excellent field assistants, with West conducting independent mapping during the latter part of the summer. We would also like to thank our expeditor, Craig Nicholson, for his exceptional help during the summer. Discussions in the field with M. Stubley, P. Thompson, A. Donaldson, W. Mueller, and B. Padgham were very useful. Critical review by P. Thompson and E. Froese helped to improve the manuscript.

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Geological Survey of Canada Project 870008-KP

Geology of the Wijinnedi Lake area: a transect into mid-crustal levels in the western Slave Province, District of Mackenzie, Northwest Territories

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Henderson, J.B. and Schaan, S.E., 1993: Geology of the Wijinnedi Lake area: a transect into mid-crustal levels in the western Slave Province, District of Mackenzie, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 83-91.

Abstract: The area consists of four lithotectonic domains. The northwestern domain consists of intermediate metavolcanic and metasedimentary rocks of the Yellowknife Supergroup, metamorphosed as low as sub-biotite grade, that are structurally underlain to the south by a granitoid gneiss domain, in part possibly pre-Yellowknife Supergroup in age. The domain structurally below these to the southeast contains migmatitic equivalents of the Yellowknife rocks and possibly those of the granitoid gneiss complex as well, together with intrusions of varied composition, age, and deformation state. The metamorphic grade rises to granulite towards the southeast. The last domain, east of a major north-trending cataclastic shear zone near the eastern margin of the area, consists of massive to weakly foliated granitoid rocks. On the basis of aeromagnetic extrapolation, the area sits at the north end of an asymmetric block uplift up to 80 km long and 30 km wide of probable Paleoproterozoic age.

Résumé : Le secteur comprend quatre domaines lithotectoniques. Le domaine nord-ouest se compose des roches métavolcaniques intermédiaires et des roches métasédimentaires du Supergroupe de Yellowknife, métamorphisées à un degré aussi faible que celui du sous-faciès à biotite, et structuralement recouvertes au sud par un domaine de gneiss granitoïde, peut-être partiellement plus ancien que le Supergroupe de Yellowknife. Le domaine structuralement sous-jacent à ceux-ci au sud-est contient des équivalents migmatitiques des roches du Supergroupe de Yellowknife et peut-être aussi des roches du complexe de gneiss granitoïde, accompagnés d'intrusions de composition et d'âge divers et plus ou moins déformées. Le degré de métamorphisme augmente jusqu'à celui du faciès des granulites vers le sud-est. Le dernier domaine, à l'est d'une grande zone de cisaillement cataclastique de direction générale nord proche de la marge orientale du secteur, se compose de roches granitoïdes massives à faiblement feuilletées. D'après les extrapolations faites à partir de données aéromagnétiques, le secteur se trouve à l'extrémité nord du soulèment asymétrique d'un bloc atteignant 80 km de long and 30 km de large, probablement d'âge paléoprotérozoïque.

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INTRODUCTION

The Wijinnedi Lake area (Fig. 1), consisting of the Ghost Lake (NTS 85O/14) and the western 10 minutes of the Dauphinee Lake (NTS 85O/15) 1:50 000 map areas, is situated near the southwestern margin of the Slave Province at the southern end of the Indin Lake supracrustal terrain (Frith, 1986). The Ghost Lake area was previously mapped by Wright (1954) with most attention being paid to the Yellowknife Supergroup rocks. His descriptions suggested that the largely undivided granitoid terrane south of the Yellowknife supracrustal rocks was highly complex and as such different from the typically large, massive, homogeneous granitoid terranes elsewhere in the Slave Province. This terrane appeared to have more in common with the heterogeneous granitoid to granitoid gneiss complexes such as the Sleepy Dragon Complex, 100 km northeast of Yellowknife, the Healey complex of the eastern Slave (Henderson and Thompson, 1980), and parts of the granitoid/granitoid gneiss complexes of the central Slave Province (Thompson et al., 1993), some of which are known to contain pre-Yellowknife Supergroup rocks. As originally documented by McGlynn (1977), pre-Yellowknife gneiss terranes are concentrated in the western half of the Slave Province

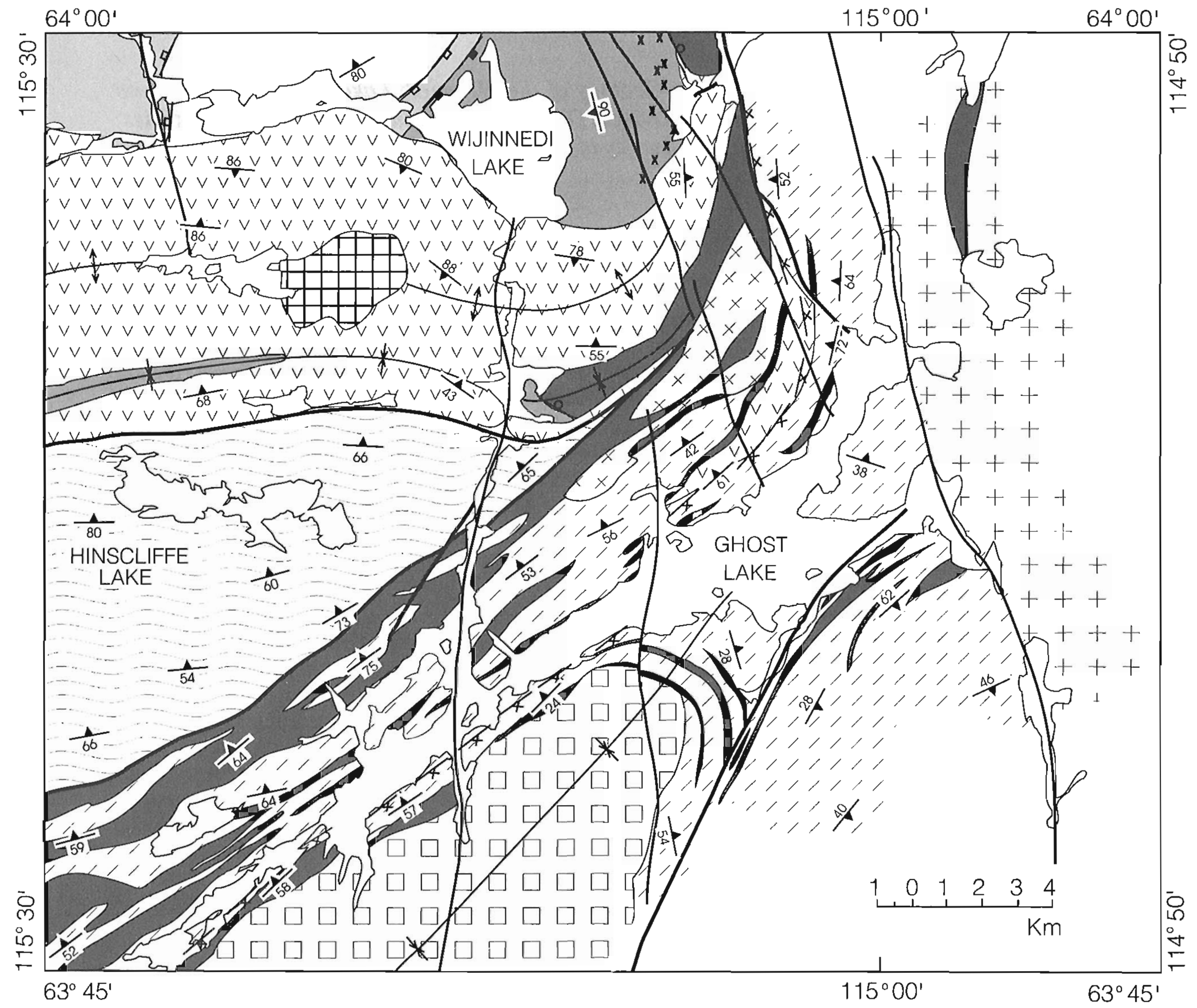
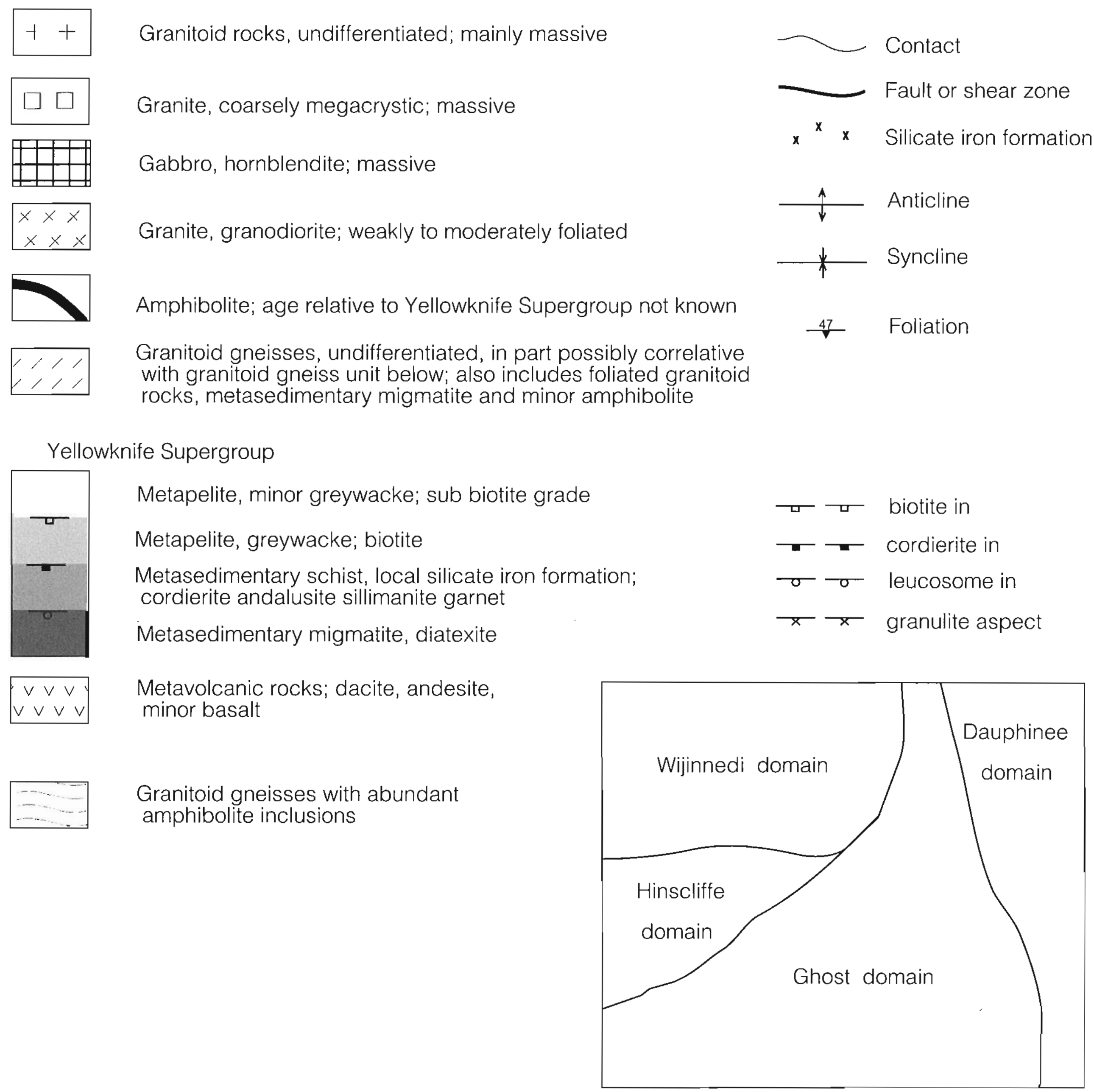


Figure 1. General geology of the Wijinnedi Lake area. The area can be divided into four lithotectonic domains. The lowest grade, least deformed rocks consisting mainly of the Archean Yellowknife Supergroup occur in the Wijinnedi domain in the northwest. It is structurally underlain by complex granitoid gneiss of the Hinscliffe domain, part of which may be pre-Yellowknife Supergroup in age. To the southeast, structurally underlying the previous two domains, is the Ghost domain consisting mainly of higher grade, more deformed equivalents of units in those domains. The metamorphic grade rises to granulite towards the southeast. In the east the Dauphinee domain, consisting of minimally deformed granitoid rocks, is separated from the high grade Ghost domain by a major greenschist cataclastic shear zone.

and include 2.99 Ga granitoid gneisses 100 km north of the area (Frith et al., 1986), 2.8-3.0+(?) Ga gneisses of the Sleepy Dragon Complex (Henderson et al., 1987; James and Mortensen, 1992) and, of course, the oldest rock in the world at almost 4.0 Ga, the Acasta gneiss, 130 km north of the map area (Bowring et al., 1989). The granitoid complex of the Wijinnedi Lake area has a distinctive high relief aeromagnetic pattern that may suggest that the complex extends as far as 40 km south of the area (Geological Survey of Canada, 1969; Fig. 2). Given the low metamorphic grade of parts of the Yellowknife succession (sub-biotite locally) and its proximity to the granitoid complex, this area appeared to be an excellent one in which to study the relationship between the Yellowknife Supergroup and the complex terranes, and so contribute to a better understanding of the geological framework and evolution of the Slave Province.



Legend to Figure 1

Most of the area was mapped at a scale appropriate for eventual publication at 1:50 000 scale. Further field work will be required in the main volcanic unit southwest of Wijinnedi Lake and in the northeast and southeast corners of the area.

GENERAL GEOLOGY

The geology of the Wijinnedi Lake area is discussed in terms of four domains, each structurally bounded (Fig. 1). The Wijinnedi domain, in the northwest quadrant of the area is dominated by moderately- to well-preserved metavolcanic and metasedimentary rocks of the Archean Yellowknife Supergroup. To the south and separated from it by a major mylonite zone is the Hinscliffe domain consisting of granitoid gneisses. The Ghost domain to the south and east is a heterogeneous terrane consisting of highly deformed and metamorphosed rocks, constituting possible lithological correlatives of the previous two domains, along with additional intrusions of a variety of ages and degrees of deformation. Further east, the Dauphinee domain, consisting of a heterogeneous assemblage of massive to weakly foliated granitoid rocks, truncates the Ghost domain at a high angle.

Dense swarms of Proterozoic diabase dykes transect the area. These are dominated by the Indin dykes that occur in a north-northwesterly-trending set that are cut by the northeasterly-trending Indin set. Also present are rare examples of Mackenzie dykes and east-northeasterly dykes, possibly correlative with the Dogrib dykes of the Yellowknife area.

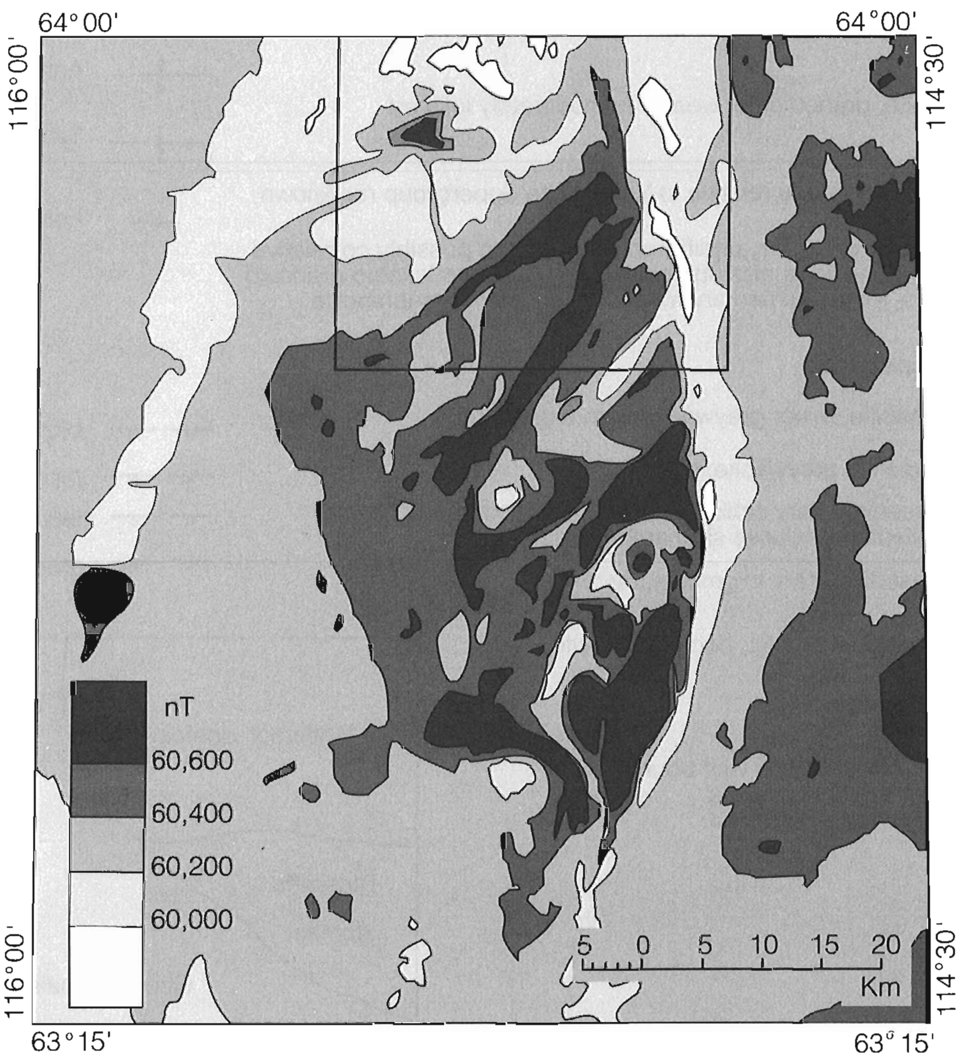


Figure 2. Regional aeromagnetic map with Wijinnedi Lake area outlined. The map area is situated at the north end of a large, triangular-shaped magnetically anomalous area that within the map area corresponds with high, largely granulite grade metamorphic rocks. The linear magnetic low on the east side of this anomalous area corresponds to the cataclastic shear zone that separates the Dauphinee domain granitic rocks from the high grade metamorphic rocks of the Ghost domain (Fig. 1). After Geological Survey of Canada (1969).

Wijinnedi domain

Volcanic rocks in this domain occur in a dome-like structure with an east-west anticlinal axis and a subsidiary syncline to the south defined on the basis of pillow facing directions and stratigraphic evidence. The main part of the unit extends beyond the map area less than 4 km to the west. To the east the units and structures change in trend and become highly attenuated as they become involved in the higher grades of metamorphism and greater deformation associated with the Ghost domain. The volcanic rocks are mainly grey to greenish grey metadacite to andesitic dacite with typical volcaniclastic textures. More mafic compositions, possibly basaltic, occur at the southern and northeastern margins of the unit. Volcanic structures and textures, including pillows, amygdules, and clasts in fine to coarse breccias are best preserved in the north. Rare carbonate layers to lenses up to 1 m thick occur on both the north and south limbs of the structure at the contact with the metasedimentary rocks, although Lord (1942) reported a carbonate unit in excess of 25 m thick at this stratigraphic position at the west end of the unit.

The dome-like shape, east-west orientation and dominantly intermediate composition of this volcanic unit contrasts strongly with the more linear, northerly-trending more mafic volcanic units of the rest of the Indin belt to the north (Frith, 1986). The volcanics south of Wijinnedi Lake have similarities to felsic volcanic centres such as those at Russell Lake, 100 km to the south (Henderson, 1985; Jackson, 1990), Clan Lake, north of Yellowknife (Hurdle, 1984), or the Back River Complex in the eastern Slave Province (Lambert et al., 1992).

Yellowknife metasedimentary rocks occur stratigraphically above the metavolcanic rocks, mainly to the north, but also within the synclinal structure near the south side of the domain; the latter connecting with the high grade metasedimentary equivalents within the Ghost domain. The metasediments are dominantly pelites to siltstones, occurring in thin-bedded, pelite dominant, graded couplets characteristic of turbidity current deposits. The greywackes that are so characteristic of the Yellowknife Supergroup sedimentary rocks elsewhere in the Slave Province are notably rare. Metamorphic grade within the domain ranges from sub-biotite west of Wijinnedi Lake to migmatite in the southeastern part of the domain. Silicate-facies iron-formation occurs in units up to several metres thick east of Wijinnedi Lake (Fig. 1). They consist of garnet-rich, amphibole-rich, and silica-rich layers together with intermediate mixtures of these end members interlayered on a centimetre scale and locally contain sulphides. These silicate iron-formations are similar in many respects to gold-bearing silicate iron-formations that occur to the south at Russell Lake (Lord, 1951; Henderson, 1985) and to the northeast between Point and Contwoyto lakes (Bostock, 1980). Some of the occurrences in the area have been explored for gold (Wright, 1950).

A large mafic igneous complex, ranging in composition from hornblendite through gabbro to anorthositic gabbro, occurs within the volcanic sequence (Fig. 1). Numerous dykes, sills, and small plugs presumably related to the complex occur both in the volcanic and sedimentary rocks to the north. Locally along the contact between the metavolcanic and metasedimentary rocks along the south shore and west of Wijinnedi Lake, are gossans due mainly to pyrrhotite mineralization. Mafic sills, invariably associated with the gossans, are also commonly mineralized. Rare, small bodies of hornblendite occur locally within the Hinscliffe and Ghost domains as well. Except for localized shears within and at their margins, the intrusive bodies are massive and metamorphosed to the prevailing regional grade.

Hinscliffe domain

The boundary between the Wijinnedi domain and the Hinscliffe domain to the south is a mylonite zone as wide as 1 km. Where exposed, the contact between the finely laminated, locally steeply down-dip lineated, fine grained, black mylonitic amphibolites of the Yellowknife volcanic rocks and the Hinscliffe granitoid mylonitic gneiss is straight, sharp, and clean.

The Hinscliffe domain consists of leucocratic pink, white to grey granitoid gneiss that is generally recrystallized to a distinct sugary texture. Layering on a decimetre to metre scale is defined by sharp variations in composition and texture. Where best preserved intrusive relations between the various phases are evident, commonly as inclusions of one phase within another. Massive to variably foliated dykes and veins of aplitic granite cut the gneiss. The complex was never a simple foliated intrusion but more likely formed through multiple injections of numerous granitoid sheets. Associated with the granitoid gneiss but in varied proportions are deformed angular blocks of amphibolite of uniform composition. They are invariably intruded by several generations of thin veins and dykes of granite that in places become so pervasive that the outcrops have textures and structures more reminiscent of metasedimentary migmatites. Where least disrupted, the amphibolite blocks occur in distinct narrow zones possibly indicating original emplacement as gabbroic sheets in the granitoid gneiss complex. At both the south and north margins of the complex are small elongate intrusions of intermediate to mafic composition that were involved in the deformation associated with the domain-margin mylonites. At the southern margin, the deformation is more intense as these intrusive bodies occur as greatly elongate, in places folded, layers. Narrow linear diatremes dominated by clasts of rounded, pink, quartz-rich granite of a type not recognized in the area, in a dark green mafic matrix, occur at two localities in the complex.

This granitoid gneiss complex with its multiple intrusive events would appear to have an extended history. With the exception of the relatively young mafic igneous complex, none of these intrusive events are represented in the adjacent Yellowknife volcanic and sedimentary rocks of the Wijinnedi domain. It is reasonable then to speculate that part and indeed perhaps much of the complex is possibly pre-Yellowknife Supergroup in age. No direct evidence of a predeformational unconformity was recognized between the complex and the Yellowknife to the north. The interpretation of mafic dykes in granitoid gneiss terranes as feeders to Yellowknife volcanic rocks has commonly been used as evidence for pre-Yellowknife Supergroup basement (Baragar, 1966; Easton et al., 1982; Lambert et al., 1992). This interpretation can not be applied to the Hinscliffe domain because the amphibolite bodies interpreted as mafic intrusions are compositionally distinct from the dacitic composition that characterizes the immediately adjacent volcanic sequence. The mafic bodies are more likely sills emplaced into the granitoid complex under deep seated conditions.

Ghost domain

As with the Wijinnedi-Hinscliffe domain boundary, the boundary between the Hinscliffe domain and the much more extensive Ghost domain to the southeast (Fig. 1) is a shear zone. Almost everywhere the contact is obscured by 10 to 20 m of overburden, except in the southwest where although exposed, the transition is obscured by a later, but still deformed, granite. Northeast of Hinscliffe domain, easterly-trending rock units of Wijinnedi domain increase in metamorphic grade and swing into concordance with the northeasterly trend of the Ghost domain. The precise nature and location of the boundary between the two remains undefined as it appears to mark little lithological or metamorphic contrast. No particular high strain zone is apparent although one may well exist, given the relatively poorer exposure in this area.

The Ghost is the most heterogeneous domain, consisting of migmatitic to diatexitic metasedimentary rocks, high grade metavolcanic rocks, granitoid gneisses, and a range of intrusive rocks that have undergone varied degrees of deformation and metamorphism. The units illustrated in Figure 1 are generalized with many of the phases described below being interlayered on a fine scale in many places. Units of the domain occur in a gently northeast-plunging, northwest-verging anticlinal structure that is cored in the south by a large, massive, homogeneous granite with dense concentrations of extremely coarse potassium feldspar megacrysts.

The northwest part of the domain is dominated by migmatite and, to a much lesser extent, diatexite derived from Yellowknife metasedimentary rocks similar to those of the Wijinnedi domain with which they are continuous. Similar rocks occur throughout the domain in lesser proportions where diatexitic equivalents are more prevalent. The migmatites, commonly with a moderate to high proportion of leucosome interlayered on a millimetre- to several centimentres-scale, form spectacular, highly plastic structures. In addition to the normal high grade pelitic assemblages made up of quartz, plagioclase, biotite, muscovite, garnet, and sillimanite, these rocks commonly contain spectacular, coarse, purplish blue cordierite, as previously reported in the region (Folinsbee, 1940; Lord, 1942; Wright, 1954). The cordierite can be over 10 cm in size and in some cases are of possible gem quality. The diatexite derived from the metasediments is a white, medium grained, even grained rock with a distinctive texture consisting of blocky plagioclase and biotite aggregates. Garnet where present is abundant and refractory inclusions of psammitic composition occur locally.

One large, mainly intermediate-composition unit of metavolcanic rock occurs north of the eastern part of Ghost Lake. It consists of flinty, grey to greenish grey, fine grained, layered on a centimetre- to decimetre-scale, laminated, feldspar-quartz-biotite-hornblende rock that is locally migmatitic, typically with less than 10% leucosome. The composition is varied and locally is amphibolitic.

By far the most abundant unit in the domain is heterogeneous granitoid gneiss. Although present throughout, it becomes increasingly dominant across the domain towards the southeast. It is generally leucocratic, grey to pinkish grey, or yellowish to rusty greenish brown at higher grades. The gneissic layering is texturally and compositionally defined. Thin laminated amphibolite layers, typically less than 10 cm thick, are common among the much more dominant quartzofeldspathic gneiss. South of Ghost Lake and north of the east end of it, these gneisses appear to be at granulite metamorphic grade based mainly on the change in colour from greys and pinks, to the yellows and greasy brownish-greens commonly seen in granulite grade terranes. This colour change is the basis for the "granulite aspect isograd" in Figure 1. The three thin sections examined contain the granulite grade assemblage plagioclase-quartz-potassium feldspar-clinopyroxene-orthopyroxene-biotite. These gneisses dominate the southeastern and eastern two thirds of the Ghost domain. In many respects these highly deformed and metamorphosed rocks resemble the lower grade granitoid gneiss of the Hinscliffe domain, with the abundant amphibolite blocks in the Hinscliffe represented in the Ghost by the thin amphibolitic layers in the gneiss.

The granitoid gneisses and migmatitic Yellowknife supracrustal gneisses contain numerous intrusions of widely varied composition, form and degree of deformation. In general, the presumed older intrusions are less potassic in composition, consisting of sill-like bodies of moderately to strongly foliated diorite to tonalite. In many areas they contain feldspar megacrysts. The younger intrusions are more felsic and less deformed. A particularly large body of pink, massive to weakly foliated, medium grained, sparsely megacrystic granite to granodiorite occurs north of the eastern part of Ghost Lake (Fig. 1). The massive coarse, densely megacrystic granite south of Ghost Lake mentioned previously, together with small irregular bodies of granite that occur northwest of Ghost Lake appear to be undeformed. Massive biotite-bearing pegmatites with common black tourmaline, apatite and, in rare instances, abundant magnetite, occur throughout the domain. Spodumene, present in some pegmatites of the southern and eastern Slave, was not noted.

The Ghost domain consists mainly of lithological elements present in the other two domains. There is little doubt that the migmatitic supracrustal gneisses are correlative with the Yellowknife Supergroup of the Wijinnedi domain. The connection between the granitoid gneiss and amphibolites of the Hinscliffe domain and those of the Ghost domain is more tenuous, but one well worth further investigation. The interlayering of metasedimentary and granitoid gneisses on a variety of scales within the domain is thought to represent a tectonic interleaving that predated the definition of the various domains.

Dauphinee domain

The boundary between the granulite grade gneisses of the Ghost domain and the massive to weakly foliated granitoid rocks of the Dauphinee domain is a prominent cataclastic shear zone up to 800 m wide that corresponds to a prominent low on the regional aeromagnetic map of the region (Geological Survey of Canada, 1969; Fig. 2) and with a relatively steep gradient on the regional Bouguer gravity map (Earth Physics Branch, 1969a, b; Fig. 3). The rocks within the zone are extensively chloritized, hematized, fractured, and contain local pseudotachylite veins. There is no through going fabric. Carbonate cemented breccias are present locally. Within the domain, parallel to the main boundary

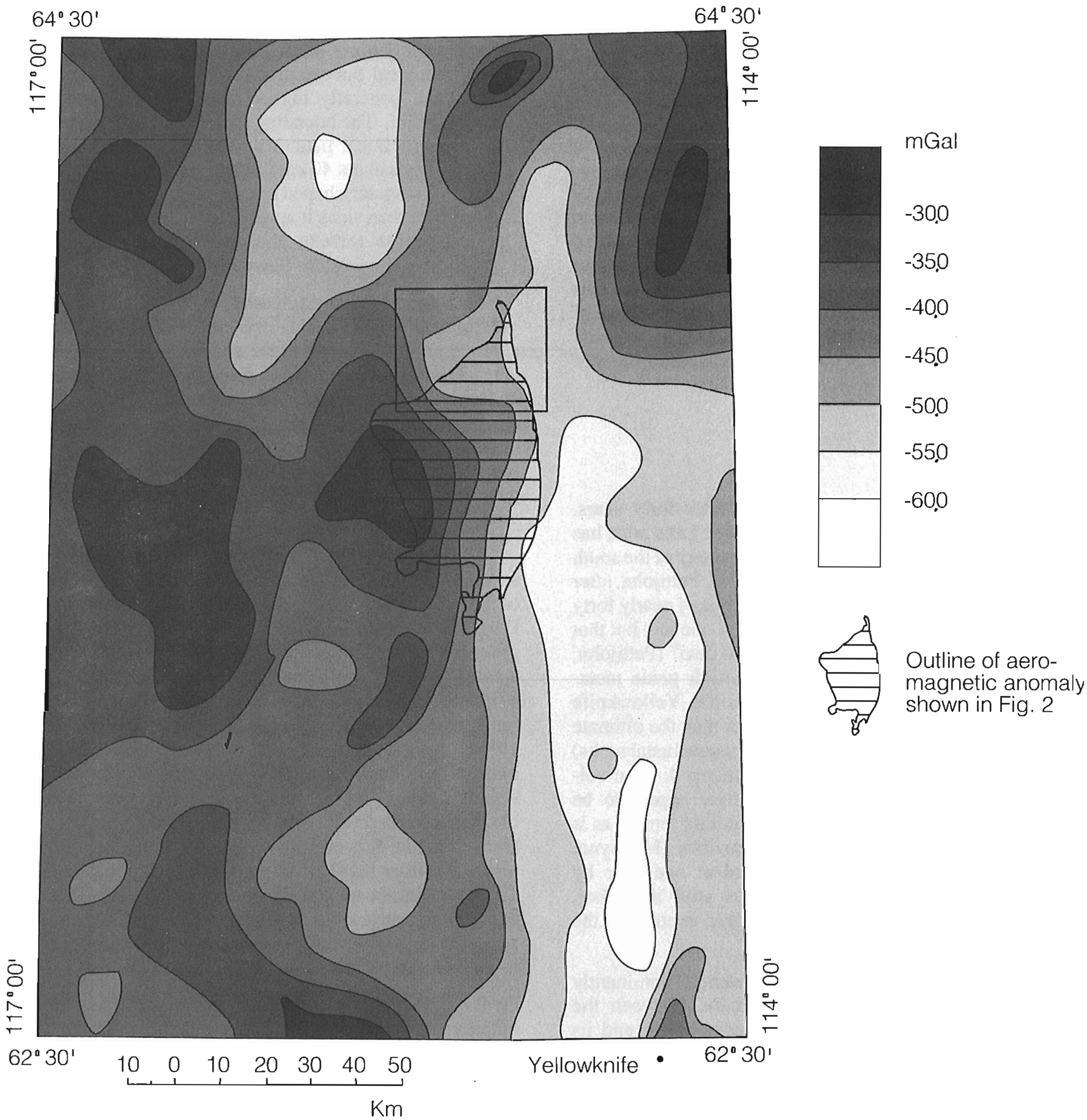


Figure 3. Regional Bouguer gravity map with Wijinnedi Lake area outlined. The map area and the magnetically anomalous area that is correlative with the high metamorphic grade rocks in the map area, correspond with a relatively steep gradient in an otherwise gently varied Bouguer gravity anomaly field. After Earth Physics Branch (1969a, b).

fault, are several crush zones in which epidote-filled fractures and gash fillings are common. Several smaller scale but similar-trending faults are mapped in the Ghost and Wijinnedi domains to the west. The amount of horizontal displacement on these structures is unknown but the vertical component may be significant. For example, across one of the faults east of Wijinnedi Lake there appears to be a metamorphic jump in pelitic metasedimentary rocks, with spectacularly developed sillimanite east of the fault but none apparent in hand specimen west of the fault. No north-northeasterly-trending transcurrent Indin faults that are so common to the north (Frith, 1986) were recognized in the area.

The Dauphinee domain itself consists of massive to weakly foliated, pale pink to grey, biotite granite to granodiorite with sparse megacrysts. East of Ghost Lake is a heterogeneous body of mainly tonalite that ranges from granodiorite to quartz diorite. It is cut by veins to small dykes of granite of several generations and is locally foliated. It has a distinctive texture that is reminiscent of some of the tonalite sills that occur in the Ghost domain. A northerly-trending zone of metasedimentary migmatite similar to the high-grade Yellowknife migmatites was mapped in the northern part of the domain (Fig. 1).

DISCUSSION

The remapping of the Ghost Lake area after forty years, together with part of the adjacent Dauphinee Lake area, has resulted in a different perspective on the geology of the south end of the Indin Lake supracrustal belt. As F.J. Pettijohn, after revisiting an Archean terrane after an absence of nearly forty years, noted: "... the rocks had not changed one iota but that our ideas about them had changed a great deal" (Pettijohn, 1970, p. 239). Thus by thinking of the high grade metamorphic rocks in terms of their protoliths (i.e. Yellowknife Supergroup metasedimentary rocks) rather than the ultimate product (granitoid rocks with remnants of supracrustal rocks) the impression of the region is greatly changed. The volcanics south of Wijinnedi Lake no longer appear to be perched at the margin of an extensive granitoid terrane as is so common in the rest of the Slave Province (McGlynn, 1977), but were more likely surrounded at one time by Yellowknife metasedimentary rocks as with the other, previously cited, felsic-dominated volcanic centres in the Slave Province.

The occurrence of a potentially old terrane of dominantly granitoid gneisses, the Hinscliffe domain, between the Yellowknife Supergroup of the Wijinnedi domain and its high-grade equivalents in the Ghost domain is intriguing; clearly some major tectonic stacking has taken place. This is supported by the fact that the boundaries between the various domains are major shear zones. In this regard aspects of James and Mortensen's (1992) model for the Sleepy Dragon Complex northeast of Yellowknife may be relevant. There the complex granitoid gneiss terrane adjacent to the Yellowknife Supergroup supracrustal rocks is interpreted as a metamorphic core complex initially overthrust and later unroofed during extension, possibly associated with the emplacement of major granite plutons.

A northwest to southeast transect across the Wijinnedi Lake area crosses increasingly high-grade rocks from subbiotite in the northern Wijinnedi domain to granulite grade gneisses in the southeastern Ghost domain, ending abruptly at the northerly-trending cataclastic fault zone that marks the western margin of the Dauphinee domain. The high grade rocks have a distinct pattern on the regional aeromagnetic map (Geological Survey of Canada, 1969; Fig. 2) characterized by a generally higher magnetic level and greater magnetic relief. The bounding fault zone on the east side has a distinct low linear pattern that along with the high-relief terrane extends about 40 km south of the area mapped to form a roughly triangular-shaped area (Fig. 2). On its northwestern and southwestern sides it appears to grade into the generally flat featureless terrane characteristic of the magnetic expression of the western Slave Province.

This area is also expressed on the regional Bouguer gravity map (Earth Physics Branch, 1969a, b; Fig. 3). There a gentle gravity gradient in the southeastern Slave Province (Fig. 3) changes from more positive in the west to more negative in the east with the steepest part of the gradient corresponding to the magnetically anomalous area described above. In addition this area marks an interference pattern formed by the north-trending low and an east-trending positive feature that ends within the magnetically anomalous area.

It is suggested that this terrane represents a block uplifted such that the deepest levels are exposed at the eastern margin of the block. Granulite grade rocks are rare in the Slave Province; the only other occurrence of metamorphic orthopyroxene-bearing rocks occurs about 100 km northeast of the area (Thompson, 1978) and their extent is unknown. The Wijinnedi Lake area is situated on the northern margin of this uplift as there is little evidence (Tremblay et al., 1953; Frith, 1986) that the high-grade rocks or bounding structure extend more than a few kilometres to the north. As can be seen in Figure 2, there is a distinct northeasterly grain within the block that corresponds to the dominant structural trends in the Ghost domain. It is of interest that this trend, parallel to the regional surface trend in the Yellowknife region (Henderson, 1985), is similar to that observed in a seismic anisotropy experiment carried out at Yellowknife that suggested the presence of parallel structures in the mantle at a depth of up to 200 km (Silver and Chan, 1988).

It is premature for speculation as to the origin of the uplift but there are some similarities between this structure and the much larger Kapuskasing uplift of the Superior Province (Percival and McGrath, 1986). The age of the uplift remains undefined although it is considered to be post-Archean as the bounding low grade cataclastic fault zone contains deformed, presumably Proterozoic dykes. As well, the Proterozoic Indin dykes appear to be somewhat more altered in the southwestern part of the area relative to that part of the area that has undergone less uplift. Consideration might be given to the idea that the uplift south of Ghost Lake is a further manifestation of the compressional event associated with the indentation into the western Churchill Province of the eastern Slave Province and part of the western Thelon Tectonic Zone during the Paleoproterozoic (Henderson et al., 1990).

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Geological Survey of Canada Project 870008-JB

Re-evaluation of field relationships in the western Point Lake transect, central Slave Province, Northwest Territories

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Abstract: Distinctly different interpretations of field relationships at Point Lake have lead to two fundamentally different views of the tectonic history of this area. This communication describes field relationships in part of the Point Lake transect, evaluates proposed tectonic models in light of the field observations, and lists questions remaining to be addressed through continued mapping and geochronology.

Résumé : Des interprétations très différentes des relations lithologiques observées sur la terrain dans la région du lac Point ont mené à l'élaboration de points de vue fondamentalement différents en ce qui concerne l'évolution tectonique de cette région. Le présent article décrit les relations observées sur le terrain dans un secteur du transect du lac Point, évalue les modèles tectoniques proposés en fonction de ces observations de terrain et dresse une liste des questions que la poursuite des travaux cartographiques et géochronologiques peut seule résoudre.

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INTRODUCTION

Location

Point Lake is located in the central portion of the Archean Slave Province, approximately 350 km north of Yellowknife, N.W.T. (Fig. 1). Wave-washed shores at Point Lake provide relatively continuous, lichen-free exposure along a >75 km east-west transect, nearly perpendicular to Archean structural trends in the region. Exposures in the western portion of the transect include a contact that separates a gneiss terrane composed of gneisses, migmatites, and granite plutons from a greenstone terrane composed of metavolcanic and metasedimentary rocks. Quality and continuity of outcrop make the Point Lake transect an unparalleled place to examine the structural characteristics of these terranes and to test models of their juxtaposition.

Previous Work

The regional distribution of various rock-types was established by Bostock (1980) and earlier reconnaissance studies (Stockwell, 1933; Fraser, 1960, 1964). Various aspects of the metamorphic, magmatic, sedimentological, and structural characteristics of rocks in the Point Lake area are described by Easton (1985, 1987), Henderson (1981, 1988), Henderson and Easton (1977), Hoffman (1986), Jackson, (1982, 1984), King and Helmstaedt (1989), Kusky, (1989, 1991), Padgham (1985) and references in these studies.

Two fundamentally different tectonic models for the development of the greenstone and gneiss terranes have arisen from this observational framework.

One tectonic scenario envisions the formation of the greenstones in fault-bounded basins which developed during extension of the older basement (Henderson, 1981, 1988; Easton, 1985, 1987). In this model, the greenstones are autochthonous and were deposited on the older gneisses. Advocates of this model emphasize the clearly depositional relationship of conglomerates within the greenstone belt on plutonic and gneissic "basement" and interpret mafic dykes and plutons present in the gneisses as hypabyssal equivalents to the mafic volcanic rocks in the greenstone belt. They regard the mylonitic contact at the base of the greenstone volcanic sequence to be the result of relatively late faulting.

Alternatively, the greenstone belt has been interpreted as an allochthonous fragment of oceanic crust obducted onto a passive margin of the gneiss terrane during collision with a west-facing volcanic arc and accretionary complex (Hoffman, 1986, 1989; Kusky, 1989, 1991). Protagonists of this model point out the ophiolite-like stratigraphy of the greenstone terrane, the large, mylonitic shear zone which structurally separates the greenstone and the gneiss terranes, and the lack of recognized extensional structures in the area. They interpreted the depositional contact cited in the first model as a synorogenic erosional surface covered by molasse shed from once-emergent, out-of-sequence thrust faults developed during the arc collision.

Purpose of Present Study

Significant differences in the models previously proposed for the tectonic evolution of the Point Lake area reflect distinctly different interpretations of field relationships at Point Lake. The purpose of this investigation is to re-examine these relationships in detail. The approach combines observations of the deformational history, metamorphism, tectonostratigraphic assemblages, and contact relationships of rocks exposed along the western Point Lake transect with geochronology in order to test the validity of existing tectonic models and/or to generate new hypotheses. This communication describes field relationships observed during the past field season, evaluates proposed tectonic models in light of the field observations, and highlights questions remaining to be addressed through continued detailed mapping and geochronology.

FIELD RELATIONSHIPS AND OBSERVATIONS

Observations in this report focus on the field relationships exposed south of Point Lake from approximately 113°23'W to 112°59'W (Fig. 1-5). This area was mapped in detail, and reconnaissance observations extend an additional 15 km to the east. The lithologies and contact relationships mapped are shown in Figures 2 and 3, and they will be described in order of increasing structural level, which corresponds generally to their order of occurrence from west to east along the Point Lake transect. Many of the observations described are similar to those recognized by previous authors (Henderson and Easton, 1977; Jackson, 1984; Easton, 1985; Hoffman, 1986, 1989; Henderson, 1988; Kusky, 1991); but several, especially regarding contact relationships in western Keskarrah Bay, are significantly different.

Gneiss Terrane

A diverse suite of interlayered and complexly deformed rock-types form the gneisses at the western end of the transect. Igneous rocks ranging in composition from basalt to diorite, tonalite, and granite, as well as metasediments including quartzite and biotite + sillimanite gneiss are interlayered and tightly folded at scales ranging from centimetres to tens of metres. For the purposes of this study, the gneisses are divided into four general lithological units: a heterogeneous gneiss unit (hgn), a granitic orthogneiss unit (ggn), an intermediate orthogneiss unit (ign), and a heterogeneous granitic mylonite unit (hmy). All of these lithologies are variably intruded by late, relatively nondeformed granite and pegmatite rich in K-feldspar megacrysts (ygr). Alternative selections of map units can be found in Easton (1985), Henderson (1988) and Kusky (1991).

Biotite + sillimanite gneiss with rare, 1 to 5 cm quartzite layers, characterize the heterogeneous gneiss unit (hgn), but at least two generations of diorite and tonalite dykes are present, and they locally dominate. Metabasalt(?) occurs as amphibolite boudins up to several metres in maximum dimension within the biotite + sillimanite gneiss.

Outcrop-scale crosscutting relationships among the various rock types in the heterogeneous gneiss unit indicate that their relative ages (oldest to youngest) are: biotite – sillimanite gneiss, amphibolite, diorite gneiss, and tonalite gneiss. All of these rock-types are well-foliated and tightly folded. The eastern boundary of the heterogeneous gneiss unit against the granitic orthogneiss unit is not exposed.

The granitic orthogneiss unit (ggn) is composed of fairly homogeneous, pink to grey granite gneiss. This gneiss is medium-grained, equigranular, and locally contains epidote-amphibolite and tonalite gneiss boudins or xenoliths up to 1 m in maximum dimension. A broad, moderately east-dipping, anastomosing shear zone forms the eastern contact of the granite orthogneiss unit against the intermediate orthogneiss

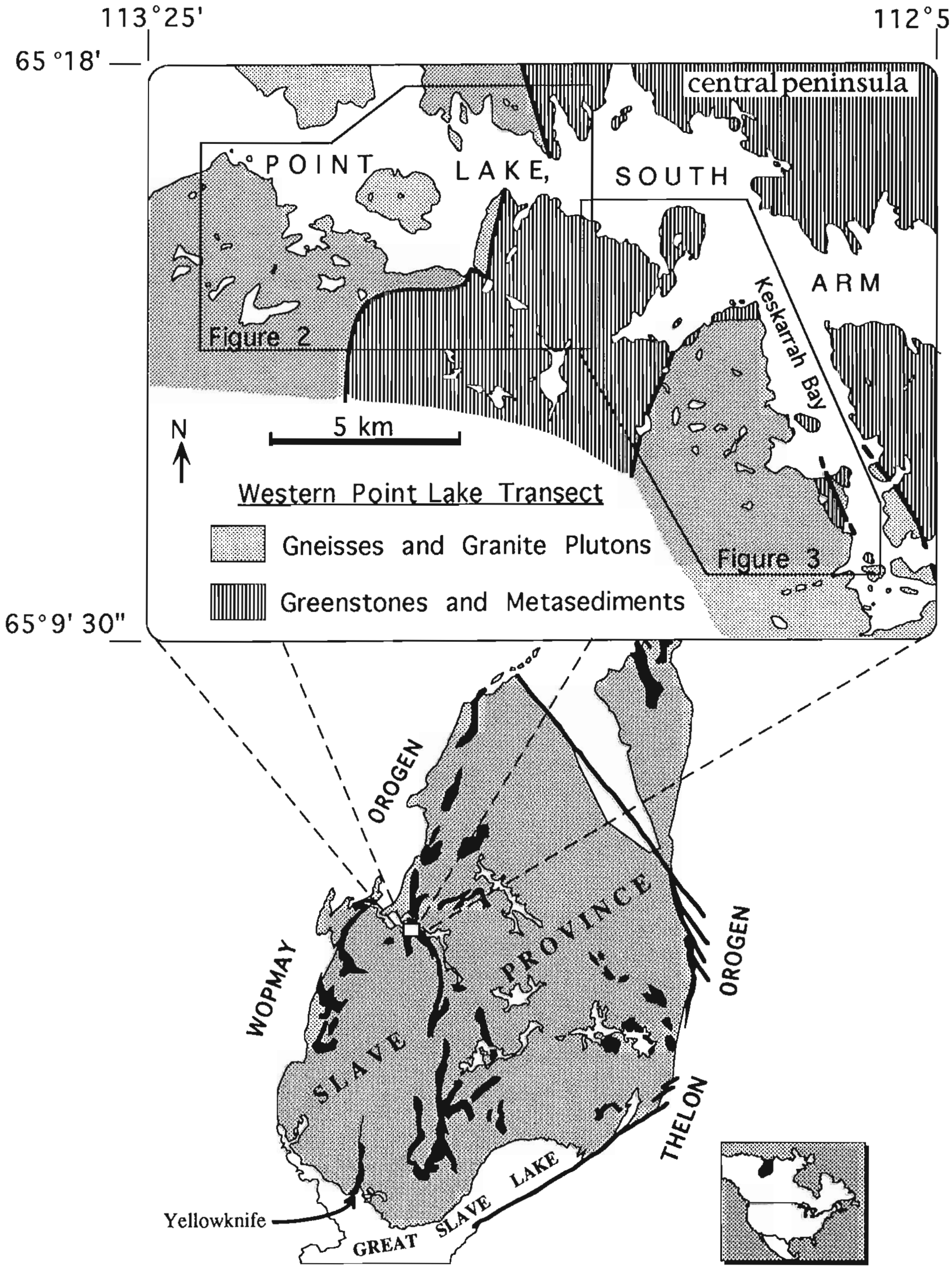

Figure 1. Location of the western Point Lake transect, and the general distributions of major tectonostratigraphic assemblages.

unit. Rare extensional shear bands and feldspar porphyroclasts with asymmetric tails occur in the shear zone, and these kinematic indicators suggest an oblique, east-over-west sense of displacement.

The intermediate orthogneiss unit (ign) is composed of heterogeneous, medium to dark grey diorite and tonalite gneiss. This unit commonly contains thin, highly strained felsic layers that are concordant to the gneissic compositional layering. Anastomosing mylonitic shear zones <2 m-wide that dip moderately to steeply to the east occur throughout the unit and grade eastward into a pervasive 0.2 to 1.0 km-wide, subvertical zone of strongly foliated and lineated, heterogeneous granitic mylonite (gmy) near the eastern contact of the gneiss terrane. The shear zones contain crystal-plastically deformed K-feldspar, and mafic zones in the mylonite contain a dynamically recrystallized amphibole + biotite + plagioclase + quartz mineralogy. These observations are consistent with amphibolite grade temperatures during the motion on the shear zones. The local compositional heterogeneity of the granitic mylonite unit may result from primary heterogeneity of the gneisses, syndeformational intrusion of magmas with variable compositions along this structural boundary, and/or structural interleaving of mylonitic gneisses with the overlying mafic mylonites. Mafic dykes up to 10 m wide are common in the mylonitic gneiss, especially north of the south arm of Point Lake. These dykes are metamorphosed to amphibolite and contain a foliation oriented parallel to the mylonitic fabric in the granitic mylonite.

Greenstone terrane

In general, the metamorphic grade of rocks in the greenstone terrane decreases from amphibolite-grade near the western margin to lower greenschist-grade in the area of Keskarrah Bay. A 0.2 to 1.0 km wide subvertical or steeply east-dipping zone of mafic mylonite (mmy) occurs at the western edge of the greenstone terrane (Fig. 2). This unit was called the Perrault Formation by Henderson (1988), and was mapped as mafic mylonite by Kusky (1991). Intensely foliated, fine grained amphibolite comprises the bulk of this lithology. The lower portion of the mafic mylonite zone contains several 0.1 to 5.0 m wide zones of chlorite + serpentine + talc schist/

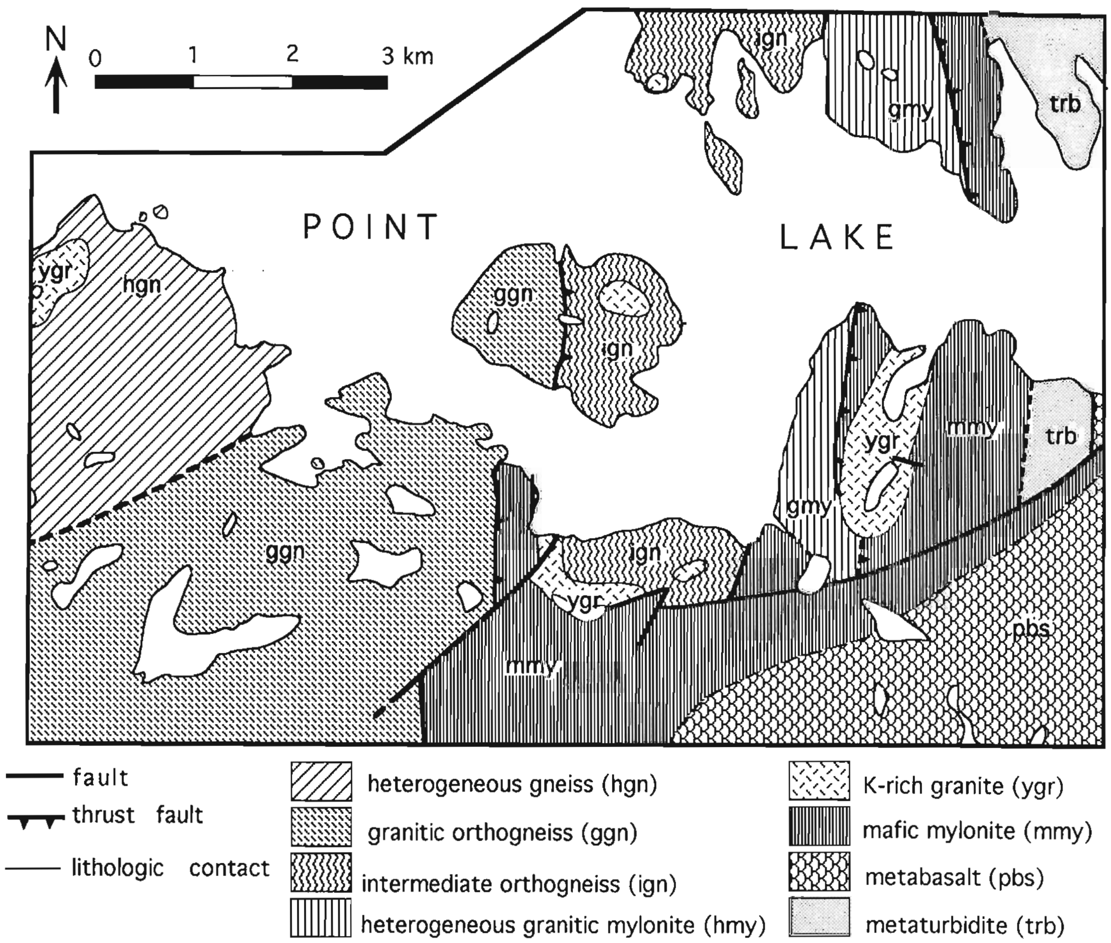

Figure 2. Simplified geological map of the western portion of the study area, including the plutons and gneisses of the gneiss terrane and the structural boundary with the greenstone terrane.

mylonite and 0.1 to 3.0 m boudins of coarse grained pyroxenite and gabbro. These features suggest that mafic and ultramafic intrusive rocks existed at the base of the greenstone terrane prior to mylonitization. In the upper portion of the mafic mylonites, metamorphic grade drops to chlorite + actinolite greenschist facies. On the south shore of the south arm of Point Lake, a small pluton of relatively nondeformed granite (ygr) intrudes the mafic mylonite (Fig. 2). South of Point Lake, at the top of the mafic mylonite zone, the strain intensity diminishes over a distance of 50 to 100 m and yields well-foliated but recognizable pillow-basalt flows.

Greenschist grade massive and pillowed metabasalt (pbs) dominate the layered volcanic sequence above the mafic mylonites, with little or no contribution from more evolved magmas or interbedded clastic sediments (Fig. 2). Where recognizable, primary layering strikes generally north, dips vertically or steeply to the east, and is subparallel to the penetrative tectonic foliation. The younging direction provided by apparent pillow facing is variable, but to what extent this reflects isoclinal folding vs. deformational modification of pillow geometry remains unclear. Zones of high strain, 1 to 10 m in width and similar in character to the mafic mylonites, are common in the metabasalt section but can rarely be traced along strike for more than 200 m. Many laterally discontinuous, <2 m-thick horizons of sulphide-bearing iron-formation which have been extensively prospected exist in the volcanic pile and are probably manifestations of synvolcanic hydrothermal activity. Henderson (1988) called these metabasalts the Peltier Formation.

The top of the mafic mylonite zone on the central peninsula of Point Lake and for approximately 1 km south of the lake is a covered or poorly exposed interval about 100 m wide. Across this interval, the lithology changes from mafic mylonite to metaturbidite (trb). These metaturbidites are at chlorite±biotite grade and contain well-preserved primary sedimentary structures. Individual beds are 10 to 200 cm thick, and are subvertical or steeply east-dipping. Graded beds demonstrate several reversals in the younging direction across the metaturbidite exposure, indicating the presence of isoclinal folds. No volcanic rocks are interbedded with the turbidites, although thin iron-formations similar in character to those in the volcanic section occur sporadically in the metaturbidites. The eastern contact of the metaturbidites is poorly exposed, but the presence of highly strained pillow-basalts east of the turbidites suggests this contact is shear zone. The metaturbidites belong to the Contwoyto Formation of Henderson (1988).

The penetrative tectonic foliation in the metaturbidites is commonly refracted within individual beds due to rheological differences between the sandy bases and the silty tops of the beds. The refracted tectonic foliation appears very similar to cross laminations produced by primary sedimentary processes, especially in places where quartz mobilized during metamorphism and deformation forms thin veins along the curved foliation, enhancing the "cross-lamination" appearance. However, close examination reveals that the foliation is defined by an alignment of metamorphic phyllosilicates, and the younging direction suggested by the pseudo-cross-laminations is sometimes opposite of that given by the primary grading in the bed. These fabric relationships are spectacularly displayed in the metaturbidite exposures along the north shore of the south arm of Point Lake, and serve as a caution against a casual interpretation of fine-scale arcurate features in deformed metasediments as primary cross-laminations.

The pillow-basalts are unconformably overlain by younger clastic sediments including arkosic quartzite (qzt) and conglomerate (cgl) (Fig. 3). These units comprise the Keskarrah Formation of Henderson (1988). Although they most likely represent different facies of the same depositional event, no clear gradation from the quartzite into the overlying conglomerate was observed, and their genetic relationship remains unclear. Bedding strikes north and dips steeply east. Trough crossbeds are prominent in the 10 to 100 cm-thick beds of quartzite near the neck of Cyclops peninsula. The quartzite beds at this location are isoclinally folded, and crossbeds demonstrate several reversals in the younging direction from west to east across the quartzite exposure.

The conglomerate is usually clast supported, and cobbles consist of variable combinations of granite, tonalitic gneiss, and mafic fragments. Mafic cobbles dominate near the base of the conglomerate, just above the quartzite, and felsic cobbles become more prevalent near the top. The matrix is composed of poorly sorted, lithologically heterogeneous material with a high proportion of mafic lithic fragments.

Several sequences of fine grained chlorite-rich rocks occur depositionally(?) interlayered with the conglomerates and quartzites on Cyclops peninsula. The chlorite-rich rocks commonly contain well-preserved pillows, and in these places it is clearly metabasalt. Locally, however, pillows are absent, and the protolith may have been a fine- to medium-grained clastic sediment composed largely of mafic detritus.

The conglomerate, quartzite, and metabasalt have been metamorphosed and deformed under greenschist conditions, producing a subvertical, N-trending foliation and a nearly vertical stretching lineation. The foliation is defined by a penetrative schistosity in the metabasalt and the matrix of the conglomerate. The extension lineation is well-defined by elongation of the cobbles in the conglomerate. Subhorizontal extension fractures filled with fibrous calcite occur commonly in the rocks in the area of Cyclops peninsula. These extension fractures are up to 2 cm wide and 50 cm long. They are generally perpendicular to the vertical foliation and extension lineation.

A 100-m-wide, greenschist-grade phyllonite zone structurally juxtaposes a distinctive, chlorite-rich metagranite against the conglomerate and metabasalt south of Point Lake (Fig. 3). This granite (Augustus granite of Henderson, 1988), dated by Krogh and Gibbins (1978) at 3.15 Ga, is lithologically identical to most of the felsic cobbles in the conglomerate and probably represents the source from which they were derived. Other spectacular exposures of conglomerates occur on the peninsula west of Keskarrah Bay, where they were clearly deposited unconformably on the granite.

Keskarrah plutonic complex

A large exposure of plutonic rocks exists west of Keskarrah Bay (Fig. 3). Several generations of intrusive igneous rocks with compositions ranging from pyroxenite and gabbro to diorite and granite are represented. In marked contrast to the intense, penetrative deformation evident in the gneisses exposed farther west, penetrative fabric in this area is only weakly to moderately developed except in relatively discrete shear zones. In addition, plutonic rocks in the Keskarrah area are metamorphosed to greenschist grade, unlike the amphibolite grade gneisses in the western exposures. Plutonic and gneissic rocks in the Keskarrah area have been divided into two map units, chloritic granite (kgr) and gabbro-diorite (kgd), but continued detailed mapping in the area may allow further subdivision.

Chloritic granite (kgr) dominates the plutonic and gneissic rocks exposed west of Keskarrah Bay. This granite (Augustus granite of Henderson, 1988) is buff-white in weathered outcrop and looks distinctly green on a freshly broken surface. Texturally, this rock is coarse grained and inequigranular, with pale 0.5 to 2.0 cm feldspar crystals in a heterogeneous quartz + biotite + chlorite matrix. Xenoliths of gabbro, diorite, and tonalite gneiss occur locally within the chloritic granite and provide evidence for a generation of rocks older than the granite.

A younger generation of gabbro-diorite (kgd) occurs as dykes and irregularly-shaped plutons that intrude the chloritic granite. The intrusive relationship between the gabbro-diorite and the chloritic granite is well exposed near the western shore of Keskarrah Bay between 65°11'N and 65°11'30"N. This lithology is common in the eastern portion of the plutonic exposure, near Keskarrah Bay (Fig. 3). The gabbro-diorite is a medium to fine grained, equigranular rock metamorphosed to a greenschist grade chlorite±amphibole± plagioclase mineral assemblage. Minor pyroxenite bodies are included in this map unit. The pyroxenite is a coarse-grained, granoblastic rock variably altered to chlorite and amphibole. It occurs as relatively rare dykes and plutons typically <50 m in maximum dimension. Recrystallization to metamorphic mineralogy is more prevalent near the fine grained margins of the pyroxenite intrusions.

Due to the fine grain size and the greenschist-grade metamorphism in this unit, hypabyssal plutonic rocks and metavolcanic equivalents would be difficult to distinguish if clearly intrusive or extrusive characteristics were absent. Consequently, this map unit could also include some massive basalt flows. In fact, a large exposure of dominantly mafic rocks in western Keskarrah Bay (113°1'W; 65°11'30"N) was previously mapped entirely as metabasalt (Henderson, 1988; Kusky, 1991) Well-developed pillows are clearly evident

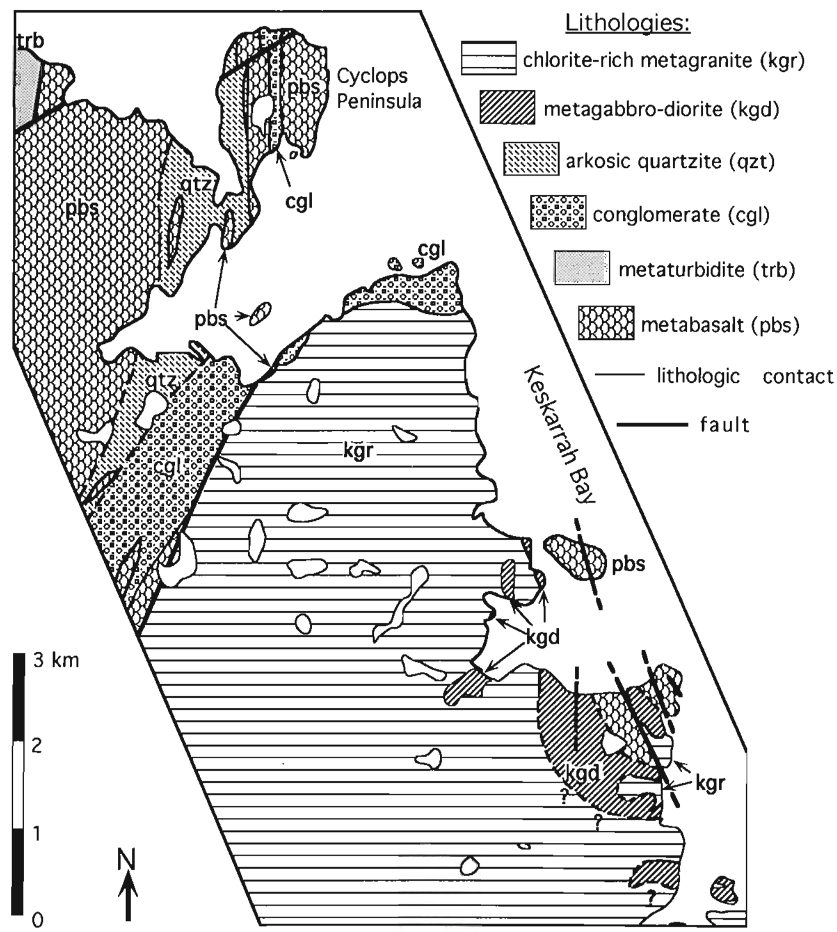

Figure 3.

Simplified geological map of the eastern portion of the area studied, including the structural inlier of plutonic rocks within the greenstone belt, west of Keskarrah Bay.

near the northeastern margin of the mafic rocks; however, they become massive, and then clearly intrusive in the southern portion of the mafic exposure.

Unfortunately, outcrop in this area is generally poor. Narrow, greenschist-grade shear zones occur within the mafic rocks, and they form the contact locally between the plutonic rocks (units kgr and kgd) and the overlying volcanic sequence. The shear senses and magnitudes of displacements on these shear zones are not clear. In other places, no distinct structural boundary between the fine grained gabbro-diorite intrusive rocks and the mafic volcanic sequence can be recognized, although local covered intervals <50 m in width potentially conceal shear zones. One structural repetition of the plutonic rocks exists on the broad peninsula on the western side of Keskarrah Bay (Fig. 3).

STRUCTURAL CHARACTERISTICS AND DEFORMATIONAL FABRIC

Figures 4 and 5 show the distributions and orientations of various structural features in the area mapped. Compositional layering (S_0-S_1) represents the earliest foliation in the gneiss terrane at the western end of the transect (Fig. 4). The original orientation of this foliation is difficult to estimate because the compositional layering is commonly folded and transposed into a second, penetrative foliation (S_2). The folds in the compositional layering are tight to isoclinal and have amplitudes of 10 cm to >50 m. Fold axial surfaces are vertical or dip steeply east and their hinges plunge 25°-50° to the northeast. The axial surfaces are subparallel to the penetrative foliation (S_2), and the fold hinges are generally parallel to the extension lineation found in minor shear zones in the gneisses and the major shear zone at the terrane boundary.

The dominant, penetrative foliation (S_2) is found throughout the area mapped, in both the gneiss and the greenstone terranes, and it is the earliest pervasive tectonic foliation recognized in rocks of the greenstone belt. S_2 is a composite fabric defined locally by transposed compositional layering and aligned phyllosilicates in the gneisses, or by mylonitic foliation in shear zones, or by schistosity in the metavolcanic and metasedimentary rocks. S_2 is subvertical or dips steeply to the east and strikes generally to the north. This foliation is pervasive and well-developed in the gneiss terrane and in the mylonite zone at the terrane boundary, but fabric intensity decreases and becomes progressively more heterogeneous in rocks in the eastern portion of the transect (Fig. 4, 5). S_2 is only moderately developed in the plutonic rocks west of Keskarrah Bay, except in relatively discrete

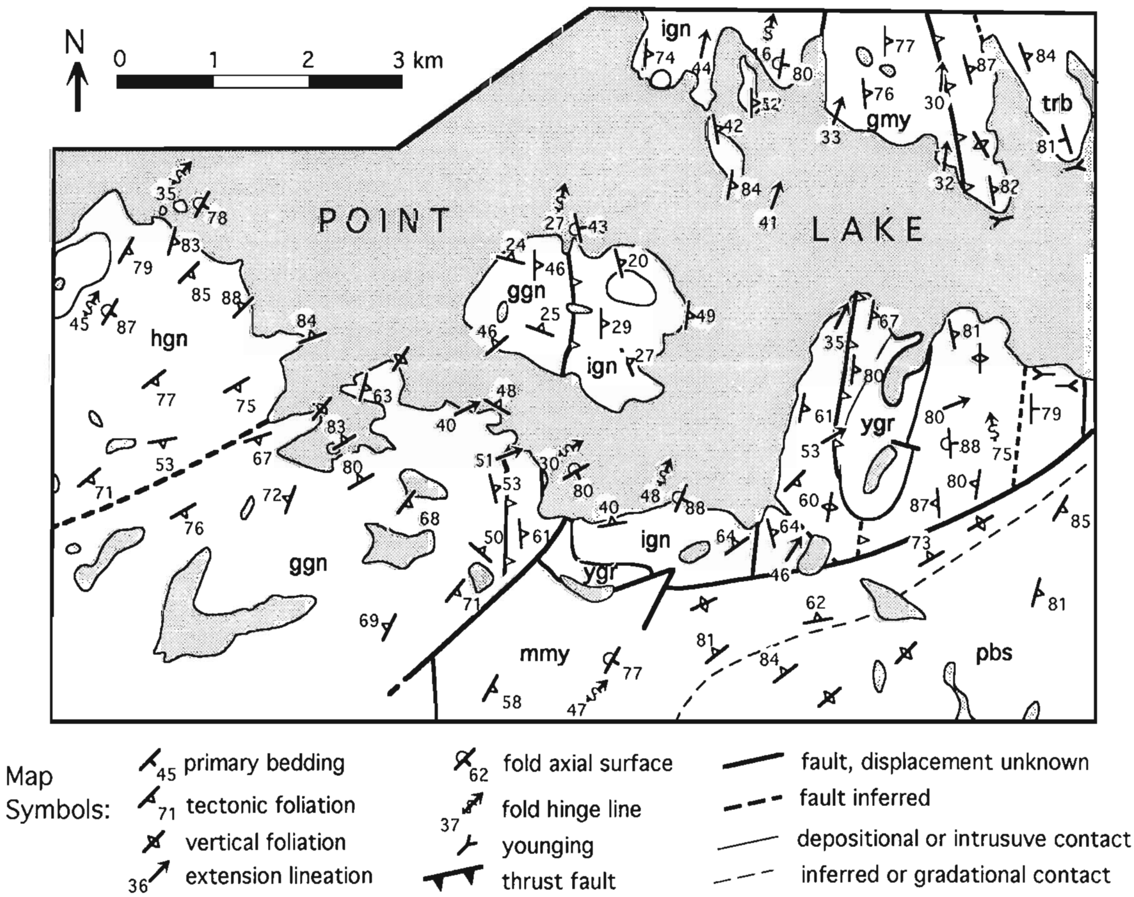

Figure 4. Orientations of tectonic structural features in the western portion of the study area.

shear zones. This foliation is absent or poorly developed in the relatively late, K-rich granite (ygr) that intrudes the gneisses and the mafic mylonite, suggesting that most of the fabric development occurred prior to the intrusion of the granite.

Shear zones

Several mylonitic or phyllonitic shear zones are present in the transect. In general, the major shear zones vary from amphibolite grade in the west to greenschist grade in the east, consistent with the gradient in peak-metamorphic conditions.

Intense foliation and grain-size reduction characterize the mafic mylonite zone at the contact between the gneiss terrane and the greenstone belt, but a well-developed mineral alignment or extension lineation is rarely observed. An extension lineation is clearly visible in granitic mylonite derived from the gneisses just beneath the mafic mylonite, and there the lineation rakes 25°-60°N on the subvertical foliation. Clear and consistent outcrop-scale shear-sense indicators are rare. Occasionally, feldspar porphyroclasts in the granitic mylonite have slightly asymmetric recrystallized "tails" that suggest an east-side-up sense of shear. Locally, however, small shear-bands and asymmetric boudins give the opposite sense of shear in the mafic mylonite a few tens of metres to the east. These observations might be explained by either a complicated motion history on the fault, or a significant component of relatively homogeneous shortening

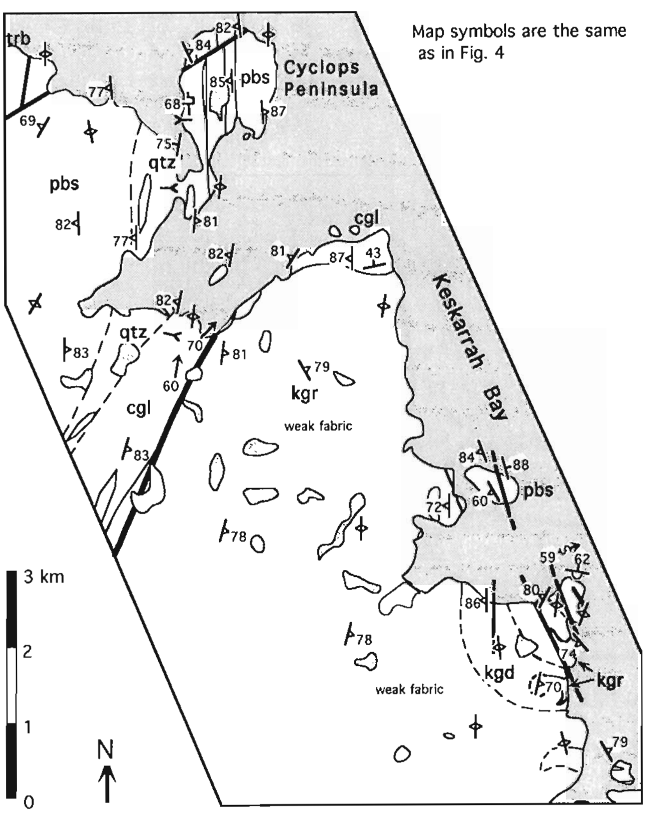

Figure 5. Orientations of structural features in the eastern portion of the study area.

in the finite strain. Subhorizontal extension fractures filled with fibrous calcite are common in the high-strain zones in the metabasalts and conglomerates on Cyclops peninsula. These fractures are perpendicular to the pervasive foliation (S_2), and their combined geometry is consistent with significant east-west coaxial shortening and vertical extension. If this shortening occurred relatively late in the deformational history, it might have overprinted and obscured earlier, simple shear fabric in the fault zones and contributed to the low degree of asymmetry observed in the finite strain. The shear sense in the mylonite zone is difficult to interpret with confidence.

Anastomosing shear zones in the gneisses beneath the mylonitic terrane boundary have a more consistent sense of shear defined by asymmetric feldspar porphyroclasts, S-C relationships, and extensional shear bands. These shear-sense indicators give an oblique, east-over-west transport. The foliation and lineation orientations of these zones are similar to those in the mylonites at the terrane boundary. If the shear zones in the gneisses developed contemporaneously with the terrane boundary shear zone, then the tectonic transport along the boundary might also be east-over-west; however, a genetic relationship between these shear zones is not firmly established, and linking their kinematic history is speculative.

EVALUATION OF EXISTING TECTONIC MODELS

Autochthonous greenstone model: rift volcanics?

As pointed out by previous authors (Hoffman, 1989; Kusky, 1989, 1991), a number of characteristics of the greenstones at Point Lake are not consistent with their formation in fault-bounded rift basins. First, no extensional structures that might be related to rifting have been recognized. In addition, continental rifts typically have a great deal of coarse clastic sediment interlayered with volcanic rocks that range in composition from basalt to rhyolite. The thick volcanic pile at the western margin of the greenstone belt is composed exclusively of metabasalt with little or no interbedded sediment.

Allochthonous greenstone model: obducted oceanic crust?

While this model adequately explains many features of the greenstone terrane, several observations are inconsistent with the structural geometry and tectonic history of the Point Lake transect suggested by Kusky (1991). The proposed geometry of Kusky (1991) interprets the fault on the western side of the Keskarrah plutonic rocks (Augustus granite of Henderson, 1988) as an out-of-sequence thrust which structurally repeats the gneiss seen farther to the west. For this interpretation to be correct, the Augustus granite should be similar in lithological, metamorphic, and deformational character to gneisses in the west, and the eastern boundary of the Augustus granite against metabasalts should be a repetition of the amphibolite-grade mylonitic terrane boundary.

Neither of these relationships, however, is obvious in the field. In fact, the plutonic rocks in the Keskarrah area are lower in metamorphic grade, and the strongly-developed, penetrative deformation characteristic of the gneisses exposed farther west are only moderately developed. In addition, mafic rocks on the eastern margin of the Augustus granite locally have an intrusive or depositional(?) relationship with the granite. In some places the volcanic sequence is faulted against the plutonic rocks by poorly exposed lower greenschist-grade shear zones that vary in width from 1 to 20 m. If these shear zones were structural repetitions of the major, amphibolite grade shear zone which forms the terrane boundary farther to the west, then they should preserve higher P-T conditions of deformation since they would restore structurally to a significantly deeper and presumably hotter level than the amphibolite mylonites exposed farther west. Clearly, the contact zone exposed along the shore and islands of western Keskarrah bay is fundamentally different in character than the >1 km-wide amphibolite-grade shear zone which forms the terrane boundary farther to the west.

IMPORTANT QUESTIONS FOR ONGOING RESEARCH

Several fundamental questions regarding the tectonic evolution of the Point Lake area arise from consideration of the field observations and the tectonic models. These questions include:

- Are the metavolcanic rocks composed exclusively of tholeiitic basalt or do they also contain more evolved magmas? This question has important implications regarding the tectonic setting of the volcanism. Volcanic rocks formed at a mid-oceanic spreading centre would be composed exclusively of tholeiite. In contrast, arc-associated volcanic piles (including those formed in intra-arc and back-arc settings) or continental rift-related volcanic rocks might contain a variable component of intermediate to felsic magmas.

- Are all of the metavolcanic rocks correlative, and did they form during a single volcanic episode or several? In the area of Keskarrah Bay, plutonic rocks and their overlying metavolcanic rocks have lithological, metamorphic, and deformational characteristics which are substantially different from those seen farther to the west – do they represent a different sequence?

- Are the pre-tectonic mafic dykes and plutons which intrude the gneissic rocks related to the metavolcanic lithologies, or do they represent an entirely different (older?) mafic magmatic event?

- What are the ages of metamorphism, faulting, folding, and general shortening experienced by the rocks in the Point Lake transect? Is the finite strain and metamorphism recorded in these rocks the result of a single, progressive deformational event, or have they experienced a polyphase tectonic history?

- What are the provenance and depositional ages of the metaturbidites, greywackes, sandstones, and conglomerates?

Clearly, these questions have important ramifications regarding the tectonic development of this area. Petrographic, geochemical, and geochronological study of samples collected in the field (combined with further detailed mapping to constrain better the contact relationships and deformational features) are critical to address these questions.

ACKNOWLEDGMENTS

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Structural reconnaissance of gneissic tectonites in the western Hepburn Island map area, northernmost Slave Province, Northwest Territories[1]

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Abstract: Gneiss tectonites of the Kangguyak gneiss belt have been affected by three phases of Archean, moderate to very high strain deformation. F1 isoclines are transposed by D2 amphibolite facies annealled mylonite zones that trend 150°. Syn-D2 L2 lineations and F2 fold axes are steeply southwest plunging. D3 produced porphyroclastic gneisses and mylonitic fabrics in granitic intrusions. A north trending zone of D3 dextral transpressive shear is marked by an east and southward increasing strain gradient. The shear zone anastomoses near the Arctic coast, preserving D2 deformation in kilometre-scale low-strain windows. Latest D3 deformation was localized at the Anialik River Greenstone Belt contact which was subsequently reactivated as a brittle-ductile normal fault.

Résumé : Les mylonites gneissiques de la zone de gneiss de Kangguyak ont été modifiées par trois phases de déformation d'intensité moyenne à très élevée au cours de l'Archéen. Les isoclinaux F1 font preuve d'une transposition par les zones mylonitiques, de direction générale 150°, foyer d'une recristallisation dans des conditions de température élevée du faciès D2 des amphibolites. Les linéations L2 contemporaines de D2 et les axes de plis F2 ont un fort plongement vers le sud-ouest. D3 a produit des gneiss porphyroclastiques et des fabriques mylonitiques dans les intrusions syntectoniques. Une zone d'orientation nord de cisaillement transpressif dextre est caractérisée par un gradient de déformation D3 qui augmente à l'est et vers le sud. La zone de cisaillement s'anastomose près de la côte arctique, en conservant la déformation D2 dans les fenêtres de faible déformation d'échelle kilométrique. La phase de déformation la pus récente, D3, était localisée au contact de la zone de roches vertes d'Anialik River qui, par la suite, a été réactivée sous forme d'une faille normale cassante à ductile.

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[1] Contribution to NATMAP – Geology of the Central Slave Province

INTRODUCTION

Two weeks of reconnaissance mapping was conducted in the western Hepburn Island map area (NTS 76M/11) to investigate the nature of a belt of mixed gneiss, informally called the Kangguyak gneiss belt (Fig. 1, 2). Reconnaissance by geologists from the Geological Survey of Canada, Department of Indian Affairs and Northern Development, and Government of the Northwest Territories recognized these rocks to be more intensely deformed and recrystallized than other gneiss units of the map area (Relf et al., 1992). The present study was undertaken to characterize these gneisses, to determine their extent and strain state, and examine their contact relations with the Anialik River Greenstone Belt.

REGIONAL GEOLOGY AND PREVIOUS STUDY

The northwestern part of the Hepburn Island map area encompasses several distinct assemblages (Fig. 2): 1) A western domain of syn- to late-syn- kinematic granitoid (Jackson, 1989); 2) the Kangguyak gneiss belt of predominantly undifferentiated gneisses (Jackson, 1989); 3) the ca. 2702-2682 Ma Anialik River Igneous Complex (Abraham et al., 1991; Padgam et al., 1983); and 4) the Anialik River Greenstone Belt (Tirrul and Bell, 1980).

Previous study of the map area includes 1:500 000 scale mapping by Fraser (1964), and 1:50 000 scale mapping by Yeo et al. (1983). The belt has been generally interpreted as higher grade, reworked equivalents of the Yellowknife Supergroup (Jackson, 1989). Orthogneisses on the Coronation Gulf coast were interpreted as mixed Yellowknife Supergroup and gneissic Anialik River Igneous Complex (Jackson, 1989).

KANGGUYAK GNEISS BELT

The Kangguyak gneiss belt can be subdivided into 3 units traceable from the Arctic coast south for 10 km (Fig. 3a). The main units are: 1) hornblende-biotite tonalite orthogneiss, 2) quartz-plagioclase-biotite gneiss, and 3) granodiorite

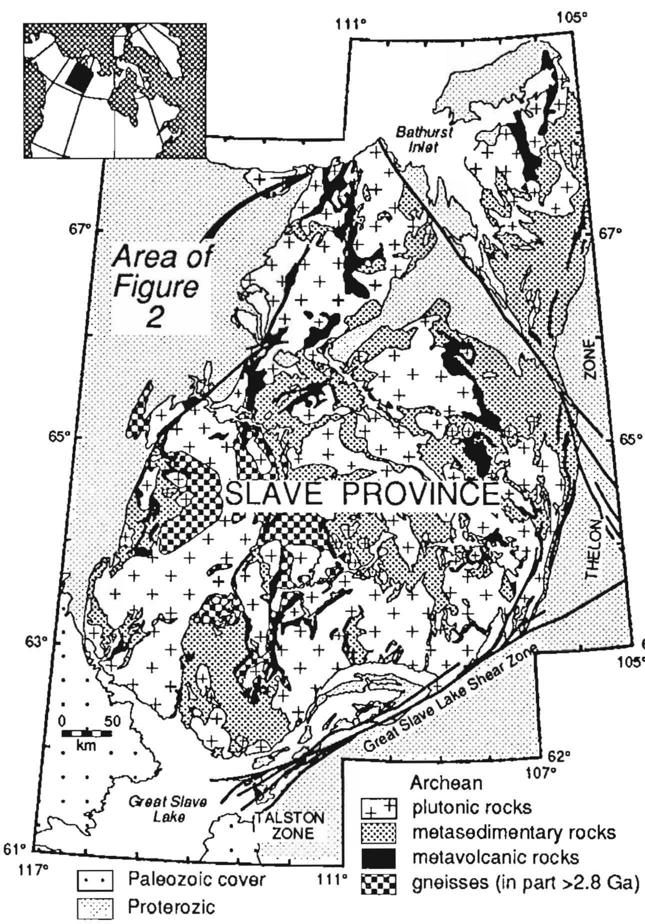


Figure 1. Location of the study area within the northern Slave Structural Province, modified from Hoffman (1989). Arrow points to the area of Figure 2.

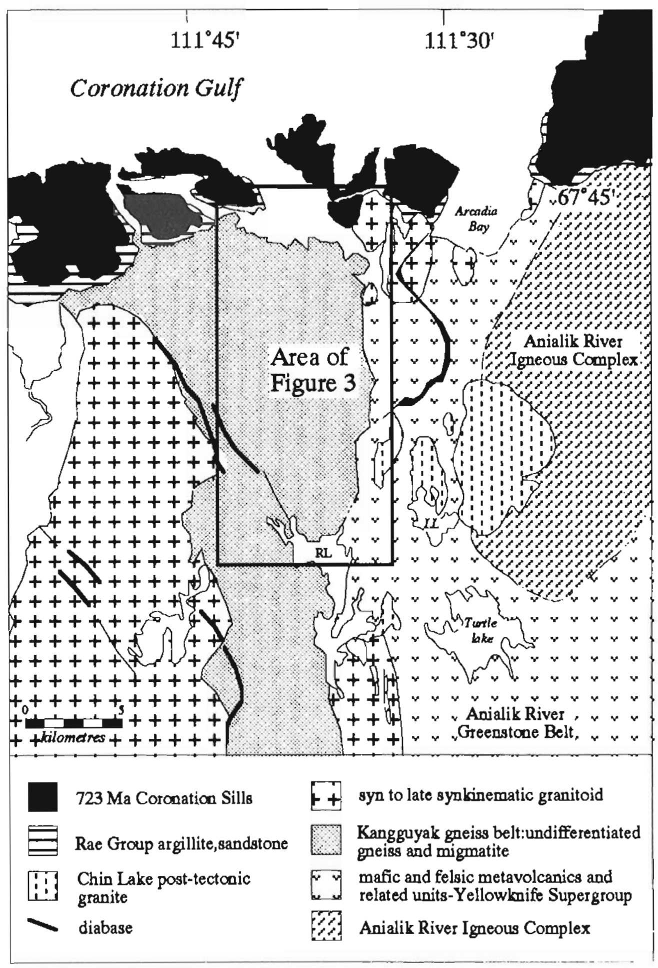


Figure 2. Generalized geology of the northwestern Hepburn Island map area (76M) after Jackson (1989). Black rectangle outlines the area of Figure 3. RL= Rhino lake, LL= Locana lake.

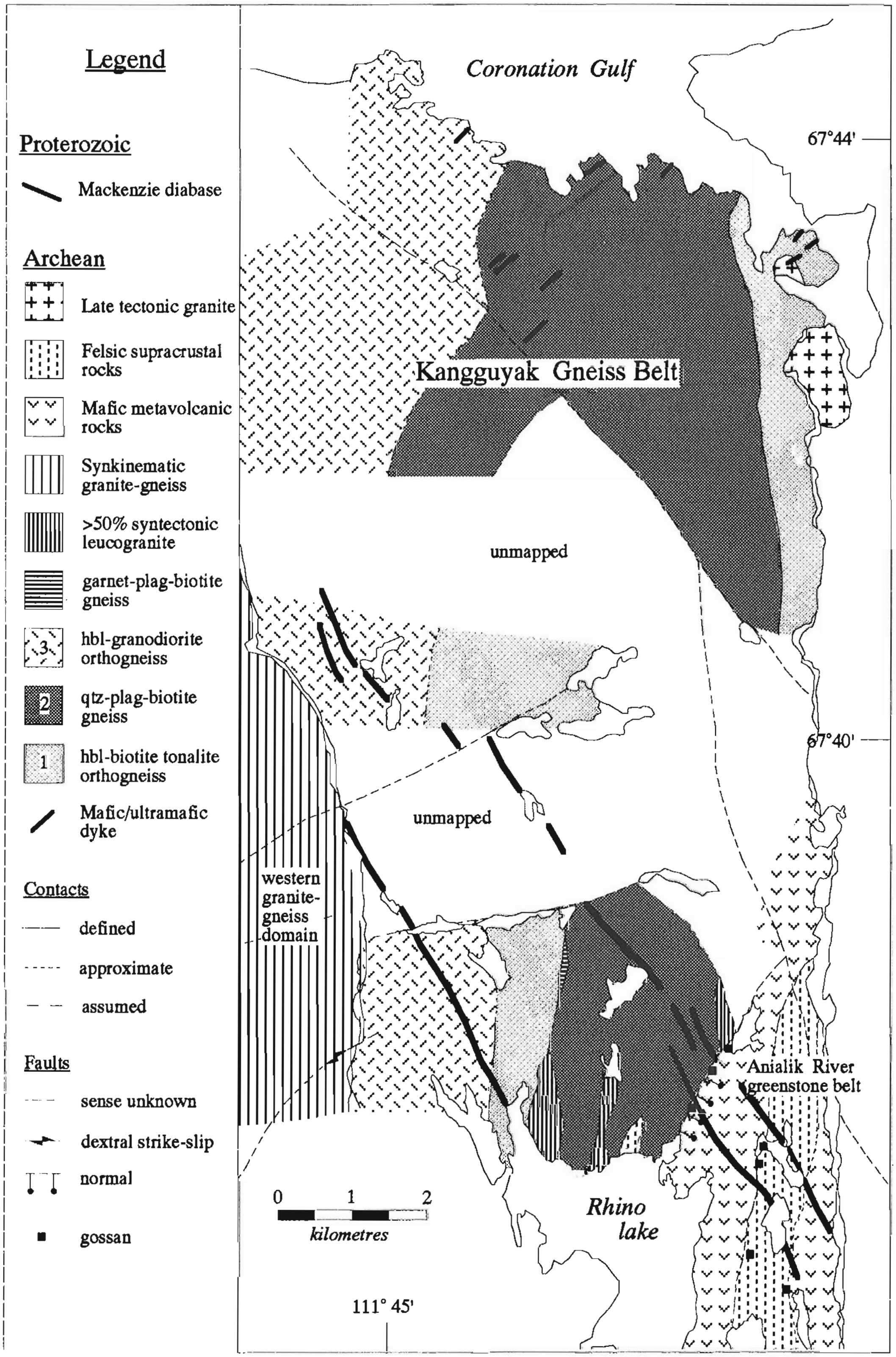


Figure 3a. Geology of the northern Kangguyak gneiss belt. Greenstone belt lithology east of Rhino lake after Jackson (1989). Western granite-gneiss domain after Barrie (1993).

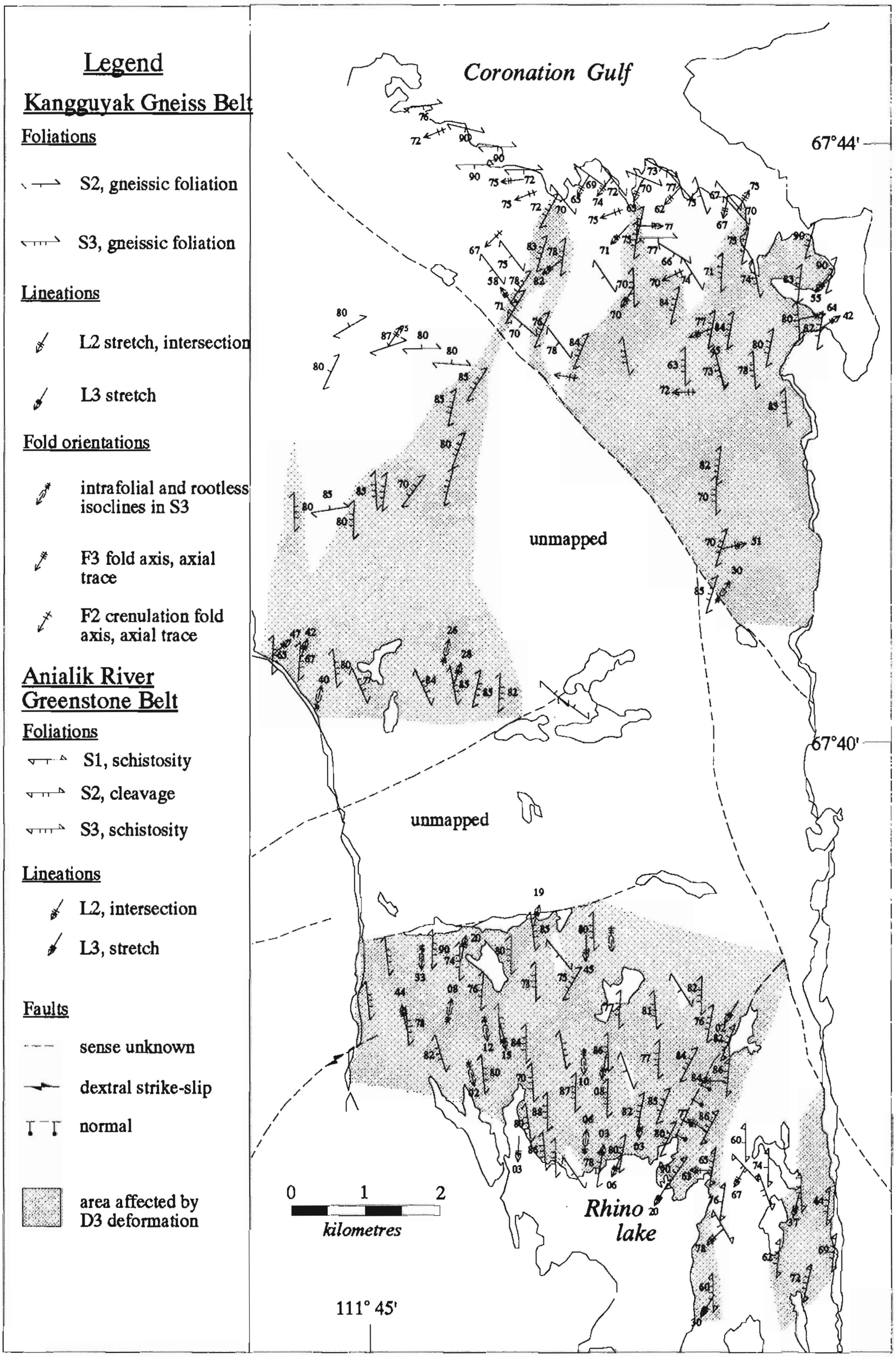


Figure 3b. Simplified structural geology of the northern Kangguyak gneiss belt and adjacent areas. Shaded area outlines zones of penetrative D3 ductile deformation.

orthogneiss. In the following sections the lithological units will be described, and the structure and deformation state of the belt outlined.

Lithological units

Unit 1: tonalite orthogneiss

This unit is a granoblastic, medium grey biotite-hornblende tonalite orthogneiss. The very strongly foliated gneiss has a coplanar compositional banding marked by concordant 0.5 to 1 cm thick bands of leucogranite and local narrow amphibolite bands. The proportion of discrete leucogranite comprising the orthogneiss varies from 10 to 40%.

The leucogranite bands are derived from attenuated and transposed syntectonic pegmatite dykes and intrusive bodies. Discrete crosscutting boudined and/or folded dykes are transposed to foliated subconcordant bodies with partly disaggregated intrusive apophyses (Fig. 4a). Where completely transposed, the pegmatites are reduced to large (2-15 cm) feldspar augen connected by 0.5 cm thick tails of recrystallized quartz and feldspar. Individual augen can separated by 3 m or more (Fig. 4b). Some or all these stages of transposition with the orthogneiss can be visible within a single outcrop.

Unit 2: quartz-plagioclase-biotite gneiss

The second unit is a dark grey, quartz-plagioclase-biotite gneiss of generally quartz diorite composition. It has a pronounced 0.5 cm wide compositional banding outlined by variation in the proportion of biotite. A locally garnet-muscovite-bearing, white leucogranite comprises 20 to 60% of the unit. Associated pegmatite occurs as veins, pods, and rafts, isoclinally folded and disrupted segments and metre-long boudins connected by centimetre-thick trails of recrystallized feldspar (Fig. 4c). Where leucogranite comprises the greater proportion of the unit the biotite gneiss is present as discrete, metre-scale boudins.

Unit 2 biotite gneiss also includes 0.5-1 m thick mafic pods, rafts, and segments. Generally the narrowest, shortest segments are altered to a distinctive bright copper green and weather recessively and rusty. These segments have an actinolite/talc/chlorite composition and may be disrupted mafic dykes.

Where the quartz-plagioclase-biotite gneiss, pegmatite, and mafic segments are transposed together, the resultant rock is an intermediate-composition gneiss tectonite with annealed mylonitic banding (Fig. 4d). South of the coast unit 2 is comprised of quartz-plagioclase-hornblende-biotite gneiss transposed with boudined white pegmatite (Fig. 4e). The pegmatite is disaggregated within the gneiss, forming a feldspar porphyroclastic gneiss tectonite with a hornblende-biotite matrix. Isolated green-weathering mafic segments are present.

Unit 3: granodiorite orthogneiss

The third unit comprises reddish gray, biotite-hornblende granodiorite orthogneiss. Gneissic banding is defined by compositional segregation of the granodiorite and by 2-15 cm thick granite and/or pegmatite veins, some of which are attenuated to trails of feldspar augen. The banding in this unit is more gradational and is markedly less rectilinear than that of units 1 and 2. Interlayered boudined amphibolite bands are common. Coarse diorite boudins, 0.5 to 2 m in length, occur in several traceable horizons and are deformed with unit 3.

Unit 3 is intruded by two, shallowly crosscutting to coplanar granite and pegmatite phases which locally comprise 20 to 30% of the gneiss. A white pegmatite phase, similar to that of unit 2, intrudes unit 3 north of Rhino lake. Dykes of this phase commonly cut the gneissic foliation at a moderate angle.

Late mafic dykes

All units at the coast are crosscut by a northeast-trending mafic dyke swarm (Fig. 3a). At least forty dyke segments were recognized, up to 30 m long and ranging in width from 0.5 to 4 m. Typically they are black, medium grained amphibolite.

Late granitoid suite

A late hornblende-biotite granite suite intrudes both the gneiss and greenstone belts. An associated mafic phase occurs as xenoliths with chilled and cuspate margins and selvedges of mixed composition. Units of the suite are mildly foliated to isotropic and carry variably oriented blocks of deformed orthogneiss (Fig. 4f).

Structure

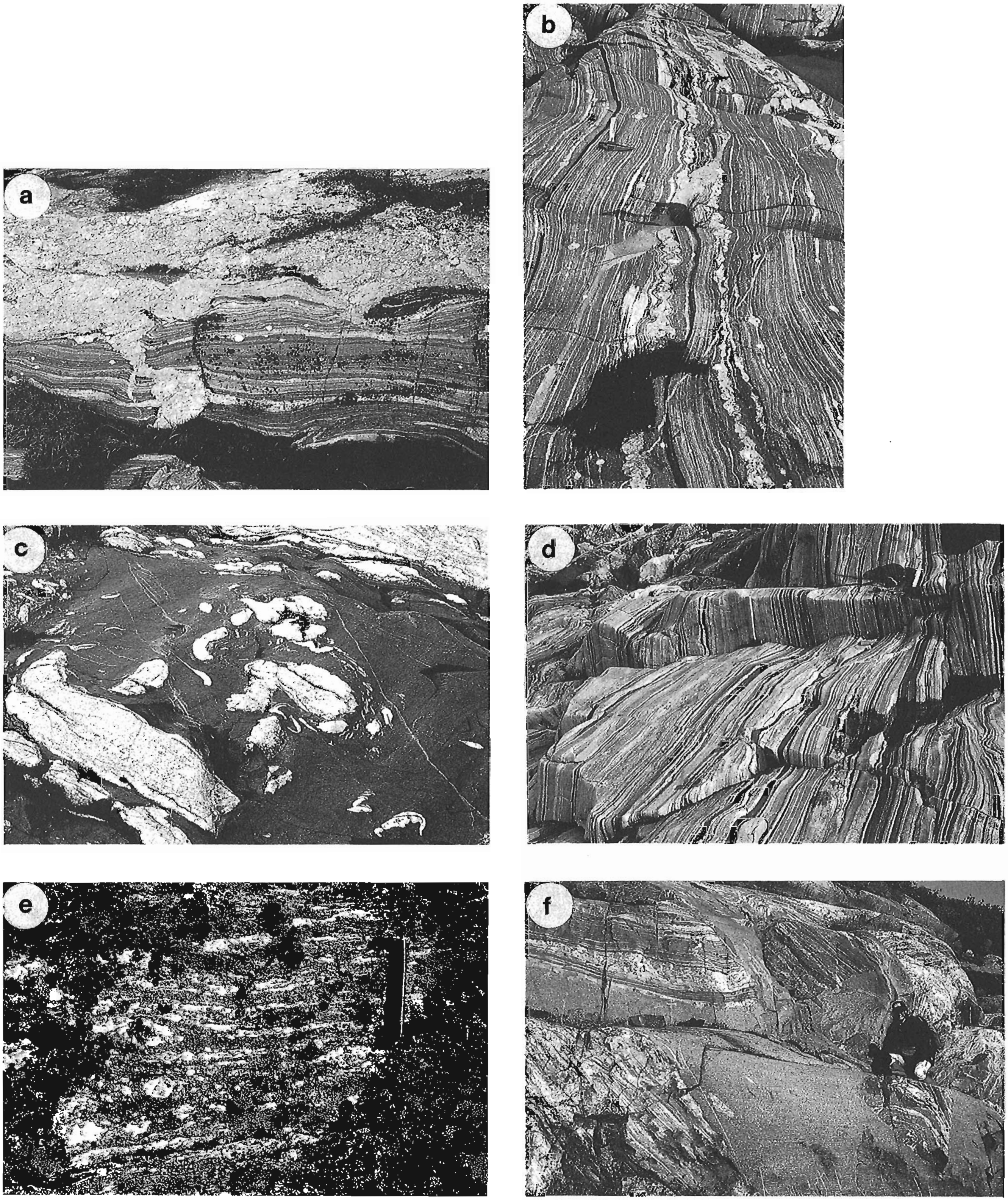
Rocks of the Hepburn gneiss belt have been affected by at least two, and possibly three, phases of moderate to very high strain Archean deformation. Data are insufficient at present to determine if the first set of structures represent a distinct D1 phase of deformation or represent an earlier phase related to the second group of structures. For clarity they will be assigned to a D1 phase of deformation, recognizing that this phase may be continuous with D2.

D1 deformation structures

The first fabric, S1, is a well developed foliation of unit 2 within boudins affected by S2. Aplite vein networks within the quartz-plagioclase-biotite gneiss are attenuated and transposed into parallelism with the S1 foliation. The first folds, F1, occur as coaxially refolded isoclines within type 3 interference patterns and as intrafolial isoclines within S2.

D2 deformation structures

The dominant fabric within the strongly recrystallized coastal units, S2, is a gneissic foliation regionally oriented at 150°/57° (Fig. 5c). Farther south, where the effects of D3 deformation are more penetrative, the 150°-trending S2 foliation is preserved in small, low-strain pods isolated by the anastomosing S3 foliation. Mappable concentrations of the pods are outlined on Figure 3b.



Parallel to the trend of S2 are high-strain zones derived from the transposition of units 1 and 2. These zones are comprised of straight gneiss tectonite characterized by: 1) extremely rectilinear gneissic banding in which the Y-Z strain section (perpendicular to the foliation and lineation) and X-Z section (perpendicular to the foliation but parallel to the lineation) both show strong attenuation of pre-existing interfaces; 2) transposition of early crosscutting features into parallelism with the shear plane; 3) extreme attenuation or disaggregation of layers; and 4) development of well defined compositional layering with very sharp interfaces.

The straight gneiss zones vary from 2-10 m in width and anastomose on the outcrop scale, bounding lower strain lozenges within which the gneiss is irregularly and complexly folded. Transposition of pegmatite dykes and bodies with the gneiss units produces a more intermediate to felsic composition gneiss banded on a centimetre scale (Compare Fig. 4c and d). The S2 straight gneiss zones are essentially annealed mylonite zones which result from post-D2 high temperature recrystallization of the mylonitic foliation.

Figure 4.

a. D2 deformed tonalite orthogneiss illustrating isolated relict feldspar porphyroclasts which are the end product of high strain transposition of syntectonic pegmatite. Shallowly oblique pegmatite is partly transposed (upper right) but clearly crosscutting the annealed mylonitic foliation, note intrusive apophysis. Further increments of high strain would reduce this body to similar isolated feldspars. GSC-1992-252G

b. Annealed mylonitic tonalite orthogneiss containing pegmatite of various stages of disaggregation. Centre field pegmatite is boudined but intact. Fist-sized feldspar augen in upper left of view are connected by a trail of recrystallized felspar and are separated by several feet. Isolated porphyroclasts are visible throughout. Hammer is 40 cm long. GSC-1992-252E

c. Boudined, folded, and torn-apart pegmatite within unit 2 biotite gneiss. Tight to isoclinal folding of disrupted pegmatite illustrates the high degree of strain. Hammer is 40 cm long. GSC-1992-252C

d. Annealed mylonite zone within unit 2, derived from the transposition of protolith as in 4c. An increasing strain gradient can be walked within 10 m from 4c to 4d. Note the rectilinear transposed layering within both the Y-Z (face-up foreground) and X-Z (behind hammer) strain sections. The shear plane, determined by the normal to the strain gradient, trends 140°. Hammer is 40 cm long. GSC-1992-252D

e. Porphyroclastic gneiss texture developed by the disaggregation of the unit 2 white pegmatite within the biotite gneiss matrix. Pen is 14 cm long. GSC-1992-252O.

f. Late granite (grey) carries variably oriented xenoliths of D3 deformed tonalite orthogneiss host rock. Field assistant for scale. GSC-1992-252A.

The S2 fabric carries a steeply-plunging mineral and aggregate stretching lineation, $L2_s$ (Fig. 5d) The mineral assemblages defining S2 include hornblende-K-feldspar-plagioclase-quartz in the orthogneiss and titanite-oligoclase-hornblende indicating deformation under amphibolite-facies conditions.

The D2 deformed gneisses are S>L tectonites which have undergone noncoaxial deformation. However, too few unequivocal shear-sense indicators have been observed in this brief reconnaissance to define the movement history.

The S2 gneissic layering is affected by tight, angular S2 folds which occur as outcrop-scale folds in units 1 and 3, and as penetrative, centimetre-scale crenulations in unit 2. All F2 folds are steeply plunging to the southwest, with a mean fold axis orientation of 253°-69° (Fig. 5d) and have an associated intersection lineation. A type 3 interference fold pattern is found where F2 refolds the F1 isoclines (Fig. 6a, Ramsay, 1967). Doubly plunging and sheath F2 folds were observed with hinges parallel to $L2_s$ (Fig. 6b).

D3 deformation structures

The third fabric, S3, is a strongly developed gneissic to mylonitic foliation with a mean orientation of 194°/75° (Fig. 5a). Near the coast, S3 is localized in anastomosing zones which bound domains of preserved D2 deformation. Farther south S3 is the predominant fabric. In contrast to the S2 mineral assemblage, the assemblage defining S3 contains hornblende and biotite in all units. At present there is not enough data to determine whether this represents regionally lower grade D3 metamorphism or continued amphibolite facies deformation under more fluid-rich conditions.

S3 fabrics are characterized by porphyroclastic feldspar textures and the development of quartz ribbons in pegmatite and granite sheets coplanar to straight gneiss foliations. The porphyroclastic texture is best developed in unit 2 where its characteristic white pegmatite pods are reoriented to the S3 190° trend. The pegmatite is variably disaggregated within the biotite gneiss matrix, reducing it to abundant, isolated porphyroclasts and connected trails of augen. The high-strain nature of the porphyroclastic tectonite is outlined by the complete disaggregation of the pegmatite to individual augen and the presence of quartz ribbons (Fig. 6c).

Near the coast, an L3 stretching lineation, defined by quartz ribbons and elongate feldspar aggregates, is moderately plunging within S3 (Fig. 4b). In the south, L3 varies from moderately north plunging north of Rhino lake to shallowly south plunging or subhorizontal closer to the volcanic belt margin.

Northwest of Rhino lake F3 folds are commonly tight, angular folds of S3 with moderately plunging fold axes parallel to L3 (Fig. 6d). Farther east several minor folds of the S3 fabric with fold axes at a high angle to L3 were observed. These have moderately to steeply south-southeast-plunging fold axes and a 'Z' asymmetry viewed down-plunge.

Unequivocal D3 shear-sense indicators were not observed near the coast. To the south, displacement associated with L3 is interpreted to be dextral based on type IIa asymmetric pull aparts (Hanmer, 1986), asymmetric extensional shear bands (Platt and Vissers, 1980), rotated inclusions and C/S fabrics (White et al., 1980). The late F3 folds, although consistent with dextral movement, are not part of a widely observed fold set and accordingly are not given kinematic significance at this time.

Brittle-ductile deformation

Straight gneiss and mylonitic leucogranite of the gneiss belt/ greenstone belt contact north of Rhino lake are strongly calcified and overprinted by brittle deformation along a north-northeast-trending fault zone. Chlorite lineations overprinting the volcanic rocks adjacent to the fault are downdip and chlorite slickensides along the fault give greenstone- belt-down sense of movement. Pseudotachylyte was noted within

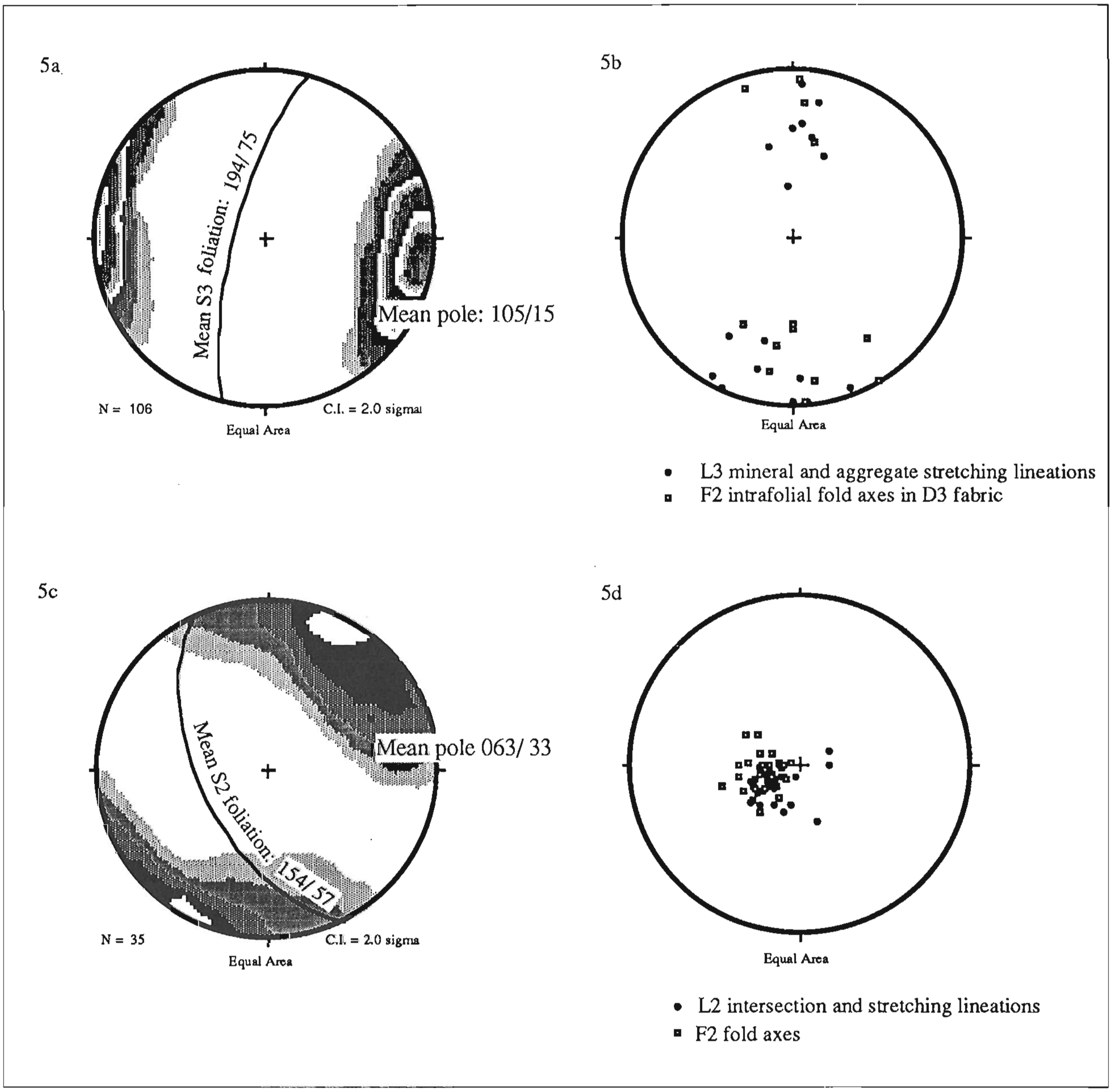


Figure 5. Stereonet projections of structural data from the Kangguyak gneiss belt. **a)** Contoured poles to S3 foliation. **b)** L3 lineations and intrafolial isoclinal fold axes in S3. **c)** Contoured poles to S2 foliation. **d)** L2 lineations and F2 fold axes. Kamb contours, lower hemisphere equal area projections.

the fault zone. In addition, granite mylonite with greenstone-belt-down shear-sense indicators (dextral C/S fabrics along 70°S plunging lineation) was observed at one locality near the Arctic coast. Brittle-ductile faults of this type in the map area are generally interpreted to be Proterozoic features (Tirrul and Bell, 1980).

ANIALIK RIVER GREENSTONE BELT

The Anialik River Greenstone Belt was examined at two localities; on the northeastern side of Rhino lake and on the Arcadia Bay coast, in order to evaluate the nature of its contact with the Hepburn gneiss belt (Fig. 2). At least 2 sets of fabrics are recognized in the volcanic belt, while 3 sets were noted near the western margin.

In the eastern part of Arcadia Bay phenocrysts in metabasalt and flattened fragments and quartz eyes in felsic porphyry are aligned in a foliation, S1, averaging 142/86°. The porphyry is also affected by folds with axial traces trending 240°. A crenulation cleavage also trending 240°, S2, overprints the foliation in the metabasalt. Towards the west the foliation in the porphyry is more strongly developed, averages 180°/75° in orientation, and contains a well developed 305°/85° lineation defined by stretched quartz eyes. Data are insufficient to determine whether this is an S3

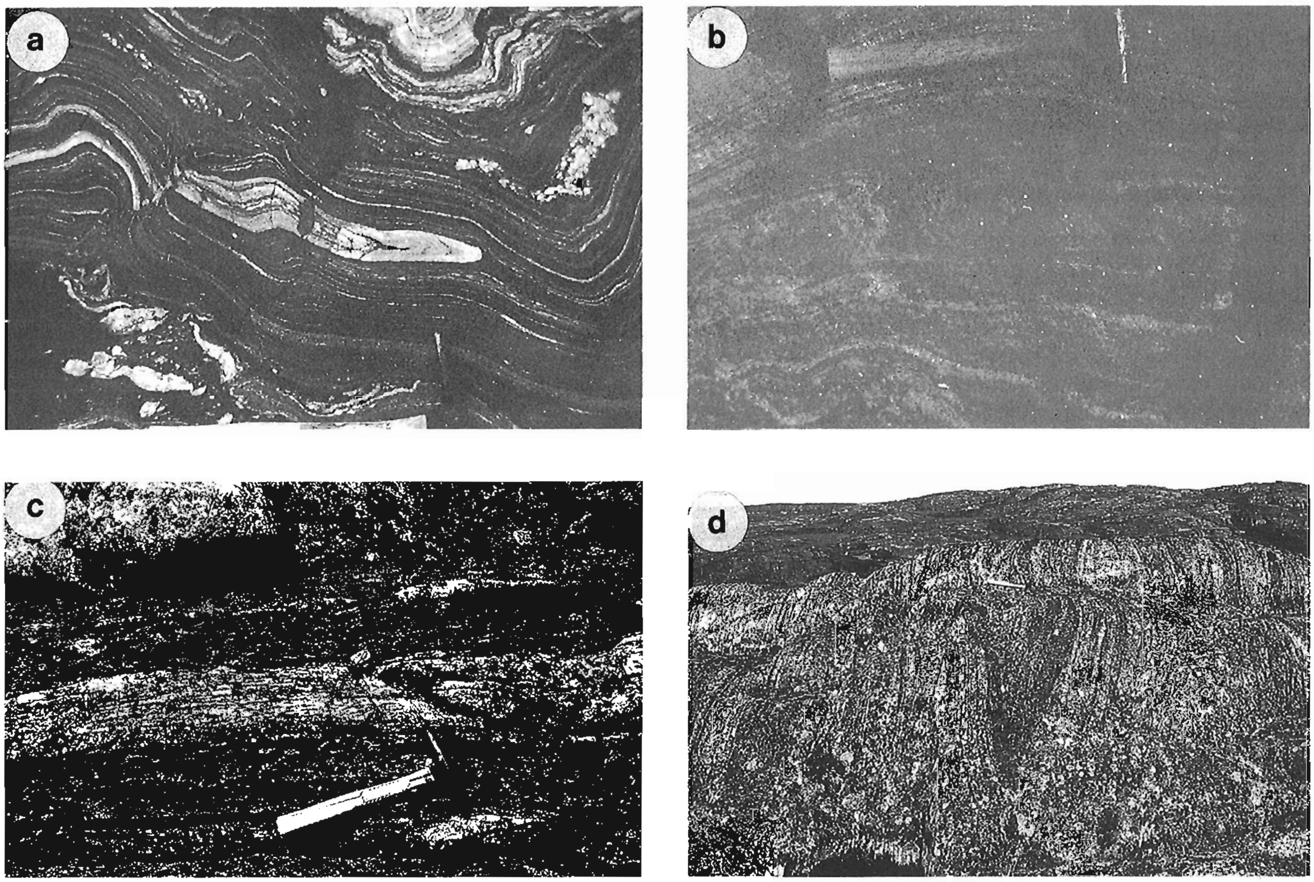


Figure 6

a) Type 3 interference fold pattern within unit 2. F1 isocline with steeply plunging fold axis parallel to the stretching lineation is refolded by steeply southwest plunging F2 folds. Hammer is 40 cm long. GSC-1992-252J

b) Sheath fold within highly strained unit 3 granodiorite gneiss. The bisector of the sheath fold is parallel to the down-dip stretching lineation. Hammer, at bottom, is 40 cm long. GSC-1992-252K

c) Highly strained white pegmatite within biotite schist (unit 2) northeast of Rhino lake. All stages from transposed dyke (note elongation of dark quartz ribbons in central band) to relict porphyroclast (centre above dyke) are present. Hammer is 40 cm long. GSC-1992-252N

d) Upright F3 fold of transposed unit 2. North-trending S3 foliation carries a moderately south-plunging stretching lineation which is subparallel to the fold hinge. View looking north. Hammer is 40 cm long. GSC-1992-252F fabric or a more highly strained S1. The fabrics in the volcanic units are cut by granites of the late suite and mafic dykes of the northeast-trending swarm.

The margin of the greenstone belt east of Rhino Lake comprises amphibolite-facies mafic volcanic schist and felsic volcanic gneiss. The first fabric, S1, is a 190° trending schistosity which dips approximately 60-70° east. A regional S2 cleavage trending 150°-160° and associated L2 intersection lineation, overprint S1. S2 is axial planar to moderately southeast-plunging Z folds of S1. A strong gneissic to mylonitic foliation, S3, trends 190-200°, parallel to the contact with the gneiss belt. Close to the contact, steeply dipping S3 anastomoses around lower strain corridors which preserve the steeply plunging L2 lineation. The associated stretching lineation, L3, plunges shallowly 20°-30°S. Extensional shear bands within highly sheared mafic volcanic rocks suggest dextral sense of shear along the shallowly oblique L3 lineation.

DISCUSSION

High strain nature of deformation

The rocks of the Kangguyak gneiss belt are gneiss tectonites which have experienced several phases of high-strain deformation. The high-strain nature of D2 is evident from the development of annealed mylonite (straight gneiss) zones and the occurrence of F2 sheath folds. In addition to the presence of straight and porphyroclastic gneiss tectonites and quartz ribbons in mylonitic leucogranite, D3 deformation is characterized as high-strain due to the ubiquitous reorientation of F2 and early F3 fold axes parallel to L3. This relationship between fold axis and stretching lineation orientations is a feature characteristic of high strain zones (Bell, 1987; Cobbold and Quinquis, 1980)

D3 strain gradient

D3 deformation has resulted in north-trending, subvertical zones of S>L transposition with an increasing strain gradient noted eastward toward the contact with the greenstone belt.

Effects of D3 are noted at least 5-6 km west of the contact where intrafolial F2 chevron and interference folds with fold axes parallel to L3 are found. Unit 3 in this area has a less well developed straight gneiss foliation which is folded by upright, tight F3 folds. Progressing eastward, F2 folds and upright F3 folds are increasingly attenuated and transposed, until only intrafolial isoclines are observed. The increased transposition of F3 folds results in upright isoclines with highly attenuated limbs visible in the Y/Z section and a completely transposed aspect in the X/Z section. With increasing strain the folds are disrupted to rootless fold hinges. In the north, penetrative D3 deformation is concentrated within 3 km west of the greenstone belt contact. Near Rhino lake porphyroclastic gneiss tectonites and quartz ribbons in leucogranite are increasingly well developed within 3-4 km west of the contact.

Localization of deformation

The occurrence of an eastward-increasing D3 strain gradient suggests that the deformation was localized in a north-trending shear zone adjacent to the greenstone belt. Preliminary study of the D3 deformation indicates it resulted from oblique dextral shear with a strong component of east-west shortening, potentially indicative of a dextrally transpressive regime.

The development of mylonitic fabrics in pegmatite suggests that latest D3 deformation localized at this margin. North of Rhino lake isotropic dykes of the suite cuts across S3 at 90°. Within several tens of metres of the contact the dykes are completely transposed within S3. In addition, there is a spatial correlation between shallowing of the L3 plunge and distance west of the contact, perhaps indicating a transition to strike-slip displacement with increased localization.

Age constraints on deformation

Previous work on the unit 1 tectonites indicated that the units might possibly be Proterozoic in origin (Relf et al., 1992). A titanite age of 2.64 Ga (M. Villeneuve, pers. comm., 1992) on a post-D2, late tectonic, pegmatite dyke shows that this unit must be Archean. In addition, the late granite suite which contains D3 deformed unit 1 xenoliths is generally correlated in age with the 2602 ± 2 Ma Chin Lake stock (Abraham et al., 1992) suggesting that all phases of moderate to high grade deformation within the belt are Archean in age. The timing of brittle-ductile normal faulting which subsequently reactivated the gneiss belt/greenstone belt contact is presently unconstrained.

Future study

The rocks of the Kangguyuk gneiss belt which continue south of Rhino lake may be an extension of the D3 deformation zone described here. More detailed study is needed to outline the distribution of the gneiss tectonites, to better characterize their deformation history, and to place them in a regional context. Preliminary examination of the greenstone belt rocks suggests that they too have seen the effects of the D3 deformation, although through a more restricted zone. Careful correlation of structures and timing of deformation is required to determine the extent to which the gneiss belt and greenstone belt share a common history.

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Geological Survey of Canada Project 870008

Initial observations on Archean and early Proterozoic deformation in the granitoid-migmatite terrane of Hepburn Island map area, northwestern Slave Province, Northwest Territories[1]

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Abstract: At least three Archean and two early Proterozoic deformation events are recorded in granitoid-migmatite terrane of the northwestern Hepburn Island map area (76M/12). To the east, S_1 gneissic fabrics are folded by kilometre-scale, D_2 south-southwest-plunging open folds. D_3 deformation produced north-trending straight gneiss with subhorizontal stretching lineations and fold axes, that everywhere cut or transpose previous fabrics. Early Proterozoic (ca. 1.9 Ga) compressional deformation under biotite grade metamorphic conditions compressed Epworth Group strata against and under the Archean basement, with horizontal shortening from tens to hundreds of metres. East-directed compression deformed the basement and created chloritic fabrics and shear zones parallel to pre-existing gneissic fabrics. The Archean basement and contact were later cut by conjugate faults parallel to those of the Wopmay Orogen.

Résumé : Au moins trois épisodes de déformation survenus pendant l'Archéen et deux épisodes de déformation survenus pendant le Protérozoïque précoce ont été relevés dans le terrane à granitoïdes et migmatites du secteur cartographique du nord-ouest de l'île Hepburn (76M/12). À l'est, les fabriques gneissiques S_1 sont déformées par des plis D_2 ouverts, d'échelle kilométrique, de plongement sud-sud-ouest. La déformation D_3 a produit un simple gneiss de direction générale nord à linéations d'étirement et axes de plissement subhorizontaux, qui partout traversent ou transposent les anciennes fabriques. Une déformation compressive survenue pendant le Protérozoïque précoce (il y a environ 1,9 Ga) dans des conditions correspondant au faciès de la biotite a comprimé les strates du Groupe d'Epworth au contact du socle archéen et au-dessous, en causant un raccourcissement horizontal de plusieurs dizaines à plusieurs centaines de mètres. Une compression dirigée vers l'est a déformé le socle et a généré des fabriques chloriteuses et des zones de cisaillement parallèles aux fabriques gneissiques antérieures. Le socle et le contact archéens ont ensuite été recoupés par des failles conjuguées parallèles à celles de l'Orogène de Wopmay.

[1] Contribution to NATMAP – Geology of the Central Slave Province

INTRODUCTION

This summary presents the initial results of lithological and structural mapping in the granitoid-migmatite terrane of the northwestern part of the Hepburn Island 1:250 000 map area, located in the northern Slave Province (Fig. 1). This work builds on previous studies in the area, including regional mapping by Yeo et al. (1983), mapping of lower Proterozoic rocks and basement contacts in the Tree River fold belt by Hoffman et al. (1984), and mapping and compilation of the Hepburn Island map sheet by Jackson (1989). Evidence is given here for three Archean and two early Proterozoic deformation events within the Archean rocks of the Slave Province. Special attention is given to early Proterozoic deformation of Archean rocks, due to the presence of, and proximity to an exceptionally well-exposed contact between lower Proterozoic strata and Archean basement in the Tree River area.

This study, part of a NATMAP Slave Province Project in the Hepburn Island map area, was conducted over four weeks during the 1992 field season. Two other studies in the Hepburn Island map sheet were initiated in 1992 field season: (1) a geochemical and isotopic study of intrusive and volcanic rocks in and adjacent to the Anialik River and High Lake volcanic belts, in conjunction with ongoing NATMAP projects by C. Relf (GNWT), J.R. Henderson (GSC) and geochronological studies by M. Villeneuve (GSC); and (2) structural, petrological and geochemical studies of granitoid-migmatite terrane in the remainder of the Hepburn Island map area. A complimentary structural study of the multiply-deformed gneissic rocks along the northwestern edge of the Anialik River volcanic belt is given by McEachern (1993).

ROCK TYPES

The geology of the area mapped in 1992 is presented in Figure 2. Granitoid and migmatitic rocks of Late Archean age (Jackson, 1989) comprise most of the area. Lower Proterozoic (1.9 Ga) Epworth Group sedimentary rocks include the Odjick and Rocknest Formations in the Tree River fold belt, part of the 'externides' of the Wopmay Orogen to the west. They are a pre-orogenic, passive margin sequence comprising shelf facies stromatolite-bearing ferroan dolomite and fine to coarse siliclistics (Hoffman et al., 1984). These rocks are cut by mafic dykes of the Mackenzie dyke swarm (1.27 Ga: LeCheminant and Heaman, 1989). Upper Proterozoic Rae Group siliciclastic rocks and the Coronation/Franklin diabase sills (0.7 Ga: Heaman et al., 1992) unconformably overlie these rocks to the north. Descriptions for the Late Archean age rocks are given below, and in McEachern (1993).

Granitoid and other intrusive rocks

Three principal granitoid rock types are included within the "granite" unit in Figure 2, and will be subdivided pending petrographic study. Alkali feldspar granite is medium- to very coarse-grained, with 25-40% quartz, and less than 10% biotite. It comprises approximately 60% of the north-trending granite body at the Epworth Group contact, with biotite monzogranite comprising the bulk of the remainder. Biotite hornblende monzogranite is medium- to coarse-grained with up to 15% biotite, and locally with up to 5% hornblende. It is more common in the southeastern quarter of the area, and the majority of the large granitoid intrusion south of "Rhino" lake (informal) is biotite hornblende monzogranite. Alkali feldspar granite and biotite hornblende monzogranite are commonly gradational with one another. Leucogranite is coarse grained to pegmatitic, with up to 5% biotite and 40% quartz. It cuts all other rock types. Leucogranite occurs as dykes, and along with alkali feldspar granite as sill-like sheets up to 200 m thick, particularly in the northwestern part of the area. In the northwestern area, the "granite" map units have gradational contacts with migmatitic rocks (except where the contact is defined by a fault); granite and migmatite rock units are divided at a 75% granitoid rock content. At greater than 75% granitoid rock content, orthogneiss and amphibolite blocks are generally randomly oriented and represent inclusions within granite, whereas at lower granitoid

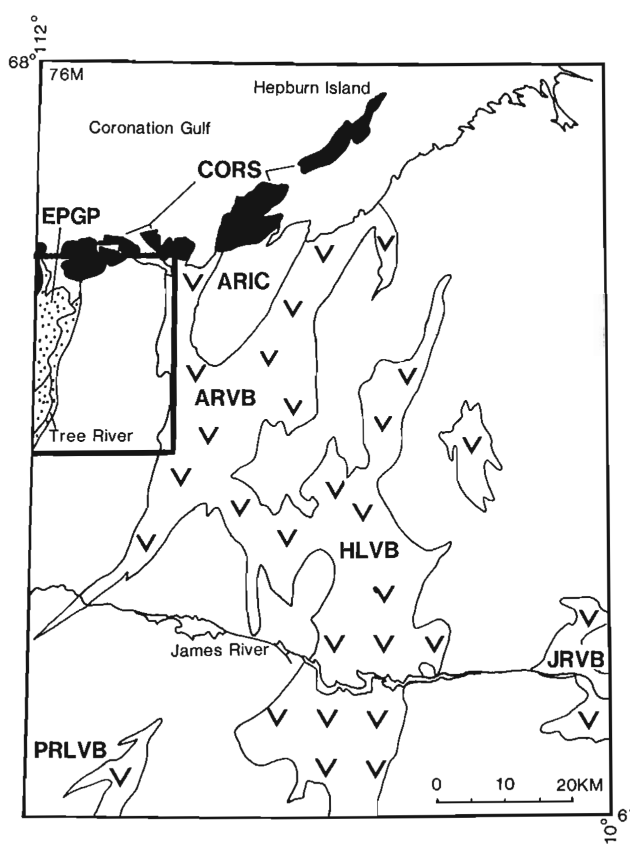


Figure 1. Major physiographic and geologic features of the Hepburn Island map sheet (76M). CORS: Coronation/Franklin sills; EPGP: Epworth Group; ARIC: Anialik River intrusive complex; ARVB: Anialik River volcanic belt; HLVB: High Lake volcanic belt (outlier to east); JRVB: James River volcanic belt; PRLVB: Pink Rock Lake volcanic belt; granitoid and gneissic rocks not patterned. Figure 2 outlined in northwest corner.

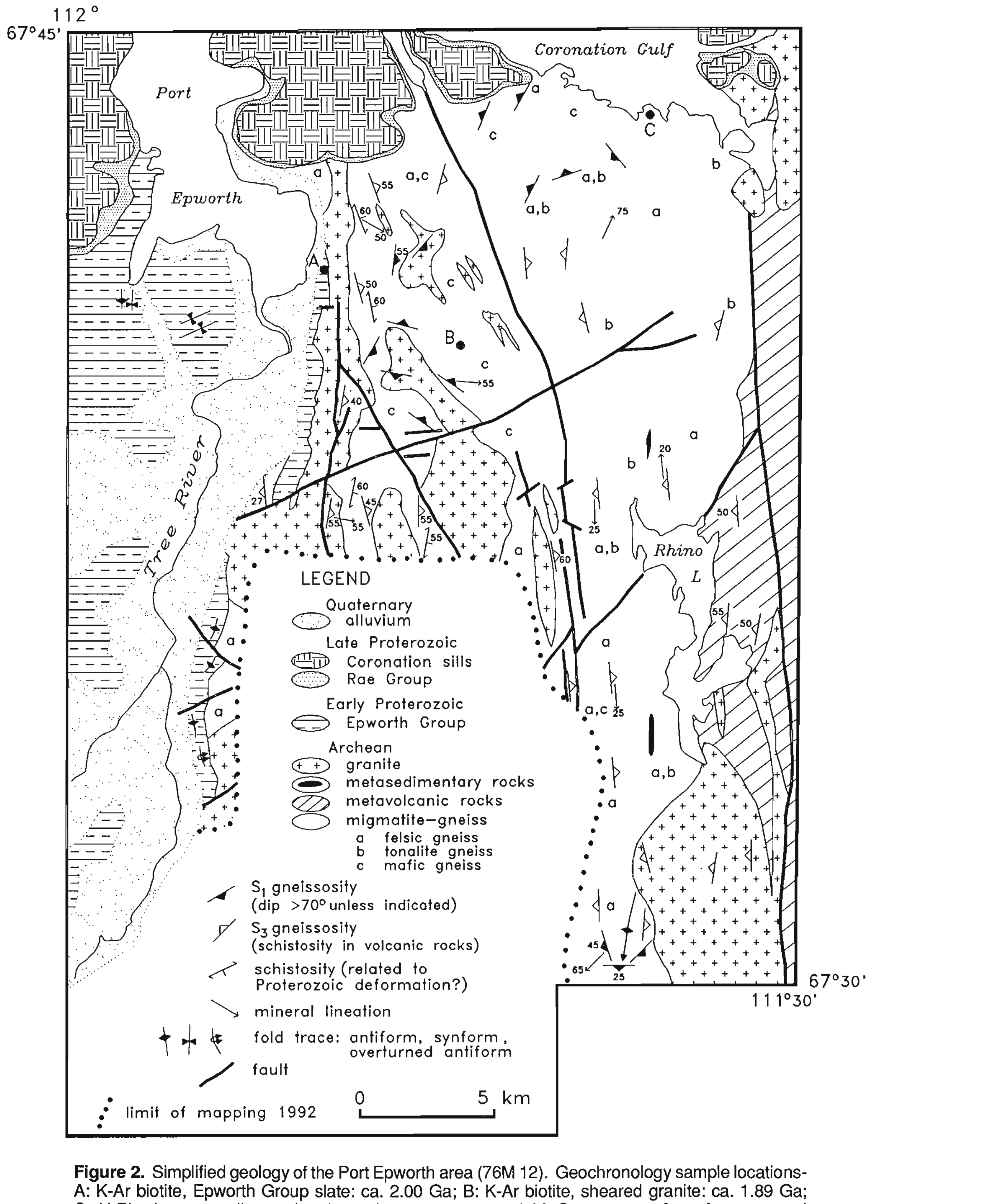


Figure 2. Simplified geology of the Port Epworth area (76M 12). Geochronology sample locations- A: K-Ar biotite, Epworth Group slate: ca. 2.00 Ga; B: K-Ar biotite, sheared granite: ca. 1.89 Ga; C: U-Pb zircon, tonalite gneiss, lower intercept age ca. 1.99 Ga; see text for references and discussion.

contents, the blocks retain a consistent attitude. In the southeast, the biotite monzogranite contacts are foliated, with dyking into the migmatite. The interior of this intrusion is generally undeformed, but exhibits weakly penetrative fabrics locally. Granite units are interpreted to be late-tectonic with respect to D_3 Archean deformation (see below).

Other intrusive rock types within the granite and migmatite rock units are much less abundant, and are too restricted to be separated as map units in Figure 2. They include: syn- to late-tectonic (with respect to D_3 Archean deformation: see below) biotite-hornblende granodiorite and aluminous leucogranite with muscovite and/or garnet; and late- to posttectonic granodiorite dykes and diorite-gabbro dykes.

Migmatite

For this study, the term migmatite is used to describe: (1) injection gneiss: angular to subangular blocks from 0.5 m to 25 m in width, of predominantly orthogneiss and amphibolite, cut by and within a matrix of variably foliated granitoid rock; and (2) straight or near-straight gneiss (McEachern, 1993), including compositionally layered, medium- to coarse-grained felsic, intermediate and mafic orthogneiss, and garnet-bearing paragneiss. The straight and near-straight gneiss is believed to be of similar parentage as the injection gneiss but with a higher degree of strain. The former type is common in the northwestern area of Figure 2; and the latter is common in the eastern half. More highly strained migmatite is present in the northwestern area locally, but cannot be separated into individual map units. Initial observations of metamorphic mineral assemblages in amphibolite and rare garnet-bearing paragneiss blocks indicate mid-greenschist to lower amphibolite facies across the map area.

In the eastern half the map area, the migmatite is commonly compositionally layered and commonly has a uniformly medium- to coarse-grained gneissic fabric. Here felsic gneiss has a granitic to granodioritic composition, with up to 15% biotite and/or hornblende, and up to 20% amphibolite inclusions as 0.5-5m, elongate, sub-rounded blocks. Intermediate gneiss has a tonalitic or dioritic composition, with up to 35% hornblende and lesser biotite, and is commonly cut by leucogranite dykes subparallel to foliation. Mafic gneiss comprises amphibolite with up to 35% gneissic leucogranite dykes up to a metre thick parallel to foliation, or elongate amphibolite blocks up to 20 m by 5 m within 35-60% gneissic granitoid matrix. The amphibolite is composed of hornblende, plagioclase and up to 10% quartz. Garnet-bearing paragneiss occurs within orthogneiss as a discontinuous unit parallel to foliation in the southeast area (Fig. 2). It is medium- to coarse-grained garnet-biotite-plagioclase-quartz schist and gneiss up to 200 m wide, with garnet-rich (60% garnet) layers up to 0.5 m in width.

Metavolcanic and metasedimentary rocks

The easternmost part of the map area contains phyllitic to schistose metavolcanic and metasedimentary rocks of the Anialik River volcanic belt, previously described by Tirrul and Bell (1980) and Relf et al. (1992). The metavolcanic rocks include massive, pillowed, fragmental and tuffaceous basaltic rock, and massive and tuffaceous, locally quartz-phenocrystic dacite and rhyolite. Metasedimentary rocks comprise biotite-rich phyllite and schist. Other metasedimentary rocks are present within the gneissic rocks west of the contact (described above). Sulphide- and oxide-bearing siliceous iron-formation is present locally within 200 m of the north-trending fault in the extreme eastern part of the map area (Fig. 2), as 0.5 m to 5 m-thick beds intercalated with metavolcanic rocks.

The contact between the metavolcanic-metasedimentary rocks and gneissic rocks north of Rhino lake is a multiply-deformed shear zone containing numerous mafic and felsic dykes that was subjected to late brittle (presumably Proterozoic) strike slip and normal faulting (McEachern, 1993). The contact is not present at the surface south of Rhino lake, due to the emplacement of a late, relatively undeformed monzogranite intrusion which separates biotite-rich mafic schist and 30% aplitic and monzogranitic dykes to the north-east and amphibolite-rich straight and near-straight gneiss to the west.

STRUCTURAL GEOLOGY

The granitoid-migmatite terrane in the map area can be divided into three structurally and lithologically distinct domains: to the northwest, northeast, and southeast. The convenient boundaries are regional faults, characterized by prominent physiographic linear features, and containing gneissic, schistose and mylonitic fabrics parallel to their trend (discussed below). These boundaries have been chosen based on empirical observation of the structural trends, and no other significance is placed on them at this time. The eastern domains are separated from the northwest domain by a north- northwest-trending fault through the centre of the area, and themselves are separated by a northeast-trending fault (Fig. 2). Granitoid rocks comprise 50% of northwest domain, whereas the eastern two domains are predominantly gneissic. A detailed description of structural geology and deformation history for the gneiss of the northeast domain, informally termed the "Kangguyak" gneiss belt, is given in McEachern (1993).

The structural fabrics and faults within the northwest and southeast domains, and the contact with the early Proterozoic Epworth Group strata are described below within a chronologic framework. Stereoplots for mineral and elongation lineations, contoured poles to foliations, and shear fabrics for these domains are given in Figures 3-5.

D_1 - D_3 Archean deformation, and granitoid emplacement

The earliest pervasive deformation fabric (S_1) is represented by gneissic fabrics that are folded in F_2 open folds on a scale of tens to hundreds of metres. The clearest example is found in the southern part of the southeast domain, where S_1 gneissic amphibolite and leucogranite fabrics form a south-southwest-plunging F_2 open antiform with a subvertical axial plane (Fig. 2). This antiform is parallel to F_2 folds noted in the Turtle lake area (Relf et al., 1992) and other folds of the Anialik River volcanic belt (Tirrul and Bell, 1980). Accompanying this fold are parasitic S- and Z-folds, and L_2 stretching lineations that parallel the fold axis (Fig. 3). In the northern two domains, there are comparable open fold patterns described by gneissic fabrics that are correlated with S_1, but they are difficult to trace along the presumed F_2 fold limbs (Fig. 4). In the northeast domain, McEachern (1993) has noted isolated, rootless fold hinges within low strain windows between more highly strained gneiss, which may represent pre-F_2 folding.

D_3 deformation is best exemplified by the pervasive, north-trending, subvertical, straight or near straight gneiss in the southeast domain. S_3 foliations dip from steep to the east in the western rocks, to moderate to steep to the west in gneisses (and adjacent metavolcanic rocks) on the eastern side. The S_3 gneissic fabrics commonly have subhorizontal L_3 stretching lineations (Fig. 3), and sub-horizontal, north- and south-trending isoclinal fold hinges. These parallel most of the lineations found in metavolcanic rocks to the east. No systematic sense of asymmetry can be discerned from the mesoscopic F_3 folds in this area. F_2 fold limbs are commonly transposed or sheared by the north-trending D_3 deformation in the eastern domains. Along the north shore of Rhino lake, layers of amphibolite and leucogranite form "type 2" mushroom-shaped interference patterns (Ramsay and Huber, 1987), resulting from the interference of F_2 and F_3. In the southeast domain, mylonitic zones in F_3 near straight gneiss with subhorizontal L_3 stretching lineations show type IIa back-rotated asymmetrical pull-aparts (Hamner and Passchier, 1991) locally. North-northwest-trending, and steep east-dipping C-S ductile shear fabrics and in tonalitic orthogneiss, and shear bands between asymmetric pull-aparts indicate a consistent, sinistral sense of displacement within one kilometre of the western boundary of southeastern domain (Fig. 5). All D_3 deformation features in the southeast domain are compatible with formation in an east-directed compressional regime.

In the northwest domain, D_3 deformation fabrics in gneissic rocks are less well-preserved than to the east, as they are commonly overprinted by chloritic fabrics, and are less abundant, as granitoid rocks comprise 50% of the domain. S_3 foliations trend to the north and north-northwest and have moderate to steep dips to the east (Fig 4).

The large biotite monzogranite body south of Rhino lake in the southwest domain cuts earlier deformed metavolcanic and gneissic rocks. It has local, foliated margins, and internal weakly developed penetrative foliation parallel to S_3 in the host rocks, and is interpreted to be late-tectonic with respect to D_3 deformation. Other granitoid rock types including garnet- and muscovite-bearing leucogranite dykes are parallel to F_3 foliation, and have fabrics that are parallel to F_3 foliation, and are also believed to be syn- to late-tectonic. The late- and posttectonic mafic and felsic dykes most clearly identified in the northeast domain (Relf et al., 1992; McEachern, 1993) were emplaced after peak D_3 deformation.

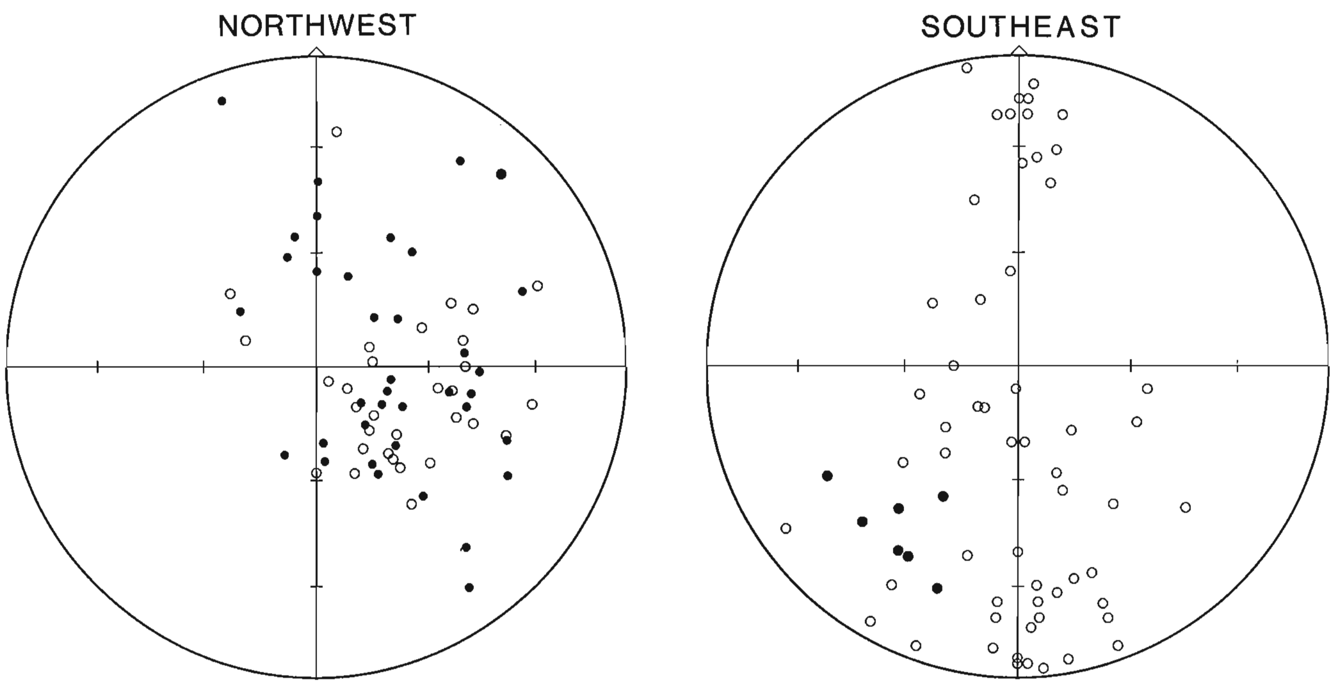
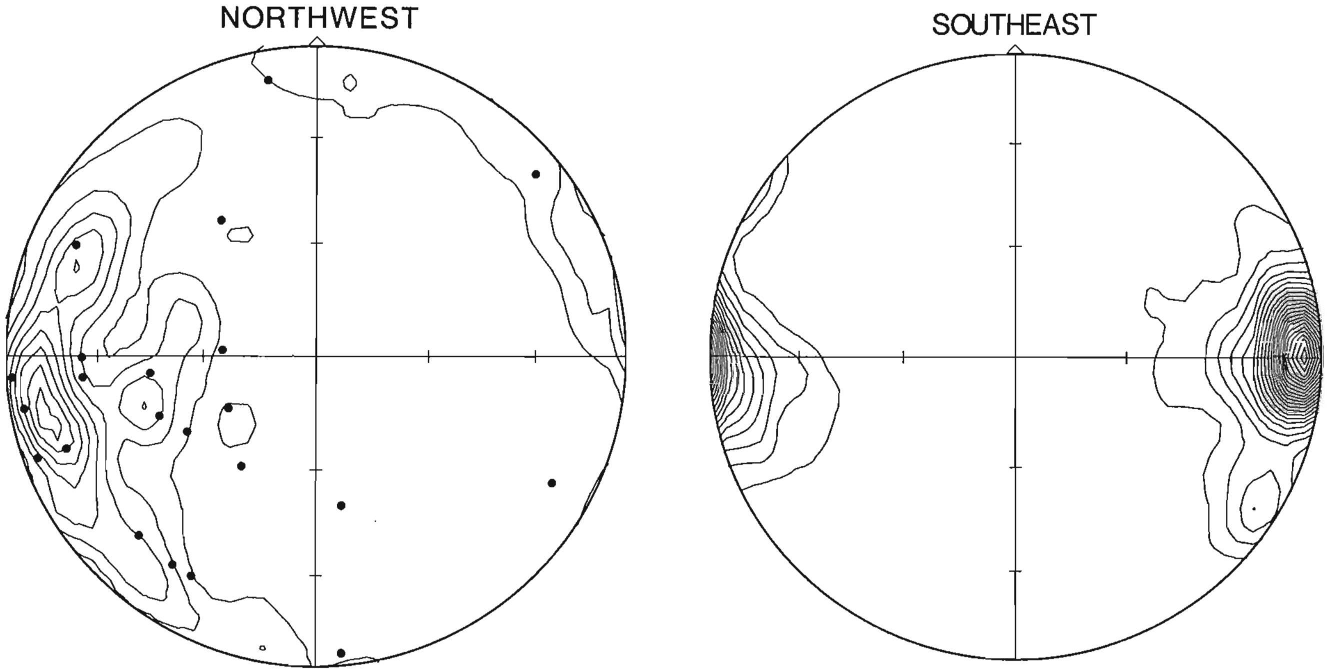


Figure 3. Mineral and elongation lineations. Northwest domain: L_2-L_3 lineations from gneissic fabrics (L_2 and L_3 not distinguishable in the field): closed circles; lineations from chloritic fabrics (early Proterozoic?): open circles. Southeast domain: L_2 lineations: closed circles; L_3 lineations: open circles.



N = 225
k = 100.00
(Peak - E)/Sigma = 15.1
Peak position : 259.4°/ 10.5°
E = 2.25
Sigma = 1.05

N = 170
k = 100.00
(Peak - E)/Sigma = 43.7
Peak position : 90.0°/ 6.2°
E = 1.70
Sigma = 0.91

Figure 4. Contoured stereoplots of foliations. Northwest domain: S_1-S_3 gneissic foliations contoured (S_1-S_3 not distinguishable in the field); and foliations from chloritic schistose fabrics: closed circles. Southeast domain: F_3 foliations contoured. Spherical Gaussian function terms: N: number of points; k: Gaussian weighting function coefficient (for k=100, counting circle has radius of 8.11°); sigma: 1 standard deviation; E: number of points expected within a fractional area. Lowest contour level = E, and higher contours step up by 2 x sigma. See Robin and Jowett, 1986.

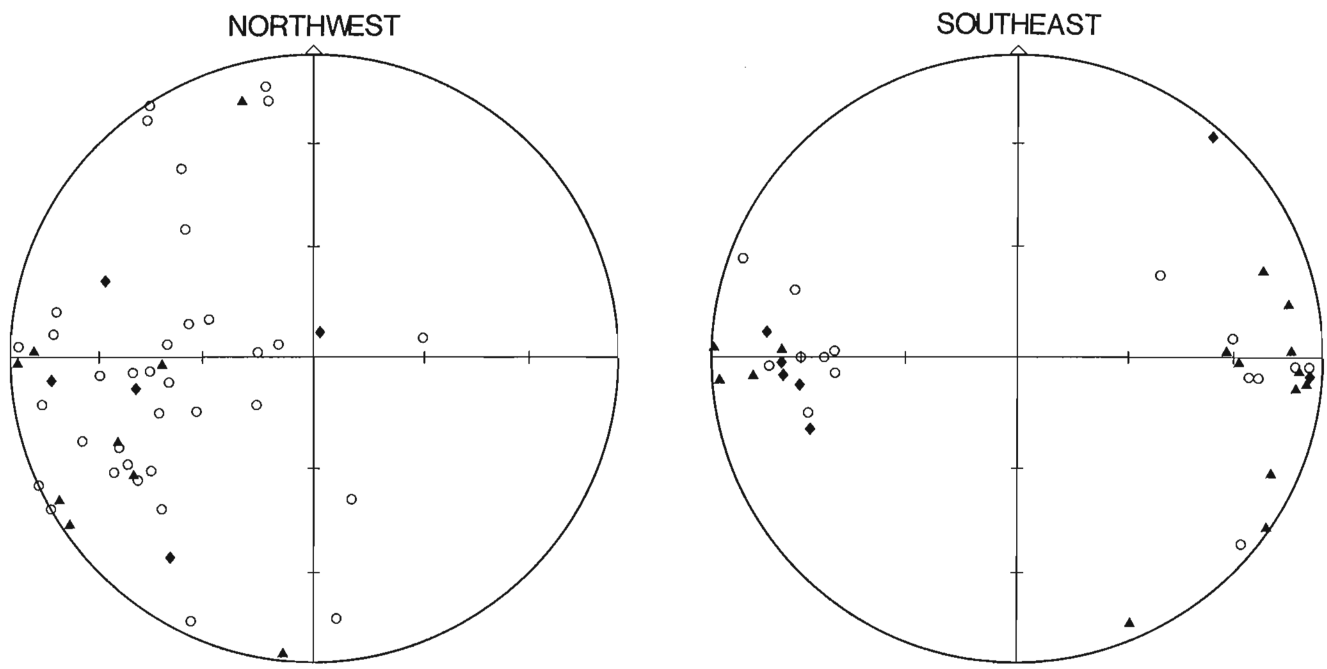


Figure 5. Stereoplots of C planes of C-S shear fabrics. Triangles- dextral sense; diamonds- sinistral sense; open symbols- no shear sense discernable. Northwest and Southeast domains.

Early Proterozoic(?) deformation

The absolute age of late fabric development in the map area is difficult to establish, for two reasons: (1) much of the late fabrics overprint, or are parallel to pre-existing gneissic fabrics, and are difficult to separate from these earlier fabrics; and (2) there is a lack of unambiguous geochronological data. However, considering the geology of the lower Proterozoic Epworth Group contact described below and in Hoffman et al. (1984), and an early Proterozoic K-Ar biotite age from a biotite chlorite schist within granite (Wanless et al., 1965: discussed below), there is reason to consider the late, chloritic fabrics in the Archean basement as early Proterozoic. These chloritic fabrics are abundant in the northwest domain, within 5 km of the contact, but are much less common to the east. Therefore this section focusses on the chloritic fabrics in the northwest domain, and the regional faults.

Chlorite- and biotite-rich partings, foliation planes and shear zones within granitic and migmatitic rocks have variable attitudes, but almost all have dips with an easterly component (Fig. 5). Their attitudes are similar to the S_2 and S_3 gneissic fabrics (Fig. 4), and also to the subvertical and east-dipping parts of the contact with Epworth Group rocks (see below). All of the shear fabric measurements are of C-S fabrics in the northwest domain are from chlorite-biotite schist. They are commonly associated with quartz veins and carbonate in zones that may extend for tens to hundreds of metres; and some are in biotite-chlorite melanosome present at the contact between amphibolite and more felsic rock. In three locations, biotite or tremolite-actinolite porphyroblasts are present on the schistose shear fabric surface. Most lineations in chlorite-biotite schist plunge moderately to the southeast, and broadly parallel L_2-L_3 lineations (Fig. 3).

There are three major north- trending faults, and numerous northwest-trending and northeast-trending faults that cut Archean rocks (Fig. 2). One north-trending fault, almost entirely within granitoid rock, is parallel to and 1 km east of the Epworth Group contact. It dips vertically or steeply to the east, and contains quartz-carbonate veins up to 2 m in thickness, with associated chlorite-carbonate-sericite alteration. In several locations, C-S fabrics in chlorite-sericite schist indicate a reverse, east-side-up sense of displacement. A second north-trending fault, dividing the northwest domain from the two domains, is a prominent physiographic depression and contains minor minor chloritic zones. The third north-trending fault is along the eastern margin of the map area, and traverses granitoid, gneissic and volcanic rocks. There is abundant brick red felsite as dykes parallel to the fault that have brittle fracturing and amethystine quartz veins locally.

In the southern part of the northwest domain, a prominent, subvertical northeast-trending fault cuts and significantly displaces the Proterozoic contact, with a minimum dextral strike separation of 200 m. Other dextral, northeast-trending faults, and sinistral, northwest-trending faults displace the contact, and displace the north-trending major faults. The northwest- and northeast-trending faults are parallel to, and have the same sense of displacement as the conjugate fault set documented in the Wopmay Orogen to the west (e.g., Hoffman et al., 1984). At least two east-trending faults are present in the southern part of the northwest domain. One contains quartz veins and quartz breccias up to 5 m thick, with open space filling and amethystine quartz.

Contact between Epworth Group and Archean basement

The Odjick Formation stromatolite-bearing ferroan dolomite, and fine to coarse siliclistic rocks are exposed along the eastern valley slope of the Tree River, at the north-trending contact with the Archean granitoid-migmatite basement (Fig. 2). In one panel between northwest-trending and northeast-trending faults, 4 km north-northeast of the fork in the Tree River (Fig. 2), basal stromatolitic dolomites rest undisturbed on massive Archean basement, with beautifully preserved, mound-like and columnar stromatolities around exfoliated slabs of granite. Here the undisturbed unconformity dips gently to the west. However most of the contact is characterized by buckled and folded Odjick Formation strata. Box folds, upright isoclinal folds and overturned tight isoclinal folds (Fig. 6a) have north-trending, sub-horizontal axes that parallel the contact and indicate east-directed shortening on a scale of tens to hundreds of metres. Much of the contact has near-vertical to moderate easterly dips, with Archean basement overhanging compressed Odjick Formation strata, (Fig. 6b). Minor, bedding- parallel shear planes in Odjick formation strata have a consistent east-over-west sense within 200 m of the contact. The contact itself is commonly schistose, with carbonate-chlorite-quartz schist that locally envelops sub-rounded, boulder-sized fragments of granite. North-trending zones of carbonate-chlorite-quartz schist up to a metre thick and tens of metres along strike are common within 200 m of the contact in the Archean basement. Where measurable, shear sense from C-S fabrics is predominantly east-side-up. Arrays of en echelon quartz veins, locally with carbonate, pyrite, chalcopyrite or malachite, are common in basement granitoid rocks near the contact (Fig. 6c). The consistent north trend of (1) the folded contact surface, (2) fold axes in the Odjick formation, and (3) shear planes in Archean rocks within 200 m of the contact, are compatible with an east-directed compressional regime in this area related to the Calderian Orogeny (e.g., Hoffman et al., 1984).

Summary of deformation events

A summary of the sequence of deformation events is as follows: 1) S_1 gneissic fabrics were deformed by F_2 open folds. This D_2 deformation event may have produced tight isoclinal folds that were later refolded and transposed in the eastern domains. It is believed to have been contemporaneous with the D_2 deformation in the Turtle Lake area (Relf et al., 1992). 2) D_3 deformation, an east-directed compressional regime, produced in the eastern domains north-trending, sub-vertical S_3 straight gneiss fabrics, with well-developed, sub-horizontal L_3 stretching fabrics and F_3 fold axes. Syn- to late-tectonic granite and aluminous leucogranite were emplaced parallel to S_3 gneissic fabrics at this time. Less deformed late-tectonic granitoid intrusions, and late- to posttectonic mafic and felsic dykes were emplaced after peak D_3 deformation. 4) Early Proterozoic, east-directed compression produced chloritic fabrics, schistose zones and shear zones containing C-S fabrics that are parallel to S_2 and S_3 fabrics. Reverse, east-over-west chlorite-carbonate-quartz shear zones characterize part of the north-trending Epworth Group – Archean basement contact, and parallel faults to the east. 5) Brittle conjugate faults, parallel to those of the Wopmay Orogen to the west, cut all Archean rocks and the Proterozoic-Archean contact.

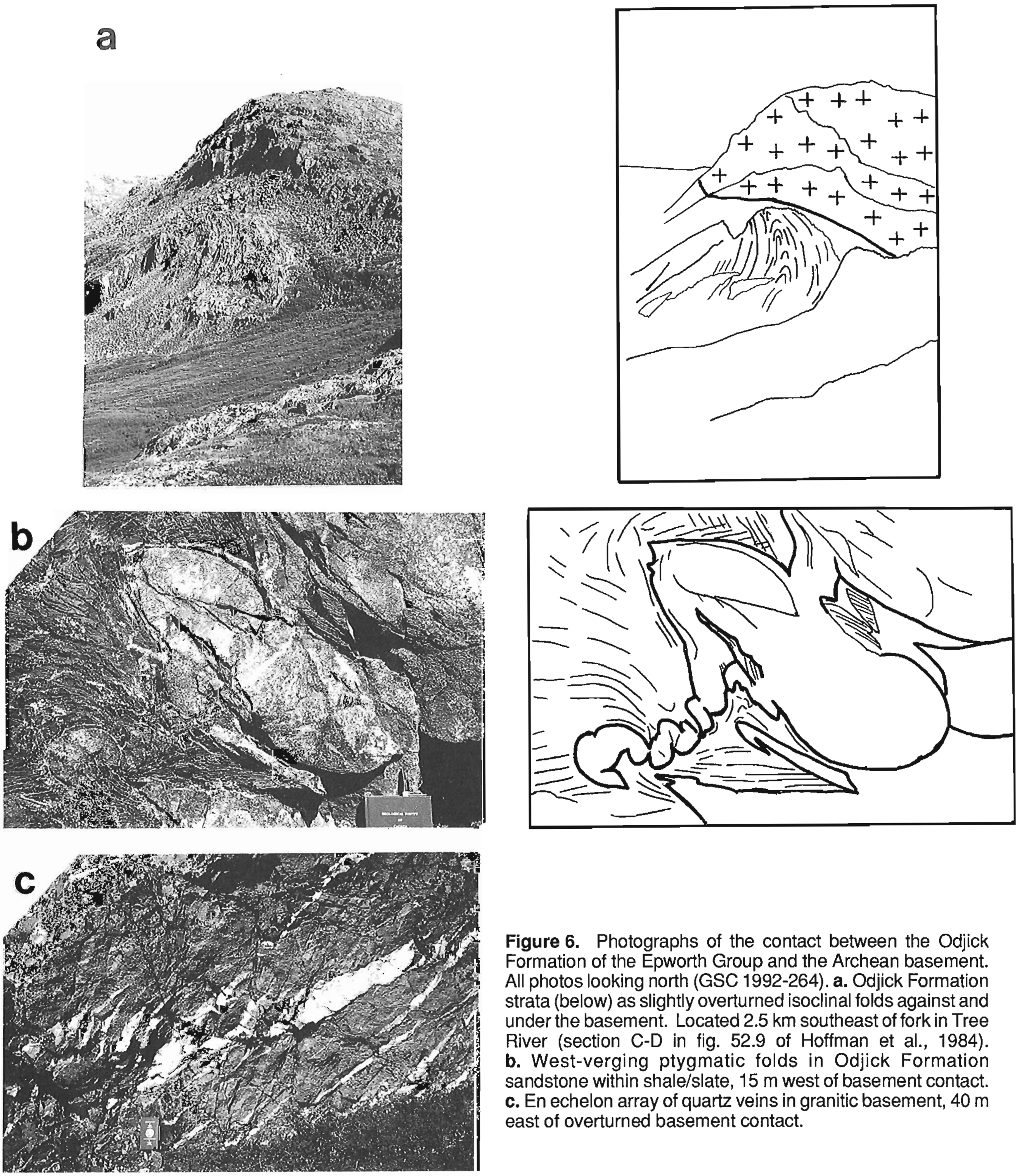


Figure 6. Photographs of the contact between the Odjick Formation of the Epworth Group and the Archean basement. All photos looking north (GSC 1992-264). **a.** Odjick Formation strata (below) as slightly overturned isoclinal folds against and under the basement. Located 2.5 km southeast of fork in Tree River (section C-D in fig. 52.9 of Hoffman et al., 1984). **b.** West-verging ptygmatic folds in Odjick Formation sandstone within shale/slate, 15 m west of basement contact. **c.** En echelon array of quartz veins in granitic basement, 40 m east of overturned basement contact.

DISCUSSION OF EARLY PROTEROZOIC DEFORMATION

Proterozoic – Archean basement contacts in the Tree River fold belt, including the contact in this study, have been described previously as mullion-like, lobe-cuspate structures, with the margins of the lobes commonly overturned (Hoffman et al., 1984). Temperatures of greater than 350°C were estimated from biotite-bearing assemblages in Coronation Supergroup rocks nearby (Lucas, 1984). A similar description of large scale, mullion-like structures in biotite grade rocks was given for the granitic basement-cover contact in the external zone of the western Alps in France (Gratier and Vialon, 1980; Tricart and Lemoine, 1986). In the western Alps, pre-existing listric normal faults are believed to have been reactivated during Alpine compression to form overhanging basement structures (Tricart and Lemoine, 1986). These descriptions of deformed basement-cover contacts infer ductile deformation in the granitic basement under biotite grade metamorphic conditions.

At the kilometre-scale, it would appear that the predominantly granitic Archean basement did undergo ductile deformation during early Proterozoic compression. The numerous shear zones in chlorite-biotite schist may have accommodated early Proterozoic compression (Hoffman et al., 1984). These shear zones parallel to pre-existing gneissic fabrics may represent discrete zones of early Proterozoic ductile deformation within largely undisturbed rock. In this study, biotite porphyroblasts were located in chloritic shear zones in Archean granite in the northwest domain, within 2 km of a K-Ar biotite sample from a chloritic shear in granite ("B" in Fig. 2) with an age of 1893 ± 70 Ma (Wanless et al., 1965: recalculated using decay constants of Steiger and Jager, 1977). This age indicates an early Proterozoic thermal setting for the ^{40}Ar in that biotite (circa 300°C (Harrison et al., 1985), and is similar to the 1.90 to 1.86 Ga U-Pb zircon ages for collision-generated Hepburn intrusive rocks of the Wopmay Orogen (Hoffman and Bowring, 1984). Deformation within ductile deformation zones may have been aided by fluids expelled from basal Epworth Group strata, analogous to Alleghenian basement deformation in the southern Appalachians (Mitra, 1979; Oliver, 1986).

Evidence for early Proterozoic metamorphism in the eastern domains is also provided by the ca. 2.0 Ga U-Pb zircon lower intercept for an orthogneiss sample from the Arcadia Bay area (Relf et al., 1992: "C" in Fig. 2). Furthermore, D_2 fabrics in the southeastern domain are consistent with a predominantly east-directed compressional regime, similar to to the early Proterozoic compressional regime. Considering the U-Pb resetting and the significantly deformed basement-cover contact, widespread early Proterozoic basement deformation in this area must be considered. One hypothesis would be that the gneiss of the eastern domains represents the eroded core of a basement antiformal structure, similar to those of the central Wopmay Orogen and western Slave Province (King, 1986; Frith, 1991).

FUTURE WORK

This work has opened questions concerning the effects of early Proterozoic deformation and metamorphic/thermal overprinting in the Archean basement adjacent to the 'externides' of the Wopmay Orogen. Continued structural mapping of the granitoid-migmatite terrane, focussing on contact relationships with the Proterozoic rocks and the Anialik River volcanic belt will be conducted, and complemented by pressure-temperature studies on metamorphic mineral assemblages. The thermal history of Archean basement rocks near the contact and in the eastern domains needs to be investigated using K-Ar, Ar-Ar and U-Pb isotope systematics. Ultimately this work will be incorporated into a new NATMAP geological synthesis of the Hepburn Island 1:250 000 map sheet.

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Geological Survey of Canada Project 87-0008TB

Geology and mineral occurrences of the southern part of High Lake greenstone belt, Slave Province, Northwest Territories[1]

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Henderson, J.R., Henderson, M.N., Kerswill, J.A., Arias, Z., Lemkow, D., Wright, T.O., and Rice, R., 1993: Geology and mineral occurrences of the southern part of High Lake greenstone belt, Slave Province, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 125-136.

Abstract: Supracrustal rocks in the belt near Hood River are mainly basalt and andesite with little felsic material and subordinate metagreywacke. Farther north, but south of James River, basalt is abundant along the western margin of the north-striking belt, and rhyolite is abundant in the centre. Thin bedded sulphidic-graphitic slate and grey siltstone form a characteristic lithofacies associated with felsic volcanic rocks. Magnetite-chert iron-formation and carbonaceous limestone also form part of this facies.

Felsic volcanic rocks extruded ca. 2.69 Ga, and are surrounded by batholiths that intruded ca. 2.60 Ga. D_1 tilting and folding were followed by D_2 folding, and overprinted by a pervasive D_3 cleavage. Andalusite, cordierite and biotite porphyroblasts, formed under static conditions in the central part of the area, predate D_3.

The area contains several gold and base metal prospects, including the gold- and arsenopyrite-rich Flood Zone on the Ulu Property (BHP Minerals Canada Ltd.)

Résumé : Les roches supracrustales de cette zone près de la rivière Hood sont principalement basaltiques et andésitiques et secondairement métagrauwackeuses. Plus au nord, mais au sud de la rivière James, du basalte est abondant le long de la bordure ouest de cette zone d'orientation nord et de la rhyolite abonde au centre. De l'ardoise sulfureuse et graphitique et du siltstone gris, forment un lithofaciès caractéristique associé aux roches volcaniques felsiques. Une formation ferrifère à chert-magnétite et du calcaire riche en carbone font aussi partie de ce faciès.

Les roches volcaniques ont fait extrusion ca. 2,69 Ga et sont entourées par des roches granitiques qui ont fait intrusion il y a environ 2,60 Ga. Le basculement et le plissement survenus pendant D_1, suivi de plissement pendant D_2, sont surimposés par un clivage pénétrant pendant D_3. Au centre de la zone, des porphyroblastes d'andalousite, cordiérite et biotite ont été formés dans des conditions statiques avant D_3.

La région contient plusieurs indices d'or et de métaux communs, y compris la zone de Flood riche en or et arsénopyrite dans la propriété Ulu (BHP Minerals Canada Ltd.).

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INTRODUCTION

Mapping of High Lake greenstone belt began in 1992 as a Federal Government component of the 1991-1996 Canada-NWT Mineral Initiatives Agreement. Objectives of the project are to map the supracrustal rocks and their contacts with adjacent granites from south of Hood River to Coronation Gulf (Fig. 1), interpret the geology, and place the mineral occurrences in a regional geological context. Mapping was completed in the southern part of the project area in parts of NTS map sheets 76L/14, 15, 16 (Kathawachaga Lake), and 76M/1, 2, and 3 (Hepburn Island).

PREVIOUS GEOLOGICAL MAPPING

The High Lake greenstone belt was mapped by Fraser (1964) as part of a helicopter-assisted reconnaissance mapping of the northwestern Canadian Shield. Baragar (1975) mapped a 12 km strip across part of the High Lake belt along James River in NTS 76M/2 and 3, and Relf (1984) mapped part of 76M/3 in the Canoe Lake area. Jackson et al. (1986a, b) mapped parts of the belt in 76L/15 and 16, and 76M/1, 2, 8, 9, and 15. Jackson (1991) published a regional compilation of previous geological mapping in 76L; compilations of previous mapping in 76M that include parts of the map area (Fig. 2) were done by Easton (1982), and Jackson (1989). The

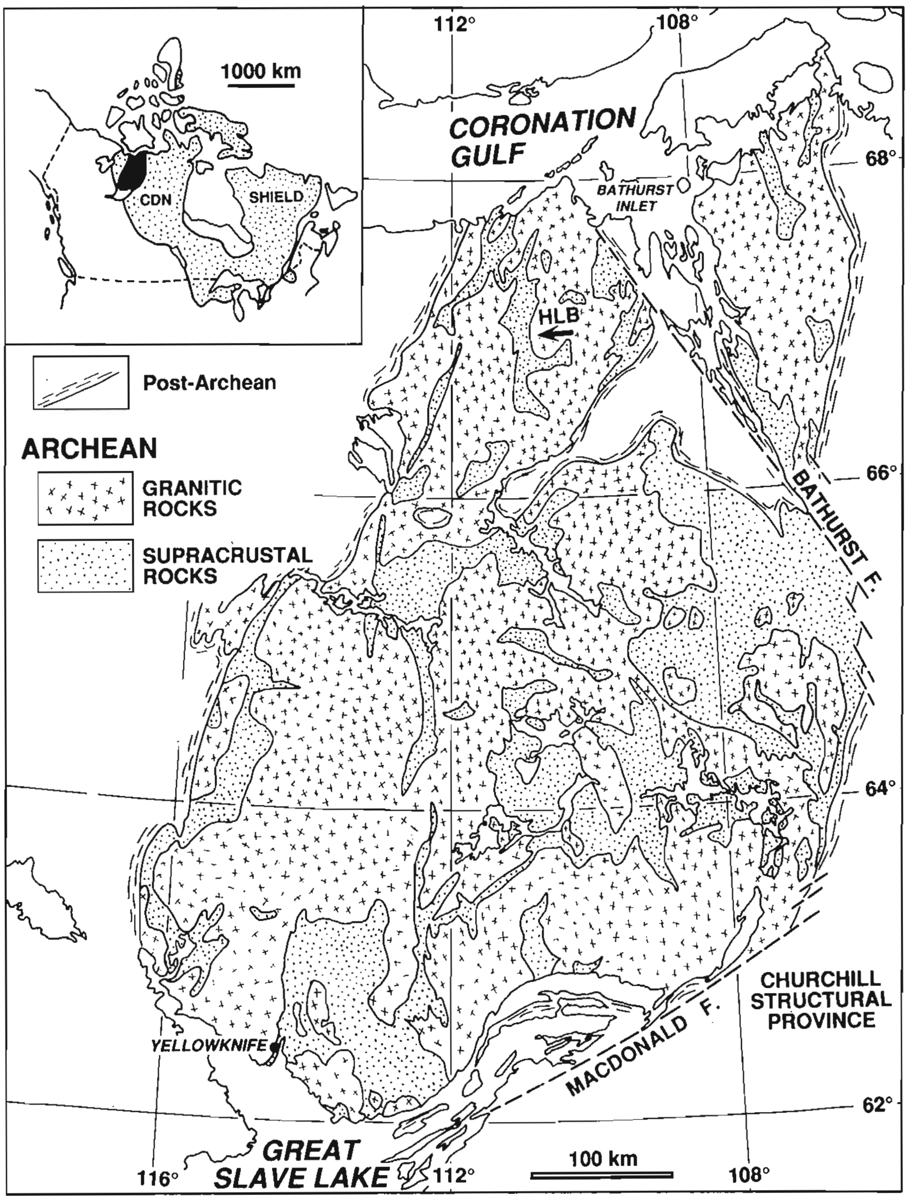

Figure 1. Location map showing High Lake belt (HLB) in relation to other major geological features in the Slave province. Some "granitic rocks" are older than "supracrustal rocks".

latter compilation includes locations and a listing of mineral occurrences, some of which are shown in Figure 2, and are included in Table 1.

REGIONAL GEOLOGICAL SETTING

The Slave structural province consists of a basement gneiss terrane, mainly in the west, containing the 3.96 Ga Acasta gneiss (Bowring et al., 1989), and younger gneisses to ca. 3 Ga (Krogh and Gibbons, 1978). Thick greywacke turbidite and basalt ranging from ca. 3.0 to 2.66 Ga (Van Breemen et al., in press) generally overlie the basement rocks. However, quartzite and ultramafic rocks suggestive of a stable-shelf assemblage are present locally (Covello et al., 1988), and stratovolcanoes penetrate the thick turbidite sequence (Lambert et al., 1992). Within the part of High Lake belt mapped (Fig. 2), U-Pb zircon dating has established a time of felsic volcanism north of James River at 2695 ± 3 Ma (M. Villeneuve, pers. comm., 1992), and also south of James

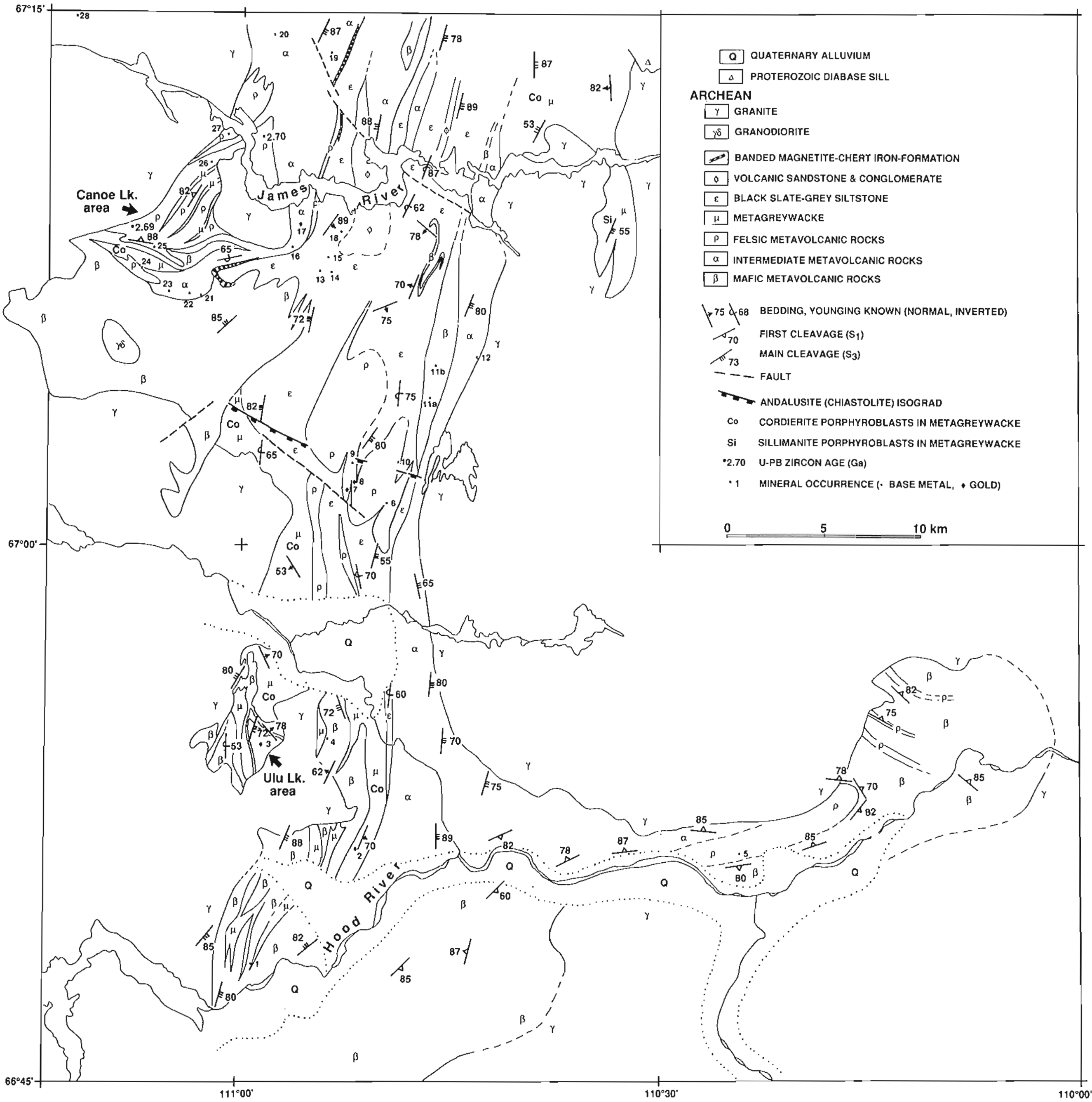

Figure 2. Geological map of the southern part of High Lake greenstone belt in the region of the Hood and James rivers. Mineral occurrences shown by numbers are listed in Table 1.

River in the Canoe Lake area at ca. 2.69 Ga (Mortensen et al., 1988). Several deformations and regional metamorphism occurred prior to, and coincident with, province-wide granite intrusion between 2.63 and 2.58 Ga (van Breemen et al., in press).

Tectonic models proposed for Slave province involve eastward subduction-accretion of single (Kusky, 1989), and multiple (Fyson and Helmstaedt, 1988) volcano-sedimentary arc complexes, as well as ensialic, intra-plate thickening models (Thompson, 1989). The province as a whole, including High Lake belt, is characterized by low P-high T regional metamorphism (Thompson, 1989).

DESCRIPTION OF MAP UNITS

Rocks forming a basement terrane to the volcanic-sedimentary assemblage in High Lake belt have not been recognized to date. The following lithological descriptions apply to units in the map legend (Fig. 2), and are presented in a preliminary quasi-stratigraphic order. All of the volcanic and sedimentary rocks mapped are metamorphosed and cleaved to various degrees, and the prefix "meta" commonly is omitted for brevity, especially where the protolith is clearly distinguishable.

Volcanic and sub-volcanic rocks

Mafic volcanic rocks

Mafic volcanic rocks are dark green to black, fine-to-medium grained amphibolites. They are massive to pillowed, foliated to unfoliated flows and sills, and well-foliated, poorly bedded volcaniclastic rocks. Flow boundaries are generally poorly defined by centimetre-thick light-coloured selvages. Amygdaloidal textures were rarely observed. Some mafic rocks have coarser grained, gabbroic textures, but have not been distinguished separately because of their gradational and irregular contacts; some of these rocks may be sub-volcanic sills. Mafic rocks in the south and east of the area mapped are at least 10 km thick, provided there are no structural repetitions.

Intermediate volcanic rocks

Intermediate volcanic rocks are pale green to grey felsite (probably andesite). Some are white plagioclase-phyric flows, quartz-amygdaloidal flows, or volcaniclastic rocks (Fig. 3). Foliation defined by clast alignment is well developed in volcaniclastic intermediate rocks, whereas felsites are more weakly foliated or unfoliated. Medium green, pillowed flows younging east were noted north of

Table 1. Mineral occurrences in the southern part of High Lake greenstone belt (Fig. 2).

ID	NTS	Principal Commodities (Reported Minerals)	Host Rock	Name	Reference
1	76L/15	Zn, Cu, As, Au, Ag (po, sph, cpy, asp)	mafic metavolcanic rocks	Ralph (Chuck)	Brophy et al. (1984)
2	76L/15	Au, As (asp, py, po)	metagreywacke	North Mare (Blackridge)	Laporte et al. (1978)
3	76L/15	Au, As (asp, po, py)	mafic metavolcanic rocks	Flood Zone, ULU Property	Flood (1991)
4	76L/15	Zn, Cu, Pb, As, Ag, Au (po, sph, cpy, ga, asp)	mafic metavolcanic rocks	Penthouse	Byrne (1970)
5	76L/16	Ag, Cu, Zn, Pb (po)	felsic metavolcanic rocks	Hog	Laporte et al. (1978)
6	76M/2	po, py	felsic metavolcanic rocks		Jackson (1989); #49
7	76M/2	Au, Zn	felsic metavolcanic rocks		Jackson (1989); #38
8	76M/2	As (asp, po, py)	felsic metavolcanic rocks		Jackson (1989); #29
9	76M/2	As	felsic metavolcanic rocks		Jackson (1989); #28
10	76M/2	py	felsic metavolcanic rocks		Jackson (1989); #48
11a	76M/2	Cu, Zn	mafic metavolcanic rocks		Jackson (1989); #30
11b	76M/2	Cu (py, cpy)	mafic metavolcanic rocks		Jackson (1989); #31
12	76M/2	Cu (cpy, py)	intermediate metavolcanic rocks		Jackson (1989); #32
13	76M/2	Pb, Zn	black slate-grey siltstone		Jackson (1989); #24
14	76M/2	Pb, Zn	black slate-grey siltstone		Jackson (1989); #23
15	76M/2	Ag, Sb, Pb	black slate-grey siltstone		Jackson (1989); #6
16	76M/2	Ag, As, Sb, Pb, Zn	black slate-grey siltstone		Jackson (1989); #7
17	76M/2	Pb, Zn	intermediate metavolcanic rocks		Jackson (1989); #22
18	76M/2	Au	intermediate metavolcanic rocks		Jackson (1989); #1
19	76M/2	Cu, Zn	intermediate metavolcanic rocks		Jackson (1989); #3
20	76M/2	Cu, Zn (py, po, cpy, sph)	intermediate metavolcanic rocks		Jackson (1989); #21
21	76M/3	Cu (py, po, cpy)	intermediate metavolcanic rocks		Jackson (1989); #27
22	76M/3	Cu, Zn, Pb, Ag, Au	intermediate metavolcanic rocks		Jackson (1989); #42
23	76M/3	Pb, Zn (py, ga, sph)	intermediate metavolcanic rocks		Jackson (1989); #26
24	76M/3	Cu, Zn, Pb, Ag	mixed metavolcanic rocks	Canoe Lake	Jackson (1989); #47
25	76M/3	py, po	felsic metavolcanic rocks		Jackson (1989); #25
26	76M/3	Pb (py, ga)	felsic metavolcanic rocks		Jackson (1989); #50
27	76M/3	Ag, Cu, Zn	felsic metavolcanic rocks		Jackson (1989); #2
28	76M/3	Cu, Zn, Pb, Mo (py, po, cpy, sph, ga, mo)	mixed metavolcanic rocks	Barb	Jackson (1989); #37

James River, but, in general, flow boundaries were rarely observed. Intermediate rocks apparently young west in the northerly-striking part of the belt east of the Ulu Lake area north of Hood River (based upon younging determinations in metagreywacke interbeds), and, provided there are no structural repetitions, are at least 5 km thick.

Felsic volcanic rocks

Felsic volcanic rocks are light grey to white, quartz and quartz-feldspar porphyries, as well as pale green muscovite-rich schists with relict quartz phenocrysts. Fragmental felsic rocks are common. Flow boundaries were rarely observed, but some flow-top breccias were seen with interstitial carbonaceous carbonate cement.

A large felsic rock unit was mapped in the centre of the area (Fig. 2) at the andalusite isograd. A breccia with carbonate cement, observed along part of the northwestern margin of this felsic unit, suggests that it is a west-younging dome cutting through about 6 km of the surrounding black slates and grey siltstones described below. Felsic units mapped elsewhere are more concordant, and are generally less than 1 km thick.

Metasedimentary rocks

Metagreywacke

Metagreywacke commonly contains massive siltstone beds, and graded siltstone-slate turbidite beds; coarse sandy beds are uncommon. One occurrence of conglomeratic metagreywacke was observed at the granite contact in the easternmost part of the belt north of Hood River. Cordierite- and andalusite-bearing pelitic tops of graded turbidite beds are more abundantly and coarsely porphyroblastic than psammitic bases of the beds, producing a metamorphic size-grading that is the reverse of primary size-grading.

Metagreywacke beds in High Lake belt appear to represent muddy turbidites, conformable with the mafic volcanic sequences, typical of the bulk of metasedimentary rocks of the Yellowknife Supergroup (McGlynn and Henderson, 1972).

Slates and siltstones

Slates and siltstones form a distinctive lithological association in the belt. These rocks are mainly thinly interbedded black, graphitic-sulphidic slates and light grey siltstones. They are spatially associated with felsic and intermediate volcanic rocks, in contrast to metagreywacke turbidites which are associated with mafic volcanic rocks in High Lake belt. Thin beds of silicate and sulphide iron-formation are present within the slate and siltstone unit. Some rusty-brown weathered slates contain very fine grained, centimetre-sized bedding-parallel ellipsoids and lenses that weather to red, granular quartz-hematite. At andalusite grade, rusty- weathered beds contain garnet and grunerite. Rare marble beds, found near contacts with some volcanic rocks, are brown to grey calcite or dolomite with thin, black cherty lenses and fine grained grey siltstone beds. Along strike at Snofield Lake, several kilometres north of the map area, Henderson (1975) reported stromatolitic carbonate beds overlying felsic volcanic rocks.

In zones of low bulk strain, soft sediment load structures, e.g. ball-and-pillow (Fig. 4) and flame structures, are preserved in some interbedded slates and siltstones. However, this unit generally is not graded, and younging indicators are not common. The origin of the grey siltstone beds is problematic. The slate and siltstone unit may be as

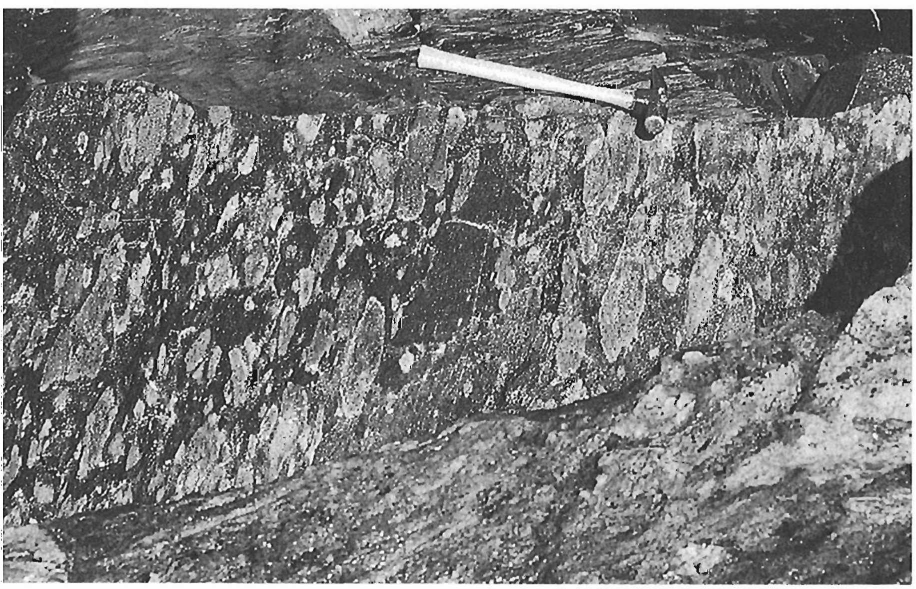

Figure 3.

Flattened grey andesitic clasts in a darker matrix. Located at falls on Hood River at tributary mouth ca. 10 km southeast of Ulu lake area (Fig. 2). Clasts interpreted as flattened parallel to S_1. (GSC 1992-254I) much as 5-10 km thick, depending upon the amount of repetition due to isoclinal folding. It pinches-out southward towards Hood River, and is interlayered with a variety of volcanic and volcaniclastic rocks north of James River.

Sandstone and conglomerate

Volcaniclastic sandstone and subordinate conglomerate form grey, medium-grained, poorly sorted, and poorly bedded units in the north part of the map area near James River (Fig. 2). Sand grains are mainly quartz and plagioclase, and felsic-volcanic rock fragments dominate the pebble fraction in conglomerate, but angular black shale clasts (Fig. 5), and rounded black chert clasts are also present. These disorganized rocks are a coarse-grained lithofacies proximal to felsic volcanic rocks, and may represent debris flows deposited near the flanks of emergent rhyolite domes.

Iron-formation

Rusty red-brown and yellow weathered oxide iron-formation forms a discontinuous map unit up to several-hundred metres wide along the northwest margin of the area (Fig. 2), stratigraphically above felsic and intermediate metavolcanic rocks. It comprises thinly interbedded fine grained magnetite-quartz and sucrosic chert. These chemically precipitated rocks may result from submarine hydrothermal activity associated with intermediate and felsic volcanism.

Granitic intrusive rocks

Granodiorite

Pale grey weathered biotite granodiorite forms several small plutons (Fig. 2, 7). Foliation is absent or developed in discrete zones several metres to 100 metres wide. Rocks of granodioritic composition are found as a subordinate component

Figure 4.

Thinly bedded, black slate-grey siltstone couplet showing load structure, due to sinking of more dense silt into hydroplastic mud (younging towards top of photograph). Black slate- grey siltstone unit located on andalusite isograd near contact with metagreywacke unit (Fig. 2). (GSC 1992-254G)

Figure 5.

Poorly-sorted volcaniclastic conglomerate with angular shale clasts. Located ca. 2 km south of "R" in James River (Fig. 2). (GSC 1992-254B).

within granitic batholiths that surround High Lake belt (Fig. 2), but have not been distinguished from granite. The age of granodiorite plutons relative to surrounding granitic batholiths is not known.

Granite

Light grey to flesh-coloured, medium- to coarse-grained biotite-muscovite granite forms batholiths (Fig. 2). Boundaries of granites are commonly discordant with the strike of layering in adjacent supracrustal rocks, and the granite is not foliated. However, where contacts are concordant with the strike of adjacent supracrustal rocks, the granite may be foliated for several hundred metres parallel to cleavage in adjacent wall rocks.

Diabase dykes and sills

Undeformed diabase dykes are common in the area, and postdate granite intrusion. They form two sets (Fig. 7, not shown in Fig. 2): a northwest-striking set, probably correlative with the 1.27 Ga Mackenzie swarm (LeCheminant and Heaman, 1989), and an east-northeast-striking set, generally plagioclase-phyric, possibly correlative with the 2.45 Ga Hearst-Matachewan swarm (Heaman, 1988). A diabase sill, partially mapped in the northeast corner of the area, probably is a Coronation Sill related to ca. 723-718 Ga Franklin igneous events (Heaman et al., 1992).

STRUCTURAL GEOLOGY

The structural history of High Lake greenstone belt is polyphase, and structural fabrics are heterogeneously developed. In places the rocks exhibit an early cleavage (D_1), folded on the map-scale by steeply plunging, tight to isoclinal folds (D_2). A period of static regional porphyroblastesis of biotite, cordierite and andalusite predated development of the "main" regional cleavage (D_3), and emplacement of granitic batholiths appears to mainly postdate D_3.

Structural fabric elements

Structural fabric elements measured in the field are defined as follows:

S_0 is bedding and primary flow layering. It is defined by differences in mineral composition observed on the mesoscale.

S_1 parallels bedding, and is locally defined by muscovite in metagreywacke. Muscovite-inclusion trails observed in centimetre-size cordierite and andalusite porphyroblasts in metagreywacke beds may be relict S_1 cleavage that has been obliterated by S_3 in the quartz-mica matrix. In the southeastern part of the belt, along Hood River, a regional foliation, defined by oriented clasts in volcaniclastic rocks (Fig. 3) as well as aligned mica, is interpreted as S_1 because it is folded on the map scale (Fig. 2, see below), although the fabric is rarely crenulated by a younger cleavage (S_3). The finite or bulk flattening plane in this area is apparently parallel to S_1, and is folded by F_2.

Macroscale tight to isoclinal folds, apparently without an axial planar cleavage, are F_2 folds. These folds are noncylindrical. They are documented best by the anticline west of Ulu Lake (Fig. 9) where bedding youngs away from the axial surface trace, and S_1 is folded in the core of the northwest-plunging fold. In the Canoe Lake area (Fig. 2), a westward-closing vertical fold in iron-formation is transected by northeast-striking S_3, and may be an F_2 fold. In the southeast part of the area, north of the Hood River (Fig. 2), an eastward-closing vertical fold may be an F_2 structure because bedding-parallel S_1 cleavage is also folded. Whether map-scale folds interpreted as F_2 for each of these areas are related to the same D_2 folding event has not been determined.

S_3 is the dominant regional cleavage (the "main cleavage"). At greenschist grade, S_3 is a slaty cleavage in slate, and a spaced cleavage in metasiltstone. It is axial-planar to some mesoscale folds. S_3 is defined by aligned muscovite in andalusite grade metasedimentary rocks, and is deflected around competent biotite, cordierite and andalusite porphyroblasts, which may exhibit a discordant internal fabric. In the northerly-striking part of the belt, flattened clasts parallel to S_3 indicate that in this area S_3 is the bulk flattening plane.

S_4 is a steeply-dipping crenulation of S_3 developed in several areas where S_3 and S_0 exhibit anomalously low dips (e.g. in the northeast of the area in Fig. 7).

Structural geometry

The geometry of bedding (Fig. 6a) in the northerly-trending part of the belt, between Hood and James rivers, indicates a dominant orientation at ca. 014°/80°, and the partial-great-circle girdle of bedding poles is interpreted to be due to isoclinal folding with a subvertical statistical mean fold axis (trend 121° and plunge 77°). F_2 fold axial surfaces are northerly-trending in this part of High Lake belt where they are most strongly overprinted by steeply-dipping, northeasterly-trending S_3 cleavage (Fig. 6b). In other parts of the belt, e.g. along Hood River and south of James River in the Canoe Lake area, S_3 cleavage is not as strong, and F_2 folds are easterly-trending and steeply plunging, suggesting that S_0 and S_1 may have been nearly vertical before D_2 folding. Possibly S_1 formed when bedding was rotated to a regional vertical attitude.

The northward-closing F_2 fold nearly surrounded by granite in the Ulu Lake area (Fig. 7 and 9) is noncylindrical. In the southern part of the area shown in Figure 7, the fold is an anticline plunging moderately northwestward. However, in the northern part of the area, the fold plunge, defined by bedding dip on the basalt closure, is steep towards the south, and several beds in the surrounding metagreywacke young inward – indicating a south-plunging syncline. This geometry is not understood at present.

The S_3 cleavage is the most penetrative structural element in the map area. It shows a strong preferred orientation at ca. 015/80 (Fig. 6b), and, on the regional scale, is subparallel to S_0 (Fig. 6a). Locally, however, S_3 crosscuts both limbs of F_2 folds, generally in a clockwise sense, as at the Ulu fold. In a few localities, S_3 appears to be an axial-planar cleavage related to steeply-plunging mesoscale F_3 folds in interbedded slate and siltstone.

In the northern part of the Ulu area (Fig. 7), S_0 and S_3 assume moderate-to-low dips and S_3 is crenulated by a vertical S_4 cleavage striking north-northeast. A vertical northwest-striking crenulation of S_3 occurs in metagreywacke east and south of the Ulu Lake area near the corner of the batholith (not shown in Fig. 2).

METAMORPHISM

Metamorphic grade increases from north to south in the area mapped. An andalusite isograd is oriented approximately east-west (Fig. 2). Blocky biotite porphyroblasts appear directly north of the isograd, and cordierite with andalusite rims is present in metagreywacke south of the isograd. Cordierite appears sporadically in metagreywacke north of the isograd in association with chlorite grade black slates. In the slate and siltstone unit, the higher metamorphic grade is characterized by occurrence of garnet and grunerite in thin, but ubiquitous, silicate-sulphide facies iron-formation beds. Sillimanite was observed only in a migmatitic lobe of metagreywacke in the batholith east of the supracrustal belt, south of James River (Fig. 2).

Cordierite and andalusite porphyroblasts preserve trails of a primary or tectonic (S_1) fabric, and are overprinted by S_3 cleavage. The trace of the andalusite isograd is not symmetrical with the batholiths bounding the belt, and the S_3 cleavage-forming event intervened between porphyroblastesis and granite emplacement. The random orientation of andalusite (variety chiastolite, Fig. 8) with textural-sector zoning indicates that metamorphism took place under lithostatic conditions (cf. Ferguson et al., 1980; Rice and Mitchell, 1991). The succession of regional static porphyroblastesis, and cleavage formation prior to batholith emplacement is similar to the sequence in Hood River supracrustal belt 50-80 km east of High Lake belt (Henderson et al., 1991).

ECONOMIC GEOLOGY

Twenty-nine mineral occurrences, compiled from public sources, are listed in Table 1, and shown in Figure 2. Base metals are the principal commodities in the majority of occurrences; they are hosted by mafic, intermediate, and felsic volcanic rocks, as well as by the slate-siltstone assemblage. Gold occurs with base metals in the Ralph and Penthouse showings, and in the Canoe Lake area. Gold is the principal commodity of interest in the Flood Zone on the Ulu property, at the North Mare (Blackridge) occurrence, and at occurrence no. 18. Arsenopyrite is strongly associated with gold in both the Flood Zone and North Mare occurrences, and is found in the base-metal rich Ralph and no. 16. occurrences. Arsenic has also been reported from two gossans within felsic metavolcanic rocks in the central part of the area (nos. 8 and 9).

The Flood Zone (Fig. 9a), on the Ulu Gold Property of BHP Minerals Canada Ltd., is the most significant mineral prospect in this part of High Lake belt. Gold and arsenopyrite are found in a silicified zone hosted by metabasalt with subordinate metagreywacke and metagabbro. The generally 2-5 m wide mineralized zone strikes about 300° for approxi-

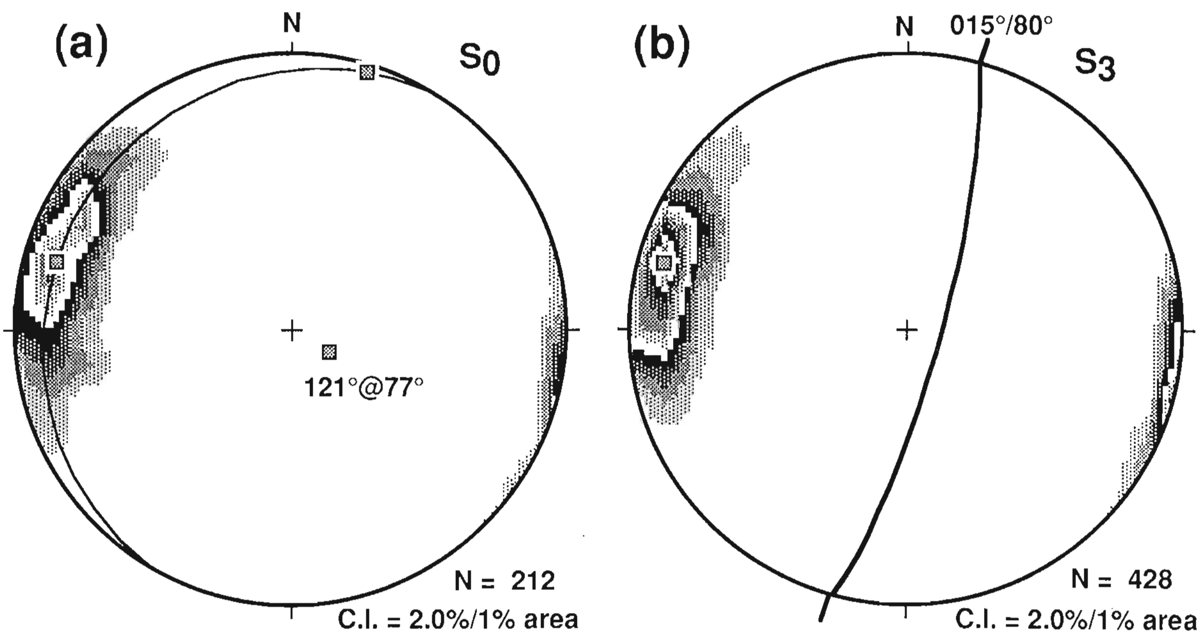

Figure 6. Orientations of S_0 and S_3 from High Lake belt between Hood and James rivers shown on lower-hemisphere Schmidt projection. (**a**) Bedding poles. Pole to best fit great circle of S_0 trends 121° and plunges 77°. (**b**) S_3 poles. Average foliation plane is oriented 015°/80°.

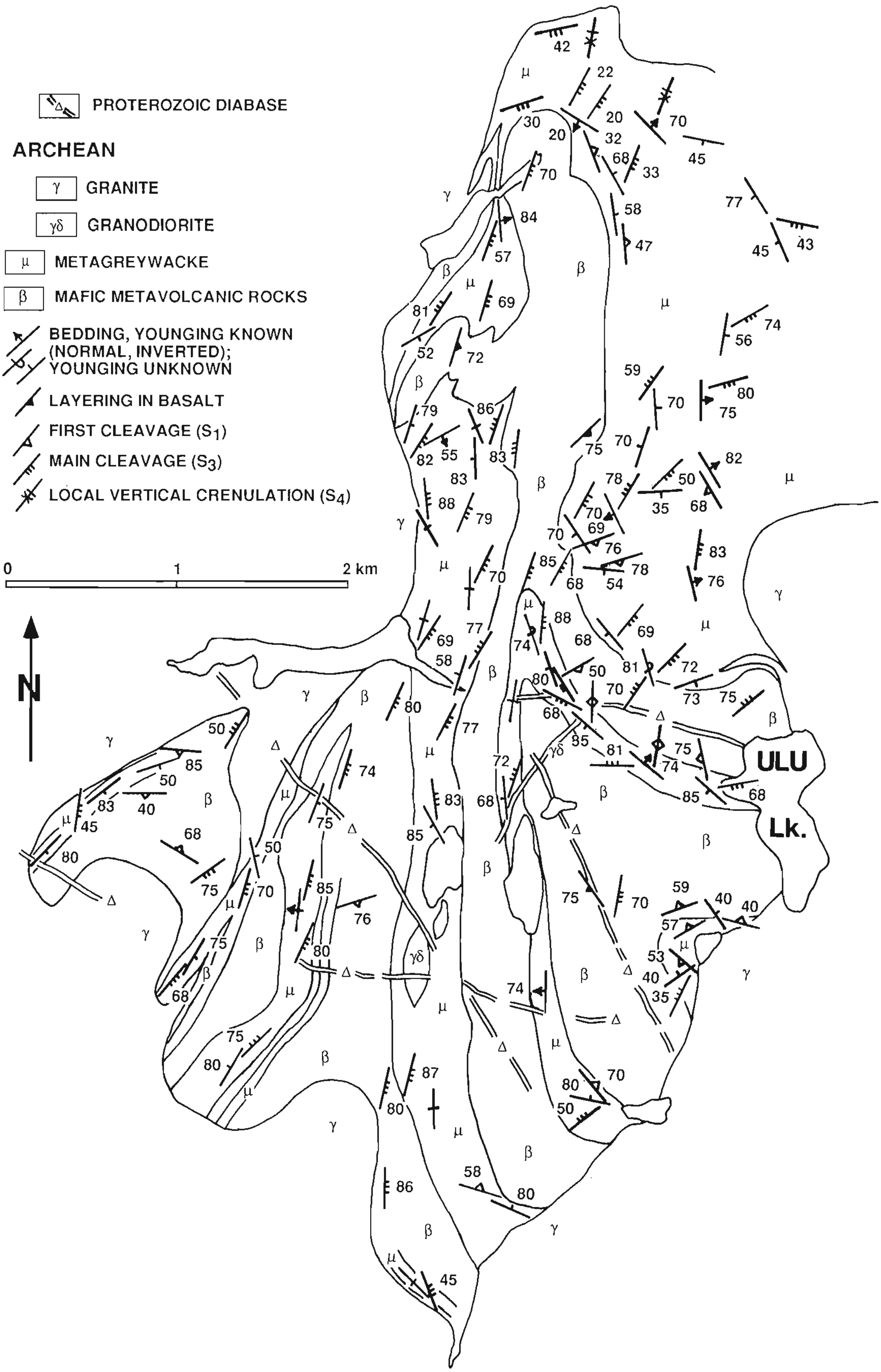

Figure 7. Geological map of Ulu Lake area.

mately 400 m on surface, and dips steeply to more than 600 m depth. Gold and arsenopyrite are accompanied by alteration minerals that include biotite, hornblende, actinolite, microcline and calcite (Flood, 1991).

Figure 9b shows the orientation of 94 quartz-actinolite vein segments measured in a stripped and washed outcrop near the northwest end of the Flood Zone; they are scattered due to original variation and open folding, but show a dominant orientation ca. 155°/66°. Quartz-actinolite veins within the Flood Zone strike more northerly than the zone itself, which is approximately perpendicular to the regional S_3, suggesting a left-stepping *en échelon* array. Figure 9c, a schematic map, indicates orientations of the basalt boundary (S_0), the Flood Zone, and a left-stepping *en échelon* vein array.

The Flood Zone occurs almost entirely within metabasalt on the west limb of the Ulu fold, near the metagreywacke-filled core. The veins resulted from dilatency within competent basalt, but the timing relative to the deformation events is uncertain. For example, if regional metamorphism can be shown to overprint the mineralized zone, it would eliminate the possibility that mineralization occurred during D_3, which postdated the metamorphic peak.

CONCLUSIONS

Mafic to intermediate flows and volcaniclastic volcanic rocks dominate the easterly-trending part of High Lake belt along Hood River, whereas sedimentary rocks are more abundant in a northerly-trending part of the belt between the Hood and James rivers. Several rhyolites, as well as numerous intermediate to mafic volcaniclastic rock units, are present within the assemblage of sedimentary rocks between the two rivers.

The structural sequence includes three cleavage and folding events. A D_1 bedding-parallel cleavage (S_1) is present in closures of several D_2 macro-scale folds (F_2), and a D_3 regional cleavage (S_3) transects most F_2 folds. Regional metamorphism occurred before D_3 cleavage, and is apparently unrelated to batholith emplacement which largely post-dated S_3/D_3. Andalusite-bearing metasedimentary rocks are present in the south half of the area mapped; the north half of the central belt of metasedimentary rocks is chlorite grade.

The Flood Zone, on the Ulu Gold Property of BHP Minerals Canada Ltd., is the most significant of more than 25 base and/or precious metal occurrences reported from this part of High Lake belt. In the Flood Zone, gold and arsenopyrite are found in a silicified zone hosted largely by metabasalt. The generally 2-5 m wide mineralized zone strikes about 300° for approximately 400 m on surface, and dips steeply to more than 600 m depth. Gold and arsenopyrite are accompanied by alteration minerals that include biotite, hornblende, actinolite, microcline and calcite (Flood, 1991).

Figure 8.

Andalusite (var. chiastolite) porphyroblasts in black slate. Black slate-grey siltstone unit located between andalusite isograd and contact with metagreywacke (Fig. 2). (GSC-1992-254F)

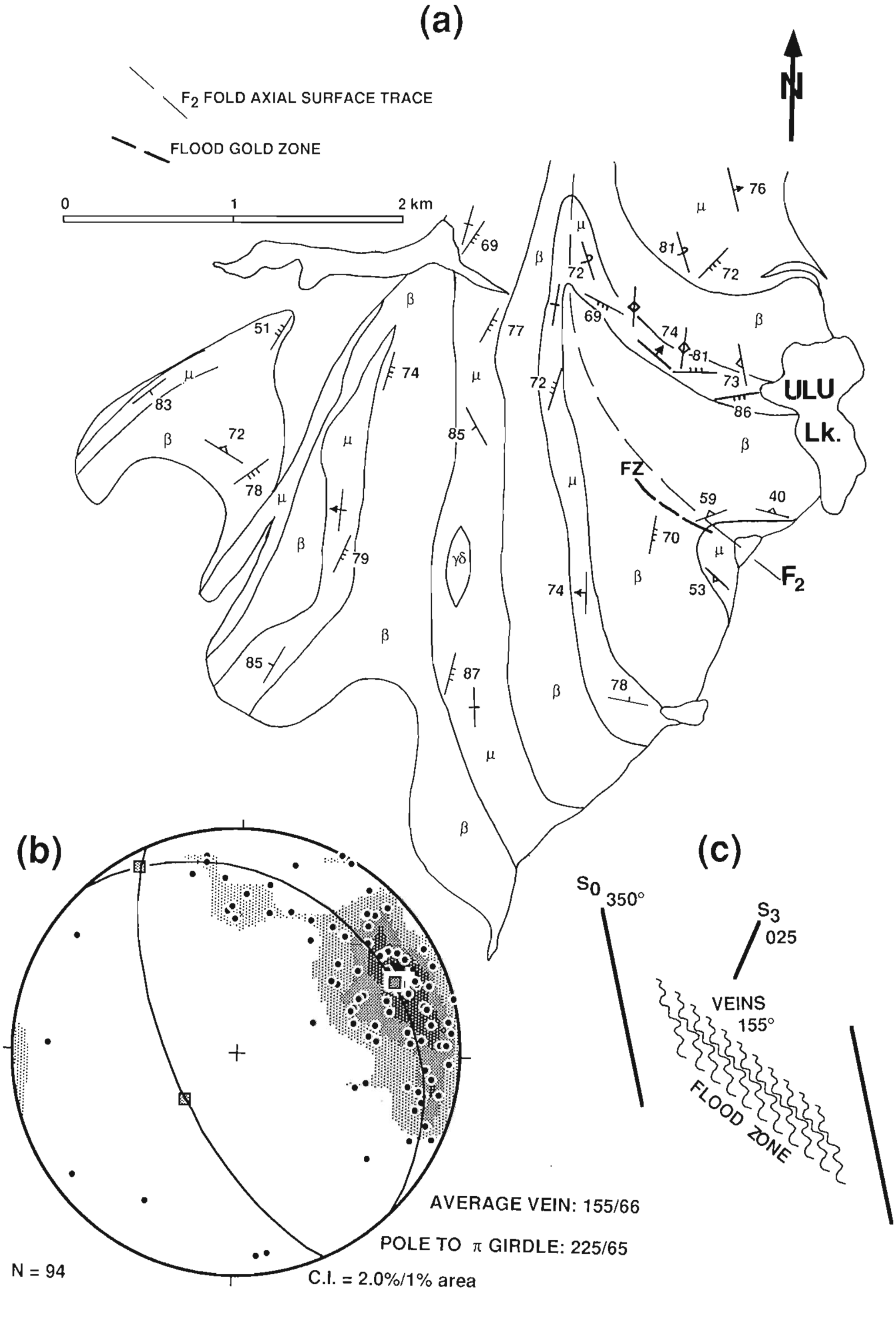

Figure 9. Structure of Flood Zone at BHP Minerals Canada, Ltd. Ulu Gold Property. (**a**) Map of Ulu lake anticline showing location of Flood Zone (FZ). Symbols of map units and structural elements are the same as Figure 7. (**b**) Equal-area plot of poles to 94 vein segments from the northwest part of Flood Zone. (**c**) Schematic map showing orientation of structural elements at the Flood Zone.

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Geological Survey of Canada Project 870008 KC

Structural investigation of high-grade rocks of the Kramanituar complex, Baker Lake area, Northwest Territories

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Abstract: Granulite-grade metaplutonic and metasedimentary rocks of the Kramanituar complex record strong to intense deformation, although moderate strain is preserved in coarse grained gabbroic anorthosite in the central part of the complex. An outward increase in strain culminates in kilometre-wide zones of ultramylonite which appear to define boundaries to the complex.

Movement on the shear zones is consistent with uplift of granulite-grade material relative to amphibolite-grade rocks exposed north and south of the complex. However, it remains to be demonstrated whether finite strain recorded by the shear zones can account for the juxtaposition of these disparate metamorphic terranes, or whether the shear zones are relatively late structures that transect a previously assembled metamorphic terrane.

Résumé : Les roches métaplutoniques et métasédimentaires du faciès des granulites du complexe de Kramanituar présentent un degré de déformation fort à intense, bien que l'anorthosite gabbroïque à grain grossier de la partie centrale du complexe soit moyennement déformée. Une intensification de la déformation à partir du centre du complexe culmine en des zones de largeur kilométrique d'ultramylonite, qui semblent définir les limites du complexe.

Le déplacement des zones de cisaillement concorde avec le soulèvement des matériaux du faciès des granulites par rapport aux roches du faciès des amphibolites qui affleurent au nord et au sud du complexe. Il reste toutefois à démontrer si la déformation finie des zones de cisaillement peut expliquer la juxtaposition de ces terranes métamorphiques disparates, ou si ces zones de cisaillement sont des structures tardives qui recoupent un terrane métamorphique s'étant constitué antérieurement.

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INTRODUCTION

Prominent linear anomalies in horizontal gravity gradient and magnetic anomaly maps of Canada (Goodacre et al., 1987; GSC, 1987) extending from central Alberta to the northwest shore of Hudson Bay correspond to discontinuous occurrences of granulite-grade rocks (highs), and metasedimentary rocks and/or paragneiss (lows) (Gibb, 1968; Wallis, 1970). Speculation on the geological nature of these anomalies across the central part of the (former) Churchill province has led to two interpretations: 1) that the Churchill Province comprises two Archean microcontinents, the northern Rae and the southern Hearne, joined along a northeast- to east-striking suture, designated the Snowbird tectonic zone (STZ) (Hoffman, 1988; Hanmer et al., 1992), which resulted from their collision in Early Proterozoic time (Walcott and Boyd, 1971; Gibb and Thomas, 1976; Thomas et al., 1988; Hoffman 1989); or, 2) that they represent an intra-plate boundary which corresponds to the northwestern limit of crustal reactivation associated with the Trans-Hudsonian orogeny (Lewry and Sibbald, 1980; Thomas and Gibb, 1985).

An early Proterozoic age for the development of the suture was proposed by Hoffman (1989), based on the apparent truncation of the circa 2.02 to 1.9 Ga Taltson magmatic arc of northeastern Alberta by the Snowbird Tectonic Zone in the subsurface; and the age of sedimentary rocks, lamprophyre dykes and related volcanic rocks of the Dubawnt Supergroup (circa 1.875 Ga), interpreted to form an overlap assemblage at the Rae-Hearne interface near Baker Lake. However, recent work along a segment of the Snowbird Tectonic Zone by Hanmer et al. (1992) has demonstrated that deformation under granulite facies conditions occurred at ca. 3.2 and ca. 2.6 Ga.

Previous studies on segments of the Snowbird Tectonic Zone

Previous work has focused on five segments of the 2800 km length of the proposed Snowbird Tectonic Zone. These include the East Athabasca mylonite zone and the Tulemalu fault zone southwest of Baker Lake; the Uvauk and Daly Bay complexes east of Baker Lake; and the Kramanituar complex near Baker Lake (Fig. 1). These studies have emphasized different aspects of the geology of each particular segment, and the structural history of the Snowbird Tectonic Zone as a whole remains enigmatic. Results of these studies are summarized briefly below.

The East Athabasca mylonite zone in northern Saskatchewan (Fig. 1) is an anastomosing system of granulite- and amphibolite-grade mylonites that developed in response to bulk pure shear during northwest-directed shortening (Hanmer et al., 1991, 1992). On a local scale, these rocks record a complex kinematic history which is interpreted to reflect partitioning of the bulk pure shear flow into coeval zones of dextral, sinistral and extensional (southwest-side-down) ductile flow. About 300 km southwest of Baker Lake, the Tulemalu fault zone (Fig. 1) contains high-pressure mafic granulite xenoliths hosted in ductilely deformed granitoid gneiss (Tella and Eade, 1986). Movement on the fault is not supported by direct field relationships but is postulated to involve reverse, west-side-up ductile displacement owing to the occurrence of granulite gneiss west of the fault; followed by normal, west-side-down faulting (Tella and Eade, 1986).

About 200 km east of Baker Lake on Chesterfield Inlet, anorthositic rocks and granulite-grade orthogneiss were reported during reconnaissance mapping by Tella et al. (1992). These rocks, designated the Uvauk complex (Fig. 1), were mapped in more detail during the 1992 field season and are described in this volume (Tella et al., 1993). The Daly Bay complex on Hudson Bay (Fig. 1) comprises felsic to mafic granulite gneiss, gabbroic anorthosite and migmatite and is interpreted as an allochthonous complex thrust northwestward from lower crustal levels (0.55 GPa, 750°C) onto amphibolite-grade gneiss during the Early Proterozoic (Gordon, 1988).

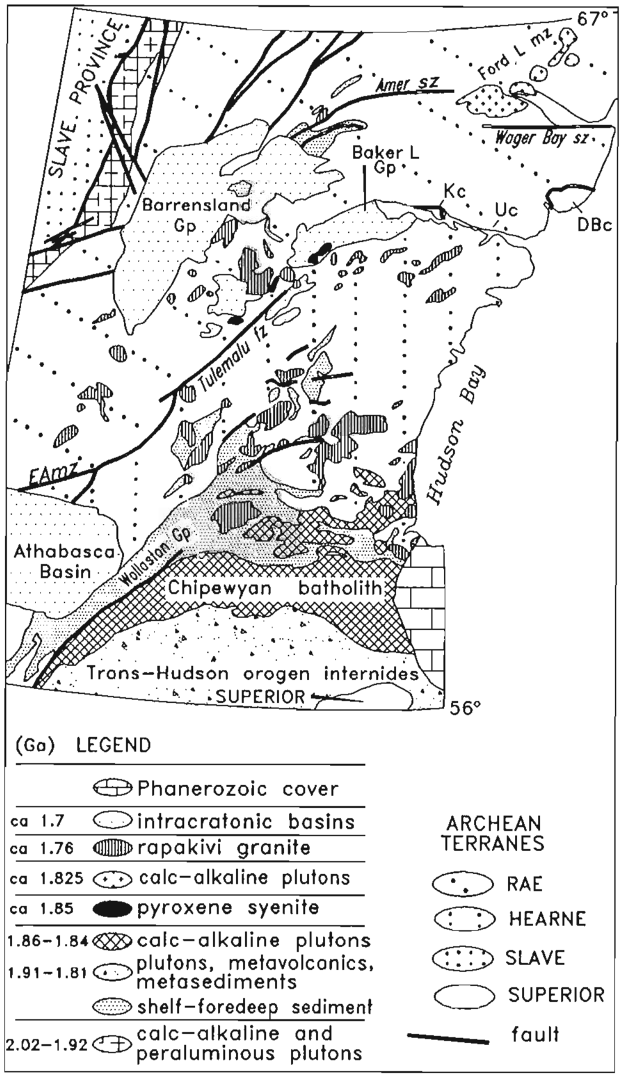


Figure 1. Geology of the western Canadian Shield after Hoffman (1989). Segments of the proposed Snowbird tectonic zone, at the interface between the Rae and Hearne provinces, include the East Athabasca mylonite zone (EAmz), the Tulemalu fault zone, the Kramanituar complex (Kc), the Uvauk complex (Uc) and the Daly Bay complex (DBc).

Layered granulites and anorthositic rocks of the Kramanituar complex near Baker Lake (Wright, 1967; Schau and Hulbert, 1977; Schau and Ashton, 1980; Schau et al. 1982; Schau, 1983) are situated close to a significant bend in the regional-scale geophysical lineament from northeast- to east-trending (Fig. 1). These high-grade rocks were described as prominently layered, strongly foliated and conspicuously lineated (Schau and Hulbert, 1977; Schau and Ashton, 1980), features that suggested these rocks had undergone significant tectonic modification.

This study

A detailed structural and petrological study of the Kramanituar complex was commenced during the 1992 field season in order to clarify the tectono-metamorphic history of this segment of the Snowbird Tectonic Zone. Previous mapping of well exposed and accessible rocks of this complex (Schau and Hulbert, 1977; Reinhardt et al., 1980; Schau and Ashton, 1980; Schau, 1983) provides an excellent geological framework from which detailed studies can proceed. The location of this complex close to a bend in an interpreted regional-scale crustal break lends added importance to this segment of the Snowbird Tectonic Zone. An understanding of the structural and metamorphic history of this segment is critical if comparisons are to be made between the Kramanituar complex and high-grade rocks of the East Athabasca mylonite zone to the southwest, or the Uvauk and Daly Bay complexes to the east (Fig. 1).

LITHOLOGICAL UNITS WITHIN THE COMPLEX

The Kramanituar complex, exposed over roughly 60 km by 30 km, was proposed by Schau and Ashton (1980) to include rocks exposed east of Baker Lake with mineral assemblages reflecting granulite facies metamorphism. The complex consists of a central suite of anorthositic rocks enveloped to the north and south by layered granulite-grade rocks. The anorthositic suite comprises mainly gabbroic anorthosite, gabbro and leucogabbro, with minor anorthosite, anorthositic gabbro and ultramafic rock. Adjacent granulite-grade rocks include paragneiss, diatexite and various metaplutonic rocks. The geology of the complex and adjacent rock units is shown in Figure 2.

Anorthositic suite

The anorthositic suite is dominated by light buff to white weathering gabbroic anorthosite (Fig. 2a) composed of sodic labradorite (Schau and Hulbert, 1977) with 12% to 20% orthopyroxene±clinopyroxene±garnet. Magnetite±ilmenite are accessory phases. Where these rocks are not penetratively strained and recrystallized, they are generally coarse grained and contain up to 60% phenocrysts of plagioclase (average 2 cm by 1 cm). Orthopyroxene porphyritic phases occur at several localities along the northeast margin of the complex and in its centre (Schau, pers. comm., 1992). Rarely are both plagioclase and orthopyroxene phenocrysts observed (Fig. 3a).

Gabbro and leucogabbro occur as 1 to 100 cm-wide layers in gabbroic anorthosite and as homogeneous units up to 2 km wide (Fig. 2a). Centimetre-scale layers are rusty brown weathering, equigranular, fine- to medium-grained (0.5-2 mm), with 30-40% pyroxene, 0-15% garnet and locally up to 2% magnetite. These rocks generally comprise 2-15% of exposures of predominantly gabbroic anorthosite. The gabbroic layers are generally tectonized, straight, uniform in composition across their breadth, with margins that typically are in discrete contact with adjacent gabbroic anorthosite. Locally, gabbroic layers are observed to thin and pinch out parallel to strike. Although these layers are tectonized, they are believed to reflect primary magmatic layering. Rarely, modal layering is present in less strained rocks (Fig. 3b).

Gabbro and leucogabbro, 1 to 2 km wide, occur as discrete units in the northeast part of the complex (Fig. 2a). These dark green to rusty brown weathering rocks contain plagioclase, 15-25% pyroxene, 2-20% garnet, and may contain <2% magnetite and <5% quartz. Patches with up to 50% garnet and 30% interstitial quartz occur locally.

Rocks of true anorthosite composition (colour index <10) form a minor component of the anorthositic suite. They occur mainly as subangular, metre-scale xenoliths of fine grained rock hosted in gabbroic anorthosite (Fig. 2a; locality A). Anorthosite (±gabbroic anorthosite) also occurs as tabular units, 10 to 30 m wide and several kilometres long, within paragneiss. These units are generally layered and locally such layering is thought to be primary. By virtue of their tabular geometry, layered nature and the presence of modal layering locally, these distinctive marker horizons are interpreted as sills and considered as part of the anorthositic suite.

Granulite suite

Granulite-grade rocks exposed north and south of the anorthositic suite are well-layered rocks composed mainly of various proportions of plagioclase+orthopyroxene+clinopyroxene±garnet±quartz±K-feldspar. These rocks are subdivided according to their protolith as paragneiss and metaplutonic rocks. Further subdivision of the metaplutonic rocks is based on textural and compositional variations.

Paragneiss

Layered paragneiss is exposed as two elongate units north and south of the anorthositic suite, and in the extreme northeast part of the complex at locality B (Fig. 2a). These rocks consist of fine grained, rusty weathering, centimetre- to metre-wide layers interpreted as feldspathic wacke which alternate with white-weathering, quartzo-feldspathic leucosome, 1-30 cm wide (Fig. 4). The feldspathic wacke (paleosome) is composed mainly of plagioclase, 15-20% biotite, 5-15% quartz and up to 8% garnets. These rocks may contain up to 10% graphite, 1% to 5% sillimanite, 2% to 10% kyanite, K-feldspar, pyrite and rutile. Locally hypersthene is present in addition to biotite, or 5% to 15% hypersthene occurs in the absence of biotite. Graphite and sillimanite appear to be fairly widespread, while kyanite-bearing rocks are restricted to two

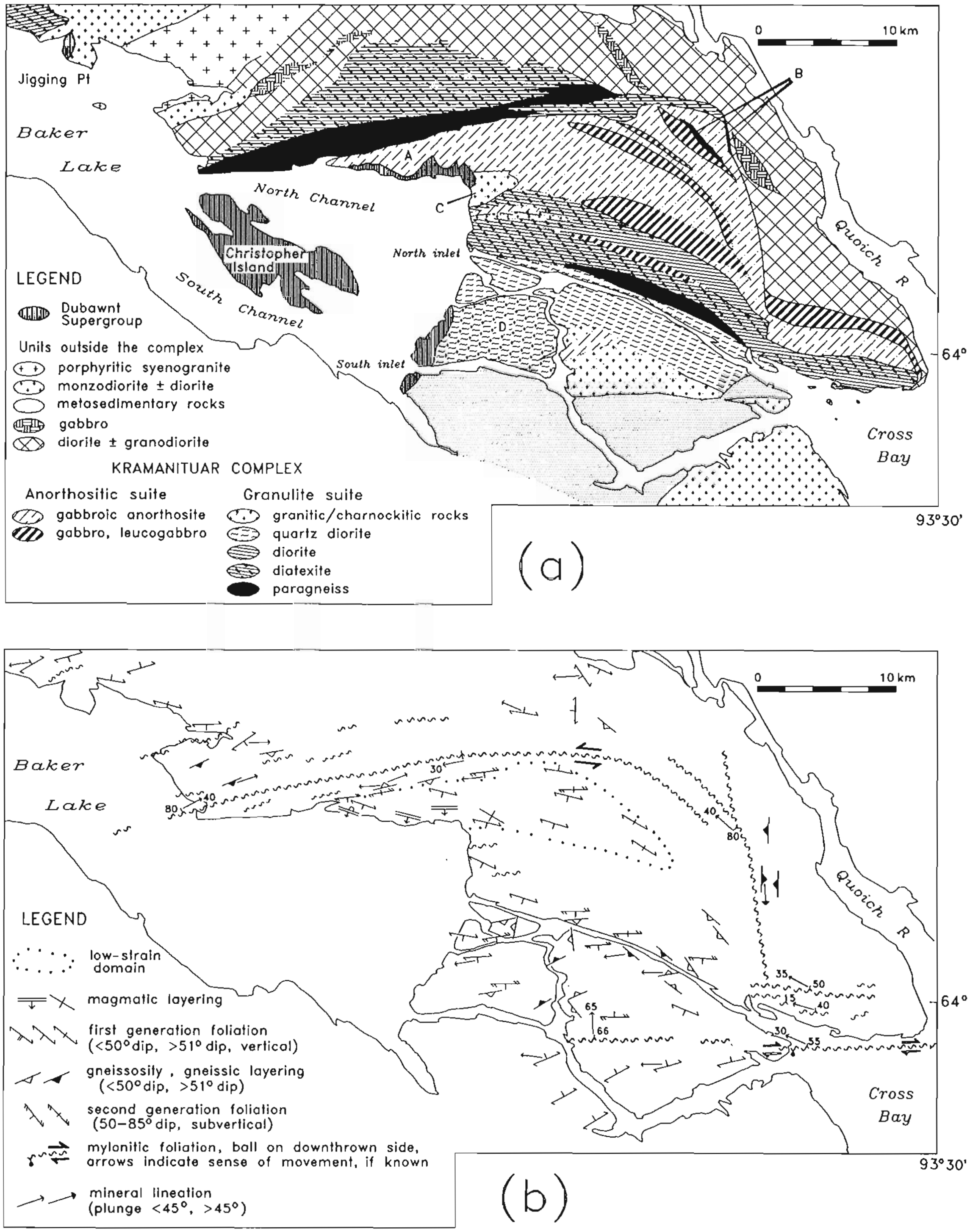


Figure 2. General geology of the Kramanituar complex and adjacent units. **a**) Lithological units modified from Schau and Ashton (1980) and Schau (1983). Localities A, B, C and D referred to in the text. **b**) Major structural elements.

narrow panels, several hundred metres wide and several kilometres long, in the northeast part of the complex (locality B in Fig. 2a). Leucosome is dominantly composed of plagioclase, 20-30% ribbon quartz, up to 10% K-feldspar and 1-10% garnet. Minor phases are graphite and sillimanite. Kyanite occurs in leucosome at locality B. Leucosome generally comprises between 2% and 20% of paragneiss (Fig. 4a), but may comprise up to 60%.

Centimetre-scale layers of pink weathering, garnet-bearing leuco-monzogranite (described below) form a minor component of the paragneiss locally. Anorthosite sills (described above) and gabbroic dykes and/or sills crosscut paragneiss.

Metaplutonic rocks

Diatexite

Pale buff weathering, quartzo-feldspathic diatexite is a distinctive unit which is widespread throughout the complex (Fig. 2a), and occurs northwest of Jigging Point where it is interpreted as an outlier of the complex (Schau et al., 1982). The diatexite is a texturally distinct unit owing to its medium to coarse grain size and pock-marked appearance created by the presence of 1-5 mm red anhedral garnets draped with ribbon quartz. These rocks contain plagioclase, 10-15% quartz (locally up to 30%), 2-15% garnet, 10-15% very fine pale grey-brown pyroxene, and up to 10% K-feldspar. Finely disseminated biotite may be present in addition to pyroxene. At several localities, rare porphyroclasts of resinous brown pyroxene <2 mm in length were observed. These rocks generally possess a mylonitic foliation. In contrast to other units, the diatexite displays a relatively homogeneous texture with indistinct centimetre- to metre-scale banding reflecting subtle variations in colour index and strain intensity. These rocks commonly occur as 5-10 m wide sheets or panels that alternate with paragneiss or garnetiferous gabbro.

Lithological relationships observed in less strained and metamorphosed rocks in the southeast part of the complex suggest that the diatexite is a composite unit of two plutonic rocks: a dioritic host cut by pervasive centimetre-wide dykes of leuco-monzogranite (described below). Metamorphism with attendant anatexis and deformation of these two units are believed to have resulted in the homogenization of these two units to produce this distinctive garnet-bearing quartzo-feldspathic diatexite. A transition from relatively unstrained,

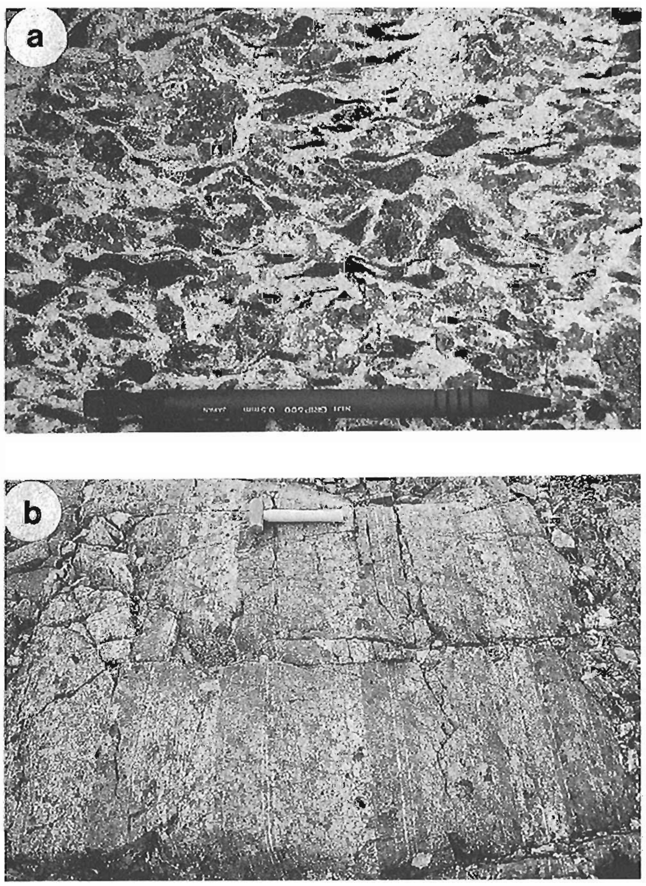


Figure 3. Anorthositic suite. **a**) Plagioclase and orthopyroxene porphyritic gabbroic anorthosite (GSC 1992-272W). **b**) Modal layering in deformed gabbroic anorthosite showing systematic decrease in mafic minerals toward right side of photo. (GSC 1992-272J)

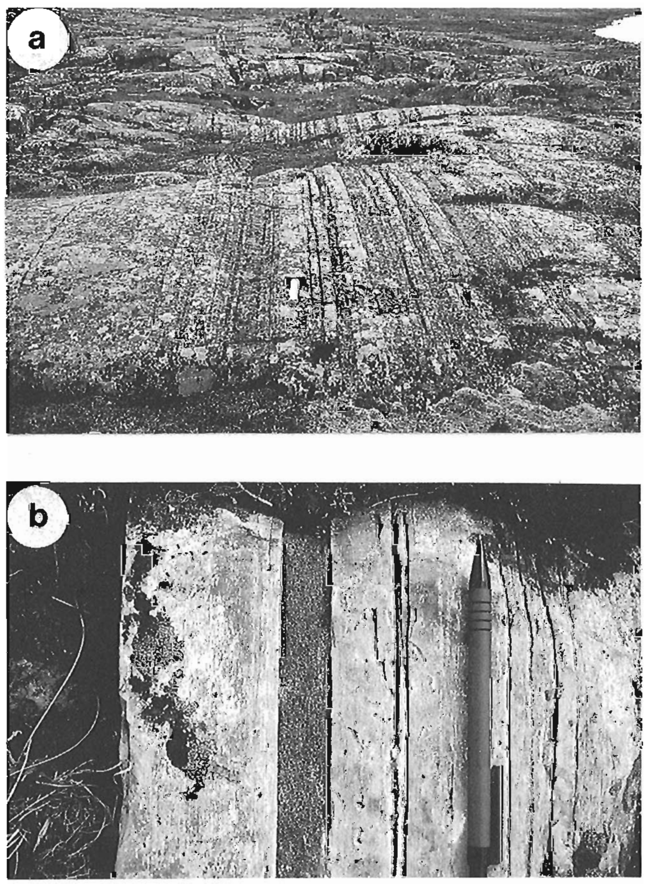


Figure 4. Tectonized, granulite-grade paragneiss. **a**) Centimetre- to metre-scale layers of feldspathic wacke alternating with white weathering leucosome (GSC 1992-272S). **b**) detail of a. (GSC 1992-272I)

unmetamorphosed plutonic protoliths to diatexite was observed northwest of Cross Bay (Fig. 2) along several across-strike traverses.

Quartz diorite and diorite

Magnetite-bearing leucocratic quartz diorite (±diorite) is widespread across the southern part of the complex (Fig. 2a). These light grey to buff-weathering, medium-grained rocks are composed of plagioclase, 5-15% pyroxene and amphibole, 5-10% quartz, and notable (2-8%) magnetite which occurs as disseminated grains and rarely as layers <1 cm wide. These rocks are moderately to strongly foliated to gneissic.

Narrow panels of fine grained diorite are exposed north of North inlet (Fig. 2a). These are dark grey-green weathering, weakly to moderately magnetic rocks with a colour index of 30 to 40. These rocks are believed to represent the dioritic component of the diatextite which occurs adjacent to, and is transitional with, these rocks.

Granitic rocks

Pink-weathering granitic rocks are exposed at two localities along the eastern shore of Baker Lake (locality C and D) and in the southern part of the complex (Fig. 2a). At locality C, quartz-monzodiorite and monzodiorite contains 5-20% ribbon quartz, 5-10% very fine grained pyroxene±biotite, plagioclase, K-feldspar±magnetite±ilmenite. K-feldspar generally makes up 10-50% of the total feldspar. These rocks are generally very highly strained. They are intruded by a swarm of moderately foliated, 5-20 m wide gabbro dykes, which limit exposures of the granitic rocks to narrow panels less than 3 m wide. At locality B, moderately strained, pink- to orange-weathering monzogranite and leuco-monzogranite (charnockite) is composed of plagioclase, K-feldspar, 20-25% quartz and 5% very fine grained mafic minerals which generally appear to be pyroxene. K-feldspar typically comprises 30-40% of the total feldspar. These rocks are commonly interlayered with dioritic to quartz dioritic rocks, however, in contrast to locality C, the granites here cut their dioritic host. The granitic rocks are interpreted to be the granitic component of the diatexite observed to the north.

Granitoid rocks in the southern part of the complex are similar in composition to those at locality B and C. These are strongly deformed rocks that appear to form a discrete plutonic body (Fig. 2a).

UNITS EXTERNAL TO THE COMPLEX

Amphibolite-grade metaplutonic and metasedimentary rocks occur external to the Kramanituar complex (Fig. 2a). North and east of the complex, foliated to gneissic metaplutonic rocks are dominant. Metasedimentary rocks are exposed south of the complex.

Grey-weathering, foliated to gneissic rocks of dioritic to granodioritic composition (Fig. 5a), exposed north and east of the complex, are designated the Ingilik Point gneiss complex (Schau et al., 1982) and interpreted to be ca. 2675 Ma old (U-Pb; Schau, 1980). These are typically hornblende-bearing rocks composed of plagioclase, 15-20% hornblende±biotite, 5-20% quartz, and K-feldspar, which comprises up to 20% of the total feldspar. These rocks are moderately foliated to gneissic. In contrast to the straight layering typified by units within the complex, these rocks display fabrics of moderate intensity and variable attitude.

Moderately foliated, amphibolitic gabbro with a colour index of 40 to 50 occurs as discrete, elongate bodies north and east of the complex (Fig. 2a). The geometry and distribution of these bodies suggest they may represent a boudinaged mafic unit(s).

Pink weathering, K-feldspar porphyritic syenogranite, east of Jigging Point (Fig. 2a), contains up to 30% blocky subhedral K-feldspar phenocrysts (up to 2 cm by 5 cm) in a medium-grained groundmass of plagioclase, 15-20% quartz

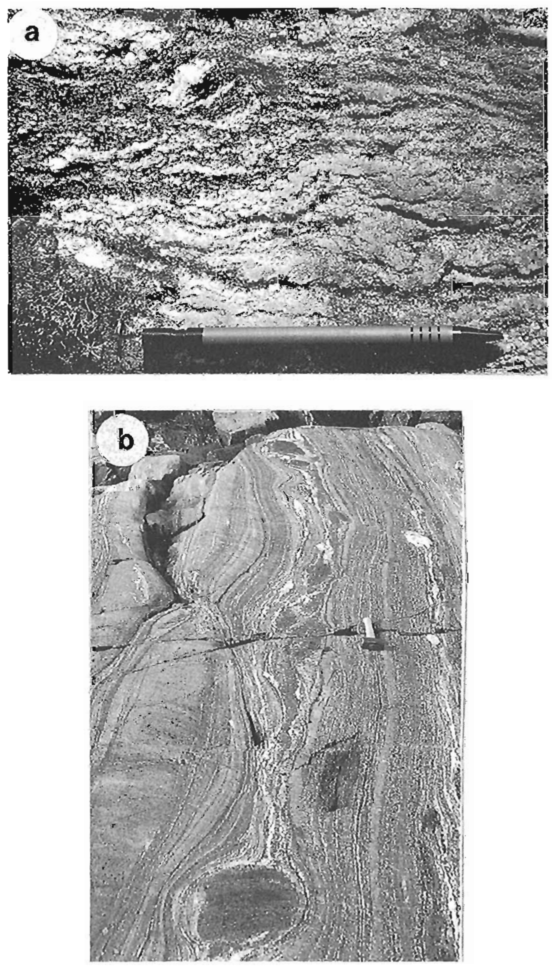


Figure 5. Lithological units external to the complex. **a)** Gneissic granodiorite north of the complex (GSC 1992-272M). **b)** Partially melted metasedimentary rocks south of the complex. (GSC 1992-272Q)

and 10-15% elliptical aggregates of biotite. Locally these rocks are cut by quartz veins with associated pyrite± chalcopyrite±fluorite.

Throughout the southern part of the map area amphibolite-grade metasedimentary rocks are widespread (Fig. 2a, 5b). These rocks form a relatively homogeneous sequence of moderately to strongly foliated, biotite- and/or garnet-bearing feldspathic wacke. These are light grey weathering rocks mainly composed of plagioclase, 10-20% black to reddish-brown biotite, 5-12% quartz and <10% pale pink garnet. Locally these rocks contain minor tremolite, talc, sillimanite and possible pyrophyllite. Magnetite ironstone and magnetite-bearing wacke occur as 1-100 cm wide units at several localities.

STRUCTURE

Planar and linear elements within the complex

Gabbroic anorthosite in the central part of the complex is relatively unstrained and displays some of the earliest fabrics recognized. These include primary magmatic and modal layering (S_0) and a weak to moderately developed tectonic foliation (S_1). S_0 strikes east-southeast and dips steeply south (Fig. 2b). Rare modal layering, showing a systematic decrease in colour index across layers toward the south, suggests that this part of the intrusion may young to the south. In these relatively unstrained rocks, S_1 strikes east-southeast, dips steeply (73-85°) south, and is subparallel to S_0 or forms a small (2-15°) counterclockwise angle. S_1 rarely displays a weak, shallow, west-plunging mineral lineation defined by aligned pyroxene.

South of the low-strain domain, gneissic layering (S_g) is pervasive throughout quartz diorite and diorite. Most commonly S_g strikes southeast and dips moderately (40-60°) to the southwest. Several fold closures (ie., domal pattern) are suggested by variations in the strike and dip of S_g (Fig. 2b), alternatively, these variations may reflect an anastomosing character of the gneissic fabric. Irrespective of the attitude of S_g, mineral lineations in these rocks plunge moderately (15-50°) to the west. The relative timing of development of S_g and S_1 is not yet established, however both of these fabrics predate, and are overprinted by S_2.

A strong to intense (mylonitic) foliation is regionally developed throughout the Kramanituar complex, with the exception of the low-strain domain in its centre. North and east of the low-strain domain, a progressive increase in strain intensity culminates in kilometre-wide zones of ultramylonite that parallel, and appear to define, the northern and eastern boundaries to the complex (Fig. 2b). This fabric is designated S_2, since an overprinting relationship with S_1 can be established at several localities.

The nature of S_2 is variable, and appears to be largely controlled by the mineralogy and relative competency of the lithological units in which it is developed. In gabbroic anorthosite, S_2 is manifest as a streaky, discontinuous foliation formed by recrystallized aggregates of feldspar, pyroxene and garnet. In paragneiss, S_2 is defined by straight, continuous, alternating layers of feldspathic wacke and leucosome (Fig. 4), and by ribbons of quartz and aligned mafic minerals within these units. In gabbro and leucogabbro, S_2 is typically manifest as a mylonitic foliation characterized by millimetre-scale laminations.

A progressive change in the attitude of S_2 can be mapped across the north and east parts of the complex (Fig. 2b). Across the northern part of the complex, S_2 strikes east-northeast to east and dips steeply to the north and south. Mineral lineations are strongly developed, and plunge shallowly to the west and east. Toward the eastern part of the complex, S_2 progressively swings in strike from east to south-southeast and dips steeply (70-85°) to the southwest. Strongly to intensely developed mineral lineations plunge west and northwest, and progressively steepen to a maximum plunge of 55° north-northwest along the eastern boundary of the complex.

South of the low-strain domain, an abrupt increase in strain intensity is revealed by a mylonitic foliation (S_2) displayed by granitic rocks at locality C (Fig. 2). These fabrics strike east, dip south (60-85°) and possess shallow (2-30°) east- and southwest-plunging mineral lineations. Farther south, a moderate to strong, east-striking, steeply south-dipping S_2 foliation overprints S_g in plutonic rocks (Fig. 2b).

In the southeast part of the complex, north of Cross Bay (Fig. 2), a strong to intense east-striking tectonic foliation is observed. In contrast to the steep dip of other high-strain parts of the complex, foliations here dip moderately (40-70°) north. Strongly to intensely developed mineral lineations plunge 15-45° to the northwest.

Ultramylonite zones

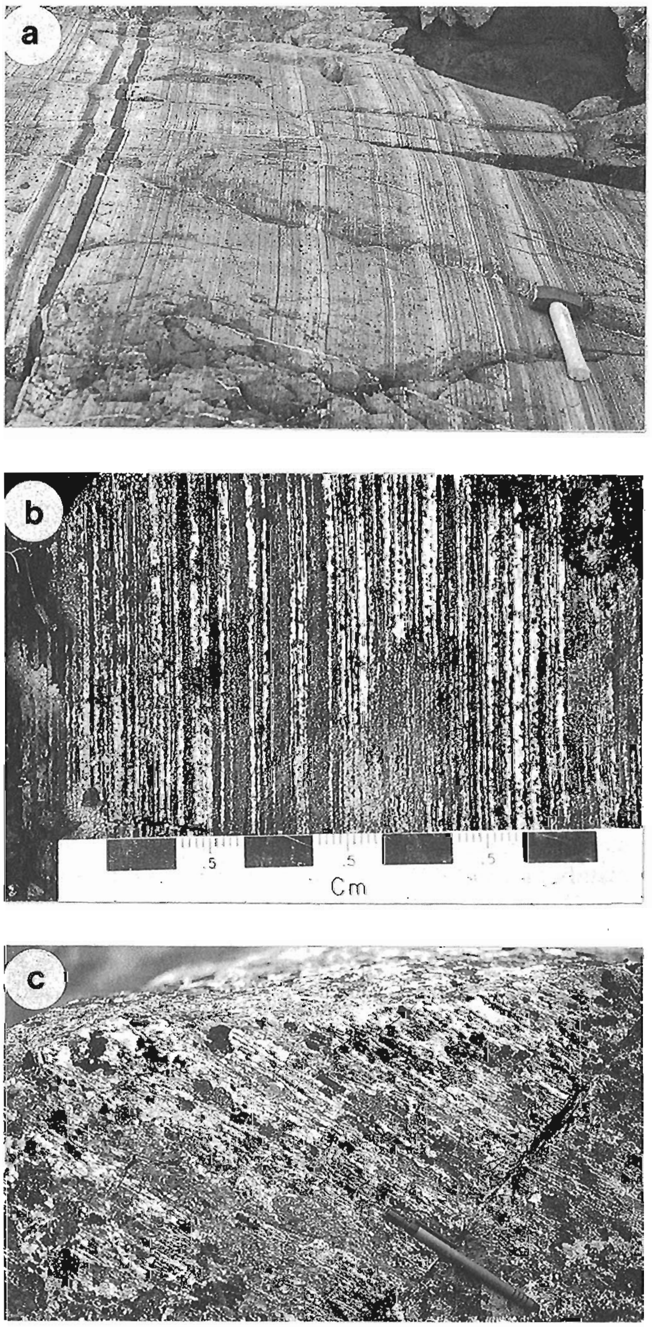
The degree of strain recorded by most units throughout the Kramanituar complex is generally very strong. Rocks characterized by intense ductile deformation appear to form discrete zones across the northern and southern margins of the complex (Fig. 2b). These zones are developed in a variety of rock types, including gabbro, gabbroic anorthosite, paragneiss and granite; all of which display ultramylonitic fabrics reflecting extreme reduction in grain size. Ultramylonitic paragneiss, gabbro, and gabbroic anorthosite occur along the northern boundary of the complex (Fig. 6), and mainly mylonitized granitic rocks occur along its southern boundary (Fig. 7). Weathered exposures of rocks within these zones are generally characterized by remarkably straight, continuous millimetre-scale laminations. Fresh surfaces are typically aphanitic and, at a hand sample scale, are virtually textureless.

The northern zone of mylonitic rocks is about 2 km wide and is curvilinear in its geometry. This zone of rocks strikes east and dips vertically in the northwest, and progressively changes to southeast- and south-striking with 70-80°SW dips in the eastern part of the complex (Fig. 2b). Strongly developed mineral lineations generally plunge shallowly to moderately to the west and northwest but may plunge shallowly east. At one exposure, east- to west-plunging lineations displayed across a single foliation surface demonstrate the curvilinear nature of the displacement vectors along this part of the zone. Kinematic indicators within the northern mylonite zone were mainly recognized in the northeast part of the complex. Asymmetric wings on garnet and pyroxene (Fig. 6d) in gabbro, and shallow plunging S-folds in paragneiss, in conjunction with generally moderate west-plunging mineral lineations indicate sinistral, south-side-up displacement.

The southern ultramylonite zone is about 500 m wide and strikes uniformly east and dips 30-65°N (Fig. 7a). Mineral lineations vary in plunge from down-dip in the west, to moderate (30°NW) in the eastern part of the zone. Kinematic indicators, including dextral offset along shear bands (Fig. 7b), asymmetric wings of K-feldspar phenocrysts (Fig. 7c), and folds of Z-asymmetry (Fig. 7d), record oblique, north-side-up, dextral shear.

Planar and linear elements external to the complex

Foliated and gneissic plutonic rocks north of the complex display southeast-striking foliations that dip moderately to shallowly (60-25°) to the southwest and northeast (Fig. 2b). Mineral lineations plunge moderately (10-35°) to the west. East of the complex, dioritic orthogneiss strikes north-northeast and dips steeply to the east and west (Fig. 2b).



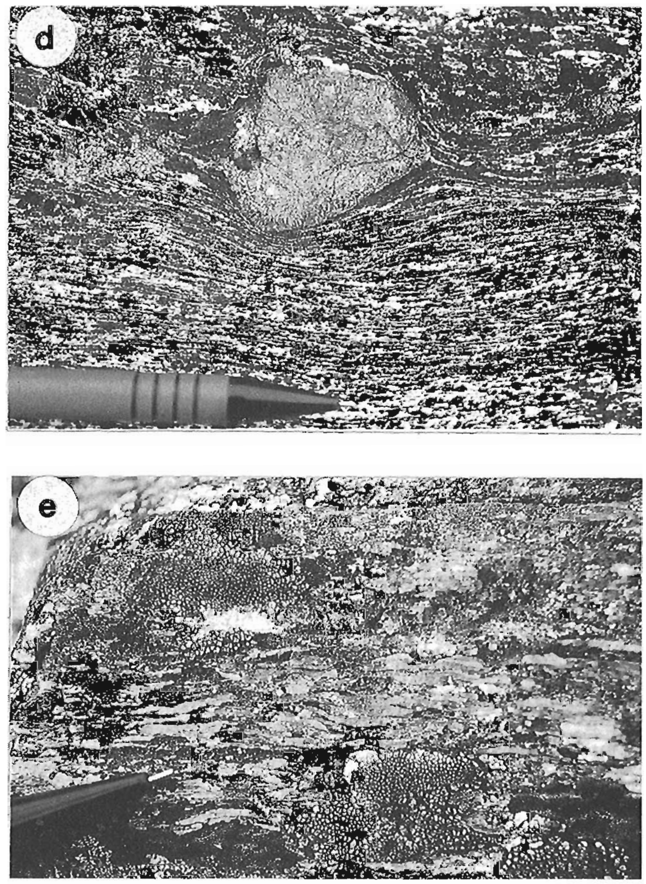


Figure 6. Mylonitic rocks along the northern margin of the complex. **a**) Tectonic layering in mylonitized gabbroic rocks (GSC 1992-272K). **b**) Gabbro ultramylonite with two pyroxene bands alternating with plagioclase bands (GSC 1992-272V). **c**) Moderately west-plunging linear fabric displayed by diatexite (GSC 1992-272Y). **d**) Sinistrally rotated orthopyroxene porphyroclast with fibrous fringe of amphibole (GSC 1992-272U). **e**) Shear bands showing sinistral shear parallel to pencil (GSC 1992-272D).

Mineral lineations plunge to the southwest and southeast at 50-80°, representing the steepest linear fabrics observed in the map area.

Metasedimentary rocks south of the complex display a moderately developed, west- to southwest-striking foliation that dips 40-60°NW. Moderately developed mineral lineations consistently plunge 15-35° to the west and northwest.

DISCUSSION

One of the most striking features of rocks of the Kramanituar complex, apart from their high metamorphic grade, is the pervasive development of strong to intense, penetrative tectonic fabrics. This study will investigate the local and regional significance of these fabrics and their kinematics, and will attempt to unravel as much of the early history of the area as possible. Several preliminary interpretations relating to the magmatic, structural and metamorphic history of the area are summarized below.

The anorthositic suite, as noted by Schau and Hulbert (1977), displays evidence of a cumulate history and is interpreted to have originated as a sill-like intrusive complex owing to the presence of magmatic layering, and particularly modal layering, in relatively unstrained parts of the intrusion. A change to more gabbroic composition toward the northeast part of the intrusion could be considered to be consistent with an interpretation of southward younging.

Two lines of evidence suggest that anorthositic magma was emplaced at some depth, and then crystallized prior to the deformation events recorded by these rocks. The first concerns the presence of leucosome, not only as a significant component of paragneiss, but also interlayered with gabbroic rocks in the northeast part of the complex. Both of these units lack textural evidence (such as restite development) of advanced partial melting required to generate such volumes of garnet±sillimanite±kyanite quartzo-feldspathic leucosome. This suggests that the leucosome was not generated in situ, but was injected (intruded) into these units. The volume of injected leucosome suggests significant melting occurred at depth. The mylonitized nature and straight gneissic layering of the paragneiss (Fig. 4) reveals that melting took place prior to significant deformation. The second line of evidence concerns the anorthositic units in paragneiss, interpreted to be sills. These sills are not fault bounded, though undoubtedly they are transposed in parallelism with the tectonic fabric

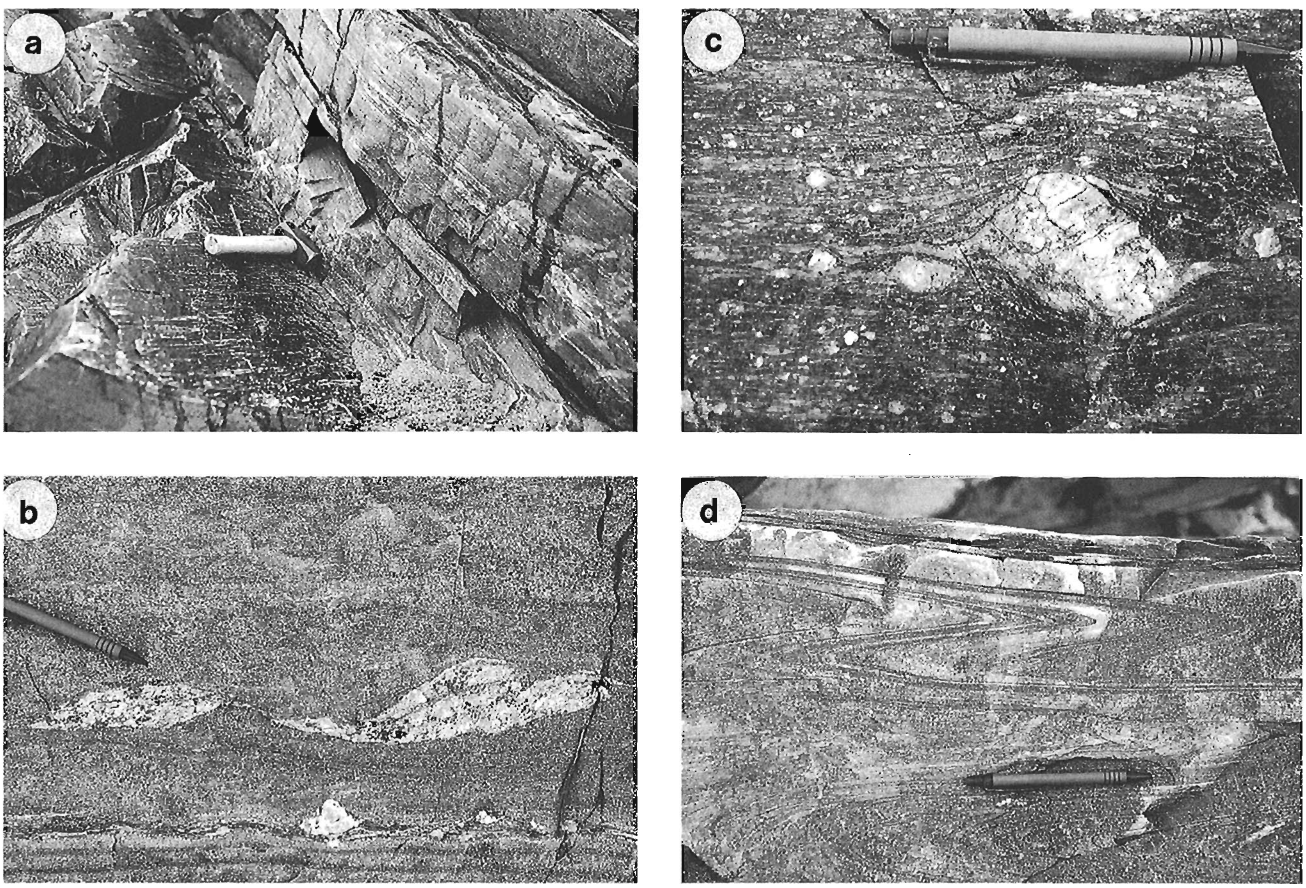


Figure 7. Mylonitic rocks along the southern margin of the complex. **a**) shallow-dipping mylonitic fabric in monzogranite, note strongly developed lineations parallel to hammer (GSC 1992-272F). **b**) back-rotated leucocratic pods resulting from dextral shear parallel to pencil (GSC 1992-272E). **c**) asymmetric wings of K-feldspar porphyroclast indicating dextral shear (1992-272A). **d**) Z-folds in granitic mylonite (GSC 1992-272D).

which penetrates the complex. Their presence strongly supports an intrusive relationship between the anorthositic suite and a metasedimentary host.

A number of lithological and structural similarities within and external to the complex were noted which should be considered, particularly in light of a tectonic model involving collision of two terranes (Hoffman, 1988). Granulite-grade rocks of metasedimentary origin are exposed within the complex, both north and south of the anorthositic suite (Fig. 2a). Amphibolite-grade feldspathic wacke, exposed south of the complex is compositionally similar (ie., quartz content) and may represent a lower grade equivalent. Some structural fabrics are similar in attitude inside and outside of the complex (Fig. 2b). For instance, moderate to shallow dipping gneissic fabrics (S_g) are similar in attitude to gneissic fabrics displayed by metaplutonic rocks north of the complex, and to foliations in metasedimentary rocks south of the complex.

Ultramylonitic rocks occur in zones that are parallel to, and appear to define the boundaries of the Kramanituar complex. Movement on the north and south mylonite zones is consistent with uplift of granulite-grade material relative to the amphibolite-grade rocks that envelop the complex. However, it remains to be demonstrated whether the finite strain recorded by the shear zones reflect movement responsible for juxtaposing these disparate metamorphic terranes, or whether the shear zones are relatively late features that dissect a previously assembled metamorphic terrane in which earlier events are cryptic in nature.

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Geological Survey of Canada Project 870008

Geology of Archean and Proterozoic supracrustal rocks in the Padlei belt, southern District of Keewatin, Northwest Territories[1]

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Aspler, L.B., Chiarenzelli, J.R., and Bursey, T.L., 1993: Geology of Archean and Proterozoic supracrustal rocks in the Padlei belt, southern District of Keewatin, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 147-158.

Abstract: Henik Group volcano-sedimentary subunits are intruded by Archean granitoids; both are unconformably overlain by the Hurwitz Group. Mapping and preliminary petrography confirm that rocks previously considered Montgomery Lake Group (unconformably between the Henik and Hurwitz groups in its type-area) are lower Hurwitz Group in the Padlei belt. The lower Hurwitz Group defines a conformable onlap sequence of radial basin expansion. Sub-Whiterock Member units represent paleovalley fill, fluvial, eolian, and lacustrine sedimentation. Near the base, interbedding of discontinuous pyritic quartz-pebble layers and pyrite-free redbed arkoses, quartz arenites, and quartz-pebble layers suggests deposition at the oxyatmoversion transition. Abrupt transition from shallow-water quartz arenites (Whiterock Member) to deep-water mudrocks (Ameto Formation) is attributed to intraplate stress-induced deepening and arching. Basement and cover were deformed by northwest-trending folds and northwest-vergent thrusts and folds that result in dome and basin interference. Highest gold values are in basement quartz veins and pyritic-chert horizons, and in lower Hurwitz conglomerates.

Résumé : Les sous-unités volcano-sédimentaires du Groupe de Henik sont traversées par des intrusions de granitoïdes archéens; le tout est recouvert en discordance par le Groupe de Hurwitz. Les études cartographiques et les études pétrographiques préliminaires confirment que des roches anciennement attribuées au Groupe de Montgomery Lake (placé en discordance entre les groupes de Henik et de Hurwitz dans sa localité type) se situent dans la partie inférieure du Groupe de Hurwitz dans la zone de Padlei. La partie inférieure du Groupe de Hurwitz définit une séquence concordante à biseaux ascendants caractéristique d'un mode d'expansion radiale dans un bassin. Les unités du membre sous-jacent au Membre de Whiterock représentent le remplissage sédimentaire d'une ancienne vallée, donc une sédimentation fluviatile, éolienne et lacustre. Près de la base, des lits interstratifiés composés de couches pyriteuses discontinues à galets quartzeux et d'arkoses de couches rouges dépourvues de pyrite, de quartzites sédimentaires et de lits de galets quartzeux, semblent indiquer que la sédimentation se serait produite pendant la période de transition à un nouvel environnement contenant l'oxygène atmosphérique. Un passage abrupt des quartzites sédimentaires accumulés en eau peu profonde (Membre de Whiterock) à des mudrocks de mer profonde (Formation d'Ameto) est attribué aux phénomènes intraplaques induits par des contraintes, d'approfondissement et de déformation en voûte. Le socle et la couverture ont été déformés par des plis de direction générale nord-ouest et par des failles chevauchantes et des plis de vergence nord-ouest qui ont donné lieu à un modèle d'interférence de dômes et de bassins. Les plus fortes teneurs en or apparaissent dans les filons quartzeux du socle et dans les horizons à chert pyriteux, et dans les conglomérats de la partie inférieure du Groupe de Hurwitz.

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INTRODUCTION

This paper summarizes the initial results of the second field season of a Canada-Northwest Territories Mineral Initiative (1991-1996) project designed to better understand the geology of Archean and Proterozoic supracrustal rocks in the south-central District of Keewatin: the Henik Group (Archean, part of the Ennadai-Rankin greenstone belt); the Montgomery Lake Group (age uncertain; in its type-area, unconformably between the Henik and Hurwitz groups) and the Hurwitz Group (Early Proterozoic). Problems outstanding from the first year's mapping (Aspler et al., 1992a) include: 1) the internal stratigraphy of the Henik Group; 2) the status of the "Montgomery Lake Group" in the Padlei belt ; 3) the origin of pyritic quartz pebble conglomerates in the Padlei belt (historical gold and uranium exploration targets); 4) the stratigraphy, depositional setting, original geometry, and tectonic significance of Hurwitz Basin; and 5) the structure of the Hurwitz Group and the extent to which Archean rocks were affected by Proterozoic deformation. Below we address these issues based on completion of 1:50 000-scale mapping in the Padlei belt (Fig. 1; Aspler et al., 1992b).

HENIK GROUP

The Henik Group is assumed to be Archean on the basis of U/Pb zircon ages determined elsewhere in the Ennadai-Rankin greenstone belt: Tavani area (2660-2680 Ma, Park and Ralser, 1991); Kaminak Lake (2697 ± 14; 2692 ± 1 Ma, Mortensen and Thorpe, 1987; 2681 ± 3 Ma, Patterson and Heaman, 1990); and northeastern Saskatchewan (2682 ± 5.9 Ma, Chiarenzelli and Macdonald, 1986; 2708 +3/-2 Ma, Delaney et al., 1990). The stratigraphic order of Henik Group map units (described in Table 1) is uncertain owing to a paucity of top indicators and poor exposure in critical contact areas. We tentatively suggest that the mixed sedimentary-volcanic unit (As) underlies the mafic volcanic dyke-sill complex (Am), based on the possible continuity between this sequence at the western end of Ameto Lake with rocks exposed in the core of the large Proterozoic anticlinorium between Montgomery and North Henik lakes (Eade, 1974; Aspler et al., 1992a), and that the section is capped by the rhyolite subunit (Af) because of consistent southwest-tops in the mafic complex on the northwest side of the Padlei belt.

HURWITZ GROUP

Status of the "Montgomery Lake Group" in the Padlei belt and stratigraphic revision of the lower Hurwitz Group

Montgomery Lake area

In the Montgomery Lake area, Eade (1964) mapped a sequence of siliciclastic rocks (his map unit 7) that is unconformably above the Henik Group and unconformably beneath the Hurwitz Group. Eade (1974) formally defined this unit as the Montgomery Lake Group, with a type section on the east side of Montgomery Lake. The geology of the Montgomery Lake Group is the topic of an M.Sc. thesis by Bursey (Carleton University). Recent mapping in the Montgomery Lake area supports Eade's (1974) conclusion of an angular unconformity separating the Montgomery Lake Group from the overlying Hurwitz Group: Montgomery Lake Group outliers are truncated by the Hurwitz Group, and local breccias in basal Hurwitz Group (bearing up to 90% Montgomery Lake Group clasts) truncate subjacent Montgomery Lake Group structure (Aspler et al., 1992a). The Montgomery Lake Group records a westward-draining fluvial system that drapes a low-relief paleotopography cut into tilted Henik Group strata. It consists of local basal breccia (talus deposits rimming paleohills); discontinuous polymictic conglomerate (fluvial channel-fill facies); local interbedded lithic sandstone and siltstone (pond facies) and pyrite-bearing cross-stratified litharenite and lithic wacke (principal fluvial system). Although a valid and useful stratigraphic unit, the distribution of the Montgomery Lake Group may be limited to its type area.

Padlei belt

In the Padlei belt, Bell (1970a) informally used the term "Montgomery Lake sedimentary rocks" to refer to a unit of predominantly subarkose and quartz arenite unconformably overlying the Henik Group, and concluded that these rocks were equivalent to Eade's (1964) map unit 7 (i.e. Montgomery Lake Group). In subsequent papers (Roscoe, 1973, 1981; Young, 1975) these rocks were referred to as the Montgomery Lake Group. For the purposes of discussion, the term "map unit X" is used herein for these rocks in the Padlei belt (Table 1; see Fig. 2 in Roscoe, 1981 for detailed subdivisions based on drilling data; Fig. 2 in Aspler et al., 1992a for measured section). The present mapping indicates that map unit X is more extensively developed on the northwest side of the Padlei belt than was previously recognized (Fig. 1; cf. Bell, 1970a).

From our initial mapping in the southern part of the Padlei belt, we questioned whether it was appropriate to designate map unit X as Montgomery Lake Group and tentatively concluded that these rocks constitute a conformable part of the lower Hurwitz Group (Aspler et al.,1992a). This conclusion is now extended to include the entire Padlei belt, based on completion of mapping and preliminary petrographic comparison between the Montgomery Lake Group in its type area, map unit X in the Padlei belt, and the lower Hurwitz Group.

Arenites within map unit X differ markedly from type-area Montgomery Lake Group, and are lithologically similar to, or indistinguishable from, rocks of the Maguse Member (compare Fig. 2A, B, C, D). In the type area, Montgomery Lake Group sandstones are texturally and compositionally immature, consisting of lithic wackes and litharenites with a high proportion of plagioclase and plagioclase-bearing clasts (Fig. 2A, B). In contrast, typical Maguse Member sandstones are relatively mature, consisting of subarkose to quartz arenite and lack detrital plagioclase (Fig. 2C). Rocks of map unit X are more mature than those of type-area Montgomery Lake Group (Fig. 2D) and consist of moderately sorted subarkose to quartz arenite, with subrounded, subspherical grains, typical of the Maguse Member. Furthermore, redbeds and quartz pebble conglomerates (with variable proportions of jasper clasts), unknown in the Montgomery Lake Group type area, yet common within the Maguse Member, are local components of map unit X.

Bell (1970a, 1971) suggested that an unconformity separates map unit X from overlying Hurwitz Group units, based on the stratigraphic thinning of map unit X that he inferred to be the result of erosional cut-off. In contrast, Young (1975) argued that the transition is conformable and that the stratigraphic thinning reflects onlapping of overlying units. Our mapping supports Young's (1975) interpretation of progressive infilling of an irregular paleotopography: 1) the boundaries between map unit X and the overlying Padlei Formation or Maguse Member are transitional (e.g. quartz arenite interbeds similar to map unit X occur within the Padlei Formation at the Noomut River measured section (Fig. 2 in Aspler et al., 1992a), local polymictic pebbly beds similar to those in map unit X occur near the base of the Padlei Formation southeast of Kinga Lake, e.g. UTM 234633; Maguse Member-like quartz arenites are within map unit X; Fig. 2C, D); 2) we have found no direct evidence of an angular discordance between map unit X and overlying units (to the contrary, in areas of good exposure such as southeast of Kinga Lake, abundant bedding data demonstrate concordance, Aspler et al., 1992b); and 3) we have found no evidence of features typical of unconformities such as are exposed at the unconformity between the Montgomery Lake Group proper and the Hurwitz Group in the Montgomery Lake Group type area (Aspler et al., 1992a).

Based on reconnaissance work, we previously suggested that quartz-rich sandstones that are cut by stockwork quartz veins west of the abandoned settlement of Padlei are probably Montgomery Lake Group (Aspler et al., 1992a). Furthermore, we suggested that although the contact between these rocks and the adjacent Padlei Formation is not exposed, the absence of stockworking in the Padlei Formation was consistent with the contact being an unconformity. We now favour an interpretation that this contact is not an unconformity but a fault (with the stockworking related to post- rather than pre-Padlei Formation deposition) and that the quartz-rich sandstones are not Montgomery Lake Group but part of map unit X. This conclusion is based on petrographic work and completion of mapping; in particular, the tracing out of these rocks to the west and south where deformation is less severe and the unit consists of quartz arenite with local polymictic pebbly sandstone beds typical of map unit X.

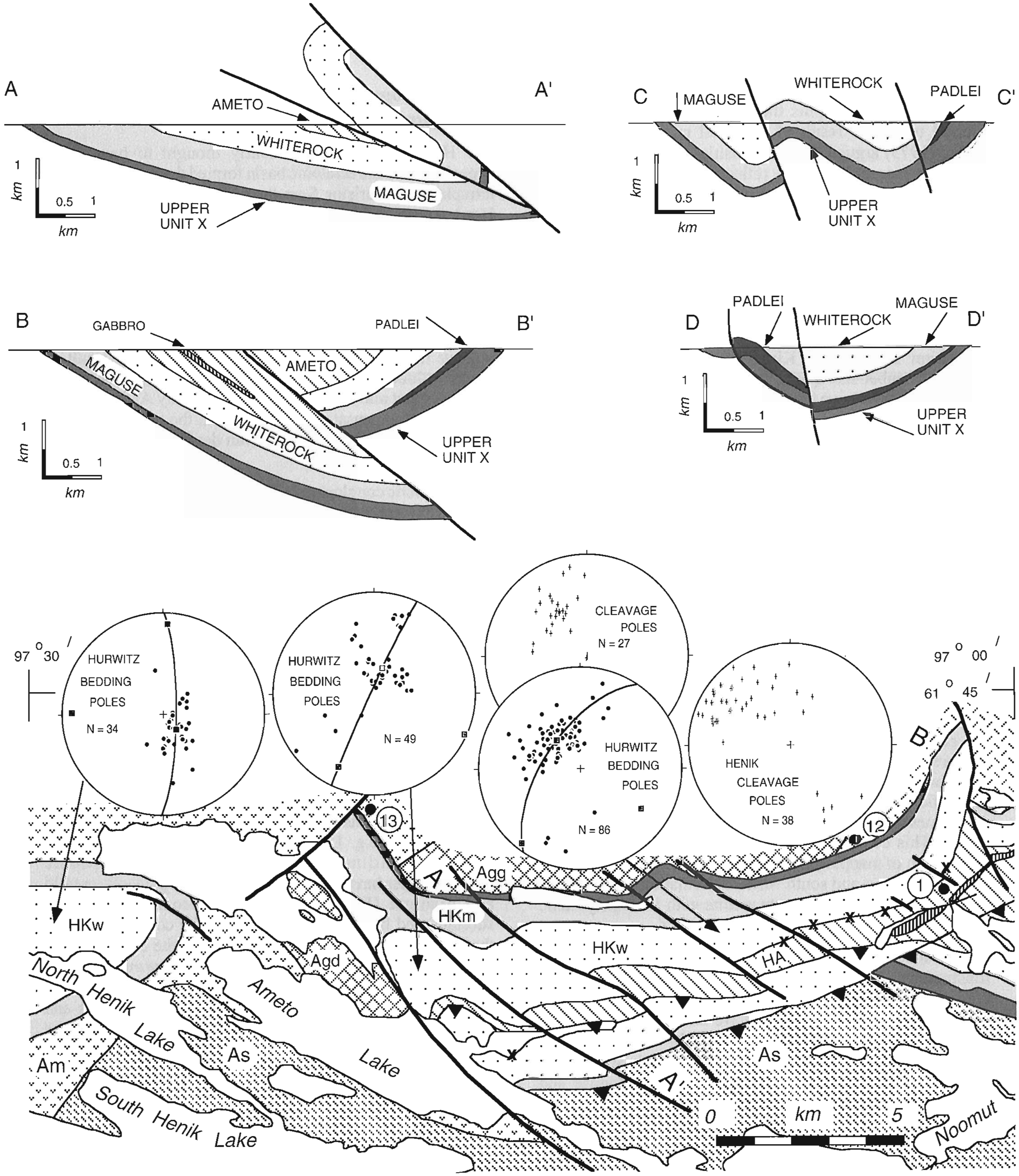
Because the physical map units of this study are functionally the same as those in Bell (1970a), we have retained the stratigraphic nomenclature as defined by Bell (1970a), with the exception that we have substituted "map unit X" for "Montgomery Lake sedimentary rocks". However, we anticipate revisions to the lower Hurwitz Group stratigraphy that will entail elevation of the Maguse and Whiterock members to formation status, with the elevated "Maguse Formation" represented by an upper member (to be named) a middle member (presently the Padlei Formation) and a lower member (map unit X; to be named). For simplicity, we have included local lenses of cobble-boulder conglomerate within the lower part of map unit X (HXl). Because similar isolated lenses are regionally developed, separation of these rocks into a distinct formation may be warranted with further study.

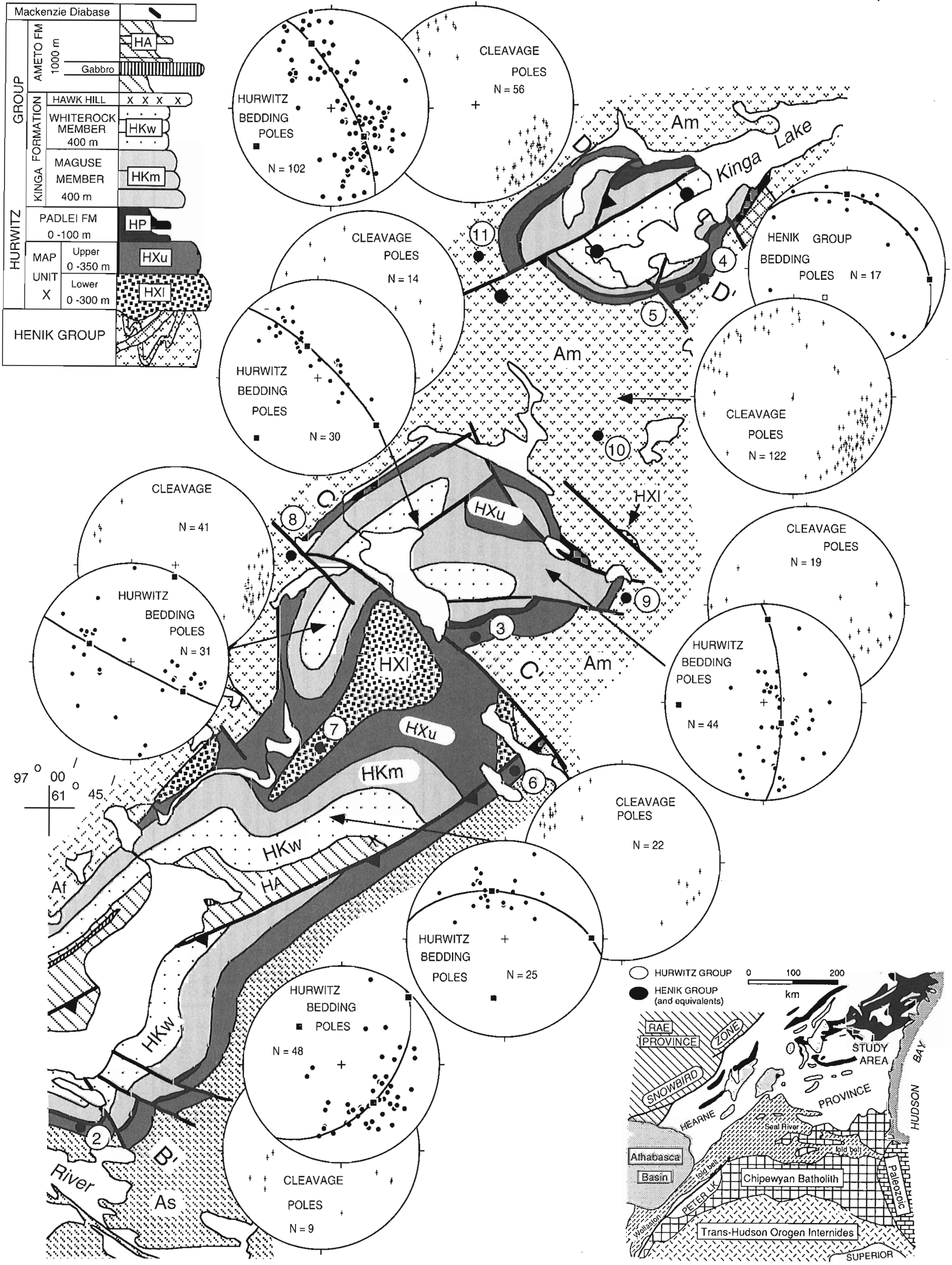
Sedimentology, sedimentation and tectonics of the lower Hurwitz Group

The Hurwitz Group is presently thought to have been deposited in an intracratonic basin formed due to compressive intraplate stress arising from the convergence of the Buffalo Head terrane with the western margin of the Churchill Province (Aspler et al.,1992a; see Hoffman, 1990; Ross et al., 1991). The lower Hurwitz Group stratigraphy is arranged in a conformable onlap sequence of radial basin expansion centered close to the Henik Lakes (Fig. 3; see also Fig. 4 in Bell, 1970b). Close coincidence between rocks of the Ennadai-Rankin greenstone belt and the Hurwitz Group may indicate that initiation of basin subsidence was the result of stress-induced viscosity decreases in the lower crust; decreases that allowed relatively high-density Archean rocks to sink (uncompensated excess mass in the upper crust; see de Rito et al., 1983; Howell and van der Pluijm, 1990; Shaw et al., 1991).

Polymictic conglomerates in the lower part of map unit X (unit HXl) are exposed as isolated lenses considered to represent paleovalleys cut into Archean basement and preserved by regional subsidence of a moderate relief paleotopography rather than by rifting. Overlying subarkoses and quartz arenites with polymictic pebbly layers, local quartz pebble beds (both pyrite-rich and pyrite-free), and rare polymictic conglomerate beds (unit HXu) are likely fluvial deposits. The contact between the lower and upper parts of map unit X is transitional, as indicated by rare HXl-like conglomerate beds within HXu (e.g. south of Kinga Lake, UTM 221623). Worthy of emphasis is that the upper part of map unit X is primarily varicoloured subarkose to quartz arenite similar to (and in part indistinguishable from) Maguse Member (Fig. 2C, D), that the polymictic pebbly beds and quartz pebble conglomerates interfinger (e.g. southeast of Kinga Lake) and that the red colours are in zones concordant with bedding (e.g. UTM 200549) as well as in zones of irregular mottling. The pyrite-bearing quartz-pebble conglomerates have long been the target for paleoplacer gold and uranium (Roscoe, 1981). A paleoplacer origin is substantiated by structures such as pyrite concentrations in the lee-side scours of pebbles (Fig. 4) and the fact that some of the pebbles are pyrite-bearing. This would signify pre-"oxyatmoversion" (Roscoe, 1973) deposition. Yet the data are contradictory: interfingering of red sandstones and pyrite-free, red-matrix, quartz pebble conglomerate with the pyrite-rich beds denotes oxic sedimentation. Typically, pyritic sandstones and redbeds are mutually exclusive (e.g. Roscoe, 1973, 1981). We tentatively suggest that the interfingering in the present example indicates sedimentation at the oxyatmoversion transition and levels of atmospheric oxygen that fluctuated, an interpretation that requires further

Figure 1. Simplified geological map and cross-sections, Padlei belt (see Aspler et al., 1992b for details). Map units as in Table 1. Stereonets are equal area, lower hemisphere projections. See Table 2 for analyses of numbered gossans.




Mackenzie Diabase
HURWITZ GROUP
AMETO FM
1000 m
HA
Gabbro
KINGA FORMATION
HAWK HILL
WHITEROCK MEMBER
400 m
HKw
MAGUSE MEMBER
400 m
HKm
PADLEI FM
0 -100 m
HP
MAP UNIT X
Upper
0 -350 m
HXu
Lower
0 -300 m
HXl
HENIK GROUP
HURWITZ BEDDING POLES
N = 102
CLEAVAGE POLES
N = 56
CLEAVAGE POLES
N = 14
HURWITZ BEDDING POLES
N = 30
HENIK GROUP BEDDING POLES
N = 17
CLEAVAGE POLES
N = 122
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HURWITZ BEDDING POLES
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1
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9
10
11
97° 00'
61° 45'
HURWITZ GROUP
HENIK GROUP (and equivalents)
0 100 200 km
RAE PROVINCE
SNOWBIRD ZONE
HEARNE PROVINCE
STUDY AREA
HUDSON BAY
Athabasca Basin
Seal River fold belt
Wollaston PETER LK fold belt
Chipewyan Batholith
Paleozoic
Trans-Hudson Orogen Internides
SUPERIOR

Table 1. Description and preliminary interpretation of map units, Padlei belt.

UNIT			DESCRIPTION	COMMENT/INTERPRETATION
Mackenzie diabase			Coarse-grained, massive, unmetamorphosed, northwest-trending	1267+/- 2 ma baddeleyite: LeCheminant & Heaman, 1989
HURWITZ GROUP		GABBRO HG	Discontinuous gabbro sills emplaced at different levels within Ameto Formation. Green weathering, coarse-grained, poikilitic. Locally well-foliated; chloritic shear zones common. Local fine-grained sills (with chilled margins) within coarse-grained sills.	Post Ducker Formation, pre-Hurwitz Group folding. Baddeleyite: 2.09 Ga (Patterson and Heaman, 1991); 2.11 Ga (Heaman and LeCheminant in press).
HURWITZ GROUP		AMETO FM HA	At base: red, blue, green slate with mm-cm-scale parallel-stratified arkose. Up-section: cm-dm-scale arkose to slate fining-upward cycles. At top: 0.1-2m cryptalgal laminates. Fine-grained, locally feldspar-phyric, 20 m thick, concordant mafic unit 80 m above base at Ameto type section.	Below wave-base sedimentation interrupted by turbidity currents; shoaling-upward to Watterson carbonate ramp. Mafic unit possibly volcanic (Happotiyik Member, Bell ,1970 a) or sill (Eade, 1974).
HURWITZ GROUP	KINGA FORMATION	Hawk Hill Mbr HKh	Discontinuous unit of bedded white chert, maroon chert and chert breccia conformable between the Whiterock Member and Ameto Formation. Discordant hematitic breccias at top of Whiterock Member	Previously considered silcretes; re-interpreted as sinters formed by surface discharge of hot springs (Aspler, in prep)
HURWITZ GROUP	KINGA FORMATION	Whiterock Mbr HKw	Supermature quartz arenite exclusively. Lower part of section commonly massive; upper 200 m with ubiquitous wave ripples. Quantitative analysis of symmetric vortex orbital ripples yields palewater depths averaging < 2 m for the unit's entire thickness and lateral extent (Aspler et al., in prep.) No evidence of tidal or evaporitic conditions.	Vast, hydrographically-open, tide-free, non-vegetated swamp-like lake (or series of lakes).
HURWITZ GROUP	KINGA FORMATION	Maguse Mbr HKm	Maroon, pink, grey subarkose to quartz arenite; interbeds of white quartz arenite at top. Local black parallel and cross-stratified heavy mineral bands; quartz grit and pebble layers; rare mudstone partings and mudcurls.	High-energy, unchanneled sheet floods draining low-relief paleotopography to form extensive sand plain (sheet-braid fluvial style of Cotter, 1978).
HURWITZ GROUP		PADLEI FM HP	Discontinuous lenses of dm-m-scale pink, grey subarkose to quartz arenite with local mudcurls and mudchips, rare ripples; interbedded with dm-m-scale mudstone with 10-50% cm-scale, subarkose to quartz arenite lenses. Local mm-cm-scale very fine sandstone-siltstone to mudstone rhythmites with micro cross stratification and graded bedding, coarse sand to pebble lonestones. Flute casts and tool marks on lonestone-bearing bedding surfaces. Rare polymictic granulestone and pebble conglomerate beds (framework-intact, coarse sand matrix).	Isolated small lakes on alluvial plain, with mixing of fluvial and lacustrine deposition. Lonestones likely from turbidity current deposition rather than from passive melt-out.
HURWITZ GROUP	MAP UNIT X	Upper HXu	Beige, grey, green and red arkose, subarkose and quartz arenite. Local polymictic pebbly sandstones (<5-30% floating clasts), rare beds of polymictic pebble-cobble conglomerate (framework-intact, coarse sand matrix); local thin (<0.2-5 m) pyrite-bearing quartz grit and pebble beds, and pyrite-free quartz grit and pebble beds.	Fluvial plain on low-relief paleotopography
HURWITZ GROUP	MAP UNIT X	Lower HXl	Lenses of massive and stratified cobble-boulder polymictic conglomerate; rare coarse sandstone interbeds. Framework-intact to disrupted; coarse sandstone (and locally mixtite) matrix. Clasts well-rounded: granitic; felsic and mafic volcanic; quartz; jasper; BIF, quartzite. Grades conformably upsection to HXu pebbly sandstones.	Both fluvial (clast-supported, sand matrix conglomerate; stratified sandstone) and mass flow (mixtite matrix-supported conglomerate) paleovalley-fill.
Ag		Granitoid	**Agg:** Well-foliated granodiorite; **Agd**: well-foliated diorite; both with xenoliths of Henik Group at margins	2512 +/- 112 Ma; Rb/Sr whole rock; Wanless and Eade 1975
HENIK GROUP	Af	Felsic volcanic	**Afr:** feldspar-phyric rhyolite. **Afra:** rhyolitic agglomerate with 30-80% ameboid-shaped clasts of rhyolite in coarse feldspar-quartz matrix	
HENIK GROUP	Am	Pillow volcanic gabbro sill/dyke complex	**Amv**: massive mafic volcanic; **Amvp:** with well-developed pillows; **Amvf:** mafic fragmental (tuff and agglomerate). Cut by **Amgb :** up to 200 m wide gabbro sills and dykes and **Amf:** 2 to 20 m wide fine-medium grained feldspar-hornblende quartz porphyry dykes (may be in part flows). **Ampc**: thinly-bedded pyrite chert.	Flows and mafic sills are probably co-magmatic.
HENIK GROUP	As	Mixed sedimentary/ volcanic	**Asi:** Dm-scale graded sandstone to mudstone; mm and cm-scale sandstone to magnetite iron formation rhythmites, locally graded. **Asft:** feldspar-quartz crystal tuff. **Asftp:** pebbly crystal tuff with 10-20% angular to subrounded felsic clasts in feldspar - quartz crystal matrix. **A1qfp :** medium to coarse - grained feldspar-quartz sills (flows?).	Below wave-base chemical sedimentation periodically interrupted by felsic pyroclastic and/or epiclastic sedimentation (mass flow deposits).

substantiation. Also contradictory is the association of compositionally mature quartz pebble conglomerates with polymictic conglomerates. Lack of clast diversity and enrichment of pyrite in the quartz pebble beds may be due, in part, to local derivation from pyrite-chert horizons in basement (unit Ampc) and need not axiomatically imply intensive weathering and/or long transport distance.

It has long been suggested that the Padlei Formation is glaciogenic (Bell, 1970a; Young, 1973; Young and McLennan, 1981), based in large part on the interpretation that sand and pebble-sized lonestones in local delicately laminated rhythmites are dropstones. Yet the presence of tool marks at the base of lonestone-bearing layers (Fig. 5) is more consistent with active current sedimentation rather than

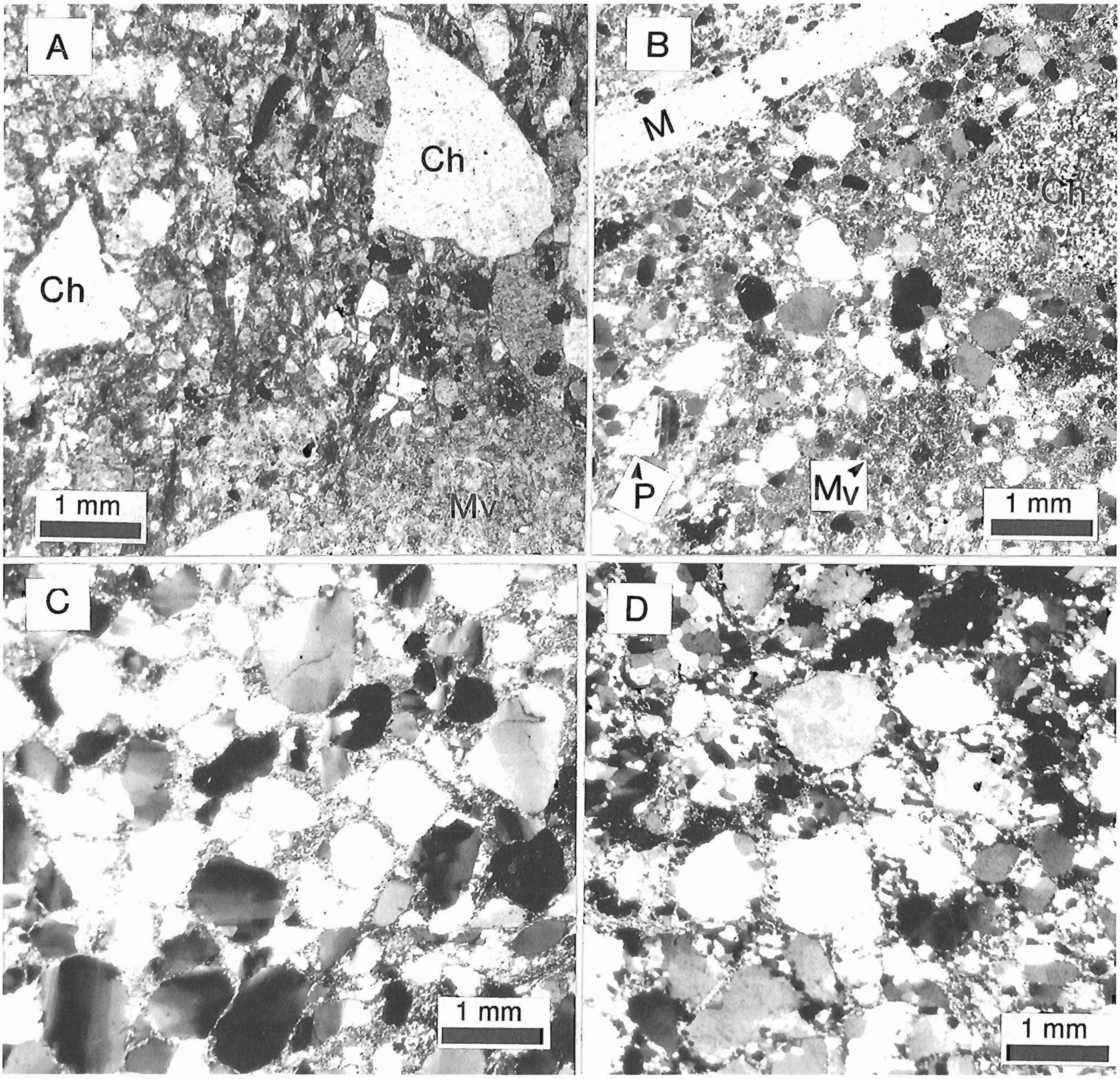

Figure 2. Petrographic comparison: Montgomery Lake Group from type area; Maguse Member and map unit X from Padlei belt. **A) and B)** Montgomery Lake Group type area: immature lithic wacke. Note poor sorting, roundness, and sphericity; Ch: chert; Mv: mafic volcanic; M: mudrock; P: plagioclase. **C)** Maguse Member: quartz arenite. Note moderate sorting; subrounded to well rounded, subspherical grains. **D)** Map unit X from Padlei belt: quartz arenite. Note moderate sorting; subrounded to well rounded, subspherical grains.

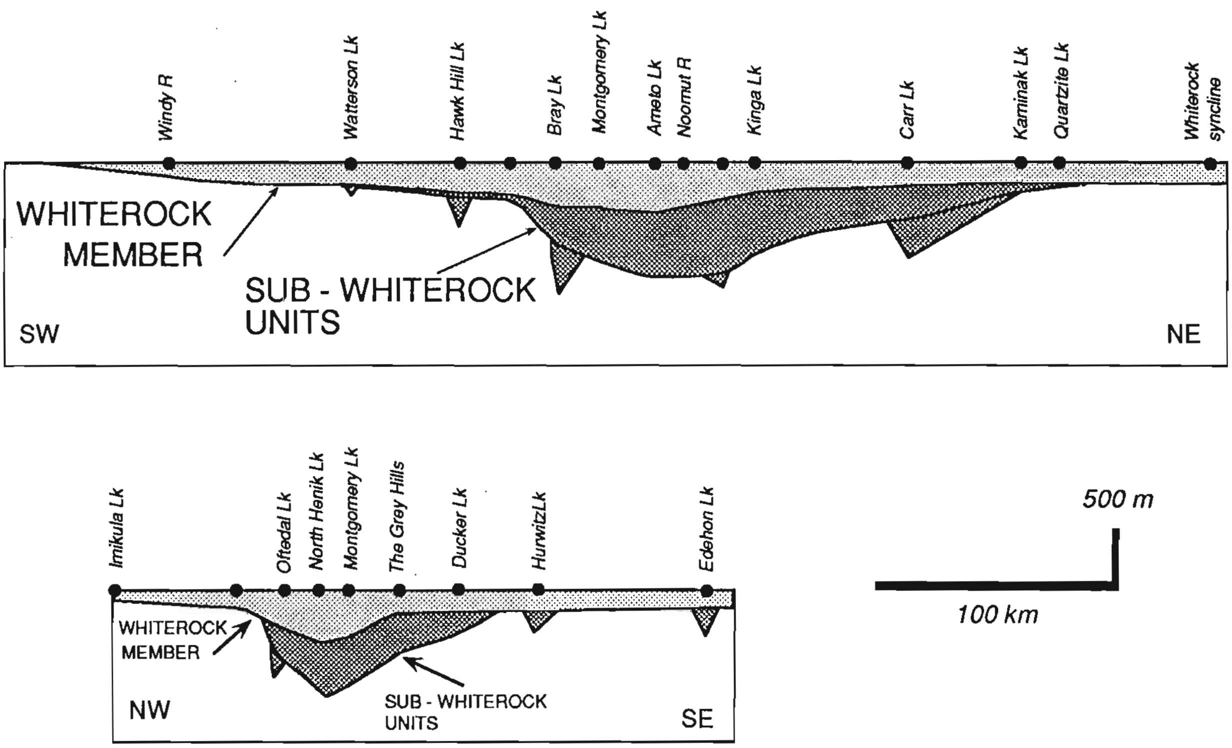

Figure 3. Stratigraphic cross-sections of the lower Hurwitz Group. Locations are plotted on a geographic base and are not palinspastically restored. Only six of the locations have measured sections (Noomut River and Montgomery, Bray, Ducker, North Henik, and Carnecksluck lakes); other sections are best estimates based on map data. Sources: Bell, 1968 (Carr, Kaminak, and Quartzite lakes and Whiterock syncline); Eade, 1973 (Hurwitz and Edehon lakes); Eade, 1974 (Bernier, Carnecksluck, and Ducker lakes); Eade, 1986 (Imikula Lake); Eade and Chandler, 1975 (Watterson Lake); Aspler et al., 1989 (Windy River); Aspler and Bursey, 1990 (Hawk Hill Lake); Aspler et al., 1992a (The Grey Hills, Oftedal, Bray, and Montgomery lakes), this study (North Henik Lake, Ameto Lake, Noomut River, Kinga Lake).

passive melt-out from icebergs. In the absence of unequivocal evidence that glaciers were present and active depositional agents (see criteria in e.g. Crowell, 1983; Mustard and Donaldson, 1987) we suggest that the Padlei Formation represents mixed fluvial/lacustrine sedimentation, with pelitic rocks deposited in small, isolated non-glacial lakes, and local conglomerates and thick sand interbeds deposited in marginal lacustrine/fluvial environments.

The maximum thickness of the Maguse Member (ca. 500 m) is near the center of Hurwitz Basin and the unit pinches out radially. During sedimentation of the Whiterock Member, Hurwitz Basin expanded to reach dimensions in excess of 500 by 200 km (Fig. 3). Similar to many other supermature quartz arenite sequences (e.g. Dott and Byers, 1981; Chandler, 1988), the Whiterock Member is thick (to 400 m) and lacks mudrocks. Yet it differs by virtue of containing a suite of sedimentary structures that consists almost entirely of short-wavelength ripple marks and parallel stratification, both formed by oscillatory flow. Application of empirical (Tanner, 1971) and analytical (e.g. Clifton and Dingler, 1984) methods to symmetric, vortex (height/wavelength <0.1), orbital (wavelength/grain diameter <400) ripples yields average water depths in the range 2 cm to 2 m (maximum 5.6 m) for the entire thickness and lateral extent of the unit. Evidence of tidal and evaporitic conditions is lacking. We envisage a lacustrine facies model that differs from the more familiar playa or deep water models described in the literature (e.g. Matter and Tucker, 1978). It entails sedimentation in a vast, hydrographically-open, tide-free, swamp-like setting in which lake levels were controlled by the long term water table, and in which continuously high rainfall in a wet equatorial climate and well-developed outlets combined to prevent raised salinities. Ultimately deposited in a wave-dominated setting, quartz arenites of the Whiterock Member owe their maturity to: 1) repeated reworking by fluvial, eolian and wave processes; 2) long-term exposure to tropical weathering (as originally suggested by Bell, 1970b; Young, 1973; see also e.g. Savage and Potter, 1991); and 3) high winds sweeping across a vegetation-free landscape and bypassing fine grained sediment out of the basin (see also Dalrymple et al., 1985). A suitable modern analogue for such a large perennial, non-saline wetland does not exist. The most comparable in geomorphological and tectonic terms, but certainly not in terms of climate and sedimentary infill, are the arid intracratonic depressions of Australia (Lake Eyre Basin; Jessup and Norris, 1971; Wopfner et al., 1974) and north-central Africa (Lake Chad Basin; Burke, 1976; Bridges, 1990).

One intriguing aspect about Lake Chad Basin emphasized by Burke (1976) is that, despite being situated 500 km from the Atlantic Ocean, a modest relative sea level rise would result in marine waters breaching the discontinuous ring of peripheral uplifts surrounding the basin, and cause a switch from lacustrine to marine sedimentation. This is similar to the history envisaged for the drowning of the Whiterock platform; the abrupt switch from shallow water lacustrine sedimentation of Whiterock Member quartz arenites to deep water mudrocks of the Ameto Formation is attributed to sudden basin deepening (induced by intraplate stress; see Cloetingh, 1991), and marine waters flooding through gaps in basement arches. With the source of quartz sand effectively cut off by drowning, the compositional maturity of arenites in overlying units decreased and carbonate rocks of the Watterson Formation were permitted to accumulate on ramps that prograded away from low relief arches.

STRUCTURAL GEOLOGY

The Proterozoic structural style of the Padlei belt is predominated by east-northeast and northeast-elongate, doubly-plunging basement-cover infolds that define a dome and basin interference pattern. Early, open, non-cleavage-bearing northwest-trending folds with steeply dipping axial surfaces are refolded by open, cleavage-bearing east-northeast and northeast-trending folds with steeply-dipping axial surfaces and east-northeast and northeast-trending thrusts (Fig. 1). Both sets of structures are cut by northwest-trending oblique-slip faults. Although Archean rocks were deformed with the Hurwitz Group, and cleavage trends in the Henik Group are broadly coincident with those of the Hurwitz Group, an earlier strain history is indicated by steep plunges of Henik Group folds (locally downward-facing) and by the presence of foliated Archean clasts in basal Hurwitz Group. The Hurwitz Group was deformed before ca. 1.75 Ga, the age of the Nueltin granite (U/Pb, Loveridge et al., 1988; cf. Rb/Sr, Wanless and Eade, 1975). How long before is uncertain; a preliminary metamorphic zircon determination of ca. 1.83 Ga (L. Heaman, pers. comm., 1992) suggests that deformation of the Hurwitz Group may be related to collisional events in Trans-Hudson

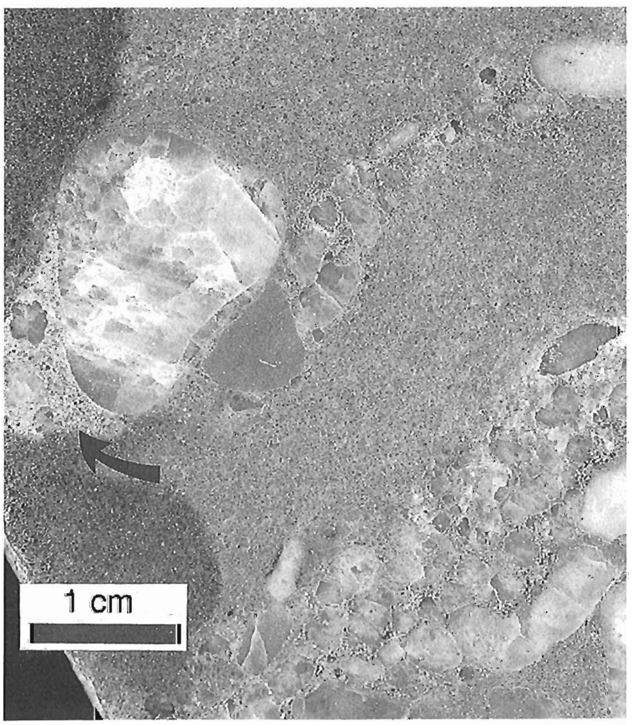

Figure 4. Pyrite concentration in lee-side scour, quartz pebble conglomerate lower map unit X. Drill core sample from S.M. Roscoe.

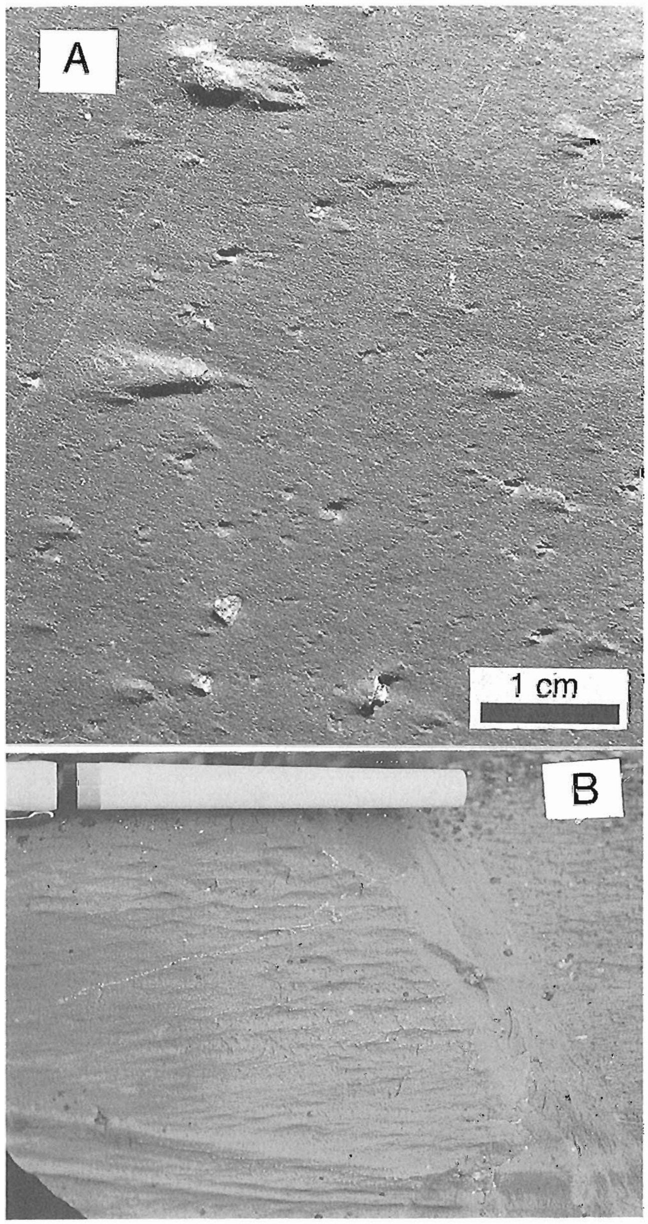

Figure 5. Sole marks preserved as moulds on the base of lonestone-bearing rhythmite bedding surfaces; Padlei Formation. **A)** Small-scale flute casts in lee of sand-sized lonestones (current from right to left). **B)** Drag marks.

Table 2. Geochemistry of grab samples. Gossan numbers are plotted on Figure 1.
All analyses by X-Ray Assay Laboratories, Don Mills, Ontario.

			ELEMENT	Au	Co	Ni	Cu	Zn	As	Mo	Ag	Cd	Sb	W	Pb
			METHOD	NA	DCP	DCP	DCP	DCP	NA	DCP	DCP	DCP	NA	NA	DCP
			DETECTION LIMIT	5 ppb	1 ppm	1 ppm	.5 ppm	.5 ppm	2 ppm	1 ppm	.5 ppm	1 ppm	.5 ppm	2 ppm	2 ppm
GOSSAN	SAMPLE	UTM	LITHOLOGY / UNIT												
1	91-54-20 C	65H/11; 038426	Quartz vein in Hurwitz gabbro	51	26	26	1410	43.7	<2	3	0.9	<1	<0.5	<2	<2
2	MS3-1	65H/10; 065387	py in HXu qtz pebble conglomerate	100	3	8	39.5	5.8	8	10	<0.5	<1	0.8	<2	<2
	MS3-2	65H/10; 066388	py in HXu qtz pebble conglomerate	39	<1	3	12.9	3.3	6	4	<0.5	<1	0.8	<2	<2
3	92-13-9	65H /15; 170523	py in HXu qtz pebble conglomerate	220	80	47	28.5	6.7	28	3	<0.5	<1	2.3	<2	<2
4	92-1-5	65H /15; 230626	py in HXu qtz pebble conglomerate	36	29	26	19.2	6.6	12	<1	<0.5	<1	1.3	2	<2
5	92-4-10 A	65H/15; 227623	py layers in HXu sandstone	18	50	35	8.9	6.9	13	<1	<0.5	<1	3.4	3	<2
	92-4-10 B		py in HXu qtz pebble conglomerate	230	42	27	8.8	5.4	17	1	<0,5	<1	1.8	2	7
	92-4-10 C		py in HXu granulestone	38	33	26	14.3	7.3	11	2	<0.5	<1	1.2	<2	3
6	92-18-25	65H/15/; 184490	sulphides in carbonate vein	<5	33	19	499	19.5	3	<1	<0.5	<1	<0.5	<2	<2
7	92-11-31	65H/15; 130491	py in HXl polymictic conglomerate	8700	130	45	447	8.7	210	8	3.7	<1	14	<2	21
8	92-19-19 A	65H/15; 123551	Ampc banded pyrite-chert	790	67	109	51.1	11.4	260	2	<0.5	<1	5.9	<2	<2
	92-19-19 B			130	44	133	41.4	6.1	150	3	<0.5	<1	2.3	<2	<2
	92-19-19 C			<5	5	8	38	14.6	4	<1	<0.5	<1	<0.5	<2	<2
9	92-14-18	65H/15; 215538	Ampc banded pyrite-chert	27	48	43	34.6	6.7	24	<1	<0.5	<1	0.7	<2	<2
10	92-3-24	65H/15; 204591	py disseminated in Am mafic rock	8	13	28	8.4	61.9	3	2	<0.5	<1	<0.5	<2	<2
	92-3-25			12	14	41	27.2	31.1	16	2	0.6	<1	0.5	<2	<2
11	92-5-2	65H/15; 175635	py clots in Am mafic volcanic	46	63	71	783	75.1	2	2	0.8	<1	0.7	3	<2
12	91-60-25	65H/11; 016434	cpy-py quartz vein; Af rhyolite	1600	139	499	1280	3.2	3	<1	2.3	<1	<0.5	<2	<2
13	92-28-20	65H/11; 887435	sulphides in qtz vein; Am dacite	180	15	18	933	4	3	<1	0.5	<1	<0.5	<2	<2

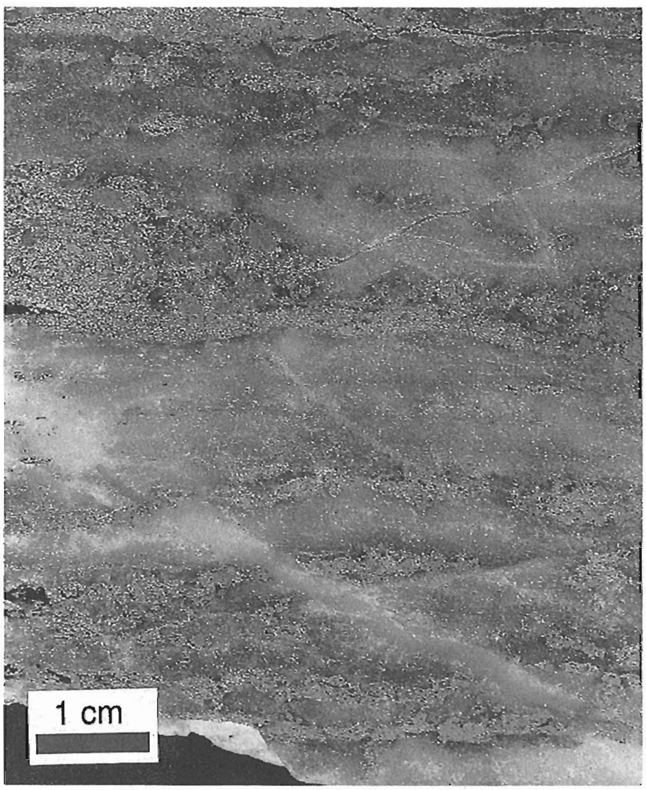

Figure 6. Banded pyrite and chert, unit Ampc; gossan 8.

Orogen, signifying a large time gap between sedimentation (pre-ca. 2.1 Ga; Heaman and LeCheminant, in press; Patterson and Heaman, 1991) and deformation.

ECONOMIC GEOLOGY

Mineral occurrences are plotted on Figure 1; UTM co-ordinates and available analyses are given in Table 2. Gossans in the Henik Group consist primarily of massive and disseminated pyrite within mafic volcanic rocks, although the maximum gold value obtained is from a quartz vein cutting rhyolites of unit Af (1600 ppb Au, gossan 12). Of particular interest are pyrite-chert horizons that consist of millimetre-to-decimetre-scale banded, nodular, and colloform py rite and chert (Fig. 6; unit Ampc; gossans 8, 9). Similar horizons are probable source rocks to auriferous polymictic conglomerates mapped east of Bray Lake (to 1600 ppb Au; gossan 21 in Aspler et al., 1992a), and likely influenced the development of pyritic quartz-pebble paleoplacers in the lower Hurwitz Group (see above). The maximum gold value from this study is from polymictic cobble-boulder conglomerates near the top of unit HXl: 8700 ppb Au (gossan 7). Pyritic quartz pebble conglomerate, granulestone, and sandstone of unit HXu have elevated gold (to 230 ppb Au, Table 2).

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Redbed copper occurrences in the Lower Proterozoic Baker Lake Group, Dubawnt Supergroup, District of Keewatin, Northwest Territories[1]

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Miller, A.R., 1993: Redbed copper occurrences in the Lower Proterozoic Baker Lake Group, Dubawnt Supergroup, District of Keewatin, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 159-170.

Abstract: In the central Churchill Province, the northeasterly-trending Angikuni subbasin is composed of continental fluvial conglomeratic to arkosic strata, alkaline mafic volcanic rocks and derived volcaniclastic rocks of the Baker Lake Group. The basement immediately adjacent to this subbasin along its northwestern margin consists of greenschist Archean volcanic rocks and intrusive monzogranite plutons. A paleosol, up to 2 m thick and locally preserved beneath the basal Baker Lake Group, records an interval of tectonic stability in the Trans-Hudson Orogen hinterland prior to deposition of the Baker Lake Group. The lower nonvolcanic-bearing equivalents of the South Channel and Kazan formations, fine basinward to the southeast and are overlain by Christopher Island Formation mafic volcanic flows, lapilli tuff, and tuffaceous arkose. Stratiform chalcocite and digenite are disseminated in carbonate- and chlorite-altered arkose and volcaniclastic sandstone. Copper is confined to the disconformity between lower sedimentary South Channel-Kazan strata and the volcanic and volcaniclastic Christopher Island Formation.

Résumé : Dans la partie centrale de la Province de Churchill, le sous-bassin d'Angikuni, de direction générale nord-est, se compose de strates fluviatiles continentales de caractère conglomératique à arkosique, des roches volcaniques mafiques de nature alcaline, et des roches volcanoclastiques dérivées appartenant au Groupe de Baker Lake. Le socle jouxtant ce sous-bassin sur sa marge nord-ouest se compose de roches volcaniques et de plutons monzogranitiques intrusifs d'âge archéen, métamorphisés dans le faciès des schistes verts. Un paléosol pouvant atteindre 2 m d'épaisseur, préservé par endroits sous la partie basale du Groupe de Baker Lake, témoigne d'un intervalle de stabilité tectonique dans l'arrière-pays de l'Orogène trans-hudsonien avant la sédimentation du Groupe de Baker Lake. Les équivalents inférieurs, dépourvus d'éléments volcaniques, des formations de South Channel et de Kazan, s'affinent en direction du bassin, au sud-est, et sont recouverts par les coulées volcaniques mafiques, les conglomérats volcaniques à lapilli et l'arkose tufacée de la Formation de Christopher Island. De la chalcocite et de la digénite stratiformes sont disséminées dans une arkose et un grès volcanoclastique carbonatisés et chloritisés. Le cuivre est confiné à la disconformité entre les strates sédimentaires inférieures de South Channel-Kazan et la Formation de Christopher Island, de caractère volcanique et volcanoclastique.

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INTRODUCTION

Between 1975 and 1982, the Dubawnt Supergroup was considered a prime uranium exploration target based on the comparison with discovered uranium deposits hosted in both lithostratigraphic sedimentary units of the northern Saskatchewan uranium province: Athabasca Supergroup (Ramaekers, 1981) and the Martin Group (Langford, 1981; Tremblay, 1972). In 1981 the author, at the invitation of companies that were exploring for uranium in the Angikuni to Yathkyed lakes area (Noranda Exploration Co. Ltd., 1981; Noranda Exploration Co. Ltd. and Agip Canada Ltd., 1979), toured numerous uranium occurrences hosted in, and located near, the margin of the northeastern Angikuni subbasin and adjacent Archean basement. Examination of sediment-hosted copper-rich showings without uranium suggested a new mineral deposit type not recognized previously in the Baker Lake Group, distinct from the spectrum of uranium occurrences. This observation in 1981 was the impetus for the present project funded under the Geoscience Initiative, 1991-1996 Northwest Territories Mineral Development Agreement.

REGIONAL GEOLOGY

Eade (1986) and Blake (1980) divided the Angikuni subbasin into sediment- and volcanic-dominated lithological packages belonging to the Christopher Island Formation of the Dubawnt Group in the Tulemalu Lake map area. This project is located along the northwestern erosional margin of the Angikuni subbasin and is positioned within the sediment-dominated portion of the basin (Fig. 1).

The stratigraphic nomenclature of the unmetamorphosed sedimentary and volcanic rocks that comprise the ~1.84 to 1.72 Ga basins in central Keewatin and MacKenzie have been revised since the original stratigraphic subdivision of Donaldson (1965). LeCheminant et al., (1979) defined the Kunwak Formation and Gall et al., (1992) proposed Dubawnt Supergroup to replace Dubawnt Group. This report will use the proposed stratigraphic nomenclature of Gall et al., (1992).

In the Baker Lake area, the presence of sedimentary rocks consisting of a conglomeratic to coarse arkosic sandstone unit (South Channel Formation), overlain by arkosic sandstone-siltstone-mudstone (Kazan Formation) was utilized as a critical feature to separate these formations from overlying mafic alkaline flows and derived sedimentary rocks of the

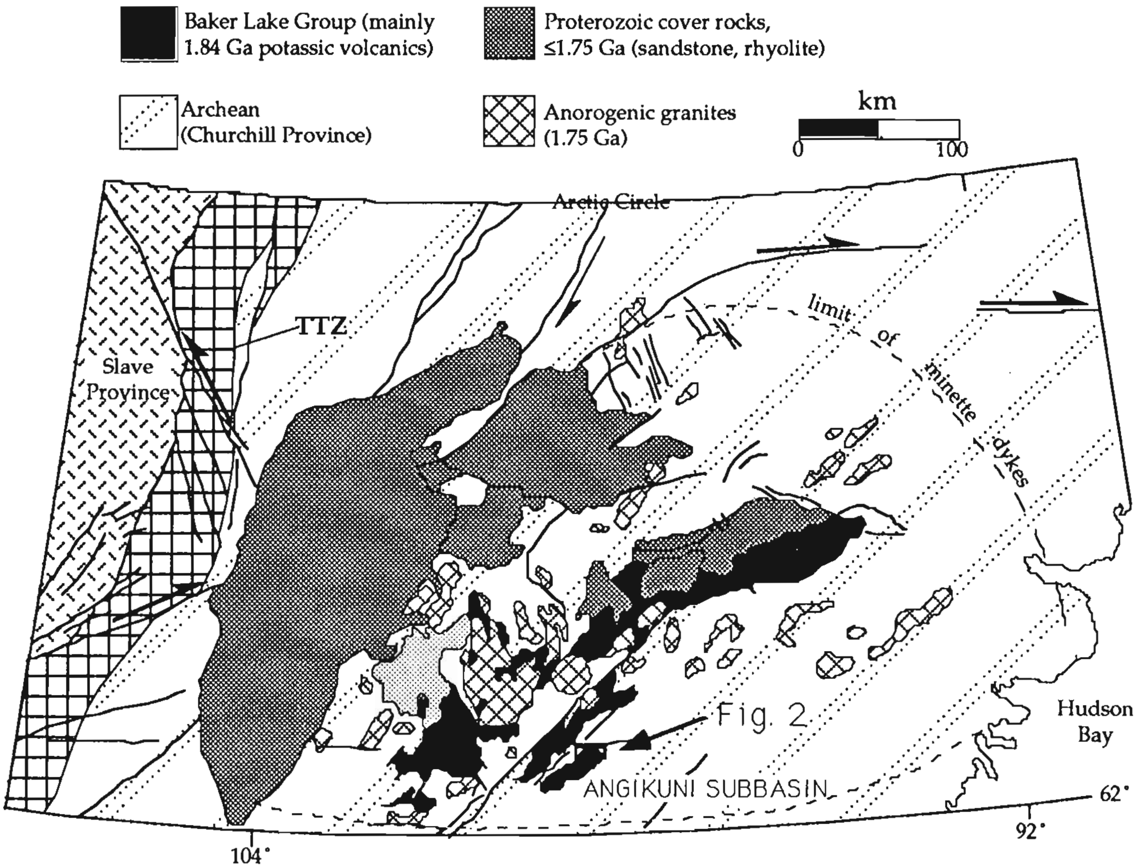

Figure 1. Location of the project area in relation to the 1.84-1.72 Ga basins of the Dubawnt Supergroup, District of Keewatin.

Christopher Island Formation. Donaldson (1965, 1967, 1969) defined the South Channel and Kazan formations as a fining-upward sequence of fluvial conglomerate through to arkosic sandstone-siltstone-mudstone.

This paper reports the results of mapping in a portion of NTS 65J/10, redefines the stratigraphy, provides preliminary descriptions of uranium and copper occurrences, and places the copper occurrences into a stratigraphic context within the Baker Lake Group, Dubawnt Supergroup.

STRATIGRAPHY AND SEDIMENTOLOGY OF A PORTION OF NTS 65J/10

Archean rocks

Archean rocks of the Henik Group (Eade, 1986) form the basement to, and are exposed in, a small inlier within the Baker Lake Group (Fig. 2). Exposure of the Henik Group is poor along the northwestern boundary of the subbasin but improves towards the northern edge of the mapped area. The basement is dominated by foliated fine grained amphibolite derived from massive and pillowed basalt flows. Mineral assemblages in foliated metabasaltic rocks include hornblende, epidote, chlorite, carbonate, and quartz indicating upper greenschist-lower amphibolite regional metamorphism. A narrow felsic metavolcanic unit is interstratified with the metabasalts. Schistose felsic volcanic rocks have been metamorphosed, sheared, and consist of variable proportions of chlorite, muscovite, and quartz. The northeasterly-trending fabric present in both metavolcanic units is subparallel to parallel to a regional fault which extends from Angikuni Lake to near Yathkyed Lake (Eade, 1986).

Intrusive into the Henik Group and deformed along with the enclosing metavolcanic rocks is a weakly foliated medium grained leucocratic muscovite-biotite monzogranite. Microscopically, quartz in this intrusion has been recrystallized into subgrains. Macroscopically, these deformed quartz grains display a rodded fabric.

Proterozoic rocks

Stratigraphy and sedimentology of the Baker Lake Group

According to Eade (1986), the northeastern end of the Angikuni subbasin is dominated by volcaniclastic sedimentary rocks. The presence of volcanic detritus within these sedimentary rocks implied by definition that they be assigned to the Christopher Island Formation. Mapping of the sedimentary strata in this area in 1992 indicated that a much wider array of sedimentary rocks is present in the northeastern Angikuni subbasin. All formations in the project area except the 1.27 Ga northwesterly-trending MacKenzie diabase dyke belong to the ~1.84 Ga Baker Lake Group, Dubawnt Supergroup. Each formation, discussed in ascending stratigraphic order, is described according to its critical sedimentological features.

South Channel Formation

The South Channel Formation has been divided into four lithostratigraphic units (Fig. 2). All four units were not observed at any one locality. This may be in part due to exposure but also to erosion during deposition of high-energy conglomeratic units. The lower three subunits, units 2a through 2c, differ markedly from unit 2d. This division is based on sediment type, sedimentary structures, and interpreted depositional environment. The former imoyd include a hematitic paleosol, locally-derived conglomerate and breccias, and low-energy distal fine grained sedimentary rocks. These three rock units are overlain by high-energy boulder to pebble conglomerate.

A paleosol, up to 2 m thick and locally preserved beneath the basal Baker Lake Group, records an interval of tectonic stability in the Trans-Hudson Orogen hinterland prior to Baker Lake Group sedimentation. Where observed in outcrop, the paleosol is developed from an Archean metabasalt protolith. This paleosol is moderately fractured and hematitized; oxidation decreases away from fracture walls (Fig. 3A). Colour variations in the paleosol range from hematitic red within and adjacent to fractures through to light pink, pinkish grey and grey-green in areas between fractures. This colour variation represents a transition in the paleosol from oxidized to weakly oxidized. These fractures parallel the northeasterly-trending structural grain in the basement and may have been present prior to initial sedimentation of the Baker Lake Group. Fracturing would have increased permeability and promoted oxidation deeper into the protolith. This texture is reminiscent of modern weathering profiles where onion-skin textured corestones are preserved in weathered rock.

In 1981, approximately 12 km to the east of the project area, a truncated but thicker paleoweathered profile was examined. This profile, also derived from Archean metabasalt, contains three subunits. The uppermost red unit, in erosional contact with conglomerate, is highly oxidized and hematite-rich. Oxidation decreases downward into a transition zone which bears a mottled red-green colour and abundant hematite-stained fractures. Oxidation decreases away from these fracture walls. This subunit in turn grades into the green to blackish-green metabasalt protolith. By comparison, in this project area the paleoweathered outcrop is truncated at a lower level than the one to the east and this erosional level corresponds to the transition zone between highly oxidized rock and the protolith.

The paleosol is overlain by a locally-derived matrix- to clast-supported sharpstone conglomerate-breccia (Fig. 3B, 3C). One variant of this unit directly overlies the paleosol (Fig. 3B) and consists of hematitized basement gneiss clasts set in a matrix of unstratified deep reddish maroon gritty siltstone. The best perserved and most widely exposed variant overlies the deformed quartz monzonite pluton in the northeastern portion of the map area (Fig. 3C). Clast compositions are generally monolithic and reflect the underlying lithology. The unit overlying the monzogranite consists of angular cobbles, up to 25 cm, of monzogranite with subordinate granitic gneiss and quartz vein pebbles. The

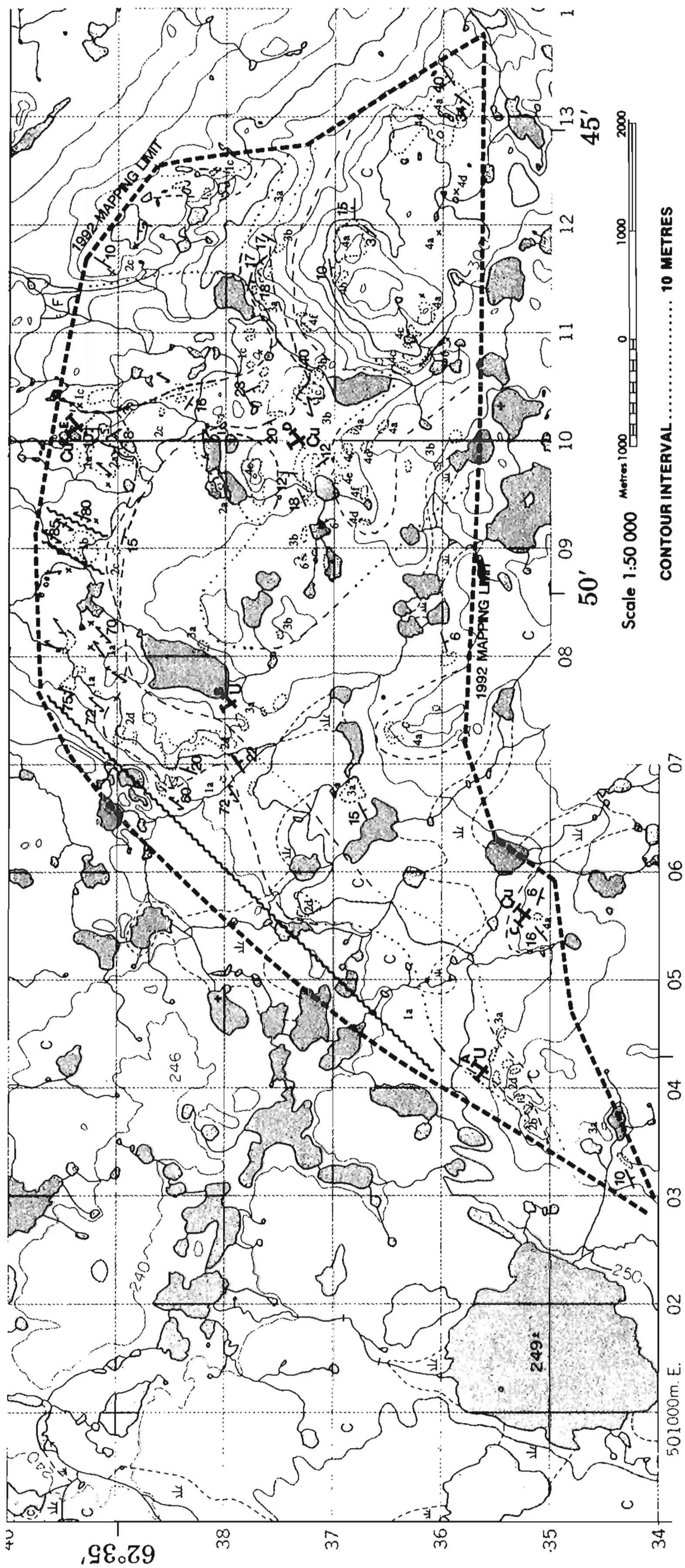

Figure 2. Geology of a portion of 65J/10.

matrix is a red to maroon, gritty siltstone which lacks stratification. The lack of maturity shown by clast shape, monolithic clast compositions, and the unstratified nature of the matrix and clast- to matrix-supported texture suggest that unit 2b may be debris flows. These may only be preserved in paleotopographic lows.

This unit is overlain by a red to maroon, fine grained arkosic sandstone-siltstone-mudstone unit (unit 2c). This triplet exhibits thin parallel bedforms, on the scale of 10 cm and less, fine- scale cross-stratification, asymmetric interference ripples and abundant mudcracks (Fig. 3D, 3E). These sedimentary features suggest a low-energy depositional environment. The interpreted depositional setting was on a topographically subdued landscape crossed by fluvial systems and possibly ephemeral lacustrine-deltaic environments. The above three lithological subunits are pre-alkaline volcanism and are placed within the South Channel Formation. These units are not present in the type area of the Baker Lake Group, south and southeast of the Baker Lake. However, locally preserved equivalents have been recognized in the Kamikuluak Lake area (Tella et al., 1981) and in the Dubawnt Lake area (Peterson et al., 1989; Rainbird and Peterson, 1990). These subunits have preserved a record of paleoweathering and sedimentation that predate regional faulting and deposition of coarse conglomeratic sedimentary rocks of the Baker Lake Group.

HADRYNIAN

6 MacKenzie Diabase

APHEBIAN

Dubawnt Supergroup: Baker Lake Group

5 Kunwak Formation
- 5: Cobble to pebble conglomerate with interbedded granulestone, minor gritty siltstone

4 Christopher Island Formation
- 4a: Volcaniclastic arkosic sandstone; siltstone
- 4b: Tuffaceous arkose; 4c: lapilli tuff; 4d: felsic flow; 4e: mafic lahar breccia; 4f: mafic flow

3 Kazan Formation
- 3b: Orange to red interbedded fine grained arkosic sandstone, siltstone, mudstone
- 3a: Pink to orange fine-to medium-grained arkosic sandstone, granulestone

2 South Channel Formation
- 2d: Pebble to boulder conglomerate, granulestone
- 2c: Red to maroon fine-grained arkosic sandstone, siltstone, mudstone
- 2b: Sharpstone conglomerate, conglomerate breccia
- 2a: Paleosol (hematitized metabasalt)

ARCHEAN

1
- 1c: Foliated monzogranite
- 1b: Chlorite-muscovite schists (metarhyolite)
- 1c: Foliated fine-grained amphibolite, pillowed metabasalt, massive metabasalt

85 Foliation

15 Bedding, tops known

Fault, assumed

Contact: defined, approximate, assumed

Outcrop, frost-heaved outcrop

Mineral occurrence: Cu, U

Legend for Figure 2

The boulder to pebble conglomerate with interbedded granulestone and gritstone, unit 2d, is preserved as a discontinuous ridge-forming unit along the northwestern edge of the map area. The conglomerate is characterized by rounded, equant to elongate clasts of tonalitic, granodioritic to granitic gneisses, vein quartz, and subordinate quantities of metavolcanic and metadiabasic clasts. Clast-supported frameworks, common near the base, form massive conglomerate with only local pebble imbrication (Fig. 3F) and interstratified pebbly granulestone and gritstone. Increased arkosic sandstone and siltstone contents have produced matrix-supported conglomerate higher in the section. Massive to stratified coarse fluvial conglomerates grade upsection into arkosic units of the Kazan Formation, similiar to relations seen in the type area, south of Baker Lake (Macey, 1973; Miller, 1980).

Kazan Formation

Granulestone, coarse- to fine-grained arkosic sandstone, siltstone and mudstone, units 3a and 3b, are interpreted as fluvial deposits that fine upsection and basinward, to the southeast. The lower unit, 3a, is gradational from the underlying South Channel conglomerate and is characterized by pink to orange granulestone, pebbly medium- to fine-grained arkosic sandstone (Fig. 4A). Subunit 3a is gradational into 3b and is based on decreasing grain size and an increasing proportion medium- to fine-bedded triplets of arkosic sandstone-siltstone-mudstone (Fig. 4B). Fine-scale cross-stratification and asymmetric ripple marks within sandstone units, and mudcracks in siltstones and mudstones support a low-energy distal fluvial environment with periodic subaerial exposure.

Christopher Island Formation

Eade (1986) grouped all rocks in the Angikuni subbasin into the Christopher Island Formation and further divided the subbasin into a northeastern volcaniclastic sediment-dominated and central and southwestern volcanic-dominated segment. Even though some of the clastic sedimentary units have been reassigned to lower stratigraphic formations, those rocks that are assigned to the Christopher Island Formation in the map area support the observation of Eade (1986) that the northern segment of the Angikuni subbasin contains a higher proportion of volcaniclastic sedimentary rocks than volcanic flows.

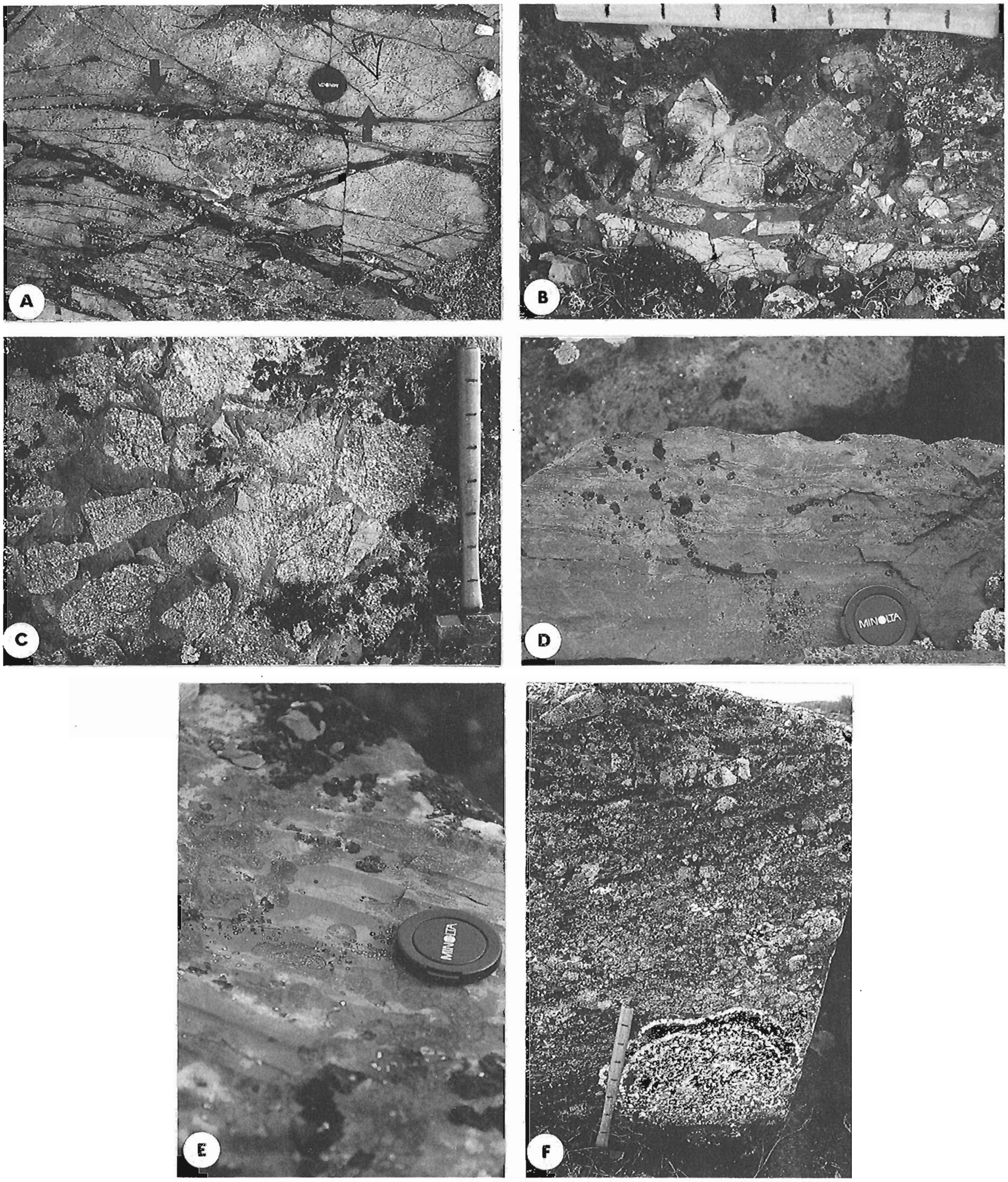

Figure 3.

A. Paleoweathered metabasalt, unit 2a. Subtle colour variations (solid arrow- strongly oxidized; open arrow-weakly oxidized) record decreasing hematitization away from fractures. (GSC 1992-261C)

B. Sharpstone conglomerate, unit 2b, directly overlying the paleosol shown in Figure 3A. Subdivisions on the hammer handle are 5cm. (GSC 1992-261H)

C. Basal angular breccia,unit 2b; interpreted as locally derived talus or debris flow deposits. (GSC 1992-261J)

D. Interbeds of fine- to medium-grained arkosic sandstone and mudstone with dessication cracks, unit 2c. (GSC 1992-261E)

E. Assymmetric current ripples, unit 2c. (GSC 1992-261D)

F. Bedded, weakly imbricated clast-supported cobble conglomerate, unit 2d. (GSC 1992-261K)

This formation records the initiation of alkaline volcanism within the Angikuni subbasin and the interplay between formation of subaerial volcanic landforms and their erosional products and mass wasting. The topopgraphic expression of Christopher Island Formation is distinctive. Rocks grouped into this formation form steep-sided isolated hills and broad flat-topped mesa-like landforms with gently sloping shoulders that punctuate relatively flat, generally lower-lying swampy areas underlain by Kazan and South Channel sedimentary units. The contact between the Christopher Island and Kazan formations is thought to be conformable and is based on the first appearance of volcanic clasts or flows. These lowermost Christopher Island sedimentary and volcanic rocks dip to the south and southeast on a paleoslope similiar to the underlying fluvial non-volcanic-bearing sedimentary units. A reverse in paleoslope is indicated by west- to northwest-dipping crossbedded volcaniclastic sediments that are located near the top of the broad upland in the southeast corner of the map area.

Figure 4.

A. Cross-stratified, medium grained arkosic sandstone and pebbly granulestone, unit 3a. (GSC 1992-261O)

B. Parallel-bedded, fine grained arkosic sandstone and siltstone with minor coarse grained sandstone, unit 3b. (GSC 1992-261N)

The base of the Christopher Island Formation is marked by fine grained felsic minette flows that are interstratified with purple to maroon pyroxene and phlogopite-phyric unsorted debris flows, possible laharic deposits (Fig. 5A), and fine grained, poorly-bedded volcaniclastic sediments. Volcaniclastic, possibly tuffaceous, sandstone and interbedded siltsone vary from parallel-bedded massive to well stratified with crossbedding and ripple marks (Fig. 5B). The lithic component is dominated by fine grained, trachytic feldspar-rich minette, but glomeroporphyritic plagioclase-phyric minette, phlogophite mafic minette and rare basement gneiss are present. The reduced basement input into these volcaniclastic sedimentary units may have been due to one or multiple factors: decreased uplift rates, distance from the basin margin, and influx of voluminous alkaline volcanic material into the sedimentary cycle. The immaturity of the volcaniclastic sedimentary rocks is demonstrated by the angular to subangular lithic clasts and framework grains. However, rounded basement granitoid clasts within the same units exhibit a higher degree of maturity and most probably had a long transport history. Preliminary petrographic examinations indicate that the alkaline volcanic and derived sedimentary strata are pervasively propylitized. Hematite, carbonate, and chlorite are present both as cements and replacements of lithic fragments. This alteration assemblage has been observed in other areas blanketed by the Christopher Island Formation (Blake, 1980; Miller, 1980; LeCheminant et al., 1979; Tella et al., 1981). This pervasive alteration may have been critical to the deposition of uranium and copper near or at the base of the Christopher Island Formation.

Kunwak Formation

This lithological unit, unit 5, is recognized at one locality, on the northern edge of the basin and is host to a low-grade copper occurrence (occurrence E, Fig. 1). This unit unconformably overlies units 1a, 2a, b and c and is presumed to overlap the monzogranite, unit 1c. Sedimentary structures along with a distinctive clast lithology distinguish this unit from the thick South Channel conglomerate (unit 2d) exposed along the northwestern margin of the subbasin. This gently south-dipping conglomerate-sandstone-minor siltstone unit fines markedly southward, upsection. This promiment fining over a narrow stratigraphic interval is distinct compared to the thick basal South Channel conglomerate. Crossbedding, lenticular beds, 5-100 cm, of interstratified conglomerate, arkosic granulestone, medium grained sandstone and siltstone suggest rapidly changing hydraulic gradients, possibly due to periodic flooding of braided river systems (Fig. 5C). Alternatively, the interstratification of coarse- and fine-grained sedimentary rocks suggest that this unit may straddle the South Channel-Kazan gradational contact. However, a unique type of clast present in this unit does not support the latter stratigraphic interpretation.

Clast lithologies within conglomeratic beds of unit 5 are highly variable and are similiar to unit 2d, the South Channel conglomerate. Clasts include granodioritic to granitic gneisses, monzogranite of unit 1c, minor metavolcanics of unit 1a, quartz vein, and a distinctive fine grained red felsic clast. The last is noteworthy because of its lack of sphericity compared to all of the other clast lithologies. These clasts are aphanitic to fine grained, red to orangy-red, angular to subrounded, appear to be rhyolitic in composition, and were not observed in the South Channel conglomerate, unit 2d. An uneven colour pattern is characteristic of this clast (Fig. 5D). Preliminary petrographic examination indicates that these clasts are composed entirely of fine grained, unoriented and unstrained feldspar. Black cores contain very fine grained hematite whereas the rims lack such hematite. Hematite removal may have occurred during diagenesis of this conglomeratic sheet. The feldspar crystals in these clasts are coarser than in the trachytic feldspar-rich clasts that are present in the Christopher Island volcaniclastic rocks. Alternatively, the feldspars are similiar to those in high-level highly fractionated felsic intrusions, possibly intrusions of bostonitic composition (LeCheminant et al., 1987).

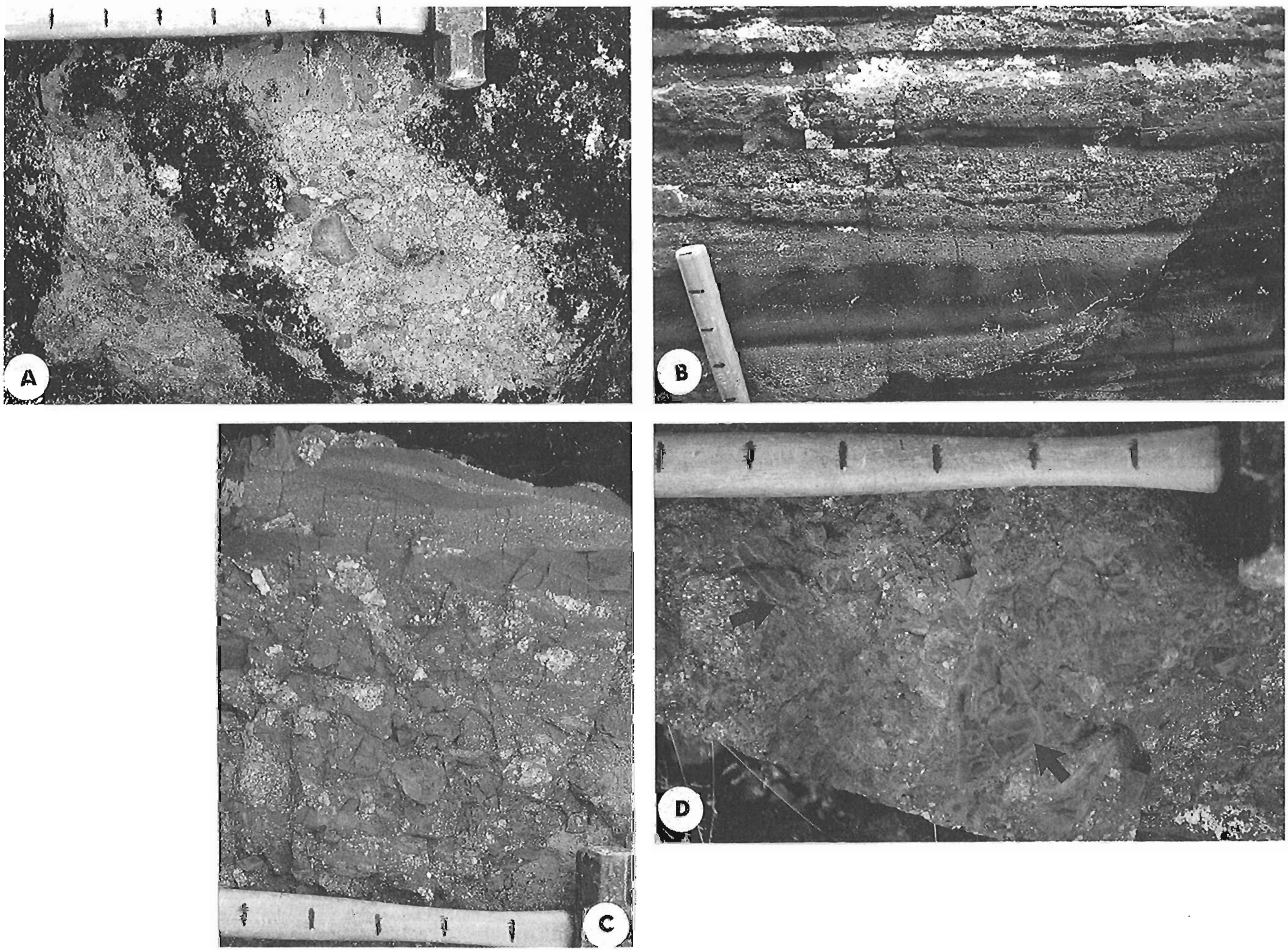

Figure 5

A. Heterolithic mafic lahar breccia deposit, unit 4e. (GSC 1992-261G)

B. Low-angle cross-stratified, ripple-marked tuffaceous sandstone, siltstone, unit 4a. (GSC 1992-261L)

C. Crossbedded Kunwak Formation conglomerate, arkosic sandstone and siltstone, unit 5. (GSC 1992-261F)

D. Conglomerate bed in Kunwak Formation. The angular mottled clasts, marked by the solid arrows, are fine grained and feldspar-rich. These are interpreted to have been derived from high-level fractionated intrusions belonging to the Christopher Island magmatic suite. (GSC 1992-261I)

Table 1. Mineral occurrence data in a part of NTS 65J/10

Occurrence	Major Element	Host Formation	Host rock	Metallic minerals	Alteration Minerals	Deposit Type
A	U	South Channel	basal breccia	uranophane, pitchblende	specularite, hematite, albite	fracture-controlled uranium
B	U	Kazan	arkosic sandstone	secondary green & yellow uranium-bearing minerals	hematite, pink carbonate	fracture-controlled uranium
C	Cu	Christopher Is.	lithic volcaniclasic sandstone	chalcocite, digenite; malachite	chlorite, porphyroblastic carbonate	redbed copper
D	Cu	Kazan	sandstone, siltstone, mudstone	chalcocite, digenite, chalcopyrite, bornite; malachite	chlorite, porphyroblastic carbonate	redbed copper
E	Cu+U??	Kunwak	conglomerate	none observed	none observed	redbed copper?

The presence of alkaline plutonic clasts indicates that this unit is stratigraphically above the Kazan Formation but the exact stratigraphic position is unclear. This could be a volcaniclastic unit near the base of the Christopher Island Formation or a post-Christopher Island Formation fluvial sedimentary unit. These coarse grained fluvial rocks indicate renewed fault activity peripheral to the Angikuni subbasin. If the interpretation is accepted that these clasts were derived from high-level fractionated intrusions belonging to the Christopher Island suite, then this conglomerate may record a late- to post-Christopher Island unroofing of the basement. This unit lies on the 240 m contour, topographically below the transition from nonvolcanic to volcanic-bearing strata. The restricted areal distribution may reflect a local channel deposit developed on lower redbeds and basement.

Unaltered northwest-trending diabase dykes, up to 5 m wide, cut the Christopher Island and Kazan formations. These dykes are correlated with the MacKenzie igneous event based on an absence of alteration and orientation.

MINERAL OCCURRENCES: CLASSIFICATION, ALTERATION, AND ORE MINERALS

The data presented in Table 1 are preliminary descriptive features based on field observations and preliminary petrographic work. Whole rock and trace element analyses are in progress but were not available in time for publication.

Copper occurrences

The lateral extent of copper-bearing sedimentary rocks at each occurrence is undetermined due to limited exposure. The stratigraphic thickness of altered and copper-bearing strata at each occurrence is 1-1.5 m at occurrence C and approximately 15 m at occurrence D.

Copper occurs at or near the contact between the uppermost Kazan and lowermost Christopher Island formations. Copper sulphides are hosted by different lithologies at occurrences C and D. Occurrence D is hosted in interbedded Kazan arkosic sandstone, siltstone, and mudstone, the last commonly displaying desiccation cracks (Fig. 6A). Occurrence C is hosted by Christopher Island crossbedded arkoic sandstone containing angular large trachytic volcanic clasts, up to 10 cm in diameter (Fig. 6B). Even though there is a difference in host rock between occurrences C and D, both occurrences lie between the 260-270 m contour, an interval that corresponds approximately to the contact between the Kazan and Christopher Island formations. This topographic and stratigraphic restriction suggests that the copper is stratabound.

The most diagnostic visual features of both occurrence C and D are the profound colour changes that accompany reduction and sulphide introduction. Variations from pink through red to maroon and purple-maroon are the norm throughout unmineralized Baker Lake Group strata. In contrast, copper-bearing rocks are visually anomalous due to distinct colours: light grey, pinkish grey to blackish grey, compared to orange, red to maroon hues in unmineralized clastic sedimentary rocks. Copper-bearing rocks commonly exhibit a spotted appearance formed by <1 mm knots of sulphide or a mixture of sulphide and chlorite. At occurrence D, approximately 15 m of volcaniclastic sandstone exhibit bleaching due to alteration. However, sulphides are not uniformly distributed throughout this stratigraphic thickness.

Preliminary ore microscope studies indicate that chalcocite and digenite are the most abundant copper sulphides but some evidence exists for precursor chalcopyrite and bornite. Chlorite, assumed to be a Mg-rich variety based on optical properties, overgrows feldspar clasts and forms poikiloblastic knots up to 0.4 mm. Chlorite knots account for the spotted texture seen in hand specimens. The carbonate content is probably higher in copper-bearing rocks. The habit

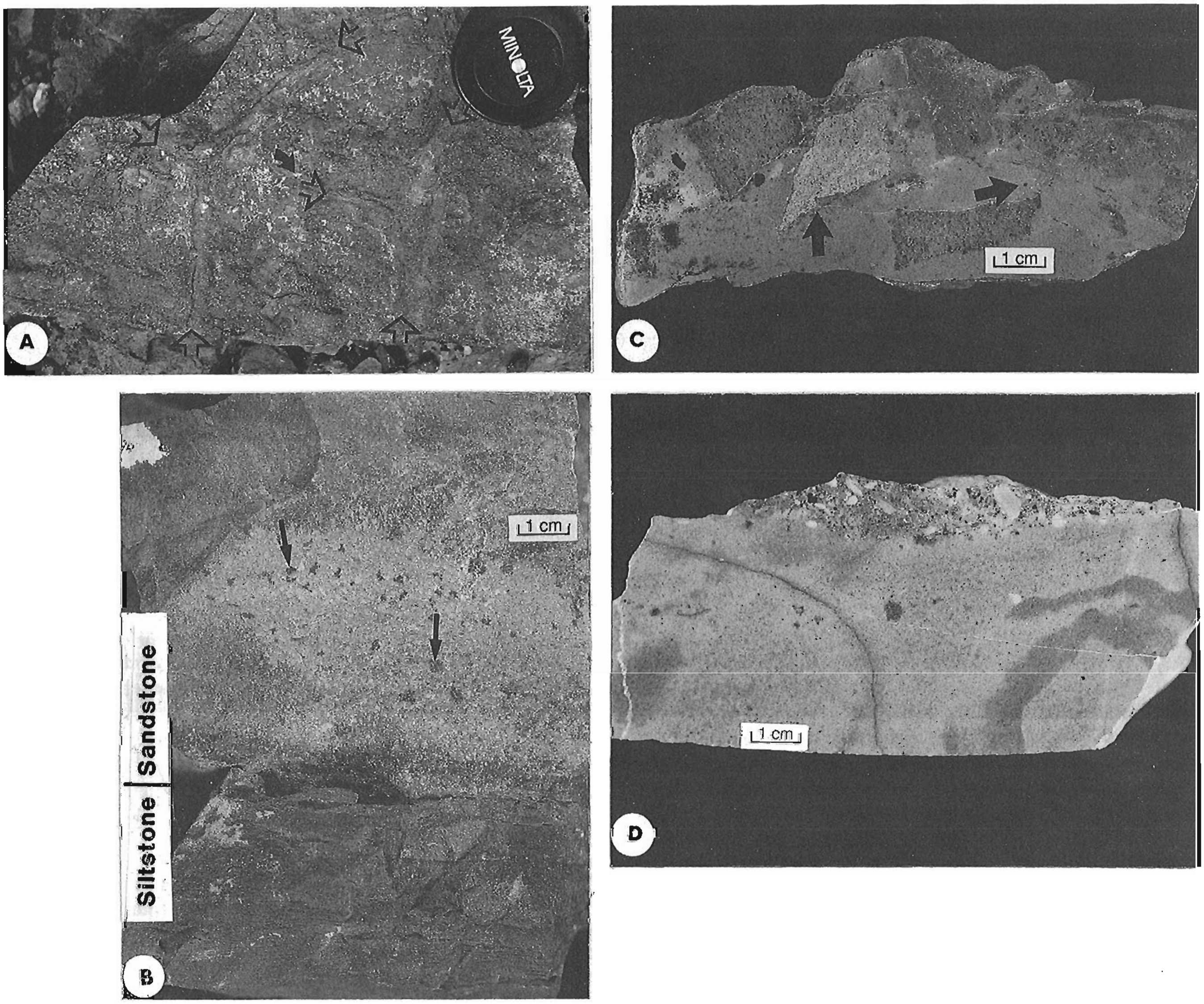

Figure 6.

A. Redbed copper occurrence 4: polygonal mud-cracked (open arrows) bedding surface in a copper-bearing interbedded arkosic sandstone, siltstone, and mudstone. The light wispy films (solid arrow) are malachite after copper sulphides. (GSC 1992-261A)

B. Redbed copper occurrence 4: strongly reduced, copper-bearing arkosic sandstone in contact with chloritic siltstone. Colour changes in reduced sandstone are due to hematite removal and formation of chlorite and copper sulphides as disseminated grains and in knots (solid arrows). These changes impart a grey hue to copper-bearing arkose.

C. Redbed copper occurrence 3: reduced medium grained sandstone with angular clasts of Christopher Island porphyritic volcanic rocks. Copper sulphides (marked by solid arrows) are localized along the redox contacts between reduced sandstone matrix and oxidized clasts and as disseminations in reduced sandstone.

D. Redbed copper occurrence 3: interface between volcaniclastic sandstone and reduced fine grained sandstone. Both units are reduced indicated by the overall light colour and contain disseminated copper sulphides (fine black flecks).

of carbonate in mineralized rock differs from unmineralized sandstone. Carbonate in copper-bearing sandstone is euhedral to subhedral and overgrows framework clasts and matrix, whereas carbonate in unmimeralized sandstone is finer grained, anhedral, and texturally appears similiar to carbonate-cemented sandstone elsewhere in the Baker Lake Group.

Chloritization is a sensitive indicator of alteration. In outcrop and hand specimen the presence of chlorite is easily recognizable considering the pink to red hues so prevalent in the clastic sedimentary rocks. Chloritized sandstone without visible sulphides was noted in the frost-heaved outcrops up to 750 m west of occurrence D. Stratigraphically, these outcrops are below occurrence 4.

Macroscopic redox boundaries are local controls for sulphide concentrations. At occurrence C, crossbedded sandstone can contain scattered coarse clasts of porphyritic volcanic rock (Fig. 6C). The interface between strongly reduced buff- to cream-coloured sandstone and altered hematitic volcanic clasts was a locus for sulphide concentration. Some volcanic clasts are concentrically altered. The outer shell is replaced by chlorite, sulphides, and euhedral coarse grained carbonate whereas the core lacks the former alteration minerals and is hematite-bearing.

Anomalous but low copper, uranium, and silver were recorded at occurrence E (Noranda Exploration Co. Ltd., 1981; Noranda Exploration Co. Ltd. and Agip Canada Ltd., 1979). Sulphides were not seen during the 1981 examination nor during the course of this study. Occurrence E, interpreted to be stratigraphically higher than occurrences C and D, does not exhibit the profound reduction features at the latter occurrences. Elevated metal contents may be in part due to the abundant clasts derived from highly fractionated feldspar-rich alkaline plutons which are enriched in uranium, thorium, and various incompatible elements (LeCheminant et al., 1987). Further petrographic studies are required to evaluate this possibility at occurrence E.

Uranium occurrences

Both uranium occurrences, numbered A and B in Table 1 are hosted by lower Baker Lake Group sedimentary rocks. Anomalous radioactivity is confined to fracture zones having very narrow widths and little apparent strike length. These fracture-controlled occurrences are similiar to structurally-controlled mineralization present in and adjacent to rocks of the Baker Lake Group.

At occurrence A, uranophane-coated fractures, trending 063/85SE, are hosted in the basal breccia, lowest subunit of the South Channel Formation. The basal breccia is similiar to that in Figure 3C but comprises gneissic diorite, granodiorite, and granite fragments infilled by red siltstone. However, the breccia texture in the radioactive zone has been variably obliterated due a pervasive iron metasomatism. Altered host rocks vary from pink to red due to ultra fine grained hematite. The most intensely altered/hematitized specimens appear to have "episyenite" type appearance. Preliminary petrography indicates the following alteration types: 1) hematitization: exhibited by a pervasive dusting of an opaque oxide, presumably hematite, throughout the breccia fragments; 2) albitization: clear albite intergrown with fine- to coarse-grained specularite occupy veinlets and fills secondary porosity; 3) minor carbonatization, and 4) chloritization (an iron-rich variety based on interference colours).

Occurrence B is hosted in fractured, grit-sized to fine grained crossbedded arkosic sandstone of the Kazan Formation. Highly oxidized discontinuous veinlets, 2-5 mm wide, contain pitchblende and its green to green-yellow oxidation products. Radioactivity is enhanced along both sets of fractures, trending 090-110° and 255-270° with the highest radioactivity present at fracture intersections. Alteration is typical of fracture-controlled occurrences. Small carbonate-cemented breccia bodies occur at the intersection of the fracture sets and the wallrock adjacent to carbonate veins was hematitized.

SUMMARY

The Baker Lake Group in the northeastern portion of the Angikuni subbasin displays many criteria that identify this subbasin as a potential stratiform copper environment. The present configuration of the Angikuni subbasin as a long linear northeasterly-trending basin bounded by regional fault zones implies a fault-controlled basin. The earliest basin fill was fault-controlled immature red clastic rocks that have abrupt lateral facies changes and fine up-section. The basin was flooded by voluminous mafic alkaline with minor felsic volcanic rocks. The geochemistry of this alkaline suite implies derivation from metasomatized upper mantle rocks and thus the volcanic rock represent a metal-rich source rock. Widespread propylitization of the volcanic pile could have provided copper and other metals to basinal brines. Tectonic distrubances may have caused basinal brines to migrate towards the basin margin and be localized along the contact between the Kazan and Christopher Island formations (Kirkham, 1989).

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Mineral Resources Division

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Abstract: Newly interpreted intrusive relationships between a multiphased intrusion and Proterozoic arenaceous sedimentary rocks previously correlated with the Hurwitz Group and the correlation of these sedimentary rocks with older clastic sequences in the Churchill Province are presented. Crossbedded drab greenish-grey pyritic sericitic arenite is correlated with the Montgomery Lake Group, the latter older than the Hurwitz Group. In the study area, arenite was intruded, thermally metamorphosed and metasomatized by the syenite complex.

The Proterozoic Heninga Lake syenite complex has three principal rock types that are considered comagmatic: (1) a main massive equigranular biotite±hornblende syenite, transitional to (2) a trachytoid megacrystic feldspar syenite and (3) a late, fine grained granite which cuts (1) and (2). The granite is restricted in area to the contact between the intrusion and Archean metabasaltic country rock. Recognition of an intrusive contact between the Heninga Lake syenite complex and sedimentary rocks correlated with the Montgomery Lake Group will allow dating of the oldest recognized Proterozoic sedimentary sequence in the Churchill Province, north of 60°.

Résumé : Dans cet article, on présente de nouvelles interprétations des corrélations établies entre une intrusion multiphasée et des roches sédimentaires arénacées d'âge protérozoïque, anciennement corrélées avec le Groupe de Hurwitz, et les corrélations établies entre ces roches sédimentaires et des séquences clastiques plus anciennes de la Province de Churchill. Une arénite pyriteuse de teinte gris verdâtre terne et à stratification oblique, est corrélée avec le Groupe de Montgomery Lake, plus ancien que le Groupe de Hurwitz. Dans la région à l'étude, l'arénite a été intrudée, thermométamorphisée puis métasomatisée par le complexe syénitique.

Le complexe protérozoïque de Heninga Lake contient trois grands types lithologiques considérés comme comagmatiques: 1) une syénite principale à biotitehornblende, massive et isogranulaire, qui passe progressivement à 2) une syénite trachytoïde à mégacristaux de feldspath et, 3) un granite tardif à grain fin qui recoupe 1) et 2). Le granite se limite au contact entre l'intrusion et la roche encaissante métabasaltique d'âge archéen. L'identification d'un contact intrusif entre le complexe syénitique de Heninga Lake et les roches sédimentaires corrélées avec le Groupe de Montgomery Lake permettra de dater la plus ancienne séquence sédimentaire protérozoïque reconnue de la Province de Churchill, au nord de 60°.

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INTRODUCTION

Ridler (1973) identified an alkaline, probably Archean, intrusive complex approximately 3 km in diameter and centred at 62°00′20″N, 96°24′W. A well-preserved carbonatite-bearing alkaline complex at Kaminak Lake (Davidson, 1970), recently dated at 2659 ± 5 Ma (Cavell et al., 1992) lies 100 km northeast of the Heninga intrusion. A possible third Archean alkaline intrusion, very poorly exposed (Ridler, 1972), lies 65 km northeast of the Kaminak Lake complex, near Last Lake. The apparent Archean age, alkaline nature and location of the three intrusions 165 km apart along the backbone of the linear structural block labelled the Hearne Province (Hoffman, 1988) suggests the possibility of an Archean alkaline province. Emplacement of the three intrusive bodies could be inferred to be controlled by a linear tectonic feature present in the late Archean granitoid-greenstone host rocks.

The primary aim of the project located northeast of the Kogtok River and straddling the boundary between NTS 65I/1 and 65H/16 (Fig. 1) was to map the various intrusive phases and contact relations of the Heninga Lake complex with Proterozoic and Archean supracrustal rocks and collect samples of the complex and other intrusions for zircon geochronology. The second objective was to evaluate the potential of the following: lode gold within Archean metasedimentary and metavolcanic rocks, paleoplacer gold in pyritic Proterozoic quartzose metasedimentary rocks, and rare-earth elements and niobium associated with the Heninga complex.

Mapping in 1992 and initial petrographic studies indicate that the Heninga intrusive complex is syenitic, not alkaline, bears no carbonatites, and may be as young as early Proterozoic rather than late Archean. On this basis, it is not directly correlative with the Kaminak Lake alkaline complex.

GEOLOGY OF ARCHEAN SUPRACRUSTAL AND INTRUSIVE ROCKS

Archean supracrustal rocks in the area belong to the Kaminak Group (Bell, 1971) and consist of a steep-dipping, northwest-facing mafic volcanic domain dominated by pillowed and massive metabasalt flows (unit 1a), and related gabbroic intrusions (unit 1d) (Fig. 1). Massive metabasalt consists of a fine grained foliated assemblage of hornblende, epidote, chlorite, plagioclase, and quartz. Pillow structures were observed throughout the project area, but were reliable top indicators only in areas of low strain. Well-exposed teardrop shaped pillows, with or without quartz±carbonate filled eyebrow structures, indicate north to northwest facings. These eyebrow structures are similiar to those in pillowed metabasalt-dominated domains in the Rankin Inlet Group (Tella et al., 1986) and in the Fat Lake area, NTS 55K/4 (Miller, 1989). Pillow reversals were observed near the contact of the tonalite pluton in the southeastern portion of the map area. A rare facing reversal may be related to this intrusion, similiar to reversals observed adjacent to plutons in the Kaminak Lake area (Davidson, 1970).

Volcaniclastic rocks (unit 1c), up to 30 m in true thickness, are interstratified with metabasalt flows throughout the pile. Southeast of the lake that constitutes a widening of the Kogtok River, the mafic volcanic pile contains numerous thin (1-2 m), discontinuous units of chloritic+calcareous tuff. Southeast of the Heninga Lake intrusion, a northeast-trending pillow breccia composed entirely of metabasaltic clasts (unit 1b), is the only continuous unit mappable within unit 1.

The proportion and diversity of metasedimentary rock types interstratified with pillowed basalt in the northwestern portion of the map area record changes in volcanism and sedimentation. Hiatuses in mafic volcanism are recorded by the presence of fine grained felsic crystal tuff and associated sediments. Low energy clastic and chemical sedimentary rocks, siliceous and chloritic schists, graphitic phyllite, and oxide iron-formation, are interstratified with feldspar crystal tuff.

Felsic tuff units contain weakly to moderately flattened anhedral to subhedral plagioclase crystals, up to 0.7 mm in size, in a matrix of very fine grained quartz, feldspar, and biotite. They commonly contain disseminated iron sulphides. Macroscopic banding in crystal tuff parallels the compositional banding in adjacent clastic sediments and iron-formation. Development of prograde chlorite, sericite, and biotite along a foliation that parallels the compostional banding suggests S_0=S_1. Bedding trends throughout the northwestern map area vary from northeast to east with steep northerly dips. The beds top to the north according to facing directions in adjacent pillowed volcanic rocks.

Iron-formation constitutes some 5-10% of interflow sedimentary rocks. In these rocks, banded oxide and silicate iron-formation are interstratified with or bounded by graphitic phyllite and siliceous biotite-bearing felsic tuff. Garnet+biotite+chlorite-bearing assemblages in phyllitic schists and grunerite+biotite±hornblende overgrowing earlier fabrics suggest regional late Archean upper greenschist-lower amphibolite metamorphism.

The northeast-trending Archean supracrustal rocks are intruded by fine- to medium-grained gabbros (unit 1d), both subparallel and discordant to stratigraphy. These intrusions commonly are strongly foliated along the margins and have medium grained porphyroblastic textures preserved in the centre of larger bodies.

Near the southeastern margin of the map area, the metabasalt sequence is intruded by a tonalitic pluton, the northern margin of a large pluton that extends from Yandle Lake to northeast of Heninga Lake (Bell, 1971). An agmatitic border zone up to 100 m wide is present along the entire mapped contact.

North and northeast of the widening of the Kogtok River in the project area, the Archean greenstone belt is intruded by a northeast-trending, fine grained microporphyritic gray granite (unit 2, Fig. 1), shown by Bell (1971) as a circular pluton of about 1-2 km diameter. This granite is intrusive into the metabasalt, metagabbro, and felsic tuff along the northern margin and contains biotite-metasomatized metabasalt roof

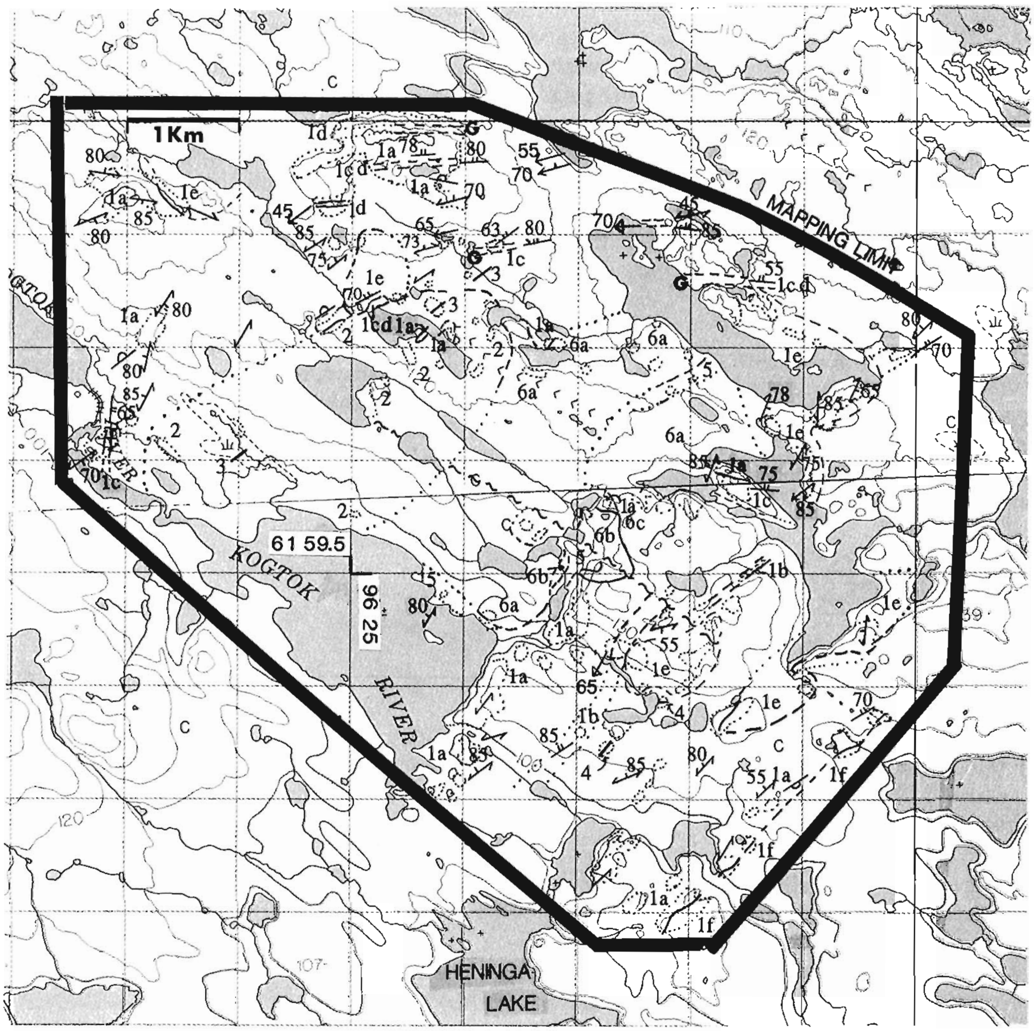


Figure 1 (above). Geological map of the Archean and Proterozoic rocks in the area of the Heninga Lake syenite complex.

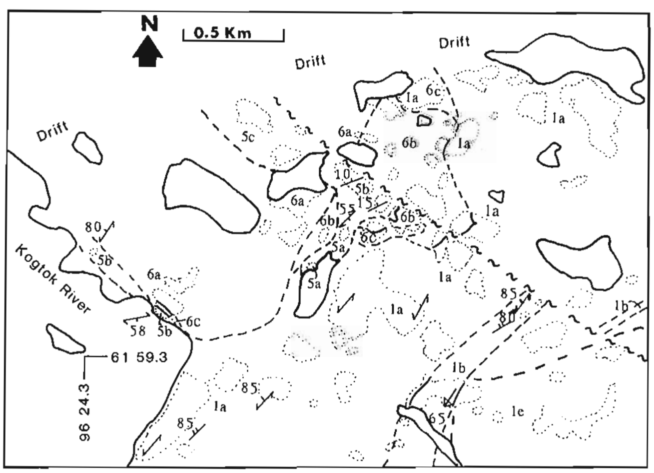


Figure 2. Detailed geology along the southern contact of the Heninga Lake complex and relationships with strata correlated with the Montgomery Lake Group. See Figure 1 for legend.

Legend for Figures 1 and 2.

PROTEROZOIC

6 Heninga Lake Syenite Complex
- 6c: Fine-grained granite
- 6b: Trachytoid megacrystic feldspar syenite
- 6a: Biotite +/- hornblende syenite

5 Montgomery Lake Group
- 5c: Coarse-grained arenite
- 5b: Pyritic sericitic arenite
- 5a: Pebbly sericite schist

4 Kaminak Dyke
- Megacrystic feldspar diabase

ARCHEAN

3 Kazan Dykes
- Fine- to medium-grained diabase

2 Fine-grained microporphyritic gray granite

1
- 1f: Granodiorite/tonalite
- 1e: Medium-grained porphyroblastic gabbro
- 1d: Oxide iron formation, graphitic phyllite garnet-chlorite schist, silicate iron formation
- 1c: Feldspar crystal tuff
- 1b: Pillow breccia
- 1a: Pillowed basalt, fine-grained basalt(amphibolite)

Bedding: tops unknown, known
Fold axis, with plunge
Pillow: facing known
Axial plane, with dip
Foliation
Mineral lineation
G Gossan
Outcrop, frost-heaved outcrop

pendants near its northeastern end. Contact relations with the Heninga Lake complex are not exposed along the eastern contact. A penetrative fabric was not observed in outcrop but a weak penetrative foliation defined by alignment of the fine grained biotite and retrogressive sericite/muscovite was observed during preliminary petrographic examination. Sericite/muscovite after feldspar is also accompanied by minor carbonate. This granite pluton is cut by a fine-grained metadiabase, trending 060°, (unit 3). This metadiabase is correlated with the Archean Kazan swarm (Eade 1986; LeCheminant et al., 1976). This dyke, approximately 22 m wide, is chilled against unit 2 and lacks a penetrative fabric. The hornblende+brown biotite metamorphic assemblage in this magnetite-bearing dyke suggests recrystallization under upper greenschist-lower amphibolite conditions. This metamorphic rank is in agreement with late Archean regional metamorphism preserved in iron-formation.

GEOLOGY OF PROTEROZOIC ROCKS

Northeasterly- to easterly-trending glomeroporphyritic megacrystic feldspar metadiabase dykes (unit 4), intruded Archean metabasalt in the southern portion of the map area. These dykes are the oldest recognized Proterozoic rocks in the project area and are correlated with the distinctive Proterozoic Kaminak dyke swarm (Davidson, 1970). Metamorphic assemblages in these Kaminak dykes include: chlorite+epidote after megacrystic feldspar, anatase+titanite after magnetite and epidote+chlorite+carbonate+actinolite+ sericite replacing the groundmass. Thus Proterozoic greenschist metamorphism overprinted higher grade Archean metamorphic assemblages, a relationship similiar to metamorphic relations to the northeast in the Kaminak Lake area (Davidson, 1970).

Proterozoic sedimentary rocks in the project area were not identified during regional mapping (Bell, 1971), probably due to the limited aerial distribution of these sedimentary rocks. Ridler (1973) recognized these sedimentary rocks, correlated them with the Hurwitz Group and stated that these sedimentary rocks rest unconformably on the Heninga Lake Complex. Mapping in 1992 and preliminary petrographic examination have shown that the sedimentary rocks do not correlate with the Hurwitz Group and were intruded by the Heninga Lake complex. Sedimentary features and opaque and silicate mineralogy support a correlation with the Montgomery Lake Group, a sedimentary sequence that underlies the Hurwitz Group locally and unconformably overlies Archean strata near Henik Lake (Eade, 1974). A schistosity dipping moderately southeast is present in all Montgomery Lake Group sedimentary rocks in the project area. This trend is subparallel to, but significantly shallower than, foliations recorded in Archean Kaminak Group rocks.

Mapping in 1992 divided the Montgomery Lake sediments into three lithostratigraphic units (Fig. 1, 2). This division is facilitated by relatively good topographic relief in the area of the southeastern contact of the Heninga Lake intrusive complex and by well preserved sedimentary structures.

The lowest stratigraphic unit (5a, Fig. 2) is a pebbly yellowish-green sericite schist, containing rounded to subangular clasts, to 5 cm in diameter, of vein quartz and foliated granite (Fig. 3A). This unit grades upward into the second unit (5b), a crossbedded drab gray-green pyritic quartz arenite. The arenite unit consists of well-sorted, 0.15-0.2 mm, subrounded to rounded, equant to elongate quartz grains in a foliated matrix of sericite and minor chlorite. The heavy minerals, which can be concentrated along crossbed laminae, consist of pyrite, ilmenite, titanite, tourmaline, and zircon.

The third unit (5c), a sequence of interbedded well- to poorly-sorted, medium grained, quartz-rich lithic arenite to grit is interpreted as the uppermost preserved sediment. Frost-heaved outcrops of unit 5c are located north of well-bedded

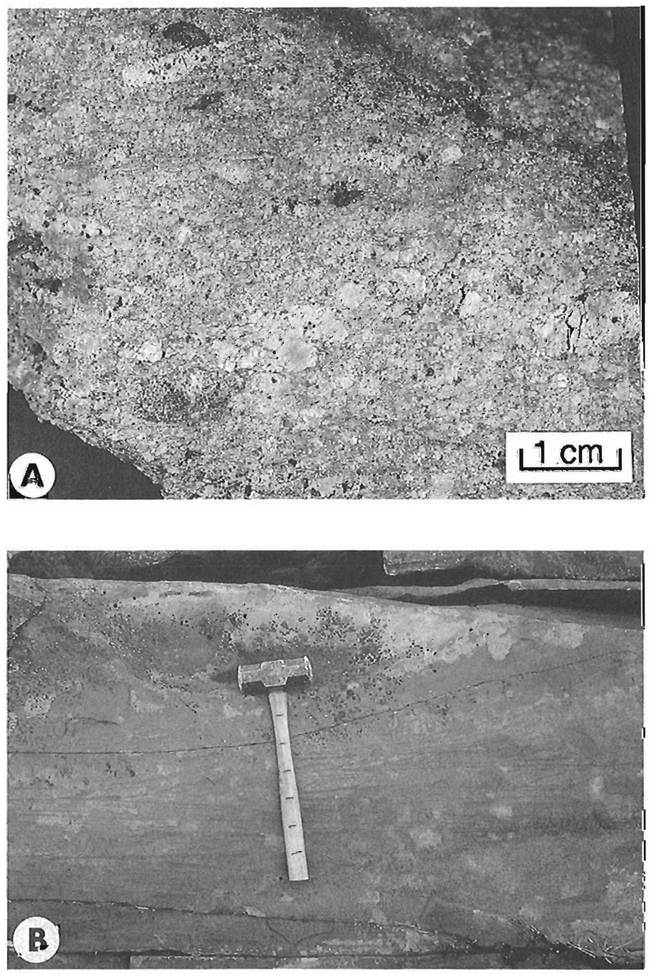


Figure 3.

A. Gritty clastic texture in pyrite-bearing sericite schist, unit 5a, lowest preserved sedimentary unit correlated with the Montgomery Lake Group. Clasts are dominantly granitic in composition.

B. Crossbedded pyritic quartz arenite, unit 5b, correlated with the Montgomery Lake Group. Pyrite is concentrated along and disseminated between crossbed laminations. Black marks on the hammer handle are 5 cm apart.

unit 5b and are positioned between the 100 and 110 m contour interval (Fig. 1). Well-rounded, strained polygranular quartz is the dominant clast. Lithic clasts contain sericite, possibly after feldspar.

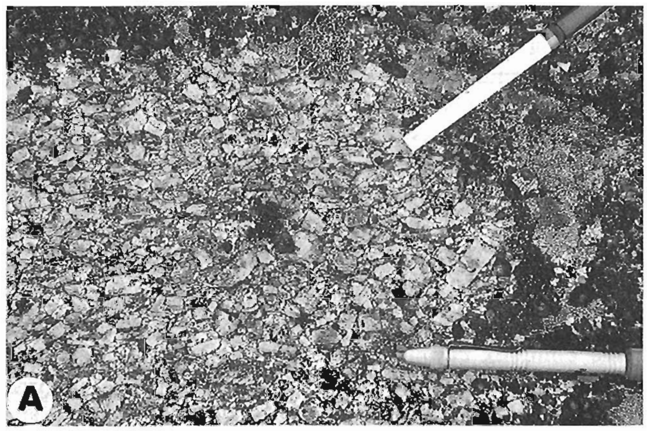
These sedimentary rocks are preserved in three separate locations near the margin of the Heninga Lake complex and at each location the sedimentary rocks are in contact with the intrusion. Intrusive relationships are best exposed along the southeastern margin of the complex (Fig. 2), adjacent to the trachytoidal phase. Relationships suggest that some of the preserved sediments are roof pendants. Megacrystic feldspar syenite is in contact with hornfelsed arenite and the late granitic phase of the complex intrudes arenite and megacrystic feldspar syenite. Macroscopic and microscopic evidence of thermal metamorphism is present in all outcrops of Montgomery Lake equivalents in contact with the Heninga Lake intrusion or at a topographic location inferred to be near the intrusive contact. In outcrop, thermal metamorphism is recognized by the change from drab grey-green to grey-black due to biotite growth in the hornfelsic sedimentary rocks. Thermally-altered sediments in contact with megacrystic feldspar syenite have been extensively recrystallized and metasomatized. In these sedimentary rocks thermal metamorphism has been responsible for recrystallization of the foliated sericite matrix, unoriented metasomatic biotite intergrown with recrystallized sericite, the absence of metamorphic chlorite, and trace quantities of anhedral porphyroblastic apatite and acicular tourmaline intergrown with metasomatic phyllosilicates. The superposition of thermal metamorphism on the regional penetrative fabric present in the Montgomery Lake equivalents implies that intrusion of the Heninga Lake complex was late- to postdeformation of these sedimentary rocks.

Ridler (1973) described the Heninga Lake intrusive complex as alkaline and possibly correlative with the carbonatite-bearing alkaline complex at Kaminak Lake (Davidson, 1970). Because there is no evidence of feldspathoids, sodic amphiboles, or sodic pyroxenes in the pluton, and in the absence at present of data on whole rock compositions from which to calculate the alkali/aluminum ratio, the intrusion is here referred to as the Heninga Lake syenite complex.

The complex, elongated along a northeasterly trend (Fig. 1), has dimensions of approximately 1.5 km in width and 3.0 km in length. The intrusion is divided into three lithological units: two syenites and a granite. All three rock units of the complex, their intrusive relationships to each other and to the arenites are best exposed and exhibited at the UTM co-ordinate 637100, NTS 65H/16.

An estimated 90% of the surface of the Heninga complex is occupied by an undeformed, equigranular, medium grained, magnetite-bearing biotite±hornblende syenite, unit 6a. Mafic minerals occur as two distinct size populations and textural forms. One is medium grained, 2-4 mm, biotite and hornblende phenocrysts. Hornblende and poikilitic medium grained biotite with exsolution needles of rutile both contain inclusions of magnetite, apatite, titanite, zircon, orange metamict phases, as well as biotite in hornblende and hornblende in biotite. The second, fine grained biotite, <0.07-0.15 mm in size, can form aggregates up to 4 mm. This textural variation may represent a second generation of biotite after either porphyritic hornblende or biotite. Subhedral to euhedral microphenocrysts of apatite and titanite, up to 1 mm in size, are common throughout this intrusive phase. Epidote, a ubiquitous assessory phase on feldspar and inclusions in biotite and hornblende, may be a late magmatic phase rather than related to retrogressive regional metamorphism.

Equigranular syenite is gradational with increasing size and quantity of feldspar crystals into the second unit. This phase, the megacrystic feldspar syenite, unit 6b, is present only along the southeastern side of the intrusion (Fig. 2). Feldspar megacrysts, up to 4 cm in length (Fig. 4A), show an increasing alignment towards the southern contact with Archean metabasalt. Melanocratic trachytoidal syenite of the megacrystic syenite unit consists of perthitic feldspar megacrysts with interstitial fine grained biotite and assessory



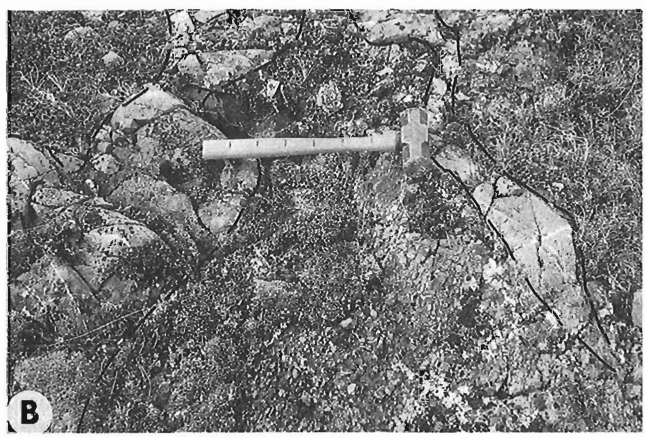


Figure 4.

A. Trachytoidal megacrystic feldspar syenite, Heninga Lake complex. Feldspar megacrysts can be aligned in two directions, see Figure 3. Rapidograph pen is 13 cm long.

B. Fine grained granite dykes, outlined by dashed line, cut trachytoidal megacrystic feldspar syenite, southeastern contact of the Heninga Lake complex.

magnetite, quartz, zircon, carbonate, and microphenocrysts of apatite to 1 mm. Sericite after feldspar may represent minor retrogression or a late magmatic crystallization. The orientation of trend lines derived from the alignment of megacrystic feldspar in the trachytoidal border phase and of biotite and hornblende in the equigranular phase of the intrusion are shown in Figure 5. Trend lines cluster between the azimuths 170-200° and 220-230°. The latter trend parallels schistosity trends present in arenaeous sediments correlated with the Montgomery Lake Group. Intrusion of the Heninga Lake complex is interpreted to have been late or post-deformation that formed the penetrative fabrics in these arenaceous rocks.

The third rock type, unit 6c, fine grained biotite granite and medium- to fine-grained granitic dykes, are areally restricted to the margin of the Heninga complex. Agmatitic intrusive textures formed by anastomosing granitic dykes in trachytoidal megacrystic syenite and in the adjacent Archean greenstone are common along the southeastern contact of the complex. Small-scale, fine- to medium-grained subvertical sheets, up to 5-10 m wide, and dykes 50 cm to 2-3 m wide, were recognized near the margin of the equigranular biotite- and hornblende-bearing syenite. The granite and related dykes and sheets display an equigranular interlocking igneous texture consisting of 0.1-0.5 mm quartz, plagioclase, and perthitic feldspar with phenocrysts of plagioclase, rare quartz and glomeroporphyritic plagioclase aggregates up to 1 mm in size. Some plagioclase phenocrysts and glomero-porphyritic aggregates of plagioclase are mantled by graphic intergrowths.

The spatial distribution of the three phases of the Heninga Lake complex coupled with textural and mineralogical transitions between these intrusive phases suggests all three phases are comagmatic.

GEOCHRONOLOGICAL IMPLICATION OF THE HENINGA LAKE SYENITE COMPLEX

The type area for early Proterozoic sedimentation in the southern Churchill Province, north of 60°, lies in the area of Montgomery-Bray-Ameto lakes (Eade, 1974; Aspler et al., 1992), In the type area, the fluvial Montgomery Lake Group is the oldest preserved sedimentary sequence to rest unconformably on the 2697 ± 14 Ma to 2681 ± 3 Ma Henik Group The Montgomery Lake Group is overlain unconformably by the Padlei Formation and Maguse member of the Kinga Formation, both of the Hurwitz Group. An age of 2094 Ma from a Hurwitz gabbro (Patterson and Heaman, 1991), intrusive into the Ameto Formation of the Hurwitz Group, establishes a minimum age for Ameto sedimentation. In the type area the lack of datable igneous bodies intrusive into the sedimentary strata of the Kinga and Padlei formations or Montgomery Lake Group precludes obtaining an older minimum age for sedimentation. Recognition of the intrusive contact between the Heninga Lake syenite complex and sedimentary rocks correlated with the Montgomery Lake Group means that dating the intrusion will provide a minimum age contraints for the oldest recognized Proterozoic-type sedimentary sequence in the Churchill Province, north of 60°.

ECONOMIC GEOLOGY

Analyses of sulphide-bearing grab samples from Archean and Proterozoic sedimentary rocks are given in Table 1. Archean interflow sediments, graphitic phyllite, silicate chlorite-rich iron-formation, and felsic tuff, contain finely disseminated pyrite, pyrrhotite, and rare chalcopyrite. Gold and silver abundances from iron-formation and associated sedimentary rocks are within the range of background values for this rock type. The potential for Archean lode gold mineralization hosted in iron-formation is low. Oxide iron-formation is thin, lacks significant thickening due to folding, and contains very little or no sulphides. Gossans are commonly developed on silicate iron-formation due to oxidation of pyrite and pyrrhotite, or iron-silicates but this type of iron formation is considered of little economic interest. Traces of pyrrhotite with or without chalcopyrite are present in felsic tuff. Copper and zinc values are highest in the tuff and interbedded sedimentary rocks but these values do not represent anomalous stratigraphic units. Trace malachite in quartz veins is present within the agmatite zone peripheral to the tonalite pluton, unit 1e.

Crossbedded, fine grained arenites correlated with Montgomery Lake Group could contain placer gold. A lack of conglomerate and the limited extent, however, significantly lowers their potential as an exploration target. On the other hand, anomalous gold contents, 20 and 32 ppb,

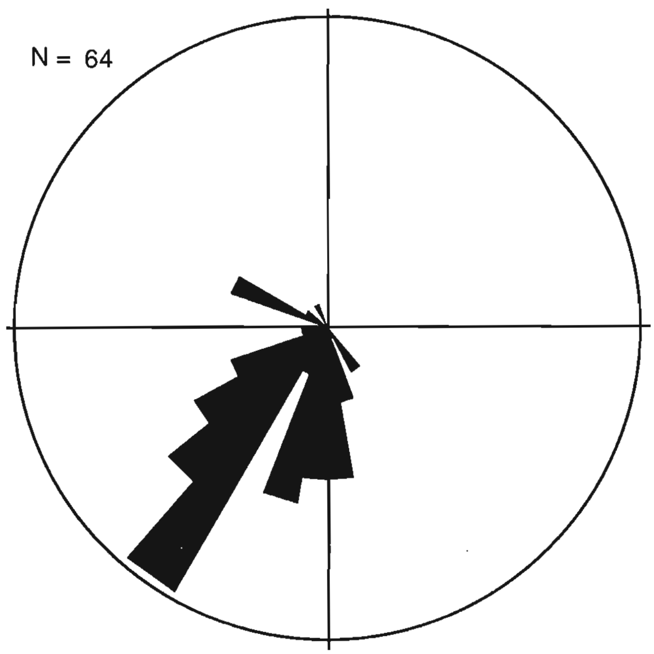


Figure 5. Rose diagram illustrating crystal alignment within biotite±hornblende and megacrystic feldspar syenites.

Table 1. Geochemistry of sulphide-earing grab samples

UTM; NTS	Sample #	Age; Rock Group	Rock type	Au ppb FADCP	Ag ppm ICP	S wt% Leco	Cu ppm ICP	Zn ppm ICP	Pb ppm ICP	Ni ppm ICP	Co ppm ICP	Cr ppm ICP	Th ppm NA	U ppm NA	Zr ppm WR	TiO_2 wt%,WR	Fe_2O_3 wt%,WR
636150; 65I/1	92MYAA 119-1	Archean; Kaminak Gp.	pyrrhotite felsic tuff	1	<0.1	*	146	90	<2	4	7	8	*	*	*	*	*
635920; 65I/1	92MYAA 123-1	Archean; Kaminak Gp.	oxide iron formation	16	0.9	*	126	585	<2	41	29	24	*	*	*	*	*
638220; 65I/1	92MYAA 133-1	Archean; Kaminak Gp.	pyrrhotite silicate iron formation	4	0.7	*	139	104	14	<1	5	15	*	*	*	*	*
638220; 65I/1	92MYAA 133-2	Archean; Kaminak Gp.	sulphide-rich iron formation	16	1.9	*	295	52.9	12	36	66	14	*	*	*	*	*
632900; 65I/1	92MYAA 172-1	Archean; Kaminak Gp.	pyrrhotite siliceous sediment	4	0.9	*	660	111	<2	116	66	97	*	*	*	*	*
632900; 65I/1	92MYAA 172-2	Archean; Kaminak Gp.	pyrrhotite, chalcopyrite felsic tuff	19	1.5	*	940	94	<2	18	19	43	*	*	*	*	*
636980; 65H/16	92MYAA 141-2B	Proterozoic; Montgomery Lake Group	pyritic arenite	32	1.1	2.58	24.2	4.9	<2	36	39	150	11	6	821	1.65	5.20
636980; 65H/16	92MYAA 141-3B	Proterozoic; Montgomery Lake Group	pyritic arenite	20	0.4	0.60	2.9	2.7	<2	8	3	117	7	<5	670	1.32	2.16
636900; 65H/16	92MYAA 182-2	Proterozoic; Montgomery Lake Group	pyritic arenite	3	<0.1	0.21	57.7	9.3	<2	28	5	37	4	<5	144	0.55	2.59

* = not analyzed

and elevated concentrations of TiO_2, Fe_2O_3, Zr, and Cr compare closely with radioactive Archean quartzose metaarenites in the Slave Province (Roscoe, 1990).

The potential for rare element and radioactive mineral enrichment was assessed by using a discriminating scintillometer during routine mapping of the syenitic and granitic phases of the Heninga Lake complex. No anomalous radioactivity was recorded but an increase in the total scintillation, from equigranular syenite into megacrystic syenite, was attributed to increasing potassium content and substantiated by increasing modal percentages of feldspar.

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Geological Survey of Canada Project 810024

Iron-formation, evaporite, and possible metallogenetic implications for the Lower Proterozoic Hurwitz Group, District of Keewatin, Northwest Territories[1]

A.R. Miller and K.L. Reading[2]
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Miller, A.R. and Reading, K.L., 1993: Iron-formation, evaporite, and possible metallogenetic implications for the Lower Proterozoic Hurwitz Group, District of Keewatin, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 179-185.

Abstract: The Early Proterozoic Hurwitz Group contains an areally extensive iron-formation in the Watterson Formation and evaporite in the Tavani Formation. Interbedded iron-rich sedimentary rocks, oolitic quartz-magnetite-hematite iron-formation and magnetite-bearing shale, siltstone, and dolostone are restricted to the W_2 member of the Watterson Formation. This rock association is compatible with the shallow subtidal to intertidal depositional environment proposed for the W_2 member. Even though gold occurrences have not been found in these iron-rich sedimentary rocks, they may represent a favourable host rock for structurally-controlled gold occurrences, analogous to known Proterozoic gold prospects.

The uppermost Tavani Formation comprises carbonate rocks that contain diagenetic albite, pseudomorphs after gypsum-anhydrite, and casts after halite. Interbedding of evaporite, with shallow marine and subaerial rocks in the upper Tavani Formation suggests a hypersaline intertidal or sabkha environment. Even though copper occurrences are not known in the Tavani Formation, the shallow marine carbonates, evaporite and subaqueous as well as subaerial arkosic strata may represent an environment favourable for stratiform sediment-hosted copper deposits.

Résumé : Le Groupe de Hurwitz, daté du Protérozoïque précoce, contient une formation ferrifère de grande étendue dans la Formation de Watterson et des évaporites dans la Formation de Tavani. Des roches sédimentaires ferrifères, une formation ferrifère oolitique à quartz-magnétite-hématite et un shale à magnétite, un siltstone et une dolomie interstratifiés se limitent au membre W_2 de la Formation de Watterson. Cette association lithologique est compatible avec le milieu sédimentaire subtidal à intertidal peu profond proposé pour le membre W_2. Bien que l'on n'ait pas découvert de venues aurifères dans ces roches sédimentaires ferrifères, ces dernières pourraient se révéler des roches hôtes favorables à la manifestation de minéralisations aurifères par suite d'un contrôle structural, minéralisations analogues à d'autres minéralisations aurifères protérozoïques reconnues ailleurs.

La partie sommitale de la Formation de Tavani englobe des roches carbonatées contenant de l'albite diagénétique, des pseudomorphes du gypse et de l'anhydrite et des moulages sur la halite. La présence d'interstratifications d'évaporites avec des roches épicontinentales et subaériennes dans la partie supérieure de la Formation de Tavani Formation semble confirmer l'existence d'un milieu intertidal hypersalin ou de sebkha. On ne connaît pas de venues cuprifères dans la Formation de Tavani, mais les roches carbonatées épicontinentales, les évaporites et les strates arkosiques subaquatiques et également subaériennes pourraient constituer un milieu favorable à la formation de gisements cuprifères stratiformes encaissés dans des sédiments.

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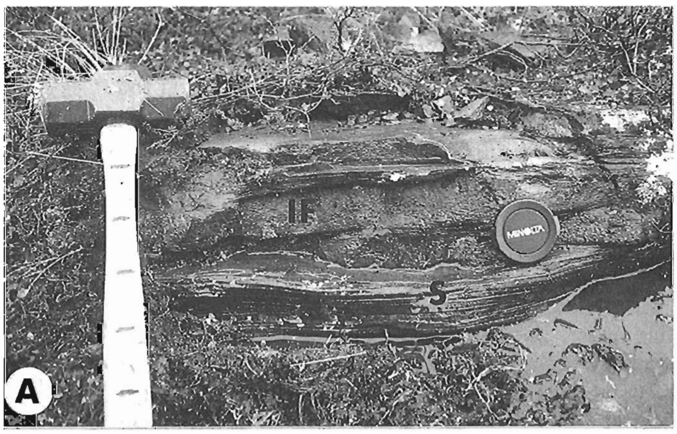
INTRODUCTION

In the 1940s the discovery of gold in the basal quartzite of the early Proterozoic Hurwitz Group, Churchill Structural Province, District of Keewatin, and the subsequent mining of gold in the early 1980s from the Shear Lake deposit signalled basal Hurwitz sedimentary rocks as a Proterozoic gold metallotect. Regional mapping of the Hurwitz Group by Eade (1974) and Bell (1968, 1970) outlined the lithological distribution and sedimentology of this cratonic basin. Renewed stratigraphic analysis of the Hurwitz Group and the discovery of gold in Hurwitz formations other than the basal quartzite (Reading, 1990; Aspler and Bursey, 1990) revived gold exploration in the Hurwitz Group.

The stratigraphic nomenclature, sedimentology, and stratigraphic relationships have been redefined during recent geological investigations. The reader is referred to the above publications in order to follow the growth in the Hurwitz data base. This paper presents new data on the type and areal distribution of iron-rich sedimentary rocks that were not recognized during the most recent mapping studies, provides new information on evaporitic sedimentary rocks, and considers the possible implications for structurally-controlled gold and sedimentary copper deposits, respectively.

IRON-FORMATION: WATTERSON FORMATION

Aspler and Bursey (1990) identified a cyclicity to Hurwitz sedimentation in the Hawk Hill, Mountain, and Griffin lakes area. They divided the Watterson Formation into three members based on the relative proportions of stromatolite, dolostone, and fine grained siliciclastic rocks. The interpreted depositional environment shallowed progressively from supratidal stromatolites during the W_1 member to subtidal to intertidal carbonate flats during the W_3 member. Fluctuations



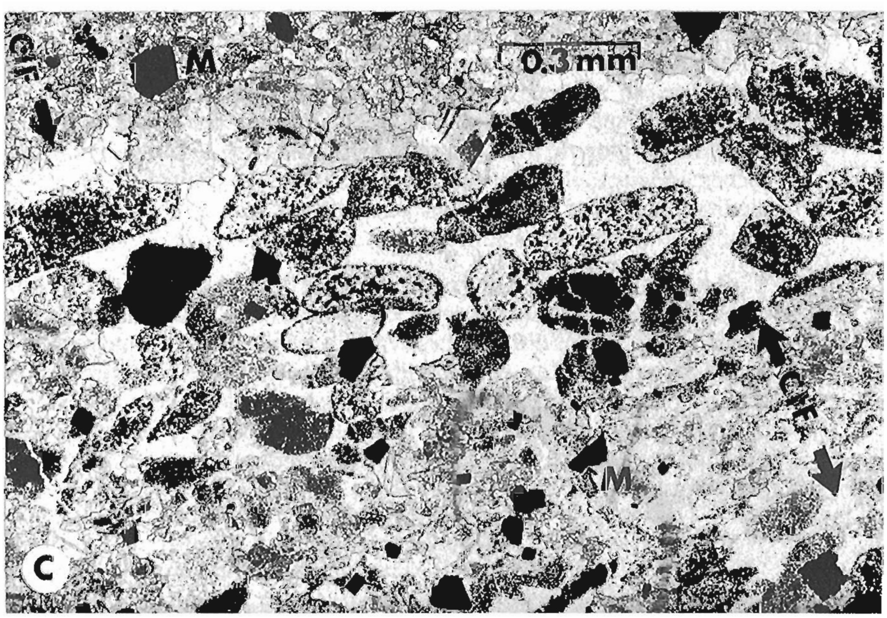


Figure 1.

A. Interbedded magnetite-hematite iron-formation (IF) with finely laminated magnetite-bearing siltstone (S), near the southeastern end of Griffin Lake. Subdivisions on the hammer handle are 5 cm apart.

B. Parallel bedded magnetite-bearing siltstone (S) with interstratified reddish weathering stromatolitic dolostone (D), near north end of Hawk Hill Lake.

C. Photomicrograph of metamorphosed oolitic magnetite-hematite-quartz laminae interbedded with magnetite (M)-bearing carbonate (CIF) iron formation, near the southeastern end of Griffin Lake.

in wave base during W_2 sedimentation resulted in the interstratification of dolostone with fine grained arkosic and pelitic siliciclastics.

The iron-rich sedimentary rocks, previously not identified in the Hawk Hill, Mountain, and Griffin lakes area, include magnetite-hematite oolitic iron-formation, magnetite-bearing carbonate, and magnetite-bearing shale-siltstone (Fig. 1A, B). These rocks are easily recognized in outcrop because the disseminated magnetite weathers in high relief; hematitic iron formation displays a fine grained clastic texture and a distinctive mottled purple-red to purple colour. They are areally extensive throughout the W_2 member in the Hawk Hill and Mountain lakes areas (Fig. 2). An apparent absence of iron-rich sedimentary rocks in the Griffin Lake area is consistent with northward thinning of the W_2 unit as suggested by Aspler and Bursey (1990). The lack of a continuous aeromagnetic signature on the west side of the synclinorium, south of Griffin Lake and north northwest of Hawk Hill Lake, may be due to higher modal hematite.

The stratigraphic thickness of iron-rich sedimentary rocks is unknown due to a lack of continous outcrop. In one area near the the southeastern end of Griffin Lake, discontinuous exposures of iron-rich rocks outcrop over a true width of 20 m. Hematitic iron-formation beds range from several millimeters to 10 cm in width. Purple magnetite-hematite iron-formation is interstratified with thin parallel-bedded, magnetite-bearing calcareous shale, siltstone, and red to reddish-orange weathering stromatolitic dolostone with or without magnetite. Minor thin beds of nonmagnetic tan to yellow weathering dolostone, cryptalgal and stromatolite dolostone, identical to the unit W_3, Watterson Formation, are interstratified with the above iron-rich sedimentary rocks.

Eade (1974) identified iron-formation in association with argillite and magnetite-bearing dolostone south and southwest of Bernier Lake. However, in the Cullaton-Ducker lakes area, iron-rich rocks in the Watterson Formation have a wider distribution than noted by Eade (1974; Fig. 2). Interstratified parallel-bedded magnetite-bearing shale, siltstone, and dolostone are similiar to the sequence in the Hawk Hill-Mountain-Griffin lakes area, but the purple magnetite-hematite iron-formation was not observed and there is also a greater proportion of magnetite-bearing dolostone to fine grained pelite. This ratio may reflect greater stability of the Watterson carbonate platform in the Cullaton-Ducker lakes area. The stratigraphic continuity of iron-rich sedimentary rocks throughout the Cullaton-Ducker

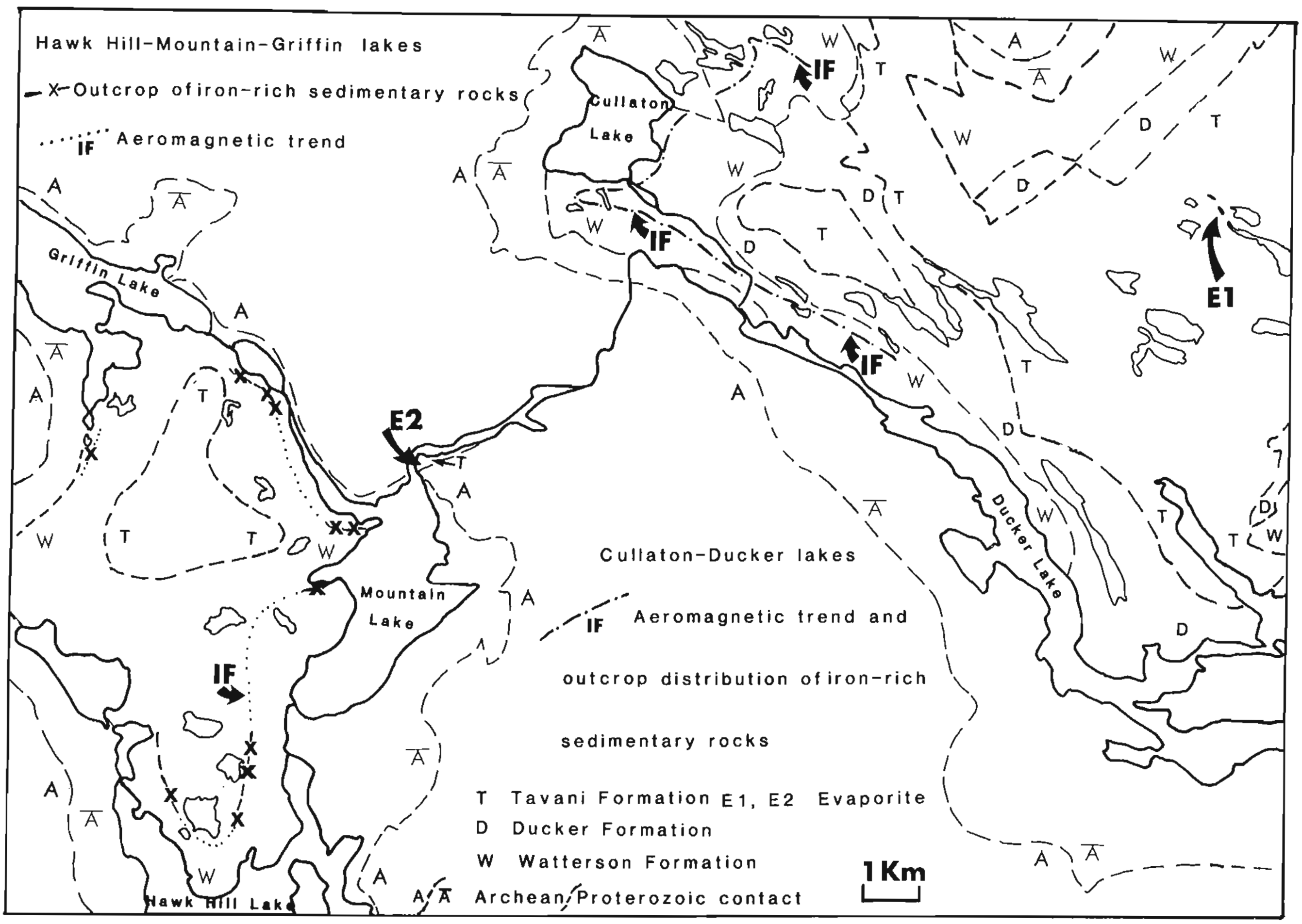


Figure 2. Distribution of iron-rich sedimentary rocks and evaporites, Hurwitz Group, Hawk Hill-Mountain-Griffin and Cullaton-Ducker lakes areas.

lakes area suggests that the Watterson Formation in this area should be divisible into members similiar to those of the Hawk Hill-Mountain-Griffin lakes area (Aspler and Bursey, 1990).

The triplet of iron-rich rocks is consistent with the paleoenvironment proposed by Aspler and Bursey (1990). Periodic deepening of the carbonate platform was associated with mass flow turbiditic sedimentation which deposited fine grained arkose, pelite, and iron-rich sedimentary rocks. The presence of oolitic textures in some iron-formation beds may indicate intervals of shoaling and reworking of iron-rich sedimentary rocks. Alternatively, oolitic textures could have a biogenic origin considering the interstratification of stromatolitic to cryptalgal dolostone with iron-rich sedimentary rocks.

Petrography of iron-rich rocks

The essential minerals of the iron-formation, quartz, opaque euhedral to subhedral magnetite, and anhedral translucent red to opaque hematite, are accompanied by minor poikiloblastic euhedral to anhedral carbonate and trace biotite/ stilpnomelane. Marked modal and textural variations characterize the finely laminated iron-formation. Some laminae contain very fine grained, <0.1 mm, equigranular quartz, disseminated euhedral magnetite, and extremely fine grained opaque to red translucent hematite. Others contain structures that are equant to round in cross-section and oblong in longitudinal section (Fig. 1C). These structures are composed of ultra fine grained quartz, possibly originally amorphous silica, and are variably replaced by translucent red hematite and carbonate. The outlines of some structures are preserved by ultra fine grained hematite even though the central areas contain more coarsely grained hematite. The matrix to these structures is dominated by a fine grained mosaic of interlocking quartz with variable magnetite, hematite, and carbonate. The ovoid structures resemble oolites that are common in Proterozoic iron-formations (Gross, 1965, Plate III). Textures and greenschist grade metamorphic assemblages in the Watterson iron-rich sedimentary rocks are similiar to textures in other Proterozoic greenschist iron-formation.

In the Cullaton Lake area, fine grained, <0.05 mm, euhedral to subhedal magnetite is typically uniformly disseminated throughout pelitic beds and units of red-brown-weathering carbonate. Magnetite in carbonate beds may form discontinuous thin laminae that parallel bedding and cleavage. Southeast of Cullaton Lake the metamorphic assemblage in magnetite-bearing calcareous pelite is sericite/muscovite, biotite, and minor chlorite and biotite replaces chlorite. The Hurwitz Group in this area lies within the biotite zone of the greenschist facies.

METALLOGENETIC IMPLICATIONS

Structurally-controlled vein-type gold deposits and occurrences are hosted in the Kinga and Watterson formations, Hurwitz Group (Miller, 1989; Aspler and Bursey, 1990). Proterozoic gold is associated with lower greenschist facies metasedimentary rocks, carbonate-bearing vein systems, carbonatization of wall rocks, and brittle rather than ductile deformation.

The Shear Lake gold deposit, a past-producer, consists of a series of subvertical pyrite-bearing fault and fracture zones hosted in Kinga quartzite. These gold-bearing structures occur above a sulphide-bearing and albitized detachment zone at or near the contact of the Hurwitz Group with Archean basement. Gold mineralization is genetically and temporally linked to development of the detachment zone, cleavage fans, metasomatism along the detachment zone, and sulphide deposition in veins (Miller, 1989). The presence of sulphates within the upper portion of pyrite-rich veins in the Shear Lake deposit are in agreement with a high-level, zoned epithermal-like system.

In the W_2 unit Watterson Formation, pyrite-gold quartz veins are hosted in calcareous arkose (Aspler and Bursey, 1990). These greenschist facies host rocks are overprinted by an alteration process that included K-, CO_2-, and S-metasomatism related to a structure that crosscuts Watterson Formation stratigraphy near the margin of the Hurwitz synclinorium (Comaplex Minerals Corp., pers. comm., 1992). In summary, the characteristics of structurally-controlled gold in the Hurwitz Group are the low grade of host metasedimentary rocks, brittle deformation, an association with reverse faults, localized near the margin of the Hurwitz synclinorium, and an association with S- and CO_2-metasomatism.

Although there are no reported occurrences of structurally-controlled gold in the Watterson Formation iron-rich rocks, the Proterozoic oxide iron-formation and associated iron-rich lithologies may represent a favourable host rock for epigenetic gold mineralization, similiar to that of the Shear Lake deposit. However, neither pyrite nor pyrrhotite have been observed in outcrop or frost heaved outcrops of bedded iron-rich rocks.

Throughout the synclinoriums in the Hawk Hill-Mountain-Griffin lakes area and in the Cullaton-Ducker lakes area, thrust and cross faults are most abundant at and near the base of the Hurwitz Group because folding was primarily by flexural slip (Aspler and Bursey, 1990; Eade, 1974; Miller, 1989). Migration of S- and CO_2-bearing hydrothermal fluids along reverse faults and escape along subvertical fault and fracture zones could provide a mechanism whereby the iron-rich rocks of W_2 member could have become altered, sulphidized, and auriferous.

TAVANI FORMATION: EVAPORITE

The Tavani Formation (Unit 16 of Eade, 1974) occurs north of Ducker Lake and southeast of Cullaton Lake. Subarkose is the dominate lithology, but Eade (1974) identified two subunits: 1) dolomite and sandy dolomite (Unit 16a), and 2) arkose with interbedded pebbly sandstone or pebble-conglomerate (Unit 16b). Both lithological units were considered by Eade (1974) to be the youngest preserved sedimentary rocks in the Tavani Formation.

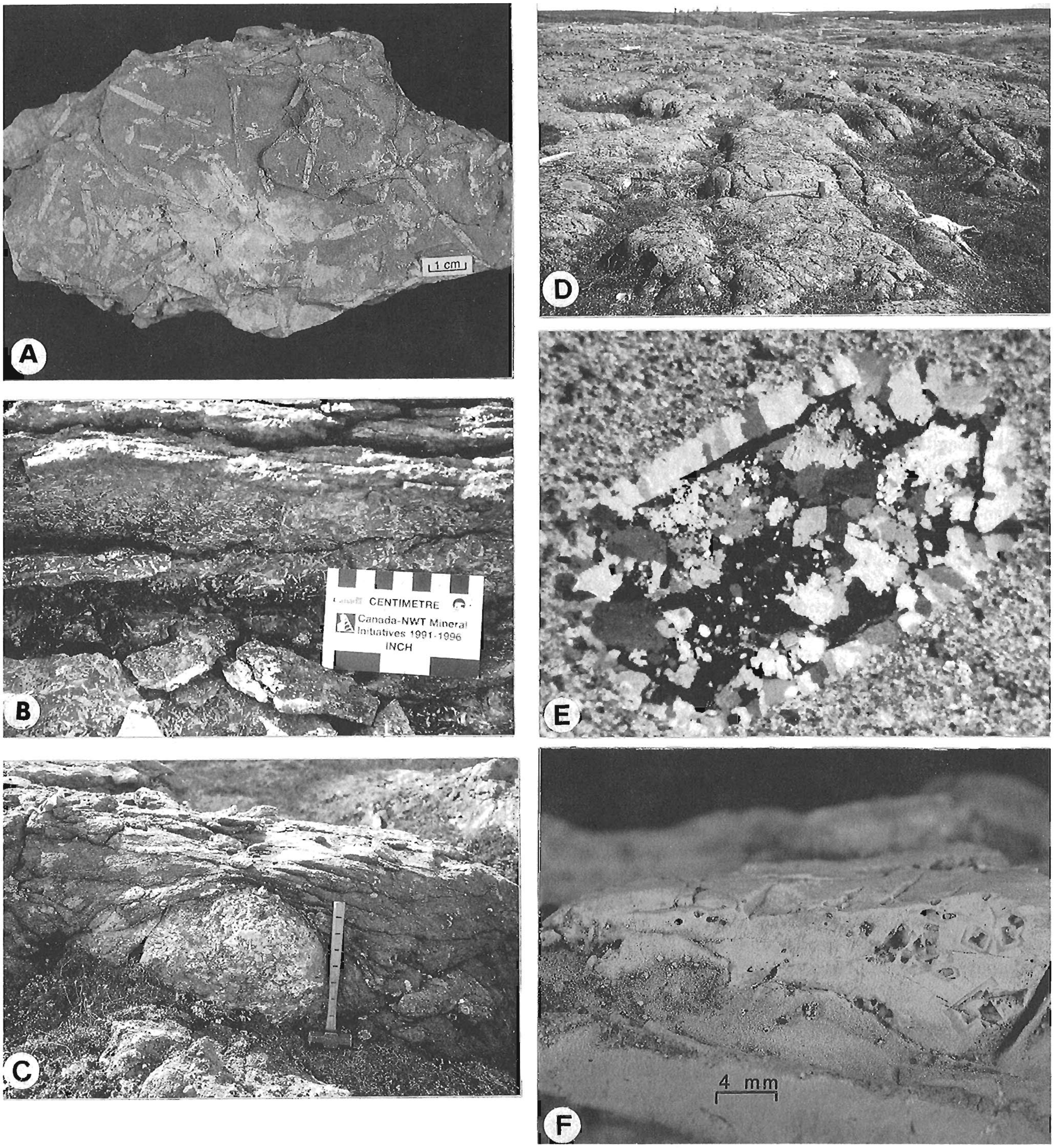


Figure 3.

A. Hand specimen of red dolostone with unoriented pseudomorphs after prismatic gypsum, Tavani Formation, locality E1.

B. Red dolostone with abundant unoriented pseudomorphs after gypsum (white blades), Tavani Formation, locality E1.

C. Bulbous stromalolite draped by pebbly muscovitic arkose. Tavani Formation, Locality E1. Subdivisions on the hammer handle are 5 cm apart.

D. Elongate stromatolite, Tavani Formation, locality E1.

E. Photomicrograph of pseudomorph after prismatic gypsum, Tavani Formation, locality E1.

F. Very fine grained siliceous dolostone with rectangular hopper-shaped casts after halite, Tavani Formation, locality E1.

During regional exploration for structurally-controlled gold in the Tavani Formation, a distinctive sequence of carbonate rocks was recognized in area E1 (Fig. 2). This area lies within Unit 16b (Eade, 1974). Following field examination of locality E1, rocks of the Tavani Formation located at the outlet of Mountain Lake, area E2 (Fig. 2). indicated that both localities have similiar lithologies and textures.

Three distinctive carbonate lithologies occur in the area of locality E1. The lateral continuity of these rocks is undetermined due to extensive till cover. The youngest unit is an interbedded sequence of tan-weathering dolostone and cryptalgal dolostone, greyish black shale, red dolostone, calcareous arkose and minor bulbous stromatolitic dolostone. These lithologies overlie two units that are dominated by different types of stromatolite. The youngest lithological package is distinguished from the other two units by the distribution and abundance of unoriented euhedral to subhedral pseudomorphs (Fig. 3A). The red, fissile, very fine grained, aluminous dolostone can contain up to 25% prismatic pseudomorphs by volume (Fig. 3B). In contrast, dense tan-coloured dolostone and shale may contain 5-10% pseudomorphs. The calcareous arkose that is interstratified with the shale and carbonate rocks is friable, granular, and has a vuggy texture. Vugs are filled with fine grained faceted crystallites. Preliminary petrography has shown that the arkose has been recrystallized and metasomatized. The clastic texture of the arkose has been overgrown by undeformed anhedral to prismatic clear albite and euhedral to subhedral medium grained carbonate.

The second lithological unit contains isolated domal stromatolites, up to 60 cm in height and 80 cm in breadth. Calcareous, pebbly, micaceous arkose fills the area between the bulbous stromatolites and commonly is draped over the stromatolites (Fig. 3C). Prismatic pseudomorphs are sparsely distributed throughout the stromatolites, but were not seen in the arkosic matrix. This unit outcrops at locality E2, but the prismatic pseudomorphs are less abundant than at locality E1. The third unit, possibly a lateral equivalent of the former unit, is comprised of elongate stromatolites (Fig. 3D). Prismatic pseudomorphs were not observed in this unit.

Petrography of evaporite

X-ray and preliminary petrographic studies indicate that the prismatic pseudomorph is a composite assemblage having variable proportions of carbonate, chlorite, quartz, and albite. Pseudomorphs up to 5 cm in long dimension can be doubly terminated, have a prismatic six-sided habit, and are rhomboid to rectangular in cross-sections (Fig. 3E). On the weathered surface, pseudomorph colours include white, green, greenish black, and red. Colour variation reflects different proportions of the above replacement minerals. Based on their morphology, these pseudomorphs are interpreted to be gypsum or gypsum after anhydrite (Adams et al., 1984). Thin bedded rose dolostone and siliceous dolostone contain rectangular hopper-shaped casts interpreted to be after halite (Fig. 3F).

The interpreted depositional environment for Tavani Formation subarkosic rocks is a shallow subaqueous to subaerial coastal plain (Aspler et al., 1992). Within this framework, the stromatolitic carbonate at locality E1 may represent a local shallow marine lagoon or tidal flat that was inundated with pebbly to medium grained immature calcareous arkosic detritus. Shoaling led to deposition of an interbedded sequence dominated by fine grained carbonate and pelitic rocks in an environment that could have been hypersaline marine or sabkha. Despite differences in metamorphic grade, these uppermost evaporitic sedimentary rocks of the Tavani Formation may be compared to probable metaevaporites within the Wollaston Group (Appleyard, 1984, 1981).

METALLOGENETIC IMPLICATIONS

Terranes that host stratiform sediment-hosted copper deposits exhibit a gradual transition in the depositional environment from subtidal to supratidal carbonates, transitional to shallow marine carbonates, and terminate with subaqueous or subaerial siliciclastic, carbonates, and evaporites. This inferred transitional depositional environment has been recognized in the Hurwitz Group (Aspler et al., 1992; Aspler and Bursey, 1990). The Tavani Formation is comprised of quartz arenite, subarkose, minor dolostone, and evaporite deposited in a shallow marine, intertidal to subaerial environment. Tavani arkosic sedimentary rocks are predominantly grey and reduced, but are in part, pink and oxidized (Eade, 1974). This diversity of fining- and shallowing- upwards of terrigenous and carbonate sediments in a cratonic depositional regime has been identified in sediment-hosted stratiform copper provinces (Kirkham, 1989). His model proposes that diagenetic oxic brines derived from evaporites will migrate along redbed aquifers, interact with rocks of the aquifer, extract ore metals, and deposit them at redox boundaries.

Although there are no reported occurrences of sediment-hosted stratiform copper in the Hurwitz Group in either the Hawk Hill-Mountain-Griffin lakes nor the Cullaton-Ducker-Bray lakes areas, the presence of evaporites, as well as an apparently favourable paleoenvironment, suggest a potential for stratiform sediment-hosted copper deposits.

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Geological Survey of Canada Project 870024

Iron-formation-hosted gold mineralization and its geological setting, Meliadine Lake area, District of Keewatin, Northwest Territories[1]

A.E. Armitage[2], S. Tella[3], and A.R. Miller[4]

Armitage, A.E., Tella, S., and Miller, A.R., 1993: Iron-formation-hosted gold mineralization and its geological setting, Meliadine Lake area, District of Keewatin, Northwest Territories: in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 187-195.

Abstract: The Meliadine Lake area is underlain by a polydeformed, greenschist grade sequence of mafic and felsic volcanic rocks, greywacke, and banded oxide facies iron-formation of the Archean Rankin Inlet Group. Syn- to posttectonic gabbro sills, granite and tonalite, and undeformed early Proterozoic lamprophyre and diabase dykes intrude the volcanic and sedimentary rocks. Visible gold with arsenopyrite, pyrrhotite, pyrite, and chalcopyrite is associated with silicification and the development of hornblende+ biotite+grunerite-bear ing assemblages in the iron-formation. The mineralization is concentrated at the hinges of pre-F_2 Z-folds. The mineralization is late syn- to posttectonic with respect to the Z-folds and is related to K, Ca, S, As, Au, and Cu metasomatism. Several smaller gold-sulphide showings, associated with brecciated quartz-ankerite±calcite veins at the contact of iron-formation with mafic volcanic rocks, are related to late brittle movements along the Pyke Fault, a reactivated pre-F_2 ductile thrust.

Résumé : Le sous-sol de la région de Meliadine Lake contient une séquence polydéformée, métamorphisée dans le faciès des schistes verts, composée de roches volcaniques mafiques et felsiques, de grauwacke, et de la formation ferrifère rubanée du Groupe de Rankin Inlet, d'âge archéen. Des filons-couches gabbroïques syntectoniques à post-tectoniques, des granites et tonalites, et des dykes non déformés de lamprophyre et de diabase, datant du Protérozoïque précoce, sont intrusifs dans les roches volcaniques et sédimentaires. De l'or visible accompagné d'arsénopyrite, de pyrrhotine, de pyrite et de chalcopyrite est associé à la silicification et à la formation d'assemblages de hornblende+biotite+grunérite dans la formation ferrifère. La minéralisation est concentrée au niveau des charnières des plis Z antérieurs à F_2. La minéralisation est syntectonique tardive à post-tectonique relativement aux plis Z, et apparentée au métasomatisme en présence de K, Ca, S, As, Au et Cu. Plusieurs indices plus petits de sulfures accompagnés d'or, associés à des filons bréchiques de quartz-ankérite±calcite au contact de la formation ferrifère avec des roches volcaniques mafiques, sont liés à des mouvements cassants survenus tardivement le long de la faille de Pyke, faille chevauchante ductile réactivée et antérieure à F_2.

[1] Contribution to Canada-Northwest Territories Mineral Initiatives 1991-1996, a subsidiary agreement under the Canada-Northwest Territories Economic Development Agreement. Project funded by the Geological Survey of Canada.

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INTRODUCTION

The Meliadine Lake area is located approximately 20 km northwest of Rankin Inlet, Northwest Territories (Fig. 1). Bedrock mapping (1:30 000 scale) during the 1992 field season covered parts of the Rankin Inlet (55K/16) and Falstaff Island (55J/13) map sheets. The objectives of this study are to describe the geology, structure, and mineralogy in and around Au-sulphide mineral showings in the Meliadine Lake area in order to provide a better tectonic framework for the mineralization and to aid future mineral exploration in this region.

Field data was processed using field-based portable PC and pocket computers with FIELDLOG (Brodaric and Fyon, 1989) and AutoCAD software applications.

REGIONAL GEOLOGY

The geology of the area surrounding Meliadine Lake has previously been investigated by Reinhardt and Chandler (1973), Reinhardt et al. (1980), Tella and Annesley (1987), Tella et al. (1986, 1989, 1990, 1992, 1993, and references therein). The region is underlain by a metamorphosed sequence of the Archean Rankin Inlet Group, comprising mafic and felsic volcanics, interflow sediments, gabbro sills, and oxide facies iron-formation, intruded by minor granite, and biotite lamprophyre and gabbro dykes of Archean and Proterozoic age. The rocks of the Rankin Inlet Group form an F_1 homocline which has been subsequently folded into a southeast-plunging F_2 syncline (Tella et al., 1986). Pre-F_2 thrusting is common within the Group. Deformation and greenschist facies metamorphism are interpreted to be Archean in age (Tella et al., 1990).

Gold mineralization in the Meliadine Lake area (Fig. 2) was discovered in the fall of 1989 (Dickson, 1991) by a two person prospecting team. Asamera Minerals Inc. in association with Comaplex Minerals Corporation, conducted extensive exploration which involved mapping (Hauseux, 1990), geophysical work, and diamond drilling during the summers of 1990-1992. They outlined a zone of Au-sulphide mineralization hosted by oxide facies iron-formation, referred to as the "Discovery" zone, and several smaller Au-sulphide showings in the area around the Discovery zone.

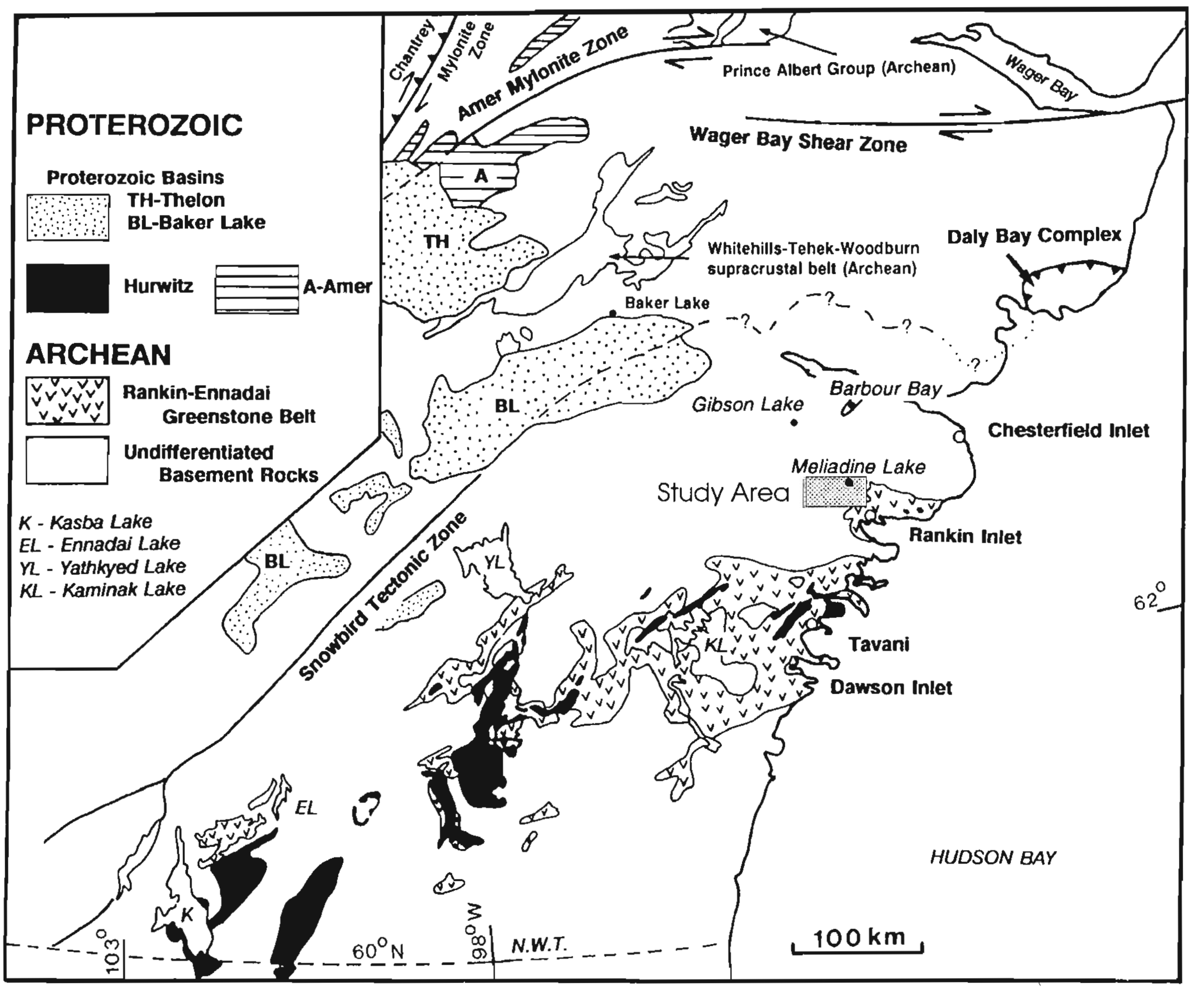

Figure 1. Map showing the location of the Meliadine Lake area, and distribution of major supracrustal belts in the District of Keewatin.

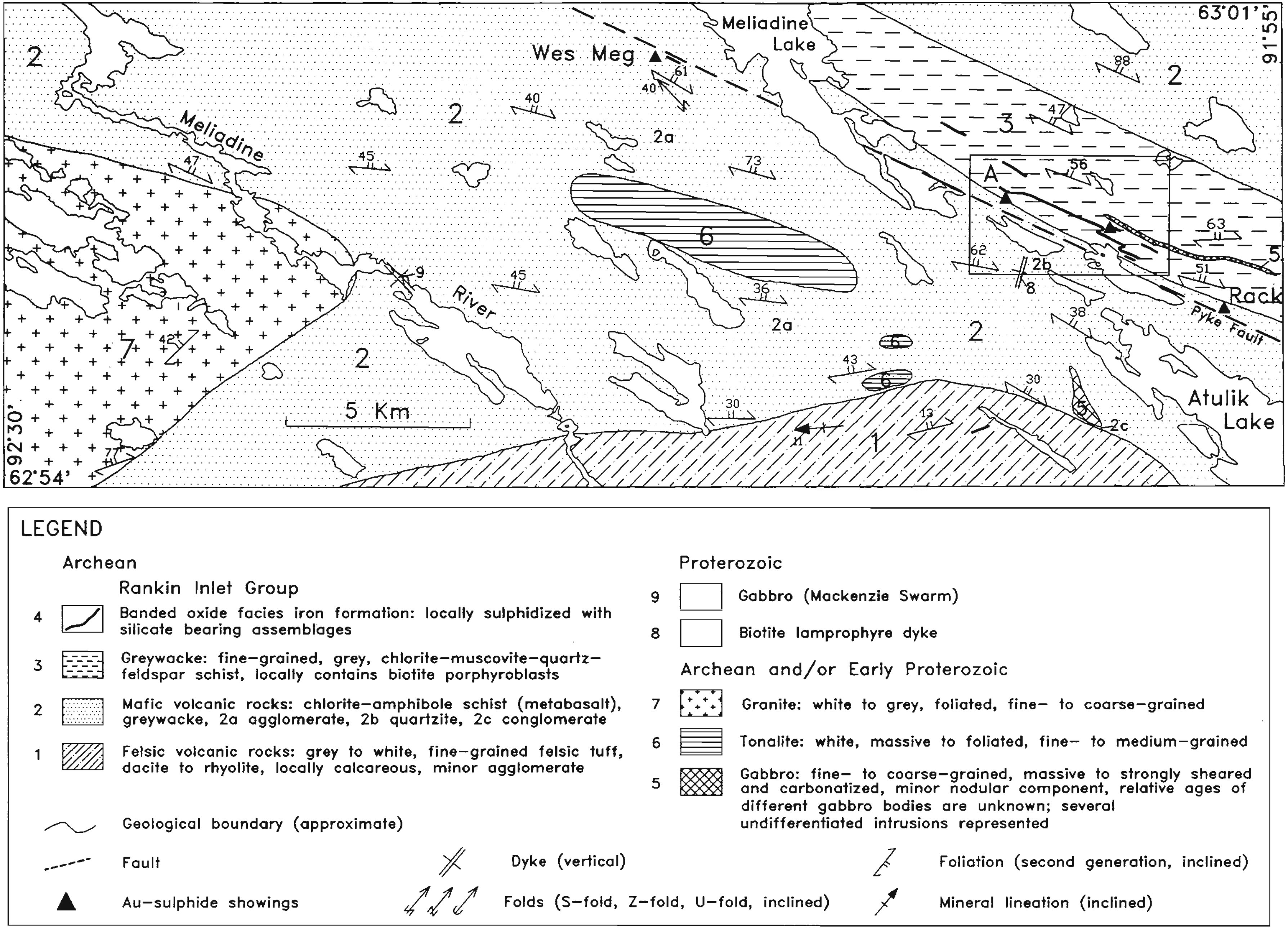

Figure 2. Geological map of the Meliadine Lake area. Area A outlines main zone of gold-sulphide mineralization hosted by banded oxide facies iron-formation (Discovery). Rack and Wes Meg (after Hauseux 1990).

GEOLOGY OF THE MELIADINE LAKE AREA

The area is underlain by an east-trending sequence of strongly deformed and metamorphosed felsic (unit 1) and mafic (unit 2) volcanics, greywacke (unit 3), and oxide facies iron-formation (unit 4), all of the Rankin Inlet Group; minor greywacke, polymictic conglomerate, quartzite, and felsic volcanics occur within the mafic unit. Syn- to posttectonic gabbro sills (unit 5), granite (unit 6), tonalite (unit 7) and late undeformed lamprophyre (unit 8) and diabase (unit 9) dykes are present throughout the region.

Felsic volcanic rocks (unit 1) consisting of dacite to rhyolite tuff and agglomerate are exposed in the southern portion of the map area (Fig. 2). The felsic tuff is light grey on fresh and weathered surfaces, fine grained, and foliated to sheared. It consists of 5-10% biotite as individual grains in a matrix of plagioclase, quartz, and minor muscovite; biotite defines the foliation. The more sheared rocks are gritty, friable, and contain abundant carbonate. Agglomerate forms relatively minor discontinuous layers in the tuff that contain clasts, up to 1 m long and 2-4 cm wide, a rock consisting of plagioclase and 10-15% amphibole in a fine grained matrix of plagioclase, quartz, and biotite. The felsic volcanic unit is conformable with the mafic volcanic unit to the north.

The mafic volcanic unit (unit 2), consists of abundant chlorite-amphibole schist and lesser amounts of tuff and agglomerate. The rocks are strongly foliated to sheared and contain lenses of quartz and carbonate with chlorite and coarse amphibole. The schist consists of fine grained chlorite, plagioclase, and carbonate with randomly oriented amphibole crystals on the foliation plane. Remnant pillow structures and flow top breccias were identified at several localities west of Meliadine Lake. West and northwest of Meliadine Lake, tuff and agglomerate (unit 2a; Fig. 3a, b), contain felsic to intermediate clasts, up to 80 cm long, in a fine grained chlorite, amphibole, plagioclase, and carbonate matrix.

A white weathering, fine grained massive quartzite (unit 2b) outcrops southeast of Meliadine Lake (Fig. 2), and lies in contact with chlorite-amphibole schist. Pods up to 2 m across consisting of coarse grained tremolite blades are present in a finer grained matrix of calcite, quartz, and phlogopite/biotite.

A polymictic conglomerate (unit 2c) is conformable with felsic volcanic rocks in the southern part of the map area (Fig. 2). It consists of pebbles and cobbles of white, unfoliated, medium grained granite and white orthoquartzite in a finer grained greywacke matrix. Isolated outcrops of conglomerate were also noted west of Meliadine Lake in the mafic volcanic unit.

Greywacke (unit 3) consisting of chlorite-muscovite-plagioclase-quartz schist is exposed east and southeast of Meliadine Lake (Fig. 2). The rocks are light to dark grey and fine grained with abundant biotite porphyroblasts giving the rock a spotted appearance. Poorly to well sorted fine grained (<1 mm) quartzofeldspathic wackes dominate this unit. Polygranular quartz and quartz+feldspar aggregates are interpreted to have been lithic clasts derived from a felsic volcanic protolith. Fine grained quartzofeldspathic wackes grade into aluminous interbeds of semipelitic to pelitic composition, but bedding-parallel transposition and recrystallization has commonly obliterated primary sedimentological features. The metamorphic assemblage in each compositional layer of the turbidite is quartz+plagioclase+muscovite+biotite±chlorite.

East of Meliadine and Atulik lakes (Fig. 2), banded oxide facies iron-formation (BIF) (unit 4), hosted by greywacke (unit 3), forms several more or less continuous layers, 10-30 m thick, with a pronounced aeromagnetic expression (Geological Survey of Canada, 1966a, b, c). Multiple layers of banded iron-formation in the greywacke sequence may in part represent repetition of one or more layers due to regional polyphase folding as evidenced by refolded folds (Tella et al., 1992). Compositional-sedimentological differences between the iron-formation bands in the Discovery area suggest that some multiple layers of iron-rich sediments were formed

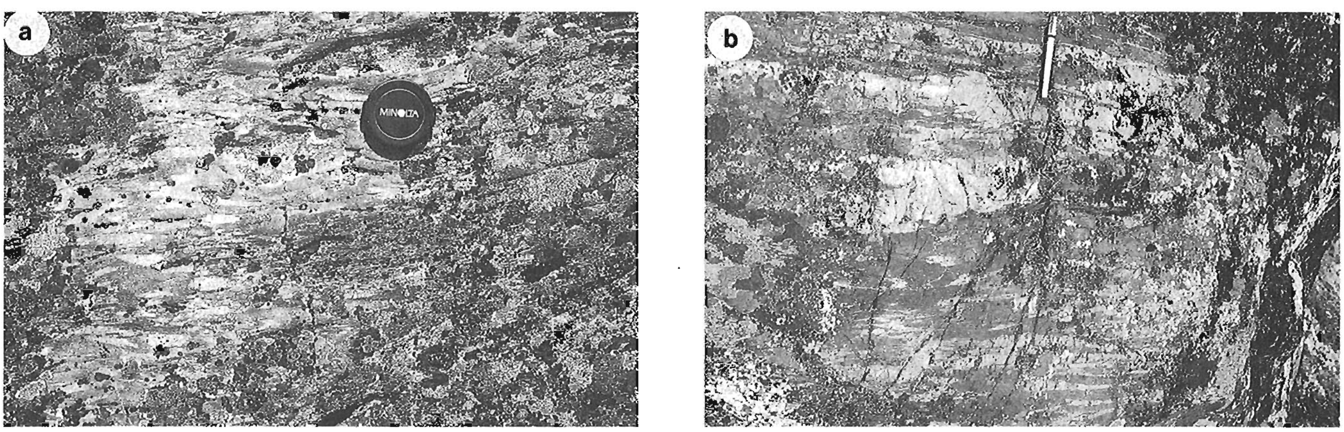

Figure 3. (**a**) Tuff and (**b**) agglomerate (unit 2a) from the mafic volcanics (unit 2) west of Meliadine Lake. GSC 1992-253Z, 253Y

during intervals of low or negligible clastic sediment influx into the basin. The iron-formation that hosts Discovery zone gold-sulphide mineralization is characterized by thin (up to 4 mm thick) banding, formed by varying proportions of quartz (recrystallized chert) and magnetite. The bands in turn are interlayered with chlorite-rich bands which have modal proportions (up to <20%) of accessory quartz and magnetite. Numerous lenticular wedges of fine grained quartzofeldspathic wacke are present in the Discovery iron-formation. In contrast, the iron-formation north of the Discovery zone displays a finer rhythmic alternation of quartz- and magnetite-rich laminae, with relatively thin inter-layers of chlorite and of sparse quartzofeldspathic sediment.

West of Atulik Lake (Fig. 2), the banded oxide facies iron-formation consists predominantly of magnetite and quartz with discontinuous 5 to 20 cm thick layers of 2-6 mm garnets, hornblende, chlorite, grunerite, and minor carbonate and pyrite. Several minor outcrops of banded oxide facies iron-formation occur along the trace of the Pyke Fault in the western part of the map area.

Gabbro sills and dykes are common in the volcanic and sedimentary units and some are shown on the map (Fig. 2). They are dark green to black, massive units with strongly foliated and sheared margins. The least deformed intrusions contain fine- to coarse-grained hornblende/pyroxene, plagioclase, and chlorite; some gabbro bodies contain 0.6 to 2 cm plagioclase phenocrysts, which may represent 20% of the gabbro. Altered and sheared equivalents contain carbonate+biotite+chlorite+epidote+albite. The relative ages of the various gabbro intrusions are unknown. West of Meliadine Lake (Fig. 2), several gabbro sills in the mafic volcanics contain glomeroporphyritic plagioclase nodules 1-20 cm in diameter, some of which are elongated parallel to the foliation. Biotite occurs at the margins of these nodules and chalcopyrite in trace amounts in the groundmass.

Southwest of Meliadine Lake (Fig. 2), the mafic volcanic rocks (unit 2) are intruded by tonalite (unit 6) which form an elongate east-trending body and smaller plugs. The tonalite is white to light grey, fine- to medium-grained, strongly to weakly foliated, and consists of plagioclase and quartz, 2-5% disseminated biotite, and traces of magnetite and pyrite. The tonalite is intruded by coarse grained massive gabbro (unit 5).

Grey to white, weakly to strongly foliated, fine- to medium-grained granite (unit 7) is exposed west of the Meliadine River (Fig. 2). It consists of 2-15% biotite with quartz, plagioclase, and minor K-feldspar and is cut by massive, up to 3 m thick salmon pink granite and pegmatite dykes. The dykes contain 2-3% biotite with quartz, K-feldspar, plagioclase, and trace magnetite. A 1 m wide, massive, white granite dyke, which may be related to unit 7 cuts mafic volcanic rocks near the northern reaches of the Meliadine River. The dyke rock contains garnets (1-2 mm in diameter) which are hosted in a fine- to medium-grained quartz, plagioclase, K-feldspar, and biotite matrix.

A 2 m wide, 195° trending, biotite lamprophyre dyke (unit 8) cuts the quartzite unit south of Meliadine Lake (Fig. 2). The lamprophyre contains medium grained biotite/phlogopite phenocrysts in a finer grained dark brown matrix. Biotite lamprophyres were also encountered in drill core from the banded oxide facies iron-formation-hosted mineralized zone. The lack of deformation in the dyke suggests that it is a part of the lamprophyre dyke swarm noted by Tella et al. (1986) in the Rankin Inlet region, and may be related to the 1.85 Ga alkaline igneous suite in the central Keewatin (Lecheminant et al., 1987).

In the central part of the map area, along the shore of the Meliadine River (Fig. 2), a northwest-trending fine grained gabbro dyke (unit 9), related to the 1267 ± 2 Ma (LeCheminant and Heaman, 1989) Mackenzie dyke swarm, intrudes the mafic volcanic rocks.

Approximately 7 km northeast of Meliadine Lake, a 10 m wide fine grained diabase dyke trends 345° and cuts the sheared mafic volcanic rocks. It contains 1-8 cm long plagioclase phenocrysts and is texturally similar to the Early Proterozoic Kaminak dykes (Davidson, 1970) that provide an upper limit to the age of penetrative deformation in the Meliadine Lake area.

STRUCTURE

In the Meliadine Lake area, Rankin Inlet Group rocks represent a portion of the northern limb of a regional F_2 fold. Several limb-parallel, ductile, high-strain zones are interpreted as pre-F_2 thrust faults that show oblique-dextral north-side-down sense of displacement (Tella et al., 1992). Two sets of planar fabrics (S_1/S_2) (Fig. 4a) at an acute angle (15°-20°) to each other were identified in the volcanic and sedimentary sequences. The northward dip of both fabrics changes from moderate to steep dipping in the north and central parts (Fig. 2), becoming shallow in the southern part of the map area. Both fabrics contain down-dip quartz and carbonate rodding lineations (Fig. 5), which suggest layer-parallel slip.

Several sets of pre-F_2 folds are best preserved in the banded oxide facies iron-formation (Fig. 6), but are rarely developed in the volcanic and sedimentary sequences. Coaxial pre-F_2 fold structures show interference patterns (Tella et al., 1992, fig. 6a). They have shallow to moderate plunges to the east (Fig. 4b) with local reversals to the west-northwest. The youngest set of pre-F_2 folds with Z-asymmetry (Tella et al., 1992, fig. 6b) plunge moderately (30°-45°) to the northeast (Fig. 4b), rarely to the northwest; a locally well developed axial planar cleavage is present in the Z-folds (Fig. 7). The Z-fold axes are parallel to a northeast-trending, regionally pervasive, pre-F_2 mineral stretching lineation. Several folds with S-asymmetry (Fig. 8) plunge moderately to the west-northwest and are identified in banded oxide facies iron-formation in the north (Fig. 2) and central parts of the map area. Pre-F_2 folding and thrusting appear to be synchronous with the development of sheath folds in the banded oxide facies iron-formation. Subsequent modification of the sheath folds have resulted in doubly-plunging fold axes and are reflected in the discontinuous aeromagnetic signatures (Geological Survey of Canada, 1966a, b, c) displayed by the banded oxide facies iron-formation. Shallow east- and west-plunging biotite mineral

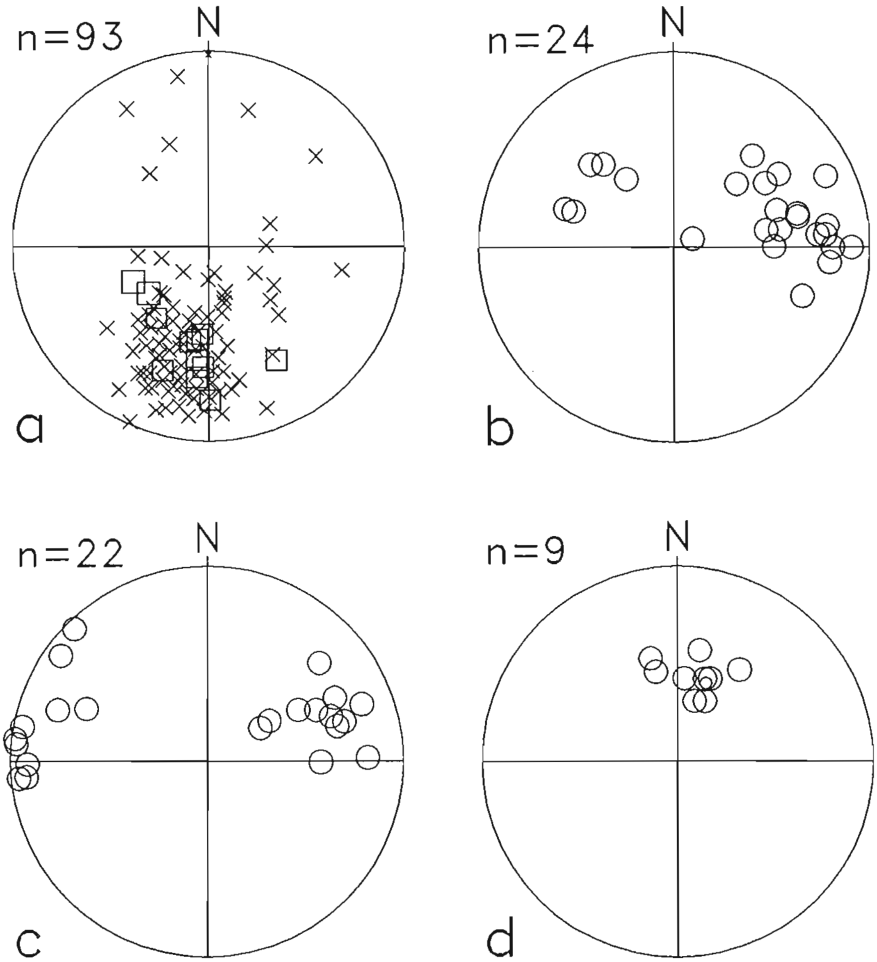

Figure 4. Equal area stereoplot of planar and linear fabrics from the Meliadine Lake area: **a)** poles to S_1 (open square) and S_2 (x) foliations, **b)** F_2 fold axes, **c)** pre-F_2 mineral and crenulation lineations, and **d)** crenulations associated with late movement on the Pyke Fault.

lineations and quartz rodding (Fig. 4c) and northeast- and northwest-plunging crenulation lineations are related to pre-F_2 deformation. A late northwest-trending brittle fault, the Pyke Fault (Hauseux 1990), transects the eastern part of the map area (Fig. 2) and appears to represent reactivation of an earlier ductile thrust. Moderate north-plunging crenulations (Fig. 4d) on the S_2 plane may be related to movement on this fault.

MINERALIZATION AND ALTERATION

The principal exploration target in the Meliadine Lake area is gold-sulphide mineralization in altered oxide iron-formation. The main gold zone ("Discovery", Hauseux, 1990) is located in the eastern part of the map area (Fig. 2). Visible gold is associated with pyrrhotite, arsenopyrite, pyrite, and minor chalcopyrite. The mineralization is concentrated at the hinges of pre-F_2 Z-folds (Fig. 5), which appear to have acted as structural traps for the mineralizing fluids carrying the mineralization. Textures indicate the

Figure 5. Steeply north-plunging quartz rodding lineation on S_2 plane in banded oxide facies iron-formation (unit 4) from the northern part of the map area. GSC 1992-253J

mineralization is late syn- to post-deformation with respect to the Z-folds. Within the mineralized zone, banding in the banded oxide facies iron-formation is sheared, discontinuous, and quartz veins cut parallel the composition layering in iron-formation (Fig. 9). Pyrrhotite is the dominant sulphide phase in the hinges of Z-folds where it occurs with coarse grained euhedral porphyroblasts (up to 1 cm) of arsenopyrite. Arsenopyrite occurs in quartz veins, as metasomatic haloes around quartz veins and as disseminated grains in sheared and altered banded oxide facies iron-formation. Chalcopyrite is invariably associated with pyrrhotite and both phases may partially or totally replace magnetite layers. Late fine grained pyrite is a minor but critical phase and is present in either pyrrhotite-rich or arsenopyrite+pyrrhotite assemblages. Pyrite stability signifies that higher sulphur activities were attained in a portion of the altered banded oxide facies iron-formation. Away from the Z-fold hinges, magnetite-quartz banding is more continuous and pyrrhotite with coarse

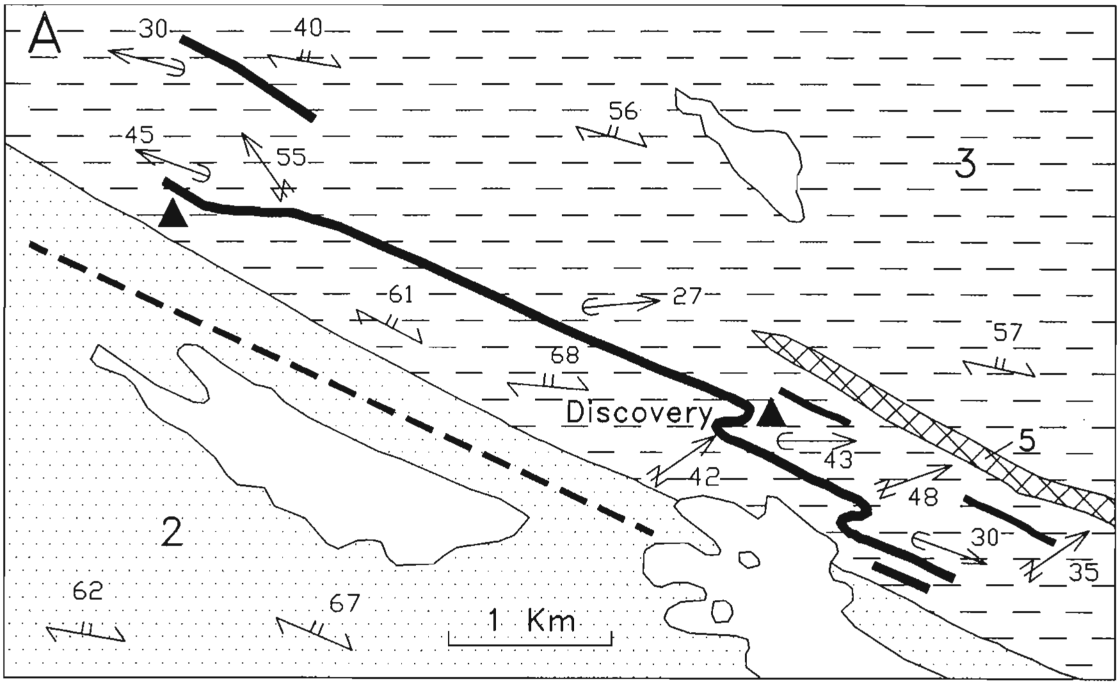

Figure 6. Enlarged map of area A on Figure 2, showing the main areas of Au-sulphide mineralization, including Discovery, and main structural features in the banded oxide facies iron-formation. (see legend Figure 2) GSC 1992-253F

Figure 7. Straight banded magnetite+quartz iron-formation on long limb of a Z-fold. Axial planes of small-scale Z-Folds are subparallel to fine quartz ribbons (Q) developed in an attenuated altered (hornblende) silicate iron-formation (S); Discovery zone.

Figure 8. Moderately northwest-plunging S-folds in banded oxide facies iron-formation (unit 4) from the northern part of the map area. GSC 1992-253F

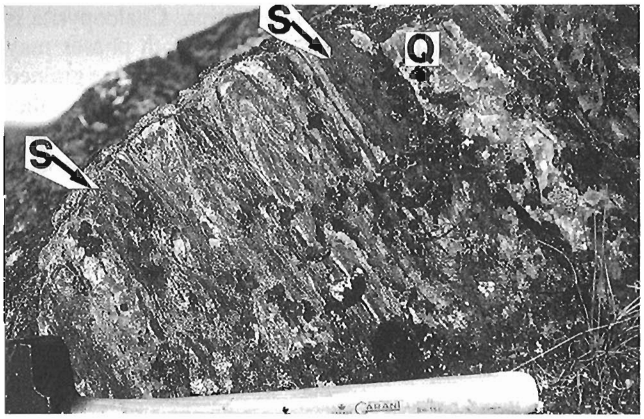

Figure 9. North-dipping tight isoclinal folds in banded magnetite iron-formation with subordinate altered silicate iron-formation (S). Quartz vein (Q) parallels the compositional layering in the iron formation; Discovery zone, hammer handle is 400 cm long.

grained arsenopyrite extends for 20-30 m along the limbs. Gold values largely correlate with arsenopyrite concentrations. Alteration in the iron-formation includes silicification, in the form of quartz veins and quartz replacement, and the development of hornblende+biotite+grunerite+calcite-bearing assemblages replacing chlorite bands; garnet was identified at one locality. Randomly oriented medium- to coarse-grained hornblende and biotite typically occur in chlorite-rich bands where they form up to 40% of these bands. Grunerite is present as rims or cores to hornblende, or as individual grains associated with calcite at magnetite-quartz layer interfaces. The intensity of alteration decreases away from the Z-fold hinges.

North and south of the Discovery zone gold mineralization, the turbidite sequence is compositionally layered but sustained variable transposition during pre-F2 folding and the development of biotite porphyroblasts appear to be related to the mineralizing event. Deformation and alteration textures in northeast-trending gabbro dykes, and subparallel orientation of dykes to Z-fold axial planes, in the greywacke-iron-formation sequence suggest that some of the dykes are coeval with Z-fold development. Chilled contacts are highly strained and altered. Potassium and carbon dioxide alteration may be profound along highly strained contacts, and may extend into the core of the intrusions, possibly due to a variation of the metasomatic process that affected the iron-formation in the Discovery zone.

Several smaller Au-sulphide showings are associated with brittle deformation along and near the Pyke Fault (Fig. 2). Two such showings are referred to as "Wes Meg" and "Rack" (Hauseux 1990; Fig. 2). Gold is associated with medium- to coarse-grained arsenopyrite with fine- to coarse-grained pyrite, chalcopyrite, pyrrhotite, and galena. The sulphide assemblage is hosted by brecciated quartz+calcite+ankerite±chlorite veins, commonly on the margins of banded oxide facies iron-formation. The alteration assemblage contrasts with banded oxide facies iron-formation-hosted gold zones. Several pyrite±pyrrhotite and arsenopyrite gossans occur along the extent of the Pyke Fault.

Pyritic gossans are common but sporadic in the mafic volcanic rocks (unit 2) and gabbro (unit 5). Some contain minor pyrrhotite and chalcopyrite.

CONDITIONS OF METAMORPHISM AND MINERALIZATION

The mineralogy of the Rankin Inlet Group volcanic and sedimentary rocks in the Meliadine Lake area indicate biotite-zone greenschist facies metamorphism. To the south in the Rankin Inlet area (Tella et al., 1986), lower greenschist facies conditions prevail, whereas to the north of Meliadine Lake metamorphism increases to the amphibolite facies (Tella et al., 1992). The gold-sulphide mineralization and related silicate alteration assemblages are confined to greenschist facies rocks. The alteration minerals indicate an epigenetic origin related to K, Ca, S, As, Au, and Cu metasomatism, which overprinted regional metamorphism. These fluids may have affected the greywacke and gabbroic rocks in this area. The sulphide mineralization associated with brittle movement on the Pyke Fault is similar to the banded oxide facies iron-formation-hosted gold mineralization (both characterized by coarse grained arsenopyrite). However, the alteration assemblage is different. This may suggest a lower temperature continuation of the same event or remobilization of the banded oxide facies iron-formation-hosted mineralization.

Garnet occurs at several localities in the Meliadine Lake area. In mafic volcanic rocks adjacent to the tonalite intrusion, west of Meliadine Lake, it may be a product of contact metamorphism. However, west of Atulik Lake, abundant garnet is associated with several lithological units and diverse mineral assemblages: garnet+hornblende+magnetite in quartz veins, garnet+biotite in felsic tuff, and garnet+hornblende in banded oxide facies iron-formation adjacent to these veins. The proximity of garnet to quartz veins and the absence of garnet from felsic volcanic rocks and banded oxide facies iron-formation suggests a hydrothermal origin rather than regional metamorphism. Rare garnet associated with the alteration assemblage in mineralized banded oxide facies iron-formation in the Discovery area is the product of hydrothermal alteration.

SUMMARY

1. The Meliadine Lake area is underlain by polydeformed mafic and felsic volcanic rocks, greywacke, and banded oxide facies iron-formation, all belonging to the Archean Rankin Inlet Group. These rocks are intruded by syn- to posttectonic gabbro and granite, and late undeformed Proterozoic diabase and lamprophyre dykes. The Rankin Inlet Group has been metamorphosed under greenschist facies conditions. Garnet identified at several localities in the study area appears to be the result of contact metamorphism and/or hydrothermal alteration.

2. The gold mineralization within the banded oxide facies iron-formation is localized at the hinges of the youngest set of pre-F_2 folds with Z-asymmetry, and appears to be late syn- to posttectonic with respect to the development of these folds. The epigenetic gold mineralization is the result of the action of metasomatic fluids carrying K, Ca, S, As, Au, and Cu.

3. Gold mineralization associated with late brittle movement on the Pyke Fault may be a lower temperature continuation of the same event or may represent a later remobilization of the banded oxide facies iron-formation-hosted mineralization.

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Geological Survey of Canada Project 850002KF

Precambrian geology and economic potential of the northeastern parts of Gibson Lake map area, District of Keewatin, Northwest Territories[1]

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Tella, S., Schau, M., Armitage, A.E., and Loney, B.C., 1993: Precambrian geology and economic potential of the northeastern parts of Gibson Lake map area, District of Keewatin, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 197-208.

Abstract: Archean lithologies in the northern portions of the area are dominated by the Uvauk complex - a layered gabbro-mafic granulite-anorthosite suite interlayered with quartzo-feldspathic gneisses deformed and metamorphosed at mid-crustal levels. The complex forms a rootless, east-northeast-trending triangular segment (30x20 km) of straight gneisses, the boundaries of which are marked by ultramylonites. It is interpreted as an allochthonous remnant of a granulite grade ductile strain zone that overlies lower grade autochthonous gneisses, and may be linked with other lithologically and tectonically similar high-strain zones: Kramanituar to the west, Hanbury Island and Daly Bay to the east.

South of Uvauk complex, polydeformed Archean quartzo-feldspathic gneisses characterized by folded high-strain zones, are interlayered with discontinous belts of amphibolite grade metavolcanic and metasedimentary rocks. Numerous gossans, some containing arsenopyrite, within the belts are potential base-metal exploration targets.

Granite and gabbro plutons, and lamprophyre dyke-swarms record Proterozoic igneous events.

Résumé : Dans les parties septentrionales de la région, les lithologies archéennes sont principalement représentées par le complexe d'Uvauk, à savoir une série stratifiée à gabbro, granulite mafique et anorthosite, interstratifiée avec des gneiss quartzo-feldspathiques déformés et métamorphisés aux niveaux intermédiaires de la croûte. Ce complexe forme un segment triangulaire sans racines, de direction générale est-nord-est (30x20 km), constitué simplement de gneiss, dont les limites sont marquées par des ultramylonites. Il s'agirait d'un vestige allochtone d'une zone de déformation ductile dans le faciès des granulites, recouvrant des gneiss autochtones moins métamorphisés, et peut-être associée à d'autres zones intensément déformées, qui seraient lithologiquement et tectoniquement semblables: Kramanituar à l'ouest, Hanbury Island et Daly Bay à l'est.

Au sud du complexe d'Uvauk, des gneiss quartzo-feldspathiques polydéformés d'âge archéen, caractérisés par des zones de plissement intensément déformées, sont interstratifiés avec des zones discontinues de roches métavolcaniques et métasédimentaires métamorphisées dans le faciès des amphibolites. De nombreux chapeaux de fer, dont certains contiennent de l'arsénopyrite, et qui se situent dans ces zones, sont des cibles possibles de travaux de prospection des métaux communs.

Des plutons granitiques et gabbroïques ainsi que des essaims de dykes lamprophyriques témoignent des épisodes protérozoïques d'activité magmatique.

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INTRODUCTION

This report summarizes preliminary results of bedrock mapping completed at a scale of 1:250 000 during the 1992 field season in the northeastern parts of Gibson Lake (55N/9,10,15,16) map area. The objectives of this study are to upgrade the reconnaissance database, to address structural and metamorphic problems, to assess the economic potential of the region, and to provide additional framework for regional correlation and tectonic synthesis of rock units in this part of the Churchill Structural Province. The study area (Fig. 1) is part of a region previously mapped at a scale of one inch to eight miles (Wright, 1967). The results of more recent work in the adjoining portions around the study area were reported by Reinhardt et al. (1980), Schau and Ashton (1980), Sanborn-Barrie (1993), Schau et al. (1982), Tella and Annesley (1987,1988), and Tella et al (1986,1989, 1992, references therein). Portable field computers and Global Positioning System (GPS) were used for data input and processing.

Acknowledgments

We extend our thanks to the following: N. Brown, G. DeSchutter, A.M. MacDonald, and M.N. Nevills for field assistance, and R. Gronas for excellent cooking; B. Brodaric, Continental Geoscience Division, for assistance with Fieldlog (Brodaric and Fyon, 1989) and AutoCad software

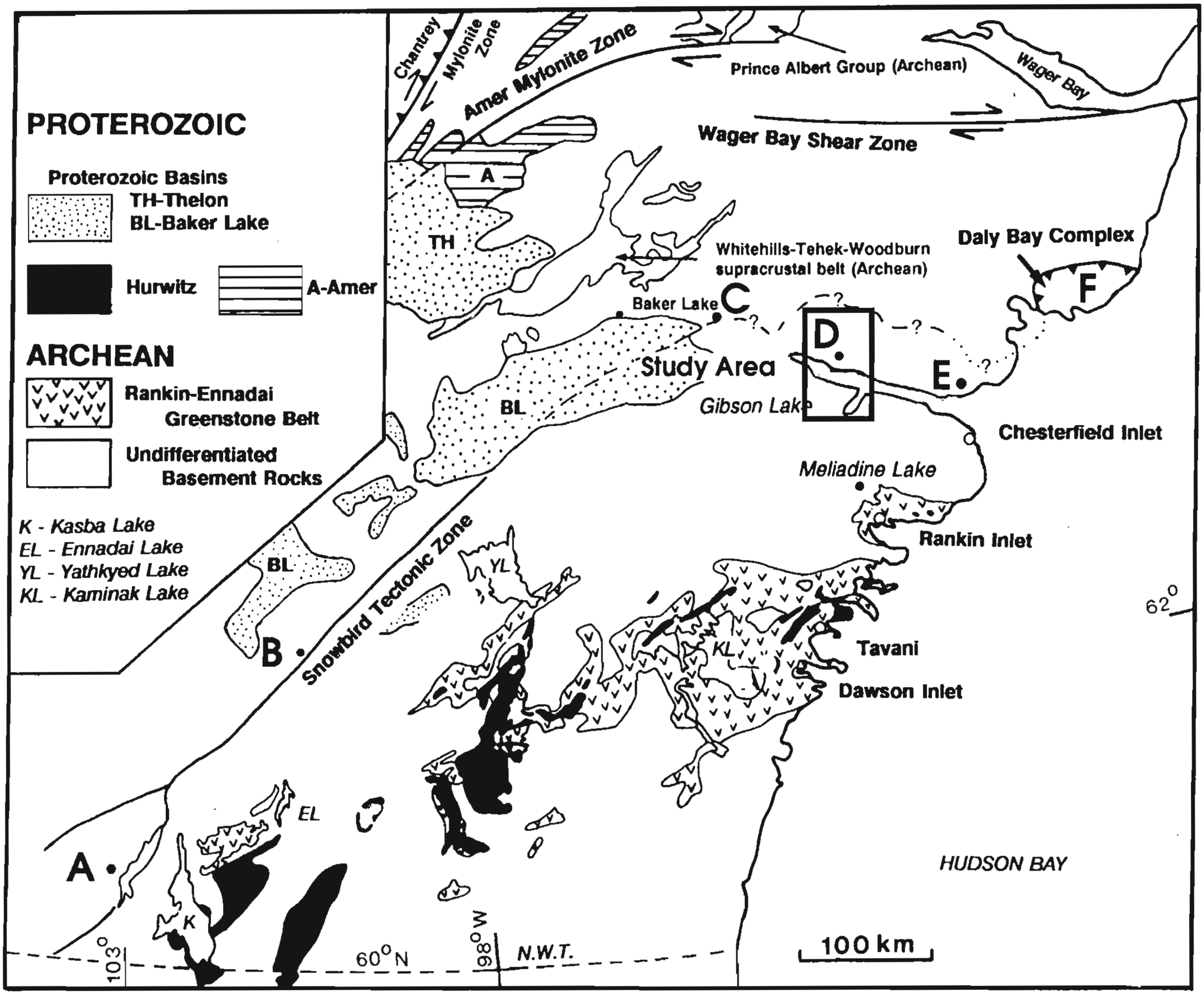


Figure 1. Sketch map showing location of the study area, and the distribution of major supracrustal belts and shear zones. A- Athabasca-Three Esker segment of the Snowbird Tectonic Zone (Hanmer and Kopf, 1993); B- High P-T granulites, Kamilukuak Lake area (Tella and Eade, 1986), C- Kramanituar Complex (Schau et al., 1982; Sanborn-Barrie, 1993), D- Uvauk Complex, this study; E- Hanbury Island Shear Zone (Tella and Annesley, 1988), F- Daly Bay Complex (Gordon,1988). Trace of Snowbird Tectonic Zone (after Hoffman, 1988).

applications; Bert Struik and coworkers at the Cordilleran Division, Vancouver for the use of the GEOF(Geological Editor Of Fieldnotes) program to facilitate data input on the outcrop using pocket PCs; the Polar Continental Shelf Project for the helicopter support; John French, Bob Burton, and Kevin Bond - our helicopter crew; M&T Enterprises Ltd., Rankin Inlet, for expediting services. Discussions and exchange of ideas on regional tectonic and metamorphic framework with T.M. Gordon, Simon Hanmer, U. Mader, and J.C. Roddick during their field visits were most valuable. J.C. Roddick and T.M. Gordon critically reviewed the manuscript.

LITHOLOGY, STRUCTURE, AND METAMORPHISM

General statement

The distribution of rock units and their regional structural framework are shown in Figures 2 and 3 respectively, and the orientations of planar and linear fabric elements are shown in Figure 4. The northern parts of the region north of Chesterfield Inlet (Fig. 2, 3), is underlain by a polydeformed and metamorphosed, allochthonous suite of rocks, the Uvauk complex, consisting of layered gabbro-mafic granulite-anorthosite (unit 1). The autochthonous rocks, distributed throughout the region, comprise relatively lower grade metavolcanics (unit 2), metasedimentary gneisses (unit 3), and quartzofeldspathic granitoid and migmatitic rocks (units 4,5). Several generations of relatively undeformed mafic intrusions (gabbro, diabase, pyroxenite, ultramafics; units 6,8,9) of Archean and/or Proterozoic age are exposed as plugs and dykes throughout the region. Post-tectonic Early Proterozoic igneous activity is recorded by several phases of granite intrusions (units 7,10,12) and by two sets of biotite lamprophyre dykes (unit 11,13). Bedrock exposures within units 8,11, and 13 are too small to be represented on the accompanying maps. Several east- and northeast-trending ductile high-strain zones (Archean ?), and at least three (east-, northeast-, and northwest-trending) sets of late (Proterozoic?) brittle faults transect the region.

UVAUK COMPLEX (unit 1)

The complex, named after Uvauk Inlet on the north shore of Chesterfield Inlet, consists of an Archean layered gabbro-mafic granulite-anorthosite suite (Fig. 2, 3, 6) interlayered with quartzofeldspathic granulites which were deformed and metamorphosed at mid-crustal levels. The Uvauk complex forms a rootless, east-northeast-trending triangular segment(30x20 km) of ductilely deformed granulite grade straight gneisses (Fig. 7), the boundaries of which are marked by well developed ultramylonites (Fig. 8) formed under mid-crustal levels. The overall geometry of the complex is that of a wrinkled sheet which overlies an amphibolite grade quartzofeldspathic gneiss terrane. The protoliths within the segment include mafic granulites, anorthosite, gabbroic anorthosite (Fig. 9A, B), and minor diatexite. The eastern margin of the segment is defined by shallow (10°-30°), west-dipping granulite-anorthosite ultramylonites; the northern and southern margins converge and terminate to the west near Uvauk Inlet where the mylonitic fabric dips (70°-85°) towards south. A shallow (10°-30°) east-northeast and west-southwest mineral stretching lineation (Fig. 4) is well developed throughout the entire length of the complex although steep plunges are noted around stiff mafic inclusions within anorthositic layers that record heterogenous ductile flow (Fig. 10, 11) at the western extremity. Rarely developed kinematic indicators suggest dextral and sinistral senses of shear for the northern and southern margins respectively.

The complex shows a long protracted history. The protolith was in part a layered and well differentiated mafic intrusive body which was subsequently deformed under granulite facies conditions. Metamorphic mineral growth and ductile deformation were episodic and intermixed as is well demonstrated by the pull-aparts of a mega-garnet segregation (Fig. 12) showing a complex tectonometamorphic history at mid-crustal levels. The high pressure and temperature mineral assemblages and the preservation of high grade ultramylonitic fabrics require extremely dry conditions and/or rapid uplift to quench the mineral textures. Annealed quartz and plagioclase ribbon-ultramylonites within the complex, on the other hand, record post-tectonic recrystallization involving release of stored strain-energy. A number of Proterozoic? gabbro, granite, and lamprophyre intrusions cut the Uvauk complex.

Figure 5 is a detailed sketch map of a portion of the Uvauk complex at the southeastern contact with the underlying autochthonous quartzofeldspathic gneisses (unit 4). The undulating nature of a relatively flat enveloping surface of the complex is well illustrated; as is a local heterogeneity manifested as a mesoscopic fold in a gabbro layer. A well developed shallow, doubly-plunging mineral lineation is present in the quartzofeldspathic gneisses (unit 4). Open, cross-folds with N-S trending axial planes overprint early structures and contribute in part to the undulating geometry of the enveloping surfaces.

At least four sets of dykes cut through the region after the tectonic emplacement of the Uvauk complex. The relationships between deformation and emplacement of various dyke-sets are well displayed in Figure 5. Northwest of Robin Hood Bay, the Uvauk Complex is cut by a set of undeformed granodiorite (unit 5) dykes and sills, whereas granitoid dykes in the underlying quartzofeldspathic gneisses (unit 4) are highly strained and sheared. Pods of amphibolitic metagabbro are found as tectonic inclusions in the quartzofeldspathic gneisses for several kilometres south of the southern margin of the Uvauk complex. They may represent dismembered parts of the Uvauk complex, and suggest that the underlying gneisses were deformed in part after the tectonic emplacement of the Uvauk complex.

One set of hornblende porphyry dykes (up to 1 m wide) cut the Uvauk complex with a planar configuration, but they cut the underlying folded quartzofeldspathic gneisses discordantly, suggesting that dyke emplacement was perhaps

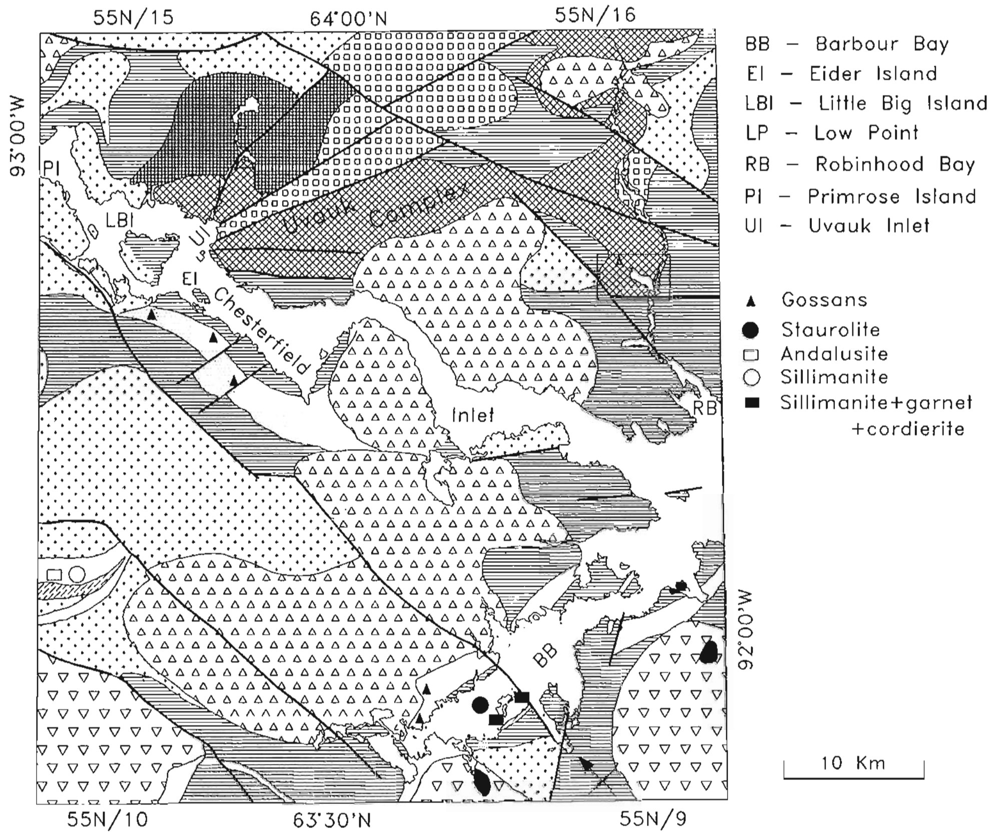
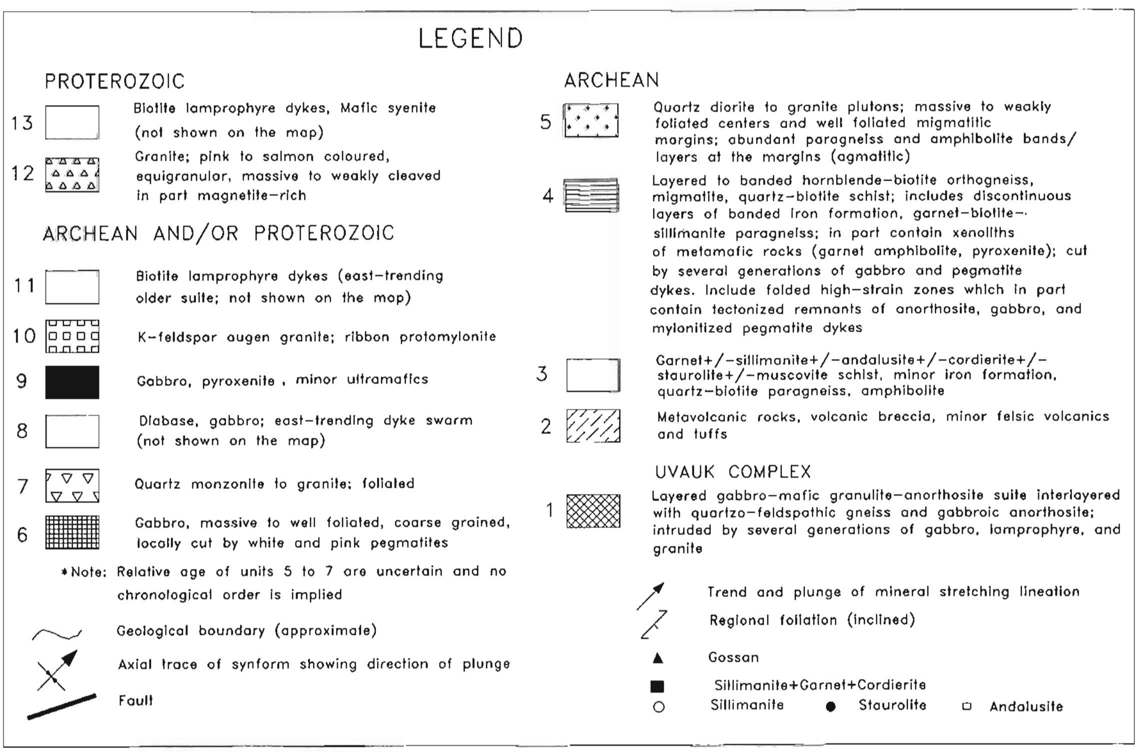


Figure 2. Simplified geological sketch map of the northeastern parts of the Gibson Lake (NTS 55N) map area. Area A refers to Figure 5.



Legend for Fig. 2 and 3.

coeval with the tectonic emplacement of the Uvauk complex. The dykes are light grey to brown and contain hornblende phenocrysts set in a fine grained quartzofeldspathic matrix.

Two sets of lamprophyre dykes (units 11, 13) – one trending 110° and the second trending 130°-140° – cut both the Uvauk complex and the underlying gneisses. The earlier set (unit 11) shows phenocrysts of olivine, clinopyroxene and orthopyroxene, and may be related to the mafic plugs exposed on Little Big Island (Fig. 2, unit 9). The later set (unit 13) is probably correlative with the regionally pervasive minette dyke swarm (Lecheminant et al., 1987).

The Uvauk complex is interpreted as an allochthonous remnant of a ductile strain zone developed under granulite facies. It was emplaced onto relatively lower grade, previously polydeformed quartzo-feldspathic gneisses, and was subsequently overprinted by later deformational events. The Uvauk complex may be linked with other lithologically and tectonically similar ductile high-strain zones that are associated with the uplift of granulite-anorthosite complexes: Kramanituar to the west, Hanbury Island and Daly Bay to the east, exposed between Baker Lake and Daly Bay (Fig. 1, points A to F). Collectively they define a broad, discontinous, east-trending deformation zone of probable Archean age. This zone may be linked with other segments of the Snowbird tectonic zone farther to the west and southwest in the District of Mackenzie (Hanmer and Kopf, 1993; Tella and Eade, 1986) on tennuous grounds. In northern Saskatchewan, the Snowbird tectonic zone contains granulite grade mylonites developed during ca. 3.2 Ga and ca. 2.6 Ga ductile deformation events, that were subsequently overprinted by Early Proterozoic brittle fault-zone fabrics (Hanmer et al., 1992).

Metavolcanic and metasedimentary rocks (units 2, 3)

Metavolcanic rocks, volcanic breccia, and minor felsic volcanics and tuff (unit 2) form two east- to northeast-trending, moderate to steep (35°-85°), north-dipping belts: one in the southwestern portion of the map area, and the second along the south shore of Barbour Bay (Fig. 2, 3). They

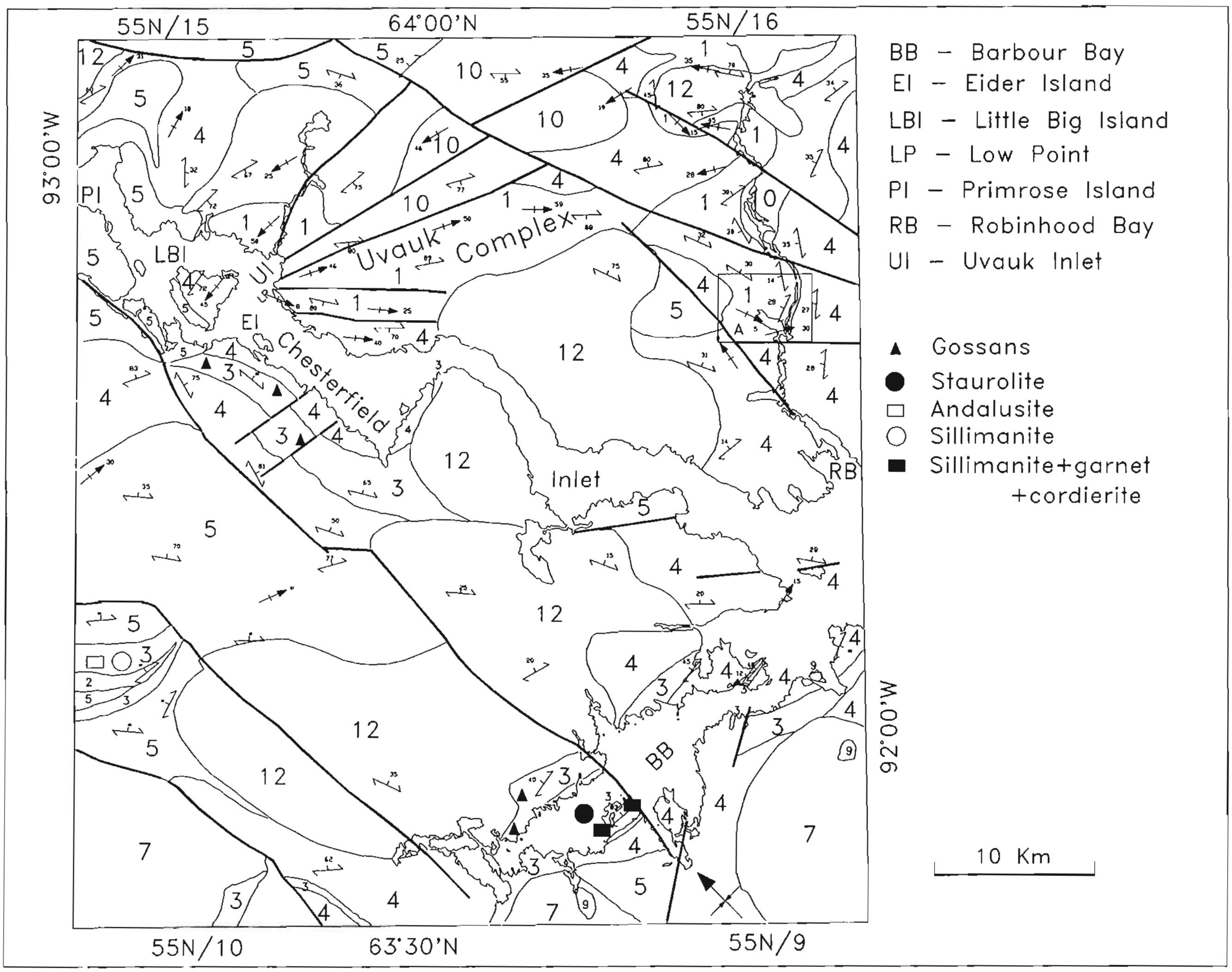


Figure 3. Simplified structural map of the northeastern parts of the Gibson Lake (NTS 55N) map area.

are believed to be higher grade equivalents of the Rankin Inlet Group metavolcanics exposed in the Rankin Inlet region 50 km to the south-southeast (Tella et al., 1986, 1992). The western belt extends west into the Gibson Lake-MacQuoid Lakes area (Reinhardt et al., 1980). Metasedimentary rocks of unit 3 structurally overlie the metavolcanic rocks.

Metasedimentary gneiss belts (unit 3), consisting of garnet+biotite ± staurolite ± andalusite ± sillimanite ± cordierite ± muscovite + plagioclase + quartz assemblages, are mostly restricted to the region south of Chesterfield Inlet. They are exposed in three regions: on both sides of Barbour Bay, and in two areas southwest of Chesterfield Inlet (Fig. 2, 3). The belts wrap around, and dip away from, domal masses of younger felsic plutons (unit 12). In the Barbour Bay region the metasedimentary gneisses structurally overlie the layered quartzofeldspathic gneiss (unit 4). Gneissosity in the metasedimentary gneiss is concordant with that in the layered gneiss for the most part, but discontinuous, layer-parallel, high-strain zones (metres to tens of metres wide) are present along the the contact zone, suggesting a tectonic break between the two units (3 and 4). The rocks in this belt are fine- to medium-grained iron-rich pelites that are compositionally well banded with quartz, quartz+feldspar, and garnet+biotite ± staurolite ± andalusite ± sillimanite ± cordierite rich layers. Sillimanite and garnet porphyroblasts up to 4 cm long are present in most parts of the belt. The metasedimentary belt extend south in to the adjoining region, where their lithological and structural characters were previously described (Tella et al. 1992). These assemblages are characteristic of low to moderate pressure-high temperature metamorphism. Such terranes record anomalously high geothermal gradients commonly attributed to magmatic transport of heat.

Southwest of Chesterfield Inlet metasedimentary rocks are exposed in two east-trending, undulating belts (Fig. 2, 3). In the northern belt, they are truncated to the northwest by an east-west trending high strain zone. To the east and west they were intruded by granitic plutons (units 5, 12), and to the south they structurally overlie quartzofeldspathic orthogneisses (unit 4). Rocks in the belt consist of biotite schists and biotite granofels with locally developed garnet layers, oxide iron-formations and amphibolite layers. The garnet rich layers contain small amounts of arsenopyrite. Mesoscopic structures in the belt are poorly developed, but minor structures include refolded, tight isoclinal folds. The rocks in the southern belt consist of garnet + biotite + andalusite + sillimanite + quartz + plagioclase assemblage.

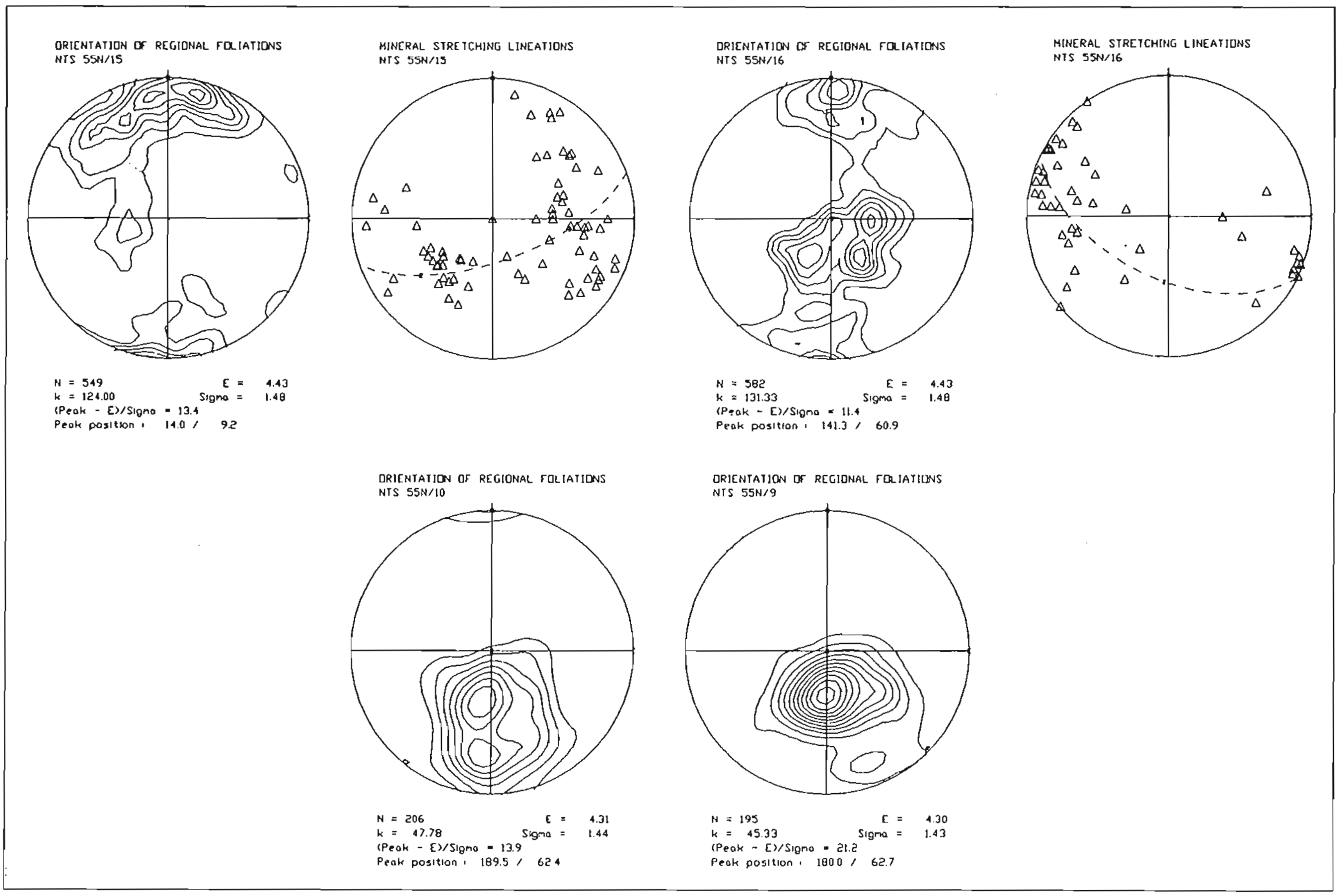


Figure 4. Orientation of selected planar and linear fabric elements from the study area (55N/9,10,15,16). Plotting software - Spheristat (1990).

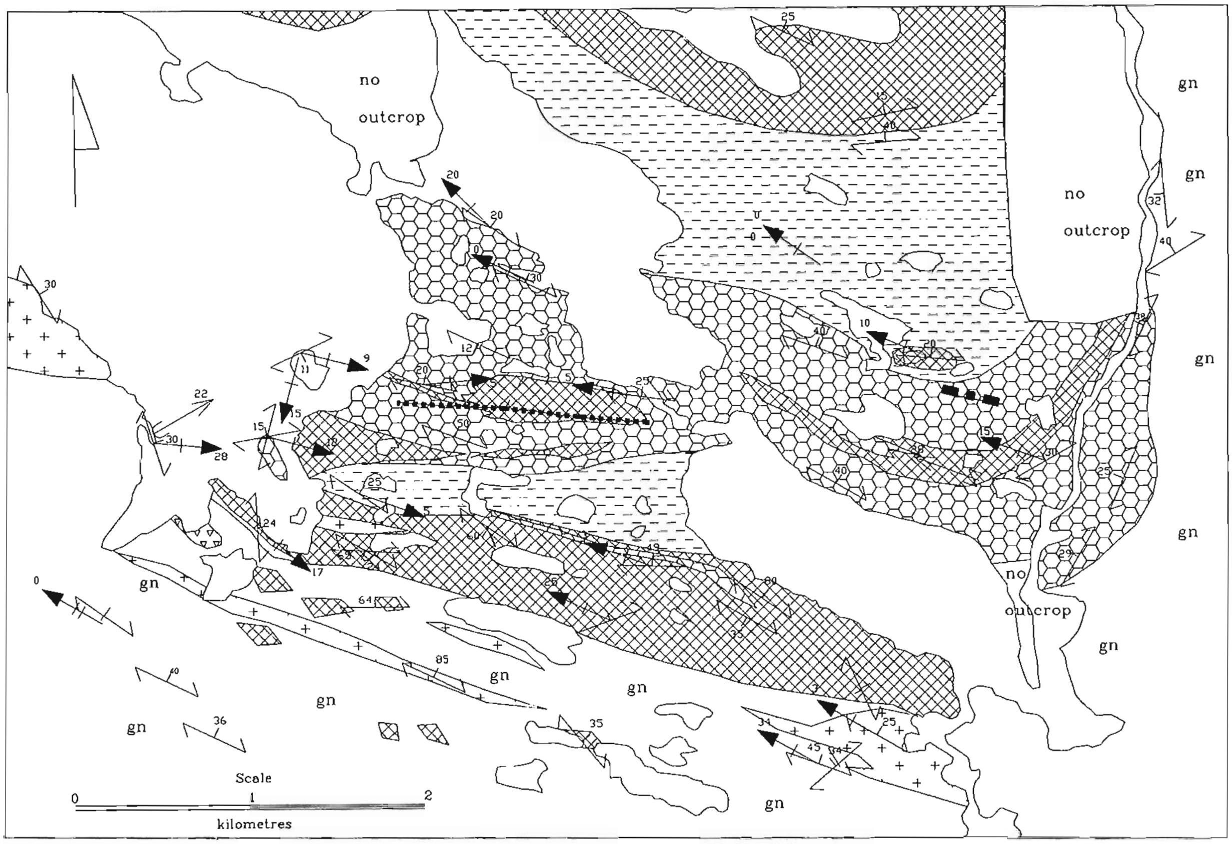
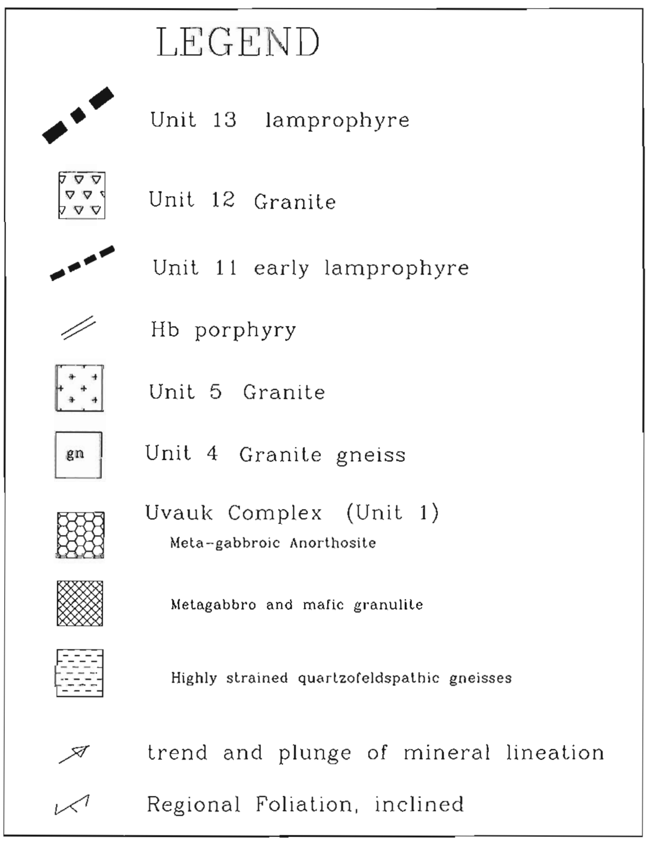


Figure 5. Sketch map showing the geology and structure in a portion of the Uvauk complex northwest of Robinhood Bay. (Location: Area-A, Fig. 2, 3).



At least four sets of folds are present in the metasedimentary belts. An early isoclinal, doubly plunging, northeast-trending fold set (F_1) is refolded by a north-northwest-plunging open to tight fold set (F_2), which in turn, is modified by moderately (<45°) west plunging open F_3 folds, and north plunging F_4 folds.The above structural observations are consistant with those reported from the adjoining regions to the south and southeast (Tella et al., 1992; Tella and Annesley, 1987).

The distribution of selected metamorphic minerals is shown in Figures 2 and 3. The presence of sillimanite, garnet, and biotite in gneisses and in adjacent quartzofeldspatic rocks, together with well developed quartzofeldspathic melt pods attest to middle to upper amphibolite facies conditions of regional metamorphism. The metamorphic mineral assemblages indicate middle to upper amphibolite facies conditions of regional metamorphism. A possible protolith for these gneisses is a hydrothermally altered sequence of acid and or intermediate volcaniclastic sediment. Staurolite occurrence in the belts is restricted to the region south of Barbour Bay. There, garnet, sillimanite, cordierite, and biotite appear to coexist as stable phases. Thermobarometric calculations, based on a number of different mineral equilibria, from the adjoining region of similar lithologies to the southeast, yielded P-T estimates of ca 3.4kb and 635°C for the assemblages in this unit (Tella et al., 1989).

Figure 6. Shallow, west-dipping layered anorthosite - mafic granulite - gabbro suite, Uvauk complex; aerial view to the south. GSC1992-253E

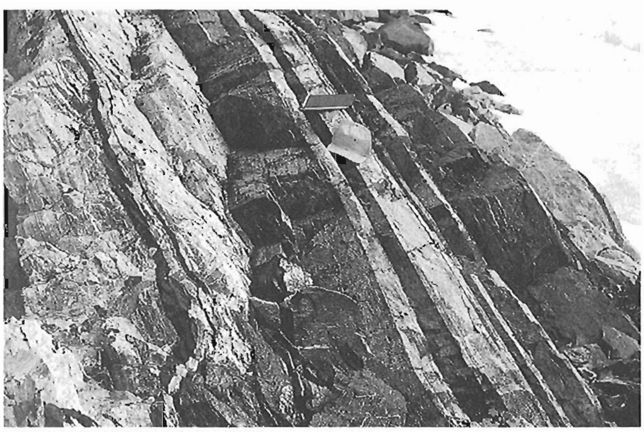


Figure 7. Well layered mafic granulite-anorthosite-gabbro suite, Uvauk complex. GSC1992-253L

Figure 8. Anorthositic gabbro mylonite at the base of the Uvauk complex displaying ribbons of plagioclase and hornblende. GSC1992-253E

Several gossan zones, some containing arsenopyrite, have been identified (Fig. 2, 3) in the metasedimentary belts. These belts are considered to be potential exploration targets for base metals.

Quartzofeldspathic gneisses (unit 4)

The rocks in this unit (Fig. 2, 3) consist of mixed assemblages of polydeformed, amphibolite grade orthogneiss, migmatite, biotite-muscovite-sillimanite ± cordierite ± garnet gneiss, minor proportions of iron-rich metasediments, and different generations of mafic dykes, now transformed into garnetiferous amphibolites. They form an autochthonous base to the Uvauk complex and contain, for the most part, amphibolite grade assemblages. These gneissic rocks are widely exposed in the region and extend into the adjoining map areas (eg. Tella et al., 1992; Tella and Annesley, 1987,1988). Granite dykes of several ages cut the rocks of this unit on all scales. Only relevant characteristics of unit 4 are presented here for an overview of lithology, structure and metamorphism. South of Barbour Bay, the orthogneiss is fine- to medium-grained, light- to dark- grey to pink, well banded and layered rock consisting of several distinct compositional phases that range from tonalite to granite. The gneissic layers contain variable proportions of quartz-plagioclase-K-feldspar-biotite-amphibole assemblages. Less deformed granite protoliths show gradational contacts with the orthogneiss. The layered gneiss contains xenoliths of metamafic rocks consisting of garnet-hornblende-clinopyroxene assemblages, agmatite, and rafts of anorthosite and gabbro. Coarse, pink pegmatite dykes, related to a large granite pluton (unit 12, Fig. 2, 3), cut all rock types in this area. Mylonitic rocks, which contain coarse K-feldspar porphyroclasts, form concordant layers (tens of metres wide) within quartzofeldspathic gneiss. The regional foliations commonly trend easterly and dip 45°-70° to the north, although shallower or steeper dips are also noted (Fig. 4).

Discontinuous, folded, ductile, high-strain zones, which display excellent mylonitic textures, form an integral part of unit 4. They are well exposed southeast of Barbour Bay. There the protoliths include deformed orthogneiss, migmatite, paragneiss, and several sets of metamafic and granite dykes. The high-strain zones are folded into tight, upright, shallow (<30°) doubly plunging isoclines that are coaxially refolded into moderate to open, northeast-plunging upright folds. Fold hinges in rootless, steep (>70°) doubly plunging folds in pelitic and mafic layers are parallel to tight isoclines developed in the high-strain zones. In some localities, the mylonitic fabric is partially or completely annealed, and is clearly overgrown by K-feldspar porphyroblasts. Mineral stretching lineations are poorly developed, but where present, they plunge shallowly (<30°) north-northeast, northeast, or southwest. The high-strain zones appear to have a complex deformational history of multiple periods of development, each separated by mafic and granitic dyke emplacement, and subsequent mylonitization. The age of deformation, metamorphism of the country rock, and subsequent development and folding of ductile strain zones, are all believed to be Archean (Tella et al., 1992).

Felsic and mafic intrusions (units 5 to 13)

Rocks of unit 5 are predominantly exposed as large plutons in the western part of the map area south of Chesterfield Inlet, and to a lesser extent in areas farther north of the Inlet, and south of Barbour Bay (Fig. 2, 3). The lithologies consist of pale-pink gneissic quartz diorite and granodiorite. The plutons clearly intrude the polydeformed, layered quartzofeldspathic gneisses (unit 4) and show well foliated migmatitic margins that contain abundant, but discontinuous, amphibolite layers, and minor amounts of sillimanite-muscovite schist. The central portions of the plutons are relatively undeformed.

South of Barbour Bay, rocks of unit 5 consist of banded quartzofeldspathic gneiss and gneissic granodiorite, minor garnet-sillimanite schist, and pink granite (Tella et al., 1992). They are fine- to medium-grained and well banded (up to 300 m wide). The banding is defined by biotite-poor and biotite-rich layers. Inclusions of biotite ± garnet schists, and garnetiferous amphibolites are common. Regional foliation strikes northeast with variable dips to the northwest. The unit is bounded to the north by a ductile, high-strain zone in unit 4, and intruded by granite (unit 7) to the south, and it is truncated to the east by a north-northeast-trending fault.

Large masses of relatively undeformed gabbro (unit 6) are exposed in the northwestern parts of the region northwest of Uvauk complex (Fig. 2, 3). The gabbro is medium to coarse grained, homogeneous, commonly massive , and shows a well developed ophitic texture. It intrudes the layered gneisses (unit 4) and is cut by white and pink granite dykes. Its relationships to units 5 and 7 are not clear. Adjacent to some of the major northeast-trending fault zones, the gabbro is intensely sheared and altered to a fine grained rock mainly consisting of hornblende, chlorite, and serpentine mixture. This unit may, in part, be coeval with the gabbroic rocks of unit 9.

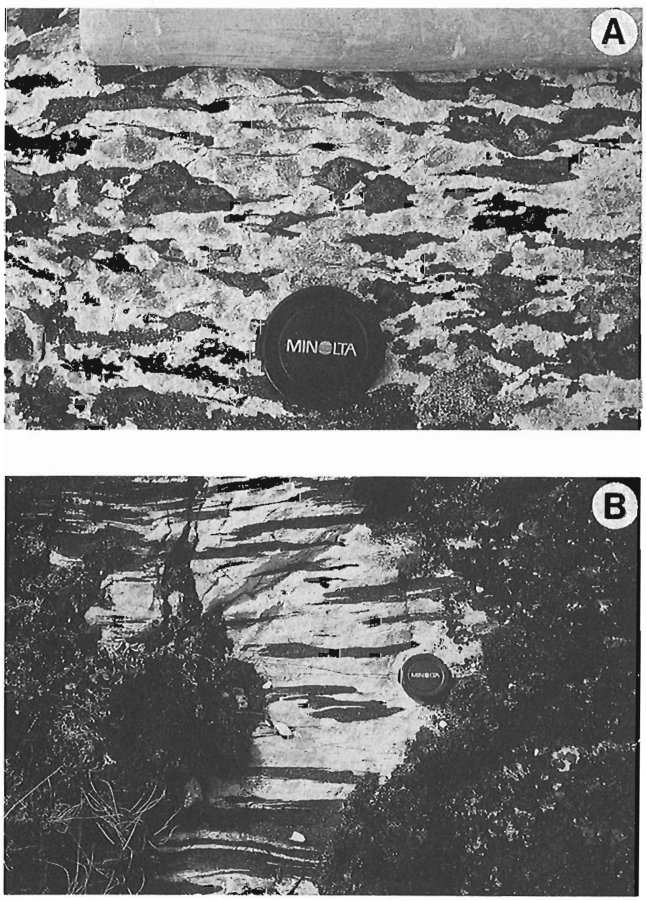


Figure 9.

A. Relatively undeformed anorthositic gabbro, Uvauk complex; note coronitic textures with pyroxene cores successively rimmed by hornblende and opaques; GSC 1992-253U.

B. Deformed equivalent at the base of the Uvauk complex; GSC1992-253V.

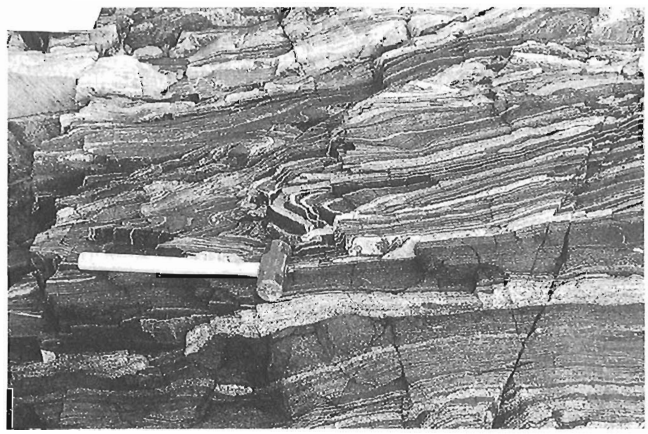


Figure 10. Heterogeneous ductile flow folds in the mafic granulite-anorthosite-gabbro suite, Uvauk complex; the tight fold at the centre plunges subverticallly. GSC1992-253R

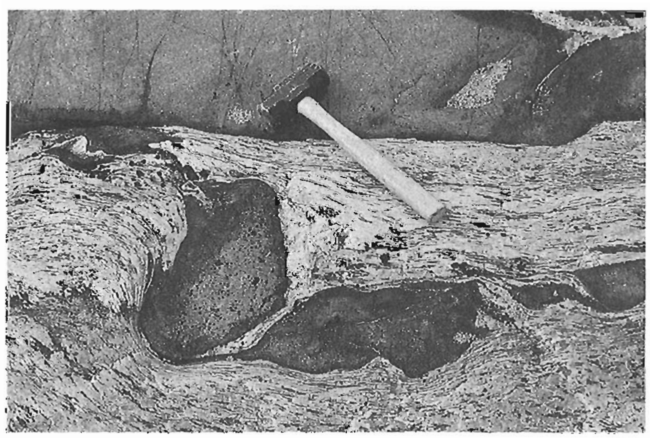


Figure 11. Ductilely dismembered mafic dyke in heterogeneously deformed anorthositic gabbro showing an apparent dextral sense of ductile shear. GSC1992-253I

In the southwestern and southeastern parts of the map area, two large masses of quartzmonzonite to granite (unit 7) intrude the pelitic and layered gneisses (units 3,4). they are massive to weakly foliated, pink to white, and contain abundant inclusions and rafts of country rocks. Although rocks in this unit (7) are lithologically similar to those of unit 12, the presence of small gabbro intrusions in unit 7 and their absence in unit 12 suggests that the two units are not coeval.

A number of east-trending gabbro and diabase dykes (unit 8), up to 300 m wide, are exposed throughout the region, but more commonly in the region close to Chesterfield Inlet (Fig. 2). They are too small to be shown on the accompanying map. The dykes clearly intrude the polydeformed rocks of units 1 to 5, and are overprinted by a weak, later deformation. They may represent multiple sets of east-trending dykes. Their relationships to rocks of units 6 and 7 are not clear. A coarse grained, massive and fresh, 150 m wide, pyroxenite dyke of uncertain age cuts gneissic rocks (unit 4) 5 km southwest of Barbour Bay. The dyke trends northerly and can be traced along strike for over 5 km to the north.

A number of medium- to coarse-grained, massive to weakly foliated pyroxenite, gabbro, and ultramafic intrusions (unit 9) of probable early Proterozoic age, are sporadically distributed throughout the region. Although they are grouped in a single map unit, they may represent several generations of mafic intrusions. Two large gabbro bodies are exposed in the region south and southeast of Barbour Bay. They are undeformed, relatively fresh, locally contain hornblende oikocrysts and numerous inclusions of older gneissic rocks (unit 4). Pink granite dykes, related to unit 12, intrude the gabbro. In the southeast portion of the map area (Fig. 2, 3), a large gabbro plug intrudes the granitic rocks of unit 7.

On either side of Chesterfield Inlet adjacent to Little Big Island (LBI, Fig. 2, 3), several small elliptical plugs of brown weathering ultramafic rocks cut the gneissic rocks (units 1 and 4). One plug on the south shore, northwest of Eider Island, contains olivine phenocrysts set in a clinopyroxene matrix (websterite ?). The phenocrysts are tabular and poorly aligned, and are in part hydrated to pseudomorphous serpentine,

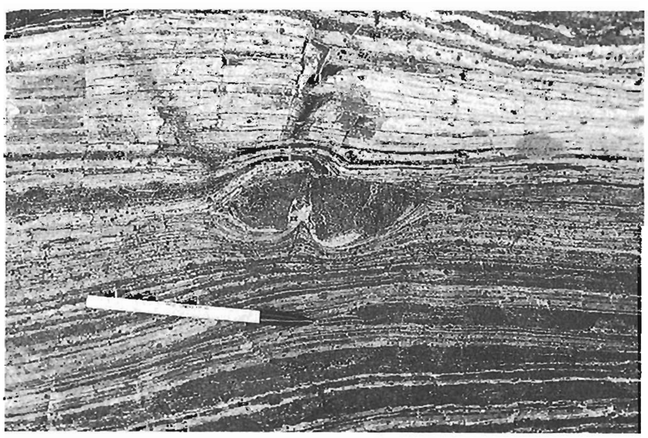


Figure 12. Pull-apart of a garnet porphyroblast in anorthosite ultramylonite, Uvauk complex. GSC1992-253T

chlorite mixture. Rafts of anorthosite and gabbro of the Uvauk complex are sparsely distributed in the plugs. No quartzofeldspathic inclusions have been noted. East of Low Point on the north side of Chesterfield Inlet and along the northeast shore of Little Big Island, similar ultramafic rocks cut the highly strained Uvauk Complex and the quartzofeldspathic gneisses (units 1 and 4), and clearly were emplaced late in the tectonic history of the region. A small ultramafic plug in the northeast portion of the Little Big Island is transected by a northeast trending, moderate to steep south dipping, anastomosing set of faults. South-southwest plunging (<40°) lineations are well developed on the fault surfaces. The fault-set is considered to be a locally reactivated pre-existing shear zone since similar fault sets did not affect the ultramafic intrusions a few kilometres to the south.

The ultamafic rocks may be correlated with dunite plugs noted on the Bowell Island (Schau and Ashton, 1980) to the west, and other small ultramafic bodies described from the adjoining regions (Gordon, 1988; Reinhardt et al., 1973).

A massive, coarse grained, pink, porphyritic (K-feldspar) augen granite (unit 10) is in tectonic contact with the Uvauk complex (unit 1) in the northern parts of the region (Fig. 2, 3). It shows intrusive relationships with rocks of units 4 to 6. The granite contains less than 2% mafics (biotite, trace magnetite), and in some places the mafics are completely absent. Apatite, zircon and sphene are present as accessories. Adjacent to two northeast trending fault zones, the granite displays excellent, well developed, quartz-ribboned protomylonite fabrics. The timing and nature of displacements on these fault zones have not been clearly established.

A number of relatively young granite plutons (unit 12) that intrude all previously described rock units are widely distributed in a northeast-trending belt west of Barbour Bay, north of Chesterfield Inlet, and in the northwestern corner of the map area (Fig. 2, 3). They are in part magnetite-rich, and show high aeromagnetic anomalies (Geological Survey of Canada, 1966). Although all plutons are grouped into one single unit, they may not all be coeval. Related pegmatite and aplite dykes are widespread throughout the region. Individual plutons range in composition from quartzmonzonite to granite. They are equigranular to massive, pink, commonly poor in mafics (<2% biotite), undeformed in the cores, and show a weak to well developed foliated margins. Migmatitic aureoles up to 100 m wide, and gradational contacts with the country rocks are common. The abundance of inclusions of country rocks increases towards the margins of the plutons. One pluton in the northeastern portion of the map area, however, has a well defined sharp intrusive contact with the anorthositic rocks of the Uvauk complex (unit 1). There the anorthositic gneisses adjacent to the contact display a fine grained, sugary textured plagioclase developed as a result of thermal overprint. The lithological characteristics of most plutons are similar to those described in the Gibson Lake west-half map area (Reinhardt and Chandler, 1973; Reinhardt et al., 1980), and to those noted in the adjoining regions to the east and to the south (Tella and Annesely, 1987; Tella et al., 1992). Most of the plutons may belong to ca. 1.82 Ga granite suite in the region (Tella et al., 1989).

At least two sets of lamprophyre dykes are widespread in the region. One set of dykes (unit 11) trends 110°. They are up to 1 m wide, brown weathering, porphyritic dykes with biotite and pyroxene phenocrysts set in a fine grained matrix. One such dyke cuts the Uvauk complex and a set of undeformed granite dykes (unit 12 ?) approximately 10 km north-northwest of Robinhood Bay (Fig. 2). A thin section shows a wide range of mineralogy with xenocrysts of orthopyroxene with olivine inclusions(8%) and individual olivine (2%) grains along with complexly zoned clinopyroxene (30 %) and poikilitic biotite (15%) phenocrysts set in a matrix consisting of alkali feldspar, trace amounts of a colourless amphibole, apatite, titanite, carbonate, and zircon.

A second set of lamprophyre dykes (unit 13) with a 130°-140° trend are also widespread in the region. They are discontinuous, 1-3 m thick, dark grey to black, relatively undeformed, medium- to fine-grained dykes with large biotite/phlogopite phenocrysts. They contain abundant inclusions of country rocks (anorthositic gabbro, granite, quartzofeldspathic gneisses) and are texturally similar to lamprophyre dykes described from the Rankin Inlet region (Tella et al., 1986; Digel, 1986).

Both sets probably are related to the ca. 1.85 Ga alkaline igneous suite in the central Keewatin (LeCheminant et al., 1987), but are separated by an event of granite dyke emplacement.

SUMMARY

1. The Uvauk complex is composed of layered gabbro-mafic granulite-anorthosite. It forms a triangular rootless segment of granulite grade straight gneisses, the boundaries of which are marked by ultramylonites.
2. Amphibolite grade, polydeformed layered quartzofeldspathic gneisses form an autochthonous base to the Uvauk complex.
3. The Uvauk complex is interpreted as an allochthonous remnant of a granulite grade ductile, high-strain zone that formed under P-T conditions representing mid-crustal levels, and subsequently emplaced on to a relatively lower grade autochthonous gneiss terrane.
4. The Uvauk complex is structurally and lithologically similar to other ductile, high strain zones in the region (Daly Bay complex, Gordon, 1988; Kramanituar Complex, Schau et al., 1982; Hanbury Island Shear Zone, Tella and Annesley, 1988) and collectively they form a discontinuous, east-trending deformation zone of probable Archean age.
5. Metasedimentary belts in the autochthonous gneiss terrane, south of Chesterfield Inlet, contain gossan zones, some with arsenopyrite. They are considered to be potential base metal exploration targets.

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Geological Survey of Canada Project 850002

Progress on geology and resource assessment of the Archean Prince Albert Group and crystalline rocks, Laughland Lake area, Northwest Territories

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Chandler, F.W., Jefferson, C.W., Nacha, S., Smith, J.E.M., Fitzhenry, K., and Powis, K., 1993: Progress on geology and resource assessment of the Archean Prince Albert Group and crystalline rocks, Laughland Lake area, Northwest Territories; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 209-219.

Abstract: The Prince Albert Group near Laughland Lake comprises 2/3 biotite psammite + phyllite + biotite schist, 1/6 komatiite + ultramafic schist + fine-grained amphibolite, 1/12 iron-formation, and 1/12 quartzarenite. These were deposited, with volcanism and hydrothermal activity, as a rifted continental margin prism.

The group is intruded in sequence by foliated tonalite (Brown River Gneiss; also includes basement?), metagabbro, layered anorthosite with related amphibolites, foliated porphyritic granite (A') and massive fluorite granite. Metamorphic grade ranges from upper greenschist to lower amphibolite. Bedding-parallel foliations, crenulations, refolded isoclines and gneissic crystalline rocks record collisional orogeny.

Extensive Quaternary cover records north-northwest ice and meltwater flow. The prime deposit types being assessed include gold in shear zones and iron-formations, base metals and platinum group elements in shales, nickel-copper in komatiites, chromite in anorthosite and paleoplacer uranium-gold in quartzarenites, and tin, tungsten and rare metals associated with fluorite granite.

Résumé : Le Groupe de Prince Albert près de Laughland Lake englobe dans une proportion de 2/3 l'ensemble psammite à biotite+phyllite+schiste à biotite, dans une proportion de 1/6 l'ensemble komatiite+schiste ultramafique+amphibolite à grain fin, dans une proportion de 1/12 une formation ferrifère, et dans une proportion de 1/12 un quartzite sédimentaire. Ces roches se sont accumulées pendant une période de volcanisme et d'activité hydrothermale, formant un prisme de marge continentale de divergence.

Le groupe est traversé en succession par une tonalite feuilletée (Gneiss de Brown River; comprend aussi le socle?), un métagabbro, une anorthosite stratifiée accompagnée d'amphibolites apparentées, un granite porphyrique feuilleté (A') et un granite massif à fluorine. Le degré de métamorphisme se situe entre le sous-faciès supérieur des schistes verts et le sous-faciès inférieur des amphibolites. Les foliations parallèles au litage, les crénulations, les isoclinaux replissés et les roches cristallines gneissiques témoignent des effets d'une orogenèse par collision.

La vaste couverture quaternaire témoigne de l'écoulement nord nord-ouest des glaces et des eaux de fonte. Les principaux types de gisements évalués comprennent les minéralisations aurifères dans les zones de cisaillement et dans les formations ferrifères, les minéralisations en métaux communs et en éléments du groupe du platine dans les shales, les minéralisations en nickel et en cuivre dans les komatiites, les gisements de chromite dans les anorthosites, les paléoplacers uranifères et aurifères dans les quartzites sédimentaires, et les minéralisations en étain, tungstène et métaux rares associées au granite à fluorine.

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INTRODUCTION

This is a progress report on a mineral and energy resource assessment (MERA) which is guided by a committee comprising senior management in Department of Indian Affairs and Northern Development, Governments of Northwest Territories and Yukon, Canadian Parks Service (CPS), and Energy, Mines and Resources. The objective of these assessments and public consultations is to ensure that park establishment criteria are met while, to the degree practicable, lands with high mineral and/or energy resource potential remain available for exploration, development and use. Canadian Parks Service first proposed a national park centred on the head of Wager Bay (Fig. 1) (Jefferson et al., 1991). East of 92°, Jefferson et al. (1991) assessed moderate to low potential for metallic minerals in Archean tonalitic and supracrustal gneisses and high potential for oil shales in lower Paleozoic strata of Southampton Island. Continuing park studies have suggested including the marine waters of Wager Bay and the headwaters of Brown River west of 92°. Assessments of the latter areas are to be completed by April 1993 and will become part of the information base for consultations.

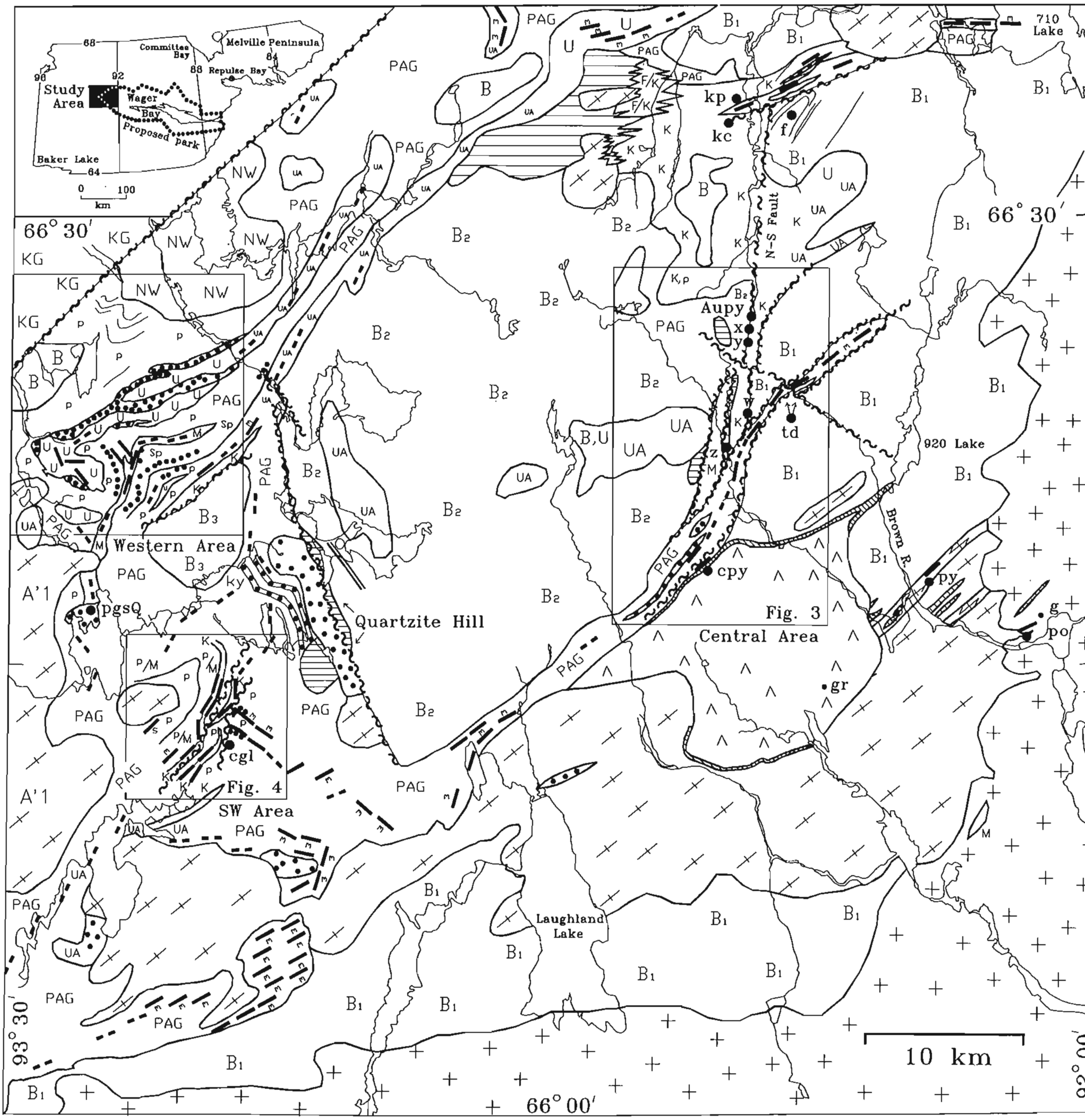

Figure 1. Index map and geology of the Laughland Lake area. Labelled large dots are localities described in text, (e.g. **Aupy** = 160 ppb gold in sulphidic sheared tonalite). Boxes locate Figures 3, 4. Legend in Fig. 2.

METHOD

We aim to upgrade the geoscience database, then determine the applicability of various mineral deposit analogues. We invite comments which might improve the database, and add mineral deposit types which we may have overlooked.

The Laughland Lake area has been partially mapped at scales of 1:1 million (Heywood, 1961), 1:250 000 bedrock (Schau, 1982), 1:250 000 surficial (Thomas, 1981) and 1:50 000 (Jefferson and Schau, 1992). Studies of similar rocks in the Woodburn Lake Group (Henderson et al., 1991), Mary River Group and Piling Group in northern Baffin Island (Jackson, in press), and around Wager Bay (references in Jefferson et al., 1991) have aided our understanding of this area.

In 1992 the authors (i) reconnoitred the southern part of the area by helicopter and foot; (ii) examined contacts between supracrustal and crystalline rocks; (iii) checked inferences by Jefferson and Schau (1992) from aeromagnetic data (Geological Survey of Canada, 1977); (iv) mapped well exposed supracrustal areas in detail; (v) sampled tills and eskers; and (vi) integrated archived field data of Heywood (1961) and Schau (1982).

Data were entered in Fieldlog version 2.83 (Brodaric and Fyson, 1989) and maps constructed in Autocad with further interpretations based on aeromagnetic data.

All rock types are listed in Figure 2. All localities mentioned in the text are shown in Figure 1 and some repeated in appropriate detailed maps (Fig. 3, 4).

GENERAL GEOLOGY

The Archean Prince Albert Group (Heywood, 1967), hereafter termed the group, extends at least 500 km to Melville Peninsula (Schau, 1982) and comprises 1/3 chlorite-actinolite schist, 1/3 phyllite, biotite schist or gneissic mica-rich rocks, and 1/6 each of iron-formation, quartzarenite and greenstone. In the study area, biotite psammite + phyllite + mica-rich rocks form about 2/3 of the group. Crystalline rocks include foliated tonalite, some of which may be basement to the group, metagabbro, anorthosite with amphibolite and porphyritic and massive granite. Metamorphic grade ranges from upper greenschist and lower amphibolite in the study area, to upper amphibolite tens of kilometres northeast of the map area.

METASEDIMENTARY ROCKS

Biotite psammite (P)

Recessive biotite psammite appears to underlie 2/3 of the area mapped by Schau (1982) as uAp and by us as "PAG". It is grey, brown to grey-weathering schistose metasandstone comprising quartz and feldspar with subordinate biotite and local garnet. Parallel bedding on a scale of millimetres to centimetres is typical. The beds are in places graded from a pale, sharp-based sand layer through recessive, darker, finer grained, locally laminated metasiltstone (Fig. 5a). In most of the area younging is unclear. One exposure of low-angle cross lamination resembles hummocky crossbedding. Subordinate facies include thick bedded granular light-buff weathering, essentially mica-free quartzose metasandstone and very thin bedded recessive schistose dark grey pelite. The predominant bedding-parallel lamination and grading are reminiscent of storm beds (Johnson, 1978) and distal turbidites (Walker, 1984). The massive, granular, light buff sandstones could be A.A.A. turbidites (Walker, 1984).

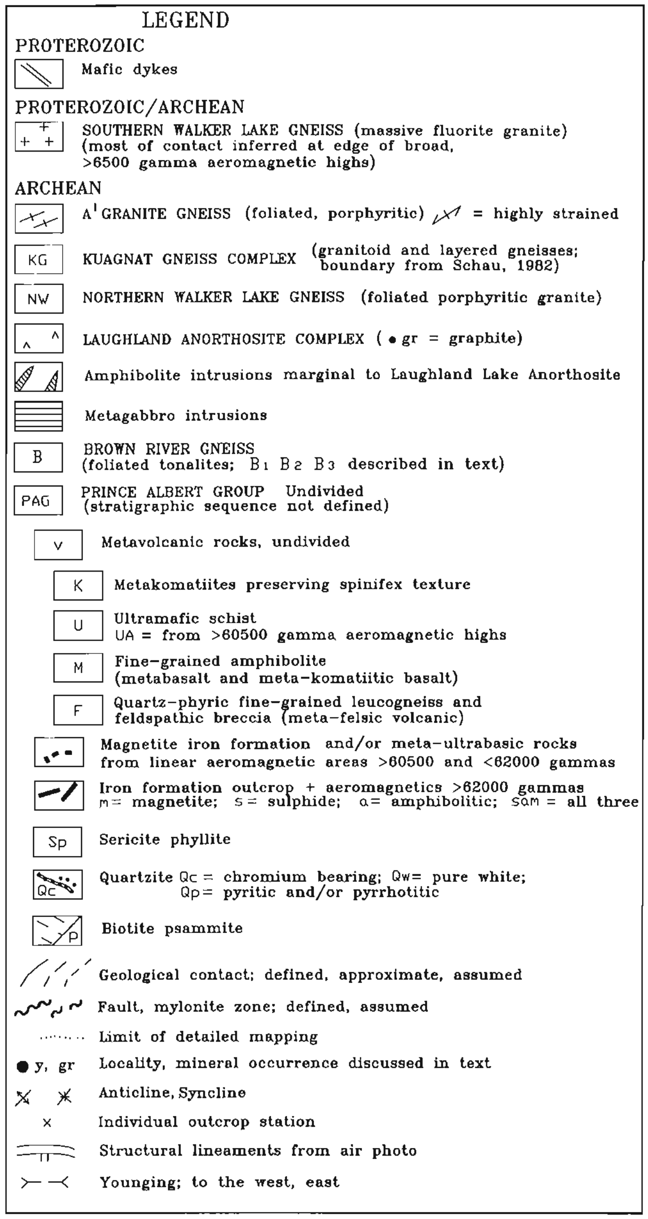

Figure 2. Combined legend for Figures 1, 3, 4.

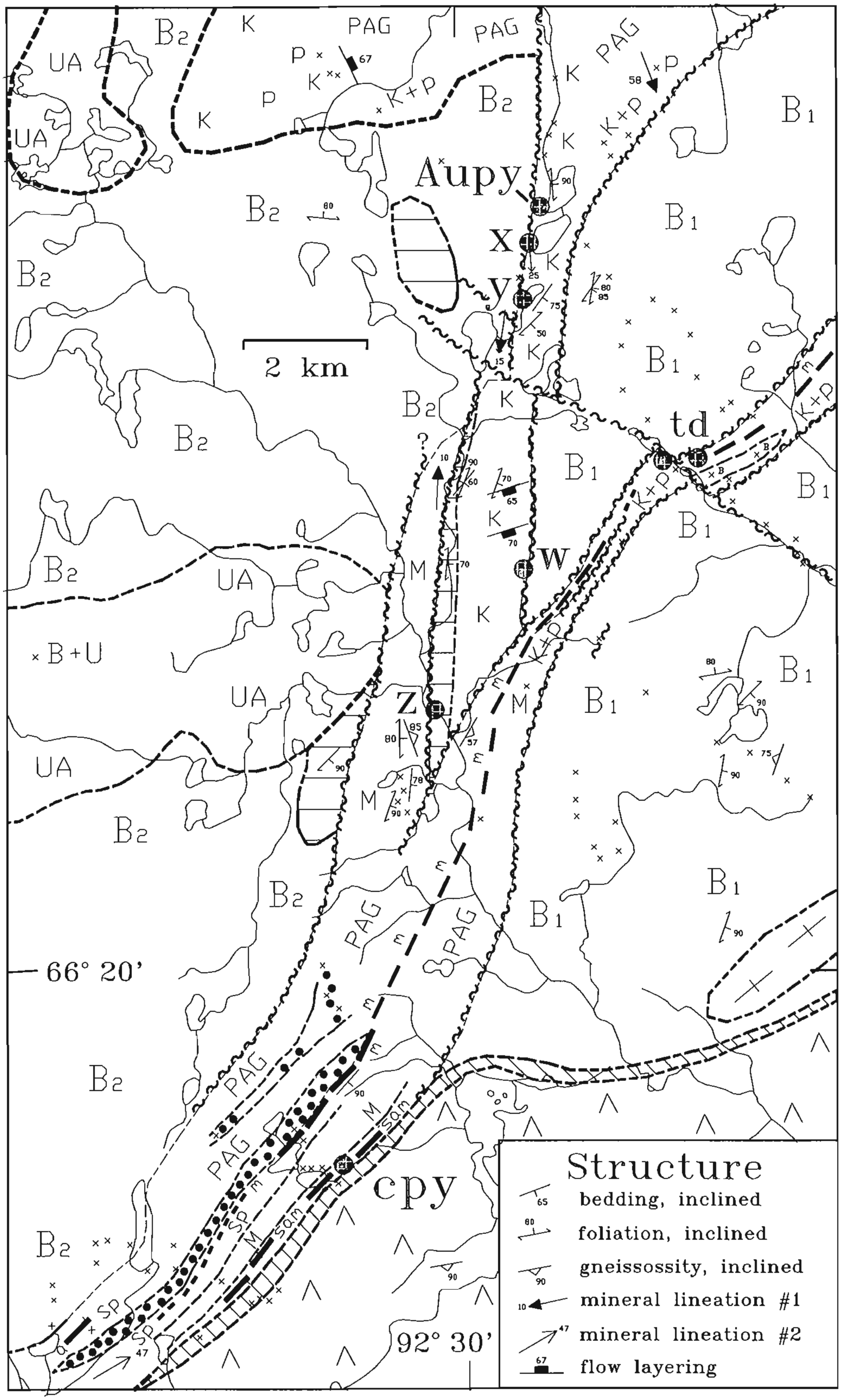

Figure 3. Detailed geology of the central area. Explanation as for Figure 1.

Sericite phyllite (Sp)

A sequence of silver- to grey-weathering phyllite with quartzarenite interbeds is 750 m thick in the western area (Fig. 1) and is mapped in the southern part of the central area (Fig. 3) where it forms 100 % of chips in frost boils. Elsewhere it is very thin or recessive. The interbedded quartzarenites and lack of dark minerals suggest that the phyllite was derived from a well weathered source area.

Quartzarenites (Q, ...)

All quartzarenites are strongly foliated and most form decimetre- to metre-scale massive beds. White quartzarenites (Qw) within large units of biotite psammite and sericite phyllite are 10 to 35 m thick and 0.1 to 6 km in length. One 1200 m-thick unit (Quartzite Hill) contains sparse muddy strata, and may be either a bent isocline (Jefferson and Schau, 1992) or as 1992 sedimentological and structural data suggest, a bent homocline with remarkably blunt lateral terminations.

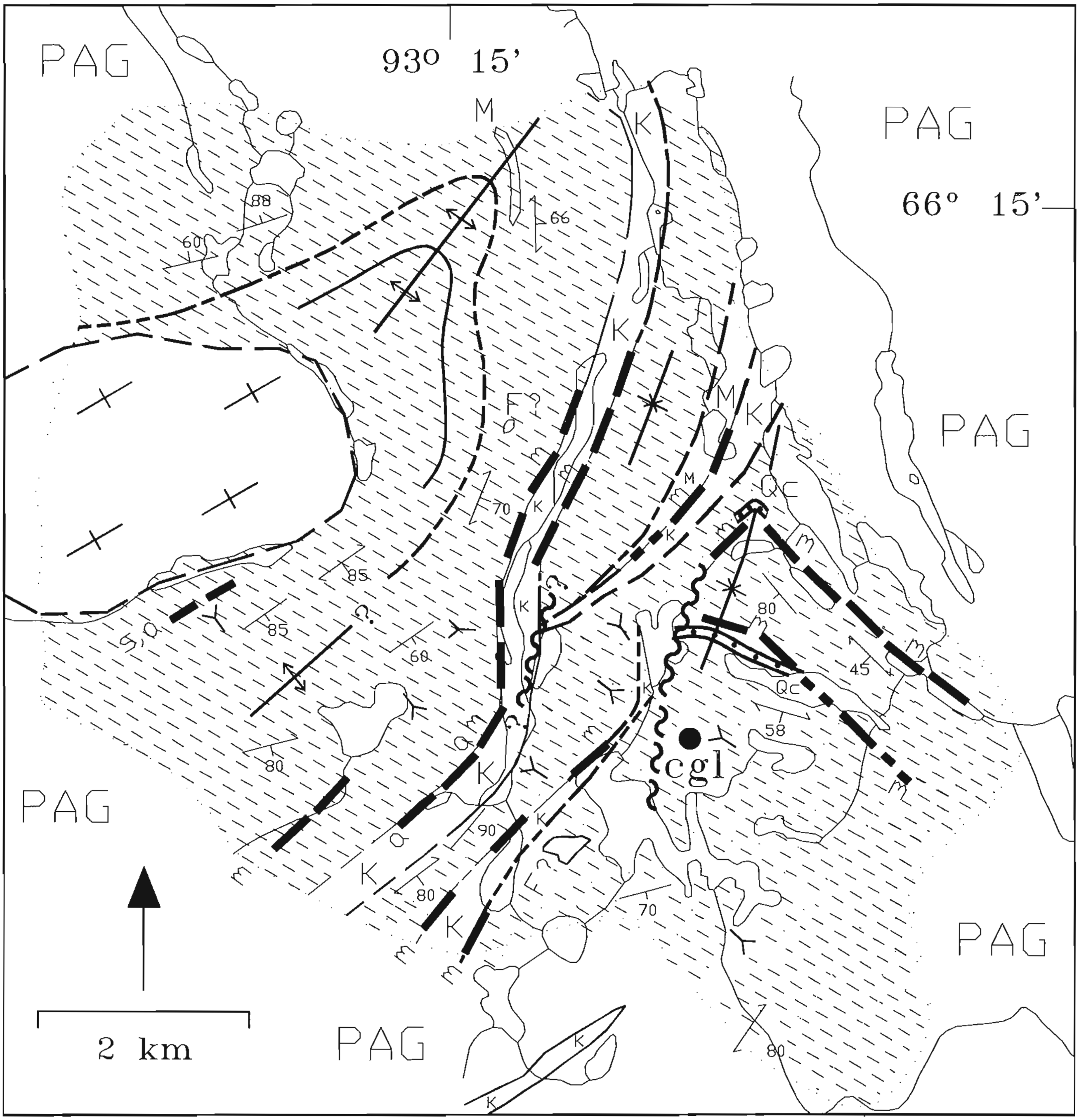

Figure 4. Detailed geology of the southwestern area (Fig. 1). Legend in Figures 2 and 3.

Crossbeds are moderately well preserved only in Quartzite Hill. Schau (1982) reported crossbeds up to 10 m thick near Laughland Lake, suggesting an eolian origin. We observed a few 1-3 m low-angle crossbeds, but mostly 5 to 25 cm tabular and trough types.

Bright green, fuchsitic quartzarenites (Qc) are closely associated with iron formations and komatiites (Fig. 4). The chromium has been interpreted as derived by paleoweathering of the komatiites (Schau, 1982, p. 28); alternatively it could have been introduced by the same hydrothermal processes which deposited iron-formation.

A rusty weathering grey pyrite-garnet-sillimanite quartzite (Qp) sequence (25 m + 15 m gap + 15 m) overlies biotite psammite at locality **pgsQ**. Garnet-biotite-rich beds (pelites?) comprise no more than 5% of this sequence. Three opposed herringbone crossbeds were observed in the 25 m part, and trough crossbeds in the 15 m part.

Insufficient large crossbeds have been found to support eolian (Walker and Middleton, 1977) genesis of the quartzarenites. Nor have mud chips, mud cracks or vertical accretion deposits of braided streams (Walker and Cant, 1984), been found. Such mud-free, extensive, cross- and planar-bedded quartz sand bodies are characteristic of marine shelves and epeiric seas (Johnson, 1978) swept by vigorous traction currents. Opposing crossbeds permit the possibility of tidal currents. Schau (1982) interpreted decimetre-scale lateral facies changes of the thinner quartzites to sericite schist and phyllite (Sp) as deposits of river, dune or offshore bars; these could also record turbiditic channel fills or dismemberment by deformation.

Quartz pebble conglomerate and pebbly mudstone (cgl)

At locality **cgl** one outcrop of stretched (4:1) quartz-pebble orthoconglomerate, 30-35 m thick and 300 m in strike length, lies within biotite psammite and contains one exposure that displays size grading. Clasts comprise only white to grey quartz and banded magnetite-quartz iron-formation, with a range of angularity (Fig. 5b). Three hundred metres northeast of **cgl**, one exposure of paraconglomerate, about 170 m thick, comprises similar clasts in a wacke matrix, and sparse diffuse beds of quartz pebble orthoconglomerate, 15 to 30 cm thick.

The psammite host, the grading and lack of continuity of these conglomerates suggest proximal turbidites and debris flows in channels cutting a deep shelf or base of continental slope (Walker, 1978, p. 946). Their rarity, compositional maturity and the absence of dropstones, contradict a glacial origin.

Carbonate units

At least 22 m of interbedded psammite and brown-weathering laminated dolomite are just east of locality **cgl**. A 40 m-thick carbonate unit mapped by Schau (1982; his Fig. 8) also appears to be metasedimentary.

Iron-formations

Schau (1982) recognized abundant magnetite iron-formation; Jefferson and Schau (1992) used linear aeromagnetic highs >62 000 gammas to extrapolate magnetite iron-formations from two outcrops and discussed the gold potential of the sulphide facies. Work in 1992 confirmed the extrapolations; exposed magnetite iron-formations are up to 60 m thick and

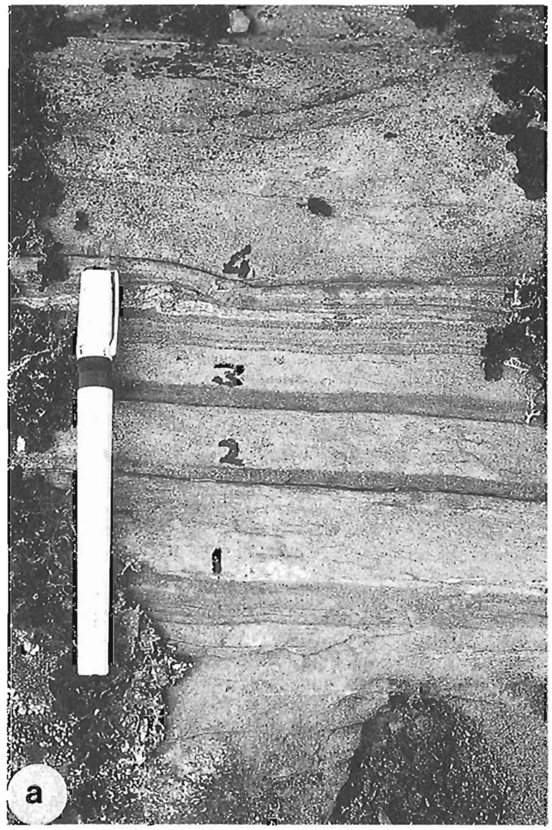

Figure 5. (**a**) Thin graded beds (numbered) in measured section located at E461388, N7326461, Grid Zone 15; typical of biotite psammite in the Laughland Lake area. (GSC 1992-259B) (**b**) Orthoconglomerate, locality **cgl**, Southwestern Area. **m** = clast of magnetite iron formation. (GSC 1992-259C)

3 km in length. Also documented were a number of gossanous, non-magnetic sulphide and silicate iron-formations, some being very fine grained sulphidic amphibolite. Many iron-formations are paired with metavolcanic rocks, a few are isolated within thick psammite sequences, and some are closely paired with chromian quartzarenites.

METAVOLCANIC ROCKS

Komatiites, ultramafic schists and amphibolites constitute <10% of Prince Albert Group, but dominate the central area. Minor quartz-phyric leucogneisses are interpreted as felsic metavolcanics in the southwestern area (F?, Fig. 4). Quartz- and feldspar-phyric metavolcanic breccias and tuffs north of the central area are interfingered with mafic and ultramafic metavolcanic rocks (stations 73SMA107-109 of Mikkel Schau). Felsic metavolcanic rocks are common several hundred kilometres to the northeast (Schau, 1977).

Ultramafic rocks (K, U, UA)

Most meta-komatiites (K) in the area have orange-weathering cumulus zones directly overlain by relatively recessive grey to green spinifex zones, with sheared interflow contacts (Fig. 6a) and lack the complete sequence typical of Archean komatiites (Fig. 6b). Rare features include columnar joints (locality **kc**) and younging indicators such as flow top breccias, differential weathering and vegetation at spinifex-cumulate contacts and foliate zones between cumulate and spinifex zones. About 30 m of flattened pillowed ultramafic flows (locality **kp**) and massive dark grey metabasaltic flows cap the komatiites in better exposures north of the central area (Fig. 3), suggesting a dominantly subaerial volcanic environment with late submergence. The predominant couplets may be sills; if so harrisite would be the appropriate textural term for the bladed (spinifex) zones (L. Hulbert, pers. comm., 1992). Strongly foliated talc-actinolite schist, locally interlayered with brown dolomite, and brecciated dolomite-talc-amphibolites are mapped as meta-ultrabasic (U) and interpreted as altered komatiite. Both komatiites and meta-ultrabasic rocks contain magnetite and are coincident with linear, aeromagnetic highs >60 500 gammas. Similar highs which have not been investigated in the field are mapped as "UA".

Fine-grained amphibolite (M)

Fine-grained black to green amphibolites interpreted as meta-basalt include minor lenses of brown dolomite with amphibolite breccia. Fine grained amphibolites containing garnet + sulphide±magnetite are interpreted as iron-formation.

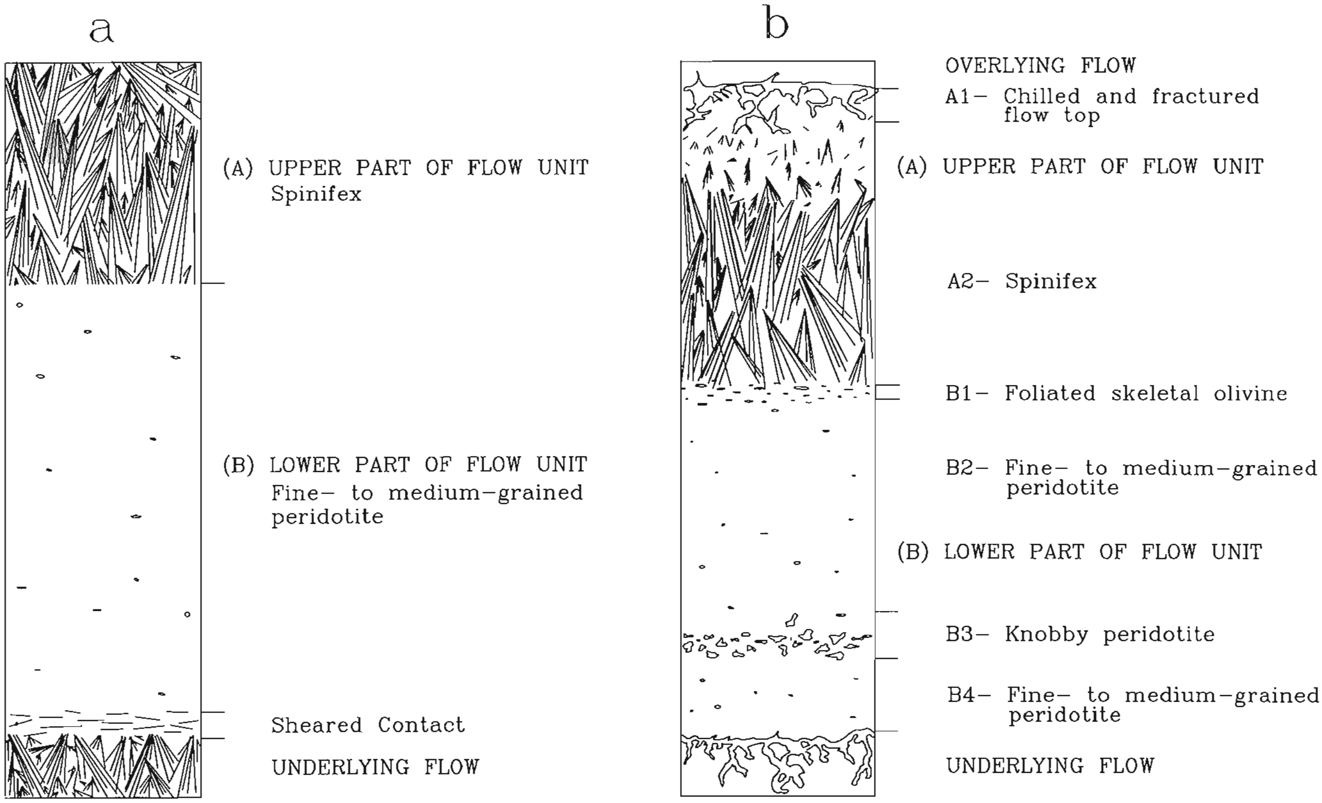

Figure 6. (**a**) Cumulate-spinifex pairs of the central area. (**b**) Type komatiite flow of Arndt et al. (1977).

CRYSTALLINE ROCKS

Crystalline rocks underlie more than 3/4 of the map area, and are listed in interpreted chronologic order in Figure 2. The Kuagnat Gneiss Complex has not been investigated in this study

Brown River Gneiss (B)

The Brown River Gneiss (Schau, 1982) includes variably foliated, uniform grey hornblende tonalites through the central and western areas, and foliated tonalites with amphibolite bands in southeastern areas. All are characterized by broad undulatory aeromagnetic lows. Tonalites truncate local structures in the Group; most exposed contacts with the Group are brittle faults or mylonite zones (e.g **py, w, x, y, z** in central area), but regional foliations in the Group wrap around larger tonalite bodies. Foliations within the tonalites are overall northeasterly, and lenticular on a mm scale, interpreted as two intersecting fabrics. Northwestof 920 Lake, anastomosing mafic dykes cutting the tonalite have two foliations (similar dykes southwest of 920 Lake also cut A' granite). Broad folds in tonalite interpreted from air photographs at locality **f** are truncated by contacts with the group and may indicate basement.

Foliated tonalite forms irregular masses (B_1) transected by gneissic and massive granites to the northeast and south of the Laughland Lake Anorthosite Complex, a large oval (B_2) in the west-central part of the area, and some smaller oval bodies (B) surrounded by the group. Pluton B_3, the foliated "western tonalite stock" of Schau (1982), is here assigned to Brown River Gneiss based on texture, composition and its mylonitic northwest contact against supracrustal rocks with no contact metamorphism. Foliations are steep toward the pluton.

Inclusions (or fold keels?) of amphibolite and fine grained biotite gneiss are associated with northeast-trending high strain zones in which the two foliations are distinct. The poorly exposed meta-ultramafic bodies (U and UA) enclosed by the central oval tonalite are interpreted as inclusions.

At Locality **td** in the central area foliated and locally mylonitic tonalite dykes transect iron formation and meta-komatiites, and one contains ultramafic schist xenoliths. Spinifex is not preserved here, but the alternating green and tan komatiitic flow layering is clear. Foliated tonalites also intrude Prince Albert Group north of the central area.

A number of different tonalite phases are suspected but could not be mapped with available logistics; some may be intrusive (perhaps part of the **A'** suite), others basement. Detailed petrology and geochronology may provide better constraints on age relationships.

Gabbroic intrusions

Oval gabbros with distinct aeromagnetic lows occur on the margins (mainly north and east) of the large foliated tonalite oval. These are medium to coarse grained and massive to intensely foliated. Some are enclosed by tonalite, others are interfingered with or laterally adjacent to the Group. It is uncertain whether the interfingering is tectonic or primary.

Laughland Lake Anorthosite (^) *and related amphibolites*

The compositionally layered (Jefferson and Schau, 1992) Laughland Lake Anorthosite Complex is exposed in rugged white hills north-northeast of Laughland Lake and underlies about 150 km^2. It is surrounded by a 20-50 m thick border phase of amphibolite interpreted as leucogabbro and metagabbro. Amphibolite apophyses obliquely transect and extend into the Brown River Gneiss to the northeast. On the west and south both anorthosite and border amphibolite are mylonitized; foliations dip inward at 45° to 85°; steep lineations and C/S fabrics suggest that the anorthosite was structurally emplaced upward and westward against the Prince Albert Group on the west. The anorthosite is cut by A' granite and affected by the same deformation.

A' granites

A' granites (Jefferson and Schau, 1992) are foliated and K-feldspar-phyric, contain foliation-parallel xenoliths of Brown River Gneiss and supracrustal remnants, and dip under and intrude the amphibolite border of the Laughland Lake Anorthosite. Large northeast-trending belts of A' granite underlie the southern and eastern parts of the map area. Intrusive contacts with Brown River Gneiss are folded and obliquely transected by the northeast foliation.

West of 93°, individual plutons form rounded hills surrounded by linear valleys. Pegmatite and quartz veins, and increased metamorphic grain size in adjacent psammites support Schau's (1982) intrusive interpretations. The westernmost A'1 pluton has a marginal phase of coarse grained feldspar-phyric to pegmatitic white-weathering massive granite. Southern outcrops of the Northern Walker Lake Gneiss (northwest in Fig. 1) resemble A' granite.

Southern Walker Lake Gneiss (+)

This salmon pink, massive, fluorite-bearing granite (Jefferson and Schau, 1992) cuts the A' granite and older tonalites in the eastern part of the area. Its outcrop area coincides with spotty, 60 500 - 62 000 gamma aeromagnetic highs (Geological Survey of Canada, 1977). Foliations in decametre to kilometer angular xenoliths (roof pendants?) are concordant with those in adjacent gneisses. Both massive and foliated (A'?) granite intrude Prince Albert Group correlatives to the south (Heywood, 1961).

STRATIGRAPHIC-STRUCTURAL RELATIONSHIPS

Complex relationships within the group and with adjacent crystalline rocks were investigated in the detailed map areas.

Central area (Fig. 3)

This structural and lithological focus is cored by a komatiite sequence with minor metabasalt and no metasediments. Komatiite outcrops are scattered, north-trending ice-sculpted knobs. Northeast-striking, steeply dipping flows are truncated in detail by north-trending outcrop margins, some of which preserve northerly high strain zones. Regionally, the northeasterly flow trends are terminated against major faults which bound tonalites on the east and west.

The komatiite flows young and become basaltic toward the northwest in the north and toward the southeast in the south, suggesting a large anticline. The flows are also intercalated on the northwest and southeast with iron-formation, quartzite, psammite and amphibolitic psammites which could be volcaniclastic. Komatiite, iron-formation and psammite extend into the foliated tonalite along several northeast-extending highly strained apophyses. At **td**, foliated tonalite dykes containing ultrabasic xenoliths crosscut iron-formation and komatiites. Axial to the **td** extension is an intensely crenulated mylonitic rock of tonalitic appearance; the same mylonitic rock outcrops just west of locality **cpy**.

Western area (Fig. 1)

In the far west a southwesterly to southeasterly curved belt comprising 1 to 2 km of ultramafic schists with intercalated iron-formation, bounded by quartzarenites and biotite psammite, is interpreted as a refolded isoclinal fold. To the southeast, the following apparently homoclinal southeast-dipping sequence is faulted to the southeast against a foliated tonalite pluton (B_3):

20 m	magnetite iron formation (at base)
250 m	covered
>15 m	ultramafic schist
>60 m	fine grained amphibolite
~750 m	gently-dipping silver to grey sericite phyllite with white quartzarenite subunits
~750 m	complex of biotite psammite with subordinate iron-formation, sericitic quartzite, dense aluminous rusty rock, interlayered recessive brown carbonate/mafic schist, and a gabbro sill (cf. detailed section by Schau, 1982, Table 4)
~500 m	southeast-facing komatiite flows pinch out to the west into biotite psammite and are possibly represented to the northwest by the brown carbonate and mafic schist (at top).

Southwestern area (Fig. 4)

The western part of Figure 4 illustrates a northeast-plunging antiform in biotite psammite, cored by an oval A' granite. The moderately dipping psammite contains minor pyrite-chert and garnet-amphibole iron-formation, and discontinuous lenses of fine grained amphibolite and quartz-phyric leucogneiss (meta-volcanics?). Foliation is bedding-parallel.

East of the antiform lie three structurally repeated, north-trending belts of steeply dipping komatiite, amphibolite (metabasalt?) and iron-formation separated by psammite. Fold closures are interpreted from map patterns and high cleavage-bedding angles in psammites.

Farther east a northerly fault truncates a syncline with limbs defined by a unit of paired iron-formation and fuchsitic? quartzarenite in psammite. A second exposure of quartz-phyric leucogneiss is south of the iron-formation. The only two conglomerate units in the study area outcrop at locality **cgl**.

Measured section of psammites

A section of well preserved biotite psammites was measured in a 2000 m spillway 15 km southwest of the map area (E461388, N7326461, Zone 15). The 590 m stratigraphic thickness comprises massive, normally-graded and laminated sand-to-mud layers 1-2 cm thick which resemble thin storm deposits or distal turbidites. Some sandstone layers sharply overlie mudstones and grade through laminated siltstone into mudstone (Fig. 5a). In the middle of the psammite sequence, a 50 m unit of rippled (?) quartzarenite may be a sand sheet deposit. A 3 m-thick laminated magnetite-chert iron-formation is 30 m below the top of the psammite section.

QUATERNARY GEOLOGY

The bedrock is extensively mantled (95%) by boulder fields, swampy meadows and a thin veneer of till. Crag-and tail, glacial flutings and rarer striae indicate north-northwest ice-flow. The tails through much of the area are remarkably long (0.5 to 3 km) and slim (~100 m), and are interpreted as formed by fast thin ice during the last stages of glaciation.

The crags expose amphibolite, quartzite, parts of foliated to massive large intrusions, komatiite, and magnetite-quartz iron-formation. The large granitoid plutons also form domal hills thinly covered by till. Recessive strata in low areas are exposed in glacial-fluvial meltwater channels recording northerly flow.

ECONOMIC GEOLOGY

The only anomalous geochemical samples from 1991 sampling are sulphidic sheared tonalite at locality **Aupy**, at the west-to-east transition from foliated tonalite, through sheared sulphidic tonalite (samples 91JP233a with 56 and 233e with 160 ppb gold) to interlaminated tonalitic and ultramafic mylonite on the west side of a small lake. A north-south gossan follows the mylonite, which has been misinterpreted several times in several places as metasediments.

In 1992, samples were systematically collected of till (62 for trace metals) and esker sand (30 for kimberlite indicators). Tills were collected both randomly and at 1-km spacings along transects located northwest (down-ice) of supracrustal belts. A variety of bedrock types, especially all known mineral showings, all sulphidic rocks (generally gossans), all facies of iron-formation, and the komatiites were also sampled for lithogeochemistry.

Locality **cpy** (central area) is a drilled showing of pyrrhotite-pyrite + 20% magnetite schlieren in an intensely strained, 20 m-thick amphibolite, transected by minor chalcopyrite veinlets (maximum 0.6 % copper in massive pyrrhotite; HAR claims, Boerner, 1972). Its intense aeromagnetic anomaly is merged with that of a magnetite-quartz iron-formation located about 500 m to the west. Boerner interpreted this amphibolite as a gabbroic marginal phase of the anorthosite. Alternatively it may be a hydrothermally altered magnetite-amphibolite iron-formation, tectonically mixed with basalt.

The following mineral deposit types (Eckstrand, 1984), listed in approximate decreasing order likelihood, are under review for their applicability to the study area:

- gold in shear zones
- gold in iron-formation
- Ni±copper in komatiites
- chromium in anorthosite
- zinc + lead + silver in shales
- zinc + platinum group elements in shales
- uranium + gold paleoplacers in conglomerates and quartzarenites.
- tin, tungsten and rare metals associated with fluorite granite.

SUMMARY OF STRATIGRAPHY AND TECTONICS

On Melville Peninsula the Prince Albert Group is typical of Archean greenstone belts; southwest toward Laughland Lake the strata are more mature and metavolcanic rocks less abundant (Schau, 1977). Campbell and Schau (1974) suggested that the group comprises a lower volcanic + greywacke + iron formation assemblage and an upper orthoquartzite assemblage.

In the Laughland Lake sparse younging indicators in the <5% exposure suggest that komatiites are low (central area) and high (western and southwestern areas) in the stratigraphy. Significant thicknesses of biotite psammite + iron-formation + komatiite appear to overlie quartzarenite + sericite phyllite in the western area. Probable structural complications make these stratigraphic suggestions very tentative.

Schau (1977, 1982, p. 32) and Schau and Ashton (1988) favoured terrestrial supracrustal deposition of the group: eruption of komatiites onto a craton; weathering of komatiites and gneissic uplands; and sedimentation of weathering-derived quartzarenites and iron-formations in lake basins. Campbell and Schau (1974) inferred shallow marine environments near Committee Bay, 200 km to the northeast.

Interpretation of the thinner quartzarenite units suggests marine shelves and epeiric seas (Johnson, 1978) and argues against a deep oceanic environment for the associated biotite psammites. But the psammite with common quartzarenite lenses and isolated lenses of conglomerate (submarine channel deposits?) suggest relatively deep water. The above interpretations can be reconciled if the group as a whole was deposited in the Archean equivalent of a modern continental prism such as is found on the east coast of North America, in compatibility with Schau and Ashton's (1988) rifting origin for the group.

The volcanic and shallow (quartzarenite) to deep marine (psammite) association documented here favours a rift-related hydrothermal exhalative origin and Superior-type classification for the iron-formations (cf. Gross, 1965).

The komatiites in the Laughland Lake area may thus be Archean analogues of continental tholeiites which are commonly associated with rifting of epicratonic quartzarenite + carbonate + Superior-type iron-formation. The central area komatiites, gabbro plugs and Laughland Lake Anorthosite are associated with major north-south faults which may record early grabens.

Despite the above broad agreement on tectonic setting, we have failed to find basement for the Prince Albert Group. The Brown River Gneiss was interpreted to be basement to the group (Schau, 1982; Jefferson and Schau, 1992) in part because strata adjacent to many of the foliated tonalites do not appear to have undergone contact metamorphism and some tonalites appear to be of higher grade than the group. On the other hand, neither has contact metamorphism been documented for the anorthosite complex which must be intrusive. Basal conglomerates have yet to be documented and contacts are faulted or mylonitic, including the inferred unconformity at locality **y** in the central area (Jefferson and Schau, 1992). Deformation recorded by steep mylonitic contacts may have telescoped contact metamorphic isograds, or been concentrated along unconformity surfaces. Subsequent metamorphism has also obscured evidence for basement versus intrusive relationships.

The intense multiple deformation in the few areas of good exposure suggests structural repetition of many of the units and, with the abundant plutonic rocks, collisional orogeny. Fabrics in the tonalites, gabbros, anorthosite and A' granites also record at least two intense deformations, one possibly during the Early Proterozoic. The youngest plutons, massive fluorite granites, are much like those around Wager Bay which have been dated at 1823 ± 3 and 1826 +4/-3 Ma by LeCheminant et al. (1987).

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Preliminary report of the geology of the southern Taltson magmatic zone, northeastern Alberta[1]

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Abstract: Northeastern Alberta is underlain by Early Proterozoic rocks of the southern Taltson magmatic zone (TMZ) and the adjacent Rae Province to the east. Taltson magmatic zone is underlain by a suite of banded gneisses of unknown age that constitute the Taltson basement complex, which is intruded by Early Proterozoic plutons of the zone.

Three principal north-south oriented shear zones dissect the southern Taltson magmatic zone. The Leland Lakes and Charles Lake shear zones are characterized by high grade mylonites of amphibolite to probable granulite grade, with mostly sinistral kinematic indicators. These are overprinted by greenschist and subgreenschist grade ductile and brittle shearing. The Andrew Lake shear zone contains southwest-dipping, well lineated ductile mylonites with kinematic indicators indicative of top-to-the-northeast shear such that Taltson magmatic zone basement is thrust in a dextral-oblique sense over Rae Province orthogneisses. Andrew Lake shear zone is a postulated tectonic boundary between Taltson magmatic zone and Rae Province.

Résumé : Le sous-sol de la partie nord-est de l'Alberta contient des roches du Protérozoïque précoce appartenant au sud de la zone magmatique de Taltson (TMZ) et aux roches de la province archéenne (?) de Rae adjacente, à l'est (74M/14, 15, 16). Le sous-sol de la TMZ contient une série de gneiss rubanés d'âge inconnu qui constituent le socle métamorphique de Taltson, traversé par des plutons de la TMZ datés du Protérozoïque précoce.

Trois grandes zones de cisaillement d'orientation nord-sud dissèquent le sud de la TMZ. Les zones de cisaillement de Leland Lakes et de Charles Lake sont caractérisées par des mylonites ductiles qui se situent dans le faciès des amphibolites et probablement jusque dans le faciès des granulites, et les indicateurs cinématiques sont surtout sénestres. Ceux-ci ont subi un métamorphisme correspondant au faciès et au sous-faciès inférieur des schistes verts, ainsi qu'un cisaillement ductile et un cisaillement cassant. La zone de cisaillement d'Andrew Lake contient des mylonites ductiles de pendage sud-ouest, à linéations bien définies et à indicateurs cinématiques révélant un cisaillement dont le sommet se situe au nord-est, de sorte que le socle de la TMZ chevauche dans un sens dextre oblique les orthogneiss de la Province de Rae. La zone de cisaillement d'Andrew Lake constitue sans doute une limite tectonique entre la TMZ et la Province de Rae.

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INTRODUCTION

This study was initiated to elucidate the tectonic, intrusive, and pressure-temperature-time histories of the southern Taltson magmatic zone of northeastern Alberta to allow further evaluation of its metallogenic potential as part of the Canada-Alberta Mineral Development Agreement. The geology of three map areas in the northeastern corner of the province has been re-evaluated by carrying out systematic 1:50 000 scale mapping in the Tulip, Leland, Mercredi, Charles, and Andrew lakes areas (74M/14, 15, 16). The mapping was supported by boat and fixed wing aircraft, and used the published geological maps of Godfrey and Langenberg (1986) and Godfrey (1966, 1986) as a basis. Results of the new mapping were reported by McDonough et al. (in press a, b, c).

GEOLOGICAL SETTING

The Canadian Shield of northeastern Alberta is mainly underlain by Early Proterozoic rocks of the southern Taltson magmatic zone, which forms the westernmost exposed part of the former Churchill Province. The exposed portion of the Taltson magmatic zone (TMZ) constitutes a 300 km long belt of granitoids, amphibolites, and metasedimentary rocks that form part of a composite Early Proterozoic Andean-type continental magmatic arc and continental collisional orogen (Hoffman, 1987, 1988; Bostock, 1988; Thériault, 1992), which extends on the surface from Great Slave Lake southward to Lake Athabasca (Fig. 1). Magmatic rocks of the zone are considered to be the southern counterpart of 2.0 to 1.9 Ga magmatic rocks of the Thelon tectonic zone, north of the Great Slave Lake shear zone (Fig. 1; Hoffman, 1987, 1988; Bostock, 1988; Culshaw, 1991; van Breemen et al., 1987; James et al., 1988).

Taltson magmatic zone has a distinctive aeromagnetic signature with strong north-trending positive anomalies that flank a central low (Geological Survey of Canada, 1987). Taltson magmatic zone magnetic anomalies are continuous beneath Phanerozoic cover as far south as the buried western extension of the Snowbird tectonic zone under the Western Canada basin (Hoffman, 1987; Hanmer et al., 1991; Ross et al., 1991).

The northern Taltson magmatic zone (north of 60°N) is characterized by 1.99 Ga I-type magmatic rocks, and a predominance of 1.96 to 1.92 Ga S-type plutonic rocks

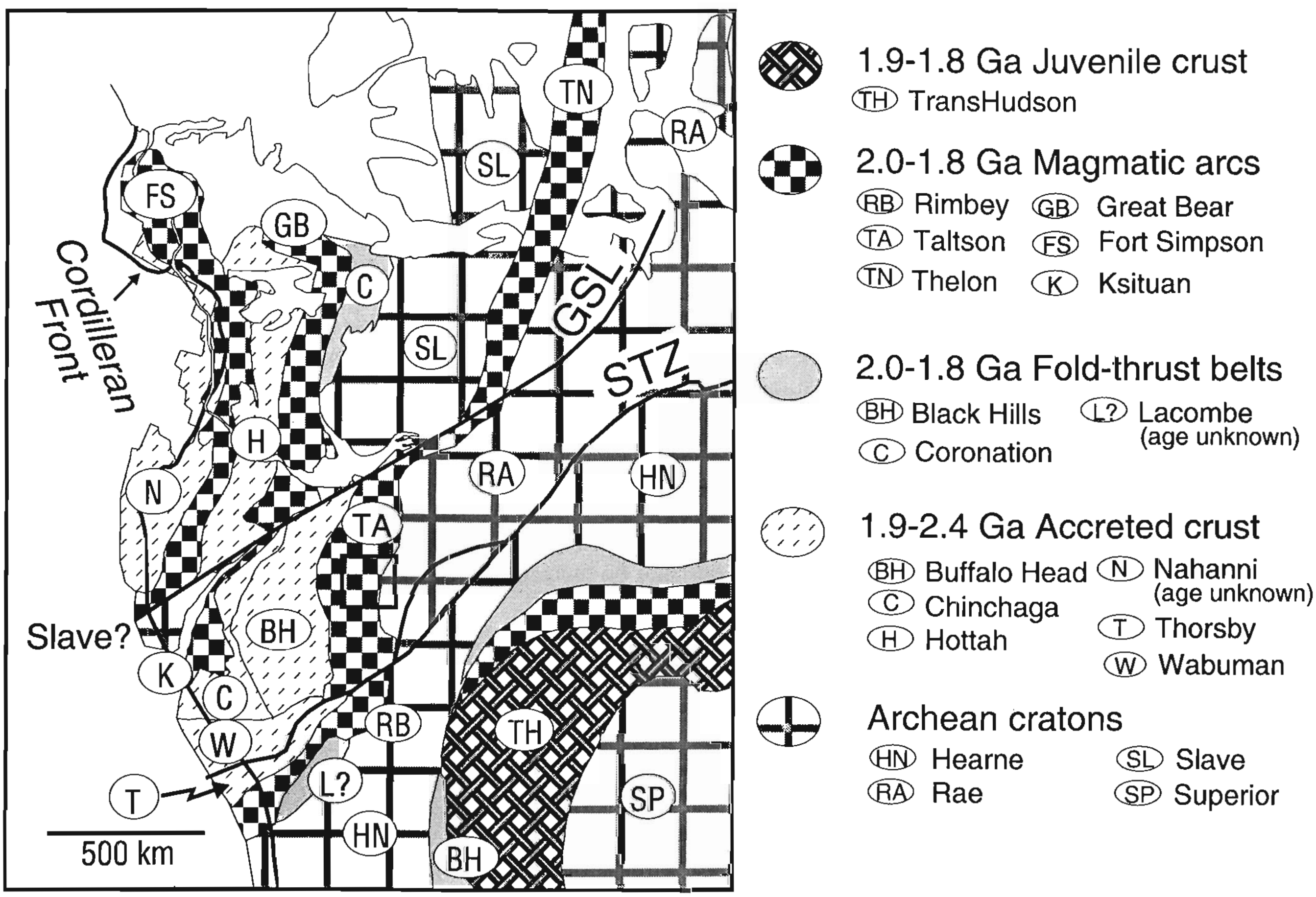

Figure 1. Generalized geological map of western Laurentia (exposed units after Hoffman, 1988, and subsurface units after Ross et al., 1991). GSL = Great Slave Lake shear zone; STZ = Snowbird tectonic zone.

(Bostock et al., 1987; Bostock and Loveridge, 1988) having Nd isotopic signatures indicative of crustal genesis (Thériault, 1992). The Benna Thy granite (1906 Ma) represents the youngest plutonism associated with deformation in Taltson magmatic zone (Bostock and Loveridge, 1988). The Thekuthili stock (1813 Ma) postdates deformation and magmatism in the zone (Bostock et al., 1991).

Taltson magmatic zone is bordered on the east by Rae Province (Hoffman, 1988, 1989), which extends over a broad area eastward to the New Quebec orogen and northward to the Arctic archipelago (Fig. 1). Rae Province is dominated by 2.8 to 2.6 Ga felsic magmatic rocks and lesser gneisses more than 3.0 Ga in age. The western margin of Rae Province adjacent to northern Taltson magmatic zone is characterized by plutonic rocks having 2.44 to 2.27 Ga U-Pb crystallization ages (Bostock and Loveridge, 1988; Bostock et al., 1991; van Breemen et al., 1992), that are possibly related to the eastward subduction of oceanic crust beneath the western edge of Rae Province (Bostock, 1989; Hoffman, 1989). In the west, the zone is flanked in the subsurface by the Buffalo Head terrane (Fig. 1), which is a dominantly magmatic terrane characterized by 2.46 to 2.82 Ga depleted mantle model ages and 2.33 to 1.99 Ga U-Pb crystallization ages (Ross et al., 1991; Thériault and Ross, 1991).

Taltson magmatic zone is transected by two shear zones of variable metamorphic grade, previously called the Warren and Allan shear zones (Godfrey, 1958; Bostock, 1982), but here called the Leland Lakes and Charles Lake shear zones (Fig. 2; see McDonough et al., 1992a, 1992b) after localities in the southern TMZ where they are well exposed. Hoffman (1988, 1989) and Hanmer et al. (1992) have postulated that the major TMZ shear zones represent zones of lateral tectonic escape that accommodated indentation of the Slave craton into the Rae craton.

LITHOLOGICAL UNITS

All of the Taltson magmatic zone and Rae Province units described below have been collected for U-Pb dating and Nd isotopic analysis, with the exception of leucogranite mylonites of the Charles Lake shear zone.

Talston magmatic zone

The southern portion of the zone is dominated by plutonic rocks of the Slave, Arch Lake, and Charles Lake suites, which have intruded an older package of gneisses, here referred to as the Taltson basement complex. The Slave granites are divided into western and eastern suites.

Taltson basement complex (TBC)

A lithologically diverse suite of banded to massive gneisses of unknown age, including hornblende-biotite granodiorite and granite gneiss, quartz diorite gneiss, amphibolite, paragneiss, and mylonitic gneiss, is exposed throughout the central part of the mapped transect (Fig. 2, 3). These banded gneisses are enveloped by and intruded by various plutonic rocks of the Taltson magmatic zone (see below), and therefore form the basement of the zone, comprising the Taltson basement complex (TBC). The principal Taltson basement complex banded gneisses were mapped as distinct hornblende and biotite granite gneiss units by Godfrey and coworkers; we have combined these into one unit.

In shear zones, banded gneisses of the complex are deformed into large oblate sheath-folds that are refolded by younger shortening (Fig. 4a). Taltson basement complex gneisses are extensively intruded by medium- to coarse-grained pink granites and pegmatites (Fig. 4b). They occur as 10 to 50 cm wide dykes that are coplanar to discordant with gneissic banding, and as small irregularly shaped intrusions, constituting up to 50% of Taltson basement complex outcrop. They are lithologically similar to granites of the Western Slave granite suite (see below), with which they are tentatively correlated. Locally, the banded gneisses are highly sheared and dismembered (Fig. 4c).

Numerous bodies of high grade pelitic and quartzose metasedimentary rocks are enveloped by Taltson basement complex banded gneisses, and occur as rafts within plutons of the zone (Fig. 2). They may have formed a cover sequence to the Taltson basement complex. Their distribution, metamorphic grade, and characteristic mineral assemblages are described by Grover et al. (1993).

Western Slave granite

A large pluton of granite and monzogranite that outcrops between Leland Lakes shear zone and Slave River (Fig. 2), mapped as Slave granite by Godfrey and Langenberg (1986), corresponds to an aeromagnetic low (Geological Survey of Canada, 1987). We remapped the eastern half of this pluton, and refer to it as the Western Slave granites. Western Slave granites comprise medium- to coarse-grained, massive to moderately foliated, quartz monzonites, monzogranites, and granites that vary in colour from white to pink. Small clots of mafic minerals are reported to be assemblages of garnet, biotite, hercynite, and cordierite (Nielsen et al., 1981).

Locally, Western Slave granites have abundant rafts of pelitic and quartzose paragneisses. Weakly to moderately foliated granites, interpreted to constitute a late phase(s) of Western Slave granites, form extensive dyke swarms in the Taltson basement complex (Fig. 4b), and intrude high grade mylonites of the Leland Lakes shear zone (Fig. 4d).

Slave granites from northern Taltson magmatic zone have U-Pb crystallization ages ranging from 1960 to 1955 Ma (Bostock et al., 1987, 1991; Hanmer et al., 1992), however the dated pluton is not continuous with the western Slave pluton we mapped that straddles the Alberta-Northwest Territories border.

Eastern Slave granite

A pluton of massive to weakly foliated, medium- to coarse-grained granite, distinguished by feldspar phenocrysts in an equigranular matrix, termed the "raisin granite" phase of the Slave granites by Godfrey and Langenberg (1986),

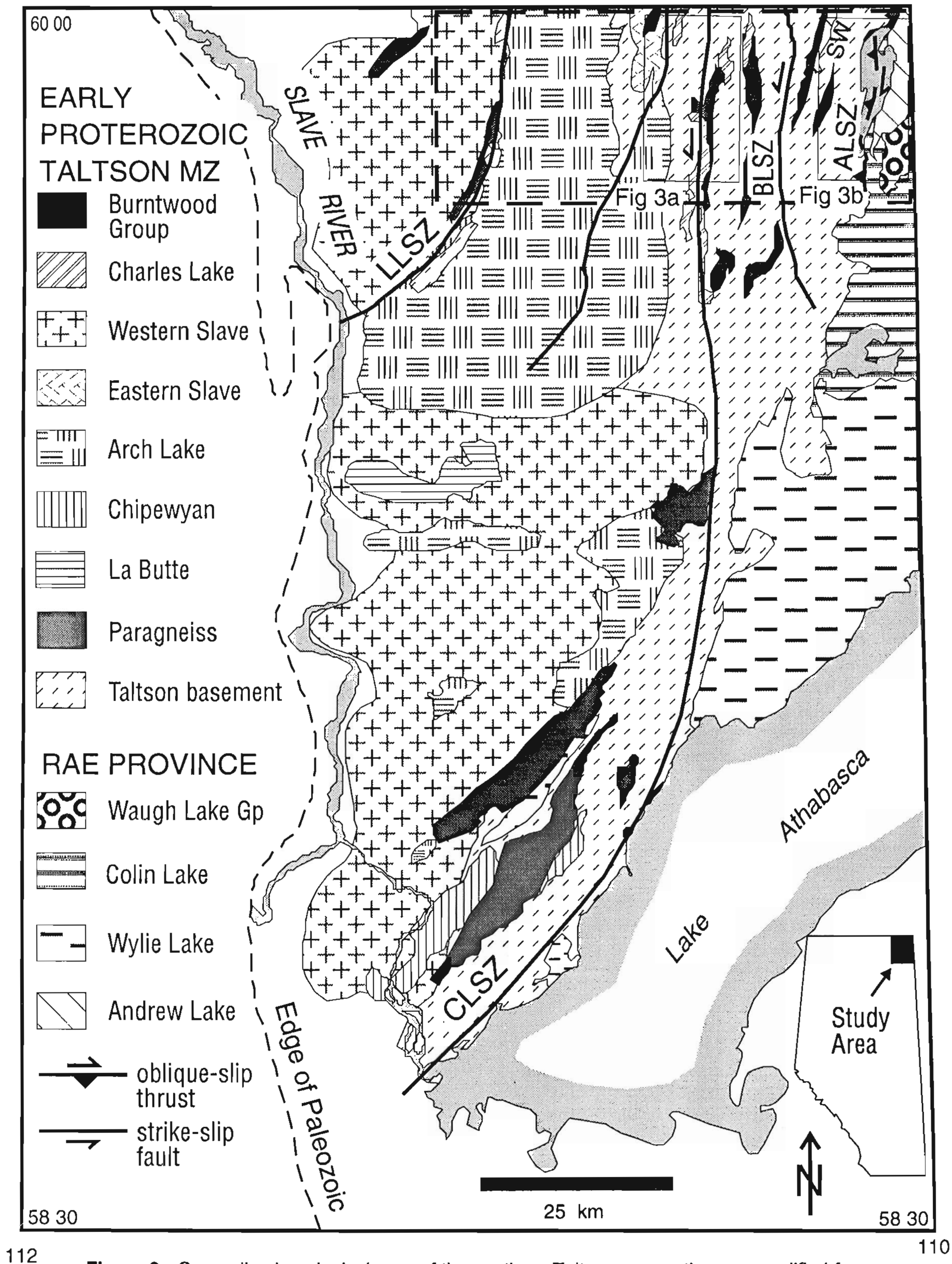

Figure 2. Generalized geological map of the southern Taltson magmatic zone modified from Godfrey (1986). ALSZ = Andrew Lake shear zone; BLSZ = Bayonet Lake shear zone; CLSZ = Charles Lake shear zone; LLSZ = Leland Lakes shear zone. Units in legend identified by a place name are granitoids. No significance is attached to the relative ages of units not discussed in the text. Dashed rectangle outlines the transect mapped in 1992; solid rectangles give locations for Figure 3.

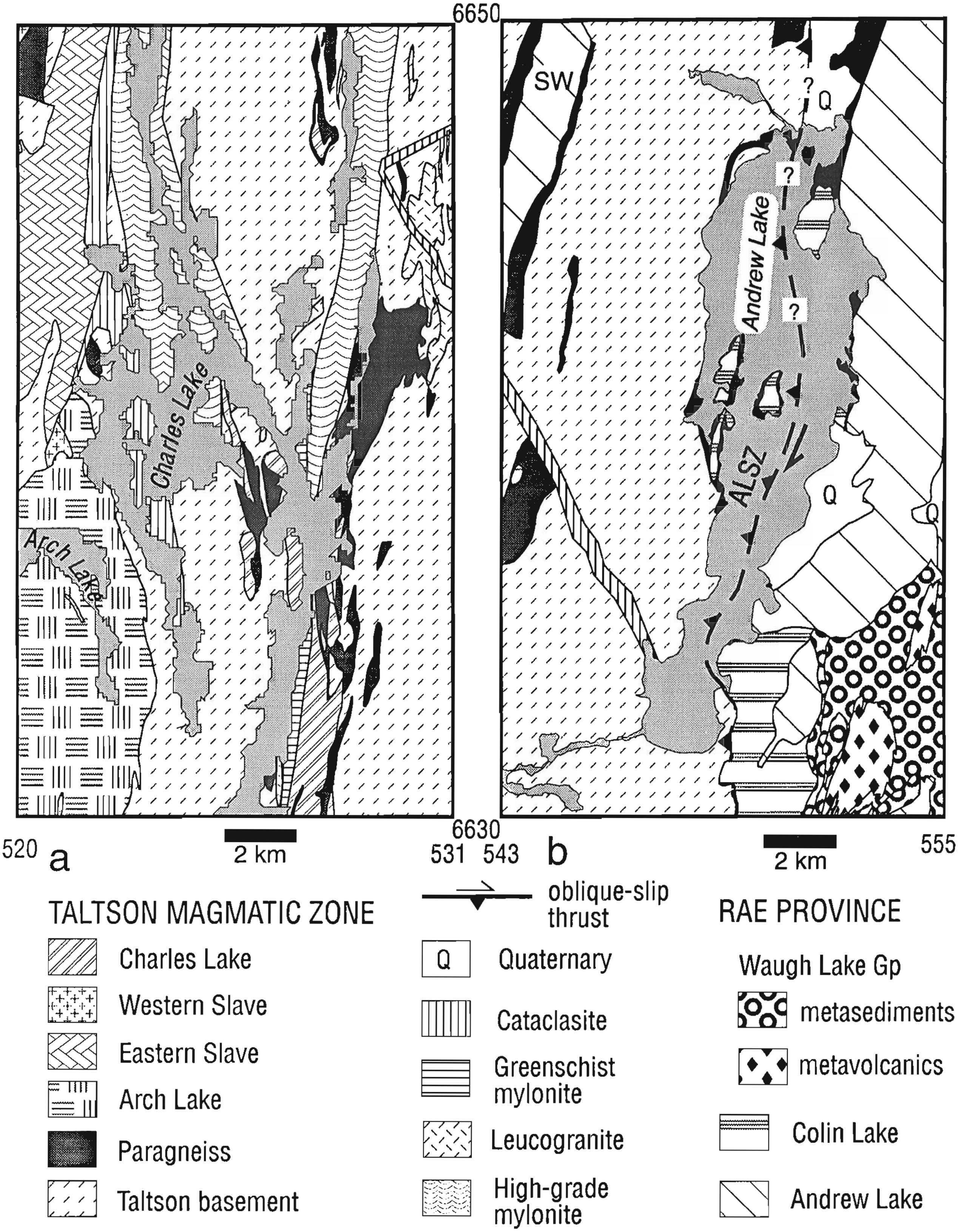

Figure 3. Geological maps of parts of the transect mapped in 1992. **a**) Charles Lake area. **b**) Andrew Lake area. See Figure 2 for locations. Coordinates are UTM grid references.

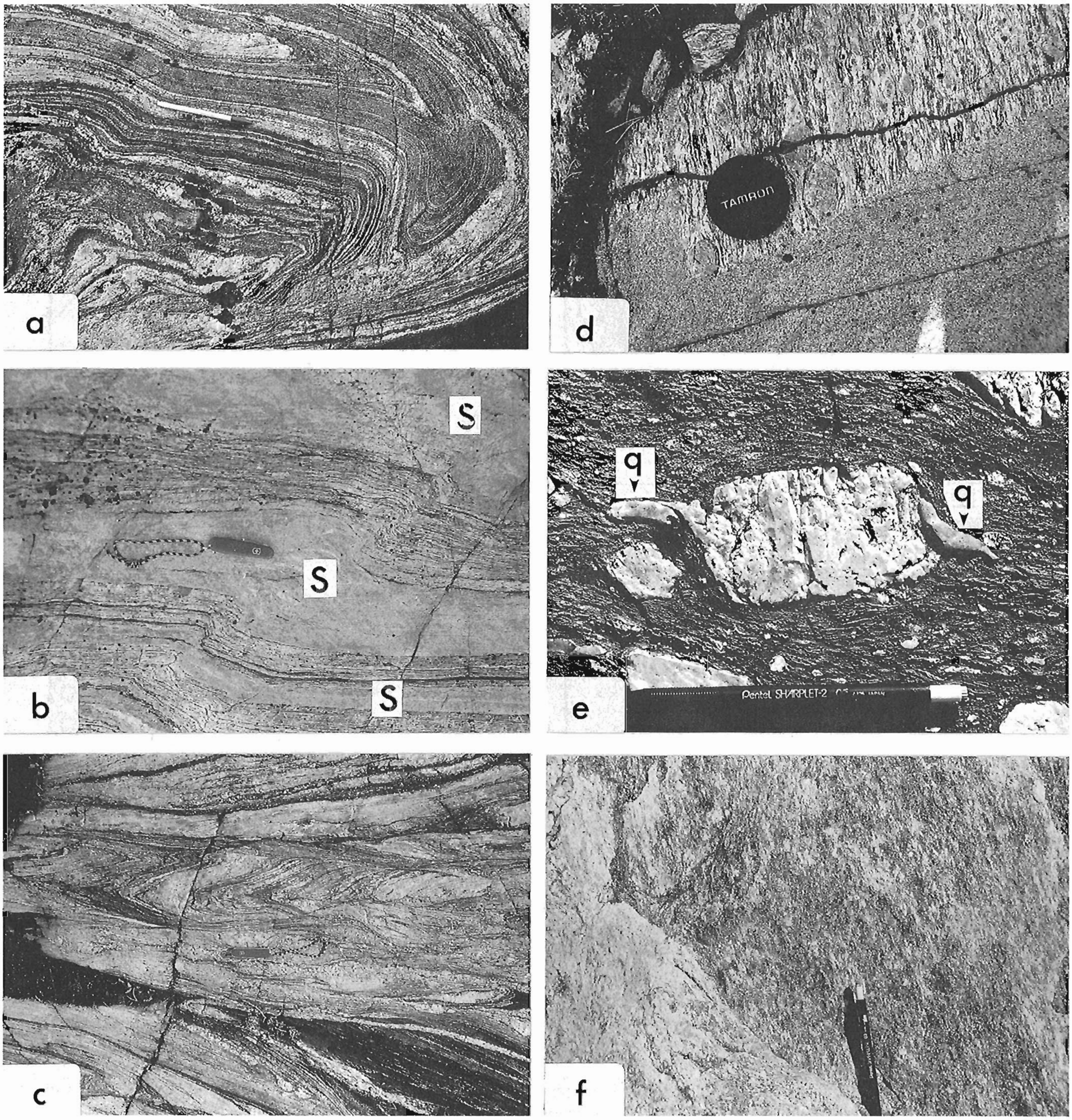

Figure 4. Outcrop photographs of basement and granitoids. **a**) Banded gneiss of Taltson basment complex deformed into refolded flattened sheath-folds. Pen is 16 cm in length. ISPG 4009-2. **b**) Mylonitic Taltson basement gneiss with coplanar to slightly discordant Slave? granite dykes (S). Knife is 9 cm in length. ISPG 4009-33. **c**) Sheared, dismembered gneiss of Taltson basement complex with tectonic wedges. Knife is 9 cm in length. ISPG 4009-32. **d**) Arch Lake granitic augen gneiss marginal to Leland Lakes shear zone, cut by weakly foliated Slave granite dyke. Lens cap is 6.7 cm in diameter. ISPG 4009-36. **e**) K-feldspar megacryst in Charles Lake granite with pseudo-tails of quartz (q). Pencil is 15 cm in length. ISPG 4009-8 (photo by K. Kelly). **f**) Lineated Andrew Lake granodiorite cut by Colin Lake white muscovite granite (lower left). Pencil (15 cm) is parallel to lineation in granodiorite. ISPG 4009-34.

outcrops west of Charles Lake shear zone (Fig. 2, 3). We call these granites the Eastern Slave granites. They contain approximately 30 to 40% equant (1-4 cm) K-feldspar phenocrysts within a fine- to medium-grained matrix of quartz, feldspar, and biotite. Garnet is abundant where significant quantities of paragneiss xenoliths are present. Eastern Slave granites contain xenolithic rafts of banded basement rocks, foliated mylonitic rocks (Charles Lake shear zone?), paragneisses, and amphibolite, and clearly intrude banded Taltson basement complex gneisses. It is not clear if the Eastern Slave granites can be correlated with any of the dated plutonic lithologies of the northern Taltson magmatic zone.

Arch Lake granite

A large pluton of variably deformed megacrystic K-feldspar syenogranite to granite (Fig. 2, 3), called the Arch Lake plutonic suite (Godfrey and Langenberg, 1986), underlies the central part of the mapped transect. The Arch Lake granitoids comprise a suite of foliated K-feldspar augen gneiss (Fig. 4d) to weakly foliated K-feldspar megacrystic biotite-hornblende granite with 30 to 50% lenticular, 1 x 3 cm K-feldspar crystals contained in a matrix of biotite, quartz, feldspar, and rare garnet. Mica inclusions in K-feldspar are rare. Blue quartz forms the lineation in L-S tectonites of the Arch Lake suite.

Goff et al. (1986) characterized the Arch Lake suite as S-type peraluminous quartz monzonites to granites. The Arch Lake suite is lithologically similar to and continuous with the megacrystic Konth syenogranite suite (1935 Ma; Bostock et al., 1987, 1991) of the northern Taltson magmatic zone, which is also an S-type peraluminous suite (Thériault, 1992).

Along the east edge of the Arch Lake pluton, 0.5 x 3 m rafts of heterolithic banded Taltson basement complex gneisses are contained in an equigranular granite that forms a marginal border phase of the pluton. These are interpreted as large xenoliths, suggesting that the Arch Lake suite intruded the Taltson basement complex. Arch Lake granites have been mapped into, and form part of, the high grade mylonites in Leland Lakes shear zone, where they are cut by Slave? granite dykes (Fig. 4d). They also form high grade L-S tectonites on the western margin of the Charles Lake shear zone.

Leucogranite

Small bodies of white to pink, medium- to coarse-grained leucogranite intrude basement gneisses in the Charles Lake area. They vary from weakly foliated granite to greenschist grade mylonite and protomylonite. Greenschist grade mylonite contains large feldspar inclusions with sinistral σ-type wings (see Hanmer and Passchier, 1991), and intrudes high grade sinistral mylonites of the Charles Lake shear zone.

Charles Lake granite

Distinctive K-feldspar megacrystic biotite-rich granite to syenogranite outcrops near Charles Lake (Granite F of Godfrey, 1966, 1986). Charles Lake granite (Fig. 4e) is typified by 10 to 20% large subhedral to euhedral K-feldspar megacrysts up to 8 x 12 cm in size, with randomly oriented inclusions of biotite that may be magmatic in origin. K-feldspar megacrysts are contained in a dark grey to black, fine grained matrix of biotite, chlorite, quartz, feldspar, and rare garnet. Locally, the matrix is coarse grained and muscovite-bearing. Charles Lake granite is deformed into transitional amphibolite to greenschist grade mylonites in the Charles Lake shear zone. Charles Lake granite lacks a dated counterpart in northern Taltson magmatic zone.

Rae Province

Andrew Lake granodiorite

A suite of well foliated, massive, granodiorite to quartz diorite feldspar augen gneiss (Fig. 4f) outcrops to the east of Andrew Lake, comprising the Andrew Lake suite (Fig. 2, 3). These rocks also outcrop near Swinnerton Lake where they are encased in banded gneiss and paragneiss of the complex, which we tentatively interpret as a tectonic window, the Swinnerton window. Orthogneisses of the Andrew Lake suite are lithologically similar to granodioritic augen gneisses of the Mixed Gneisses of Rae Province adjacent to the east margin of northern portion of the zone (e.g., Yatsore augen gneiss), which have U-Pb crystallization ages ranging from 2.27 to 2.44 Ga (Bostock and Loveridge, 1988; Bostock et al., 1991; van Breemen et al., 1992).

Andrew Lake suite orthogneisses are locally intruded by weakly deformed, muscovite-bearing, white leucogranites (see below; Fig. 4f), and by dykes of medium grained pink granite similar to dykes that intrude Taltson basement complex gneisses.

Colin Lake white granite

In the Andrew Lake area, massive, medium- to coarse-grained, weakly foliated, white muscovite-bearing granite intrudes orthogneisses of the Andrew Lake suite (Fig. 4f), as well as Taltson basement complex gneisses. These granites were mapped as part of the Colin Lake suite (Rae Province) by Godfrey (1961, 1963, 1986). Field relationships clearly indicate that Colin Lake granites intrude older Andrew Lake granodiorite augen gneisses, and truncate both planar and linear ductile mylonite fabrics in the Andrew Lake shear zone (Fig. 4f). Colin Lake granites probably correlate with the 1935 Ma Natael muscovite granite of northern Taltson magmatic zone (Bostock et al., 1987).

Waugh Lake Group

A sequence of lower greenschist facies, metasedimentary and metavolcanic rocks form a 40 km^2 outlier in the western Rae Province in the vicinity of Waugh Lake (Watanabe, 1961; Godfrey, 1963, 1986; Godfrey and Langenberg, 1978). Metasediments include chlorite phyllite and schist, quartzose chlorite phyllite, quartzite, and foliated feldspathic pebble conglomerate. Metavolcanics include massive chlorite and chlorite-biotite schist. Relic pyroxene phenocrysts in metavolcanics are altered to epidote plus actinolite.

North of the west arm of Waugh Lake, the contact between Waugh Lake Group sediments and mylonitic Rae gneisses is characterized by volcanic breccias with large angular blocks of mylonitic gneiss enveloped by massive chloritized volcanics, forming an angular unconformity. The basal volcanics are overlain by thin bedded sandstones and siltstones with abundant small scale extension faults. At 1:50 000 scale, the contact is coplanar to a zone of thermally annealed mylonites in Rae gneisses.

A K-Ar biotite date of 1760 Ma from Waugh Lake metavolcanics (Baadsgaard and Godfrey, 1972) provides a younger age limit for development of the Waugh Lake basin.

Shear zones

Three major north-south oriented high grade mylonite zones transect the southern Taltson magmatic zone. From west to east these are the Leland Lakes, Charles Lake, and Andrew Lake shear zones (Fig. 2, 3). A fourth shear zone, the Bayonet Lake shear zone (Fig. 2), is a splay of the Charles Lake shear zone (see Bostock, 1982). Godfrey and Langenberg (1978) suggested that shear zone deformation in northeast Alberta occurred at greenschist grade because the mineralogy of the mylonites is largely quartz+feldspar, and diagnostic high grade mineral assemblages generally are not present. However, we postulate that high grade mylonitization in Leland Lakes and Charles Lake shear zones initially took place at upper amphibolite to possibly granulite facies conditions for the following reasons: 1) feldspars in L-S tectonites show evidence of extreme ductile flow, which probably took place at temperatures in excess of 600°C; 2) sparsely preserved ductile stretching lineations in mylonites probably formed under high grade conditions and were overprinted and largely destroyed by younger (greenschist and sub-greenschist grade) recrystallization and cataclasis; 3) syntectonic leucosome development during mylonitization suggests high grade conditions; and 4) pelitic paragneisses with assemblages of garnet+sillimanite+K-feldspar±cordierite are involved in the Leland Lakes shear zone (Grover et al., 1993). High grade ductile deformation in the western two shear zones is characterized by both sinistral and dextral, bulk, non-coaxial shear, based on kinematic indicators observed in 1992. Greenschist grade, brittleductile, sinistral and dextral shearing, and shortening, overprint parts of the Leland Lakes and Charles Lake shear zones.

In the northern portion of the zone, the boundary between Taltson magmatic zone and Rae Province is defined by a series of mylonite zones near Gagnon Lake, interpreted by Bostock (1988, 1992) to be a composite zone of ductile dextral shear that separates Taltson magmatic zone plutons from mixed gneisses of the western Rae Province. This zone lies along strike of a zone of sinistral mylonites that form the northern continuation of the Charles Lake shear zone (see Bostock, 1982, 1988; Bostock et al., 1991). In the southern portion of the zone, we have mapped the Taltson-Rae boundary farther to the east in the vicinity of Andrew Lake (Fig. 2, 3). There, the boundary is a newly recognized shear zone that is largely covered and obscured by younger intrusions, here called the Andrew Lake shear zone. This shear zone separates Taltson basement complex gneisses on the west shore from granodioritic augen gneisses of the western Rae Province on the east shore.

Leland Lakes (Warren) shear zone (LLSZ)

The Leland Lakes shear zone (LLSZ) is a 3 to 5 km wide zone characterized by vertical to subvertical high grade mylonites with subhorizontal lineations (strike-lineated mylonites) developed at amphibolite to probable granulite grade. High grade mylonites were overprinted by sinistral, greenschist grade, strike-lineated mylonites on the western margin of the shear zone (Fig. 5).

The high grade mylonites have undergone a considerable degree of recrystallization and as a result generally lack lineation. Preserved quartz rods are subhorizontal, but have a reoriented internal quartz grain fabric that plunges downdip. Pull-aparts of amphibolite layers that now form pods spaced 3 to 5 m apart are the best remaining evidence of subhorizontal extension. These are interpreted to indicate a strong component of horizontal bulk stretching during high grade deformation in Leland Lakes shear zone.

Using sparse subhorizontal lineations and amphibolite pull-aparts as a guide, kinematic indicators in high grade mylonites of the shear zone are both sinistral and dextral (Fig. 5a, b). C-S fabric, σ-type winged inclusions, shear bands, and small localized shear zones associated with granitic dykes are predominantly sinistral. Dextral kinematic indicators include σ-type and σ-type winged inclusions. Dextral strike-slip duplexes (e.g., Woodcock and Fischer, 1986) laterally imbricate high grade mylonite.

Arch Lake granites (Konth equivalent, 1935 Ma) are deformed into strike-lineated mylonites in the shear zone. Western Slave granite, including a local dip-lineated (quartz) graphic granite phase, postdates deformation associated with the high grade mylonites in Leland Lakes shear zone. Along the western margin of the shear zone, Slave granites are deformed into greenschist facies, strike-lineated mylonites with sinistral C-S fabrics (Fig. 5c).

Charles Lake (Allan) shear zone (CLSZ)

The Charles Lake shear zone (CLSZ) is a 5 to 8 km wide zone composed of anastomosing zones of vertical, high grade (upper amphibolite to possible granulite facies) mylonites with largely sinistral kinematics, overprinted by narrow zones of amphibolite to greenschist grade mylonites, and younger brittle faulting. As with Leland Lakes shear zone, mylonites in Charles Lake shear zone were largely recrystallized during greenschist grade deformation. In addition, brittle faulting has induced significant cataclasis of older mylonites and, as a result, linear stretching fabrics are rarely preserved.

Sparsely preserved, subhorizontal, stretching lineations and amphibolite pull-aparts suggest a substantial subhorizontal bulk stretch parallel to the Charles Lake shear zone, permitting kinematic significance to be attached to asymmetrical planar fabrics. Kinematic indicators in high grade mylonites are largely sinistral (Fig. 5d, e), including

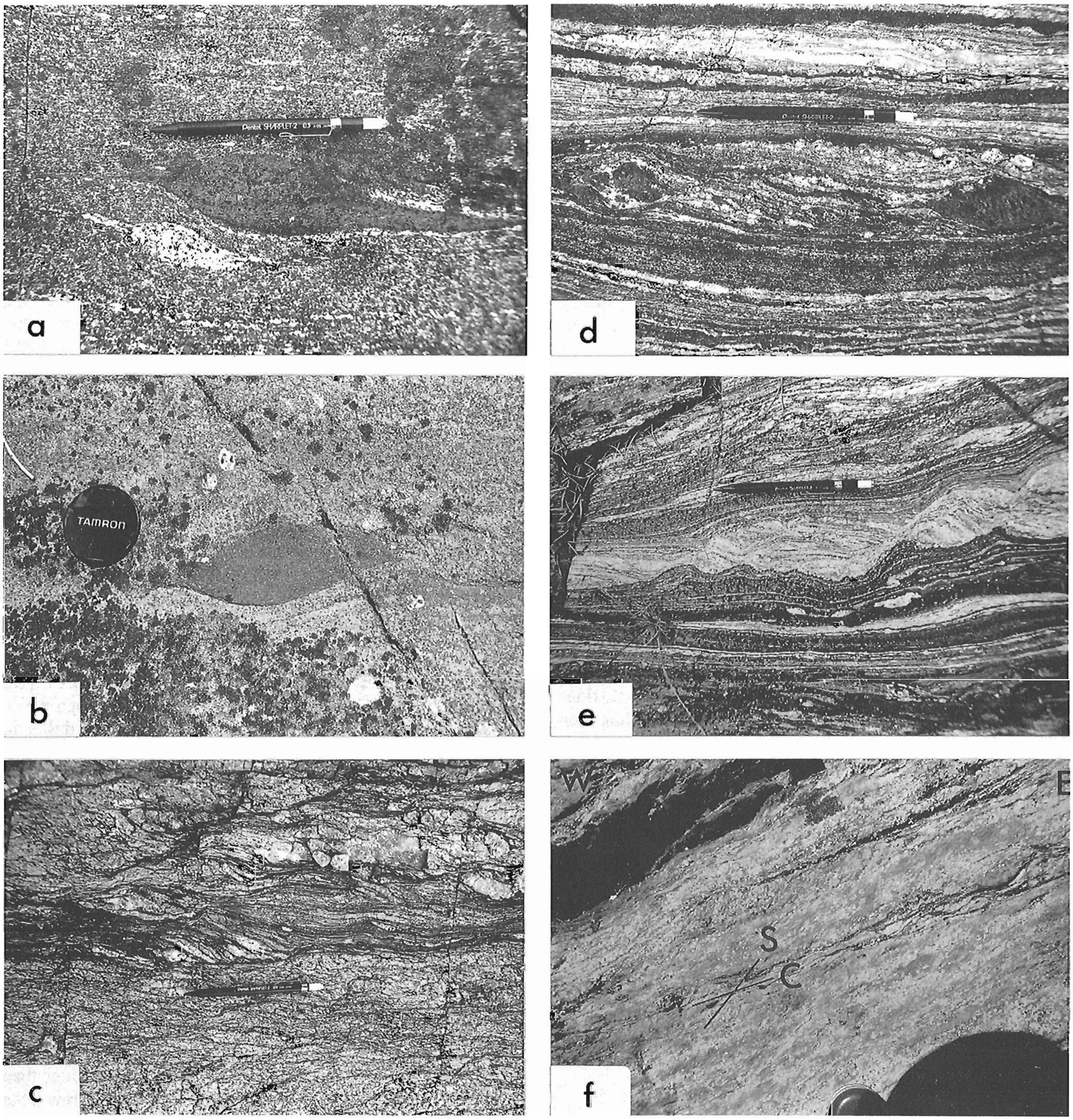

Figure 5. Outcrop photographs of shear zones. **a**) Sinistral σ-type winged inclusions of amphibolite and pegmatite in mylonite of Leland Lakes shear zone. Pencil is 15 cm in length. ISPG 4009-38. **b**) Dextral σ-type inclusion of amphibolite in Leland Lakes shear zone. Lens cap is 6.7 cm in diameter. ISPG 4009-30. **c**) Strike-lineated greenschist grade mylonite with sinistral C-S fabric, western margin of Leland Lakes shear zone. Pencil is 15 cm in length. ISPG 4009-26. **d**) High grade mylonite with sinistral foliation fish and σ-type inclusions in Charles Lake shear zone. Pencil is 15 cm in length. ISPG 4009-16. **e**) Sinistral asymmetrical shears in leucosome boudins in high grade mylonite of Charles Lake shear zone. Pencil is 15 cm length. ISPG 4009-22. **f**) View to north of vertical fracture surface oriented parallel to southwest plunging stretching lineation in Andrew Lake shear zone showing top-to-the-northeast C-S fabric. Width of knife tip is 1.5 cm. ISPG 4009-18.

C-S fabric, σ-type winged inclusions, shear bands, strike-slip duplexes, and small localized shear zones associated with granitic dykes and leucosomes. High grade mylonites in this shear zone also exhibit lateral tectonic wedging of oblate isoclinal sheath-fold closures with sinistral and dextral displacement on either side of the wedges.

Arch Lake granite (Konth equivalent, 1935 Ma) forms high grade L-S tectonite near the western margin of the shear zone. Charles Lake granite (undated) was deformed into transitional greenschist to greenschist mylonites (Fig. 3), with the grade of mylonite decreasing toward the eastern margin of the zone. Greenschist grade deformation outlasted intrusion of Charles Lake granite. Greenschist grade mylonitic leucogranite (undated) intruded the highest grade (granulite?) mylonites in the zone.

Brittle crush zones with intense cataclasis are found in the shear zone. The rocks in these zones are extensively silicified, indicating large volumes of fluid infiltration.

Andrew Lake shear zone (ALSZ)

At three mapped localities within and southeast of Andrew Lake, Taltson basement complex banded gneisses and Andrew Lake granodiorite have been strongly deformed into very well lineated, high grade (upper amphibolite?) mylonites, defining the Andrew Lake shear zone (ALSZ; Fig. 2, 3). Planar fabrics dip moderately to the southwest, and have a pronounced downdip stretching lineation with a dextral obliquity. C-S fabrics and σ-type winged inclusions observed in the field and in slabs cut parallel to the lineation indicate that Taltson magmatic zone basement has been thrust to the northeast in a dextral oblique sense over granodioritic gneisses of the Andrew Lake suite (Fig. 5f). We postulate that the Andrew Lake shear zone forms the boundary between the Taltson magmatic zone to the west, and mixed gneisses of the Rae Province on the east. The Andrew Lake shear zone may merge with the Tazin River shear zone in the northern Taltson magmatic zone (Bostock, 1982; H.H. Bostock, pers. comm., 1992).

High grade mylonitization in Andrew Lake shear zone predates intrusion of white muscovite granite of the Colin Lake suite, a probable correlative of the 1935 Ma Natael granite (Bostock et al., 1987) of northern Taltson magmatic zone.

ECONOMIC GEOLOGY

Disseminated pyrite locally forms from 2 to 4% of Taltson basement complex gneisses, younger intrusions (Slave granite?) in basement complex gneisses, and fine grained feldspar porphyry, a body that is a probable equivalent of Charles Lake granite.

Cataclasis of mylonites in the Charles Lake and Leland Lakes shear zones is accompanied by silicification. Rusty weathering quartz breccias and quartz stockwork related to silicification are locally important in Leland Lakes and Charles Lake shear zones. The occurrences studied to date are devoid of sulphides, however, large volumes of fluid infiltration are probable, suggesting that these zones could bear important sulphide occurrences along strike.

DISCUSSION

A package of strongly deformed granitic gneisses outcrops on either side of the Charles Lake shear zone, forming the oldest rocks of the Taltson magmatic zone (Fig. 2; Godfrey, 1966, 1986; McDonough et al., in press b, c). This package is intruded by Taltson plutons forming the basement of the southern portion of the zone; therefore, we refer to it as the Talston basement complex. Our preliminary interpretation of the location of the Taltson-Rae boundary along the Andrew Lake shear zone differs significantly from the boundary mapped to the north by Bostock (1982, 1989). He placed this boundary along the northern continuation of the Charles Lake shear zone, and thus, by his interpretation, the basement complex gneisses would form part of the western margin of Rae Province. Two equally valid tectonic interpretations for the Taltson basement complex are: 1) Taltson basement complex is a distinct tectonic entity separated from Rae Province by the Andrew Lake shear zone; or 2) Taltson basement complex is part of western Rae basement upon which the Taltson continental magmatic arc evolved.

The field relationships discussed above indicate that the Arch Lake granites in the Leland Lakes and Charles Lakes shear zones were deformed into strike-lineated, high grade mylonites and were intruded by Slave(?) granitic dykes in Leland Lakes shear zone. The weakly foliated nature of dykes and small intrusions in Leland Lakes shear zone and the Taltson basement complex suggests that these granites may represent a younger phase(s) of Slave granites. U-Pb crystallization ages for Slave granite of northern Taltson magmatic zone range from 1960 to 1955 Ma, and are older than ages for Konth syenogranite (Arch Lake equivalent; 1935 Ma; Bostock et al., 1987, 1991; Bostock and Loveridge, 1988). Correlation of the Arch Lake pluton with the Konth pluton is probably more certain than the correlation of younger granite dykes with Slave granites of northern Taltson magmatic zone, suggesting that high grade mylonites in Leland Lakes shear zone formed in part after 1935 Ma. High grade sinistral mylonites of Charles Lake shear zone are also younger (in part) than the Arch Lake suite, and older than undated leucogranite and Charles Lake granite. Andrew Lake granodiorite is deformed into mylonite in Andrew Lake shear zone, which is cut by Colin Lake white granite (correlated with Natael granite; 1935 Ma). Therefore, Andrew Lake shear zone is probably older than 1935 Ma. New geochronology of plutonic rocks of southern Taltson magmatic zone will provide further age constraints on shear zone evolution.

Close spatial and temporal relationships of north-south zones of sinistral and dextral shear have been suggested as a model by which the tectonic escape of slices of Rae crust away from the lateral face of the the Slave Province indentor was accommodated (Hoffman, 1988; Bostock, 1989; Hanmer et al., 1992). Hanmer et al. (1992) suggested that early sinistral zones of displacement may have later yielded to dextral displacement on the trailing edge of the same crustal block(s) to accommodate an eastward progression of the locus of Slave Province indentation and subsequent southward escape of Rae craton. Two observations suggest that this model may apply to the southern Taltson magmatic zone: 1) overprinting of dextral strike-slip duplexes on high grade sinistral mylonites in Leland Lakes shear zone; and 2) tectonic wedging of isoclinal sheath-fold closures in Taltson basement complex and Charles Lake shear zone. Such wedging requires that sinistral displacement be accommodated by synchronous dextral displacement on the opposite side of the wedge, and as such may provide an outcrop-scale model for tectonic escape of Taltson magmatic zone crustal slices. Future detailed field, petrological, and geochronological work is required to fully understand the spatial and temporal relationships of dextral and sinistral displacments in southern Taltson magmatic zone.

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Preliminary report of the metamorphic geology of Taltson magmatic zone, Canadian Shield, northeastern Alberta[1]

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Grover, T.W., McDonough, M.R., and McNicoll, V.J., 1993: Preliminary report of the metamorphic geology of Taltson magmatic zone, Canadian Shield, northeastern Alberta; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 233-238.

Abstract: Paragneiss and amphibolite are common in the Canadian Shield of northeastern Alberta. They are enveloped by basement orthogneisses and younger plutonic rocks of the Taltson magmatic zone and eastwardly adjacent Rae Province. Pelitic and quartzose paragneisses occur as north-south elongate rafts and lenses, varying from metres to kilometres in length, within granitoids and granitic gneisses. Amphibolite bodies occur as small lenses (less than 50x100 m) in the basement to the Taltson magmatic zone, and are particularly abundant in mylonite zones.

Mineral assemblages in the pelitic paragneisses provide constraints on the metamorphic evolution of the region. The assemblage garnet+biotite+sillimanite+K-feldspar+quartz±plagioclase±cordierite was identified in the field. This assemblage was found in rocks deformed in the Leland Lakes and Charles Lake shear zones, suggesting ductile deformation was at or near granulite facies conditions. Texturally late chlorite+epidote, associated with cataclastic rocks in these shear zones, indicate that displacement under greenschist facies conditions was superposed on high grade mylonites.

Résumé : Les paragneiss et amphibolites sont très répandus dans la région du Bouclier canadien qui se situe dans le nord-est de l'Alberta. Ils sont enveloppés par des gneiss du socle et des roches plutoniques plus jeunes de la zone magmatique de Taltson et de la Province de Rae qui la jouxte à l'est. Des paragneiss pélitiques et quartzeux forment, dans des granitoïdes et des gneiss granitiques, des blocs et lentilles allongés dans une direction nord-sud dont les dimensions varient de quelques mètres à plusieurs kilomètres. Des corps amphibolitiques se manifestent sous forme de petites lentilles (de moins de 50 x 100 m) dans le socle de la zone de Taltson et sont particulièrement abondants dans les zones mylonitiques.

Les assemblages minéraux présents dans les paragneiss pélitiques renseignent les chercheurs sur l'évolution métamorphique de la région. L'assemblage grenat+biotite+sillimanite+feldspath potassique+quartz±cordiérite±plagioclase a été identifié sur le terrain. La présence de cet assemblage a été observée dans des roches déformées des zones de cisaillement de Leland Lakes et de Charles Lake, ce qui porte à croire que la déformation ductile a eu lieu dans les conditions ou à peu près dans les conditions associées au faciès des granulites. Un assemblage chlorite+épidote, montrant une texture tardivement formée, et associé à des roches cataclastiques dans ces zones de cisaillement, indique qu'un mouvement survenu dans les conditions associées au faciès des schistes verts s'est surimposé à la déformation intense des mylonites.

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INTRODUCTION

This report summarizes the preliminary findings of the metamorphic part of a multidisciplinary research project on the geological evolution of a portion of the Canadian Shield exposed in northeastern Alberta. This part of the project concentrates on determining the number of metamorphic episodes, their respective pressure-temperature (P-T) conditions, and the relative timing between periods of metamorphism and deformation. Particular attention is being paid to discerning the prevailing metamorphic conditions during multiple phases of shearing which resulted in several north-trending mylonite belts.

The project was initiated in June 1992 and approximately seven weeks were spent in the field during July and August. Fieldwork was concentrated along several north-trending shear zones, shown on the Tulip Lake (74M/14), Mercredi Lake (74M/15), and Andrew Lake (74M/16) map sheets (Fig. 1). Paragneiss and metabasic lithologies were preferentially collected for further petrographic and microprobe studies because these lithologies commonly contain the most important mineral assemblages for determining the P-T conditions of metamorphism. Where possible, the same lithology was sampled both within and away from shear zones.

The results presented herein are based largely on relationships observed in the field and mineralogy deduced from hand samples. Mineral assemblages observed in the field are compared with published data, in particular that of Godfrey and Langenberg (1978), Nielsen et al. (1981), and Langenberg and Nielsen (1982), who report mineral assemblage and microprobe data from a large number of samples from the northeastern Alberta area. Pressure-temperature estimates are calculated using previously published compositional analyses and modern thermochemical data.

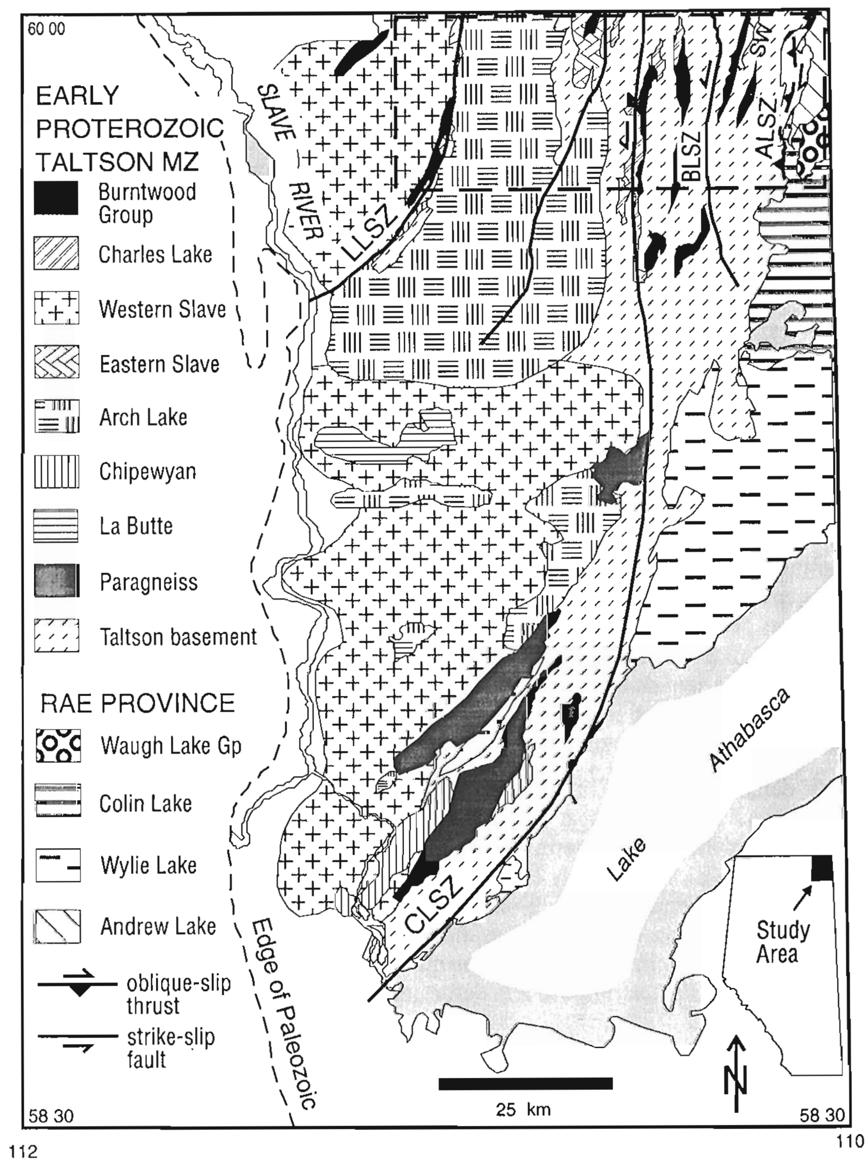

Figure 1.

Generalized geological map of the southern Taltson magmatic zone (modified from Godfrey, 1986). Dashed rectangle outlines the transect mapped during 1992. ALSZ = Andrew Lake shear zone; BLSZ = Bayonet Lake shear zone; CLSZ = Charles Lake shear zone; LLSZ = Leland Lakes shear zone. Units in the legend identified by a place name are granitoids.

REGIONAL GEOLOGY

The Canadian Shield exposed in northeastern Alberta is composed of rocks of the southern Taltson magmatic zone (TMZ) and Rae Province, to the east (McDonough et al., 1993; Ross et al., 1991; Hoffman, 1988). The Talston magmatic zone and the western Rae Province are separated from the Slave Province to the north by the Great Slave Lake shear zone, and bounded to the south by the Snowbird tectonic zone, two continental scale structural discontinuities (see Fig. 1, McDonough et al., 1993). The Buffalo Head terrane, a wholly subsurface terrane identified through aeromagnetic anomaly and drill core data (Ross et al., 1989), borders the Talston magmatic zone to the west. Ross et al. (1991) speculate that the 2.0 to 1.9 Ga magmatic rocks of the zone (Bostock et al., 1987, 1991) formed during the eastward subduction of oceanic lithosphere flanking the Buffalo Head terrane beneath the Archean Rae Province. Subduction and magma generation may have been contemporaneous with movement along the Great Slave Lake shear zone. The relationship between the tectonometamorphic evolution of the Talston magmatic zone in northeastern Alberta and movement along the Great Slave Lake shear zone and the Snowbird tectonic zone is poorly understood.

The portion of the southern Taltson magmatic zone mapped during the summer of 1992 is underlain by basement gneisses and various plutons with subordinate amounts of amphibolite and paragneiss (McDonough et al., 1993, in press a, b, c; Godfrey, 1963, 1966, 1986; Godfrey and Langenberg, 1986). The southern Talston magmatic zone is deformed by several shear zones having polymetamorphic displacement histories. In these north-trending zones, granitic basement gneisses, paragneisses, amphibolitic gneisses, and plutonic gneisses are deformed into high grade mylonites, greenschist grade mylonites, and low grade cataclasite (McDonough et al., 1993). Field relationships and descriptions of the shear zones and granitoids are described in McDonough et al. (1993). Observations and field relationships of the paragneisses and amphibolites are described in this report.

Paragneisses

Paragneiss is present throughout the area mapped although it comprises less than 10% of the exposed bedrock (Fig. 1). It occurs as elongate lenses and rafts within the granitoids and granitic gneisses, varying in length from less than a metre to several kilometres. Although paragneisses are present as rafts and xenoliths in all granitoids, the larger exposures are associated with the Taltson basement gneisses. Figure 2a-c illustrates some of the typical characteristics of the paragneisses and their relationships with enveloping granites.

Paragneisses were recognized in the field by pronounced layering and the presence of diagnostic minerals. Minerals such as biotite, sillimanite, and garnet suggest a pelitic protolith, especially when associated with quartzite layers (Fig. 2d). Paragneisses are often migmatitic with segregation of minerals into irregularly shaped light and dark layers. They are intruded by plutonic rocks of the Slave, Arch Lake, and Charles Lake intrusive suites, and by multiple phases of pegmatites. High grade metamorphism and associated deformation, mylonitization, and injection of granitic melts, altered the composition of the original protoliths and obliterated primary sedimentary structures often making positive identification of paragneiss impossible.

Paragneiss lithologies include quartzite, biotite+quartz+feldspar±sillimanitegarnet±gneisses, and some probable calc-silicate gneisses containing green amphibole and/or clinopyroxene+quartz+feldspar. Cordierite was not positively identified in the field but is reported to be commonly associated with garnet in the pelitic paragneisses (Langenberg and Nielsen, 1982). Orthopyroxene was not identified in any of the high grade paragneisses. Coarse grained muscovite was found only in the vicinity of the Charles Lake shear zone (CLSZ, Fig. 1). This coarse grained muscovite forms part of the foliation in the rock suggesting it is part of the high grade metamorphic mineral assemblage and not secondary in nature. Fine grained muscovite was found elsewhere in the area, however textures suggest it grew later in the paragenitic sequence, perhaps as an alteration product of K-feldspar.

High grade garnet+biotite+sillimanite+K-feldspar± muscovite were found in the Leland Lakes, Charles Lake, and Andrew Lake shear zones (LLSZ, CLSZ, and ALSZ respectively, Fig. 1). These high grade assemblages are locally overprinted and obliterated by the later growth of chlorite+ epidote assemblages. Chlorite+epidote growth is commonly associated with deformed granitic rocks containing cataclastic textures including brittly deformed porphyroclastic feldspar in a very fine grained, highly strained matrix.

Amphibolite

Amphibolite bodies are found throughout the mapped transect but they comprise less than 5% of the exposed rocks. They also occur as lenses and rafts within the granitic gneisses and granitoids. Amphibolites are most common in the granitic gneisses of the Taltson basement and are particularly abundant in the high grade mylonites of the Leland Lakes and Charles Lake shear zones. On average the amphibolite bodies are less than 50 m long and 10 m wide and are too small to be mappable units at 1:50 000 scale. Fabric in amphibolites is generally concordant with that of the surrounding granitoids. The amphibolites are commonly boudinaged and/or crosscut by younger, weakly foliated granites. Field occurrences of amphibolite are illustrated in Figure 2e and f.

The predominant mineralogy of the amphibolites is hornblende+feldspar+quartz+biotite. Orthopyroxene has also been identified in several hand samples of amphibolite. Only one garnet-bearing amphibolite was found; orthopyroxene was not identified at that outcrop.

METAMORPHISM

Mineral assemblages in the pelitic paragneisses and metabasic lithologies are the most informative for documenting the metamorphic evolution of the area. In the metapelitic rocks, the assemblage Grt+Bt+Sil+Kfs+Pl+Qtz was documented throughout the region, while Hbl, Hbl+Bt, and

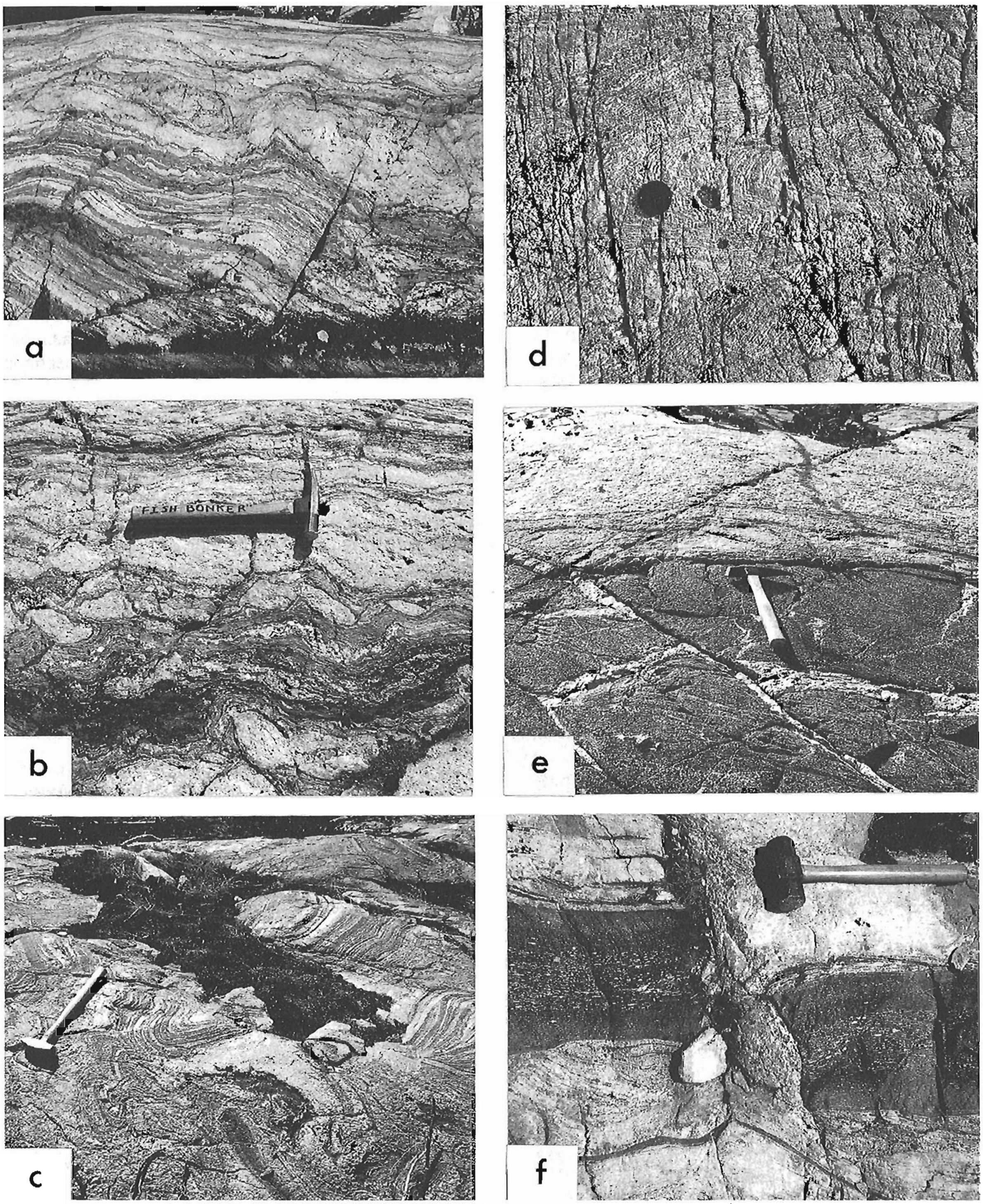

Figure 2. Field photographs illustrating contact relationships between paragneiss and granitoid (a-d), and amphibolite and granitoid (e-f). **a.** Lakeshore exposure of primarily paragneiss on the bottom and granitic gneiss on top. View is approximately 10 m across. **b.** Closeup of the outcrop shown in Figure 2a illustrating the intermingling of granitic material and paragneiss. **c.** Outcrop of well layered paragneiss broken up and injected by granite. **d.** Folded, layered paragneiss with 62 mm diameter lens cap for scale. The roughly circular shape adjacent to the lens cap is a garnet crystal. Garnet often weathers low in these rocks. **e.** Amphibolite in contact with granitic gneiss and crosscut by pegmatites. **f.** Pulled apart and displaced amphibolite dyke(?). The dark margins of amphibolite may be relict chilled margins.

occasionally Hbl+Opx, were found in the more mafic rocks (mineral abbreviations after Kretz, 1983)[1]. The assemblage Grt+Opx was not observed in any hand samples of the metapelitic or metabasic lithologies. The occurrence of the same high grade mineral assemblages across the mapped transect suggests that there was little or no gradient in the metamorphic conditions during this high grade episode.

The mineral assemblages found in hand samples are characteristic of upper amphibolite to granulite facies metamorphic conditions. This is consistent with the migmatitic character of much of the paragneiss. Langenberg and Nielsen (1982) reported that cordierite is commonly associated with garnet in the metapelitic rocks. This would indicate that the mineral assemblage Grt+Crd+Kfs is present in the paragneiss, which is characteristic of granulite facies metamorphism. This suggests that the metamorphic grade equalled or exceeded the P-T conditions of the model reaction:

Bt+Sil+Pl+Qtz = Crd+Grt+Kfs+L

where L represents a melt formed by dehydration melting. The position of this equilibrium is illustrated schematically in Figure 3, based on data in Thompson (1982) and Pattison and Tracy (1991). The exact position of this equilibrium in P-T space is dependent on mineral and melt compositions. The apparent lack of orthopyroxene in the metapelitic paragneisses suggests that the P-T conditions were below those of the model reaction:

Bt+Grt+Pl+Qtz = Kfs+Crd+Hy+L

also shown schematically in Figure 3.

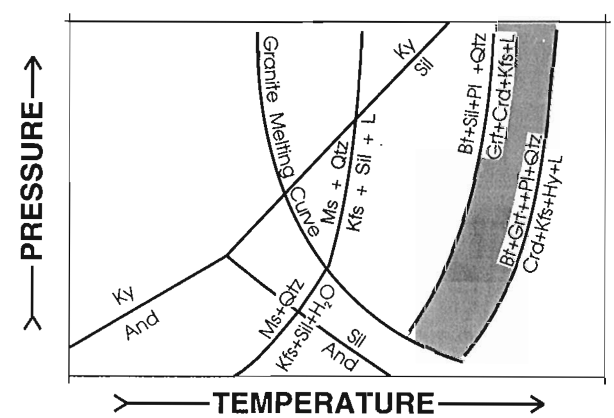

Figure 3. Schematic P-T diagram illustrating the relative position of equilibria involving the appearance/disappearance of diagnostic phases in high grade metapelitic rocks. The shaded area indicates the stability field of mineral assemblages identified in a hand sample.

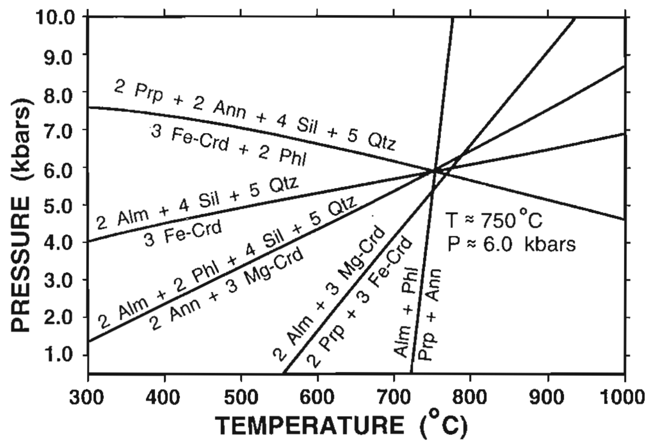

Figure 4. P-T diagram showing the position of several garnet-cordierite-biotite bearing equilibria. The diagram was constructed using data reported in Langenberg and Nielsen (1982) using GeO-Calc (Brown et al., 1988).

GEOTHERMOBAROMETRY

Little well constrained compositional data exist for rocks from the Canadian Shield in northeastern Alberta. Langenberg and Nielsen (1982) provide some compositional data for garnet, cordierite, and biotite in rocks, coexisting with sillimanite, K-feldspar, quartz, and plagioclase. Several equilibria may be used to constrain the P-T conditions recorded by these rocks. Using the data reported by Langenberg and Nielsen, the position of these equilibria were calculated using the GeO-Calc program of Brown et al. (1988). The calculations for many samples resulted in a wide scatter of estimated pressures and temperatures of metamorphism, suggestive of disequilibrium. This is not surprising given the problems commonly encountered in geothermobarometry of granulite facies rocks (Frost and Chacko, 1989). However, several samples show relatively consistent pressures and temperatures in the range of 750 to 780°C and 5 to 6 kbars (Fig. 4). These results are consistent with the observed mineral assemblages (Pattison and Tracy, 1991) and suggest that meaningful compositional data can be extracted from these rocks.

DISCUSSION

The metamorphic rocks exposed in the Canadian Shield in northeastern Alberta are the products of a complex history including high grade metamorphism, deformation, mylonitization, and emplacement of granitic plutons. Field observations reported herein are pertinent to the tectonometamorphic evolution of this area. At this time, however, several working hypotheses are possible.

[1] Abbreviations are: Alm = almandine; And = andalusite; Ann = annite; Bt = biotite; Crd = cordierite; Grt = garnet; Hbl = hornblende; Hy = hypersthene; Kfs = K-feldspar; Ky = kyanite; Ms = muscovite; Opx = orthopyroxene; Phl = phlogopite; Pl = plagioclase; Prp = pyrope; Qtz = quartz; Sil = sillimanite.

High grade metamorphic rocks characterized by Grt+Sil+Kfs+Crd assemblages reported in the metapelitic rocks and Opx+Hbl assemblages in the more mafic lithologies are found throughout the mapped transect. Granulite grade metapelitic rocks are found in the Leland Lakes shear zone (LLSZ, Fig. 1) suggesting that movement along this shear zone was at least rooted to mid-crustal levels with temperatures in the 750 to 800°C range. Similar granulite grade metapelitic rocks were also found in the vicinity of the Charles Lake shear zone. However, Ms+Sil+Qtz assemblages were also documented in the area. If Ms+Sil+Qtz represents the stable mineral assemblage at the time of deformation, it may have some important implications regarding the relative timing of peak metamorphism and shear zone deformation.

Godfrey and Langenberg (1978), Nielsen et al. (1981), and Langenberg and Nielsen (1982) reported a polymetamorphic history for the Canadian Shield of northeastern Alberta. They observed spinel±corundum inclusions wholly enveloped by cordierite and correctly interpreted this to imply multiple phases of recrystallization. They postulated that the spinel+corundum+cordierite assemblages were the result of a high grade (900°C, 7.5 kbars) Archean(?) granulite facies event. This was then followed by a pervasive Proterozoic granulite-amphibolite facies event, and a younger greenschist facies event manifested by late chlorite+epidote growth in shear zones.

Our future work will be towards understanding the complex tectonometamorphic history of this area and testing the ideas presented by Langenberg and Nielsen (1982), among others. Several questions regarding the relative and absolute timing of metamorphism, emplacement of various plutons, and shear zone movement need to be answered. Detailed petrography will establish the sequence of mineral growth relative to deformation, and equilibrium mineral assemblages. Microprobe studies in conjunction with geothermobarometry will be conducted to determine the P-T conditions of metamorphism. These data will be combined with data from ongoing structural, geochronological, and isotopic studies to develop an integrated model for the geological evolution of this region.

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Progress report on gravity surveys in Cormorant Lake map area, Manitoba-Saskatchewan

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Thomas, M.D., Williamson, B.L., Williams, R.P., and Miles, W., 1993: Progress report on gravity surveys in Cormorant Lake map area, Manitoba-Saskatchewan; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 239-247.

Abstract: The Cormorant Lake map area covers a large segment of the Flin Flon-Snow Lake Belt, a geological domain having significant mineral potential located in the Trans-Hudson Orogen. Several ongoing geoscientific programs funded by NATMAP, Lithoprobe and federal-provincial Mineral Development Agreements are directed towards a better understanding of the belt and its southward extension beneath adjacent Phanerozoic sedimentary cover. Geophysical methods are playing a prominent role in achieving objectives. Gravity surveys conducted in March and June, 1992 produced 369 new gravity stations positioned to provide detailed information relating to several targets. Geological targets included the Reed Lake granite, the Reed Lake gabbro, volcanic sections of the Amisk Group, and a small basin of Missi Suite metasediments near Flin Flon. Geophysical targets, defined by earlier gravity surveys, included a belt of steep gravity gradients that partly coincides with the margin of the shield, and a prominent gravity high located over Phanerozoic sediments.

Résumé : La région cartographique de Cormorant Lake couvre un vaste segment de la zone de Flin Flon-Snow Lake, domaine géologique ayant un potentiel minéral considérable dans l'Orogène trans-hudsonien. Plusieurs programmes géoscientifiques en cours financés dans le cadre du CARTNAT, de Lithoprobe et des Ententes fédérales-provinciales sur l'exploitation minérale, doivent aider à mieux expliquer la nature de la zone et de son prolongement sud sous la couverture sédimentaire adjacente, d'âge phanérozoïque. Les méthodes géophysiques aident dans une mesure très importante à atteindre les objectifs recherchés. Les levés gravimétriques effectués en mars et en juin 1992 ont permis d'établir 369 nouvelles stations gravimétriques positionnées de façon à fournir de l'information détaillée sur plusieurs cibles. Les cibles de la prospection géologique comprenaient le granite de Reed Lake, le gabbro de Reed Lake, les portions volcaniques du Groupe d'Amisk, et un petit bassin de métasédiments de la série de Missi près de Flin Flon. Les cibles de la prospection géophysique, définies par des levés gravimétriques antérieurs, comprenaient une zone de gradients gravimétriques prononcés coïncidant partiellement avec la marge du bouclier, et un important maximum gravimétrique situé au-dessus de sédiments d'âge phanérozoïque.

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INTRODUCTION

Prior to 1991 gravity coverage of the Cormorant Lake map area (NTS 63K) was limited to stations spaced generally 10 to 15 km apart, exceptions being a small grid of stations near Flin Flon spaced at intervals of about 6 km, and observations along Highway 10 spaced approximately 3 km apart. In 1991 gravity coverage was enhanced by 1 km stations along LITHOPROBE Trans-Hudson seismic lines (Hearty and Gibb, 1992), which in the Cormorant Lake area followed sections of Manitoba highways 10 and 39, secondary roads east and northeast of Flin Flon and Saskatchewan Highway 106 (principal highways are outlined in Fig. 1).

Blair et al. (1988) reporting on Project Cormorant, supported under a Canada-Manitoba Mineral Development Agreement and initiated to map the buried extension of the Flin Flon – Snow Lake Belt beneath Phanerozoic cover, cited aeromagnetic data and drill core data as their principal tools. Recent studies indicate that gravity data, also, can make a significant contribution to the program. For example, granitic and mafic volcanic units, both of which might be characterized by strong magnetic signatures, but which have radically different densities, may be discriminated on the basis of their gravity signature.

In March 1992 a gravity survey was conducted within the Cormorant Lake map area to investigate specific gravity features outlined by the existing regional coverage. Most of these features were located over the Phanerozoic sediments, with an exception being a prominent gravity low associated with the Reed Lake granite east of Reed Lake (Fig. 1). Another target on the shield was a gravity high over the Reed Lake gabbro on the western shores of Reed Lake, selected for detailed gravity survey in support of geological investigations by the Mineral Resources Division of the Geological Survey of Canada.

In June 1992 gravity observations were made on the shield to investigate the structure of volcanic rocks of the Amisk Group north of Flin Flon and a basin of Missi Suite sediments near Flin Flon (Fig. 1), and to augment gravity coverage in the western part of Reed Lake.

GEOLOGY OF THE FLIN FLON-SNOW LAKE BELT

The Flin Flon-Snow Lake Belt is a metavolcanic belt which, on the basis of its constituent lithologies and their geochemistry and environment of deposition, has been compared to Cenozoic intraoceanic island arcs (Syme, 1990). A simplified geological map is presented as Figure 1.

Amisk Group

Much of the belt comprises an assemblage of mainly subaqueous volcanic rocks, synvolcanic intrusions and associated sedimentary rocks, which together form the Amisk Group (Galley et al., 1990). Generally the group is dominated by mafic volcanic flows and volcaniclastic rocks ranging in composition from basaltic to andesitic. Although locally attaining significant thicknesses (Galley et al., 1990; Gordon et al., 1990), rhyolitic flows and felsic volcaniclastic rocks, greywacke and mudstone are relatively minor components of the group. Rocks of the Amisk Group are the oldest in Flin Flon-Snow Lake Belt, U-Pb dating yielding an age of 1886 Ma (Gordon et al., 1990; Syme, 1990).

Missi Suite

The Amisk Group is unconformably overlain by rocks of the Missi Suite, which have limited distribution within Flin Flon-Snow Lake Belt. The suite consists predominantly of magnetite-bearing, quartz-rich gneisses derived from fluvial and alluvial sandstones, and subordinate amphibolites and felsic gneisses derived from volcanic rocks (Zwanzig, 1990). Stauffer (1990) has interpreted these rocks as representing a syntectonic molasse deposit. The Missi Suite has yielded a U-Pb age of 1832 Ma (Gordon et al., 1990).

Burntwood Suite

The Burntwood Suite is restricted to the northern margin of the area (Fig. 1) and is included in the Kisseynew Gneiss Belt (Zwanzig, 1990). It consists of high grade paragneisses derived largely from turbidites deposited along the northern margin of the Flin Flon-Snow Lake Belt.

Plutonic intrusions

Both the Amisk Group and Missi Suite are intruded by a variety of plutonic intrusions, which collectively form a significant part of the belt. Felsic to intermediate plutonic intrusions of various sizes occur throughout Flin Flon-Snow Lake Belt, some coalescing to form large composite batholiths. Mafic and mafic-ultramafic plutonic intrusions are also present, but are more restricted in development. U-Pb ages determined on the felsic to intermediate bodies (Gordon et al., 1990; Bailes et al., 1991) indicate that the intrusions were emplaced between about 1889 Ma and 1830 Ma ago. Gordon et al. (1990) noted that a younger age of 1820 Ma has also been reported.

Metamorphism

The Flin Flon-Snow Lake Belt has been affected by a regional metamorphic event that peaked at 1815 Ma in the core of the Kisseynew Gneiss Belt and imposed the generally east-west pattern of metamorphic isograds (Bailes and McRitchie, 1978) on the Kisseynew Gneiss Belt and Flin Flon-Snow Lake Belt (Gordon et al., 1990). Within the Flin Flon-Snow Lake Belt the metamorphic grade ranges from subgreenschist to upper amphibolite facies, and generally increases northward across the belt (Gordon et al., 1990). Froese (pers. comm., 1991) has noted that locally the metamorphic grade also increases southward.

Structure

The Flin Flon-Snow Lake Belt has been subjected to several periods of deformation associated with folding and faulting. At least five phases of folding have been recognized in some areas (Syme, 1990). Major faults transect all parts of the Flin Flon-Snow Lake Belt and are of economic interest because of their spatial association with gold occurrences (Galley et al., 1989). Many faults converge on a southwest-trending feature, called the Namew Lake structure (Gordon et al., 1990), near the western shores of Athapapuskow Lake (Figure 1).

Phanerozoic cover rocks

The Flin Flon-Snow Lake Belt disappears southward under a blanket of gently dipping Phanerozoic sedimentary rocks. Ordovician dolomites, dolomitic limestones, and limestones form most of the cover in the map area; these pass upwards into Silurian dolomites along the southern margin of the Cormorant Lake map area, where the succession attains a maximum thickness of about 105 m (Taiga, 1986).

GRAVITY SURVEYS

A map showing the distribution of old (pre-1992) and new (1992) gravity stations in the Cormorant Lake map area is shown in Figure 2.

SIAL Geosciences Inc. carried out a gravity survey under contract in March 1992 using helicopter transport. Observations were made at 317 stations, 14 on the frozen surfaces of lakes and 303 on land. They were distributed along Traverses 1 to 8 (Fig. 2) and spaced generally 0.5 or 1 km apart, and on a grid pattern (2.5 km spacing) centred on Reed Lake. Measurements were made with LaCoste and Romberg Geodetic gravity meters No. G172 and G278, and tied to control stations of the National Gravity Net. Horizontal positions were determined using a Global Positioning System (GPS) and are estimated to be accurate to within ±55 m based on non-differential GPS. Vertical positions were obtained using altimetry tied to federal benchmarks, and are estimated to be accurate to within ±1 m. The overall accuracy of the Bouguer anomalies, calculated using a uniform density of 2.67 g/cm^3 and a sea level datum, is estimated to be ±0.21 mGal.

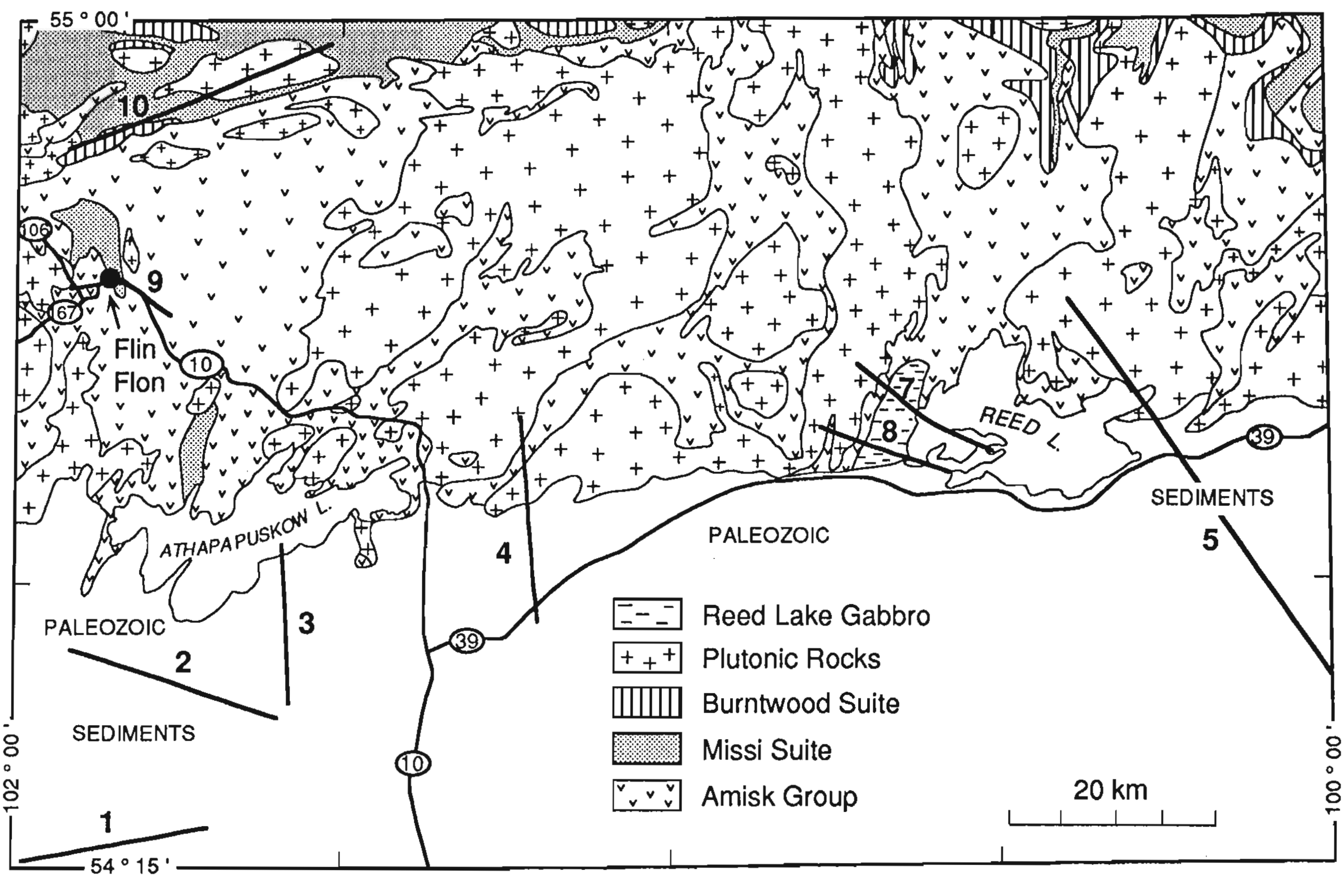

Figure 1. Simplified geological map of the Flin Flon-Snow Lake Belt. Plutonic rocks comprise mainly granitic, granodioritic and tonalitic varieties and less commonly dioritic, gabbroic and ultramafic varieties. The unit labelled Burntwood Suite includes metasediments of the File Lake Formation. Gravity profiles 1-5 and 7-10 discussed in text are indicated by solid black lines; profile 6 falls partly outside the figure and is not shown (see Figure 2 for location).

In June 1992 the authors established another 52 gravity stations on the shores and islands of Kisseynew (Traverse 10) and Reed lakes (southeast extensions of Traverses 7 and 8) and along a short traverse crossing the town of Flin Flon (Traverse 9). Measurements were made with LaCoste and Romberg Geodetic gravity meter G790 and were tied to control stations of the National Gravity Net. Horizontal positions were scaled from 1:50 000 scale NTS topographic maps and have an estimated accuracy of ±25 m. Elevations were determined by direct measurement from the lake levels and by altimetry tied to local federal benchmarks; all elevations are considered accurate to within ±1 m. Although rough water conditions resulted in degraded data for certain stations on Kisseynew Lake, the accuracy of Bouguer gravity anomalies for all traverses is estimated to be within ±0.5 mGal.

Terrain corrections have not been calculated for either survey, but are not expected to exceed a few tenths of a milligal. It is estimated that all Bouguer anomalies computed for the 1992 surveys are accurate to within ±1.0 mGal.

BOUGUER GRAVITY FIELD

The Bouguer gravity field contoured at a 5 mGal interval is displayed in Figure 2. Contour trends are variable across the map area, but generally strike north-northwest to north-northeast over the Flin Flon-Snow Lake Belt, where they mimic trends of major faults and geological units, and west-northwest over Phanerozoic sediments south of the belt. The

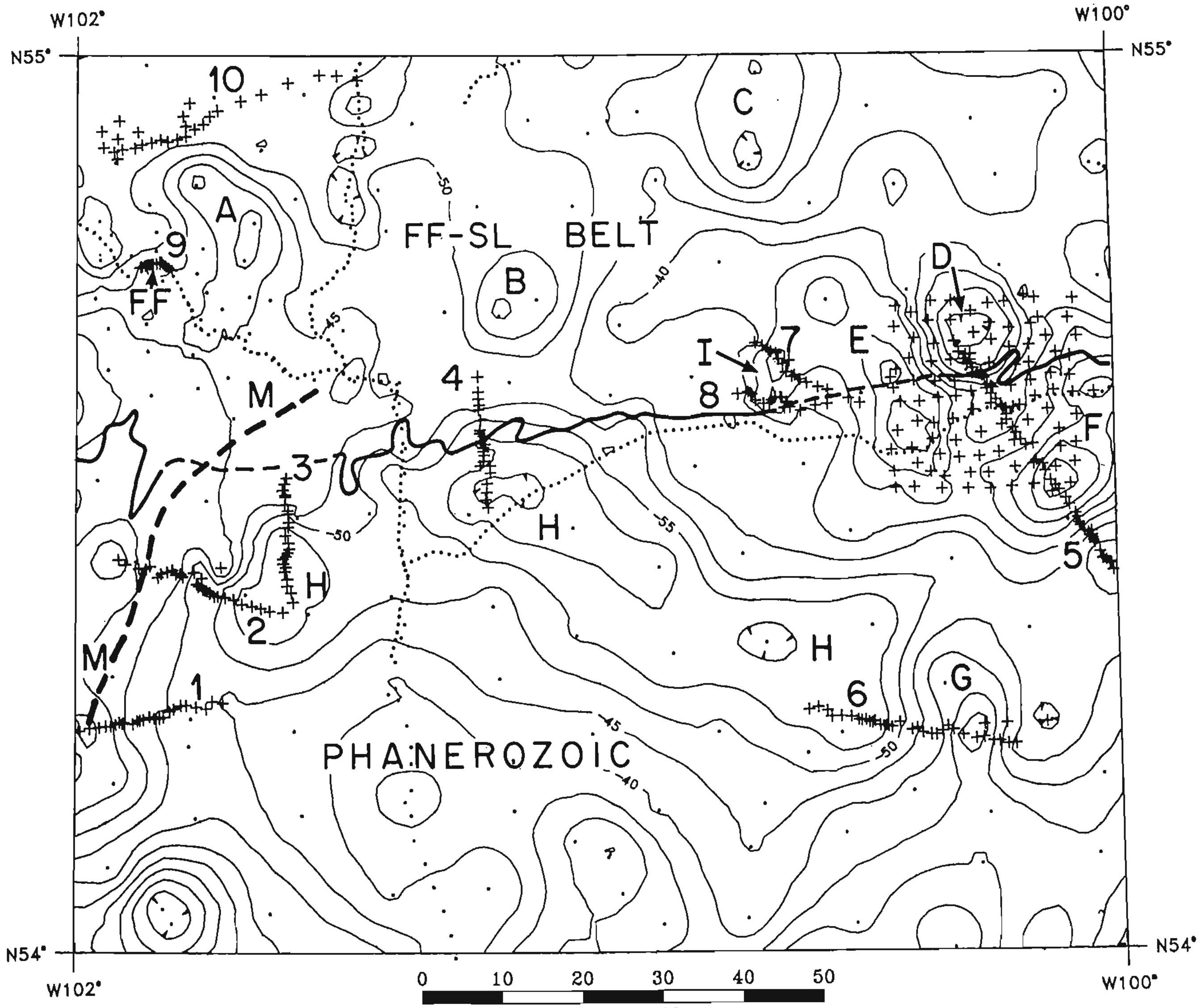

Figure 2. Bouguer gravity anomaly map (5 mGal contour interval); small teeth inside closed contours indicate gravity low. Dots indicate gravity observations made before 1992; crosses are observations made in 1992. The 1992 gravity traverses are numbered 1 to 10. A to H, gravity anomalies discussed in text. FF-SL, Flin Flon-Snow Lake Belt; FF, town of Flin Flon; thick dashed line labelled M, magnetic lineament. The Precambrian Shield-Phanerozoic boundary is marked by a continuous, medium width line (dashed where it crosses Athapapuskow and Reed lakes).

difference in trends over the two regions is somewhat surprising, because an interpretation of sub-Phanerozoic geology based largely on aeromagnetic and borehole data (Blair et al., 1988) indicates that north to north-northeast geological trends predominate. It is likely that the west-northwest trends over the Phanerozoic deposits and the extensive gravity low 'H' (amplitude ≈ 15 mGal), which the trends partially define, are related to a large 'granitic' batholith. A magnetic domain defined by Blair et al. (1988) is spatially coincident with the eastern half of gravity low 'H' and was interpreted by these authors to be underlain by a granitoid complex. The gravity map suggests that this complex extends considerably farther west than indicated by the magnetic domain. Drillcore extracted from the granitoid complex has yielded an age of 1845 Ma (Blair et al., 1988) making it younger than the Amisk Group. The age of the complex and its cross-cutting relationship with respect to interpreted north- and north-northeast-trending magnetic domains are consistent with emplacement of a discordant intrusion and explain the aberrant trend of anomaly 'H'.

Other notable anomalies in the area (Fig. 2) are:

- positive anomaly 'A' (amplitude ≈ +16 mGal) over Amisk Group volcanics near Flin Flon,
- negative anomalies 'B', 'C' and 'D' (amplitudes ≈ -12, -12, and -16 mGal, respectively) corresponding with areas dominated by granitic rocks,
- positive anomalies 'E' and 'F' (amplitudes ≈ +16 and + 25 mGal, respectively), which span the shield margin, and which are attributed to mafic intrusions on the basis of spatial relationships observed on the shield,
- positive anomaly 'G' (amplitude ≈ +18 mGal) which falls entirely over Phanerozoic sediments, but which has a probable source in buried Amisk Group mafic volcanics and/or a mafic intrusion(s),
- positive anomaly 'I' (amplitude ≈ +14 mGal) which correlates closely with the Reed Lake gabbro.

These relationships between gravity anomalies and geological bodies provide a means for deriving quantitative information on the size and shape of the bodies through a process known as 'gravity modelling'. Crustal sections obtained in this fashion augment surface geological data and can lead to an enhanced understanding of the structure and tectonic history of a region. Gravity models are not presented in this report, but a preliminary appraisal of the new gravity data is provided.

PRELIMINARY INTERPRETATION OF THE NEW GRAVITY DATA

Profiles 1 and 2

Gravity traverses 1 and 2 are located over Phanerozoic sedimentary cover and were positioned to investigate a belt of steep gradients defined in the regional gravity field. Corresponding gravity profiles, together with aeromagnetic profiles, are shown in Figure 3 (magnetic profiles in this report portray the total magnetic field with the International Geomagnetic Reference Field removed). The amplitude of the change across the gravity gradient is ≈ 16 and ≈ 22 mGal, respectively, for Profiles 1 and 2. Such gradients, or 'step' anomalies, are typical of a change in rock density across a planar geological boundary, and given the convergence of several faults towards the gradient, it was anticipated that detailed gravity information might provide information on the dip and depth extent of any fault. The belt lies immediately east of a north-northeast-trending magnetic domain boundary (Blair et al., 1988) located along the western margin of the Cormorant Lake map area. This boundary (M in Fig. 2 and 3) is defined by the truncation of north- to northwest-trending magnetic features west of the boundary along a gently curvilinear belt of north- to north-northeast-trending features east of the boundary. This boundary is probably marked by a fault. One of the magnetic features west of, and truncated at, the domain boundary is interpreted to define the sub-Phanerozoic path of the Sturgeon Weir fault (Blair et al., 1988).

Profile 2 demonstrates clearly that the magnetic domain boundary has little expression in the gravity field. The same appears to hold true for Profile 1 also, although detailed gravity information is lacking west of the boundary. It is speculated that the gravity gradient might be related to a gently to moderately dipping contact between a basement granitoid complex to the east and sedimentary-volcanic rocks of the Amisk Group to the west. The eastern portion of

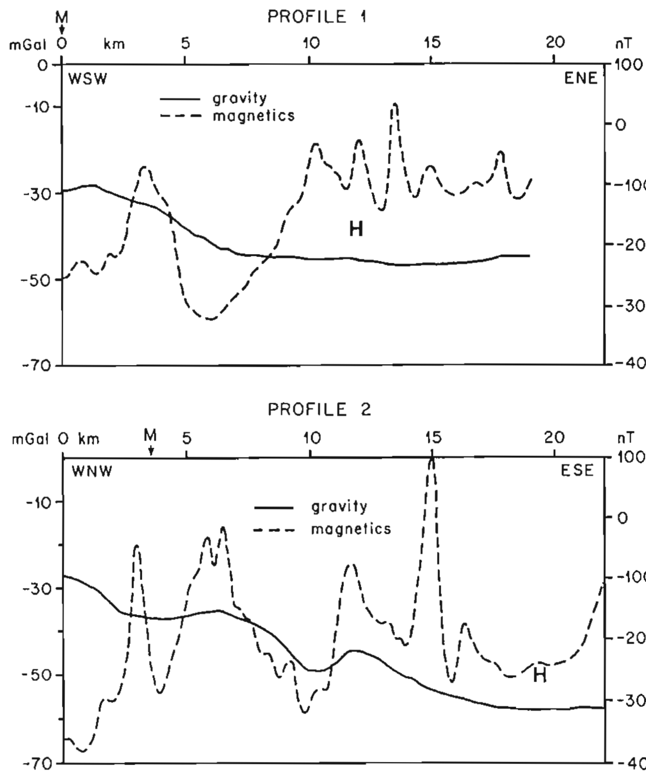

Figure 3. Gravity and aeromagnetic profiles along Traverses 1 and 2. M, magnetic domain boundary; H, gravity low.

Profile 2 falls within the extensive negative anomaly 'H', part of which correlates with the large granitoid complex interpreted by Blair et al. (1988). Several boreholes occurring near the eastern ends of both profiles indicate the presence of different rock types in the basement. If a granitoid body is indeed the source of the gradient and of gravity low 'H' in this area, it is probably buried within the basement.

Profiles 3 and 4

Traverses 3 and 4 were selected to examine the northeastward extension of the gradient crossed by Traverses 1 and 2, as it swings gradually into a west-east orientation along the shield margin. The corresponding profiles (Fig. 4) have amplitudes of ≈ 18 and ≈ 21 mGal, respectively, similar to those of Profiles 1 and 2. Both profiles extend northwards from the gravity low 'H', and so the interpretation of the gradient is the same as that for Profiles 1 and 2, namely that it is a manifestation of a gently to moderately dipping contact between a granitoid complex to the south and sedimentary-volcanic assemblages of the Amisk Group to the north.

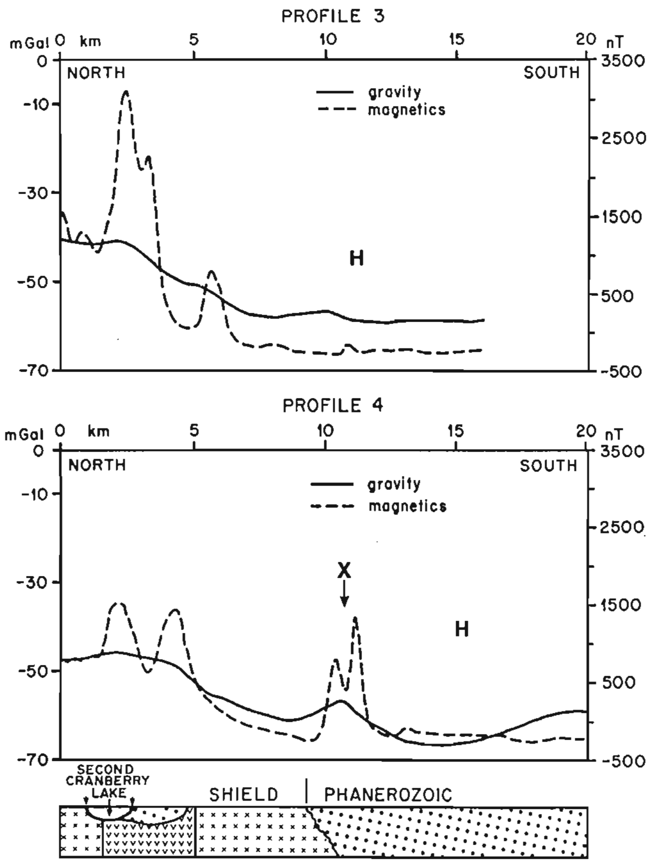

Figure 4. Gravity and aeromagnetic profiles along traverses 3 and 4. X indicates gravity and magnetic highs discussed in text; H, gravity low. Legend for simplified, schematic geological section provided for Profile 4: x pattern, tonalite-granodiorite; v pattern, Amisk Group volcanics; dot pattern, Phanerozoic sediments.

The character of the aeromagnetic profile for Traverse 3, which is located entirely on Phanerozoic sediments, changes dramatically from being relatively flat in the south to being replaced by intense positive anomalies across the gravity gradient. These anomalies can be traced east-northeast to the margin of the shield where they correlate with Amisk Group volcanics. In situ measurements of magnetic susceptibility conducted on outcrops of Amisk Group volcanic rocks along Highway 10 some 15 km to the northeast of the north end of the traverse show these rocks to be highly magnetic with susceptibilities frequently attaining over 100 x 10^{-3} SI, up to ≈300 x 10^{-3} SI.

Traverse 4 crosses the shield margin. In the corresponding gravity profile the upper level of the 'step' anomaly corresponds with tonalite-granodiorite and Amisk Group volcanics, whereas the lower level lies over Phanerozoic sediments. The gradient or 'step' itself correlates with a tonalite-granodiorite body. The close coincidence of short wavelength gravity and magnetic highs at position 'X' along the profile points to the presence of a narrow belt of magnetic Amisk Group volcanics in the underlying basement.

Profile 5

A prominent aeromagnetic lineament lying close to the shield margin and running approximately east-west across much of the area is interpreted to be the expression of a major fault. Near Reed Lake the lineament splays into several lineaments, of which one coincides with the Berry Creek Fault. Traverse 5 was positioned to investigate these faults based on the proximity of the southern margin of the Reed Lake granite to the Berry Lake Fault (Fig. 5, 6). It was anticipated that the margin would be associated with a significant density contrast, which would

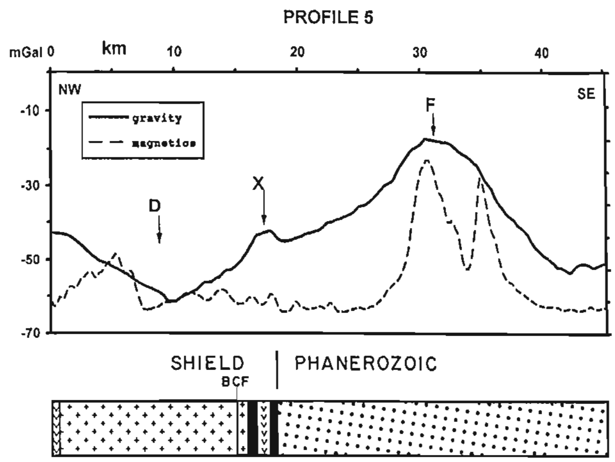

Figure 5. Gravity and aeromagnetic profiles along traverse 5. D, gravity low associated with Reed Lake granite; X, small gravity high associated with gabbro and Amisk Group volcanics; F, gravity high attributed to buried quartz dioritic to gabbroic rocks. Legend for schematic geological section: BCF, Berry Creek fault; + pattern, Reed Lake granite; black, gabbro; v pattern, Amisk Group volcanics; dot pattern, Phanerozoic sediments.

produce a distinct gravity anomaly. Modelling of this anomaly might then provide information on the dip and depth extent of the fault.

However, apparently, the gravity profile corresponding to traverse 5 (Fig. 5), does not contain a signature that can be unequivocally related to a steep fault. The profile is generally smooth with values increasing gradually southward from the low 'D' over the Reed Lake granite to the gravity high 'F' over the Phanerozoic sediments (see also Fig. 2). A small positive perturbation is present between 'D' and 'F' (X in Fig. 5) just south of the Berry Creek Fault and is related to Amisk Group volcanics and gabbroic rocks. Modelling studies are required to determine whether these rocks have faulted boundaries. The positive anomaly 'F' runs northeast to the shield where it coincides with a large intrusive body comprising quartz diorite, diorite, and gabbro.

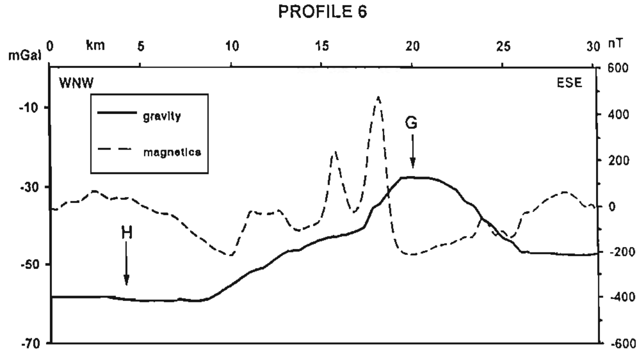

Figure 7. Gravity and aeromagnetic profiles along traverse 6. H and G, gravity low and high, respectively, discussed in text.

Profile 6

Traverse 6 is located on Phanerozoic sediments and runs WNW-ESE from gravity low 'H' across the pronounced gravity high 'G' (see profile in Figure 7). Mafic metavolcanics and metasediments are amongst basement rock types that have been retrieved from boreholes near the traverse and the high is tentatively attributed to a relatively thick development of these rocks, although a contribution from mafic intrusive rocks cannot be ruled out.

Profiles 7 and 8

Profiles 7 and 8 were designed to provide a cross section of the Reed Lake gabbro intrusion (Fig. 1). The intrusion is one of several bodies of metagabbro occurring in the Reed Lake-Iskwasum Lake area, intruded into isoclinally folded volcanic rocks of the Amisk Group. The mafic volcanic host rocks extend for about 15 km to the east of the intrusion, but only about 1 km to the west before encountering a granitic batholith.

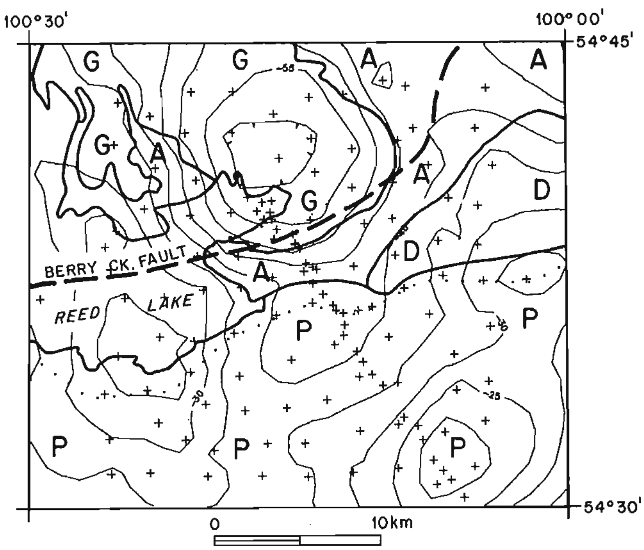

Figure 6. Bouguer gravity field (5 mGal contour interval) over the Reed Lake granite (G). Pre-1992 and 1992 gravity stations are indicated by dots and crosses, respectively. Geological legend: A, mainly Amisk Group volcanics and small mafic intrusions; G, Reed Lake granite; D, quartz diorite, diorite, gabbro; P, Phanerozoic sediments.

The intrusion consists of a variety of gabbroic and ultramafic rocks, ranging from anorthosite to pyroxenite. The lower (eastern) part of the section is a mixture of anorthositic, leuco-, normal and mela-gabbros and pyroxenites. The middle and upper parts of the section comprise a mixture of leuco- gabbro to melagabbro, with quartz-bearing gabbros occurring throughout. Layer contacts strike approximately north- northwest and dip between 70° west and vertical. The intrusion has undergone regional metamorphism to upper greenschist facies, and much of the original mineralogy has been altered.

Aeromagnetic data indicate the presence of three principal north-south belts of positive anomalies within the bounds of the intrusion, parallel to the strike of both the body itself and its internal layer contacts. The strongest anomaly is at the centre of the body (see Fig. 8, Profile 8), and swings sharply to the east at its southern end. Weaker anomalies occur just within the upper (western) and lower (eastern) contacts. The basal anomaly consists of a series of elongate, narrow zones which individually strike slightly oblique to the other magnetic highs, but collectively are strike parallel. The aeromagnetic anomalies can be correlated on the ground with more mafic rocks (melagabbro to pyroxenite) which attract a swing magnet.

Gravity profiles 7 and 8 (Fig. 8) indicate the presence of a 14 mGal positive Bouguer anomaly centred on the intrusion. The peak of the gravity anomaly coincides closely with the central, coextensive major aeromagnetic anomaly. The coincidence of a gravity high and magnetic high suggests that ultramafic rock should be encountered here. In fact, geological studies along traverse 8 have established the presence of strongly magnetic pyroxenites, but peridotites have not yet been found.

The gravity anomaly terminates fairly abruptly at the intrusion's south end, as evidenced by a steep gradient, across which values change by about 15 mGal over the 3 km which

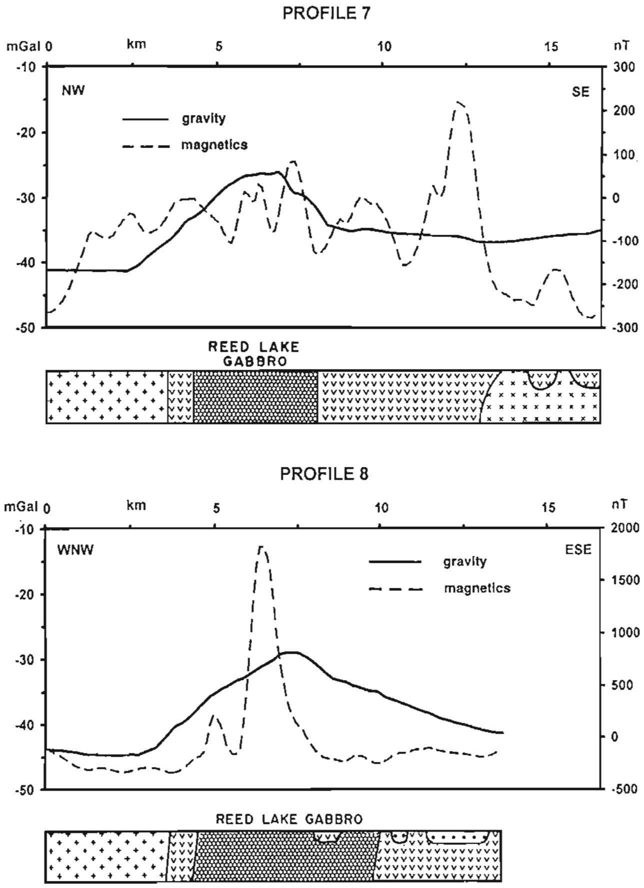

Figure 8. Gravity and aeromagnetic profiles along traverses 7 and 8. Legend for schematic geological sections: + pattern, granite; x pattern, tonalite and granite; v pattern, Amisk Group volcanics; honeycomb pattern, Reed Lake gabbro; dot pattern, Phanerozoic sediments.

separates Profile 8 from the Lithoprobe traverse along Highway 39. Bouguer values along the highway traverse, which is coincident with the shield margin, vary by no more than a few milliGals over tens of kilometres and indicate that the intrusion does not extend under the Phanerozoic cover. It is uncertain whether this is a result of an intrusive contact (fortuitously coincident with the shield margin) or fault-displacement of an originally more extensive intrusion. The sharp eastward deflection of the south end of the central magnetic anomaly, the presence of a prominent magnetic lineament, the gradual decrease of the gravity anomaly eastward on Profile 8 and the occurrence of scattered outcrops of gabbro on small islands in Reed Lake support faulting, although a fault zone has yet to be observed in outcrop.

In both Profiles 7 and 8 the gravity field is noticeably higher to the east of the Reed Lake gabbro than to the west. This is a consequence of the higher densities measured for mafic volcanics (mean = 2.84 g/cm^3) compared to an estimated density of about 2.65 g/cm^3 for the granite to the west. The gabbro itself has a mean density of about 3.05 g/cm^3 based on measurements to date.

Profile 9

A small basin of Missi Suite metasediments extends north-south through Flin Flon and is in faulted and unconformable contact with the Amisk Group, which comprises largely mafic and intermediate volcanics. Because the Missi Suite sandstones and conglomerates are less dense than surrounding Amisk Group volcanics, 2.71 g/cm^3 compared with 2.84 g/cm^3, it was assumed that a corresponding gravity anomaly (negative) would be present. Modelling of same would enable the size and shape of the basin to be determined. Traverse 9 was positioned to define the anticipated anomaly; gravity stations were placed between about 150 and 400 m apart to provide the required detail. The resulting profile (Fig. 9) shows that a small, yet distinct negative anomaly (amplitude ≈-2.5 mGal) is indeed associated with the Missi Suite.

Profile 10

Amisk Group volcanics near Flin Flon are associated with a strong positive gravity anomaly ('A', Fig. 2). This anomaly 'noses' towards the northwest where the -50 mGal contour defines the limit of the anomaly.

Prior to 1992 very few stations existed in this region, and as defined by those stations the -50 mGal 'nose' extended some 8 km farther to the northwest. It also extended approximately the same distance beyond the boundary of the Amisk Group volcanics, into a region containing relatively high grade metamorphic rocks of diverse affiliation. The northern boundary of the Amisk Group is linear in this area and trends

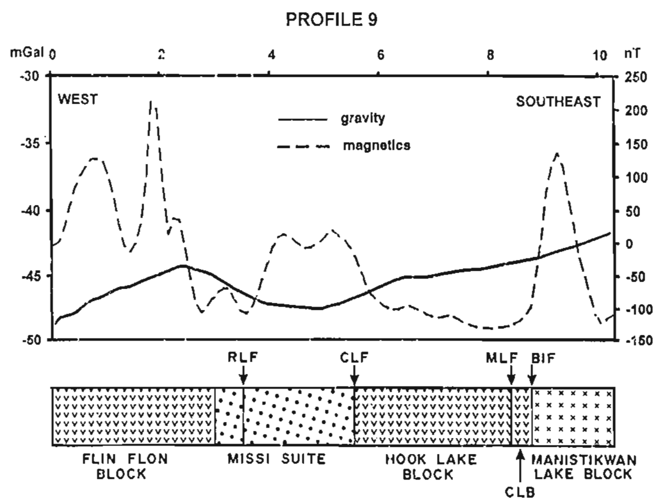

Figure 9. Gravity and aeromagnetic profiles along traverse 9. Schematic geological section based on Bailes and Syme (1989); RLF, Ross Lake fault; CLF, Cliff Lake fault; MLF, Manistikwan Lake fault; BIF, Big Island fault; CLB, Cope Lake Block; v pattern, mafic volcanics of Amisk Group; dot pattern, metasediments of Missi Suite; x pattern, mafic volcanics and diverse intrusive rocks.

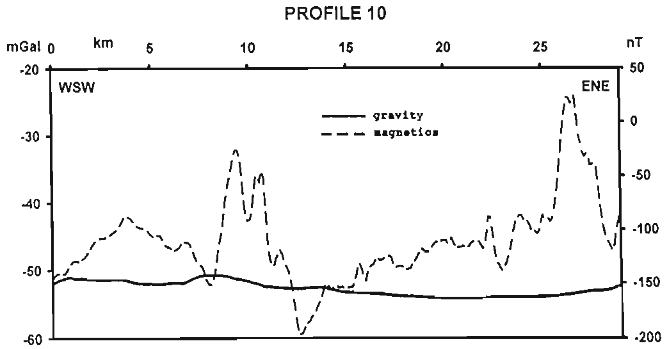

Figure 10. Gravity and aeromagnetic profiles along traverse 10.

west-southwest (Fig. 1). One interpretation of this relationship was that the gravity high signified the presence of the Amisk Group beneath the high grade metamorphic rocks, thereby implying that the boundary might be a north-dipping thrust.

Further investigation of this possibility required better definition of the northwestern extremities of the anomaly. A traverse oriented northwest-southeast across the northwestern flank would have been ideal, but suitable road or lake access was not available. The best alternative was a traverse (number 10) positioned along Kisseynew Lake where gravity observations were made on the shores and islands. The new data (Fig. 10, Profile 10) define a fairly flat and featureless gravity field along the length of the lake. This indicates that gravity high 'A' does not extend beyond the northern boundary of the Amisk Group, which in turn suggests that Amisk Group volcanics do not extend at depth beyond the boundary.

SUMMARY

Detailed gravity data obtained from surveys completed in 1992 in the Cormorant Lake map area indicate that several anomalies are closely associated with geological bodies, thereby offering the potential to derive information on their shape and size. In the region covered by Phanerozoic sediments gravity data provide, in addition to aeromagnetic and drill core data, another constraint for mapping the sub-Phanerozoic basement. Future studies will utilize density and magnetic susceptibility data collected during the summer in modelling the described gravity anomalies and accompanying magnetic anomalies.

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Geological Survey of Canada Projects 840050, 870049, 91003

Sub-Paleozoic geology of the Flin Flon Belt from integrated drillcore and potential field data, Cormorant Lake area, Manitoba and Saskatchewan[1]

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Leclair, A.D., Scott, R.G., and Lucas, S.B., 1993: Sub-Paleozoic geology of the Flin Flon Belt from integrated drillcore and potential field data, Cormorant Lake area, Manitoba and Saskatchewan; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 249-258.

Abstract: A combination of regional drillcore mapping and potential field data interpretation have been used to develop a preliminary geological map that extends known elements of the Flin Flon Belt into the subsurface and delineates four new geological-geophysical domains. The principal elements of the sub-Paleozoic Precambrian basement are: an arcuate, fault-bounded belt of metabasalts (Athapapuskow Domain); linear sequences of intercalated, dominantly amphibolite-facies, mafic metavolcanic, and metasedimentary rocks intruded by felsic and mafic plutons (Clearwater Domain); a heterogeneous complex of variably-deformed granitoid rocks interlayered with mixed mafic, psammitic, and pelitic gneisses (Namew Gneiss Complex); and a large (60 x 25 km) late-tectonic granite batholith (Cormorant Batholith). The Athapapuskow Domain possibly correlates with the MORB-like Athapapuskow Assemblage exposed at the shield margin, while the Clearwater Domain may represent the subsurface continuation of the Flin Flon Assemblage rocks exposed in the File Lake-Snow Lake area.

Résumé : On a combiné des travaux de cartographie régionale à partir de carottes de sondage et l'interprétation de données sur le champ potentiel pour dresser une carte géologique préliminaire indiquant le prolongement en subsurface d'éléments connus de la zone de Flin Flon et permettant de délimiter quatre nouveaux domaines de nature géologique et géophysique. Les principaux éléments du socle précambrien sous-paléozoïque sont les suivants: une zone arquée, limitée par des failles, composée de métabasaltes (domaine d'Athapapuskow); des séquences linéaires de roches métavolcaniques et métasédimentaires interstratifiées, principalement métamorphisées dans le faciès des amphibolites, et traversées par des plutons felsiques et mafiques (domaine de Clearwater); un complexe hétérogène de roches granitoïdes plus ou moins déformées, interstratifiées avec des gneiss mixtes de caractère mafique, calcareux, psammitique et pélitique (complexe gneissique de Namew); et un grand (60x25 km) batholite granitique tarditectonique (batholite de Cormorant). On pourrait peut-être corréler le domaine d'Athapapuskow avec l'assemblage d'Athapapuskow du type basaltes de dorsale médio-océanique (MORB) affleurant sur la marge du bouclier, tandis que le domaine de Clearwater pourrait représenter le prolongement en subsurface des roches de l'assemblage de Flin Flon affleurant dans la région de File Lake-Snow Lake.

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INTRODUCTION

The multi-disciplinary and multi-institutional NATMAP Shield Margin Project was undertaken to generate new perspectives on the geology and evolution of the volcano-plutonic Flin Flon Belt in Manitoba and Saskatchewan (Lucas et al., 1992). The Shield Margin Project area spans from the Thompson Belt (99°W) to the Tabbernor Fault (103°W) and the margin of the Kisseynew Gneiss Belt (55°15'N) to 54°N (Fig. 1). A fundamental component of this project is the development of an interpretive geological map for the subsurface extension of geological elements of the Flin Flon Belt beneath Paleozoic cover rocks that form the eastern edge of the Western Canada Sedimentary Basin. Two complimentary approaches are being undertaken to achieve this objective: (1) an extensive relogging program of Precambrian core recovered during mineral exploration; and (2) interpretation of potential field data (aeromagnetics and gravity). The integration of the new drillcore information and potential fields data analysis forms the basis for the delineation and characterization of rock units and major lithotectonic domains in the subsurface. This paper presents the initial results of this integrated study carried out in the Cormorant Lake sheet (NTS 63K) of the NATMAP project area (Fig. 1)

Under the auspices of 1984-89, Federal-Provincial Mineral Development Agreements, a number of high resolution total magnetic field and gradiometer surveys were carried out over the Flin Flon Belt and its sub-Paleozoic continuation to assist base metal and gold exploration in Manitoba and Saskatchewan (Kiss, 1990). Preliminary interpretion on the sub-Paleozoic geology of the Cormorant Lake sheet integrated the results of these surveys with drillcore data from approximately 100 boreholes (Blair et al., 1988) resulting in a magnetic domain map of the Precambrian basement. An earlier interpretation of the sub-Paleozoic geology of the same area, utilizing aeromagnetic data and a compilation of drill logs from about 1000 boreholes, yielded a pseudo-geological map at 1:250 000 scale (Taiga Consultants Ltd., 1986). A lack of critical evaluation of drill logs originated from numerous authors, however, undermined the utility of the map.

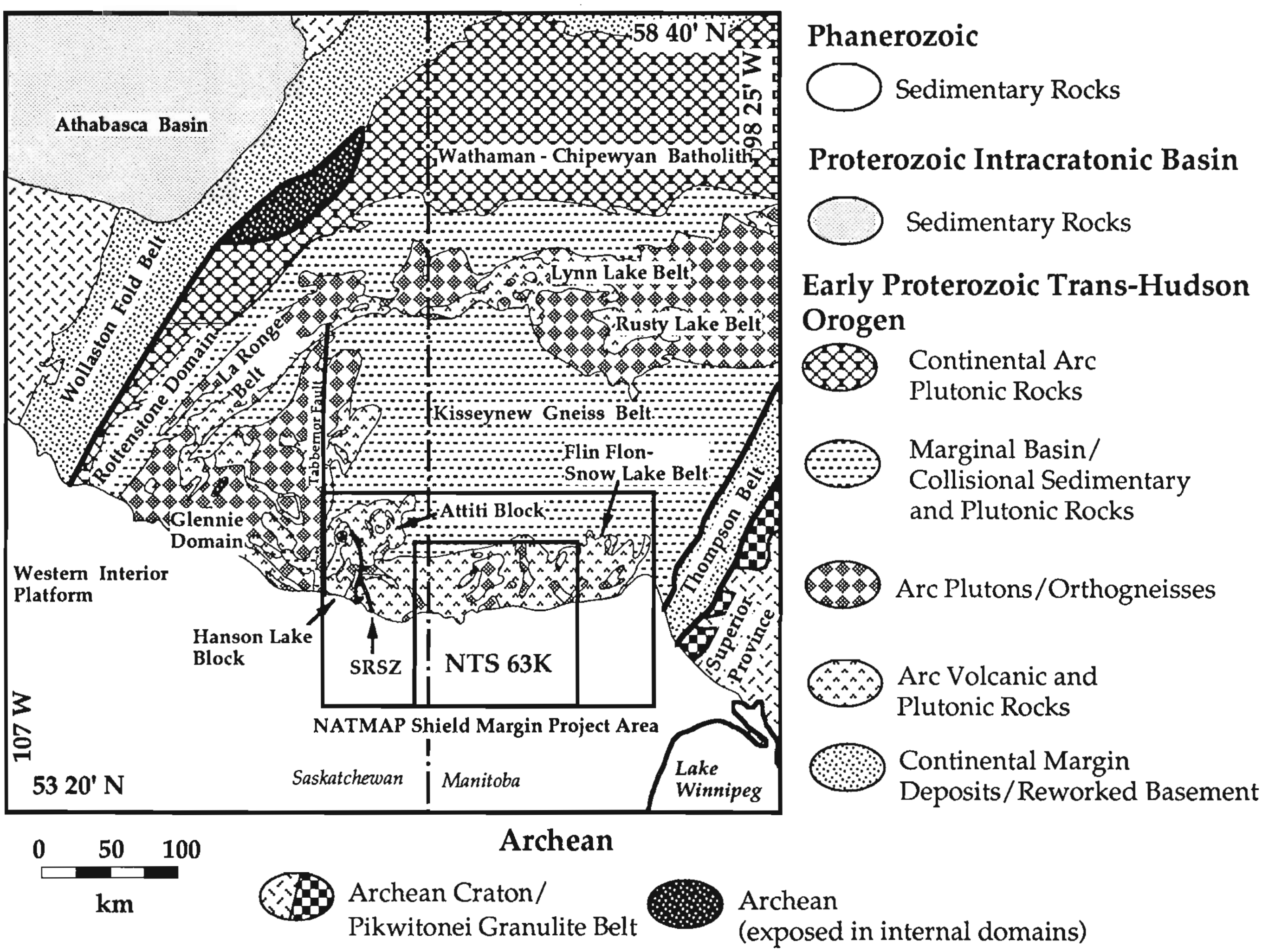

Figure 1. Geological map showing the principal tectonic domains of the Trans-Hudson Orogen in Manitoba and Saskatchewan (after Hoffman, 1988; Lewry et al., 1990) and the location of the NATMAP Shield Margin Project area and the Cormorant Lake sheet (63K). SRSZ – Spruce Rapids Shear Zone.

REGIONAL GEOLOGICAL SETTING

The Flin Flon Belt is part of the Trans-Hudson Orogen, a 500 km wide collision zone of mainly juvenile Early Proterozoic rocks flanked by Archean cratons (Hoffman, 1988). The belt consists of an assemblage of arc and back-arc volcanic and volcaniclastic rocks and associated sedimentary rocks (Amisk Group), a sequence of continental metasedimentary and subordinate metavolcanic rocks (Missi Group) in structural/unconformable contact with the Amisk Group, and a voluminous suite of felsic to mafic intrusive rocks (syn-Amisk to post-Missi). The Flin Flon Belt is bounded to the north by the Kisseynew Gneiss Belt, which includes metamorphosed equivalents of the Amisk and Missi groups (Bailes, 1980a; Ashton et al., 1987; Ashton and Leclair, 1991), and is overlain to the south by flat-lying Ordovician dolomitic limestones and lesser sandstones (Fig. 1). The along strike continuity of aeromagnetic and gravity anomalies indicates that the Flin Flon Belt may extend beneath the Paleozoic cover south to at least 46°N (Green et al., 1985). The Flin Flon Belt has experienced a complex and protracted deformation history marked by pre- to post-Missi structures which accommodated crustal thickening and transpressional tectonics. Deformation was accompanied by subgreenschist- to amphibolite-grade regional metamorphism (Gordon et al., 1990; Lewry et al., 1990).

MAPPING THE SUB-PALEOZOIC PRECAMBRIAN BASEMENT

Methods

A two stage process was employed to generate a new, better constrained interpretation of the Precambrian geology beneath the Paleozoic cover in the Cormorant Lake sheet. First, detailed analysis of the high resolution aeromagnetic data, integrated with the regional gravity data, led to the production of a series of magnetic susceptibility maps (Fig. 2). The sub-Paleozoic geological interpretation was primarily based on the extrapolation of correlations determined between exposed Precambrian geology and the character of potential field anomalies. The maps provided a general framework for the second stage of the process which involved the examination of Precambrian core from more than 300 diamond drillholes. Systematic drillcore mapping of public-domain and confidential Precambrian core recovered by industry in the course of mineral exploration was carried out over a three-month period, following re-examination of the core samples acquired in the Manitoba "scout" drilling project (1984-89 MDA program). The "scout" drillholes were selected to provide a reconnaissance collection of subsurface Precambrian rock types, in contrast to most of the exploration drilling programs which targeted electromagnetically conductive zones. In this study, boreholes were carefully selected for core logging so as to obtain maximum regional coverage and a representative sampling of basement rocks. The initial emphasis of this study has been on the western half of the Cormorant Lake (63K) sheet, accounting for approximately 80% of the Precambrian drillcore examined to date (see Fig. 5). Sub-Paleozoic interpretation was done at 1:50 000 scale by combining drillcore mapping results with potential field data displayed on magnetic susceptibility maps (Fig. 2), high resolution aeromagnetic total field and vertical gradient maps (series of 1:50 000 scale maps available at the Geological Survey of Canada).

Aeromagnetic anomalies

The geological interpretation of aeromagnetic anomalies is potentially complex owing to the non-uniqueness of the signatures (Hinze and Zeitz, 1985; Goodacre, 1991). However, some empirical observations can be made by careful examination of the high resolution aeromagnetic maps for both exposed and buried Flin Flon belt rocks. The observation that distinct anomalies can be traced unbroken from the exposed shield into the Paleozoic-covered area indicates that the thin (<100 m) south-dipping Paleozoic cover is magnetically transparent. In addition, the diminishing magnitude of the anomalies toward the south, with increasing sediment cover thicknesses, suggests that the causative bodies are of basement sources. In general, the aeromagnetic maps show that areas underlain by Amisk Group rocks have moderate to high positive signatures and a distinctive, striated to corrugated texture defined by small wavelength anomalies. Map-scale mafic to ultramafic intrusions are associated with local gravity highs and are characterized by subcircular to elliptical, high amplitude, positive aeromagnetic anomalies with curvilinear internal structure. Granitoid rocks produce either aeromagnetically low-relief neutral regions with little internal fabric, or moderate-intensity positive aeromagnetic highs displaying relatively short wavelength, sinuous internal anomalies.

SUBSURFACE LITHOLOGICAL UNITS

Amisk Group supracrustal rocks and related gneisses

Av: *Mafic volcanic and volcaniclastic rocks and related mafic intrusions* occur in the westernmost part of the study area within a narrow (<12 km wide) fault-bounded, south-trending belt that extends from the shield margin to, at least, the southern edge of the Cormorant Lake sheet (Fig. 3). These rocks are predominantly in greenschist-facies but grade to amphibolite-facies near gabbro and granite plutons in the southwest corner of the area. They consist of dark green to light grey-green, fine grained, laminated to finely-layered and homogeneous metabasites. The rocks probably represent plagioclase-phyric and aphyric lava flows, sills, and volcaniclastic sediments of basaltic to intermediate composition. Related mafic intrusions are medium grained, homogeneous diorite and gabbro, some of which may in part represent subvolcanic feeder dykes. The belt also includes minor felsic volcanic and volcaniclastic rocks and associated intrusive rocks, commonly containing quartz and plagioclase phenocrysts. Rusty staining of both mafic and felsic rocks is attributed to oxidation of disseminated sulphides. Black, graphitic and sulphidic argillite and phyllite occur as minor intercalations within this volcanic package. The **Av** unit is well represented in industry cores, presumably due to the presence of electromagnetic conductive zones.

At the shield margin, the basaltic Athapapuskow Assemblage of MORB affinity (Syme, 1990) has a prominent positive aeromagnetic expression with an internal linear fabric that trends southwestward. The interpreted mafic extrusive and intrusive rocks of unit **Av** are contiguous with the Athapapuskow Assemblage along this distinctive, striated aeromagnetic anomaly. A geochemical study of unit **Av** core samples is underway to test this correlation.

Aa: *Metamorphosed volcanic, volcaniclastic, and sedimentary rocks* form linear, south-southwest-trending units which are intruded by many felsic and mafic plutons. These rocks make up a large part of the Precambrian basement around the central Cormorant Batholith, especially in the eastern half of the sheet (Fig. 3), and are correlated with the Amisk Group (Flin Flon Assemblage) on the exposed shield. The unit **Aa** appears to grade westward into gneisses (subunit **Aag**) which are interpreted as the highly-deformed and metamorphosed equivalents of **Aa**. **Aa** consists of intercalated sequences of predominantly amphibolite-facies mafic metavolcanic and metasedimentary rocks. The mafic rocks comprise hornblende-biotite±garnet schists and gneisses; more homogeneous garnet amphibolite is interpreted to be the high-grade equivalent of mafic volcanic and volcaniclastic rocks and related diorite and gabbro sills. Metasedimentary rocks include psammites and pelites and contain biotite±garnet±sillimanite±graphite metamorphic assemblages. These rocks are interpreted to be derived from greywacke, siltstone, and shale.

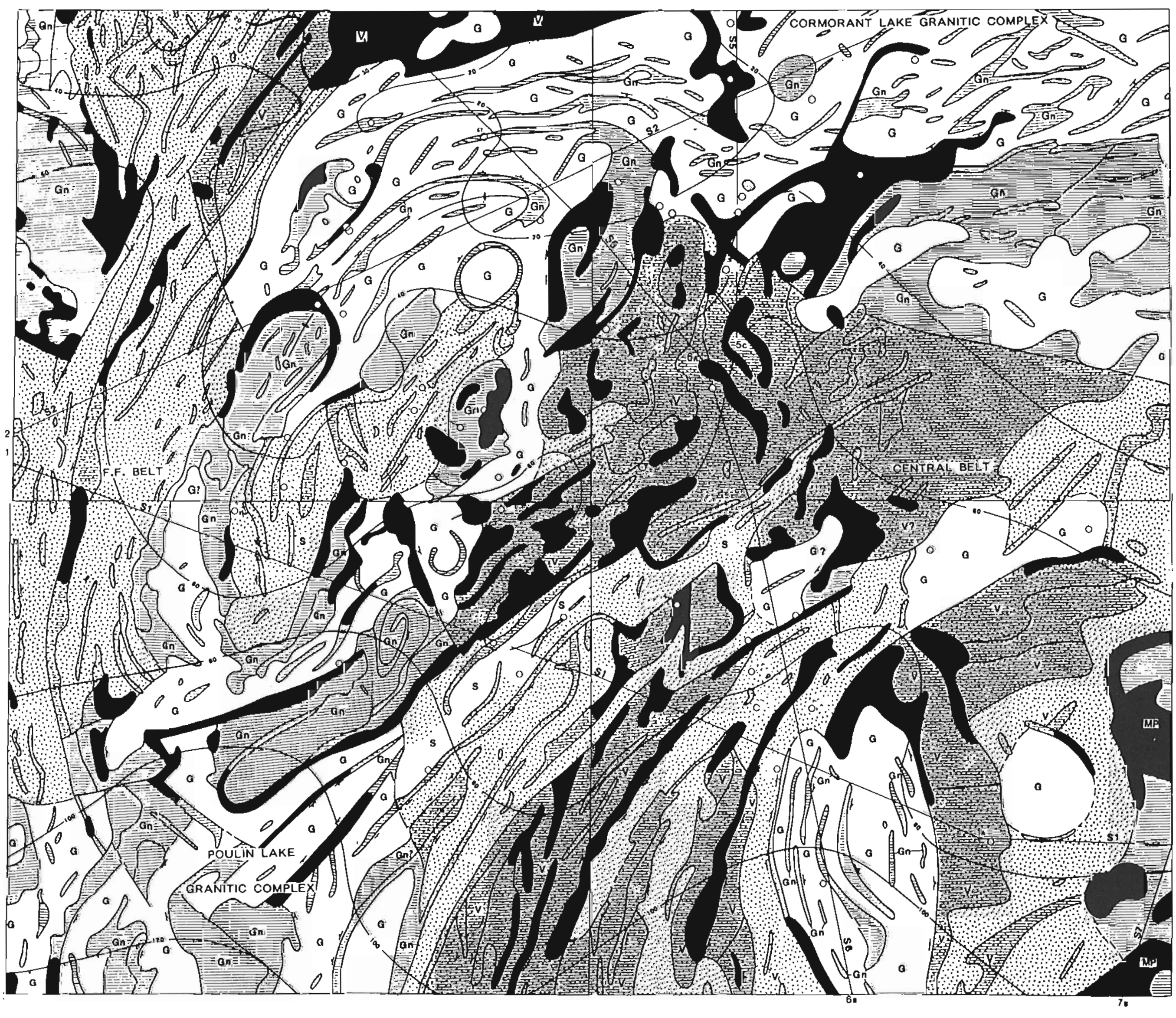

Figure 2. Map of the southwestern quadrant of the Cormorant Lake (63K) sheet (NATMAP Shield Margin Working Group) showing the distribution of four magnetic susceptibility units and their preliminary lithological interpretation. Borehole data is taken from public domain information.

Aag: *Mixed metasedimentary, mafic, and granitoid gneisses* make up a subunit of **Aa** in the western part of the map area (Fig. 3). These heterogeneous gneisses are fine- to medium-grained and are layered on the scale of a few centimetres to tens of metres. They occur mainly as thin (<3 km wide) elongated units within engulfing tonalite, granodiorite, and quartz diorite bodies (subunit **Ta**; see below). **Aag** is principally composed of amphibolite-facies mafic, psammitic, and pelitic gneisses of Amisk Group affinities, with subordinate layers of tonalite-granodiorite and quartz diorite which are inferred to be lit-par-lit injections.

The dark green mafic gneiss metamorphic assemblage is hornblende-plagioclase±biotite±garnet±clinopyroxene. It grades into abundant grey-green gneisses of more Ca-rich composition, which contain a relatively higher proportion of clinopyroxene and Ca-amphiboles, including hornblende and tremolite, and locally graphite and epidote. Pale green layers composed of clinopyroxene-plagioclaseætitanite are associated with the mafic and relatively Ca-rich gneisses. Ashton and Leclair (1991) have interpreted similar layers to represent the high-grade equivalents of the pervasive epidote-altered mafic volcanic-volcaniclastic rocks of the Flin Flon belt.

MAGNETIC SUSCEPTIBILITY UNITS

highly magnetized zone (200 to 3000 gammas)

moderately magnetized zone (10 to 100 gammas)

reversely magnetized zone (-20 to -200 gammas)

non-magnetic zone

LITHOLOGICAL UNITS

	inferred lithology	magnetization
G	granitic rocks	nil
Gn	granitic rocks	moderate
MP	mafic intrusive rocks	high
S	metasedimentary rocks	nil
Sm	metasedimentary rocks	moderate to high
V	metavolcanic rocks	low to moderate

magnetic bedding attitude

magnetic contact

O borehole location

1——1A location of interpreted geological section

—80— countour of depth to Precambrian basement (in meters) based on magnetic depth determinations and drillcore data

km

0 5

Legend to Figure 2

The psammitic gneisses generally contain graphite and are characterized by biotite-rich assemblages with or without hornblende and garnet. They grade into less abundant schists and gneisses of more aluminous composition with muscovite-biotite±garnet±sillimanite assemblages. These pelitic rocks are graphite-bearing, and can contain up to about 30% garnet porphyroblasts with a typical lavender colour.

The heterogeneity of the gneissic rocks of subunit **Aag** suggests that they were derived from a compositionally-diverse assemblage including mafic to intermediate volcanic and related tuffaceous rocks, calcic wackes, pelites, and arenites. More homogeneous amphibolite layers may represent synvolcanic diorite and gabbro sills. The combined effects of high-grade metamorphism and deformation, however, preclude definitive characterization of all precursors, as well as a clear distinction between inferred calcic wackes and mafic volcaniclastic sediments. The presence of graphite generally supports a sedimentary precursor for pelites, psammites, and Ca-rich mafic gneisses. The presumed volcanic, volcaniclastic, and sedimentary character of these gneisses is an integral part of their interpretation as metamorphosed and deformed equivalents of Amisk Group supracrustal rocks. Furthermore, the gneiss package appears to resemble the high-grade Amisk Group rocks found in the Attitti Block and the southern flank of the Kisseynew Gneiss Belt (cf. Ashton and Leclair, 1991; Ashton et al., 1987; Bailes, 1980a).

Mafic and ultramafic intrusions

Ga: *Gabbro-diorite* occurs as distinct, subcircular plutons marked by positive aeromagnetic highs. In drillcore, the unit is dark to brownish green, medium- to coarse-grained, massive, and homogeneous. It contains hornblende and biotite± clinopyroxene and orthopyroxene, and has accessory apatite, tourmaline, titanite, and magnetite. Subophitic texture and igneous layering are locally preserved. While **Ga** is dominated by mafic gabbros and diorites, it also includes leucogabbro and melanocratic quartz diorite. These latter units appear to be associated with a suite of isolated layered intrusions which include ultramafic phases.

Um: *Pyroxenite-melagabbro/diorite* forms crosscutting to concordant plutons and sills with moderate to high aeromagnetic signatures. These intrusions occur sporadically over the entire area, although only the larger, map-scale bodies are delineated on Figure 3. The rocks of unit **Um** are in general dark green, medium- to coarse-grained, massive, and locally magnetiferous. They show various degrees of alteration, with phlogopite, Ca-amphiboles, epidote, chlorite, talc, carbonate, and serpentine representing the principal alteration products. Aggregates of these minerals may be pseudomorphing primary olivines and pyroxenes. Cumulate-textured layers with oikiocrystic orthopyroxene, clinopyroxene, and phlogopite have been observed in drillcore.

Felsic and intermediate intrusions

T: *Tonalite-granodiorite-quartz diorite* occur as several subcircular to elongate plutons emplaced in the supracrustal assemblage **Aa** (Fig. 3). Unit **T** is subdivided into subunit **Ta** in order to distinguish orthogneiss associated with a dominantly intrusive complex in the western part of the area (Fig. 3) from the less deformed plutonic rocks of unit **T**. The hornblende-biotite tonalite, granodiorite, and quartz diorite of unit **T** are light grey to grey, fine- to medium-grained, leucocratic to mesocratic, and heterogeneously foliated. Unit **T** also contains fine grained, equigranular leucotonalite containing less than 10% biotite.

Ta: *Gneissic tonalite and quartz diorite with mafic and metasedimentary enclaves* form a large complex in the western part of the map area (Fig. 3). The complex consists mainly of variably-deformed plutonic rocks interlayered with numerous mafic and metasedimentary enclaves, some of which occur as lenticular map-scale units (**Aag**; Fig. 3). Subunit **Ta** is marked by structural and lithological heterogeneity compared to unit **T**. Underground examination of subunit **Ta** at the Namew Lake Ni-Cu mine (Fig. 3) highlighted the presence of multiple plutonic phases, perhaps representing several magmatic episodes. The plutonic rocks are mainly hornblende-biotite±epidote±titanite±magnetite tonalite and quartz diorite, but also include relatively minor

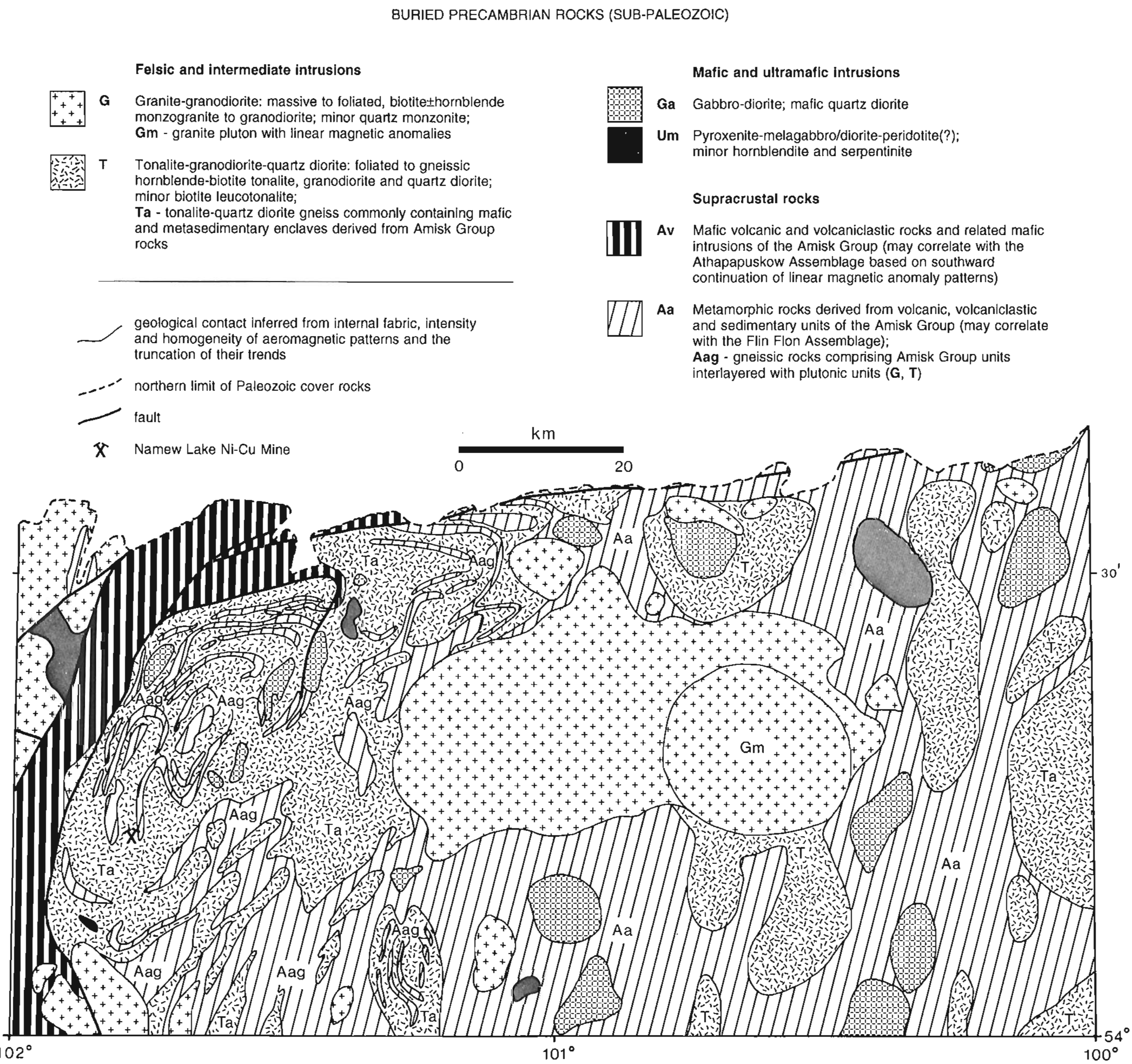

Figure 3. Simplified geological compilation map of subsurface Precambrian basement rocks in the Cormorant Lake (63K) sheet.

granodiorite. They vary texturally from fine- to medium-grained, and display a range of planar tectonic fabrics from a heterogeneously-developed foliation to a relatively homogeneous gneissosity. Well-layered mesocratic to melanocratic tonalite and diorite gneiss is locally important in the gneiss complex, although the diorite end-member is less important than the tonalite. The layering in the tonalite-diorite gneiss is on the centimetre-scale, and is defined by varying proportions of hornblende and biotite. Preliminary petrographic examination indicate that the tonalite contains up to 35% quartz and rare K-feldspar, with epidote, titanite, apatite, and zircon as accessory phases.

The orthogneisses surrounding the Namew Lake orebody yielded U-Pb zircon ages that range from ca. 1890 to 1860 Ma (Cumming and Krstic, 1991). Initial U-Pb geochronological and Nd isotopic results elsewhere within the complex suggest that tonalites associated with subunit **Ta** are possibly synvolcanic and isotopically similar to exposed Amisk Group rocks (R. Stern, pers. comm., 1992). These results suggest a parallel magmatic evolution between the gneiss complex and the volcano-plutonic sequence of the Flin Flon belt. The amphibolite-grade, penetratively deformed nature of the gneiss complex further suggests that it may represent a mid-crustal segment of the Flin Flon arc.

G: *Granite-granodiorite* forms intrusive bodies with dimensions ranging from batholith-scale down to centimetre-scale dykes and veins. The most conspicuous unit **G** body is the roughly rectangular-shaped Cormorant Batholith which occupies an area of approximately 60x25 km in the south-central part of the Cormorant Lake sheet (Fig. 3). Unit **G** is dominated by monzogranite, but also includes granodiorite, syenogranite, and minor quartz monzonite. These light grey to pink, homogeneous, medium- to coarse-grained rocks are only locally foliated. Mafic phases are biotite±magnetite, with minor hornblende and epidote, which altogether make up less than 10% of the rock. Accessory minerals include apatite, zircon, and rare titanite and garnet. Some monzogranite bodies are porphyritic, with K-feldspar megacrysts up to 4 cm long. Dykes and veins commonly have pegmatitic textures, although medium grained, equigranular phases have been observed.

In general, the intrusive rocks of unit **G** truncate the trend of aeromagnetic anomalies and crosscut all other rock types in drillcore. These field relations imply that they represent the youngest intrusive rocks and were emplaced late in the tectonomagmatic history of the area. A U-Pb zircon age of 1845 +10/-8 Ma for the Cormorant Batholith (Blair et al., 1988) substantiates this relative chronology.

Gm: *Granite with linear aeromagnetic anomalies*, a subunit of **G**, comprises one subcircular body which forms an integral part of the southeastern Cormorant Batholith (Fig. 3). This body is characterized by a semiconcentric aeromagnetic pattern on its margins and discrete linear north-northwest-trending anomalies within its core. This peculiar geophysical expression may reflect primary igneous layering derived from multiple intrusions. Drillcore data from a single borehole in the southern margin of the **Gm** body indicates a pink, fine- to medium-grained, commonly pegmatitic, sphene-bearing granite (Weber, in press), similar to granites observed elsewhere in the Cormorant Batholith.

PHYSICAL PROPERTIES OF THE DRILLCORE

Density

Density measurements were obtained from 667 representative core samples using a Mettler BB3000 high-precision digital balance. The results (Fig. 4) were subdivided into six main lithological groups. In general, the felsic to intermediate metavolcanic and plutonic rocks have similar low densities, with >95% of the densities lying in the range 2.6-2.9 gcm^{-3} (Fig. 4a). The higher density tail to these distributions (2.8-3.1 gcm^{-3}) are attributed to more mafic tonalite-diorite units. Pelitic to psammitic rocks have density values between 2.7 and 3.1 gcm^{-3}, the mean value lying between 2.8 and 2.9 gcm^{-3} (Fig. 4a). Mafic metavolcanic/ volcaniclastic rocks, mafic intrusions, and compositionally equivalent mafic

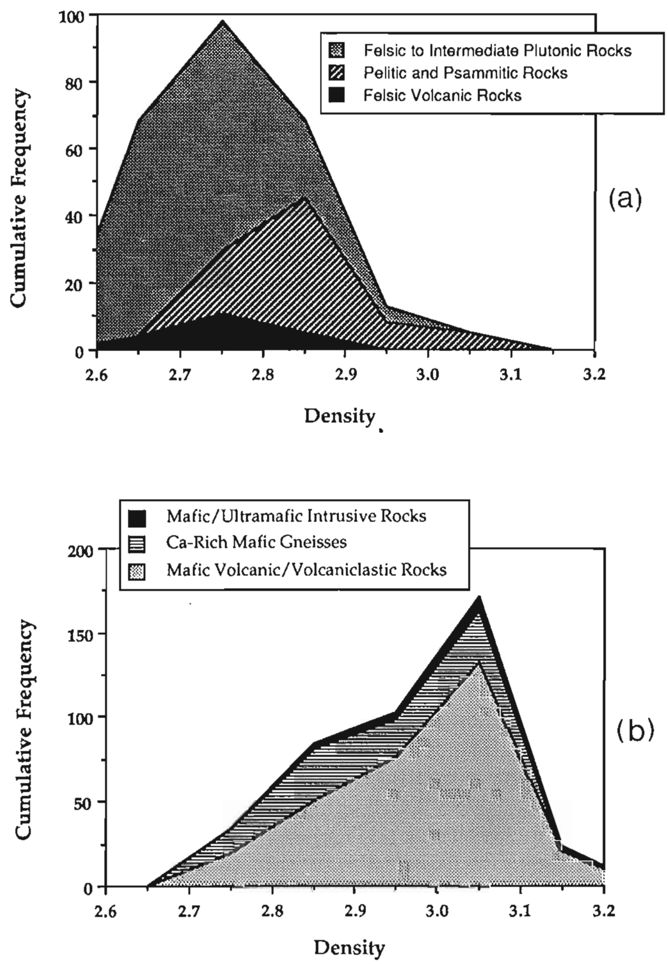

Figure 4. Cumulative frequency plots comparing the density of six main lithological groups of the sub-Paleozoic Precambrian basement in the Cormorant Lake sheet. (**a**) Lower density metavolcanic, metasedimentary, and plutonic rocks; (**b**) Higher density metavolcanic, metasedimentary, and intrusive rocks.

gneisses yield relatively high densities (highest frequencies between 2.9 and 3.1 gcm^{-3}; Fig. 4b). The tail toward lower values on the frequency distribution of densities may reflect a gradation towards more intermediate compositions. Unaltered ultramafic rocks and gabbro-diorite occupy the upper portion of the density spectrum (2.9-3.2 gcm^{-3}; Fig. 4b). Ca-rich mafic gneisses have a bimodal density distribution, with a peak between 2.8-2.9 gcm^{-3}, and a second peak 3.0-3.1 gcm^{-3} matching the peak in the density distribution for the mafic rocks (Fig. 4b).

The density measurements are comparable to Gibb's (1968) database derived for Precambrian rocks from northern Manitoba. Densities will be used in the interpretation and modelling of new high resolution gravity data for Paleozoic-covered areas which will be acquired in support of the NATMAP project. Gravity surveys represent an important exploration tool for locating high density ultramafic bodies associated with deposits such as the Namew Lake Ni-Cu orebody.

Magnetic susceptibility

Magnetic susceptibility measurements were made on all core lithologies, with 8-10 measurements per lithology, using a hand-held KT-5 Exploranium instrument with a sensitivity of 1×10^{-5} SI units. In general, granitoid rocks exhibit relatively low susceptibility ($<2\text{-}3 \times 10^{-3}$ SI units, without correction for cylindrical shape of drillcore). Mafic and ultramafic rocks normally have high susceptibility values (up to $\sim 30\text{-}40 \times 10^{-3}$ SI units). Other rock types display a wide range of magnetic susceptibility. Disseminated magnetite and pyrrhotite can occur in all rock types throughout the region and thus correlation of lithologies with magnetic anomalies must be done cautiously and with the integration of other data sets (e.g., gravity).

STRUCTURE

Primary structures are difficult to recognize in core samples and have been largely obliterated by the combined effects of deformation and amphibolite-facies regional metamorphism. They comprise rare amygdules and pillow margins in metavolcanic rocks, igneous layering in mafic intrusions, and quartz and plagioclase phenocrysts in subvolcanic porphyries. The most prominent structural element observed in core samples is a foliation defined by an alignment of biotite and/or hornblende. The foliation appears to become a gneissic layering as the proportion of intrusive veins and sheets increases (e.g., units **Ta**, **Aag**), while the inferred primary layering in metavolcanic and metasedimentary rocks appears to have been transposed into this foliation. As most exploratory drillholes have inclinations of 60° and azimuths normal to the direction of the targeted linear geophysical anomaly, measurements of core-axis/foliation angles indicate that foliations generally have steep to moderate dips. This is consistent with a steeply-dipping magnetic bedding inferred from the analysis of the high resolution aeromagnetic data (Fig. 2).

The regional structural "grain" of the subsurface Precambrian rocks trends south-southwest, as inferred from the predominant trend of aeromagnetic anomalies (Fig. 5). This generally mimics structural trends north of the shield margin and suggests that exposed geological elements and domains of the Flin Flon Belt can be traced along strike beneath the Paleozoic cover. Curvilinear aeromagnetic anomalies appear to outline a series of north-trending tight folds within some supracrustal sequences (e.g. southwest end of Clearwater Domain, Fig. 5).

SUBDIVISION OF THE BURIED PRECAMBRIAN BASEMENT

The combination of regional drillcore mapping and potential field data interpretation permits the preliminary subdivision of the buried Precambrian rocks into four broad geological-geophysical domains that in part correlate with elements of the Flin Flon Belt. These domains are: Athapapuskow Domain, Clearwater Domain, Namew Gneiss Complex, and Cormorant Batholith (Fig. 5).

Athapapuskow Domain

The Athapapuskow Domain is characterized by the fault-bounded mafic rocks of unit **Av** which occur along a large south-trending positive aeromagnetic anomaly (Fig. 5). Preliminary interpretation suggests that it represents the subsurface continuation of the MORB-like Athapapuskow Assemblage exposed at the shield margin. The arcuate Athapapuskow Domain is bounded on its west by the Namew Lake Structure (Syme, 1988) and on its east by the inferred sub-Paleozoic continuation of the South Athapapuskow Lake Fault (Fig. 3, 5). The inferred sub-Paleozoic extension of the southeast-striking Spruce Rapids Shear Zone (Fig. 1; Ashton, 1990) appears to be truncated against the Namew Lake Structure (Fig. 5).

Clearwater Domain

The Clearwater Domain is a broad, south-southwest-trending volcano-plutonic domain with a generally corrugated aeromagnetic pattern (Fig. 5). This lithotectonic domain appears to include sequences of intercalated metavolcanic and metasedimentary rocks of unit **Aa** and intrusions that belong to units **Ga**, **T**, and **G**. It extends southward from the shield edge to the bottom of the Cormorant Lake sheet and is intruded on its western side by the Cormorant Batholith (Fig. 5). Although drillcore control is currently sparse in the area, the Clearwater Domain is tentatively interpreted to represent the subsurface extension of the supracrustal-plutonic rock package of the Flin Flon Assemblage that is exposed in the File Lake-Snow Lake area (cf. Bailes, 1980b). This suggestion is supported by the unbroken continuation of aeromagnetic and gravity anomalies between the exposed and buried parts of the Flin Flon belt in the eastern part of the study area.

Namew Gneiss Complex

The Namew Gneiss Complex is a region dominated by aeromagnetically-neutral, variably-deformed tonalite, quartz diorite, and minor diorite and granodiorite (subunit **Ta**) mixed with subordinate supracrustal rocks (subunit **Aag**) (Fig. 5). The **Aag** bands form curvilinear, cusp-shaped aeromagnetic highs. This domain is inferred to be in tectonic contact with the Athapapuskow Domain to the west, and appears to be transitional into the Clearwater Domain towards the east. It also appears to be crosscut by the Cormorant Batholith (Fig. 5). The complex may have been built up from multiple intrusion of tonalite-diorite sheets at mid-crustal levels. Preliminary geochronological and isotopic evidence suggests that the Namew Gneiss Complex may represent part of the Flin Flon magmatic arc (R. Stern, pers. comm., 1992).

Cormorant Batholith

The Cormorant Batholith is probably the most conspicuous geological element of the sub-Paleozoic Precambrian basement (Fig. 5). It is characterized by an overall aeromagnetically-neutral pattern which locally displays subtle semi-concentric and linear anomalies. The massive granite batholith and satellite intrusions clearly crosscut the structural trend of the Clearwater Domain and Namew Gneiss Complex, indicating that their emplacement may have been relatively late in the tectonic history of the area.

ACKNOWLEDGMENTS

Werner Weber is gratefully acknowledged for providing logistical assistance and stimulating discussions over the entire drillcore mapping program. Ken Ashton, Al Bailes, Eric Syme, and Dave Thomas are thanked for leading superb field trips in the Flin Flon and Kisseynew gneiss belts. We are grateful to Ted Baumgartner (Hudson Bay Exploration and Development Co. Ltd.) and John Pearson (Cominco Ltd.) for their co-operation with this study. Michael Lesher, John Pearson, Jim Pickell, Rob MacQuarrie, Thomas Menard, and Richard Stern are thanked for discussions about the subsurface geology, and Namew Lake mine geologists Peter Oshust and J.J. O'Donnell are thanked for an interesting underground tour. Fred Hyndman and Dave Prouse are acknowledged for their help while working at the core library in The Pas. The manuscript was reviewed by Richard Stern and Werner Weber.

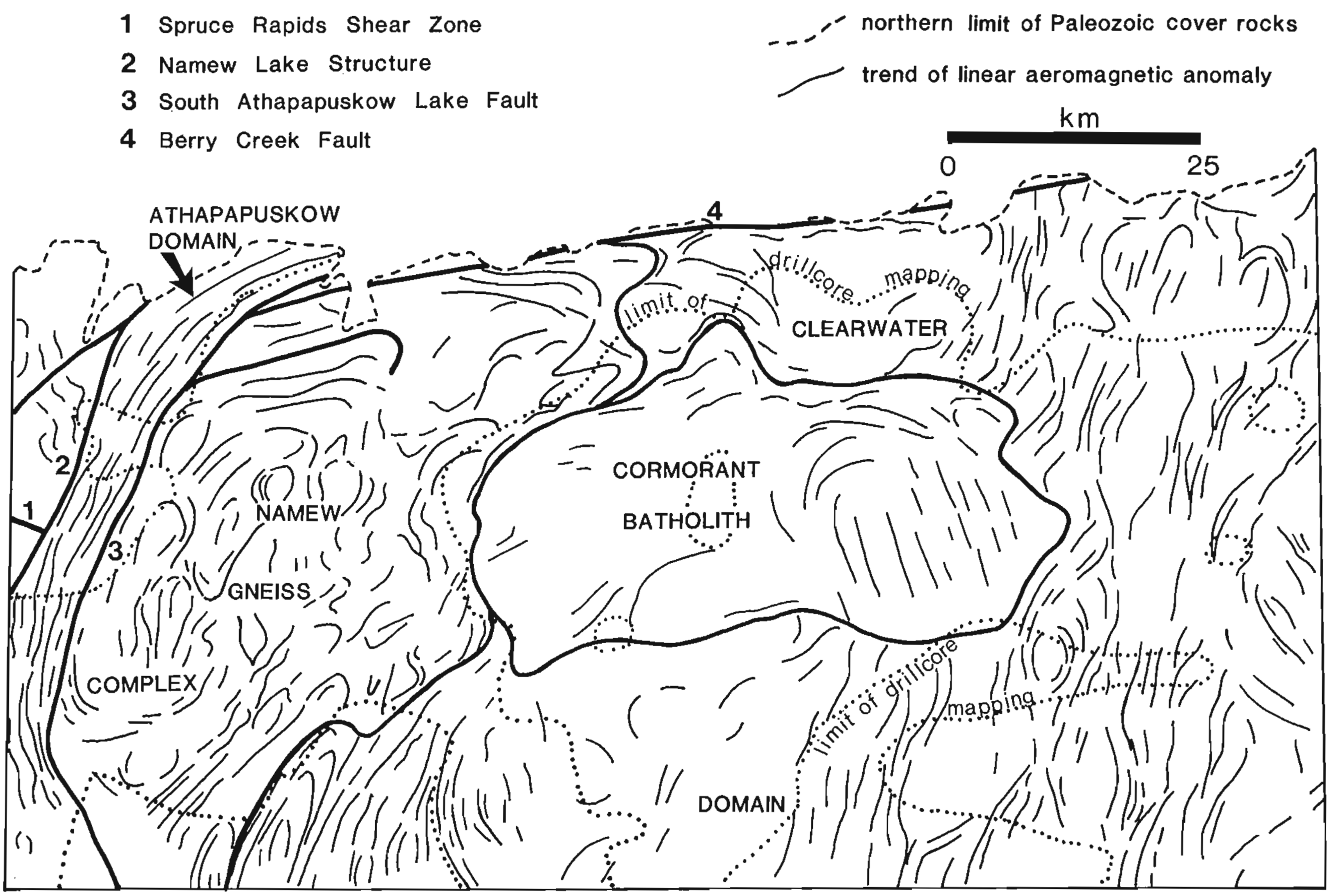

Figure 5. Geological-geophysical domains of sub-Paleozoic Precambrian basement in the Cormorant Lake sheet.

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Geological Survey of Canada Project 910035

Komatiites in the Garner Lake-Beresford Lake area: implications for tectonics and gold metallogeny of the Rice Lake greenstone belt, southeast Manitoba[1]

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Brommecker, R., Scoates, R.F.J., and Poulsen, K.H., 1993: Komatiites in the Garner Lake-Beresford Lake area: implications for tectonics and gold metallogeny of the Rice Lake greenstone belt, southeast Manitoba; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 259-264.

Abstract: Previously unrecognized komatiites have been discovered in the Garner Lake-Beresford Lake area. They contain 21 to 25 wt.% MgO (anhydrous), display spinifex and polysutured textures, and are intercalated with magnesian tholeiitic basalt (8-10% MgO) and calc-alkalic metabasalt (4-7% MgO). This northward-facing sequence occurs stratigraphically above the Garner Lake layered peridotite-pyroxenite intrusion which is inferred to represent a magma chamber for the ultramafic flows. The presence of komatiitic rocks in the Rice Lake belt is significant in that they may be correlative with pre-2.8 Ga supracrustal rocks in the Red Lake belt to the east. The Garner Lake komatiites are separated from more voluminous 2.73 Ga felsic and intermediate volcanic rocks by the major north- northwest-striking Moore Lake-Beresford Lake shear zone. This early structure may have played an important regional control on gold deposits in the Rice Lake belt.

Résumé : Des komatiites jusque-là non identifiées ont été découvertes dans la région de Garner Lake-Beresford Lake. Elles contiennent 21 à 25 % de MgO (anhydre) en poids, présentent des textures de type spinifex et polysuturées et sont intercalées avec des basaltes tholéiitiques magnésiens (8 à 10% de MgO) et des métabasaltes calco-alcalins (4 à 7 % de MgO). Cette séquence orientée vers le nord se situe stratigraphiquement au-dessus de l'intrusion stratifiée de péridotite et pyroxénite de Garner Lake, intrusion que l'on suppose pourrait correspondre à l'une des chambres magmatiques dont seraient issues les coulées ultramafiques. La présence de komatiites dans la zone de Rice Lake est significative, celles-ci se laissant peut-être corréler avec des roches supracrustales mises en place il y a plus de 2,8 Ga dans la zone de Red Lake à l'est. Les komatiites de Garner Lake sont séparées des roches volcaniques felsiques et intermédiaires, plus volumineuses, datées de 2,73 Ga, par la grande zone de cisaillement de Moore Lake-Beresford Lake, de direction nord-nord-ouest. Cette ancienne structure a peut-être joué un rôle régional important du point de vue de la genèse des gîtes aurifères de la zone de Rice Lake.

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[2] Westminer Canada Ltée, Chibougamau

INTRODUCTION

The Archean Rice Lake greenstone belt is located in the western part of Uchi subprovince in Superior Province. In the central part of Uchi subprovince (Red Lake, Confederation Lake) the greenstone belts are composed of juxtaposed slices of pre-2.8 Ga volcanic sequences that locally contain komatiitic rocks, quartzite, and iron-formation and post-2.8 Ga sequences composed mainly of felsic and intermediate volcanic rocks. The only volcanic rocks for which radiometric ages are available in the Rice Lake belt are of the second type (Turek et al., 1989)

Geological mapping by the senior author in the 1989 field season revealed the presence of meta-pyroxenitic rocks of komatiitic composition in the Garner Lake-Beresford Lake portion of the belt. As part of an ongoing compilation of the structure of the Rice Lake belt and its gold deposits, a more detailed re-examination of some of the key outcrops in the 1992 field season resulted in the recognition of textures in the ultramafic rocks identifying them as volcanic flows. The recognition of these komatiites lends new importance to the Moore Lake-Beresford Lake Shear Zone which separates them from the mafic, intermediate and felsic volcanic rocks in the central part of Rice Lake belt.

GEOLOGICAL SETTING

The Garner Lake-Beresford Lake area lies in the eastern part of the Rice Lake belt (Fig. 1a). New mapping by the senior author has shown that the rocks in this area occur in three distinctive fault-bounded domains (Fig. 2). The folded rocks west of the Moore Lake shear zone comprise a relatively coherent stratigraphic package of basalt, dacitic volcanic breccia, tuff, and siltstone that defines the Bidou Lake Subgroup of the Rice Lake Group. The volcanic rocks are overlain by metagraywacke and argillite of the Edmunds Lake Formation. The rocks between the Moore Lake and Beresford Lake shear zones are more highly strained and transposed and are correlative with the upper part of the Bidou Lake volcanic rocks and the metagraywacke of the Edmunds Lake Formation. This eastward-facing panel of volcanic and sedimentary rocks is detached from the western mass of Bidou Lake and Edmunds Lake units. The rocks east of the Beresford Lake shear zone are composed of a variety of lithologies. Although most of these rocks are shown on compilation maps (Weber, 1971a) to be correlative with metagraywacke units of the Edmunds Lake Formation, they are in fact composed of a north- to northwestward-facing sequence of mainly metabasalt, minor banded iron-formation, and abundant gabbroic sills. Some of the mafic rocks, however, have been previously included in the Banksian Lake Formation of the Gem Lake Subgroup of the Rice Lake Group (Weber, 1971b). The Garner Lake layered ultramafic sill intrudes metasedimentary rocks that are inferred to underlie the mafic rocks (Fig. 2).

The structural history of this area involves three increments of deformation (Brommecker et al., 1989). Ductile D1 deformation is recorded most intensely in and to the east of a kilometre-wide structural zone, herein referred to as the Moore Lake-Beresford Lake shear zone (Fig.2). It is bounded by narrower discrete structures, the Moore Lake shear zone on the west and the Beresford shear zone on the east. Within it and to the east of it, steep S1 foliation and numerous layer-parallel D1 shear zones are common. D2 deformation records shortening about a northeast-southwest axis across the entire area and is represented by intense steep northwest-striking cleavage containing extension lineations that plunge steeply southeast, by upright F2 folds and by abundant ductile shear zones. D3 deformation, represented by east- northeast-trending minor folds, axial planar crenulation cleavage, conjugate kink bands, and brittle-ductile faults, is likely related to dextral movement on the Wanipigow Fault.

KOMATIITIC ROCKS

The komatiites occur east of the Beresford Lake shear zone in a series of large outcrops 800 m north of the northwest corner of Garner Lake. North- and northwesterly-striking D1 and D2 shear zones are prominent here and are responsible for considerable disruption of the volcanic sequence (Fig. 2). Pillowed and massive mafic flows, with rare thin interflow

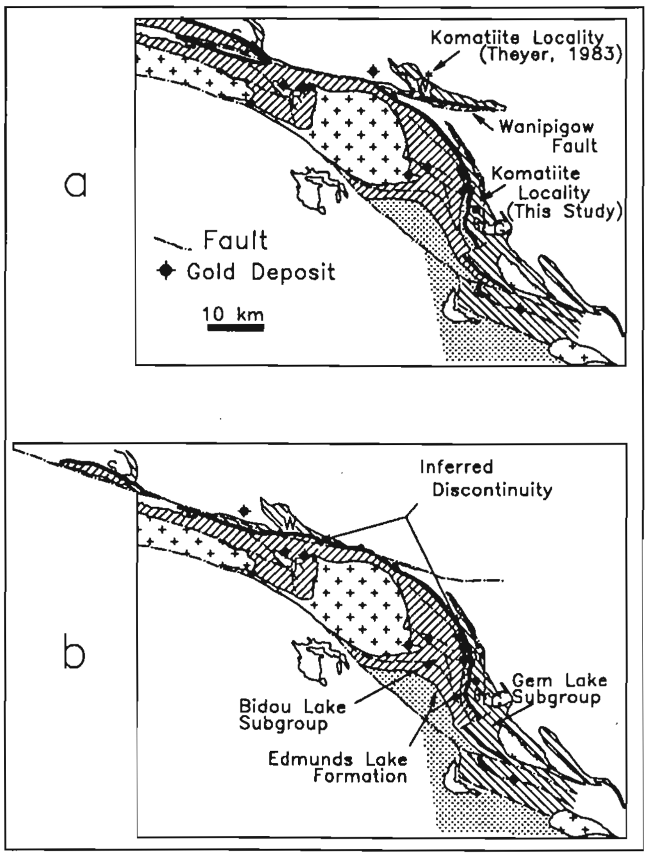

Figure 1. a) Geological sketch of the Rice Lake greenstone belt showing the regional setting of komatiitic rocks and gold deposits. Identified lakes include Saxton (S), Rice (R), Wallace (W) and Garner (G). **b)** As in a) but with restoration of 22 km of dextral strike slip offset along the Wanipigow Fault.

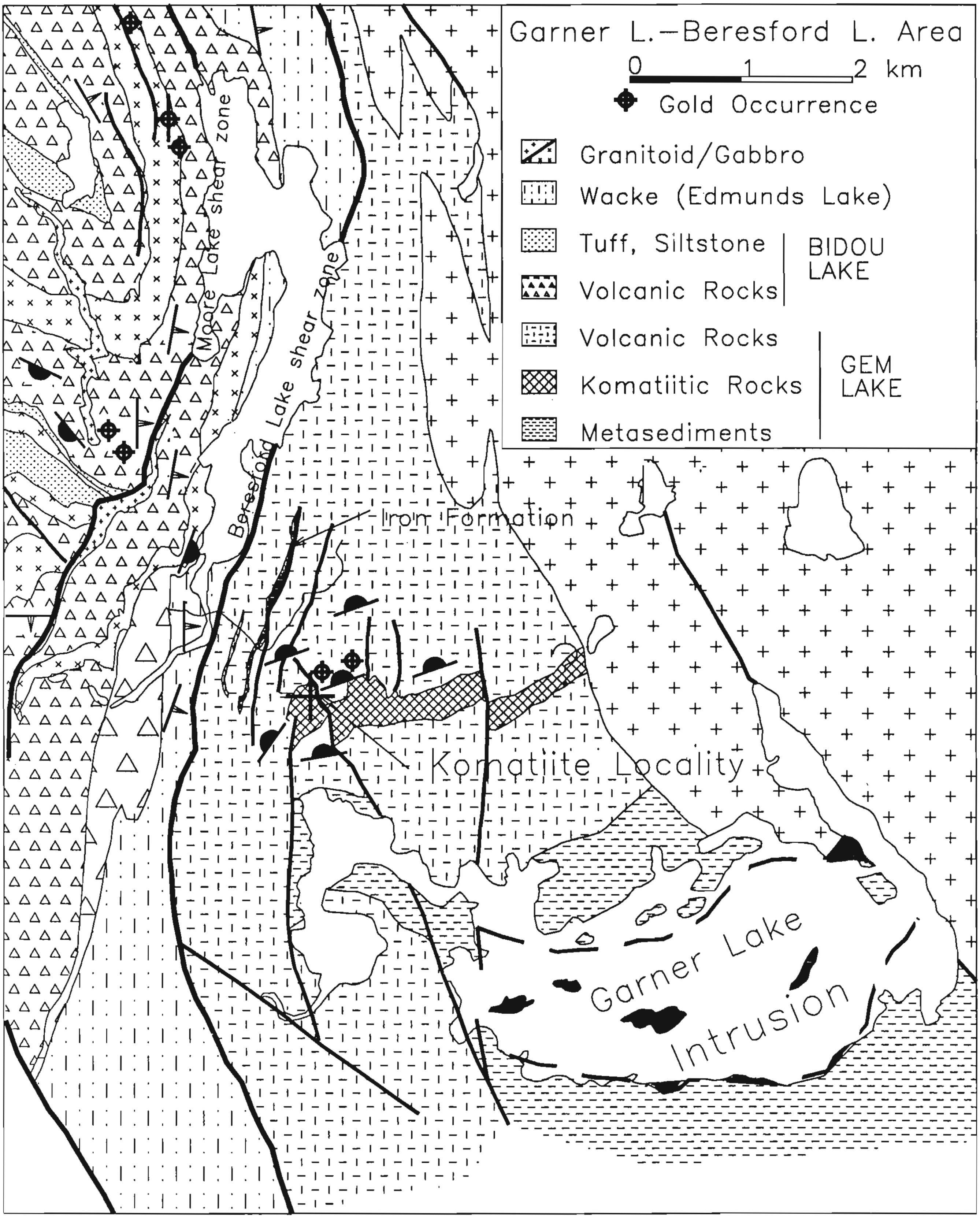

Figure 2. Revised geology of the Garner Lake-Beresford Lake area, Rice Lake greenstone belt. Tops of pillowed flows and sedimentary beds are shown by conventional symbols.

sedimentary units of black mudstone, siltstone, and iron-formation, constitute most of the outcrops in this area. The mafic flows compositionally include both calc-alkalic and high-Mg tholeiitic varieties (Fig. 3). The former contain 4-7% MgO whereas the latter contain 8-10% MgO. Some mafic flows are locally variolitic and several pillows contain good examples of drainage cavities that, along with pillow shape, indicate that the volcanic sequence here faces northward.

The komatiites occur as thin units intercalated with the basaltic flows (Fig. 2). They have been metamorphosed to assemblages consisting mainly of tremolite-actinolite and minor amounts of chlorite, carbonate, serpentine, talc, and iron oxides. These rocks contain approximately 24% MgO (anhydrous) and are compositionally komatiites (Fig. 3). Three observed textural features indicate that they are volcanic flows. First is the presence of spinifex texture. One 10-15 cm thick spinifex textured unit was traced along strike for 10 m parallel to an inferred flow contact. It is characterized by a lower weakly foliated zone overlain in turn by a zone of random mineral orientation and an upper zone of upward converging bladed crystal pseudomorphs (Fig. 4). This sequence corresponds to the B1-A2 zones commonly reported for komatiitic flows (Pyke et al., 1973). A second indication of a probable volcanic derivation is the presence of breccia units up to 1 m thick. The clasts, 5 to 15 cm in diameter are set in a finer grained matrix of similar ultramafic composition (Fig. 5). The angular shapes of the clasts and altered salvages on their margins suggest quenching by, and reaction with, water. The third distinctive feature of the komatiitic rocks is polyhedral jointing (Fig. 6). Two such units have been recognized (Fig. 2) but neither contains spinifex textured material.

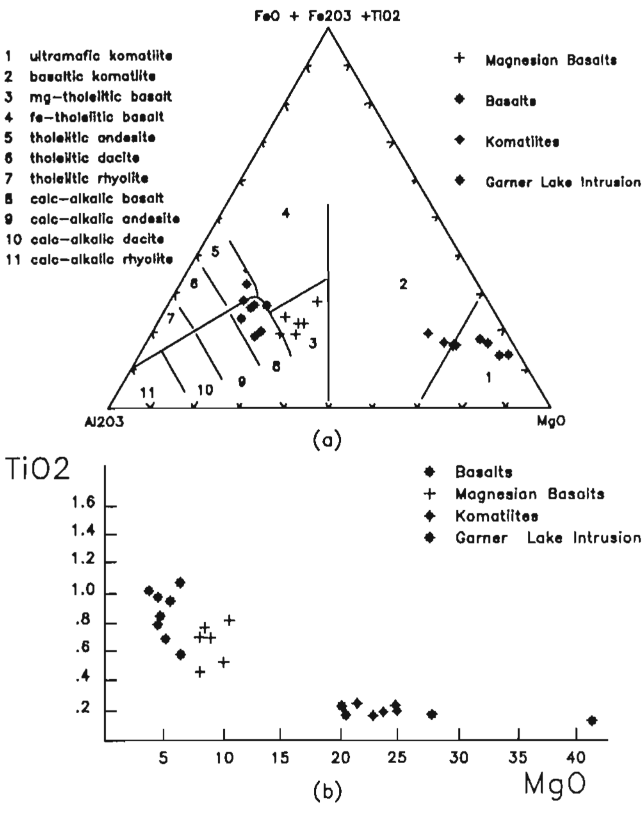

Figure 3. Chemical composition of Garner Lake komatities and related rocks portrayed on **a**) the Jensen diagram and **b**) the TiO_2 vs. MgO plot.

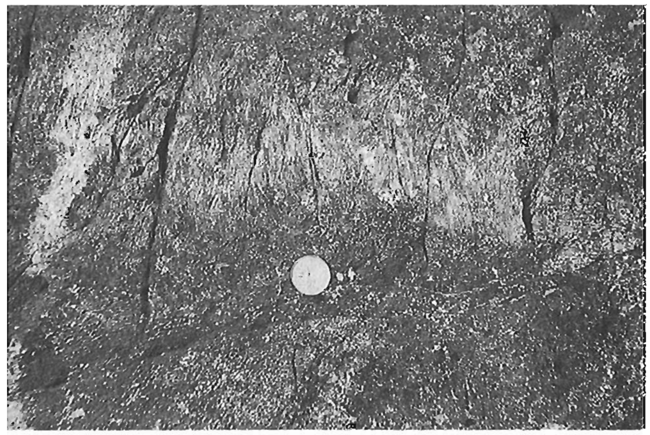

Figure 4. Spinifex texture in komatiite flow. The diameter of the coin is 2.7 cm.

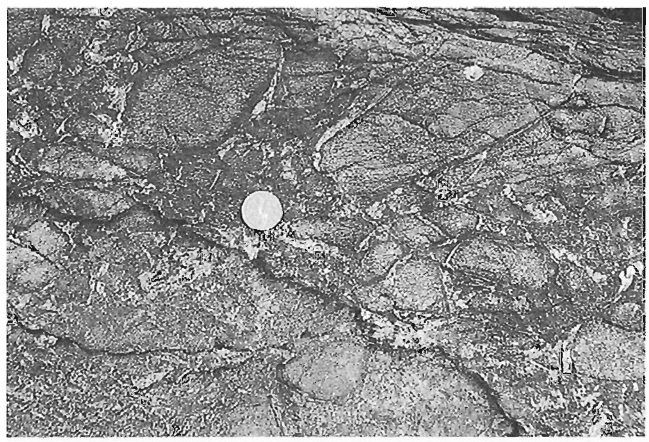

Figure 5. Volcanic breccia and hyaloclastite. The diameter of the coin is 2.7 cm.

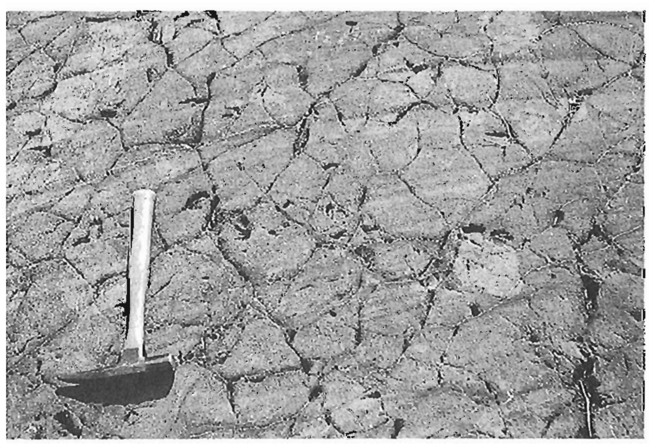

Figure 6. Polysutured komatiitic flow.

DISCUSSION

The komatiites described above are not particularly remarkable by comparison with those of other localities in Superior Province but their presence in the Garner Lake area is of considerable local significance in terms of the interpretation of the geology of the Rice Lake greenstone belt.

Although the full extent and many details of the komatiitic rocks remain to be determined, their presence stratigraphically above the Garner Lake layered intrusion is of some importance. The Garner Lake peridotite-pyroxenite body contains rocks which are compositionally similar to the komatiites described here (Fig. 3). The intrusion was previously interpreted to be a relatively late-tectonic stock (Scoates, 1971) but the strong possibility now exists that it represents the fault-bounded remnant of a subvolcanic magma chamber for the overlying komatiitic flows.

The sequence in which the komatiitic rocks are found is distinctive by comparison with the rest of Rice Lake greenstone belt. It forms part of what Weber (1971a, b) termed the Gem Lake Subgroup of the Rice Lake Group and inferred to be younger than the dominantly intermediate volcanic rocks of the Bidou Lake Subgroup. A comparison of these subgroups with stratigraphic assemblages elsewhere in Uchi subprovince suggests the reverse, however. The rocks of the Gem Lake Subgroup can now be characterized by the dominance of basalt and the presence of komatiitic rocks, layered ultramafic intrusions, and thick volcanic-hosted iron-formation. This lithological assemblage compares favourably to pre-2.8 Ga sequences in the Red Lake district (Balmer and Ball assemblages) to the east whereas the Bidou Lake Subgroup corresponds closely to the 2.73-2.75 Ga Confederation assemblage at Red Lake (Stott and Corfu, 1991).

The correlations of the type suggested above have been previously proposed for the rocks of the Wallace Lake area north of the Wanipigow Fault (Fig. 1a). Theyer (1983) noted the presence of spinifex textured ultramafic rocks stratigraphically below thick iron-formation at Wallace Lake and Weber (1988), in reviewing the quartz arenite-marble-iron-formation sequence of the Conley Formation, suggested a pre-Bidou Lake age for these rocks and their possible correlation with iron-formation in the Beresford Lake area. These suggestions are strongly supported by the observations reported here and indicate the probable presence of two distinctive volcanic components in the Rice Lake belt analogous to the pre- and post-2.8 Ga sequences elsewhere in Uchi subprovince.

The exact relationship between the two types of sequences in the Rice Lake remains to be determined but new mapping in the Garner Lake-Beresford Lake area confirms that they are separated by a profound structural discontinuity (Fig. 1, 2), the Moore Lake-Beresford Lake shear zone. This broad structural zone, the Beresford Lake Deformation zone of Brommecker et al. (1989), is defined to encompass the area between the Moore Lake and Beresford Lake shear zones and likely contains many other narrow discontinuities within it. It extends northward from the study area to a point where it appears to merge with the Wanipigow Fault (Fig. 1a). To the east of this structure are the basalts, komatiites and iron-formation that distinguish the domain containing the Gem Lake Subgroup. The precise location of the boundary of these rocks and the Bidou Lake Subgroup is somewhat cryptic but is taken to coincide with the facing reversal that is observed across the Beresford Lake shear zone (Fig. 2). The former continuity of the Gem Lake Subgroup rocks with similar ones at Wallace Lake (Fig. 1b) is best appreciated once approximately 22 km of dextral offset of units is restored along the Wanipigow Fault. This provides a hypothetical view of the pre-D3 distribution of units in the Rice Lake belt and illustrates how the original discontinuity between the Gem Lake and Bidou Lake domains has likely been incorporated into, and obscured by, the Wanipigow Fault. D2 folds, observed to overprint the D1 structures at a small scale may also be important in displacing them at a larger scale. We speculate further that the northwestern extension of the D1 structural-stratigraphic discontinuity may deviate from the trace of the Wanipigow Fault into the Saxton Lake area (Fig. 1) where the presence of thin ultramafic fault slices is well documented (Scoates, 1971). There remains a pronounced flexure in the trace of the discontinuity after dextral offset along the Wanipigow Fault is accounted for. This could be either a primary indentation of the southern Bidou Lake volcanic domain into the older Gem Lake domain or the result of an earlier stage of sinistral D2 shearing along the trace of the Wanipigow Fault. In either case it appears that the average trace of the discontinuity strikes northwesterly, oblique to the predominant east-west structural trend of the Rice Lake belt.

Of considerable practical interest is the fact that the inferred cryptic discontinuity may be of much greater importance to the distribution of gold deposits (Fig. 1) than the more conspicuous Wanipigow and Manigotogan faults. This, too, is a point of comparison with the Red Lake area where a similar discontinuity between the Balmer and Confederation assemblages passes through the Madsen and Campbell-Dickenson mine areas. At Rice Lake the presence of intense D1 and D2 structures in the Moore Lake-Beresford Lake shear zone is consistent with deposit-scale observations that gold is temporally linked to minor structures of D1 and D2 generation in the eastern part of the belt (Brommecker et al., 1989).

CONCLUSIONS

The presence of a cryptic but major structural-stratigraphic discontinuity within the Rice Lake belt is invoked to account for the juxtaposition of a presumed older sequence containing komatiite and iron-formation and a younger 2.73 Ga intermediate volcanic sequence. The predicted age distinction must however await confirmation by radiometric dating. The inferred discontinuity is likely the regional-scale locus for the gold deposits of the Rice Lake greenstone belt.

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Geological Survey of Canada Project 90019

Observations on the Maberly shear zone, a terrane boundary within the Central Metasedimentary Belt, Grenville Province, Ontario

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Davidson, A. and Ketchum, J.W.F., 1993: Observations on the Maberly shear zone, a terrane boundary within the Central Metasedimentary Belt, Grenville Province, Ontario; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 265-269.

Abstract: The Maberly shear zone, boundary between Sharbot Lake and Frontenac terranes, is the locus of high strain and probably of northwest-directed thrust displacement. This deformation occurred after emplacement of plutonic rocks that show some petrographic affinity with 1170 Ma plutonic rocks in Frontenac terrane, but pre-dated intrusion of a number of small plutons, some of which are similar to ≈1080 Ma syenitic plutons elsewhere in the Central Metasedimentary Belt. Confirmation of this interpretation awaits future petrological and geochronological studies.

Résumé : La zone de cisaillement de Maberly, qui constitue la limite des terranes de Sharbot Lake et de Frontenac, est le foyer d'intenses déformations et probablement de charriages de direction nord-ouest. Cette déformation a eu lieu après la mise en place de roches plutoniques présentant certaines affinités pétrographiques avec des roches plutoniques de 1 170 Ma situées dans le terrane de Frontenac, toutefois plus anciennes que l'intrusion de plusieurs plutons de petite taille, dont certains sont semblables à des plutons syénitiques de ≈1 080 Ma présents ailleurs dans la Zone métasédimentaire centrale. Avant de pouvoir confirmer cette interprétation, il faudra attendre les résultats de futures études pétrologiques et géochronologiques.

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INTRODUCTION

When Wynne-Edwards (1972) defined the Central Metasedimentary Belt in his widely quoted synthesis of the Grenville Province, he also divided it into specific regions. The term "terrane" for these and similar subdivisions was first put forward by Moore (1982; Brock and Moore, 1983), and their distinctive features and distribution were elaborated by Davidson (1986). Rivers et al. (1989, p. 64) advised that this term should not necessarily imply plate tectonic connotations, but should simply describe "... a primary rock grouping based on shared lithotectonic characteristics." However, some recent studies (e.g. van der Pluijm and Carlson, 1989; Cosca et al., 1991; Hanmer and McEachern, in press; Mezger et al., 1991, in press) are providing information that may allow the Central Metasedimentary Belt to be interpreted in terms of accretionary tectonics, followed by extensional relaxation.

One of the terrane boundaries of fundamental interest in this regard separates Frontenac and Sharbot Lake terranes, extending north-northeasterly across the northwest part of the Frontenac Arch (Fig. 1). Although this boundary passes through several areas for which detailed maps are available, only the recent map of Wolff (1985) identifies a major shear zone along it; twenty years earlier, however, Wynne-Edwards (1965) had emphasized a marked contrast in the geology across a zone extending southward from Wolff's shear zone, and had noted occurrences of mylonitic rocks in the same vicinity. Along strike to the northeast, a change in the assemblage of rock units is apparent on the geological maps (Dugas, 1961; Reinhardt, 1973). In defining the southeastern limit of Sharbot Lake terrane, Easton (1988) traced this boundary to the edge of Paleozoic cover north of Carleton Place, and also southwestward as far as Tichborne, south of which its position remained conjectural. This report presents some new field observations on the nature of the

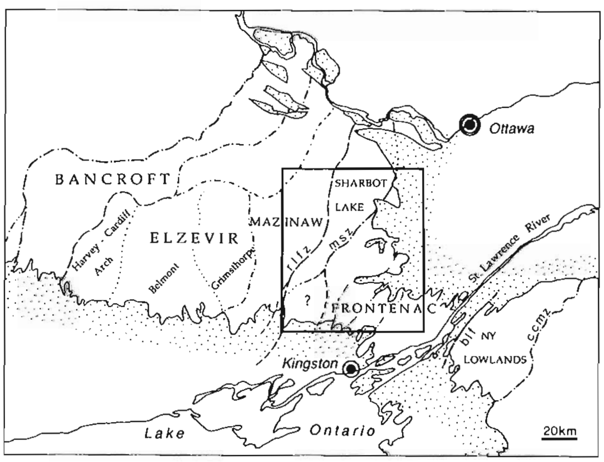

Figure 1. Terranes and domains in the Central Metasedimentary Belt of southeastern Ontario, in part after Easton (in press). rlfz - Roberston Lake fault zone; msz - Maberly shear zone; blf - Black Lake fault; ccmz - Carthage-Colton mylonite zone.

Frontenac/Sharbot Lake terrane boundary, here referred to as the Maberly shear zone on account of its excellent exposure at the village of Maberly (Fig. 2).

Sharbot Lake and Frontenac terranes are distinguished on the basis of differences in supracrustal rock assemblage, grade of metamorphism, and nature and age of plutonic rocks. Marble is common to the supracrustal assemblages in both terranes. In Sharbot Lake it is generally well layered or bedded and is associated with metavolcanic rocks, whereas in Frontenac it is generally a tectonic mélange associated with pelitic and feldspathic granulite and quartzite. Easton (in press) stated that supracrustal rock units cannot be traced across the boundary. Metamorphic grade in Sharbot Lake terrane varies from upper greenschist facies in the northeast (Reinhardt, 1973) to granulite facies in the southwest (Wynne-Edwards, 1965); Frontenac terrane is characterized by granulite facies of relatively low-pressure type (Anovitz and Essene, 1990), witnessed by the fact that aluminous rocks commonly carry cordierite. The metamorphic gradient from upper greenschist to granulite facies across the northeast segment of the terrane boundary would seem to be best explained in terms of tectonic telescoping, although it is not yet certain that the ages of metamorphism in the adjacent terranes are the same.

If the Hinchinbrooke orthopyroxene-bearing tonalite (leucoenderbite) complex is included in Sharbot Lake terrane (see below), the oldest plutonic rocks in this terrane are similar in age (≈1270-1240 Ma) and composition to those in Elzevir terrane to the northwest (Easton, 1986; Lumbers et al., 1990). In Frontenac terrane, apart from one occurrence along the St. Lawrence River of a granitoid dated at ≈1415 Ma (McLelland et al., 1988), most of the dated plutonic rocks are ≈1170 Ma (van Breemen and Davidson, 1988; Marcantonio et al., 1990), and range from pyroxene monzonite to biotite granite. The Rideau Lakes plutons in Frontenac terrane (R in Fig. 2), one of which has been dated at 1077 Ma (Marcantonio et al., 1990), belong to a younger suite of monzodioritic, syenitic and granitic plutons that also occur in Sharbot Lake, Mazinaw, and Elzevir terranes (Lumbers et al., 1990), and which are age-correlative with syenitic plutons in the Central Metasedimentary Belt of Quebec (Corriveau, 1990).

NEW OBSERVATIONS

Fundamental to understanding the role of shear zones along terrane boundaries is an evaluation of their displacement histories. In his summary of findings on the Maberly shear zone, Easton (1988) did not mention identification of kinematic indicators that might imply a sense of displacement. Although the moderately to steeply southeast- dipping foliation in the mylonitic rocks of this zone commonly carries an oblique mineral stretching lineation (raking to the east), no unequivocal sense of displacement was observed during the present study, either in outcrop or in oriented thin sections from this zone. The metasedimentary gneisses within the northwesternmost part of Frontenac terrane, however, are mylonitic in places, as was noted by Wynne-Edwards (1965)

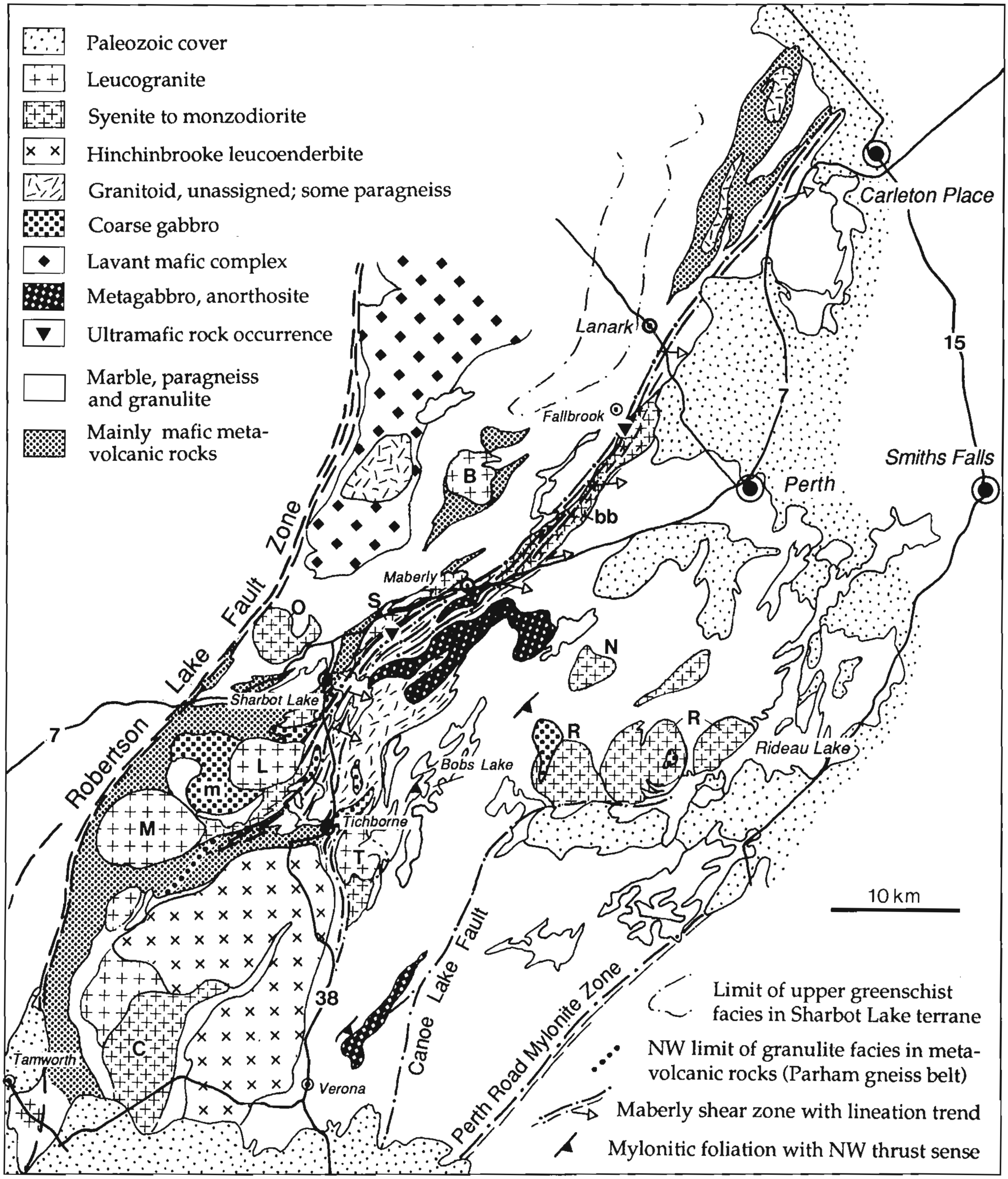

Figure 2. The Maberly shear zone and the distribution of plutonic rocks in adjacent parts of Frontenac and Sharbot Lake terranes. See text for names of plutons designated by letters.

among the islands in Bobs Lake. Northwest-directed thrust displacement has been confirmed at two such localities, and may be related to development of the Maberly shear zone.

Whether the Maberly shear zone beyond Tichborne extends to the southwest or south is not yet resolved. Mylonitic foliation strikes southwestward at Eagle Lake; it affects mafic granulites and syenitic plutonic rocks for some 10 km west of Tichborne (Fig. 2), but farther west it appears to dissipate. Equating this structure with the terrane boundary requires that the enderbitic plutonic rocks of the Hinchinbrooke complex and metavolcanic rocks which lie north and west of it are in Frontenac terrane. An alternative extension, southward toward Verona, east of the Hinchinbrooke complex, would allow the lithological contrast across the boundary to be maintained (metavolcanic rocks/marble in Sharbot Lake terrane *vs.* pelite/quartzite/marble in Frontenac), but would require an explanation for the marked increase in metamorphic grade southward within Sharbot Lake terrane; a mylonite zone was not identified along this course, although it is possible that a high strain zone may lie unrecognized within coarsely recrystallized marble tectonite. Mylonites with northwest thrust sense are well developed within and along the base of the southeast-dipping Richardson anorthositic gabbro sheet, a few kilometres northeast of the projected boundary near Verona (Fig. 2).

Whereas some plutonic rock masses are clearly involved in the deformation that produced mylonitic fabrics along the Maberly shear zone, others equally clearly cut these fabrics, a feature recognized by Easton (in press) and confirmed in the present study. Plutonic rocks of several different types are present, but with the exception of the Hinchinbrooke complex (1254 Ma; Wallach, 1974), no isotopic dating has yet been done that would allow assignment of individual plutons to particular suites defined by age. Two plutons composed predominantly of massive biotite leucogranite (McLean and Barbers Lake, M and B in Fig. 2) lie within Sharbot Lake terrane, and two others (Leggat Lake and Tichborne, L and T in Fig. 2) cut rocks with mylonitic fabric related to the Maberly shear zone. Although these plutons were assigned to the ≈1250 Ma "alaskite suite" by Lumbers et al. (1990), they appear to be post-tectonic and thus probably do not belong to this suite. Both the McLean and Leggat Lake plutons also contain tracts of massive hornblende-biotite monzonite grading to leucodiorite with relict augite, rock types that make up much of the nearby undeformed Oso pluton (O in Fig. 2) in Sharbot Lake terrane; similar rocks occur as well within the variably deformed Silver Lake pluton (S), adjacent to the Maberly shear zone, and in the foliated, locally mylonitic Bennett Bay (bb) and North Crosby (N) plutons in Frontenac terrane. The Oso pluton also includes syenitic and leucogranitic phases, and a cogenetic phase of augite-biotite diorite rich in apatite; this association makes the Oso pluton strikingly similar to some plutons of the ≈1080 Ma potassic suite of Corriveau (1990), in particular the Skootamatta pluton in Elzevir terrane, Ontario, and the Kensington pluton in Quebec.

Gabbroic and anorthositic rocks are also present in various states of repair within and adjacent to the Maberly shear zone, and probably represent at least three separate suites. The semi-circular Mountain Grove pluton (m) between the McLean and Leggat Lake granites is composed of coarse, massive, biotite-bearing noritic leucogabbro; small masses of gabbro of identical appearance occur south and east of Sharbot Lake village, and lie within the Maberly shear zone, which they may post-date. The southern part of the Lavant mafic complex in central Sharbot Lake terrane appears to be associated with metavolcanic rocks and is recrystallized, mainly to hornblende gabbro; replacement of primary olivine by coroniform fibrous amphibole and chlorite suggests greenschist-facies metamorphism. A number of lenses of basic rock, ranging from ultramafic to anorthositic, lie within Frontenac terrane adjacent to and in the Maberly shear zone. Where not sheared, these rocks are variably recrystallized; anorthositic rocks south of Maberly, for example, are composed of granular aggregates of basic plagioclase with minor metamorphic orthopyroxene and porphyroblasts of garnet and corundum, and ultramafic rocks near Fallbrook and southwest of Maberly (Fig. 2) contain hercynite and orthopyroxene within aggregates of fibrous amphibole. East and south of Sharbot Lake, anorthositic rocks are spatially associated with ortho- and clinopyroxene-bearing quartz diorite and quartz monzodiorite (unassigned granitoid unit, Fig. 2). Equivalent rocks within the mylonitic zone near Sharbot Lake are retrograded to hornblende- and biotite-bearing assemblages, as are gabbroic rocks within the shear zone at Maberly and to the northeast. Although Wynne-Edwards (1965) grouped these rocks with the Hinchinbrooke complex, they are sufficiently more varied in composition and state of deformation to warrant separate consideration

The plutonic rocks which lie west of the Hinchinbrooke tonalite are equally varied. Leocogranitic and syenitic rocks south of the McLean pluton give way southward to variably foliated quartz monzodiorite and minor hornblende diorite (Chippewa pluton, C in Fig. 2). Anorthosite reported by Harrison (1944) near the northern contact of this pluton is quite unlike the anorthositic rocks near Maberly and Verona; it appears to be a crowded aggregate of rounded, coarse anorthosite blocks and large plagioclase xenocrysts in a sparse granitoid matrix; isolated, dark plagioclase xenocrysts are present in the adjacent granitoid rocks. Farther southwest, near the village of Tamworth, isolated outcrops in poorly exposed terrain are of massive augite monzodiorite similar to some rocks of the Oso pluton.

CONCLUSIONS

Although unequivocal kinematic indicators were not identified in the Maberly shear zone itself, the abrupt juxtaposition of granulite facies in Frontenac terrane structurally above lower grade rocks in northeastern Sharbot Lake terrane suggests that this zone resulted from compressive tectonism. This interpretation is supported by the presence of thrust-sense mylonites in adjacent, smaller scale shear zones in Frontenac terrane, and by lack of correlation of supracrustal assemblages across the boundary. The nature and extension of the shear zone south or southwest of Tichborne remains to be documented.

The plutonic rocks in the vicinity of the Maberly shear zone cannot yet be assigned unequivocally to the established suites defined by petrological affinity or age. Some of the deformed plutonic rocks within the zone, however, are petrologically similar to the ≈1170 Ma suite characteristic of Frontenac terrane farther southeast. At least some of the undeformed plutonic rocks which cut mylonitic fabric associated with the boundary may belong the ≈1080 Ma syenite suite whose plutons occur in various terranes of the Central Metasedimentary Belt. Geochemical and geochronological studies are planned to test these contentions. If Frontenac terrane is exotic with respect to Sharbot Lake and other terranes to the northwest, the Maberly shear zone may conceal a suture; if so, dating the deformation along this zone may also date the docking of Frontenac terrane, unless there has been subsequent reactivation, not yet recognized. If the age assignments suggested above prove correct, amalgamation occurred at some time between the initiation (≈1190 Ma) and cessation (≈1060 Ma) of ductile thrust tectonism along the boundary between the Central Metasedimentary Belt and the gneiss terranes to the northwest. If on the other hand some of the younger, post- shear plutons belong to the ≈1170 Ma Frontenac suite, then amalgamation may have occurred before the earliest deformation so far recognized along the Central Metasedimentary Belt boundary tectonic zone.

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Grenville Front studies in the Sudbury region, Ontario

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Abstract: Several controversial aspects of the geology bearing on understanding the nature of the Grenville Front south and east of Sudbury are addressed. West of Ontario highway 69, ductile fault splays diverge from the front and penetrate the Chief Lake batholithic complex. Recognition of specific, albeit attenuated, Huronian stratigraphy within different fault panels in the complex may answer why direct correlation across the front is so difficult. Previously unmapped diabase dykes of the ≈1235-Ma Sudbury swarm are offset at faults and cut mylonitic fabrics between faults on both sides of the front, providing an age constraint on early "Grenvillian" deformation. To the northeast, the Wanapitei mafic complex is interpreted as a map-scale tectonic inclusion within ductile gneisses southeast of the front.

Résumé : On examine plusieurs détails géologiques controversés concernant la nature du front de Grenville au sud et à l'est de Sudbury. À l'ouest de l'autoroute 69 en Ontario, des réseaux de failles divergentes rayonnent à partir du front et pénètrent dans le complexe batholithique de Chief Lake. L'identification d'une stratigraphie huronienne spécifique mais amincie dans différents panneaux de failles du complexe, peut expliquer la raison pour laquelle il est si difficile d'établir une corrélation directe entre les deux côtés du front. Des dykes de diabase jusque-là non cartographiés, anciens de ≈1 235 Ma, et appartenant à l'essaim de dykes de Sudbury, sont décalés à l'emplacement des failles et recoupent les fabriques mylonitiques entre les failles des deux côtés du front, ce qui permet de situer avec plus de précision dans le temps les premiers épisodes de déformation «grenvillienne». Au nord-est, selon les interprétations, le complexe mafique de Wanapitei serait une inclusion tectonique visible à l'échelle de la carte reposant au sein des gneiss ductiles au sud-est du front.

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INTRODUCTION

In the Sudbury region, the boundary between the Southern and Grenville provinces for some distance east of Coniston is a brittle, normal fault, the Wanapitei fault (Fig. 1). It coincides with the eastward extension of the Murray fault and was reactivated in post-Grenvillian time (Davidson, 1992). Immediately south of Coniston, this fault truncates intensely mylonitized Southern Province rocks (middle Huronian formations, Nipissing gabbro and younger granitoids) in the footwall of a ductile fault. The hangingwall includes mylonitic schist, gneiss and migmatite at much higher metamorphic grade than the footwall (La Tour, 1981; Davidson et al., 1990). This distinctive Grenville Front mylonite zone is easily followed to Ontario highway 69 and for a few kilometres beyond, but farther southwest it is less readily identified (Fig. 2). The reason for this is two-fold: 1) the Huronian rocks and Nipissing gabbro of the Southern Province footwall are voluminously intruded by grantoid rocks of the Chief Lake complex, and included remnants of them close to the front are so metamorphosed and deformed as to be difficult to distinguish from hangingwall gneisses; 2) the granitoid rocks themselves are deformed penetratively, though inhomogeneously, to protomylonite and mylonite associated with west-southwest-trending fault splays diverging from the main Grenville Front mylonite zone. A similar situation pertains in the Tyson Lake area, some 40 km to the southwest, although there the fault splays diverge in the other direction, namely northerly (Bethune and Davidson, 1988); farther southwest, toward the coast of Georgian Bay, the Grenville Front again becomes a single, sharply defined mylonite zone (Davidson, 1986a).

In addition to the main Grenville Front mylonite zone southwest of Coniston, other well-defined mylonite zones occur within the hangingwall (La Tour and Fullagar, 1986), outlining panels of less mylonitized although highly folded and disrupted gneisses. The isograd marking the first appearance of sillimanite as mapped by Kwak (1968) is

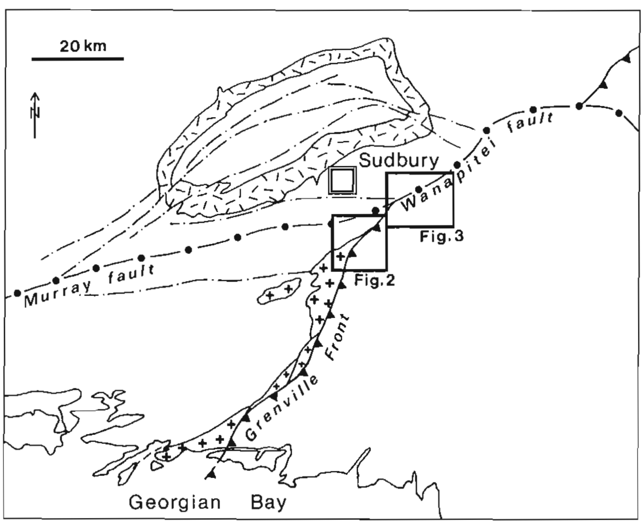

Figure 1. Location map.

parallel to this regional structure, which converges with and is truncated by the Wanapitei fault in the area west of the Wanapitei mafic igneous complex (Fig. 3). East of this complex, metasedimentary gneisses immediately south of the front include kyanite-rich units and are disposed about east-trending, southeast-plunging folds.

Two unrelated aspects of Grenville Front geology were examined in detail in 1992: the potential for correlation of distinctive Huronian strata and other rock units across the front, and the tectonic setting of the Wanapitei mafic complex. These are treated in separate sections below.

PART 1: CORRELATION ACROSS THE GRENVILLE FRONT

One of the unresolved controversies involving the Grenville Front in this region is the question whether or not Southern Province rocks are represented in deformed and metamorphosed state in the adjacent Grenville orogen. Inability to trace Huronian units directly across the front, or to recognize them with confidence southeast of it, prompted early workers to explain the 'disappearance of the Huronian' by processes of granitization (Quirke and Collins, 1930; Quirke, 1940; Phemister, 1960). More recently, Lumbers (1975) suggested that the metasedimentary gneisses southeast of the front represent less mature, deeper water facies of the lower part of the Huron Supergroup, supporting this contention with the interpretation that greywacke, siltstone and minor conglomerate southeast of Sudbury are facies of the Mississagi Formation; he also stated, however, that "Detailed stratigraphic correlations between metasediments of the Grenville and Southern Provinces ... are unjustified, ..." (op. cit., p. 49).

It is the lack of 1), sufficiently thick units of metasedimentary rock types that might represent individual, distinctive Huronian formations, and 2), successions of metasedimentary rock units comparable to any part of the well known Huronian succession, that make it difficult to accept correlation across the front. Close to the base of the hangingwall between Alice and Brodill lakes, and also farther east immediately south of the Wanapitei fault, a variety of metasedimentary gneisses is present, generally interlayered in units a few tens of metres thick or less; rock types include pelitic and semipelitic gneiss (garnet-biotite gneiss with and without kyanite, muscovite and rare staurolite), thinly layered biotite-hornblende gneiss and amphibolite with or without garnet, variable biotite-feldspar-quartz gneiss, feldspathic and pure quartzite, and rare calc-silicate marble. Relatively thick lenses and layers (several tens of metres) of massive amphibolite and rare, smaller pods of metamorphosed ultramafic rock are associated with the paragneisses. Also present are tracts of relatively uniform migmatitic biotite quartzofeldspathic gneiss with parallel leucosomes. In most places within a few hundred metres of the front, outcrop surfaces in relatively less strained zones show mesoscopic structures indicative of a high degree of ductility, such as disaggregated or drawn out pegmatite seams with rotated feldspar porphyroclasts, isolated boudins of mafic rock, and sheath folds and refolded folds of attenuated layers. Mylonite and ultramylonite are developed in the intervening high strain zones, forming moderately southeast-dipping planar units that truncate older folds in the lower strain panels. A variety of both meso- and microscopic shear-sense indicators implies northwest thrust displacement across the internal mylonite zones, the same sense that is found at the front itself. In general, strain decreases gradually up structural section into the overlying panels. The zone of mylonite-bounded panels in the hangingwall can thus be interpreted as a stack of thin, interleaved lenticular thrust sheets, and it is little wonder that, if the metasedimentary gneisses and massive amphibolite lenses represent Huronian sediments and Nipissing gabbro, there is now no semblance of stratigraphic order or maintenance of thick, uniform units.

A mid-way stage in the development of such an attenuated stack may be illustrated by the arrangement of Huronian strata within the Chief Lake batholith west of highway 69 (Fig. 2). It became apparent during reconnaissance in this area that there are several shortcomings in the information available on existing geological maps. The early mapping of

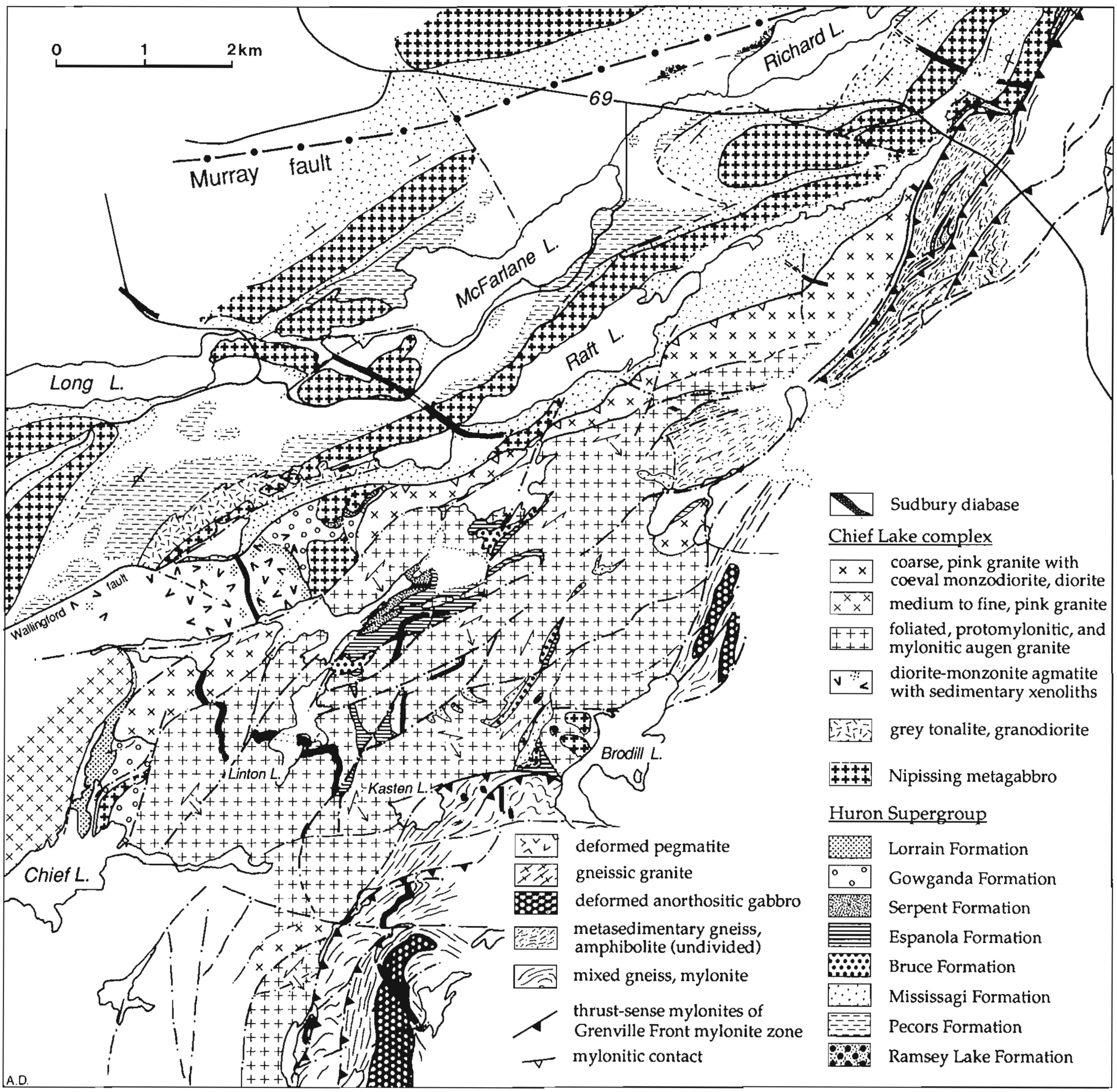

Figure 2. Geology of the northeastern Chief Lake complex and the adjacent Grenville Front, 12 km south of Sudbury.

Phemister (Thomson, 1962) assigns lithologic rather than formational status to the Huronian sedimentary rocks, and portrays the Chief Lake granitoid rocks as feldspathized equivalents of them; in addition it does not distinguish between Nipissing gabbro and other types of mafic igneous rock. Detailed remapping by Henderson (1967) corrected this situation to some extent, but in a later compilation for the smaller scale map of the Burwash area (Lumbers, 1975), all the metasedimentary rafts in the Chief Lake batholith were assigned to the Mississagi Formation, as well as the metawacke, slate, knotted schist and conglomerate that occupy a large tract north of the batholith, and he interpreted all the mafic masses in the marginal part of the Chief Lake batholith to be inclusions of Nipissing gabbro. A combination of these interpretations was incorporated into the most recently published map of the Sudbury region (Dressler, 1984). The results of mapping in 1992 and the required reinterpretations are summarized in Figure 2.

Probable Pecors Formation in the McFarlane-Long Lake area

Re-examination of the area around Richard Lake (Davidson, 1992) resulted in the suggestion that Collins' (1936) original identification of Ramsey Lake conglomerate in the Richard Lake area south of the Murray fault may be correct. In Collins' day, the Pecors Formation of greywacke, siltstone and shale, intervening between the Ramsey Lake and Mississagi formations, had not been described and therefore was not part of the established stratigraphic succession; thus the significance of argillite and slate stratigraphically beneath Mississagi sandstone on the south side of Richard Lake was not appreciated. To the southwest around McFarlane Lake, large areas are underlain by knotted schist (andalusite ± staurolite, commonly altered to mica and chlorite) derived from thin-bedded shale and siltstone. Other outcrops in this area are composed of fine grained, greenish to creamy grey metasandstone interbedded with argillaceous and, more rarely, gritty beds; unlike typical crossbedded Mississagi Formation, the sandstone beds have features of turbidites. These rocks are disposed in the core of an anticlinorium defined by outward-facing Mississagi sandstone on the flanks of the ridges that border this topographically subdued area. Knotted schist and interbedded sandstone are exposed along strike for several kilometres to the southwest, but are truncated by a major fault (Wallingford fault of Card, 1978) south of Long Lake (Fig. 2). Except for their higher metamorphic grade, the sub-Mississagi rocks are similar in every way to the Pecors Formation exposed south of Mikkola, about 12 km due west (Card, 1978).

Huronian formations and Nipissing Gabbro within the northeastern Chief Lake complex

Although usually referred to as the Chief Lake batholith, the term complex is preferred because the 'batholith' is composed of several different plutonic rock types forming small units and enclosing rafts and septa of Huronian metasedimentary rocks and Nipissing metagabbro. The plutonic rocks, predominantly granite, are variably deformed, and the complex is cut by numerous faults. Krogh and Davis (1970) obtained a preliminary U-Pb zircon age of ≈1.7 Ga for Chief Lake granite.

A major topographic lineament extends southwest from Highway 69 through Raft Lake and joins the Wallingford fault south of Long Lake (Fig. 2). It is flanked to the north by a continuous ridge underlain by a sill-like intrusion of Nipissing gabbro, which appears to cut gradually down section to the southwest and is intruded, along with knotted schists, by an irregular mass of light grey, medium grained biotite granodiorite. Southward-dipping Mississagi Formation exposed along the south side of the valley is tectonically thinned, being represented in many places by flaggy, quartz-rich mylonite. In the Raft Lake area, this mylonite is structurally overlain by pink, coarse grained biotite granite of the Chief Lake plutonic complex, mylonitic at the contact, along what appears to be a thrust fault. Farther southwest, however, the granite diverges from the mylonite and is in intrusive contact with metasedimentary rocks identifiable as feldspathic quartzite of the Serpent Formation and muscovite-biotite schist and hornfels of the Gowganda Formation (including phases in which thin bedding or matrix-supported granite pebbles are preserved), both intruded by Nipissing metagabbro. These rocks are invaded by a complex of irregular, interconnected dykes ranging from hornblende gabbro to quartz diorite. The intrusive component becomes more voluminous to the west and the Gowganda component is reduced to the status of rafts and inclusions. Inclusions yet farther west change successively to pink meta-arkose, green and yellow muscovitic quartzite, and white orthoquartzite, representing the succession typical of the Lorrain Formation. The stratigraphic succession thus faces southwest and does not seem to have been significantly disrupted, despite the amount of intrusive material. Bedding preserved in the rafts strikes northwest and dips steeply northeast; it is cleanly truncated by the contact with mylonitic Mississagi sandstone, which is not penetrated by the intrusive dyke complex. Lorrain and Mississagi quartzites are juxtaposed at this moderately south-dipping contact, suggesting a net, south-side-down displacement of at least 3 km, with complete change in orientation of structure across the fault.

A westerly-facing succession of Gowganda and Lorrain formations (Cobalt Group), also intruded by Nipissing metagabbro, occurs both north and south of Chief Lake, some 3 km to the southwest, where it is intruded by or faulted against three different types of granitoid rock (Fig. 2). At the western contact of the Chief Lake complex, some 8 km farther southwest, Card et al. (1975) mapped a complete succession ranging from Mississagi to Lorrain formation that faces northeastward. Taken together, all these occurrences of upper Huronian formations define the remnants of a structural basin engulfed in the Chief Lake complex and subsequently sliced by faults. If the north bounding fault against mylonitic Mississagi Formation is now a thrust, the presence of stratigraphically higher formations in the hangingwall requires an earlier foundering, perhaps at the time of emplacement of the Chief Lake intrusions.

The fault-bounded strip in which the Cobalt Group remnants occur, along with lenses of coarse grained granite, is in contact to the southeast with medium to fine grained granite which ranges from mildly foliated to mylonitic. In the area between Linton Lake and an unnamed lake just southwest of Raft Lake (Fig. 2), numerous remnants of metasedimentary rocks include a large proportion of outcrops of metamorphosed, thinly bedded calcareous siltstone whose Huronian counterpart can only be the Espanola Formation. This identification is consistent with the presence of massive calc-silicate rock at its contacts with feldspathic quartzite representing Serpent Formation, and calcite-bearing hornblende hornfels at its contacts with biotite-muscovite metawacke containing quartz granules and scattered, grey granitoid pebbles, representing Bruce Formation. To the southeast again, particularly northwest of Brodill Lake, metasedimentary relicts include Espanola, Bruce and Mississagi formations, the latter represented by meta-sandstone with rusty parting surfaces and very rarely preserved crossbedding. Small masses of Nipissing metagabbro are also present. All of these occurrences are engulfed in fine grained, variably protomylonitic pink, violet and grey granitoid rocks whose foliation trends northeasterly and carries a pronounced stretching lineation that dips moderately southeast. Where augen structure is present, rotated feldspar porphyroclasts, C-and-S fabric and shearband foliation are commonly developed, and all indicate penetrative thrust-sense displacement.

Phemister (1961) identified 'feldspathized' gabbro masses within the northeastern part of the Chief Lake complex, which Lumbers (1975) equated with Nipissing gabbro. Detailed re-examination of these occurrences has resulted in a different interpretation. The mafic rock does indeed have the form of large and small included masses within coarse grained granite, and is commonly intimately veined by thin, discontinuous seams of granite-like material characterized by feldspar megacrysts. It is composed mainly of hornblende and plagioclase with some small K-feldspar and bluish quartz megacrysts, and it does not look at all like the Nipissing metagabbro found with the Huronian metasedimentary rocks. In addition it also occurs as sharply bounded dykes within the granite. These dykes are compound in that they have self-contained, anastomosing granite veinlets which enclose rounded mafic masses. Similar compound dykes intrude the Gowganda and Lorrain formations and Nipissing metagabbro southwest of Raft Lake. Such rocks may be best interpreted as having crystallized from two coexisting magmatic phases; the mafic component is part of the Chief Lake complex, not Nipissing metagabbro.

Southwest extension of the Grenville Front mylonite zone

The intensely mylonitic rocks that overlie a thin sliver of highly deformed Chief Lake granite at Highway 69 can be traced more or less continuously for 3 km southwest along the southeast side of a swampy valley. Derived mainly from metasedimentary rocks, they are separated from well-preserved Chief Lake plutonic rocks by a thin unit of mafic mylonite laced with brittle fractures and quartz seams. Southwest of Brodill Lake, mylonitic rocks lying south of recognizable Huronian remnants, and separated from them by shallow south-dipping thrust faults, appear to have been derived from a migmatitic granitoid parent. The map pattern in this area suggests that the continuity of the main mylonite zone has been dislocated by late faults, but south of Kasten Lake, a prominent south-trending, thrust-sense mylonite zone separates east-dipping, variably foliated quartzofeldspathic rocks from an overlying unit of mixed felsic and mafic, migmatitic gneisses which has no obvious counterpart to the northwest. A broader mylonite zone just to the east carries a distinctive unit of intensely deformed and metamorphosed gabbro in its hangingwall. This unit is characterized by an anorthositic phase containing crowded plagioclase megacrysts from 2 to 5 cm in diameter. It is cut by small amphibolite dykes and laced by granitoid veinlets that have become the loci of mylonitization during deformation; in many places the rock is an aggregate of rounded leucogabbro blocks in an anastomosing network of granitoid or pegmatite mylonite carrying large, rotated plagioclase porphyroclasts. The underlying shear zone is composed of fine grained, straight-layered, uniform to augen mylonite with consistent thrust-sense indicators.

The Grenville Front between Brodill Lake and the mylonite zone traced from highway 69 is difficult to evaluate on the basis of identifying a Huronian/Chief Lake footwall and a mylonitic gneiss hangingwall, and needs further investigation. Remnants of Huronian strata define variably oriented, complex folds within straight-foliated mylonitic granitoid rocks, but give way southeastward to dark, folded metasedimentary and amphibolitic mylonites, cut by deformed pegmatite dykes oriented parallel to the main trend of mylonitic foliation in the fold limbs. However, the main metamorphic break, heralded by the sudden appearance of coarse garnet, occurs a little farther southeast (Henderson, 1967; Hsu, 1968).

Diabase dykes of the 1235-Ma Sudbury swarm

Apart from a dyke exposed on the road north of Long Lake, no occurrences of Sudbury diabase had been identified previously within the mapped area of Figure 2. Four other small diabase dyke segments mapped by Hsu (1968) and Lumbers (1975) were not assigned to the Sudbury swarm. Reasons for this are that Sudbury dykes south of the Raft Lake – Wallingford fault lineament do not follow predictable courses, are not everywhere well exposed, and are composed in part of chloritized or amphibolized diabase, whereas to the north the dykes follow regular southeasterly courses, can be traced by their aeromagnetic expressions where not exposed, and are of fresh diabase which has a characteristic rusty, spheroidal weathering.

The distribution of Sudbury diabase dykes shown in Figure 2 illustrates the irregular nature of those dykes found so far within the Chief Lake complex. They show pronounced apparent offsets along prominent faults, but it may be that some of these are partly or wholly the result of natural dyke termination controlled by pre-existing structure.

In the area immediately east and south of Kasten Lake, dyke segments lie within panels between intense mylonite zones. Exposure is complete enough in some places to be certain that the dykes do not cut through these mylonite zones, yet deformed diabase was not found within them, nor does the closest diabase show any evidence of increased strain. At the locality east of Kasten Lake a north-trending dyke terminates with chilled contact within southeast-dipping protomylonitic gneiss that grades down section into flaggy mylonite. An important observation throughout this region is that the dykes cut across pre-existing mylonitic foliation within the fault-bounded panels. Their irregular courses appear to be primarily a function of inability of magma to propagate straight fractures across rocks that record a marked variation in strain, not because they have been subsequently folded. Adjacent country rock foliation does not display deflection commensurate with bends in the dykes, and the diabase itself, even where hydrated, is not penetratively foliated. However, small, widely spaced shears within the dykes and, south of Kasten Lake, distortion of primary plagioclase laths attest to a certain amount of post-intrusion adjustment; it is certainly possible that original irregularities could have been enhanced by subsequent compression (c.f. Bethune and Davidson, 1988; Bethune, 1989).

Summary

There is no doubt that Sudbury diabase dykes were intruded across the Grenville Front mylonite zone, and therefore that this regional structure was in place by 1235 Ma. The only constraint of pre-diabase deformation in this area is the 1.7 Ga age of the Chief Lake granite, although similar relationships farther southwest are constrained more firmly to 1.45 Ga by younger granite and pegmatite (van Breemen and Davidson, 1988). Fault-bounded panels containing deformed granitoid rocks and distinctive parts of the Huronian stratigraphic succession northwest of the front may have developed at an early stage. These may be represented in more highly attenuated form by quartzofeldspathic and metasedimentary gneisses and schists in the hangingwall, within narrower, mylonite-bounded panels. Finding post-Huronian detrital zircons in quartzitic units among these rocks would disprove the Huronian connection. The hangingwall also contains rock units, such as the deformed anorthositic gabbro south of Kasten Lake, that have no obvious counterparts in the adjacent footwall, although these could conceivably be reworked equivalents of the earliest Proterozoic mafic igneous rocks at the base of the Huron Supergroup.

PART 2: TECTONIC SETTING OF THE WANAPITEI COMPLEX

In the Grenville Province east of the junction between the Grenville Front mylonite zone and the Wanapitei fault at Alice Lake (Fig. 3), an assemblage of migmatitic quartzofeldspathic gneisses of probable plutonic origin, metasedimentary gneisses including quartzite and pelite, amphibolite, and gneissic anorthositic gabbro encloses an oval mass of predominantly mafic igneous rock known as the Wanapitei complex (Grant et al., 1962; Lumbers, 1975; Rousell and Trevisiol, 1988). Lumbers (op. cit., p. 101) considered this complex to have been "... emplaced relatively late in the tectonic and metamorphic history of the Grenville Province." However, a recent U-Pb zircon age of 1745 Ma for gabbroic rock from the complex suggests otherwise (Prevec, in press). Re-examination of the complex and its surrounding rocks was undertaken in an attempt to clarify this question.

The Wanapitei complex underlies an elliptical area 6 km long by 2 km wide, oriented northeast. The southwestern two-thirds are very well exposed, except for a central, sand-covered area. A large swamp obscures the northeastern one-third, except for a small area adjacent to the northeast contact. Most of the southwestern part is composed of an intimate mixture of metamorphosed gabbroic rock and younger granodiorite and granite of more than one generation. Field relationships suggest that the gabbro was extensively fractured and heated during intrusion of the earlier granitoid veins. Both phases deformed together, resulting in an agmatite composed of metagabbro blocks ranging from 10 cm to 100 m size within a smeared-out granitoid matrix. This complex rock is cut by northeast- to east-trending, vertical, porphyritic granite dykes, some of which show mingling with mafic magma in much the same way as the compound dykes within parts of the Chief Lake complex. In fact, the later granite phase looks very similar to some Chief Lake granite.

The agmatite is in sharp contact with massive hornblende noritic gabbro that occupies a north-south strip across the central part of the complex. The contact is the locus of mylonitization in the agmatite unit. Locally the noritic gabbro has primary igneous layering that dips and faces moderately to the northwest. It grades down section into mafic olivine norite characterized by well developed coronitic texture in which olivine is partly or wholly replaced by orthopyroxene surrounded by pargasite-spinel symplectite (c.f. Rousell and Trevisiol, 1988). East-northeast-trending, compound granite dykes cross the agmatites' eastern contact and intrude both hornblende and olivine norite. All of these rocks are cut by narrow, east-trending dykes of amphibolite. A varied mixture of different types of metagabbro occurs along the southeast contact southwest of the main swamp, and occurs again in the northeast contact area, where xenoliths of metasedimentary origin are common.

Not previously reported, north-south, vertical metadiabase dykes, up to 15 m wide, cut the agmatitic unit, granite and amphibolite dykes in the southwest part of the complex. The metadiabase is fine grained and has an equant, mottled texture given by small heterads of primary clino- pyroxene; its groundmass contains very fine metamorphic garnet between plagioclase laths. Black plagioclase xenocrysts occur near the dyke margins. The dykes are very similar in appearance to Sudbury metadiabase dykes elsewhere in the Grenville Front tectonic zone. In a few places they are cut by irregular, nondeformed pegmatite masses, adjacent to which they are converted to amphibolite.

The outer contact of the Wanapitei complex is not cut by either the granite or metadiabase dykes, and is clearly tectonic. The contact along the west and northwest sides displays an abrupt transition from agmatite with internally irregular structure to a narrow carapace in which the agmatite is drawn out parallel to the contact, is recrystallized to coarser grain size, and develops streaky leucosome veins parallel to foliation. All resemblance to the parent rock is lost within a few metres of the first development of foliation that swings dextrally into parallelism with the contact. At one place along the west contact, remobilized gneiss penetrates the agmatite in a hook-shaped structure, and several smaller penetrations occur along the northwest contact. The migmatitic carapace gneiss grades abruptly into metasedimentary gneiss, including quartzite and coarse garnet-bearing gneiss with amphibolite boudins. The carapace rocks are strongly foliated, but are not mylonite *per se*, and do not exhibit lineations or shear-sense indicators.

The southeast contact is less well exposed. Part of it appears to be structurally overlain by metasedimentary gneiss in the form of a synform outlined by a narrow quartzite unit. Farther northeast, a narrow sliver of similar gneiss in part separates the complex from an elongate lens of anorthositic gneiss which is folded about east-plunging folds near the eastern contact. Although not continuous in outcrop, this anorthositic unit may correlate with the Red Deer anorthosite mass farther east (Lumbers, 1975), itself a tectonic enclave probably related to the 2.4 Ga River Valley anorthosite (Davidson, 1986b; Ashwal and Wooden, 1989). Fine grained, coronitic metadiabase similar to that of the north-south dykes within the complex occurs as tectonically disrupted dyke segments within these rocks, again intruded by and amphibolized next to late pegmatite.

In summary, the Wanapitei complex appears to be a large, resistant tectonic enclave within gneisses that have undergone severe ductile flow. The distribution and

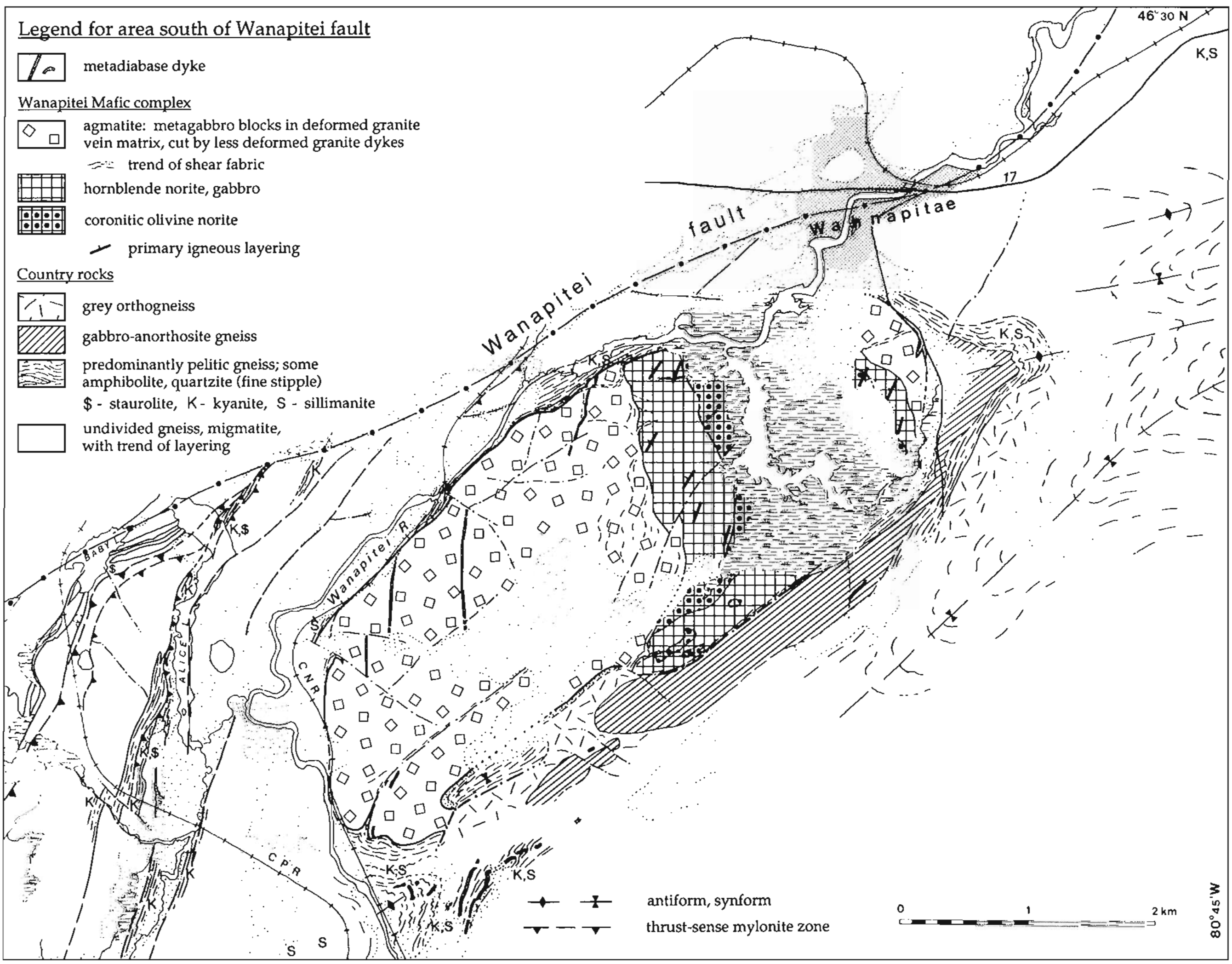

Figure 3. Geology of the Wanapitei mafic complex, 15 km east of Sudbury. The Wanapitei fault is the northwest margin of the Grenville Province in this region.

disposition of the least disturbed gabbroic rocks in the central part of the complex suggest that it was once part of a larger mass. Whether or not it has been displaced a significant distance from its origin cannot be ascertained. However, it does not have the form of a thrust slice like the anorthositic gabbro panel south of Kasten Lake, previously described, and it may not have moved far. If the essentially undisturbed metadiabase dykes within the complex are indeed Sudbury dykes, their southerly trend rather than east-southeasterly, as is typical in the Southern Province, may signify either wholesale rotation of the complex after dyke intrusion, or a regional difference in stress field orientation across the front at the time of their intrusion.

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Electrical resistivity and porosity of core samples from the Sudbury Structure, Ontario

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Abstract: Electrical resistivity and porosity have been measured for 52 drill core samples representing major lithologies exposed in the Sudbury Structure, in order to assist interpretation of low resistivity anomalies observed at depth by surface electromagnetic (EM) surveys carried out across the structure by the Lithoprobe program.

Results indicate that many samples from Onaping and Granophyre formations show high resistivities (10 000-60 000 Ω • m), and some from the Chelmsford Formation show lower resistivities (1-30 Ω • m), somewhat consistent with former findings. However, low resistivities in the order of 7-20 Ω • m have been unexpectedly found in the lower section of the Onaping Formation. These findings are significant in relation to the deep anomalies found by surface EM surveys.

Résumé : On a mesuré la résistivité électrique et la porosité de 52 échantillons de carottes de forage qui représentent les principales lithologies exposées dans la structure de Sudbury, pour mieux interpréter les faibles anomalies de résistivité observées en profondeur au moyen de levés électromagnétiques (EM) de surface effectués à travers la structure dans le cadre du programme Lithoprobe.

Les résultats indiquent que de nombreux échantillons provenant des formations d'Onaping et de Granophyre ont des valeurs de résistivité élevées (10 000 à 60 000 Ω • m), et que quelques-uns provenant de la Formation de Chelmsford ont des valeurs de résistivité plus faibles (1 à 30 Ω • m), résultats qui concorde dans une certaine mesure avec les résultats antérieurs. On a toutefois observé, contrairement à toute attente, de faibles valeurs de la résistivité (de l'ordre de 7 à 20 Ω • m) dans la partie inférieure de la Formation d'Onaping. Ces résultats sont importants lorsqu'ils sont considérés en fonction des anomalies profondes établies à l'aide des levés EM de surface.

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INTRODUCTION

Electrical resistivity, effective porosity and bulk density have been measured for 52 drill core samples representing major lithologies exposed in the Sudbury structure. The purpose is to obtain data required for interpretation of surface electromagnetic (Boerner et al., 1992) and logging data obtained across the Sudbury Structure as part of the Lithoprobe program. This paper contains results of these measurements and descriptions of methods used to obtain them.

The Sudbury Structure has been a significant source of massive sulphide deposits since their discovery in the late 1800s. Since that time, considerable effort has been made to determine the origin and configuration of this structure and its ore deposits. The Sudbury Structure is early Proterozoic in age (Dietz, 1964), is roughly elliptical in shape, and is filled with a well developed sequence of igneous and sedimentary rocks. Recent interpretations (Gupta et al., 1984) have contradicted the conventionally accepted "impact" theory as the origin of the structure. In addition, deep seismic reflection (Milkereit et al., in press; Salisbury et al., in press) and electromagnetic profiles (Boerner et al., 1992) recently conducted across the Sudbury Structure by the Lithoprobe Project have produced unexpected results. The seismic data show prominent reflectors that can be traced from the outcrop in the northern rim of the structure (the North Range) to great depth beneath the southern rim (the South Range) of the structure, and the EM data indicate a south dipping conductor under the South Range. Laboratory petrophysical measurements are essential because they are likely to provide significant information regarding the origin of the Sudbury Structure.

A brief description of the geology related to the samples used in this study can be found in Salisbury et al. (in press). The sequence of igneous and sedimentary rocks that fill the Sudbury Basin is about 5 km thick at the centre of the basin. The lower half of the sequence, the Sudbury Igneous Complex (SIC), is magmatic in origin consisting of felsic norite or gabbro overlain by a thick unit of granophyre. The norite contains the well known massive sulphides of Sudbury. Overlying the granophyre is the Onaping Formation, a thick tuff or fallback breccia containing clasts of Archean country rock and melt bodies at its base. This is overlain by the

Table 1. Sample depth in borehole and in formation sequence of the Sudbury Structure

Samples	Borehole Depth (m)	Sequence Depth (m)	Samples	Borehole Depth (m)	Sequence Depth (m)
4301-1	6	200	M9-1	2.7	2740
4301-2	75	280	M9-2	91	2810
4301-3	188	390	M9-3	183	2900
4301-4	188	390	M9-4	274	2970
4301-5	279	480	M9-5	488	3220
4301-6	284	480	M9-6	549	3280
4301-7	278	480	M9-7	640	3370
4301-8T	371	570	M9-8	731	3460
4301-8B	371	570	M9-9	823	3550
4301-9	408	620	M9-10	914	3640
			M9-11	1006	3740
3502-1	104	1050	M9-12	1097	3830
3502-2	183	1120	M9-13	1189	3920
3502-3	261	1200	M9-14	1280	4010
3502-4	323	1260	M9-15	1372	4110
3502-5	362	1300	M9-16	1463	4190
			M9-17	1554	4280
5601-1	145	1590	M9-18	1645	4380
5601-2	255	1700	M9-19	1737	4470
5601-3	359	1810	M9-20	1829	4560
5601-4	450	1900	M9-21	1012	4740
5601-5	544	1910	M9-22	2103	4830
5601-6	646	2100	M9-23	2195	4920
			M9-24	2286	5020
52847-1	9	2390	M9-25	2377	5110
52847-2	70	2450	M9-26	2469	5200
52847-3	97	2480	M9-27	2542	5260
52847-4	186	2570			
52847-5	270	2650			
52847-6	307	2690			
52847-7	336	2720			

Table 2a. Dimensions of specimens cut out from the Sudbury samples for electrical measurements: Boreholes 4301, 3502, 5601 and 52847

Samples	r_D (cm)	ℓ (cm)	W (g)	K_G (m)	δ (g/cm^3)
4301-1	2.468	0.654	8.9232	0.073	2.85
4301-2	2.467	0.675	8.9589	0.071	2.78
4301-3	2.469	0.713	9.4714	0.067	2.77
4301-4	2.460	0.643	8.5549	0.074	2.80
4301-5	2.465	0.669	8.7264	0.071	2.73
4301-6	2.470	0.447	6.1107	0.107	2.85
4301-7	2.466	0.813	10.9511	0.059	2.82
4301-8T	2.465	0.575	7.6057	0.083	2.77
4301-8B	2.465	0.754	10.1551	0.063	2.82
4301-9	2.464	0.710	9.3346	0.067	2.76
3502-1	2.467	0.867	11.0284	0.055	2.66
3502-2	2.465	1.774	22.8226	0.027	2.70
3502-3	2.466	0.799	10.6389	0.060	2.79
3502-4	2.460	0.693	9.4175	0.069	2.86
3502-5	2.462	0.625	8.2935	0.076	2.79
5601-1	2.472	0.568	7.4572	0.085	2.74
5601-2	2.473	0.684	9.1623	0.070	2.79
5601-3	2.469	0.660	8.8696	0.073	2.81
5601-4	2.470	0.701	9.3330	0.069	2.78
5601-5	2.470	0.635	8.5261	0.075	2.80
5601-6	2.470	0.683	9.1487	0.070	2.80
52847-1	2.466	0.517	6.9490	0.092	2.81
52847-2	2.459	0.559	7.6838	0.085	2.89
52847-3	2.465	0.611	8.1144	0.078	2.78
52847-4	2.464	0.595	7.8940	0.080	2.78
52847-5	2.461	0.644	8.6975	0.074	2.84
52847-6	2.463	0.492	6.9066	0.097	2.95
52847-7	2.465	0.598	7.7633	0.080	2.72

r_D = diameter
ℓ = thickness
W = weight
K_G = geometric factor
δ = bulk density

Onwatin Shale, a pelagic unit in gradational contact with the overlying Chelmsford Formation, which consists of greywacke and minor interlayered shale. The 52 core samples were obtained from 5 boreholes located in the North Range of the Sudbury Structure, at depths ranging from 336 to 2542 m (Table 1). They represent the top five major formations: Chelmsford greywacke, Onwatin Shale, Onaping, granophyre and gabbro-norite formations.

METHOD OF INVESTIGATION

Sample preparation

Cylindrical core samples of 2.54 cm (1 inch) in diameter were cut out, in the horizontal direction, from approximately vertical drill cores of 2.54 (1 inch) to 3.81 cm (1.5 inches) in diameter, at the Dalhousie University. At least two disc specimens were cut from each of the cylindrical samples for the laboratory measurements of electrical resistivity, porosity and bulk density. The geometric characteristics of the specimens used for electrical measurements are listed in Tables 2a and b, their dimensions ranging from 2.4-2.5 cm in diameter and 0.4-1.8 cm in thickness. The geometric factor, K_G, calculated for all specimens used for the electrical measurements is

$$K_G = r_D 2 \pi / \ell \qquad (1)$$

where r_D is the diameter and ℓ is the thickness of the specimen. The specimens for effective porosity measurements were selected from those not used for the electrical measurements. The specimens used for electrical measurements were also used for bulk density measurements.

Table 2b. Dimensions of specimens cut out from the Sudbury samples for electrical measurements: Borehole M-9

Samples	r_D (cm)	ℓ (cm)	W (g)	K_G (m)	δ (g/cm^3)
M9-1	2.470	0.659	8.5919	0.073	2.72
M9-2	2.470	0.528	6.9192	0.091	2.73
M9-3	2.466	0.559	7.3830	0.085	2.77
M9-4	2.460	0.590	7.7945	0.081	2.78
M9-5	2.458	0.581	7.6166	0.082	2.76
M9-6	2.464	0.563	7.4611	0.085	2.78
M9-7	2.469	0.695	8.9081	0.069	2.68
M9-8	2.464	0.596	7.8165	0.080	2.75
M9-9	2.467	0.685	8.9687	0.070	2.74
M9-10	2.455	0.640	8.3133	0.074	2.74
M9-11	2.465	0.563	7.4131	0.085	2.76
M9-12	2.468	0.584	7.6170	0.082	2.73
M9-13					
M9-14	2.466	0.680	8.9014	0.070	2.74
M9-15	2.464	0.683	9.0612	0.070	2.78
M9-16	2.466	0.651	8.8628	0.073	2.85
M9-17	2.465	0.654	9.0242	0.073	2.89
M9-18					
M9-19	2.464	0.574	8.0735	0.083	2.95
M9-20	2.469	0.637	8.8345	0.075	2.90
M9-21	2.464	0.655	8.9123	0.073	2.85
M9-22	2.465	0.598	8.1310	0.080	2.85
M9-23	2.465	0.588	8.1137	0.081	2.89
M9-24	2.464	0.625	8.6065	0.076	2.89
M9-25	2.466	0.641	8.9167	0.075	2.91
M9-26	2.463	0.676	9.5658	0.070	2.97
M9-27	2.465	0.748	9.8337	0.064	2.75

r_D = diameter
ℓ = thickness
W = weight
K_G = geometric factor
δ = bulk density

Bulk density measurements

The caliper method (American Petroleum Institute, 1960) was used to determine the bulk density, δ, of the samples. This method requires measurement of dimensions and weight of the specimens and is derived from

$$\delta = (r_D 2 \pi \ell / W)^{-1}, \qquad (2)$$

where W is the weight of the room dry (humidity 40%) specimen.

Effective porosity measurements

Effective porosity in principle includes the pore volume of all interconnected pores, and is determined from the difference in weight between the oven-dried and water saturated rock specimen:

$$\phi_E = \delta (W_w - W_D) / (\delta_W W_D) \qquad (3)$$

where ϕ_E is the effective porosity, W_w, W_D and δ are the wet weight, dry weight and bulk density of the rock sample, respectively, and δ_W is the density of the pore water (usually considered as unity). Details of the technique, including its advantages and limitations are described in Katsube et al. (1992a) and Katsube and Scromeda (1991). The recommended practice for core-analysis procedures (American Petroleum Institute, 1960) has generally been followed in these papers.

The degree of saturation, S_r, which is expressed as follows:

$$S_r = (W_r - W_D) / (W_W - W_D) \qquad (4)$$

where W_r is the weight of the specimen when it reaches a constant value under vacuum drying, represents the degree of moisture bound to the pore surfaces. Further details of the technique are described in Katsube et al. (1992a).

Electrical resistivity measurements

The techniques used for obtaining electrical resistivity measurements have previously been described in Katsube (1975, 1981), Katsube and Walsh (1987), and more recently in Katsube et al. (1991) and Katsube and Salisbury (1991). First, the complex resistivity, ρ^*, is measured:

$$\rho^* = \rho_R + i\, \rho_I, \qquad (5)$$

where ρ_R is the real resistivity and ρ_I is the imaginary resistivity, from impedance Z (Θ) measurements:

$$r^* = K_G Z(\Theta), \quad (6)$$

where Θ is the phase angle. Then the electrical resistivity, ρ_r, is determined from the Cole-Cole plots of the complex resistivity (ρ^*), by the method described in Katsube (1975) and Katsube and Walsh (1987), or in more recent publications (e.g., Katsube, et al., 1991; Katsube and Salisbury, 1991).

Measurement errors

Effective porosity and complex resistivity measurements were made at room temperature. The errors for these measurements are estimated to be generally in the ranges of 9-10 per cent and 10-20 per cent, respectively.

Table 3a. Results of the effective porosity measurements: Boreholes 4301, 3502, 5601 and 52847

Samples	δ (g/cm³)	W_w (g)	W_D (g)	S_r (%)	ϕ_E (%)
4301-1	2.85	5.3178	5.3071	14.0	0.57
4301-2	2.78	5.3206	5.3117	16.9	0.47
4301-3	2.77	4.6193	4.6139	11.1	0.32
4301-4	2.80	3.6619	3.6523	10.4	0.74
4301-5	2.73	5.4545	5.4455	18.9	0.45
4301-6	2.85	5.5292	5.5220	19.4	0.37
4301-7	2.82	6.1578	6.1491	25.3	0.40
4301-8T	2.77	4.9663	4.9545	24.6	0.66
4301-8B	2.82	5.5546	5.5444	27.5	0.52
4301-9	2.76	3.9889	3.9605	23.2	1.98
3502-1	2.66	5.0608	4.9910	27.5	3.72
3502-2	2.70	2.9512	2.9056	13.2	4.24
3502-3	2.79	2.6831	2.6672	18.9	1.66
3502-4	2.86	3.4202	3.4133	21.7	0.58
3502-5	2.79	4.3596	4.3514	25.6	0.53
5601-1	2.74	4.7880	4.7794	32.6	0.49
5601-2	2.79	3.3277	3.3133	30.6	1.21
5601-3	2.81	4.9429	4.9331	32.7	0.56
5601-4	2.78	3.0110	3.0049	29.5	0.56
5601-5	2.80	4.3246	4.3163	27.7	0.54
5601-6	2.80	4.5291	4.5240	35.3	0.32
52847-1	2.81	4.0192	4.0148	52.3	0.31
52847-2	2.89	6.9306	6.9278	57.1	0.12
52847-3	2.78	2.7134	2.7091	41.9	0.44
52847-4	2.78	5.0472	5.0426	34.8	0.25
52847-5	2.84	5.7005	5.6970	40.0	0.17
52847-6	2.95	3.3918	3.3879	18.0	0.34
52847-7	2.72	8.5444	8.5332	39.3	0.36

W_w = wet weight
δ = bulk density
W_D = dry weight
ϕ_E = effective porosity

EXPERIMENTAL RESULTS

The results of bulk density (δ) determination using Equation (2) are listed in Tables 2a and b. The results of the effective porosity (ϕ_E) and degree of saturation (S_r) determinations, using Equations (3) and (4), are listed in Tables 3a and 3b.

Two sets of complex resistivity measurements were carried out on each specimen from the 52 samples. The results of the electrical resistivity (ρ_r) determinations are listed in Tables 4a and b. Only two samples (52847-5 and 52847-6) showed irregular shaped arcs deviating from semi-circles, which suggest a slightly reduced accuracy due to limitations of the measuring system (Katsube et al., 1992b). These samples display very high resistivities, and are marked with an asterisk in Table 4a. Concern for reduced accuracy caused by extremely low resistivities also exists, because the system used for these measurements is designed for high resistivities (Katsube, 1975). Samples that may fall in this category are those with electrical resistivity values below 10 Ω • m, and are marked with two asterisks in Table 4a.

Table 3b. Results of the effective porosity measurements: Borehole M-9

Samples	δ (g/cm³)	W_w (g)	W_D (g)	S_r (%)	ϕ_E (%)
M9-1	2.72	5.6836	5.6712	40.3	0.59
M9-2	2.73	3.4003	3.3935	27.9	0.55
M9-3	2.77	2.4070	2.4027	32.6	0.50
M9-4	2.78	3.4777	3.4719	36.2	0.46
M9-5	2.76	4.2579	4.2487	43.5	0.60
M9-6	2.78	2.4791	2.4721	21.4	0.79
M9-7	2.68	4.2975	4.2791	44.0	1.15
M9-8	2.75	3.9968	3.9864	26.9	0.72
M9-9	2.74	4.7733	4.7610	32.5	0.71
M9-10	2.74	4.4782	4.4679	16.5	0.63
M9-11	2.76	4.9964	4.9789	36.6	0.97
M9-12	2.73	5.1539	5.1412	33.1	0.67
M9-13					
M9-14	2.74	4.6236	4.6038	29.3	1.18
M9-15	2.78	3.3412	3.3262	31.3	1.25
M9-16	2.85	4.2225	4.2132	22.6	0.63
M9-17	2.89	6.7468	6.7263	36.6	0.88
M9-18					
M9-19	2.95	4.5800	4.5654	7.5	0.94
M9-20	2.90	5.0872	5.0752	15.8	0.69
M9-21	2.85	5.9206	5.9105	10.9	0.49
M9-22	2.85	4.4562	4.4455	9.4	0.69
M9-23	2.89	3.7964	3.7886	10.3	0.59
M9-24	2.89	5.5435	5.5295	7.1	0.73
M9-25	2.91	5.6334	5.6215	6.7	0.62
M9-26	2.97	5.6403	5.6272	1.5	0.69
M9-27	2.75	4.3972	4.3836	2.2	0.85

W_w = wet weight
δ = bulk density
W_D = dry weight
ϕ_E = effective porosity

Results of the three major parameters (δ, ϕ_E, ρ_r) are compiled in Tables 5a and 5b.

DISCUSSION AND CONCLUSIONS

The bulk density (δ) is in the range of 2.66-2.97 g/cm^3 with an average of 2.80 g/cm^3 for these samples, results generally consistent with previous measurements (Katsube and Salisbury, 1991) which were in the range of 2.63-2.86 g/cm^3.

There is a black-grey tuff boundary in the Onaping Tuff Formation (Katsube and Salisbury, in press), just between samples number 52847-1 and 52847-2. Both effective porosity (ϕ_E) and electrical resistivity (ρ_r) are relatively constant with depth for samples from formations below the black-grey boundary, with effective porosities (ϕ_E) in the range of 0.12-1.25% generally increasing with depth, and electrical resistivities (ρ_r) in the range of 7000-53 000 $\Omega\cdot$ m generally decreasing with depth. These values are consistent with previous measurements (Katsube and Salisbury, 1991).

Both parameters show considerable scatter with no consistent relationship with depth for formations above the black-grey boundary, with effective porosities (ϕ_E) and electrical resistivities (ρ_r) varying over the ranges of 0.3-4.2% and 0.6-30 000 $\Omega\cdot$ m, respectively. These electrical resistivity (ρ_r) and effective porosity (ϕ_E) values are generally consistent with previous laboratory measurements (Katsube and Salisbury, 1991), except for the very low resistivity values (0.6-16 $\Omega\cdot$ m) in the Onaping Formation just above the black-grey boundary. Previous measurements have shown the Onaping Formation to have resistivities in the range of 12 000-49 000 $\Omega\cdot$ m (Katsube and Salisbury, 1991), values similar to those of the rocks below the black-grey boundary. No previous data similar in nature are known to exist for the Onwatin Shale, which the effective porosities (ϕ_E) and electrical resistivities are in the ranges of 1.6-4.2% and 0.6-1.6 $\Omega\cdot$ m, respectively.

There are three low resistivity zones (Katsube and Salisbury, in press) in this group of samples from the five formations of the Sudbury Structure. These are

1. in the Chelmsford Greywacke Formation,
2. in the Onwatin Shale associated with increased porosity, and
3. in the Onaping Formation just above the black-grey boundary.

Table 4a. Complex electrical resistivity measurements: Boreholes 4301, 3502, 5601 and 52847

Sample/ Specimen	Bulk Electrical Resistivity ρ_r (10^3 Ω·m)		
	Measurement (#1)	Measurement (#2)	Mean
4301-1	17.7	16.4	17.1 ± 0.6
4301-2	7.67	7.75	7.7 ± 0.05
4301-3	9.37	10.9	10.1 ± 0.8
4301-4	0.0165	0.0157	0.016 ± 0.0005
4301-5	0.0286	0.0320	0.030 ± 0.002
4301-6	0.692	0.816	0.75 ± 0.066
4301-7	0.241	0.244	0.24 ± 0.004
4301-8T	0.0175	0.0038	0.011 ± 0.007
4301-8B**	0.0052	0.0052	0.0052 ± 0.0000
4301-9**	0.0015	0.0014	0.0015 ± 0.0001
3502-1**	0.0011	0.0014	0.0013 ± 0.0001
3502-2**	0.0005	0.0006	0.0006 ± 0.0001
3502-3**	0.0016	0.0015	0.0016 ± 0.0001
3502-4	30.6	29.9	30.2 ± 0.4
3502-5	0.944	0.933	0.94 ± 0.004
5601-1	1.63	1.98	1.81 ± 0.17
5601-2	1.67	1.65	1.66 ± 0.01
5601-3**	0.0092	0.0072	0.0082 ± 0.001
5601-4**	0.0085	0.0065	0.0075 ± 0.001
5601-5**	0.00997	0.0091	0.0094 ± 0.0003
5601-6	0.428	0.428	0.43 ± 0.00
52847-1	0.013	0.024	0.019 ± 0.005
52847-2	54.2	51.8	53.0 ± 1.2
52847-3	20.3	19.4	19.8 ± 0.5
52847-4	16.0	19.7	17.8 ± 1.9
52847-5*	47.7	53.4	50.6 ± 2.8
52847-6*	65.1	66.0	65.6 ± 0.4
52847-7	14.3	22.0	18.2 ± 3.8

Measurement (#1) : Measurement after 24 hours of saturation.
Measurement (#2) : Measurement after 48 hours of saturation.
* and ** : Somewhat reduced measurement accuracy.

Table 4b. Complex electrical resistivity measurements: Borehole M-9

Sample/ Specimen	Bulk Electrical Resistivity ρ_r, (10^3 Ω·m)		
Samples	Measurement (#1)	Measurement (#2)	Mean
M9-1	20.5	30.6	25.6 ± 5.0
M9-2	17.1	22.0	19.5 ± 2.5
M9-3	18.0	21.0	19.5 ± 1.5
M9-4	35.9	40.8	38.3 ± 2.5
M9-5	8.06	12.1	10.1 ± 2.0
M9-6	15.6	21.0	18.3 ± 2.7
M9-7	8.97	12.4	10.7 ± 1.7
M9-8	16.3	22.3	19.3 ± 3.0
M9-9	7.64	13.8	10.7 ± 3.1
M9-10	13.5	19.0	16.2 ± 2.8
M9-11	5.65	11.4	8.5 ± 2.9
M9-12	7.38	11.0	9.2 ± 1.8
M9-13			
M9-14	5.64	10.0	7.8 ± 2.2
M9-15	4.94	9.63	7.3 ± 2.3
M9-16	20.0	27.9	24.0 ± 4.0
M9-17	3.95	7.11	5.5 ± 1.6
M9-18			
M9-19	11.5	15.1	13.3 ± 1.8
M9-20	5.13	8.82	7.0 ± 1.8
M9-21	6.70	12.2	9.5 ± 2.7
M9-22	7.02	12.6	9.8 ± 2.8
M9-23	8.39	15.6	12.0 ± 3.6
M9-24	9.06	17.1	13.1 ± 4.0
M9-25	8.84	12.6	10.7 ± 1.9
M9-26	10.9	13.1	12.0 ± 1.1
M9-27	2.36	5.18	3.8 ± 1.4

Measurement (#1) : Measurement after 24 hours of saturation.
Measurement (#2) : Measurement after 48 hours of saturation.

Table 5a. Electrical resistivity and porosity of the core samples from the Sudbury Basin: Boreholes 4301, 3502, 5601 and 52847

Samples	Formation	Density δ (g/cm^3)	Porosity ϕ_E (%)	Resistivity ρ (x10^3 Ω·m)
4301-1	Chlm.-Wck	2.85	0.57	17.1 ± 0.6
4301-2		2.78	0.47	7.7 ± 0.05
4301-3		2.77	0.32	10.1 ± 0.8
4301-4		2.80	0.74	0.016 ± 0.0005
4301-5		2.73	0.45	0.030 ± 0.002
4301-6		2.85	0.37	0.75 ± 0.066
4301-7		2.82	0.40	0.24 ± 0.004
4301-8T		2.77	0.66	0.011 ± 0.007
4301-8B		2.82	0.52	0.0052 ± 0.0000
4301-9		2.76	1.98	0.0015 ± 0.0001
3502-1	Onwt.-Sh	2.66	3.72	0.0013 ± 0.0001
3502-2		2.70	4.24	0.0006 ± 0.0001
3502-3	-Arg	2.79	1.66	0.0016 ± 0.0001
3502-4	Onp.	2.86	0.58	30.2 ± 0.4
3502-5		2.79	0.53	0.94 ± 0.004
5601-1	Onp.	2.74	0.49	1.81 ± 0.17
5601-2		2.79	1.21	1.66 ± 0.01
5601-3		2.81	0.56	0.0082 ± 0.001
5601-4		2.78	0.56	0.0075 ± 0.001
5601-5		2.80	0.54	0.0094 ± 0.0003
5601-6		2.80	0.32	0.43 ± 0.00
52847-1	Onp. -bl	2.81	0.31	0.019 ± 0.005
52847-2		2.89	0.12	53.0 ± 1.2
52847-3	Onp. -gr	2.78	0.44	19.8 ± 0.5
52847-4		2.78	0.25	17.8 ± 1.9
52847-5	-Bsl	2.84	0.17	50.6 ± 2.8
52847-6		2.95	0.34	65.6 ± 0.4
52847-7	Grp.	2.72	0.36	18.2 ± 3.8

Petrophysical Parameters	Formations	Rock Types
δ = Bulk Density	Chlm.= Chelmsford	Wck = Wacke
ϕ_E = Effective Porosity	Onwt.= Onwatin	Sh = Shale
ρ = Electrical Resistivity	Onp. = Onaping	Arg = Argillite
		Bsl = Basal
		Grp.= Granophyr

Table 5b. Electrical resistivity and porosity of the core samples from the Sudbury Basin: Borehole M-9

Samples	Formation	Density δ (g/cm^3)	Porosity ϕ_E (%)	Resistivity ρ (x10^3 Ω·m)
M9-1	Grp.	2.72	0.59	25.6 ± 5.0
M9-2		2.73	0.55	19.5 ± 2.5
M9-3		2.77	0.50	19.5 ± 1.5
M9-4		2.78	0.46	38.3 ± 2.5
M9-5		2.76	0.60	10.1 ± 2.0
M9-6		2.78	0.79	18.3 ± 2.7
M9-7		2.68	1.15	10.7 ± 1.7
M9-8		2.75	0.72	19.3 ± 3.0
M9-9		2.74	0.71	10.7 ± 3.1
M9-10		2.74	0.63	16.2 ± 2.8
M9-11		2.76	0.97	8.5 ± 2.9
M9-12		2.73	0.67	9.2 ± 1.8
M9-13				
M9-14		2.74	1.18	7.8 ± 2.2
M9-15		2.78	1.25	7.3 ± 2.3
M9-16		2.85	0.63	24.0 ± 4.0
M9-17	Trs. Gb.	2.89	0.88	5.5 ± 1.6
M9-18				
M9-19	Fls. Nrt	2.95	0.94	13.3 ± 1.8
M9-20		2.90	0.69	7.0 ± 1.8
M9-21		2.85	0.49	9.5 ± 2.7
M9-22		2.85	0.69	9.8 ± 2.8
M9-23		2.89	0.59	12.0 ± 3.6
M9-24		2.89	0.73	13.1 ± 4.0
M9-25	Drc. Nrt	2.91	0.62	10.7 ± 1.9
M9-26	Ftwl	2.97	0.69	12.0 ± 1.1
M9-27		2.75	0.85	3.8 ± 1.4

Petrophysical Parameters	Formations	Rock Types
δ = Bulk Density	Chlm.= Chelmsford	Wck = Wacke
ϕ_E = Effective Porosity	Onwt.= Onwatin	Sh = Shale
ρ = Electrical Resistivity	Onp. = Onaping	Arg = Argillite
		Bsl = Basal
		Grp.= Granophyr

The existence of the third low resistivity anomaly in the Onaping Tuff Formations just above the black-grey tuff boundary appears to be new information. These data may help explain the source of the low resistivity EM anomaly observed at depth under the South Range of the Sudbury Structure.

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Platinum group element concentrations in the Sudbury rocks, Ontario – an indicator of petrogenesis[1]

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Abstract: As part of the geochemical study of the Sudbury Structure, low-level PGE have been determined for both the Sudbury Igneous Complex (SIC) and its country rocks using the ICP-MS method. Except for the Nipissing diabase dykes, all the country rocks contain very low PGE (Pt 2.6-5.3 ppb, Pd <3.9 ppb, Ir 0.02-0.15 ppb, and Ru <0.01-0.15 ppb). The SIC samples have a large range of PGE concentrations (0.4-12.4 ppb Pt, 0-7.4 ppb Pd, 0.025-2 ppb Ir, and 0-0.1 ppb Ru). The Pt content increases from felsic norite of the lower zone to near the transitional zone and decreases sharply in the granophyre of the upper zone; Pd seems to decrease from the lower zone upwards.

It is concluded that: 1) the impact melting of the Sudbury regional rocks could not have formed the Sudbury Igneous Complex and its sulphide ores under normal conditions; 2) granophyre is distinct from the norite in terms of PGE contents, suggesting either the result of magma differentiation or magma immiscibility.

Résumé : Dans le cadre de l'étude géochimique de la structure de Sudbury, on a déterminé la présence de faibles concentrations d'ÉGP dans le complexe igné de Sudbury (SIC) et dans ses roches encaissantes par la méthode de la spectrométrie de masse du plasma couplé par induction (ICP-MS). Sauf en ce qui concerne les dykes de diabase de Nipissing, toutes les roches encaissantes ont des teneurs très faibles d'ÉGP (2,6 à 5,3 ppb de Pt, ,<3,9 ppb de Pd, 0,02 à 0,15 ppb d'Ir et <0 ,01 à 0,15 ppb de Ru). Les échantillons du SIC présentent une large gamme de concentrations d'ÉGP (0,4 à 12,4 ppb de Pt, 0 à 7,4 ppb de Pd, 0,025 à 2 ppb d'Ir et 0-0,1 ppb de Ru). La teneur en Pt augmente entre la norite felsique de la zone inférieure et les abords de la zone de transition, et diminue fortement dans le granophyre de la zone supérieure; il semble que la teneur en Pd diminue vers le haut, à partir de la zone inférieure.

On en conclut que: 1) il est peu probable que la fusion des roches régionales de Sudbury sous l'effet d'un impact soit à l'origine du SIC et de ses minerais sulfurés dans des conditions normales; 2) le granophyre est distinct de la norite du point de vue de sa teneur en ÉGP, ce qui pourrait s'expliquer soit par la différenciation soit par l'immiscibilité du magma.

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INTRODUCTION

The Sudbury structure is one of the most scientifically and economically important geological features in the world (Fig. 1). It comprises the Sudbury Igneous Complex (SIC), the Sudbury Basin, and the brecciated basement rocks of the Superior and Southern provinces (Dressler, 1984). A series of large Ni-Cu-PGE deposits occur along the base of the SIC, making it the largest Ni-Cu camp in the world.

The ring structure of the Sudbury Igneous Complex, widespread distribution of Sudbury Breccia and the presence of shatter cones in the footwall rocks led Dietz (1964) to suggest that the Sudbury Structure was the site of a major meteorite impact. This hypothesis was soon adopted by petrologists and geochemists to explain the silicate-rich nature of the Sudbury Igneous Complex and the triggering of sulphide segregation by mixing of mantle magma and impact melt (Stevenson and Colgrove, 1968; Naldrett and McDonald, 1980). Rare-earth element data and radiogenic isotopic studies all point to a substantial contribution from the crust to the Sudbury Igneous Complex (Gibbins and McNutt, 1975; Kuo and Crocket, 1979; Faggart et al., 1985; Naldrett et al., 1986). Based on a comparative study of meteorite impact sites in the world and calculation of the size of the Sudbury Structure, Grieve et al. (1991) proposed a hypothesis of complete melting of the crustal rocks to form the Sudbury Igneous Complex, as well as the sulphide ores. Therefore, the question has become: does the Sudbury Igneous Complex represent crystallization of a mantle derived magma, a mixture of mantle magma with crustal melt, or a completely crustal melt?

The present study represents part of a project undertaken to understand better the origin of the Sudbury Structure and the potential of its Ni-Cu-PGE resources. This project consists of two subprojects: the geochemical study of the Sudbury Igneous Complex, and the study of the sulphide ores along a vertical profile through the Sudbury contact mineralization.

The objective of the geochemical study is to model the possible source of the Sudbury Igneous Complex rocks and its related sulphide ores. This will be pursued through a systematic geochemical investigation of the Sudbury Igneous Complex and its country rocks by analyzing and comparing a spectrum of major and trace elements, REE and PGE, in both the SIC and its country rocks.

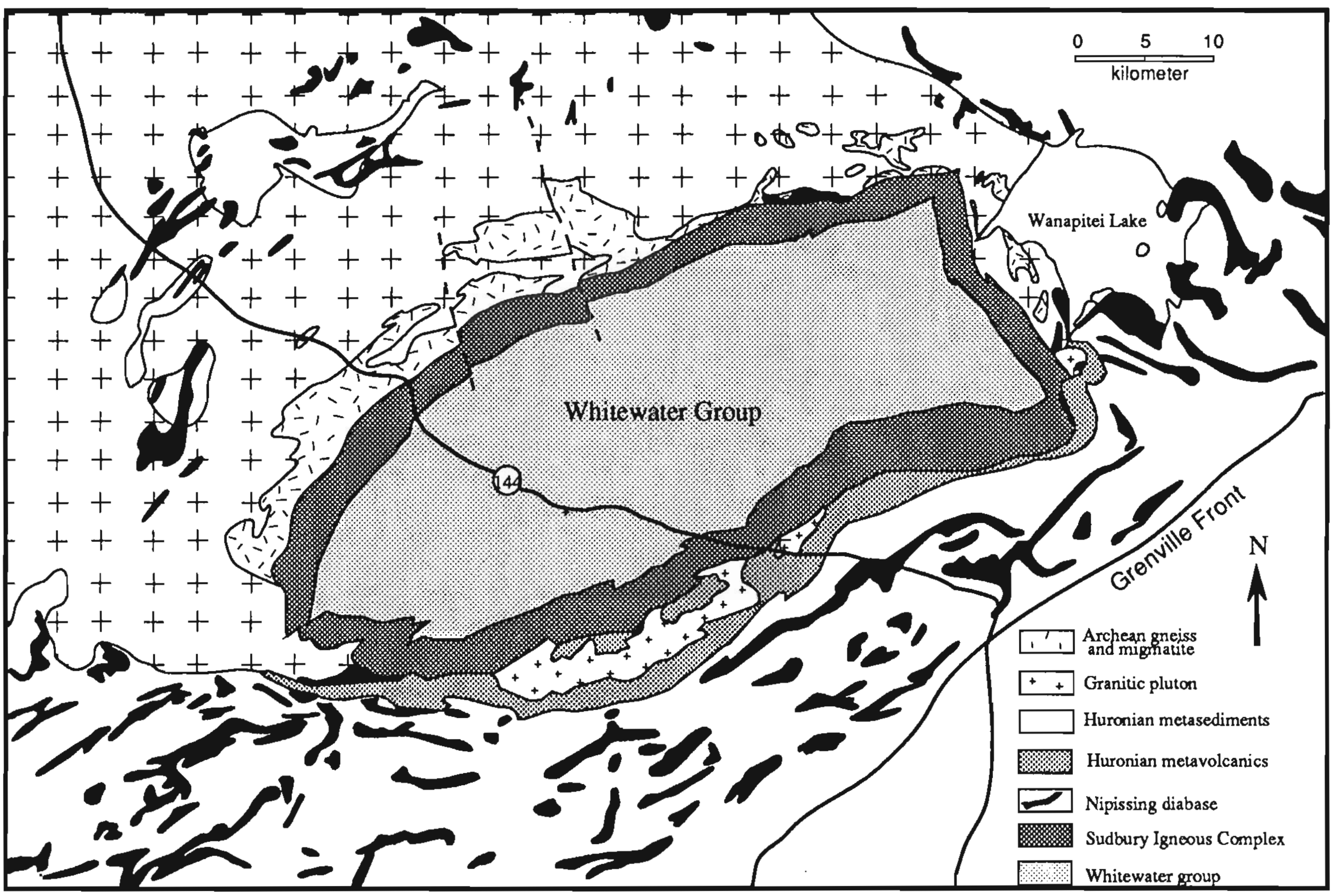

Figure 1. Simplified geological map showing the relationship of the Sudbury Structure to specific lithological units.

This paper reports part of the geochemical study. We present the preliminary results of PGE analysis by ICP-MS and their implications for the origin and differentiation of magma.

FIELD SAMPLING

Sampling was carried out in June of 1992. A total of 65 samples have been collected at the first stage. Thirty-seven samples were taken from the Sudbury Igneous Complex along two sections of the Sudbury Igneous Complex, one across the north range and another across the south range, mostly along highway 144. A few samples were taken around the Strathcona mine to compensate for the lack of norite at the contact at highway 144. Twenty-six samples were taken from the country rocks to the Sudbury Igneous Complex, representing seven major units older than the Sudbury event. These samples are taken from several localities of each unit and are considered to be representative of the rocks at the impact site. In previous studies, the composition of the crustal component was portrayed by data representing the average composition of the Superior basement rocks (Kuo and Crocket, 1979; Grieve et al., 1991) or simply by one sample from the Levack gneiss (Walker et al., 1991), or Huronian group (Dickin et al., 1992). The representativeness of these compositions is questionable. Our samples were taken from all units of the country rocks to the Sudbury Igneous Complex, and should more realistically represent the possible source rocks to the complex.

All the samples were thoroughly washed, and only clean, fresh material was taken for pulverization. The samples were crushed to half inch size by a manganese plate. These coarse grains were then ground to <6 mesh by ceramic disks. 40 g of rock powder were further pulverized with ceramic balls to <200 mesh.

ANALYTICAL PROCEDURES

The low-level PGE analysis was carried out in the chemical laboratories of the Geological Survey of Canada, using the ICP-MS method developed by Grégoire (1988). The following describes the analytical procedures:

Table 1. Analytical results for blank standard and detection limits

Sample No.	Ir (ppb)	Ru (ppb)	Pd (ppb)	Pt (ppb)
Blank 1	0.143	0.166	0.748	0.080
	0.148	0.157	0.787	0.034
	0.133	0.149	0.830	0.127
Blank 2	0.154	0.134	0.889	<0.047
	0.165	0.115	0.950	<0.047
	0.164	0.111	1.048	<0.047
Blank 3	0.146	0.135	0.456	<0.047
	0.126	0.139	0.386	<0.047
	0.143	0.124	0.488	<0.047
Blank 4	0.139	0.141	1.127	<0.047
	0.135	0.130	1.116	<0.047
	0.134	0.134	1.284	<0.047
Blank 5	0.138	0.163	1.161	<0.047
	0.140	0.107	1.152	<0.047
	0.148	0.141	1.092	<0.047
Blank 6	0.137	0.094	0.493	<0.047
	0.137	0.140	0.473	<0.047
	0.140	0.100	0.502	<0.047
2σ	0.010	0.021	0.300	0.047
Average	0.135	0.125	0.789	0.081

1. Dissolution of silicates: 5 g of rock powder were dissolved in 20 ml of concentrated hydrofluoric acid, and 50 ng of enriched isotope spike was added to each sample. Six blank standards were also made using the same amount of isotope spike. After evaporation of the acid, 25 ml of aqua regia was added and evaporated. Then addition of 10 ml of concentrated HCl was followed with the same evaporation procedure. 20 ml of 1M HCl were added and the solution was transferred to a 50 ml plastic centrifuge tube. The mixture was centrifuged until supernate was clear, and the solution was transferred to a 250 ml beaker. The residue was centrifuged again and the supernate was mixed with the previous one. The final volume of solution was made up to 100 ml.

2. PGE precipitation: 5 ml of tellurium solution containing 1 mg per ml Te was added to the solution, and the solution was brought to boiling. Then 5 ml of $SnCl_2$ solution was added. The solution was further heated until the Te metal precipitate coagulated into large particles or clumps. Then a small amount of either solution was added until no precipitation took place. The solution was filtered through 0.45μm Millipore filtration apparatus. The residue was dissolved with aqua regia and the solution was carefully evaporated to near dryness. Three ml of 1.5M HNO_3 were added, and the sample solution was transferred to a 15 ml centrifuge tube. The final volume was made up to 5 ml using 1.5M HNO_3.

3. ICP-MS analysis: The analysis of the solution was performed on a SCIEX ELAN 5000 instrument. A sample intake rate of 0.5 ml/min was used. Electrothermal vaporization was carried out using the conditions described by Grégoire (1988). Each sample aliquot was measured three times and a blank reagent was always used for at least 30 seconds to avoid any contamination. The instrument operation, ion counting, and data reduction were controlled by an IBM microcomputer with the software provided by the manufacturer.

ANALYTICAL RESULTS

Four PGE: Pt, Pd, Ir, and Ru have been determined for both the blank standards and the rock samples. The results of the blank standard are listed in Table 1. They are all very low

Table 2. Analytical results of the Sudbury samples

Sample No.	Rock type	Ir (ppb)	Ru (ppb)	Pd (ppb)	Pt (ppb)	Pt/Pd
92-cs-1	Huronian Metasediments	0.10 (0.01)	0.15 (0.01)	1.90 (0.20)	2.66 (0.09)	1.40
92-cs-2	Nipissing Diabase	0.06 (0.01)	0.08 (0.01)	19.34 (0.10)	20.25 (0.15)	1.05
92-cs-3	Huronian Metavolcanics	0.14 (0.01)	0.07 (0.01)	3.89 (0.06)	3.90 (0.04)	1.00
92-cs-4	Murray Granite	0.10 (0.03)	0.02 (<0.01)	<0.3	5.29 (0.04)	
92-cs-5	Archean Granite	0.02 (0.02)	<0.02	<0.3	3.78 (0.06)	
92-cs-6	Levack Migmatite	0.06 (0.02)	0.03 (0.01)	2.02 (0.11)	2.60 (0.06)	1.29
92-cs-7	Levack Gneiss	0.07 (0.01)	0.02 (0.01)	2.20 (0.11)	5.10 (0.11)	2.31
92-sic-1	Felsic Norite	0.03 (0.01)	0.03 (<0.01)	0.66 (0.32)	5.54 (0.04)	8.38
92-sic-2	Felsic Norite	0.05 (0.02)	<0.02	7.43 (0.20)	3.89 (0.04)	0.52
92-sic-3	Felsic Norite	0.03 (0.01)	<0.02	4.35 (0.10)	1.53 (0.03)	0.35
92-sic-4	Felsic Norite	0.05 (0.01)	0.03 (0.01)	2.58 (0.09)	0.76 (0.02)	0.29
92-sic-5	Felsic Norite	0.04 (0.02)	<0.02	2.81 (0.10)	2.33 (0.07)	0.83
92-sic-6	Felsic Norite	0.06 (0.03)	0.05 (<0.01)	4.15 (0.08)	2.35 (0.02)	0.57
92-sic-7	Felsic Norite	0.02 (<0.01)	<0.02	1.99 (0.13)	2.14 (0.01)	1.08
92-sic-8	Felsic Norite	0.07 (<0.01)	<0.02	0.85 (0.04)	12.41 (0.04)	14.61
92-sic-9	Felsic Norite	0.10 (0.01)	0.06 (0.01)	2.30 (0.03)	6.29 (0.09)	2.73
92-sic-10	Felsic Norite	0.04 (0.01)	<0.02	1.05 (0.03)	6.87 (0.18)	6.54
92-sic-11	Granophyre	0.18 (0.02)	0.07 (0.01)	0.93 (0.05)	1.60 (0.05)	1.72
92-sic-12	Granophyre	0.04 (0.02)	<0.02	<0.3	0.42 (0.06)	
92-sic-13	Granophyre	0.20 (0.01)	0.11 (0.01)	0.50 (0.04)	1.41 (0.03)	2.80
92-sic-14	Granophyre	0.08 (0.01)	<0.02	<0.3	0.39 (0.11)	

(<1 ppb) and show closely similar values for each particular element, indicating good reproducibility of the method. The detection limit of this method is determined by the average of standard deviation. The detection limits for Pd, Ir, Pt, and Ru are 0.30, 0.01, 0.05, and 0.02 ppb, respectively.

Table 2 shows our results for samples of both the SIC and the country rocks. The precision of the ICP-MS analysis is also given with the results as 2 s.d. of three measurements of the same sample aliquot.

Platinum group element concentrations in the Sudbury Igneous Complex are generally low. The Pt and Pd contents are at ppb level, and Ir and Ru are at sub-ppb level. Their variations across the igneous stratigraphy are shown in Figure 2. Generally, Pd decreases from 5.4-7.43 ppb at the base of the norite to around 2 ppb in the rest of the norite zone and to below 1 ppb in granophyre. Platinum seems to increase slightly from the base of the norite to a maximum of 12.41 ppb in the upper part of the norite, and then starts to decrease in the granophyre. Iridium is low in norite, slightly above detection limit, and erratically variable in granophyre, with values up to 0.2 ppb. The Ru values are mostly below detection limit, thus convey little information about any systematic variation. Platinum shows a tendency toward inverse relationship with Ir (Fig. 3). This is to be expected because of their different behaviour. However, this trend is quite different from the common observation that Ir is a compatible element and should decrease with magma differentiation, and that Pd is the reverse (Brügman et al., 1987). This departure from the normal can be explained by two possible effects: 1) iridium behaves compatibly only when olivine and spinel are the cumulus minerals that can accommodate some Ir; 2) the erratic variation of Ir in granophyre may be a result of heterogeneous distribution of this element due to local contamination.

Most of the country rocks, except the Nipissing diabase, contain less than 5.3 ppb Pt and 3.89 ppb Pd. Their Ir and Ru contents are below 0.14 ppb. The granitic rocks contain almost no Pd although their Pt content seems high relative to other rock types. Generally, the mafic-felsic volcanic rocks of the Huronian group contain high Ir, and reasonably high Pt and Pd. Nipissing diabase has the highest Pt and Pd concentrations, but its Ru and Ir contents are very low, less than 0.08 ppb.

DISCUSSION

Kuo and Crocket (1979) studied the rare-earth element (REE) concentrations of the Sudbury Igneous Complex rocks. They found that the irruptive rocks were characterized by high absolute REE and a light rare-earth element-enriched chondrite normalized pattern compared to other layered intrusions. This led them to conclude that mixing of a silicic component of high La/Yb ratio with a normal basaltic melt was responsible for the distinctive REE characteristics of the irruptive magma. This mixing model of the Sudbury Igneous Complex was supported by the radiogenic isotopic studies of the Nd-Sm and Rb-Sr systems. Naldrett et al. (1986) obtained crustal-like ϵ_{Nd} (-5 to -9) isotopic results, which they attributed to crustal contamination of a mantle-derived basaltic melt. The addition of high SiO_2 melt was the trigger for the segregation of sulphide melt. Walker et al. (1991)

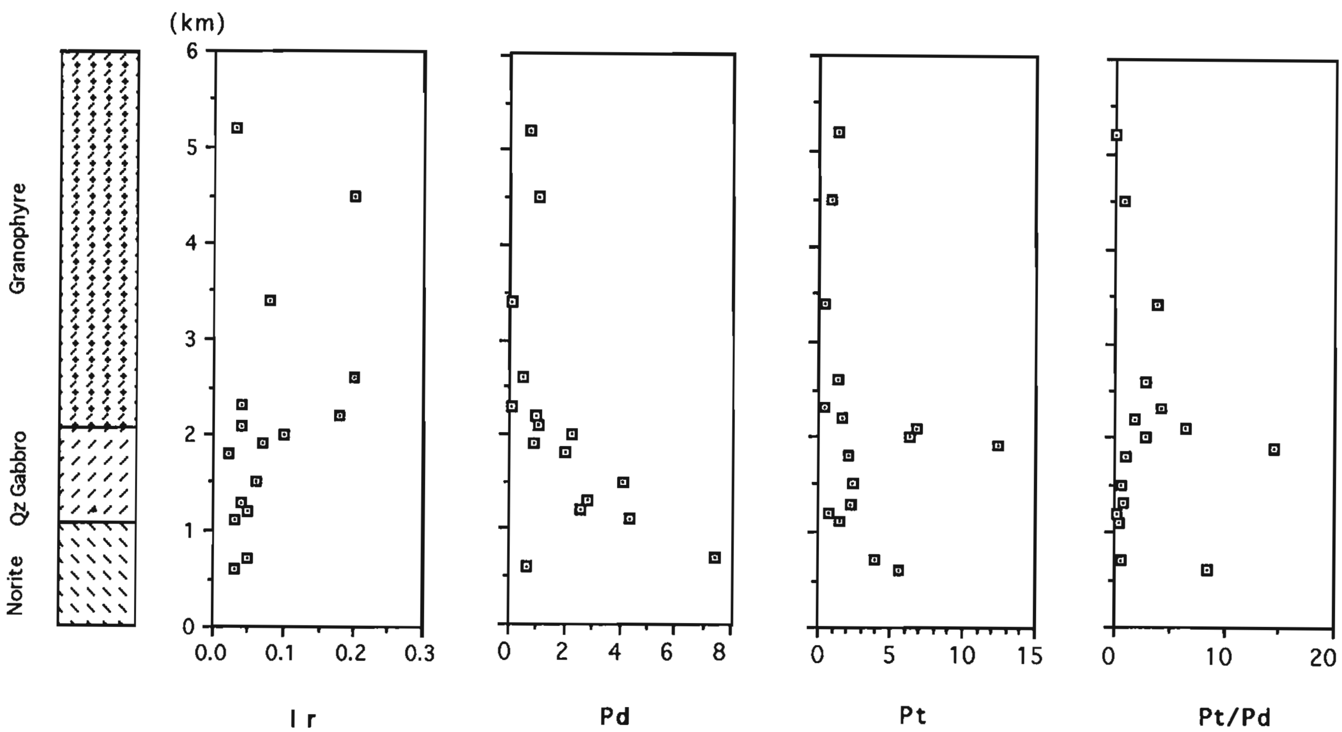

Figure 2. Distribution of Ir, Pd, Pt, and Pt/Pd through the lithostratigraphic sequence of the Sudbury Igneous Complex.

studied the Re-Os isotopic system of the Sudbury Ni-Cu ores. They obtained high initial $^{187}Os/^{188}Os$ and estimated that >75% of sulphide ores were contributed by ancient crust.

However, total melting produced by meteorite impact to form the Sudbury Igneous Complex magma is also proposed based on similar information. Faggart et al. (1985) reported results (average $e_{Nd} = -7.5$), similar to Naldrett et al. (1986), for the Sudbury Igneous Complex norite and gabbro, but interpreted them to mean that the magma was derived solely from meteorite-impact-induced melting of the Superior province basement rocks. Based on the calculation of the size of the Sudbury Structure, Grieve et al. (1991) found that the predicted meteorite impact melt is sufficient to generate the Sudbury Igneous Complex, and that its major elements can be modeled by a combination of Archean greenstones, granites, and Huronian sedimentary rocks. They concluded that the Sudbury Structure and its sulphide ores are entirely the result of melting of the crustal rocks.

Our PGE results are preliminary at this stage, and cannot provide sufficient evidence to lead to a definite answer as to whether and how much a mantle component contributed to the formation of the Sudbury Igneous Complex and its sulphide ores. However, this limited data does put some constraints on the hypothesis of complete crustal melting and has implications for the origin of the complex.

1. Generally, the PGE contents in the Sudbury Igneous Complex and its country rocks are similar if the Nipissing diabase is not considered. The Huronian metavolcanics contain slightly higher PGE than the average Sudbury Igneous Complex rocks. It would be possible to form the Sudbury Igneous Complex from crustal rocks in term of PGE concentration if the PGE in the sulphide ores are not included. In order to form the enormous amount of the sulphide ores with PGE in sulphide content around 3000 ppb for Pd or Pt (Coats and Snajdr, 1984), a highly depleted residual magma, i.e. the Sudbury Igneous Complex, would be expected. This is not the case at Sudbury based on our data. Furthermore, the Pt/Pd values of the country rocks are all greater than 1, while most of the norite that contains higher PGE have Pt/Pd values much below 1. If the country rocks were the source, then the Ni sulphide around the Sudbury basin should contain higher Pt than Pd. Coats and Snajdr (1984) compiled PGE data for the McCreedy West deposit. Although their values are higher than those of Hoffman et al. (1979), the Pt/Pd values are all less than one. Similarly, Li and Naldrett (1992) obtained PGE with Pt/Pd values less than one for different zones in the Strathcona mine. Naldrett (1984) listed average PGE values for the Sudbury deposits based on the available data. Almost all show less Pt than Pd.

Possibilities remain for the crustal melting model if two conditions are met: 1) The volume proportion of granophyre was much larger than that of norite; in this case the Sudbury

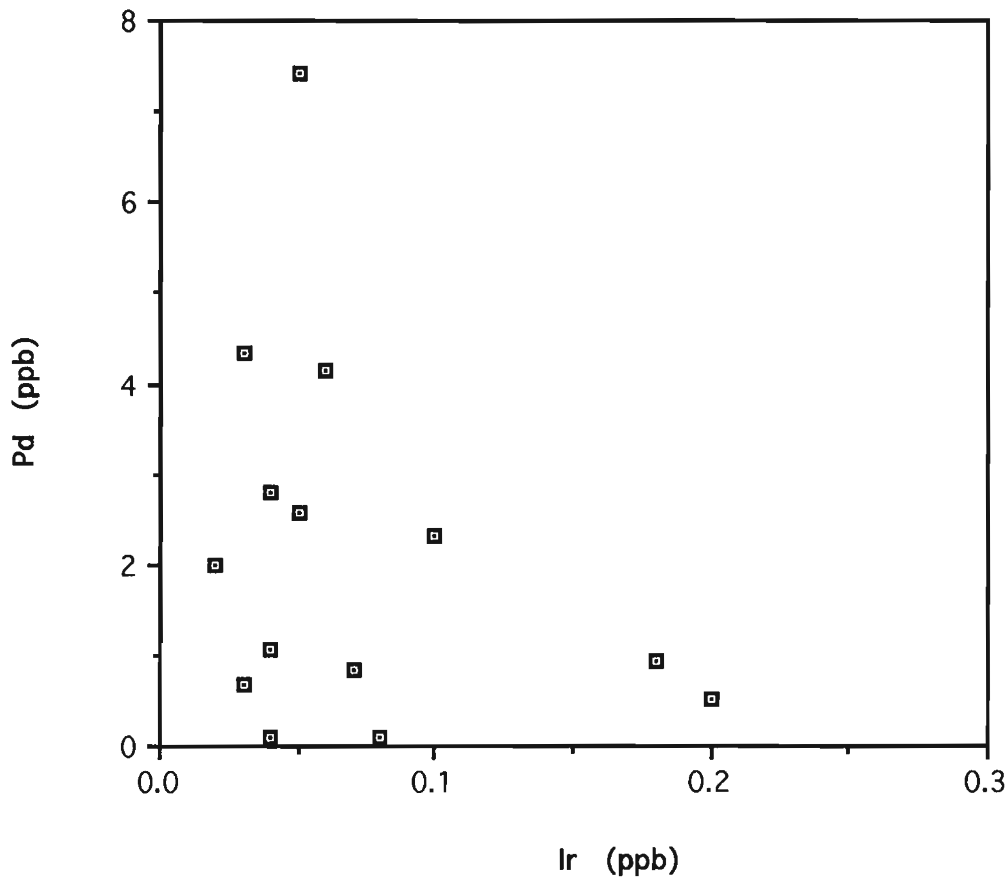

Figure 3.

Plot of Ir against Pd for the Sudbury Igneous Complex rocks, showing a crude inverse relationship.

Igneous Complex magma would be depleted in PGE related to the country rocks; 2) There are more Pt-enriched ore bodies in the Sudbury basin that can account for the Pt discrepancy.

There may be a certain contribution of PGE by the Nipissing diabase and the metavolcanic rocks if the volume of the primary melt was considerably larger than estimated. This will require more quantitative modelling of the process. Again, the Pt/Pd ratios of these rocks are greater than one, making it difficult to explain that of the Sudbury Igneous Complex rocks and sulphide ores. Moreover, there is little difference between the PGE characteristics of mantle magma and Nipissing diabase. Therefore, our tentative conclusion is that the Sudbury Igneous Complex rocks and its ores are not the result of total melting of crustal rocks.

2) The PGE content varies through the Sudbury Igneous Complex sequence. Pt and Pd decreased substantially from norite to granophyre, to much lower than in any of the country rocks. The depletion of Pt and Pd in granophyre indicates that the granophyre was either a result of fractionation of the Sudbury Igneous Complex magma or an immiscible portion of it. Furthermore, the fact that Pt and Pd contents of the granophyre are much lower than those of the country rocks indicates that during crystallization of the magma, at least at the stage when granophyre started to crystallize, there was little crustal contamination. Therefore, when the magmatic system was established after the meteorite impact, the Sudbury Igneous Complex magma probably behaved as a closed system.

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Geological Survey of Canada Project 840050-FG

Regional geology, structure, and mineral deposits of the Archean Swayze greenstone belt, southern Superior Province, Ontario[1]

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Abstract: The Swayze project area encompasses about 12 000 km^2 southwest of Timmins centred on the Swayze greenstone belt and includes a significant proportion of the surrounding granitoid terranes. The belt is the westward extension of the Abitibi greenstone belt and has high potential for mesothermal Au, volcanogenic massive sulphide Cu-Zn-Pb, mafic and ultramafic intrusion hosted Ni-Cu-PGE and komatiite-hosted Ni-Cu deposits, as well as a wide variety of industrial minerals and building stones. The program consists of reconnaissance, synoptic-style lithological and structural mapping augmented by geochemical and geochronological work, with emphasis on better understanding of the complex tectonic history of the region and its relationship to the various mineral deposit types. Preliminary results include: (a) additions and modifications to the previous geological maps both within the greenstone belt and the surrounding granitoids; (b) an interpretation of the structural chronology and its influence on the present distribution of lithologies; and (c) implications of (a) and (b) for mineral exploration.

Résumé : La région du projet Swayze englobe un secteur d'environ 12 000 km^2 au sud-ouest de Timmins, centré sur la zone de roches vertes de Swayze, qui inclut une proportion importante des terranes granitoïdes environnants. Cette zone est le prolongement vers l'ouest de la zone de roches vertes de l'Abitibi et présente un potentiel élevé quant aux minéralisations mésothermales en Au, aux gisements volcanogéniques de sulfures massifs (VMS) de Cu-Zn-Pb, aux minéralisations en Ni-Cu-ÉGP logés dans des intrusions ultramafiques, aux gisements de Ni-Cu logés dans des komatiites, et à la présence d'un grand nombre de divers minéraux à caractère industriel et divers types de pierre de construction. Le programme comprend des étapes de cartographie synoptique de reconnaissance de la lithologique et de la structure, accompagnées d'études géochimiques et géochronologiques, et doit contribuer à une meilleure connaissance de l'évolution tectonique complexe de la région et ses liens avec les divers types de gîtes minéraux. Les résultats préliminaires sont notamment: a) des ajouts et modifications aux cartes géologiques auparavant établies, à la fois dans la zone de roches vertes et dans les granitoïdes environnants; b) une interprétation de la chronologie structurale et de son influence sur la répartition actuelle des lithologies; et c) les implications de a) et de b) au niveau de la prospection minérale.

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INTRODUCTION

The Swayze Project is a bedrock mapping program being conducted by the Geological Survey of Canada under the auspices of the Northern Ontario Development Agreement (NODA) for minerals. The project area encompasses approximately 12 000 km^2 southwest of Timmins centred on the Swayze greenstone belt (SGB) and surrounding granitoid terranes (Fig. 1, 2, and 3).

Access to the area is provided by highways 101, 129, 144, and the Sultan Industrial road (Fig. 2). A large portion of the area has been logged providing an extensive network of roads and new bedrock exposures.

The project consists of 1:100 000 and 1:50 000 scale reconnaissance and synoptic style lithological and structural bedrock mapping augmented by geochemical and geochronological work. The objectives of this first field season were to: 1) become familiar with the access and logistics; 2) assess the density and distribution of outcrop; 3) provide a reconnaissance-level understanding of the major lithologies and structures; 4) assess the regional distribution of strain; 5) assess the recent assemblage subdivisions (Jackson and Fyon, 1991) within the Swayze greenstone belt; 6) identify the mineral deposit types that might be expected to occur in the area; and 7) identify geological problems to be addressed during the subsequent years of the project. Suites of metavolcanic and plutonic rocks representing the major lithological packages were sampled for whole-rock geochemistry. In addition, a total of one felsic metavolcanic and six felsic plutonic rock samples were selected for preliminary U-Pb zircon geochronology investigations. Field results were entered into the Fieldlog digital database and plotted in Autocad. The overall objective of the project is to provide a systematic, regional update of the geology and structure of the Swayze greenstone belt in the context of assessing the mineral potential and aiding mineral exploration.

GENERAL GEOLOGY

The Swayze greenstone belt is located within the western Abitibi Subprovince of the Superior Province, a Neo-Archean granitoid-greenstone terrane that developed between 2.8 and 2.6 Ga. (Jackson and Fyon, 1991). The belt is bounded to the: a) west by the Kapuskasing Structural Zone; b) east by the Kenogamissi Batholith; c) north by the Nat River granitoid complex; and (d) south by the Ramsey-Algoma granitoid complex. The belt is connected to the Abitibi greenstone belt by narrow septa of metavolcanic-metasedimentary rocks that wrap around the north and south margins of the Kenogamissi Batholith (Fig. 1 and 3).

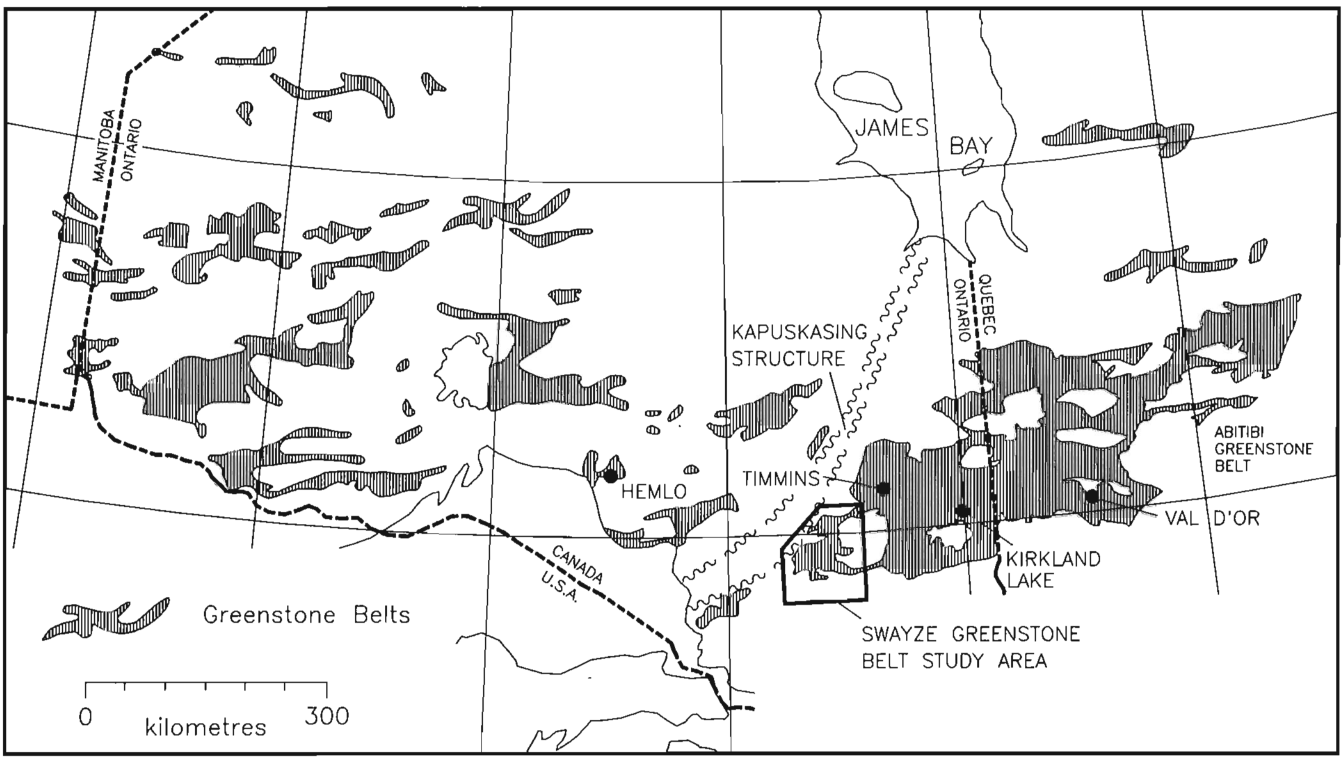

Figure 1. Regional setting of the Swayze greenstone belt study area.

Supracrustal rocks

Ultramafic metavolcanic rocks

The distribution of extrusive ultramafic metavolcanic rocks in the Swayze greenstone belt is poorly documented on published geological maps. Much of the early mapping (e.g., Rickaby, 1935; Bannerman, 1933) did not involve whole-rock geochemistry and during more recent mapping (e.g., Donovan, 1968; Milne, 1972) most of the ultramafic rocks were interpreted as intrusions. Recently, Ayer and Puumala (1991) and Ayer and Theriault (in press) have documented massive peridotitic and dunitic intrusions and spinifex textured komatiite flows in Foleyet, Ivanhoe, and Keith townships (Fig. 2 and 3).

The sequence of ultramafic metavolcanic rocks in Heenan and Newton townships (Fig. 2 and 3) , within the Halcrow-Swayze assemblage of Jackson and Fyon (1991), were first recognized by Bateman (1976) and subsequently mapped by Innes (1977) and Cattell (1985). Current mapping has traced this package of mafic and ultramafic rocks

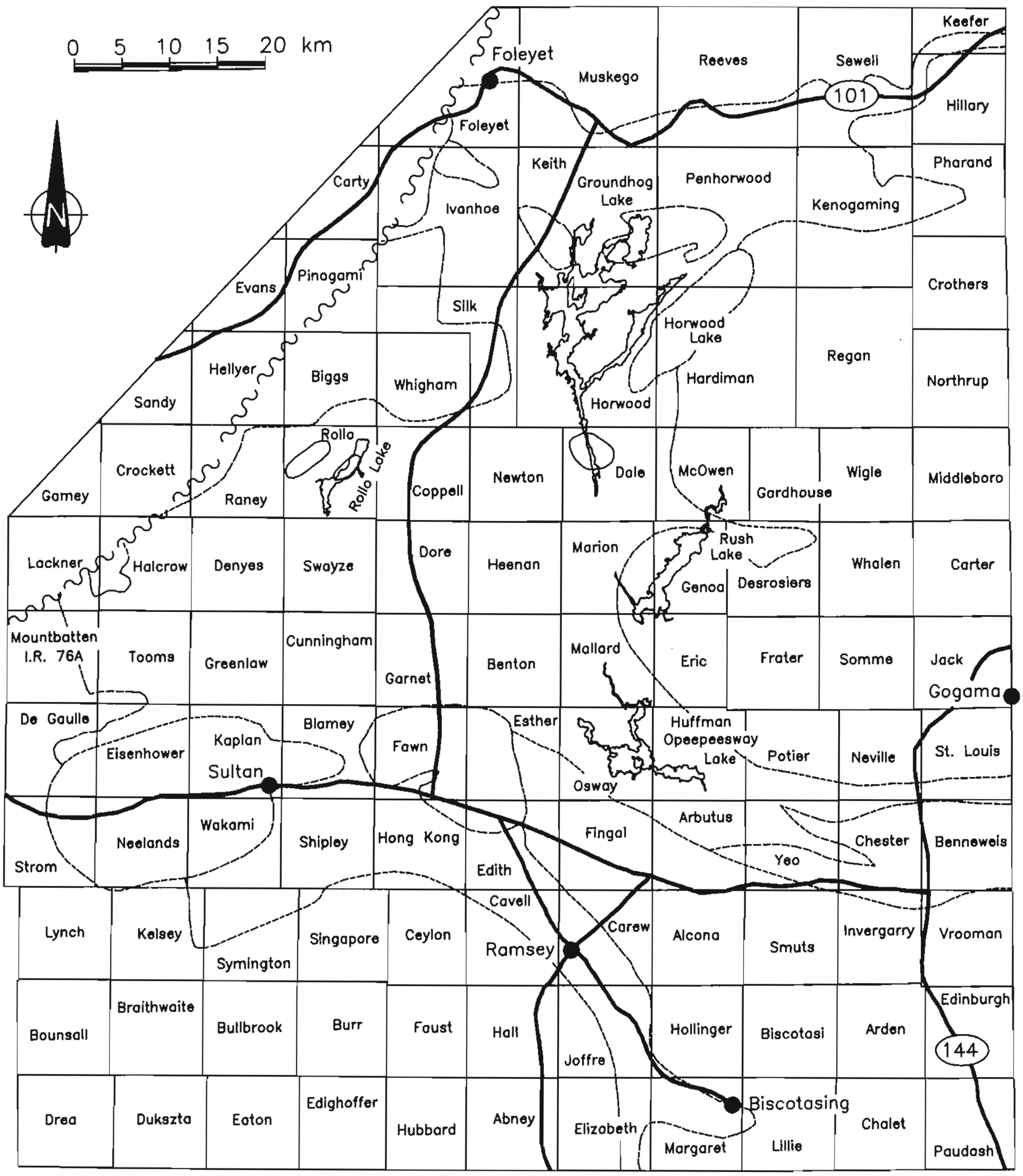

Figure 2. Townships, major lakes, and roads (heavy black lines) for the Swayze project area. The distribution of supracrustal rocks and major intrusions is indicated by the fine dashed lines. This figure is to be used in conjunction with Figure 3.

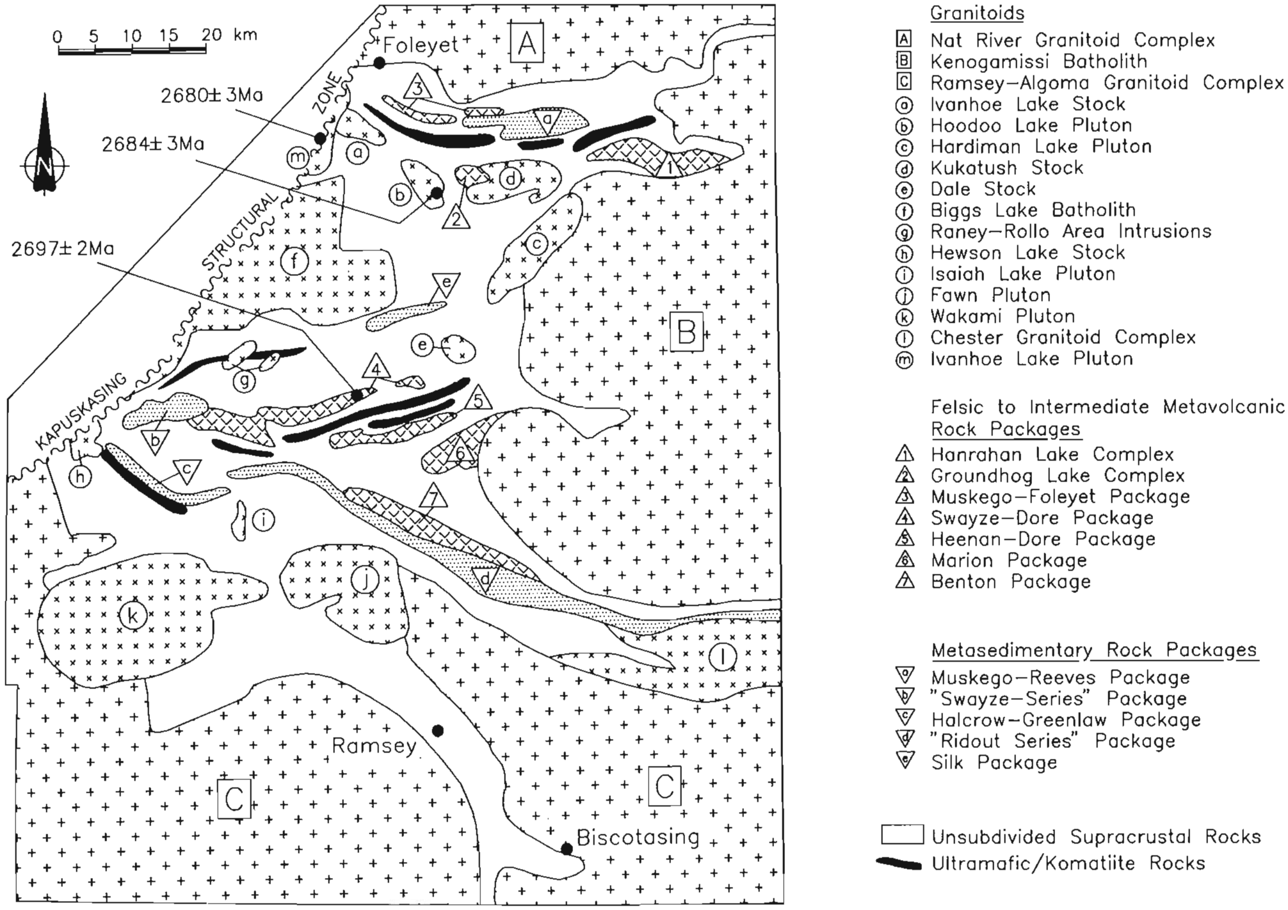

Figure 3. Simplified geology of the Swayze project area. Only the major rock packages and intrusions are depicted. Late diabase dykes have been omitted for clarity. This figure is to be used in conjunction with Figure 2.

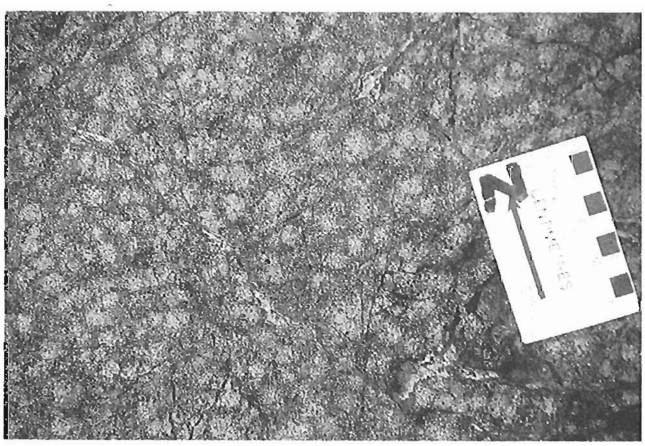

Figure 4. Glomeroporphyritic feldspars within a layered tholeiitic intrusion spatially associated with komatiite and Mg-tholeiite metavolcanic rocks, northern Dore Township. Scale card indicates north. GSC 1992-290B

westward into eastern Swayze Township (Fig. 2 and 3) immediately northwest of Pepperbell Lake. In northern Dore Township the package is dominated by distinctive orange-brown-weathering, massive and pillowed mafic metavolcanic rocks and medium- to coarse-grained diorite and gabbro intrusions. The massive flows have crudely developed polyhedral cooling joints, while the pillowed flows tend to consist of large (up to several metres in size) mattress- and balloon-shaped pillows with no tails. Vesicular and/or varolitic pillows are common, as is hyaloclastic breccia in the pillow interstices. Locally, the pillowed flows grade into spectacular pillow flow-top breccias. The diorite and gabbro intrusions are crudely layered, ranging from fine- to medium-grained and equigranular to coarse grained and locally feldspar glomeroporphyritic (Fig. 4). Intercalated with these flow and intrusive rocks are narrow horizons of polysutured and spinifex-textured, komatiitic metavolcanic rocks.

A new occurrence of ultramafic metavolcanic rocks was found in central Raney Township (Fig. 2 and 3), approximately 3 km west of Ridley Lake and immediately south of Biggs Creek. The rocks in this area are part of the Horwood assemblage which does not have any documented occurrences of extrusive ultramafic rocks (Jackson and Fyon, 1991). The rocks are massive to polysutured flows that have poorly developed spinifex textured (?) zones. Strong serpentinization and magnetite alteration imparts a black colour to the fresh rock, while the surface weathers a distinctive orange-brown. The rocks are strongly magnetic with magnetite occurring as both pervasive alteration and infilling of the polyhedral joints. These ultramafic rocks have a corresponding magnetic high anomaly (Ontario Geological Survey, 1982b) which is one of a series of semicontinuous magnetic-high anomalies that can be traced for nearly 20 km (Ontario Geological Survey, 1982a, b and c), from west of Raney Lake in southwestern Raney Township eastward to north of Rollo Lake in northeastern Rollo Township. Particularly interesting, in terms of komatiite-hosted Ni-Cu mineralization, are several two-channel electromagnetic anomalies with coincident magnetic anomalies (Ontario Geological Survey, 1982a, b) found in the vicinity of these strongly altered ultramafic rocks.

Mafic and intermediate metavolcanic rocks

Mafic metavolcanic rocks are widely distributed throughout the Swayze greenstone belt and include Fe-tholeiitic, Mg-tholeiitic, and calc-alkalic basalts (Fig. 2 and 3) which consist of massive, pillowed, pillow breccia, variolitic and amygdaloidal flows, as well as fine- to medium-grained sills and dykes.

During the current mapping program an attempt was made to distinguish the intermediate (i.e., dacite to andesite) metavolcanic rocks from the mafic and felsic metavolcanic rocks. Some of the mafic metavolcanic packages mapped by previous workers are composed predominantly of intermediate metavolcanic rocks. For example, in northeastern Garnet and south-central Dore townships there is a package of intermediate, massive and pillowed flows intercalated with volcanic breccias and lapilli tuffs. The massive flows are typically feldspar-phyric and are grey-green on the fresh surface. The volcanic breccias and lapilli tuffs contain heterolithic fragments and abundant feldspar crystals (not quartz) both within the fragments and the groundmass.

Intermediate metavolcanic rocks are also common in the northern part of the Swayze greenstone belt (Ayer and Puumala, 1991; Ayer and Theriault, in press). During the current study, intermediate pyroclastic rocks were mapped immediately north of the Kenogamissi Batholith in Kenogaming Township (Fig. 2 and 3). In the southern portion of the belt, intermediate flows and pyroclastic rocks were encountered in each of Cunningham, Arbutus, Yeo, and Chester townships.

Felsic metavolcanic rocks

There are several large sequences of felsic to intermediate metavolcanic rocks within the belt (Fig. 2 and 3). The Swayze-Dore package, one of the largest in the belt (Fig. 2 and 3), consists of felsic metavolcanic rocks intercalated with clastic metasedimentary rocks of the "Swayze Series" (discussed below) and intermediate metavolcanic rocks. Rocks of intermediate composition (i.e., andesite and dacite) are common, while rhyolitic rocks are less abundant. Feldspar±quartz porphyritic flow and intrusive rocks are common, as are ash-tuffs, lapilli tuffs, and volcanic breccias (both monolithic and heterolithic varieties). A felsic volcanoclastic rock (interpreted by Cattell et al., 1984 as a lapilli tuff) from this package of rocks yielded a U-Pb zircon age of 2697 ± 3 Ma. (Cattell et al., 1984).

The Heenan-Dore (Fig. 2 and 3) package consists of massive, feldspar-phyric flows and flow breccias, feldspar-quartz-bearing tuffs, rhyolitic feldspar-quartz porphyries (FQP) and associated lapilli tuffs and volcanic breccias. In general, the Heenan-Dore package contains more felsic rocks and less intermediate ones than the Swayze-Dore package. In addition, no metasedimentary rocks, equivalent to the "Swayze Series", are associated with the felsic rocks of the Heenan-Dore package. A sample of massive, rhyolitic feldspar-quartz porphyry, spatially associated with volcanic breccias and lapilli tuffs (Fig. 5), was collected from central Heenan Township for U-Pb zircon geochronology.

Clastic metasedimentary rocks

Historically, the clastic metasedimentary rocks have been subdivided into two major types: 1) older sequences associated and intercalated with the metavolcanic rocks; and 2) younger sequences, referred to as the "Ridout Series", unconformably overlying the older metavolcanic and metasedimentary rocks.

The older metasedimentary rocks are distributed throughout the Swayze greenstone belt and have a close temporal and tectonic relationship to the metavolcanic rocks. They range from narrow interflow sediments to larger, more extensive packages of mixed metasedimentary rocks. Several large packages of metasedimentary rocks have been document in the belt and will be informally referred to as: a) the

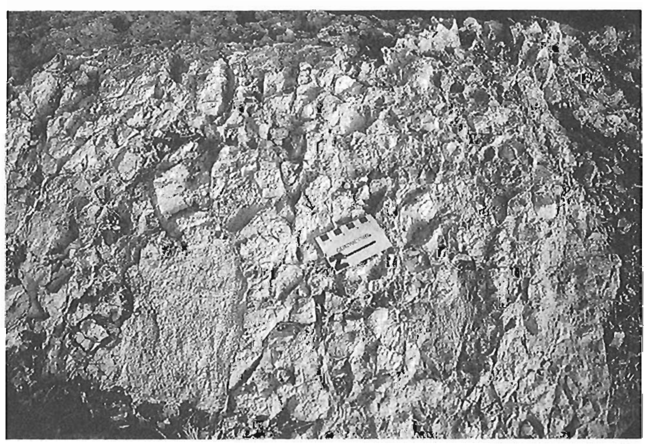

Figure 5. Weakly deformed, monolithic rhyolite volcanic breccia exhibiting subangular to rounded fragments of quartz-feldspar phyric porphyry. From the Heenan-Dore felsic metavolcanic rock package, Heenan Township. A sample of massive, feldspar-quartz porphyry adjacent to this fragmental unit was collected for U-Pb zircon geochronology. Scale card indicates north. GSC 1992-290D

Muskego-Reeves package; b) the "Swayze Series" (Emmons and Thomson, 1929); c) the Halcrow-Greenlaw package; and e) the Silk package (Fig. 2 and 3).

The Muskego-Reeves package is located within the northern Swayze greenstone belt (Fig. 2 and 3) and consists of two major and several minor metasedimentary units intercalated with the metavolcanic rocks (Ayer and Puumala, 1991; Ayer and Theriault, in press). These units consist of polymictic conglomerates, sandstones, siltstones, and argillites. Cleavage-bedding reversals, small-scale folds and Ramsey-type 3 refolded folds indicate that this is not a simple, homoclinal sequence of metasedimentary rocks.

The "Swayze Series" rocks (Fig. 2 and 3) are intimately intercalated with the Swayze-Dore package of felsic metavolcanic rocks. Many of these felsic metavolcanic rocks resemble felsic-derived metasedimentary rocks and may be more volcanoclastic than volcanic.

The Silk package is a previously unmapped, 1.5 km wide unit, of intercalated wackes, argillites, arkosic siltstones, and minor felsic to intermediate tuffs in southern Silk Township (Fig. 2 and 3). These metasedimentary rocks are finely laminated, however isoclinal folding and transposition have obliterated the primary sedimentary features. A distinctive fragmental or clastic rock consisting of poorly sorted, angular to subrounded, siliceous clasts in a quartz- and feldspar-rich, sandy matrix occurs on the southern side of the Silk metasedimentary package. The clasts are up to 1.5 cm in size and consist of: massive, fine grained, white and grey rhyolite; feldspar- and quartz-phyric rhyolite; and rare dark grey to black siliceous argillite (?). This rock may be an immature volcanoclastic sediment proximal to a felsic volcanic centre. An identical rock type occurs within the Swayze-Dore felsic metavolcanic package on the north shore of Crossley Lake, near the Coppell-Dore township boundary (Fig. 2 and 3). The Silk metasedimentary package is similar to the "Swayze Series" metasedimentary rocks which are closely associated with rocks of the Swayze-Dore felsic metavolcanic package.

The younger metasedimentary rocks referred to as the "Ridout Series" (Emmons and Thomson, 1929; Furse, 1932; Jackson and Fyon, 1991) have been interpreted as the temporal and tectonic equivalents of the Timiskaming assemblage in the Kirkland Lake area (Meen, 1944; Moorhouse, 1951; Jackson and Fyon, 1991). The "Ridout Series" is mappable (Rickaby, 1935; Meen, 1944; Laird, 1932, 1936) over a distance in excess of 150 km in the southern Swayze greenstone belt and consists of intercalated polymictic conglomerates, sandstones, siltstones, and minor argillites (Fig. 2 and 3). Two distinctive features of the "Ridout Series" metasedimentary rocks are the preponderance of granitoid and iron-formation clasts within the polymictic conglomerates and the well developed trough crossbedding within the sandstones. A wide variety of clast types representing the majority of rock units within the belt are found within the polymictic conglomerates. Many of the granitoid clasts are foliated biotite tonalite similar to that seen within the Kenogamissi Batholith and the Ramsey-Algoma granitoid complex (Fig. 2 and 3). Most of the clasts are typically flattened parallel to a strong foliation (i.e., axial planar cleavage to schistosity), however many of the granitoid clasts still retain a round shape. The foliation within the tonalite clasts is at different orientations relative to the strongly developed schistosity in the matrix surrounding the clasts which suggests that the tonalites were foliated prior to their incorporation. Intercalated with the clastic metasedimentary rocks are peculiar fragmental textured rocks that may represent: a) intermediate to felsic lapilli tuffs and volcanic breccias; b) volcanoclastic sediments; or c) folded and strongly transposed bedded sediments.

The "Ridout Series" rocks are structurally modified as evidenced by: a) numerous reversals in cleavage-bedding relationships from clockwise to counterclockwise, across the strike of the "Ridout Series"; b) reversals in facing direction, from north to south, as determined from grain-size gradition and trough crossbedding; c) east- and west-facing fold closures with a strong east-southeast-striking axial planar cleavage; and d) varing degrees of transposition along the east-southeast-striking axial planar cleavage (Fig. 6). An important characteristic of Timiskaming-like metasedimentary rock sequences is their unconformable relation to the metavolcanic and older metasedimentary rock sequences. In the Swayze greenstone belt, an unconformable relationship between the "Ridout Series" metasedimentary rocks and the metavolcanic rocks to the north and south has not been unequivocally demonstrated. Superimposed deformation has produced high-strain zones along the north and south contacts of the "Ridout Series" which have likely obliterated any primary stratigraphic or unconformable relationships.

Figure 6. Serrated and interdigitated sandstone-conglomerate contact resulting from strong folding and transposition, within "Ridout Series" metasedimentary rocks, Yeo Township. The transposition is occurring along a pronounced, east-southeast-striking schistosity/cleavage (S_c) that is axial planar to isoclinal folds seen elsewhere in the area. Scale card indicates north and the blue pencil is parallel to the strong schistosity. GSC 1992-290I

Granitoid rocks

The large granitoid terranes bounding the Swayze greenstone belt (Fig. 2 and 3) consist of a complex, yet systematic sequence of intrusions including early foliated, hornblende and biotite tonalites that are cut by massive, nonfoliated biotite granites and granodiorites (Algoman granites). A compositionally diverse suite of granitoid batholiths and stocks, ranging from foliated to massive and nonfoliated, occur within the Swayze greenstone belt supracrustal rocks.

Nat River granitoid complex

Poorly exposed granitoid rocks north of the Swayze greenstone belt (Fig. 1 and 3) will be informally referred to as the Nat River granitoid complex which consists of foliated hornblende-biotite tonalite to granodiorite and associated foliated aplite dykes. The percentage of hornblende and biotite varies from 5 to 15% with hornblende being twice as abundant as biotite. Massive to weakly foliated biotite granodiorite contains occasional, small inclusions of this foliated, hornblende±biotite tonalite. Weakly foliated pegmatite and aplite dykes cut the biotite granodiorite. Locally, massive, nonfoliated granite dykes (Algoman) cut the biotite granodiorite and the pegmatite and aplite dykes.

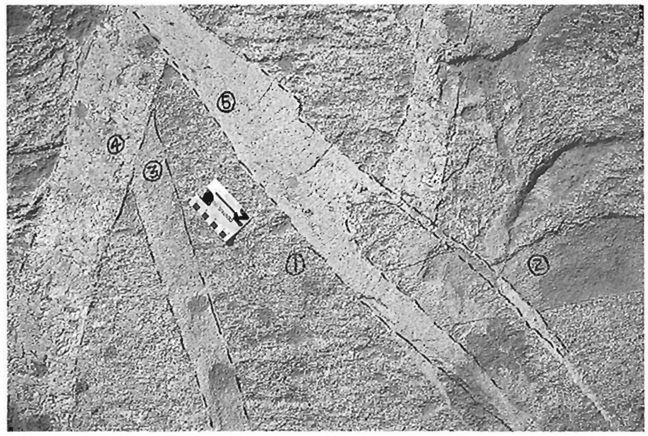

Figure 7. One of the many outcrops exhibiting excellent crosscutting relationships between different intrusive phases within the Kenogamissi Batholith, Crothers Township. Here an early, foliated biotite granodiorite (1) is cut by narrow, foliated salt-and-pepper textured biotite granodiorite dykes (2). Aplitic dykes (3) cut rock types (1) and (2) and are cut and weakly buckled by an east-southeast-striking foliation. Narrow, white pegmatite dykes (4) crosscut all of the earlier dyke types. Late, nonfoliated quartz-feldspar-bearing aplite dykes (5) cut and locally offset all of the previous intrusive phases. Scale card indicates north. GSC 1992-290H

Kenogamissi Batholith

The Kenogamissi Batholith is a large (>4000 km), elliptical granitoid complex that separates the Swayze greenstone belt from the Abitibi greenstone belt (Fig. 1 and 3). Preliminary reconnaissance mapping has documented a complex, yet systematic sequence of crosscutting intrusive rock phases (e.g., Fig. 7). From oldest to youngest these phases are: a) remnant xenoliths of foliated mafic amphibolite; b) foliated hornblende diorite to monzodiorite and hornblende monzonite to quartz monzonite; c) foliated hornblende± biotite tonalite to granodiorite and associated pegmatite and aplite dykes; d) foliated biotite tonalite to granodiorite and associated pegmatite and aplite dykes; e) foliated mafic gabbroic dykes; f) massive to foliated, potassium feldspar megacrystic, hornblende granodiorite to granite, and associated dykes; and g) massive, nonfoliated, biotite granite to granodiorite and associated pegmatite and aplite dykes. Representative samples of rock types c), d), f), and g) were selected for U-Pb zircon geochronology investigations.

Ramsey-Algoma granitoid complex

The Ramsey-Algoma granitoid complex (Jackson and Fyon, 1991) is one of the largest areas of granitoid rocks in the southern Superior Province. Card (1979) subdivided this region into a northern Ramsey Gneiss Domain, a central Algoma Plutonic Domain and a southern Levack and Algoma Gneiss domains. The southern portion of the current study area includes parts of the Ramsey Gneiss and Algoma Plutonic domains. During the current study a complex chronology of crosscutting igneous rock types was documented within the Ramsey-Algoma granitoid complex, similar to that described for the Kenogamissi Batholith. From oldest to youngest these rock types are: a) remnant xenoliths of mafic amphibolite; b) foliated hornblende± biotite tonalite; c) gneissic to foliated biotite tonalite; d) foliated biotite tonalite to granodiorite and associated aplite and pegmatite dykes; e) massive to foliated mafic dykes; and f) massive, nonfoliated biotite granite to granodiorite and associated pegmatite and aplite dykes (Fig. 8).

Granitoids internal to the supracrustal rocks

Several discrete plutons and stocks of widely variable composition have intruded the supracrustal rocks (Fig. 2 and 3).

Biggs Lake Batholith

The large granitoid batholith occupying southeast Hellyer, southeast Pinogami, southwest Ivanhoe, Biggs, Whigham, and Silk townships will be formally referred to as the Biggs Lake Batholith (Fig. 2 and 3). In Biggs and Silk township the Biggs Lake Batholith consists of massive, nonfoliated, medium grained, equigranular, biotite granodiorite with occasional small inclusions/autoliths of hornblende diorite to quartz diorite. The rock contains 5 to 10% biotite and approximately 1% disseminated magnetite and locally pyrite. The mafic metavolcanic rocks on the southern margin of the batholith are amphibolitized.

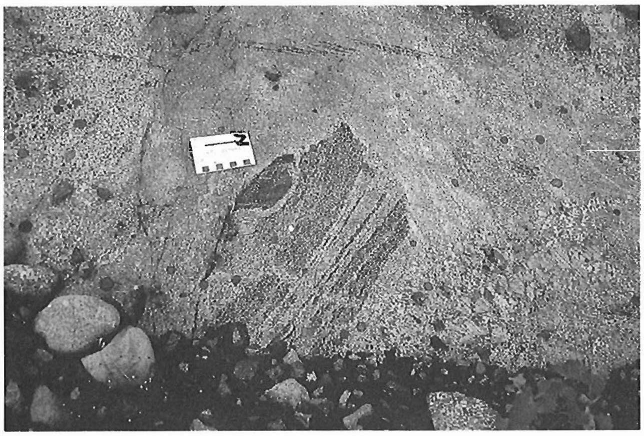

Figure 8. Ramsey-Algoma granitoid complex, Abney Township. An inclusion of foliated biotite granodiorite to tonalite with early crosscutting aplite dykes subparallel to the foliation. The inclusion is engulfed within massive, non-foliated Algoman biotite granite and its associated coarse pegmatite patches. Scale card indicates north. GSC 1992-290G

Raney and Rollo township area intrusions

Several small- to medium-sized intrusions occur in Raney and Rollo townships (Fig. 2 and 3). A small intrusion (<3 km^2) in northeastern Raney Township consists of early biotite (10%) granodiorite which is crosscut by later, massive, equigranular, nonfoliated, biotite-rich (up to 25%) granodiorite. Crosscutting these granodioritic phases are fine grained granite dykes that are related to a coarse grained, nonfoliated, hematite-stained, potassium feldspar megacrystic (up to 2 cm in size) granite which contains rare inclusions of hornblendite (up to 5 cm in size). The granodioritic phases are of similar texture and composition to those documented within the Biggs Lake Batholith and therefore may be an apophysis off that intrusion.

Rickaby (1935) mapped two granitoid intrusion; one centred on Ridley and Little Ridley lakes and the second centred on Rollo Lake. Both of these intrusions consist of strongly hematized, massive, nonfoliated, coarse grained, locally potassium feldspar megacrystic (up to 1 cm in size), biotite granite to granodiorite containing inclusions of hornblende diorite and hornblendite. There are also small isolated outcrops of massive, nonfoliated, medium grained, hornblende monzodiorite that contains inclusions of the hornblende diorite, however the relationship of these hornblende-bearing phases to the potassium feldspar megacrystic granite are not known. Perhaps the hornblendite inclusions described earlier are autoliths related to these hornblende-bearing dioritic phases. These granitoid rocks are of similar texture and composition to those documented in southern Biggs Township associated with the Biggs Lake Batholith. Recent airborne magnetic data (Ontario Geological Survey, 1982b) indicate that the Ridley Lake and Rollo Lake intrusions are physically joined by narrow apophyses.

Chester granitoid complex

Mapping in Yeo, Chester, and Benneweis townships revealed a complicated series of crosscutting intrusive phases that are informally referred to here as the Chester granitoid complex (Fig. 2 and 3). The complex is a composite, possibly layered, trondhjemite-diorite intrusion that can be subdivided into two major zones: a northern leucocratic, trondhjemite-dominated zone (Chester Township pluton of Siragusa (1981)) and a southern melocratic, diorite-dominated zone ("granite-diorite complex" of Laird, 1932; mafic migmatite of Siragusa 1981). The trondhjemitic zone consists of biotite trondhjemite to tonalite, leucocratic, quartz-rich trondhjemite, and feldspar porphyritic trondhjemite. The dioritic zone consists of hornblende diorite, hornblende quartz diorite, leucocratic diorite, and hornblende-plagioclase pegmatite. Both the trondhjemitic and dioritic phases contain up to 1% disseminated magnetite. Locally, the trondhjemite contains a chlorite-sericite-, magnetite-filled, brittle fracture stockwork.

Hornblende diorite and quartz diorite are intimately related and show a wide variety of intrusive textures, including weak magmatic layering and quartz diorite intrusion breccia with diorite fragments (Fig. 9). The fine- to coarse-grained dioritic rocks range from massive to weakly foliated. The quartz diorite is characterized by distinctive blue opalescent quartz eyes which are also found within some of the trondhjemitic rocks to the north. Hornblende-plagioclase pegmatite dykes and irregular patches, with hornblende phenocrysts up to 3 cm long, are spatially associated with the hornblende diorite and quartz diorite rocks. The fine- to coarse-grained trondhjemitic rocks are massive to locally foliated and rarely contain more than 5% biotite.

Complex and enigmatic crosscutting relationships and textures, suggestive of magma mixing, between the dioritic and trondhjemitic include: 1) abundant diorite, quartz diorite,

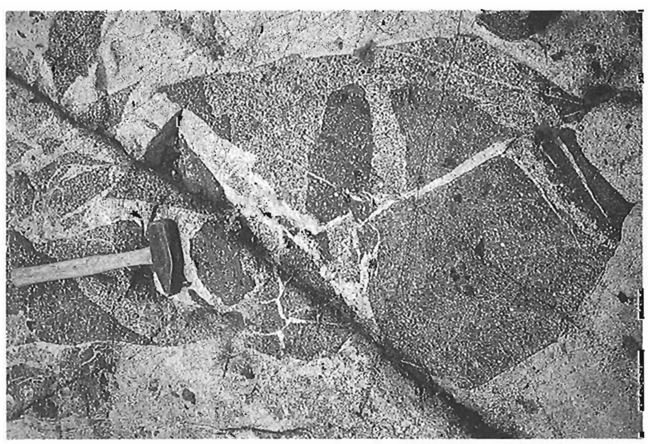

Figure 9. Complex intrusive relationships typical of the Chester trondhjemite-diorite granitoid complex, Benneweis Township. Early diorite and quartz diorite phases are included within each other and as blocks within a more trondhjemitic igneous matrix. Hammer handle indicates north. GSC 1992-290K

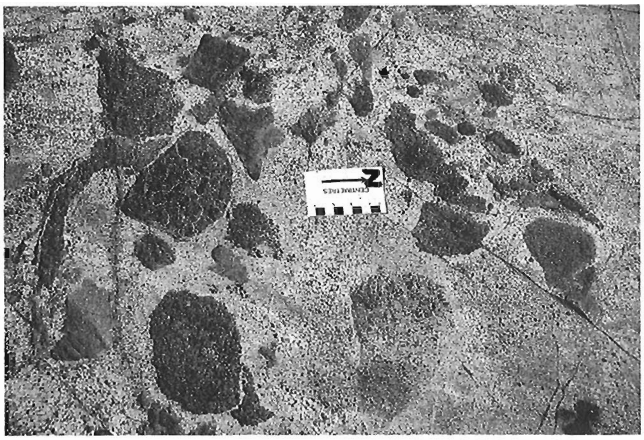

Figure 10. Diorite and quartz diorite inclusions within trondhjemite of the Chester granitoid complex, Chester Township. Note the diffuse contacts and the "frothy" and net-textured character of the dioritic inclusions. Scale card indicates north. GSC 1992-290C

and trondhjemite inclusions of variable size (<1 cm up to tens of metres) and texture (i.e., fine- to coarse-grain size) within the trondhjemite (Fig. 9 and 10); 2) inclusions are rounded, subrounded, and angular with both sharp and diffuse contacts with the host rock (Fig. 9 and 10); 3) many of the inclusions have a frothy or pitted net-texture (Fig. 10); 4) a glomeroporphyritic-feldspar diorite dyke with a 30 cm wide chilled margin adjacent to the trondhjemite in one location, is crosscut and included within the same trondhjemite a short distance away. Reconnaissance mapping has documented similar rocks as far west as Arbutus Township and as far east as Champagne Township (J. Ireland, pers. comm., 1992), a strike extent of over 50 km (Fig. 2 and 3).

STRUCTURAL GEOLOGY

Important advances in the understanding of the structural geology were achieved. A chronology of overprinting foliations and folds was documented that suggests that the current distribution of lithologies is the result of a complex interplay of polyphase folding (i.e., refolded folds), regional transposition, and faulting. The unraveling of this complex structural history, in conjunction with U-Pb geochronology and whole-rock geochemistry, is critical to reconstructing the stratigraphy and understanding where favourable packages of rocks, such as Cu-Zn bearing felsic metavolcanics, Ni-Cu bearing komatiitic metavolcanics, or gold-bearing structures, may occur.

Foliations, folds, and lineations

Historically, most of the geological maps for the Swayze greenstone belt area have depicted only one generation of foliation and it has been assumed that this foliation was produced during a single deformation event. Preliminary results from the current study indicate that at least four generations of foliation, herein designated S_a, S_b, S_c, and S_d, can be documented throughout the belt. Similarly, folds that are associated with these foliations are designated as F_a, F_b, F_c, and F_d.

The earliest foliation now recognized is a schistosity, S_a, that is tightly to isoclinally folded (F_b) about a west- to west-southwest-striking foliation (S_b). The earliest foliation (S_a) may be axial planar to an early fold generation (F_a) that has yet to be unequivocally documented. The steeply-dipping S_b foliation varies from a spaced cleavage to a penetrative schistosity along which the earlier bedding (S_0) and/or foliation (S_a) are modified and locally transposed. The S_1 foliation documented by Ayer and Theriault (in press) in Muskego and Keith townships is correlative with the S_b foliation discussed in this report. The S_b foliation and transposed zones may form axial planar high-strain zones to regional F_b folds such as the northwest-plunging Hanrahan Lake antiform of Milne (1972), the Woman River anticline (antiform, this study) of Goodwin (1965) and the Dore Township syncline (synform, this study) of Donovan (1965). The small-scale F_b fold axes have highly variable plunges and azimuths which is attributed to later refolding. Small-scale F_b folds and associated S_b foliations are locally refolded by F_c folds and overprinted by an associated east-southeast-striking foliation (S_c). The steeply-dipping S_c foliation varies from a spaced cleavage to a penetrative schistosity along which the earlier foliations (i.e., S_a, S_b) and folds (i.e., F_a, F_b) are modified and locally transposed (Fig. 6). The segment of supracrustal rocks extending from eastern Garnet Township southeastward to Chester Township is dominated by S_c foliations that overprint and modify earlier foliations and folds.

All of the early foliations are locally overprinted by a northeast-striking foliation (S_d) and associated small-scale folds (F_d). The S_d foliation varies from a spaced cleavage to a pressure solution crenulation cleavage associated with Z-shaped F_d crenulation folds (Fig. 11). Unlike the earlier fold generations (i.e., F_a, F_b, and F_c), the F_d folds do not appear to be manifested as megascopic, map-scale folds. The

Figure 11. An example of the strong northeast-striking crenulation cleavage (S_d) overprinting and Z-crenulating an east-southeast-striking schistosity (S_c), Chester Township. GSC 1992-290E

F_d folds typically plunge moderately to steeply to the northeast. In addition, the S_d foliation appears to be preferentially localized within earlier schistose and well-layered rocks (i.e., finely bedded or compositionally banded) and hence its distribution, although of regional extent, is in detail very sporadic.

A subhorizontal, north-dipping crenulation cleavage ($S_?$) cuts a subvertical schistosity (S_b/S_c) within both the Slate Rock Lake deformation zone and subvertical high-strain zones in the Joburke mine area (Ayer and Theriault, in press). The northeast-striking S_d crenulation cleavage is weakly developed within both of these zones, however its chronology relative to the subhorizontal cleavage ($S_?$) could not be determined.

Intersection lineations (e.g., between S_c and S_d) are by far the most common type of lineation encountered in the Swayze greenstone belt. Mineral (e.g., feldspar), dimensional (e.g., clasts), crinkle/crenulation, shear, and slickenside lineations are also being systematically recorded, however the polyphase folding and foliation development has produced a complex distribution.

IMPLICATIONS FOR MINERAL EXPLORATION

The new discovery of serpentinized komatiite flows with coincident magnetic and electromagnetic anomalies in Raney Township offers potential for Ni-Cu sulphide mineralization. The magnetic expression of this unit has an apparent strike length of over 20 km (Ontario Geological Survey, 1982a, b). Ultramafic and mafic rocks in Heenan, Newton, Dore, and Swayze townships are similar to those described by Jackson and Fyon (1991) in the Kidd-Munro assemblage, which hosts several komatiite related Ni-Cu deposits, as well as the Kidd Creek Cu-Zn deposit. Systematic mapping and separation of felsic metavolcanic from intermediate metavolcanic rocks, supplemented with whole-rock geochemistry, will help identify favourable rhyolites, such as those in central Heenan Township, for future Cu-Zn-Pb exploration. U-Pb zircon geochronology of the felsic metavolcanic packages will allow comparisons to be made to the base-metal rich felsic metavolcanic rocks in the Abitibi greenstone belt. The Chester granitoid complex has similar characteristics to other layered mafic intrusions and hence may offer potential for Ni-Cu-PGE mineralization. Systematic documentation of the deformation intensity of each outcrop may allow delineation of regional zones of high strain, typically associated with mesothermal gold mineralization.

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Lithotectonic setting of mineralization in the Manitouwadge greenstone belt, Ontatrio: preliminary results[1]

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Zaleski, E. and Peterson, V.L., 1993: Lithotectonic setting of mineralization in the Manitouwadge greenstone belt, Ontario: preliminary results; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 307-317.

Abstract: The Cu-Zn deposits of the Manitouwadge greenstone belt lie within an upper amphibolite-facies, highly deformed remnant of supracrustal rocks near the Wawa-Quetico subprovince boundary. The belt includes mafic and felsic volcanic and volcaniclastic rocks, metasomatically altered rocks, and iron-formation. Mineralization is associated with iron-formation and alteration. Three alteration types, characterized by metamorphic assemblages, include: 1) orthoamphibole-garnet-cordierite, 2) muscovite-sillimanite or sillimanite-garnet-biotite±cordierite, and 3) hornblende-garnet±epidote±diopside. The belt has been interpreted previously as a nappe structure, subsequently refolded to its present synformal shape. Our observations of associated map-scale truncations and sheared rocks suggests that repetition of units within the belt may be due to early thrusting. Subsequent folding, during peak metamorphism, produced strong northeasterly-plunging mineral lineations and the dominant foliation. Later dextral motion resulted in the map-scale synformal shape of the belt, local kink folds, and crenulation cleavage.

Résumé : Les gisements de Cu-Zn de la zone de roches vertes de Manitouwadge se situent dans des restes intensément déformés de roches supracrustales, situées dans le sous-faciès supérieur des amphibolites, et proches de la limite de la sous-province de Wawa-Quetico. La zone englobe des roches volcaniques et volcanoclastiques mafiques et felsiques, des roches ayant subi une métasomatose, et une formation ferrifère. La minéralisation est associée à la présence de la formation ferrifère et à l'altération. Trois types d'altération, caractérisés par la présence d'assemblages métamorphiques, sont les suivants: 1) ortho-amphibole-grenat-cordiérite, 2) muscovite-sillimanite ou sillimanite-grenat-biotitecordiérite, et 3) hornblende-grenat±épidote±diopside. Selon les interprétations antérieures, la zone serait une structure de type nappe de charriage, qui aurait subi un nouvel épisode de plissement avant de prendre sa configuration actuelle de synforme. Les troncatures coïncidentes à l'échelle de la carte et les roches cisaillées observées dans la région semblent indiquer que la répétition des unités dans la zone pourrait résulter d'un chevauchement précoce. Le plissement ultérieur survenu au cours de la phase de métamorphisme maximal a généré des linéations minérales bien définies plongeant vers le nord-est et la foliation dominante. Un mouvement dextre ultérieur a donné à la zone sa configuration de synforme à l'échelle de la carte et généré, par endroits, des plis en chevrons (kink folds) et une schistosité de crénulation.

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INTRODUCTION

The Cu-Zn deposits of the Manitouwadge area, including the Geco mine, lie within a highly metamorphosed and deformed remnant of supracrustal rocks comprising the Manitouwadge greenstone belt. Within the last 20 years, most studies of the massive sulphide deposits in the area have accepted a volcanogenic origin for the mineralization. However, sulphide mineralization and associated alteration were modified by regional metamorphism at upper amphibolite grade and by multiple episodes of deformation; hence, the tectonometamorphic effects have been equally important in determining the features and setting of the rocks as they are today.

HISTORY AND PREVIOUS MAPPING

In 1931, J.E. Thomson, Ontario Department of Mines, investigated descriptions by Ojibway Indians of rusty rocks near Manitouwadge Lake, and found sulphide mineralization (Thomson, 1932; Brown, 1963). Moses Fisher, Thomson's Ojibway guide on the 1931 expedition, staked the showing in 1943, but was unable to generate interest in his claim (Pye, 1957). In 1953, Roy Barker and William Dawd noted the sulphide zones on Thomson's map and staked the showing again. Within the year, drilling intersected an ore body and Geco Mines Limited was incorporated. In the ensuing staking rush, the Willroy, Nama Creek, and Willecho deposits were discovered. The Geco and Willroy mines went into production in 1957.

The Geco mine is one of the largest base-metal deposits in Canada, having produced nearly 47 million tonnes of ore as of 1989, grading 1.87% copper, 3.83% zinc, and 46.9 g/tonne silver (H. Lockwood, pers. comm., 1992). At present, it is the only producing mine in the Manitouwadge synform and has reserves for approximately five years.

Pye (1957) and Milne (1974) mapped the main area of supracrustal rocks including the massive sulphide deposits. Regional mapping by Williams and Breaks (1990a) covered extensions of the belt and its enclosing rocks.

REGIONAL GEOLOGY

The Manitouwadge greenstone belt lies in the Archean Superior province, within the Wawa subprovince, near its northern boundary with migmatitic and granulitic metasedimentary rocks of the Quetico subprovince (Fig. 1). The regional structure of the belt is dominated by a steeply inclined northeasterly-plunging D3 synform (Williams and Breaks, 1989, 1990a, b). Thickening of the hinge region with respect to the limbs has been attributed mostly to the shallow plunge of the fold axis. Supracrustal rocks are transitional to screens and inclusion trains in the enclosing gneissic plutonic rocks (Williams and Breaks, 1989, 1990a, b). Regional mapping (Williams and Breaks, 1990a), suggests that the northern and southern limbs of the synform are folded around the Blackman Lake and Banana Lake antiforms, respectively (Fig. 1). Interpretation of the southern limb is complicated by the observation that a strong aeromagnetic anomaly, associated with the Geco mine horizon, trends east-northeasterly toward the Quetico subprovince boundary (H. Lockwood and G. Charlton, pers. comm., 1992) and does not appear to be folded southward around the Banana Lake antiform.

The Black Pic batholith forms the southern boundary of the Manitouwadge greenstone belt, and the northern boundary of the Schreiber-Hemlo greenstone belt to the southwest. Inclusion trains of supracrustal rocks suggest that the two belts may be related and were structurally dismembered (Williams et al., 1991). This interpretation is supported by the U-Pb zircon age of sericitic rhyolite from the Manitouwadge belt at 2720 ± 2 Ma (Davis, pers. comm., 1992), which is within error of the age of felsic volcanism near the Winston Lake mine (Schreiber-Hemlo belt) at 2723 ± 2 Ma (Schandl et al., 1991). The Quetico subprovince to the north has been interpreted to be an accretionary complex, somewhat younger in age than the Manitouwadge belt (Percival and Williams, 1989).

OBJECTIVES AND SCOPE

Our objective is to gain an understanding of mineralization and alteration in the Manitouwadge belt, with respect to their origin, setting, and subsequent modifications. The belt is unusual in that, although relatively small, it hosts a giant Cu-Zn deposit (Geco), and has undergone high grade regional metamorphism and a complex structural history. It presents a unique perspective for understanding the petrogenetic development of mineralized greenstone belts.

Detailed mapping (1:5000) and sampling has been started in the hinge area and along the southern limb of the synform (generalized to 1:30 000 in Fig. 2). Mapping will be extended in subsequent field seasons. We have made use of exploration maps and reports provided by Noranda Minerals Inc. (Geco Division) and Granges Inc., and field work was greatly facilitated by grids cut by these companies. Part of the southwestern map area (Fig. 2) represents a compilation of our preliminary mapping and detailed mapping by Warren Bates and Steve Masson of Granges Inc.

DESCRIPTION OF UNITS AND CONTACT RELATIONS

Supracrustal rock units

Mixed mafic and felsic gneiss

The innermost supracrustal unit (with respect to the synform) is a heterogeneous package of interlayered mafic, felsic, and leucocratic quartzofeldspathic (leucofelsic) gneiss. The generally more voluminous mafic rocks are typically dark, fine grained, well foliated hornblende-plagioclase± magnetite±quartz±cummingtonite±biotite±garnet gneisses. The northernmost exposures, west of Fox Creek (Fig. 2), include a thick package of mafic gneiss with abundant quartz veins, epidote-clinopyroxene-hornblende-plagioclase calc-silicate lenses and knots and, locally, deformed pillows. Layers rich in biotite, cummingtonite, or garnet are most common near the southern contact with altered gneiss. Intercalated felsic and leucofelsic gneiss are quartz-rich with plagioclase, biotite and, locally, magnetite, garnet, and quartz phenocrysts. West of Fox Creek, the felsic gneiss locally contains elongate hornblende-richer patches or 'clasts' with a fine mesh-like substructure.

Hornblende-plagioclase gneiss

Hornblende-plagioclase±epidote gneiss is present near the inner hinge area of the Manitouwadge synform. The rocks have a diffuse, semi-continuous layered to lenticular structure (0.5 to 30 cm thick) defined by the abundance and grain size of fine- to medium-grained plagioclase and epidote and fine- to coarse-grained hornblende.

Quartz-eye felsic gneiss

There are three mappable belts of felsic gneiss with abundant quartz phenocrysts on the southern limb and in the hinge region of the Manitouwadge synform. These are well foliated felsic to leucofelsic quartz-plagioclase-biotite gneisses, commonly with magnetite porphyroblasts. Quartz-phyric rocks grade to aphyric rocks in some belts. Preliminary petrographic observations show that, in the less strained rocks, some quartz phenocrysts have subhedral or nearly euhedral boundaries and many contain fine rutile needles. The abundance of microcline varies dramatically between and within samples, suggesting secondary processes (potassic alteration?). The gneiss commonly contains zones of garnet-hornblende±epidote±magnetite calc-silicate alteration, which is increasingly pervasive toward contacts with iron-formation.

Various types of altered breccias are present, including volcaniclastic and flow breccias, or possibly tectonic breccias, in which the matrix shows preferential calc-silicate alteration. In some cases, the brecciated appearance may be an artifact of differential alteration.

Iron-formation

Three main types of iron-formation were observed: quartz-magnetite, quartz-sulphide, and silicate facies. Quartz-magnetite iron-formation, the most voluminous of these, is characterized by layers of coarsely recrystallized quartz, typically with thinner interlayers of magnetite-grunerite±hedenbergite±garnet. The relative proportions of iron-rich phases are highly variable and even magnetite is absent in a few cases. In the vicinity of ore bodies, quartz-magnetite grades into quartz-sulphide iron-formation in which the dominant iron-rich phases are pyrite and pyrrhotite. In some cases, sulphidic iron-formation grades laterally into massive sulphide bodies (Timms and Marshall, 1959).

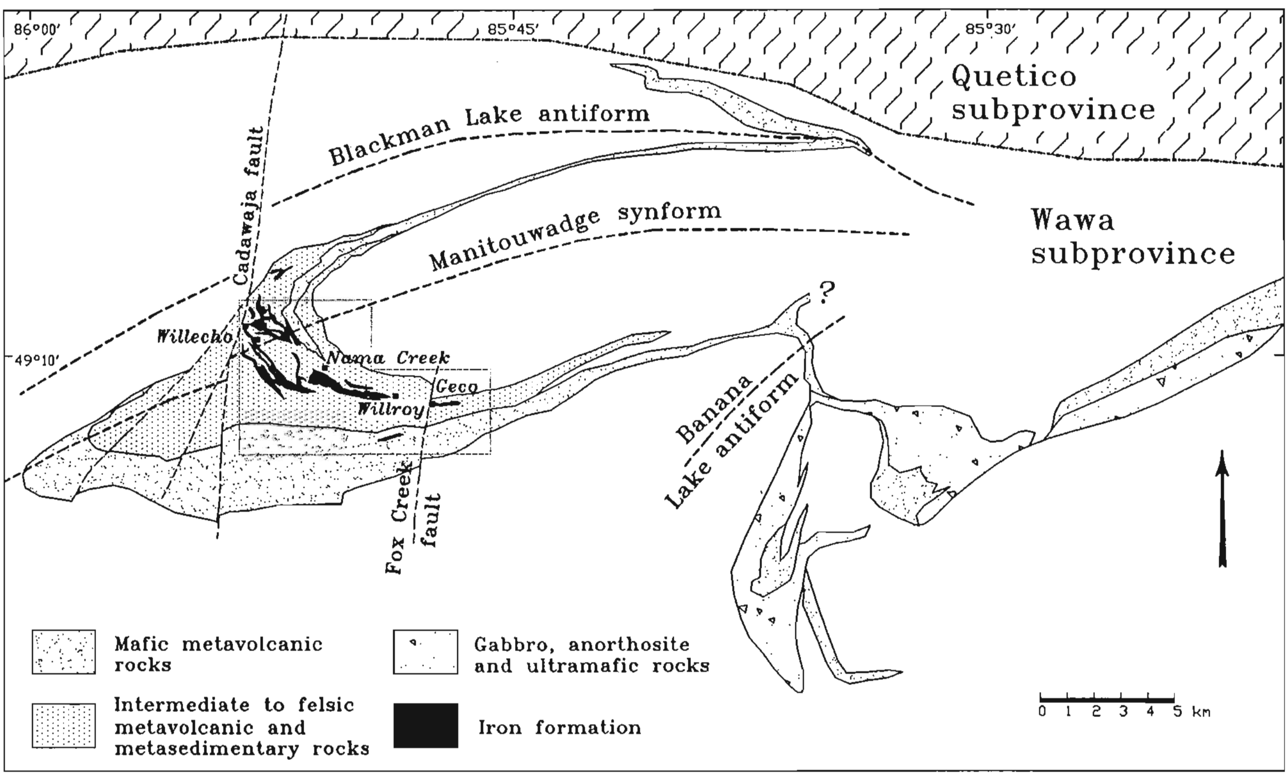

Figure 1. Regional setting of the Manitouwadge greenstone belt and location of Figure 2. Granodioritic to tonalitic gneisses are unornamented. Adapted from Williams and Breaks (1990a).

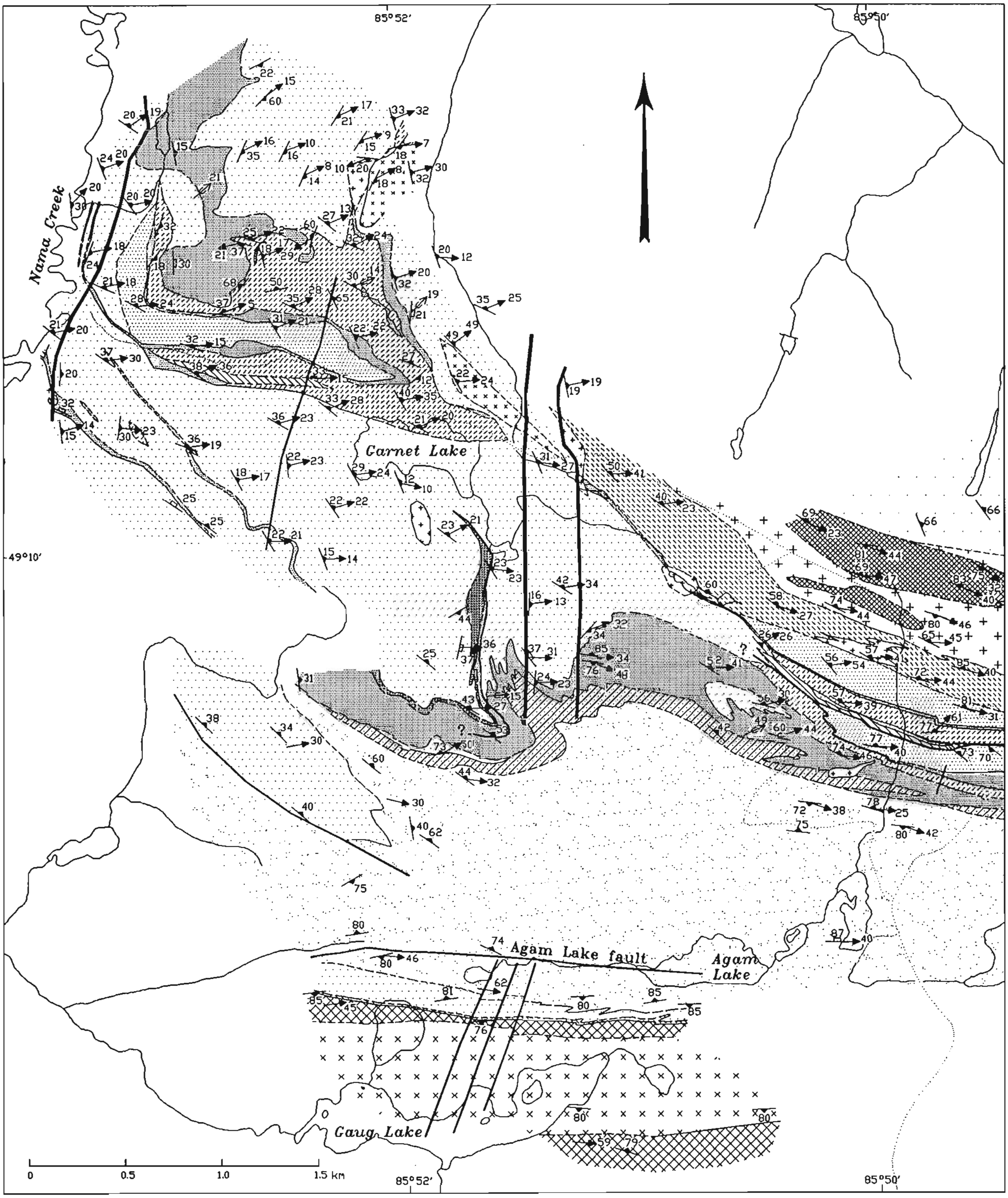

Figure 2. Geology of part of the hinge and southern limb of the Manitouwadge synform (Fig. 1), generalized from 1:5000 mapping.

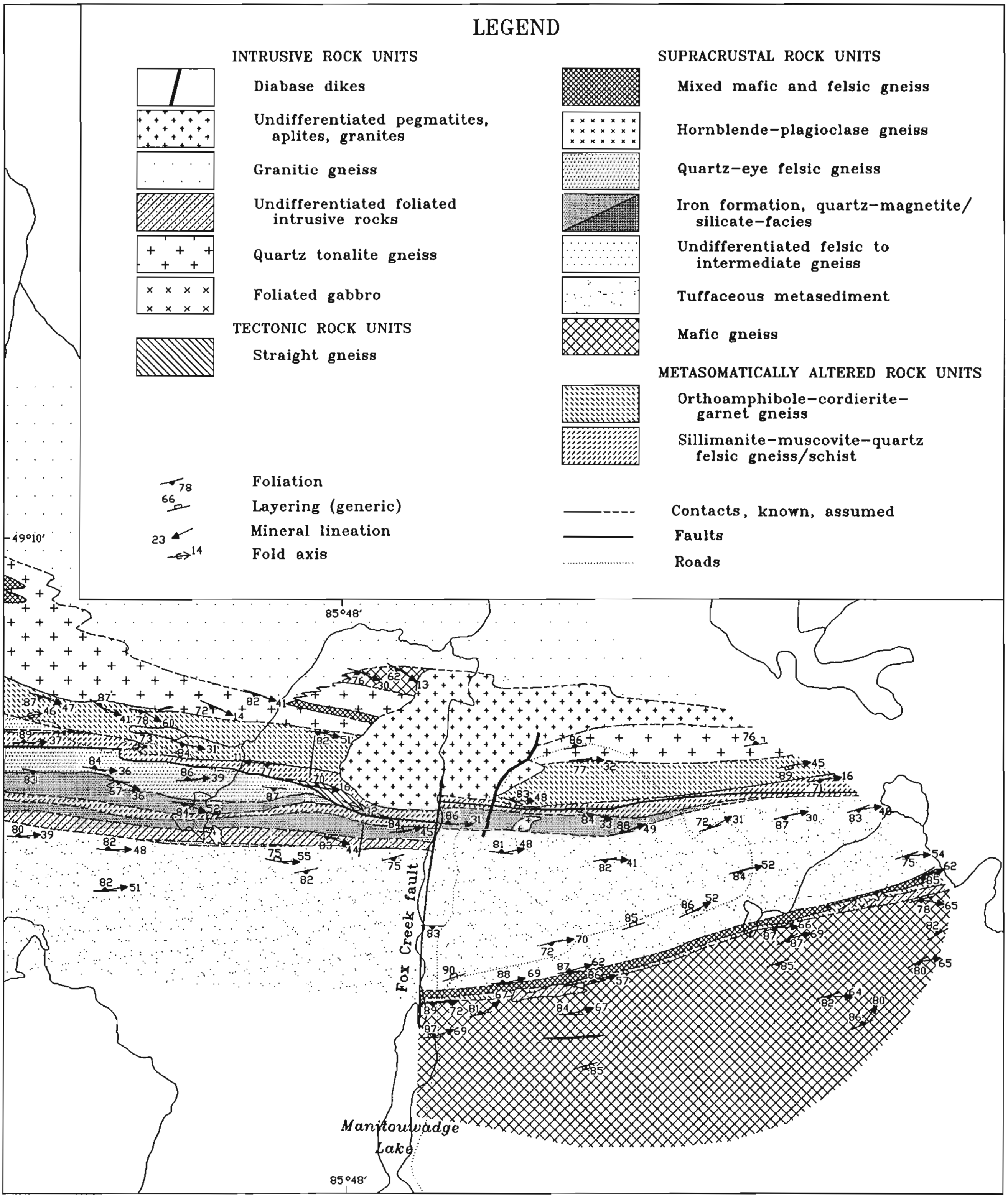

Figure 2. (cont.)

Silicate-facies iron-formation is similar to magnetite-grunerite±hedenbergite±garnet layers in quartz-magnetite iron-formation, but typically has coarser and more abundant garnet. Massive quartz is sparse or absent, except in rare fragments of quartz layers. Within quartz-magnetite iron-formation, thickened iron-rich layers in the hinge regions of folds resemble silicate-facies iron-formation. In many cases, a matrix of iron-rich phases encloses segmented quartz layers that apparently formed by tectonic brecciation (Fig. 3). Silicate-facies iron-formation may represent larger-scale thickening and remobilization of material derived from iron-rich layers in quartz-magnetite iron-formation.

Figure 3. Folding and tectonic brecciation in iron-formation. The quartz-rich fragment below the hammer preserves a fold hinge. (GSC 1992-265A)

Figure 4. Soft-sediment folding in thinly bedded tuffaceous metasediment. (GSC 1992-265B)

Along the south limb of the synform, iron-formation is exposed in four main belts. The relatively thick southernmost belt is heavily invaded by a foliated felsic tonalite, much of which cannot be shown at the scale of Figure 2. West of the Geco mine, silicate-facies iron-formation is common along the northern contact of this southernmost belt. South of Garnet Lake, silicate-facies iron-formation is relatively thick along this contact and may in part be separated from the main southern belt of iron-formation by an early backthrust (see below).

In the northwest (Fig. 2), belts of iron-formation are involved in map-scale folds or appear to be tectonically thinned, ultimately forming boudins or pinching out entirely. The iron-formation is mainly of the quartz-magnetite type. In the hinge region of the Manitouwadge synform, magnetite-grunerite layers are thick relative to quartz layers, and the grunerite is coarse grained (1 to 5 cm in length). In contrast, north of the hinge region, iron-formation is dominated by white massive quartz and is locally nonmagnetic.

Undifferentiated felsic to intermediate gneiss

This unit encompasses a variety of leucofelsic, felsic, and intermediate rock types that defy subdivision due to subtle or transitional differences in rock types, limited extent of subunits, absence of marker units, and poor exposure. In addition to unremarkable felsic gneiss, the unit includes breccias, straight gneiss, tonalitic rocks, and minor muscovite-sillimanite felsic schist. The breccias generally have aphyric felsic clasts in an intermediate to mafic matrix, containing biotite±hornblende±garnet±magnetite, similar to calc-silicate alteration found in quartz-phyric felsic rocks. The structure is typically lenticular and, in part, the brecciated appearance may be an artifact of tectonism or alteration. However, where the matrix is more felsic (less altered?), the clasts are better defined, and the breccias are more convincingly of volcaniclastic or epiclastic origin. Between Agam and Gaug lakes, interlayered fine to coarse breccias with angular to subrounded fragments (up to 30 cm) were interpreted as transitions between distal and proximal volcaniclastic deposits (W. Bates, pers. comm., 1992).

Felsic rocks from Garnet Lake west to Nama Creek consist of a complex association of fine grained leucocratic quartzofeldspathic (leucofelsic) gneiss, straight gneiss, muscovite-biotite and muscovite-sillimanite schist, and minor calc-silicate-bearing felsic rocks. Between the two layers of iron-formation west of Garnet Lake, felsic rocks locally contain zoned sillimanite knots identical to those described below.

Tuffaceous metasediment

The dominant rock type is a homogeneous grey, foliated, fine grained, granular, poorly to well layered, biotite-quartz-feldspar schist. Locally, the rock is moderately aluminous, contains garnet and/or sillimanite, and shows layering suggestive of graded bedding. In the Geco area, the southern part of the belt is locally quartz-rich with green-grey layers rich in calc-silicate minerals. Layers or sills of plagioclase-phyric, plagioclase-quartz-biotite gneiss are common throughout. There is evidence of soft-sediment deformation in isolated layers characterized by chaotic folding (Fig. 4) or by brecciation.

Mafic gneiss

This fine grained, well foliated, hornblende-plagioclase± biotite±magnetite±garnet gneiss is typically either homogeneous or finely layered with interlayers richer in plagioclase. Epidote is common in calc-silicate lenses or boudins. Flattened and slightly elongated deformed pillows can be identified locally (Fig. 5). West of Gaug Lake, W. Bates (pers. comm., 1992) interpreted northward younging from less deformed pillows.

East of Fox Creek (Fig. 2), near the northern contact with metasedimentary rocks, mafic gneiss is intimately interlayered with felsic gneiss. In the west, near Gaug Lake, there is a more abrupt transition to felsic gneiss that includes volcanic breccias (see above). A thin belt of altered gneiss is present near the transition. The altered rock includes coarse garnet-rich hornblende-biotite±magnetite gneiss and orthoamphibole-plagioclase±garnet±magnetite±hornblende gneiss, as well as some quartz-magnetite iron-formation in the west. Foliated intrusive rocks present near the altered gneiss include hornblende or biotite tonalite and granodiorite, and plagioclase-rich gabbro with hornblende phenocrysts. Strongly sheared, thinly laminated gneiss is also common near this transition.

Metasomatically altered rock units

Orthoamphibole-cordierite-garnet gneiss

This belt has been interpreted as a zone of alkali depletion and Fe-Mg enrichment in the stratigraphic footwall to the Cu-Zn mineralization (Freisen et al., 1982). It varies both along and across strike. The northern contact is transitional to mixed mafic-felsic gneiss through patchy zones of incipient alteration. Orthoamphibole-bearing gneiss is commonly intercalated with sillimanite-garnet-biotite gneiss, similar to that described below, on a scale of 0.5 to several metres. Thinner, semi-continuous quartz-rich layers (1-20 mm) are overgrown by coarse grained orthoamphibole (up to several centimetres long) and garnet (up to 3 cm, Fig. 6). Interstitial cordierite, is commonly pinitized, but also occurs in rare clear crystals up to 30 cm in diameter. Staurolite is mantled by cordierite. Local zones of coarse biotite pseudomorphs after orthoamphibole indicate potassic metasomatism during retrogression.

Sillimanite-muscovite-quartz felsic gneiss/schist

A number of distinctive rock types are grouped in this unit because of similar mineralogy, or because they are intimately interlayered or have gradational contacts. The unit is characterized by abundant muscovite and/or sillimanite and quartz, with or without plagioclase, biotite, K-feldspar, garnet, cordierite and magnetite, which are generally more abundant in sillimanite-rich gneisses.

Within the southern limb of the synform, two belts of sillimanite-rich felsic gneiss merge to the east. The sillimanite is predominantly in elongate knots that range from a few millimetres up to 1 cm in diameter. The northern belt grades easterly to quartz-muscovite-sillimanite schist composed of fine interlayers (1-5 mm) of quartz and coarse muscovite. Sillimanite typically forms fine sprays on the foliation surface. Quartz-muscovite-sillimanite schist envelopes the Geco and some Willroy ore bodies. The schists have been variously interpreted as altered metasedimentary rocks (Pye, 1957), fault zones (Milne, 1974), and altered volcanogenic rocks (Freisen et al., 1982).

Figure 5. Deformed pillows with bleached margins. (GSC 1992-265C)

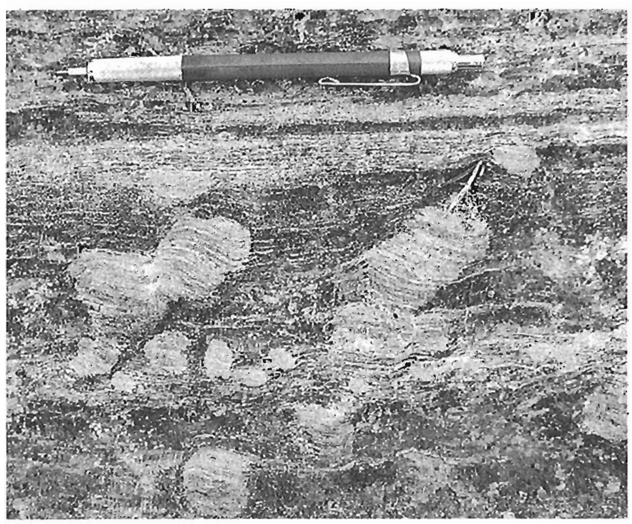

Figure 6. Quartz layers overgrown by differentially rotated garnet porphyroblasts with dextral sense of motion in orthoamphibole-cordierite-garnet gneiss. (GSC 1992-265D)

In the eastern map area, quartz-muscovite-sillimanite schist grades southward over a few metres into a sulphide-rich, muscovite-biotite quartzite. The quartzite contains streaky garnet-hornblende-rich patches and has a highly strained appearance in the field. The quartzite grades westward to a thinly layered quartz-muscovite-rich felsic gneiss, that is also sulphide-rich in the vicinity of the Geco mine. Layered felsic gneiss can be traced for several kilometres westward between the two southernmost belts of iron-formation.

In the northwest, felsic rocks with sillimanite knots are present as part of a distinctive association that includes felsic gneiss, felsic straight gneiss, and sheared pegmatite. The sillimanite knots generally range from 1 to 8 cm in diameter, varying from nearly equant in cross-section, to strongly flattened and elongated. They are zoned, with greenish or bluish cores, mantled by white fibrous sillimanite with a pinkish rim. Partial replacement by coarse grained muscovite is common. The matrix is generally composed of fine- to medium-grained quartz, plagioclase, microcline, and biotite. Garnet porphyroblasts (up to 2 cm in diameter), mantled by sillimanite, are present locally. In two northern localities, thinly bedded, quartz-phyric tuffs with minor garnet are graded from quartz- to sillimanite-rich.

Tectonic rock units

Straight gneiss

Finely laminated gneiss has been identified within the map area in isolated exposures and as semi-continuous map-scale layers or lenses, interpreted as shear zones, near the southern margin of the western and northern belts of quartz-muscovite-sillimanite schist (Fig. 2). Felsic gneiss and pegmatite appear to be the protolith to this rock, which is typically fine grained, quartz-feldspar-rich, thinly laminated on a millimetre to sub-millimetre scale, hard, and flinty. The thin continuous-to-streaky laminae are defined by segregated quartz and feldspar. Fine grained mica-rich or hornblende-rich interlayers are common at some localities. In many places, the gneiss is pinkish, possibly due to the presence of fine microcline. Preliminary petrography suggests that the gneiss has been annealed, but remains fine grained. These rocks have the characteristics of straight gneisses (Hanmer, 1988) and probably represent annealed mylonites. A transition from sillimanite-bearing pegmatite to straight gneiss is exposed along the wall of the Willecho mine pit. In this transition zone, porphyroclasts of pegmatite are enclosed by finely laminated, sheared gneiss with a strong sillimanite lineation.

Intrusive rock units

Granitic gneiss

A strongly foliated, medium- to coarse-grained granitic gneiss, with abundant coarse quartz, marks the northern boundary of supracrustal rocks in the core of the Manitouwadge synform. Locally, a finer grained equigranular pink granitic gneiss is present near the contact.

Figure 7. Folded garnet-orthoamphibole alteration in quartz tonalite. (GSC 1992-265F)

Quartz tonalite gneiss

A leucocratic tonalite gneiss invades the mixed mafic and felsic gneisses along the northern margin of the synform. It is composed of quartz (30-40%) and sodic plagioclase, with minor biotite, magnetite, and local garnet. It is typically medium- to fine-grained, commonly with coarse quartz grains, up to 5 mm in diameter. The abundance of biotite and garnet increases toward the southern contact with ortho-amphibole-cordierite-garnet gneiss, and the quartz tonalite gneiss hosts garnet-orthoamphibole alteration along narrow seams (Fig. 7).

Foliated gabbro

Near Gaug Lake, medium- to coarse-grained foliated gabbro, composed of hornblende and plagioclase with accessory biotite and magnetite, is thickly interlayered with fine grained mafic gneiss. A smaller body of foliated gabbro is present within the same mafic gneiss unit near its northern contact, east of Fox Creek.

Undifferentiated foliated intrusive rocks

Homogeneous fine- to medium-grained rocks, ranging in composition from granodiorite to diorite, intrude the supracrustal rocks as dykes, sills, and irregular bodies. The unit may include rocks of various ages. Tonalites are most common, with or without plagioclase phenocrysts, and with accessory biotite and/or hornblende. Tonalite pervades the southernmost belt of iron-formation and the northern part of the tuffaceous metasediment. In several outcrops, foliated tonalite dykes cut mylonitic layering or early folds.

Undifferentiated pegmatites, aplites, granites

This group of leucocratic, medium- to coarse-grained dykes, sheets, and intrusive bodies includes both foliated and unfoliated rocks. They are composed of quartz, K-feldspar, and plagioclase, with accessory muscovite and/or biotite. Garnet is present locally, especially in pegmatites, as a thin selvedge near contacts with iron-formation. In many cases, pegmatites show tectonic fabrics only near contacts or within their finer grained aplitic variants. As previously described, some pegmatites were involved in shearing, and sheared contacts and anastomosing shear surfaces coated by strongly aligned sillimanite are common. In contrast, there are also discordant straight pegmatite dykes and dyke swarms that show little evidence of deformation, apart from quartz-filled gashes at a high angle to the contacts.

Diabase dykes

Northerly-trending massive, fine- to medium-grained, diabase dykes of probable Proterozoic age cut across all rock units and tectonic fabrics.

STRUCTURAL GEOLOGY

In the area of the Geco and Willroy deposits, the presence of altered rocks immediately north of the mineralization, and Cu- to Zn-rich zoning within the deposits from north to south, have been interpreted as evidence for southerly younging (Suffel et al., 1971; Freisen et al., 1982). In the southwestern map area, primary features indicate northerly younging (W. Bates, pers. comm., 1992; Williams and Breaks, 1990b). The change in younging direction and a crude symmetry of rock units across the southern limb of the Manitouwadge synform led Touborg (1973) and Robinson (1979), among others, to propose that the supracrustal rocks are repeated in a synclinal nappe structure, that was subsequently refolded into a synform. Our observations do not give direct evidence bearing on the proposed early syncline.

Map- and outcrop-scale structural features can be separated into four generalized phases of deformation. Large-scale, low-angle thrusts, suggested by map-scale repetition and truncation of rock units, represent the earliest deformation (D1) and could be associated with nappe formation. A distinctive sequence of units; quartz-muscovite-sillimanite schist, iron-formation, quartz-eye felsic gneiss, is repeated across a possible thrust surface or shear zone (shown as a fault in Fig. 2) near the southern contact of the northernmost quartz-muscovite-sillimanite schist, west of Fox Creek. Iron-formation and other rock units are truncated or slivered along this contact and discontinuous belts of straight gneiss (shear zones) are present for much of its length. The geometry of the zone suggests possible westward transport during thrusting with a map-view separation of about 3 km. The same sequence of rock units is present in the hinge region of the Manitouwadge synform. Similarly, straight gneiss is intimately associated with sillimanite-knot gneiss at the base of the sequence, suggesting that the package may represent another thrust slice. Subsequent complex deformation and areas of poor exposure hinder pre-thrusting reconstruction. The finger of iron-formation extending into felsic gneiss south of Garnet Lake (Fig. 2) may result from backthrusting. This interpreted fault roots south and west into iron-formation and repeats the south-to-north sequence: quartz-magnetite iron-formation, silicate-facies iron-formation (not continuous at map scale), felsic gneiss. It is apparently east-directed with approximately 1 km of map-view separation. Fabrics associated with D1 deformation include transposed gneissic and mylonitic layering. In most areas, these are difficult to distinguish from D2 planar fabrics.

D2 structural features include the dominant mineral lineations and planar fabrics observed throughout the map area (Fig. 8), the dominant outcrop-scale folds, and many of the map-scale folds. Among the map-scale folds, is the counter-clockwise fold in iron-formation, southeast of Garnet Lake (Fig. 2). Its sense of rotation is not consistent with what would be expected for a parasitic fold related to the Manitouwadge synform. Digitations on the fold are related to changes in vergence of outcrop-scale minor folds within the iron-formation. The dominant northeasterly-plunging mineral lineation is parallel to the fold axes of the minor folds. In one area, D1 mylonitic foliation is folded, with an axial planar foliation parallel to the dominant foliation in the area. Other map-scale D2 structures include the isoclinal folds in iron-formation and mylonitic gneiss in the northwest (Fig. 2). Many of these D2 structures now have a sinistral rotational sense, but may have been west-directed prior to formation of the Manitouwadge synform. D2 folding could represent the culmination of westward transport initiated by D1 thrusting.

The dominant D2 planar fabric is typically a pervasive gneissosity or schistosity. Peak metamorphic minerals, such as sillimanite, orthoamphibole, cummingtonite, hornblende, and biotite, define the orientation of the ubiquitous D2 lineation. In many places, sillimanite and orthoamphibole are present in random sprays, commonly within the same outcrop as lineated grains. This juxtaposition of lineated and randomly oriented grains may reflect patterns of strain partitioning. It may also suggest that D2 deformation occurred near peak metamorphic conditions.

The planar fabrics define a girdle with a moderate to shallow easterly-plunging axis (Fig. 8a), generally parallel to the Manitouwadge synform axis. The apparent bimodal distribution of planar fabrics may indicate a discontinuity or rotation between the southern limb and hinge region of the synform, although we cannot discount a sampling bias due to uneven exposure. D2 mineral lineations from the hinge region have a shallower, more northeasterly plunge than those on the southern limb (Fig. 8b). It appears that the Manitouwadge synform and related major D3 folds (Fig. 1) are nearly coaxial with D2 folds. Formation of D3 folds by dextral motion is suggested by their crude map-scale asymmetry and by the pattern of aeromagnetic anomalies. Smaller map-scale folds with dextral asymmetry may also be a product of D3 deformation.

Porphyroclast tails and rotated boudins were observed in the field, but provided conflicting information on shear sense. Within the hinge region of the synform, both east-side-up and

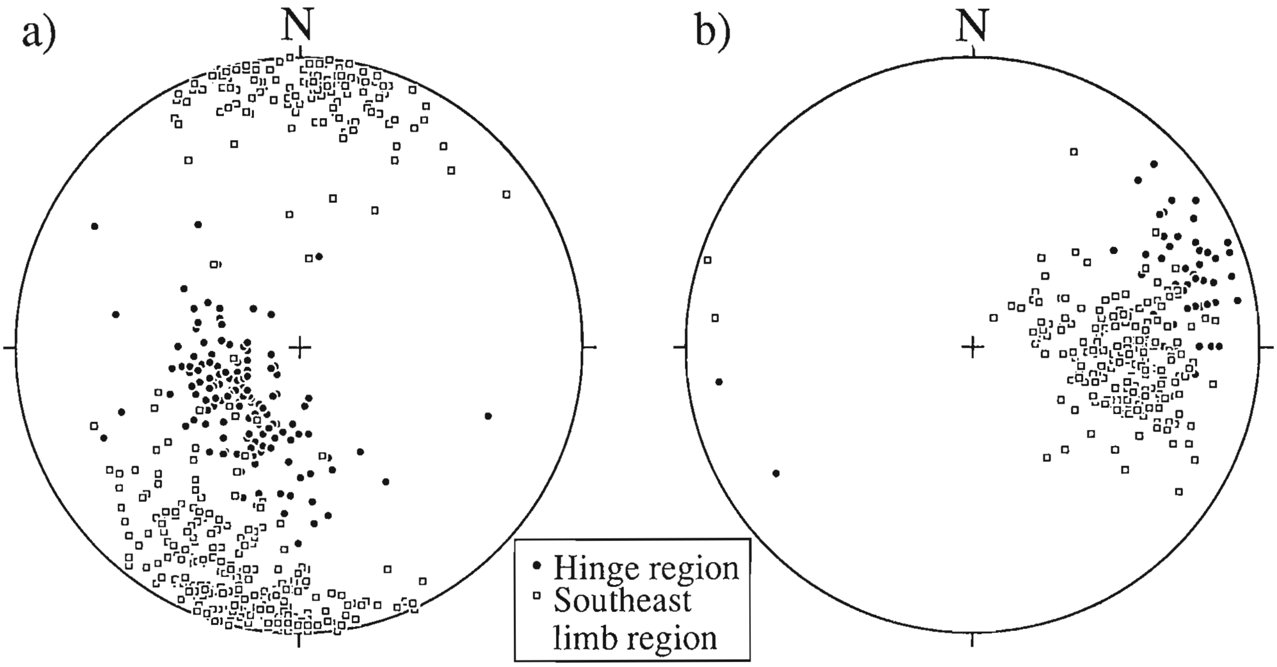

Figure 8. Equal area projections of structural data from the hinge and southeast limb of the Manitouwadge synform. **a)** Poles to S0, S1, and S2 planar data include bedding, layering, gneissosity, mylonitic layering, and foliation. **b)** Orientations of dominant (L2) mineral or stretching lineations.

east-side-down kinematics are indicated locally. Similarly, in steeply dipping rocks on the southern limb of the synform, dextral indicators are dominant (Fig. 6), but sinistral indicators are also observed. These kinematic indicators could be products of both D2 and D3 deformation.

Kink folds, crenulation cleavages, and intersection lineations are developed locally throughout the area. In general, related planar fabrics have a moderate dip, either to the northwest or southeast, and the linear features plunge either easterly or westerly. Some of these features may be related to formation of the synform, but most are interpreted to have formed later. Along the south limb, most late structural features have dextral asymmetry. In the hinge region of the synform, the westerly plunge of some L2 lineations appears to be due to rotation by kink folds that postdate the synform. More work is needed to better evaluate late-stage fabrics and kinematic indicators.

The Agam Lake and Fox Creek faults (Fig. 2) are among a number of late brittle-ductile to brittle faults, previously identified by Pye (1957), Milne (1974) and others. The Agam Lake fault forms a pronounced lineament within the tuffaceous metasediment and is associated with strong L-tectonites and focused brittle deformation, suggesting a ductile and brittle history. The Fox Creek fault shows an apparent left-lateral separation of 100 m. Outcrop-scale brittle faults with centimetre offsets may be associated with the Fox Creek and other late faults.

SUMMARY

The massive sulphide deposits in the Manitouwadge belt are generally considered to be of volcanogenic origin, as supported by their association with quartz-phyric felsic rocks and breccias, iron-formation and altered rocks. However, amphibolite-facies metamorphism, intrusive activity, and deformation have modified the textures, structures, distribution, and disposition of the mineralization and host rocks. Massive sulphides have been remobilized during deformation; for example, in the Geco mine, they are thickened in the hinge regions of folds (Friesen et al., 1982). Our preliminary evidence for early thrusting raises the possibility of structural repetition of mineralized zones, and of dismemberment in high strain zones. An understanding of this early deformation is critical to the development of structural and genetic models with predictive value for exploration.

There is a wide spectrum of metasomatically altered rocks, including orthoamphibole-garnet-cordierite gneiss and muscovite-sillimanite-quartz schist, interpreted as the product of metamorphosed synvolcanic alteration. Conversely, widespread calc-silicate alteration may have had a long history, possibly extending from synvolcanic to post-peak metamorphic. All of the altered rocks are characterized by metamorphic, or postmetamorphic minerals and, particularly the micaceous rocks, tend to focus deformation. Ambiguity in the identification of protoliths stems from the problem of distinguishing synvolcanic from synmetamorphic and later features.

Further work, using petrographic relationships, geochemical analyses and geochronology should improve our understanding of timing and relationships between volcanism, alteration, mineralization, deformation, and metamorphism. On a more regional scale, continued detailed studies within and adjacent to the Manitouwadge belt should be useful in evaluating the nature of the Quetico-Wawa subprovince boundary and its relationship to events in the synform.

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Archean unconformity in the Vizien greenstone belt, Ungava Peninsula, Quebec

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Abstract: Detailed mapping in the Vizien greenstone belt has revealed a steeply dipping angular unconformity between mafic dyke-bearing tonalitic basement of 2940 Ma age and a younger, variably foliated, middle amphibolite facies sequence of conglomerate, greywacke, and volcanic rocks with a maximum age of 2708 Ma (U-Pb zircon age from granite cobble). A regolith of quartz-muscovite grit beneath the unconformity is overlain by a sedimentary succession up to 170 m thick of lower pebble and boulder conglomerate and upper thick-bedded greywacke. The sedimentary unit is conformably overlain by a 70 m-thick mafic unit of volcanic and intrusive rocks, in turn structurally overlain by a 210 m-thick ultramafic unit. The unconformity-based package is the youngest of at least three discrete supracrustal assemblages juxtaposed in the Vizien belt that include >2940 Ma enclaves in tonalitic basement and 2724 Ma rhyolite in a mafic-to-felsic sequence.

Résumé : La cartographie détaillée de la zone de roches vertes de Vizien a révélé l'existence d'une discordance angulaire fortement inclinée entre le socle tonalitique, âgé de 2 940 Ma et contenant des dykes mafiques, et une séquence plus jeune, plus ou moins feuilletée, située dans le faciès moyen des amphibolites, et composée de conglomérat, de grauwacke et de roches volcaniques qui ont un âge maximum de 2 708 Ma (datation U-Pb sur le zircon provenant de galets de granite). Un régolithe composé de grès grossier à quartz et muscovite, et situé au-dessous de la discordance, est recouvert par une succession sédimentaire pouvant atteindre 170 m d'épaisseur, constituée d'un conglomérat inférieur à galets et blocs et, au-dessus, d'une grauwacke formant des lits épais. L'unité sédimentaire est recouverte en concordance par une unité mafique de roches volcaniques et intrusives de 70 m d'épaisseur, elle-même recouverte structuralement par une unité ultramafique de 210 m d'épaisseur. L'ensemble dont la base est la discordance est le plus récent d'au moins trois assemblages supracrustaux discrets juxtaposés dans la zone de Vizien et englobant des enclaves > 2 940 Ma dans un socle tonalitique et une rhyolite de 2 724 Ma dans une séquence mafique à felsique.

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INTRODUCTION

The Vizien greenstone belt (Percival and Card, 1992a) was discovered in 1991 during reconnaissance mapping of a 100 by 400 km transect across the Ungava Peninsula (Percival et al., 1990, 1991). This mapping served to define the outline of the belt, and provided preliminary information on its lithological and structural character and mineral potential (Percival and Card, 1992b).

The belt consists of several lithologically diverse, fault-bounded panels of supracrustal and intrusive rocks. Although most contacts between panels are tectonic, a wide range of ages determined by preliminary U-Pb geochronology indicated that unconformable relationships could also exist between or within panels. Consequently, at the invitation of a mining exploration venture active in the region, the writers revisited the Vizien belt for one week during the 1992 field season to examine key contacts. In the following, we describe a well-preserved unconformity between tonalitic basement and sedimentary-volcanic cover within panel B and discuss its significance in the evolution of the Vizien belt and the northeastern Superior Province.

REGIONAL SETTING

The Vizien greenstone belt is located in central Ungava, northern Quebec. It forms part of the north-northwest-trending Goudalie domain (Percival et al., 1991, in press) of mainly amphibolite-facies tonalitic rocks that forms a central spine to the 500 km-diameter Minto block. Numerous supracrustal relics, including the well-preserved Vizien belt, distinguish the Goudalie domain from the adjacent Lake Minto domain to the west and Utsalik domain to the east. Lake Minto domain consists mainly of pyroxene-bearing granodiorite, diatexite, and granite, with minor granulite-facies supracrustal rocks, whereas Utsalik domain is composed mainly of hornblende- and pyroxene-bearing granodiorite with minor mafic enclaves. Granitoid intrusions in the Lake Minto and Utsalik domains range in age from 2725 ± 5 to 2690 ± 5 Ma (Percival et al., in press; Stern et al., in press). Older tonalitic rocks within the Goudalie domain include units dated at 3010 (Percival and Card, 1992b) and 2940 Ma (J.K. Mortensen, unpub. data).

The Goudalie, Lake Minto, Utsalik, and other domains of the Minto block are similar in scale and lithological diversity to the east-trending subprovinces of the southern Superior Province but contrast in their northerly structural trends and aeromagnetic patterns. The Goudalie domain, where defined by mapping in the Leaf River area, is characterized by a low aeromagnetic anomaly which continues to the northwest toward the Cape Smith belt and to the southeast toward the La Grande River area. Flanking plutonic high-grade gneiss domains have characteristic magnetic highs which also continue to the north and south of the Leaf River transect.

GEOLOGY OF THE VIZIEN GREENSTONE BELT

Well-preserved supracrustal rocks within the Vizien greenstone belt include metavolcanic and metasedimentary rocks, metamorphosed to mid-amphibolite facies. Along with metaplutonic rocks, the supracrustal units occur in four lithotectonically distinct panels which are bounded by faults and together constitute a northwesterly tapering belt up to 10 km wide and 40 km long (Fig. 1). Primary features such as sedimentary stratification and volcanic textures are locally well preserved. Contacts with some bounding tonalitic rocks are high-strain zones.

Panel A, on the western edge of the belt, is bounded to the west by quartz-phyric tonalite. This body contains quartz phenocrysts with square crystal outlines, suggestive of inverted tridymite and consequently a subvolcanic origin for the unit. The structurally basal unit of the panel is plagioclase-phyric andesite, overlain by basaltic rocks, which contain garnet-grunerite iron-formation and alteration assemblages including tourmaline, cummingtonite, and anthophyllite-cordierite. The mafic rocks are structurally overlain by plagioclase-phyric andesite that grades into felsic volcanic rocks made up of mixed dacite and quartz-phyric rhyolite. Rhyolite contains zircons of 2723.6 ± 1.3 Ma age, and is interlayered with minor siltstone, as well as thin andesitic layers. The felsic unit is overlain structurally by metasedimentary rocks made up mainly of muscovite schist, grading upward to thinly-layered quartz-rich siltstone and minor conglomerate. Units of panel A are truncated to the east by a fault carrying the structurally basal units of panel B.

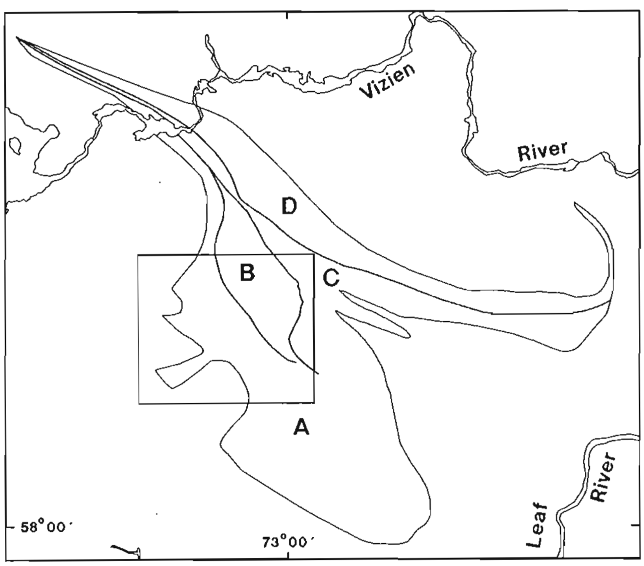

Figure 1. Simplified geological map of the Vizien belt showing location of structural panels and their bounding faults. The box indicates the area portrayed in detail in Figure 2.

The basal unit of panel B, concordant to the A-B fault, is a prominent ridge of serpentinized peridotite, with some talc schist. To the east is a unit of interlayered ultramafic and mafic rocks, intruded on the east by a body of variably plagioclase-phyric, muscovitic, foliated tonalite which contains numerous kilometre-scale septa of the mafic-ultramafic unit. Minor folds, as well as rodding and mineral lineations plunge moderately southeast. Tonalite is cut by 2-5 m wide mafic dykes. The 5 km-long eastern margin of the tonalite, with its contained inclusions and dykes, is truncated by a well preserved and exposed unconformity with a regolith beneath it, described in detail below. The unconformity is overlain by a sequence of supracrustal rocks including a basal conglomeratic unit containing pebble- to boulder-sized clasts of granitoid and metagabbroic rocks in a matrix of biotitic wacke. Conglomerate is interlayered with and overlain by biotitic wacke, which is depositionally overlain by a unit of mafic flows and sills, in turn structurally overlain by an ultramafic unit. The upper contact of the supracrustal sequence is a fault carrying units of panel C.

The B-C boundary fault truncates the southwestern limb of a northwest-plunging antiformal structure that makes up panel C (Fig. 1). The tight fold has a core of tonalitic rocks which is structurally overlain by felsic and mafic volcanic rocks. Although deformed and metamorphosed, several features of these volcanic rocks suggest possible subareal deposition: 1) massive, fine grained textures in mafic rocks over layer thicknesses of hundreds of metres (no pillows or pillow breccias); 2) thick sequences of massive, fine grained felsic rocks with sporadic feldspar-phyric zones, rare flow layering, and local fragmental textures; and 3) absence of alteration zones or sulphide mineralization. Panel C forms a northwest-tapering wedge with arms of mafic and dacitic schist that extend through migmatitic zones to the southeast and east. The hinge zone of the fold, plunging ~50°NW, has a strong axial planar foliation which appears to have localized pegmatite and quartz veins.

Panel D, to the east of the regional fold, is a steeply northeast-dipping unit of deformed, schistose rocks with a sporadic, moderately northerly-plunging elongation lineation. Dominantly mafic in composition, the panel also contains fine grained felsic schists of probable volcanic, tonalitic, and granitic/pegmatitic origin, as well as discordant pods of peridotite and pyroxenite. Mafic components probably also represent rocks of extrusive and intrusive origin. The 3 km wide panel of intensely deformed, mixed rocks gives way to the northeast to gneissic and homogeneous tonalites with few mafic enclaves. Pods of ultramafic rock (serpentinite, pyroxenite) are common in tonalite within 300 m of the northeastern contact of the belt.

GEOLOGY AND GEOCHRONOLOGY OF PANEL B

Bounded by steeply-dipping faults to the northeast and southwest, panel B outlines a lenticular form 3 by 11 km (Fig. 1, 2). It consists principally of foliated to massive biotite±muscovite tonalite containing enclaves of mafic and ultramafic schist as well as minor intermediate tuff-breccia representing an older supracrustal unit.

Tonalite is homogeneous, and medium grained to quartz- or plagioclase-phyric. It is cut by at least two intrusive phases. Numerous sharp-walled, linear bodies 3-5 m wide of homogeneous medium grained, weakly foliated amphibolite are interpreted to be metamorphosed dykes. Later white pegmatite containing assemblages of muscovite±garnet occurs as dykes and pods.

Evidence of alteration and metamorphism is widespread in several units of panel B. Mafic rocks commonly contain assemblages including the minerals anthophyllite, cordierite, cummingtonite, Mg-chlorite, and rare staurolite. Tonalitic rocks contain abundant muscovite and rare garnet or cordierite-anthophyllite. Aluminous schists contain the local low-variance assemblage staurolite-garnet-andalusite- sillimanite-biotite-plagioclase-quartz-graphite, defining metamorphic conditions of 550°C, 0.3 MPa.

The commonly strong foliation, defined by 5-10% aligned muscovite, strikes generally northwesterly, with vertical or steep dips to the northeast. In the central region trends are variable, reflecting the distribution of mafic-ultramafic schist enclaves (Fig. 2).

An angular unconformity, described in detail below, occurs in the eastern part of the panel between tonalite, with its enclosed mafic and ultramafic bodies, and an overlying sequence of conglomerate, greywacke, and volcanic and intrusive rocks of mixed mafic and ultramafic composition. The unconformable package is bounded above by a fault.

Structures bounding the panel were observed in a few locations. At the southeastern margin, metasedimentary schists of panel A are in sharp contact with ultramafic schist of the southwestern unit of panel B. The southwestern contact is inferred to be a fault because several units belonging to panel A terminate against the continuous, southwestern ultramafic unit of panel B (Fig. 2). Similarly, on the northeastern margin, interlayered mafic and intermediate volcanic units of panel C are truncated at the contact with panel B.

Radiometric ages are available for two units in panel B. Tonalite contains zircon of apparent igneous origin which yielded a preliminary U-Pb age of 2940 ±5 Ma. Two clasts from the conglomerate have been dated: 1) a tonalite boulder with zircons of ~2955 Ma age; and 2) a large boulder of graphic granite with a heterogeneous population of zircons giving single-grain Pb/Pb ages from 2708 to >3000 Ma (J.K. Mortensen, unpub. data), interpreted as a mixture of inherited and igneous zircons. The youngest grain, even if also inherited, provides a maximum age for crystallization of the granite, and hence for deposition of the conglomerate. The sedimentary-volcanic package is therefore younger than 2708 Ma. Rare dykes of tourmaline-bearing granitic pegmatite, of similar composition to pegmatites elsewhere in the Vizien belt, cut conglomerate and biotite schist parallel to the foliation. The presence in the sedimentary-volcanic rocks of a strong foliation parallel to regional attitudes as well as the crosscutting pegmatites, establishes an Archean age for the unconformity-based package.

THE VIZIEN UNCONFORMITY

The angular unconformable relationship is exposed over a strike length of about 5 km (Fig. 2, 3). The unconformity terminates at the B-C boundary fault in the north and becomes deformed beyond recognition in a high-strain zone in the south. In the intervening area the steeply dipping unconformity is remarkably well preserved and exposed (Fig. 3). At the map scale it truncates several mappable units in the basement. At the outcrop scale, older foliation, quartz veins, and mafic dykes in tonalite are locally truncated at the contact, indicating that deformation and mafic intrusive events separate tonalite intrusion from conglomerate deposition. Relief on the ancient erosion surface of at least 1 m is evident, and mappable embayments of conglomerate into tonalite of tens to hundreds of metres (Fig. 3, 4) may either be depositional or a result of faulting.

Regolithic surface

Tonalitic rocks become progressively altered eastward toward the unconformity. The proportion of muscovite, quartz, and albite increase at the expense of biotite and plagioclase, as the texture changes from foliated igneous to coarse sedimentary grit (Fig. 5). Mafic units within the basement proximal to the unconformity are altered to massive, medium grained sandstone with minor centimetre-scale clasts. Basement alteration and conversion to grit are interpreted to result from surficial weathering. Geochemical studies are underway to assess compositional changes associated with the weathering process.

STRATIGRAPHY OF THE <2708Ma PACKAGE

The supracrustal sequence above the unconformity consists of a basal conglomeratic unit (maximum 25-70 m thick), a middle biotite schist unit (70-100 m), and an upper unit of mafic metavolcanic and intrusive rocks (~70 m thick)(Fig. 3, 6, 7). The variably deformed sedimentary unit can be traced

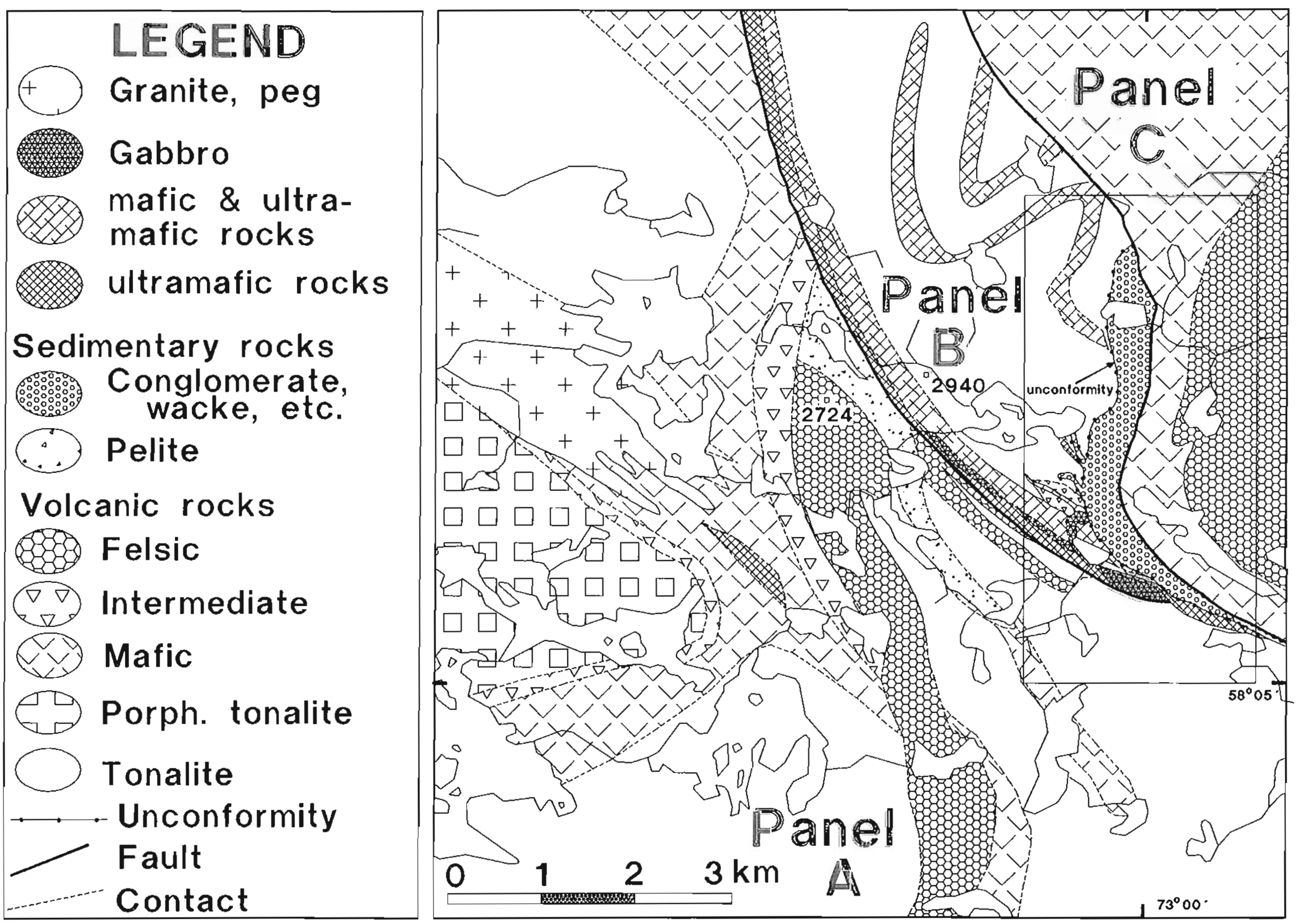

Figure 2. Detailed map of the southern part of panel B, showing location of the Vizien unconformity and the <2708 Ma sedimentary-volcanic package. The box outlines an area shown in more detail in Figure 3. Locations of U-Pb zircon ages (Ma) are indicated by small square symbols.

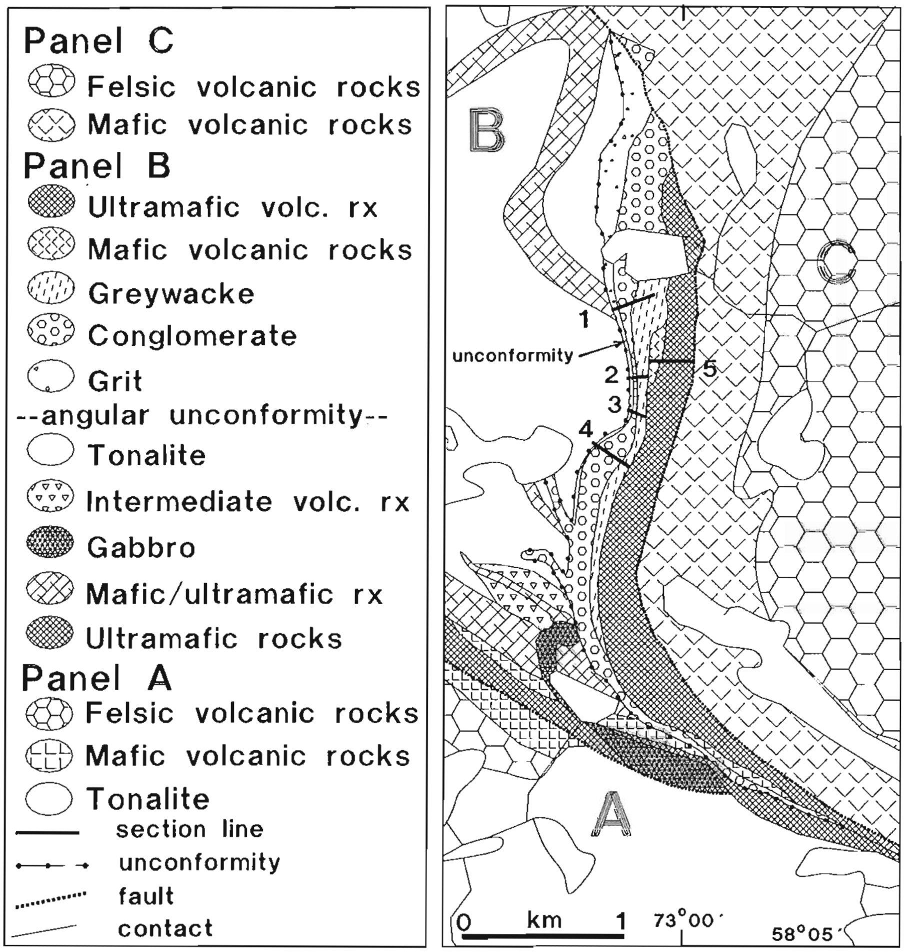

Figure 3. Detailed map of the unconformity-bound sedimentary-volcanic package. Lines 1 to 5 refer to measured sections portrayed in Figures 6 and 7.

Figure 4. Outcrop photograph of the unconformity surface showing tonalitic basement (bottom) and matrix-supported conglomerate (top); note significant relief on unconformity. Width of photograph is 1 m. GSC 1992-249J

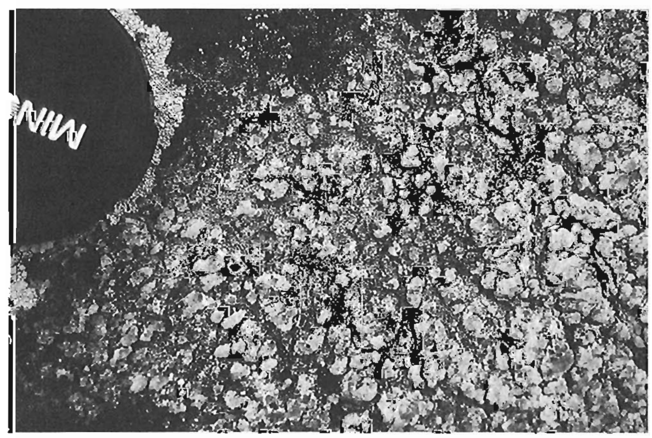

Figure 5. Coarse quartz-muscovite grit in regolith derived from tonalite. The regolith has variable thickness. Lens cap scale in this and subsequent photos is 5 cm in diameter. GSC 1992-249I

over a strike length of 5 km, from the B-C boundary fault in the north to a highly attenuated (<1 m) tail between panels A and C in the south.

Sedimentary rocks of the <2708 Ma package

The basal conglomeratic unit is a heterogeneous assemblage of polymictic orthoconglomerate (Fig. 8), paraconglomerate, quartz-rich grit, and quartz pebble conglomerate. The clast population comprises, in approximate order of abundance, tonalite, gabbro, mafic and felsic volcanic rocks, granite, and ultramafic rocks. The tonalite and gabbro clasts resemble lithologies in the subjacent basement. Clasts are most commonly 2 to 10 cm in size, with local occurrences of tonalite boulders up to 1 m. Conglomeratic units are recognizable even where intensely deformed (Fig. 9).

Stratigraphic sections through the sedimentary unit were measured at four locations over 1 km of strike length (Fig. 3, 6) and indicate considerable lateral variation. In section 1 (Fig. 6), much of the lower part of the unit consists of tonalite pebble conglomerate. The matrix is generally biotite-rich metagreywacke, with local occurrences of white to greenish quartz-rich grit. Quartz grit with quartz pebble conglomerate lenses forms a member approximately 18 m thick in section 3 (Fig. 6). These rocks are relatively mature texturally.

Coarse metamorphic assemblages occur locally near the base of the sequence. In section 2, metre-scale layered schists have variable ferromagnesian assemblages including garnet-biotite, staurolite-biotite, and cordierite-anthophyllite-biotite. These units, which overlie gritty-textured micaceous tonalite, and are in turn overlain by grit rich in quartz, muscovite, and garnet, may represent either aluminous, Fe- and Mg-rich sediments produced by weathering of mafic units in the basement, or older, metasomatically altered rocks included within the tonalite. If sedimentary, their nature suggests a fine grained (pelitic) protolith, with considerable compositional variety.

A unit of metasedimentary schist overlies and interfingers with conglomerate. It consists mainly of medium grained, moderately foliated schist with biotite-plagioclase-quartz± garnet assemblages. Local centimetre- to decametre-scale layering may be transposed bedding. The bulk composition appears to be similar throughout the unit, with the exception

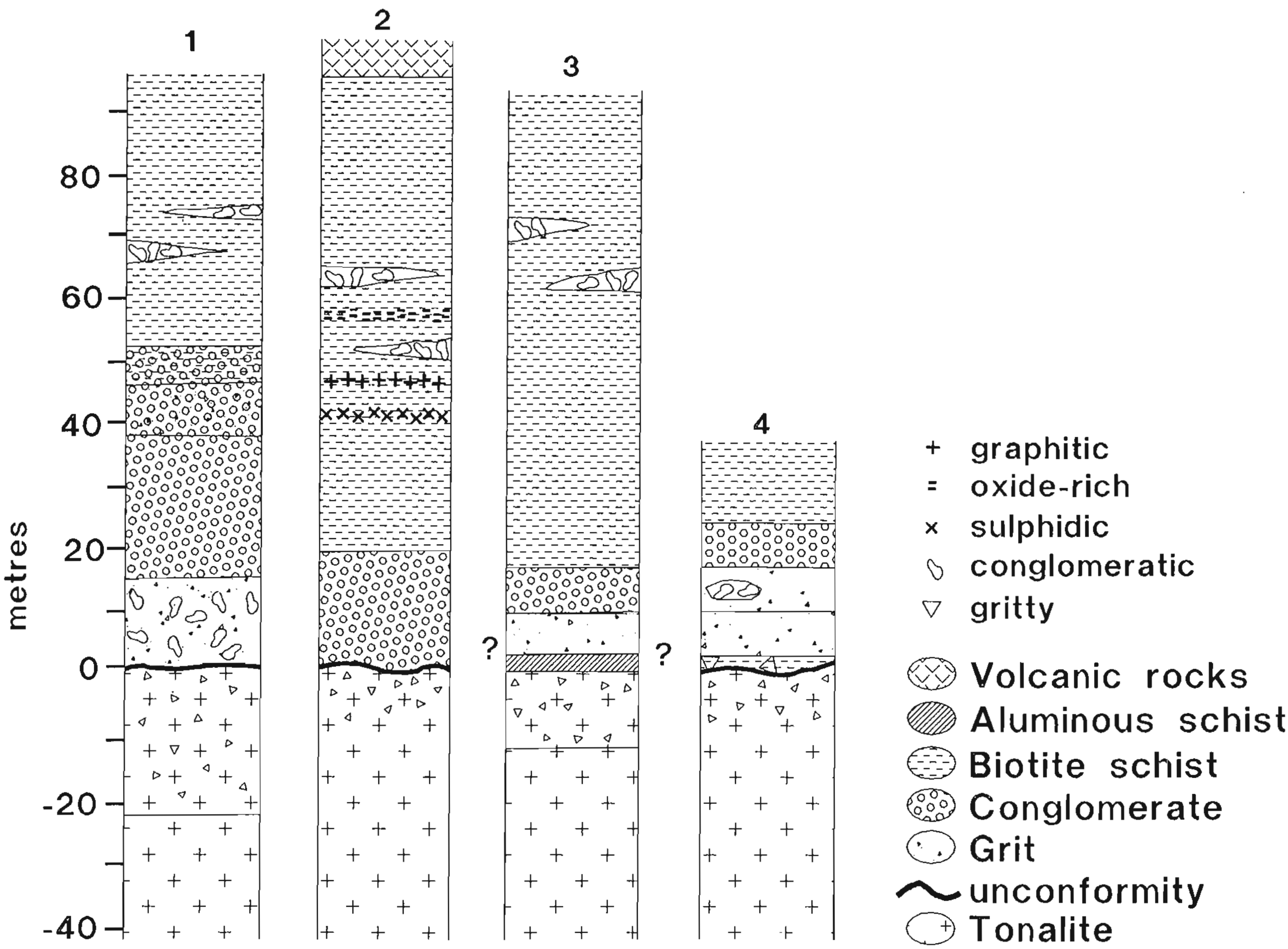

Figure 6. Stratigraphic sections 1-4 through the sedimentary unit in locations shown in Figure 3.

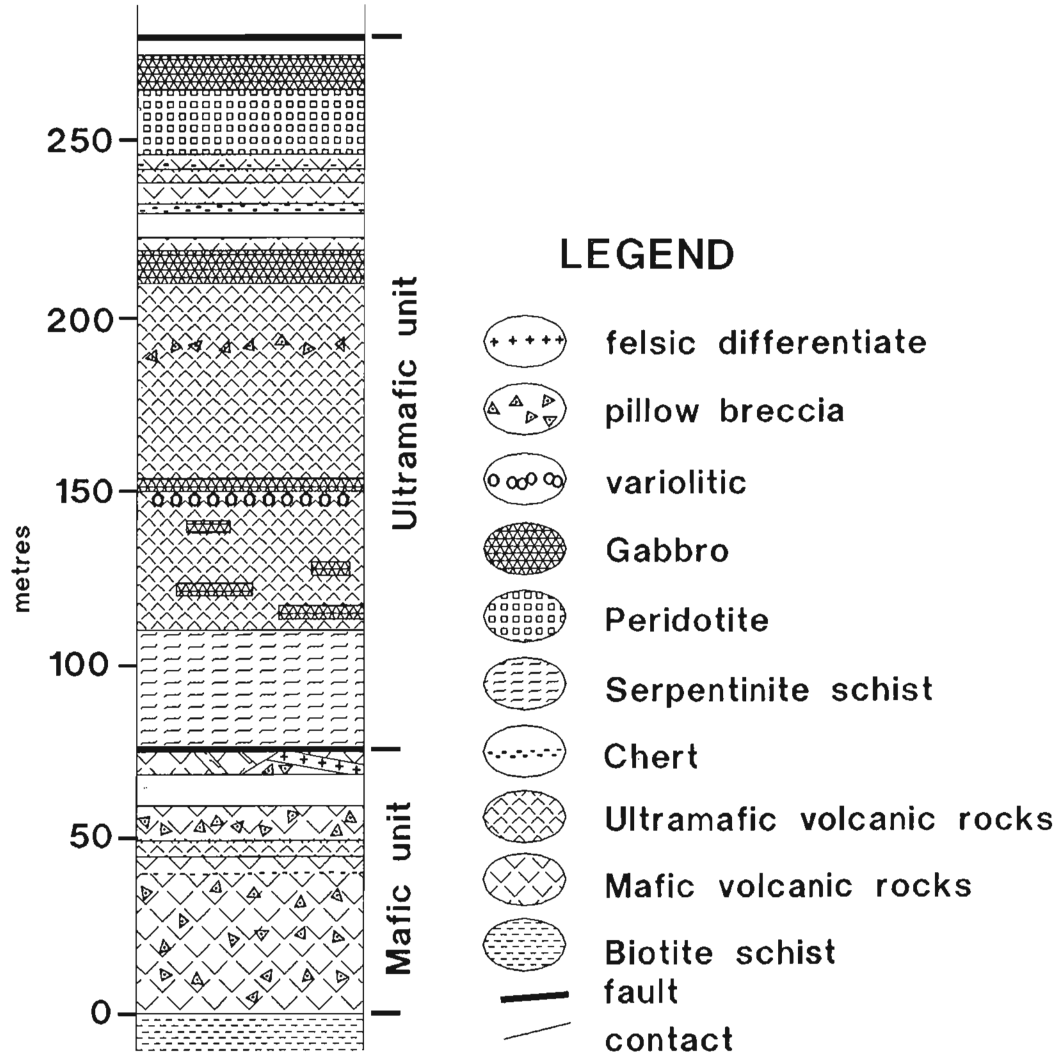

Figure 7. Stratigraphic section through the volcanic unit (section 5, shown in Fig. 3).

Figure 8. Clast-supported conglomerate with rounded cobbles of tonalite and gabbro and minor matrix of biotite schist. GSC 1992-249D

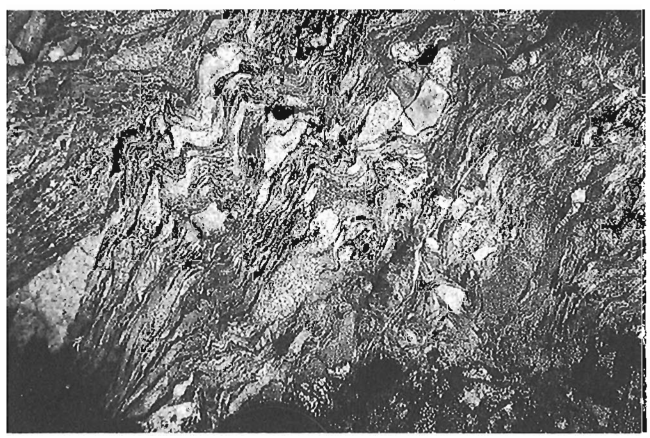

Figure 9. Strongly foliated conglomerate in a deformation zone adjacent to a fault near the northern end of the unit. GSC 1992-249F

of local lenses of conglomerate and sulphide-, magnetite- and graphite-bearing schist (Fig. 6, section 2). These characteristics suggest that the unit for the most part represents metamorphosed thick-bedded or unbedded greywacke.

Together, the measured sections illustrate considerable lateral variation in the nature of the sedimentary section above the unconformity. They also demonstrate the variability in the thickness of the regolithic zone and show significant local relief on the ancient erosion surface.

Volcanic and associated rocks of the <2708 Ma package

Unbedded greywacke in the upper part of the sedimentary unit is overlain by two units of volcanic and associated rocks of mafic and ultramafic composition. At one locality (section 5, Fig. 3), a unit dominated by mafic volcanic rocks lies directly and conformably on the sedimentary unit. However, over most of the length of the package, a unit consisting mainly of ultramafic flows and sills lies in tectonic contact with the structurally lower greywacke or mafic unit. The locally preserved, vertical to steeply-dipping mafic unit ranges up to 70 m thick (Fig. 7). The structurally overlying ultramafic unit is fault-bounded to the east against the subaerial mafic and felsic volcanic rocks of panel C.

Mafic unit

A heterogenous assemblage of mafic volcanic and intrusive rocks, with minor ultramafic and felsic units makes up the mafic unit. A section through the sequence consists of several distinct layers (Fig. 7). The basal 2 m-thick flow has a chilled flow base, a medium grained interior and fine grained top overlain by pillow breccia (Fig. 10). In some locations pillows are well preserved in the sequence. A 2 m-thick ultramafic (flow?) unit has centimetre-scale layering. Mafic pillow breccias and gabbro comprise a heterogeneous unit ~35 m thick that is overlain by a thin (10 cm) layer of chert. The upper 35 m of the mafic unit is composed mainly of mafic pillow breccia, with metre-scale interlayers of schistose ultramafic rock. The uppermost 10 m is a structurally complex "mélange" of metre-scale blocks of pillow breccia, gabbro and layered gabbro in which internal layering is discordant to that in adjacent blocks. Layered gabbro locally

Figure 10. Well preserved pillow breccia in a basaltic flow, basal mafic unit. GSC 1992-249C

Figure 11. Serpentinite schist in the deformation zone at the base of the ultramafic unit; note the strong crenulations. GSC 1992-249H

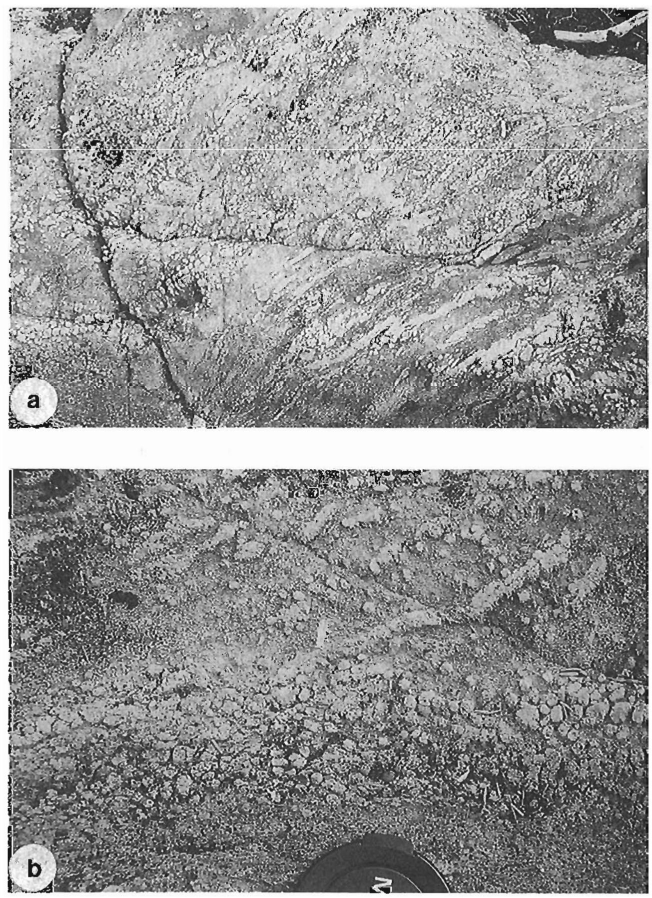

Figure 12. Ultramafic volcanic unit showing light-coloured varioles in an ultramafic matrix. **a)** outcrop photograph showing discrete varioles as well as coalesced zones forming pseudolayering. GSC 1992-249A; **b)** close-up of variolitic texture. GSC 1992-249E

contains differentiated layers of quartz dioritic composition (Fig. 7). The structurally complex nature of this uppermost "mélange" suggests that it is a syndepositional feature, possibly related to emplacement of the overlying ultramafic unit.

Ultramafic unit

The ultramafic unit has a basal serpentinite layer approximately 30 m thick. It is characterized by a strong foliation and crenulation folds forming a steep, strong west-plunging lineation (Fig. 11). Serpentinite is overlain by a 35 m-thick unit of ultramafic schists with gabbroic pods and layers oriented parallel to the basal contact of the volcanic unit. A thin (5 m) layer of variolitic ultramafic rock overlies the ultramafic schist. Variolitic units (Fig. 12) up to 15 cm thick, although somewhat distorted and serpentinized, are generally parallel to the basal contact of the unit. Overlying the variolitic unit is a 3 m-thick medium grained mafic sill or flow, which is in turn overlain by a 55 m-thick unit of fine grained ultramafic rocks which contains a 1 m layer of variolitic hyaloclastite (Fig. 13). Stratigraphically above is a 10 m-thick gabbro body, overlain by 3 m of mafic pillow lava and pillow breccia. Beyond a 10 m exposure gap is a 2 m wide unit of chert and garnetiferous amphibolite, probably representing silicate-facies iron-formation. A second, 5 m-thick, cherty layer containing some sulphides, occurs above intervening locally fragmental amphibolite (7 m) and fine grained ultramafic schist (5 m). Above the second chert layer is a prominent 15 m-wide ridge of coarse grained peridotite of probable intrusive origin that extends along strike for ~2 km. It is overlain by a 10 m-thick gabbroic unit which is in fault contact with fine grained mafic rocks of panel C. A stratigraphic column is presented in Fig. 7.

The basal serpentinite of the ultramafic unit (Fig. 7, 11) probably marks a zone of dislocation. In some locations, it rests directly on the sedimentary unit. It cuts up section through layering of the mafic unit along strike and has conformable layering with that in the mafic unit elsewhere.

Figure 13. Texture of variolitic hyaloclastite, central ultramafic unit. Note delicate fine-scale layering, outlined by variole-rich horizons, within hyaloclastite fragments. Pen barrel is 7 mm in width. GSC 1992-249B

The ultramafic schist is generally strongly foliated, with tight crenulation folds of "w" or "z" geometry. Where contacts with underlying units are exposed, the structurally inferior biotite schist or mafic unit generally do not possess the crenulation fabric. These observations suggest that the ultramafic unit is allochthonous with respect to the biotite schist and mafic units. As it is also fault-bounded against panel C, the ultramafic unit represents an additional discrete lithotectonic panel.

DISCUSSION

At least three ages of supracrustal rocks are represented in the Vizien greenstone belt. The oldest comprises undated inclusions of felsic tuff-breccia in panel B tonalite of 2940 ± 5 Ma age. Volcanic rocks of panel A, including rhyolite with a date of 2724 ± 1 Ma, represent an intermediate age group. The unconformity-based package younger than 2708 Ma is the youngest. Although the relationship between the oldest and youngest package is evident within panel B, rocks of the intermediate-age group do not appear to be represented in panel B, except perhaps as conglomerate clasts. Therefore there is no lithological or age linkage between panels A and B.

The relationship between the lithotectonic panels in the Vizien belt is not clarified by the present study. It is evident that panels A and B formed at different times and had discrete lithotectonic histories prior to amalgamation. The subareal depositional environment postulated for volcanic rocks of panel C distinguishes these rocks as well from both the 2724 and <2708 Ma sequences. Geochronology in progress on dacite from panel C may establish age linkages with rocks in other panels.

The timing of amalgamation of the panels constituting the Vizien belt is loosely constrained on the basis of field and geochronological relationships established to date. As rocks younger than 2708 Ma are involved in the assembly of panels, this event occurred after deposition of these rocks. It is possible that the juxtaposition occurred immediately following deposition of the mafic unit of the <2708 Ma package, if the "mélange" at the top of the mafic unit is correctly interpreted as a depositional unit related to over-riding of a thrust sheet, the ultramafic unit. The age of that particular stacking event could then be determined by dating the mafic unit. A date in progress on pegmatite cutting conglomerate will provide a minimum age for assembly.

REGIONAL SIGNIFICANCE AND IMPLICATIONS

Long history and crustal recycling

The presence of rock units in the Vizien belt with ages spanning ~300 Ma (3010 to <2708 Ma) and the unconformable relationship described above demonstrate a lengthy, complex history in this part of the Superior Province. Elsewhere in the Minto region, much of the history has been eradicated by widespread plutonic recycling (Stern et al., in press).

In its complex history, the Vizien belt is similar to some greenstone belts of the northwestern Superior Province, for example the Uchi (Nunes and Thurston, 1980), Island Lake (Stevenson and Turek, in press) and Cross Lake (Corkery et al., in press) belts. Unconformable relationships occur in these belts among rock units in the 3.1-2.7 Ga range.

Sedimentary-ultramafic associations

In the northwestern Superior Province, an association between relatively mature quartz-rich sedimentary rocks and mafic-ultramafic rocks occurs in rock sequences >2.8 Ga (Thurston and Chivers, 1990). Based on the tectonic nature of contacts between the minor quartz-rich sedimentary rocks and ultramafic rocks and of the Vizien belt, there appears to be little similarity between the Vizien sequence and the quiescent platform associations of northwestern Ontario.

The relatively young age of the unconformity in the Vizien belt suggests possible comparisons with conglomeratic deposits of Timiskaming type that occur sporadically within greenstone belts throughout the Superior Province. These relatively young, fault-bounded sequences of variably deformed sedimentary and volcanic rocks differ from the Vizien occurrence in several aspects. Timiskaming sequences generally occur on a volcanic-plutonic basement which is only slightly (<50 Ma) older. They occur in linear, fault-bounded basins and have broadly symmetrical stratigraphy with respect to their margins. Finally, the volcanic rocks associated with Timiskaming deposits have characteristic alkaline compositions. These differences suggest that the setting of the unconformable package in the Vizien belt is unlike that of Timiskasming sequences.

In the western Wabigoon subprovince of Ontario, the Steep Rock Group and its basal unconformity closely resemble the Vizien occurrence. There, tonalitic basement of ~3.0 Ga age, cut by mafic dykes, is truncated by an unconformity and overlain by coarse basal conglomerate and volcanic rocks including the ultramafic "ashrock" (Wilks and Nisbet, 1988). A tectonic upper contact separates the Steep Rock Group from unrelated volcanic rocks.

Possible extensions of the Goudalie domain

An unconformity and younger stratigraphic sequence in the Sakami Lake greenstone belt of the La Grande River subprovince (Roscoe and Donaldson, 1988) have similarities to the Vizien unconformity. There, basal quartz arenite with some quartz-pebble conglomerate and ultramafic sills overlie gneisses representing probable basement. The quartz-rich rocks are overlain by calc-silicate rocks and oxide-silicate-facies iron-formation, capped by basalt.

Sparse geochronology in the La Grande River belt indicates comparable ages to those from the Vizien belt. Tonalitic gneiss has an age of 2811 ±2 Ma and a granodiorite pluton, 2712 Ma (Mortensen and Ciesielski, 1987). A possible connection between the east-trending La Grande belt and north-trending Goudalie domain is indicated by their common low aeromagnetic anomaly. The change in trend occurs near the junction between the La Grande, Ashuanipi, and Minto subprovinces, in the region east of Lac Bienville.

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Geology and structural development of the Archean to Paleoproterozoic Burwell domain, northern Torngat Orogen, Labrador and Quebec[1]

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Continental Geoscience Division

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Abstract: The Archean gneisses and Paleoproterozoic meta-igneous rocks of the Burwell domain, Torngat orogen, are deformed in the south by the older, sinistral Abloviak shear zone (ASZ), and in the east by the younger, sinistral, east-side-up Komaktorvik zone (KZ). A $\leq$2 km wide dextral mylonite along the northern contact of the ASZ and which cuts it, may represent a slip plane that resulted from folding of the Abloviak shear zone during counterclockwise rotation of Burwell domain. The southerly end of the mylonite zone also marks the end of the Komaktorvik zone, suggesting that it formed contemporaneously with the dextral mylonite and the folding of the Abloviak shear zone in response to regional sinistral transpression. Preliminary U-Pb geochronologic data show that deformation within the Komaktorvik zone occurred at ca. 1.79 Ga, the same time as west-side-up mylonites in the southern part of Torngat orogen, suggesting overall scissor movement along the orogenic front at this time.

Résumé : Les gneiss archéens et les roches métamagmatiques (Protérozoïque précoce) du domaine de Burwell dans l'orogène de Torngat sont déformés, au sud, par la zone de cisaillement sénestre plus ancienne d'Abloviak (ASZ) et, à l'est, par la zone sénestre de Komaktorvik (KZ), zone plus jeune à compartiment est relevé. Des plis résultant d'un glissement strate sur strate sont apparus dans l'ASZ à la suite de la rotation anti-horaire du domaine de Burwell, ce qui a peut-être entraîné la formation de la zone dextre de mylonites de $\leq$2 km de largeur, longeant et recoupant la partie nord du contact avec l'ASZ. La partie méridionale de la zone à mylonites marque également la limite de la KZ, soutenant ainsi l'hypothèse selon laquelle cette dernière serait contemporaine de la zone dextre de mylonites et au plissement de l'ASZ, à la suite d'un épisode de transpression sénestre à l'échelle régionale. Des résultats préliminaires de datation par la méthode U-Pb indiquent que la déformation au sein de la KZ a eu lieu il y a 1,79 Ga, soit à la même époque que la formation des mylonites au compartiment ouest relevé dans la partie méridionale de l'orogène de Torngat, ce qui suggère que des déplacements en ciseaux se seraient produits d'un bout à l'autre du front orogénique au même moment.

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INTRODUCTION

The Torngat project is aimed at mapping the Labrador peninsula north of 59°15'N. In addition to geologists from federal and provincial surveys, the project also involves several university research groups. Mapping was carried out in northernmost Labrador, northeastern Quebec and parts of the Northwest Territories (Button Islands, Killiniq Island) through helicopter-supported traverses.

The Paleoproterozoic Torngat orogen lies between and involves rocks of the Archean Nain and Rae provinces in the east and west, respectively (Fig. 1). In the map area, Torngat orogen splays into two distinct belts of strongly deformed rocks: the east-striking portion of the Abloviak shear zone in the south, and the north-striking Komaktorvik zone in the east, which bound Archean to Paleoproterozoic rocks of the Burwell domain (formerly terrain or terrane: Fig. 1, 2). As described previously (Van Kranendonk and Scott, 1992; Wardle et al., 1992), this part of the Torngat orogen is unique in that a widespread suite of Paleoproterozoic diorite-tonalite-granitic rocks (DTG suite; equivalent to the DQT suite of Wardle et al., 1992) intrude Nain gneisses.

In this report we describe the geology of the area between 59°15'N and 60°N, and examine the structural relationships around the triangular region formed by the "big bend" in the Abloviak shear zone and the Komaktorvik zone (Fig. 1, 2). Readers are referred to reports by Wardle et al. (1992, in press) for detailed descriptions of rocks north of 60°N, and to the companion report by Scott et al. (1993) which describes preliminary U-Pb geochronologic data.

LITHOTECTONIC ELEMENTS

The distribution of major rock units is shown in Figure 2. The units are grouped into several lithotectonic elements, described below, that are defined on the basis of crosscutting relationships, associations with other groups of rocks, structural complexities, and metamorphic grade.

Archean Nain gneisses

Archean gneisses interpreted as the northern continuation of the Nain Province occur along the eastern margin of the map area (Fig. 2). The gneiss complex consists of 70-80% migmatitic tonalitic gneisses and 20% of inclusions of layered mafic and ultramafic intrusive rocks and of supracrustal gneisses. The layered enclaves consist of mafic gneisses, minor ultramafic layers and meta-leucogabbros, and local anorthosite. Paragneiss units dominated by garnet-rich quartzofeldspathic gneiss form ≤25% of the complex in the inner part of Kangalaksiorvik Fiord (Fig. 2). East of Komaktorvik Fiord, paragneisses with prominent quartz-rich layers containing irregular fragments of clinopyroxenite are associated with mafic gneisses and with garnet-biotite felsic gneisses of possible clastic sedimentary origin.

Meta-anorthositic rocks and anorthositic gneiss form a major unit up to 10 km wide and 100 km long within the eastern part of the area. Similar rocks occur in lenses over 325 km, south to Okak Bay (Fig. 1, 2). Gabbroic anorthosite compositions dominate and textures vary from pristine cumulate rocks to gneissic tectonites. Locally, meta-anorthositic rocks are spatially related to layered metagabbros, both of which are disrupted by adjacent tonalitic orthogneiss containing numerous ultramafic inclusions (see also Wardle et al., 1992, in press).

Archean gneisses, infolded with, and locally intrusive into the western contact of the meta-anorthositic rocks have similar characteristics as those to the east and are also interpreted to be derived from Nain protoliths. All Archean rocks are cut by one or more sets of variably-deformed mafic dykes (see below). An exploratory Pb-Pb whole-rock isotope study now in progress (D. Bridgwater) suggests that a major part of the northern Nain Province east of the meta-anorthosite differentiated from mantle-like sources at about 3.0 Ga. Results from a small area around outer Seven Islands Bay (Fig. 2) point to an older (>3.3 Ga) component; these rocks are tentatively correlated with the Uivak gneisses of the Saglek area (Fig. 1, Bridgwater and Collerson, 1976).

Mafic dykes

Archean rocks and structures of Nain Province are cut by one or more sets of metamorphosed mafic dykes. Set 1, referred to as Avayalik dykes, are most prominent. They are characterized by large black plagioclase feldspars in a diabasic matrix and garnet-clinopyroxene assemblages where host Archean gneisses are also at granulite facies. Where Archean gneisses are retrogressed to amphibolite facies, mafic dykes contain white feldspars in a schistose hornblende-plagioclase±garnet matrix. Avayalik dykes are cut by the DTG suite and are therefore older than ca. 1.91 Ga (see below), but younger than late Archean deformation and metamorphism.

On the islands north of 60°N, a second set of dykes was recognized that cuts the DTG suite. Also containing garnet, this suite of dykes lacks the black feldspars and can locally be distinguished from the Avayalik dykes on the basis of a lighter-green colour and equigranular groundmass. In places, the weakly deformed nature of set 2 dykes within strongly sheared rocks suggests they were syntectonic with Paleoproterozoic deformation within the Komaktorvik zone (S. Hanmer, pers. comm., 1992).

Archean Rae(?) gneisses

Several outcrops south of the Tasiuyak gneiss (Fig. 2) consist of tonalitic orthogneiss and strongly deformed mafic dykes. These white-weathering, leucocratic gneisses contain centimetre to decametre layering, but do not contain a leucosome phase, distinguishing them from their Nain counterparts. Dykes are fine- to medium-grained and equigranular. On the basis that the gneisses are cut by mafic

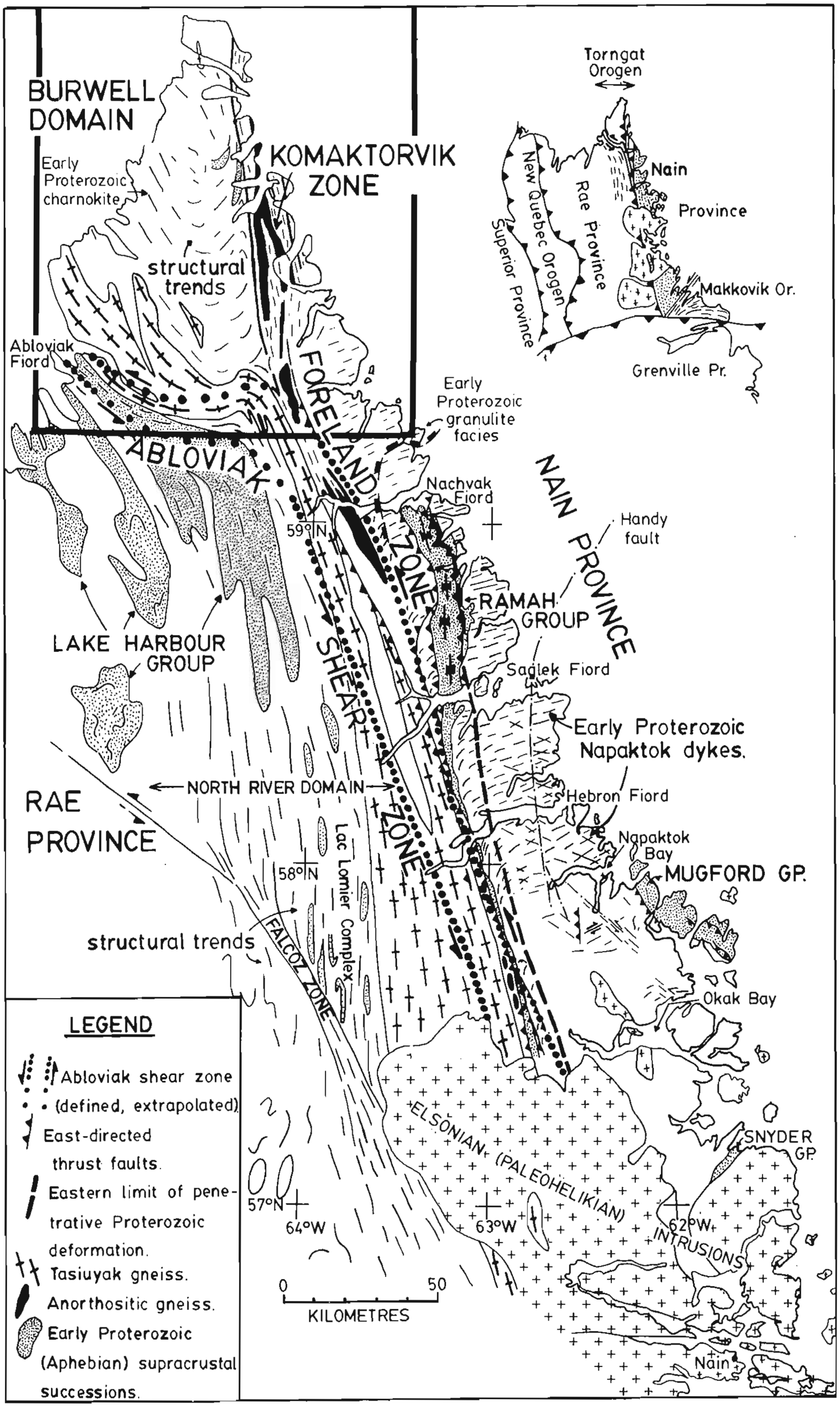

Figure 1. Geological map of the Torngat orogen, northern Labrador and Quebec. Major lithotectonic elements and subdivisions are shown, along with the location of the present mapping project (in box), shown in detail in Figure 2.

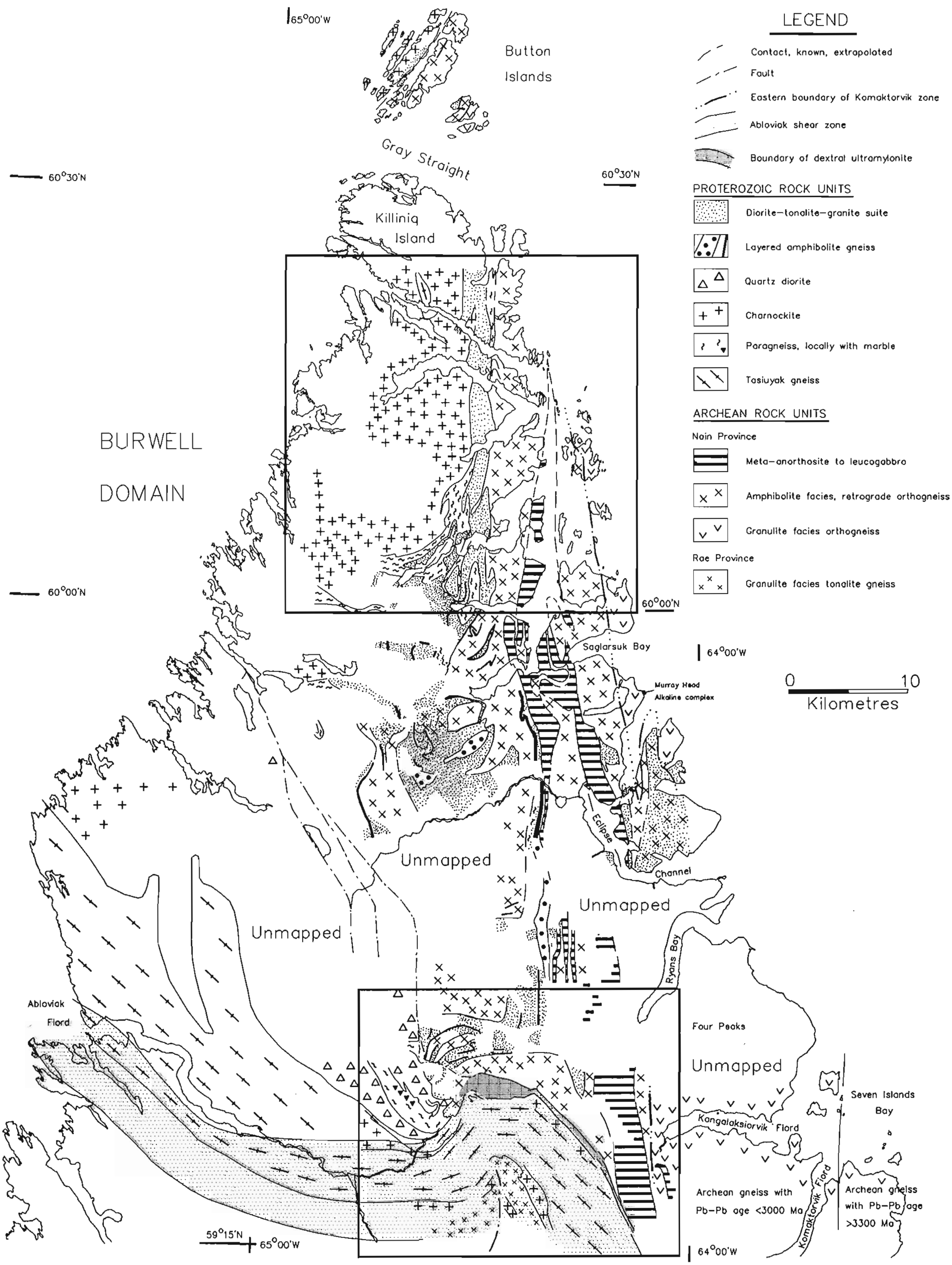

Figure 2. Geological compilation map of the Burwell domain in northern Labrador, Quebec and N.W.T. Area of Figure 4 indicated by southern box: area described in Wardle et al. (1992, in press) indicated by top box. Northwestern limits of Tasiuyak gneiss from Taylor (1979), and of the Abloviak shear zone from Goulet and Ciesielski (1990).

dykes, they are interpreted as Archean in age, and because they occur south of the Tasiuyak gneiss, are inferred to belong to the Rae Province.

Tasiuyak gneiss and paragneiss

Tasiuyak gneiss is a rusty-weathering, white unit of garnet-biotite±sillimanite paragneiss and diatexite that forms a prominent band, 6-20 km wide, along the boundary between the Rae and Nain provinces (Fig. 1, 3F: Taylor, 1979). It is intruded by charnockitic rocks and has experienced several periods of tight to isoclinal folding (see below).

North of the Abloviak shear zone, the Tasiuyak gneiss with its lilac-coloured garnets grades into grey, migmatitic paragneiss with red garnet. The latter locally contain ≤15 m thick, graphite-bearing calc-silicate and marble units (Fig. 2, 4). A similar association of paragneiss, also with local marble lenses, occurs near 60°N (Fig. 2).

Layered amphibolite gneiss

Included within, and tectonically interleaved with, Archean gneisses and rocks of the DTG suite are a series of 10-200 m wide sheets of centimetre to decametre layered amphibolites (Fig. 3A, 2, 4). These thin units generally contain hornblende-plagioclase assemblages, with local relict clinopyroxene cores and garnet-clinopyroxene-plagioclase veins. Layering is defined by modal variation of hornblende and plagioclase. Volcanic textures have not been identified.

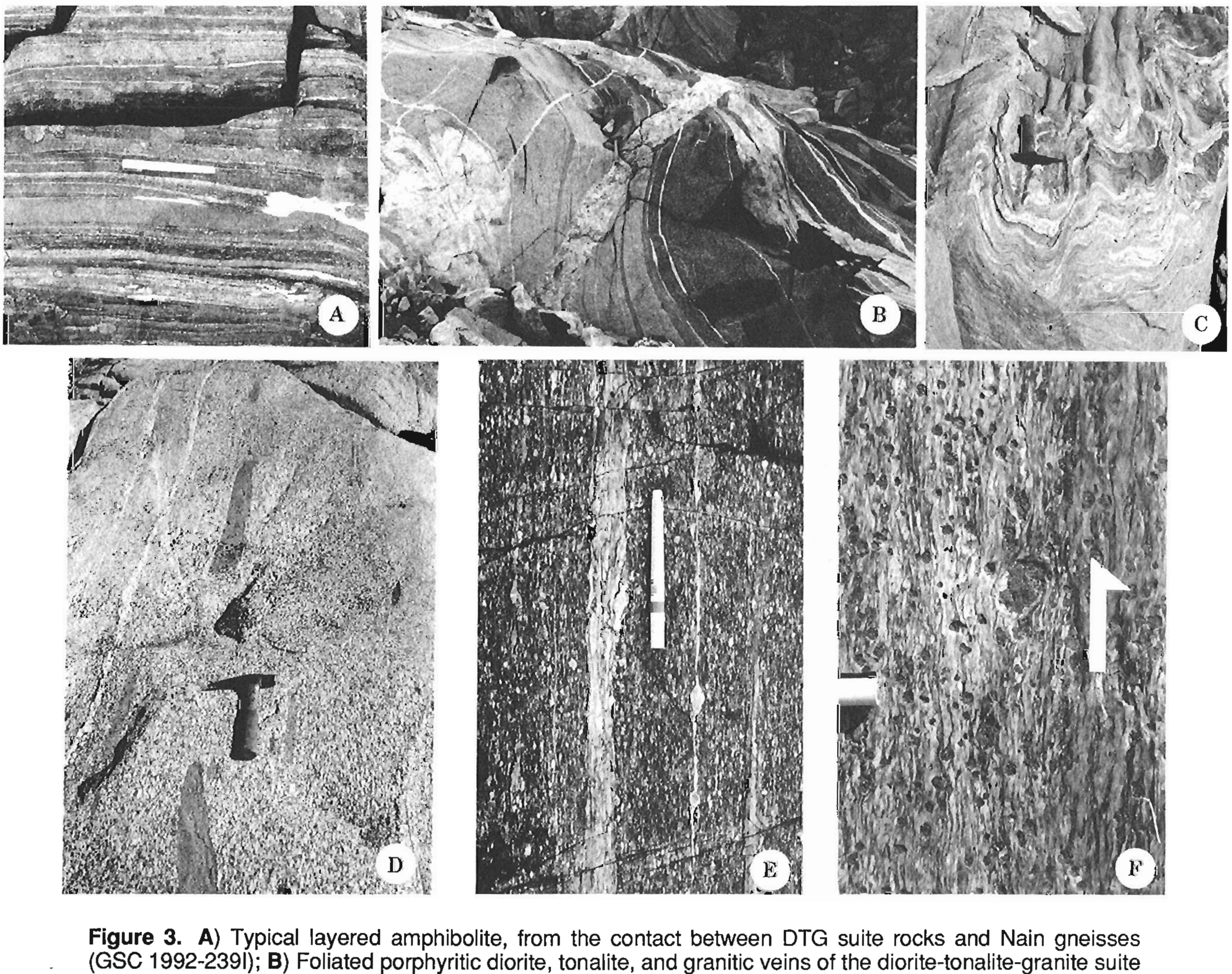

Figure 3. A) Typical layered amphibolite, from the contact between DTG suite rocks and Nain gneisses (GSC 1992-239I); **B)** Foliated porphyritic diorite, tonalite, and granitic veins of the diorite-tonalite-granite suite (GSC 1991-243A); tonalite from this site has been dated by U-Pb on zircon as 1910 ± 2 Ma (Scott, in press); **C)** Folded tonalite migmatite of the DTG suite, west of Ryan's Bay (GSC 1992-239C); **D)** View northwest of cognate xenoliths in grey quartz diorite (GSC 1992-239A); **E)** Mylonitized, porphyroclastic quartz diorite in the Abloviak shear zone (GSC 1992-239O); **F)** Mylonitic Tasiuyak gneiss in the Abloviak shear zone, showing sinistrally rotated garnets and shear bands: pencap is 1 cm wide (GSC 1992-239G). Pen in A) and E) is 13 cm long; hammer in C) and D) is 30 cm long.

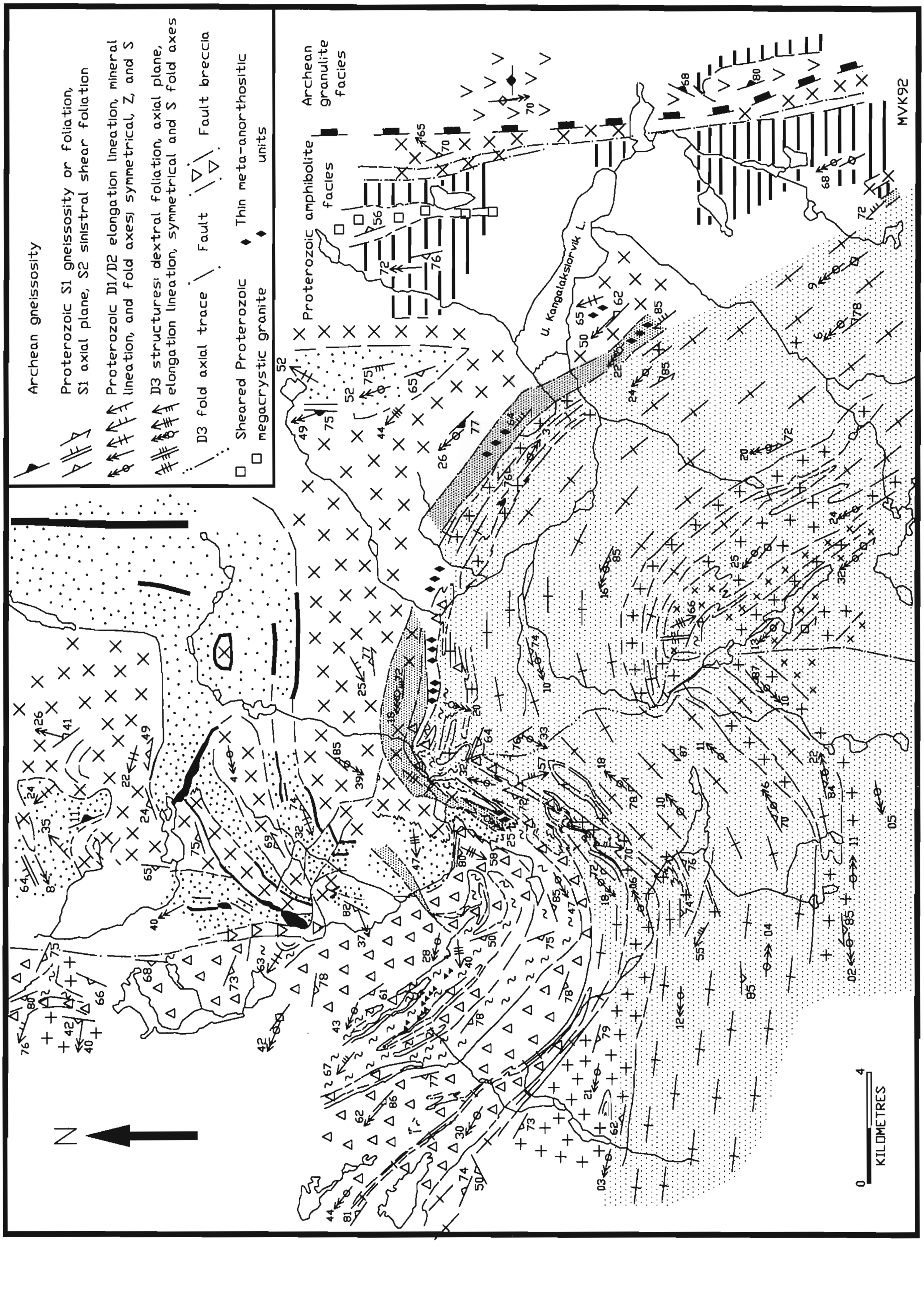

Figure 4. Geological map of the triangular region defined by the fold of the Abloviak shear zone and the southern part of the Komaktorvik zone. Legend is the same as for Figure 2, which also shows location of the map.

The primary relationship of these amphibolites to the DTG suite and Archean gneisses is enigmatic due to highly-strained contacts which obscure intrusive relationships. Relatively simple fabrics in these rocks suggest a Paleoproterozoic age. They are thus considered to represent early, gabbroic phases of the diorite-tonalite-granite suite.

Diorite-tonalite-granite (DTG) suite

Much of the eastern half of the map area is underlain by a suite of grey, amphibolite-facies, Paleoproterozoic rocks ranging from porphyritic leucogabbro and mesocratic diorite-quartz diorite, through tonalite and granodiorite, to granite (Fig. 2; see also Van Kranendonk and Scott, 1992; Wardle et al., 1992). Diorite-tonalite-granite suite rocks intrude Archean Nain gneisses, mafic dykes of the Avayalik suite (Van Kranendonk and Scott, 1992; Wardle et al., 1992) and paragneiss. Tonalite is most common, ranging from homogeneous, weakly foliated rocks, to leucocratic migmatite and gneiss (Fig. 3B, C). These rocks typically have simple fabrics relative to Archean orthogneisses and can be compositionally homogeneous over large areas (up to 600 km^2). Tonalitic-granodioritic rocks commonly occur as lit-par-lit sheets within Archean gneisses, with a wispy, discontinuous mafic layering inherited from the older rocks.

In the absence of mafic dykes, distinction between Archean and Proterozoic gneisses is often difficult. U-Pb geochronology indicates an age of 1910 ± 2 Ma for the tonalite of Figure 3B, and a possible range in age for the suite of ca. 1.91-1.86 Ga (Scott, in press).

Charnockitic rocks

Orthopyroxene-bearing granitoid rocks, generally granodioritic and referred to as the charnockite suite, occupy the western part of the map area (Fig. 2). These massive to foliated, brown-weathering rocks are medium- to coarse-grained, with equigranular to feldspar-megacrystic texture. Rafts of Tasiuyak gneiss and paragneiss occur throughout the suite. A charnockite from Killiniq Island yielded a U-Pb zircon age of 1895 ± 3 Ma (Scott, in press).

Light to dark grey, equigranular to plagioclase-megacrystic quartz diorite intrusions occur across the northern margin of the Tasiuyak gneiss (Fig. 2). They contain 15-20% hornblende-biotite aggregates, sometimes with relict orthopyroxene cores, and cognate xenoliths of fine grained dioritic material (Fig. 3D). The typically porphyritic texture and more mafic composition of this unit help distinguish it from charnockitic rocks, even where strongly deformed within the Abloviak shear zone (Fig. 3E). Although compositionally more characteristic of rocks within the diorite-tonalite-granite suite, the restriction of this unit to the northern margin of the Tasiuyak gneiss and western part of the map area suggests it may be related to the charnockitic suite.

AGE RELATIONSHIPS BETWEEN PROTEROZOIC IGNEOUS ROCKS

Contact relationships between the charnockitic and DTG suites are equivocal and remain to be mapped in detail. In one locality, charnockitic veins cut DTG suite tonalites. Elsewhere, however, these rocks are interlayered, showing gradational contacts suggestive of a metamorphic transition. Preliminary U-Pb geochronology suggests that the diorite-tonalite-granite suite may have been intruded over a period of at least 50 Ma. The 1895 Ma age of a charnockite suggests that rocks of this suite may be, at least in part, coeval with the diorite-tonalite-granite suite (Scott, in press; Scott et al., 1993). Major and trace elements and Sm-Nd and Pb-Pb isotopic ratios are being analyzed from both suites and adjacent gneisses in order to determine if the suites are distinctly different from each other or co-magmatic, whether they are juvenile or contaminated by older material, and to identify the potential sources of contamination.

METAMORPHISM

Nain gneisses and Avayalik dykes along the Labrador coast contain pristine or partly retrogressed garnet- and clinopyroxene-bearing assemblages indicative of high-pressure, granulite-facies metamorphism. Mineral assemblages within gneisses are considered to be late Archean, but the presence of similar assemblages within the dykes suggests that rocks in this zone were also affected by granulite-facies metamorphism in the Paleoproterozoic.

West of the Archean granulites to east of the charnockites, granulite-facies assemblages are locally preserved through a pervasive amphibolite-facies overprint. Many of the rocks in the diorite-tonalite-granite suite contain hornblende-biotite-epidote assemblages, with no evidence of granulite facies. Mafic dykes are recrystallized to hornblende-white plagioclase±garnet assemblages. Preliminary geothermobarometry from Archean gneisses north of 60°N indicates high P-T conditions in their cores (P≈8 kbars, T≈700-800°C) and rims with conditions of P≈6-7 kbars and T≈600-700°C (F. Mengel, unpub. data).

Charnockitic rocks within and northwest of the Abloviak shear zone contain equigranular orthopyroxene-hornblende± garnet±clinopyroxene assemblages, for which geothermobarometry suggests P≈8-9 kbars and T≈800°C (F. Mengel, unpub. data). Sillimanite is the stable aluminosilicate in metasedimentary rocks. A U-Pb age of 1843 ± 3 Ma on a metamorphic zircon overgrowth from the 1895 Ma charnockite sample is suggested as a minimum age for granulite-facies metamorphism (Scott et al., 1993).

STRUCTURAL GEOLOGY

The map area contains the following principal structural elements, from east to southwest (Fig. 2):

1. Archean gneisses with east to north-northeast striking Avayalik dykes unaffected by Paleoproterozoic deformation;
2. The north-south Komaktorvik zone of Paleoproterozoic sinistral, east-side-up shear deformation at amphibolite facies;
3. Magmatically and tectonically interlayered Archean gneisses and DTG suite rocks within a region of large-scale, north-trending folds, including the bend in the Abloviak shear zone;
4. Foliated charnockitic rocks with northwest-trending fold axial planes;
5. A zone of dextral mylonite and ultramylonite located along the northern margin of the bend in the Abloviak shear zone;
6. the Paleoproterozoic Abloviak shear zone of granulite-facies sinistral shear.

The most prominent feature of the map area is the triangular region in the southeastern corner, defined by the southeast-striking limb of the Abloviak shear zone and the north-striking Komaktorvik zone. Mapping in this area was aimed at investigating the relationship between these zones and unravelling their structural evolution.

Archean fabric elements

Archean structures in granulite-facies Nain gneisses include a folded centimetre to decametre scale gneissic layering, a north-northwest, vertical foliation, axial planar to folds of the gneissosity, and strong south-southeast-plunging lineations defined by orthopyroxene, quartz, and other mineral aggregates. At least one phase of granitoid magmatism separates the formation of the gneissic layering and of subsequent penetrative foliations and lineations. These structures are cut by both sets of mafic dykes and collectively referred to as D_n.

Paleoproterozoic structures

Four sets of Paleoproterozoic structures were identified:

D_{n+1}: Regional foliations, locally developed into a migmatitic layering with intrafolial isoclinal folds.

D_{n+2}: Structures associated with formation of the Abloviak shear zone.

D_{n+3}: Ductile, amphibolite-facies structures which post-date the Abloviak shear zone.

D_{n+4}: Late, narrow mylonite zones, with predominantly west-side-up kinematics.

D_{n+1}: Regional foliations

D_{n+1} structures within Paleoproterozoic igneous rocks vary from a weak foliation to a tightly-folded gneissosity in strongly deformed migmatites (compare Fig. 3B and C). The charnockitic suite is generally foliated to weakly gneissic, but rarely migmatitic.

In the Tasiuyak gneiss and paragneisses to the north, a migmatitic gneissosity with intrafolial isoclinal folds is the earliest set of structures observed, although it is likely that an earlier set (thrusts, folds?) related to crustal thickening is present (e.g. Van Kranendonk, 1990: N. Goulet, pers. comm., 1992).

D_{n+2}: The Abloviak shear zone

Structures within the 8-15 km-wide Abloviak shear zone include subvertical mylonitic foliations, strong subhorizontal mineral lineations and kinematic indicators of sinistral shear (Fig. 3E, F; 4; 5A-B, E-F). Highly elongate quartz blades have typical aspect ratios of 30-20:5:1. These D_{n+2} fabrics deform compositional and migmatitic layering in the Tasiuyak gneiss and a poorly developed gneissosity in charnockitic rocks. The zone is primarily developed within the Tasiuyak gneiss, but transgresses lithologic contacts; it deforms Rae gneisses in the southern and southwestern parts of the map area (cf. Goulet and Ciesielski, 1990), and Nain gneisses south of, and possibly also within the map area (Fig. 1).

D_{n+2} and/or D_{n+3} linear structures

In the northwestern corner of Figure 4, tight, V-shaped folds of gneissosity that have angular hinges and upright to westerly-inclined axial planes, plunge moderately to steeply to the west (Fig. 5C). A gradation from these steeply west-plunging folds and mineral lineations (sillimanite-biotite-quartz) to the shallow-plunging ribbon lineations of the Abloviak shear zone was observed from north to south across this area, but it is uncertain whether these structures formed contemporaneously during D_{n+2}, or whether the more steeply-plunging elements formed in response to D_{n+3} (see below).

D_{n+3} structures

Grouped under D_{n+3} are several different structural elements, including large-scale folds, shear fabrics within the Komaktorvik zone, belts of amphibolite-facies straight gneiss, and a zone of dextral mylonite. Although not neccessarily contemporaneous, they postdate D_{n+2} and are believed to be related through progressive deformation. Preliminary U-Pb zircon and titanite geochronology indicate that deformation at amphibolite facies within the Komaktorvik zone occurred at between 1.79-1.71 Ga (Scott, in press).

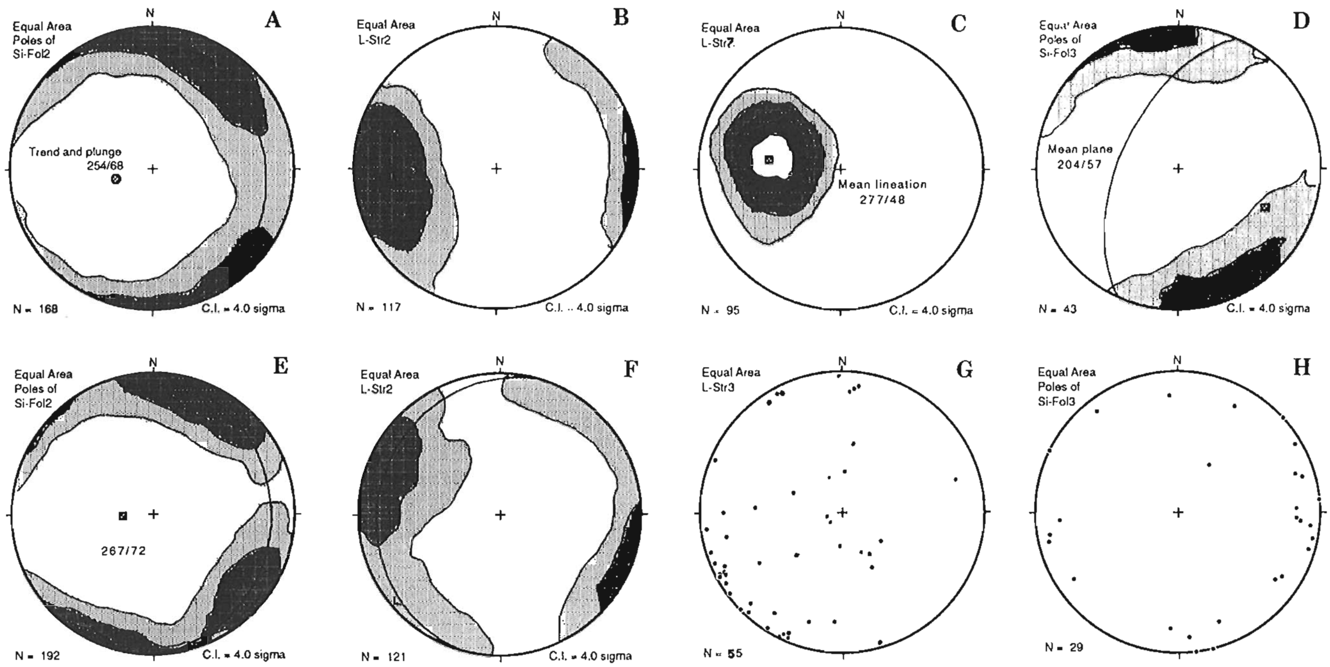

Figure 5. Stereoplots of D_{n+2} and D_{n+3} structures from the area of the bend in the Abloviak shear zone. **A**)-**D**) are from the western half of Figure 4, whereas **E**)-**H**) are from the eastern half. Plots **A**)-**F**) have a contour interval of 4σ. **A**) and **E**): D_{n+2} foliations in the Abloviak shear zone, showing their folded nature around steeply west-plunging poles; **B**) and **F**): D_{n+2} lineations of the Abloviak shear zone, with maximum concentrations at 20→260 for **B**) and 02→294 for **F**); **C**): D_{n+2} lineations from the northwest part of Figure 4, north of the Abloviak shear zone; **G**): D_{n+3} lineations and fold axes; **D**) and **H**): D_{n+3} foliations.

Large-scale folds

In the bend of the Abloviak shear zone, mylonitic D_{n+2} structures are deformed by open to tight D_{n+3} folds which locally develop strong elongation lineations on mylonitic foliations within their hinges. West of the D_{n+3} fold axial trace shown in Figure 4, D_{n+3} folds have a consistent Z-asymmetry, whereas immediately east of this, they have an S-asymmetry. This geometry suggests that the bend in the Abloviak shear zone formed as a result of buckling. The orientation of D_{n+3} folds and lineations in this area are concentrated within the southwest quadrant (Fig. 5G): D_{n+3} foliations are generally oriented north-northwest (Fig. 5H, H). The inner arcs of both the northward- and the southward-closing folds of the Abloviak shear zone are marked by brittle faults (Fig. 4) which accommodate intense shortening at these locations.

An oblique fold profile extends northward from the bend in the Abloviak shear zone to Killiniq Island, along the western side of the anorthosite body. Prominently visible on aeromagnetic maps of the region, these folds have south-southwest trending axial traces that merge into the Komaktorvik zone to the east, and shallow to intermediate southwesterly-plunging axes.

Komaktorvik zone

The Komaktorvik zone is a north-south zone of shear at amphibolite facies, characterized by subvertical, proto-mylonitic foliations, south-southeast plunging lineations, tight folds, and kinematic indicators of east-side-up, sinistral shear (Van Kranendonk and Scott, 1992). Deformation is concentrated within Archean gneisses along the eastern margins of the relatively competent meta-anorthosite and the charnockitic rocks north of 60°N. The eastern margin of the zone is well defined along the transition between granulite-facies Archean gneiss with undeformed Avayalik dykes, and amphibolite-facies gneisses with penetrative Paleo-proterozoic foliations and deformed dykes (Fig. 2). However, the western boundary of the zone is not well defined, as deformation south of 60°N splays into southwesterly-trending zones of straight gneiss.

Straight gneiss belts

On the southwestern limb of large-scale, D_{n+3} folds, belts of amphibolite-facies straight gneiss, ≤150 m wide, are typified by a straight centimetre-scale gneissosity and granoblastic, equigranular textures (Fig. 6A). They postdate the formation of a migmatitic layering in some diorite-tonalite-granite rocks, and typically occur along margins of thin units of layered amphibolite located between panels of Archean gneiss and diorite-tonalite-granite suite rocks (Fig. 2, 4). Very high strain features such as sheath folds (Fig. 6B) suggest significant translation of panels across these zones, which vary from steeply northwest-dipping to moderately north and northeast-dipping. All zones contain moderately-plunging lineations trending between 240-280°, subparallel to D_{n+3} folds and lineations. Rare kinematic indicators show reverse, southward movement and, locally, a component of sinistral shear.

Dextral mylonite

Along the northern margin of, and cutting D_{n+2} fabrics within the Abloviak shear zone, is a 500 m to 2 km wide zone of amphibolite-facies dextral mylonite and ultramylonite (Fig. 4, 6C, D). Rocks within this zone contain a penetrative, subvertical schistosity and moderately-plunging mineral lineations and fold axes (Fig. 4). The mylonite transgresses the lithologic boundary between the interlayered rocks of the DTG suite and Archean gneisses, and the association which includes charnockitic rocks, quartz diorites, Tasiuyak gneiss, and paragneiss (Fig. 4). The dextral mylonite is widest where it strikes east, narrowing both to the southeast and southwest.

Northwest of the dextral mylonite, northwest-striking brittle faults with a dextral sense of displacement (Fig. 4) pass into a north-striking zone of cataclastic fault breccia which probably continues to the northwest for a distance of 45 km (Fig. 2; N. Goulet, pers. comm., 1992). These faults are considered to be related to formation of the dextral mylonite because they do not crosscut it and form within the dilational field of the zone as predicted from theoretical studies of discontinuous faults (e.g. Burgmann and Pollard, 1992).

D_{n+4}: Late mylonite zones

Narrow, north-northwest-striking zones of west-side-up mylonite and ultramylonite cut D_{n+3} structures in the Komaktorvik zone. These mylonites are typically $\leq$5 m wide, but rarely form zones up to 50 m wide.

Ductile mylonitic structures within both the Abloviak shear (D_{n+2}) and dextral mylonite zones (D_{n+3}) are cut by brittle faults that show transcurrent displacements and by pseudotachylite veins. Such brittle faults have a sinistral sense of displacement both along the inner arc of the southeastern limb of the bend in the Abloviak shear zone and in the southwest-striking portion of the dextral mylonite zone (Fig. 6E).

PROTEROZOIC TECTONIC EVOLUTION

The map area is characterized by a widespread suite of Paleoproterozoic meta-igneous rocks (diorite-tonalite-granite suite) emplaced within the Nain Province between ca. 1.91-1.86 Ga. This suite contains a similar range of rock types

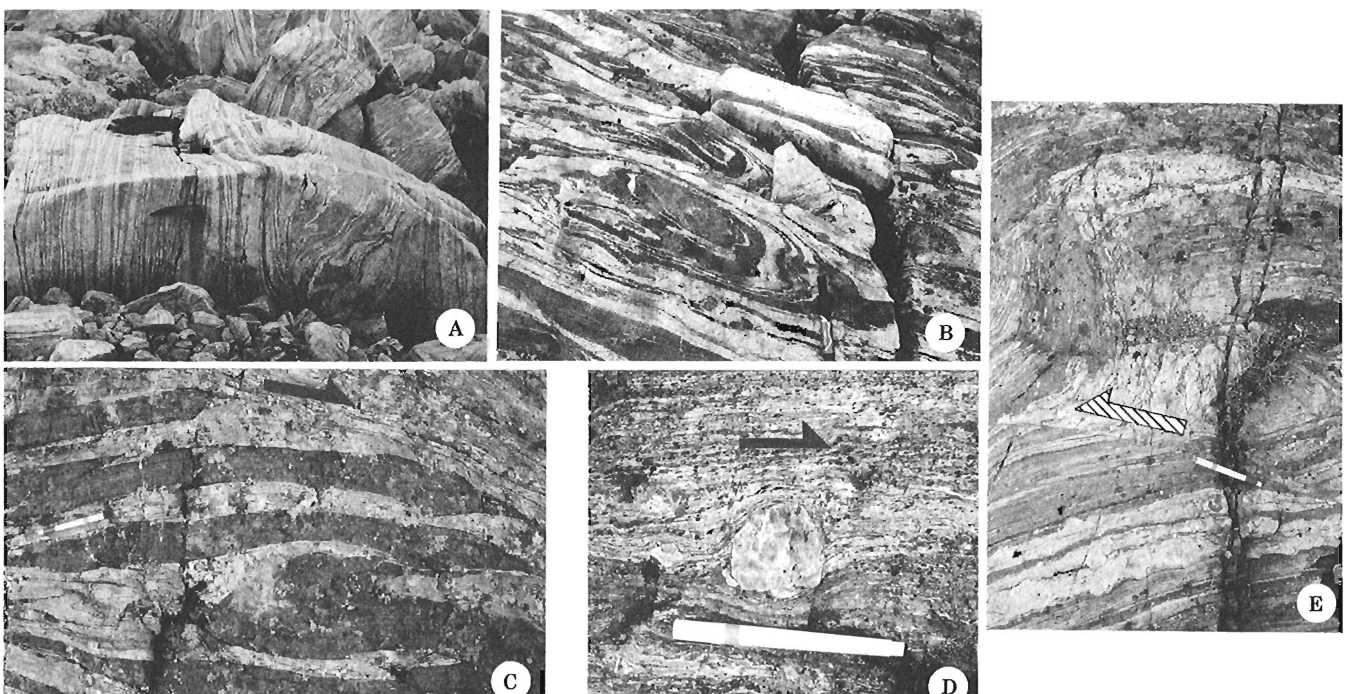

Figure 6. A) View southwest of subvertical straight gneiss in Proterozoic tonalite migmatite; hammer is 30 cm long (GSC 1992-239D); **B)** Sheath folds within tonalite gneiss along the high-strain contact with layered amphibolite, north of the bend in the Abloviak shear zone: marker is 14 cm long (GSC 1992-239Q); **C)** Dextral strike-slip duplex of mafic dyke in Nain gneisses, dextral mylonite zone (GSC 1992-239L); **D)** Rotated K-feldspar porphyroclast in dextral mylonite zone (GSC 1992-239K); **E)** Sinistral strike-slip fault and fold overprinting dextral mylonitic fabrics (GSC 1992-239M). Pen in **C)**-**E)** is 13 cm long.

as calc-alkaline arc suites of Phanerozoic to Archean age worldwide and is therefore considered to represent a magmatic arc which was emplaced within, and perhaps in part marginal to, the Nain Province. The relationship of this suite to the charnockitic rocks which were emplaced within the Tasiuyak gneiss and Rae Province remains enigmatic. Further mapping and chemical analyses will hopefully resolve this question.

The formation of regional structures postdated development of both the diorite-tonalite-granite and charnockite suites. Based on regional considerations, D_{n+1} structures are possibly related to Nain-Rae continental collision, dated as ca. 1.86 Ga in the area west of Okak Bay (Bertrand et al., 1992). These early structures were deformed within the Abloviak shear zone during D_{n+2}, dated as between 1845-1822 Ma in the Okak area (Bertrand et al., 1992), an age similar to that of granulite-facies metamorphism in charnockite in the map area.

Abloviak shear fabrics were folded during D_{n+3}. The west-northwest-striking arm of the Abloviak shear zone is characterized by Z-folds of the D_{n+2} fabrics, suggesting a period of dextral shear subsequent to sinistral Abloviak deformation. Dextral D_{n+3} shear also occurs within the 0.5-2 km wide zone along the northern margin of the Abloviak shear zone. This dextral mylonite thins to the southeast and roots near Komaktorvik Fiord, also the apparent root of the sinistral, east-side-up Komaktorvik zone (Fig. 2).

The relative age of V-shaped folds and lineations to the north of the southward-closing fold of the Abloviak shear zone (Fig. 4) is unknown. The gradational nature of the transition from these fabrics into those of the Abloviak shear zone suggests that the former may represent D_{n+2} structures developed along the northern margin of the Abloviak shear zone (cf. Van Kranendonk, 1990). Alternatively, they may have resulted from pinching of units and earlier structures within the inner arc of the southward-closing, D_{n+3} fold. A third possibility is that the folds north of the shear zone represent D_{n+2} structures that were tightened during D_{n+3} deformation.

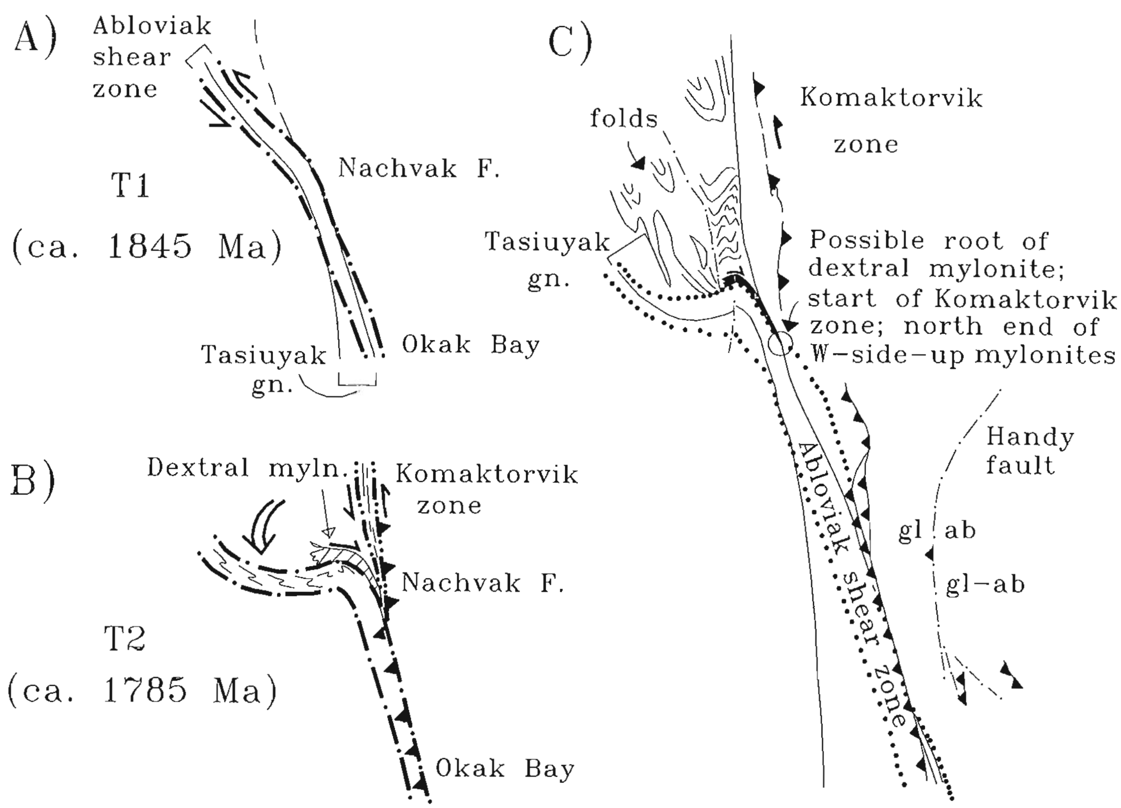

Figure 7. Interpreted structural development of the bend in the Abloviak shear zone and of the Komaktorvik zone, Torngat orogen. **A)** Formation of the Abloviak shear zone (ASZ), at ca. 1845 Ma during D_{n+2}; **B)** Counterclockwise rotation of Burwell domain resulted from northward movement of the Nain Province, which caused the contemporaneous formation of D_{n+3} structures at ca. 1.79-1.71 Ga, including the Komaktorvik zone and straight gneiss belts, folding of the Abloviak shear zone and rocks west of the meta-anorthosite, and the dextral mylonite; **C)** Sketch map of structures within Torngat orogen, showing a possible correlation between D_{n+3} structures in the map area and west-side-up mylonites and the Handy fault to the south: gl = granulite-facies rocks; ab = amphibolite-facies rocks.

In a working hypothesis to be substantiated by further U-Pb geochronology, it is suggested that all D_{n+3} structures resulted from counterclockwise rotation of Burwell domain and the formation of the bend in the Abloviak shear zone (see Fig. 7). In this hypothesis, the D_{n+3} Z-folds are interpreted to have formed in response to rotation of the Abloviak shear zone from an original northwest orientation (Fig. 7A). As rotation progressed, buckling of the Abloviak shear zone commenced, coincident with the formation of the dextral mylonite zone which formed as a slip plane during folding (Fig. 7B). The observation that both the Komaktorvik and dextral mylonite zones terminate around Komaktorvik Fiord strongly suggests that they were coeval. Sinistral movement on the Komaktorvik zone is, at first glance, inconsistent with the hypothesis of Burwell domain rotation if the Nain and Rae provinces were fixed. However, we speculate that northward movement of Nain Province at this time may have caused rotation of Burwell domain and led to the formation of the sinistral, east-side-up deformation in the Komaktorvik zone.

U-Pb dating of rocks within the Komaktorvik zone indicate that they were deformed at the same time (ca. 1.79 Ga) as west-side-up ultramylonites were developed to the south of Nachvak Fiord (Bertrand et al., 1992; Fig. 1). The kinematics of these contemporaneous zones indicate large-scale, scissor-like movement along the eastern margin of the Torngat orogen, pinned at the southern end of the dextral mylonite and Komaktorvik zones (Fig. 7C). We speculate that the scissor movement which occurs across the Handy fault (Bridgwater and Collerson, 1976) may have occurred at this time (Fig. 7C).

ECONOMIC GEOLOGY

Gossan zones of rusty-weathering, sulphide-rich material within, or adjacent to metasedimentary rocks were found in charnockite on the south shore of Killiniq Island. Assay results from one sample revealed 450 ppm Ni and 330 ppm Cu: values of Pt, Pd, and Au were <15 ppb. Analysis of other gossan zones are in progress.

An assay sample of rusty Archean paragneiss from a thin unit (<200 m) east of the meta-anorthositic body and south of Saglarsuk Bay (Fig. 1) showed 600 ppm Ni and 250 ppm Zn.

Sulphide gossans associated with Archean meta-anorthosites are described in Wardle et al. (1992, in press).

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Geological Survey of Canada Project 910034

A preliminary report of U-Pb geochronology of the northern Torngat Orogen, Labrador[1]

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Scott, D.J., Machado, N., Van Kranendonk, M., Wardle, R., and Mengel, F., 1993: A preliminary report of U-Pb geochronology of the northern Torngat Orogen, Labrador; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 341-348.

Abstract: This report summarizes the initial results of an ongoing U-Pb geochronology project in the northern part of the Paleoproterozoic Torngat Orogen, northern Labrador. Voluminous granitoid plutonic rocks range in age from ca. 1.91 Ga to ca. 1.86 Ga. The Komaktorvik shear zone appears to have been active from ca. 1.79 to 1.71 Ga. Granulite facies conditions may have existed in the northern part of the area at ca. 1.84 Ga. Analytical work in progress will contribute to the understanding of the timing of events in this complex area.

Résumé : Le présent rapport résume les résultats préliminaires d'une étude géochronologique (méthode U-Pb) en cours sur des échantillons provenant du nord de l'orogène de Torngat (Labrador) d'âge Protérozoïque précoce. Les roches granitiques les plus répandues dans la région ont livré des âges entre ca 1,91 et 1,86 Ga. La zone de cisaillement de Komaktorvik était probablement active entre 1,79 et 1,71 Ga. Dans la nord de la région, le métamorphisme au faciès des granulites date probablement de 1,84 Ga. Les travaux en cours contribuemont à une meilleure compréhension de la chronologie des événements géologiques relevés dans cette région complexe.

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INTRODUCTION

The Torngat Orogen lies between the western margin of the Nain Province and the eastern margin of the Rae Province, located in northernmost Labrador and northeastern Quebec (e.g. Hoffman, 1990; Van Kranendonk et al., 1993). The northern part of the orogen preserves, from east to west, pristine and variably-reworked Archean tonalitic gneisses (Schiøtte et al., 1990) and anorthosites, both intruded by a previously unrecognized Paleoproterozoic suite of dioritic, quartz-dioritic and tonalitic, and rare granitic intrusive rocks which are referred to as the DQT suite (Van Kranendonk and Scott, 1992; Wardle et al., 1992a). Further west are the extensive orthopyroxene-bearing tonalitic- to granodioritic "charnockitic" rocks of the Burwell batholith (Taylor, 1979). Previous examinations of the study area have been limited to reconnaissance-scale operations (Taylor, 1979; Wardle, 1983) or transects (Wilton and Wardle, 1990). Detailed U-Pb geochronological studies undertaken to the south of the present area indicate that much of the deformation in that part of the orogen occurred during the period 1.86-1.78 Ga (Bertrand et al., 1992). Prior to the present study, no precise U-Pb ages exist for rocks in the map area, and as such, much detailed work remains to be done (e.g. Scott, in press).

Recent regional geological mapping in the study area (Van Kranendonk and Scott, 1992; Wardle et al., 1992a) has revealed important structural and magmatic differences relative to the southern parts of the orogen. A regionally extensive belt of strongly reworked Archean tonalitic gneisses and anorthosites, referred to as the Komaktorvik zone (Korstgård et al., 1987), shows evidence of east-side-up dip-slip movement (Van Kranendonk and Scott, 1992), which has not been recognized in the south. Neither the absolute timing of movement within the Komaktorvik zone, nor its relationship to deformation in the Abloviak shear zone to the south were known prior to the present mapping (Van Kranendonk and Scott, 1992; Wardle et al., 1992a). Rocks of the diorite-quartz diorite-tonalite suite intrude the western margin of the Nain Province in the study area (Van Kranendonk and Scott, 1992; Wardle et al., 1992a), but have not been recognized in the Nain Province in the southern part of the orogen. The genetic relationship between the diorite-quartz diorite-tonalite suite and the Burwell charnockitic rocks in the west is uncertain. The DQT suite and/or the Burwell charnockitic rocks may represent a composite, subduction-related continental magmatic arc (Van Kranendonk and Scott, 1992; Wardle et al., 1992a). This speculative interpretation is based only on compositional range observed in the field, and must be tested geochemically and isotopically. Knowledge of the age range of this suite is important to our understanding of the evolution of the orogen.

The present U-Pb study has two principal objectives: 1) to characterize the voluminous meta-igneous rocks which occur in the map area; and 2) establish an absolute framework of the timing of deformation. Samples representative of the various metaplutonic rocks and those which appear to have been intruded during ongoing deformation were collected the first season (1991) of regional mapping in an area of approximately 1500 km^2, in order to further these objectives. The purpose of the present contribution is to introduce the reader to the complexity of the tectonomagmatic history and the U-Pb systematics in the area, based on a discussion of preliminary analyses from these samples. As such, they may pose more questions than they resolve. Additional samples collected in 1991 but not yet analyzed, combined with samples collected during 1992, will build on this preliminary dataset. A more complete presentation of data and interpretation will be published elsewhere.

ANALYTICAL METHOD

The analytical methods used here are essentially those of Krogh (1973, 1982), as followed at the GEOTOP U-Pb Laboratory at the University of Quebec at Montreal (Machado et al., 1991), involving magnetic sorting of grains using a Frantz isodynamic separator, hand-picking of highest-quality grains, air abrasion, and sample dissolution in Teflon digestion vessels with addition of a mixed ^{205}Pb-^{233}U-^{235}U tracer spike solution. Analyses were carried out on a VG Instruments Sector mass spectrometer equipped with a Daly detector.

U-Pb RESULTS

Timing of igneous activity

1. Tonalite, Tellialuk Arm (VOD-135-91)

Leucocratic hornblende-biotite tonalite is an important component of the diorite-quartz diorite-tonalite suite. The foliated to locally gneissic tonalite intrudes tonalitic orthogneisses of the Nain Province which preserve retrograde, amphibolite-facies metamorphic assemblages, and sequences of metasedimentary rocks (Wardle et al., 1992a; Van Kranendonk and Scott, 1992). Xenoliths of tonalitic gneiss similar in appearance to that of the Nain Province are common near the margins of intrusions. In a lichen-free outcrop several kilometres west of the western end of Tellialuk Arm (locality 1, Fig. 1), fine- to medium-grained leucocratic tonalite intrudes darker, fine- to medium-grained porphyritic diorite. The sample analyzed was selected from a portion of the outcrop free of recognizable xenoliths. Hornblende and biotite comprise less than 10% of the rock, and minor amounts of apatite and magnetite are accessory phases. Whereas both the tonalite and diorite were sampled, to date only the tonalite has been analyzed.

The majority of zircon grains separated from this sample are metamict and of low quality, the only high-quality zircon occurring as colourless euhedral overgrowths on cloudy metamict cores. Four such crack- and inclusion-free euhedral overgrowths were removed from their respective cores, heavily abraded (>66 h @ 3 psi), and analyzed separately. Two of the overgrowths are concordant at 1909 ± 2 Ma and 1910 ± 2 Ma, a third overgrowth is 0.8%, with a ^{207}Pb/^{206}Pb age of 1909 Ma. The fourth overgrowth is 0.9% discordant, and has a ^{207}Pb/^{206}Pb age of 1920 Ma. The fourth grain, optically indistinguishable from the others, contains an older

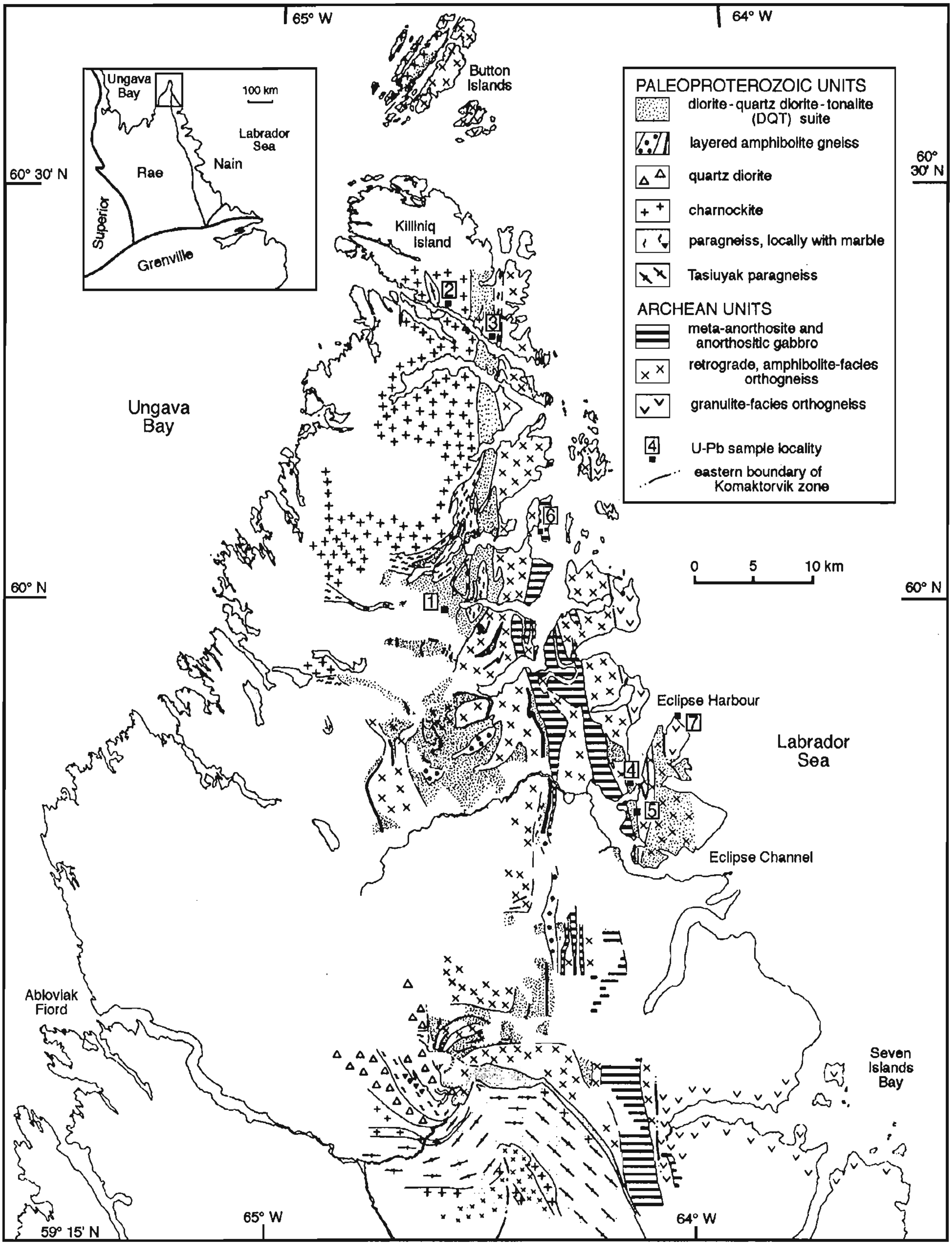

Figure 1. Geological map of the northern part of the Torngat Orogen (modified from Van Kranendonk et al., 1993). Numbers within boxes refer to locations of samples described in text.

component of radiogenic lead, which may represent unrecognized core material which was not removed during hand-picking or subsequent abrading.

The crystallization age of the the overgrowths which yielded concordant points is 1910 ± 2 Ma, interpreted as the crystallization age of the rock. As this tonalite body intrudes Nain province tonalitic gneisses and locally contains gneissic xenoliths, much of the metamict zircon observed in this sample may have been inherited; this remains to be tested by direct analysis of the cores. The possibility that the overgrowths represent a post-crystallization metamorphic event cannot here be ruled out. As such, 1910 ± 2 Ma should be considered as a minimum age of crystallization for this sample.

2. Charnockite, Killiniq Island (VOD-124-91)

Buff- to brown-weathering orthopyroxene-bearing granitoid rocks which are weakly- to moderately foliated underlie much of the northwestern part of the map area. These distinctive rocks, which range in composition from tonalite to granite, have been termed "charnockite" as a general mapping term (Taylor, 1979; Wardle et al., 1992a; Van Kranendonk et al., 1993). Small mafic inclusions in various stages of assimilation, xenoliths of tonalitic gneiss, and rafts of metasedimentary rocks are common in the charnockitic rocks. A representative, leucocratic sample of granodioritic composition was selected from the southern shore of Killiniq Island along McLelan Strait (locality 2, Fig. 1). The sample contained orthopyroxene, clinopyroxene, garnet, abundant magnetite, and trace amounts of zircon.

A wide variety of shapes and sizes of variable-quality zircon was present in each of the magnetically separated fractions. Dominant throughout were small (20-40μm), well-faceted needle-like grains. A crack- and inclusion-free selection of these grains which did not display visible core-overgrowth relationships were abraded, and three fractions were analyzed. Each of the analyses are concordant within two-sigma analytical error, yielding ages of 1894, 1895, and 1896 Ma. Some of the larger grains in each of the magnetically separated fractions consisted of cloudy and cracked, partially metamict cores, surrounded by colourless, inclusion-free euhedral overgrowths. One of these overgrowths, removed from its core and heavily abraded, was concordant at 1843 ± 3 Ma.

The crystallization age of the needle-like grains which are the dominant morphotype in this sample is 1895 ± 3 Ma, the average age of the three concordant analyses. The well-faceted nature of these grains, absence of internal core-overgrowth features, and their abundance in this sample lead us to take 1895 Ma as the age of crystallization of the granodiorite. The 1843 Ma overgrowth may represent a post-crystallization metamorphic event, possibly related to the presence of orthopyroxene in this sample, and indicative of granulite-facies conditions. The wide variety of metamict and rounded grains present in this sample suggests that they may be inherited; this is consistent with the observed presence of metasedimentary and tonalitic gneiss xenoliths.

3. Megacrystic granite, Killiniq Island (VOD-122-91)

Distinctive pink-weathering, decametric sheets of garnet-bearing potassium-feldspar megacrystic granite occur along the eastern margin of the charnockitic rocks and the western limit of reworked Nain gneisses (Wardle et al., 1992a; Fig. 2), and are thought to be part of the diorite-quartz diorite-tonalite suite. These intrusions range from mildly foliated to mylonitic, generally have sheared margins, and can be traced along strike for several kilometres. A sample was collected from a weakly foliated sheet which intrudes an orthopyroxene- and clinopyroxene-bearing tonalite associated with the charnockitic rocks on the south shore of Killiniq Island (Locality 3, Fig. 1). In addition to zircon, this sample contained trace amounts of apatite and magnetite.

The majority of zircon grains in each of the magnetic fractions separated from this sample were medium- to dark-brown and well-faceted. Variably metamict subhedral to anhedral light-coloured grains were visible as cores in virtually all of the brown zircon grains. Four brown overgrowths which did not contain visible inclusions or fractures were removed from their cores prior to heavy abrasion (up to 100 h); each of the grains was analyzed separately. One was concordant at 1864 ± 2 Ma, the remaining three were <1% discordant, and yielded a range of $^{207}Pb/^{206}Pb$ ages; 1791, 1795, and 1850 Ma, the age obtained increasing with total time of abrasion. These four analyses regress to yield a discordia with an upper intercept of 1877 +19/-13 Ma, a lower intercept of 1566 ± 55 Ma, and an overall probability of fit of 92% (Davis, 1982). The large errors on the intercepts are due to the extremely shallow angles of intersection between the discordia and concordia.

The widespread occurrence of these overgrowths and their euhedral nature suggests that they may have grown during crystallization of this rock. The apparently excellent fit of the four analyses to a discordia suggests that the effects of recent lead-loss have been successfully removed (Krogh, 1982). The rock may have crystallized at 1877 Ma, the zircons then subjected to a lead-loss event roughly 300 Ma later. In this case, the concordance of the overgrowth at 1864 Ma may not be sufficient evidence to interpret this as the crystallization age of the rock; 1864 ± 2 Ma should however be regarded as the minimum age of the brown zircon overgrowths. Analyses of additional overgrowths are required to more fully understand this sample.

4. Mafic diorite, Eclipse Island (VOM-71B-91)

Sheet-like bodies of mafic diorite to tonalite commonly occur within tonalitic Nain orthogneiss, reworked at amphibolite-facies conditions within the Komaktorvik shear zone. The intrusions range from metre-scale sheets to map-scale intrusive bodies (Van Kranendonk and Scott, 1992), and are included in the diorite-quartz diorite-tonalite suite. The sample is from a body of mafic diorite several hundreds of metres across that is locally cut by swarms of centimetre-wide, white granitic veins (see Fig. 3d of Van Kranendonk and Scott, 1992). The diorite intrudes deformed orthogneiss at the southern end of Eclipse Harbour (locality 4, Fig. 1). Zircon, magnetite, and titanite were separated from this sample.

A wide range of zircon grain shapes are present. Most are white and metamict, and very few of these have well-developed crystal faces. Light-brown zircon showing euhedral habit development occurs as thin overgrowths and tips on many of the colourless and metamict grains, particularly those in the more-magnetic fractions. To date, three of these overgrowths have been analyzed; the most heavily abraded is concordant at 1891 ± 2 Ma, two less-abraded overgrowths have $^{207}Pb/^{206}Pb$ ages of 1883 and 1880 Ma respectively, and are <1% discordant. Titanite extracted from this sample is light-brown; the fraction analyzed is concordant at 1710 ± 4 Ma.

The ubiquitous occurrence of brown zircon in this sample, and its euhedral habit suggests in situ growth. The present preliminary data suggest that the age of crystallization of the brown zircon is 1891 ± 2 Ma; although the preferred interpretation is that this represents the igneous age of the rock, the brown overgrowths may be of metamorphic origin, and thus 1891 Ma must be treated as a minimum age. Titanite growth suggests that the sample underwent retrograde amphibolite-facies conditions at 1710 Ma.

5. Diorite, Eclipse Channel (VOM-12-91)

Similar to the setting described above, this dioritic sample occurs as metre- to decametre-scale sheets in a zone of tonalitic orthogneiss deformed at amphibolite-facies conditions (locality 5, Fig. 1; darker, homogeneous component in Fig. 3e of Van Kranendonk and Scott, 1992). The present sample, considered part of the diorite-quartz diorite-tonalite suite, was collected from a lichen-free outcrop, where a ~10 m wide exposure of diorite was free of visible xenoliths.

A wide variety of euhedral to subhedral colourless zircon grains were recovered from this sample, in addition to honey-brown fragments of titanite. At least three populations of euhedral grains may be defined on the basis of morphology: 1) square-section stubby prisms; 2) flat, hexagonal crystals; and 3) hexagonal-section elongated prisms which have minor surface embayments. Three multiple-grain fractions of square-section prisms were analyzed, their $^{207}Pb/^{206}Pb$ ages increasing from 1830 Ma to 1859 Ma with increasing duration of abrasion, up to 70 h. The fraction of flat, hexagonal prisms analyzed has a $^{207}Pb/^{206}Pb$ age of 1847 Ma. Two fractions of hexagonal-section elongated prisms abraded for similar amounts of time but containing 13 and 24 grains have $^{207}Pb/^{206}Pb$ ages of 1846 and 1856 Ma respectively. Each of these six analyses is ~1% discordant. A selection of ~40 titanite fragments free of visible inclusions are 0.8% discordant at 1750 Ma. The six zircon analyses and the titanite define a discordia with an upper intercept of 1997 +47/-38 Ma, a lower intercept of 1714 +23/-29 Ma, and a total probability of fit of 95% (Davis, 1982).

The crystallization age of this sample is not immediately apparent from the present analyses. The apparently excellent fit of the seven analyses to a discordia suggests that the rock may have crystallized at 1997 Ma, the zircons and the titanite subsequently subjected to a lead-loss event(s), between ca. 1750 and 1714 Ma. The large errors on the intercepts are the result of the shallow angle between the discordia and concordia. It is interesting to note that the previous sample (VOM-71B-91), located 2.5 km from the present sample, contains titanite with an age of 1710 ± 4 Ma. A minimum crystallization age of ca. 1859 Ma is suggested by the oldest, square-prism zircon fraction. Further work is required to determine the crystallization age of this sample.

6. Anorthosite, Hutton Peninsula (VOD-132-91)

A distinctive feature of the western margin of the Nain Province in the map area is the presence of an anorthositic body, up to several kilometres wide, and at least 100 km along strike (Wardle et al., 1992b). Similar rocks occur as tectonically isolated blocks in orthogneiss along the eastern margin of the orogen up to 200 km south of the present area (Ermanovics et al., 1989), and may be related to this less dissected body. Field evidence suggests that these rocks may be Archean in age; tonalitic partial-melt segregations containing orthopyroxene are cut by amphibolite-facies mafic dykes interpreted as Paleoproterozoic in age, all affected by Proterozoic deformation (Wardle et al., 1992a). In order to test this hypothesis, a sample of leucocratic, clinopyroxene- and garnet-bearing gabbroic anorthosite and material from a tonalitic melt segregation were collected from the Hutton Peninsula (locality 6, Fig. 1). The melt segregation did not yield sufficient high-quality zircon for analysis.

Very little zircon was recovered from the ~60 kg gabbroic anorthosite sample. A wide assortment of shapes of dominantly colourless grains were accompanied by light brown, square-section prisms with up to 10:1 aspect ratios. As these peculiar grains were the only euhedral form in the entire population, seven inclusion-free examples were selected for analysis. After abrasion, the grains were divided into two fractions for analysis. The smaller fraction is concordant at 1779 ± 2 Ma, the larger has a $^{207}Pb/^{206}Pb$ age of 1781 Ma and is 0.7% discordant. A third fraction, consisting of faint pink, stubby, subhedral prisms with well-developed faces and smooth, concave embayments, possibly indicating resorption, is 0.7% discordant, with a $^{207}Pb/^{206}Pb$ age of 1748 Ma.

Field relationships (see above) suggest that the age determined for the elongated euhedral grains, 1779 ± 2 Ma, is unlikely to represent an igneous crystallization age. The most reasonable interpretation of this age is that it may date the metamorphism and/or deformation at amphibolite facies. The ca. 1748 Ma age of the third fraction may be indicative of subsequent metamorphism. Additional samples of melt segregations within the anorthosite, and of a granitic vein which crosscuts these segregations but is cut by a mafic dyke were collected during 1992 fieldwork in order to provide a minimum emplacement age for the anorthosite, and more fully understand its metamorphic history.

Timing of deformation

A considerable effort was made during fieldwork to identify intrusive rocks which appear to have been intruded synchronously with ongoing deformation, and which might readily be dated using the U-Pb system in accessory minerals. The principal targets were ~1 m wide granitic veins in wave-washed coastal exposures where crosscutting relationships with their deformed host-rocks could unequivocally be determined. Two such samples were collected during 1991 fieldwork. In both cases, the granitic veins transect amphibolite-facies mylonitic fabrics, but are themselves deformed by ductile deformation processes. The veins have sharp intrusive contacts (i.e. do not have strain gradients at their margins), and planar and linear fabric elements in orientations identical to their deformed host-rocks. This is thought to indicate that although the veins intruded after a strong deformation fabric had been developed in the host rocks, deformation continued and imparted identical fabric elements to the granitic veins. Although it cannot be unequivocally demonstrated that deformation of the veins did not occur long after they were intruded during renewed deformation, the strong physical similarity of deformation fabrics and their orientations in both the host-rocks and the veins is consistent with the interpretation that the veins were intruded relatively late into an ongoing, continuously deforming zone. Analytical results for the first of these two samples are described below.

7. Granitic vein, Mt. Bache Point (VOM-21-91)

A granitic vein interpreted as synkinematic is located on the headland of Mt. Bache Point (locality 7, Fig. 1). The <0.5 m vein is a medium- to coarse-grained granite which weathers pink to white and can be traced along the coastal outcrop for >20 m. It truncates amphibolite-facies mylonitic fabric (~340°/80°, well-developed down-dip stretching lineation) in the host tonalitic orthogneiss, and is itself folded and necked (see Van Kranendonk and Scott, Fig. 4a). It is moderately foliated, and contains a well-developed mineral lineation in a similar orientation to the stretching lineation in the host tonalitic gneisses. A narrow, folded mafic dyke also truncates the mylonitic fabric in the tonalitic orthogneiss, but is cut by the granitic vein.

The zircon separated from this sample comprises three physically distinct types. Most numerous are small (20-40 μm), equant, colourless subhedral grains that are largely free of inclusions. Cloudy, metamict grains, larger and more varied in shape, are also common. In some of the lower-quality, more magnetic fractions, thin brown overgrowths were commonly observed on all other grain types. Three abraded, multiple-grain populations of the gem-like colourless zircons without visible overgrowths, and three individual brown overgrowths were analyzed. The three brown overgrowths define a discordia with an upper intercept of 1719 ± 2 Ma and a lower intercept of 56 Ma. The gem-like colourless zircons are discordant, and have $^{207}Pb/^{206}Pb$ ages which range from 2685 to 2815 Ma.

The age of crystallization of the brown zircon, 1719 ± 2 Ma, is interpreted as the age of emplacement of this vein. No evidence was found of zircon growth between 1719 Ma and 2685 Ma, nor of resetting after 1719 Ma. The colourless, subhedral gem-like grains are interpreted as xenocrysts. Using 1719 Ma as a lower intercept to construct discordia lines through the youngest and oldest inherited grains, the upper intercepts are 2.78 to 2.87 Ga, interpreted as ages of inheritance.

DISCUSSION

The results presented here are summarized in Figure 2. Based on the present, limited data set, Proterozoic magmatic activity ranges from 1910 Ma to ca. 1859 Ma. These limits are

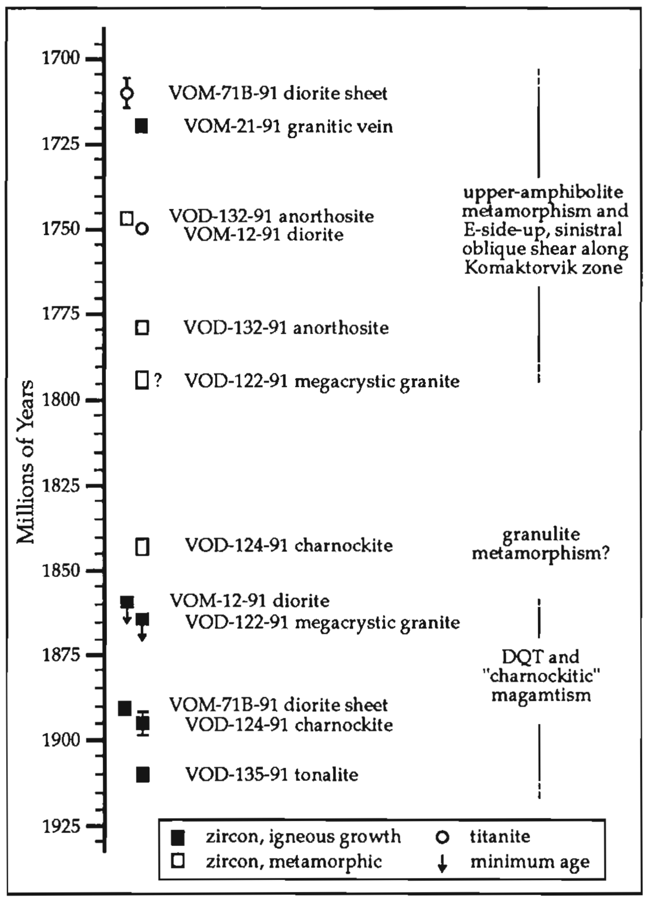

Figure 2. Summary of U-Pb geochronological data for the northern Torngat Orogen. Samples and the significance of their ages are described in text.

minima, as rocks of the Proterozoic suite are known to both predate and postdate the samples which have thus far been analysed. These rocks are up to 50 Ma older than other Proterozoic igneous rocks in the Labrador segment of the orogen, south of the present area (Bertrand et al., 1992). They are significant in that extensive intrusion of the Nain Province by similar igneous rocks has not been observed elsewhere. This may be indicative of a distinct tectonic regime in the present area.

An outstanding question remains as to the nature of the relationship between the Proterozoic diorite-quartz diorite-tonalite suite and rocks of the extensive "charnockitic" Burwell batholith. The present preliminary results suggest at least partial overlap in the timing of intrusion of these distinctive rocks, indicating that they are coeval. It is not known to what extent the present single age determination, 1895 ± 3 Ma (VOD-124-91), represents the wide variety of granitic igneous rocks which comprise the western part of the map area. Further insight into these rocks awaits geochemical and isotopic characterization of both the charnockitic and diorite-quartz diorite-tonalite rocks, and additional age determinations.

The timing of initiation of the dominantly east-side-up, sinistral deformation associated with the Komaktorvik zone is not well-defined by the present data. Numerous samples contain zircon interpreted as metamorphic in origin ca. 1.75-1.79 Ga (Fig. 2), suggesting that deformation may have been underway by ca. 1.79 Ga. A granitic vein (VOM-21-91), interpreted as syntectonic with amphibolite-facies deformation, is 1719 ± 2 Ma. If the field interpretation of this rock is correct, then deformation within the Komaktorvik zone occurred over a period of up to 70 Ma. It is not known whether this deformation was continuous or episodic, however samples with ages in the range ca. 1.71 to 1.79 Ga support the idea of extensive tectonic activity in the Komaktorvik zone throughout this interval. It is interesting to note that this deformation is in part younger than that documented in the southern part of the orogen (Bertrand et al., 1992), where rocks associated with late, west-side-up vertical movement near the eastern orogenic front have ages of ca. 1.78 Ga. These preliminary data suggest that ductile deformation within the Komaktorvik zone was in part synchronous, but outlasted comparable deformation in the southern part of the Torngat Orogen.

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Geological Survey of Canada Project 910034AP

Preliminary report on the geology of the Mugford Group volcanics, northern coastal Labrador[1]

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Hamilton, M.A., 1993: Preliminary report on the geology of the Mugford Group volcanics, northern coastal Labrador; in Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 349-357.

Abstract: The Mugford Group is one of three principal Lower Proterozoic supracrustal successions in northern Labrador, deposited unconformably on amphibolite-facies Archean orthogneisses of the Nain craton. The group consists of basal siliciclastic sediments overlain by pillowed basalt, tuff, and agglomerate. Detailed sampling for petrological, geochemical, and Nd, Sr, and Pb isotopic investigations was carried out in pillowed and massive flows and breccias of the lower volcanic unit, in crystal-rich and mafic members of the middle tuff, and in pyroclastic agglomerates, breccias, and massive flows in the upper volcanic unit. Pegmatitic zones in gabbroic sills present in the basal sedimentary units were sampled for U-Pb geochronology in order to determine a precise age for early Proterozoic volcanism in northern Labrador. An ultramafic sill is found to be more laterally extensive than previously known. A possible minor vent source is identified high in the lower volcanic unit.

Résumé : Le Groupe de Mugford, l'une des trois principales successions supracrustales du Protérozoïque inférieur du Labrador septentrional, a été mis en place en discordance sur des orthogneiss archéens du craton de Nain, situés dans le faciès amphibolitique. Le groupe se compose à la base de sédiments silicoclastiques recouverts de basaltes en coussins, de tufs et de brèches volcaniques. On a effectué, en vue d'analyses pétrologiques, d'analyses géochimiques et d'analyses isotopiques de Nd, Sr et Pb, un échantillonnage détaillé de coulées, aussi bien en coussins que massives, et de brèches situées dans l'unité volcanique inférieure, de membres riches en cristaux et éléments mafiques du tuf intermédiaire, et de brèches pyroclastiques, brèches volcaniques et coulées massives situées dans l'unité volcanique supérieure. Les zones pegmatitiques de filons-couches gabbroïques présents dans les unités sédimentaires basales ont fait l'objet d'échantillonnages, en vue d'études géochronologiques par la méthode U-Pb qui aideront à dater avec précision les épisodes de volcanisme du début du Protérozoïque dans le nord du Labrador. On a constaté qu'un filon-couche ultramafique s'étendait latéralement beaucoup plus loin qu'on ne croyait et on a confirmé la présence d'une cheminée volcanique d'importance mineure à un niveau élevé de l'unité volcanique inférieure.

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INTRODUCTION

This report summarizes the results of a first season of field mapping and geochemical sampling aimed at investigating the petrology, age and geochemical characteristics of the basic volcanic rocks of the early Proterozoic Mugford Group (Fig. 1). Geological mapping was done at 1:50 000 scale in NTS map areas 14E/16, 14F/12, and 14F/13. This program follows a three year mapping project which studied Archean and early Proterozoic crustal evolution in the North River (NTS 14E) and Nutak (NTS 14F) map areas (Ermanovics et al., 1988, 1989; Van Kranendonk and Ermanovics, 1990).

Much of this season's fieldwork was accomplished on Cod and Grimmington islands. Partial helicopter support permitted preliminary regional sampling and reconnaissance on the mainland and the Finger Hill Island sections of the Kaumajet Mountains. Mapping in the 1993 field season will focus primarily on more detailed work in these latter areas.

The earliest geological description of the Mugford Group dates back to Daly's visit to the Kaumajet Mountains in 1900 (Daly, 1902). Work by Coleman (1921), Kranck (1939), Christie (1952), and Douglas (1953) described the general geology of the Mugford Series, named by Daly (who also used the term Kaumajet Mountain Group). Kranck used the term Cape Mugford Beds. This usage was discontinued after a redefinition by Taylor (1970), who suggested the name Mugford Group. Preliminary paleomagnetic work was done by Murthy and Deutsch (1972). Barton (1975) presented the first major and trace element geochemistry on a selection of the lower volcanic members. The most detailed stratigraphic and structural work on the Mugford Group was done by Smyth (1976), whose mapping serves as a foundation for the geochemical, petrological, and age investigations launched with this present study. Smyth (1976) divided the group into a lower sedimentary unit, a lower volcanic unit, a 'middle' sedimentary unit, and an upper volcanic unit. He suggested a maximum thickness of 1225 m for the entire group. The units were then revised slightly and given formation names – the Sunday Run, Cod Island, Calm Cove, Shark Gut, and Finger Hill formations by Smyth and Knight (1978). Reconnaissance mapping in northern Labrador by Taylor (1979) presented further information on the geology of the Mugford Group.

This report constitutes part of a program, funded by the Canada-Newfoundland Cooperation Agreement on Mineral Development, to investigate the volcanic and sedimentary units of the Mugford and other early Proterozoic groups in northern Labrador for their baseline geochemical and economic mineral potential.

GENERAL GEOLOGY

The project area lies within the Nain Province of northern coastal Labrador. Much of this region is underlain by amphibolite- to granulite-facies quartzofeldspathic orthogneiss (and lesser supracrustal gneisses), typical of the northern Labrador segment of the North Atlantic craton which have elsewhere yielded U-Pb ages in excess of 3.8 Ga (Schiøtte et al., 1989).

The Mugford Group is one of three Aphebian supracrustal successions in northern Labrador deposited unconformably on Archean gneisses and migmatites, the others being the Ramah and Snyder groups (Morgan, 1975; Speer, 1982). The Mugford Group is preserved in a graben-like structure bounded on its western side by a series of northwest-striking brittle and ductile faults which bring underlying amphibolite-facies basement in juxtaposition with granulite-facies basement to the west (Ermanovics et al., 1989). Generally, the Mugford Group defines a northwest-trending syncline with relatively shallow dips of limbs (10-40°) and, locally, a shallow northwest plunge (Smyth, 1976; Ermanovics et al., 1989; this work). Almost flat-lying bedding is common in the interior portions of the islands. Steep, axial planar cleavage is visible in many of the shales and tuffaceous and argillaceous sediments, and is related to the large-scale upright folding. East-vergent thrust faults locally offset lower members of the Mugford Group by as much as a few hundred metres (Smyth, 1976; Ermanovics et al., 1989). A west-vergent thrust which repeats the basal unconformity is present on the west side of Finger Hill Island (Smyth, 1976). Another low-angle, west-directed thrust was found this summer in the lower part of the sequence on the southwest side of Mugford Tickle. The estimated offset on this fault is a few tens of metres. Most of the Mugford Group rocks are preserved at subgreenschist grade, but locally reached low greenschist grade in response to the effects of Early Proterozoic Torngat orogenesis, whose front lay some 60 km to the west (Van Kranendonk, 1990). A system of north- to northwest-trending high angle normal faults cuts all levels in the Mugford Group and locally appears to have outlasted all exposed volcanism and earlier thrusting events. Estimates of dip-slip displacement on these faults is on the order of tens of metres (Ermanovics et al., 1989). On the south end of Grimmington Island, multiple normal faulting of this magnitude, spaced at a few hundred metre intervals, is responsible for an irregular apparent basement topography. At the head of Anchorstock Bight, on Cod Island, west-side-up displacement along one of these faults has brought interlayered pillowed and massive lavas of the Calm Cove formation into contact with polydeformed, leucocratic Archean orthogneiss, a feature also noted by Douglas (1953).

The age of deposition of the Mugford Group is still poorly constrained. Although the group unconformably overlies Archean basement which has early Archean age components (>3.46 Ga, Lost Channel; Hurst, 1974), the regional gneisses also record late Archean deformation and metamorphism between 2.7 - 2.8 Ga, as well as the local emplacement to the north and west of the group of posttectonic granites at roughly 2.56 Ga (see Schiøtte et al., 1989, for a review). A suite of early Proterozoic (2.4-2.2 Ga?) tholeiitic diabase dykes, the Napaktok dykes (Ermanovics et al., 1989), intrude the Archean basement, but have not been observed as transgressive to the Mugford Group. Minor deformation of the Mugford Group is ascribed to Torngat Orogeny (1.86-1.78 Ga; Bertrand et al., 1990). The depositional age of the Mugford Group is therefore indirectly constrained between 2.56 and 1.78 Ga, but likely lies in the range 2.4-2.2 Ga. This is in general agreement with a Rb-Sr whole-rock isochron on fresh and altered volcanics and an ultramafic sill from the group reported by Barton (1975), which yielded an age (corrected)

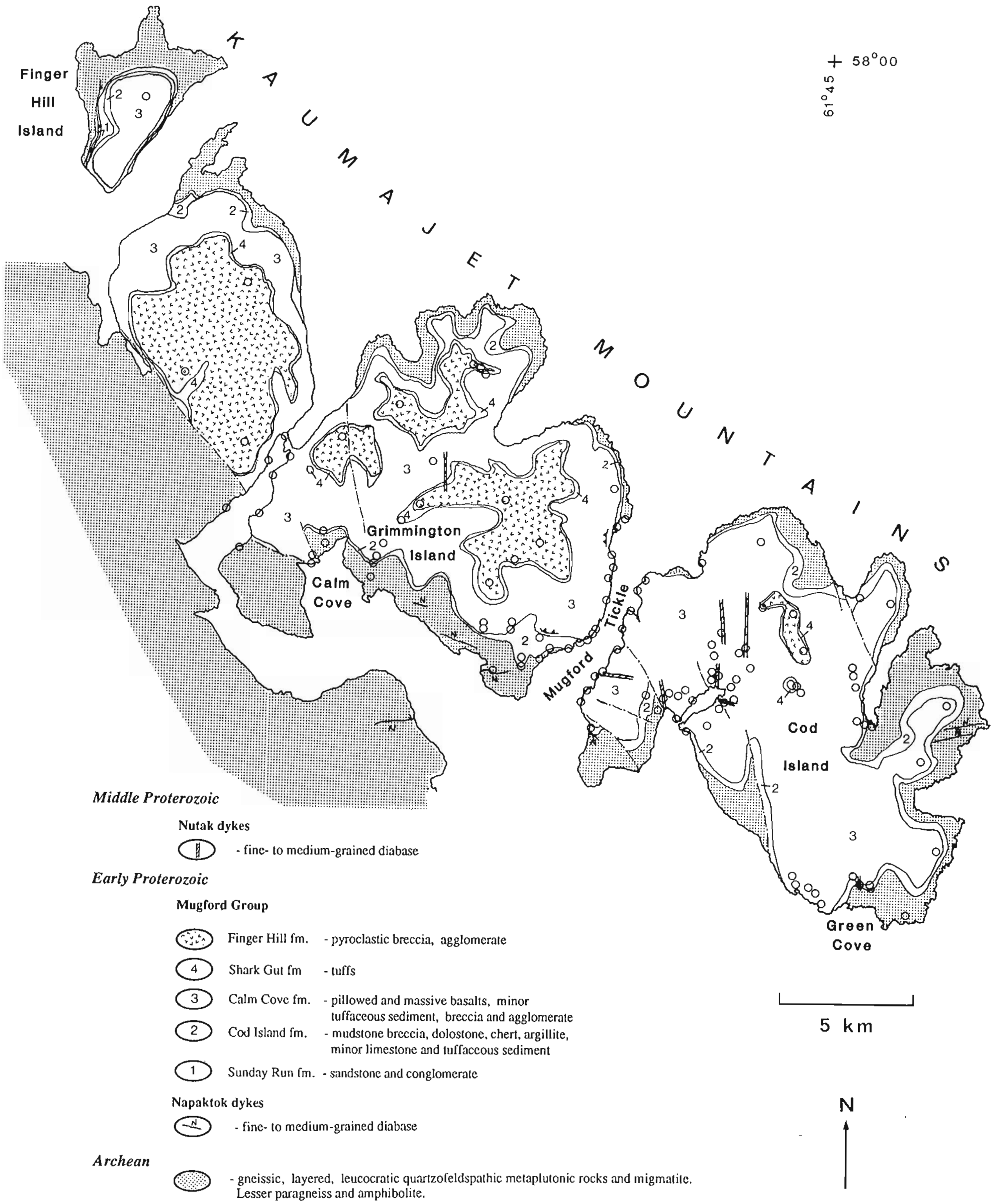

Figure 1. Geological sketch map of the Kaumajet Mountains area, Nain Province, Labrador. Geology of the Mugford Group after Smyth (1976). Principal 1992 sampling areas are shown in open circles.

of 2319 ±54 Ma. The isochron age is heavily weighted on a single analysis of the ultramafic sill and is therefore poorly constrained.

A brief description of the lithologies encountered during the 1992 field season is presented below. Detailed stratigraphic sections through several transects of the group are currently in progress. A more complete account of the lithological variations within each of the Mugford Group formations is provided by Smyth (1976) and Smyth and Knight (1978).

Archean basement

Units of the eastern amphibolite terrane of Ermanovics et al. (1989) comprise dominantly gneissic, layered, leucocratic, pale grey to light pink quartzofeldspathic metaplutonic rocks, and directly underlie the Mugford Group. Sheets, pod-like amphibolite lenses, and minor paragneiss layers are not uncommon in these rocks. Immediately west of the Mugford Group, these gneisses are slightly more migmatitic, and contain a higher proportion of late, deformed, pinkish, granitic pegmatite. Gneissic layering in this area dips moderately to subhorizontally. Minor biotite and hornblende are the chief mafic silicates. Granulite-facies gneisses, described northwest of the Mugford Group by Ermanovics et al. (1988), were not identified in the areas south and east of Lost Channel.

Early Proterozoic mafic dykes

A number of fine- to medium-grained, grey, weakly-deformed diabase dykes (Napaktok dykes; Ermanovics et al., 1989) clearly intrude the Archean orthogneisses. These define a conjugate set which are oriented at 250-260° and 320-330°. Typically, these dykes carry grey to very pale green subequant plagioclase phenocrysts up to 2.5 cm long. The dykes are 5 to 15 metres thick and can often be traced along strike for several hundred metres. Epidote mineralization is frequently visible on joint surfaces.

Mugford Group

Unconformity

Thin regolithic horizons (<5 m) between the basal units of the Mugford Group and the underlying Archean gneisses were best observed on the southeast side of Mugford Tickle and on the north shore cliffs of Cod Island. The regolith has a strongly schistose subhorizontal fabric which is subparallel to the layering in unweathered gneisses stratigraphically below. The unit is pale-orange to buff-weathering, and grades into overlying basal sediments.

Sunday Run formation

Smyth (1976) and Smyth and Knight (1978) have described the oldest unit of the group as a thin (<30 m) sequence of basal grey, well-sorted, crossbedded sandstones, gritty sandstones, pebble and cobble conglomerates, in turn overlain by thin, purplish laminite which becomes interbedded upwards with fine black argillites and shales. This unit, the Sunday Run Formation, is exposed principally on Finger Hill Island and on the north shore of the mainland portion of the Kaumajet Mountains, and thins toward the south. Two thin basaltic flows are exposed in the sandstones on Finger Hill Island (Smyth, 1976). Although these outcrops were not visited this season, a poorly-exposed, but distinct light grey-weathering cobble conglomerate bed was observed just above the basement unconformity on the southeast side of Mugford Tickle on Cod Island. This unit consists predominantly of well-rounded, grey tonalite cobbles up to 20 cm, with lesser amphibolite clasts set in a matrix of grey, dolomitic(?) sandstone or grit which also carries well-formed pyrite cubes up to 0.5 cm. It is possible, therefore, that the Sunday Run formation either extends farther south than earlier noted, or that local conglomeratic horizons developed in restricted basins during the initial stages of sedimentation.

Smyth and Knight (1978) note the presence of a thin unit of impure dolostone at the top of the Sunday Run formation.

Cod Island formation

The Cod Island formation consists of a roughly 120 m-thick sequence of shale- and chert-dominated siliciclastic sediments. The unit has a basal section of black shales containing thin dolostone beds, decimetre- to metre-scale beds of chert, pyritiferous mudstones and shales, overlain by more black shales and numerous metre-scale interbeds of calcareous argillite and grey and black cherts. On the western side of Cod and Grimmington islands, thin units of the basal shales and dolostones are laterally extensive and the latter offers potential use as a stratigraphic marker within the Cod Island formation. Minor, thin, grey limestone beds occur near the top of the formation at the south end of Cod Island. Many of the thinly-bedded sediments (particularly the argillites) show evidence of crossbedding, soft-sediment deformation and, locally, listric normal faulting at the centimetre scale; the latter features presumably relating to progressive basinal extension, and infilling with overriding lavas of the Calm Cove formation.

A roughly 30 m thick laharic breccia, related to slope instabilities during sedimentation, is best exposed on the northwest side of Mugford Tickle and the west side of Mugford Harbour (Smyth, 1976). Preliminary mapping suggests this may thin to the south and west and grade from an unsorted, coarse mudstone breccia to a more regularly bedded, distal(?) intraformational breccia. The laharic mudflow contains angular to subrounded clasts of chert, shale, volcanic fragments, and fine argillite in a fine grained mudstone matrix. On the south side of Grimmington Island, a mudstone breccia in dark grey shales several meters above the basement unconformity consists of poorly-sorted angular clasts of up to 0.5 m of shale and subordinate quantities of rounded cobbles of granodioritic basement gneiss, and minor subangular basalt fragments in an open, mud-supported framework.

Most of the calcareous argillites present near the top of the Cod Island formation are interpreted as volcanogenic tuffs (see also Smyth, 1976; Smyth and Knight, 1978). These are very fine- to fine-grained, pale-green to green, well-laminated and often crossbedded units on the order of a few metres thick and are commonly interbedded with chert horizons of similar dimension. They are interpreted to be ash-fall pyroclastic rocks deposited in a shallow marine environment.

Numerous buff- to brown-weathering basaltic or gabbroic sills intrude the Cod Island formation (Smyth, 1976). These sills are typically several metres thick, but on the south side of Grimmington Island one sill just above the basement unconformity intrudes a thin (<10 m) section of basal pyritic shale and dolostone and is more than 30 m thick. The sill was thick enough to develop a relatively coarse grained interior, and in at least two locations has pegmatitic gabbroic segregations which were sampled for U-Pb dating (baddeleyite). The sill also shows internal differentiation features related to modal layering of plagioclase and pyroxene (and olivine?). A knobbly, reddish-brown-weathering mafic cumulate layer approximately 2 m thick stands out distinctly from the overlying and underlying gabbro (Fig. 2).

An ultramafic sill, intruded into well-bedded, laminated dark cherts and argillites near the top of the Cod Island formation was first investigated by Barton (1975) who analyzed the unit and determined it to be a basaltic komatiite flow. This was reinterpreted by Smyth (1976) to be a sill, as the upper contact shows evidence of having ingested the overlying cherts and argillites, and the unit shows numerous millimetre- to centimetre-scale internal layering features. This black-weathering, fine- to medium-grained peridotite was also found in a number of other locations on the west side of Grimmington Island, all of which are undoubtedly the same sill traceable over a strike length of at least 5 km. Coarse grained, fresh clinopyroxene was noted in a few localities although much of the unit is heavily serpentinized. An isolated ultramafic sill was also found on the east side of Lost Channel, but its relation to the main sill is uncertain. Although no other ultramafic sills were found on Cod Island, a greater distribution of ultramafic sills in the Mugford Group is now recognized, and provides important information regarding petrologic models and magma chemistry for the group as a whole.

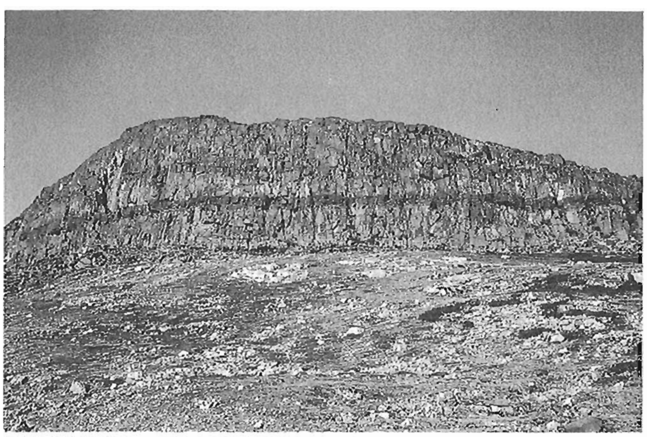

Figure 2. Gabbroic sill forming a hill at the south end of Grimmington Island. Note 2 m thick mafic cumulate portion approximately 14 m up from the base. Base of cliff is thin shale unit directly overlying light-coloured Archean gneisses (foreground). GSC 1992-247H

Most well-exposed sills were comprehensively sampled so that petrographic and chemical relationships within and between individual bodies might be investigated, and that comparisons with the chemistry of overlying volcanics could be made. This will provide a basis for determining Sr, Nd, and Pb isotopic characteristics of the suite of sills, which has the potential for yielding at least Nd isotopic age information provided the internal compositional variations are large enough. Several of the sills will be investigated for U-Pb isotopic age determinations.

Calm Cove formation

The Calm Cove formation consists essentially of basaltic pillowed flows overlain by massive basalts and pillow breccias. Although pillowed lavas dominate the basal part of the formation, each pillowed flow contains a massive upper section with or without an associated pillow breccia zone. This sequence repeats many times in the lower part of the formation, each flow being up to a few tens of metres thick. The relative proportions of pillowed, massive, and brecciated zones within individual flows vary, and are nonsystematic. In contrast to the roughly 100 m of pillowed lavas estimated by Smyth (1976), a considerably greater relative volume of pillowed lavas is estimated in this study. Pillowed flows on Grimmington Island, in fact, persist to the top of the Calm Cove formation.

Thick, continuous, pillowed flows often contain restricted massive sections, as has been noted elsewhere by Baragar (1984). Most of the pillowed basalts have individual pillow diameters on the scale of a metre or less, though isolated flows have pillows up to 2 m in diameter. No relationship between pillow-rim thickness and flow thickness or colour has yet been ascertained. Pillowed flows are buff-weathering, fine grained, grey to light green, and may be amygdaloidal. Amygdales are typically 1-2 mm in diameter and filled with chlorite or calcite or both (with chorite linings). Interpillow voids contain pods of quartz+epidote+calcite, or have finely-laminated calcareous argillite, at times crossbedded between pillows. Several flows near the base of the formation contain pale greyish white to light green plagioclase phenocrysts up to 1.5 cm in size.

Spectacular pillowed basalts exposed in three dimensions are visible near shoreline at two localities – just north of the entrance to the salt-water cove at Anchorstock Bight, and along the south shore of Cod Island just west of Green Cove (Fig. 3). The pillows are elongate, tubular in shape and interconnected and together form a matted surface (see also Baragar, 1984). In several interconnected lava tubes, lava drainage cavities filled with quartz, calcite, and epidote were observed (Fig. 4). Beautifully preserved ropy or pahoehoe features are also visible in isolated localities (Fig. 5).

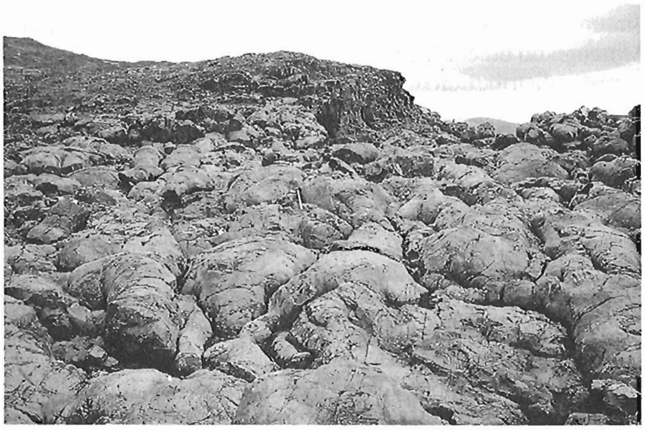

Figure 3. Pillow lava tubes, west of Green Cove, Cod Island. Hammer for scale. GSC 1992-247D

Figure 4. Basaltic pillow with multiple drainage cavities, filled with quartz+calcite+epidote. West of Green Cove, Cod Island. GSC 1992-247B

Fine- to medium-grained, dark- to light-grey to light-green, buff-weathering massive basalts are generally blocky, or less commonly, columnar jointed. These units may be plagioclase-phyric much as their pillowed counterparts in the lower sections of the formation. Many of the massive sections of pillowed flows were sampled for chemical analysis throughout the entire stratigraphy.

Pillow breccias ordinarily mark the upper portions of individual flows, but intraflow breccias are also quite common. These may be composed of angular basalt fragments commonly <4 cm in size, which often weather to a rubbly mass (fragments in positive relief), or are subrounded or elongate rimmed pillow fragments up to a metre in size. Many pillow breccias have characteristics of both (Fig. 6). Occasionally, pillow breccias are interspersed with abundant angular hyaloclastic fragments. Rare localities contained frothy or vesicular basaltic fragments in the pillow breccia matrix (e.g., south side of Mugford Tickle).

On the southwest side of Grimmington Island, at 550 m elevation, a thin (<2 m), grey, fine grained tuff or ash layer drapes over a chaotic zone of pillows and pillow breccia. This apron of pyroclastic material contains dark grey to black angular scoriaceous fragments ranging from 1 cm to 15 cm in diameter overlain and wrapped by ash (Fig. 7). The nature of the highly disordered pillow breccias, locally steep bedding surfaces, and presence of frothy scoria ejecta suggest the local proximity of a vent source.

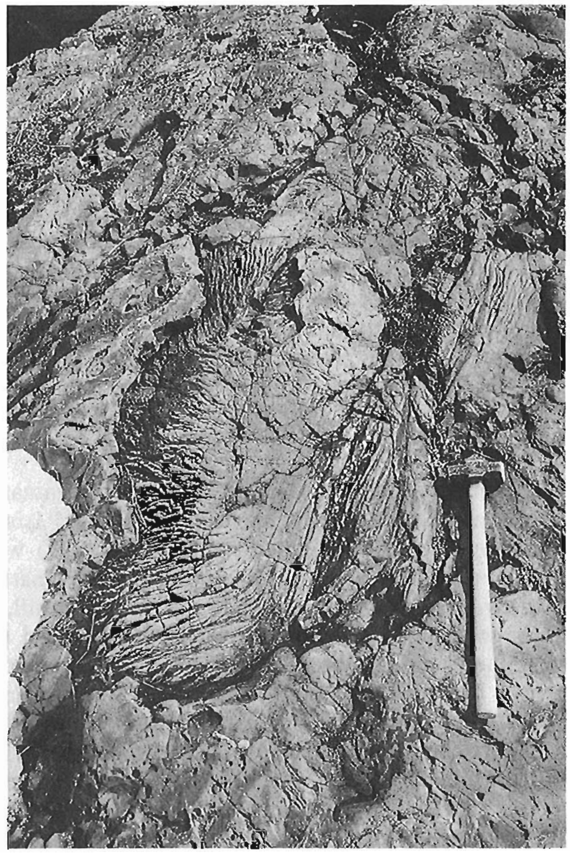

Figure 5. Ropy features among pillowed basalts, typical of pahoehoe flows. Anchorstock Bight area, Cod Island. GSC 1992-247E

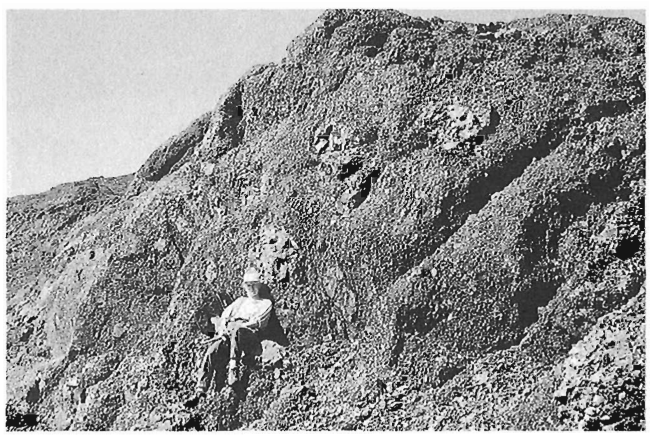

Figure 6. Rubbly pillow breccia near the top of the Calm Cove formation, Grimmington Island. GSC 1992-247C

Figure 7. Angular, black scoriaceous ejecta in fine grained, grey ash layer. Calm Cove formation, Grimmington Island. GSC 1992-247I

Minor, thin (<5 m), white- and brown-weathering, pale and dark green basaltic agglomerates are present throughout the Calm Cove formation. These are composed of largely angular, fragmental basalt clasts in a fine matrix of tuff, similar to agglomerates of the Finger Hill formation. Layered flows are distinctly absent.

Shark Gut formation

Fine grained, generally pale grey-weathering, thinly laminated (millimetre to centimetre scale), normally graded, crossbedded, calcareous tuffs and argillites characterize the Shark Gut Formation (Smyth, 1976; this work). Exposures studied this season were on Cod and Grimmington islands. Dark greenish-brown mafic tuffs and crossbedded crystal tuffs are locally present. Massive, thin (<5 m), columnar jointed basaltic sills were observed within the formation. On the ridge due east of Bishop's Mitre, a metre-thick section of grey lapilli tuff contains abundant subangular to subrounded centimetre-scale fragments of grey-green basalt and grey tuff clasts and millimetre-scale grains and crystals. Elsewhere on Grimmington Island, fine grained, white-weathering Shark Gut basal tuffs are draped over the brecciated top of a pillowed flow at the top of the Calm Cove formation (Fig. 8).

Finger Hill formation

Smyth (1976) estimated that the Finger Hill formation consists of at least 600 m of mafic agglomerates and breccias, minor tuffs, and basaltic sills. This unit was visited during the 1992 field season on the higher peaks of Cod Island, Grimmington Island, and the mainland portion of the Kaumajet Mountains. Despite the heterogeneous nature of the agglomerate unit, it is remarkably uniform from area to area. The breccias are composed of thick individual flow units, are poorly sorted, contain buff-weathering (at times vesicular) angular to rounded basaltic fragments that are less than a centimetre to over 2 m in size and, like many basal layers of pyroclastic flows, locally show reverse grading (Fig. 9). Subordinate clasts and blocks of earlier breccia and agglomerate, white-weathering tuffaceous material and even rare millimetre to centimetre scale quartz pebble conglomerate clasts are found in the laterally extensive Finger Hill breccias. That terrigenous clastics from the basal sedimentary sequence (Sunday Run formation?) occur as ejecta within the breccia flows attests to the ease with which accidental fragments have been brought upwards through the entire lower volcanic sequence, as the large volume of basaltic fragments also suggests. Sampling emphasis was placed on the volumetrically minor, but ubiquitous massive flows.

Middle (?) Proterozoic diabase dykes

The Mugford Group is cut in many places on Cod and Grimmington islands by undeformed, fine- to medium-grained, grey-weathering subvertical diabase dykes which are up to 15 m thick. These dykes are oriented north-south with the exception of at least two dykes which are oriented approximately east-west. Minor greyish to pale white plagioclase phenocrysts up to 1 cm are common in both dyke sets. Epidote mineralization is widespread coating joint surfaces. Several dykes were sampled for paleomagnetic work and

Figure 8. Basal tuffs of the Shark Gut formation (top) draping over a brecciated pillow flow top of the uppermost Calm Cove formation, Grimmington Island. GSC 1992-247A

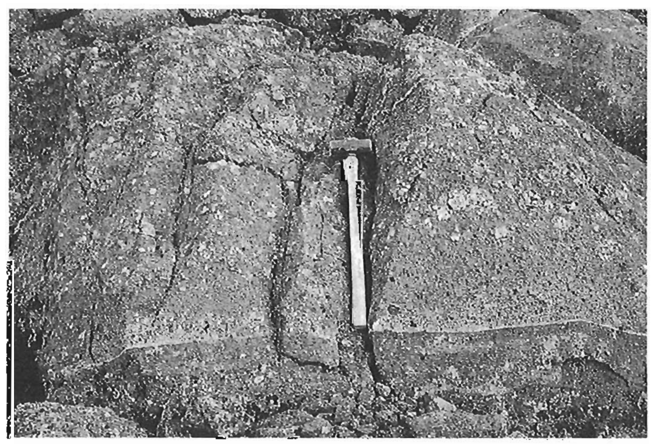

Figure 9. Reverse grading in uppermost of two mafic pyroclastic agglomerate beds, Finger Hill formation, Finger Hill peak, Kaumajet Mountains. GSC 1992-247F

U-Pb geochronology to constrain the timing of this magmatism in relation to Mugford Group volcanism. They are similar in orientation to the Nutak dykes of Ermanovics et al. (1989), who suggests that the latter are Neohelikian in age.

ECONOMIC GEOLOGY

The basal black shales of the Cod Island formation are locally pyritiferous. Pyrite mineralization is also common at the base of gabbroic sills where they intrude these sediments. Chalcopyrite and pyrrhotite are present as disseminations and blebs throughout much of the volcanics, and larger blebs occur in pods up to 1.5 cm and on surfaces in the agglomerates and volcanic breccias of the Finger Hill formation. Seams of fine grained magnetite 1-2 mm thick are present along joint surfaces in some of the ultramafic sill localities in the Cod Island formation. Magnetite and pyrrhotite are found in the late diabase dykes.

DISCUSSION

A significant observation was made on the southeast side of Mugford Tickle in a small outlier of the group which exposes the Archean basement unconformity, regolith, basal conglomerates, dolostone and interbedded cherts, and overlying gabbroic sill. A steep northeast-dipping 5 m diabase dyke trending roughly 320°, cuts the basement and basal sediments (Fig. 10). It could not be seen if the dyke transgressed the gabbroic sill. The general orientation and appearance of this dyke, which contains abundant coarse plagioclase phenocrysts up to approximately 2.5 cm, is similar to the Early Proterozoic Napaktok dykes described above. However, these dykes have generally been regarded as predating deposition of Mugford Group sediments and its attendant volcanism (Ermanovics et al., 1988, 1989). These authors have also noted a subset of breccia dykes which may be associated with, but exploit earlier Napaktok dyke trends just west of the Mugford Group. Ermanovics et al. (1989) and Van Kranendonk (pers. comm., 1992) have postulated on indirect evidence that younger intrusive episodes of these dykes may also transgress the Mugford Group. A possibility exists then that Napaktok dyke intrusion spans an extended length of time, the youngest episode of which may be evidenced by the dyke on Cod Island. Although it could not be seen if the dyke cuts the overlying volcanic series, one may consider that the dyke is a feeder to the Mugford Group volcanics, and that the regional Napaktok dyke swarm may not be just fortuitously adjacent to this early Proterozoic volcanic complex.

Most traverses allowed detailed field investigation and sampling of the lowermost sedimentary units and the basaltic flows of the lower volcanic units (Cod Island and Calm Cove formations, respectively), as well as the gabbroic sills that intrude them. On Cod and Grimmington islands, sampling higher topographic and structural levels is generally hampered by inaccessibility from sea level due to steep cliffs. A complete sampling transect was run from the Archean basement up through the Cod Island, Calm Cove, and Shark Gut formations and a few hundred metres into the overlying Finger Hill formation on a mountain section northeast of Calm Cove. Elsewhere on these islands, sampling of different stratigraphic levels within the Shark Gut and Finger Hill formations was aided by helicopter-assisted traverses.

Figure 10. A 5 m thick Napaktok(?) diabase dyke (right; strike 320°) cutting interbedded dolostone and cherts (left) of the Cod Island formation, southeast side of Mugford Tickle. GSC 1992-247G

An extensive collection of the Mugford Group samples from this field season is being analyzed petrographically and for comprehensive major and trace element geochemistry. Samples of some of the coarser sills and dykes are being processed for U-Pb geochronology, as are some of the volcanic flows for Nd, Sr, and Pb geochronology and tracer geochemistry. One of the aims of this work is that it will provide a basis for interpretation of the relative influences of mantle and crustal processes on the evolution of an Early Proterozoic rift volcanic sequence, and shed light on possible tectonic scenarios regarding Nain-Rae Province geometries leading up to Torngat Orogeny.

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Geological Survey of Canada Project 730044AQ

Interprétation préliminaire des écoulements glaciaires dans la région de la Petite rivière de la Baleine, région subarctique du Québec

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Parent, M. et Paradis, S.J., 1993 : Interprétation préliminaire des écoulements glaciaires dans la région de la Petite rivière de la Baleine, région subarctique du Québec; dans Recherches en cours, Partie C; Commission géologique du Canada, Étude 93-1C, p. 359-365.

Résumé : L'examen des surfaces rocheuses striées à quelque 145 sites répartis sur environ 7 400 km^2 dans la région de la Petite rivière de la Baleine indique que l'écoulement glaciaire dominant était dirigé vers l'ouest-nord-ouest (285°). Toutefois, à quelques endroits le long du littoral de la baie d'Hudson, en position abritée par rapport à ce mouvement, il subsiste des surfaces striées plus anciennes formées par un mouvement glaciaire dirigé vers le nord-nord-ouest. Ce mouvement ancien pourrait hypothétiquement être relié à un courant glaciaire formé dans le nord de la baie d'Hudson. Le long de la côte, entre Kuujjuarapik et le lac Guillaume-Delisle, la plupart des surfaces striées montrent que le mouvement dominant vers l'ouest-nord-ouest a subi une déflexion vers le sud-ouest, présumément en réponse à la formation d'un courant glaciaire dans la partie sud de la baie. L'existence même de cette déflexion tardiglaciaire démontre que le mouvement régional dominant vers l'ouest-nord-ouest s'était amorcé bien avant l'invasion tyrrellienne dans le bassin de la baie d'Hudson.

Abstract: Striated bedrock surfaces were investigated at about 145 sites over an area of approximately 7400 km^2 in the Petite rivière de la Baleine region. The dominant regional ice-flow event was towards west-northwest (285°). However, at a few sites along the shore of Hudson Bay, small striated surfaces lying in sheltered positions relative to the main glacial movement provide evidence for an older ice-flow event directed towards north-northwest. This older glacial movement may tentatively be related to glacial streaming in northern Hudson Bay. Along the shore, between Kuujjuarapik and Lake Guillaume-Delisle, most striated outcrops show that the dominant westward movement was deflected towards the southwest, presumably in response to the formation of an ice stream in southern Hudson Bay. The very existence of this late-glacial re-orientation of ice-flow demonstrates that the dominant regional movement towards the west-northwest had formed well before the incursion of the Tyrrell Sea in Hudson Bay basin.

INTRODUCTION

Puisque la région de la Petite rivière de la Baleine est située à proximité du centre géographique de l'Inlandsis laurentidien, les levés de géologie du Quaternaire dans cette région peuvent contribuer à une meilleure compréhension de la dynamique globale de l'inlandsis ou, tout au moins, apporter certaines contraintes aux reconstitutions paléogéographiques (Shilts, 1980; Prest, 1984; Dyke et Prest, 1987) et aux modélisations glaciologiques (Fisher et al., 1985; Boulton et al., 1985; Hughes, 1987) de la dernière glaciation et déglaciation. La dynamique des écoulements glaciaires régionaux est certes l'un des aspects clé sur lesquels ces reconstitutions et modèles doivent s'appuyer et à l'aide desquels ils peuvent être validés ou testés.

Les objectifs de cet article sont 1) de présenter sommairement les séquences de mouvements glaciaires observés lors d'une première saison de terrain dans la région de la Petite rivière de la Baleine et 2) d'en discuter les implications relativement à la dynamique de l'Inlandsis laurentidien dans la partie orientale de la baie d'Hudson.

RÉGION ÉTUDIÉE

La région étudiée (fig. 1) est située à l'est de la baie d'Hudson et couvre un territoire d'environ 7 400 km^2 s'étendant de 55°30′ à 56°00′ de latitude nord et de 75°00′ à 77°30′ de longitude ouest. Cette région fait l'objet d'une étude pilote multidisciplinaire visant à caractériser les environnements et processus géomorphologiques et géochimiques dans une région où d'importantes infrastructures de développement hydro-électrique sont censées être aménagées (complexe Grande-Baleine). La carte de localisation montre aussi les limites des bassins hydrographiques de la Grande rivière de la Baleine, de la Petite rivière de la Baleine et de la rivière Nastapoka. Les travaux de terrain effectués à l'été 1992 visaient principalement à réaliser des levés de géologie du Quaternaire dans la région pilote ainsi qu'un levé géochimique de reconnaissance des formations superficielles.

Le substratum rocheux de la région de la Petite rivière de la Baleine est constitué de deux domaines géologiques distincts (Eade, 1966; Ciesielski, 1991; Chandler et Schwarz, 1980) : 1) celui de la sous-province archéenne de Bienville, composé surtout d'orthogneiss et sous-jacent à la très grande majorité du territoire étudié, et 2) celui de la couverture sédimentaire protérozoïque, composé de roches sédimentaires arénacées et carbonatées coiffées de roches basaltiques et formant une étroite bande côtière marquée par des cuestas dont le front est tourné vers le continent. La région n'a toutefois pas fait l'objet de levés détaillés de sorte que les seules cartes géologiques disponibles sont des cartes de reconnaissance (Eade, 1966; Ciesielski, 1991).

Si l'on fait exception de l'étroite zone des cuestas côtières, le modelé de la région étudiée est celui d'une pénéplaine ondulée dont les replats rocheux sommitaux oscillent autour

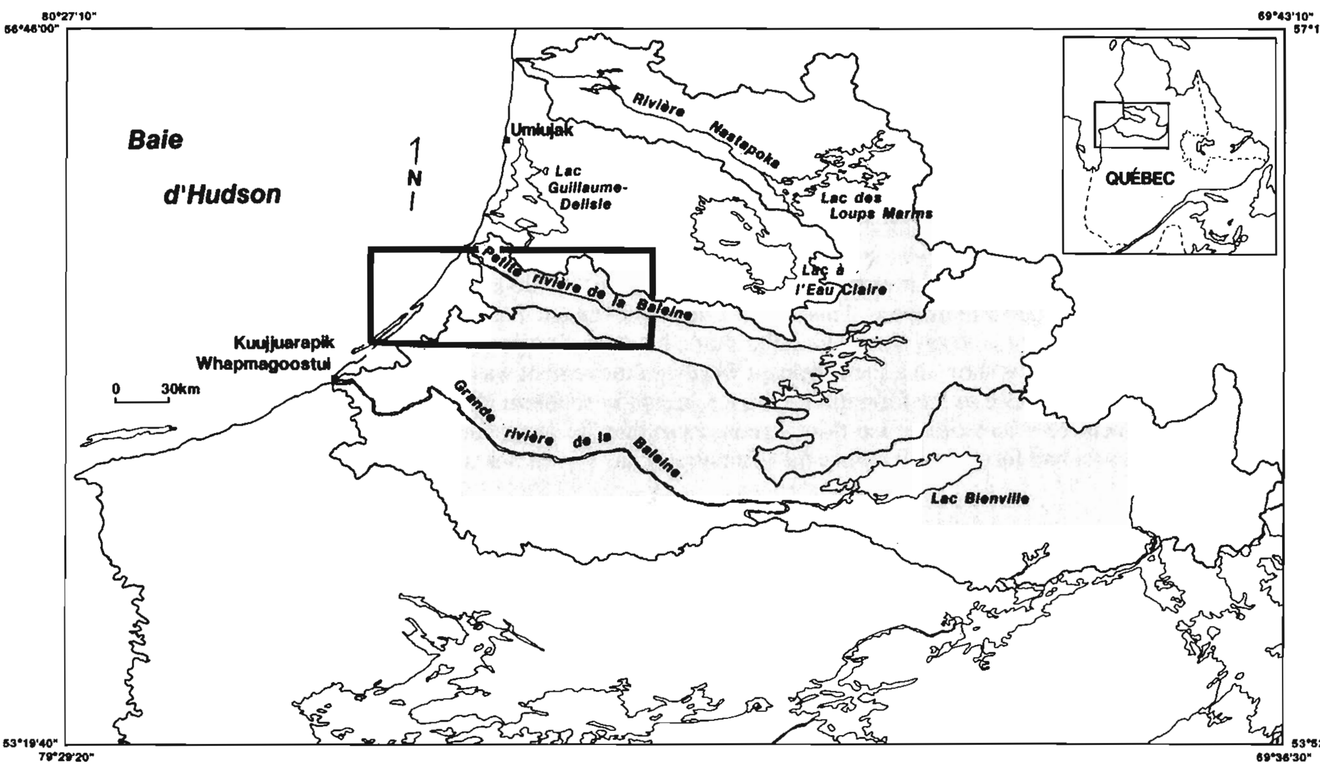

Figure 1. Localisation de la région de la Petite rivière de la Baleine. La carte montre aussi les limites des bassins hydrographiques de la Grande rivière de la Baleine, de la Petite rivière de la Baleine et de la rivière Nastapoka.

de 300 m d'altitude et dont le grain topographique est orienté est-sud-est-ouest-nord-ouest, parallèlement à la direction structurale de la sous-province de Bienville (Ciesielski, 1991). Quoique les dénivellations ne dépassent que rarement 100 m, la densité de la dissection est élevée, en particulier sur les premiers 80 km vers l'intérieur des terres, de sorte que le relief y est essentiellement composé de replats hectométriques à kilométriques bordés par des versants raides, et même subverticaux dans les secteurs de forte érosion glaciaire. Plus à l'intérieur des terres, le relief s'adoucit et le socle rocheux y est masqué sur de grandes superficies par une couverture de till généralement drumlinisé ou fuselé. Sous la limite maximale de la Mer de Tyrrell, dont l'altitude varie régionalement d'environ 275 m (Archer, 1968) à 315 m (Hillaire-Marcel, 1976), les dépressions sont partiellement comblées par les sédiments marins alors que les replats sommitaux et les versants sont constitués de surfaces rocheuses délavées par l'action des vagues.

TRAVAUX ANTÉRIEURS

Bien que la région de la Petite rivière de la Baleine n'ait pas fait l'objet d'un levé systématique des formations superficielles, certains aspects de la géologie du Quaternaire ont été couverts dans le cadre de quelques études thématiques. À part les stries compilées sur la Carte glaciaire du Canada (Prest et al., 1968) et celles observées par Hillaire-Marcel (1976) montrant un écoulement glaciaire orienté en moyenne vers l'ouest, les directions régionales d'écoulement glaciaire sont très méconnues. Par contre, l'histoire de l'invasion marine et celle de l'émersion subséquente ont fait l'objet de plusieurs publications (Archer, 1968; Hillaire-Marcel, 1976, 1980; Allard et Tremblay, 1983; Vincent et al., 1987).

MÉTHODOLOGIE

Les surfaces rocheuses striées ont été soigneusement observées à quelque 145 localités réparties dans toute la région (fig. 2). La grande majorité de ces localités sont situées à l'intérieur du périmètre de la région pilote identifiée à la figure 1. Étant donné les meilleures conditions d'atterrissage sur les replats sommitaux où les arbres sont plus épars ou même absents, il est probable que les localités de ce type sont un peu surreprésentées dans cet échan- tillonnage. Toutefois, un bon nombre de surfaces striées ont été observées sur les berges de lacs et de rivières et les directions d'écoulement glaciaire mesurées y étaient les mêmes que celles mesurées sur les replats avoisinants. Comme la plupart des points d'atterrissage devaient aussi servir à l'échantillonnage du till, les surfaces striées observées étaient souvent de dimension restreinte et devaient être dégagées sous une mince couche de till. De toute façon, l'altération des surfaces rocheuses dénudées est telle dans la région que les surfaces striées y sont pour la plupart trop dégradées pour faire l'objet de mesures d'orientation.

Sauf dans le cas de quelques localités à l'intérieur des terres, presque toutes situées dans le secteur du lac à l'Eau Claire, les endroits où plusieurs générations de stries ont été observées sont localisés sur les affleurements de roches protérozoïques le long de la côte de la baie d'Hudson (fig. 2). Il faut aussi signaler que des affleurements rocheux de même nature lithologique et de même taille ont été observés attentivement à plusieurs autres endroits sous la limite marine et que ces derniers ne comportaient qu'un groupe de stries. Selon toute évidence, la présence de plusieurs générations de stries glaciaires sur les affleurements côtiers ne résulte pas d'un biais méthodologique mais réflète bel et bien l'histoire glaciaire régionale.

Afin d'établir la polarité des stries glaciaires observées, deux grands types de critères ont été utilisés : 1) ceux fondés sur les micro-formes d'érosion glaciaire, telles les queues-de-rat (fig. 3), les stries en tête de clou, les broutures et les fractures de broutage; et 2) ceux fondés sur le profilage et le façonnement général des affleurements, l'exemple typique étant celui des roches moutonnées (fig. 4). Comme le montre la figure 2, on a pu déterminer la polarité de toutes les surfaces striées observées. À l'intérieur des terres, les critères les plus communément utilisés étaient évidemment le profilage des affleurements : en effet, l'intensité de l'érosion glaciaire a été telle dans la région que presque tous les affleurements rocheux montrent soit de belles surfaces d'abrasion, bien polies et bien striées, faisant face à l'amont glaciaire, soit des surfaces de débitage tournées vers l'aval glaciaire, ou soit les deux. Dans toutes les localités de la zone des cuestas côtières ainsi qu'à plusieurs localités intérieures, diverses micro-formes d'érosion glaciaire viennent s'ajouter aux critères de profilage général des affleurements.

La chronologie relative des écoulements glaciaires s'établit à partir des recoupements et surimpositions observés sur les surfaces striées. Dans le cas d'un recoupement, une surface striée antérieure est préservée, en position relativement abritée, de l'érosion causée par un mouvement glaciaire subséquent. Typiquement, le recoupement est marqué par une troncature séparant deux surfaces striées selon des directions distinctes (Veillette, 1983), la plus ancienne étant celle qui est située en contrebas de la troncature (fig. 3). Dans le cas d'une surimposition, les stries plus récentes sont formées à même une surface striée antérieure, mais sans qu'il y ait oblitération complète de cette dernière. Dans ce cas, ce sont les stries de la première génération qui occupent l'essentiel de la surface de l'affleurement rocheux alors que dans le cas d'un recoupement, ce sont les stries de la deuxième génération qui occupent presque toute la surface de l'affleurement.

MOUVEMENTS GLACIAIRES RÉGIONAUX

Mouvement glaciaire dominant

La cartographie des directions généralisées d'écoulement glaciaire permet de mieux dégager les patrons régionaux (fig. 5). Le mouvement glaciaire dominant qui est orienté à environ 280° à l'intérieur des terres s'infléchit progressivement vers 300° en approchant de la côte de la baie d'Hudson. Cette tendance se maintient d'ailleurs depuis les environs de Kuujjuarapik jusqu'à la hauteur du lac Guillaume-Delisle. Compte tenu du relief de la région, la

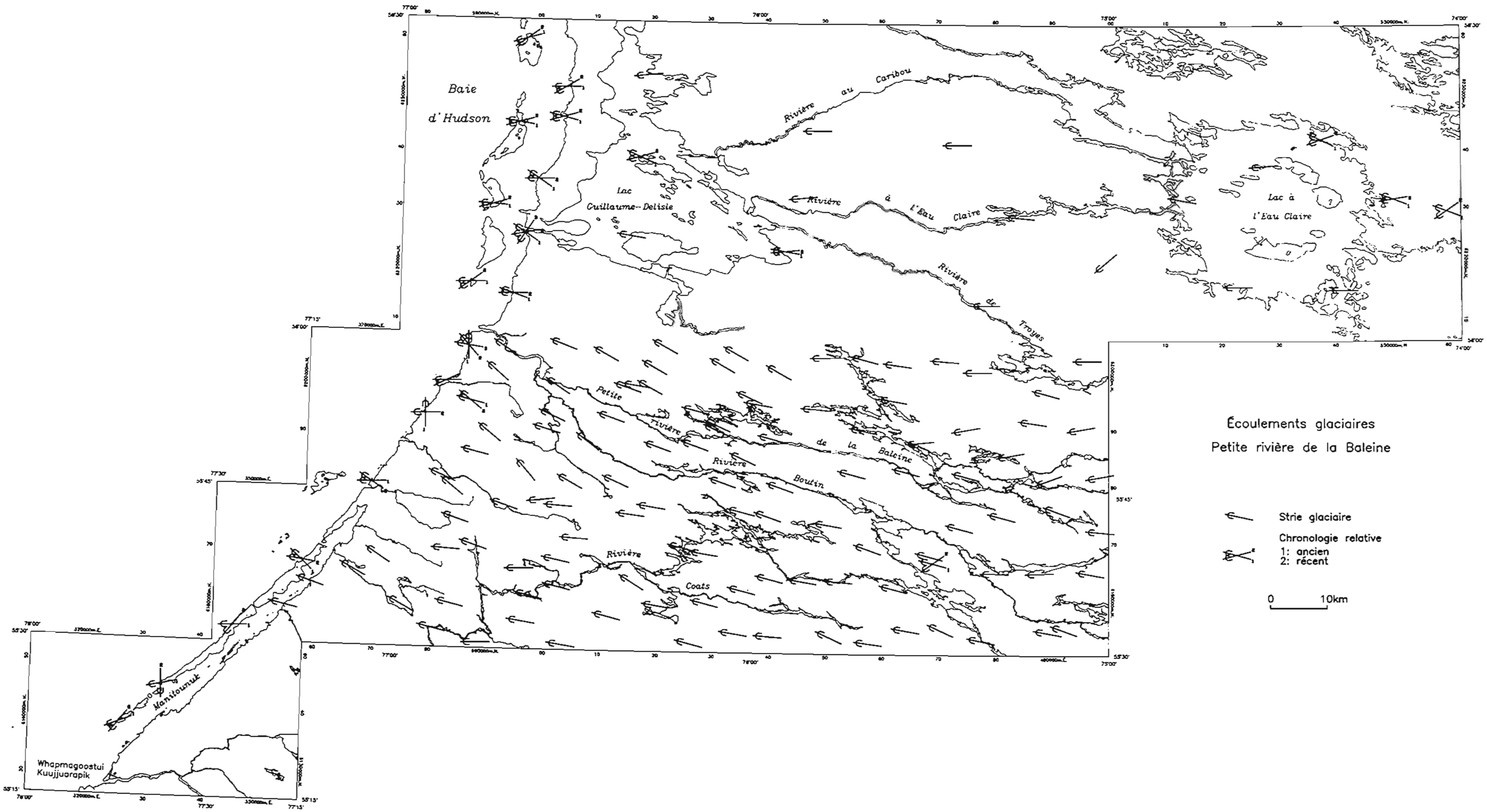

Figure 2. Carte des stries glaciaires levées dans la région de la Petite rivière de la Baleine et du lac Guillaume-Delisle.

dispersion des orientations par rapport à la tendance générale de ce mouvement semble très faible. C'est à ce mouvement dominant que sont associées les grandes formes glaciaires de la région, comme les rochers profilés, les rochers dissymétriques, les roches moutonnées, les drumlins et drumlinoïdes et les moraines de fond fuselées.

Partout le long de la côte et à basse altitude, ce mouvement glaciaire dominant est suivi d'une déflexion vers le sud-ouest (fig. 2, 5). Cette déflexion de l'écoulement tardiglaciaire est mise en évidence pour la première fois dans la région. Dans la plupart des localités observées, la réorientation de l'écoulement est de l'ordre de 30° à 40°.

Mouvement glaciaire ancien

À deux localités côtières comprises entre la tête du détroit de Manitounuk et l'embouchure de la Petite rivière de la Baleine (fig. 2, 3), un mouvement glaciaire ancien dirigé vers le N est enregistré sur des surfaces striées sises en position abritée par rapport au mouvement régional dominant. Malgré d'attentives recherches dans le but de trouver d'autres indices de ce mouvement glaciaire ancien sur les autres affleurements du littoral, ces deux localités demeurent les seules où ce mouvement a pu être retrouvé.

DISCUSSION

Une question importante que l'on doit se poser concerne la signification du mouvement régional dominant par rapport à la dynamique d'ensemble de l'Inlandsis laurentidien. Dans un premier temps, faut-il signaler que les données recueillies dans la région étudiée ne supportent aucunement les modèles d'inlandsis présentant un dôme central au-dessus de la baie d'Hudson lors du dernier maximum glaciaire (Hughes, 1987). En effet, non seulement n'existe-t-il aucun indice d'écoulement ou de transport glaciaire vers l'est mais toutes les données concordent pour indiquer que le mouvement glaciaire vers l'ouest a été de longue durée. À ce sujet, deux observations s'imposent : 1) tout le modelé du socle rocheux est relié à ce mouvement glaciaire vers l'ouest; 2) un bloc de formation ferrifère provenant de la fosse du Labrador, et donc transporté vers l'ouest sur une distance minimale de 400 km, a été trouvé à la surface du till dans la partie est de la région. En admettant une vitesse moyenne de transport de 10 m/an, déjà excessive, ce bloc aurait mis au moins 4 000 ans à atteindre la région. Or, d'après les modèles préconisés par Hughes (1987), la période de temps disponible ne dépasse pas 2 000 ans, tout au plus. Les observations supportent plutôt la présence d'une ligne de partage située loin à l'est de la région durant presque tout le dernier hémicycle glaciaire (Prest, 1984; Dyke et Prest, 1987).

D'après les reconstitutions de Dyke et Prest (1987), l'écoulement vers l'ouest dans la partie orientale de la baie d'Hudson n'aurait vraiment débuté qu'après 9 000 ans, et son existence serait partiellement reliée à la formation de courants glaciaires dans le sud de la baie d'Hudson. Or, la déflexion vers le sud-oeust de l'écoulement glaciaire dans la région de la Petite rivière de la Baleine est fort probablement reliée à la formation de ces courants glaciaires, d'autant plus que le fait que l'on ne retrouve les traces de cette déflexion qu'à très basse altitude semble indiquer que c'est surtout l'écoulement de la partie basale du glacier qui était réorienté. En effet, ce type de réorientation semble correspondre davantage à ce qui pourrait se réaliser en bordure d'un courant glaciaire qu'au voisinage d'une éventuelle baie de vêlage où l'écoulement glaciaire serait progressivement réorienté. Que l'on retienne l'hypothèse du courant glaciaire ou de la baie de vêlage pour expliquer la déflexion tardiglaciaire vers le sud-ouest, il n'en demeure pas moins que la chronologie relative des mouvements glaciaires indique clairement que l'écoulement régional vers l'ouest s'était amorcé longtemps avant que n'aient lieu les réorientations tardiglaciaires. Selon toute évidence, ce mouvement régional vers l'ouest semble bien davantage correspondre à une caractéristique intrinsèque de l'Inlandsis laurentidien qu'à un phénomène de réorientation de l'écoulement glaciaire en réponse à l'invasion de la Mer

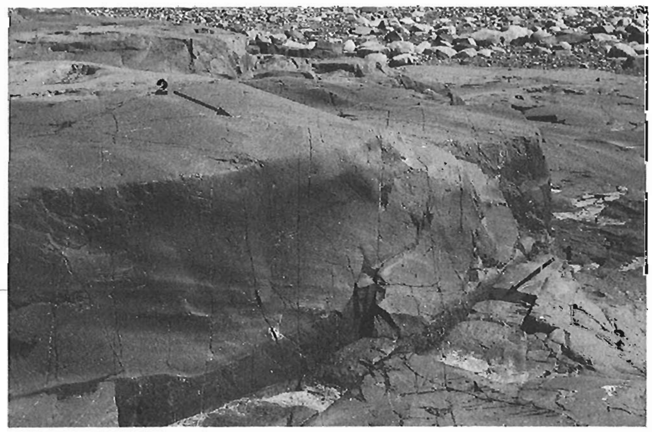

Figure 3. Bel exemple de surface striée (1) préservée en position abritée par rapport au dernier mouvement glaciaire régional (2). Le mouvement glaciaire ancien (1) est orienté vers 360° alors que le mouvement subséquent est orienté vers 270°. Littoral de la baie d'Hudson, environ 15 km au sud de la Petite rivière de la Baleine.

Figure 4. Roches moutonnées formées dans des basaltes protérozoïques sur la plus haute colline au sud-est du lac Guillaume-Delisle. Écoulement glaciaire vers la droite.

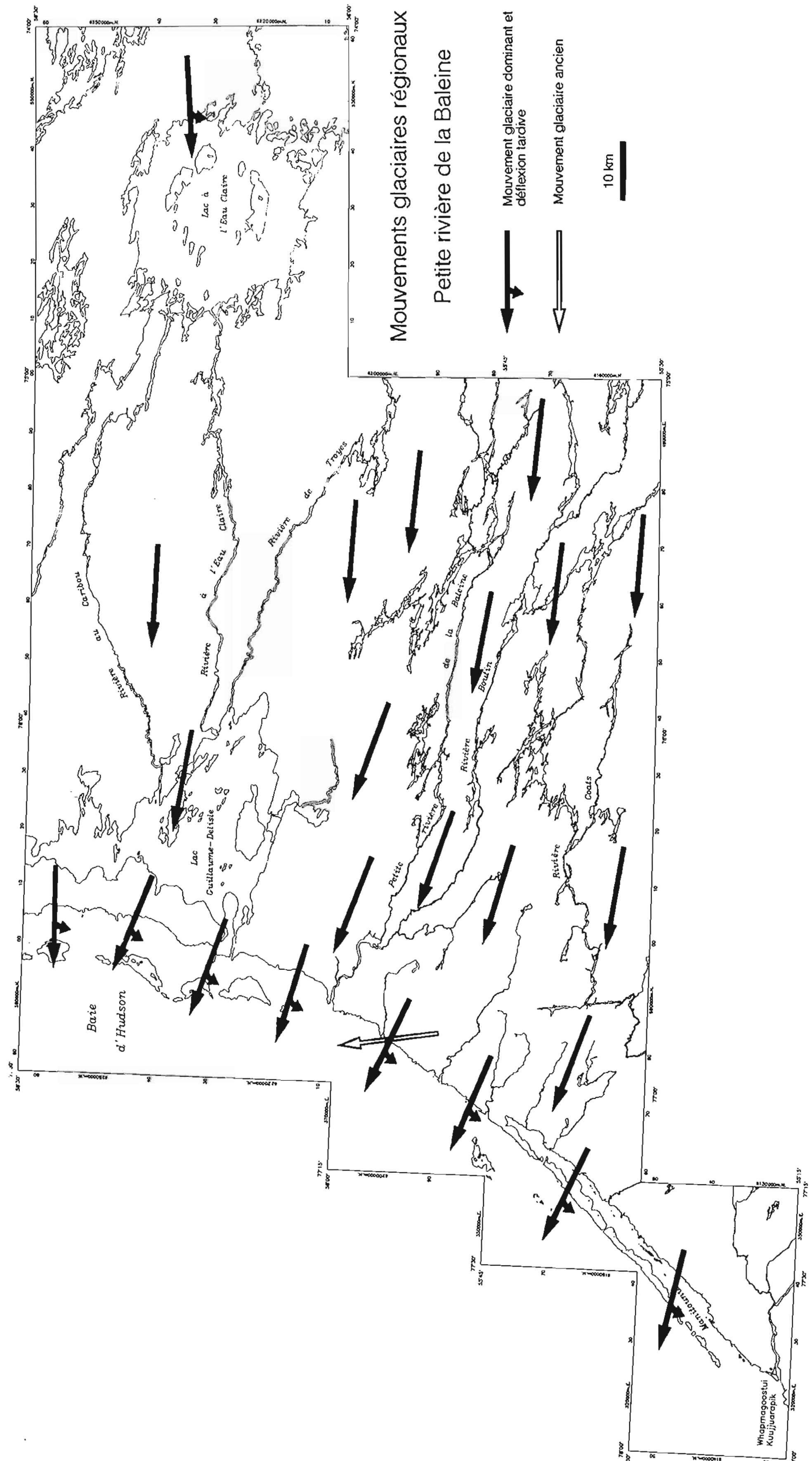

Figure 5. Mouvements glaciaires régionaux dans la région de la Petite rivière de la Baleine.

de Tyrrell dans la baie d'Hudson, ainsi que l'ont interprété nombre d'auteurs (Andrews et Falconer, 1969; Hillaire-Marcel, 1976; Hughes, 1987).

Il existe aussi une déflexion vers le sud-ouest de l'écoulement tardiglaciaire dans la région du lac à l'Eau Claire. En raison de la faible densité des observations dans ce secteur, il semble préférable de n'avancer qu'une interprétation des plus hypothétiques : cette réorientation pourrait être associée à la rejuvénation d'un centre de dispersion tardiglaciaire centré sur l'Ungava. Il faut toutefois aussi reconnaître que la réorientation pourrait être associée à l'épisode glaciolacustre postulé par Allard et Séguin (1985) dans le bassin du lac à l'Eau Claire.

Étant donné le peu d'observations relatives au mouvement ancien vers le nord-nord-ouest, il est difficile d'en saisir toute la signification : il pourrait en effet être relié à la formation de courants glaciaires dans le nord de la baie d'Hudson aussi bien qu'à celle d'une ligne de partage glaciaire au sud de la région de la Petite rivière de la Baleine. Ce mouvement pourrait bien être corrélatif d'un mouvement dirigé vers le nord-nord-ouest dans les îles Ottawa (Andrews et Falconer, 1969), quelque 400 km au nord de la région étudiée.

Il se pourrait aussi que ce mouvement soit en partie relié au mouvement ancien vers le nord-ouest identifié par Veillette et Pomarès (1991) dans la région de Matagami-Chapais, quelque 500 km au sud de la région de la Petite rivière de la Baleine.

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