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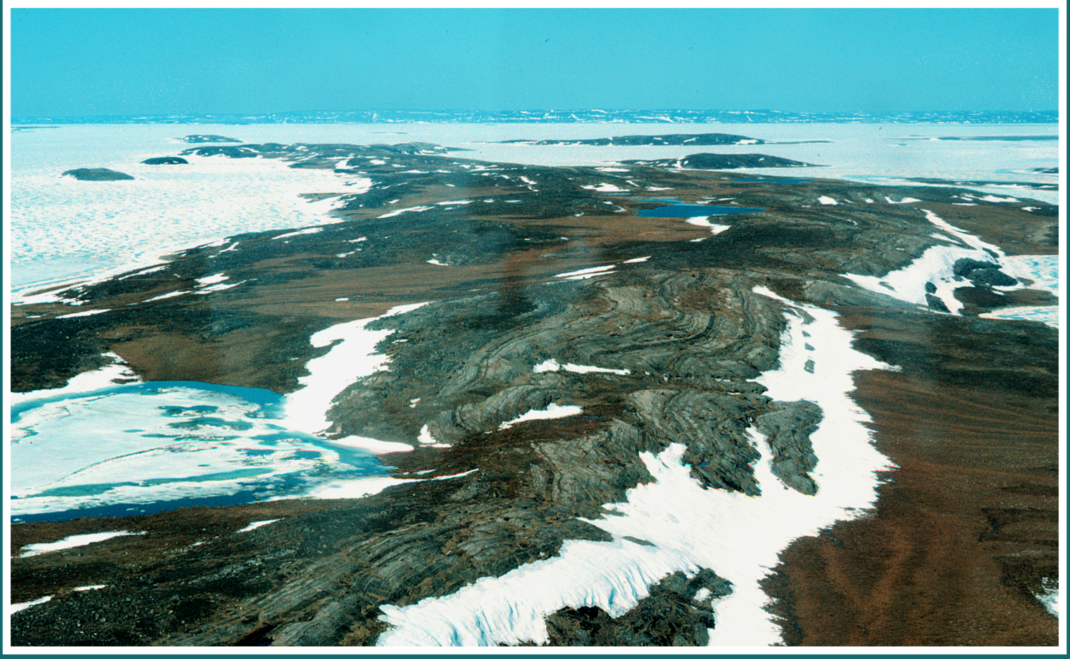
T. Frisch



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Cover illustration

Steeply dipping marble and pelitic schist of the early Proterozoic Chantrey Group at the head of Chantrey Inlet. The islands in the middle distance are underlain by Archean granitic rock. The view is northeastward, toward Victoria Headland on the eastern shore of Chantrey Inlet. Photograph by T. Frisch. GSC 2000-007

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PRECAMBRIAN GEOLOGY OF IAN CALDER LAKE, CAPE BARCLAY, AND PART OF DARBY LAKE MAP AREAS, SOUTH-CENTRAL NUNAVUT

Abstract

The map areas are located southwest and east of Chantrey Inlet in the Churchill Province of the Canadian Shield and take in a late Archean granitoid complex and several amphibolite-grade supracrustal belts, all with northeasterly structural trends. The granitoid complex consists largely of foliated granodioritic to tonalitic plutonic rocks, of which a 2.6 Ga, microcline augen-rich variety is particularly abundant. Low- to medium-pressure, granulite-facies gneiss and K-feldsparphyric orthopyroxene granite (2.6 Ga) exposed in northern Ian Calder Lake and southern Cape Barclay map areas apparently belong to a single, (?)fault-bounded belt 10–15 km wide. Straight gneiss that may belong to the Queen Maud block is bordered by the Chantrey fault zone, which runs through northwestern Ian Calder Lake map area. The Barclay supracrustal belt, which lies east of Chantrey Inlet, has attributes of a greenstone belt and is interpreted as being intruded by 2.6 Ga granodiorite. Outcropping in northeastern Ian Calder Lake map area, the highly deformed Franklin belt comprises metasedimentary rock, schist and amphibolite and is correlated with the Barclay belt. Northeast of, and on strike with, the Franklin belt, lies the Chantrey belt, a doubly plunging, locally thrust-faulted synclinorium 175 km long and 3–7 km wide. It consists entirely of sedimentary rock (carbonate, pelite, psammite, and conglomerate) metamorphosed at 0.3 GPa or less and 470°–570°C (generally slightly lower than the temperatures estimated for (?)Aphebian metamorphism of the Barclay belt). The Montresor belt, in southern Ian Calder Lake map area, is an open synclinal succession, about 1800 m thick, of mainly meta-arenaceous rock. The Chantrey and Montresor belts are interpreted to represent intracontinental, shallow-water deposits. As both belts are intruded by pegmatite dated in the Chantrey belt at 1.7 Ga (K-Ar muscovite), they are regarded as early Proterozoic.

Résumé

La région couverte par ces cartes s'étend au sud-ouest et à l'ouest de la baie Chantrey dans la Province de Churchill du Bouclier canadien et comprend un complexe de granitoïdes de l'Archéen tardif ainsi que plusieurs ceintures de roches supracrustales du faciès des amphibolites, qui affichent tous une direction structurale nord-est. Le complexe de granitoïdes se compose en grande partie de roches plutoniques granodioritiques à tonalitiques à structure foliée, dont une variété oeillée à microcline remontant à 2,6 Ga est particulièrement abondante. Des gneiss du faciès des granulites formés dans des conditions de pression faible à moyenne et un granite à orthopyroxène à phénocristaux de feldspath potassique (2,6 Ga) qui affleurent dans la partie nord de la région cartographique de Ian Calder Lake et dans la partie sud de celle de Cape Barclay appartiennent vraisemblablement à une même ceinture large de 10 à 15 km, limitée par des failles(?). Des gneiss droits, qui pourraient appartenir au bloc de Queen Maud, sont limités par la zone de failles de Chantrey qui traverse la partie nord-ouest de la région cartographique de Ian Calder Lake. La ceinture de roches supracrustales de Barclay, à l'est de la baie Chantrey, présente les attributs d'une ceinture de roches vertes et il semble qu'elle soit recoupée par des granodiorites remontant à 2,6 Ga. La ceinture de Franklin, qui consiste en roches métasédimentaires, schistes et amphibolites très déformés affleurant dans la partie nord-est de la région cartographique de Ian Calder Lake, est mise en corrélation avec la ceinture de Barclay. La ceinture de Chantrey, qui s'étend au nord-est de la ceinture de Franklin et dans le prolongement de celle-ci, est déformée en un synclinorium à double plongée long de 175 km et large de 3 à 7 km qui présente par endroits des failles de chevauchement. Elle consiste exclusivement en roches sédimentaires (roches carbonatées, pélites, psammites et conglomérats) métamorphisées dans des conditions de pression de 0,3 GPa ou moins et de température de 470 à 570 °C (températures légèrement inférieures à celles estimées dans le cas du métamorphisme aphébien(?) dans la ceinture de Barclay). Dans la partie sud de la région cartographique de Ian Calder Lake, la ceinture de Montresor est formée d'une séquence épaisse d'environ 1 800 m de roches métamorphiques principalement dérivées de faciès arénacés qui est déformée en un synclinal ouvert. Les ceintures de Chantrey et de Montresor seraient constituées de dépôts d'eau peu profonde de milieu intracontinental. Puisque les deux ceintures sont recoupées par de la pegmatite qui a été datée à 1,7 Ga (K-Ar sur muscovite) dans la ceinture de Chantrey, elles sont rapportées au Protérozoïque précoce.

SUMMARY

The study area comprises two 1:250 000 scale map areas and part of a third in south-central Nunavut: Ian Calder Lake, southwest of Chantrey Inlet, and Cape Barclay and northern Darby Lake, east of Chantrey Inlet. These areas encompass several northeast-trending Archean and Aphebian supracrustal belts set in an Archean granitoid terrane of the Churchill structural province.

Most of the study area is underlain by amphibolite-grade granodioritic and tonalitic gneiss and massive to gneissic granitoid rocks with northeasterly structural trends. Particularly widespread is a relatively mafic hornblende-biotite granodiorite gneiss with K-feldspar augen, which locally grades into more or less massive porphyritic rock. It underlies most of the central part of Ian Calder Lake map area and extensive regions elsewhere. Augen gneiss from southern Chantrey Inlet has a U-Pb zircon age of 2596 +13/-10 Ma and a Sm-Nd T_{DM} model age of 2.68 Ga. Mylonite is common over much of the study area but especially so in the vicinity of the Chantrey fault zone, which trends northeast in northwestern Ian Calder Lake map area. The fault zone separates grey granodiorite gneiss with amphibolite sheets ((?)metamorphosed dykes) in the east from lithologically similar but more deformed straight gneiss (Nd model age 2.79 Ga) in the west. Although the latter appears to be an equivalent of the grey gneiss, it may belong to the easternmost Queen Maud block, an upfaulted, older, deep-level terrane. Partly retrograded, low- to medium-pressure, granulite-facies rocks, including orthopyroxene-bearing orthogneiss, K-feldspar-phyric orthopyroxene granite, and garnet-sillimanite-cordierite paragneiss, outcrop in easternmost Ian Calder Lake and southern Cape Barclay map areas. They apparently belong to a single, easterly to northeasterly trending belt, 10–15 km wide and presumably fault bounded. Zircon from the orthopyroxene granite gives a U-Pb age of 2587 +9/-7 Ma, which is interpreted to be the time of intrusion under granulite-facies conditions.

Lower amphibolite-grade supracrustal rocks known or inferred to be Archean occur in three areas. In northern Cape Barclay map area, a supracrustal assemblage of greenstone-belt affinity comprising schist with andalusite, cordierite, sillimanite, staurolite, and/or garnet and subordinate metavolcanic rocks, marble, calc-silicate rocks, amphibolite, and ironstone constitutes the Barclay belt. The belt is 2.5–4 km wide, steeply dipping, mylonitized at contacts, and disrupted by granitoid rock. Weakly gneissic granodiorite interpreted as intrusive into schist has a U-Pb zircon age of ca. 2.59 Ga. The Barclay belt is also cut by pegmatite of

SOMMAIRE

La région d’étude, située dans le centre sud du Nunavut, couvre deux cartes à l’échelle de 1/250 000 et une partie d’une troisième. La carte de Ian Calder Lake couvre la région qui s’étend au sud-ouest de la baie Chantrey, alors que la carte de Cape Barclay et la partie nord de la carte de Darby Lake, celle qui est située à l’est de cette baie. Ces régions renferment plusieurs ceintures de roches supracrustales de l’Archéen et de l’Aphébien qui s’étirent dans une direction nord-est au sein de terrains granitoïdes de l’Archéen de la Province de Churchill.

La plus grande partie de la région d’étude se compose de gneiss granodioritiques et tonalitiques du faciès des amphibolites et de roches granitoïdes à structure massive à gneissique affichant des directions structurales nord-est. L’une des lithologies particulièrement abondantes se compose de gneiss granodioritiques à hornblende-biotite relativement mafiques à amas oeillés de feldspaths potassiques qui passent par endroits à des roches porphyriques plus ou moins massives. Cette lithologie couvre la plus grande partie du secteur central de la région cartographique de Ian Calder Lake ainsi que des régions étendues ailleurs. Les gneiss oeillés au sud de la baie Chantrey ont été datés par la méthode U-Pb sur zircon à 2 596 +13/-10 Ma et ont livré un âge modèle de manteau appauvri (T_{DM}) de 2,68 Ga par la méthode Sm-Nd. Des mylonites sont couramment présentes dans une grande partie de la région d’étude, mais surtout abondantes dans les environs de la zone de failles de Chantrey qui recoupe dans une direction nord-est la partie nord-ouest de la région cartographique de Ian Calder Lake. La zone de failles sépare des gneiss granodioritiques gris avec feuillets d’amphibolite (dykes métamorphisés?), à l’est, de gneiss droits semblables quant à la lithologie, mais davantage déformés (âge modèle Nd de 2,79 Ga), à l’ouest. Bien que ces derniers semblent équivalents aux gneiss gris, ils pourraient appartenir au bloc de Queen Maud qui occupe la position la plus à l’est et constitue un terrain plus ancien et de plus grande profondeur, qui aurait été amené à la surface par le jeu de failles. Des roches du faciès des granulites composées entre autres d’orthogneiss à orthopyroxène, de granite à orthopyroxène et à phénocristaux de feldspath potassique et de paragneiss à grenat-sillimanite-cordiérite, qui ont été métamorphisées dans des conditions de pression faible à moyenne et qui ont subi en partie les effets d’un métamorphisme rétrograde, affleurent à l’extrémité est de la région cartographique de Ian Calder Lake et dans la partie sud de la région cartographique de Cape Barclay. Ces lithologies semblent appartenir à une même bande de direction est à nord-est et large de 10 à 15 km qui serait limitée par des failles. Une analyse U-Pb sur zircon du granite à orthopyroxène a livré un âge de 2 587 +9/-7 Ma, lequel correspondrait au moment de la mise en place de cette intrusion dans des conditions du faciès des granulites.

Des roches supracrustales du faciès des amphibolites inférieur dont on sait, ou dont on déduit, qu’elles datent de l’Archéen ont été observées dans trois régions. Dans le nord de la région cartographique de Cape Barclay, la ceinture de Barclay consiste en un assemblage de roches supracrustales rappelant ceux des ceintures de roches vertes qui se composent de schistes à andalousite, à cordiérite, à sillimanite, à staurotide ou à grenat et de quantités secondaires de roches métavolcaniques, de marbres, de roches calco-silicatées, d’amphibolites et de roches ferrugineuses. Cette ceinture d’une largeur de 2,5 à 4 km aux bordures mylonitisées présente des unités fortement inclinées dont l’organisation est dérangée par des massifs de granitoïde. Une

probable Aphebian age so was probably metamorphosed twice. Pressure of (?)Aphebian metamorphism is inferred to have been about 0.25 GPa and temperature, estimated from garnet-biotite thermometry, ranged from 510°C to 564°C. The Franklin belt outcrops in a high-strain zone in northeastern Ian Calder Lake map area. The belt, 32 km long and up to 7 km wide, comprises metaconglomerate, fuchsitic orthoquartzite, marble and calc-silicate rocks, amphibolite, and mica schist, tectonically interleaved with, and disrupted by, granitoid gneiss. Deformed amphibolite dykes cut the belt. Although on strike with the Aphebian Chantrey supracrustal belt to the northeast, the Franklin belt is correlated with the Archean Barclay belt, on the basis of the presence of amphibolite, severe deformation, and dismemberment by Archean granitoid rock in both belts. In southern Ian Calder Lake map area, cordierite-andalusite and (fibrolite-)garnet mica schists, quartzite, marble, and calc-silicate rock structurally underlie the Aphebian Montresor Group. Foliation in the supracrustal rocks is concordant with foliation in the adjacent granitoid basement but discordant to bedding in the Montresor Group; however, no contacts were found. Because they are intruded by deformed granitic sheets and discordant to the Montresor Group, the supracrustal rocks are inferred to be Archean.

granodiorite à faible gneissosité que l'on croit recouper les schistes a été datée par la méthode U-Pb sur zircon à environ 2,59 Ga. La ceinture de Barclay est également traversée par une pegmatite qui remonte probablement à l'Aphébien et qui a donc par conséquent subi deux épisodes métamorphiques. On déduit que le métamorphisme aphébien(?) s'est déroulé dans des conditions de pression d'environ 0,25 GPa et de température de l'ordre de 510 à 564 °C, d'après le géothermomètre biotite-grenat. La ceinture de Franklin affleure dans une zone de forte déformation dans la partie nord-est de la région cartographique de Ian Calder Lake; d'une longueur de 32 km et d'une largeur pouvant atteindre 7 km, celle-ci se compose de métaconglomérats, de grès quartzeux à fuschite, de marbres et de roches calco-silicatées, d'amphibolites et de micaschistes qui sont tous interstratifiés tectoniquement avec des gneiss granitoïdes et dérangés par ceux-ci. Des dykes d'amphibolite déformés recoupent cette ceinture. Bien que située dans l'alignement de la ceinture de roches supracrustales de Chantrey de l'Aphébien au nord-est, la ceinture de Franklin est mise en corrélation avec la ceinture de Barclay de l'Archéen d'après l'existence dans les deux ceintures d'amphibolites, d'une intense déformation et d'un démembrement par des roches granitoïdes de l'Archéen. Dans la partie sud de la région cartographique de Ian Calder Lake, des micaschistes à cordiérite-andalousite et à (fibrolite-)grenat, des quartzites, des marbres et des roches calco-silicatées reposent structuralement sous le Groupe de Montresor de l'Aphébien. La foliation dans les roches supracrustales est concordante avec celle qui existe dans le socle granitoïde adjacent, mais discordante par rapport à la stratification dans le Groupe de Montresor; aucun contact n'a cependant été observé. Puisqu'elles sont recoupées par des feuillets de granite déformés et discordantes par rapport au Groupe de Montresor, on déduit que les roches supracrustales datent de l'Archéen.

Proterozoic supracrustal rocks are confined to two belts of Aphebian age. The Chantrey Group forms a belt, 175 km long and 3–7 km wide, extending northeastward from near the head of Chantrey Inlet and terminating in northern Darby Lake map area, where it trends easterly. Wholly sedimentary in origin, the Chantrey Group consists of marble and calc-silicate rocks (locally stromatolitic), quartzite (including orthoquartzite and ferriferous varieties), andalusite- and garnet-bearing, locally graphitic, mica schist, and metaconglomerate, which is cut by pegmatite with a ^{39}K-^{40}Ar muscovite age of 1684 Ma. The calcareous rocks occur in the lower half, and the metaconglomerate appears to be at the top, of the succession but the stratigraphic order is uncertain. The Chantrey Group is tightly folded into a canoe-shaped synclinorium plunging 20° at each end. Contacts with basement are typically strongly sheared and regarded as tectonic. Intensity of deformation and metamorphism varies along the length of the belt. Thrusting to the northwest and north has caused local overturning of the southern Chantrey Group–basement contact and recumbent folding in the supracrustal rocks. Fibrolite occurs near the western end of the belt. As in the Barclay belt, metamorphism was of low-P,T Buchan type but temperatures were generally lower, ranging from 468°C to 510°C (but rising to 572°C in fibrolite-bearing rock) at an estimated pressure of 0.25 GPa. The protolith of the

Des roches supracrustales du Protérozoïque sont confinées à deux ceintures d'âge aphébien. Le Groupe de Chantrey forme une ceinture longue de 175 km et large de 3 à 7 km, qui s'étire vers le nord-est depuis un point situé près du fond de la baie Chantrey pour se terminer dans le nord de la région cartographique de Darby Lake où elle adopte une direction est. Entièrement d'origine sédimentaire, les lithologies du Groupe de Chantrey se composent de marbres et de roches calco-silicatées (stromatolitiques par endroits), de quartzites (incluant des grès quartzeux et des variétés ferrifères), de micaschistes à andalousite et à grenat, graphitiques par endroits, et de métaconglomérats; toutes ces lithologies sont recoupées par une pegmatite datée par la méthode ^{39}K/^{40}Ar sur muscovite à 1 684 Ma. Les roches calcareuses se trouvent dans la moitié inférieure alors que les métaconglomérats semblent occuper la partie supérieure de la succession, mais l'ordre stratigraphique est incertain. Le Groupe de Chantrey est déformé par des plis serrés définissant un synclinorium en forme de canot dont chacune des extrémités plonge à 20°. Les contacts avec le socle sont de manière caractéristique fortement cisaillés et seraient de nature tectonique. L'intensité de la déformation et du métamorphisme varie le long de la ceinture. Dans la partie sud de la zone d'affleurement du Groupe de Chantrey, des mouvements de charriage vers le nord-ouest et le nord ont engendré par endroits un renversement du contact de cette unité avec le socle, ainsi que la formation de plis couchés dans les roches supracrustales. On trouve de la fibrolite près de l'extrémité ouest de la ceinture. Comme dans la ceinture de Barclay, un métamorphisme de type Buchan s'y est déroulé dans des conditions de faible pression et de faible température, mais les températures y

Chantrey Group is interpreted as having been deposited in a shallow inland sea. This basin, the site of which is now marked in part by the steep Murchison fault, probably formed by Proterozoic reactivation of the shear zone in which the Franklin belt lies.

ont généralement été inférieures, s'échelonnant de 468 à 510 °C (mais atteignant 572 °C dans la roche renfermant de la fibrolite), et la pression y est estimée à 0,25 GPa. Les protolites du Groupe de Chantrey auraient été déposés dans une mer intérieure peu profode. Ce bassin, dont la position correspond aujourd'hui en partie à celle de la faille de Murchison à fort pendage, s'est probablement formé par réactivation au Protérozoïque de la zone de cisaillement dans laquelle est présente la ceinture de Franklin.

The other Aphebian supracrustal sequence is the Montresor Group, exposed in a synclinal belt 65 km long and 8 km wide and in a neighbouring, more open structure up to 11 km across, north of the Back River in Ian Calder Lake map area. The Montresor Group succession is about 1800 m thick and comprises meta-arenite (including meta-arkose and orthoquartzite) and subordinate marble, mica schist, and phyllite, which are concentrated near the base of the section. Disseminated fibrolite in muscovite-bearing orthoquartzite indicates lower amphibolite-facies metamorphism. The main Montresor belt is a northeast-trending, open syncline, the axis of which is warped and plunges 15° alternately southwest and northeast. Contacts are rarely exposed but the Montresor Group appears to unconformably overlie Archean granitoid and supracrustal rocks and is cut by (?)Aphebian pegmatite. The Montresor Group, like the Chantrey Group, is probably of shallow-marine origin.

L'autre séquence de roches supracrustales de l'Aphébien est le Groupe de Montresor. Cette unité affleure dans une zone synclinale longue de 65 km et large de 8 km ainsi que dans une structure voisine plus ouverte d'une largeur pouvant atteindre 11 km au nord de la rivière Back, dans la région cartographique de Ian Calder Lake. La succession du Groupe de Montresor est d'une épaisseur d'environ 1 800 m et se compose de méta-arénites (incluant des méta-arkoses et des grès quartzeux) et, en quantités secondaires, de marbres, de micaschistes et de phyllites qui sont concentrés près de la base de la coupe. De la fibrolite disséminée dans des grès quartzeux à muscovite indiquent des conditions de métamorphisme propres au faciès des amphibolites inférieur. La ceinture de Montresor principale est déformée en un synclinal ouvert de direction nord-est dont l'axe est gauchi et qui plonge de 15° successivement vers le sud-ouest et le nord-est. Les contacts affleurent rarement, mais le Groupe de Montresor semble reposer en discordance sur les roches granitoïdes et les roches supracrustales de l'Archéen et est recoupé par une pegmatite de l'Aphébien(?). Les roches du Groupe de Montresor, comme celles du Groupe de Chantrey, ont probablement été déposées en milieu marin peu profond.

Undeformed pegmatite bodies, intrusive into granitoid and supracrustal rocks, presumably are offshoots of Aphebian granitic intrusions at depth, none of which was recognized in outcrop.

Des masses de pegmatite non déformées, qui recoupent à la fois les roches granitoïdes et les roches supracrustales, sont sans doute des apophyses d'intrusions granitiques de l'Aphébien sises en profondeur mais qui n'ont pas été observées en affleurement.

Sparse diabase dykes of the same age as Mackenzie dykes (1.27 Ga) trend northwest and represent the youngest bedrock in the area.

Des dykes de diabase épars du même âge que les dykes de Mackenzie (1,27 Ga) présentent une direction nord-ouest et constituent les lithologies du substratum rocheux les plus récentes dans la région.

Rocks of potential economic interest occur at the Montresor Group–basement contact and in the Chantrey belt. At one locality at the western end of the Montresor belt, arsenopyrite is present in basement leucogranite adjacent to rusty Montresor Group schist containing trace native copper. Dark, graphitic and pyritiferous mica schist in the upper part of the Chantrey Group characteristically weathers rusty, contains numerous gossans, and locally shows elevated metal contents.

Les lithologies pouvant présenter un intérêt du point de vue de la présence de minéralisations utiles se trouvent au contact du Groupe de Montresor avec le socle dans la ceinture de Chantrey. En un endroit, à l'extrémité ouest de la ceinture de Montresor, de l'arsénopyrite a été observée dans un leucogranite du socle tout à côté de schistes oxydés du Groupe de Montresor renfermant des traces de cuivre natif. Les micaschistes graphitiques et pyriteux de couleur foncé de la partie supérieure du Groupe de Chantrey prennent de manière caractéristique une couleur rouille à l'altération, renferment de nombreux chapeaux de fer et présentent par endroits des teneurs élevées en métaux.

INTRODUCTION

This report describes the geology of the Canadian Shield in an area straddling upper Chantrey Inlet in south-central Nunavut. The study area comprises two 1:250 000 scale map areas and part of a third: Ian Calder Lake, lying southwest of Chantrey Inlet, and Cape Barclay and part of Darby Lake, east of the inlet (Fig. 1). A focus of the study was the early Proterozoic, metasedimentary Chantrey belt, which lies in a northeast-trending structural zone that runs diagonally through the area. Two 1:250 000 scale geological maps accompany the report (maps 1779A, 1780A, in pocket; the reader will note that because they were printed at different times, this bulletin reflects the new name of the area, Nunavut, whereas the maps were printed with the old name, Northwest Territories).

Location, access, and wildlife

Ian Calder Lake map area (NTS 66-I) lies astride the Arctic Circle (66°30′ N) and is bounded by latitudes 66°N and 67°N and longitudes 96°W and 98°W. The bounds of Cape Barclay map area (NTS 56 M), which includes upper Chantrey Inlet, are 67°N and 68°N, 94°W and 96°W. The study area also includes part of northern Darby Lake map area (NTS 56 N) between 67°40′ N and 68°N and the southeastern corner of Sherman Basin map area (NTS 66 P) at 67°N, 96°W. The names of Ian Calder Lake and Cape Barclay map areas were formerly Montresor River and Lower Hayes River, respectively. Although the name changes were approved in the latter part of the 1970s, the new names did not appear in print until the mid-1980s.

The study area is uninhabited. The human settlement nearest to Ian Calder Lake map area is Baker Lake, 240 km south of the centre of the map area. The area of upper Chantrey Inlet is most easily reached from Gjoa Haven on King William Island, nearly 170 km north of the confluence of the Back and Hayes rivers at the head of Chantrey Inlet. Baker Lake and Gjoa Haven enjoy regular air service and are annually supplied by sealift. A sport fishing camp operates briefly during summer on the Back River east of Franklin Lake.

Caribou, singly and in small groups, are common throughout the region. A few muskoxen were seen near the Hermann River in southeastern Ian Calder Lake map area. Lake trout, Arctic char, and, locally, grayling abound in the lakes and streams. Queen Maud Gulf Bird Sanctuary extends into northwestern Ian Calder Lake map area and hosts abundant breeding geese and swans.

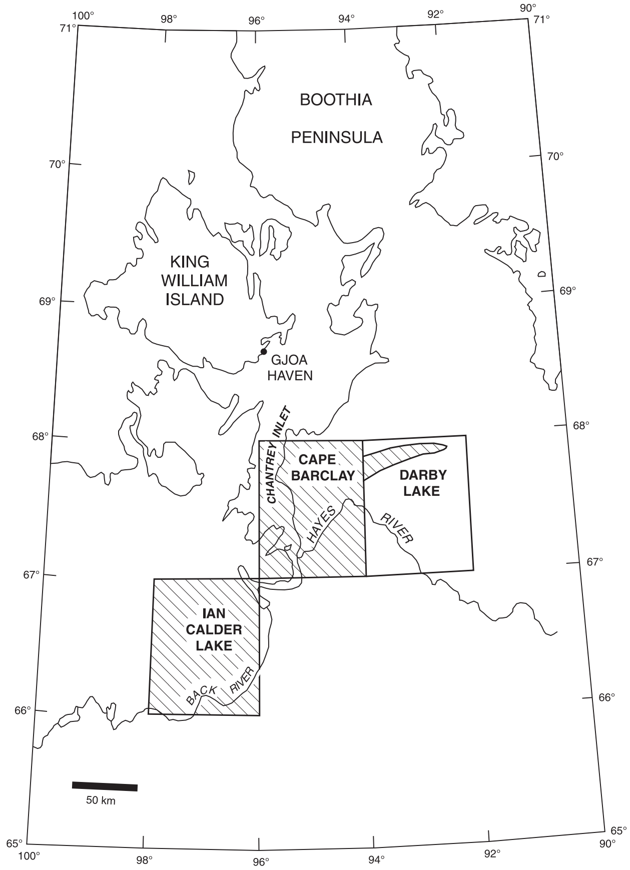

Figure 1. *Location of the study area. Areas hachured are covered by the two geological maps in pocket, Map 1779A Cape Barclay and part of Darby Lake, and Map 1780A Ian Calder Lake.*

Physiography

The study area overlaps two physiographic subdivisions of the northwestern Canadian Shield (Dyke and Dredge, 1989): Ian Calder Lake map area lies within the eastern Back lowland, whereas Cape Barclay and Darby Lake map areas belong to the western Wager plateau. There is, however, no significant physiographic break at the boundary between the two subdivisions, one grading into the other. Ian Calder Lake is predominantly a lowland with relief typically less than 60 m and myriad lakes connected by shallow streams. The lakes commonly are linear and aligned north-south, reflecting the trend of ice flow during the last glaciation. Relief is slightly

Figure 2. *Badlands topography in unconsolidated glacial marine sands and silts resting on the Archean granitoid complex in the valley of a tributary of the Hayes River. View to the south. Photograph by T. Frisch. GSC 204635-L*

higher (100 m) south of the Back River, which flows across the southern part of the map area in a reasonably well defined valley.

Cape Barclay map area is dominated by upper Chantrey Inlet in the west and the coalescing deltas of the Back and Hayes rivers in the south (*see* Frontispiece). South of the Hayes River, the edge of the Wager plateau rises to elevations exceeding 300 m above sea level. The valleys of the Hayes River and its tributaries are filled by thick deposits of glacial silt and sand fluvially eroded to a badlands topography (Fig. 2). North of the Hayes River, the terrain of the Chantrey supracrustal belt and its borders is a gently rolling upland at an elevation of about 150 m a.s.l., with the Chantrey belt itself occupying an almost imperceptible, trough-like depression containing linear lakes controlled by the northeast-trending bedrock structure (Fig. 3). Towards the east, the terrain of the Chantrey belt rises to more than 300 m a.s.l. north of Darby

Figure 3. *View eastward across the central part of the Chantrey belt, which occupies a subdued, trough-like depression marked by linear lakes. In the right distance is the 'great bend' of the Hayes River, outlined by light-coloured glacial deposits (Fig. 2). NAPL T447L-7*

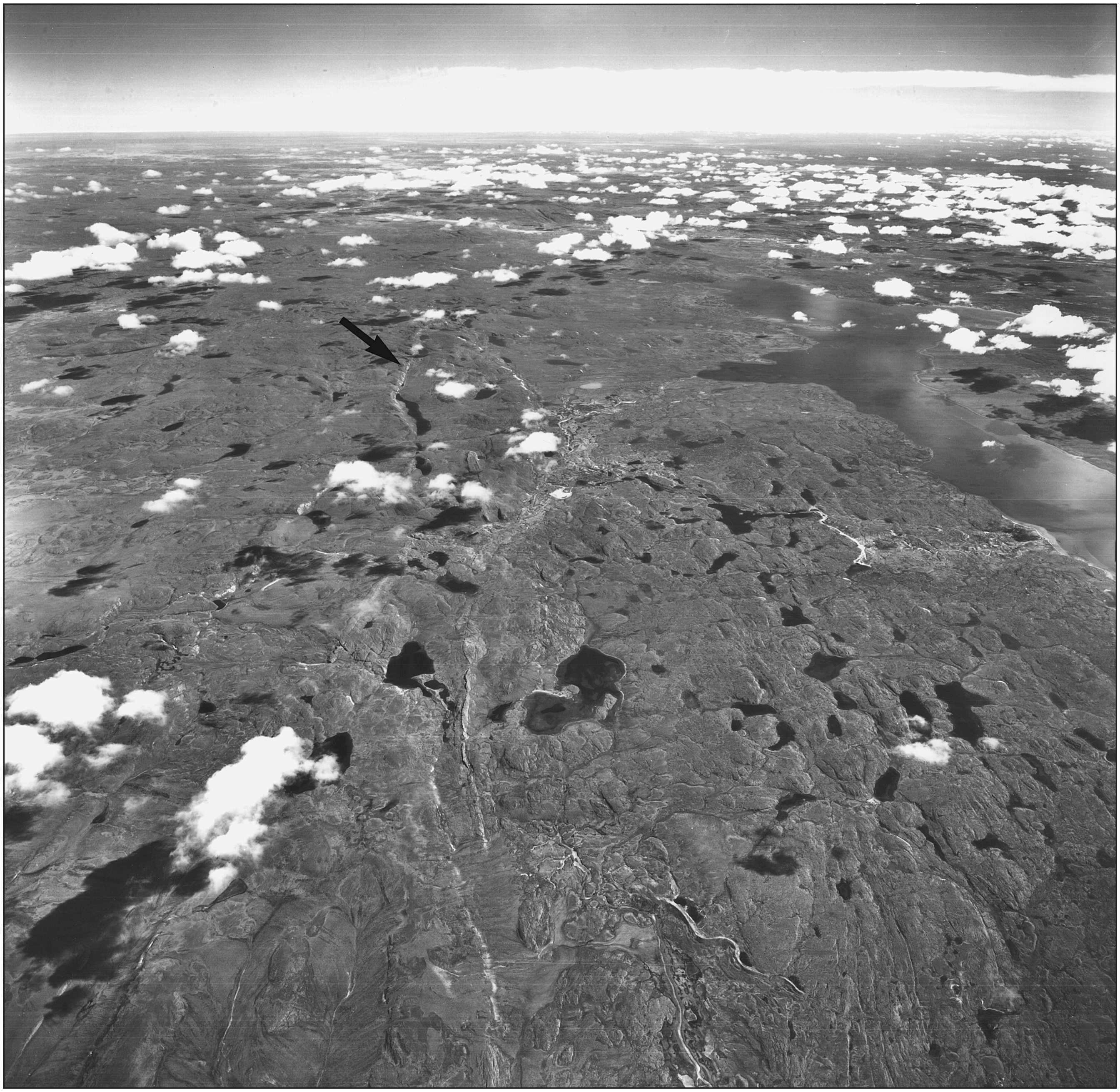

Figure 4. *Eastward view of the eastern end of the Chantrey belt, with Darby Lake on the right. The belt is partly outlined by light-coloured marble and calc-silicate rock. The arrow indicates the area shown in Figure 23. NAPL T469R-92*

Lake (Fig. 4). The northern part of Cape Barclay map area is a heavily drift-covered lowland predominantly 100 m or so above sea level.

Highlights of early travel and exploration

Chantrey Inlet, strategically located on the central Arctic mainland coast and providing access to the interior via the Back River, attracted nineteenth century explorers, particularly search parties looking for survivors of expeditions to the Northwest Passage. In 1834, the British naval officer, George Back, travelled down the river that was subsequently to be named after him and the length of Chantrey Inlet before retracing his route home. Back named many geographical features in the area and his maps of the Back River remained the only ones in use until as recently as 1948 (Holland, 1994). Twenty-one years after Back's journey, James Anderson and James G. Stewart of the Hudson's Bay Company followed a similar route, looking for relics of the Franklin expedition, which had come to grief north of King William Island and whose survivors attempted to reach the Back River on foot in 1848 (Holland, 1994). In 1879–1880, Frederick Schwatka, a United States Cavalry officer, led a Franklin search expedition on a monumental sledge journey of 5300 km in 50 weeks from the west side of Hudson Bay to King William Island and back via the west side of Chantrey Inlet and the Back River (Holland, 1994). On the outbound leg, Schwatka discovered and named the Hayes River, from whose "great bend" (Fig. 3) he and his party struck out westwards across country to Chantrey Inlet (Gilder, 1881).

Previous work

The study area was included in the 90 000 km^2 area of Operation Back River, a helicopter-borne geological reconnaissance in the District of Keewatin in 1960 by the Geological Survey of Canada (Heywood, 1961). Among the results of this work were bedrock and surficial geological maps at scales of about 1:500 000 and 1:1 000 000, respectively (Heywood, 1961; Craig, 1961).

Airborne magnetic surveys of the study area and surrounding region were conducted in the 1970s and the results were published in the form of contoured aeromagnetic maps on the NTS 1:250 000 scale base (Geological Survey of Canada, 1977a, b, c, d, e).

The surficial geology of the study area was mapped on a reconnaissance scale in 1976 and 1977, as part of a feasibility study for a proposed eastern Arctic gas pipeline (Thomas, 1977, 1982; Thomas and Dyke, 1982a, b). The distribution of Quaternary deposits shown on the geological maps accompanying this report is taken from Thomas (1982) and Thomas and Dyke (1982a, b).

Present work

Reconnaissance bedrock mapping of the study area was undertaken over the course of the field seasons of 1982 and 1984, each season lasting 9–10 weeks from mid-June to late August. In 1982, Ian Calder Lake map area was mapped by the writer, assisted by Judith G. Patterson, by means of ground and helicopter traverses, operating from a base camp on the north bank of the Back River northwest of Mount Meadowbank (from mid-June to late July) and a base camp 60 km to the north (late July to late August). For the 1982 work, the helicopter, an Enstrom Model 280C, was shared with two other field parties working to the south.

Mapping in the Cape Barclay and Darby Lake map areas and minor follow-up work in northeastern Ian Calder Lake map area were accomplished in 1984 from a base camp on Irby and Mangles Bay on the east shore of Chantrey Inlet. A Bell Helicopter Textron 206B was dedicated to the field party for the entire season. Fieldwork in the Chantrey and Barclay supracrustal belts was mainly done by ground traversing by Irvine A. Annesley and Catherine A. Gittins and, to a lesser extent, the author. Mapping of the Archean granitoid complex in the study area was done almost entirely by the author on helicopter traverses.

Chemical analyses of rocks were performed by the Analytical Chemistry section of the Geological Survey of Canada in Ottawa by XRF and rapid chemical methods.

Mineral analyses were made by the writer with a Camebax electron microprobe, using automated, wavelength-dispersive techniques, at the Geological Survey of Canada, Ottawa.

Rock and mineral analyses and the locations of chemically and isotopically analyzed samples are tabulated in the Appendix.

Acknowledgments

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Thanks are due B. and E. Kotelewitz for expediting services in Baker Lake in 1982; C.R. Dickie, Executive Service Officer, N.W.T. Government, in Gjoa Haven in 1984, for invaluable help and hospitality; G. Qitsualik of Gjoa Haven for his hard work caching camp supplies in the field; and the staff of the Shepherd Bay DEW Line station for handling food shipments from the south in 1984.

The 1984 field program was generously supported by the Polar Continental Shelf Project. Polar Continental Shelf Project furnished 53 hours of Twin Otter and 50 hours of helicopter flying time; both aircraft were on charter to PCSP. Polar Continental Shelf Project also provided accommodation for the field party at Resolute before and after the 1984 fieldwork. Particular thanks are due F.P. Hunt, B. Hough, and staff of PCSP for numerous courtesies at Resolute and C. Zberg (Twin Otter pilot-in-command) and his co-pilots of Bradley Air Services for indefatigable flying and sheer hard work.

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GENERAL GEOLOGY

The study area is underlain entirely by crystalline bedrock of the Churchill structural province of the Canadian Shield and lies in the west-central part of the Archean Rae Province of Hoffman (1989). The bedrock consists overwhelmingly of late Archean, amphibolite-grade, gneissic granitoid rock of plutonic aspect and with northeasterly foliation. Straight gneiss in the northwest corner of Ian Calder Lake map area may belong to the easternmost Queen Maud block (Heywood and Schau, 1978). An easterly to northeasterly trending, 15 km wide belt of Archean, granulite-grade rocks is exposed in northeastern Ian Calder Lake and southeastern Cape Barclay map areas.

Four northeast-trending, amphibolite-grade supracrustal belts, 5–10 km wide, are present. Two are highly deformed and disrupted by granitoid rocks and are inferred to be Archean and correlative: the Franklin belt in northeastern Ian Calder Lake map area and the Barclay belt in northern Cape Barclay map area. Each contains schist, quartzite, marble, metaconglomerate, and amphibolite and the Barclay belt includes acid to intermediate metavolcanic rocks and minor iron-formation, suggestive of greenstone belt affinity.

The other two supracrustal belts, the Chantrey and Montresor, are Aphebian and consist entirely of sedimentary rock metamorphosed in lower amphibolite facies, forming doubly plunging synclinal successions. The Chantrey Group, comprising pelitic and psammitic schist, quartzite, marble, and metaconglomerate, forms a tightly folded and locally thrust-faulted belt, extending 175 km from southwestern Cape Barclay map area into northern Darby Lake map area. The Montresor belt crosses the southern half of Ian Calder Lake map area and consists primarily of quartzite and meta-arkose and subordinately of marble and schist, folded into an open syncline with a gently warped axis; maximum thickness of the succession is estimated to be 1800 m. Apparently unconformably underlying the Montresor rocks are amphibolite-grade supracrustal rocks presumed to be of Archean age. Both Chantrey and Montresor groups are intruded by pegmatite but no other granitoid rock of post-Archean age has been recognized in the study area.

The youngest bedrock in the study area is diabase in the form of northwest-trending dykes presumed to be of the same age as Mackenzie dykes.

Unconsolidated Quaternary cover of glacial origin is locally extensive. Flow direction in the Keewatin Ice of the Laurentide Ice Sheet that covered the area in Late Wisconsinan (late Pleistocene) time was generally northward (Thomas, 1977). Of particular note among glacial features are stratified deposits of silt and fine sand, up to 75 m thick, in the valleys of the Hayes River and its tributaries (Fig. 2) and at its confluence with the Back River in Cape Barclay map area (*see* Frontispiece). These deposits are distal marine sediments laid down during final retreat of the Keewatin Ice from the area stretching from south of Committee Bay to south of Chantrey Inlet (Dyke and Dredge, 1989).

ARCHEAN GRANITOID COMPLEX

The bulk of the study area is an Archean plutonic terrane in the northwestern Churchill structural province or western Rae province of Hoffman (1989). This infracrustal terrane of mainly granitoid rocks encloses and presumably underlies the Archean and Proterozoic supracrustal belts; Archean supracrustal rocks are locally intruded by rock assigned to the granitoid complex. Plutonic ages obtained from the complex are predominantly late Archean and Proterozoic igneous activity appears to have been very limited but it must be emphasized that our knowledge of the northwestern mainland Churchill Province is far from complete.

In this report, ‘gneiss’ refers to quartzofeldspathic rocks in which mineral grains are arranged in planar structures, i.e. foliated quartzofeldspathic rocks. The term thus includes slightly foliated, homogeneous granitoid rocks that are simply deformed plutonic bodies but which could not always be distinguished from gneiss *sensu stricto* during reconnaissance mapping.

Mixed gneisses (unit At*)*

This unit includes a variety of gneisses, mainly granitoid, east of the Montresor belt in the southern part of the Ian Calder Lake map area. They trend northeasterly and generally dip northwesterly and are commonly mylonitic near the southern border of the map area.

The most common rock type is an augen gneiss in which the augen are predominantly of pink microcline, similar to much of the rock of unit **A**k. The augen average 0.5–1 cm in length and are white on fresh surfaces and pink on weathered surfaces. Generally, the matrix has an irregular grain size (fine to medium) and is recrystallized. It is typically composed of brown, slightly chloritized biotite, oligoclase, microcline, and quartz; allanite and sphene are common accessory minerals. Colour index of the augen gneiss ranges from 20 to 25, significantly lower than that of unit **A**k. Another common type of augen gneiss, with a similar colour index, is one in which the augen of K-feldspar are accompanied by smaller (up to 5 mm) ones of normally zoned plagioclase $An_{25–30}$.

Tonalitic varieties of gneiss, present locally, particularly near the southern border of Ian Calder Lake map area, have crystals, up to 4 mm long, of oligoclase (generally about An_{25}) in a matrix of grass-green hornblende, brown biotite (generally partly chloritized), oligoclase, a little microcline, and quartz; sphene is a major accessory mineral. The tonalite gneiss tends to be darker than the microcline-rich gneiss, having a colour index of 30–35.

All these gneiss types commonly are veined by pink or white granite and, especially, pegmatite. The granite veins tend to parallel the gneissosity and are themselves gneissic, whereas the pegmatite veins are slightly to markedly discordant.

Mylonitic structure is prominent in the gneiss at several localities, particularly in the southeastern corner of Ian Calder Lake map area, which is skirted by the northeast-trending Amer mylonite zone (Fig. 5; Tella, 1994).

Also included in unit **A**t are amphibolite and anorthosite gneiss. Amphibolite occurs as isolated outcrop-sized bodies concordant with the surrounding gneiss. It weathers black and consists mainly of olive-green hornblende, commonly altered to pale green amphibole, and well twinned calcic

Table of Formations

Eon	(Sub)Era	Period	Formation	Unit	Lithology
	Cenozoic	Quaternary		Q	Gravel, sand, silt
Unconformity					
Proterozoic	Neohelikian		Mackenzie	db	Diabase dykes
	Intrusive contact				
	Aphebian				Pegmatite
		Intrusive contact			
			Montresor Group	*A*Ma	Meta-arkose, quartzite, metapelite
				*A*Mt	Marble, calc-silicate rock
				*A*Mo	Orthoquartzite
			Chantrey Group	*A*Cc	Metaconglomerate
				*A*Cp	Dark pelitic schist
				*A*Co	Quartzite, orthoquartzite
				*A*Cs	Metasiltstone/mudstone
				*A*Cq	Impure quartzite, schist
				*A*Cm	Marble, calc-silicate rock
(?)Unconformity					
Archean				**A**a	Amphibolite, metagabbro
				Ad	Marble, calc-silicate rock
				Aq	Orthoquartzite
				Ap, **A**m	Schist, quartzite, calc-silicate rock, ironstone
				Ac	Metaconglomerate, schist
				Agr	K-feldsparphyric granite
				Amg	Garnet-muscovite granite
				Agd	Hornblende-biotite granodiorite gneiss
				Ak	K-feldspar augen gneiss
				At	Hornblende-biotite tonalite gneiss
				Ao, **A**g	Orthopyroxene gneiss and granite, garnet-cordierite-sillimanite gneiss
				Ab, **A**s	Hornblende-biotite granodiorite gneiss, augen gneiss, (?) amphibolite metadykes
Stratigraphic order of Archean and Aphebian units is uncertain					

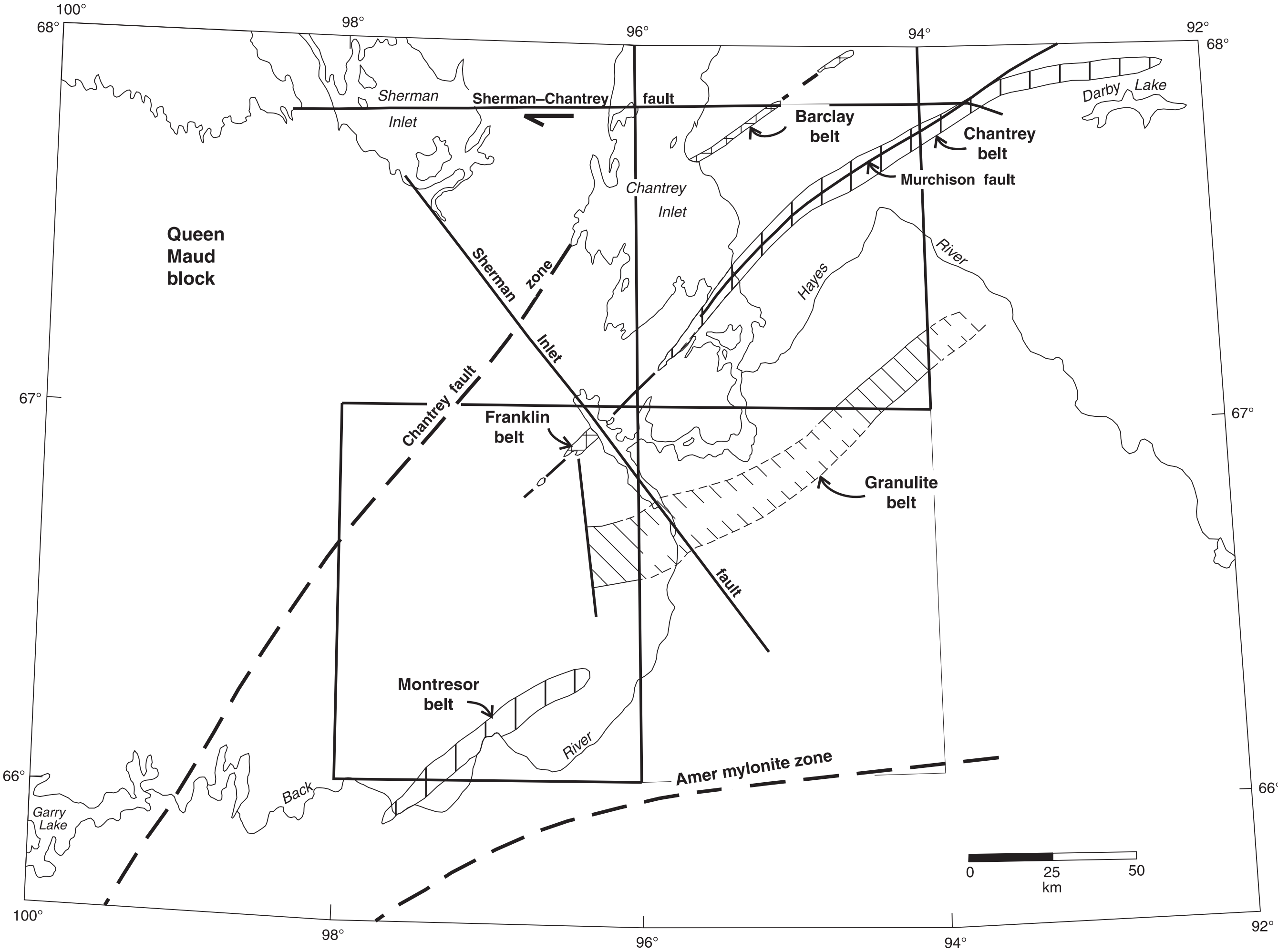


Figure 5. Major structural elements of the study area and surrounding region, based on data from the present work, Heywood and Schau (1978), Hoffman (1989), and Tella (1994). Ian Calder Lake and Cape Barclay map areas heavily outlined, Mistake River map area lightly outlined.

plagioclase in a granoblastic texture showing good equilibrium triple junctions between the grains. Veining by pegmatite is locally evident.

Anorthositic rocks were found at a helicopter stop near the Back River, 27 km east-southeast of Mount Meadowbank (Ian Calder Lake map area). The outcrops examined are composed of well layered, highly deformed, very light-weathering gneiss with amphibolite layers up to 1.5 m wide, all cut by dykes of biotite granite. Two gneiss samples were examined in thin section. One is a biotite-hornblende tonalite gneiss consisting of plagioclase (An_{35}), grass-green hornblende with a little associated epidote, brown biotite, and seven volume per cent quartz; grain size averages 1 mm and colour index is 30. The other rock is hornblende anorthosite gneiss composed of well twinned labradorite, pale-green hornblende (?Mg-rich) partly altered to nearly colourless chlorite, and a little clinozoisite; grain size averages 1.5 mm and colour index is less than 10. Both rocks show excellent equilibrium triple-junction texture characteristic of recrystallized igneous rocks.

The rocks of unit **A**t appear to be of entirely igneous origin. There are no recognizable metasedimentary rocks interlayered with the gneiss. Phenocrysts of plagioclase with well developed normal zoning and the euhedral tablet form of microcline megacrysts in many rocks provide good evidence of igneous processes. Lastly, the amphibolite and anorthosite gneiss are indisputably meta-igneous rocks.

Augen gneiss (unit Ak*; unit* Ab *in Cape Barclay and Darby Lake map areas)*

Most of the central part of Ian Calder Lake map area is underlain by a granodioritic gneiss that is commonly mafic and characterized by centimetre-sized K-feldspar augen. Its area of outcrop is notably featureless in the aeromagnetic map of the region (Geological Survey of Canada, 1977a). Augen gneiss is also a major component of unit **A**b of the Cape Barclay and Darby Lake map areas.

Figure 6. *Mylonite in augen gneiss of unit* **A**k, *12 km southeast of the Chantrey fault zone, western Ian Calder Lake map area. Although dextral shear is evident in this outcrop, no predominant movement direction was recognized in the area of the fault zone. Photograph by T. Frisch. GSC 204075*

The most common variety of augen gneiss consists of 2–4 cm long K-feldspar augen, white weathering to pink, in a medium-grained matrix rich in biotite. Colour index ranges from 20 to 40, with most values exceeding 30. The rock typically is darker than the augen gneiss of unit **A**t. Deformation has generally been strong, evidenced by stretched and flattened augen, cataclastic texture in the matrix, and thin mylonitic layers of fine-grained gneiss. Locally, particularly in and near the Chantrey fault zone (Fig. 5), deformation of the augen gneiss was high enough for thin layers of extremely fine-grained mylonite to form (Fig. 6). The latter is especially noticeable in thin sections, which typically show mortar texture of quartzofeldspathic matrix and ribbons of quartz grains. The K-feldspar of augen and matrix is microcline perthite. Plagioclase, which may occur in small augen as well as smaller grains in the matrix, is sericitized oligoclase. Biotite is dark brown in thin section, commonly partly chloritized, and associated with abundant accessory sphene. Some rocks contain a little late muscovite, probably formed from alteration of feldspar.

Essentially massive, coarse-grained, porphyritic biotite granite–granodiorite outcrops east of the Montresor River between Bromley Lake and Ian Calder Lake and appears to grade into, in all directions, augen gneiss. This rock must be considered a likely protolith for at least part of unit **A**k.

Augen gneiss containing large plagioclase, in addition to microcline, augen are found in units **A**k and **A**b but are very subordinate to the potassic variety in unit **A**k. Plagioclase-microcline augen gneiss has bluish-green hornblende as well as brown biotite and shows deformation features similar to those in the K-feldspar augen gneiss.

Both main varieties of augen gneiss locally contain concordant mafic lenses of fine-grained biotite gneiss, which appear to be metapelitic and therefore of supracrustal origin. Almost ubiquitous is pegmatite, in concordant and discordant veins and in pods.

Despite its locally well preserved intrusive igneous aspect, the augen gneiss is certainly one of the older rock units in the study area. Rocks seen in contact with the augen gneiss appear to postdate it, as intrusive relations were not observed. In fact, unit **A**k may be exceeded in age only by the biotite gneiss that occurs as inclusions in it. A sample (FS82-15) of massive, porphyritic granodiorite of unit **A**k, lying between Bromley and Ian Calder lakes, has yielded a Sm-Nd T_{DM} age (“crust formation” age based on the depleted mantle model of DePaolo (1988)) of 2.78 Ga (E. Hegner, pers. comm., 1991). Augen gneiss from western Chantrey Inlet, presumably coeval with the gneiss of unit **A**k, has given late Archean U-Pb zircon and Sm-Nd model ages of 2.6 Ga and 2.7 Ga, respectively (Frisch and Parrish, 1992); these are discussed below in the section on granitoid rocks of Cape Barclay map area.

Relict porphyritic texture, largely uniform quartzofeldspathic mineralogy, and relatively high mafic mineral content suggest an igneous origin for the augen gneiss. Uranium-lead zircon ages from the Chantrey Inlet area and elsewhere in this part of Nunavut (LeCheminant and Roddick, 1991) indicate major felsic magmatism occurred at 2.6 Ga.

Gneissic to massive granodiorite-diorite and grey biotite gneiss (unit **A**gd*)*

Rocks assigned to this unit are restricted to the area around Franklin Lake in the northeast corner of Ian Calder Lake map area. They are distinguished from the typical augen gneiss by their grey colour, generally equigranular texture, and lower colour index.

Most of unit **A**gd west of Franklin Lake consists of gneissic, felsic granodiorite containing bluish-green hornblende and dark brown biotite as the chief mafic minerals; quartz; and oligoclase predominating over microcline. Coarser grained, more or less massive diorite to granodiorite forms several bodies up to several hundred metres long, concordant with the enclosing gneiss.

The granitoid complex east of Franklin Lake in Ian Calder Lake map area comprises mainly grey biotite gneiss and metagabbro and amphibolite bodies (**A**a). The biotite gneiss, like the granitoid rocks on the opposite side of Franklin Lake, is relatively leucocratic and readily distinguishable from the typical augen gneiss elsewhere in the map area. Locally, the gneiss includes lenses of amphibolite or mafic schist, some of which may be deformed metagabbro; these are probably remnants of the larger bodies that were mapped separately as unit **A**a.

Along the borders of the Franklin belt, the supracrustal belt that crosses Franklin Lake, the granitoid rocks are strongly sheared to mylonitic. They exhibit crush texture, smearing out and reddening of feldspar, and chloritization of mafic minerals. Interleaving of gneissic and supracrustal rocks occurs on all scales and, on an outcrop scale, may result in gneiss appearing to intrude supracrustal rock. The intense deformation, however, leaves little doubt that such contacts are tectonic.

Grey gneiss with amphibolite, northern Ian Calder Lake map area (units **A**b, **A**s*)*

The bulk of these rocks are well layered, grey granodioritic gneiss with abundant amphibolite sheets, which commonly outline tight isoclinal folds. Units **A**b and **A**s lie within the Slave–Chantrey mylonite zone as defined by Heywood and Schau (1978) and characteristically are strongly strained.

Unit **A**b consists mainly of grey, (hornblende-)biotite granodiorite gneiss with concordant sheets of amphibolite (Fig. 7). The gneiss is typically medium grained and extremely well layered and locally mylonitic. The layering is probably due largely to tectonic processes, such as flattening and stretching. In places, granitic gneiss is thinly interlayered with the granodiorite gneiss and pink K-feldspar porphyroblasts may be developed. Rarely, granitic gneiss, without significant amounts of associated amphibolite, predominates but is invariably as deformed as the granodioritic gneiss.

In thin section, the nonmylonitic granodiorite gneiss is seen to have a modified granoblastic-polygonal texture, with straight-sided grains 0.2–0.5 mm in diameter. Well twinned plagioclase (An_{30-35}) greatly exceeds microcline in abundance and both are fresh. Dark brown biotite is ubiquitous and is generally accompanied by grass-green hornblende and epidote, which commonly mantles allanite.

Amphibolite bodies range from metre-thick sheets traceable for many tens of metres to swirly lenses and shreds only a few centimetres long. Some of the larger sheets can be followed around fold noses (Fig. 7) and some smaller lenses are actually fold hinges. Discordant contacts and margins finer grained than interiors, visible here and there, suggest that at least some of the amphibolite bodies are dykes. The amphibolite is composed of granoblastic, grass-green hornblende similar to that in the gneiss, interstitial brown biotite, abundant accessory Fe-Ti oxide, and minor sphene.

The northwest corner of Ian Calder Lake map area is underlain by straight gneiss of unit **A**s, which appear to be the highly strained equivalent of the granodiorite gneiss just described, characterized by a dense, highly linear aeromagnetic pattern (Geological Survey of Canada, 1977a). Despite the intense deformation (Fig. 8), the characteristic amphibolite sheets can be recognized and are even slightly discordant in a few places. Nevertheless, the lithological similarities between units **A**b and **A**s may be coincidental and the equivalency of the units is not proven.

In thin section, the less deformed rocks of unit **A**s are virtually identical to their counterparts in unit **A**b but are finer grained. In high-strain zones the mafic minerals, particularly biotite, are strongly chloritized and epidote is a major mineral, suggestive of significant alteration.

On the accompanying geological map of Ian Calder Lake area, the contact of units **A**s and **A**b is drawn at the break in the aeromagnetic pattern (Geological Survey of Canada, 1977a). In the *Metamorphic map of the Canadian Shield* (Fraser et al., 1978), this boundary is shown as marking the eastern edge of a large granulite terrane, which Heywood and Schau (1978) named the “Queen Maud Block” and which they considered was separated from a higher crustal level terrane, the “Committee Bay Block”, to the east by the

Figure 7. *An unusually dense concentration of folded amphibolite sheets (metamorphosed dykes) in grey gneiss of unit* **A**b *in northern Ian Calder Lake map area. Photograph by T. Frisch. GSC 204075-A*

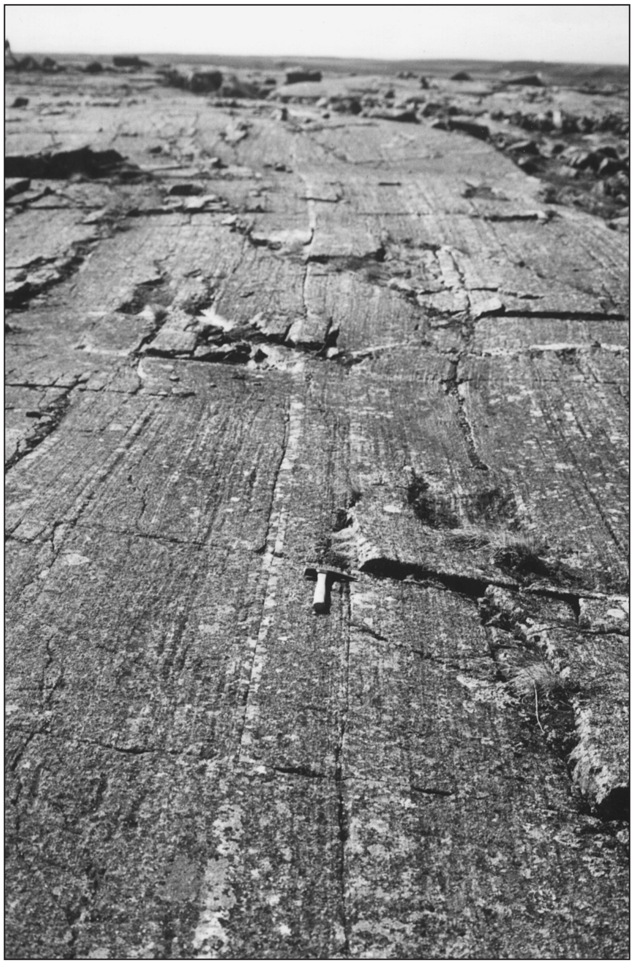


Figure 8. *Highly ductilely sheared straight gneiss of unit* **A**s *(?Queen Maud block) near the Chantrey fault zone in northwestern Ian Calder Lake map area. Photograph by T. Frisch. GSC 204074-Z*

Slave–Chantrey mylonite zone (Chantrey fault zone in Fig. 5). In the present study, no evidence of granulite-facies metamorphism was found in rocks of unit **A**s, with the possible exception of granoblastic-polygonal texture, a common feature of granulite-grade rocks. Perhaps the evidence of granulite-grade metamorphism has been nearly obliterated by retrogression, which might be expected at the margin of a highly deformed block. It is possible, however, that unit **A**s was never at granulite grade and merely represents unit **A**b deformed in the Slave–Chantrey mylonite zone. Isotopic data are too scanty to unequivocally differentiate between the two terranes. Granodioritic gneiss of unit **A**s (sample FS82-105) has yielded a Sm-Nd T_{DM} age of 2.79 Ga (E. Hegner, pers. comm., 1991), which is equal to or only slightly older than the other Nd model ages from the study area (2.79–2.68 Ga). The only other Sm-Nd data available from the Queen Maud block are six Nd model ages from its western side (west of 102°W), which range from 3.6–3.1 Ga (Thériault et al., 1994). In summary, the interpretation of the contact between units **A**s and **A**b in Ian Calder Lake map area as a terrane boundary, although plausible, has yet to be proven.

Farther east of the contact, near the northern border of the Franklin belt, grey gneiss of unit **A**b is interleaved with belts up to 6 km long, of white, gneissic to massive granodiorite with muscovite, biotite and, locally, fibrolite. Mineralogically similar but intensely deformed rocks occur within the supracrustal belt. The massive variety of white granodiorite is characterized by muscovite 'books' several centimetres across. Thin sections show well developed granoblastic-polygonal texture and felted masses of fine sillimanite needles associated with muscovite and chestnut-brown biotite. Sillimanite-bearing rocks are rich in well twinned oligoclase ($An_{27–30}$) but microcline becomes more abundant at the expense of plagioclase in sillimanite-free rocks. Gneissic rocks of similar mineralogy (presumably the deformed equivalents of the massive white granodiorite) display feldspar porphyroclasts in a mortared groundmass of feldspar, quartz, muscovite, and biotite; sillimanite was not seen.

The aluminous mineralogy of, and presence of two micas in, the white granodiorite suggest an S-type granitoid, formed by melting of metasedimentary supracrustal rock. Conceivably, however, the rock is metamorphosed regolithic material formed by alteration of the supracrustal rocks. A somewhat similar rock, garnet-muscovite granite (unit **A**mg), occurs at the northern margin of the Montresor belt (*see* below).

Garnet-muscovite granite (unit **A**mg*)*

This rock forms the basement to the metasedimentary rocks of the Montresor Group at the southern border of Ian Calder Lake map area. It is a massive, porphyritic, leucocratic granite with abundant muscovite, scattered small, pink garnet crystals, and patchy pegmatite.

As seen in thin section, the granite has a granitic texture modified by deformation. Most of the grains are 0.5–1 mm but aggregates of finer grained material and subhedral feldspar laths 3 mm long are common. The feldspar in both phenocrysts and matrix is sericitized oligoclase (An_{26}) and perthitic microcline. Many grains are cracked and show bent twin lamellae. Quartz is invariably strained and has undulose extinction. Muscovite forms laths up to 2 mm long intergrown with, and penetrating, the other major minerals; it lacks preferred orientation and clearly grew later than quartz and feldspar. Garnet forms poikiloblastic subhedral grains, 1–2 mm in diameter, rich in inclusions of quartz and feldspar. Many garnet crystals are imperfectly developed and this mineral, also, seems to be late in the paragenesis.

As noted above, the granite shows some similarity to the white granodiorite exposed near the Franklin belt in northern Ian Calder Lake map area. The main difference is the absence of biotite and higher microcline:plagioclase ratio in the granite. These differences aside, and given its aluminous character and proximity to a supracrustal belt, the granite may have the same origin as the two-mica granodiorite: intrusive rock derived from the melting of metasedimentary material or metamorphosed regolithic material.

Pink granite (unit **A**gr*)*

Gneissic to massive, largely homogeneous, pink granite outcrops on the eastern reaches of the Montresor River in Ian Calder Lake map area.

The rock is a potassic granite, coarse grained and inequigranular, with microcline phenocrysts up to 2 cm in length very locally developed. Much of the granite is moderately gneissic but in many places foliation, defined by biotite flakes, is very weak. Magnetite crystals up to 6 mm long are conspicuous in some outcrops and probably account for the distinct aeromagnetic expression of the granite (Geological Survey of Canada, 1977a). In thin section the granite is seen to consist largely of perthitic microcline, quartz, and subordinate oligoclase ($An_{21–23}$). Greenish-brown biotite and accessory magnetite make up a few per cent by volume of the rock. Subhedral to anhedral microcline has overgrown all other minerals and is by far the major mineral.

One kilometre east of the approximate contact with the augen gneiss unit **A**k, trains of augen gneiss and lenses parallel the foliation in the granite. Thus the granite postdates the augen gneiss but its age relative to the granulite belt on its northern border is unknown.

Granulite facies rocks and associated gneiss (units **A**o, **A**g*)*

Pyroxene-bearing granitoid rocks and associated garnetiferous gneiss form a belt, 10–15 km wide, that apparently extends eastwards from south of Franklin Lake and reappears in southeastern Cape Barclay map area (Fig. 5). Only those parts of the belt exposed in Ian Calder Lake and Cape Barclay map areas have been examined but, judging from the aeromagnetic map (Geological Survey of Canada, 1977d), the belt runs through Mistake River map area.

In Ian Calder Lake map area, the high-grade belt occupies an area about 11 x 11 km south of Franklin Lake and is designated as unit **A**o. The rocks of the belt are characterized by an east-west trend, interlayering on a scale of hundreds of metres of dark-weathering, relatively mafic, tonalitic gneiss and garnetiferous leucogneiss, and high strain.

Although foliation tends to run easterly, a number of northerly foliations and varying direction and degree of dip indicate a complex structure for unit **A**o in Ian Calder Lake map area. To the west, the gneisses are apparently truncated by a fault, inferred to extend from the Franklin belt. To the north, they appear to be in structurally conformable contact with units **A**k and **A**gd; to the south they are presumably intruded by the granite of unit **A**gr, but no contact was seen.

The dark-weathering gneiss (colour index 15–30) is medium grained, with an inequigranular-granoblastic texture commonly modified by crushing, which has produced fine-grained quartzofeldspathic material interstitial to coarser grains of plagioclase and quartz. Plagioclase ($An_{25–30}$) has blocky antiperthite texture and bent twin lamellae and may be accompanied by a little K-feldspar. The main mafic minerals are weakly pleochroic orthopyroxene and dark brown biotite, which together define the foliation. The orthopyroxene is associated with, and partly replaced by, olive-green hornblende. Some rocks contain, in addition, a little pale green clinopyroxene. Subordinate dark-weathering gneiss is of granodioritic composition and lacks pyroxene. It consists largely of oligoclase, microcline, grass-green hornblende, and minor chloritized biotite.

The leucogneiss weathers bright white, is fine to medium grained, and is made up mainly of antiperthitic oligoclase ($An_{26–28}$), perthitic untwinned K-feldspar, and abundant quartz, intergrown in an inequigranular-granoblastic texture. Pale pink garnet and red-brown biotite together make up less than ten volume per cent of typical leucogneiss.

In Cape Barclay map area, the apparent continuation of the high-grade belt forms a northeast-trending tract (unit **A**g) in the southeast. The northeasterly foliation within the belt conforms to the general structural trend in the map area. Exposure is only moderately good and the contacts of the belt must be interpolated between outcrops (some as much as a kilometre apart) of granulite- and amphibolite-facies rocks. In contrast to its extension south of Franklin Lake, the high-grade belt in Cape Barclay map area contains abundant meta-intrusive granitoid rock.

Strongly gneissic rocks include orthogneiss and paragneiss. Orthogneiss, which contains orthopyroxene, is typically brown weathering and medium grained with local development of porphyroblastic feldspar and garnet. Paragneiss tends to be leucocratic, even white weathering, and rich in garnet. In thin section, all the gneisses show granoblastic textures modified by deformation and a few display granoblastic-elongate textures with lenticles of xenoblastic quartz, so characteristic of high-grade metamorphic tectonites. Paragneiss is commonly fine grained with trains of tiny garnets defining a wavy foliation.

The pyroxene-bearing orthogneiss is tonalitic to granodioritic with colour index 10–20. Plagioclase invariably exceeds K-feldspar in abundance and discrete K-feldspar is absent altogether from some rocks. Plagioclase (oligoclase-andesine, with Ca content higher in the more mafic gneisses) exhibits bent and kinked twin lamellae and minor blocky antiperthite. K-feldspar in the granodioritic gneiss is finely perthitic without any visible microcline twinning. Orthopyroxene occurs in anhedral grains (commonly formed by the breakup of larger grains), weakly pleochroic and slightly altered. Biotite is reddish brown in tonalitic gneiss and dark brown in granodioritic gneiss. Granodioritic gneiss may also contain a little olive-green hornblende occurring as discrete grains bearing no particular relationship to the other mafic minerals. Although uncommon in the orthogneiss, garnet occurs sparsely in a few rocks in grains 2–3 mm across.

Paragneiss is found as layers, rarely more than a few metres thick, and lenses, some only a metre long, in orthogneiss. It is much less abundant than orthogneiss in Cape Barclay map area. It typically is fine grained and mylonitic and characterized by abundant garnet and absence of pyroxene. In thin section, inequigranular-elongate texture predominates, defined by lenticles of quartz and trains of granulated garnet crystals 1–4 mm in diameter. Plagioclase and perthite make up, with quartz, the matrix but locally also occur as sparse augen. In some paragneisses, garnet and red-brown biotite, generally associated spatially, are the only major mafic minerals but in others are joined by abundant cordierite intergrown with quartz and feldspar. Graphite is finely disseminated in some cordierite-bearing gneisses.

Cordierite-rich paragneiss from the southern border of Cape Barclay map area has porphyroclastic texture, with abundant rounded and deformed porphyroclasts of cordierite and hair perthite in a very fine-grained, granoblastic matrix of quartz, plagioclase (An_{29}), perthite, and cordierite. Some cordierite porphyroclasts are cored by anhedral garnet in what appears to be a replacement texture and most are fringed by fine sillimanite needles, which probably precipitated by some form of pressure solution. In this rock, cordierite is fresh but in another, from the central part of the high-grade belt, retrogression has resulted in heavy pinitization of cordierite and the replacement of sillimanite by muscovite.

The third major rock type in the high-grade belt of Cape Barclay map area is porphyritic orthopyroxene-bearing granite. It occurs throughout the belt about equal in abundance to orthogneiss and, as it intrudes, or contains inclusions of, both granulite-facies and amphibolite-facies gneiss, is among the younger rocks of the Archean granitoid complex.

Nonretrograded granite weathers brown and on fresh surfaces has the green colour and greasy lustre characteristic of granulite-facies granitoid rocks. Strongly retrograded granite weathers pink or grey and is grey on the fresh surface. Most outcrops are highly porphyritic (up to 50% phenocrysts), with tabular, pink, Carlsbad-twinned K-feldspar, 1–4 cm long, as the phenocryst phase. Phenocrysts are commonly aligned to varying degree, imparting a strong to weak gneissic structure to the rock. In strongly gneissic rocks, the matrix is foliated. Structure and texture commonly vary across an outcrop. For example, gneissic structure may be enhanced in a shear zone, the density of feldspar phenocrysts may differ in different parts of the outcrop, and inclusions may be restricted to a particular layer.

Inclusions are common in the porphyritic granite. They range from lenses several centimetres long to rafts on the metre scale (Fig. 9), composed of granulite-grade pyroxene gneiss and paragneiss and augen gneiss apparently derived from the lower grade granitoid complex bordering the high-grade belt.

Under the microscope, the K-feldspar phenocrysts are commonly seen to be cracked and partly recrystallized to a mosaic of small grains. The feldspar is very finely perthitic without microcline grid twinning and is relatively inclusion-free. The matrix is highly inequigranular granoblastic and extensively recrystallized. Very fine-grained quartzofeldspathic material borders feldspar phenocrysts and locally has invaded them in cracks. The large (1–3 mm) crystals of the matrix consist of perthite, plagioclase (about An_{25}) with bent twin lamellae, and subordinate garnet; plagioclase and garnet are very inhomogeneously distributed. Quartz occurs in fine-grained, recrystallized aggregates and poorly developed lenticles and is highly strained. The main mafic silicate is red-brown biotite, some of which has partly replaced garnet. In the six samples examined in thin section, all orthopyroxene has been replaced by fine-grained, greenish biotite-chlorite pseudomorphs typical of retrograded granulite-facies granitoid rocks. Apatite and zoned zircon are common accessory minerals.

Figure 9. *A lensoid inclusion of granulite-facies paragneiss in porphyritic orthopyroxene granite of unit* **A**g *in southeastern Cape Barclay map area. Photograph by T. Frisch. GSC 204635-S*

As already stated, no contact was observed between units **A**o or **A**g and other rocks. In southeastern Cape Barclay map area, a large exposure of the porphyritic granite on a major tributary of the Hayes River contains a septum (or large inclusion) of flattened mafic augen gneiss, indistinguishable from that outside the high-grade belt. Furthermore, smaller masses of amphibolite-grade augen gneiss occur as inclusions in the granite. Evidently the granite is younger than much of the surrounding granitoid complex. The granite is definitely younger than the bordering high-grade gneiss, which occurs as inclusions in it. Also, the granite locally becomes finer grained at contacts with gneiss, suggestive of chilling.

A sample (FS84-45A) of retrograded porphyritic garnet-orthopyroxene granite from an outcrop on the tributary of the Hayes River was dated by U-Pb zircon and Sm-Nd whole-rock methods (Frisch and Parrish, 1992). Three discordant zircon fractions were analyzed and gave an age of 2587 +9/-7 Ma, which is interpreted as the age of intrusion (Frisch and Parrish, 1992). The Sm-Nd T_{DM} age of the same sample, determined by E. Hegner (pers. comm., 1991), is 2.79 Ga, which may be the time at which the felsic source material of the granite was emplaced in the crust (Frisch and Parrish, 1992).

By virtue of their high metamorphic grade, the rocks of units **A**o and **A**g are clearly out of context with respect to the bordering amphibolite-facies terrane. No appropriate metamorphic gradient nor heat source is evident in the study area that would account for the presence of granulite-facies rocks. The granulite belt may be an upfaulted block with origins in Mistake River map area. Yet the evidence from inclusions in the orthopyroxene granite that normal amphibolite-grade rock underlies the granulite belt is puzzling in that the granite is surely genetically related to the granulite belt and does not postdate emplacement of the belt.

Amphibolite-facies granitoid complex, Cape Barclay and Darby Lake map areas (unit **A**b *in Map 1779A)*

The scale of geological mapping in Cape Barclay and Darby Lake map areas in 1984 did not permit meaningful subdivision of the granitoid complex beyond delineation of the granulite-facies belt in the Cape Barclay map area. Thus all Archean granitoid rocks outside the high-grade belt in this part of the study area are included in a single map unit, **A**b.

Most, if not all of the rocks in this unit find their counterparts in Ian Calder Lake map area. The most abundant type is a biotite-bearing granodioritic gneiss with augen of K-feldspar, generally microcline perthite. Typically, colour index ranges from 10–15, feldspar augen are 1–3 cm long, and mylonitic and cataclastic textures are prominent in many outcrops. In places the rock tends toward the massive and the large feldspars are tabular in outline. Leucocratic layers and veins are locally abundant and give rise to migmatite in places. An exposure in a northeasterly trending shear zone near the mouth of the Back River is depicted in Figure 10. Locally, the augen gneiss contains centimetre- to metre-sized inclusions of biotite gneiss, mafic hornblende-biotite gneiss and amphibolite, generally aligned parallel to the foliation of the host. At the head of Chantrey Inlet, 15 km southwest of

Figure 10. *Highly deformed, veined, biotite-rich granodiorite gneiss of unit* **A**b *in a ductile shear zone near the mouth of the Back River, southern Cape Barclay map area. Photograph by T. Frisch. GSC 204635-F*

Backhouse Point, garnetiferous metasedimentary inclusions appear to have partially melted to form garnet-muscovite granite pods and veins, which in places are sufficiently abundant to form migmatite. As in Ian Calder Lake map area, pink pegmatite commonly forms discordant sheets and segregations in the augen gneiss.

In thin section the matrix of typical augen gneiss is inequigranular granoblastic with numerous indications of strain such as undulose extinction in quartz, deformed twin lamellae in plagioclase ($An_{25–27}$), and fine-grained quartzofeldspathic crush zones. Biotite is dark brown and hornblende, rarely present, is grass-green or bluish green. Sphene in euhedral crystals and zircon included in biotite are abundant accessory minerals.

Other locally abundant gneissic rocks are felsic to mafic, plagioclase-rich granodioritic varieties in which plagioclase, both as augen or megacrysts and as smaller components of the matrix, predominates over K-feldspar. These rocks may contain hornblende and epidote as well as biotite.

Mylonitic mafic gneisses with small augen (possibly porphyroclasts) are prominent between the Hayes River and the high-grade belt. They have a colour index of 30–35 but also contain biotite much in excess of hornblende.

Weakly gneissic, grey (muscovite-)biotite granodiorite occurs locally in bodies up to a few kilometres long and several hundred metres wide. It appears to be particularly common north of the Chantrey belt in Cape Barclay map area. Near the northern border of the map area, muscovite-biotite granodiorite intrudes dark (?metavolcanic) schists of the Barclay supracrustal belt (*see* below). A large body of similar rock at the northern margin of the Chantrey belt to the south shows no intrusive relations with the Chantrey rocks.

On his reconnaissance map Heywood (1961) showed the granitoid terrane north of the Chantrey belt to be different in lithology from that to the south: the belt separates a northern region underlain mainly by massive granitoid rocks from mixed gneisses in the south. In the present work no such distinction could be drawn and there is no reason to suppose that the Chantrey belt marks the site of a contact between two different crustal blocks.

Hornblende-biotite augen gneiss (sample FS84-6) from the western shore of Chantrey Inlet has given a U-Pb zircon age of 2596 +13/-10 Ma (Frisch and Parrish, 1992). The same rock yielded a Sm-Nd T_{DM} model age of 2.68 Ga (determined by E. Hegner *in* Frisch and Parrish, 1992). Weakly gneissic muscovite-biotite granodiorite (sample FS84-88), apparently intrusive into the Barclay belt 39 km east of Cape Hay, has yielded an imprecise U-Pb zircon age of ca. 2590 Ma and a Nd model age of 2.70 Ga (Frisch and Parrish, 1992). Nearly massive granodiorite (sample FS84-A292) from a body near the northern margin of the Chantrey belt gave a Nd model age of 2.74 Ga (E. Hegner, pers. comm., 1991). The two U-Pb ages are interpreted as indicating the time of granitoid intrusion, presumably not long after emplacement of the source material in the crust at ca. 2.7 Ga (Frisch and Parrish, 1992).

Amphibolite (unit **A**a*)*

Black or dark green amphibolite bodies other than those forming thin layers and small lenses in gneiss are generally either metamorphosed dykes or larger, typically up to 40 x 30 m, irregularly shaped intrusions into granitoid gneiss. They occur throughout the study area but are nowhere abundant, although several bodies some hundreds of metres across are concentrated east of Franklin Lake. Ranging from fine- to medium-grained and massive to foliated, the bodies consist of grass-green hornblende, plagioclase, minor brown biotite, and trace amounts of quartz. Metamorphosed dykes show foliated margins and more massive interiors with metagabbroic texture and tend to be of limited extent and broken up by remobilized gneiss. The larger amphibolite bodies, commonly metagabbroic, have preserved apophyses into bordering gneiss, which may contain inclusions of an older generation of amphibolite. The amphibolite bodies are cut by pegmatite.

Whereas the abovementioned amphibolites undoubtedly represent metamorphosed basaltic intrusions of Archean and/or Aphebian age, amphibolites in two particular occurrences may have metavolcanic affinities. One occurrence is on the west slope of Mount Meadowbank overlooking the Back River in Ian Calder Lake map area. There, amphibolite is bordered by highly siliceous grey gneiss, which in turn is flanked by mafic biotite schist. If the grey gneiss and mafic rocks in this restricted exposure are felsic and mafic metavolcanic rocks, respectively, they may constitute a remnant of a greenstone belt, the only one so far recognized in Ian Calder Lake map area.

Possibly equivalent rocks form similarly restricted outcrops at Cape Hay, on the eastern shore of Chantrey Inlet. There also, amphibolite and possible felsic metavolcanic rock are associated. The amphibolite is black, fine grained, schistose, and folded and contains minor layers of pale grey, fine-grained gneiss. These rocks form isolated exposures near the shore and no further exposures were found inland, where drift cover is heavy. They may represent a relict northeast-trending greenstone belt presumably of Archean age.

ARCHEAN SUPRACRUSTAL ROCKS

Rocks associated with the Montresor Group (unit **A**m*)*

Metamorphosed supracrustal rocks assigned to the Archean structurally underlie the Aphebian Montresor Group in the southern part of Ian Calder Lake map area. They are most abundant at the northern end of the Montresor belt. The rocks comprise mainly knotted mica schist and impure quartzite and subordinate marble and calc-silicate rock. They are cut by abundant granitic-pegmatitic veins, sheets, and dykes.

Lithology

The mica schist is of two varieties, cordierite-andalusite-biotite-muscovite and (garnet-)muscovite-biotite. Feldspar and quartz are abundant in both varieties and tourmaline may be a common accessory mineral. Cordierite occurs as highly to completely altered, ragged porphyroblasts, bluish in hand specimen, and andalusite porphyroblasts typically exhibit helicitic inclusion trails discordant to the mica foliation, which wraps around the porphyroblasts. The biotite is nut-brown.

The (garnet-)muscovite-biotite schist consists mainly of brown biotite, muscovite, feldspar, and quartz and, if garnetiferous, carries only minor garnet in tiny crystals. A schist from the northern end of the Montresor belt contains fibrolite included in quartz and in coarse muscovite flakes.

The other major rock type of unit **A**m is quartzite, both orthoquartzite and impure quartzite. White, glassy, medium-grained orthoquartzite with scattered, 1 mm long actinolite prisms and intruded by numerous granite sheets outcrops on the Montresor River near the northern end of the Montresor belt. Phyllite is locally thinly interbedded with the orthoquartzite. Quartzite units in the supracrustal suite adjacent to the Montresor belt are mainly impure, grey or pink varieties. These rocks are subarkose in Pettijohn's (1975) classification, with up to 20 volume per cent microcline, muscovite, epidote, actinolite, and garnet. Grain size is fine to medium and quartz grains, although markedly sutured, are only locally strongly strained. Microcline, the most abundant constituent after quartz, is sericitized. Pale green actinolite, the major mafic mineral, occurs in euhedral prisms up to 2 mm long and, in one sample from the northern end of the Montresor belt, is intergrown with tiny euhedral crystals of colourless garnet. Epidote tends to be associated with actinolite. Like the other constituents, muscovite is randomly distributed without preferred orientation.

Impure quartzite at the northern end of the Montresor belt is locally interlayered with white dolomite and calc-silicate rock. Another minor member of the supracrustal suite, outcropping at the southern end of the belt, is a dark, laminated, fine-grained actinolite-plagioclase schist, originally possibly a volcaniclastic rock.

Contact relations

The Archean supracrustal rocks and the Montresor Group have not been found in contact. At the northern end of the Montresor belt, the supracrustal rocks structurally underlie Montresor Group rocks and the markedly discordant trends suggest an unconformity.

No depositional contact between the supracrustal rocks and the Archean granitoid complex was seen. Foliation in the supracrustal rocks is generally concordant with that in the adjacent granitoid rocks.

The supracrustal suite is intruded by, and locally interlayered with, granitic rock of two types, of unknown age. One type (?Archean) forms fine-grained, gneissic to mylonitic sheets concordant, and presumably in tectonic contact, with the supracrustal rocks. The other type makes up coarser grained to pegmatitic, muscovite-bearing granite in concordant sheets and crosscutting dykes. Along with lithology, it is the intimate association with granitic rock that distinguishes this supracrustal suite from the Montresor Group.

Age and correlation

The stronger deformation, higher metamorphic grade, and breakup by granitic rock indicate that the supracrustal suite predates the rocks of the spatially associated and structurally overlying Montresor Group, which is considered to be Aphebian but the age of which has not been determined.

The supracrustal rocks may be correlative with metasedimentary and metavolcanic schist and quartzite known or suspected to be Archean in regions to the south of the study area. In Amer Lake map area, immediately south of Ian Calder Lake map area, granite apparently intrusive into orthoquartzite, which has intercalations of schist and phyllite, has yielded 2.6 Ga U-Pb zircon ages and felsic metavolcanic rock, associated with chlorite schist, metagreywacke, and other supracrustal rocks, has given a 2.8 Ga U-Pb zircon age (Tella, 1994).

Franklin belt (units A*c,* A*q,* A*d,* A*a)*

A narrow belt of highly deformed supracrustal rocks, herein named the Franklin belt, occurs in northeastern Ian Calder Lake map area. These rocks outcrop on both sides of Franklin Lake and can be traced southwest as far as 25 km from the lake. The belt lies at the southern end of the northeast-trending Murchison fault (Heywood and Schau, 1978). To the northeast, the Murchison fault follows the Chantrey belt for most of the length of the belt. It was originally thought (Frisch and Patterson, 1983) that the rocks of the Franklin belt belonged to the Chantrey Group but subsequent fieldwork cast doubt on this correlation (*see* below).

Lithology

At the southern margin of the Franklin belt, the rock in contact with the Archean granitoid complex is generally a dark green, fine-grained muscovite-biotite schist with rare clasts of white quartz and granite. The clasts are up to several centimetres long and augen shaped and the fine grain size of the matrix is due to mylonitization. Quartz clasts consist of a mosaic of sutured grains and may be rimmed by coarse (up to 4 mm) muscovite blades, which clearly grew after shearing ceased. The marginal schist is up to 20 m wide but generally much less.

Structurally overlying the schist is stretched quartz-pebble metaconglomerate (unit **A**c), the most abundant supracrustal rock in the belt. The metaconglomerate typically consists of abundant white quartz pebbles, 10–15 cm long, in a green matrix of chlorite-muscovite-biotite schist (Fig. 11). The pebbles are stretched parallel to the regional northeasterly lineation and flattened in the plane of foliation of the matrix but retain an overall cigar shape. In areas of exceptionally high strain, such as the western end of the belt, pebbles attain length:width ratios of 12:1. Locally, the metaconglomerate is polymictic, containing pebbles of granitic and mafic rocks of various shapes and sizes. Such rocks are virtually identical to metaconglomerate at the southern margin of the Chantrey belt in Chantrey Inlet (*see* below). Quartz veining is commonly prominent and locally (for example east of Franklin Lake) intense.

***Figure 11.** Cigar-shaped quartzose clasts in highly deformed metaconglomerate (unit* A*c) of the Franklin belt east of Franklin Lake, northeastern Ian Calder Lake map area. Photograph by T. Frisch. GSC 204075-G*

Fifteen kilometres southwest of Franklin Lake, where, due to severe deformation, the belt begins to peter out, supracrustal outliers, which consist of metaconglomerate, are most easily recognized; these occur a further 10 km southwest.

West of Franklin Lake, highly strained to mylonitic granitoid gneiss (unit **A**b) separates metaconglomerate from the rest of the supracrustal sequence to the north, which comprises orthoquartzite, marble and calc-silicate rock, and amphibolite.

The orthoquartzite (unit **A**q) is typically white, massive, medium grained, and fuchsitic. Although highly recrystallized, strongly sutured grain boundaries being visible in thin section, scattered larger quartz grains of primary aspect are common in outcrop. Dark muscovite-biotite schist, sporadically calcareous and containing scattered quartz clasts, occurs interlayered with the orthoquartzite but is everywhere subordinate to it. Apart from being more deformed and fuchsitic, these two rock types resemble those in unit **A**m below the Montresor Group. Fuchsitic quartzite in the lower Chantrey Group is generally more feldspathic.

Farther north, southeast-dipping orthoquartzite is structurally underlain by interlayered amphibolite (unit **A**a) and marble (unit **A**d); contacts are sharp and concordant. The amphibolite is black or dark green on weathered and fresh surfaces, medium-grained, and foliated. It consists of pale olive-green, probably magnesian hornblende, slightly altered to orangey-brown phlogopitic biotite, and calcic plagioclase. Its mineralogy distinguishes the amphibolite from that forming bodies in gneiss, including metamorphosed dykes, which typically contain grass-green hornblende, chestnut-brown biotite, and sphene. Also, the amphibolite does not appear to extend into the surrounding gneiss. Thus it seems likely that the amphibolite is an integral part of the Franklin belt and distinguishes the Franklin belt from the Chantrey belt, the latter being devoid of mafic rock.

Marble and calc-silicate rock (unit **A**d), which adjoin both amphibolite and the granitoid complex, outcrop poorly along the northern margin of the Franklin belt. These are grey, white or pale yellow, well layered and disharmonically folded, medium-grained rocks with abundant tremolite and quartz. Amphibolite boudins occur locally and appear to be remnants of layers and lenses that were disrupted during deformation.

Contact relations

The contact between the Franklin belt and granitoid complex is rarely exposed and is invariably a high-strain zone. The best exposure known of the contact is at the eastern end of the belt, 4 km inland from Franklin Lake. The contact there is interdigitating: dark muscovite-biotite schist with sparse augen-shaped clasts of white quartz and granite and a few carbonate lenses alternates with mylonitic, grey biotite-muscovite granodiorite gneiss (unit **A**gd). Fine grain size and abundant quartz lend the gneiss the appearance of a feldspathic quartzite. The gneiss is actually a blastomylonite, with much, if not all, of the quartz having recrystallized and relatively coarse (up to 0.3 mm) muscovite having grown in a largely randomly oriented fashion. There seems little doubt that the interleaving of gneiss and schist, on the scale of less than a metre, is entirely tectonic. Indeed the contact between Franklin belt and granitoid complex, in most places at least, is probably faulted. To what degree the Franklin belt is allochthonous is unknown.

East of Franklin Lake, the supracrustal rock–gneiss contact is structurally conformable (dipping inward to the belt) whereas west of the lake, the (synformal) dips of the supracrustal rocks are opposite to those of the gneisses.

Dykes and other discordant bodies of amphibolite are seen locally in the Franklin belt. An instructive example occurs associated with metaconglomerate of a supracrustal outlier 20 km along strike of the belt from the west shore of Franklin Lake. A 1 m wide amphibolite dyke in metaconglomerate trends approximately north perpendicular to the pebble elongation direction, which is nearly east in this outcrop, probably as a result of rotation of the outlier from the normal northeasterly trend of the Franklin belt. The dyke is sheared along planes parallel to pebble elongation. It may be an offshoot of a nearby layered amphibolite body, lying between metaconglomerate and gneiss, that is discordant to the metaconglomerate but has foliation striking east concordant with foliation in the gneiss and pebble elongation in the metaconglomerate. In that it consists largely of a pale green amphibole, the amphibolite in these exposures resembles more the amphibolite of unit **A**a of the Franklin belt than the typical amphibolite in the granitoid complex, which has grass-green amphibole. Whatever its origin, the amphibolite is younger than the metaconglomerate but predates shear deformation in the Franklin belt.

Little-deformed quartz veins and irregular pegmatitic quartz-feldspar segregations were found in metaconglomerate east of Franklin Lake but are not abundant and were not seen elsewhere in supracrustal rocks of the Franklin belt.

Age and correlation

Because of the absence of intrusive relations, the Franklin belt appears to postdate emplacement of granitoid rocks (except pegmatite). The belt is, however, cut by sheared amphibolite dykes. Such dykes have not been observed in the Chantrey and Montresor belts. Furthermore, metabasaltic rock, such as the amphibolite of the Franklin belt, does not occur in either the Chantrey or the Montresor belt but is found in the Barclay belt, which appears to be Archean (*see* below). Thus, although the Franklin belt is essentially on strike with the Chantrey belt, not far distant to the northeast, it is considered more likely to be correlative with the Barclay belt and therefore Archean.

Barclay belt (units Ap, Ad*)*

The Barclay belt consists of a sequence of metamorphosed supracrustal rocks that extends northeastwards from 10 km south of Cape Barclay on the eastern shore of Chantrey Inlet. The belt, 2.5–4 km wide, is well exposed at the coast and, although exposure progressively deteriorates, can be traced inland discontinuously for 30 km. Farther northeast along strike, what appear to be outliers of the belt outcrop near the northern border of Cape Barclay map area. Much of the belt is presumably hidden beneath the extensive drift cover in this region.

The belt consists mainly of folded, steeply dipping pelitic and psammitic schist units, (?)volcaniclastic rocks, marble and calc-silicate rocks, and subordinate volcanic and mafic intrusive rocks, all metamorphosed to amphibolite-facies grade.

Lithology

Metapelitic and metapsammitic rocks (unit **A**p*)*

At the coast, the Barclay belt largely comprises pelitic and psammitic schistose rocks interlayered on a scale of tens of metres to several hundreds of metres.

Pelitic schist is particularly abundant in the northern half of the belt at the coast. It is commonly dark grey due to disseminated graphite and/or abundant biotite; schist with less or no graphite is light grey or silvery. The schist is invariably porphyroblastic, the main porphyroblast minerals being biotite, andalusite, cordierite, garnet, and staurolite. Generally only two or three of these minerals coexist as porphyroblasts but some rocks contain all five, in crystals 3–4 mm and locally 1–2 cm in size. Commonly, staurolite (Appendix, Table 5) has been replaced by andalusite (Fig. 12) and garnet is unstable. Garnets are almandine with 5–10% of the pyrope molecule (Appendix, Table 1) and grains are slightly zoned, Mn decreasing and Fe increasing from core to rim. Andalusite and cordierite, typically in elongate, spongy crystals, are invariably syn- to postkinematic and commonly have

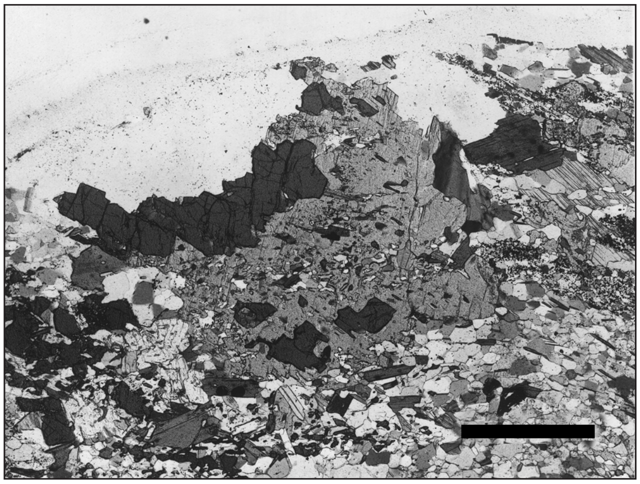


Figure 12. *Relict staurolite (in partial extinction) enclosed in an andalusite porphyroblast in pelitic schist of unit* **A**p *of the Barclay belt, east shore of Chantrey Inlet, Cape Barclay map area. The replacement of staurolite by andalusite was a common reaction in both the Barclay and Chantrey supracrustal belts. Bar scale is 1 mm. GSC 1995-237D*

overgrown the foliation. Andalusite is characteristically of the chiastolite variety, graphitic material forming a cruciform pattern in crystal cross-sections (Frisch et al., 1985, Fig. 33.6). Cordierite (X_{Fe}=0.44; Appendix, Table 6) is heavily to completely altered to pinite or sericite in almost all rocks but was clearly a common phase throughout the belt. It forms overgrowths on late porphyroblastic biotite in some rocks; in others it appears to have partly replaced andalusite. Chestnut-brown or reddish-brown biotite (X_{Fe} ~0.6; Appendix, Table 2) tends to occur both as a constituent of the matrix and as porphyroblasts, which typically are randomly oriented and rich in inclusions with pleochroic haloes. Chloritization of porphyroblastic biotite is locally heavy but generally erratically developed. Chlorite after biotite in a garnet-free andalusite schist is ripidolite with X_{Fe} almost identical to that of the biotite (Appendix, Table 4). Muscovite occurs in many rocks, both as part of the matrix and as later, coarser blades. Muscovite in a garnet-free andalusite-biotite schist is Mg-free and low in the paragonite molecule (Appendix, Table 3). Sillimanite is fairly widespread in the pelitic schists, typically as fibrolite rimming or replacing biotite. Rarely, sillimanite occurs in discrete needles cutting across andalusite. The remainder of the matrix of the pelitic schist consists of plagioclase (oligoclase) and quartz.

Very dark grey to black, highly garnetiferous biotite schist forms a distinctive unit in the pelitic schist and underlies the highest land in the Barclay belt at the coast. These are well layered, even finely laminated rocks with layers marked by differences in garnet content (Fig. 13). The rocks consist of subhedral, pink garnet porphyroblasts, 1 mm to 3 cm (average 2–8 mm) across, with helicitic inclusion trails in cores, in a well foliated, mafic matrix of greenish-brown biotite, andesine ($An_{22–24}$), quartz, and rare tourmaline. The coarsest layers are those richest in garnet (indeed, some layers are practically garnetites) and probably represent the fine-grained, shaly upper parts of size-graded deposits; recrystallization during metamorphism resulted in a grading inversion.

Minor calcareous interbeds and lenses occur in the dark garnetiferous schist. They weather rusty, are coarse grained, and contain helicitic, pink garnet, up to 1 cm across, colourless tremolite in blades 1–2 mm long, reddish-brown biotite, green chlorite, and a little carbonate.

At the southern border of the main belt of dark garnetiferous schist, interbeds of light-weathering psammitic schist become increasingly abundant southward (Fig. 14). Typically, the psammitic schist is composed of muscovite, biotite, quartz, and subordinate plagioclase. Cordierite occurs locally in anhedral poikiloblasts up to 3 mm across. A pale grey, fine-grained psammitic schist from near the northern boundary of Cape Barclay map area consists of greenish-brown biotite (relatively Ti-poor and Mg-rich; Appendix, Table 2, sample FS84-39C), quartz, minor plagioclase (An_{48}), and a little skeletal Mg- and Mn-bearing almandine garnet (Appendix, Table 1).

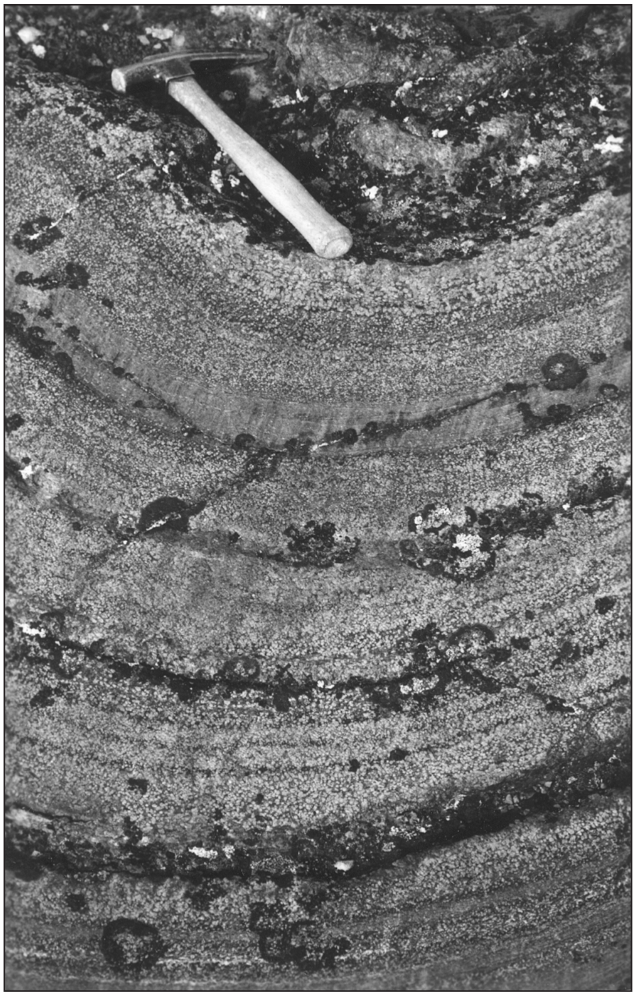


Figure 13. *Inverse size grading in a metamorphosed psammitic-pelitic schist (unit* **A**p*) of the Barclay belt, east shore of Chantrey Inlet: coarse-grained (garnet-rich) layers probably represent originally fine-grained argillaceous material, whereas the now fine-grained layers were once coarser sandy material. Photograph by T. Frisch. GSC 204073-E*

Figure 14. *Dark, garnet-biotite schist interlayered with light, psammitic muscovite-biotite schist in the Barclay belt, east shore of Chantrey Inlet. Photograph by T. Frisch. GSC 204073-O*

Figure 15. Psammitic (light) and pelitic (dark) schists interpreted to comprise a metamorphosed turbidite succession in the Barclay belt, east shore of Chantrey Inlet. Stratigraphic tops are presumed to be towards the viewer. Photograph by T. Frisch. GSC 204073-K

Figure 16. Rounded to elliptical masses of metapsammitic rock in psammitic schist of a putative metaturbidite succession in the Barclay belt probably represent concretionary structures. The large, dark porphyroblasts on the left are andalusite pseudomorphs of staurolite. East shore of Chantrey Inlet. Photograph by T. Frisch. GSC 204635-W

The interlayering of metapelite and metapsammite is locally sufficiently regular as to suggest a turbidite sequence as protolith (Fig. 15). Sharp contacts between psammitic and pelitic units are taken to be bedding surfaces; gradational contacts presumably mark the transition from sand to mud within a graded bed, which in these outcrops is generally 1–1.5 m thick.

Also noted in outcrops near the southern margin of the belt was what appear to be concretionary structures. Isolated, rounded to elliptical masses, 20–40 cm across, of light-coloured psammitic rock lie in a semipelitic matrix (Fig. 16).

Metamorphosed volcanic, volcanogenic, and mafic rocks (unit **A**p*)*

Inland about 7 km from the coast, schistose mafic to intermediate rocks occur in isolated outcrops at the northern boundary of the Barclay belt. These rocks are medium to dark grey, well layered on a centimetre scale, fine grained, and locally pyritic. Their contact with rocks to the south is not exposed.

The darker schist consists largely of amphibole, biotite, and plagioclase. One pyritic variety contains euhedral needles of very pale green amphibole (possibly cummingtonite) and grass-green amphibole (Ca-hornblende), brown biotite, minor pale pink garnet in euhedral poikiloblasts (up to 2 mm in diameter, confined to Ca-hornblendic layers), and granular, untwinned plagioclase. Some of the pale amphibole prisms are zoned to green calcic hornblende at the edges.

Closely associated with the two-amphibole schist is a feather amphibolite, in which a wavy foliation defined by tiny, dark brown biotite flakes is overgrown by poikiloblastic green amphibole forming spongy, ragged blades, about 1 mm long, singly and aggregated. The mafic minerals lie in a matrix of finely granular plagioclase.

Another associated rock type has subhedral plagioclase crystals (0.7–1 mm), in part recrystallized to a mosaic substructure, and scattered, ragged, poikiloblastic garnets in a very finely granular matrix of plagioclase, quartz, and brown biotite. The larger plagioclase grains are thought to be relict phenocrysts.

All the above rocks — rich in amphibole, poor in quartz — are interpreted to be ultimately volcanogenic. The amphibole-bearing varieties may be metamorphosed sediments derived from the weathering of mafic volcanic rocks but the protolith of the rock with the relict (?)phenocrysts was probably a volcanic rock of intermediate composition.

Mafic rocks outcrop at the southern border of the Barclay belt 20–25 km inland. These include black-weathering, medium-grained amphibolite and metagabbro, both of which are commonly rusty and locally contain irregular, pyritiferous gossan zones. Some of the gossans are developed on meta-ironstone, which is consistently pyrite-rich and locally magnetite-rich and appears to form narrow lensoid masses exposed over a distance of at least a few tens of metres.

A sample (FS84-38C) of meta-ironstone studied in thin section consists of intergrown colourless cummingtonite with Fe/Fe+Mg=0.67 and bluish-green, ferro-tschermakitic hornblende with Fe/Fe+Mg=0.77 (Appendix, Table 7); spongy almandine garnet with 12% of the grossular molecule (Appendix, Table 1); quartz; pyrite; and magnetite. The iron-silicate minerals are closely intergrown but Ca-rich amphibole tends to have overgrown the Ca-poor variety and both are enclosed by garnet. Magnetite is fine grained and disseminated throughout the rock. The proportions of magnetite, pyrite, and quartz in the ironstone vary greatly from outcrop to outcrop.

Metacarbonate rocks (unit **A**d*)*

Marble and calc-silicate rock outcrop over a strike length of several kilometres near the northern border of the Barclay belt 20 km from the coast. Grey impure marble, with tremolite and quartz, is interlayered with tremolite-bearing quartzite; the layers are about 1 m thick and dip steeply northwest. The carbonate and quartzitic rocks are locally intensely contorted, showing disharmonic folding and flowage of marble around fragments of disrupted quartzitic layers. This deformation predates pegmatite intrusion.

The marble-quartzite sequence is bordered on the south by dark grey and dark green metasiltstone with variable argillaceous content and scattered epidote-rich lenses; the northern contact is with Quaternary deposits.

Contact relations

The contact of the Barclay belt with the Archean granitoid complex is rarely exposed. Judging from the mylonitization evident in rocks on both sides of the contact, the contact is invariably one of high strain. In all but one of the few localities where it is exposed, the contact is sharp and structurally conformable, hence almost certainly tectonic. Sense of movement, however, is unknown. Near the southern margin of the belt at the coast, a slice of gneiss with an outcrop width of about 100 m is concordantly bordered on both sides by steeply dipping supracrustal rocks, suggestive of thrusting.

The only nontectonic contact found is one near the northern boundary of the map area, 39 km east of Cape Hay. At this locality, lenses of schist are enclosed in weakly gneissic, granoblastic two-mica granodiorite (U-Pb zircon age 2.6 Ga, *see* below). The schist of the lenses is identical to that forming a large outcrop nearby, which is on strike with similar outcrops to the northeast and southwest. All these rocks are on strike with undisputed Barclay belt exposures to the southwest, on the other side of a wide area of drift. Although the schist lenses are not diagnostic of Barclay belt rocks, they closely resemble some of them and so are considered to belong to the belt. The texture of the granodiorite — euhedral plagioclase crystals in a finer grained quartzofeldspathic matrix — indicates a magmatic origin. Thus the granodiorite is thought to have intruded the Barclay belt.

Pegmatite sheets form largely concordant intrusions that are locally concentrated but overall sparsely developed in the Barclay belt supracrustal rocks. They have been described by Černý et al. (1988), who found that most contain beryl in addition to the characteristic cleavelandite, quartz, and muscovite; rarely present are columbite and triphylite. The age of the pegmatite bodies is not known but their undeformed nature suggests they are post-Archean and correlative with the Aphebian pegmatite of the Chantrey belt.

Origin and environment of deposition

The association in the Barclay belt of mafic rocks, intermediate to acid volcanic rocks, psammitic and pelitic (including turbiditic) metasedimentary rocks, and minor meta-ironstone suggests affinity with the classical greenstone belt. There is even a suggestion of characteristic greenstone belt stratigraphy: from east to west, mafic igneous rocks at the margin of the belt followed (?overlain) by more felsic volcanic rocks and various sedimentary rocks. However, the analogy with greenstone belts should not be taken too far. Mafic rocks are not abundant and no pillow basalts have been recognized. Moreover, the stratigraphic sequence in the Barclay belt is unknown. On the basis of lithology, the Barclay belt could be considered an Archean variant of Condie's (1982) Proterozoic quartzite-pelite-carbonate association.

Age and correlation

The Barclay belt has not been dated directly. As described above, what appears to be the northern end of the belt in the Cape Barclay map area is intruded by weakly gneissic, magmatic-textured granodiorite. Zircon from the granodiorite has given a U-Pb age of 2596–2590 Ma, which is taken to approximate the time of intrusion (Frisch and Parrish, 1992). If the identification of the intruded rocks as belonging to the Barclay belt is correct, the belt must be Archean. Support for an Archean age is lent by the severe deformation of the belt and by the fact that the belt predates gneissic granitoid rock, all of which appears to be Archean in this part of Nunavut.

Differences in lithology, degree of deformation, and age relative to granitoid rocks militate against correlation of the Barclay and Chantrey belts. The Chantrey belt is generally less deformed and disrupted, lacks metavolcanic and mafic rocks, and postdates granitoid rock (other than pegmatite).

The Barclay belt lies approximately on strike with the Chantrey fault zone west of Chantrey Inlet (Fig. 5); no outliers of the belt are known on that side of the inlet. How far the belt extends north of Cape Barclay map area is unknown. It is likely that the Barclay belt correlates with the Franklin belt in northeastern Ian Calder Lake map area. There too, granitoid rocks are intimately associated with supracrustal rocks and metamorphosed mafic rocks occur within the belt.

PROTEROZOIC SUPRACRUSTAL ROCKS

*Montresor Group (units A*Mo*, A*Mt*, A*Ma*)*

Montresor Group is the name given to metamorphosed sedimentary rocks that form a belt trending northeast across the southern part of Ian Calder Lake map area. The belt was discovered during the Geological Survey of Canada's helicopter reconnaissance, Operation Back River (Heywood, 1961). It was subsequently named the "Back River Belt" by Bell

(1970, Fig. 1) but Frisch and Patterson (1983) proposed this name be replaced by Montresor belt to avoid confusion with the Back River greenstone belt in the Slave structural province.

The Montresor belt is a synclinal structure, 8 km wide, extending 73 km from just south of the Back River in northeastern Amer Lake map area (Tella, 1994) almost to the Montresor River in southeastern Ian Calder Lake map area. Dips of bedding are commonly 10–20°, steepening to 60° and more at margins of the belt. A poorly exposed outlier of the Montresor Group, oval in outline (longest dimension 11 km), occurs in southwestern Ian Calder Lake map area.

The Montresor Group consists mainly of arenite and subordinate carbonate rock metamorphosed to the lower amphibolite facies; pelitic schist and phyllite are common near the base of the section. A schematic stratigraphic column is presented in Figure 17. No sections were measured but maximum thickness of the Montresor Group is estimated to be 1800 m.

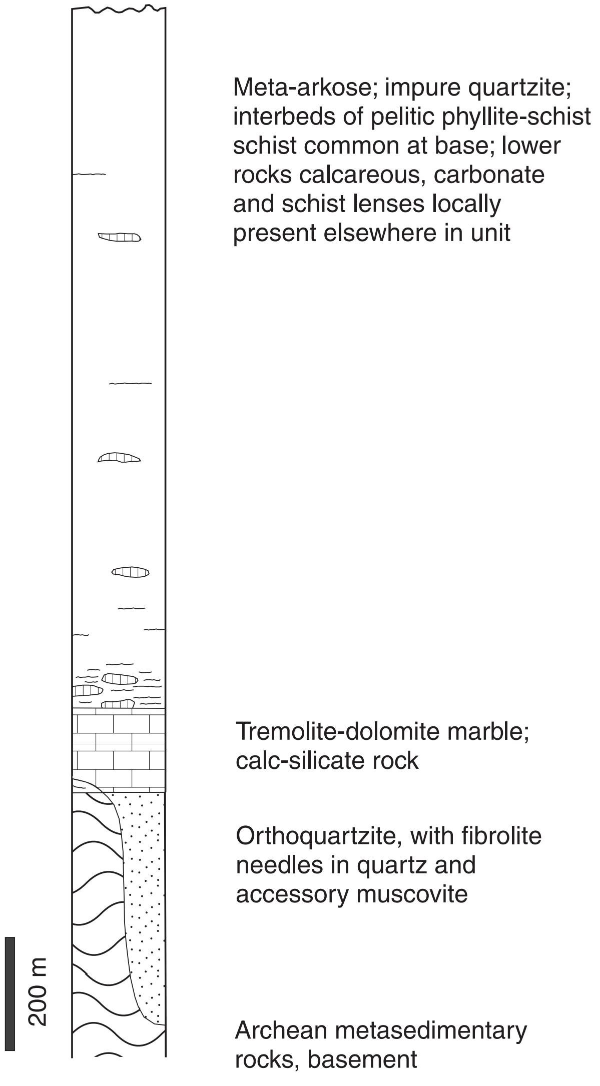


Figure 17. *Generalized stratigraphic section of the Montresor Group. Thicknesses shown are approximate.*

Lithology

*Orthoquartzite (unit A*Mo*)*

At the eastern end of the Montresor belt, the basal Montresor rock is a bright white orthoquartzite forming resistant ridges. Nowhere was it seen in contact with basement. Orthoquartzite was not found in the Montresor Group west of 96°50′ W.

The orthoquartzite is a glassy, medium-grained, strongly recrystallized rock. Although macroscopically slightly micaceous locally, it is characteristically of uniformly pure appearance. Lack of lithological variability makes bedding difficult to discern but, where visible, bedding is thick (1–1.5 m). Thin sections show highly strained quartz grains with sutured borders, accessory amounts of tiny muscovite flakes, and the nearly ubiquitous presence of fibrolite needles in quartz. The muscovite flakes define a distinct foliation, which is also marked in highly strained rocks by aligned quartz grains. A few samples have minor, reddish, altered plagioclase and fresher microcline generally accompanied by muscovite in flakes 2 mm long. These more schistose rocks have little or no fibrolite.

*Marble and calc-silicate rock (unit A*Mt*)*

Except at its eastern end, the border rock of the Montresor Group is commonly marble with abundant quartzose layers. In the east, marble is a fairly persistent unit between orthoquartzite (unit *A*Mo) and meta-arkose (unit *A*Ma). The contact between basal marble and gneiss is almost nowhere preserved (*see* below).

Marble is easily eroded and outcrops poorly in the area (Fig. 18). It may also have been thinned and squeezed out during deformation in the Montresor Group–granitoid complex contact zone. Thus marble outcrops sporadically and commonly exhibits intricate folding and boudined quartz layers. It is grey, white, or buff and locally weathers orange; fine grained; and thinly layered or contains layers and lenses of

Figure 18. *Typical exposure of siliceous marble (unit A*Mt*) at the northern margin of the Montresor belt, west of Mount Meadowbank, southern Ian Calder Lake map area. Photograph by T. Frisch. GSC 204075-L*

quartz or siliceous calc-silicate rock. Crossbedding is locally evident, indicative of a carbonate sand protolith. The carbonate is dolomite, locally limey, and commonly characterized by coarse tremolite in single blades up to 5 cm long or in sprays of crystals. Ophicalcite occurs rarely, associated with tremolite-bearing calc-silicate rock consisting of intergrown carbonate and quartzofeldspathic material.

In thin section, marble typically is seen to consist of unoriented, colourless tremolite needles in an equigranular dolomite matrix with a grain size of 0.5–1 mm. A few samples contain pale yellow phlogopite intergrown with tremolite. Olivine in ophicalcite is invariably heavily serpentinized. Diopside has not been found in the Montresor Group.

*Meta-arkose and metapelite (unit A*Ma*)*

The bulk of the Montresor Group consists of grey or pink meta-arkose, which overlies the marble unit. The two units are commonly separated by up to several tens of metres of dark phyllitic to schistose rock, which grades into overlying and underlying rock. Where marble is absent, meta-arkose is separated from basement by phyllite-schist.

The dark phyllite-schist below the meta-arkose may be interlayered with carbonate or calc-silicate rock or is itself calcareous. Typically, it is black or dark green, fine grained and laminated, consisting of brown biotite flakes, tremolite blades up to 2 mm long, and granular epidote in a quartz-rich quartzofeldspathic matrix with a grain size of 0.1–0.3 mm and a relict clastic texture. Most samples show good alignment in the foliation plane of all the main minerals, which clearly grew or recrystallized during metamorphism. Interstitial carbonate is abundant in the matrix of some samples and pyrite is a significant accessory in rusty varieties. The phyllite-schist is interpreted to be calcareous, argillaceous metasedimentary rock.

Figure 19. *Dark biotite phyllite filling depressions in the Archean granitoid complex adjacent to the northern margin of the Montresor belt, southern Ian Calder Lake map area. The phyllite is identical to typical schistose rock at the base of the Montresor Group. Rucksack (arrowed) for scale. Photograph by T. Frisch. GSC 204075-J*

Metapelitic rock resting directly on granitoid gneiss, where the carbonate unit is absent (Fig. 19), is a dark brown-weathering, well foliated phyllite consisting of brown biotite flakes (0.2 mm) in parallel alignment in a matrix of feldspar and quartz (0.1 mm).

Well foliated metapelite also forms layers, 10–20 cm thick, in meta-arkose, particularly in the lower part of unit *A*Ma (Fig. 20). Biotite is invariably present as poorly developed, brown or greenish-brown porphyroblasts (0.3–0.5 mm across), which may be accompanied by rounded grains (?porphyroblasts, 0.3 mm across) of plagioclase and microcline in a fine-grained, pyritic sericite(-biotite) matrix.

The meta-arkose is grey or pink, fine grained, and thick bedded (0.7–1.2 m). Trough crossbedding (Fig. 21) and straight-crested ripple marks, generally asymmetric, are common. Convolute bedding was noted in scattered outcrops and lenticular bedding, expressed as thin pelitic lenses, is present locally in the lower meta-arkose beds. Elsewhere in the lower part of the unit, carbonate or calcareous arkose lenses are common locally; they are rare in higher beds.

Figure 20. *Meta-arkose (unit A*Ma*) with thin interbeds of biotite schist-phyllite in the lower part of the Montresor Group at the northern margin of the belt, northwest of Mount Meadowbank. Photograph by T. Frisch. GSC 204074-W*

Figure 21. *Crossbedding in meta-arkose of the lower Montresor Group. Note the argillaceous layers. Photograph by T. Frisch. GSC 204075-H*

Typical meta-arkose is seen in thin section to be inequigranular and of obviously clastic texture. Angular grains of quartz and feldspar (microcline and plagioclase) lie in a very fine grained matrix of sericite, feldspar, quartz, and, locally, minor green-brown biotite. The largest grains are generally 0.2–0.3 mm across and rarely exceed 1 mm. Feldspar:quartz ratios range approximately from 50:50 to 40:60. Mica, especially biotite, and carbonate, as an interstitial component of the matrix, are more abundant in the lower beds, which also are slightly more recrystallized than the upper ones. Laminated meta-arkose consists of alternating laminae of mica-poor and mica-rich rock, which may differ in grain size.

Conglomerate occurs in the meta-arkose unit near the western margin of the Montresor belt about 2 km north of the southern border of the Ian Calder Lake map area. The conglomerate forms a lens a few hundred metres long and up to 20 m thick. It consists of well rounded clasts, up to 20 cm across, of white quartz and pink granite in a matrix of grey calcareous arkose. The matrix is inequigranular with clasts, 1–5 mm across, of quartz, feldspar (chiefly microcline), and aggregates of sutured quartz grains in a feldspathic and sericitic groundmass; carbonate is patchily distributed in the matrix. Rocks stratigraphically above and below the conglomerate are calcareous arkose. Conglomerate has not been found elsewhere in the meta-arkose unit.

Contact relations

The contact between Montresor Group and the granitoid complex was seen at only two places. At the southern margin of the Montresor belt near the Back River west of Mount Meadowbank, carbonate fills a 10 cm wide crack in gneiss. The carbonate may be an original deposit in fractured basement or may represent redeposited material derived from marble during later erosion and alteration. The other Montresor Group–basement contact is at the western margin of the belt 4–5 km due north of the southern border of Ian Calder Lake map area. There, massive garnet-muscovite granite (unit **A**mg) is overlain by southeasterly dipping, rusty brown-weathering, dark biotite phyllite containing rounded clasts, up to 5 cm long, of white quartz and granitic rock and crosscut by quartz veins. The phyllite is identical to that filling depressions in the granitoid rocks nearby (Fig. 19). Overlying the conglomeratic phyllite is grey meta-arkose with abundant, long pelitic lenses and layers, followed by the typical pink and grey rocks of the meta-arkose unit. These lower beds dip 45° to the southeast.

The lack of strong deformation and the presence of well preserved basal conglomerate suggest a depositional contact of the Montresor Group at this locality. This is, however, an exceptional outcrop as elsewhere an unconformity can only be inferred from the structural divergence of Archean gneiss or supracrustal rock (unit **A**m) and Montresor Group. Nowhere was deformation intense and nowhere is basement intercalated with Montresor rocks. The close spatial association of the Montresor Group with the older metasedimentary rocks (unit **A**m) suggests that both units were deposited in basins at different times at the same site, and strengthens the argument that the Montresor Group is autochthonous.

Pegmatite intrusive into the Montresor Group is uncommon and confined mainly to its basal part. Irregular veins and sheets, typically up to one metre wide, of pink, coarse-grained muscovite-feldspar-quartz rock were found cutting argillaceous quartzite and marble. None of the pegmatite bodies could be extrapolated into the nearby granitoid terrane.

Environment of deposition

The Montresor Group was deposited in shallow water in a continental environment. Ripple marks and crossbedding are common; carbonate rock is prominent as a thin, basal unit; K-feldspar occurs abundantly and widely in the clastic rocks; and no volcanic or hypabyssal igneous rocks other than minor pegmatite have been found. Whether the depositional environment was marine or lacustrine, or a combination of both, cannot be established on present evidence. Orthoquartzite in the lower part of the mainly shallow-marine, Early Proterozoic Hurwitz Group (*see* below), not unlike the orthoquartzite of unit *A*Mo of the Montresor Group, was interpreted by Aspler et al. (1994) on the basis of ripple mark configuration as having been deposited in a vast, shallow, freshwater lake during a period of marine regression. Probably, the bulk, at least, of the Montresor Group was deposited in a shallow inland sea. In Condie's (1982) classification of Proterozoic supracrustal successions, the Montresor Group is an example of lithological assemblage I: quartzite-carbonate-shale, characteristic of deposition at stable continental margins or in intracratonic basins.

Age and correlation

The Montresor Group has not been dated and only upper and lower limits can be set to its age. Augen gneiss of the type on which the Montresor rocks have been deposited has been dated, in the nearby Cape Barclay map area, at 2596 Ma. If the pegmatite that intrudes the Montresor Group is of the same age as that intruding the Chantrey Group, the Montresor Group must be at least 1.7 Ga old.

The nearest lithostratigraphic correlative of the Montresor Group is the Amer Group, 50 km to the south (Patterson, 1986; Tella, 1994) and the Montresor Group rocks have been referred to as the Montresor belt of the Amer Group (Frisch and Patterson, 1983). The stratigraphic succession in the Amer Group of two clastic packages separated by a carbonate unit is similar to that at the eastern end of the Montresor belt but is some 500 m thicker and includes volcanic rocks (Patterson, 1986). The Amer Group has not been dated directly but is intruded by 1.85 Ga syenite (Tella, 1994). Thus the age of the Montresor Group probably falls between 2.5 Ga and 1.85 Ga.

Another possible correlative to the Montresor and Amer groups is the Hurwitz Group, which outcrops west of Hudson Bay. The Hurwitz Group overlies Archean basement and is characterized by a thick basal unit of orthoquartzite, locally underlain by conglomerate, and overlain by pelitic and psammitic rocks, carbonate rocks at several levels, and, in the lower part of the group, basaltic volcanic rocks and gabbro sills; most rocks are at least weakly metamorphosed (Aspler and Chiarenzelli, 1997). The Hurwitz Group proper is confined to a belt extending southwest from Rankin Inlet, well south of Baker Lake and thus is at least 400 km south of the Montresor belt. Although far removed, and stratigraphically and lithologically significantly different, from the Montresor Group, the Hurwitz Group may be of similar age. Baddeleyite from a gabbro sill has yielded an age of 2111.2 ± 0.6 Ma, which is a minimum age for the lower part of the Hurwitz Group intruded by the sill (Heaman and LeCheminant, 1993). An early Aphebian age may be widely valid for the supracrustal rocks overlying Archean basement north of Baker Lake.

*Chantrey Group (units A*Cm*, A*Cq*, A*Cs*, A*Co*, A*Cp*, A*Cc*)*

The Chantrey belt, 3–7 km wide and extending 175 km northeastward from near the head of Chantrey Inlet, is the largest of the four supracrustal belts in the study area. It was first recognized on the Geological Survey of Canada's airborne survey, Operation Back River, in 1960 and its constituent rocks were named the "Chantrey group" by W.W. Heywood, who stated: "Crystalline limestone and quartzite predominate, with lesser amounts of conglomerate, greywacke, slate, chert, and basic volcanic rocks" (Heywood, 1961, p. 5). Although lacking a marked topographic expression (Fig. 3), the Chantrey belt is a coherent unit, lithologically quite distinct from the granitoid complex, lying along the trace of the Murchison fault (Heywood and Schau, 1978). Thus even small-scale reconnaissance mapping such as was carried out on Operation Back River resulted in accurate delineation of the Chantrey belt on Heywood's (1961) geological map.

The rocks of the Chantrey belt are entirely of sedimentary origin (psammite, pelite, carbonate, and conglomerate). Although Heywood (1961) stated that greenstones derived from basic volcanic rocks occur, albeit rarely, in the Chantrey belt, no rocks of volcanic origin were recognized during the present study. The major rock types of the Chantrey Group are pelitic and psammitic schist, quartzite, marble, and metaconglomerate, representing metamorphosed shallow-water sediments. The common occurrence of andalusite and the sporadic presence of cordierite and fibrolite indicate low-pressure, amphibolite-grade metamorphism.

The Chantrey Group has been deformed into a canoe-shaped synclinorium comprising tight to isoclinal, more or less upright folds. The intensity of metamorphism and deformation varies along the length of the belt. Locally, especially at the western end of the belt, northwestward thrusting has refolded the Chantrey strata into recumbent folds with northeast-trending axes and has brought granitoid rock over the Chantrey Group along the invariably strongly sheared southern margin of the belt. In fact, all contacts of the belt with the granitoid complex are probably tectonic, although not necessarily sites of major movement. Southward dips and inverted stratigraphy of the supracrustal rocks at a number of places along the southern contact indicate overturning of the synclinorium to the northwest.

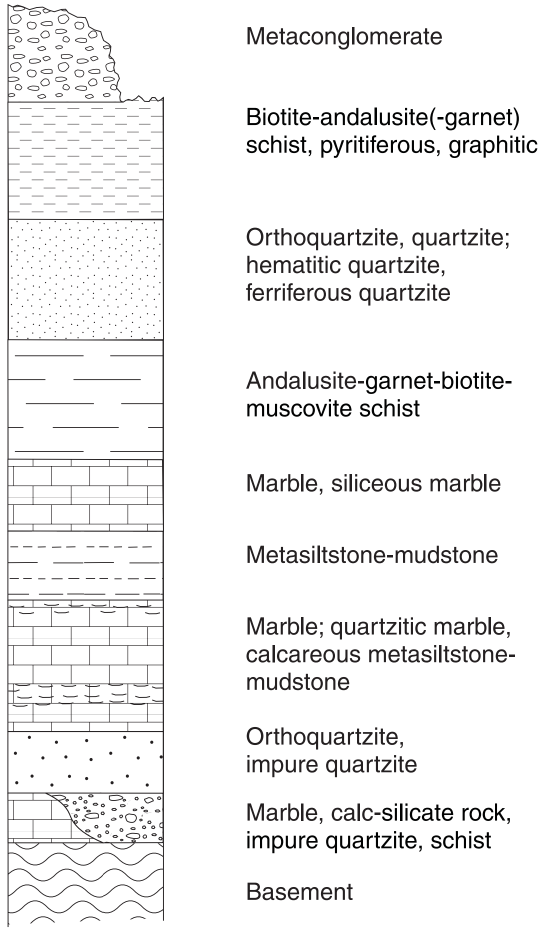


Figure 22. *Highly schematic and generalized stratigraphic section, not to scale, of the Chantrey Group.*

Metamorphism and severe deformation have rendered the stratigraphic sequence in the Chantrey belt uncertain. Tectonism has undoubtedly caused repetition and disappearance of units but the overall regularity and symmetry in the outcrop pattern of the Chantrey Group suggest a reasonably intact metasedimentary sequence. A generalized stratigraphic column is shown in Figure 22 and the units are described below in approximate order of deposition.

Lithology and geochemistry

*Marble and calc-silicate rock (unit A*Cm*)*

Metamorphosed carbonate rocks occur in the basal part of the Chantrey Group, both at the margins of the belt (presumably the base of the succession) and farther in (higher in the succession). They are dolomitic and generally impure and include a variety of calc-silicate rocks. Although commonly flow folded, the carbonate rocks, where less deformed, display well preserved stromatolites.

Much of the Chantrey belt in Darby Lake map area is bordered by a resistant unit of siliceous marble, commonly only a few metres thick (Fig. 23). The rock is white, medium grained and recrystallized, and locally rich in rosettes of coarse, pale

Figure 23. *Steeply dipping, light-coloured siliceous marble and calc-silicate rock (unit* ACm*) abuts basement at the northern margin of the Chantrey belt near its eastern end, northern Darby Lake map area (*see *Fig. 4). The low terrain is developed on poorly exposed quartzite of unit* ACo*. View is northeastward. Photograph by T. Frisch. GSC 204072-W*

Figure 24. *Close-up of a variety of the calc-silicate rock shown in Figure 23, showing coarse sprays of white tremolite, which locally coexists with cleavelandite feldspar. Photograph by T. Frisch. GSC 204635-G*

green tremolite (Fig. 24) and sprays of coarse, white cleavelandite feldspar crystals several centimetres long; clearly, growth of these two minerals postdated deformation. No original sedimentary features are preserved in these thin border units of marble. All structures within the marble parallel the contact with the granitoid complex.

Thicker exposures of the border rocks consist of dolomitic-calcitic marble, siliceous marble, and impure quartzite interlayered on the scale of centimetres to metres and commonly intensely deformed (Fig. 25). The calcareous rocks are white or pale grey; the quartzitic ones are grey or pale brown. Pale green to colourless tremolite-actinolite is prominent and some of the quartzites, which are finely laminated and feldspathic, carry two varieties of pale tremolite-actinolite and poikiloblastic diopside, accompanied by minor sphene.

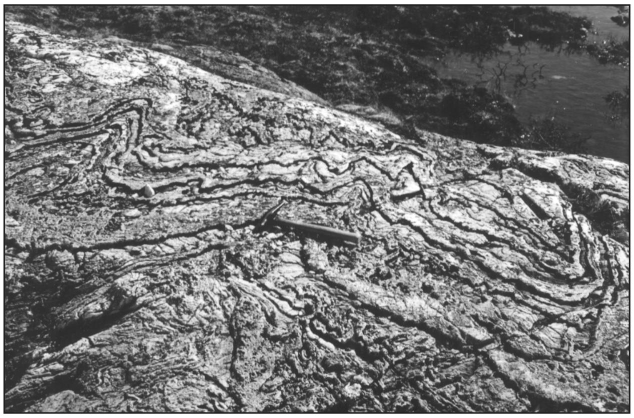


Figure 25. *Intense deformation in siliceous marble (unit* ACm*) at the northern border of the Chantrey belt near its western end, southwestern Cape Barclay map area. Photograph by T. Frisch. GSC 204635-H*

Carbonate rocks higher in the stratigraphic succession, though still in the lower half of the Chantrey Group, are thickest at the western end of the belt and display the best preserved primary structures in the central part of the belt.

At the western end of the belt, there appear to be two major calcareous units separated by a pelitic unit. Deformation in this area, however, is particularly intense and it is entirely possible that a single calcareous unit has been repeated by folding and faulting (presumably northwest-directed thrusting).

The outer (?lowest) major calcareous unit comprises four subunits, which locally are interlayered (possibly structurally repeated) and grade into each other. Deformation makes thickness estimates of these steeply dipping strata difficult but the maximum thickness of the calcareous unit is thought to be about 700 m. Outermost is a light- to medium-grey dolomitic metasiltstone/mudstone with an overall massive appearance but locally it is finely laminated and has a gradational contact with the dark grey, noncalcareous metapelite that borders the impure quartzite at the southern margin of the Chantrey belt. The second calcareous subunit is a resistant, white dolomitic quartzite-siliceous dolomite, thick bedded to massive, and 30–50 m thick. The third subunit consists of several types of carbonate rock, predominantly light coloured, dolomitic calcite marble but ranging from thick-bedded dolomite marble to schistose and flaggy, tremolite-bearing, dolomitic calcite marble, with abundant siliceous interbeds. Intense deformation, evidenced by flow folding, boudinage, and tectonic breccia, is pervasive and the only primary structure recognizable is bedding, defined by lithological contrasts. The fourth and final subunit is a finely laminated, calcareous metamudstone/siltstone with alternating pink and green laminae.

The outer major calcareous unit is separated from the inner one by a noncalcareous metasiltstone/mudstone unit about 100–300 m thick.

Overall, the inner calcareous unit, about 250 m thick, resembles the third subunit of the outer calcareous unit. It consists of grey, white, and buff, texturally variable dolomitic-calcitic marble with siliceous layers giving a strongly layered appearance to outcrops. Wave-washed exposures in Chantrey Inlet (*see* Frontispiece) show abundant disharmonic folding with shallow to moderate plunge, as well as boudinage.

East of Chantrey Inlet, marble generally occurs at or near the borders of the Chantrey belt and appears to be least deformed in that part of the belt bounded by latitudes 94°30′W and 95°W. It is here that stromatolites are best preserved (Fig. 26). They are of the noncolumnar type, flat-laminated or laterally-linked domal. Domal types are elliptical in horizontal section with long axes from 20 cm to 2 m in length; the high degree of preservation suggests the elliptical shape is primary. These stromatolites and cross-bedding, locally present in quartzitic interbeds in marble, provide way-up indicators. Poorly preserved stromatolites are found farther east and west in the Chantrey belt (but not at its extremities), but are too deformed to be used as way-up indicators.

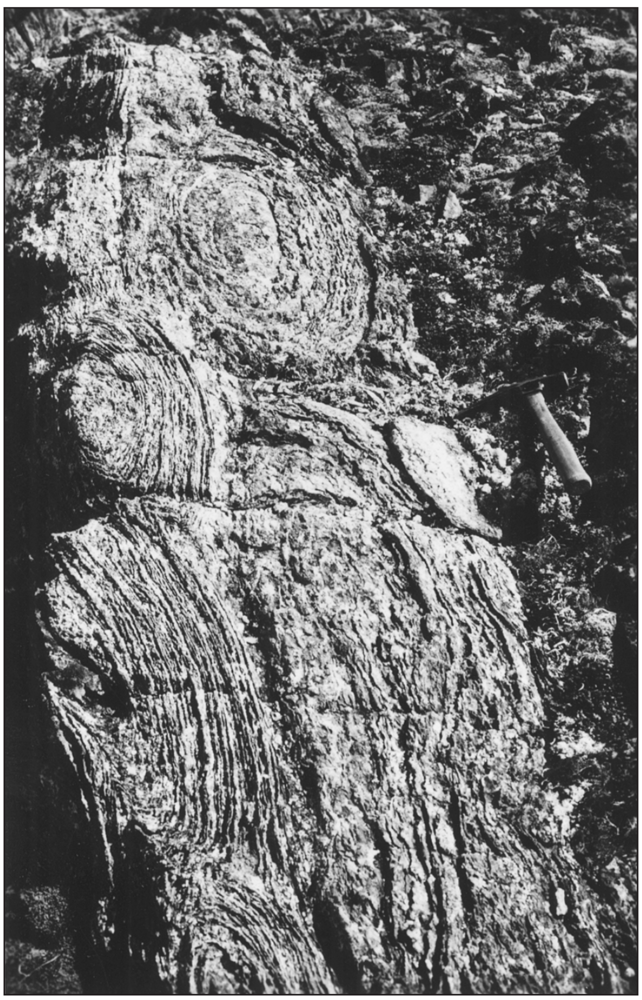


Figure 26. *Exceptionally well preserved stromatolites in marble of unit A*Cm *near the northern margin of the Chantrey belt, 28 km along strike from the east shore of Chantrey Inlet. Photograph by T. Frisch. GSC 204073*

*Impure quartzite; minor schist and metaconglomerate (unit A*Cq*)*

This unit forms the border rock to the Chantrey belt at its western end and in the region where the Murchison River cuts the belt. At the eastern end of the belt, the unit is separated from the gneisses by a thin layer of siliceous marble, described above.

Lithologically, the unit is heterogeneous but consists mainly of grey, pink, or white, fine-grained quartzite containing a little mica and/or chlorite and feldspar; small (up to 5 cm long) clasts of quartz and feldspar are locally present. At and near the contact with gneiss, the quartzite is strongly mylonitic and commonly mica laminated and/or chlorite laminated; in places, the mica is green (?fuchsite). At the western end of the belt, muscovite-biotite quartz schist and schistose quartzite, 3–5 m thick, occurs adjacent to gneiss, which itself is mylonitic to cataclastic. The schistose rocks contain completely recrystallized quartz clasts barely recognizable in thin section as diffusely bordered augen up to 15 mm long; S-C fabric is locally developed.

At the southern border of the Chantrey belt, 20 km from its eastern end, the ‘basal schist’ is noticeably conglomeratic in outcrop. Recrystallized quartz clasts appear as augen up to several centimetres long in a dark, biotite-rich schistose matrix. The biotite occurs in tiny brown flakes that grew perpendicular to the foliation defined by alternating micaceous and quartzose laminae.

Deformational fabric in the quartzite weakens with distance from the basement contact and ripple marks and cross-bedding are recognizable here and there.

*Metasiltstone/mudstone (unit A*Cs*)*

This very fine-grained semipelitic unit is generally found inboard of the inner marble unit, thus is thought to stratigraphically overlie it. Although in detail lithologically variable, it can be characterized as a grey, laminated, muscovite-biotite-quartz schist, commonly with porphyroblasts of garnet and andalusite and, rarely, biotite and cordierite. Laminations generally are from a few millimetres to 2 cm thick and defined by variations in quartz and mica content. Psammitic laminae are lighter coloured than pelitic ones but some rocks are so finely laminated that they are virtually massive. Cross-bedding and ripple marks, which commonly are poorly preserved, were seen locally. Rusty, pyrite-rich lenses occur sporadically.

In thin section, the rocks are seen to consist of, typically, approximately equal amounts of biotite and muscovite in a quartzofeldspathic matrix in which quartz greatly predominates over feldspar (andesine, where determinable). Quartz and feldspar grain size is about 0.2–0.3 mm; the mica grains are about 0.4–0.5 mm long. Matrix biotite, with X = nearly colourless and Y = medium brown, and muscovite appear, for the most part, to have grown contemporaneously in the foliation plane but there is minor development of cross mica. Quartz and feldspar grains are straight sided and the feldspar is only slightly sericitized. The few plagioclase composition determinations possible range from An_{23} to An_{26}. Porphyroblastic minerals include pink garnet, 1–3 mm across, andalusite, commonly 6–8 cm but up to 25 cm long, biotite, up to 1 mm long, and cordierite, 3–5 mm across; the last two minerals, however, are rare. Idioblastic almandine garnet porphyroblasts in a rock from the central Chantrey belt (FS84-59A in Appendix, Table 1) are zoned, with Fe and Mg increasing, and Mn and Ca decreasing, from core to rim. Coexisting biotite porphyroblasts (0.5 mm) have X_{Fe}=0.65 and TiO_2=1.88% (Appendix, Table 2); matrix muscovite has Na/Na+K=0.07 (Appendix, Table 3). Andalusite forms spongy porphyroblasts with small inclusion-free cores. Cordierite is invariably completely altered and has not been found coexisting with garnet. Like andalusite, it formed late, overgrowing the foliation.

*Orthoquartzite, quartzite, and associated rocks (unit A*Co*)*

More or less pure quartzitic rocks are associated, or even interlayered, with pelitic units throughout the Chantrey belt; they are particularly abundant at the eastern end of the belt but outcrop is generally poor (Fig. 23).

The quartzite units are white, grey, pink, or purple, fine grained, and thick bedded to massive. Ripple marks and crossbedding are relatively uncommon. Thin pelitic layers occur locally within the quartzite, particularly near contacts with pelitic units. Contacts of quartzite with other rocks may be gradational or sharp.

Where deformation was weak but recrystallization nonetheless took place, quartzite shows excellent relict clastic texture: little-strained quartz grains, 0.2–0.6 mm in diameter, are bordered by flakes of muscovite and biotite, which has X = nearly colourless and Z = brown or greenish brown. Muscovite predominates over biotite in the white quartzite, whereas the reverse holds true in the darker grey varieties. Purple quartzite contains finely disseminated hematite and a little muscovite. In strongly deformed quartzite, annealed texture is characteristic and mica flakes are aligned at a shallow angle to the layering defined by the quartz grains, forming an S-C fabric.

Pelitic layers in quartzite are typically a few centimetres to 30 cm thick and consist of dark (cordierite-andalusite-)biotite-muscovite schist. At the western end of the Chantrey belt, spongy andalusite crystals, up to 10 cm long, are heavily altered to muscovite and minor, altered cordierite occurs in skeletal crystals, 2 mm across, intergrown with quartz; the matrix consists of quartz, biotite, muscovite, and accessory tourmaline.

On the small island underlain mainly by quartzite in Chantrey Inlet just off Victoria Headland, purple quartzite is interlayered with rusty-weathering, dark garnet-mica schist, which contains sillimanite. The schist and quartzite layers are 7–20 m thick. The schist consists of porphyroblasts of garnet, 1–2 mm across, and felted masses of fibrolitic sillimanite associated with biotite in a graphitic biotite-muscovite-quartz matrix. The garnets are almandine and slightly zoned ($Alm_{91}Prp_6$... in the core to $Alm_{94}Prp_4$... in the rim; Appendix, Table 1, FS84-14A). Garnets are cracked and muscovite flakes (Na/Na+K=0.12; Appendix, Table 3), which are up to 4 mm long, are strongly kinked. Much of the biotite, nearly colourless to chestnut-brown with X_{Fe}=0.74 and TiO_2=1.82% (Appendix, Table 2), has been replaced by fibrolite, as shown by fibrolite rims with interdigitate borders against biotite and fibrolite wisps penetrating biotite. Besides fibrolite, sillimanite also occurs in scattered aggregates of tiny needles, confirming that the rock entered the stability field of sillimanite during metamorphism. This is the only known occurrence of sillimanite in the Chantrey belt. Finely disseminated graphite doubtless accounts for the dark colour of the rock.

A ferriferous quartzite with exceptionally iron-rich minerals occurs within metapelite of unit *A*Cp in the middle part of the Chantrey belt, about 11 km west of the boundary between the Cape Barclay and Darby Lake map areas. Exposure is poor in this region and the ferriferous quartzite appears to form large lenses in the metapelite, several metres thick and traceable discontinuously over a distance of 5 km. This variety of quartzite has not been recognized elsewhere in the Chantrey belt. Chemical analyses of ferriferous quartzite and associated iron-rich pelitic-psammitic rocks are given in the Appendix (Table 9).

Due to a high content of mafic minerals, the ferriferous quartzite in hand specimen superficially resembles an igneous (gabbroic) rock more than a metasedimentary one. Dark brown aggregates (1–2 mm) of amphibole, commonly cored by tiny garnets, and magnetite grains (0.2–0.5 mm) are set in a fine-grained matrix of quartz. Under the microscope, the amphibole-garnet aggregates in thin section are resolved into sprays of polysynthetically twinned, pale brown amphibole intergrown with, and locally, overgrown by, pink garnet (Fig. 27). Both these

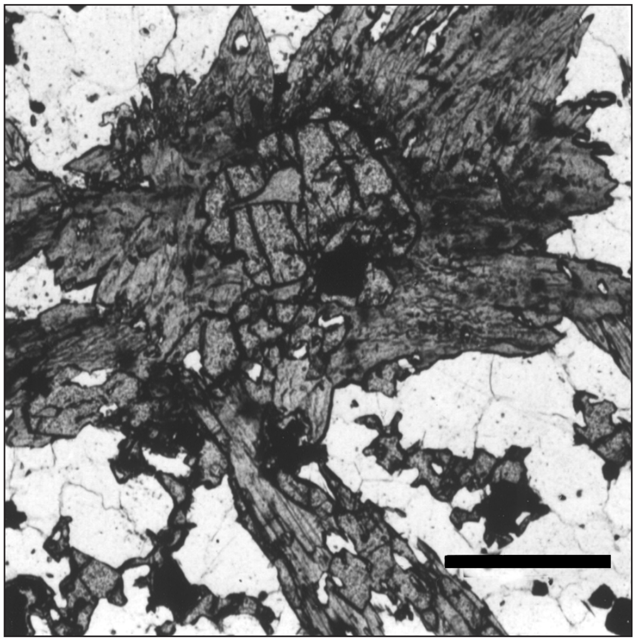


Figure 27. *Photomicrograph of an aggregate of grunerite (X_{Fe}=0.96–0.99) cored by near end-member almandine (X_{Fe}=1.00) in ferriferous quartzite of unit A*Co *of the Chantrey Group. Black grains are magnetite. Sample FS84-86B (Appendix, Tables 1, 7). Bar scale is 0.5 mm.*

minerals are nearly pure iron end-members of their respective compositional series: grunerite with X_{Fe}=0.96–0.99 and almandine with X_{Fe}= 1.00 (Appendix, Tables 7, 1; samples FS84-86A, FS84-86B). Almandine of identical composition also occurs as thin rims on magnetite. With no detectable magnesium and a little calcium (up to 0.71% CaO) as the only significant impurity, the almandine is, to the writer's knowledge, the most iron-rich natural garnet yet reported. A little grass-green chlorite (X_{Fe}=0.96; Appendix, Table 4) occurs associated with the mafic silicate minerals and as discrete flakes. The quartz grains (0.3–0.6 mm) have sutured borders and tiny apatite grains are disseminated throughout the matrix.

A garnet- and biotite-rich, magnetite-poor layer in ferriferous quartzite consists of spongy garnet porphyroblasts, 2–3 mm across ($Alm_{93}Prp_5Grs_1Adr_1$; Appendix, Table 1, FS84-87B), and partly chloritized, pale brown biotite (X_{Fe}=0.64, TiO_2=0.95%; Appendix, Table 2) in a quartzose matrix.

*Dark pelitic schist (unit A*Cp*)*

Dark grey, pyritiferous, pelitic schist, characterized by coarsely porphyroblastic biotite, forms a major component of the central and eastern Chantrey belt. A hallmark of this unit is glossy, black porphyroblasts of biotite commonly about 1 cm, but locally 2 cm long. Andalusite (chiastolite) porphyroblasts, up to 15 cm long and commonly skeletal, occur in some layers but garnet is relatively sparse and crystals rarely exceed 2 mm in diameter. The matrix is fine grained. Rusty-weathering patches and lenses are abundant throughout. The schist is thick bedded to massive and locally psammitic and ripple marked.

Thin sections show abundant randomly oriented biotite porphyroblasts ranging from 0.5 mm to 1 cm in length. Some porphyroblasts are composite, consisting of aggregated flakes of different size. All biotite flakes are spongy and have ragged edges and innumerable pleochroic haloes. Pleochroism is very pale brown or neutral to brown. Biotite analyzed in three samples of schist (Appendix, Table 2, FS84-87B, FS84-A85, FS84-A142) have X_{Fe}=0.55–0.60, TiO_2= 0.94–1.85%, and significant MnO, 0.11–0.23%, consonant with the spessartine-rich nature of the coexisting almandine garnet (*see* below).

Other porphyroblast minerals are garnet and andalusite. Garnet occurs as euhedral to subhedral crystals, typically 1–2 mm in diameter and overgrown by porphyroblastic biotite. Garnet analyzed in two schists (Appendix, Table 1, FS84-A85, FS84-A142) are Mn-rich almandine with normal prograde zoning, $Alm_{68}Sps_{23}Prp_5$ (core) to $Alm_{72}Sps_{17}Prp_6$ (rim). Andalusite is characteristically skeletal with crystal cores that are free of inclusions. Crystal margins tend to be rich in inclusions and complex, with arms or inclusion-free zones interdigitated with matrix and graphite fringes on crystal terminations (Fig. 28). The graphite accumulated as the graphite crystal grew out into the matrix. In contrast, biotite appears to have incorporated graphite as it grew, resulting in the 'dirty' appearance of the heavily dusted flakes. Crystallization of andalusite outlasted that of biotite and garnet as both

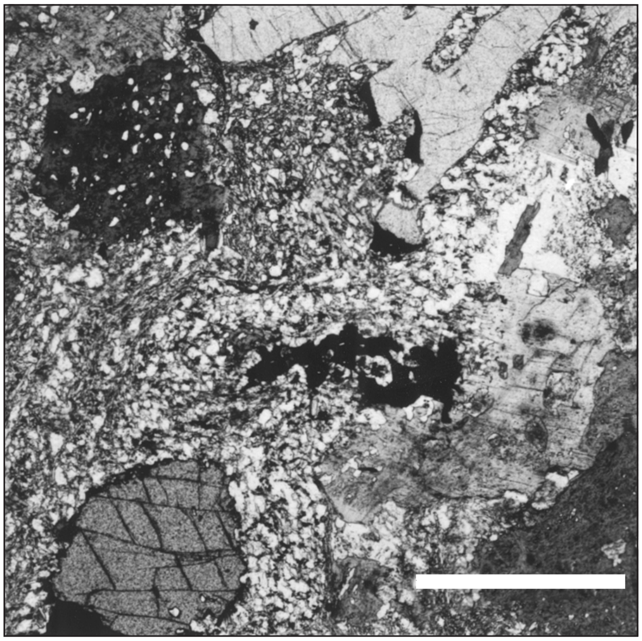


Figure 28. *Photomicrograph of typical graphitic schist of unit A*Cp *of the Chantrey Group. Porphyroblasts of biotite (characteristically spongy), skeletal andalusite, and garnet are visible. Note the graphite accumulations at the ends of andalusite 'arms', formed as the crystal grew out into the graphitic matrix. Bar scale is 1 mm. GSC 1995-238F*

these minerals occur as inclusions in andalusite. Opaque accessory minerals are ilmenite in tablets 0.2–0.3 mm long and subordinate pyrite. In outcrop, pyrite is locally concentrated, giving rise to rusty zones.

The matrix of the schist is extremely fine grained and consists of quartz and graphite alone in some rocks, accompanied by white mica or chlorite in others. Chlorite was analyzed in two schists, one of them garnet-bearing. It is ripidolite with appreciable manganese (MnO=0.25–0.39%; Appendix, Table 4).

Twenty-five samples of unit *A*Cp were chemically analyzed. Seven of them are compositionally aberrant with respect to shale (Appendix, Table 10), e.g. high (>71%) or low (<49%) SiO_2, high (>4%) MgO, high (>3%) CaO, and/or high (>19%) total iron as Fe_2O_3.

The remaining 18 samples, tabulated in Appendix, Table 10, show the following compositional ranges for selected oxides (in weight per cent) and trace elements (in ppm): SiO_2=49.1–70.9, TiO_2=0.34–1.54, Fe_2O_{3T}=7.0–12.2, MgO=1.86–3.64, CaO=0.12–2.48, Na_2O=0.2–3.2 and K_2O=2.50–9.69, Nb=16–66, Y=30–73, Cr=29–130, and Ni=20–85. In Appendix, Table 10, the average composition of the 18 samples is compared with average compositions of shale — the presumed protolith of unit *A*Cp — compiled by Cameron and Garrels (1980) and Condie (1993). Major differences include higher Fe (10.4% total Fe as Fe_2O_3) and Nb (41 ppm) and lower Cr (76 ppm) and Ni (37 ppm) in the Chantrey Group rocks; three shale averages have 6.3–6.4% total Fe_2O_3, up to 16.8 ppm Nb, 104–125 ppm Cr, and 52–58 ppm Ni.

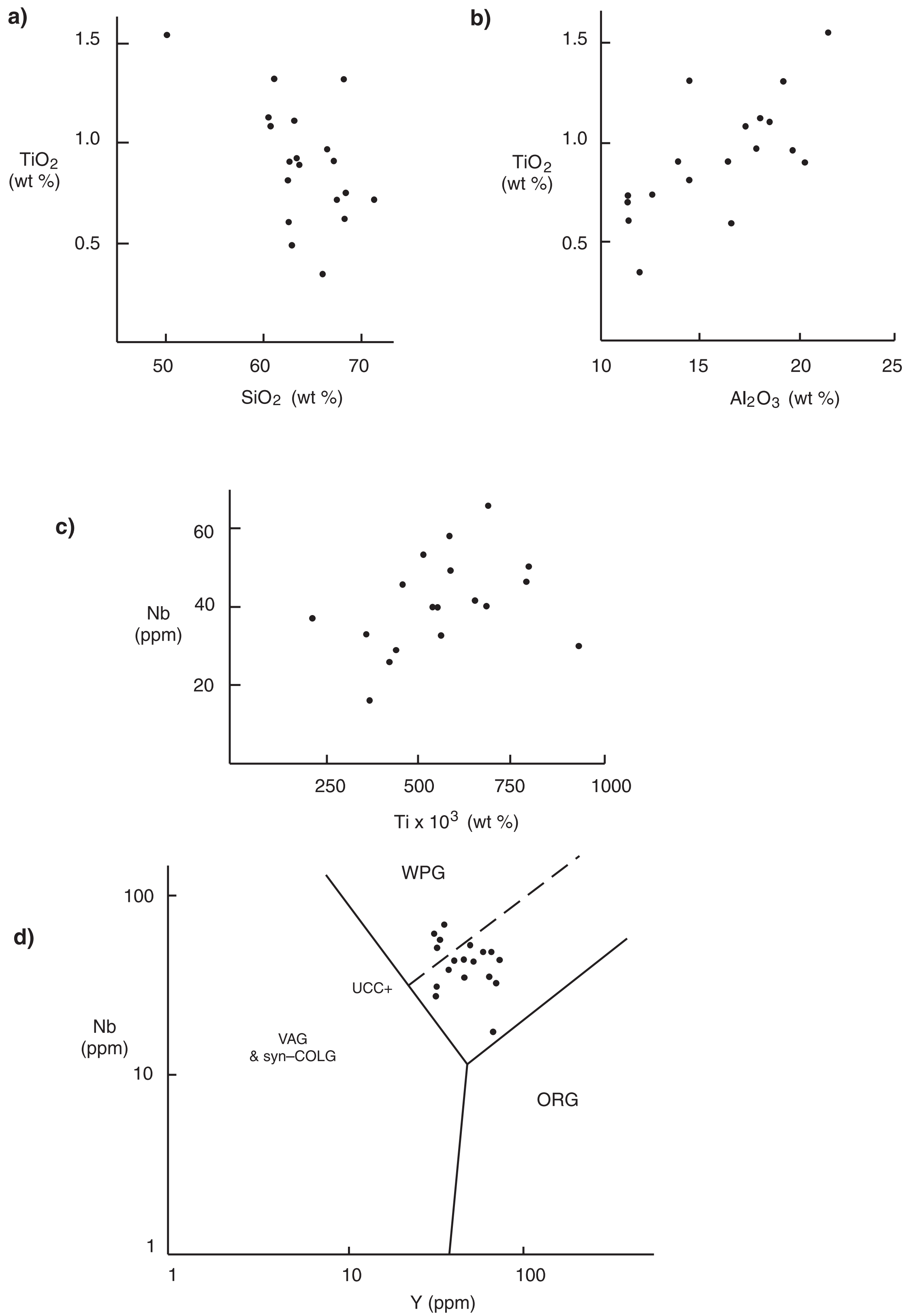


Figure 29. *Chemical variations in pelitic schist of unit A*Cp *of the Chantrey Group. Chemical analyses in Appendix, Table 10.* ***a)*** *TiO_2 vs. SiO_2;* ***b)*** *TiO_2 vs. Al_2O_3;* ***c)*** *Ti vs. Nb;* ***d)*** *Nb-Y discriminant diagram after Pearce et al. (1984). WPG = within-plate granites, VAG = volcanic-arc granites, syn-COLG = syncollisional granites, ORG = oceanic-ridge granites. Dashed line marks the compositional limit of ORG from anomalous ridge segments. UCC = average upper continental crust (Taylor and McLennan, 1985).*

Although the Chantrey Group metapelite units are iron-rich, their other chemical characteristics indicate the rocks were derived from a felsic source. Contents of Ni and Cr are low and those of the relatively immobile large-ion elements Nb and Y are high. As TiO_2 correlates negatively with SiO_2 and positively with Al_2O_3 (Fig. 29a, b), the positive correlation between Nb and Ti (Fig. 29c) suggests that Nb was contained in the clay-sized, rather than the sand-sized, fraction of the original sediment. Thus the Nb content of the Chantrey Group metapelite samples should reflect the composition of the source rocks. In the Nb-Y discriminant plot of Pearce et al. (1984) (Fig. 29d), the metapelite samples fall entirely within the field of within-plate granites, whose composition overall is probably not unlike that of the granitic terrane adjacent to the Chantrey belt.

*Metaconglomerate (unit A*Cc*)*

Metaconglomerate of this unit of the Chantrey Group is confined largely to that part of the belt immediately east of Chantrey Inlet. From the shore of the inlet eastward for about 25 km metaconglomerate forms most of the exposed Chantrey belt. It also underlies a fairly large area near the eastern border of the Cape Barclay map area. Aside from that exposure, metaconglomerate has not been found east of approximately 94°45′W nor west of Chantrey Inlet.

The metaconglomerate consists of pebbles and cobbles, rarely boulders, of white quartz, subordinate granitoid rock and very subordinate mafic schist in a medium- to fine-grained matrix of quartz, feldspar, and mica (Fig. 30a). The matrix is medium to dark grey, weathering to a characteristic pink or pale orange colour. Bedding is difficult to discern; the rocks typically having an overall massive appearance. Locally, clasts are concentrated in distinct layers (Fig. 30b). Where deformation is weak, clasts tend to be moderately rounded but angular ones are common and size sorting is poor. Clast abundance ranges from 40% to 10% by volume of rock and averages perhaps 20–30%. Clast sizes of 2–3 cm predominate but 10 cm long cobbles are abundant and clasts 20–30 cm long are present locally. In general, clast size and abundance, and the proportion of mafic clasts, appear to decrease with distance from the margins of the main metaconglomerate exposure near Chantrey Inlet.

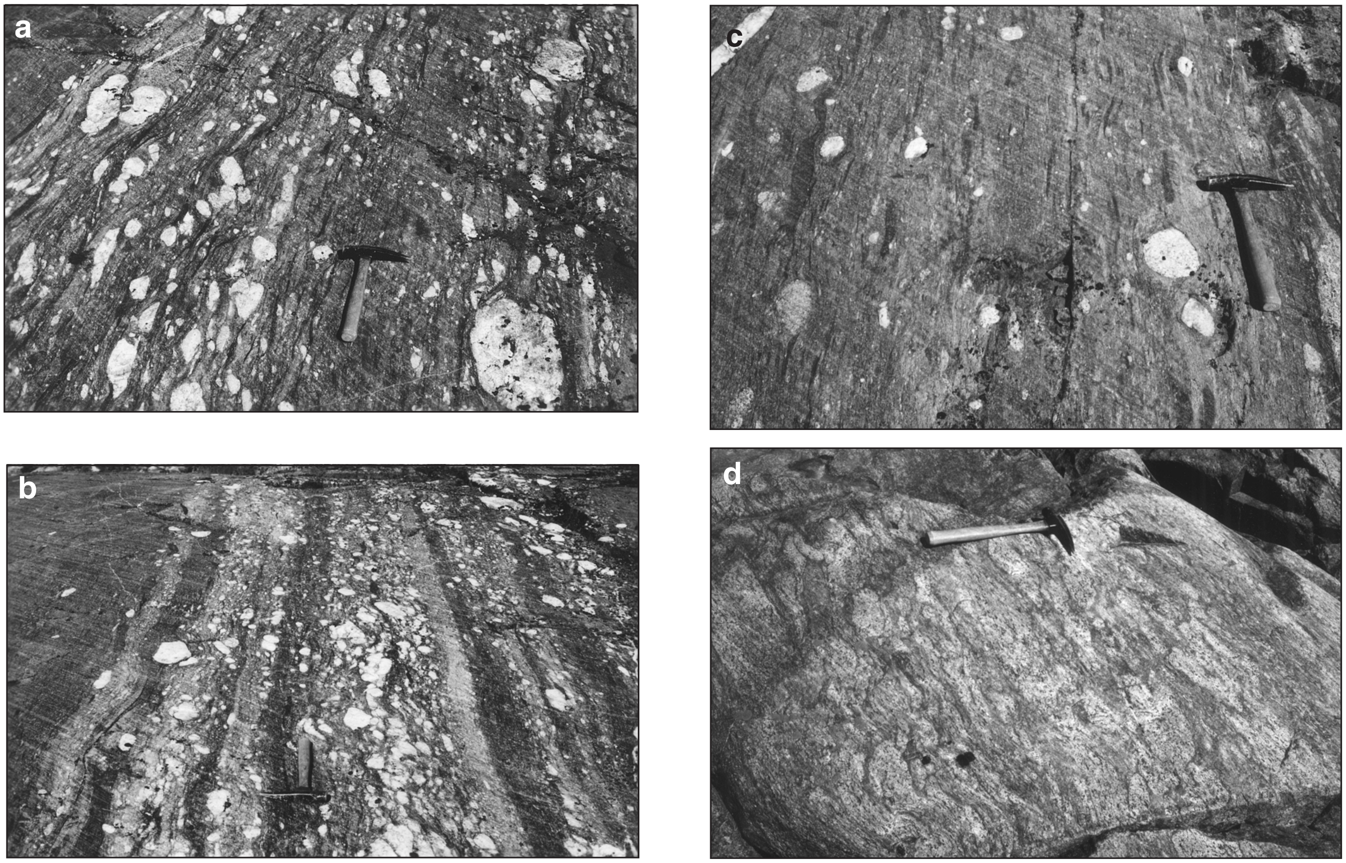


Figure 30. *Metaconglomerate (unit A*Cc*) at the southern margin of the Chantrey belt, as exposed on the islets in Chantrey Inlet southwest of Victoria Headland.* ***a****) Note the boulders and flattened, schistose mafic clasts. Photograph by T. Frisch. GSC 204072-U* ***b****) The layered appearance of this outcrop is unusual for unit A*Cc *in general. Photograph by T. Frisch. GSC 204072-Y* ***c****) Glacially striated outcrop of metaconglomerate relatively poor in quartzose clasts but rich in attenuated mafic clasts. Photograph by T. Frisch. GSC 204635-E* ***d****) Metaconglomerate in which intense deformation has blurred the clast outlines. Photograph by T. Frisch. GSC 204635-U*

The matrix of the metaconglomerate is seen in thin section to be a very fine-grained biotite gneiss, containing abundant plagioclase and microcline in addition to quartz and well aligned flakes of brown biotite; muscovite and Fe-Ti oxide are present in accessory amounts.

Pelitic interbeds between metaconglomeratic layers locally contain andalusite pseudomorphs after staurolite.

Along most of the contact, the metaconglomerate is deformed, as shown by flattened clasts and a strongly gneissic matrix. Rocks of the margins of the metaconglomerate unit are well exposed on wave-washed islets off Victoria Headland in Chantrey Inlet (Fig. 30) but it should be emphasized that these rocks are not necessarily representative of the bulk of the metaconglomerate unit. The metaconglomerate units are more heterolithic than those farther from the contact and are rich in mafic schist clasts, which commonly are highly attenuated parallel to bedding (Fig. 30c). On the southernmost islet, presumably very close to the eastern margin of the Chantrey belt, the metaconglomerate is strongly deformed to a gneissic rock and clast outlines are blurred (Fig. 30d). The matrix of this rock, however, preserves a good clastic texture: rounded quartz clasts averaging 1 mm in diameter, consisting of a mosaic of sutured grains, in a fine-grained, equigranular groundmass rich in feldspar and quartz and with tiny flakes of brown biotite and minor muscovite; all grains are well aligned and the small grain size may be due to mylonitization, followed by recrystallization.

Contact relations

Despite significant deformation and metamorphism, contacts between the Chantrey belt and the Archean granitoid complex are everywhere abrupt. No evidence of intrusion into the Chantrey Group by granitoid rock (except pegmatite) was found, nor has any undisputed depositional contact been identified. Although in many places the contact may well be nonconformal, it appears most commonly to be tectonic.

Generally, the actual contact between Chantrey Group and Archean granitoid rocks is not exposed but it is observable at a number of localities, particularly at or near the eastern and western ends of the Chantrey belt.

At Victoria Headland, on the east shore of Chantrey Inlet, metaconglomerate of unit *A*Cc is in conformable, steeply north-dipping (65°) contact with sheared, cataclastic granitoid gneiss. The highly deformed conglomerate outcropping on the islets in Chantrey Inlet, on strike with the conglomerate-gneiss contact at Victoria Headland, may represent the basal deposit of the Chantrey Group at this locality, close to a depositional contact, but the granitoid complex is not exposed. A few kilometres inland from Victoria Headland, however, both metacongomerate and gneiss dip moderately south and a thrust-faulted contact is inferred but not necessarily associated with significant displacement.

At the south side of the extreme western end of the Chantrey belt, schistose, locally conglomeratic quartzite (unit *A*Cq) dips steeply (60–70°S) and conformably under mylonitic-cataclastic granitoid gneiss and the contact is demonstrably a high-angle reverse fault. A similar situation obtains on the south side of the Chantrey belt near its eastern end in Darby Lake map area. There, the supracrustal rocks against granitoid complex are either schistose quartzite or marble–calc-silicate rocks. Variations in lithology of the Chantrey Group are particularly common at the southern margin of the belt and provide circumstantial evidence of tectonism that locally cuts units out.

The northern margin of the Chantrey belt is more regular than the southern and the contact, along the entire length of the belt, is predominantly between granitoid complex and marble–calc-silicate rock, dipping steeply southward (Fig. 23). Nevertheless, gneiss and carbonate rocks are typically highly sheared and the contact is largely, if not solely, tectonic. Deformation at the western end of the belt was intense enough to have resulted in disruption of the Chantrey Group. Just north of the north margin of the belt, at 67°15′N, cataclastic granitoid rock encloses an approximately 200 m x 500 m lens of white, fine-grained, recrystallized quartzite (unit *A*Cq), oriented parallel to the gneiss–Chantrey Group contact. Part of the straight fault scarp marking the latter is well shown by the Frontispiece.

Environment of deposition

Due to the high degree of metamorphism and deformation of the Chantrey Group, deciphering its depositional environment is not straightforward. Furthermore, any analysis of the origin of the Chantrey Group rocks must account for the following fundamental features:

- The Chantrey Group is constituted entirely of metamorphosed sedimentary rocks.
- Although the site of the Chantrey belt is a fault zone, there is no evidence that the zone marks a break between two different crustal blocks, such as a suture, terrane boundary, or other deep rift.
- The Chantrey Group is an intracratonic supracrustal succession apparently developed on and from continental crust.
- The narrow, linear shape of the Chantrey belt and the roughly symmetrical distribution of rock units within it.

The Chantrey Group consists of a psammite-pelite succession with calcareous strata in its lower part. Although possessing attributes of assemblage I (quartzite-carbonate-shale) in Condie's (1982) classification of Proterozoic supracrustal successions, the Chantrey Group includes atypically large proportions of carbonate and argillaceous rocks. In parts of the Chantrey belt, a conglomerate unit is inferred to overlie the psammite-pelite succession. Sedimentary structures characteristic of shallow-water deposits, such as small-scale cross-bedding and ripple marks are found throughout the main Chantrey succession. Rocks formed from sandy and silty to muddy sediments alternate on both large and small scales. The evidence thus favours a shallow-water origin for the entire Chantrey Group.

The calcareous rocks in the lower part of the Chantrey succession are interpreted to be shoreline deposits, common at the margins of inland seas and some lakes (Picard and High, 1981). If the alternation of carbonate and clastic rocks is real, and not the result of tectonic repetition, it suggests a shallow body of water whose shoreline position fluctuated due to climatic variations, with concomitant facies changes. Stromatolites are widely known from both shallow-marine and lacustrine environments (Dean and Fouch, 1983; Sellwood, 1986). The considerable thickness and lateral extent of carbonate rocks in the Chantrey belt accord more with a littoral marine origin.

The fine-grained, dark grey metapelite bodies of units *A*Cs and *A*Cp are confined largely to the inner parts of the Chantrey belt. Both units are carbonaceous, especially unit *A*Cp, which is also pyritiferous. Fine-grained, carbonaceous clastics are likely to settle in areas where water circulation and aeration are restricted, typically deep water. However, in the inland sea and lacustrine environments, where wind fetch is limited, wave base is much shallower than in open-marine shoreline areas and sediments of deep-water aspect may be deposited near shore. Whether the silty and muddy protoliths of units *A*Cs and *A*Cp were deposited in bays and estuaries of a shoreline or in relatively deeper water offshore, conditions were likely oxygen-depleted, conducive to accumulation of carbonaceous material and H_2S.

The lithological associates of the metapelite bodies suggest a shallow-water environment. The (?overlying) metaconglomerate unit *A*Cc is interpreted to be derived from an alluvial fan built on shore, possibly from a source area now occupied by Chantrey Inlet. Iron-bearing sediments, such as those represented by the hematitic quartzite and magnetite-rich quartzite lenses of unit *A*Co, are generally thought to have been deposited in shallow water (Pettijohn, 1975; Blatt, 1992).

Outcrop of the Chantrey Group is restricted to a sharply defined, long and narrow belt, much of which occupies a fault zone. It is lithologically distinct from other northeast-trending supracrustal belts in the study area. Thus, although the Chantrey belt may be a fortuitously preserved single remnant of deposits that once covered a much wider area, more likely its shape reflects a confined depositional basin whose length significantly exceeded its width. The basin may have been created by reactivation of the structurally weak zone in which the Franklin belt to the southwest was deposited and was perhaps marked by a number of parallel faults, now reduced to the narrow Murchison fault zone. Relief may have been low and the basin floor covered by the waters of a shallow inland sea.

Age and correlation

The only radiometric age available from the Chantrey belt is a ^{39}K-^{40}Ar determination on muscovite from a "pegmatite sill" intrusive into metaconglomerate (unit *A*Cc) near the southern margin of the belt about 15 km from Chantrey Inlet (Heywood, 1961; GSC 61-91 *in* Lowdon et al., 1963). Recalculated using modern decay constants, the age obtained is 1684 Ma. Thus the Chantrey Group is at least 1.7 Ga old. A maximum age for the Chantrey Group can only be inferred from circumstantial evidence such as the absence of intrusion by granitoid rock other than pegmatite and the far lesser disruption (by intrusion or tectonism) of the Chantrey belt compared to the Barclay and Franklin belts, the former of which is known to be Archean. The Chantrey Group is probably Aphebian and was metamorphosed at least 1.7 Ga ago.

The most likely correlatives of the Chantrey Group are the Penrhyn Group on Melville Peninsula, east of Chantrey Inlet, and the Wollaston Group of northern Saskatchewan. All three groups comprise highly deformed and metamorphosed, early Proterozoic, psammite-pelite-carbonate successions resting on Archean basement. The Penrhyn Group consists of a lower clastic and carbonate sequence and an upper carbonaceous shale and arkosic wacke sequence, all deposited in a shallow, stagnant basin in the early Proterozoic (Henderson, 1983). A reliable estimate of the depositional age of the Penrhyn Group is given by a U-Pb zircon age of 1.88 Ga on a synsedimentary sill intrusive into the Piling Group of central Baffin Island, which is correlated with the Penrhyn Group (Henderson and Henderson, 1994). Deformation into a northeasttrending thrust-and-fold belt was closely followed by upper amphibolite-grade metamorphism (≥0.5 GPa, 700–750°C) at 1.80 Ga (Rb-Sr whole-rock isochron age) (Henderson, 1983). In the Wollaston Group, pelitic rocks occur mainly in the lower part (a graphitic unit commonly lies at the base), psammitic rocks dominate the upper part, and, although calcareous rocks are present, marble is not abundant (Annesley, 1989). Metamorphism attained upper amphibolite to lower granulite grade 1.81–1.82 Ga ago (U-Pb monazite ages) (Annesley, 1989; Annesley et al., 1996). Both the Penrhyn and Wollaston successions include metamorphosed mafic rocks (amphibolite) but these are not voluminous (Henderson, 1983; Annesley, 1989).

A less likely correlative of the Chantrey rocks is the Montresor Group, which itself is correlated with the Amer Group (Frisch and Patterson, 1983). The Chantrey and Montresor groups may well be coeval but they differ significantly in lithology.

APHEBIAN PEGMATITE

Pegmatite intrusive into the Aphebian supracrustal rocks is not distinguishable in the field from much of the pegmatite in Archean rocks of the study area. It was not studied in detail, is not abundant, and no bodies large enough to be mapped were seen. However, pegmatite cutting the Montresor and Chantrey groups is considered separately here because of its importance as an indicator of both the age of the host rock and the likely presence at depth of parental granitic bodies.

Typically, the pegmatite consists of pink K-feldspar in crystals up to 4 cm across, quartz, and coarse muscovite and forms sheets, 1–5 m thick, that are more or less concordant with the enclosing rock; an example is the radiometrically dated body in the Chantrey belt (*see* above). No folded or otherwise deformed pegmatite was seen in the Montresor and Chantrey belts but the common lack of strong discordance between pegmatite and host rock suggests that intrusion was more late tectonic than post-tectonic.

The ^{39}K-^{40}Ar age of 1684 Ma from pegmatite in the Chantrey belt is the only age available for pegmatite of the study area. Having been determined on coarse muscovite from an undeformed, low-temperature magmatic rock, the age may closely approximate the time of pegmatite formation. However, the age falls within a limited range of K-Ar mica ages, 1750–1642 Ma (recalculated using modern decay constants), from 10 (of 11) samples collected during Operation Back River (Heywood, 1961; Lowdon et al., 1963). Such a tight cluster of ages from diverse, mainly infracrustal rocks widely dispersed over a very large area must reflect a regional cooling event in which the K-Ar closure temperature for micas, about 300–350°C (Hanes, 1991), was reached. As all U-Pb zircon ages from the present study area are late Archean, a 1.7 Ga K-Ar age may have little significance as to the time of crystallization. Nevertheless, the genetic link between pegmatite and granite, in that granitic pegmatite is a volatile-rich fractionate of a granitic melt, is unequivocal (Jahns and Burnham, 1969) and the conclusion that granitic magmatism occurred at some time in the lower Proterozoic in the study area, even if the evidence for it is, for the most part, literally buried, seems inescapable.

DIABASE DYKES (UNIT db)

Unmetamorphosed diabase dykes are uncommon in the study area and outcrops sparse. All such dykes — outcropping and inferred from aeromagnetic data — trend northwest, cut granitoid and supracrustal rocks, and are regarded as being of Mackenzie age (1.27 Ga).

Two major gabbro dykes, up to 100 m thick, follow the western shoreline of Franklin Lake and their emplacement was probably related to the fault zone occupied by the lake. One of the dykes cuts the Franklin supracrustal belt at the northern edge of Ian Calder Lake map area.

A second locality at which a thick diabase dyke outcrops is at the east end of the Chantrey belt in Darby Lake map area. The eastern side of the dyke outcrops adjacent to impure marble (unit *A*Cm) at the north margin of the Chantrey belt but the actual contact — undoubtedly intrusive — is weathered out. The dyke exposure is about 40 m wide and ranges from medium- to fine-grained gabbro as the contact is approached. The southern extension of the dyke does not outcrop but can be traced on aeromagnetic maps to the vicinity of Darby Lake.

Farther west in the Chantrey belt, a weathered diabase dyke outcrops in a transcurrent fault zone near the eastern border of Cape Barclay map area at about 94°13′W.

Two major dykes in the study area are recognizable only by their aeromagnetic expression as no outcrops have been found. One runs from the granulite belt in southern Cape Barclay map area to Cockburn Bay and beyond into Chantrey Inlet. The other trends across Ian Calder Lake.

METAMORPHISM

Metamorphic grade is either amphibolite or granulite and is essentially uniform over the entire study region, both in Archean and Proterozoic rocks. Metamorphic pressures are consistently low, as signified by the common presence in metapelitic rocks of andalusite and cordierite in amphibolite facies and cordierite in granulite facies.

The amphibolite-facies rocks of the Archean granitoid complex are essentially devoid of diagnostic mineral assemblages. Judged by its deep green colour, amphibole in these rocks appears to belong to the Ca- and Fe-rich hastingsitic group. Biotite is brown or greenish brown.

In the Archean granulite belt, exposed in northeastern Ian Calder Lake area and southeastern Cape Barclay area, garnetiferous rocks are fairly common and quantitative estimates of metamorphic pressure and temperature may be obtained, although unaltered orthopyroxene coexisting with garnet is rare. Two rocks from the granulite belt in Cape Barclay area yield data on metamorphic conditions in the late Archean.

The first is a garnet-orthopyroxene-biotite-two feldspar-quartz gneiss. Garnet occurs sporadically in this rock and may be out of equilibrium with the orthopyroxene. Compositions of orthopyroxene and garnet, widely separated in the thin section, and plagioclase (Appendix, Tables 1, 8, sample FS84-43B) indicate a pressure of 0.53 GPa at an assumed temperature of 750°C, according to the geobarometer recommended by Fonarev et al. (1991) in their geothermobarometry program TPF. Temperatures derived from several versions of the garnet-orthopyroxene geothermometer, at 0.5 GPa, range from 540° to 570°C, which seem too low for a granulite-facies assemblage and further suggest that these minerals are not in equilibrium. Application of Kleemann and Reinhardt's (1994) garnet-biotite geothermometer gives a temperature of 606°C at 0.5 GPa, also low and probably indicative of cooling following the metamorphic climax.

The other rock, sample FS84-83C from a site 3 km south of sample site FS84-43B, is sillimanite-garnet-biotite-cordierite-K-feldspar-quartz gneiss. The gneiss is highly strained — practically mylonitic — and rich in cordierite. Much of the sillimanite, which occurs as tiny needles and prisms, fringes porphyroblastic garnet that is intergrown with quartz. Most of the biotite is also associated with garnet, which it appears to replace. Metamorphic conditions deduced from garnet-cordierite-sillimanite-quartz equilibria (Fonarev et al., 1991) are 0.45 GPa, from the reaction

$$\text{2 pyrope + 4 sillimanite + 5 quartz} \leftrightharpoons \text{3 Mg-cordierite}$$

and 766°C, from the reaction

$$\text{2 pyrope + 3 Fe-cordierite} \leftrightharpoons \text{2 almandine + 3 Mg-cordierite.}$$

Garnet-biotite temperatures are, as might be expected from the paragenesis of the two minerals, markedly lower: 555°C (recommended value at 0.4 GPa of Fonarev et al. (1991)) and 519°C at 0.4 GPa (Kleemann and Reinhardt, 1994).

A third rock analyzed, sample FS84-73, comes from the eastern part of the easterly trending granulite facies belt in Ian Calder Lake map area. The rock is a garnet-biotite-two feldspar-quartz gneiss, with antiperthitic plagioclase $An_{32–34}Ab_{64–66}Or_2$. Garnet porphyroblasts are strongly embayed and locally intergrown with biotite. The biotite is chestnut brown and much of it is closely associated with (?replacing) garnet.

Garnet rims are Mg and Fe richer and Ca poorer than cores (Appendix, Table 1). Biotite shows some intergrain variability in iron content (Appendix, Table 2). Temperatures obtained with Kleemann and Reinhardt's (1994) geothermometer range from 683°C (using garnet core composition) to 652°C (garnet rim) at 0.4 GPa. Although the core-to-rim compositional variation in the garnet resembles that found in prograde growth zoning of medium-grade garnet (Yardley, 1989), the garnet-biotite temperatures indicate retrograde conditions more typical of cooling following high-grade metamorphism.

In summary, limited thermobarometric data suggest low to medium paleopressures, 0.4 to 0.5 GPa, in the granulite-facies terranes of the study area. The highest temperatures recorded approach 770°C but most values are 600°C or less, doubtless indicative of retrograde re-equilibration following the metamorphic climax in the late Archean.

Turning now to the amphibolite-facies supracrustal rocks, we have analytical data only from minerals of the Barclay and Chantrey belts in Cape Barclay map area. Judged from the mineral assemblages, however, metamorphic grade in the Franklin belt and certainly the older metasedimentary rocks of the Montresor belt is comparable to that in the Barclay and Chantrey belts. The metamorphic grade of the Montresor Group is not readily apparent due to a scarcity of pelitic lithologies but the widespread fibrolite in the orthoquartzite unit *A*Mo indicates amphibolite facies.

In the Barclay belt, the widespread occurrence in metapelitic rocks of andalusite and cordierite porphyroblasts and subordinate fibrolite characterizes this belt as belonging to a low-pressure, low-temperature (Buchan) facies series (Miyashiro, 1994). Also notable is the coexistence in some rocks of cordierite and almandine-rich garnet, normally found together only in medium- to high-pressure regional metamorphism (Miyashiro, 1994). The association is, however, stable, albeit uncommon, in contact and low-pressure regional metamorphism (Miyashiro, 1973; Pattison and Tracy, 1991).

Cordierite is present in quartz-rich semipelitic, as well as micaceous pelitic, schist; in the latter it commonly coexists with andalusite. It is invariably altered and commonly has been completely replaced by sericite or similar fine-grained phyllosilicate minerals ("pinite"), whereas andalusite porphyroblasts, although sieved, are fresh. The scarcity or absence of primary chlorite and muscovite in rocks with andalusite and cordierite suggests operation of the univariant reaction

$$\text{chlorite+muscovite+quartz} \rightleftharpoons \text{cordierite+} \\ \text{aluminosilicate+biotite+}H_2O$$

(Fig. 31, curve 1) or separate reactions involving these phases to produce cordierite and andalusite in parallel.

Another important reaction boundary that was crossed is that represented by the replacement of staurolite by andalusite, exemplified by inclusions of the former in the latter:

$$\text{staurolite+muscovite+quartz} \rightleftharpoons \text{aluminosilicate+} \\ \text{biotite+}H_2O.$$

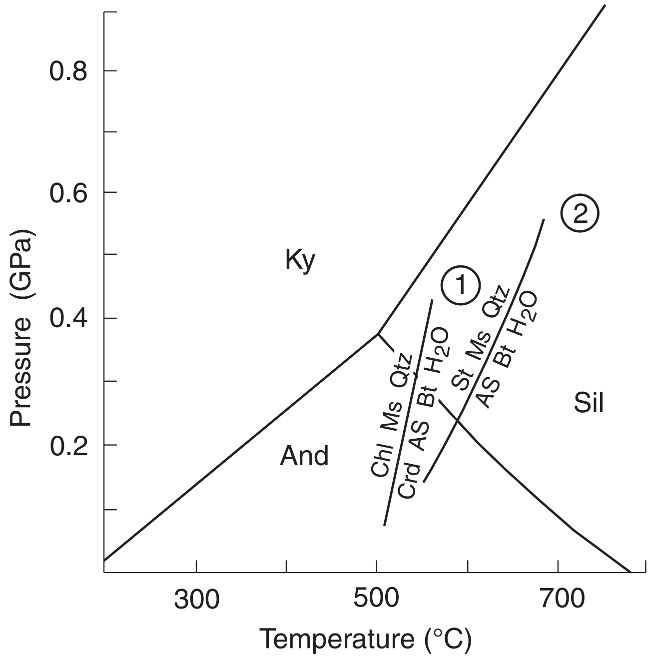


Figure 31. *Pressure-temperature diagram showing the stability fields of the aluminium silicate minerals (*after *Holdaway and Mukhopadhyay, 1993) and two reaction boundaries pertinent to metamorphism of the Barclay and Chantrey groups: curve 1 (Hirschberg and Winkler, 1968) and curve 2 (Hoschek, 1969). Ky=kyanite, Sil=sillimanite, And=andalusite, Chl=chlorite, Ms=muscovite, Qtz=quartz, Crd=cordierite, AS=aluminosilicate, Bt=biotite, St=staurolite.*

This reaction is divariant and its position depends on bulk rock composition. The experimentally determined boundary shown in Figure 31 (curve 2) is for a composition with Mg/Mg+Fe=0.4 and, because reversals were not achieved, is a maximum stability limit for the reactants (Hoschek, 1969).

Clearly, the pressure-temperature conditions of metamorphism of the Barclay belt were bounded largely by the limits of the andalusite stability field. However, the presence of significant amounts of fibrolitic sillimanite in a number of rocks suggests that the sillimanite stability field was entered locally. The aluminosilicate stability relations followed here are those determined by Holdaway (1971) and refined by Holdaway and Mukhopadhyay (1993) (Fig. 31), work which has gained wide acceptance among metamorphic petrologists (e.g. Spear, 1993; Kretz, 1994). Maximum metamorphic pressure in the Barclay belt therefore could not have much exceeded 0.3 GPa and a value of 0.25 GPa was chosen for temperature estimates.

Two analyzed samples from the Barclay belt, FS84-39B and FS84-62B (Appendix, Table 1) have the requisite mineral assemblage for application of the garnet-aluminosilicate-plagioclase-quartz geobarometer based on the reaction

$$\text{3 anorthite} \rightleftharpoons \text{grossular + 2 aluminosilicate + quartz}$$

(Newton and Haselton, 1981). Because this geobarometer has been calibrated experimentally for high P-T assemblages containing kyanite or sillimanite and more calcic garnet, Reymer et al. (1984) calculated a modification appropriate to a low-pressure regime in which andalusite is the stable aluminosilicate polymorph. Application of Reymer et al.'s (1984) formulation, however, gives pressures that generally seem too high, particularly for garnet core compositions (0.40 and 0.56 GPa; Appendix, Table 1). Use of the garnet rim composition in sample FS84-39B yields a reasonable result of 0.29 GPa. In addition to problems associated with theoretical calculations of the equilibria, garnet cores may not have formed at the same time as andalusite and perhaps record different metamorphic conditions.

Marginally better, but similarly inconsistent, results were obtained using geothermobarometers based on the biotite solution models of Patiño Douce et al. (1993). Simultaneously derived temperatures and pressures of the garnet-biotite assemblage in sample FS84-39B are 598°C, 0.39 GPa (garnet core composition) and 574°C, 0.41 GPa (garnet rim) and, in sample FS84-62B, are 513°C, 0.27 GPa (garnet core) and 537°C, 0.30 GPa (garnet rim). Use of the cordierite-biotite geothermobarometer on sample FS84-39B, however, gave values that are certainly too high: 820°C and 0.67 GPa.

Quantitative estimates of metamorphic temperature in the Barclay belt are derived almost exclusively from the biotite-garnet thermometer of Kleemann and Reinhardt (1994). At a pressure of 0.25 GPa, temperatures range from 510° to 564°C (Appendix, Table 1). For sample FS84-62B, in which garnet zoning follows the classic prograde pattern (from core to rim, increasing Fe and Mg, decreasing Ca and Mn), the rim temperature, 524°C, is indeed slightly higher than the core temperature, 510°C. The garnet core-to-rim temperature trend is the opposite in the cordierite-bearing sample FS84-39B, 556° to 540°C, but compositional zoning is slight and Mg is actually lower in the rim than in the core. Garnet-cordierite temperatures recommended by Fonarev et al. (1991) agree satisfactorily with the garnet-biotite temperatures: 554°C (core), 536° (rim). This suggests the three minerals formed in equilibrium. Considering that reaction curve 2 in Figure 31 marks the maximum stability limit for andalusite+biotite, the geothermometry-derived temperatures appear reasonable.

Mineral assemblages indicate that metamorphic conditions in the Chantrey belt were broadly similar to those in the Barclay belt. In the Chantrey rocks, the chief (practically sole) aluminosilicate is andalusite; cordierite (invariably altered) occurs sporadically; and, at the western end of the belt, fibrolite has been found and andalusite has locally replaced staurolite.

Two samples from the Chantrey belt have the mineral assemblage required for application of the garnet-biotite geothermobarometer of Patiño Douce et al. (1993) but neither yields wholly satisfactory results. Sample FS84-A85 (Appendix, Table 1) is a graphitic andalusite-garnet-biotite-muscovite-quartz schist and gives a temperature of 446°C and a pressure of 0.14 GPa; the pressure value appears to be excessively low. Sample FS84-14A (Appendix, Table 1) is a fibrolite-garnet-muscovite-biotite-quartz schist with secondary chlorite. The fibrolite is closely associated with the chestnut-brown biotite and the garnet is slightly zoned (the major change from core to rim is a 28% decrease in Mg). Indicated temperatures and pressures are 628°C and 0.14 GPa for garnet cores, and 527°C and 0.18 GPa for garnet rims. Again, the pressure values are too low. For the Chantrey belt, then, as for the Barclay belt, a metamorphic pressure of 0.25 GPa is assumed.

Metamorphic temperature in the Chantrey belt, estimated using the garnet-biotite geothermometer of Kleemann and Reinhardt (1994), ranges from 468° to 510°C in fibrolite-free rocks; garnet core and rim temperatures in the fibrolite-bearing rock sample FS84-14A are 572° and 522°C, respectively (Appendix, Table 1). In two other rocks, samples FS84-59A and FS84-A142 (Appendix, Table 1), both of them garnet-biotite-muscovite schist in which garnet shows prograde zoning, temperatures for garnet rims are, appropriately, higher than those for cores. Except for the fibrolite-bearing rock, paleotemperatures in the Chantrey belt were lower than in the Barclay belt, in accord with the wider distribution of sillimanite in the latter.

The Chantrey and Montresor groups were metamorphosed in the Aphebian, probably as a result of the intrusion at depth of granite whose differentiates are the pegmatite bodies now exposed. Judged by textural features such as large, undeformed andalusite and biotite porphyroblasts, metamorphism of the Chantrey Group outlasted, if not followed, the major deformation. Metamorphism in the Barclay belt, of the same type as in the Chantrey belt but higher temperature, presumably occurred in the Archean, as the belt is intruded by Archean granodiorite. The Barclay belt rocks may have been metamorphosed again in the Aphebian, giving rise to disequilibrium among minerals which might explain their commonly complex textures and cause problems in estimating paleopressures. Other possible evidence of a later metamorphism in the Barclay belt is the presence of coarse, undeformed andalusite and garnet. Metamorphism of the Franklin belt and the older metasedimentary rocks of the Montresor belt probably first occurred in the Archean.

STRUCTURAL GEOLOGY

The study area, which lies roughly astride Chantrey Inlet and is underlain entirely by Precambrian crystalline bedrock, has the northeasterly 'grain' typical of the Canadian Shield northwest of the mouth of Hudson Bay. The gross structures are the Archean granitoid complex, several narrow supracrustal belts, a granulite belt, a crustal-scale mylonite zone, and possibly a short segment of the boundary of the Queen Maud block, all trending mainly northeasterly. Greatly subordinate structural features are northwest- and north-striking faults and diabase dykes. Some of the major structural elements of the region are shown in Figure 5.

Little work was done on the structure of the granitoid complex, which, for the most part, is not amenable to structural analysis. At least some of the gneisses display evidence of a complex history (Fig. 10), presumably confined largely to the Archean and requiring a focused study to unravel. Because of its prevalence over a vast area of the Archean craton northwest of Hudson Bay, the northeasterly structural trend in the granitoid terrane of the study area is considered to have originated in the Archean. Later tectonism took advantage of pre-existing zones of weakness, so many Proterozoic structures similarly trend northeasterly.

In Ian Calder Lake map area, a central body mainly of augen gneiss (unit **A**k) is bordered to the northwest and southeast by more heterogeneous, layered gneiss. The northwestern part of the augen gneiss, rocks to the north (units **A**b and **A**s), and the Franklin belt are characterized by high strain and, in part, a distinctly linear aeromagnetic pattern. This region of Ian Calder map area falls within the Slave–Chantrey mylonite zone of Heywood and Schau (1978), who deemed the zone to be up to 50 km wide and to separate the "deep level" Queen Maud block from the higher level Committee Bay block to the east (Fig. 5). In contrast, Hoffman (1989), to designate the eastern border of the Queen Maud "uplift", included the edge of the straight-gneiss belt (unit **A**s) in a "Chantrey fault zone". Hoffman (1989, Fig. 22) showed the fault as a reverse type with dextral movement. Because no consistent shear sense was observed, the dextral nature of the Chantrey fault could not be confirmed during the present study but Tella (1994) reported dextral slip along the fault (which he termed "Chantrey mylonite zone") in Pelly Lake and Deep Rose Lake map areas to the south. However, from the present work, units **A**s and **A**b may be equivalent and apparently were not formed at different crustal levels and unit **A**s, instead of belonging to the Queen Maud block, may simply be unit **A**b deformed in the Slave–Chantrey mylonite zone, in which case the edge of the Queen Maud block may lie farther west. Although northeast-trending shear-mylonite zones are common over much of northern Ian Calder Lake map area, the Slave–Chantrey mylonite zone is here restricted to the boundary zone of the straight-gneiss unit, i.e. it essentially coincides with Hoffman's (1989) Chantrey fault zone. East of Chantrey Inlet, Heywood and Schau (1978) showed the Slave–Chantrey mylonite zone as a broad zone running parallel to, and just north of, the Murchison fault and suggested that the Chantrey Group might be affected by movements in the zone. The present work has shown that evidence of major shear deformation east of Chantrey Inlet is confined essentially to the Barclay and Chantrey belts and that neither belt marks a change in granitoid complex lithology or structure.

In southeastern Ian Calder Lake map area, the northeast-trending Montresor supracrustal belt separates the central augen gneiss terrane from tonalitic-granodioritic gneisses. This boundary — well defined aeromagnetically (Geological Survey of Canada, 1977a) — may imply a significant tectonic break and the Montresor belt may be yet another example of a sedimentary sequence deposited in a fault zone in this part of Nunavut, additional to those listed by Heywood and Schau (1978). Forty kilometres to the southwest, the trace of the Amer mylonite zone skirts the southeast corner of the map area (Tella, 1994) and the anorthosite described under unit **A**t may represent xenolithic rock brought up from depth.

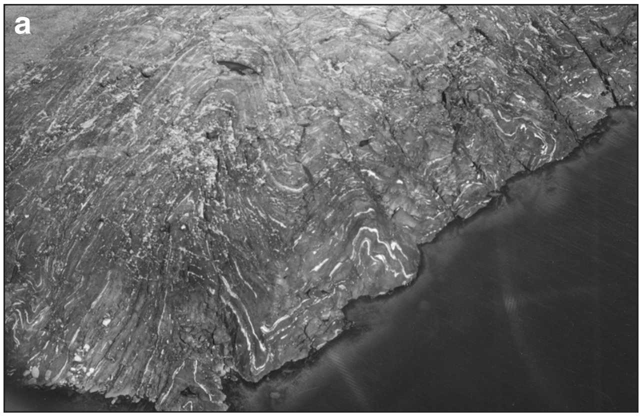
The rocks of units **A**o (the granulite belt) and **A**k in northeastern Ian Calder Lake map area form the western end of a belt aeromagnetically prominent in Mistake River map area (Geological Survey of Canada, 1977d) to the east. At its western end the belt is terminated by an inferred north-south fault. East of approximately 95°W longitude, the belt, according to the aeromagnetic pattern, swings northeasterly and part of it is exposed as a zone of granulite-facies rock (unit **A**g) in southeastern Cape Barclay map area. The granulite belt presumably is an uplifted terrane in tectonic contact with the amphibolite-grade granitoid complex.

The Franklin belt, exposed west and east of Franklin Lake, occurs in a major northeast-trending fault zone, named the Murchison fault by Heywood and Schau (1978). The Franklin belt is the most deformed (in terms of strain) of all the supracrustal sequences in the study area. West of the lake, the various lithological units of the belt are poorly exposed in linear arrays of outcrops and are interleaved with granitoid rock. The rocks dip predominantly southeast at 50–60° and all are sheared to mylonitized; lithological contacts are all probably tectonic. Stretched clasts in metaconglomerate are cigar shaped parallel to strike, indicating that movement was primarily strike-slip. Farther west along strike, only remnants of the supracrustal rocks (mainly metaconglomerate) are preserved in isolated outcrops surrounded by granitoid rock. On the east shore of Franklin Lake, the Franklin belt is represented only by metaconglomerate and subordinate schist underlying an area 4 km by 1.3 km and dipping consistently northwest at 50–65°. Eastward along strike lenses of quartzite and schist occur as tectonic inclusions in gneiss. The deformation of the Franklin belt probably occurred principally in the Archean, perhaps contemporaneously with movements in the Chantrey fault zone.

Supracrustal rocks interpreted to be correlative with those of the Franklin belt constitute the Barclay belt extending northeastward from the eastern shore of Chantrey Inlet. These also are highly deformed and broken up by granitoid rocks, as well as being apparently intruded by Archean granodiorite. Attitudes of minor folds indicate that the rocks have been folded tightly to isoclinally with steep plunges (about 60–65°) to the northeast. Rare, more open crossfolds plunging both east and west at 35° overprint the earlier structures.

As Heywood and Schau (1978) pointed out, the Chantrey belt is associated with the Murchison fault zone, an apparently long-lived structure. The Archean Franklin belt rocks at the southern end of the fault were presumably deposited in some kind of (?linear) depression presently on strike with the Aphebian Chantrey belt, which is similarly interpreted as a restricted-basin fill. Assuming they have not been transported from their original sites of deposition, the Franklin and Chantrey belts outline a long, linear zone of weakness in the crust that was initiated and subsequently deformed in the Archean, was probably reactivated, widened, and deformed again in the Aphebian and is now marked by the Murchison fault zone. If the main deformation of the Franklin belt indeed took place in a shear zone in the Archean, reactivation of such a zone would promote formation of a rift basin (Daly et al., 1989).

The Chantrey Group has been deformed into a canoe-shaped syncline, 5 km wide and 175 km long, plunging 20° at each end. At least two, probably closely related folding events occurred, the more pervasive resulting in early, tight, more or less upright folds with northeasterly trending axial planes (Fig. 32a). Local overturning of strata at the southern margin



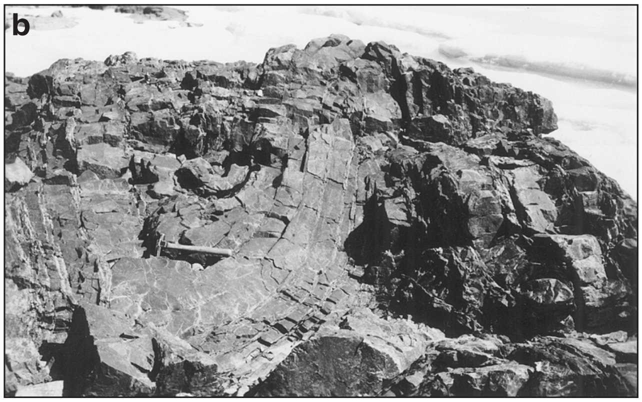


Figure 32. *Folding in the Chantrey Group.* ***a****) Aerial view of early, gently northeast-plunging folds in calcareous metasiltstone/mudstone of unit* ACs *at the northern border of the Chantrey belt, about 8 km east of Irby and Mangles Bay. About 400 m of shoreline are shown. Photograph by T. Frisch. GSC 204635-M* ***b****) Early folds plunging 35° to the northeast, away from the viewer, in unit* ACo *on small island in Chantrey Inlet, west-northwest of Victoria Headland. Photograph by T. Frisch. GSC 204072-T* ***c****) Recumbent fold with northeast-trending axis in pelitic-psammitic schist of unit* ACo *on small island near the western end of the Chantrey belt (*see *Frontispiece). Strata of lower limb at shoreline are right side up; upper limb is overturned. View northeastward along the strike of the belt. Photograph by T. Frisch. GSC 204635-Y* ***d****) Lower limb of a recumbent fold, closing to left, in marble (unit* ACm*) near the western end of the Chantrey belt. Photograph by T. Frisch. GSC 204635-V* ***e****) Aerial view of late, steeply northwest-plunging crossfolds in unit* ACm *at the northern margin of the Chantrey belt, east of Chantrey Inlet at about 95°5′W. Photograph by T. Frisch. GSC 204072-Z*

of the Chantrey belt, due to northwestward thrusting of the Archean granitoid complex, regarded as basement to the Cantrey Group, renders at least part of the belt asymmetric. At the Chantrey Group–basement contact, bedding in the Chantrey Group and foliation in the gneiss are invariably concordant. Clast elongation in metaconglomerate, while not as extreme as in the Franklin belt, indicates considerable shear locally, at least, along the basement contact. In fact, the Chantrey Group–basement contact is regarded as being everywhere tectonic, although not necessarily with significant associated movement.

Although the gross structure of the belt is relatively simple, the degree of deformation varies along its length. At the extreme western end, early folds plunge 20°NE but on the small islands in Chantrey Inlet proper they steepen to 35–40° (Fig. 32b). Between these two localities, recumbent folds with northeast-trending axes have been superimposed on the early upright folds, probably as a result of particularly severe overthrusting of basement (Fig. 32c, d).

East of Chantrey Inlet, in the more central parts of the Chantrey belt, deformation appears to have been less severe. Clasts in the metaconglomerate unit *A*Cc are predominantly more or less equant; only near the contact with basement are clasts elongate parallel to strike (length:width up to 8:1 but generally much less). Although part of the southern contact of the belt is a high-angle reverse fault like that at the western end of the belt, nappe-style folds were not recognized. Relatively open crossfolds plunging steeply (>50°) northwest occur in the marble unit *A*Cm at the northern contact at about 95°5′W (Fig. 32e). Between 94°30′ and 94°45′W, the units constituting the Chantrey belt are symmetrically distributed about a northeast-trending synclinal axis.

Still farther east along the belt, outcrop is only moderately good to poor, particularly in the area of the boundary between Cape Barclay and Darby Lake map areas. East of the major west-northwest-trending fault that cuts the Chantrey belt in Darby Lake map area (*see* below), the trend of the belt swings to east-west away from the Murchison fault as defined by Heywood and Schau (1978). The reorientation of the Chantrey belt may be due to movement along an inferred fault running due east from Sherman Inlet and here named the Sherman–Chantrey fault (Fig. 5). This fault is well expressed aeromagnetically in and east of Sherman Inlet (Geological Survey of Canada, 1977e), where aeromagnetic anomalies are dextrally offset by up to 15 km. The west-northwest-trending fault along which the Chantrey belt is dextrally offset in Darby Lake map area may be related to the Sherman–Chantrey fault. Structures become very complex at the eastern end of the Chantrey belt. Much of the southern contact there is a south-dipping high-angle reverse fault and the tightly folded Chantrey strata wrap around a major fold closure (Fig. 4). As at the western end of the belt, repetition of lithological units may be due to thrusting following folding. Further work is needed to decipher the structural history of the Chantrey belt.

In contrast to the structural complexity of the Chantrey belt, the possibly coeval Montresor belt in Ian Calder Lake map area has a simple structure. The Montresor belt, 65 km long in the map area and extending a further 12 km into Deep Rose Lake map area, has the form of a northeast-trending, doubly plunging, open syncline. Gentle warping has caused the synclinal axis to plunge alternately southwest and northeast along its length; the axis plunges 15°SW at its eastern end. Steeper dips of bedding at the northern contact of the belt (40–80°) than at the southern (25–40°) lend a slight asymmetry to the syncline. Scattered exposures of the Montresor Group and an aeromagnetic anomaly outline an open synclinal structure, up to 8 km wide, west of the main Montresor belt. Although there is evidence of a depositional contact of Montresor Group rocks with granitoid basement at the western end of the belt, elsewhere the contact probably is generally faulted, if only as a result of relatively minor slippage between basement and supracrustal rock. Unlike the Chantrey Group–basement contact, that between Montresor Group and basement is nowhere overturned and lacks evidence of thrusting, with the possible exception of the fault block of northwest-dipping carbonate and quartzitic rock northeast of Mount Meadowbank. If the stratigraphic repetition here is due to thrusting, the movement direction (southeast) would be opposite to any known from supracrustal belts in the study area and neighbouring areas, including the correlative Amer belt (Patterson, 1986; Tella, 1994). Also, bedding in the Montresor Group and foliation in the basement are commonly discordant. Similarly discordant is the structural relationship between the Montresor Group and the more highly deformed and metamorphosed rocks of unit **A**m, implying that the latter unconformably underlie the Montresor Group.

Based on orientation, late faults and lineaments fall mainly in two groups: northwesterly and northerly. In Ian Calder Lake map area, a major northwest-trending lineament is the site of Franklin Lake and several Mackenzie diabase dykes and was named the Sherman Inlet fault by Heywood and Schau (1978). At its northern end, the fault displaces lower Paleozoic strata east of Sherman Inlet mainly by dip-slip movement (Heywood and Schau, 1978). In the study area, the fault appears to truncate part of the Franklin belt, as only metaconglomerate and minor schist occur east of Franklin Lake. If these rocks are basal in the Franklin belt, reverse (east side up) movement is indicated. Similarly, dip-slip rather than strike-slip movement is inferred to have occurred along strike of the fault to the northwest and southeast, where aeromagnetic lineaments and anomalies are not noticeably offset. Interestingly, a magnitude 6.0 earthquake whose epicentre was 66.6°N, 94.8°W was recorded early in 1992 (Bent and Cassidy, 1993). The epicentre lies approximately in the centre of Mistake River map area, roughly along the strike of the Sherman Inlet fault. Bent and Cassidy (1993) concluded that the earthquake was a consequence of both reverse and strike-slip movements on a northwest-striking plane.

Northerly trending faults for the most part are minor features recognized in all four supracrustal belts of the study area. In Ian Calder Lake map area, however, a major north-south fault runs south from the Franklin belt and probably truncates the granulite-grade gneiss units near the eastern border.

ECONOMIC GEOLOGY

Two subjects of potential economic interest are the Montresor Group–granitoid complex contact at the western end of the Montresor belt and the pyritiferous metapelite unit *A*Cp of the Chantrey Group.

Near the southern border of Ian Calder Lake map area, at the northern margin of the Montresor belt, rusty-brown-weathering biotite schist with rounded quartz and granite clasts and veined by quartz rests on garnet-muscovite leucogranite (unit **A**mg) that bears traces of arsenopyrite. Overlying the biotite schist is grey quartzite with long lenses and layers, commonly 0.6 m thick, of very fine-grained dark schist containing trace native copper. Less than 50 m away eastward along strike, the Montresor Group–basement contact is heavily silicified. Massive, glassy white quartzite, sheathed in highly sheared, rusty-weathering, pyritiferous quartzite, forms a 50 m long, 30 m wide body separating the granitic rocks and pink and grey quartzites of the Montresor Group.

As far as is known, the mineral showings are confined to that part of the granitoid complex consisting of garnet-muscovite granite (unit **A**mg). This leucocratic, peraluminous granite may represent a metamorphosed weathered zone or regolith developed on basement.

A second suggested exploration target is the commonly rusty-weathering and gossanous, dark pelitic schist of unit *A*Cp of the Chantrey Group. The sedimentary protolith is interpreted to have been derived from granitic and gneissic rocks and deposited in shallow water under oxygen-deficient, reducing conditions favourable to the precipitation of metallic sulphides.

In the Zambian Copperbelt, a world-class copper deposit in central Africa, the ore is hosted principally by the 'ore shale', a carbonaceous, sulphide-rich, euxinic sediment and also by siltstone, arkose, and quartzite, in which conditions immediately below the water-sediment interface during sedimentation were anoxic and reducing (Guilbert and Park, 1986). All the sediments were deposited in late Precambrian time along an ancient shoreline constructed on Archean granitic basement of low relief, which was the source of the ore (Guilbert and Park, 1986). Tectonism and metamorphism (greenschist grade) did not cause significant remobilization of the sulphide ore (Sweeney et al., 1991).

In the writer's opinion, the relevance of the Copperbelt model to the Chantrey belt warrants a closer analysis of the economic potential of the Chantrey rocks. Although, admittedly, the metapelite of unit *A*Cp in general show no anomalous metal contents (except iron), two samples, FS84-60 and FS84-G65D are markedly enriched in one or more of the elements Fe, Co, Cu, V, and Zn (Appendix, Table 10).

Brief work of a prospective nature was done in 1988 on pegmatite and associated rocks in the Barclay and Chantrey belts in Chantrey Inlet (Černý et al., 1988, 1989). In the Barclay belt, Černý and his co-workers identified beryl as a common constituent of pegmatite and found scattered occurrences of rare-element minerals, such as columbite and tantalite, and phosphates, such as triphylite, in other pegmatite bodies at the west end of the belt (Černý et al., 1988, 1989). In a cursory examination of exposures of the Chantrey Group off Victoria Headland, Černý et al. (l988) found the pegmatites barren. No economic concentrations of ore minerals in pegmatite were found anywhere but a potential for beryl and Li-aluminosilicate minerals was identified in the Barclay belt, at least (Černý et al., 1989). Attention was also drawn to the abundance of garnet and the chiastolite variety of andalusite in some units of the Barclay and Chantrey belts for possible use as souvenir material (Černý et al., 1989; Gebert, 1993).

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APPENDIX

Mineral and rock analyses

Table 1. Microprobe analyses of garnet.

Sample	Barclay belt						Chantrey belt										Granitoid complex			
	38C	39B		39C	62B		14A		59A		86A	86B	87B	A85	A142		43B	73		83C
		C	R		C	R	C	R	C	R					C	R		C	R	
SiO_2	37.30	36.08	36.13	37.48	37.31	37.11	36.52	36.64	37.38	36.95	35.58	36.10	36.91	37.19	36.57	36.27	37.21	37.28	36.92	38.23
TiO_2	0.10	0.04	0.02	0.04	0.05	0.04	0.03	0.02	0.06	0.04	0.05	0.06	0.05	0.05	0.09	0.08	0.05	0.08	0.07	0.03
Al_2O_3	20.30	19.87	19.76	21.19	20.78	20.85	20.78	21.18	20.61	20.53	19.80	19.99	20.77	20.74	20.72	20.51	20.40	20.60	20.66	21.84
Cr_2O_3	0.02	0.03	0.02	0.03	0.03	0.03	0.01	0.03	0.06	0.04	0.07	0.04	0.05	0.04	0.01	0.07	0.07	0.04	0.02	0.09
FeO_t	35.18	37.87	38.16	29.52	38.34	39.65	40.60	41.18	34.06	37.33	43.21	44.00	41.23	31.35	30.49	32.06	32.55	29.04	33.13	32.95
MnO	1.00	0.44	0.39	8.38	0.50	0.08	1.12	0.96	6.77	3.24	0.01	0.03	0.02	10.04	9.79	7.34	2.77	0.83	0.84	0.53
MgO	0.85	2.31	2.16	2.87	1.25	1.50	1.49	1.08	0.95	1.37	-	-	1.27	1.26	1.22	1.50	2.75	4.32	5.18	6.93
CaO	5.04	1.64	1.30	2.31	3.01	1.79	0.26	0.25	1.86	1.30	0.56	0.46	0.71	0.94	1.97	1.89	3.88	7.04	2.32	0.75
Total	99.79	98.28	97.94	101.82	101.27	101.05	100.81	101.34	101.75	100.87	99.28	100.68	101.01	101.61	100.86	99.72	99.68	99.23	99.14	101.35
Fe_2O_3(calc)	0.84	1.19	1.00	0.72	0.41	0.27	0.51	-	0.64	0.56	1.01	1.13	0.21	0.48	0.75	0.66	0.87	1.50	1.33	0.42
Cations per 12 O																				
Si	3.026	2.984	3.001	2.972	3.001	2.995	2.972	2.972	3.006	2.998	2.976	2.979	2.994	2.996	2.966	2.970	3.001	2.969	2.961	2.969
Ti	0.006	0.002	0.001	0.002	0.003	0.002	0.002	0.001	0.004	0.002	0.003	0.004	0.003	0.003	0.005	0.005	0.003	0.005	0.004	0.002
Al	1.941	1.938	1.935	1.981	1.970	1.984	1.994	2.026	1.954	1.963	1.953	1.945	1.987	1.970	1.982	1.980	1.939	1.934	1.953	1.999
Cr	0.001	0.002	0.001	0.002	0.002	0.002	0.001	0.002	0.004	0.003	0.005	0.003	0.003	0.003	0.001	0.005	0.004	0.003	0.001	0.006
Fe^{3+}	0.051	0.074	0.063	0.043	0.025	0.016	0.031	-	0.039	0.034	0.063	0.070	0.013	0.029	0.046	0.041	0.053	0.090	0.080	0.024
Fe^{2+}	2.336	2.546	2.588	1.915	2.554	2.660	2.732	2.794	2.252	2.499	2.959	2.966	2.785	2.083	2.023	2.155	2.142	1.844	2.142	2.115
Mn	0.069	0.031	0.027	0.563	0.034	0.005	0.077	0.066	0.461	0.223	0.001	0.002	0.001	0.685	0.673	0.509	0.189	0.056	0.057	0.035
Mg	0.103	0.285	0.267	0.339	0.150	0.180	0.181	0.131	0.114	0.166	-	-	0.154	0.151	0.147	0.183	0.331	0.513	0.619	0.802
Ca	0.438	0.145	0.116	0.196	0.259	0.155	0.023	0.022	0.160	0.113	0.050	0.041	0.062	0.081	0.171	0.166	0.335	0.601	0.199	0.062
Σ	7.971	8.007	7.999	8.013	7.998	8.001	8.013	8.013	7.993	8.000	8.010	8.009	8.001	8.001	8.014	8.013	7.997	8.013	8.017	8.015
Fe/Fe+Mg	0.96	0.90	0.91	0.85	0.95	0.94	0.94	0.96	0.95	0.94	1.00	1.00	0.95	0.93	0.93	0.92	0.87	0.78	0.78	0.72
Pyrope	3.49	9.53	8.92	11.40	5.00	6.02	6.08	4.39	3.81	5.52	-	-	5.12	5.04	4.96	6.15	11.03	17.24	20.88	27.00
Almandine	79.30	85.24	86.30	64.38	85.26	88.74	90.96	93.96	75.39	83.29	96.64	96.27	92.91	69.47	68.06	72.43	71.51	62.02	72.24	71.21
Grossular	12.20	0.27	0.65	2.92	7.32	4.14	-	-	3.22	1.82	-	-	1.02	0.92	1.72	1.77	8.31	15.29	2.39	-
Spessartine	2.33	1.03	0.92	18.92	1.08	0.07	1.29	1.49	15.44	7.42	-	-	-	22.84	22.63	17.11	6.26	0.54	0.16	0.18
Uvarovite	0.07	0.10	0.07	0.09	0.10	0.10	-	0.10	0.19	0.13	-	-	0.16	0.13	0.03	0.23	0.22	0.13	0.06	0.28
Andradite	2.61	3.71	3.14	2.17	1.24	0.81	0.67	-	1.95	1.70	1.53	1.18	0.64	1.45	2.32	2.06	2.66	4.54	4.06	1.24
Schorlomite	-	0.12	-	0.12	-	0.12	0.09	0.06	-	0.12	0.16	0.19	0.15	0.15	0.28	0.25	-	0.24	0.21	0.09
Skiagite	-	-	-	-	-	-	0.91	-	-	-	1.67	2.36	-	-	-	-	-	-	-	-
T1		556	540	564	510	524	572	522	472	510			476	472	468	492	606	683	652	519*
T2		554	536																	766
T3																	540-570			
T4		598	574		513	537	628	527						446						
P1		0.40	0.29		0.56	0.37														
P2																	0.53			
P3																				0.45
P4		0.39	0.41		0.27	0.30	0.14	0.18						0.14						

Assemblages (all samples numbers prefixed by FS84-)

38C Cum-Hbl-Qtz-Grt-Ox
39B Qtz-Pl($An_{23\text{-}26}$)-Chl-And-Crd-Bt-St-Grt
39C Qtz-Bt-Pl(An_{48})-Grt
62B Qtz-Ms-Bt-Grt-And-Pl($An_{24\text{-}25}$)-Crd(altered)-St
14A Qtz-Grt-Bt-Fb-Ms
59A Qtz-Ms-Bt-Pl-Grt
86A Qtz-Gru-Grt
86B Qtz-Grt-Gru-Chl-Mag
87B Qtz-Bt-Grt
A85 Qtz-Bt-Ms-And-Grt-Gr
A142 Qtz-Ms-Bt-Grt-Chl
43B Qtz-Pl($An_{38\text{-}39}$)-Kfs-Bt-Opx-Grt
73 Qtz-Pl($An_{32\text{-}34}$)-Bt-Grt
83C Qtz-Crd-Grt-Bt-Sil-Kfs ($Ab_{15}Or_{85}$)

T1: Garnet-biotite temperature (°C) from geothermometer of Kleemann and Reinhardt (1994) at P=0.25 GPa for samples from Barclay and Chantrey belts, at P=0.5 GPa for sample 43B, at P=0.4 GPa for 73 and 83C.
* Recommended value of Fonarev et al. (1991) at P=0.4 GPa is 555°C.
T2: Garnet-cordierite temperature (°C) recommended by Fonarev et al. (1991) at P=0.25 GPa for sample 39B, at P=0.45 GPa for 83C.
T3: Garnet-orthopyroxene temperature (°C) at P=0.5 GPa according to several calibrations (Fonarev et al., 1991).
T4: Garnet-biotite temperature (°C) from biotite solution geothermobarometer of Patiño Douce et al. (1993).
P1: Pressure (GPa) from geobarometer of Reymer et al. (1984) at temperature T1, with X_{An}=0.245.
P2: Pressure (GPa) from garnet-orthopyroxene-plagioclase-quartz geobarometer recommended by Fonarev et al. (1991) at T=750°C.
P3: Pressure (GPa) from garnet-cordierite-sillimanite-quartz geobarometer recommended by Fonarev et al. (1991) at T=766°C.
P4: Garnet-biotite pressure (GPa) from biotite solution geothermobarometer of Patiño Douce et al. (1993).

Cum=cummingtonite
Hbl=hornblende
Qtz=quartz
Grt=garnet
Ox=Fe-Ti oxide
Pl=plagioclase
Chl=chlorite
And=andalusite
Crd=cordierite
Bt=biotite
Ms=muscovite
St=staurolite
Fb=fibrolite
Gru=grunerite
Mag=magnetite
Gr=graphite
Kfs=K-feldspar
Opx=orthopyroxene
Sil=sillimanite

Table 2. Microprobe analyses of biotite.

Sample	Barclay belt				Chantrey belt						Granitoid complex		
	39B	39C	55	62B	14A	59A	87B	A85	A142	A143	43B	73	83C
SiO_2	35.00	35.32	35.31	34.32	33.86	33.94	34.88	34.75	35.35	34.63	36.10	35.33	37.25
TiO_2	2.08	1.45	2.21	1.95	1.82	1.88	0.95	1.85	0.94	1.76	4.14	4.69	4.65
Al_2O_3	19.65	17.83	20.08	19.79	20.50	19.83	17.39	20.62	20.26	20.34	14.47	16.59	16.19
Cr_2O_3	0.06	0.10	0.06	0.05	0.02	0.05	0.04	0.04	0.01	0.06	0.17	0.06	0.15
FeO_t	21.30	19.45	21.08	23.29	25.48	23.20	25.27	21.76	21.34	21.72	21.69	17.56 (16.8-18.5)	12.09
MnO	-	0.19	0.20	-	-	0.06	-	0.23	0.11	0.18	0.08	0.01	0.01
MgO	8.18	11.28	9.00	6.94	5.06	7.10	7.97	8.84	9.66	8.31	10.23	11.11	16.04
K_2O	8.66	8.68	8.69	8.39	8.19	8.45	8.38	8.36	8.57	8.28	9.25	9.01	9.16
Total	94.93	94.30	96.63	94.73	94.93	94.51	94.88	96.45	96.24	95.28	96.13	94.36	95.54
Cations per 11 O													
Si	2.683	2.712	2.655	2.662	2.645	2.643	2.734	2.624	2.668	2.647	2.766	2.696	2.729
Ti	0.120	0.084	0.125	0.114	0.107	0.110	0.056	0.105	0.053	0.101	0.239	0.269	0.256
Al	1.776	1.614	1.780	1.810	1.888	1.820	1.607	1.835	1.803	1.833	1.307	1.493	1.398
Cr	0.004	0.006	0.004	0.003	0.001	0.003	0.002	0.002	0.001	0.004	0.010	0.004	0.009
Fe	1.366	1.249	1.326	1.511	1.665	1.511	1.656	1.374	1.347	1.388	1.390	1.121	0.741
Mn	-	0.012	0.013	-	-	0.004	-	0.015	0.007	0.012	0.005	0.001	0.001
Mg	0.935	1.291	1.009	0.802	0.589	0.824	0.931	0.995	1.086	0.946	1.168	1.264	1.751
K	0.847	0.851	0.834	0.831	0.817	0.840	0.838	0.806	0.825	0.808	0.904	0.878	0.856
X_{Fe}	0.59	0.49	0.57	0.65	0.74	0.65	0.64	0.58	0.55	0.60	0.54	0.47 (0.45-0.49)	0.30

<u>Assemblages</u> (all sample numbers prefixed by FS84-)
55 quartz-muscovite-biotite-chlorite-andalusite-plagioclase(An_{28-34})-tourmaline
A143 quartz-muscovite-biotite-chlorite-andalusite-plagioclase-graphite-pyrite
For other assemblages *see* Table 1

Table 3. Microprobe analyses of muscovite from the Barclay and Chantrey belts.

	Barclay belt	Chantrey belt		
	55	14A	59A	A85
SiO_2	45.70	45.38	45.54	46.76
TiO_2	0.65	0.45	0.56	0.57
Al_2O_3	36.98	36.67	36.80	37.05
Cr_2O_3	0.03	0.02	0.03	0.02
Fe_2O_{3t}	0.95	1.01	1.30	0.72
MnO	0.01	0.01	0.04	0.02
MgO	-	-	0.03	-
Na_2O	0.18	0.78	0.48	0.34
K_2O	10.00	8.92	9.47	9.36
Total	94.50	93.24	94.12	94.84
Cations per 11 O				
Si	3.038	3.046	3.034	3.079
Ti	0.033	0.023	0.028	0.028
Al	2.899	2.901	2.891	2.876
Cr	0.002	0.001	0.002	0.001
Fe	0.048	0.051	0.065	0.036
Mn	0.001	0.001	0.002	0.001
Mg	-	-	0.003	-
Na	0.023	0.101	0.062	0.043
K	0.849	0.764	0.805	0.787
Na/Na+K	0.03	0.12	0.07	0.05

Sample numbers prefixed by FS84-.
\- = below detection limit.

Table 4. Microprobe analyses of chlorite.

	Barclay belt	Chantrey belt		
	55	86B	A142	A143
SiO_2	25.33	21.40	22.84	23.06
TiO_2	0.07	0.13	0.05	0.14
Al_2O_3	19.83	20.56	23.30	23.31
FeO_t	29.79	46.16	26.77	27.97
MnO	0.27	0.02	0.25	0.39
MgO	13.47	1.07	13.54	12.47
CaO	0.02	0.02	0.03	0.02
K_2O	0.07	0.03	0.07	0.01
Total	88.85	89.39	86.85	87.37
Cations per 28 O				
Si	5.409	5.010	4.927	4.967
Ti	0.011	0.023	0.008	0.023
Al	4.992	5.673	5.924	5.920
Fe	5.321	9.038	4.829	5.040
Mn	0.048	0.004	0.046	0.072
Mg	4.288	0.373	4.354	4.003
Ca	0.005	0.005	0.007	0.004
K	0.018	0.009	0.019	0.003
Fe/Fe+Mg	0.55	0.96	0.52	0.56

Sample numbers prefixed by FS84-.

Table 5. Microprobe analyses of staurolite from the Barclay belt.

Sample	39B	62B
SiO_2	28.00	26.60
TiO_2	0.46	0.51
Al_2O_3	54.64	55.14
Cr_2O_3	-	0.08
FeO_t	14.84	13.39
MnO	0.02	0.03
MgO	0.87	1.13
ZnO	0.59	0.29
Total	99.42	97.17
Cations per 23 O		
Si	3.846	3.715
Ti	0.048	0.054
Al	8.846	9.078
Cr	-	0.009
Fe	1.705	1.564
Mn	0.002	0.003
Mg	0.178	0.235
Zn	0.060	0.030

Sample numbers prefixed by FS84-.
\- = below detection limit.

Table 6. Microprobe analyses of cordierite from the Barclay belt (FS84-39B) and granitoid complex (FS84-83C).

Sample	39B	83C
SiO_2	47.51	49.32
TiO_2	0.01	0.01
Al_2O_3	31.94	33.38
FeO_t	9.97	7.70
MnO	0.02	-
MgO	7.01	9.21
K_2O	0.04	0.01
Total	96.50	99.63
Cations per 18 O		
Si	5.020	4.992
Ti	0.001	0.001
Al	3.978	3.982
Fe	0.882	0.652
Mn	0.001	-
Mg	1.105	1.390
K	0.005	0.002
Fe/Fe+Mg	0.44	0.32

Sample numbers prefixed by FS84-.
\- = below detection limit

Table 7. Microprobe analyses of amphibole.

Sample	Barclay belt		Chantrey belt	
	38C		86A	86B
	Colourless	Green		
SiO_2	49.69	40.19	46.88	46.14
TiO_2	0.03	0.33	0.02	0.06
Al_2O_3	0.56	12.88	0.11	0.23
Cr_2O_3	0.04	0.07	0.01	0.04
FeO_t	34.35	24.72	48.65	48.28
MnO	0.10	0.03	0.01	0.01
MgO	9.40	4.18	0.37	1.10
CaO	0.57	11.12	0.04	0.04
Na_2O	-	0.16	-	0.03
K_2O	0.04	0.65	0.02	0.05
Total	94.78	94.33	96.11	95.98
Cations per 23 O				
Si	7.941	6.457	7.969	7.864
Ti	0.004	0.040	0.003	0.008
Al	0.106	2.441	0.022	0.046
Cr	0.005	0.009	0.001	0.005
Fe	4.591	3.322	6.917	6.881
Mn	0.014	0.004	0.001	0.001
Mg	2.239	1.001	0.094	0.279
Ca	0.098	1.914	0.007	0.007
Na	-	0.050	-	0.010
K	0.008	0.133	0.004	0.011
Fe/Fe+Mg	0.67	0.77	0.99	0.96

38C colourless: cummingtonite
green: ferro-tschermakitic hornblende
86A,B: grunerite
Sample numbers prefixed by FS84-.
\- = below detection limit.

Table 8. Microprobe analysis of orthopyroxene from the granitoid complex.

Sample	FS84-43B
SiO_2	50.99
TiO_2	0.05
Al_2O_3	0.88
Cr_2O_3	0.04
FeO_t	34.52
MnO	1.19
MgO	13.16
CaO	0.32
Na_2O	-
Total	101.15
Cations per 6 O	
Si	1.993
Ti	0.001
Al	0.041
Cr	0.001
Fe	1.129
Mn	0.039
Mg	0.767
Ca	0.01
Na	-
Ca	1
Fe	59
Mg	40
- = below detection limit.	

Table 9. Chemical analyses of ferriferous quartzites and associated metapelitic-psammitic rocks (unit *A*Co), Chantrey belt.

Sample	87A	87B	87C	87D
SiO_2	61.6	62.0	82.2	78.7
TiO_2	0.20	0.35	0.15	0.10
Al_2O_3	6.7	11.0	3.4	2.9
Cr_2O_3	0.01	0.01	0.01	0.01
Fe_2O_3	4.5	3.8	2.0	3.4
FeO	22.6	15.3	9.6	12.1
MnO	0.12	0.02	0.10	0.01
MgO	2.13	2.90	0.75	0.40
CaO	0.63	0.43	0.57	0.57
Na_2O	0.0	0.0	0.0	0.0
K_2O	0.00	1.89	0.00	0.06
H_2O	1.3	2.1	0.4	0.7
CO_2	0.2	0.1	0.2	0.9
P_2O_5	0.26	0.21	0.33	0.38
S	0.23	-	0.01	-
Less O≡S	0.12	-	-	-
Total	99.7	100.2	100.4	100.1
Ba	22	66	55	90
Nb	13	2	11	18
Rb	20	22	6	112
Sr	20	36	0	0
Y	43	41	39	46
Zr	148	95	291	369
Be	2.0	1.9	0.6	2.5
Co	3	0	11	16
Cr	66	62	21	49
Cu	6	6	13	0
La	8	11	3	17
Ni	6	1	11	20
V	100	220	230	310
Yb	2.0	2.2	1.8	1.8
Zn	7	11	57	56

Sample numbers prefixed by FS84-.
- = below detection limit.

Table 10. Chemical composition of rocks of unit *A*Cp (Chantrey Group) and Proterozoic shale units.

Sample	60	A68	A91	A109B	A130A	A213	A246	A250	A252	A254	A290	A291	G64A	G65D	G70	G76	G115A	G118	G192	1	2	3
SiO_2	43.0	66.6	62.6	67.9	60.4	60.0	49.1	68.1	62.1	67.5	62.5	63.2	70.9	66.8	59.7	62.0	65.7	65.7	62.2	63.8	63.10	65.00
TiO_2	1.62	0.87	1.10	0.61	1.30	1.07	1.54	0.73	0.94	1.28	0.92	0.89	0.69	0.72	1.12	0.58	0.34	0.95	0.82	0.86	0.64	0.76
Al_2O_3	21.5	13.4	18.1	11.2	18.4	16.7	20.7	12.3	19.3	13.7	15.8	19.9	10.8	10.9	17.6	16.2	11.7	17.3	14.1	15.4	17.50	16.22
Cr_2O_3	0.03	0.01	0.01		0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01		
Fe_2O_3	19.5*	0.8	1.0	1.9	1.3	1.3	1.9	2.0	1.8	1.8	1.4	1.5	1.1	1.3	1.7	3.6	4.0	1.0	1.6	1.8	6.28*	6.37*
FeO		8.1	5.4	9.2	5.7	7.1	9.3	7.3	6.0	6.2	7.0	5.6	8.3	7.3	7.7	2.3	11.0	5.6	9.4	7.1		
MnO	0.78	0.07	0.11	0.09	0.06	0.07	0.07	0.16	0.16	0.13	0.12	0.09	0.13	0.18	0.57	0.07	0.17	0.14	0.13	0.14		
MgO	4.57	1.99	1.85	2.09	2.20	2.62	3.64	2.46	1.89	2.23	1.82	1.80	2.01	2.56	2.65	2.53	2.23	1.86	3.33	2.32	2.20	2.51
CaO	1.44	0.43	0.34	0.82	1.32	1.72	1.18	0.59	0.35	0.32	2.48	0.33	1.62	2.11	1.34	0.74	0.12	0.74	0.42	0.94	0.71	0.52
Na_2O	0.7	0.9	0.7	1.2	1.5	3.2	1.3	0.9	0.7	0.8	2.4	0.7	0.3	0.2	1.0	0.7	0.2	1.0	0.7	1.0	1.06	1.44
K_2O	2.82	3.80	4.92	3.05	4.07	3.57	6.26	3.47	4.54	3.75	3.37	3.47	2.5	2.61	2.59	9.69	3.01	3.64	4.44	4.04	3.62	4.83
H_2O		2.2	2.7	1.7	2.1	1.8	2.9	1.9	2.3	2.0	2.3	2.0	1.6	2.5	2.3	1.6	1.6	1.9	2.4	2.1		
CO_2	1.8	0.5	1.3	0.3	5.1	3.0	3.6	0.9	1.7	1.0	1.6	1.8	0.4	3.6	6.2	0.1	0.1	1.1	0.2	1.8		
C					1.4	0.8	1.0		0.5	0.3	0.4	0.5		1.0	1.7			0.3		0.4		1.68
P_2O_5	0.29	0.05	0.09	0.15	0.08	0.07	0.13	0.07	0.06	0.08	0.05	0.12	0.26	0.20	0.02	0.8	0.03	0.11	0.08	0.10	0.12	
S	2.01	0.15		0.40	0.14	0.28	0.02	0.01			0.19	0.03	0.03	0.75	0.23			0.01	0.14	0.13		0.65
Less O≡C,S	1.00	0.08		0.20	3.80	2.28	2.68		1.34	0.8	1.16	1.35	0.02	3.04	4.65			0.80	0.07	1.14		
Total	99.1	99.8	100.2	100.4	101.3	101.0	100.0	100.9	101.0	100.3	101.2	100.6	100.6	99.7	101.8	100.5	100.2	100.6	99.9	100.8	95.23	99.98[1]
Ba	704	486	1036	400	764	349	1047	601	931	613	426	531	82	277	467	668	264	1135	753	602	642	
Nb	42	40	66	16	50	41	30	45	58	46	33	40	26	29	40	33	37	49	53	41	16.8	
Rb	126	164	188	148	163	200	258	179	168	167	149	166	86	125	114	258	183	149	219	171	165	174
Sr	54	96	42	101	111	94	60	49	58	48	258	51	81	38	104	84	35	104	82	83	108	69
Y	120	51	34	66	48	45	71	57	30	65	64	73	31	32	39	45	37	31	33	47	35	25
Zr	148	287	321	254	229	313	398	298	229	546	210	195	235	335	144	191	259	232	342	279	196	
Be	2.1	3.1	1.7	1.4	2.1	6.4	1.7	1.2	2.9	0.9	3.8	3.6	0.9	1.6	2.8	1.9	2.3	1.2	1.9	2.3		4.4
Co	92	26	18	22	22	14	30	25	17	20	21	18	25	22	24	16	22	19	32	22	18	22
Cr	150	50	78	75	100	94	120	48	82	83	68	89	64	63	130	50	29	76	67	76	115	105
Cu	310	32	9	140	11	14	1	7	5	7	61	12	20	230	57	6	7	14	30	37		75
La	99	22	51	42	33	18	26	21	39	58	62	55	21	28	14	30	22	49	31	34	38.0	43
Ni	240	39	33	35	39	20	45	30	30	23	32	26	24	23	85	51	58	29	42	37	52	57
V	350	94	89	160	100	74	130	61	86	65	96	88	70	77	190	57	110	110	71	96	100	188
Yb	11	1.5	3.4	2.4	1.6	1.3	2.3	2.0	2.9	2.6	2.0	2.2	2.5	2.6	3.4	1.1	2.4	3.5	2.5	2.3	2.86	
Zn	170	74	53	100	65	130	110	71	60	59	72	60	110	210	90	41	22	44	70	80		114

All sample numbers except 1, 2, and 3 prefixed by FS84-.
Oxides, C, and S in weight per cent, minor elements in ppm.
*Total iron as Fe_2O_3.
[1]Recalculated to 100% by weight.
1. Average of rocks of unit *A*Cp except FS84-60.
2. Average Proterozoic cratonic shale (Condie, 1993).
3. Average Aphebian shale (Cameron and Garrels, 1980).

Table 11. UTM co-ordinates of chemically and isotopically analyzed samples.

Sample	Easting	Northing	Sample	Easting	Northing
FS82-15	568000	7358200	FS84-A109	427400	7502200
105	553500	7410200	A130	461800	7516700
FS84-6	391300	7477000	A142	430600	7502600
14	397400	7475400	A143	431500	7502700
38	411000	7522800	A213	487400	7533200
39	439700	7539600	A246	491200	7535400
43	435800	7441600	A250	438800	7509400
45	441000	7440400	A252	440700	7505100
55	406400	7521000	A254	442000	7511400
59	438200	7504500	A290	426600	7501600
60	438100	7505400	A291	426700	7502000
62	401400	7517800	A292	429500	7505300
73	631000	7394600	G64	417900	7496100
83	435100	7438700	G65	415200	7494500
86	445000	7510200	G70	416400	7494700
87	445800	7511300	G76	418100	7496700
88	431300	7538100	G115	448600	7511500
A68	414700	7493900	G118	445300	7507000
A85	423600	7498300	G192	453400	7509000
A91	430300	7500200			

All locations are in grid zone 15, except those of FS82-15 and -105 and FS84-73, which are in zone 14.