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GEOLOGY OF A TRANSECT THROUGH THE SOUTHERN MARGIN OF THE FOXE FOLD BELT (MAINLY NTS 27B), CENTRAL BAFFIN ISLAND, DISTRICT OF FRANKLIN

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ABSTRACT

The study area lies on the southern margin of the Foxe Fold Belt, a major Hudsonian structural feature on central Baffin Island which is defined by the presence of Aphebian (Early Proterozoic) metasedimentary rocks of the Piling Group.

The stratigraphic sequence is (bottom to top): 1) quartzofeldspathic basement complex (BC) of possible Archean age; 2) quartz-rich and pelitic metasediments of the Dewar Lakes Formation (DLF) which form the basal unit of the Piling Group; 3) mafic and ultramafic rocks of the Bravo Lake Formation (BLF) with closely associated sulphide-rich metasediments of the Astarte River Formation; and 4) turbidites and equivalent high grade paragneisses and schists of the Longstaff Bluff Formation (LBF).

The DLF is a transgressive shallow water sequence deposited on the previously exposed BC. Collapse and deepening were related to extrusion of the BLF. The LBF is a turbidite deposit which infilled the resulting basin.

The stratigraphic sequence has been subjected to four phases of deformation. D1 resulted in the tectonic imbrication of units by thrust faulting. D2 was a synmetamorphic event involving the formation of a pervasive subhorizontal schistosity and a number of macroscopic nappes. D3 and D4 were primarily compressional in nature with the related folds having generally upright axial surfaces and a divergence of only 40° between their axial surface traces (110-290° versus

070-250°). The 110-290° phase is considered to be D3 on the basis of structural observations and on the apparent association of that set of structures with limited gneiss doming.

In the central and southern parts of the area metamorphic grade reached amphibolite facies with temperatures between 580° and 720°C and pressures between 300 and 380 MPa. The distribution of grade and the intersection sense of isograds indicate the presence of a thermal dome to the southeast. This metamorphic event may have been responsible for the intrusion of large, possibly BC-derived plutonic bodies along the southern margin of the belt.

Intrusion of post-tectonic diabase dykes of the Hadrynian Franklin swarm is the last event recognized in the Precambrian evolution of the area.

INTRODUCTION

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Finally I would like to thank the Geological Survey of Canada for introducing me to Baffin Island in 1974 and for providing me with the opportunity to work in this remote area through various contracts and other logistical support.

Background

A portion of central Baffin Island was mapped at a scale of 1:50 000 during the summers of 1977 and 1978. The field work as supported by the Geological Survey of Canada through contracts 1977-93564 and 1978-94753 to the author. A minor amount of additional mapping was conducted during 1979 as part of the final 1:250 000 mapping of the region.

The region described in this report comprises N.T.S. 1:50 000 sheets 27B/4, 27B/5, 27B/12 and 27B/13 as well as parts of 27B/11 and 37A/8 (Fig. 1). Mapping coverage is variable, ranging from the incorporation of reconnaissance data only (for example in the northwestern corner of 27B/13) to detailed examination. Dewar Lakes (D.E.W. Line site FOX 3) was used as a base of operations during the field work.

The study area is underlain by part of the southern margin of the Foxe Fold Belt which is a major Precambrian structural feature in the northeastern Churchill Province of the Canadian Shield (Jackson and Taylor, 1972). Aphebian (Early Proterozoic) metasedimentary and metavolcanic rocks in the fold belt were deformed, together with their presumably Archean basement, during the Hudsonian Orogeny. The study was undertaken to determine the Precambrian stratigraphic, structural and metamorphic evolution of the area and to help provide some understanding of the evolution of the Foxe Fold Belt as a whole.

Figure 2 is an informal compilation of the early results of regional mapping based on Morgan et al. (1975; 1976), Henderson et al. (1979), Henderson and Tippett (1980), and Tippett (1978; 1979). It is intended to provide only the general regional setting. More recent

maps, specifically those contained in Morgan (1981), Henderson (1982), Tippett et al. (1980) and Jackson (1984) have not been incorporated. The reader is referred to those open files for the most up-to-date information and interpretation.

Initial results from this project were reported by Tippett (1978, 1979). Data from the study area collected during 1968 (Jackson, 1971; G.D. Jackson and W.C. Morgan, personal communications (1976 and 1977)), 1975 (Morgan et al., 1976; Chernis, 1976) and 1978 (Henderson et al., 1979) were incorporated into the study. This combined database was used to construct Figs. 3 through 9.

Most geographical names on the figures contained in this open file have not been officially approved but are under consideration by the Canadian Permanent Committee on Geographic Nomenclature.

The study area straddles the transition from low to high grade metamorphic rocks on the southern margin of the Foxe Fold Belt. Examination of thin sections provided information which together with field observations were used to construct a map (Fig. 11) showing metamorphic isograds. The assemblage data were interpreted with the aid of a petrogenetic grid (Fig. 12) to produce a model for the origin of the metamorphic zonations (Fig. 10).

These studies formed the basis of a doctoral dissertation at Queen's University at Kingston, Ontario (Tippett, 1980). This open file presents some of the data used in the thesis as well as the major conclusions pertaining to the study area. The reader is referred to G.S.C. Open File 1111 for additional data and to Tippett (1980) for more discussion

and speculations on the origin of Foxe Fold Belt. This open file is cross-referenced to Open File 1111 in detail using the convention: O.F. 1111, Appendix B or O.F. 1111, Fig. A5 etc.

STRATIGRAPHY

Introduction

Detailed mapping was initially conducted on the basis of lithological characteristics. Upon compilation it became apparent that there was a consistent set of lithological assocations which were then used to set up a relatively simple stratigraphy for the area. The resulting geological map is shown in Figs. 3 and 4. Many structural complications are recognized on the basis of departures from this inferred stratigraphy. It remains true that some of these "structures" could be the result of stratigraphic complications.

Three of the stratigraphic units identified in the study area have been correlated with formations of the Piling Group defined elsewhere in the Foxe Fold Belt (Morgan, 1981 and personal communications, 1979; 1983), namely the Bravo Lake, Astarte River and Longstaff Bluff Formations. The type area of the fourth, the Dewar Lakes Formation, occurs in the study area near Dewar Lakes dome (W.C. Morgan, personal communications, 1979; 1983). The Flint Lake Formation, the fifth Piling Group formation on a regional scale is absent from the study area, being restricted to the northern margin of the Foxe Fold Belt. The term "Clyde River" Formation which was used in Tippett (1980) is obsolete and has been replaced with the term Astarte River Formation (W.C. Morgan, personal communication, 1983). The basement complex on the northern and southern margins of the Foxe Fold Belt may be of the same origin but this has not been established.

In general terms the arrangement of these stratigraphic units in the study area is as follows. A granitoid basement complex is overlain by a variable thickness of Dewar Lakes Formation strata, which are in turn overlain by the closely associated Bravo Lake and Astarte River Formations. The Longstaff Bluff Formation blankets the underlying units. In the southern part of the area the Longstaff Bluff Formation gradually onlaps the basement complex directly as the three intervening formations stratigraphically pinch out. Although the map units undergo significant changes in metamorphic grade, they retain their identities and usefulness in mapping.

Details of each unit are presented below. Modal data, by lithological type, are contained in O.F. 1111, Appendix A. Igneous rocks, interpreted as being intrusive into this pre-existing stratigraphic sequence, are described in the following section.

Basement Complex (Archean)

Dominantly granitoid rocks of the basement complex are exposed in the cores of domal structures throughout the central part of the area (Figs. 3, 4). These rocks are among the most southwesterly exposures of the basement on the southern margin of the fold belt (see Fig. 2).

Granitoid gneisses

Granitoid basement rocks are leucocratic, ranging through pink, orange and grey on fresh surfaces with a distinctive yellowish-white weathered surface. They are usually flaggy to blocky where planar

fabrics are developed. Average grain size varies from fine to medium, generally increasing across the area towards the south. In terms of their major constituents - quartz, alkali feldspar and plagioclase - they are predominantly granitic but include rocks from virtually the entire nonfeldspathoidal igneous plutonic spectrum (O.F. 1111, Fig. A1; Streckeisen, 1973). Individual domes contain a similar range of rock types. Different compositions are commonly closely interlayered so that the labelling of one rock type as "typical" of the basement complex is difficult.

Accessory minerals generally make up less than 10% by mode of the granitoid gneisses. Biotite is most common; hornblende is locally accompanied by colourless amphibole. Identification of the basement is usually based on the absence of metasedimentary index minerals such as muscovite, sillimanite, garnet, cordierite, graphite and iron oxide. Locally, close to contacts with the overlying metasediments, sparse garnet, muscovite (at low grade) and sillimanite (at high grade) occur, probably reflecting the effects of pre-Piling Group weathering of the basement gneisses. At such locations the contact may appear transitional. In general, however, the contact between the basement and metasediments is sharp and distinctive.

Gneissosity (compositional layering) of variable continuity (locally grading into a schlieren fabric) is on the scale of from 1 to 10 cm. It is most strongly developed near the domal contacts with the metasediments, to which is is generally parallel. Towards the centres of the domes, gneissosity decreases in strength leaving only a penetra-

tive schistosity (preferred planar orientation of mineral grains) or, more rarely, a penetrative lineation. With the exception of a small area on the north margin of Dewar Lakes dome, where the layering is interpreted as a relict of pre-Piling Group deformation, pre-existing inhomogeneities which might have been transposed to produce this marginal gneissosity do not appear to be present. If the marginal gneissosity does result from the transposition of older gneissosity throughout the area, its absence in the cores of the domes is puzzling.

Foliated plutonic rocks

Nonlayered granitoid rocks, which contain only fabrics also found in the Piling Group, make up a significant proportion of the basement complex.

These rocks commonly contain ribbon quartz and lenticular feldspar augen which locally have been flattened to form a discontinuous millimetre-scale gneissosity. These fabrics suggest that a coarser grained parent rock underwent a general reduction in grain size through recrystallization during deformation. These fabrics are most apparent in the more northerly domes and decrease in strength to the south, where new grain growth caused by higher metamorphic grade apparently dominated deformation-induced polygonization.

Variation in terms of composition and accessory minerals is similar to that of the granitoid gneisses with which they are gradational.

Amphibolites

Amphibolite layers and boundins up to 20 m thick locally form

up to 10% of the basement complex. They are generally concordant with the gneissosity of the enclosing basement gneisses and have a fine internal gneissosity. They are composed primarily of hornblende and plagioclase with accessory biotite, garnet, pyroxene and sphene. Amphibolites locally serve to differentiate the basement complex from the Piling Group but because amphibolites also occur, although to a generally lesser extent, in all formations of the Piling Group, their presence is not necessarily diagnostic of the basement complex.

Ultramafic rocks

Pods of ultramafic rock in the core of Harris Lake dome have high normative components of orthopyroxene (O.F. 1111, Appendix B) and may be metaorthopyroxenites. In outcrop they are generally massive. Their mottled fabric is due to the presence of coarse blades of tremolite-actinolite crosscutting a finer tremolite-actinolite matrix which also contains clinopyroxene and olivine porphyroblasts. Partially serpentinized olivine pods locally give the outcrops a spotted appearance.

Granitoid veins and dykes

Granitoid veins make up approximately 10% of the basement complex. They average 5 cm in thickness and are coarse grained, massive and generally pinkish, in contrast to the grey host gneisses. Their close interdigitation with the gneisses, in addition to the general correspondence of accessory minerals between vein and host, suggest that they originated through partial melting and are not far removed from

their point of generation. Their range of compositions is similar to that of the basement complex gneisses (O.F. 1111, Fig. A1). Their limited development at high metamorphic grade, in contrast to adjacent highly migmatized metasediments, may result from a lower fluid content due to a generally anhydrous igneous parent, due to the absence of phases favourable for prograde water-generating reactions or due to dehydration during a metamorphic episode which predated the Piling Group. That this contrast in response to high metamorphic grade is related to the absence of fluids is confirmed by an increase in the proportion of veins to up to 60-80% of the outcrop exposure in some zones marginal to highly migmatized, presumably water-saturated metasediments. At the highest metamorphic grade, dykes of pinkish, apparently basement-derived mobilizate, crosscut the base of the metasedimentary succession which contains in contrast generally white mobilizate.

Discordant K-feldspar-quartz-muscovite-garnet dykes crosscut Dewar Lakes dome but their systematic distribution suggests they were not generated in situ.

Piling Group (Aphebian)

Within the study area the Piling Group is made up of four formations, namely the Dewar Lakes, Bravo Lake, Astarte River and Longstaff Bluff Formations.

Dewar Lakes Formation

Distribution

The dominantly quartzitic Dewar Lakes Formation overlies the basement complex and encircles most domes (Figs. 3 and 4), although it forms the core of Kakiartok Rapids dome, probably due to a slightly higher level of exposure. The formation varies markedly in thickness, in a relatively unsystematic manner, although there is a first order thickening from south to north. Thickness variability may reflect the infilling of original erosional irregularities on the Piling Group-basement complex unconformity. The absence of quartzite above the basement complex in some areas towards the south suggests that the basement was locally emergent during deposition of the formation.

The formation as a whole does not contain internal markers or consistent internal variability. It grades into the overlying impure clastics of the Astarte River and Longstaff Bluff Formations over some tens of metres.

Quartzites and schists

The physical appearance of strata of the formation in the field is controlled by the scale of interlayering of its two dominant components: fine- to medium-grained, greyish-white quartzite and greyish mica schist. Layers of true orthoquartzite are rare. Locally, nebulous centimetre-thick layers are produced by finely disseminated iron oxide. Some exposures are made up of pinkish quartzite in which millimetre- to centimetre-thick laminations with varying mica and feldspar contents produce a colour banding and fissility. The scale of interlayering of

mica-rich schists begins at the millimetre-thick laminations in the quartzite and grades through the metre scale until the schists form entire outcrops. The appearance of the outcrops undergoes a parallel change from blocky, through flaggy to fissile and recessive weathering. Rock fragments have not been recognized in the formation although their original presence may have been obscured by deformation. Pebbles identified by Morgan et al. (1976) are interpreted as deformed veins and dykes.

If the mica content of these rocks can be assumed to represent the original matrix, then the formation contains a complete gradation of quartz arenites, subarkoses, arkoses, quartz wackes, arkosic wackes and feldspathic greywackes (O.F. 1111, Fig. A2a; Pettijohn, 1975). The formation as a whole has a relatively high maturity index (quartz versus feldspar plus rock fragments). Compositional data at low grade relevant to the response of these rocks to partial melting and muscovite breakdown are presented in O.F. 1111 Figs. A2b and A2c, respectively.

The dominance of muscovite as the accessory mica (O.F. 1111, Fig. A2d) is responsible for the light colour of the formation in contrast to the dominance of biotite in the darker Longstaff Bluff Formation. The formation locally contains abundant faserkiesel (quartz-sillimanite pods) which give it a pseudoconglomeratic appearance. Accessory minerals include andalusite, sillimanite, garnet, cordierite, staurolite, graphite, tourmaline and opaques.

Amphibolite

Minor amphibolite occurs near Dewar Lakes dome. Along the southeast flank of Harris Lake dome amphibolite bodies up to 10 m thick locally make up more than 10% of the exposures of the formation.

Sedimentary features

Sedimentary textures have been completely destroyed by recrystallization. Layering is thought to be primary, representing transposed bedding. Crossbedding on a scale of 20-30 cm is rare. Graded bedding and base-of-bed structures were not observed.

Depositional environment

In less deformed areas transgressive quartz-rich deposits are generally interpreted as sublittoral sand sheets and barrier island complexes. Intense deformation and metamorphism in this area have, however, removed essentially all primary evidence of the depositional environment of the formation. A shallow water environment is inferred from the stratigraphic position of the unit (onlapping the previously exposed basement complex) and from its apparent maturity.

Astarte River Formation

Distribution

The Astarte River Formation is dominated by rusty-weathering schists and paragneisses with pyrite-pyrrhotite contents on the order of 2 to 3%. The formation has not been indicated as a separate unit in

Figs. 3 and 4 but is combined with the Bravo Lake Formation due to their close spatial association. Within the study area the Astarte River Formation is well developed where the Bravo Lake Formation is present and is poorly developed to nonexistent in its absence. This relationship contrasts with that present on the northern margin of the fold belt (Morgan et al., 1975; Morgan, 1981) where the Astarte River Formation is well developed and the Bravo Lake Formation is absent. Strictly speaking, correlation of these rusty rocks in the study area with the Astarte River Formation proper may not be correct despite their similar stratigraphic positions.

Paragneisses and schists

A complete gradation appears to exist between clastic-dominated and sulphide-dominated metasediments. Sulphides are locally concentrated to the extent that the term lean sulphide facies iron formation is applicable, for example southwest of Casson Lake (68°32'30"N, 71°57'30"W; station 971), where very friable and fine grained, black graphitic schists with up to 50% pyrrhotite occur.

Massive sulphide layers are rare but have been observed on the northern margin of FitzGerald Lake dome at 68°27'N, 71°52'W (station 823) and on the northern margin of South Jackson Lake dome at 68°33'N, 71°14'W (station 38). At the latter location, the layer is 20 to 30 cm thick with an uncertain lateral extent and is contained within a sequence of rusty schists and paragneisses, close to a major amphibolite of the Bravo Lake Formation. The massive sulphides are dominated by pyrrhotite

with a minor gangue component of quartz, feldspars and micas. Pyrite, magnetite and chalcopyrite are each locally present in amounts up to 2 to 3%. The dominance of pyrrhotite over pyrite may be related to the metamorphic grade. The sulphides are totally recrystallized with a massive, medium grained fabric. Compositional layering was not recognized.

In the area, one small occurrence of oxide facies iron formation has been noted (station 977, south margin of Dewar Lake dome) in which magnetite makes up 20% and hematite 5% of a thin, otherwise clastic, layer laminated on the scale of 2 to 5 mm. Chernis (1976, p. 18) also described lean oxide facies iron formation from the structurally complex zone north of Dewar Lakes dome. That zone contains in addition rusty quartz-rich metasediments with abundant iron-rich amphiboles (cummingtonite-grunerite; Chernis, 1976) which may constitute a silicate facies of the iron formation.

Hematitic alteration of sulphide facies iron formation is weakly developed. Malachite staining of calc-silicate gneiss on the northern margin of South Jackson Lake dome at 68°33'N, 71°19'30"W (station 204) may be the result of the weathering of chalcopyrite in the adjacent rusty metasediments.

Depositional environment

The close association in the area of the sulphide-rich metasediments of the Astarte River Formation with mafic volcanic rocks of the Bravo Lake Formation suggests a genetic link between the two

specifically that those rocks are volcanogenic sulphide-rich sediments. The more massive sulphide occurrences may be similar in nature to Besshi-type ore deposits.

Bravo Lake Formation

Distribution

The Bravo Lake Formation is a stratiform assemblage of chiefly mafic volcanic rocks which occurs above the Dewar Lakes Formation and below of the Longstaff Bluff Formation. The formation is thickest along the western margin of the area and thins to the east, generally being poorly developed east of Dewar River. It also appears to thin northwards, being absent north of Thomson Lake, and southwards, being poorly developed to absent around Kakiartok Rapids dome. Thick stratiform rocks of this association, other than rare amphibolite to ultramafic layers in the Longstaff Bluff Formation, are interpreted as being present at only one level in the stratigraphy. Their occurrence higher in the section, for example north of Dewar Lakes dome (Chernis, 1976), is interpreted here to be the result of structural complications.

The formation is made up of mafic-rich rocks ranging from amphibolites with subequal hornblende and plagioclase, through hornblende and tremolite-actinolite-rich schists, to ultramafic rocks containing abundant olivine and clinopyroxene.

Mafic rocks

Amphibolites and hornblendites are commonly closely associated and locally have well developed compositional layering which may be

primary or deformation-induced. In places amphibolites are interlayered with metasediments on the scale of metres. More rarely interlayering is on the 1 to 5 cm scale, for example at stations 55 and 73 (both on the southwestern margin of Varley Lake synform). Near station 562 (on the northern margin of Harris Lake dome) more massive amphibolites are replaced along strike by mafic-rich metasediments. Fine scale interlayering is interpreted to indicate that some of the amphibolites are metasedimentary in origin and were derived from weathering of the mafic igneous rocks combined with a variable terrigenous component. Such thin bedding and lateral changes have not been observed in the more mafic-rich rocks of the formation.

Amphibolites commonly contain cummingtonite in optical continuity with hornblende. Clinopyroxene is a common accessory and the hornblende-cummingtonite association locally shows a correlation with the distribution of clinopyroxene on the thin section scale. Sphene, biotite and garnet are the other dominant accessory phases.

Ultramafic rocks

The origin of ultramafic rocks is more problematical. The only possibly pillowed outcrop observed (station 14, on the southern margin of Dewar Lakes dome) is made up of ultramafic rocks. Their spotted appearance in most outcrops results from selective weathering of olivine and clinopyroxene porphyroblasts which are interpreted to be metamorphic. An igneous origin is precluded by the lack of layering in these rocks and by the fact that the porphyroblasts locally overgrow a

fine schistosity. The overgrowth by clinopyroxene of green spinel of metamorphic origin confirms this interpretation. The development of the porphyroblasts is controlled by the bulk composition of the rock and does not represent a primary texture.

Calc-silicate gneisses

Calc-silicate gneisses composed of millimetre- and centimetre-thick layers rich in calcite, diopside, tremolite-actinolite and hornblende are closely associated with the mafic and ultramafic rocks of the formation, even more so than the rusty metasediments of the Astarte River Formation. This suggests that they too are volcanogenic. These rocks are interpreted as primarily chemical sediments with a variable terrigenous component. They were mapped as part of the Bravo Lake Formation in the area.

At station 166 (within Casson Lake nappe north of Tullugak Lake dome) a "melange" of flattened to subangular carbonate blocks in an amphibolitic matrix is present. The origin of this association is unknown.

Origin

A suite of 45 hand specimens of mafic and ultramafic rocks from the area were chemically analysed by the Geological Survey of Canada in order to ascertain the types of rocks included in this and adjacent formations as well as to get some clues concerning their origin. The analytical results are presented in O.F. 1111, Appendix B, together

with a number of diagrams (O.F. 1111, Figs. B1 to B8). Rock types have been determined on the basis of classifications used by Irvine and Baragar (1971). The results indicate that mafic-rich ingneous rocks ranging from subalkaline tholeiitic basalts to ultramafic rocks form the bulk of the Bravo Lake Formation. The presence of minor alkalic tendancies is here considered to be the result of contamination during diagenesis and metamorphism but could also be a primary signature.

A recognizable stratigraphic arrangement of these lithologies generally appears to be absent, tending to discount the possibility that the formation is actually an ophiolitic complex (an idea suggested by Jackson (1971) who mentioned "ophiolite-like" rocks along the southern margin of the fold belt). In addition, compositional variations in the igneous rocks of the formation are similar to those of amphibolitic and ultramafic rocks in the underlying Dewar Lakes Formation and basement complex. to be specific, the presence of ultramafic rocks in the basement complex which are orthopyroxenites in terms of their normative mineralogy correlate with chemical trends in the Bravo Lake Formation which are compatible with orthopyroxene fractionation. This conclusion is based on a number of chemical variations which are discussed in detail in Tippett (1980). These compositional trends are interpreted to be related to fractional crystallization of an intrusive and extrusive suite of basaltic magmas which were contaminated by assimilation of crustal material with the dominant precipitating phase being orthopyroxene.

This similarity of compositional trends together with the absence of a pseudostratigraphy indicates that the Bravo Lake Formation

originated as an authochthonous submarine flood basalt with differentiates rather than as an obducted ophiolite sequence.

Figure B9 of O.F. 1111 is an attempt to determine the general tectonic setting of these rocks based on their major element chemistry by using a diagram developed by Pearce et al. (1977). The non-ultramafic points are mainly scattered over the continental and the ocean island fields. It may be significant that these settings are intraplate in nature, as opposed to the plate margin-related ocean ridge and floor, orogenic and spreading centre island fields which have only one representative in this suite.

Longstaff Bluff Formation

Distribution

The Longstaff Bluff Formation, the uppermost stratigraphic unit of the Piling Group, underlies a greater part of the study area than any other formation (Figs. 3 and 4). It is difficult to obtain data relevant to its primary nature for two reasons. It is highly migmatized in the southeastern half of the area. To the northwest, exposures are generally poor and where present, for example along the shores of North Jackson Lake, intense deformation has obscured all but compositional layering.

On a regional scale, the formation appears to be extremely homogeneous. Well preserved sedimentary features at the type area near Longstaff Bluff are compatible with the identification of the formation as a thick sequence of turbidites. The strata have similarily been

interpreted as a sequence of variably metamorphosed distal turbidites by Henderson et al. (1979) and by Henderson and Tippett (1980). It is assumed that this is the origin of the formation in the study area. Marker horizons of significant extent were not observed, making detailed correlations impossible.

Low metamorphic grade area

At low metamorphic grade in the area the contrast between two bed types, quartz- and feldspar-rich versus mica-rich, is generally the only primary sedimentary structure preserved with the exception of local biotite-rich laminations. The "gritty" quartzofeldspathic beds give the impression of being thicker on average and appear to range in thickness from 0.1 to 3 m but numerical data to support this generalization are not available. Grading of grain size was only rarely observed. In thin section, these rocks are seen to contain angular polycrystalline quartz clasts and fresh, well-twinned subrounded microcline grains. The two minerals were not observed together in polycrystalline clasts. In most cases, grains are small and tend to be monocrystalline making their distinction from the recrystallized matrix difficult. In general the formation is made up of dark to speckled, "salt and pepper", massive to slightly laminated (1 to 20 mm scale) metasiltstones. North of Dewar Lakes dome the formation lacks this strongly bedded aspect and is more massive and gradationally laminated. The relationship between the two types of stratification is now known. Calc-silicate nodules have been observed throughout the area.

Assuming the mica content to be representative of the original matrix content, these rocks are generally arkoses to arkosic arenites and arkosic wackes to feldspathic greywackes with a low to nonexistent component of recognizable rock fragments (O.F. 1111, Fig. A3a; Pettijohn, 1975). The lower maturity index and greater mica contents distinguish these rocks from the quartz-rich rocks of the Dewar Lakes Formation. The predominance of plagioclase over K-feldspar (O.F. 1111, Fig. A3b (cf. A2b)) is noteable. Figure A3c (O.F. 1111) illustrates the pre-muscovite breakdown variability in terms of quartz-muscovite-K-feldspar. The predominance of biotite as an accessory mineral is responsible for the dark coloured nature of the unit (O.F. 1111, Fig. A3d (cf. A2d)).

High metamorphic grade area

At higher metamorphic grade the formation is transformed into a sequence of generally brown to rusty paragneisses and schists. Compositional layering of blocky to flaggy quartz-feldspar-biotite paragneiss and schistose pelites is interpreted as relict, though transposed, sedimentary layering. In some cases, the development of variable concentrations of metamorphic minerals seems to mimic original grading. Pervasive deformation has often produced a secondary layering but locally blocks containing original lamination may be observed in a deformed matrix.

The transformation of the formation into paragneisses and schists is generally accompanied by the appearance of significant amounts of white, medium- to coarse-grained, generally massive, quartz-plagioclase-K-feldspar granitoid material. These rocks are mainly granitic and

granodioritic in composition (O.F. 1111, Fig. A4; Streckeisen, 1973). Although this range of composition is the same as that of the basement complex gneisses and the southern plutons, the colour, texture and physical relationship to the paragneisses and schists of these granitoid rocks serve to set them apart.

The close association of the granitoid material with the metasediments is considered to be diagnostic of its generation by partial melting (anatexis). The resulting migmatites are usually well layered with the scale of association of restite (residual) and leucosome (crystallized melt) ranging from millimetres and centimetres through several metres. Locally thick concordant bodies are up to tens of metres in thickness. Such concentrations of leucosome confirm its mobility which is further supported by local crosscutting features which give rise to agmatitic zones of paragneiss and mobilizate. Such discordant bodies are generally limited to the zone in which in situ mobilizate also exists, indicating that injection into the overlying nonmigmatitic sediments was not common.

At high grade the formation appears to contain fewer calc-silicate nodules than at lower grade. The high grade areas also locally contain zones of greenish paragneiss with closely associated reddish mobilizate, often near major amphibolite bodies. On the basis of petrographic evidence these zones appear to be made up of highly retrograded equivalents of the normal dark grey to brown paragneisses and white mobilizate.

Mafic and ultramafic rocks

Within the area, the Lonstaff Bluff Formation contains scattered amphibolites and rare ultramafic rocks. As a rule these rocks occur as pods on the scale of 1 to 3 metres in the metasediments and may be interpreted as disrupted basaltic and ultramafic(?) dykes although discordance with primary compositional layering has not been observed. Locally such units can be traced for up to a kilometre. Two assemblages of mafic rocks in the formation are more substantial. A band of mafic and possibly ultramafic rocks up to 50 m in thickness is particularly well exposed at station 683 north of Holgate Lake. Although it is difficult to prove, it appears to be a conformable unit. Likewise east of Lismer Lake an amphibolite layer up to 40 m in thickness was traced for over 6 km and appears to be folded concordantly with the enclosing paragneisses and schists. These occurrences, as well as minor calcsilicate gneiss observed near station 920 southeast of Harris Lake dome, are interpreted to be the result of minor pulses of igneous activity during the deposition of the formation, analogous to those which led to the extrusion of the Bravo Lake Formation. The smaller amphibolites mentioned above may be feeder dykes to these stratiform units.

Depositional environment

Turbidites are presently interpreted in terms of submarine fan environments rather than simply by the variable development of the Bouma cycle. In spite of the lack of fine primary sedimentary features in the study area, the preservation of primary bedding indicates that the turbidites of the Longstaff Bluff Formation lack the lateral variations

in bed thickness, the cyclicity and the coarse clastic sediments characteristic of the channelled portions of fans. The thickness of the beds appears to eliminate the outer fringe of the fan and the basin away from the fan as possibilities. These observations, together with the generally fine-grained nature of the sediments suggests that deposition on a suprafan lobe of the outer fan is the most likely setting. Deposits of suprafan lobes should have a cyclicity related to the lateral movement of feeder channels on the middle fan. The apparent lack of such cyclicity may simply reflect the incomplete nature of presently available data.

The Longstaff Bluff Formation is made up of quartzofeldspathic turbidites for which the source areas are unknown. Given the capability of far-travelled turbidites to transport evidence of their source it appears unlikely that a carbonate shelf or bank, for example the carbonate-rich Flint Lake Formation, was the facies equivalent of the formation. Rather, the formation seems to have overlapped the entire underlying stratigraphy as a result of a major regional subsidence which was accompanied by small intrusions and extrusions of basaltic magmas. It would seem logical to suggest that whatever was responsible for this subsidence was also responsible for the uplift of the corresponding area which was capable of producing a huge volume of quartzofeldspathic detritus. Unfortunately neither facies changes nor sedimentary structures have yet yielded any clues as to the location of this uplifted area.

Intrusive map units

Lismer Lake, Nulliarek Lake and Carmichael Lake plutons (Aphebian)

Distribution

The main body of this type is Carmichael Lake pluton which occurs along the southern edge of the area (Fig. 4) and appears to merge with a terrane dominated by similar rock types extending to the south. This generalization is based on the distinctive homogeneous appearance of these plutons as seen on aerial photographs which, although crisscrossed by lineaments of regional extent, contrast with the strong linear grain of areas underlain by Piling Group metasediments.

Phases of the plutons

Megacrystic granitoids, mainly granites and granodiorites (O.F. 1111, Fig. A5; Streckeisen, 1973), make up 80 to 90% of Carmichael Lake pluton. The nongenetic adjective "megacrystic" is considered most appropriate because of the uncertain origin of the large K-feldspar crystals (up to 2 by 8 cm in cross section) which characterize the pluton on the outcrop and hand specimen scales. A strong preferred orientation of the crystals appears to be tectonic in origin and parallels a penetrative schistosity which is irregular only to the extent that it weaves among the relatively coarse grains of the rock. Anastamosing dikes of equigranular, medium-grained, garnetiferous granitoid (O.F. 1111, Fig. A5) discordantly crosscut this schistosity. Small blocks of fine- to medium-grained, equigranular, nongarnetiferous granitoid material occur as enclaves in the megacrystic granitoids and texturally contrast

sharply with them (O.F. 1111, Fig. A5). Some zones are characterized by fine grained, rounded to subangular, mafic-rich blocks in a matrix of megacrystic granitoid. Such blocks generally have a preferred orientation parallel to the schistosity, probably as a result of rotation during deformation. The two enclave types appear to be gradational in composition but are commonly strongly contrasting.

These observations are generally valid for Nulliarek Lake and Lismer Lake plutons as well, however, the megacrystic fabric of the dominant rock type is progressively less well developed to the north, paralleling a decrease in the size of the bodies.

Of critical importance in the interpretation of these rocks are the accessory minerals. In general, biotite is the dominant accessory in all three plutons. It is often accompanied by garnet, not just near the contacts with metasediments of the Piling Group, as is the case with the quartzofeldpathic basement complex to the north, but throughout the outcrop area of the plutons. Hornblende is locally present. Clinopyroxene is present only in mafic-rich enclaves in Carmichael Lake pluton. Charnockitic (hypersthene-bearing) rocks are absent in the study area. Traces of apatite, sphene, epidote, zircon and opaques are scattered throughout. Rusty-weathering outcrops are not related to the presence of abundant sulphides or graphite but rather to the weathering of biotite. There does not seem to be any difference in the accessory minerals among the three main rock types with the exceptions of clinopyroxene which occurs only in the mafic-rich enclaves and which is absent in the more felsic enclaves.

The megacrystic granitoid rocks in the study area were initially interpreted to be recrystallized in situ basement gneisses with the equigranular granitoid rocks being dykes, a pattern similar to that of gneiss and mobilizate in the basement complex to the north. Garnet in the granitoid fractions was considered to be a result of the partial breakdown of biotite through a mechanism including dehydration. The enclaves were interpreted as restites and amphibolites whose block-like form in their massive granitoid hosts was the result of deformation and disruption.

Lismer Lake, Nulliarek Lake and Carmichael Lake plutons are presently interpreted as intrusions, derived by melting of the basement complex at slightly deeper crustal levels. The absence of pelitic index minerals is thus expected and the regional discordance is explained by the intrusive nature of the magmas. Within the study area the position of the plutons below the metasediments and the local presence of basal Piling Group lithologies near their contacts suggests that penetration into the metasediments has not been significant.

Age relationships

The exact timing of the emplacement of the plutons is difficult to determine. The penetrative schistosity which they contain is interpreted to be axial planar to the major regional nappes and to the transposition of layering during the peak of metamorphism in the area. As such, these bodies predate the peak of metamorphism but to what extent is not known. It is considered most likely that they are both

tied to the same overall thermal event.

The absolute age of these intrusions has been established as late Aphebian, but it is possible that superficially similar rocks have a variety of ages on a regional scale. Jackson and Morgan (1978) indicate an age of 1670±30 Ma (Rb-Sr whole rock isochron) for charnockitic granite, a representative of their quartz monzonitic-monzocharnockitic plutonic group, from southwest of Cape Hooper (Fig. 2). A high initial strontium isotope ratio was thought to be indicative of its crustal origin. Pidgeon and Howie (1975) dated megacrystic charnockite from Pangnirtung at 1900±20 Ma (zircon, concordia diagram) and interpreted the granulite facies metamorphism of the region to have overprinted these originally igneous rocks. Results from the western part of the island have given a minimum age of 1826 Ma on zircon (Wanless, 1979). It would appear that intrusion and reworking of these plutonic rocks has been either diachronous or long lived on a regional scale.

Diabase dyke swarm (Hadrynian)

Distribution

The distribution of diabase dykes in the area is illustrated in Figs. 3 and 4 although the thickness of the dykes has been slightly exaggerated to allow them to be visible. Closely associated dykes locally form sets which can be traced across the entire area and for over 100 km on a regional scale. Within these sets the pinching out of one dyke is compensated by the widening of another. In the southern part of the area these clusters seem to occur at roughly 20 km intervals.

Trend and attitude

The dykes cut across all structures in the area and are definitely post-tectonic. Due to their resistant nature they form major lineaments in areas of sparce outcrop. Two subtrends are locally present, namely 110-290° and 130-310° (for example along the crest of Dewar Lakes dome), but in general a trend of 120-300° is representative. The presence of joint sets with a similar trend may suggest that the two are related although locally the dykes seem to have been intruded along even older zones of weakness, such as at station 485 where the FitzGerald Lake dome "break" seems to control the trend of one dyke. Available evidence including observed contacts and columnar joints indicates that the dykes are vertical.

Structure

Dykes range from centimetre scale dykelets, through the metre scale and up to major dykes of over 40 m in thickness. In general, the maximum thickness of individual dykes seems to correlate with their total length. Dikelets are sometimes quite irregular with matching irregularities along their margins implying an origin by intrusion during tension.

The presence of relatively abundant magnetite gives the dykes a strong magnetic expression which dominates aeromagnetic maps of the area.

Columnar joints were observed at stations 106 (central Dewar Lakes dome) and 260 (southwestern flank of South Jackson Lake dome).

Small marginal apotheses are locally present. Evidence of multiple intrusions, such as crosscut chill margins, is absent. Rare xenoliths of the immediately adjacent country rock up to one centimetre in diameter are confined to the margins of the dykes. Wall rock alteration is limited to local reddening although at stations 378 and 381 (both north of Kakiartok Rapids dome) a small amount of orange melt appears to have been developed.

Texture and mineralogy

The diabase is massive, dark grey to black on fresh surfaces and a distinctive chocolate brown on weathered surfaces. Grain size variation is most evident near the sharp margins of the dykes where aphanitic chill margins grade over 10-30 cm into fine- and medium-grained diabase. The term gabbro is locally applicable for the cores of the thickest dykes. Interlocking plagioclase laths, clinopyroxene and magnetite occur in a generally ophitic to subophitic texture. Hornblende, minor hematite and saussuritic alteration have been observed. The dykes are locally slightly vesicular or amygdaloidal (up to 1-2%) on a millimetre scale, such as at stations 260 (southwest flank of South Jackson Lake dome) and 403 (north of Kakiartok Rapids dome). Evidence of neither flow differentiation nor fractionation and crystal settling during crystallization was observed, suggesting that the basaltic liquid was intruded in a completely liquid state and was too viscous during cooling to permit the movement of crystals. There is every reason to believe that the dykes are made up of a single homogeneous magma type but this

has not been proven chemically.

Age

These dykes are believed to form part of the Franklin dyke swarm (Fahrig et al., 1971). An age of 675 Ma is generally accepted for the swarm, on the basis of both K-Ar dates and paleomagnetism.

STRUCTURAL GEOLOGY

Introduction

The structural evolution of the area is interpreted in terms of five phases of deformation. The first is restricted to the basement complex and may be Archean. The remaining four phases (D1 to D4) followed deposition of the Piling Group. Each is characterized by significantly different sets of structures which are discussed in inferred order of their development below, following a brief discussion of the structural terminology used in this study.

Structural terminology

It is appropriate to define a number of fabric terms used in this report in order to avoid confusion. Schistosity is intended to mean a preferred planar orientation of platy and/or acicular mineral grains, as opposed to gneissosity which means a compositional layering. Lineation means a preferred linear orientation of platy and/or acidular mineral grains. In neither case is there any inference as to grain size.

"D" refers to the phase of deformation (D1, D2 etc.). "S" refers to an axial planar feature formed during the folding of that phase (S2 etc.). "S" can either be the geometric axial plane of a fold or a schistosity which is inferred to have formed parallel to such an axial plane. "L" indicates a linear feature related to the folding of that phase (L2 etc.). "L" can refer to the hinge of a fold of layering or older schistosity, the trace of which on the axial plane having the

same significance. It can also refer to a mineral lineation which formed during that phase. In some cases both types of "L" may be formed during the same phase at high angles to each other and may occur in the same outcrop. The set of structures associated with each event has its own variations.

A comment should be made about the selection of the domain boundaries used in the display of structural data on stereonets in Fig. 9. The attempt was made to break down the area into a number of zones in which the structural trends are relatively homogeneous. Given the gradations in orientation present in the area, however, this process is bound to be arbitrary. In recognition of this the data have also been combined in Fig. 9 into a number of composites which would correspond to larger domains.

The types of structures associated with D1 to D4 in the area are described below.

Pre-Piling Group deformation

There is very limited evidence for an older structural history in the basement complex. The single fabric element which is assigned to this history is a gneissosity observed along the northern margin of Dewar Lakes dome and at single localities within Varley Lake dome, Dewar Lakes dome and Harris Lake dome. This gneissosity is almost totally transposed parallel to the schistosity which also occurs in the Piling Group, being separable from it only in restricted hinge zones. It has not been assigned a fabric name. This older gneissosity is different

from the gneissosity which is increasingly well developed towards the margins of some domes in that it is folded with the development of crosscutting axial planar schistosity whereas the marginal gneissosities are everywhere concordant with the schistosity and the basement complex - Piling Group contacts. Some of this margin-parallel gneissosity may be the older gneissosity completely transposed into this orientation but, unless the basement complex is viewed as regionally having had a zone of well layered gneisses beneath the Piling Group contact, the two gneissosities should be considered separately.

D1

Introduction

The earliest phase of deformation to have affected the Piling Group in the study area is recognized strictly on the basis of the map pattern of stratigraphic units. It appears to have involved the imbrication of these units by thrusting; in one case with involvement of the basement complex, and in another case, above a decollement at the top of the Dewar Lakes Formation. There do not appear to be any fabrics, such as cataclastic zones, associated with this phase. In Figs. 3 and 4 the term "defined thrust fault" is used to indicate the presence of a break in the normal stratigraphic sequence which is interpreted to be related to this thrust faulting. The two main examples of the type of field relationships which define this phase of deformation are described in detail below. The sequence of units in each of these areas should be compared to a "type" stratigraphic section, for example that present

immediately above the basement complex along the southern and northern margins of Dewar Lakes dome.

Varley Lake synform imbricate zone

Figure 7 is a detailed map of an area just north of Dewar Lakes dome which is based on data collected by Chernis (1976). According to the interpretation of Chernis (1976), the sequence of mafic, ultramafic and metasedimentary rocks in this area form part of an "upper sequence" of such rocks in the Piling Group. They were not considered to be related to the Bravo Lake Formation which elsewhere in the region is generally the only such assemblage and directly overlies the Dewar Lakes Formation. Within this "upper sequence", five stratiform divisions were recognized, each of which is made up essentially of a basal mafic to ultramafic unit with minor calcareous sediment, overlain by rusty quartzofeldspathic gneiss and schist, and then by metagreywacke (legend of Fig. 7).

This complex sequence is here interpreted as a series of thin thrust sheets involving the Bravo Lake, Astarte River and Longstaff Bluff Formations. It apparently developed above a decollement at the top of the Dewar Lakes Formation. This interpretation is illustrated in Figs. 3 and 7. Quartz-rich metasediments associated with the "upper sequence" are considered to be part of the Astarte River Formation rather than slices of Dewar Lakes Formation because of their rusty nature, their lack of pelitic index minerals and of related non-rusty muscovite schists, and their association with zones containing abundant

iron-rich amphiboles. These latter zones are interpreted as silicious chemical sediments, possibly including some silicate facies iron formation.

Eastern margin of FitzGerald Lake dome

The second macroscopic example of the effects of D1 exists along the eastern and northern edges of FitzGerald Lake dome (Fig. 4). The stratigraphy in this area can be considered as being made up of the sequence basement complex - thin Dewar Lakes Formation - thin Bravo Lake Formation - Longstaff Bluff Formation. The structure of the dome margin is dominated by three thin thrust sheets, each of which carries basement complex gneisses at its base. The outcrop pattern cannot be interpreted in terms of basement complex-cored D2 folds because of the normal stratigraphic order of units above the base of each basement complex sheet and the absence of overturned section.

The two lower thrusts root in the main basement complex exposures near the southeastern corner of the dome where they presumably place basement complex on basement complex. They cut higher in the section to the northwest and progressively lose the basement complex, the Dewar Lakes Formation and the Bravo Lake Formation in their hangingwalls, although stratigraphic absence reduces the number of steps in this sequence in some places. The thrusts disappear into Logan Lakes synform where they place Longstaff Bluff Formation on Longstaff Bluff Formation.

The uppermost thrust sheet is not obviously rooted in the autochtonous basement complex in that it loses its basal basement complex

sliver at both ends where its cover of Piling Group lithologies comes into contact with that of the underlying sheet. This truncation is accomplished at its northwestern termination by placing Longstaff Bluff Formation of the uppermost sheet on Longstaff Bluff Formation of the middle sheet, the Dewar Lakes and Bravo Lake Formations being stratigraphically absent above the uppermost basement sheet at that locality. At the southeastern termination, the Dewar Lakes Formation of the uppermost sheet is cut at the same location as the basement complex, resulting in the superimposition of Bravo Lake Formation of the uppermost sheet on Bravo Lake Formation of the middle sheet. In Fig. 5, (section C-C", D-D", E-E", F-F", G-G") the geometry of the thrust sheets in vertical section has been drawn schematically to give some idea of the presence of these structures, rather than to imply detailed knowledge of their actual configuration.

D2

Introduction

This event involved the regional development of sub-horizontal schistosity (S2). The rare minor folds to which S2 is axial planar are isoclinal and of unknown sense. The parallelism of compositional layering (probably related to original bedding) and S2 over large areas indicates a wholesale transposition of older layering during D2. A penetrative lineation termed L2 lies within S2 on the crests of some of the major D3 gneiss domes in the area. The reason for its parallelism with the long axes of these later domes is poorly understood. Henderson et al. (1979)

reported a similar pattern to the east. Tippett (1980) speculated on possible origins.

The degree to which macroscopic folds were developed during this transposition is not obvious in areas such as the southern migmatitic paragneiss zone (highly metamorphosed Longstaff Bluff Formation only) but where a variety of stratigraphic units is present, it is possible to test for the presence of such structures. Excluding those zones in which the stratigraphic sequence has been repeated by D1, removal of the effects of later compressional folding (D3-D4) on the attitudes of S2 results in the observation that in a significant number of cases the original stratigraphic order of units had been reversed prior to D3. This has led to the recognition of a number of large, originally recumbent, structures. The detailed geometries of two of the largest of these structures are discussed in more detail below.

Casson Lake nappe

In the central and west-central parts of the area (Fig. 3), a significant thickness of amphibolites and ultramafic rocks occurs in a seemingly anomalous position. These rocks can, however, be traced back to the south side of South Jackson Lake dome into a position consistent with their identification as Bravo Lake Formation (Fig. 4). Given that this formation is largely made up of volcanic rocks, it is conceivable that the apparent crosscutting of the Longstaff Bluff Formation could be an intrusive feature. There is, however, evidence that it is the result of the emplacement of a major northerly overturned and originally

recumbent, isoclinal D2 fold, here named Casson Lake nappe, whose present outcrop pattern is the result of a combination of D2, D3 and D4 folding.

The key to this structural interpretation is the presence of quartzites and schists of the Dewar Lakes Formation at several locations within this zone of mafic and ultramafic rocks. Unlike the silicious metasediments associated with the mafic rocks of the "upper sequence" north of Dewar Lakes dome, these silicious rocks are not rusty and are made up of interlayered quartzite and aluminous pelites containing andalusite, sillimanite, garnet and biotite, typical of the Dewar Lakes Formation. At station 204, north of South Jackson Lake dome, the quartzite has a gritty appearance, probably due to the presence of relict clastic grains. Southwest of Casson Lake, the zone contains two bands of quartzite and aluminous pelite, and is there interpreted as a large anticline with a medial syncline (Fig. 3). The Bravo Lake Formation in this vicinity has a two-fold pseudostratigraphy with well foliated tremolite-rich schist stratigraphically overlying massive mafic rocks. The map pattern of these units confirms the presence of a medial syncline indicated by the distribution of the Dewar Lakes Formation. The double fold structure continues at least four kilometres along strike and probably much farther given the consistent outcrop width of this portion of the nappe. At station 204, only a single quartzite zone occurs and there the structure can be considered as having a single closure. Other than in these two areas, however, the Dewar Lakes Formation is not exposed in the core of the nappe which is made up entirely of Bravo Lake Formation lithologies. Its presence in these

few locations, however, is considered diagnostic in the interpretation of the structure as a whole.

There is no evidence that the basement complex is involved in this nappe. The detailed cross-section through North Jackson Lake synform (Fig. 8) is consistent with the nappe being developed solely within the Piling Group in that the medial position of the main closure of the synform between the two margins of the adjacent domes shows that the basement complex - Piling Group interface has been folded into a simple, though slightly inclined, syncline between Dewar Lakes dome and South Jackson Lake dome.

D2 structures in Harris Lake dome

The outcrop pattern of map units in the vicinity of Harris Lake dome is unusual (Fig. 4). Three features are relevant to the interpretation of D2 structures in the area.

The first feature is the band of basement complex gneisses which is contained in a D3 synform within the unusually wide zone of Dewar Lakes Formation along the southern margin of the dome (Fig. 4). The geometry of the zone is reasonably well constrained along its western and central portions but only loosely constrained to the east. In this eastern portion, the Longstaff Bluff Formation - Dewar Lakes Formation contact and the main basement complex - Dewar Lakes Formation contact are plotted on the basis of air photo interpretation and limited outcrop data, combined with the established contacts from the north and south margins of the dome. These constraints indicate that the synform cored

by basement gneisses terminates near the eastern end of the dome where the outcrop belt of the Dewar Lakes Formation begins to thin. The synformal structure containing these gneisses is unusual because its presence implies that the gneisses overlay the Dewar Lakes Formation before D3.

The second unusual feature is the interfolding of the basement complex and Dewar Lakes Formation near the narrows between Harris Lake and Lismer Lake (Fig. 4). The prong of basement complex gneisses which extends out from the main body forms the core of a D3 synform, implying that prior to D3 the gneisses overlay the Dewar Lakes Formation there as well. This is the opposite of the case just farther west where the original stratigraphic order appears to have been preserved.

The third unusual feature is the presence of a thin sliver of the Dewar Lakes Formation within the dome between station 585 and 586 on the eastern shore of Harris Lake. This sliver appears to have been folded into the basement complex during D2 with S2 axial planar.

These three features are combined with the variation in S2 measurements in the dome to produce parts of sections F-F", I-I" and J-J" (Fig. 5) and section L-L" (Fig. 6). The interpretation involves two basement complex-cored D2 nappes of which the upper is much larger and extends from the northern margin of the dome to past the southern keel of basement complex gneisses. Both structures were originally recumbent towards the south. The small keel of Dewar Lakes Formation within the dome is considered to mark the axial surface trace of the recumbent "syncline" which separated the basement complex-cored "anticlines". The

hook in the main basement complex - Dewar Lakes Formation contact is due to the presence of the smaller underlying structure which has been folded about an easterly-plunging D3 synform. There is no evidence that the Bravo Lake or Longstaff Bluff Formations were involved in these structures.

D3 and D4

Introduction

The presence of two post-D2 structural trends is obvious from the macroscopic map pattern of units on a regional scale and within the study area (Figs. 1,3,4). The surface traces of one fold set trend at roughly 110-290° and those of the other at roughly 070-250°, although in some areas characteristic trends are altered by interference effects. In most of the study area there is a continous range of minor fold axial plane strikes between these two orientations. As a result it was decided to arbitrarily make 090-270° the dividing line between S3 and S4. Because the separation is arbitrary, however, the data have been recombined in Fig. 9 where the S3-S4 composites of folds of schistosity (S2) indicate the lack of a bimodal distribution for many domains. Similar composites of L3-L4 (Fig. 9) show a similar lack of differentation.

In Tippett (1980) the two regional sets of structures were attributed to two phases of deformation, namely D3 (110-290°) and D4 (070-250°). It is nomenclature compatible with this relative sense of timing which appears on the various figures accompanying this text. There is, however, some controversy concerning the relative timing of

the two sets of structures (cf. Henderson and Tippett, 1980). The geometrical attributes of each of the two sets are described below.

Structures trending 110-290° ("D3")

Megascopic antiforms with this orientation are cored with basement gneisses and are spaced at roughly 15 km intervals thoughout the central part of the area (Figs. 3, 4) where they have been described as domes. Schistosity axial planar to D2 structures (S2) is gently arched across the central parts of these structures. At the margins of the intervening synforms (so called because locally the folded stratigraphy was inverted during D2) S2 becomes progressively steeper in orientation.

Textures within the synforms are dependent on the level above the basement contact to which S2 was formed during D2. In those zones where it was developed high into the sequence of overlying metasediments, axial planes of structures trending 110-290° have an upright to vertical orientation and are defined by a crenulation of S2. Where S2 was not developed high into the sequence, the axial planes are likewise upright to vertical but are characterized by the presence of a penetrative axial planar schistosity with no hint of a crenulation of an older fabric. An intervening zone of crenulated S2 between the domes and the synforms may or may not be present in these cases. Measurements from three synformal zones of upright D3 schistosity are contained in Fig. 9 (Domains E, K and O).

The variation in flattening planes from horizontal on the crests of the antiforms to vertical in the cores of the synforms (compare

Domains D and E), together with the presence of significantly more minor folds with cascading vergence than would be expected in a strictly compressional setting (Figs. 3, 4), is interpreted to be the result of movement driven by density instability (gravity tectonics) during the formation of these structures. This conclusion is supported by local overturning of contacts to form a mushroom-shaped geometry (Harris Lake dome - Fig. 5, sections C-C", D-D" and Fig. 6, sections F-F" and I-I"). Because the flattening plane associated with such gneiss doming is horizontal in the crestal regions of the domes, the planar fabric there should probably be labelled S2-S3 as the two are coplanar.

The evolution of these alteration domes and synforms is considered to have been as follows. Weak lateral compression is thought to have produced low amplitude antiforms and synforms with the characteristic 110-290° trend. The cores of the antiforms then began to rise on their own as a result of gravitational instability. Horizontal shortening in the intervening synforms is interpreted to be result of a combination of true lateral compression and strain related to gneiss doming. Penetrative folding in areas removed from the domes as well as local upright folds on the crests of the domes, both having axial surface traces of 110-290°, are interpreted to indicate, however, that compressional deformation accompanied and locally outlasted "diapiric" unwelling (for example within Harris Lake dome). Given such behaviour the rocks can be assumed to have been relatively hot during the formation of these structures.

Specific gravity determinations were carried out on 223 hand specimens from the various stratigraphic units in the area. The results

are summarized in O.F. 1111, Appendix C (Fig. C1). The results of statistical analysis of the data (using two sample t-tests with pooled estimates of the common variance and excluding mafic and ultramafic components) indicate a 95% confidence level for the statements that the basement complex and the Dewar Lakes Formation form one population and that it is distinct from the Longstaff Bluff Formation, whether migmatitic or nonmigmatitic, the latter formation being approximately 0.05 (specific gravity) heavier. Combined with the presence of the dense Bravo Lake Formation (O.F. 1111, Fig. C1) at the base of the Longstaff bluff Formation and the probable effects higher temperatures at depth in accentuating this density instability it is not unreasonable to suggest that gravitationally-induced gneiss doming was a viable tectonic process during D3.

Structures trending 070-250° ("D4")

Mesoscopic folds of this orientation are generally well defined only in the central and marginal parts of Piling Basin, including the northwestern corner of the study area (as inferred from the characteristic grain of the scattered exposures in that area), and in the southern migmatitic paragneiss zone. Within the central gneiss dome belt structures with axial surface traces oriented at 070-250° are generally limited to the mesoscopic scale.

Some insight into the nature of the 070-250° phase may be obtained by examining the geometry of such structures elsewhere in the belt. At Longstaff Bluff on the western coast of Baffin Island (Figs.

1, 2) rare vertical cuts through this succession of folded Longstaff Bluff Formation strata (Morgan et al., 1975) reveal that folds of this generation have a penetrative axial planar cleavage whose surface trace is parallel to the general 070-250° trace of bedding (personal observations 1974). The very pronounced grain of Piling Basin there is the result of folding of laterally continuous beds of the Longstaff Bluff Formation into tight upright folds (Morgan et al., 1975) with subvertical axial planes and subhorizontal fold axes.

In most of the study area, where subhorizontal S2 was developed, folds of this generation are, in contrast to the case at Longstaff Bluff, characterized by the crenulation of S2. The latter type of deformation is particularly penetrative in the southern migmatitic paragneiss zone. There is little evidence, apart from the arch of schistosity in the core of Kakiartok Rapids dome (which may have originated during D3) that there was a component of diapiric upwelling of the basement complex during D4. This suggests a slightly cooler environment in which the rocks were less ductile and in which all folds are related only to lateral compression.

The relative age of post-D2 structures

Determining the relative age of the two sets of structures described above has been problematical for a number of reasons. In those zones where S2 was previously developed, both later deformations are expressed as a crenulation of S2. Interference between the two sets of generally upright structures is expressed as an irregularity of axial

surfaces since the flattening planes associated with the two sets are separated by only 40°. In other words this separation of axial surface traces is not sufficient to produce refolding, as opposed to rotation, unless the dip of the earlier axial surface becomes quite shallow. Stereonet plots (S3-S4 composites in Fig. 9) confirm the lack of bimodality of orientations of minor folds in zones such as the southern migmatitic paragneiss zone where overprinting might be expected. This relationship is interpreted to be due to the reorientation through rotation of folds of whichever generation is older, as well as to the nonlinearity of stress trajectories produced by the various domes and plutonic bodies.

In the study area there are only two localities at which the axial surface of one schistosity-deforming fold has been observed to fold the axial surface of another such fold. The presence of such an overprinting relationship has been noted in Figs. 3 and 4 and is reviewed below.

At station 179 north of South Jackson Lake dome, an isoclinal fold of schistosity has an axial planar orientation of 258 14 (strike and dip with strike azimith measured with dip to the right) and is folded by an open fold with an axial plane orientation of 240 80. Although the relationships in this area are complicated by the inferred presence of Tullugak Lake dome in the subsurface, producing a deviation in the axial surface traces of the shallowly-dipping earlier structures, the sequence of folding is clear. At station 284, south of Harris Lake dome, a fold with an axial plane orientation of 117 68 is deformed by an

open fold with an axial plane orientation of 055 77. In both cases the indicated sequence of superimposition is that the trend of 110-290° is older and is overprinted by folds with a trend of 070-250°.

In those zones where S2 was not previously formed, outcrops in the study area indicate the presence of only one penetrative axial planar cleavage, for example in the core of North Jackson Lake synform (regional trend 110-290°, Figs. 3 and 8) or the northwest corner of the study area (trend 070-250°). It would appear that the angle of only 40° between the upright axial planes precludes the crenulation of the older post-D2 cleavage, whichever it is, during the younger event. The response would logically be the rotation of the older cleavage toward the flattening plane of the later deformation.

In addition to these rare refolding relationships observed within the study area there is both indirect and circumstantial evidence to support the relative sense of timing proposed in Tippett (1980). The indirect field evidence is presented first, followed by the circumstantial.

The location of folds with the trend 070-250° relative to macroscopic structures trending 110-290° may provide some insight into the problem of relative timing. Between stations 955 and 956, near the narrowest constriction of Piling Group units between Harris Lake and FitzGerald Lake domes (Fig. 4), crenulations with axial plane orientations of 250 60 and 251 58 were observed. Just to the west near stations 822 and 823, similar crenulations with axial planes oriented at 255 47 and 257 58 were observed. All these crenulations are developed in rocks which must have been penetratively deformed during the formation of

Logan Lakes synform which trends at roughly 110-290°. It is considered unlikely that these crenulations would have been preserved in their original orientation during the formation of the synform and it is therefore concluded that they were imposed on a pre-existing macroscopic geometry. This indicates that the 110-290° trend predates that of 070-250°.

Near the southeastern corner of FitzGerald Lake dome (Fig. 4), a macroscopic synform exists with associated minor folds of the appropriate asymmetries having axial plane orientations of 073 75, 040 72 and 075 58. This structure deforms the edge of the dome, as indicated by the presence of minor folds with axial surface traces parallel to the dome margin that maintain a consistent relationship to the dome margin around the closure of the synform. The late synform also appears to be associated with the presence of a generation of minor folds with similar orientations within FitzGerald Lake dome and in the eastern portion of Logan Lakes synform, that is to the west, northwest and north. It has been suggested above that the presence of large areas of gently dipping basement gneisses, such as the eastern end of FitzGerald Lake dome, is related to a component of diapiric upwelling during the generation of structures trending 110-290°. The presence of some minor folds trending 110-290° within this zone is attributed to minor compression after active diapirism but it is considered unlikely that the folds and crenulations with the trend 070-250° could have survived through the entire diapiric phase if that trend were older. These folds and crenulations are therefore interpreted as having been superimposed entirely after diapirism, again indicating that

the 070-250° trend is younger.

The circumstantial evidence in support of the interpretation that the 110-290° trend is older is based on the assumption of a simple thermal history. D2 was a synmetamorphic event which appears to have occurred, based on textural relationships, while the rocks were at their highest temperatures. A variety of structural features (discussed above) indicate that gravity tectonics produced a limited degree of gneiss doming during the 110-290° event. This involved the buoyant rise of the less dense basement gneisses into their mantle of denser metasediments. From this it can be concluded that the rocks must have been relatively hot to have ductilities which would permit such a process. The geometry of the 070-250° structures where observable cannot be demonstrated to be related to any type of gravity tectonics. In addition, pervasive retrogression in those areas characterized by high grades of metamorphism is often associated with zones penetratively folded during the 070-250° event. Thus for the sake of simplicity in the thermal history of the belt, one can argue that the two high temperature events (D2 and 110-290°) were probably followed by a lower temperature event (070-250°). The alternative would be a scenario in which the rocks were heated (synmetamorphic D2), cooled (070-250° event), reheated (110-290° event with gneiss doming) and finally cooled. This is not to say that such a sequence is impossible but rather that it introduces complications which could otherwise be avoided.

In conclusion, it can be stated that most direct and indirect field evidence as well as circumstantial evidence support the sequence

of events suggested here. The figures contained in this open file therefore have been annotated with the 110-290° event being D3 and the 070-250° event being D4. These conclusions are not definitive, however, and the alternative hypothesis that the 070-250° event is indeed older should certainly be considered a reasonable possibility.

FitzGerald Lake dome "break"

There appears to be a major fault expressed as a strong air photo lineation, which truncates the southern margin of FitzGerald Lake dome. This conclusion is supported by the discordance of structural trends (Fig. 4) and of metamorphic zonation (Fig. 11) across the break. Well developed retrograde mineralization (epidote, chlorite) and alteration of feldspars characterize the fault zone which is considered to be post-D4 in age.

TIMING OF METAMORPHISM RELATIVE TO DEFORMATION

Introduction

Fabric relationships relevant to the timing of metamorphism with respect to deformation indicate that a single phase of prograde metamorphism accompanied nappe emplacement (D2) and the development of its related axial planar schistosity (S2). This is in accordance with both textures observed in thin section and relationships observed in the field. There does not appear to be any fabric associated with D1 although this could be interpreted as being the result of the severity of D2.

Porphyroblast growth

The relationships of porphyroblast growth relative to deformation is not obvious as porphyroblasts generally do not contain well developed inclusion trails. In cases where such trails are present, their relationships to the external schistosity are often not clear-cut.

The timing of porphyroblast growth relative to D2 is difficult to determine as S2 and S3 are often roughly coplanar on the tops of domes whereas towards the flanks S2 tends to become highly crenulated. Only in zones where D2 structures have been relatively isolated from the effects of D3 and D4 can the timing be well established. In order to obtain an unambiguous answer it is also necessary to observe the fabric relationships of mineral growth at an isograd. A zone meeting both qualifications is the closure of Casson Lake nappe southwest of Casson

Lake. The relationship existing at station 971 is that andalusite and sillimanite are involved in a reaction based on the inversion of aluminosilicate polymorphs. The development of sillimanite at only the contact points of adjacent andalusite porphyroblasts, where slightly higher pressures can reasonably be expected to have existed during D2, indicates that the end of D2 and the peak of metamorphism were virtually synchronous. This conclusion is based on the reasoning that if deformation had outlasted the peak of metamorphism, the development of sillimanite would not be expected, while if metamorphism had outlasted deformation, sillimanite would not show a spatial correlation with the contact points of adjacent porphyroblasts.

It was observed that mineral grains which nucleated prior to the peak of metamorphism which are within the zone of stability of their assemblage were strained during D2 whereas new mineral grains which were growing at an isograd at the peak of metamorphism were not strained during D2. The available evidence (also see Tippett (1980)) indicates, therefore, that prograde porphyroblast growth was syn-D2 and ceased at the end of D2.

Syntectonic recrystallization and the growth of retrograde porphyroblasts occurred during D3 and D4. Evidence is the presence of syn-D3 muscovite which has overgrown crenulated S2. Mineral growth during D3, including the formation of axial planar schistosity in Domains E, K and O (Fig. 9) is interpreted to be the result of a combination of the nucleation of strain-free grains at the expense of strained grains and the actual retrogression of prograde assemblages under conditions less than those of the peak of metamorphism.

Relationship of mobilizate to deformation

Partial melting is an integral part of prograde metamorphism in the area. The timing of the related mobilizate generation with respect to deformation is therefore of interest and constitutes a check on the timing based on porphyroblast growth.

Ductile boundinage of mobilizate in the crestal regions of D3 structures is difficult to evaluate as it could be the result of either D2 or D3. More diagnostic is the relationship of allochthonous dykes of mobilizate in Domains E, K and O (Fig. 9) where, because the rocks did not experience D2, the timing of intrusion gives a definite indication of the timing of mobilizate generation relative to D3. In these domains, mobilizate dykes were intruded before D3 and were folded during D3, in some cases with the development of angular boudins. This is definite proof that metamorphism predated D3 as there is little reason to doubt that these dykes are related to anatexis at slightly deeper levels.

The relationships along the northern margin of South Jackson Lake dome could be interpreted as being the result of post-D2 metamorphism in that partial melting of the Longstaff Bluff Formation is well developed only under Casson Lake nappe whose emplacement would therefore seem to predate the metamorphic peak. An alternate and favoured interpretation of this relationship is that the nappe developed at the top of the zone of migmatization whose presence represented a major mechanical weakness in the stratigraphy and facilitated movement of the nappe.

The second diagnostic feature with respect to the timing of partial melting is the identity of the schistosity developed in the

mobilizate. Throughout the main part of the area that experienced partial melting, mobilizate did not develop a schistosity during D2 but rather occurs as texturally massive segregations between folia of restite which contain a penetrative schistosity (S2). In highly migmatized zones such assemblages of massive mobilizate and interleaved restite were folded over D3 and D4 closures accompanied only by the development of crenulation cleavage in the restite. The superimposition of a penetrative schistosity on the mobilizate during post-D2 deformation occurred only in those areas marginal to nonmigmatized zones. Schistosity (S3 and S4) in these rocks is labelled on the basis of the same arbitrary criterion used to evaluate the identity of D3 and D4 minor folds (the 090-270° break). The presence of such schistosities in deformed mobilizate confirms that the peak of metamorphism did predate both D3 and D4.

Retrogression

Retrogression is widespread and includes extensive pinitization of cordierite, chloritization of biotite, saussuritization of plagioclase and serpentinization of olivine.

Unannealed strain and brittle deformation

In virtually every thin section examined from the area there is evidence of unannealed strain in mineral grains. Undulose and polygonized quartz, locally accompanied by the development of slip lines and serrate to lobate grain boundaries, is of regional distribution. Kinking of phyllosilicates and brecciation of plagioclase feldspar, hornblende,

epidote and garnet have been observed. Some calcite grains contain deformation twins.

Relationship of retrogression and strain

The phenomena of retrogression and late strain are closely related in the area. On the basis of thin section observations there appears to be an empirical correlation between the degree to which mineral grains are strained and the degree to which they are altered.

In many areas of migmatitic paragneiss, there is a strong field correlation of highly retrograded greenish (as opposed to the usual grey to black) "anomalous" paragneiss and associated red (as opposed to the usual white) mobilizate with bodies of mafic and ultramafic rocks. Although the relationship may be metamorphic in origin, it more likely results from the concentration of strain along the margins of these rigid bodies during D3 and D4. The presence of chlorite and epidote in areas of the southern migmatitic paragneiss zone where D4 folding has been particularly penetrative, for example marginal to the major folded amphibolite east of Lismer Lake, suggests that between D3 and D4 the temperature had dropped to that of greenschist facies conditions.

METAMORPHIC PETROLOGY

Introduction

Assemblage data for pelitic and quartzofeldspathic rocks of the Piling Group are presented in Fig. 11. One "isograd", based on the appearance of closely associated granitoid segregations in the metasediments, is defined by the distribution of these segregations in the field. Based on these data, it has also been possible to draw two other lines, more properly termed isograds according to common practice, across which the incompatibility of mineral pairs can be demonstrated. Several other assemblages are incompatible with those more commonly observed but are too scattered in distribution, probably as a result of bulk composition controls, to permit the drawing of an isograd.

The assemblages have been interpreted in terms of pressure and temperature using Fig. 12.

Partial Melting

The first major isograd to be considered is identifiable in the field and involves the appearance of segregations of white, medium- to coarse-grained granitoid material in close association with the metasediments (Fig. 11). The composition of these granitoid material is depicted in O.F. 1111, Fig. A4.

As the zone in which migmatization exists is approached from nonmigmatized rocks (which contain rare and obviously discordant dykes) there is usually a transitional band within which such granitoid material

exists but has the form of well defined dykes and sills of variable thickness. These appear to have been intruded into the metasediments. The isograd therefore must be defined on a subjective basis, namely at the point at which the metasediments and granitoid material are so intimately intermixed that an origin by injection is unlikely.

Appearance of coexisting K-feldspar and sillimanite

This isograd is defined on the basis of the first appearance of coexisting aluminosilicate and K-feldspar. Within the area, the relevant aluminosilicate is always sillimanite. The appearance of K-feldspar and sillimanite is interpreted to be related to the breakdown of muscovite in the presence of quartz. The isograd is place at the first occurrence of the product assemblage.

It is not possible to identify the isograd in the field due to the fine grain size of some of the rocks involved and the identical white colour of K-feldspar and plagioclase in the restite, particularly in the Longstaff Bluff Formation. The control points were therefore established on the basis of the identification of sillimanite in thin section and of K-feldspar in a stained slab from the same sample. Contact between K-feldspar and sillimanite has not been documented in most cases but given the presence of extensive retrogression, this is probably not common in any case. The coexistence of K-feldspar and sillimanite at the peak of metamorphism is taken to be established by the presence of sillimanite in a rock whose matrix contains significant and well distributed K-feldspar.

Knots or pods of quartz containing fine sillimanite ("faserkiesel") have been observed in many metamorphic belts but their origin is enigmatic. In this area, these pods are locally well developed in the quartzofeldpathic beds of the Longstaff Bluff Formation where their variation in abundance helps to define bedding. They locally make up 30% of these rocks. The more pelitic portions of beds contain a higher concentration of the segregations, producing a pseudograding with "tops" in the opposite sense to that defined structurally and stratigraphically. Similar pods are locally well developed in the Dewar Lakes Formation.

Within the faserkiesel, whisps of sillimanite are intergrown with quartz. Coarse muscovite often overgrows the pods and may contain small clusters of sillimanite isolated in quartz. In hand specimen, the pods have a strong preferred alignment in S2 and are developed on a variety of scales. In some rocks the pods are associated with muscovite. These rocks grade into those containing only coarse porphyroblastic muscovite.

Between the K-feldspar-sillimanite zone and the muscovite-quartz zone there is a zone in which all four phases coexist. By definition this falls within the K-feldspar isograd. It is these areas which are characterized by the presence of faserkiesel. In addition, rocks in certain areas just down grade from the isograd also contain this texture or textures apparently gradational with it.

It was suggested in Tippett (1980) that the presence of faserkiesel indicates the former presence of a stable K-feldspar-sillimanite assemblage which has been subjected to retrograde metamorphism through a

mechanism involving paired ionic equilibria. The former prograde position of the K-feldspar-sillimanite isograd can therefore be estimated on the basis of the distribution of this faserkiesel texture, as indicated on Fig. 11.

Appearance of coexisting garnet and cordierite

This isograd is defined by the first occurrences of coexisting garnet and cordierite. Although cordierite is often highly to completely retrograded, its distinctive alteration products allow it to be identified as an original phase.

This isograd is less continuous than the previous two and the high grade products are interpreted to occur in four separate zones. The appearance of coexisting garnet and cordierite is thought to be related to the following reaction:

sillimanite + biotite + quartz + plagioclase =
garnet + cordierite + K-feldspar + melt + H_2O

Pressure and temperature conditions

Assemblages observed in the area permit the pressure and temperature of metamorphism to be estimated on the basis of Fig. 12. The isograds defined in the previous sections and illustrated in Fig. 11, together with several more scattered observations, delimit the metamorphic conditions in the central and southern portions of the area. Assemblages observed within the Dewar Lakes Formation indicate that the breakdown of staurolite (reaction 5, Fig. 12) occurred in one case

within the stability field of andalusite (station 971, Casson Lake nappe, southwest of Casson Lake) and in another case within the stability field of sillimanite (stations 15 and 55, near Dewar Lakes dome). The conditions at those locations would be constrained to roughly 550°C and 300 MPa. According to the terminology of Carmichael (1978), this relationship indicates that the boundary between bathozones 2 and 3 occurs somewhere in the area.

In the case of anatexis and of the appearance of coexisting sillimanite and K-feldspar, isograds constructed on the basis of the available data are reasonably well constrained within South Jackson Lake synform. The isograds cross near South Jackson Lake (Fig. 11), producing two alternate sequences of transformations (Fig. 10). To the northwest of the intersections anatexis (reaction 1) preceded muscovite breakdown (reaction 2), while to the southeast of the intersections anatexis (reaction 1) either has not occurred or occurred only after muscovite breakdown (reaction 3). This relationship constrains the conditions at the intersections of the isograds to roughly 660°C and 350 MPa. According to Carmichael (1978), as modified by St-Onge and Carmichael (1979), these intersections are points which lie on the isobaric plane separating bathozones 3a and 3b and the conditions at them are invariant in terms of Fig. 12.

The only restriction on the high end of the metamorphic grade in the area is that the reaction of biotite and garnet to produce cordierite and hypersthene (reaction 6, Fig. 12) has not occurred. This would appear to limit the temperature in the study area at the time of metamorphism

to less than 720°C at a projected pressure of 380 MPa.

Phase relations in mafic, ultramafic and calcareous rocks are compatible with the conditions from pelitic assemblages but in general do not provide as accurate a calibration.

Geometry of metamorphic isograds

Examination of Fig. 11 illustrates that there is an increase in metamorphic grade from northwest to southeast across the study area. The grade increases on the northwest flank of the D3 dome belt, as expected from the progressively deeper stratigraphic level, but continues to increase to the southeast in spite of the fact that the stratigraphic level within the southern migmatitic paragneiss zone is higher than that in the gneiss dome belt. As such the isograds can be inferred to dip towards the northwest. Reference to Fig. 10 (adapted from Thompson, 1977) indicates that such a dip of both melting and dehydration reactions can be expected on the flank of a thermal dome produced by a higher geothermal gradient to the southeast than to the northwest.

The presence of a thermal dome is confirmed by reference to Fig. 10 and the map distribution of isograds presented in Fig. 11. In Fig. 10 it is apparent that not only do the isograds dip away from the thermal culmination but they intersect as well, producing different sequences of reaction, as seen in vertical section, across the margin of the thermal dome. It has been demonstrated that such a contrast of sequences exists in the area. This pattern is well defined in the case of South Jackson Lake synform and is suggested on the basis of faserkiesel

distribution in North Jackson Lake and Logan Lakes synforms as well. Rocks in which the appearance of coexisting K-feldspar and sillimanite precedes anatexis occur to the southeast of the area in which the more general reverse sequence occurs. With reference to Fig. 10 and in particular its schematic portion, it can be seen that the pre-anatexis appearance K-feldspar-sillimanite occurs closer to the centre of a thermal dome.

Postmetamorphic deformation has produced the present distribution of isograds. The existence of low grade metasediments in Piling Basin appears to be the result of a combination of a primary northwestward dip on the isograds and postmetamorphic D4 compressional deformation.

REGIONAL TECTONIC EVOLUTION

The reader is referred to Tippett (1980) for a discussion of the significance of those features observed in the study area in the context of the regional tectonic setting. It is there concluded that the possible role of plate tectonic forces in the evolution of the Foxe Fold Belt remains enigmatic.

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