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PRECAMBRIAN GEOLOGY OF PARTS OF TAVANI, MARBLE ISLAND, AND CHESTERFIELD INLET MAP AREAS, DISTRICT OF KEEWATIN: A PROGRESS REPORT

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PRECAMBRIAN GEOLOGY OF PARTS OF TAVANI, MARBLE ISLAND, AND CHESTERFIELD INLET MAP AREAS, DISTRICT OF KEEWATIN: A PROGRESS REPORT

Abstract

The region is predominantly underlain by the Rankin Inlet Group, an (?) Archean greenstone-metasedimentary sequence comprising at least two major cycles of volcanism with intervening turbidite deposition. Ultramafic and gabbroic sills occur within the volcanic sequence. Three periods of deformation and one event of metamorphism can be recognized within the Rankin Inlet Group. The rocks within the group form an east dipping F_1 homocline that is folded into a southeast plunging F_2 upright syncline. Shallow, northeast and southwest plunging pre-F_2 mineral stretching lineations are developed in the metavolcanic rocks on the limbs of the F_2 syncline.

Strain ratios based on selvage thicknesses from deformed lava pillows of the Rankin Inlet Group range from 2.0 to 6.0; the high ratios commonly coincide with northwest trending, southwest verging ductile shear zones – tentatively interpreted as high angle (? thrust) faults. Paleohorizontal markers such as "lava-shelves" are reliable bedding and facing markers where axial ratios in pillows are less than 4:1.

Stratigraphic and structural facing reversals suggest two structural discordances, not coincident with ductile shear zones, within the upper volcanic cycle of the Rankin Inlet Group. The discordances may be interpreted as two thrusts with opposing vergence (northeast and southwest) that juxtapose rocks of different structural geometry. The orientation of pre F_2 mineral lineations supports the model, but further field testing is required.

Migmatitic paragneiss with kyanite, sillimanite, and staurolite assemblages, mafic and felsic schists, and metamorphosed iron formation are exposed in the region north and northeast of Rankin Inlet. They are believed to be the metamorphic equivalents of the Rankin Inlet Group. A weakly deformed two-mica granite intrudes the gneiss complex.

A (?) Lower Proterozoic orthoquartzite sequence, tentatively correlated with the Hurwitz Group, is exposed on Kudlulik Peninsula and on Marble Island 45 km east of Rankin Inlet in Hudson Bay. The contact with the older rocks on Marble Island is a high-strain zone, and is interpreted to be a low angle, south verging fault.

Résumé

La région repose en grande partie sur le groupe de Rankin Inlet, une suite de roches vertes et roches métasédimentaires archéennes(?) qui comporte au moins deux grands cycles de volcanisme séparés par une période de sédimentation turbiditique. Des filons-couches ultramafiques et gabbroïques se trouvent dans la suite volcanique. Trois périodes de déformation et une période de métamorphisme sont identifiées dans le groupe de Rankin Inlet. Les roches du groupe forment une structure monoclinale F_1 est qui a été plissée pour former un synclinal droit F_2 à inclinaison sud-est. Les roches métavolcaniques sur les flancs du synclinal F_2 font voir des linéations minérales pré-F_2 peu profondes, à inclinaison nord-est et sud-ouest.

Le taux de déformation fondé sur l'épaisseur de la salbande des laves en coussins déformées du groupe de Rankin Inlet varie de 2,0 à 6,0; les chiffres élevés coïncident généralement avec des zones cisaillées ductiles, à déversement sud-ouest et à direction nord-ouest, qui sont provisoirement interprétées comme étant des failles (chevauchantes?) à angle presque vertical. Les repères paléohorizontaux comme les «plates-formes de lave» sont des repères fiables de la stratigraphie et du regard, là où les paramètres cristallographiques des coussins sont inférieurs à 4:1.

L'inversion du regard stratigraphique et structural porte à croire qu'il existe, dans le cycle volcanique supérieur du groupe de Rankin Inlet, deux discordances tectoniques qui ne coïncident pas avec les zones cisaillées ductiles. Ces discordances représenteraient deux chevauchements à déversement opposé (nord-est et sud-ouest) qui ont juxtaposé des roches à géométrie structurale distincte. L'orientation des linéations minérales pré-F_2 confirme ces observations, mais il faudra entreprendre d'autres travaux sur place.

Un paragneiss migmatique composé d'assemblages de cyanite, de sillimanite et de staurotide, des schistes mafiques et felsiques et une formation de fer métamorphisée affleurent dans la région au nord et au nord-est de Rankin Inlet. Ces roches seraient l'équivalent du groupe de Rankin Inlet, sauf qu'elles ont été métamorphisées. Un granite à deux micas faiblement déformé fait intrusion dans le complexe gneissique.

Une suite de quartzite sédimentaire du Paléozoïque inférieur(?), provisoirement mise en corrélation avec le groupe de Hurwitz, affleure dans la péninsule Kudlulik et dans l'île Marble, 45 km à l'ouest de Rankin Inlet dans la baie d'Hudson. Dans l'île Marble, le contact avec les roches plus anciennes est une zone fortement cisaillée qui représenterait une faille à angle faible et à déversement sud.

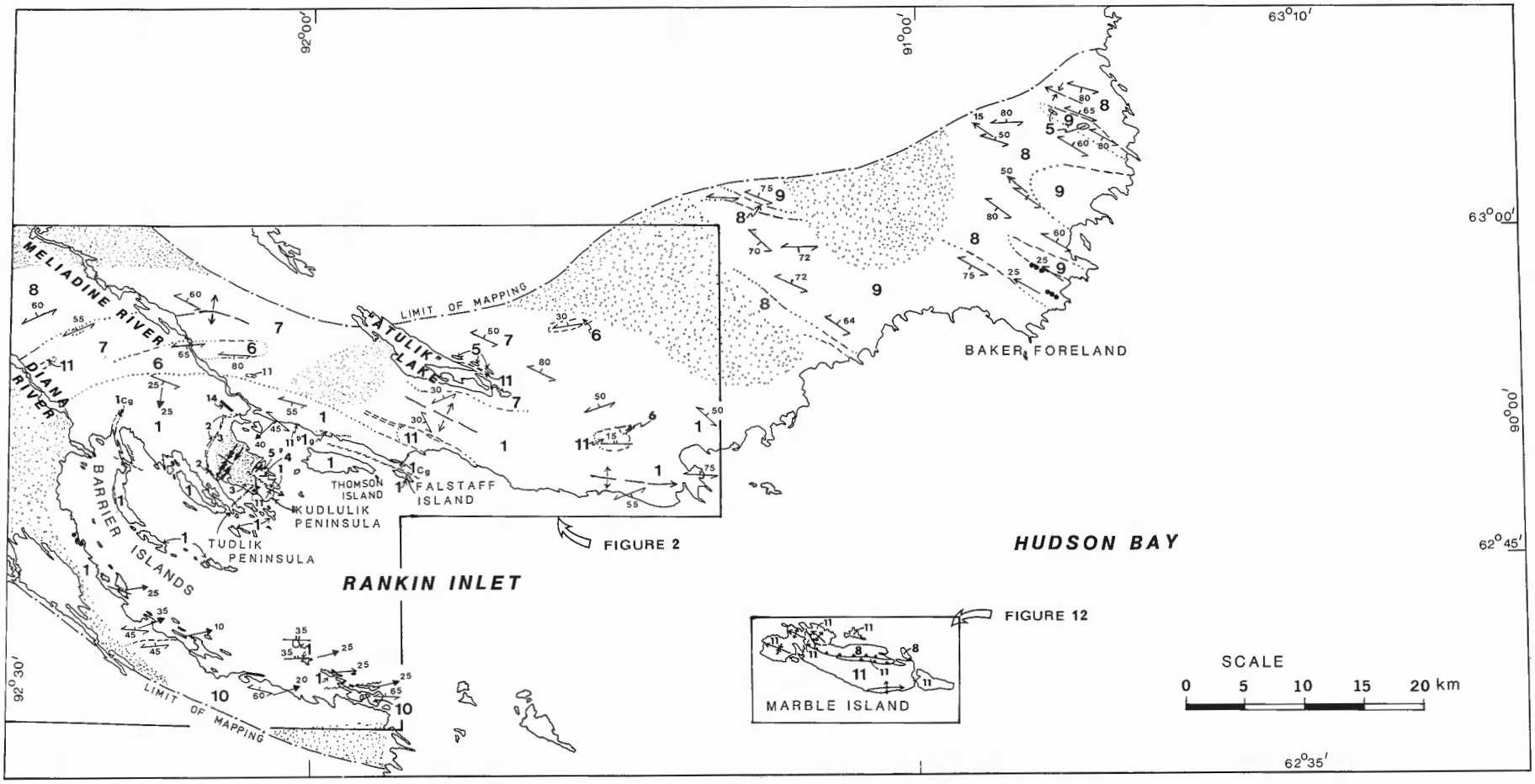

Figure 1. Geological sketch map of parts of Tavani Marble Island and Chesterfield Inlet map areas, District of Keewatin (see Figure 2, in pocket, for legend).

INTRODUCTION

Bedrock mapping in parts of Tavani (NTS 55 K/9, 16), Marble Island (55 J/11 to 15), and Chesterfield Inlet (55 O/2,3) map areas, compiled at scales of 1:250 000 and 1:50 000, was completed during the 1985 field season. Marble Island was mapped at a scale of 1:20 000. The objective was to provide a comprehensive study of the map areas in order to refine and update the previous reconnaissance mapping of the region, and to examine in detail conflicting structural/stratigraphic interpretations of the Rankin Inlet region by previous workers. This report summarizes preliminary results of the first of three seasons field work. G.J. Borradaile concentrated on strain determination in tectonically deformed pillow structures from the Rankin Inlet Group metabasalts.

Previous work

Early geological investigations in the region were carried out by Bell (1885, 1887) and Tyrrell (1898). The map areas are part of a region mapped by Lord (1953), Weeks (1932) and Wright (1955, 1967), and published at a scale of one inch to eight miles. Heywood (1973) mapped the Tavani map area at a scale of 1:250,000. The geology of the Rankin Inlet region (55 K/9, 16; J/12, 13) was the subject of masters' theses by Bannatyne (1958) and by Laporte (1975). Bedrock mapping and other topical studies concerning stratigraphy, structure, and economic potential of the Rankin Inlet region were carried out by several workers (Bell, 1968, 1970a, b; Drybrough, 1931; Hudec, unpublished; Laporte, 1983; Laporte and Frape, 1973). The most recent comprehensive review of previous geological investigations in the region is given by Laporte (1983).

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Part 1: LITHOLOGY, STRUCTURE AND STRATIGRAPHY, RANKIN INLET REGION

Subhas Tella, I.R. Annesley, and J.R. Henderson

GENERAL STATEMENT

The distribution of rock units in the study area is shown in Figures 1, 2, 12, and 13, which are accompanied by a common legend (see Fig. 2, in pocket). The area is underlain by an (?) Archean greenstone-metasedimentary sequence, the Rankin Inlet Group (units 1 to 5) and its metamorphic equivalents (units 6 and 7), and by (?) Lower Proterozoic gneissic, schistose, and migmatitic rocks and younger granitoid intrusions (units 8 to 10). A (?) Lower Proterozoic orthoquartzite sequence (unit 11) unconformably overlies the migmatitic rocks. Minor granitic intrusions (unit 12), and Helikian biotite lamprophyre (unit 13) and gabbro (unit 14) dykes transect the region. Some of the stratigraphic relationships and structural interpretations of the region presented in this report, differ from those of previous workers (Bannatyne, 1958; Hudec, unpublished; Laporte, 1973, 1983), and some conclusions drawn are preliminary.

RANKIN INLET GROUP (UNITS 1 TO 5)

The name Rankin Inlet Group was proposed by Bannatyne (1958) for the metavolcanic and metasedimentary rocks exposed near Rankin Inlet (Fig. 2). According to Bannatyne (1958) it is composed of two volcanic sequences, lower and upper, each consisting of various proportions of basaltic, andesitic, and dacitic flows and interflow sediments. The two sequences are separated by a conformable sedimentary sequence consisting of dolomite, locally ripple-marked quartzite, argillite, and derived schists which is exposed in an arcuate belt on Kudlulik Peninsula (Fig. 2). The upper volcanic sequence is exposed on Kudlulik Peninsula southeast of Nipissak Lake, and on Thomson and Falstaff islands (Fig. 2). The age of the Rankin Inlet Group is controversial. Wright (1967) and Bell (1968) believed that rocks of both Archean and Aphebian age are present in the Rankin Inlet region. Bannatyne (1958) and Heywood (1973) considered the Rankin Inlet Group to be Archean, whereas Hudec (unpublished) and Laporte (1975, 1983) assigned an Aphebian age. Heywood (1973) correlated these rocks with the Archean Kaminak Group (Davidson, 1970a). Based on whole rock major element geochemical studies Laporte (1975, 1983) concluded that the volcanic rocks of the Rankin Inlet Group are similar to oceanic tholeiites. Laporte noted no significant chemical differences between the lower and upper volcanic sequences of Bannatyne (1958), and he correlated all of the metavolcanic rocks with the Aphebian Hurwitz Group of Bell (1968) on the basis of chemical similarities to the volcanic rocks of the Hurwitz Group (Ridler, 1974), and on stratigraphic/structural interpretations. Laporte (1975, 1983) postulated two phases of folding – a first phase characterized by northwest trending recumbent isoclinal folds formed by gravitational sliding, and a second phase characterized by east-southeast plunging upright folds. This structural interpretation of the region was primarily based on facing directions determined from the orientation of "lava-shelves" ("eyebrow"[1] structures of Laporte, 1975, 1983) in pillowed metabasalt, structures which are believed to represent geopetal paleohorizontal markers (lava-levels or drain-away structures; Ballard and Moore, 1977). A discussion of these "lava-shelf" structures is presented in Part 2 of this report where their significance and limitations as paleohorizontal and stratigraphic-facing markers are noted. Our interpretation of the stratigraphy and structure of the Rankin Inlet Group differs from that of the previous workers. For a detailed account of lithologic and petrographic descriptions of various rock units the reader is referred to Bannatyne (1958) and Laporte (1975, 1983). The revisions suggested in this report, based on field observations only, are summarized in the sections to follow, together with the preliminary conclusions. The areal distribution and structure of the Rankin Inlet Group is shown in Figures 1 and 2. The orientation of fabric elements is shown in Figure 6. It should be pointed out that the structural geometry of the region has not been fully resolved.

The Rankin Inlet Group is composed of deformed and metamorphosed successions of massive and pillowed mafic volcanic flows (unit 1 m,p; Fig. 2) and interflow sediments (greywacke, 1 g; quartz-magnetite iron formation, 1if), and undifferentiated mafic and felsic tuffs, pyroclastics and volcanic breccia. At least two major cycles of volcanism with an intervening cycle of turbidite deposition is recognized in the Rankin Inlet region.

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[1] The term "eyebrow" was previously used by Burrows and Rickaby (1935) to document a "peculiar" structure consisting of crescent-shaped quartz-filled bodies in acid volcanic rocks of the Sudbury Nickel Field. To avoid confusion we will use the term "lava-shelf" instead of "eyebrow" structure.

The metavolcanic rocks intercalated with thin turbidite beds, and minor banded iron formation, exposed to the west and south of Barrier Islands, Buttress Islands, and Smooth Islands, and to the north of Thomson Passage, represent the lower volcanic cycle. The distribution of iron formation is shown by a linear magnetic anomaly (Geological Survey of Canada, 1966a). These rocks appear to be conformably overlain by a highly sheared, discontinuous unit of polymictic conglomerate (unit 1cg, Fig. 2, 3) consisting of abundant clasts of granite, white orthoquartzite, and dolomite, with minor amounts of slate, tuff, and black chert set in a relatively finer grained greywacke-type matrix. The conglomerate defines a discontinuous arcuate band which is exposed along the west shore of Barrier Islands, northwest of Prairie Bay, and north of Falstaff Island. The conglomerate (unit 1cg) locally grades upwards into a well bedded and graded greywacke. Abundant blue-quartz grains are disseminated througout the conglomeratic greywacke, making it a distinctive marker horizon in the region. The best exposures of turbidites are observed on the southern margins of Ahigik, Kango, Hump, and Buff islands (Fig. 2, 4).

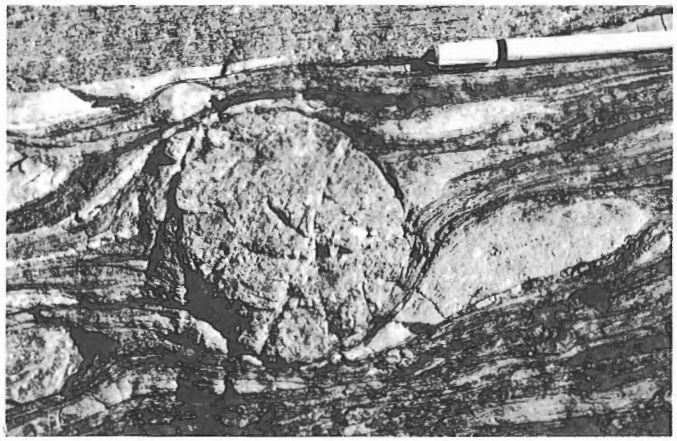


Figure 3. Polymictic conglomerate (unit 1cg) composed of rounded granite, and ellipsoidal quartzite clasts in a sheared matrix. View looking east-southeast at a vertical face. Stretching lineation is subparallel to the face (trends 015-025). Location: west shore north of Barrier Islands (see Fig. 2). GSC 204308-S

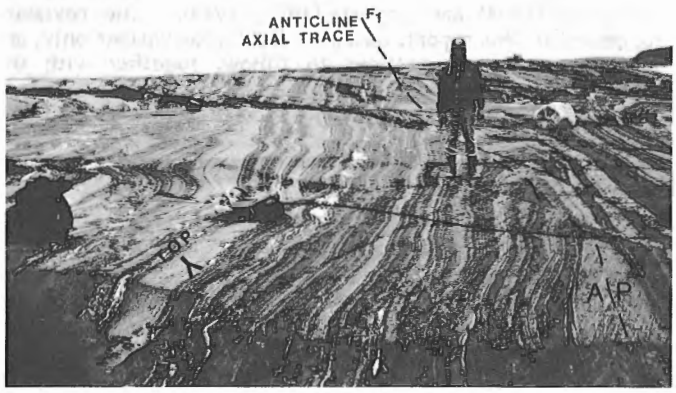


Figure 4A. Tightly folded turbidite sequence (unit 1g) on Buff Island, Rankin Inlet Group. View looking east-northeast (see Fig. 2 for location). AP-axial plane trace of an anticline. GSC 204308-V

The second or upper cycle of massive and pillowed volcanic rocks and interflow sediments overlies the conglomerate and greywacke (1 cg,g) with apparent conformity. Rocks within the sequence are extensively exposed between Barrier Islands and Falstaff Island, and the contact between the volcanics and interflow sediments with the underlying conglomerate and greywacke is well exposed along the west shore of Barrier Islands. The contacts between the rocks of the upper volcanic sequence and the conglomerate-greywacke, exposed on numerous islands south of Kudlulik Peninsula, are highly sheared and it is not possible to correlate the units with certainty. The pillowed flows in the upper volcanic cycle contain abundant "lava-shelves" ("eyebrow" structures of Laporte, 1975, 1983). Although pillowed flows are present in the rocks of the lower volcanic cycle on Broken Islands, and on the islands northwest of Pangertot Peninsula (Fig. 2), none contain recognizable "lava-shelves". Pillowed flows with "lava-shelves", however, do occur structurally below the conglomeratic greywacke (1 cg) north of Falstaff Island.

Northwest of Melvin Bay, the massive and pillowed volcanic rocks (unit 1 m,p; Fig. 2) are conformably overlain by an interbedded sequence of dolomite (unit 2) and laminated impure quartzite, slate, argillite, minor quartz-pebble conglomerate and calcareous sedimentary rocks (unit 3). Primary structures in pillowed volcanics, and trough crossbeds in impure quartzite suggest that the sequence is right-way up and faces to the northeast. East of Melvin Bay the crossbedded (Fig. 5A) impure quartzite (unit 3) is overturned and faces southwest. The trend of the impure quartzite and the greywacke (units 3, and 1 g) is apparently truncated to the east by the greenstone-sedimentary sequence (unit 1m, p), and the contact appears to be tectonic. "Ramp" structures (Fig. 5B) transecting the bedding are well developed in the impure quartzite (unit 3) east of Melvin Bay, and abundant secondary carbonate concentrations are present along the planes of the "ramps".

Layered intrusions of serpentinite (unit 4) and gabbro (unit 5) occur as sills within the mafic volcanic sequence of the Rankin Inlet Group. Bannatyne (1958) described an ore deposit (now an abandoned mine) consisting of magmatically segregated Cu-Ni sulphides that is associated with one of the ultramafic sills near the settlement of Rankin Inlet (Fig. 2).

Laporte (1975, 1983) interpreted the structure of the Rankin Inlet Group to be a series of five northwest trending isoclinal recumbent folds that were refolded by east-southeast trending upright folds. The isoclinal character of

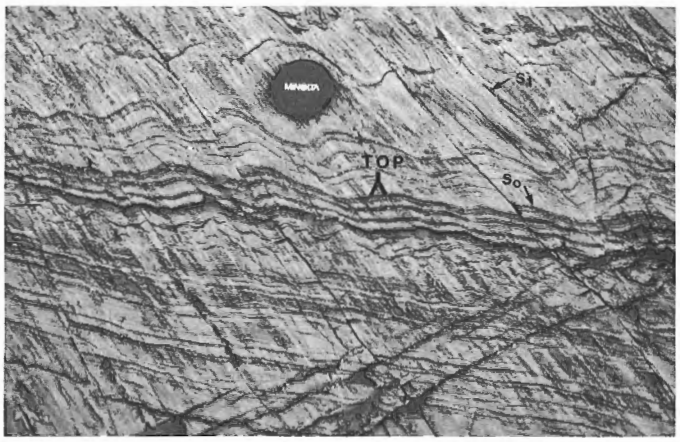


Figure 4B. Bedding (S_0), cleavage (S_1) relations in the turbidite sequence (unit 1g) on Buff Island. View looking east; orientation of erosion surface subhorizontal (see Fig. 2 for location). GSC 204308-F

the first folds was determined from the orientation of "lava-shelves". The facing (top) directions of the flows were determined from these structures without regard to the strain undergone by the pillows. That the "lava-shelves" (Fig. 5C) are reliable stratigraphic-facing indicators, at low strains, can be demonstrated in the Rankin Inlet Group where interflow sediments such as lapilli tuffs (Fig. 5D) give an independent accurate top determination and confirm the stratigraphic way-up directions in the pillows independently of "lava-shelf" orientation. At strains higher than 4:1, however, the "lava-shelves" progressively flatten and rotate into the plane of regional schistosity and do not provide reliable top determinations although the pillow may still yield the structural-facing. In Part 2 of this paper it is shown that the strain undergone by the Rankin Inlet Group rocks is high enough in most places to render the orientation of "lava-shelves" unreliable for top determinations, and hence the structural interpretation of Laporte (1975, 1983) is unsubstantiated.

Equal-area lower-hemisphere projections of fabric elements from metavolcanic and metasedimentary rocks of the Rankin Inlet Group are shown in Figure 6 (A to E). The regional distribution of lithologies is determined by an F_2 upright syncline (Fig. 2) that plunges shallowly towards the southeast (Fig. 6A). Stratigraphic-facing reversals and F_1 isoclinal fold hinges were rarely observed, and structural data suggest that prior to F_2 the greenstone-metasedimentary sequence of the Rankin Inlet Group was an F_1 homocline dipping gently to the southeast. The F_2 fold axes have a variable trend and plunge (Fig. 6C). On Barrier Islands the structures face to the east and northeast (Fig. 6D). F_1 minor folds are relatively rare. They are present in a few places within the turbidites (unit 1g) and dolomite (unit 2), and commonly plunge towards east-northeast at shallow angles. F_1 major fold closures have not been identified in the region, but pre-F_2 mineral (hornblende, quartz, and feldspar rodding) stretching lineations are commonly present within the metavolcanic rocks (Fig. 2, 6E). Their orientations have been modified by F_2 folds. The lineations generally trend southwest and east-northeast with shallow to moderate (10°-35°) plunges. Minor variations in trend towards the northwest and southeast are noted in a few places on Barrier Islands (Fig. 2). The tectonic significance of this lineation is yet to be evaluated. The metavolcanic and associated

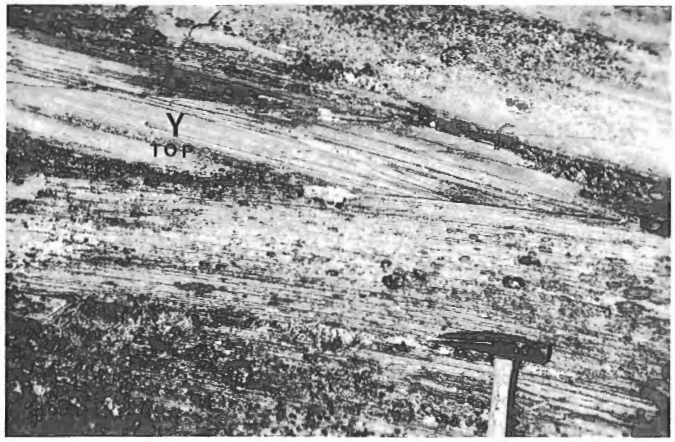


Figure 5A. Finely laminated and crossbedded impure quartzite (unit 3). View looking northeast; erosion surface dips steeply towards the observer, and the bedding (So) dips steeply away from the observer. Tops to the bottom of the photograph. (see Fig. 2 for location). GSC 204039-O

Figure 5B. "Ramp" structures in quartzite (unit 3). Dotted line marks the trace of the ramp. Note the carbonate concentration along this plane. The location, and the orientations of erosion and bedding (S_0) surfaces are same as in Figure 5A. GSC 204039-Q

Figure 5C. Multiple "lava-shelves" in a pillow, Rankin Inlet Group metabasalt; Location: small island southeast of Kudlulik Peninsula. Dotted line represents the bedding plane (So) trace on subhorizontal erosion surface; view looking east. (see Fig. 2 for location). GSC 204308-P

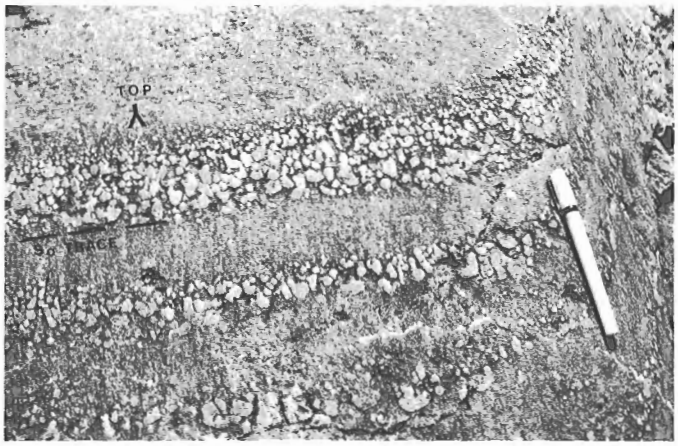
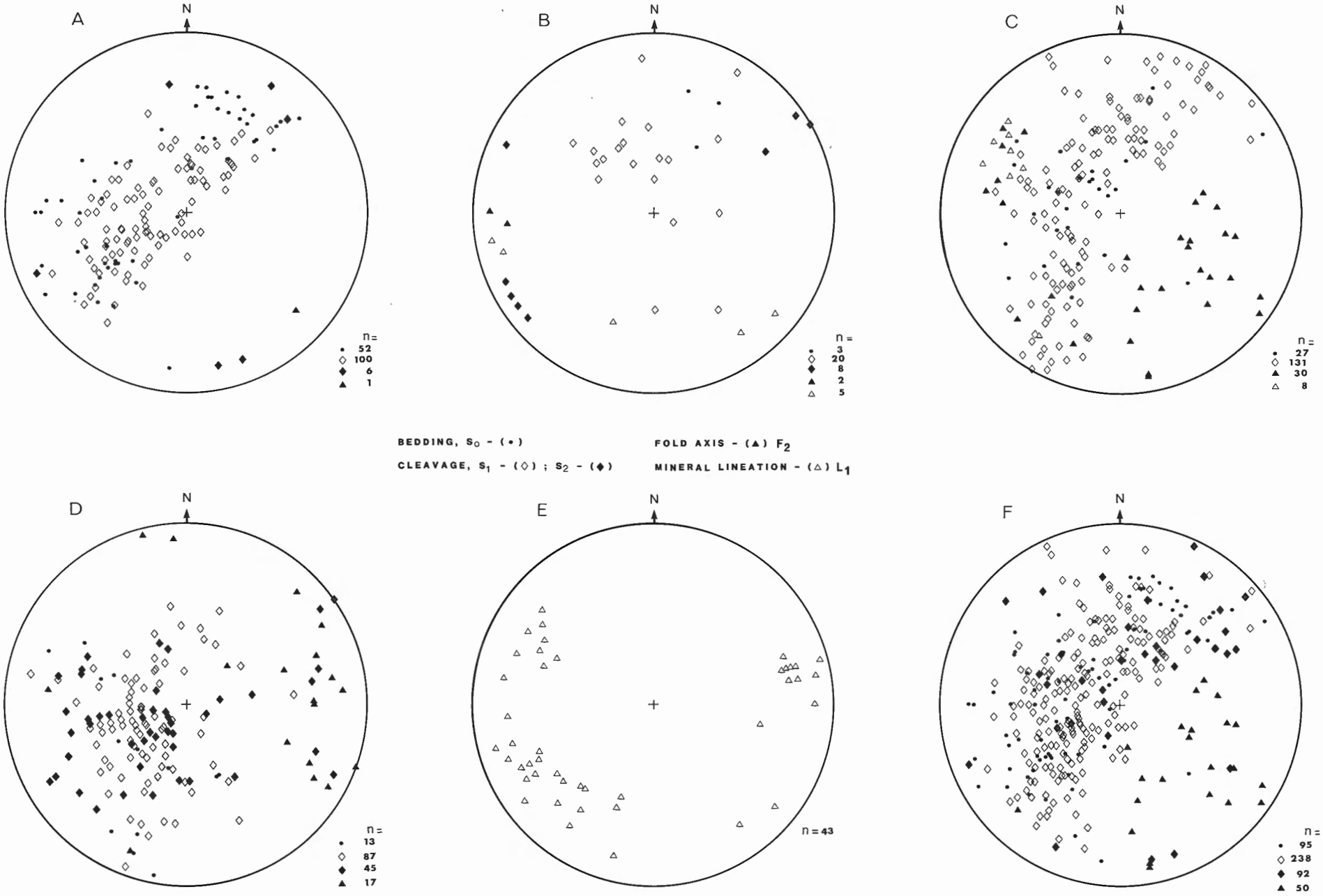


Figure 5D. Graded lapilli tuff, Rankin Inlet Group; Location same as in Figure 5C. Subhorizontal erosion surface; view looking southeast. GSC 204308-N



A. Kudlulik Peninsula and Thomson Island;
B. North of Kudlulik Peninsula;
C. Tudlik Peninsula Ahigik, Kango, Slab, Hump, Buff, and Kresic islands;
D. Barrier Islands;
E. North and South of Rankin Inlet;
F. Combined plot (A to D).

Figure 6. Equal area lower hemisphere projections of fabric elements from the metavolcanics and interflow sediments, Rankin Inlet Group.

sedimentary rocks, exposed on both sides of Thomson Passage are highly sheared and locally lineated (quartz-feldspar rodding) suggesting that a ductile shear zone occurs within the passage.

On Tudlik Peninsula and in the region northwest of Nipissak Lake (Fig. 2) the cleavage (S_1) intersects the bedding (S_0) at low angles. Bedding on the southwest limb of the F_2 syncline is consistently more steeply dipping than the S_1 cleavage (Fig. 6A), yet the sequence is not overturned and youngs to the east and south. The bedding-cleavage relationships are intriguing, because the S_1 cleavage vergence (Bell, 1981) is downward, yet the sequence youngs upward (Fig. 7). The significance of this geometry in the Rankin Inlet region has not been fully determined, but similar structural relationships have been reported from the Dalradian rocks near the Highland Boundary Fault (Shackleton, 1957) where they are associated with overturned limbs of recumbent fold nappes.

On Thomson Island (Fig. 2, 7) the cleavage and bedding trends are parallel, but the dip of cleavage is shallower than the bedding, and the structural facing and cleavage vergence is downward. Southwest of Johnston Cove the northwest trending cleavage (S_1) is steeper than the bedding, and the structural facing remains downward. The bedding-cleavage geometry in the Rankin Inlet Group (Fig. 7) suggests that the F_1 structures are discordant between Thomson Island and Kudlulik Peninsula, and between Kudlulik and Tudlik peninsulas. These relations are illustrated in Figure 8, a down-plunge F_2 fold profile sketch. It is shown that the structural facing directions are refolded. Since the structural facing on Thomson Island is downwards it follows that there should be a tectonic contact between Thomson Island and Rankin Inlet. Prior to refolding the volcanic sequence on Thomson Island faced up to the southeast and the vergence was towards the southwest. On Kudlulik Peninsula (Rankin Inlet) the sequence faced up to the northeast and the vergence was to the northeast. These structural facing and stratigraphic younging relationships suggest two possible tectonic contacts – one between Thomson Island and Rankin Inlet, and the second between Rankin Inlet and Tudlik Peninsula (Fig. 8). These tectonic contacts may be explained by a thrusting model in which two opposing tectonic motions occur on interleaving wedges (or ? thrust sheets). If the southwest and northeast trending mineral stretching lineations (L_1) observed in the Rankin Inlet region are indicative of tectonic transport directions, they appear to offer some support to the thrusting model; the Thomson Island sequence moving from the northeast, and the Rankin Inlet sequence moving from the southwest. Further structural studies are required to substantiate these preliminary conclusions.

West and north of Tudlik Peninsula, and on Ahigik, Kango, Slab, Hump, Buff, and Kresic islands (Fig. 2, 6B,C) the cleavage is steeper than the bedding, and the structural facing is commonly towards the east-southeast, with minor local deviations.

The volcanic rocks on the southern tip of Kudlulik Peninsula, and Thomson and Falstaff islands constitute the upper volcanic unit according to Bannatyne (1958), whereas according to Laporte (1975, 1983) and Hudec (as cited by Laporte, 1983) there is only one volcanic sequence that is repeated by an F_1 isoclinal fold. From our field work we conclude that these volcanic rocks and associated sediments are part of the second (or upper) volcanic cycle of this study

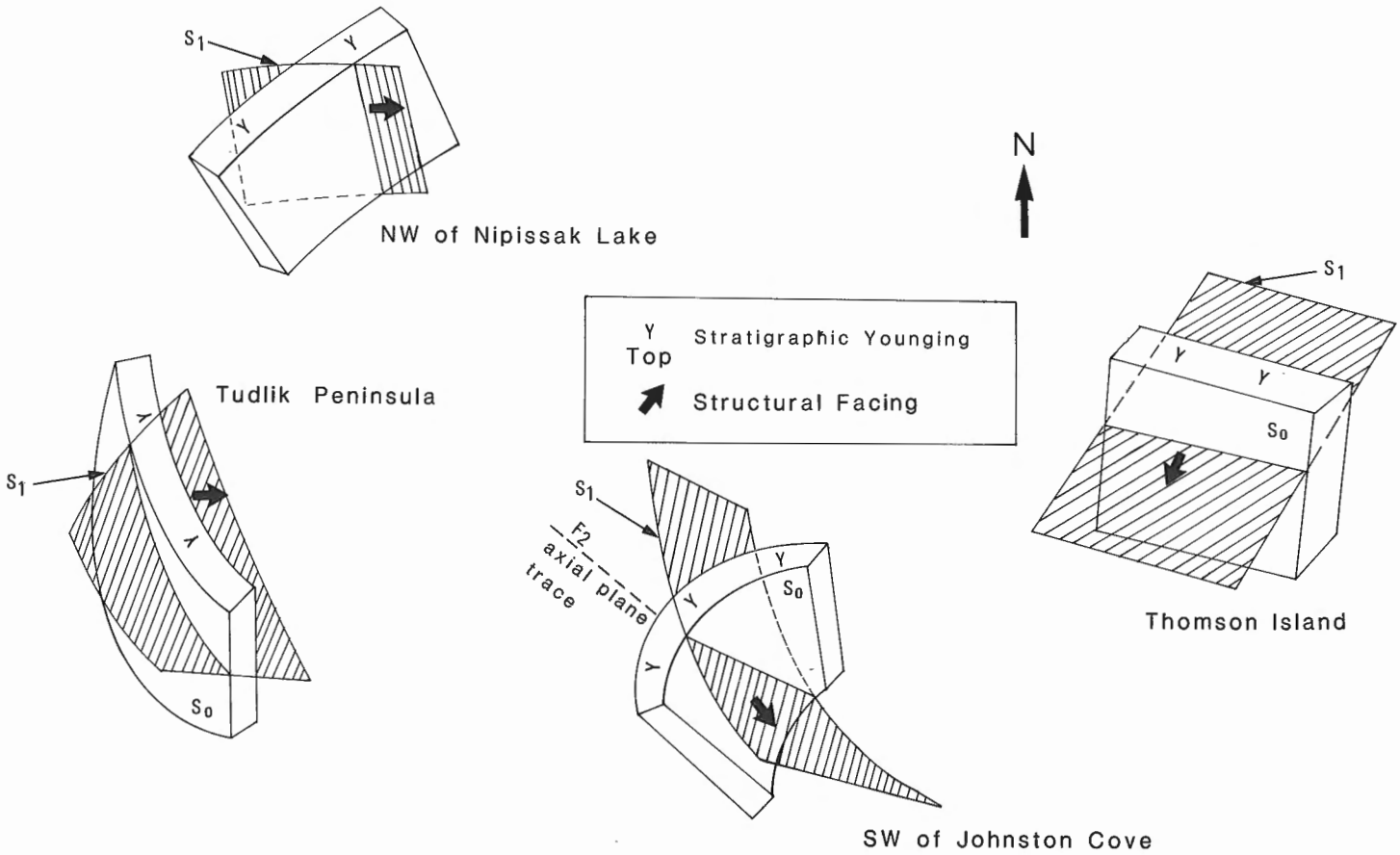


Figure 7. Schematic representation of bedding (S_0)-Cleavage (S_1) relations, Rankin Inlet Group.

and postulate that they have been tectonically transported above a thrust fault. If the "ramps" noted in unit 3 are related to this thrusting their orientation suggests a "northerly" tectonic transport direction. The thrusts have been modified by the F_2 syncline. The extent and effects of the postulated pre-F_2 thrusting have not been fully resolved. Additional structural data are required to document the direction of tectonic transport with certainty.

We favour an Archean age for the Rankin Inlet Group on the basis of lithological and structural similarities to the Archean Kaminak Group (Davidson, 1970a; Heywood, 1973) which is extensively exposed in the region Southwest of Rankin Inlet. Further geochronological and geochemical studies are planned to resolve the ages of the rocks. The folding may be Archean and/or Proterozoic.

Numerous northwest- and north-trending faults transect the region between Barrier Islands and Falstaff Island. Only a selected few are shown in Figure 2 and the reader is referred to Laporte (1983) for a more complete documentation of the regional distribution of faults. Between Ahigik and Buff islands metavolcanic rocks are extensively sheared along northwest trending zones a few metres to 100 m wide. Strain estimates reveal the presence of northwest trending high-strain zones, possibly representing high angle (? thrust) ductile shear zones within the greenstone-metasedimentary sequence. Deformed pillow selvages in the volcanics show strain ratios from 4 to 6 within and adjacent to these shear zones. Additional strain studies are planned to further define and resolve the structure of the Rankin Inlet Group.

FELSIC AND MAFIC SCHISTS (UNITS 6 AND 7)

Felsic volcanic rocks consisting of dacite to rhyolite porphyry (unit 6A) and felsic tuff (unit 6B) are sporadically distributed in the region east and west of Meliadine River (Fig. 1). They are commonly sheared and metamorphosed to quartz-mica schists, but some relict igneous textures are locally preserved. Laporte (1983) considered these rocks to have been derived from the greywacke sequence (unit 1g), but the relict igneous textures indicate a felsic volcanic protolith. Contact relationships with other rock types in the region are obscure, but the felsic rocks are believed to be the metamorphic equivalents of the Rankin Inlet Group (Heywood, 1973; Wright, 1967).

Metamafic rocks (unit 7) consisting of actinolite-chlorite schist, hornblende-biotite-chlorite schist, amphibolite gneiss, and mafic tuff are exposed in east to east-southeast trending bands east of Meliadine River, and north and south of "Atulik" Lake (Fig. 2). Their distribution and petrologic character have been described in detail by Heywood (1973) and Laporte (1983). The unit is predominantly actinolite-chlorite schist which consists of highly sericitized plagioclase, and fine grained quartz aggregates (0.5-1.0 mm) set in a finer grained quartz, feldspar, chlorite, actinolite matrix (Laporte, 1983). The abundance of quartz aggregates in an otherwise chlorite-amphibole rich rock together with the presence of compositional banding suggest a mafic tuff protolith for these schists. Laporte (1983) reported mafic hornfelses that developed as a result of granite intrusion northwest of Kudlulik Peninsula (Fig. 1). However, we found no evidence of static or contact metamorphism in the area.

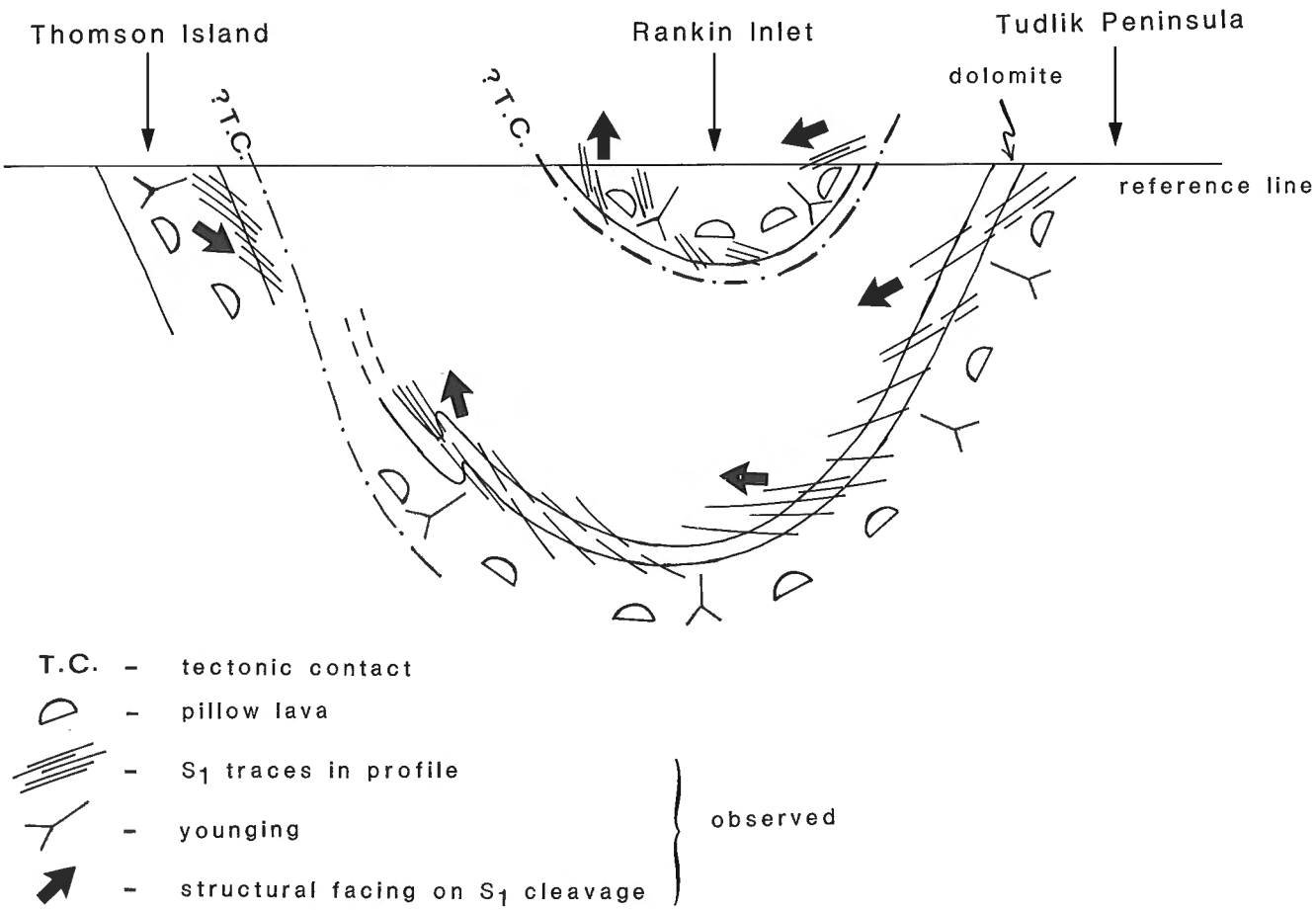


Figure 8. Down-plunge profile sketch of the Rankin Inlet structure; view to southeast

LAYERED GNEISS COMPLEX (UNIT 8):

This is a mixed unit of uncertain age consisting of highly deformed and metamorphosed migmatitic gneisses (Fig. 9) and paragneiss with garnet, biotite, ± staurolite, ± cordierite ± sillimanite, and ± kyanite assemblages. The rocks within the unit are widely exposed in the region north of Rabbit Island and Baker Foreland (Fig. 1), and to a lesser extent on Marble Island (Fig. 12). They contain abundant inclusions and layers of metamafic rocks and felsic schists possibly derived, in part, from the Rankin Inlet Group. The rocks within the unit also include minor amounts of metamorphosed oxide facies banded iron-formation which shows a pronounced aeromagnetic expression (Geological Survey of Canada, 1966b). At least two periods of deformation and an event of regional metamorphism can be recognized within the rocks of this unit. The orientations of fabric elements are included in Figure 10. Axial surfaces of early tight, upright folds commonly trend northwest parallel to S_1, and fold axes plunge shallowly either towards west-northwest or towards east-southeast. The axial surfaces are defined by a regionally pervasive planar fabric (S_1) that trends northwest and dips steeply (60°-80°) to the northeast and to the southwest. A penetrative doubly plunging (15°-30°) hornblende lineation is present in migmatitic rocks. The geometry of the folds is similar to that of the Rankin Inlet Group F_2 structure. Later folds (F_3) are commonly open, upright structures with north-northeast trending axial surfaces. Preliminary petrographic data suggest that the regional metamorphic grade is within the upper amphibolite facies with some local greenschist facies overprinting effects. The gneiss complex is intruded by a two-mica leucogranite (unit 9).

TWO-MICA LEUCOGRANITE (UNIT 9)

Coarse- to medium-grained, grey-to white weathering, weakly- to well foliated biotite-muscovite leucogranite is widely exposed in the region north and northwest of Baker Foreland (Fig. 1). The regional foliation trends northwest and dips steeply (60°-80°) both to the northeast and southwest. The granite commonly consists of biotite (ca. 12%), muscovite (5%), quartz (25%), alkali feldspar (55%), and rare amphibole. Porphyritic phases are locally present where augens of pale pink, K-feldspar phenocrysts up to 2 cm long, and large books of muscovite (? xenocrysts) are set in a finer grained, well foliated matrix composed of quartz, feldspar, biotite, muscovite, and rare amphibole. Quartz grains are commonly translucent in hand specimen. Pink garnet porphyroblasts (? xenocrysts) up to 3 cm diameter are present at a few localities. The granite contains abundant inclusions of layered gneiss (unit 8) and is cut by pink granite dykes and veins, ranging from a few centimetres to one metre thick. Small inclusions of garnetiferous metagabbro are present within the unit north of Baker Foreland.

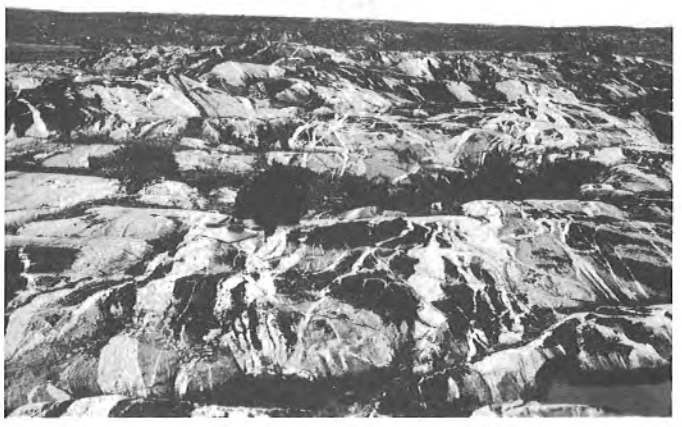


Figure 9. Layered gneiss complex (unit 8, Fig. 1; unit 1 in Fig. 2), Marble Island. This unit is overlain unconformably by the ?Hurwitz Group rocks to the south of this locality (see Fig. 2). GSC 204308-T

GRANODIORITE TO QUARTZ MONZONITE (UNIT 10)

Massive to weakly foliated porphyritic granodiorite to quartz monzonite is exposed on Pangertot Peninsula along the south shore of Rankin Inlet (Fig. 2) south and southeast of Longspur Island (Heywood, 1973; Laporte, 1983). Microcline phenocrysts up to 6 cm long occur in a medium grained quartzofeldspathic matrix. Mafic minerals, mainly biotite and chlorite, constitute less than 10 per cent of the rock. Sphene, apatite, and magnetite are present as accessory phases. Minor amounts of well foliated orthogneiss, and abundant inclusions of metavolcanic and metasedimentary rocks (unit 1 ?) form enclaves in this unit.

The contact between the quartz monzonite and the Rankin Inlet Group is not exposed. South of Kind Islet (Fig. 2) the granitoid rocks are highly sheared along the shoreline and show well developed mylonitic textures, suggesting a tectonic contact between the two units. A penetrative mineral (hornblende) stretching lineation is well developed within the orthogneiss-granite unit that plunges shallowly (ca. 20°) towards east-northeast. The lineation is parallel to the linear elements (fold axes and mineral lineation) noted in the greenstone-turbidite sequence just

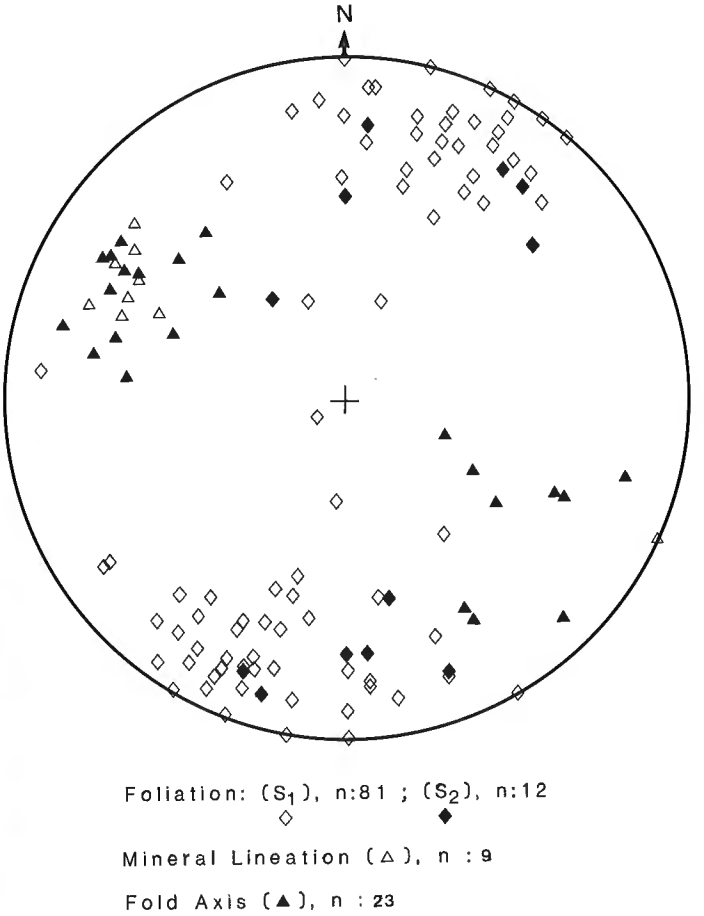


Figure 10. Equal area lower hemisphere projection of fabric elements from units 8 and 9, north of Baker Foreland.

north of Kind Islet. Although Laporte (1983) implied a faulted unconformity between the quartz monzonite and the Rankin Inlet Group rocks, the presence of numerous inclusions of layered amphibolite, metasedimentary and metavolcanic rocks (similar in composition and texture to the rocks of the Rankin Inlet Group) suggest that the quartz monzonite is younger, perhaps similar in age to that of unit 9. These observations are consistent with those suggested by Heywood (1973), and may be confirmed by radiometric dating.

ORTHOQUARTZITE (UNIT 11)

Deformed, thin-to thick bedded, predominantly ripple-marked, white orthoquartzite is exposed along the east shore of Melvin Bay on Kudlulik Peninsula (Fig. 2), and on Marble Island (Fig. 12, unit 4). They were named Marble Island Quartzites by Bell (1885, 1887) and considered to be of Lower Proterozoic age (Bell, 1968; Laporte, 1973, 1983; Tyrrell, 1898; Wright, 1967). Our interpretation of the stratigraphy and structure of the quartzites at the above localities differs from that of previous workers (eg., Bannatyne, 1958; Hudec, unpublished; Laporte (1975, 1983).

On Kudlulik Peninsula the quartzite is locally overlain conformably by thin beds of argillite, buff dolomite, and siliceous dolomite (unit 11a). The contacts with the structurally overlying and underlying rocks of the Rankin Inlet Group are drift covered, but previous workers (Bannatyne, 1958; Bell, 1968; Laporte, 1973, 1983) considered them to be conformable. Our investigations, however, favour tectonic contacts. The northern contact is inferred to be a northeast trending fault of unknown displacement that, on Tudlik Peninsula across Melvin Bay (Fig. 2), shows sinistral oblique-slip movement within the metavolcanic rocks of the Rankin Inlet Group. East of Melvin Bay, a southwesterly overturned succession of Rankin Inlet Group metasedimentary rocks (unit 3) and the upright orthoquartzite sequence (unit 11) are juxtaposed along this fault. The eastern contact of the quartzite (unit 11) is a postulated thrust over which the Rankin Inlet Group strata have been transported. The direction of transport has not been determined, but may be northerly based on tenuous evidence from the orientation of "ramp" structures in the laminated quartzite (unit 3; Fig. 5B).

On Marble Island orthoquartzite beds define a series of east- to east-southeast trending, upright to south verging, tight to open, shallowly plunging F_1 folds (unit 4, Fig. 12). These F_1 folds are not necessarily correlative with the F_1 folds in the Rankin Inlet Group. A thin quartz-mica schist (unit 3, Fig. 12) is locally present beneath the quartzite and adjacent to a layered gneiss basement complex (unit 1, Fig. 9, 12). For the most part the contact with the basement complex is marked by east trending, southerly directed, high angle imbricate thrusts. Adjacent to the thrusts, in a few places, the bedding is locally overturned to the south.

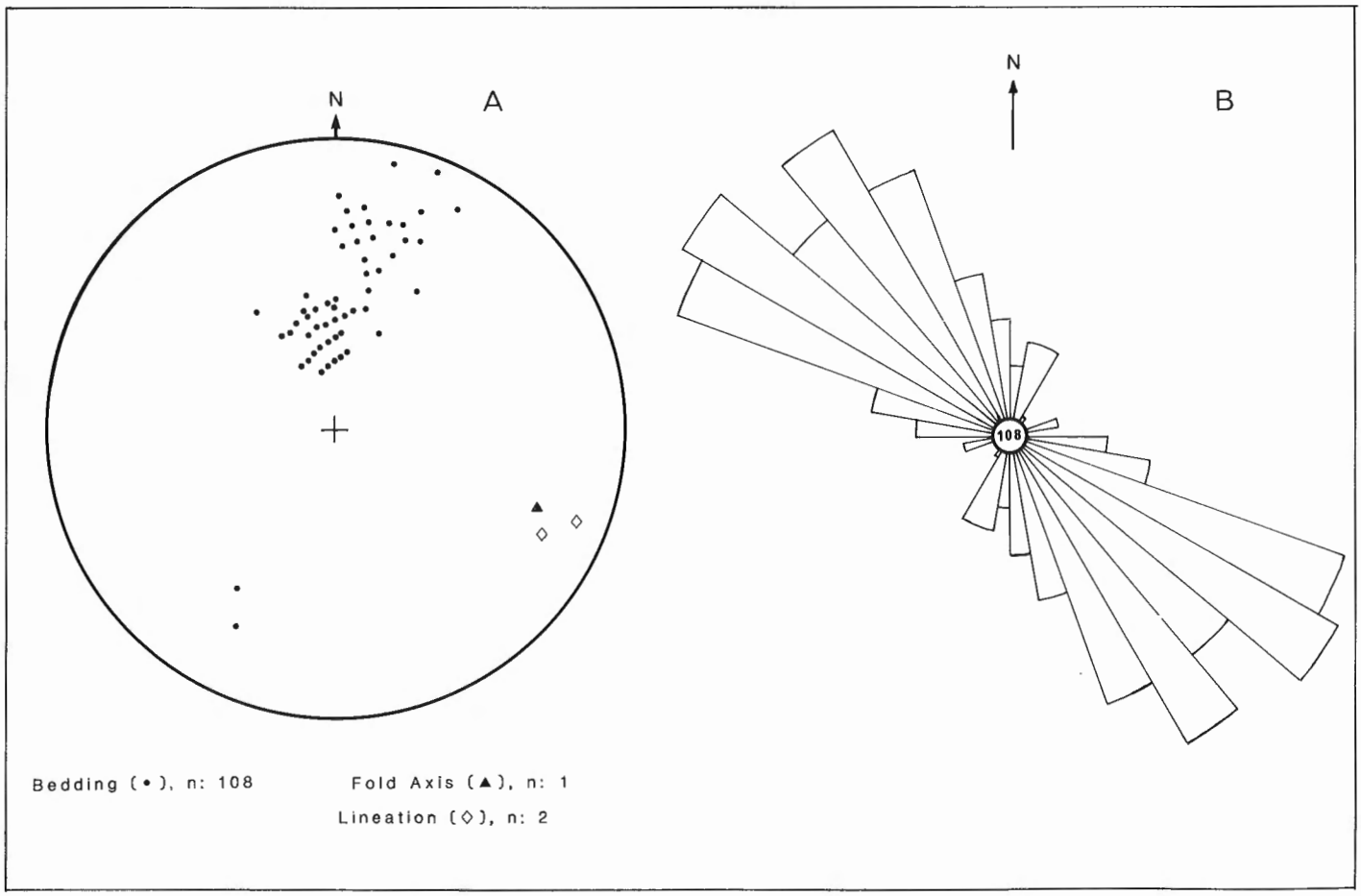


Figure 11. A. Equal area lower hemisphere projection of fabric elements from orthoquartzite (unit 11), Kudlulik Peninsula. B. Orientation of Ripple Axes in orthoquartzite (unit 11), Kudlulik Peninsula.

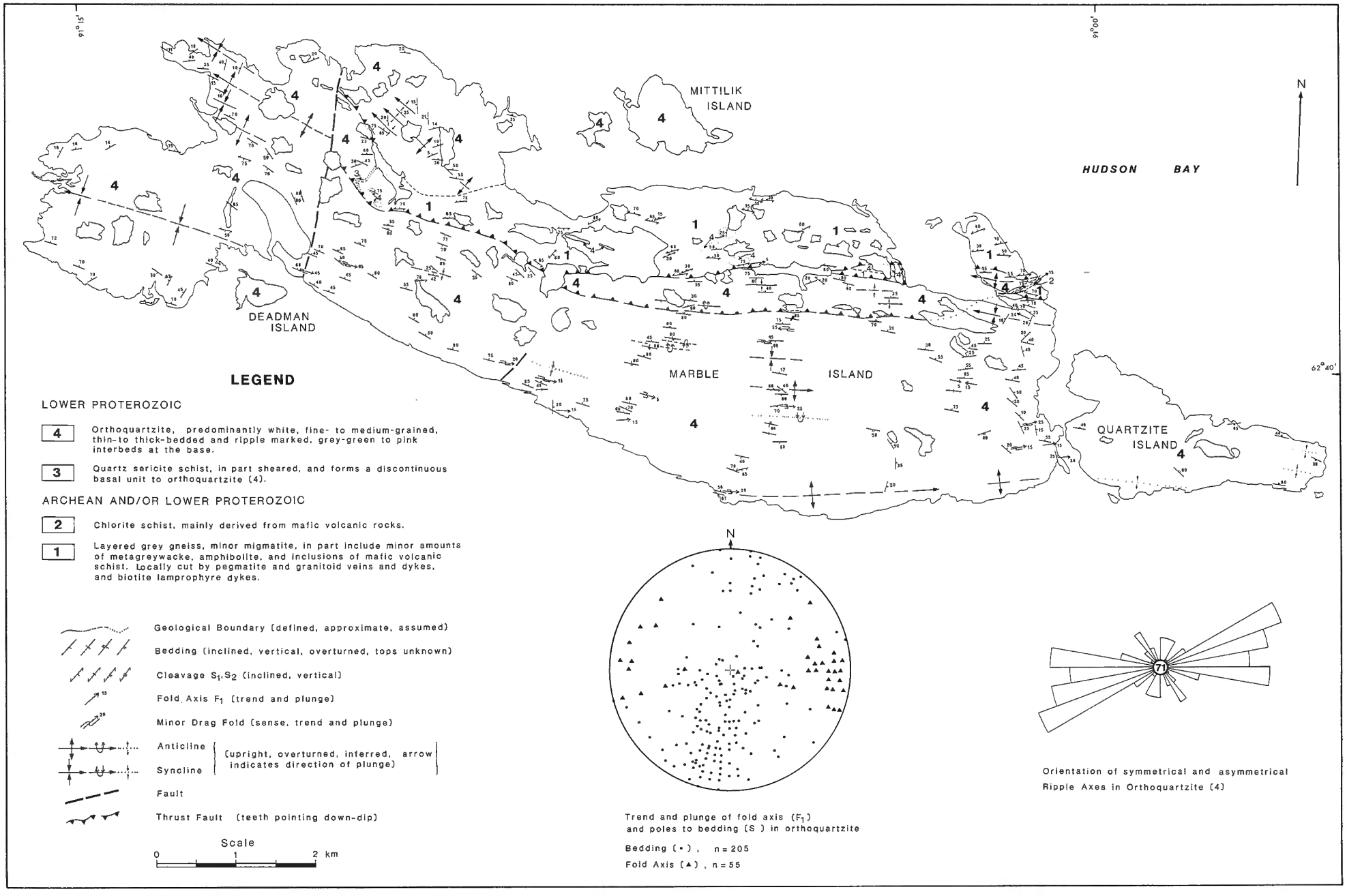


Figure 12. Geology of the Marble Island area.

The orientations of ripple-axes in orthoquartzite on Kudlulik Peninsula (Fig. 11B), and from Marble Island (Fig. 12) suggest sediment transport direction from the southwest and from south-southeast respectively. The data are consistent with the observations of Bell (1968), and Laporte (1983).

Scattered outcrops of deformed orthoquartzite (unit 11) are also exposed in narrow linear belts north of Diana River, east of Meliadine River, north and south of "Atulik" Lake, and northwest of Scarab Point (Fig. 2). The contacts with the Rankin Inlet Group and its metamorphic equivalents (units 6 and 7) are commonly obscured by heavy drift cover or by high degree of deformation. At one locality, northeast of Thomson Passage (Fig. 2), the quartzite rests on a steeply dipping (S_1) amphibolite schist (unit 7, Fig. 2) with an apparent structural discordance. Although the contact is somewhat sheared, the quartzite dips gently (20°-25°) to the south suggesting an unconformable relationship. These scattered quartzites are tentatively correlated with the Marble Island Quartzites of Bell (1885, 1887) on the basis of their lithological similarities and structural position. The Marble Island Quartzites are tentatively correlated with the White Rock Member of the Lower Proterozoic Hurwitz Group (Bell, 1968) on the basis of lithological similarities.

GRANITE (UNIT 12)

Scattered outcrops of massive to weakly foliated, equigranular medium grained pink granite occur between Diana Lake and Meliadine River in the northern parts of the region (Fig. 1). Sills and dykes of the granite cut the layered gneiss complex (unit 8). The age of the granite with respect to orthoquartzite (unit 11) is uncertain, but it cuts the rocks of units 8 and 9.

BIOTITE LAMPROPHYRE DYKES (UNIT 13)

A 130° to 160° trending dyke swarm composed of biotite lamprophyre cuts all the rocks described previously with the exception of orthoquartzite (unit 11). The distribution of selected dykes in the Rankin Inlet region is shown in Figure 13. The dykes range from a few centimetres to eight metres thick, and locally contain abundant, 1 to 20 cm size, rock fragments of granite, volcanic rocks, orthoquartzite, vein quartz, and carbonate rocks (Fig. 14A). The lamprophyre is commonly a black to dark grey, medium- to fine-grained rock consisting of large phenocrysts of zoned biotite (and/or phlogopite) set in a finer grained groundmass composed of K-feldspar, plagioclase, biotite, quartz, carbonate, and blue-green actinolitic hornblende.

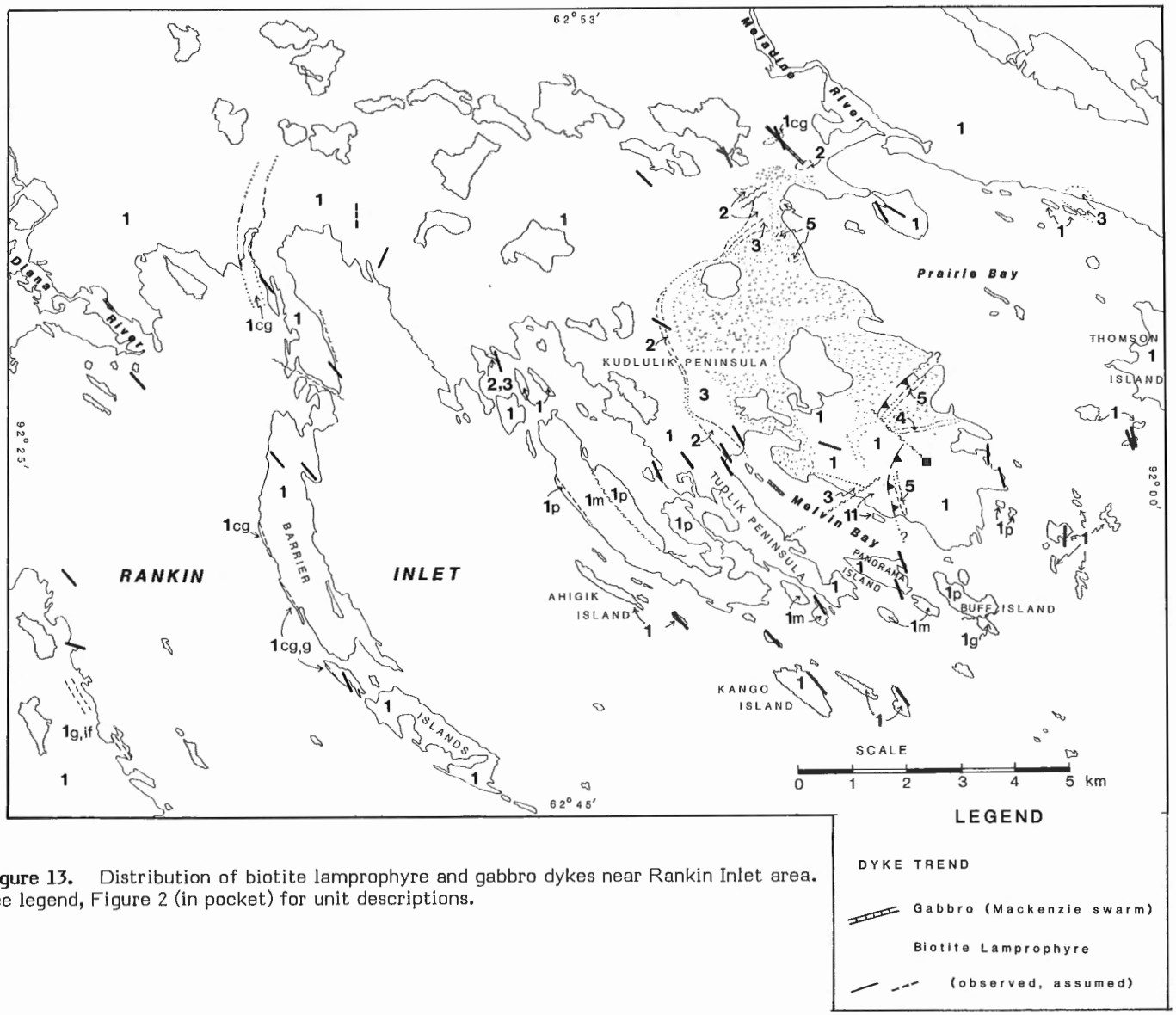


Figure 13. Distribution of biotite lamprophyre and gabbro dykes near Rankin Inlet area. See legend, Figure 2 (in pocket) for unit descriptions.

Sphene, opaques, apatite, epidote and chlorite represent accessory minerals. Zircon and fluorite are present in trace amounts in two of the twenty thin sections examined.

The age of the dykes in the Rankin Inlet Region is unknown, but they are compositionally and texturally similar to the Helikian lamprophyre dykes described in the Kaminak Lake map area (Davidson, 1970a) to the southwest. Post-emplacement deformation of dykes has been observed at two localities- one on Ground Squirrel Island (Fig. 2, 14B) and the other on a small island northeast of Panaroma Island (Fig. 2, 14C) where the dyke was shortened by approximately 20 per cent. The deformation appears to be localized along narrow shear zones. Lamprophyre dykes are not known to cut the orthoquartzite (unit 11) on Marble Island or Kudlulik Peninsula. Their apparent absence in this ? lower Proterozoic sequence together with localized deformation of some dykes in the Rankin Inlet region, raises a possibility that some dykes may be older than the Helikian lamprophyres noted in the Kaminak Lake area and elsewhere in the District of Keewatin (Davidson, 1970a; Wright, 1967).

GABBRO DYKES (UNIT 14)

Two northwest trending, fine- to medium-grained, massive gabbro dykes, possibly related to the Mackenzie Swarm, are noted in the Rankin Inlet area (Fig. 2). Northwest of Prairie Bay, a 10 m thick, massive dyke with chilled margins intrudes the metavolcanic and metasedimentary rocks (units 1g, 1cg) of the Rankin Inlet Group. This dyke also cuts a narrow (50 cm thick) biotite lamprophyre dyke at the same locality.

Part 2: INTERPRETATION OF REGIONAL STRUCTURE USING PILLOW-BASALT STRUCTURES AT RANKIN INLET

G.J. Borradaile

INTRODUCTION

The region studied poses special problems in structural mapping due to the absence of bedding-markers over large areas. Without orientation data from the large areas of pillowed meta-basalts the structure of many parts of the area would be unsolvable and an integrated picture of the whole tectonic pattern could not be realized. Nor are such problems of limited importance. Pillows are undoubtedly the most common macroscopic primary rock structure at the surface of the solid earth and they were at least as ubiquitous, if not more common, at the surface of the solid earth during the Archean and early Proterozoic eons. Whenever pillow structures are tectonically deformed we encounter structural problems at the most fundamental level: what are the orientations of bedding and way-up? (Facing is not used as a synonym for way-up in this text, to avoid confusion with "structural facing" which is a further concept of greater practical use in folded terranes).

The problem unique to pillowed sequences owes its special nature to the initial geometry of pillow structures. (The plural is used here because the range of geometries of "pillows" is very great; see Ballard and Moore, 1977). The problem arises because as a marker of primary geometry a pillow forms a "closed loop" in all sections. Upon tectonic deformation approximating passive distortion a closed loop deforms to produce a special class of strain marker whose new long axis is in a different orientation from the line which was the original long axis. The originally most distant parts of the loop are no longer the most separated parts after deformation. Note that this and ensuing arguments do not

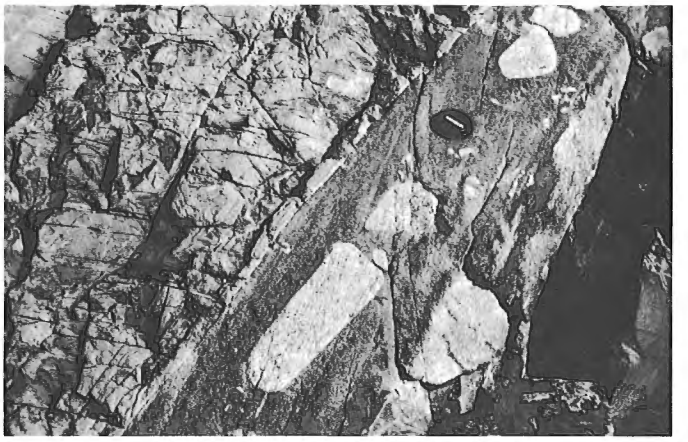


Figure 14A. Biotite lamprophyre dyke (unit 13) with rounded to angular granitoid fragments cutting massive metabasalt of the Rankin Inlet Group. View looking towards east-southeast (110); subhorizontal erosion surface. (see Fig. 2 for location). GSC 204308-U

Figure 14B. Folded biotite lamprophyre dyke (unit 13) with chilled margins cutting massive volcanics (unit 1) on Ground Sqirrel Island. View looking towards northwest; subhorizontal erosion surface. (see Fig. 2 for location). GSC 204308

Figure 14C. Folded biotite lamprophyre dykelet in a sheared metavolcanic unit. View looking northwest; subhorizontal erosion surface. (see Fig. 2 for location). GSC 204039-P

require that the loop is elliptical in shape. Indeed, in most cross-sections pillows have a more complex shape involving externally directed cusps due to sagging into the hollows between underlying pillows and re-entrant lobes due to incomplete budding.

The originally longest line in an undeformed pillow cross-section will normally be a paleohorizontal or "bedding" indicator. Certainly this is true for the average longest dimensions of a group of pillows. The angular position of this line can be related to the position of the final longest line,after a tectonic deformation, by the Rf-ϕequations (Ramsay 1967, eq. 5-22 and 5-27; here equations 1 and 2)

$$\tan 2\phi = \frac{2 R_T^{\frac{1}{2}} (R_o - 1) \sin 2\theta}{(R_o + 1)(R_T - 1) + (R_o - 1)(R_T + 1)\cos 2\phi} \qquad (1)$$

$$R_T = \frac{\tan^2\phi\,(1 + R_o \tan^2\theta) - R_T(\tan^2\theta + R_o)}{R_T \tan^2\phi\,(\tan^2\theta + R_o) - (1 - R_o \tan^2\theta)} \qquad (2)$$

Where, $\sqrt{R_T}$ = Strain ellipse ratio

$\sqrt{R_o}$ = Ratio of ascribed ellipse to original pillow shape

$\sqrt{R_T}$ = Ratio of ascribed ellipse to deformed pillow shape

θ = Attitude of original long axis, with respect to maximum elongation

ϕ = Attitude of long axis after strain, with respect to maximum elongation

Figure 15 indicates the relations of these elements.

In this context it should be noted that the strain is considered to be homogeneous. The extent to which this criterion is not met will be mentioned later. Furthermore the pillow outline is considered to behave passively. That is to say, we consider that the outline can be stretched and rearranged like a rubber band so that points on the outline may appear to drift around the outline with respect to the termini of the pillow. For example, in Figure 15A grains located at the ends of the line A, B appear to shift from the ends of the pillow in the deformed state (Fig. 15B).

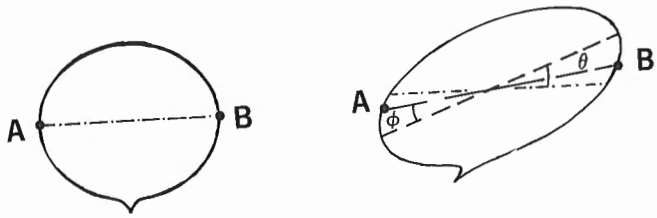


Figure 15. Relative position of points A and B and the longest axis of a pillow before (a) and after (b) homogeneous passive deformation.

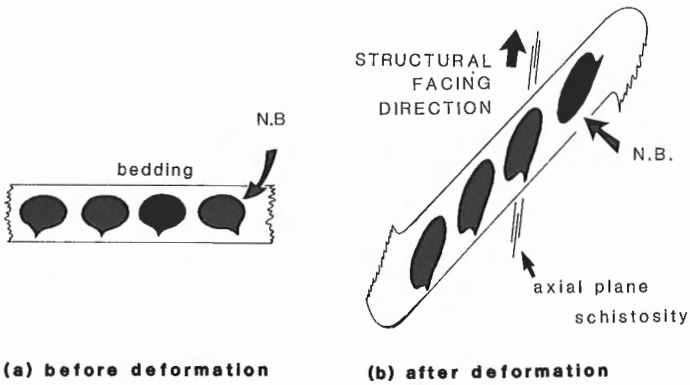


Figure 16. Relationship between the shape of a deformed pillow and the bedding, younging, and structural facing directions before (a), and after (b) deformation.

The practical consequence of this fundamental problem is that

a. the long axis of a deformed pillow is not the bedding orientation,
b. the cusp no longer points in the opposite direction to way-up.

Figure 16 illustrates the relationship between the shape of a deformed pillow and the bedding, younging and structural facing directions. For a fuller definition of structural facing, the younging component on the axial planar schistosity, see Borradaile (1976) and Bell (1981).

SPECIAL PILLOWS WHICH GIVE YOUNGING DIRECTLY IN THE DEFORMED STATE

Before the problem is addressed as to how one may deal with the general case of a deformed pillow some optimism can be expressed concerning special initial geometries. Such special original cases include the flat based pillow and the pillow with an original paleohorizontal marker (Borradaile, 1984). These special cases have been found in the Rankin Inlet area, and the latter case is unusually common and will be dealt with at some length later on.

The flat-based pillow appears to have formed where a pillow has been extruded onto interflow sediment, ash, massive lava or hyaloclastite with a flat surface. Consequently, after homogeneous deformation the flat base remains planar and provides an indication of bedding. In this case bedding is known anyway from the underlying structure. However, the bun shaped top indicates the younging direction, perpendicular to the flat base (Fig. 17A). The multicusp pillow forms where a sack-like pillow overlaps several smaller pillows so that multiple cusps are formed where it drapes in between them. The multiple cusps provide reference points which restrict the number of possibilities for the younging direction but the bedding direction may not be well constrained (Fig. 17B).

The paleohorizontal pillow contains an originally horizontal lava-level resulting from the partial draining of the pillow or lava tube due to subsequent budding (Borradaile, 1985, fig. 3 after Ballard and Moore, 1977). The planar lava surface remains planar after homogeneous strain and provides a bedding indicator after deformation. The way-up points away from the lava level toward the side with the convex-upward amygdaloidal cavity (Fig. 17C).

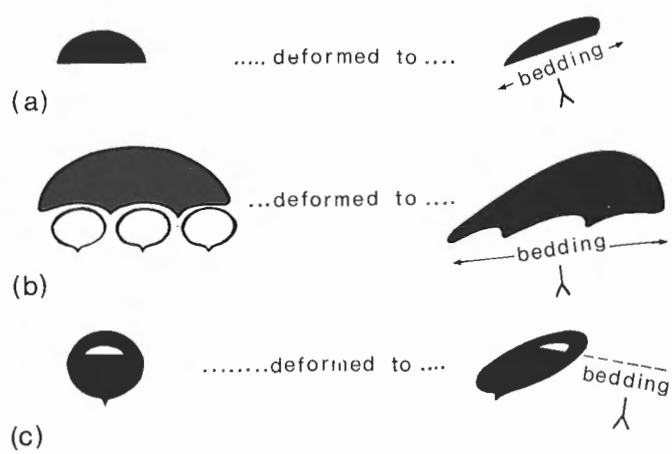


Figure 17. Examples of special pillows which give younging in the deformed state.

Figure 18A. Heterogeneously strained pillow with molar tooth shape, Rankin Inlet Group. Note that the cusp becomes indistinguishable from the adjacent terminus of the pillow. (see Fig. 2 for location). GSC 204040

SPECIAL PILLOWS WHICH GIVE THE STRUCTURAL FACING DIRECTION BUT NEITHER BEDDING NOR WAY-UP

Upon heterogeneous strain, occuring at a strain beyond an ellipse ratio of about 4:1 in most cases (Borradaile, 1982, fig. 17) the cusp becomes indistinguishable from the pear-shape adopted by the nearest terminus on the pillow (Fig. 18A). As a result the pillow takes the shape of a molar tooth which from consideration of Figure 18B gives the structural facing direction. While the younging and the bedding directions are unknown in this case it can be of considerable value to know the component of the younging direction on the axial planar schistosity, i.e. the structural facing direction. In the present study it was often valuable to know this as it gave the sense of younging of the entire folded stratigraphic sequence. In many parts of Tudlik and Kudlulik peninsulas it was only possible to determine the structural facing direction, bedding and way-up being unknown.

STRAIN DETERMINED FROM PILLOWS

Studies of undeformed pillows both in the metamorphosed and unmetamorphosed states show no significant, consistent variations in selvage thickness when traced around the pillow rim (Borradaile and Poulsen, 1981; Borradaile, 1985). However, when deformed, such selvages show consistent thickness variations with maximum selvage thickness in directions nearly parallel to schistosity or in the maximum extension direction which may be indicated by the shapes of deformed amygdules. The minimum thicknesses of selvage are in the minimum extension direction.

Studies involving large numbers of pillows in individual, homogeneously strained outcrops give consistent selvage thickness ratios which are believed to be close to the true strain ellipse axial ratios for the selvages in the plane of measurement. In Archean terranes, the plane of measurement is mostly the horizontal peneplane and the present study is no exception. In one study from the Superior Province it was recently shown that in pillows with several "concentric" selvages, slight differences can exist between

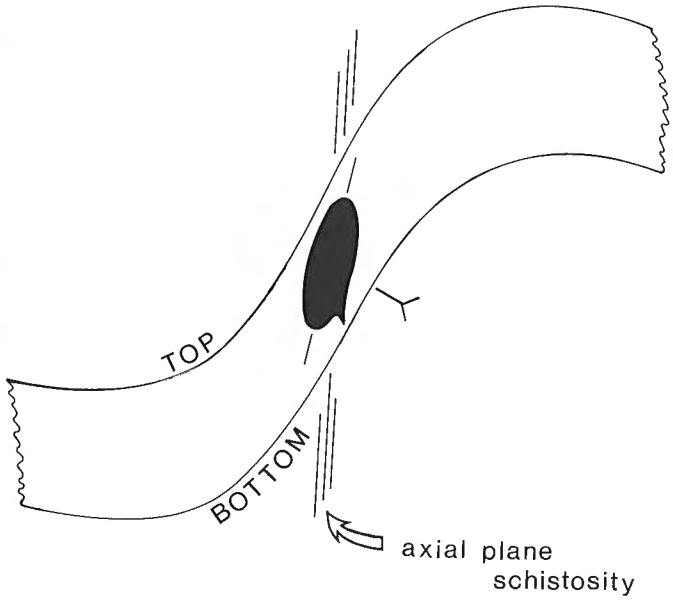


Figure 18B. Illustration of component of younging direction on the axial plane schistosity (ie., structural facing). Structures face to the top of the diagram.

Table 1. Comparison of outer (dark) selvages, inner (light) selvages and pillow shapes, Swamp Creek, Shebandowan Belt Northern Ontario

Pillow shape ratio	Outer selvage ratio	Inner selvage ratio
(± s.e.)	(± s.e.)	(± s.e.)
5.34 ± 0.60 (n = 50)	4.85 ± 0.49 (n = 50)	3.27 ± 0.12 (n = 50)
s.e. – standard error	n – number of measurements	

inner and outer selvages, indicative of the strain differences which arise due to partitioning between different lithologies (see Table 1). While these slight differences do indicate that selvage strain is not synonymous with pillow strain it should be realized that the cut-effect of the selvage renders only an approximate value of the strain from the selvages (Borradaile, 1985). Provided that the selvages are measured in a section through the central third of the pillow, there is a ± 10 per cent margin of error in the selvage ratio estimate, and this is of the same order of magnitude as the error due to rheological differences between the inner and outer selvages (Table 1). Greater errors should not occur, provided that obviously heterogeneously strained selvages (pear-shaped or folded, Borradaile 1982, fig. 15, 16) are avoided.

The regional distribution of strain ellipses determined from pillow selvages in plan view (or corrected so as to be rendered in plan view) are shown in Figure 19. Additional data may make it possible to remove the effects of strain from the stratigraphy, and to determine original stratigraphic thicknesses as percentages of the present values assuming that the deformed rocks have remained as a continuum.

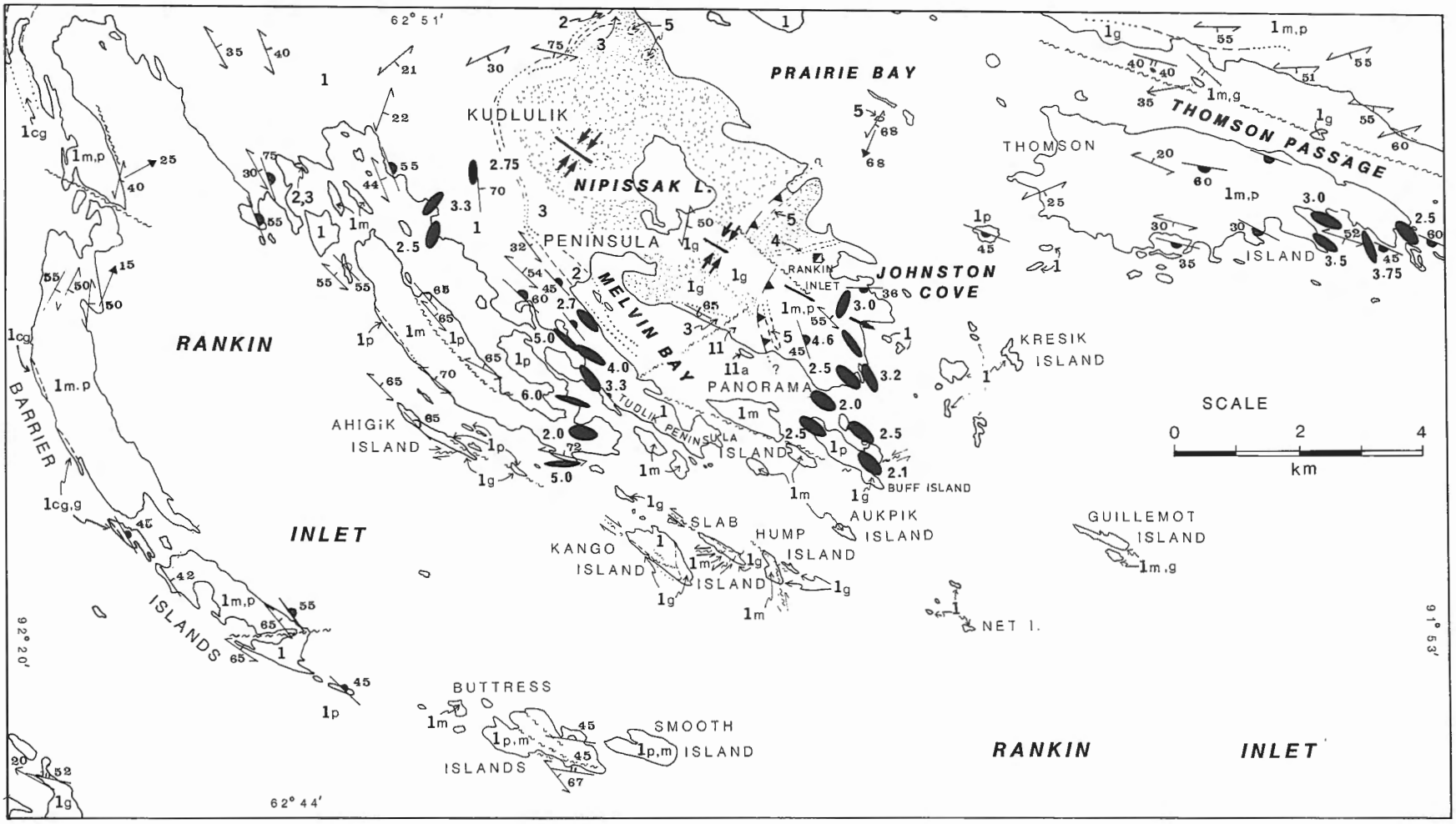


Figure 19. Provisional map of strain ellipses from deformed pillows, Rankin Inlet Group. See legend (Figure 2, in pocket) for unit descriptions.

Figure 20A. "Lava-shelves" in a pillow filled with calcite. Discoidal cavities with planar bases and convex tops. (see Fig. 2 for location). GSC 204040-A

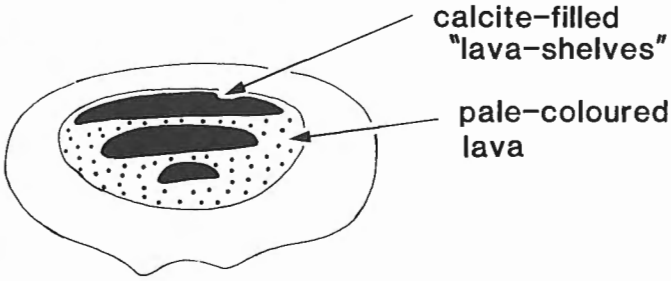


Figure 20B. Generalized sketch illustrating the internal texture of the "lava-shelves" shown in A.

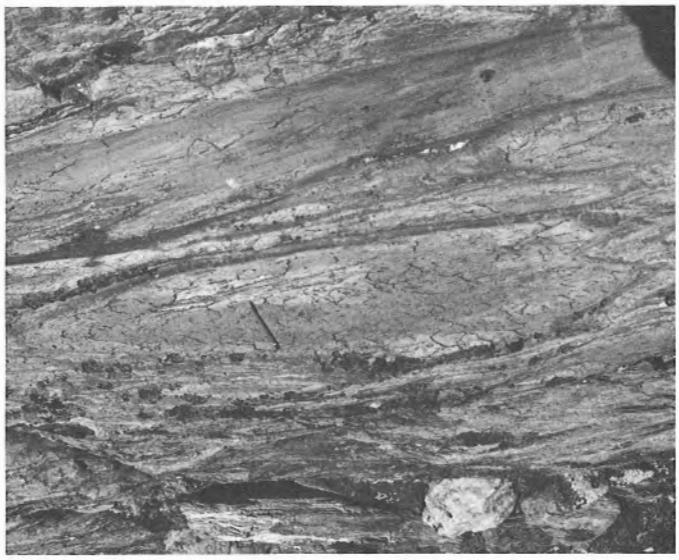


Figure 21. Highly deformed and transposed pillow from the Rankin Inlet Group. Convex surface of the "lava-shelf" is indeterminable at this strain. Molar tooth shape and the cleavage orientation suggest structural facing to the left of the photograph. (See Fig. 2 for location). GSC 204040-B

DETERMINING WAY-UP AND BEDDING FROM PILLOWS WITH MORE COMMON OUTLINES AND FROM PILLOWS WITH INTERNAL PALEOHORIZONTAL MARKERS

One technique exists for restoring the original shape of a tectonically deformed pillow, as long as the strain is homogeneous. The technique requires a knowledge of strain (for example, from pillow selvages) in the plane of observation. By graphically removing the effects of strain it is possible to reproduce the original shape and orientation, and therefore the true bedding and way-up directions (see Borradaile 1982, fig. 13). During the present study it has been possible to utilize (and indeed to test) this method because special pillows exist that contain internal paleohorizontal markers. These were discovered by Laporte (1975, 1983). He termed such structures "eyebrows", but we refer them as "lava-shelves" for the reasons mentioned previously.

LAVA-SHELVES

The most useful evidence comes from an unnamed island of low strain southeast of Kudlulik Peninsula (Fig. 19). The strain ellipse ratio determined from selvages of pillows in plan view was 2.52 ± (δ) 0.55 (n=14). On this small, reef-like island bedding is well shown by 10-20 cm thick lapilli tuffs (Fig. 5D) which are stratified with pillowed flows. The tuffs are graded and therefore provide good way-up evidence as well as the orientation of stratification.

Within many pillows there occur sub-elliptical or amoeboid lenses of paler coloured lava, within which there are one or more elongate cavities filled with calcite (Fig. 20A). In all cases these cavities are discoidal in shape with planar bases and convex tops. Some plano-convex lenses are filled with a combination of pale vesicular lava and calcite, in which case they are embedded in somewhat darker green meta-lava. These general features are shown in the sketch (Fig. 20B).

Regardless of the mechanism of formation of these "lava-shelves" it is clear from the field relations on the island that

a. they are parallel to bedding at that level of strain

b. they have the convex side of the lense pointing in the local way-up direction.

On Tudlik Peninsula (Fig. 19), and on the large island west of Tudlik Peninsula, it appears that such "lava-shelves" rapidly transpose towards the cleavage and pillow long axis direction as the strain approaches an ellipse ratio of 4:1 in the plane of observation. In a previous study it was shown that pillow basalts in the Superior Province record signs of heterogeneous strain at a ratio of 4.5:1. Characteristics of heterogeneous strain found there and in this study include the following:

a. the pear-shape of pillow termini causing unusually large strain of the selvage.

b. the folding of pillow outlines

c. heterogeneous strain and "streaking" of amygdule inclusions in the pillow.

Clearly, where any of the aforementioned features occur it is likely that the orientation of "lava-shelves" will not truly indicate the bedding or way-up because they may have experienced excessive (rigid body) rotation beyond that of the degree of passive rotation dictated by homogeneous strain. Figure 21 shows the degree of deformation and transposition of pillows where it becomes impossible to use "lava-shelves" for bedding determination. Even the convex surface of the "lava-shelf is indeterminable at strains of 5:1 and nothing can be determined except the overall structural facing direction from the majority of molar tooth pillows (cf. Fig. 18).

THE AGREEMENT BETWEEN STRAIN DETERMINED FROM PILLOW SELVAGES AND "LAVA-SHELF" ORIENTATIONS

It has been shown that the strain ellipse for a section passing through the central third of a pillow is given by the selvage ratios (Borradaile, 1985). Another more complex strain analysis is possible if the pillow contains a paleohorizontal indicator and if the pillow is pod-shaped rather than tubular. Investigators in the Rankin Inlet region have been unable to identify any tubular pillow structures with "lava-shelves". The degree of exposure is sufficient to ensure that the pillows measured (and most others observed) were pod-shaped.

Where a "lava-shelf" occurs as shown in Figure 22 the tectonic strain ellipse ratio $R_T^{\frac{1}{2}}$ is given by

$$R_T = \frac{\tan\theta^1 (\tan^2\phi + R_T) - \tan\phi (R_T - 1)}{\tan\theta^1 (R_T \tan^2\phi - 1) - \tan\theta^1 \tan\phi (R_T - 1)}$$

where $R_T^{\frac{1}{2}}$ is the pillow's shape ratio in the deformed state. The strain ratios determined from this "single particle" and from the selvage ratio method agree closely (see Fig. 22).

THE ORIGIN OF THE "LAVA-SHELVES" AND THEIR USE IN STRUCTURAL MAPPING

Structures like "lava-shelves (Fig. 5C, 20A) reported by Laporte (1975, 1983) have been documented several times in the literature. Before reviewing their interpretation and use attention should be drawn to the important items of concern. Firstly, are the "lava-shelves" truly paleohorizontal markers? Fortunately, the "lava-shelves exposed near the village of Rankin Inlet occur with interbedded sedimentary rocks. In several cases graded tuffaceous sediments and lapilli tuffs confirm that the planar base of the "lava-shelf" is parallel to bedding in weakly deformed rocks. Also, the convex surface of the "lava-shelf" does point to the original top of the local sequence. Secondly, do the "lava-shelves" occur in pod-like lava pillows or tubular pillows? Investigations of good three-dimensional outcrops show that in the present area tubular pillows cannot be identified. This is important in strain analysis because the single-particle strain method of Borradaile (1984), discussed in the previous section and in Figure 22, will not work due to an unfavourable cut-effect of tubes. Laporte's (1983) observations were that "lava-shelves" occured only in pod-like pillows in the Rankin Inlet area.

The first reports of structures like "lava-shelves" appear to have been made by Fuller (1931) while Waters (1960), Wells et al. (1979) and Ballard and Moore (1977) have also documented them. More recently, such structures have been reported by Sawyer et al. (1983) from the Cape Smith Fold Belt, and by Baragar (1984) from the Aphebian volcanic rocks of the Ottawa Islands and Belcher Islands in eastern Hudson Bay. Some of these authors refer to "lava-shelves" as lava-levels or drain-away structures implying a genesis through partial drainage of a tube or pillow to produce the plano-convex cavity in cross-section. The cavity is presumably filled with water initially (it is filled with calcite in the field examples at Rankin Inlet). In Ballard and Moore's diagram and in examples from the Ottawa Islands and Belcher Islands (Baragar, 1984) it is clear that they recorded examples in tubular pillows while the other authors did not specifically note this. Laporte (personal communication, 1985) indicated that the drainage of a pod-like pillow may occur where a pillow buds a smaller pillow. We have not been able to verify independently that part of the hypothesis but drainage of pillows is confirmed by the presence of lava drips with a "stalactite" form hanging from the lower surface of lava shelves (Fig. 23A). The plano-convex form of the "lava-shelf" confirms the way-up and bedding directions (Fig. 23B), the paler-coloured lava may indicate the refilling of a pod with lava later (Fig. 23C) or,since chilled surfaces are absent, it may indicate an immiscible central phase, or it could represent an alteration halo (Baragar, personal communication, 1985). The multiple "lava-shelves" (Fig. 23A) are a common feature but are difficult to explain in any of the tentative hypotheses that Laporte or the present authors have been able to form.

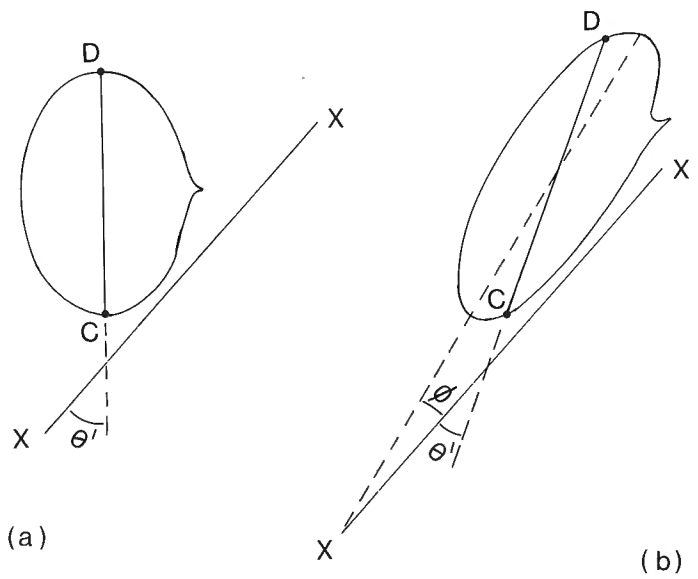


Figure 22A & B. Empirical relationships between material lines and points before (a), and after (b) homogeneous strain. Line CD refers to a planar base of an eyebrow structure; line XX is an arbitrary reference line.

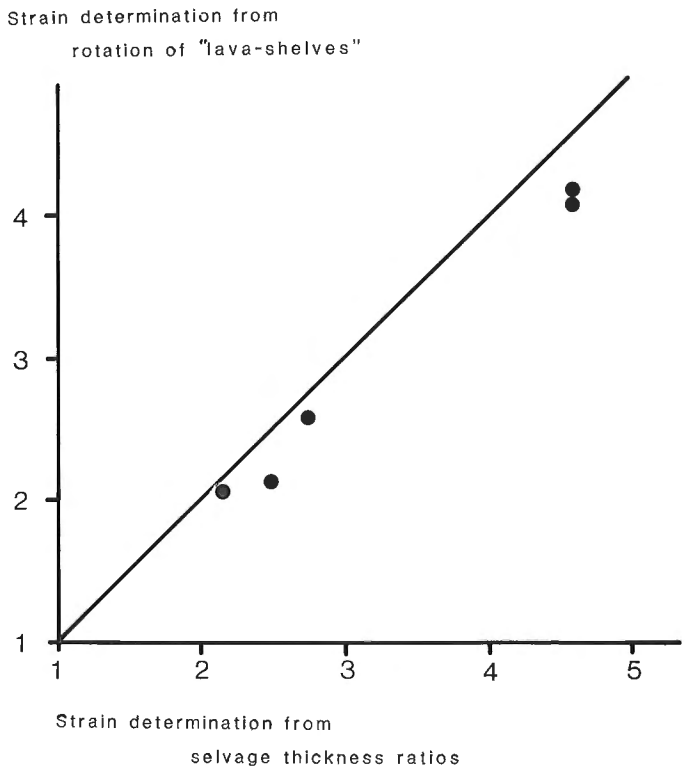


Figure 22C. Correlation between the strain ratio determined from pillow selvages and from orientation of "lava-shelves".

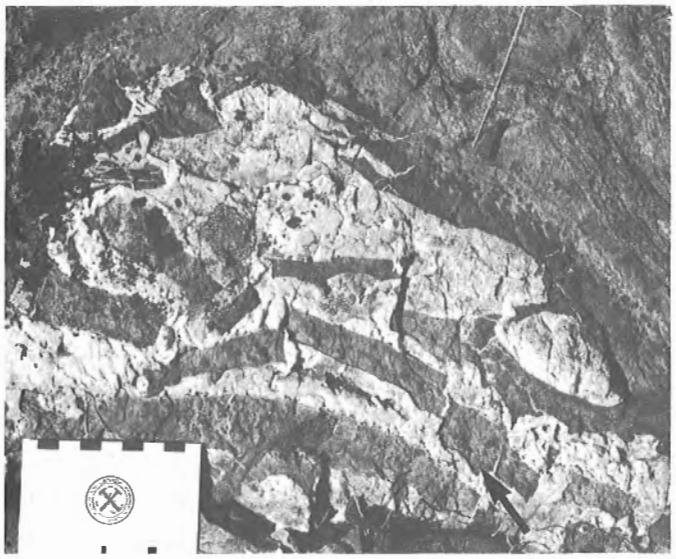


Figure 23A. Multiple "lava-shelves" showing "stalactite" form lava-drips from the lower surface of lava shelves. (The shelves have been somewhat disrupted subsequently but a lava-drip is arrowed). (see Fig. 2 for location). GSC 204040-C

Figure 23B. Plano-convex form of a "lava-shelf" showing way-up (to top left) and bedding direction. (see Fig. 2 for location). GSC 204040-D

Part 3: CONCLUSIONS

The rocks within the Rankin Inlet Group record at least two cycles of ?Archean volcanism. A discontinuous, but apparently conformable, unit consisting of polymictic conglomerate marks a hiatus between the two volcanic cycles. The structural data suggest that the greenstone-metasedimentary rocks of the Rankin Inlet Group represent an east dipping F_1 homoclinal sequence that has been subsequently modified by a southeast plunging F_2 upright syncline. The stratigraphic younging and structural facing relationships in the metavolcanic rocks indicate structural discordance between Thomson Island and Kudlulik Peninsula, and between Kudlulik Peninsula and Tudlik Peninsula. An attempt is made to explain this discordance by proposing two tectonic contacts between the three regions, and by invoking a thrusting model in which two opposing tectonic motions (northeast and southwest) between two thrust sheets may

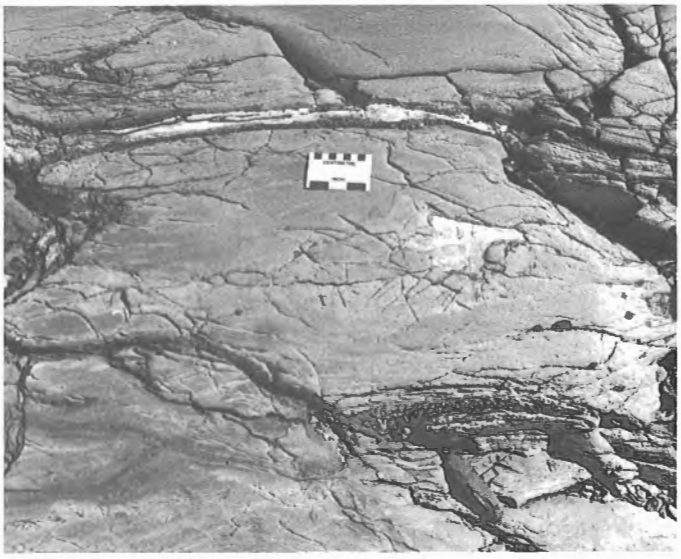


Figure 23C. Pillow in metavolcanics of the Rankin Inlet Group showing paler coloured lava in the central part of the pillow. (see Fig. 2 for location). GSC 204040-E

juxtapose rocks of different structural geometry. The orientation of mineral stretching lineations (L_1) offers some support to the model, but further structural studies are required to substantiate the model.

Strain ratios based on deformed pillow selvage thicknesses from lava pillows of the Rankin Inlet Group range from 2.0 to 6.0; the high ratios commonly coincide with northwest trending and southwest verging ductile shear zones- tentatively interpreted as high angle (?thrust) faults.

It is pointed out that "lava-shelves", the paleohorizontal and stratigraphic facing markers, are unreliable as primary facing markers when the axial strain ratios in pillows are greater than 4:1.

The paleocurrent data from a ?Lower Proterozoic orthoquartzite sequence indicate sediment transport direction from the southwest and from the south-southeast on Kudlulik Peninsula and Marble Island respectively. The contacts of the orthoquartzite sequence with the older rock units are interpreted as possible thrusts.

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