A REINTERPRETATION OF THE STRATIGRAPHY AND DEFORMATION OF THE AILLIK GROUP, MAKKOVIK, LABRADOR

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STRATIGRAPHY AND DEFORMATION
OF THE AILLIK GROUP,
MAKKOVIK, LABRADOR

by

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ABSTRACT

The Aillik Group of Aphebian volcanic and sedimentary rocks is situated in the Makkovik Subprovince of Labrador, north of the Grenville Front. Its lithology, in the vicinity of Makkovik, has been completely reinterpreted and the first formal stratigraphic subdivision of the group defined, with introduction of new names. The previously unrecognised sequence of deformational events and associated metamorphism has been delineated.

The 8500 m. (25,000 ft.) thick Aillik Group consists of cratogenic flow-banded and porphyritic rhyolite lavas, (previously mapped as quartzites), minor interbedded basalt lavas, polymictic and rhyolite conglomerates,and quartz-poor arkoses (also previously mapped as quartzites). The group is subdivided into six formations. The lowest (Nesbit Harbour Formation), comprising 800 m. to 2000 m. (2400 - 6000 ft.) of conglomerate with minor arkoses and basalt lavas, is conformably overlain by the Makkovik Formation of 2000 m. to 7000 m. (6000 - 21,000 ft.) of rhyolite lavas, with minor tuffs, conglomerates and arkoses. This is conformably overlain by two stratigraphically equivalent (?) formations:- the 2000 m. (6000 ft.) thick Pomiadluk Formation of interbedded rhyolite, conglomerate and tuff in the northeast; and the 2000 m. (6000 ft.) thick Manak Bay Formation of arkose and minor basalt lava

in the southeast.

West of this sequence, and separated from it by a major tectonic slide (and the associated, structurally defined, Ranger Bight complex of amphibolites), is the 3000 m. (9000 ft.) thick Big Island Formation of interbedded rhyolite, arkose, conglomerate and basalt lava.

The Aillik Group is intruded by several major syntectonic granitic to gabbroic plutons including a large, rhymically layered, hornblende gabbro (appinite?) intrusion, the Adlavik Igneous Complex.

Four deformations are recognised, of which the second produced penetrative fabrics and large, upright, northeast trending, folds. The others developed only local structures and fabrics, and one major interference dome. The second deformation resulted in different amounts ($r = 1$ to 5) and types ($k = 1$ to ∞) of strain, (which is not lithologically controlled), in different areas. The entire area is shown to possibly occur within a major zone of strike-slip (simple-shear) movement.

Metamorphism has remained in the amphibolite and greenschist facies throughout tectonism.

The Aillik Group is unconformably overlain by the Jurassic Ford's Bight Conglomerate.

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Dr. E. R. W. Neale, Professor and Head of the Department of Geology;

Dr. K. D. Collerson, Assistant Professor of Geology;

Dr. A. R. Berger, Assistant Professor of Geology;

Mr. R. Stevens, Assistant Professor of Geology;

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CHAPTER I

INTRODUCTION

Location

The area of this study is located in the Makkovik region on the Coast of Labrador between longitudes 58° 55' W and 59° 25' W, and latitudes 54° 55' N and 55° 10' N. The study area comprises approximately 250 square kilometers (100 square miles) bounded by Makkovik Bay on the west and northwest, by the Labrador Sea between Pomiadluk Point and Adlavik Bay on the east, and by an irregular line from the head of Makkovik Bay eastwards towards Adlavik Bay (Figs. 1-1, 1-2 and Plate 1).

The settlement of Makkovik, with a population of approx. 350 persons is in the north of the area, and is the only settlement in the vicinity. Makkovik has a general store, Moravian Mission church and International Grenfell Association nursing station, and is served in the summer by the C.N. coastal boat service.

Physiography

The entire area has been heavily glaciated leaving a topography of rounded hills and hillocks, with a few sharper peaked hills over 300 m. (1000 ft.) above sea level, separated by lakes, ponds and bog in till-filled valleys.

Fig. 1-1. Map showing location of area and outline of structural provinces of Labrador.

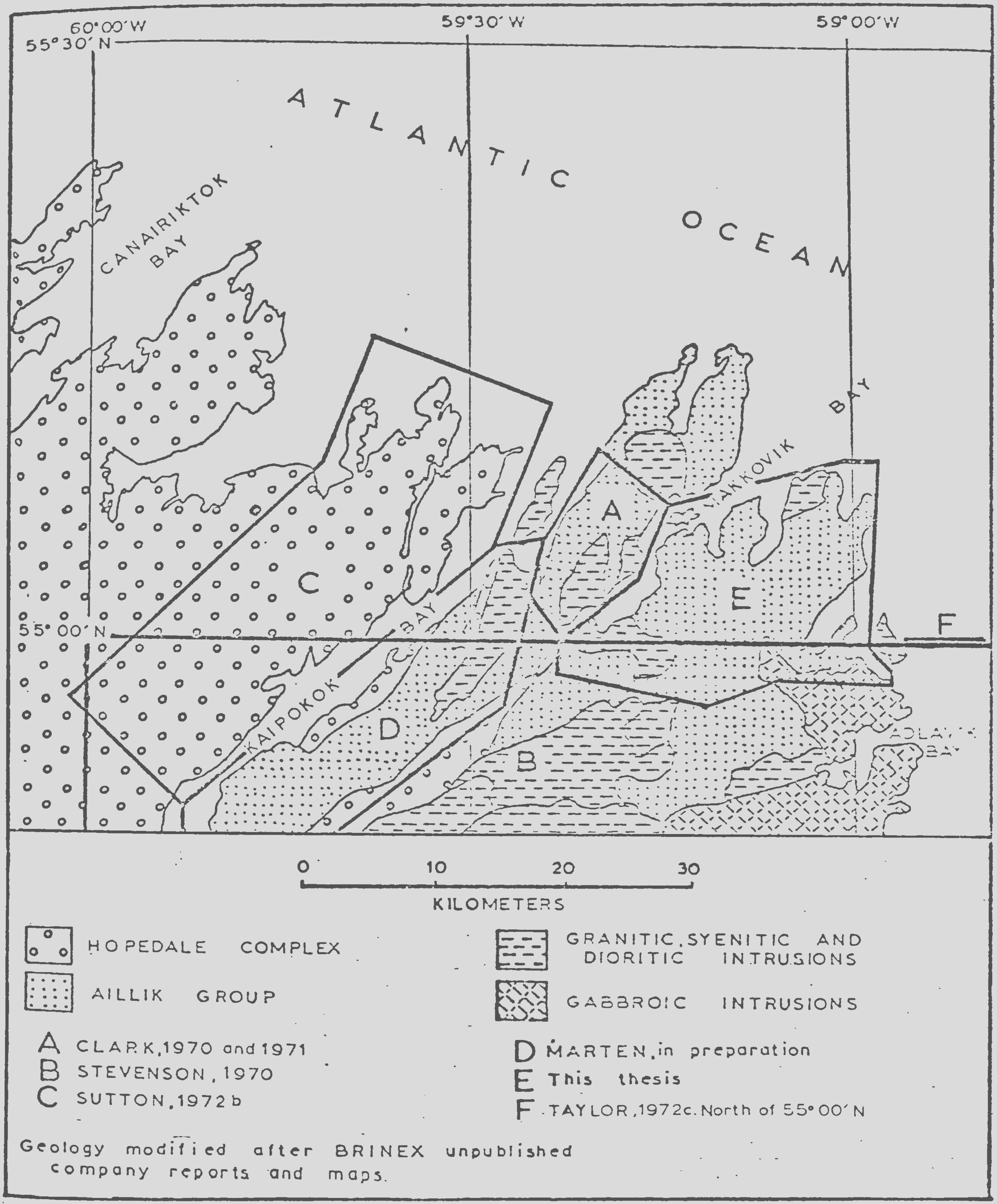

Fig. 1-2. Map showing location of thesis area and other recent work in the region.

Drainage is generally poor, though locally waterfalls and rapids are developed. Multiple raised beaches are common, with the highest up to 30 m. (100 ft.) above sea level. Vegetation is of arctic type, composed, for the most part, of stunted evergreens in a carpet of hardy shrubs, moss and lichen. However, good stands of timber do occur in sheltered areas. Outcrops, though abundant, are heavily encrusted with lichen.

Geographic Nomenclature

The correct spelling of Aillik, according to the Gazetteer of Canada, (Dept. of Energy, Mines and Resources, 1968) is the spelling used in this thesis. However, the spelling 'Ailik' is used by some authors, and also occurs on the 1:50,000 scale Provisional Topographic Map of the area (Dept. of Mines and Technical Surveys, 1963) and on the earliest known map of Labrador (O'Hara, 1872).

The following place names are not officially recognised:

Cross Lake
Round Pond
Falls Lake
Grampus Cove
Present Lake

However, most of these names have been regularly used by the people of Makkovik or have been introduced by BRINEX geologists over the last two decades.

General Geology

The thesis area is situated within the structurally defined Makkovik Sub-province of the Nain Province (Fig. 1-1; Taylor, 1971, 1972a & b). The area is underlain by the Aillik Group of early Proterzoic age (Gandhi, et al., 1969) which overlies the Hopedale Complex of banded gneisses (Sutton, 1972b) to the northwest and probably also the Domino gneiss (Kranck, 1953; Stevenson, 1970) to the southeast. The Aillik Group comprises an assemblage of lithologies of essentially shallow-water cratonic, as opposed to orogenic, affinities and consists of at least 8500 m. (25,500 ft.) of metamorphosed sedimentary and volcanic rocks--primarily arkoses, conglomerates and rhyolitic lavas, but also including mafic lavas. Most of the rocks in the group are genetically related to one another and formed a composite acid/basic volcanic terrane with locally derived sediments, which was deformed and metamorphosed to the greenschist/amphibolite facies during the Hudsonian orogeny (Gandhi, et al., 1969). Lithologies previously described as sedimentary quartzites (Gandhi, et al., 1969) are shown to consist almost entirely of rhyolitic lavas, with minor quartz-poor arkoses, and tuffs, as previously suggested by Bridgwater (1970) and Taylor (1971).

The Aillik Group has been intruded by major syn- and post-tectonic intrusions of granitic to gabbroic composition and by dykes of all ages ranging from pegmatite to hornblendite, and including Helikian and Hadrynian diabases and Cambrian lamprophyres (Gandhi, et al., 1969).

The area has undergone several phases of deformation during the Hudsonian orogeny, only one of which, the second, is widespread and resulted in penetrative fabrics. This deformation developed broad, large-scale, open to close (Fleuty, 1964a) folds throughout the area, whereas the other deformations resulted in local, mainly small-scale, folding only. The major structural features developed during the second deformation are: a large central doubly-plunging, northeasterly-trending anticline, slightly smaller folds to east and west of this anticline, and a major tectonic slide of unknown displacement in the west of the area.

Previous Geological Work

The first geological reconnaissance of the area was carried out as part of voyages of discovery by Steinhauer (1814), Lieber (1860), Packard (1891), Daly (see Delabarre, 1902) and others. Observations were of a very general nature and were based on brief investigations of randomly scattered areas along the coast. Daly was the first to mention the geology of the Makkovik Region when he noted the large number and wide range of compositions of the dyke rocks intruding the sediments at Aillik Bay.

More systematic investigations were begun by Wheeler (1933 and 1935) who studied the petrology of some diabase dykes and an amazonite aplite from the region. The sediments of the Makkovik/Aillik coastal area were more thoroughly mapped by Kranck (1939 and 1953) who introduced the term "Aillik series" for these sediments, and who carried out mineralogical and chemical analyses of some of the sedimentary and igneous rocks. Later work has essentially confirmed Kranck's original outline of the geology.

The broad regional relationships of the Makkovik area, both within Labrador and with Greenland, have been reported by Douglas (1970), Bridgwater (1970), Greene (1972), Greene and McKillop (1972), Sutton (1972a and b; Fig. 1-2) and Sutton, et al., (1972). Large scale regional geological mapping has been carried out under the auspices of the Geological Survey of Canada since the early 1950's and has lead to the publication of regional maps with descriptive notes by Douglas (1953), Christie, et al., (1953), Stevenson, (1970) and Taylor (1972c) who noted the large amount of acidic volcanic material in the Aillik Group. Approximately 50 miles inland along strike mapping has been carried out in the Croteau Group by Williams (1970).

The Geological Survey of Canada has also carried out radiometric age-dating as published in Lowden (1961), Leech, et al., (1963), Stockwell (1964) and Wanless, et al., (1970 and 1972), with the purpose of subdividing the Canadian Shield into provinces (Stockwell, 1968). Further work on definition, subdivision and discussion of the structural provinces has been reported by Taylor (1971 and 1972a and b), Fahrig, et al., (1972) and Douglas, et al., (1972).

Other radiometric ages are reported in Grasty, et al., (1969) and Gandhi, et al., (1969).

The most comprehensive map and report of the Geology of the area is by Gandhi, et al., (1969) in which the stratigraphy, lithology and igneous petrology of the different rock-types of the area are described. Cooper (1951), Moore (1951) and Riley (1951) have written master's theses based on investigations of the syenites, granites, igneous rocks and, to a lesser extent, the sediments of part of the area. Other master's theses are by King (1963b) on the lithology, metamorphism and stratigraphic succession on the Cape Makkovik peninsula; by Gill, (1966) on the petrography of the molybdenite-bearing gneisses; by Barua (1969) on the geology, mineralisation and geochemistry of the rocks of part of the Cape Makkovik peninsula; and by Clark (1970 and 1971) on the structure and lithology of the main body of the Makkovik peninsula. Stoeterau (1970) has written a bachelor of science (honours) thesis on the lithology and structure of part of the area discussed in this thesis. B. E. Marten (Memorial University of Newfoundland) and M. V. White (McGill University) are both completing theses (Ph.D. and M.Sc. respectively) on the rocks of the Hopedale Gneiss/Aillik Group contact area. Marten is working on the structural and tectonic relationships of the basement and cover rocks, whereas White is undertaking chemical analyses of the respective lithologies.

British Newfoundland Exploration Limited (BRINEX) has been conducting a thorough mineral exploration and geological reconnaissance program in the Makkovik region since 1953 (Beavan, 1958), but ceased work in 1971. Other work included an investigation of the basement Hopedale Complex (Sutton, 1972b) and its relationships to the Aillik Group (Sutton, et al., 1971).

Purpose of Present Study

Previous work in the Makkovik region (Kranck, 1953; Gandhi, et al., 1969) indicated the region was underlain by a sedimentary and volcanic sequence of quartzites, conglomerates and mafic lavas, which had undergone simple tectonic deformation resulting in, for the most part, gentle folding. However, work carried out by the author in an adjacent area (Clark, 1970 and 1971) indicated the Aillik Group consisted of arkoses rather than quartzites, which were derived from an, as then unrecognised, acid volcanic terrane. Furthermore, it was also shown that the area had undergone complex polydeformational tectonics resulting in both large and small scale refolded upright and recumbent folds. Thus, it was shown that both the lithology and the structure (and hence also the stratigraphy) were in doubt. As the area contains a few scattered sub-economic uranium and molybdenum mineral showings and is on strike with potential economic mineral deposits to the southwest towards McLean and Seal Lakes, a thorough understanding of the structure, lithology and stratigraphy would be both interesting in itself and of possible economic importance in determining the controls of ore deposition.

Methods of Investigation

The area was mapped in the field on a scale of 1" to 2000' (1:24,000) using topographic maps and aerial photographs of the same scale. The topographic maps used were enlargements of 1:50,000 scale Provisional Topographic maps compiled by the Surveys and Mapping Branch of the Department of Mines and Technical Surveys and published in 1963.

Modal analyses were carried out on a point-counter using stained thin-sections to facilitate distinguishing between untwinned K-feldspar, untwinned plagioclase and quartz (Fig. 1-3). A total of 2000 to 4000 counts were made on each thin-section, and from one to six or eight (but usually two or three) thin-sections were made of each lithological unit. The staining method employed was that described by Baily and Stevens (1960). Plagioclase compositions were determined by comparing the refractive indices with K-feldspar, quartz and balsam. Where necessary, the 2V and absolute refractive indices were determined using a universal-stage and refractive index liquids, and the composition read from the graphs in Slemmons (1962).

Where the opaque mineralogy is specified, it was determined using normal reflection (ore) microscopy methods (Cameron, 1961 and Schouten, 1962).

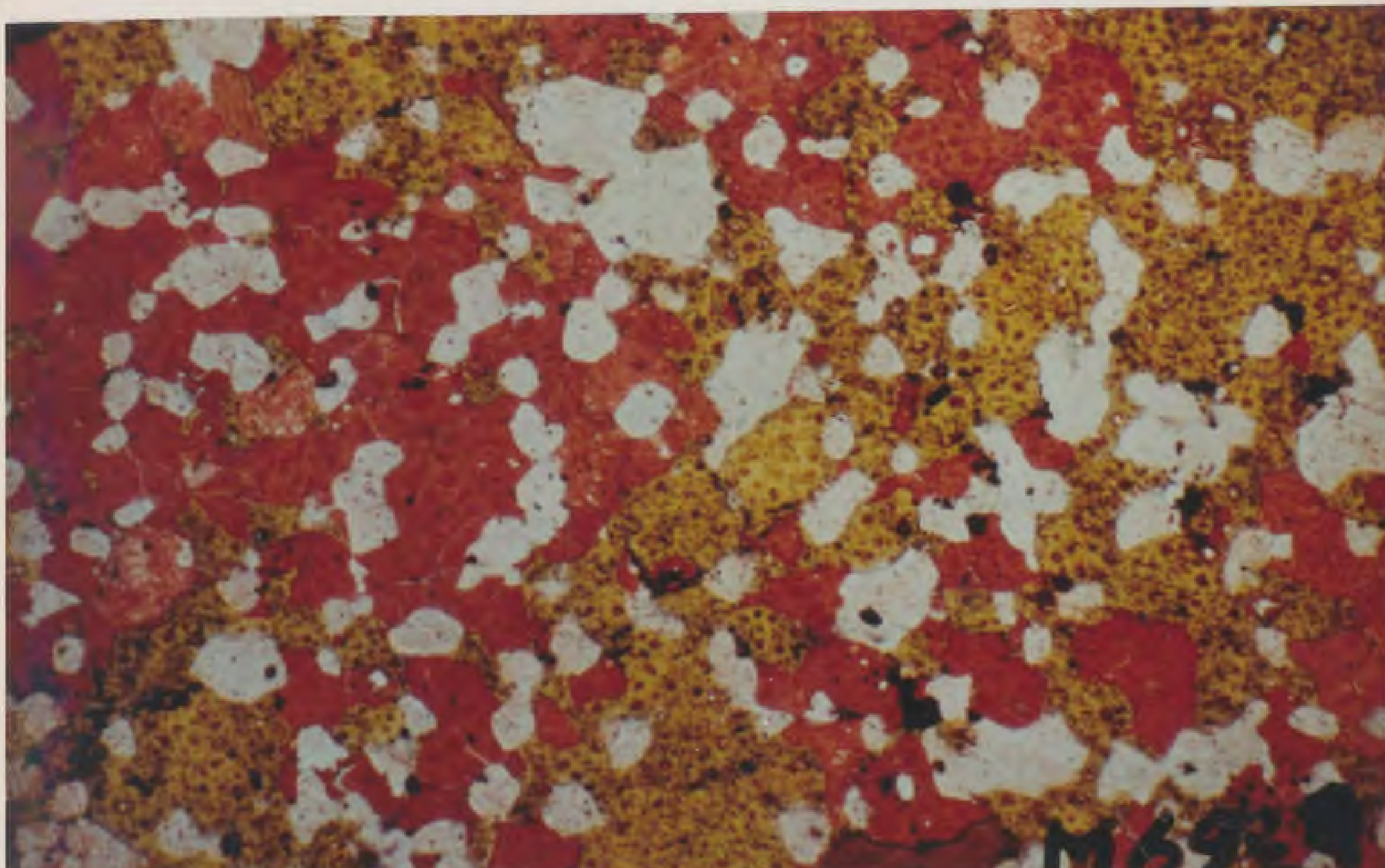

Fig. 1-3. Photomicrograph of stained thin-section of arkose showing staining colours: yellow--K-feldspar, red--plagioclase, white (unstained)--quartz. (Red and brown spots on K-feldspar are the results of the staining-method and not due to the mineralogy). Plane polarised light, x 300.

The structural and petrofabric investigations were carried out following the ideas, theories and methods of Flinn (1962 and 1965) and Ramsay (1967).

Summary of Results of Study

The previous impression (gained through work in an adjacent area; Clark 1970 & 1971) that the sediments are arkoses and not quartzites and are probably derived from an acid volcanic terrane, is shown to be correct. However, it is also shown that the majority of the rocks previously mapped as quartzites are in fact rhyolite lavas and subsidiary tuffs.

It is also shown that the tectonic history of the area is complex, as previously envisaged, but that the tectonism has resulted in a simple structural pattern of upright open to close folds, with one major tectonic slide-zone.

The simple structural pattern has facilitated a more detailed stratigraphic investigation and several formations of the Aillik Group are described and formally defined for the first time.

It has not been possible to determine the controls of uranium and molybdenum mineral deposition from the few widely scattered, low-grade occurrences in the area.

CHAPTER 2

LITHOLOGY AND STRATIGRAPHY

Introduction

The thesis area is underlain by the approximately 8 km. (4 mi.) thick Aillik Group of Proterozoic acid volcanic rocks (previously mapped as quartzites) and associated sediments and minor basaltic lavas.

One very small outcrop of previously unofficially recorded Jurassic conglomerate has been recognised in Ford's Bight, and is the only pre-Quarternary Phanerozoic sediment known from this part of Labrador.

No basement rocks to the Aillik Group have been recognised in the area, though the Hopedale Complex (Hopedale Gneiss, Gandhi, et al., 1969; Sutton, et al., 1971; Sutton, 1972b) forms the basement further west at Kaipokok Bay.

Stratigraphic Nomenclature

The Code of Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature, 1970) has been used in naming the various rock-stratigraphic units. All the formal stratigraphic names are new except for "Aillik Group" (King, 1963; Stevenson, 1970). All stratigraphic names that have been discarded were of unofficial status according to the stratigraphic code, and most of the names were descriptively misleading. The units informally termed 'members' in this thesis were mapped on a sclae of 1 : 24,000 and therefore could qualify for formational status (American Commission on Stratigraphic Nomenclature, 1970, Article 6d). However, in order to prevent the occurrence of isolated informal units in the group (op. cit., Article 9a), or the occurrence of a large number of formational names (op. cit., Article 7b) the present scheme is preferred. All type-sections are marked on the map (Plate 1).

The terms 'young' and 'face' are used to refer to the direction in the sediment in which the units get younger (Shackleton, 1957).

The term 'band' (American Geological Institute, 1962) is used, where possible, with a qualifying term. The genesis of banding of various types is discussed in Chapter 4 (Structure).

Aillik Group(Proterozoic)

GENERAL DESCRIPTION

The Aillik Group of early Proterozoic acid and basic volcanic rocks and related sediments has a maximum thickness of 8500 m. (25,500 ft.) in the thesis area and has previously been mapped and sub-divided into several sedimentary units by Gandhi, et al., (1969), the most extensive of which are acid volcanic rocks as previously suggested by Bridgwater (1970) and Taylor (1972c). These units (members) have been grouped together here into six formations (Fig: 2-1). This grouping is based partly on the occurrence of major tectonic breaks and partly on mapping convenience, rather than on any major differences in depositional environment. The oldest is the Nesbit Harbour

Table 2-I

Table showing approximate comparison of nomenclature.
(Not in stratigraphic order)

Clark (1970 & 1971) (in adjoining area to west)	Gandhi, et al., (1969)	This Thesis
Porphyroclastic psammite	Feldspar porphyroblastic arkosic quartzite	Quartz-porphyritic rhyolites
	Lineated grey feldspathic quartzite	Microcline-porphyritic rhyolites, and Makkovik Formation tuff.
Variable psammite	Feldspathic quartzite (variable lithology)	Flow-banded rhyolites.
	Thin-bedded quartzite	Arkoses, tuffs, plagioclase-porphyritic rhyolite & some flow-banded rhyolites.
Amphibolite	Mafic lava (amphibolite)	Metabasalts, amphibolites & Ranger Bight Complex.
Conglomerate	Conglomerate	Conglomerates
Cross-bedded psammite & Banded psammite	Banded quartzite (varicoloured)	Arkoses and tuffs
	Granoblastic feldspathic quartzite	Round Pond Granodiorite

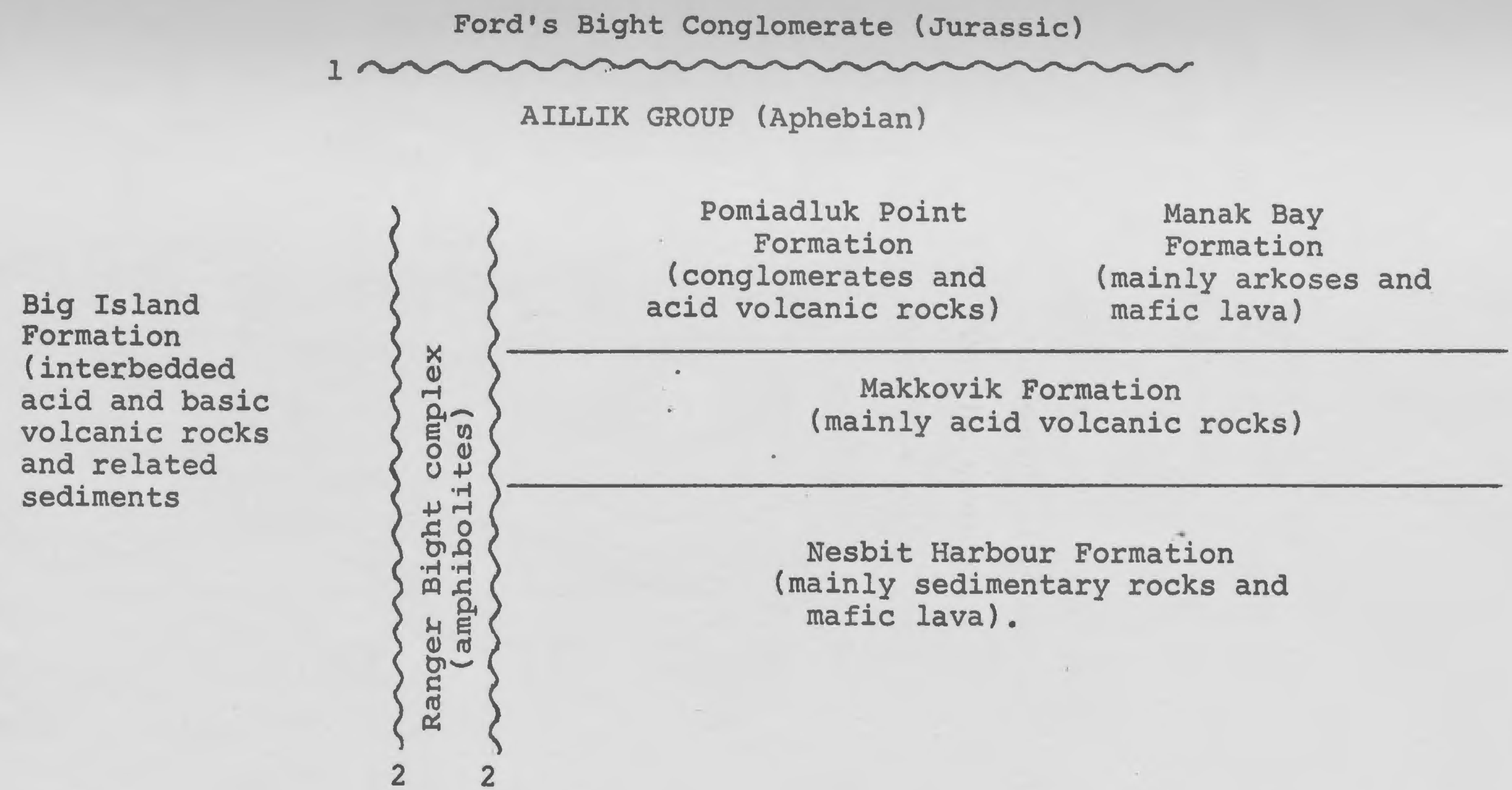

Fig. 2-1. Schematic diagram showing the relationship of formations in the area to one another.

[1]Major angular unconformity

[2]Tectonic contact.

Formation (composed primarily of arkose, conglomerates and mafic lava) which is conformably overlain by the Makkovik Formation (composed primarily of acid volcanic rocks previously mapped as quartzites). The Makkovik Formation is overlain by both the Pomiadluk Point Formation in the northeast (consisting primarily of acid volcanic rocks, also previously mapped as quartzites, and polymictic conglomerates), and the apparently stratigraphically equivalent Manak Bay Formation in the southeast (composed primarily of arkoses and mafic lava). To the west of these formations is the Big Island Formation of acid and mafic volcanic rocks, conglomerates and arkoses which is separated from the other four formations by a major tectonic slide zone containing the Ranger Bight complex of amphibolitic rocks.

Only one disconformity (between the Makkovik and Manak Bay Formations) and no angular unconformities have been recognised between members and formations of the Aillik Group and contacts are therefore termed "conformable". However, metamorphic recrystallisation and tectonic flattening undergone by the group may well have obliterated evidence of disconformity.

Grain sizes mentioned for members of the different formations of the Aillik Group refer, in all cases, to the present grain size which is due to secondary metamorphic recrystallisation.

All the arkoses in the Aillik Group have very low quartz content, intermediate microcline content, and fairly high to very high albite content suggesting possible soda-metasomatism. However, the general lack (except for one very minor occurrence at Big Bight) of sodium-rich amphiboles or other sodium-rich minerals suggests soda-metasomatism was not the cause of the high albite content. Also, sodium in the albite is usually in the approximate proportion to calcium in the calcium-bearing minerals (andesine, epidote, calcite, etc.) that would be expected if oligoclase had undergone metamorphic differentiation (peristerite separation; Brown, 1962) and subsequent recrystallisation of the constituents into different mineral species, suggesting that it is more an apparent than a real enrichment in albite. Furthermore, if soda-metasomatism had occurred, the rhyolites, and possibly the early intrusive rocks (Long Island Gneiss and Kennedy's Cove Gneiss), would also have been affected, but they have normal albite contents. Soda-metasomatism is not considered, therefore, to have taken place in the Aillik Group in this area.

AREA EAST OF RANGER BIGHT SLIDE

Nesbit Harbour Formation

The name Nesbit Harbour Formation is proposed for a mixed sedimentary and volcanic sequence of arkoses, conglomerate, rhyolite lava and mafic lava that occurs due north of Monkey Hill (the location of the type-section is plotted on Plate 1). The base of the formation has not been recognised as its location is obscured by the intrusive Round Pond Granodiorite. The top of the formation is defined as the top of the metabasalt member (Table 2-II and Figs. 2-2 and 2-3).

The Nesbit Harbour Formation is the oldest formation in the thesis area and occurs as an annular outcrop pattern around Round Pond in the center of the area. It varies in true thickness from 800 m. to 2000 m. (2400 to 6000 ft.) and crops out over an area of approximately 14 km. (7 mi.) long and 6 km. (3 mi.) wide.

The formation was deposited under subaqueous conditions for the most part as shown by cross-bedding in the arkoses, and probably under subaerial conditions during the extrusion of the mafic lavas, as shown by a lack of pillow-structures (which are extensively developed in other formations).

i. Lower Arkose Member: The Lower arkose member is poorly exposed as an incomplete annular outcrop around Round Pond, with a gap, due to intrusion of the Round Pond Granodiorite, where Makkovik Brook drains from Round Pond. The base of the member has not been recognised.

The member is most clearly exposed on Monkey Hill where it is approximately 200 m. (600 ft.) thick. The total outcrop length appears to be several kilometers, but the recognised outcrops are widely separated.

The arkose (Williams, et al., 1958) is a fine-grained, polygonally textured metasediment, and is bedded on a 1 to 8 cm. (1/2 to 4 in.) scale. This bedding is defined

Table 2-II

Table showing the approximate thickness of members of the Nesbit Harbour Formation

	m.	ft.
Metabasalt member	0-300	0-900
Upper arkose member	0-500	0-1500
Plagioclase-porphyritic rhyolite member	0-200	0-600
Conglomerate member	200-1000	600-3000
Lower arkose member	0-200	0-600
Nesbit Harbour Formation	800-2000	2400-6000

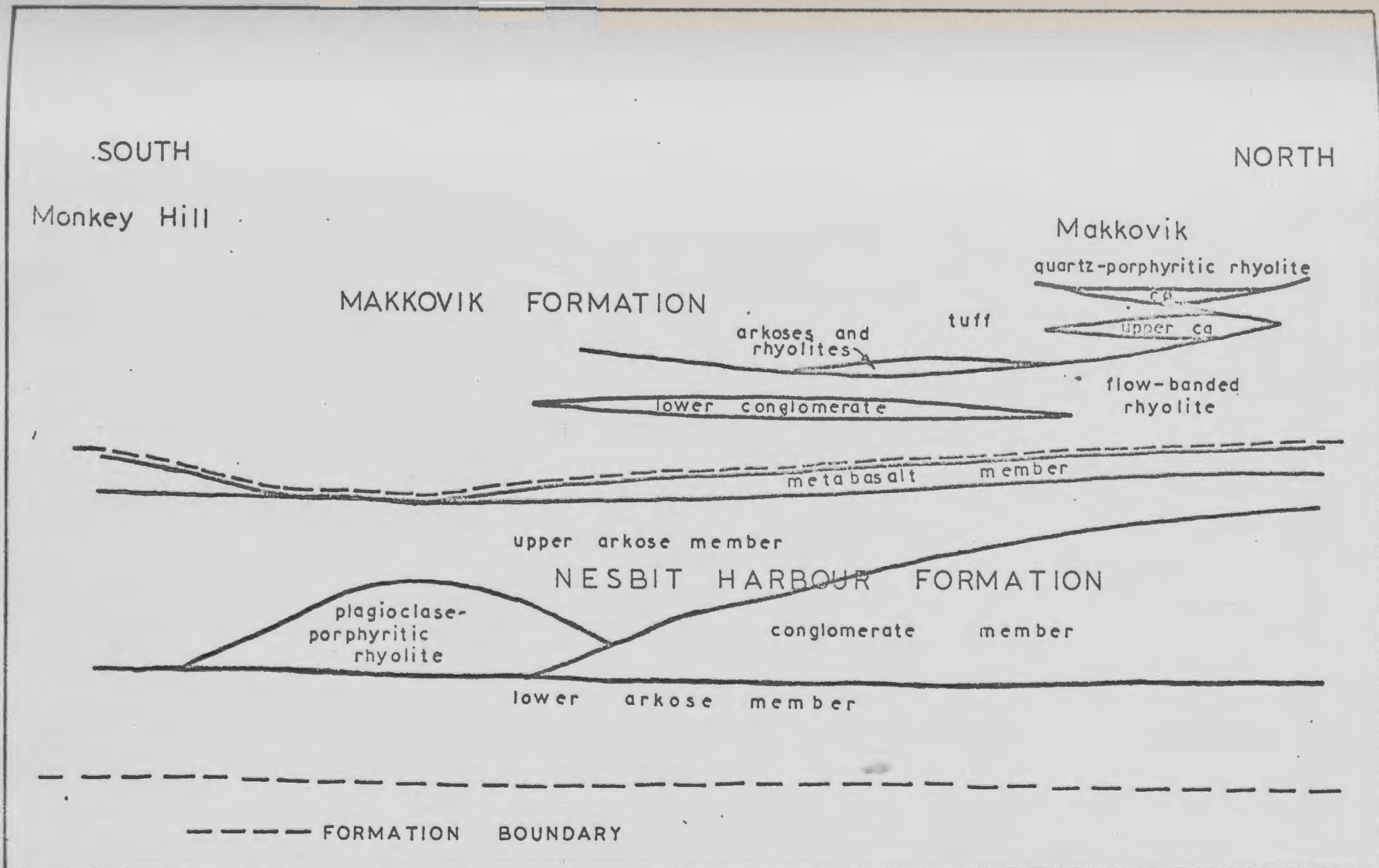

Fig. 2-2. Stratigraphic Diagram of the Nesbit Harbour and Makkovik Formations from Makkovik to Monkey Hill.

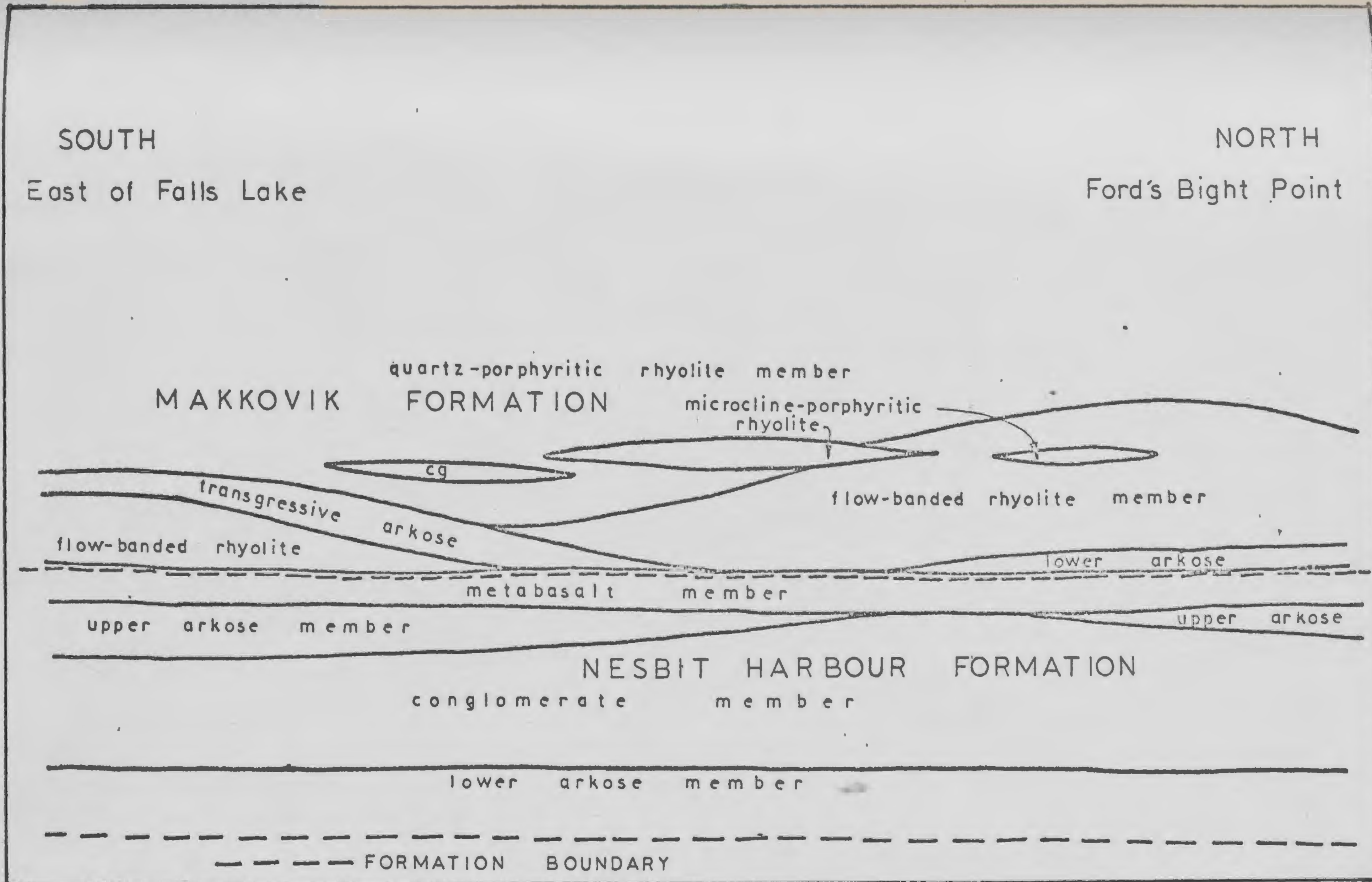

Fig. 2-3. Stratigraphic Diagram of the Nesbit Harbour and Makkovik Formations from Ford's Bight Point to east of Falls Lake.

by varying concentrations of mafic minerals and shows cross-bedding (on a 10 cm. (6 in.) scale) on Monkey Hill. Throughout most of the area the unit shows little variation in field appearance.

The unit is composed of:

albite	50%
oligoclase/andesine	20%
diopside	20%
phlogopite[1]	5%
microcline	2%

and accessory sphene and opaque minerals. No quartz has been recognised. In the vicinity of the Long Island Gneiss on Monkey Hill microcline porphyroblasts up to 2 mm. in diameter occur evenly distributed in the arkose. A thin graphite schist bed occurs on the eastern arm of Monkey Hill.

The regular banding suggests deposition under fairly stable rhythmic conditions. The lack of quartz and the high feldspar content suggest possible derivation from acid volcanic rocks (Williams, et al., 1958), followed by sedimentary sorting, or derivation from a dioritic source of which there is no evidence at present.

ii. Conglomerate Member: The conglomerate member occurs as a 20 km. by 5 km. (10 x 2 1/2 mi.) annular outcrop pattern about Round Pond, situated with sharp contact conformably

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[1]As phlogopite and common biotite have almost identical optical properties in thin-section (Deer, et al., 1970), the term "phlogopite" is reserved for pale brown to honey-brown micas, and "biotite" for the darker varieties.

above the lower arkose member and averaging from 200 m. to 1000 m. (600 ft. to 3000 ft.) in thickness. The conglomerate is most clearly exposed on the east face of Monkey Hill. It is polymictic and contains tectonically deformed, roughly ellipsoidal clasts set in a fine- to medium-grained matrix (Fig. I-1). The present range of clast sizes is approximately 2 cms. by 6 cms. to 15 cms. by 0.5 m. (1 in. x 3 in. to 6 in. x 18 in.), suggesting a predeformation clast diameter of 4 cms. to 30 cms. (2 in. to 1 ft.). The clast size does not vary in any regular manner within the conglomerate. There is generally no indication of pre-deformation angularity to the clasts, and the rock was, therefore, a cobble conglomerate (Pettijohn, 1957).

The most common clast-type, which forms 60-80% of the clasts, is a pale pink, yellow or grey, fine-grained, quartzo-feldspathic rock, generally showing no banding or bedding, and identical in appearance to the unbanded parts of the flow-banded rhyolite sequences (q.v.). Other rock types are fine- to medium-grained amphibolite and fine-grained grey micaceous pelite (?). In places pebbles of quartz-porphyritic rhyolite and of a grey fine-grained pumiceous rock are common (Fig. 2-4).

The matrix of the conglomerate is predominately quartzo-feldspathic in composition, but dark minerals or calcite can make up 20% of it in places. Where fragments of quartz-porphyritic rhyolite predominate, the matrix is composed of recognisable fragments of the clasts, including porphyroclasts.

Fig. 2-4. Pebbles and boulders of quartz-porphyritic rhyolite in quartz-porphyroclastic tuffite (?) matrix--Nesbit Harbour Formation conglomerate.

Fig. 2-5. Pillow-like structures in quartz-porphyritic rhyolite on eastern slope of Monkey Hill. Bedding and S_2 tectonite fabric parallel the long axes of the structures.

In thin-section the most common clasts are seen to be composed of:

quartz	20-40%
microcline	30-50%
plagioclase	40-60%

with minor and accessory amounts of biotite, hornblende and opaque minerals. The matrix has the same general composition, though carbonate is commonly present.

Sandy beds from 10 cm. to 1 m. (6 in. to 2 ft.) thick are distributed throughout the conglomerate. These beds increase in number and thickness towards the top of the unit, and are upright as shown by poorly developed cross-bedding. Transportation directions could not be determined.

The polymictic character of the conglomerate indicates a lithologically varied source. The quartz-porphyritic rhyolites indicate older acid igneous rocks which do not occur in the area.

iii. <u>Plagioclase-porphyritic Rhyolite Member</u>: The plagioclase-porphyritic rhyolite member outcrops over an area of approximately 8 sq. kms. (2 sq. mi.) on the southeast slopes of Monkey Hill, and directly overlies the conglomerate member. The most complete outcrop is from Falls Lake due west-northwest, though exposure is poor everywhere.

The rhyolite is a fine- to medium-grained, pale grey to pale brown rock with 2-5 mm. (1/8 to 1/4 in.) diameter plagioclase phenocrysts and aggregates of dark minerals up to

2 mm. by 10 mm. (1/8 to 1/2 in.) in size, the latter imparting a gneissic foliation to the rock. In thin-section the rock is seen to consist of polygonised quartz and feldspar grains in which are set a few aggregates of larger plagioclase grains derived from phenocrysts. The rock is composed of:

quartz	25 - 35%
albite & oligoclase/andesine	45 - 60%
microcline	15 - 25%

and minor amounts of opaque minerals, carbonate, fluorite, chlorite and sphene. Diopside, hornblende and green biotite occur (comprising less than 5%), and garnet was seen as a vein filling with opaque minerals in one sample.

No evidence of chill-margins has survived the later metamorphic recrystallisation, and no apophyses of porphyry in the sediments were seen so that it is not known whether the rock is intrusive into the Aillik Group as a sill or small lopolith, or whether it is extrusive onto it and therefore part of the Group. The conformable nature of the unit and lack of apophyses suggest the latter interpretation is correct and it will, therefore, be considered an extrusive acid igneous rock.

iv. Upper Arkose Member: The upper arkose member forms an incomplete annular outcrop pattern about Round Pond. The arkose has a total strike length of approximately 22 kms. (11 mi.) and a maximum true thickness of approximately 500 m. (1500 ft.).

The member has a sharp, conformable contact with the underlying conglomerate and rhyolite. No interfingering of the rhyolite and the overlying arkose was recognised. The rhyolite is everywhere overlain by the arkose indicating deposition of the arkose continued after the end of rhyolite deposition.

The arkose is a fine-grained metasediment and shows black and grey 5 mm. to 4 cm. (1/4 to 2 in.) thick beds outlined by varying concentrations of mafic minerals. The bedding-plane orientation is easily discerned, though the individual beds grade from one to another. Cross-bedded units up to 30 cms. (1 ft.) thick were seen in a few places, but were not sufficiently well exposed for current directions to be determined. The cross-beds indicate the unit youngs upwards.

The arkose is predominantly composed of:

albite & oligoclase	80%
diopside	10%
and hornblende	7%

with minor amounts of microcline (2%), quartz, biotite, chlorite, sphene and opaque minerals. Bedding is defined by variations in amount of dark minerals, especially diopside and hornblende. The texture is equigranular and polygonal, with no indication of preferred mineral orientation. Towards the top of the unit, on the south side of Monkey Hill where cross-bedding is common and bedding is thinner, the amount of quartz increases to approximately 30% of the total and the unit apparently grades into the locally overlying tuffites and rhyolites of the Makkovik Formation. The high feldspar content of the rock suggests derivation from an acid volcanic terrane (Williams et al., 1958).

v. Metabasalt Member: The metabasalt member is exposed as an annular outcrop around Round Pond. It has a 40 to 50 km. (20 to 25 mi.) circumferential length and a maximum true thickness of approximately 300 m. (900 ft.). The metabasalt is composed of several flows which are partly interbedded with the underlying arkose member. The top of the member defines the top of the Nesbit Harbour Formation.

The clearest exposure of the metabasalt is northwest of Falls Lake where four non-pillowed lava flows occur. Each flow is several meters thick and they are separated from one another by thin, conformable, arkosic layers. The flows may be traced for approximately 300 m. (900 ft.) along strike and similar flows are clearly exposed in the gorge at the Makkovik Brook outlet from Falls Lake. The base of each flow shows poorly developed autobrecciation, but the remainder of the flow appears uniform and unfractured suggesting fluid-flow of the lava. Bands of amygdales have been previously reported (Morris, 1959), but were not seen by the author, though isolated calcite and epidote-filled blebs occurring in a few places may represent metamorphosed amygdales.

Pillow lavas have not been recognised by the author though possible pillow structures have been reported (Morris, 1959; Stoeterau, 1970).

The metabasalt is highly variable in appearance, but is mainly a fine-grained, dark green to black amphibolite which commonly shows a moderately developed tectonic fabric. In areas where schistosity is not apparent, the grain size is much coarser (up to 2 mm. (1/8 in.)) giving the rock a speckled black (amphibole) and white (feldspar) appearance. Small (up to 5 mm. (1/4 in.)) feldspar phenocrysts have been seen, but are rare. Coarse scapolite, epidote or plagioclase porphyroblasts are developed in the more strongly metamorphosed units.

In thin-section the amphibolite is seen to have undergone complete recrystallisation, with the development of complex tectonic fabrics formed by hornblende or biotite mineral alignment, and hornblende, scapolite, epidote or plagioclase porphyroblasts. The rock is fine-grained throughout, except for the porphyroblasts. Scapolite and epidote porphyroblasts often appear coarse-grained in hand-specimen, reaching a size of 1 cm. (1/2 in.) diameter, but in thin-section are seen to consist of aggregates of medium-grained crystals.

The rock is composed of:

plagioclase	20-60%
hornblende	5-40%

biotite	0-30%
scapolite	0-30%
epidote	0-30%
carbonate	0-10%

with accessory amounts of opaque minerals, chlorite, sericite, muscovite, zoizite, sphene and diopside. The scapolite occurs as porphyroblast aggregates on its own, and also as porphyroblastic growths around epidote aggregates.

The high birefringence (up to upper first order to lower second order) of the scapolite indicates a high meionite (calcic scapolite) content (Deer, et al., 1970). The plagioclase is usually albite, though where scapolite and epidote are not developed andesine may also be developed. Where scapolite and epidote are developed they predominate in amount over albite, indicating the original rock had a fairly high Ca : Na ratio. The plagioclase composition was, therefore, probably originally labradoritic or bytownitic, and the lava, therefore, basaltic rather than andesitic in composition.

v. Other Units: A small 2 km. (1 mi.) long, 60 m. (180 ft.) wide, amphibolite unit occurs conformably between the conglomerate and the upper arkose member between Round Pond and Makkovik.

On the eastern flank of Monkey Hill, unusual large, 20 to 50 cms. by 1 to 2 m. (1 to 2 ft. x 3 to 6 ft.), elongate, angular blocks of quartz porphyry are set in a "matrix" of slightly softer, but otherwise identical rock

(Fig. 2-5). The structures are not glacial features as the glacial striae are oriented approximately 30° to their length, but they are oriented parallel to the local S_2 tectonic fabric and also to bedding in the vicinity (the structures were only seen in two dimensions). However, they are only weakly deformed, as shown approximately by a visual inspection of the relative distribution of the phenocrysts (see Appendix I), and are, therefore, probably primary in origin and shape. They are assumed to be a form of syn- or early post- emplacement brecciation or autobrecciation, but their exact nature is unknown.

Makkovik Formation

The name Makkovik Formation is proposed for a sequence of rhyolite lavas, with minor tuffs, rhyolitic conglomerates and arkoses, which occurs from Ford's Bight to south of Wild Bight (the location of the type-section is plotted on Plate 1). The base of the formation is defined as the base of the lower arkose member and the top of the formation as the top of the quartz-porphyritic rhyolite member (Table 2-III and Figs. 2-2 and 2-3).

The Makkovik Formation conformably rests on the Nesbit Harbour Formation and extends for a distance of 20 km. (10 mi.) in a north/south direction and up to 10 km. (5 mi.) in an east/west direction, is 2000 m. to 7000 m. (6000 to 21,000 ft.) thick, and forms a domal. annular structure around the Nesbit Harbour Formation.

Table 2-III

Table showing the approximate thickness of members of the Makkovik Formation.

	m.	ft.
Quartz-porphyritic rhyolite member	2000-4000	6000-12,000
Upper conglomerate member	0-500	0-1500
Tuff member	0-500	0-1500
Lower conglomerate member	0-500	0-1500
Flow-banded rhyolite member	200-1500	600-4000
Trangressive arkose member	0-700	0-2000
Lower arkose member	0-200	0-600
Makkovik Formation	2000-7000	5000-21,000

The Makkovik Formation is much more homogeneous than the Nesbit Harbour Formation, being composed almost entirely of rhyolite lavas with only minor, genetically associated, interfingering sediments and tuffs indicating an acid volcanic environment with local derivation of sediments and subaqueous (presumably shallow water) deposition.

The boundary between the Nesbit Harbour and Makkovik Formations is apparently conformable, as are all boundaries within the Makkovik Formation.

i. Lower Arkose Member: The lower arkose member is only locally developed west of Ford's Bight, and is best exposed on the hill west of Nesbit Harbour. The arkose is lenticular in shape, 4 to 8 kms. (2-4 mi.) long and up to 200 m. (600 ft.) thick. The unit conformably overlies the Nesbit Harbour Formation and may be stratigraphically equivalent to the transgressive arkose member.

The rock is a fine- to medium-grained, highly weathered, friable, grey-brown arkose in which poorly defined 2 to 4 cm. (1 to 2 in.) compositional banding is recognised. This banding is conformable with the upper and lower contacts of the unit and is, therefore, probably bedding.

The rock is composed of:

quartz	30%
feldspar	60%
dark minerals (mainly hornblende)	10%

The quartz content is considerably higher than that of the arkose of the Nesbit Harbour Formation, but still low enough to suggest derivation from an acid volcanic source (Williams, et al., 1958).

ii. Transgressive Arkose Member: The transgressive arkose member occurs conformably above the Nesbit Harbour Formation southeast of Round Pond. The unit wedges out northwards from a position due east of Round Pond, and extends southwards to south of Falls Lake. The total strike length of the member is approximately 12 km. (6 mi.) and the maximum apparent thickness approximately 700 m. (2000 ft.). The arkose is conformably transgressive to the south up into the normally overlying flow-banded rhyolite member and is most clearly exposed due east of Falls Lake.

The transgressive arkose is a light grey, cream or white, fine- to medium-grained, laminated (1/2 to 2 cm. (1/4 to 1 in.)) very slightly micaceous quartzo-feldspathic rock (Fig. 4-10). The lamellae are discontinuous over a distance of approximately 0.5 m. and are defined by variation in colour, due to variation in dark-mineral and opaque-mineral content. Towards the north, the units become finer grained, tectonic disruption of less competent lamellae has taken place, and a few micaceous lenses showing a well developed schistosity occur as interlayers within the beds. No sedimentary structures are preserved.

The arkose is composed of:

albite	45%
quartz	36%
orthoclase	13%
opaque minerals	3%
carbonate	2%

with accessory amounts of green biotite and muscovite. Towards the north, the unit becomes very much more micaceous, with phlogopite as the major dark mineral and diopside occuring as a minor dark mineral. Minor crossite was recognised in one sample. A well developed tectonite fabric is defined by inequant mica, diopside and amphibole grains set in a polygonal textured quartzo-feldspathic groundmass. In the vicinity of pre-tectonic amphibolite dykes diopside, sphene and hornblende are common accessories. A 1 mm. wide post-tectonic vein of sphalerite and pyrite was seen in one sample.

Although no sedimentary structures were found in the unit, the similarity of field appearance and mineralogy to other known sedimentary arkoses in the area suggests a normal sedimentary origin for this arkose. The quartz content of the transgressive arkose is considerably higher than in the arkoses of the Nesbit Harbour Formation, but is still low enough to suggest possible derivation from a volcanic source (Williams, <u>et al</u>., 1958).

iii. Flow-banded Rhyolite Member: The flow-banded rhyolite member occurs as a wide outcrop around Round Pond and conformably overlying the lower arkose member and the Nesbit Harbour Formation. It also generally conformably overlies the transgressive arkose member, but towards the southeast the arkose is transgressive up into the rhyolite. The total outcrop length from southwest to northeast of the dome is approximately 25 kms. (12 mi.), with the maximum apparent width of 4 to 5 kms. (2 to 2 1/2 mi.) occurring at the northern closure of the dome. The maximum true thickness of the member is approximately 1 to 2 kms. (1/2 to 1 mi.). The rhyolite is most fully represented northeast of Round Pond, where both the upper and lower contacts are visible, but is more clearly exposed south of Makkovik, where well developed primary flow-bands and flow-folds may be seen. Several tuffs, conglomerate and porphyritic rhyolite members are partly or completely interbedded and enclosed in the flow-banded rhyolite (Fig. 2-2 and 2-3), and are discussed later under separate headings. Contacts are, for the most part, sharply defined and conformable, but contacts with the associated rhyolitic conglomerates are usually gradational over about 30 cms. (1 ft.).

The flow-banded rhyolite is a white to pale-pink, grey or yellow, very fine-grained rock with alternating 1 to 2 mm. (1/20 to 1/10 in.) banding defined by slight colour differences and resistance to erosion. These bands are discontinuous, have an average length of 5 to 10 cms. (2 1/2 to 5 in.) and are poorly developed or non-existent in places. Primary folding is uncommon, and where developed is usually modified by later tectonic flattening.

The rhyolite is composed of:

quartz	25 - 35%
microcline	30 - 50%
albite	15 - 50%

with minor opaque minerals, and chlorite. The banding is defined by medium-grained microcline-quartz bands in fine-grained, polygonal-textured quartz, microcline and albite, in which are set a few recrystallised, euhedral, quartz phenocrysts composed of strained sub-grains with sutured boundaries. The origin of the banding is discussed more fully under a separate heading (p. 107). In the vicinity of the Monkey Hill Granite, diopside, hornblende and sphene occur, and may form up to 10% of the rock, and garnet is locally developed.

Autobrecciation has been recognised in the rhyolite west of Wild Bight. The autobrecciation consists of 5 to 10 cm. (2 1/2 to 5 in.) diameter fragments of banded rhyolite set in an unbanded rhyolite matrix. The fragments are angular and rotated relative to one another.

iv. <u>Lower Conglomerate Member</u>: The lower conglomerate member occurs north of Monkey Hill. The conglomerate has a strike length of 8 km. (4 mi.) and thickness of approximately 1/2 km. (1/4 mi.). The conglomerate occurs as a conformable lens within the flow-banded rhyolite member, to which its clasts and matrix are apparently directly related. No interbedded sediments or volcanic rocks have been recognised. The clasts are tectonically flattened and usually 1 cm. by 2 cm. to 8 cm. by 16 cm. (1/2 x 1 in. to 4 x 8 in.) in size. They show rounded outlines which are presumably due to sedimentary processes, but may partly be due to tectonic flattening and formation of augen-structures.

In thin-section the conglomerate shows well-developed polygonal fabrics of quartz and fèldspar aggregates or, in the case of small pebbles, of pure quartz or pure microcline aggregates, set in a finer grained quartz-feldspar matrix in which opaque menerals, calcite, hornblende and diopside may occur as accessories.

v. Tuff Member: The tuff member is best exposed on Big Island, but occurs from Big Island through Ranger Bight to Makkovik, and also south over a distance of approximately 15 kms. (8 mi.). The unit varies in thickness from approximately 100 m. (300 ft.) on Big Island to 500 m. (1500 ft.) at Makkovik. The apparent thickness is highly variable due to the effects of repetition by folding about the syncline through Ranger Bight. The tuff is conformably underlain by and partly interbedded with the flow banded rhyolite to the east, but in the west is in tectonic contact with the amphibolites of a major slide zone, and partly intruded by amphibolite dykes occurring in the slide zone.

The tuff is very variable in appearance and composition, but on Big Island is a finely laminated, pale grey to cream coloured, fine-grained rock in which 5 mm. (1/4 in.) long pink, dark grey and black flattened feldspathic and opaque mineral clots, and a few euhedral quartz phenoclasts occur. Within this unit are a few 5 to 30 cm. (2 1/2 in. to 1 ft.) thick pebble-conglomerate horizons with small (up to 3 cm. (1 1/2 in.) diameter) medium grey quartzo-feldspathic pebbles (Fig. 2-6 and 2-7), and 1 to 2 cms. (1/2 to 1 in.) diameter dark (commonly greenish) coloured quartzo-feldspathic pebbles containing epidote. South and east of Ranger Bight a fine-grained, dark and light grey, banded (on a 1 to 5 cm. (1/2 to 2 1/2 in.) scale) member occurs interbedded and overlying the other laminated and pebbly types (Fig. 4-62). This dark and light grey, banded type gives way higher up in the sequence to a pale grey and pink banded type. Feldspar phenoclasts are abundant in places and give the rock a pink speckled appearance in hand-specimen.

Nowhere in the unit have any sedimentary structures such as cross-bedding, ripplemarks or graded bedding been recognised. Primary folding has been recognised south and southeast of Ranger Bight (Fig. 4-60, 4-61 and 4-62; Chapter 4).

Fig. 2-6. Lamellae and pebble-beds in tuff member of Makkovik Formation on Big Island.

Fig. 2-7. Lamellae and magnetite clots (black) in tuff member of Makkovik Formation on Big Island.

The laminated tuff on Big Island is composed of:

quartz	20%
albite	60%
microcline	10%

with accessory magnetite, biotite, diopside, zircon, fluorite, garnet, tremolite, sphene, chlorite, apatite and, in one sample, crossite. The biotite defines a tectonite fabric, as do the flattened feldspar and quartz aggregates. These flattened aggregates are of three types, microcline aggregates derived from K-feldspar phenoclasts, plagioclase aggregates from plagioclase phenoclasts (which comprise the majority of phenoclasts), and quartz-microcline-plagioclase-biotite aggregates presumably derived from recrystallisation of lithic or pumice fragments. Quartz aggregates from quartz phenoclasts do not generally show much flattening, and the outline of the original crystal is commonly clearly visible. All the aggregates and the grounmass are composed of unstrained equant grains showing polygonal texture, with the groundmass grains being slightly smaller than the aggregate grains. This variation in grain-size commonly gives the appearance of fine- and medium-grained bands in the rock. Where large microcline grains are common, they are usually perthitic to mesoperthitic. Southeast of Ranger Bight the tuff contains up to 10% calcite. Many of the large plagioclase grains (determined by staining) show a "chequer-board" type of albite twinning. The grains are biaxial negative with a refractive index less than quartz, and must, therefore, be oligoclase, as plutonic ("low-temperature") albite (Stemmons, 1962) is biaxial positive and volcanic ("high-temperature") albite could not exist in these recrystallised amphibolite facies rocks.

vi. <u>Upper Conglomerate Member:</u> The upper conglomerate member occurs southwest of Makkovik. The member is lens-shaped and outcrops over a strike distance of approximately 4 kms. (2 mi.) amd a width of 1/2 km. (1/4 mi.). The conglomerate is situated within the tuff member with which it appears to have gradational reworked contacts. However, where the southern contact of the unit outcrops on the coast it is seen to be in contact with the flow-banded rhyolite. The conglomerate is similar in field and thin-section appearance to the rhyolitic lower conglomerate except that it also contains a few amphibolite boulders, and the clast size ranges up to 30 cms. (1 ft.) diameter.

vii. <u>Quartz-porphyritic Rhyolite Member</u>: The quartz-porphyritic rhyolite member occurs as a 4 to 6 km. (2 to 3 mi.) wide band extending for 34 kms. (12 mi.) northeasterly through the eastern part of the area from east of Falls Lake to Wild Bight and also north and northeast of Makkovik. The rhyolite is most clearly exposed between Ford's Bight and Wild Bight. No folding has been recognised within the member, except for west and south of Big Bight and at Ranger Bight, and the thickness is therefore estimated to be of the order of 2 to 4 kms. (1 to 2 mi.). The rhyolite is conformably underlain by the flow-banded rhyolite member, except in the extreme south, at Falls Lake where it is underlain by the transgressive arkose member, and at Ranger Bight where it is underlain by conglomerates and tuffs.

The quartz-porphyritic rhyolite is a massive (non-banded), pale grey to pale yellow, fine-grained porphyritic rock, with quartz, microcline and plagioclase phenocrysts up to 2 mm. in diameter evenly distributed throughout the rock and comprising approximately 20% to 40% of the whole (Fig. 2-8). The phenocrysts are commonly euhedral to subhedral, but later tectonic flattening has deformed them and caused annealing recrystallisation to an aggregate of subgrains, and they may be lensoid in shape as a result (Fig. III-3, III-4, III-5 and III-8; Clark, 1971, Plate 1). Where the feldspar phenocrysts are smaller that the quartz phenocrysts, they are usually aggregated with both albite and microcline occurring in a single aggregate. The microcline is perthitic indicating inversion from a higher temperature form of K-feldspar.

The porphyry is composed of:

quartz	20% - 40%
microcline	30% - 60%
albite and oligoclase/ andesine	20% - 40%

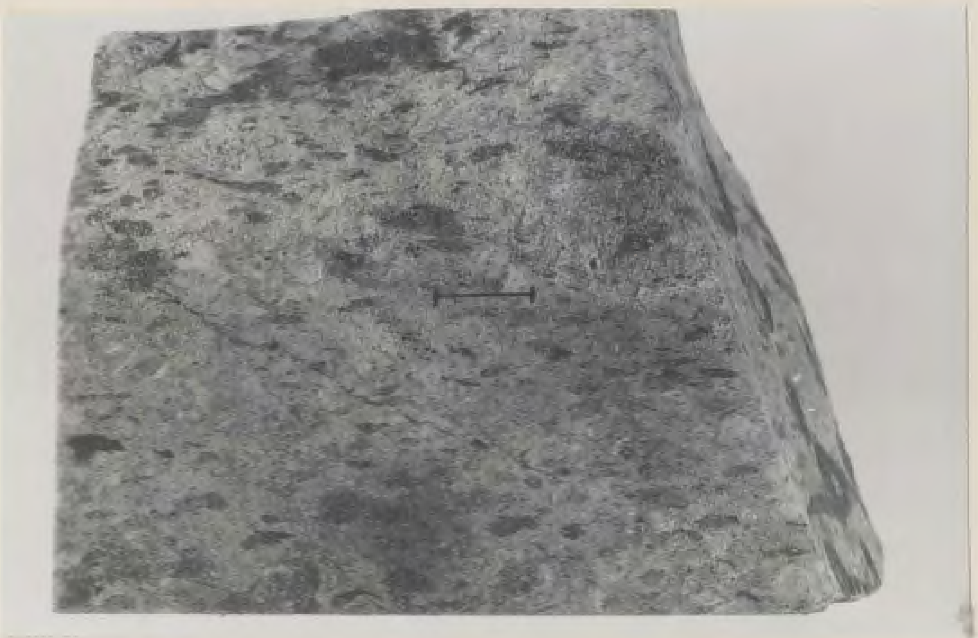


Fig. 2-8. Quartz-porphyritic rhyolite. Quartz phenocrysts are dark, feldspar phenocrysts are lighter with hazy outlines.

Fig. 2-9. Quartz-porphyry xenoliths in quartz-porphyritic rhyolite. East coast of Wild Bight.

and accessory amounts of sphene, carbonate, hornblende, garnet, pyrite, magnetite, haematite (after pyrite and magnetite) and very sparse chalcopyrite. Variations within the formation are very minor and mainly caused by variation in percentage of phenocrysts (20% to 40%), ratio of quartz to feldspar phenocrysts (from 100% to 20% quartz), and size of phenocrysts (from 1 mm. diameter up to 3 mm., with the feldspar phenocrysts usually being slightly larger than the quartz). However, in the vicinity of Poodle Pond, a few 4 to 30 cms. (2 in. to 1 ft.) thick layers containing up to 10% carbonate occur oriented parallel to local tuffite horizons. Apart from the higher carbonate content, these layers are identical to the rest of the rhyolite. At the south end of the estuary at the head of Wild Bight, the rhyolite is purple in colour and contains up to 40% pink microcline phenocrysts and 40% grey quartz phenocrysts. It is noticeably different from the surrounding grey, less porphyritic rhyolite, though it appears to grade into it. On the east coast of Ranger Bight and at the east end of Big Island the rhyolite is medium-grained, equigranular, with no evidence of phenocrysts, though elsewhere in the vicinity it is a normal porphyry.

In the few places on the east coast of Wild Bight angular platy xenoliths up to 15 cms. by 30 cms. (6 in. by 1 ft.) in size occur in the porphyry. These xenoliths are almost identical in composition to the host porphyry, except for a slightly paler matrix (Fig. 2-9). The xenoliths show a preferred orientation that is different to the tectonite fabric orientation and is, therefore, thought to be a primary (igneous) fabric, but it is not parallel to the assumed S_o ("bedding") orientation in the area.

Within the rhyolite conformable tuff beds up to 50 m. (150 ft.) thick, but more usually 10 cm. to 2 m. (4 in. to 6 ft.) thick occur (Fig. 2-10). Tuffs northeast of Poodle Pond occur as two long thin 50 m. by 3 km. (150 ft. by 1 1/2 mi.) subvertically dipping beds. The beds are laminated on a 1 mm. to 2 cm. (1/20 in. to 1 in.) scale. The bedding is shown to be due to a variation in grain-size (0.1 mm. to 1 mm.) and colour (dark grey; pale green; yellow) reflecting a variation in mineral content (opaque minerals; epidote and diopside; quartz and feldspar respectively).

The tuffs are composed of:

microcline	40%
albite	20%
quartz	20%
carbonate	5% - 10%
opaque minerals	5%

and accessory diopside, phlogopite, actinolite and garnet. The finer-grained beds are composed of quartz, albite, opaque minerals and minor microcline, and show microscopic graded-bedding (Fig. 2-11). The coarser-grained beds contain recrystallised phenocrysts of quartz set in a matrix

Fig. 2-10. Tuff horizon in quartz-porphyritic rhyolite. West coast of Wild Bight.

Fig. 2-11. Photomicrograph of microscopic graded-bedding in tuff bed in quartz-porphyritic rhyolite. Between Ford's Bight and Wild Bight. Plane polarised light, x 10.

of microcline, albite, quartz, carbonate and the accessory minerals. Porphyroblasts of albite have grown in the vicinity of the small gabbro body southwest of Wild Bight.

The extremely fine-grained nature of the deposit on deposition is indicated by the very fine-scale, delicate cross-bedding. The initial fine-grained nature of the rock, its quartzo-feldspathic composition, the phenoclasts of quartz and the association of the unit with acid volcanic rocks indicate the lithology is probably a tuff. The fairly high carbonate content may indicate contemporaneous carbonate deposition.

Biotite-diopside schist beds also occur in the rhyolite and are 50 m. to 100 m. (150 to 300 ft.) thick and up to 1 km. (1/2 m.) long.

They are composed of:

plagioclase	40 - 50%
biotite	20 - 30%
diopside	5 - 20%
hornblende	0 - 10%
carbonate	0 - 10%

and accessory opaque minerals. These schists have the appearance of a recrystallised biotite-calcite lamprophyre (?), but as they are conformable with bedding in the associated tuffs, are interbedded with the tuffs, and also have flattened fragments of quartz porphyritic rhyolite, they are assumed to be mafic or pelitic metasediments. They are probably the former as they are not rich in alumina-bearing minerals and as pelitic rocks have not been recognised elsewhere in the thesis area.

viii. <u>Microcline-porphyritic Rhyolite Member</u>: The microcline-porphyritic rhyolite occurs from east of Falls Lake, northwards for a strike distance of approximately 10 to 13 km. (5 to 6 mi.). The member appears to dip steeply eastwards, wedges out northwards and has a maximum true thickness of approximately 1 km. (1/2 mi.). The member occurs within the flow-banded rhyolite, and transgresses up it towards the south, where it occurs in the overlying quartz-porphyritic rhyolite. The margins of the microcline-porphyritic rhyolite could not be accurately discerned and it appeared to grade into the enclosing quartz-porphyritic rhyolite.

The rhyolite is a massive, unbanded, grey, fine-grained rock with 2 mm. to 1 cm. (1/8 to 1/2 in.) diameter, pink microcline phenocrysts, and quartz phenocrysts up to 2 mm. (1/8 in.) in diameter. The microcline phenocrysts, predominate over the quartz phenocrysts. In thin-section the quartz phenocrysts are seen to form elongate slivers composed of an aggregate of small quartz grains, resulting from tectonic flattening and recrystallisation, whereas the microcline phenocrysts are only slightly deformed.

The rock is composed of:

quartz	30 - 45%
microcline	20 - 30%
oligoclase/ andesine	25 - 45%

with accessory amounts of opaque minerals and biotite. The groundmass has fine-grained polygonal texture, the microcline phenocrysts are now composed of an aggregate of polygonal grains, and the quartz phenocrysts usually occur as aggregates of strained grains with sutured boundaries, though polygonal aggregates occur in some specimens.

An almost identical, though less deformed and less quartz-rich, microcline porphyritic rhyolite occurs west of Wild Bight and may be a continuation of this member.

ix. Other Units: Several minor conglomerate, tuff, arkose and rhyolite lenses occur within the Makkovik Formation, and are lithologically similar to members already described.

Pomiadluk Point Formation

The name Pomiadluk Point Formation is proposed for a sequence of interfingering acid and basic lavas, tuffs, conglomerates and arkoses which occur on the Pomiadluk Point peninsula east of Wild Bight (the location of the type-section is plotted on Plate 1). The base of the formation is defined as the base of the lower rhyolite member (Table 2-IV and Fig. 2-12). The top of the formation is not exposed.

The Pomiadluk Point Formation occurs over a strike length of 10 kms. (5 mi.) and has a true thickness greater than 2000 m. (6000 ft.) The lithologies comprising the

Table 2-IV

Table showing approximate thickness of members of the Pomiadluk Point Formation.

	m.	ft.
Upper rhyolite member	1000+	3000+
Upper conglomerate member	0-500	0-1500
Upper tuff member	100-200	300-600
Lower conglomerate member	0-1000	0-3000
Lower tuff member	0-200	0-600
Middle rhyolite member	0-900	0-2700
Lower rhyolite member	0-1000	0-3000
Rhyolite and tuff unit to south	100-200	300-600
Pomiadluk Point Formation	2000	6000

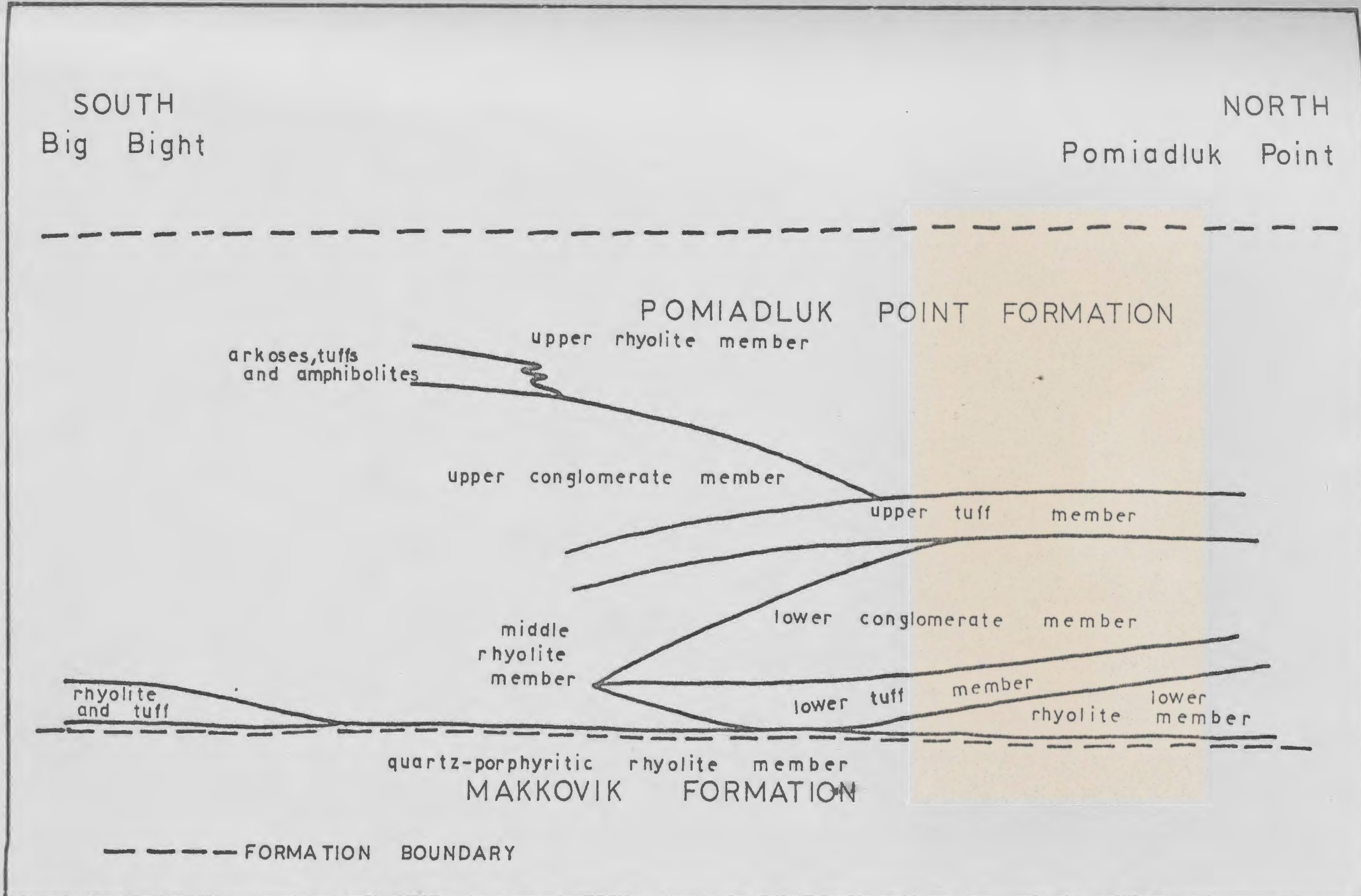


Fig. 2-12. Stratigraphic Diagram of the Pomiadluk Point Formation from Pomiadluk Point to Big Bight.

formation indicate a similar acid volcanic environment of deposition to that of the Makkovik Formation, with mainly local derivation of rock-types. However, the occurrence of two major conglomerate sequences indicate greater sedimentary control, such as during deposition of the Nesbit Harbour Formation. Contacts between this formation and the underlying Makkovik Formation are conformable, as are all contacts within the Pomiadluk Point Formation.

A rhyolite-tuff unit west of Big Bight is tentatively included in the Pomiadluk Point Formation for convenience.

i. Lower Rhyolite Member: The lower rhyolite member occurs along the east coast of Wild Bight, and is most clearly exposed along the coast. The member dips steeply eastward and is approximately 5 kms. (2 1/2 mi.) long and 1 km. (1/2 mi.) thick at its maximum exposure. However, as the rhyolite is wedge-shaped and partially submerged the true maximum thickness could not be determined. The member conformably overlies the Makkovik Formation. The contact with the underlying quartz-porphyritic rhyolite member is straight on all scales from a few millimeters to 4 kms. (2 mi.) (Fig. 2-13 and 2-14; Plate 1), and shows no chill margin or other diagnostic feature such as fossil erosion surface or mylonite.

The rhyolite shows well developed banding and primary (flow) folds identical in appearance to that in the flow-banded rhyolite of the Makkovik Formation (Fig. 2-15).

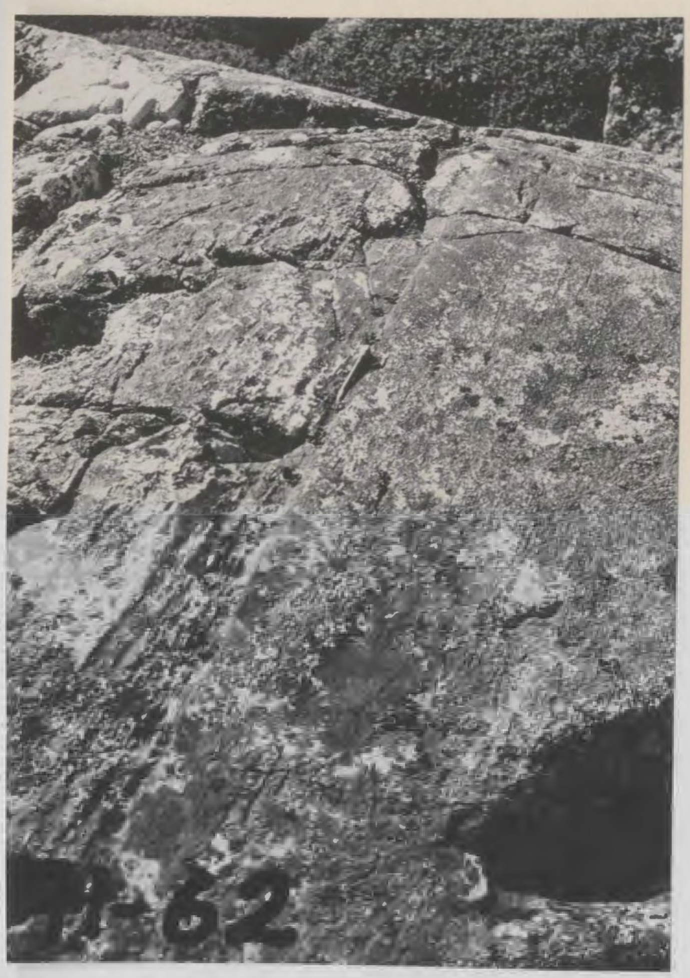


Fig. 2-13. Contact (marked by pencil) between the quartz-porphyritic rhyolite member (Makkovik Formation, right) and the overlying lower (flow-banded) rhyolite member (Pomiadluk Point Formation, left). East of Wild Bight.

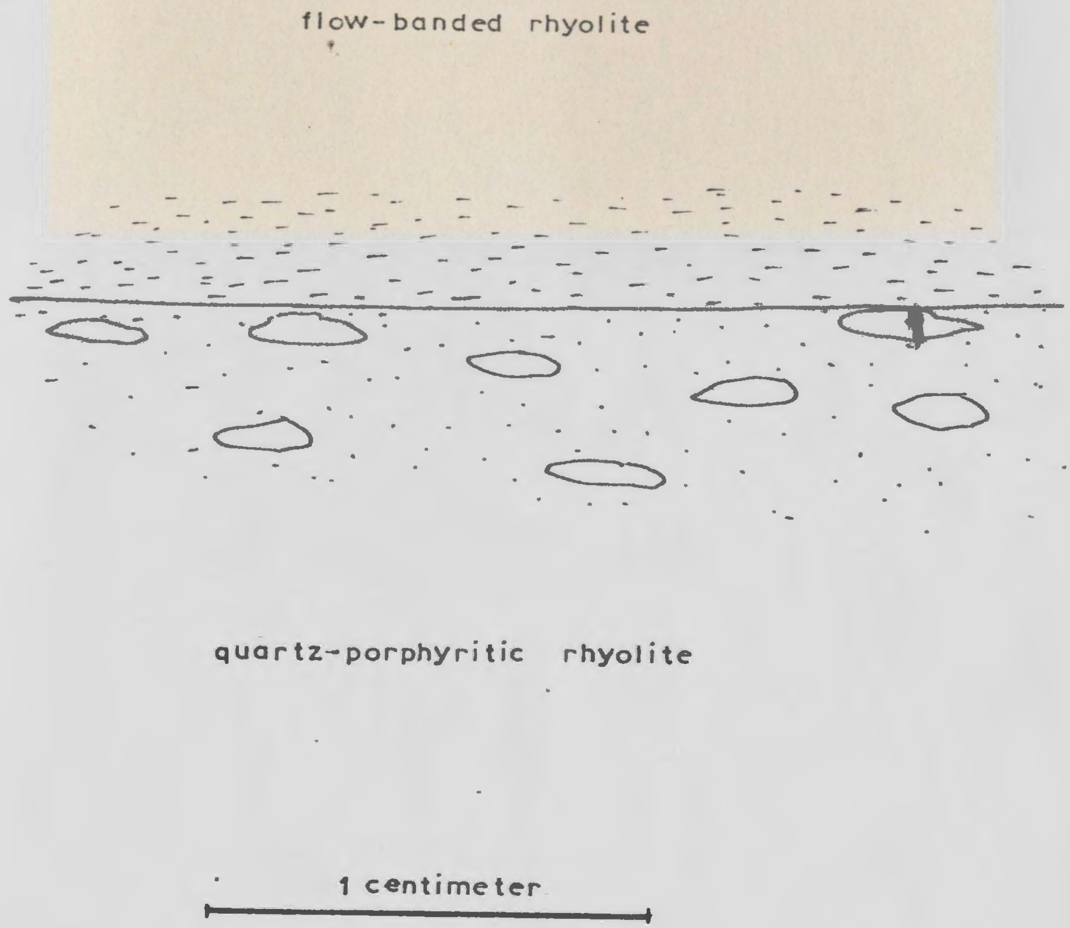


Fig. 2-14. Sketch from part of a thin-section of the contact between the quartz-porphyritic rhyolite member (Makkovik Formation) and the lower (flow-banded) rhyolite member (Pomiadluk Point Formation), showing the straight nature of the contact.

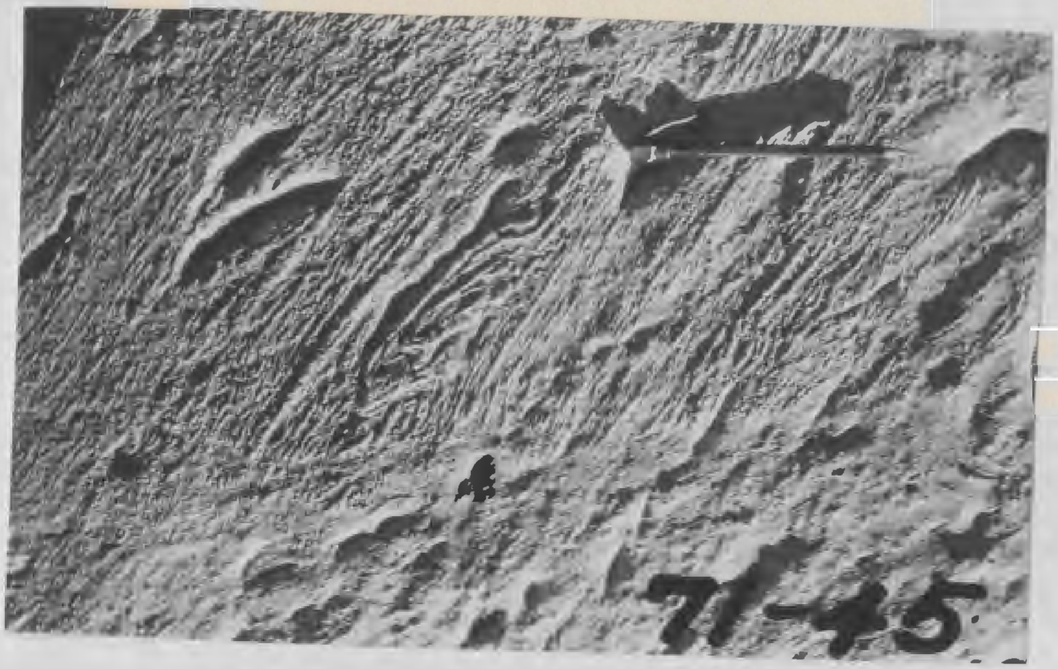


Fig. 2-15. Flow-banding and primary folds in lower rhyolite member (Pomiadluk Point Formation), at Pomiadluk Point.

Fig. 2-16. Autobreccia in lower rhyolite member, (Pomiadluk Point Formation). Northeast coast of Wild Bight.

This banding is seen throughout the member except where autobrecciation occurs. Autobrecciation has been recognised from two localities on the east coast of Wild Bight. The autobreccia fragments are 5 to 10 cms. (2 1/2 to 5 in.) in diameter and consist of banded and unbanded rhyolite set in a rhyolite matrix (Fig. 2-16). The fragments are usually angular and only slightly rotated, but where rotation is more pronounced they show a subangular to subrounded outline.

The rhyolite is almost identical in composition to the flow-banded rhyolite member of the Makkovik Formation, but also contains minor andalusite. The banding is composed of quartz-rich 1/2 mm. (1/40 in.) wide layers showing very fine-grained polygonal texture, surrounded by an elongate "envelope" of slightly coarser grained, polygonal textured microcline (Fig. 2-17). These bands are separated from one another by very fine-grained, polygonal textured albite. Accessory amounts of (metamorphic?) andalusite, opaque minerals (haematite?), green hornblende, sphene, fluorite, tourmaline and zircon generally occur in the microcline-rich parts of the rock.

ii. Middle Rhyolite Member: The middle rhyolite member occurs north of Big Bight. The member is interbedded with a major conglomerate and tuff unit, and is stratigraphically equivalent, at its base, to the lower rhyolite member, as it also conformably overlies the Makkovik Formation, though it extends to a higher stratigraphic level than the lower rhyolite. The member is steeply dipping eastward, has a

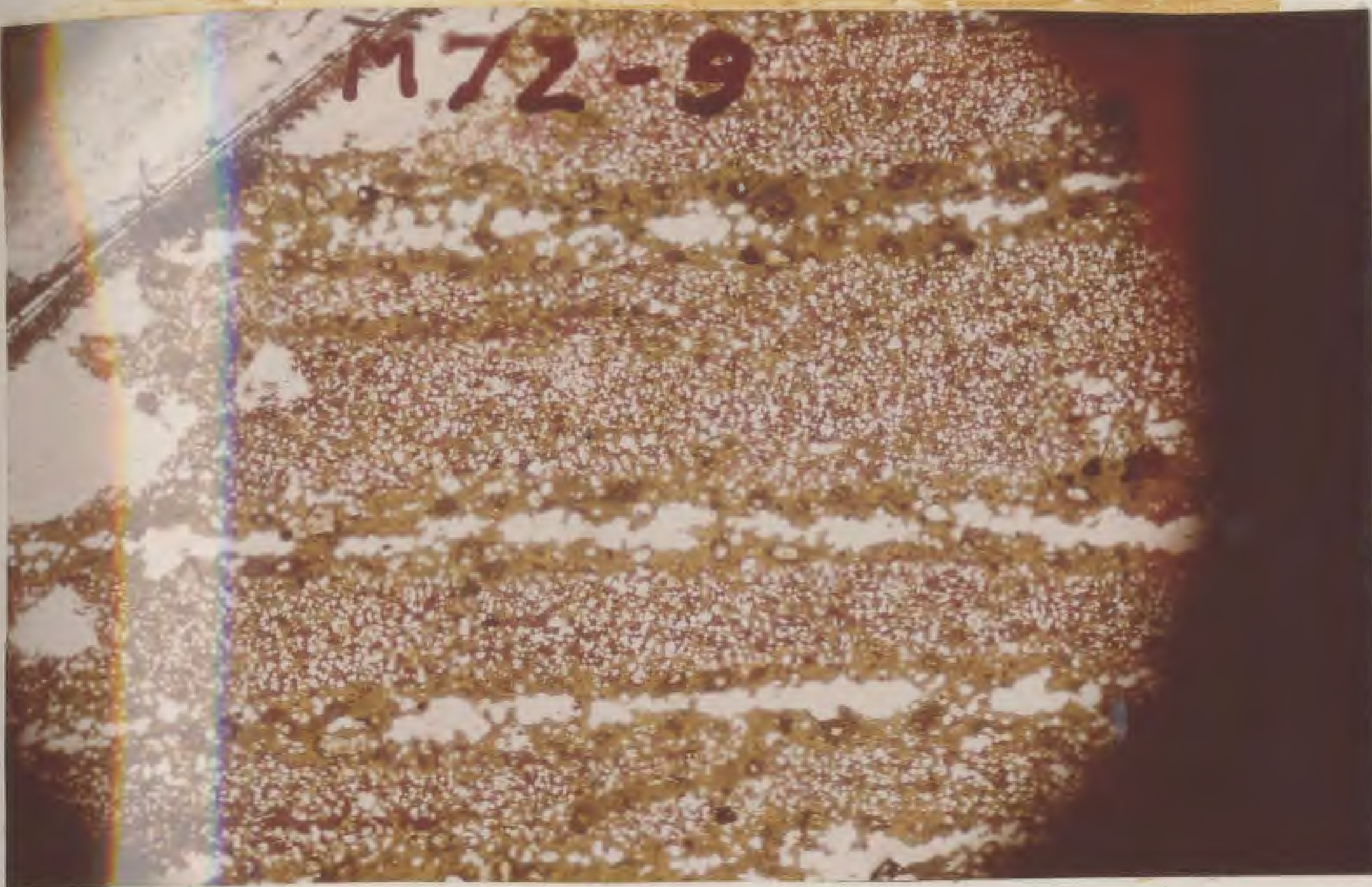


Fig. 2-17. Mineralogical banding in the lower (flow-banded) rhyolite member of the Pomiadluk Point Formation. White-quartz; red-plagioclase; yellow-K-feldspar. Plane polarised light, X20.

strike length of approximately 5 kms. (2 1/2 mi.) and a maximum thickness of approximately 900 m. (2700 ft.) at its southern end, but wedges out northwards.

The rhyolite is identical in both field and thin-section appearance to the lower rhyolite member, except for the lack of andalusite and the development of up to 10% muscovite and 5% biotite in places. Flow-banding is moderately to well developed in the north, but is generally absent to the south.

iii. Lower Tuff Member: The lower tuff member strikes south-southwest from Pomiadluk Point, has a strike-length of 8 kms. (4 mi.), a maximum thickness of 200 m. (600 ft.), and dips subvertically. It conformably overlies the lower rhyolite member in the north and is interbedded with the middle rhyolite member in the south.

The tuff varies from a grey and white, very fine grained, finely laminated (0.1 mm. to 1 mm. (0.005 to 0.05 in.) scale) rock with easily recognisable crystal and lithic fragments, the latter increasing in size towards the top of the member where it grades into the overlying lower conglomerate member over a width of approximately 30 to 60 cm. (1 to 2 ft.). Very fine scale cross-bedding may be seen in the finer grained beds, but is difficult to recognise due to the small angles subtended by the lamellae as a result of tectonic flattening sub-parallel to bedding.

In thin-section the finest grained members of the unit consist of 80% quartz with polygonal texture, 10% phlogopite and 10% muscovite forming a well developed tectonite fabric, and accessory amounts of apatite, blue-grey tourmaline, epidote, opaque minerals and interstitial K-feldspar. The lamellae are defined by a very slight variation in grain size and mica content. A few porphyroclasts of quartz and K-feldspar occur in some lamellae.

In contrast where porphyroclasts form a large proportion of the rock plagioclase is abundant, and the lamellae are defined by a variation in K-feldspar/plagioclase ratios, giving discontinuous K-feldspar and plagioclase lamellae in which are set euhedral to subhedral porphyroclasts of quartz and plagioclase (now mainly altered to epidote) (Fig. 2-18). Epidote bands are common and are commonly associated with high opaque-mineral content. Phlogopite and muscovite occur as accessories, but may form as much (combined) as 10% of the mineral species.

iv. Lower Conglomerate Member: The lower conglomerate member outcrops from Pomiadluk Point southwards for a distance of 8 kms. (4 mi.) and an outcrop width of 1 km. (1/2 mi.). At Pomiadluk Point the clast size varies considerably from approximately 2 mm. (1/8 in.) pebbles to 80 cm. by 15 cm. (2 ft. by 6 ft.) boulders, though the larger boulders are usually broken down by boudinage and the smaller flattened

Fig. 2-18. Banding and phenoclasts in the lower tuffite member of the Pomiadluk Point Formation. White-quartz; red-plagioclase; yellow-K-feldspar. Large quartz phenoclast on left and epidote (after plagioclase?) on the right. Plane polarized light, X10.

(Figs. 2-19, 4-14 and 4-36). The clasts are all subrounded though sizes vary regularly across strike, with medium to large sized clasts making up the majority in the west (the base of the member), and large clasts in the center grading down to small clasts in the east (the top of the member). Bedding is generally poorly developed, but graded beds, which young eastwards, occur in the central area and in the east (Figs. 2-20, 2-21 and 2-22), and a few beds of white (quartz-feldspar porphyritic) rhyolite tuffs and banded arkose up to 3 m. (10 ft.) wide occur in the west. Both upper and lower contacts of the conglomerate with the overlying and underlying tuffs are gradational, and show a gradual change of clast size over a distance of about 30 to 60 cms. (1 to 2 ft.). Similarly, upper and lower contacts of the various graded and ungraded beds within the conglomerates are also commonly gradational, (Fig. 2-20), suggesting a smooth transition from high to low energy environments, and vice versa. However, in the center of the conglomerates, graded beds show sharply defined transitions at their upper contacts suggesting a hiatus in deposition, with possible minor reworking or erosion (Fig. 2-22).

Clast types in the conglomerate vary considerably from west to east, with 80% of the clasts in the west consisting of dark orange feldspar porphyritic rhyolite, and the remainder consisting of grey 'pelite', coarse-grained granite with specular haematite, light pink feldspar porphyritic rhyolite and

Fig. 2-19. Lower conglomerate of the Pomiadluk Point Formation, showing flattening of boulders and consequent development of boudins.

Fig. 2-20. Graded bed with gradational upper and lower boundaries. Lower conglomerate member of the Pomiadluk Point Formation. Note tectonic flattening of the pebbles.

Fig. 2-21. Graded bed in lower conglomerate of Pomiadluk Point Formation. Bed dips down to right, plane of pebbles dips subvertically (parallel to S_2), indicating a tectonic origin for the pebble shape.

Fig. 2-22. Graded bed with sharply defined boundaries. Lower conglomerate member of Pomiadluk Point Formation.

light grey rhyolite. Towards the center of the area the coarse-grained granite boulders increase in proportion to approximately 50%, and then decrease again towards the east, giving way to grey pelite (?) clasts which form up to 70% of the clasts in the east, and a few green epidotitic clasts. The matrix also varies from a light grey, fine- to medium-grained quartzo-feldspathic composition in the west, to a medium grey, fine-grained pelite with more mafic minerals in the east.

In thin-section the feldspar porphyritic rhyolite is seen to contain microcline phenocrysts and to be identical in mineralogy to the microcline-porphyritic rhyolites, and the light grey rhyolite to the unbanded parts of the flow-banded rhyolites. The granite has undergone strain-recrystallisation to give sutured boundaries to sub-grains in the original large quartz and feldspar grains, but is otherwise a normal biotite-muscovite granite in which the biotite is partly altered to chlorite. No source of the granite pebbles has been recognised in the area. The 'pelite' pebbles consist of fine-grained quartz, feldspar, biotite and opaque minerals, which latter give them their grey pelitic look. The green-coloured clasts consist of epidote, calcite, chlorite and biotite, with minor quartz and feldspar. The matrix consists of fine- to medium-grained quartz and feldspar with quartz and feldspar phenoclasts and minor amounts of biotite in the west, and increasing opaque minerals and biotite content with decreasing quartz and feldspar phenoclasts towards the east (top) of the member.

v. Upper Tuff Member: The upper tuff member occurs on the east coast of Pomiadluk Point southwards for a distance of 12 kms. (6 mi.). The member dips subvertically, has a maximum thickness of approximately 200 m. (600 ft.). The tuff is in gradational contact with the underlying lower conglomerate member, but is in sharp, conformable contact with the underlying middle rhyolite member. The tuff is identical in appearance and mineralogy to the lower tuff member, except for the occurrence, north of Big Bight, of ellipsoidal quartz and feldspar bodies up to 20 cms. by 10 cms. by 5 cms. (10 in. by 5 in. by 2 1/2 in.) in size (Fig. 2-23). These bodies are of two origins, those in the upper portion of the tuff are deformed granite clasts which increase in number towards the overlying conglomerate, whereas the remainder are boudins developed from pegmatite or aplite veins. The latter commonly may be seen to form rows of boudins, though each boudin is usually separated from the others by up to 1/2 m. (1 1/2 ft.). The veins were originally pure quartz or quartz-feldspar which have undergone recrystallisation and now have coarse-grained quartz rims around medium-grained quartz-feldspar cores (Fig. 2-24).

vi. Upper Conglomerate Member: The upper conglomerate

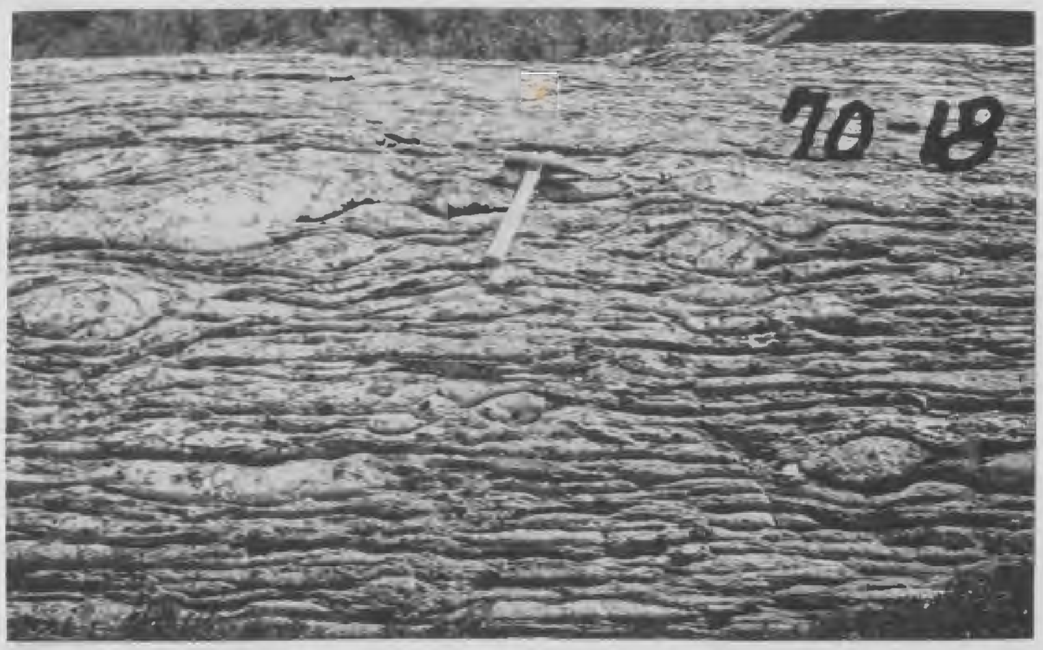


Fig. 2-23. Flattened granite clasts and boudins of pegmatite or aplite veins. Upper tuff member of Pomiadluk Point Formation.

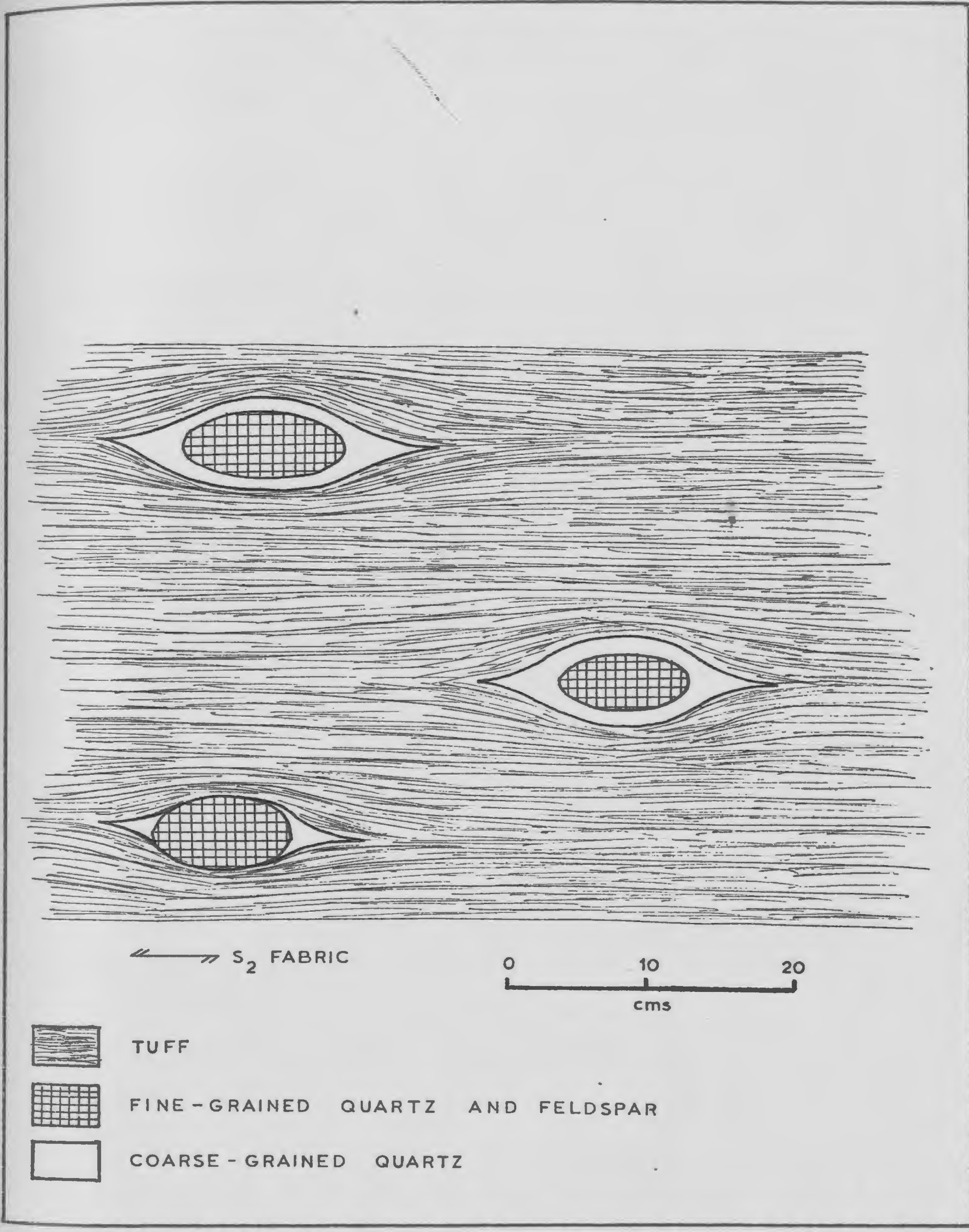


Fig. 2-24. Schematic diagram of quartzo-feldspathic boudins in upper tuff member of the Pomiadluk Point Formation. North of Big Bight.

member occurs north of Big Bight to the west of October Harbour. The unit has a strike length of approximately 8 km. (4 mi.), and an outcrop width of 1/2 km. (1/4 mi.) which is increased up to 2 km. (1 mi.) by folding. The member is in gradational contact with the underlying upper tuffite member. The most complete section occurs southwest of October Harbour, but the lithology is most clearly exposed north of Big Bight.

The conglomerate is very similar to the lower conglomerate member in clast composition, though the clast sizes do not vary as much as at Pomiadluk Point and are usually 4 cm. by 8 cm. to 8 cm. by 16 cm. (2 by 4 in. to 4 by 8 in.) in size and subrounded, though tectonically flattened, in shape. The matrix is composed of fine-grained quartz and feldspar with minor biotite, and phenoclasts of quartz and feldspar. The unit also contains some minor sandy and tuffaceous beds, but no graded beds were recognised.

vii. Upper Rhyolite Member: The upper rhyolite member extends for 5 km. (2 1/2 mi.) east of Pomiadluk Point to October Harbour. The rhyolite is steeply dipping eastward and has a maximum true thickness of approximately 1 km. (1/2 mi.), west of October Harbour, though the top of the member has not been seen. The member is conformably underlain by the upper conglomerate member in the south, and by the upper tuff member in the north. It contains an interbed of amphibolite (mafic lava?) at October Harbour, and several minor quartz-porphyritic rhyolite and tuff lenses. The rhyolite is intruded by the October Harbour Granite along its eastern margin, and as a result no stratigraphically overlying lithology has been recognised. The rhyolite is a normal flow-banded rhyolite similar to that of the Makkovik Formation.

viii. <u>Rhyolite-tuff Unit West of Big Bight</u>: The rhyolite-tuff unit west of Big Bight occurs as a 2 km. (1 mi.) long, 100 to 200 m. (300 to 600 ft.) wide outcrop. A small rhyolite outcrop immediately east of this on the coast of Big Bight is considered to be part of the same unit. The rhyolite conformably overlies the Makkovik Formation, but its upper stratigraphic contact is not seen as it is intruded to the east and south by the Adlavik Igneous Complex and by the Monkey Hill Granite. The rhyolite is a fine-grained, imperfectly banded, quartz-porphyritic rhyolite, and contains tuffaceous beds in the southwest, in which slump structures and a few "pseudoconglomerate" pebbles resulting from syndepositional slumping, occur (Pettijohn, 1957). The pebbles contain up to 20% aegerine-augite, and are generally coarser-grained than the surrounding rhyolitic matrix.

The situation of the rhyolite, conformable above the Makkovik Formation, and its slight similarity to the middle rhyolite member of the Pomiadluk Point Formation suggests it may be a southward extension of the middle rhyolite, and it is therefore, tentatively included in the Pomiadluk Point Formation.

ix. Other Units: Various minor amphibolites occur at the base of the upper rhyolite member and also associated with arkose, tuff and porphyritic metabasalt units near Big Bight.

Manak Bay Formation

The name Manak Bay Formation is proposed for a conformable sequence of arkoses which occurs at Manak Bay (the location of the type-section is plotted on Plate 1). A sequence of arkoses, mafic lava and rhyolite lava at the head of Big Bight is separated from the Manak Bay sequence by an intrusion of the Adlavik Igneous Complex, but is correlated with the Manak Bay sequence, based on the similarity of the arkoses (Table 2-V and Fig. 2-25). The formation is disconformably transgressive onto the underlying Makkovik Formation. The base of the lowest member of the formation has not been recognised, and neither has the top of the highest member.

The formation has a maximum strike length of approximately 4 kms. (2 mi.) and a maximum true thickness of approximately 2000 m. (600 ft.), though repetition by folding increases the apparent thickness considerably.

The Manak Bay Formation is a shallow-water deposit

Table 2-V

Table showing approximate thickness of members of the Manak Bay Formation

	m.	ft.
Flow-banded rhyolite	200+	600+
Metabasalt member	200	600
Arkose member	2000	6000
Subgreywacke	10+	30+
Manak Bay Formation	2000	6000

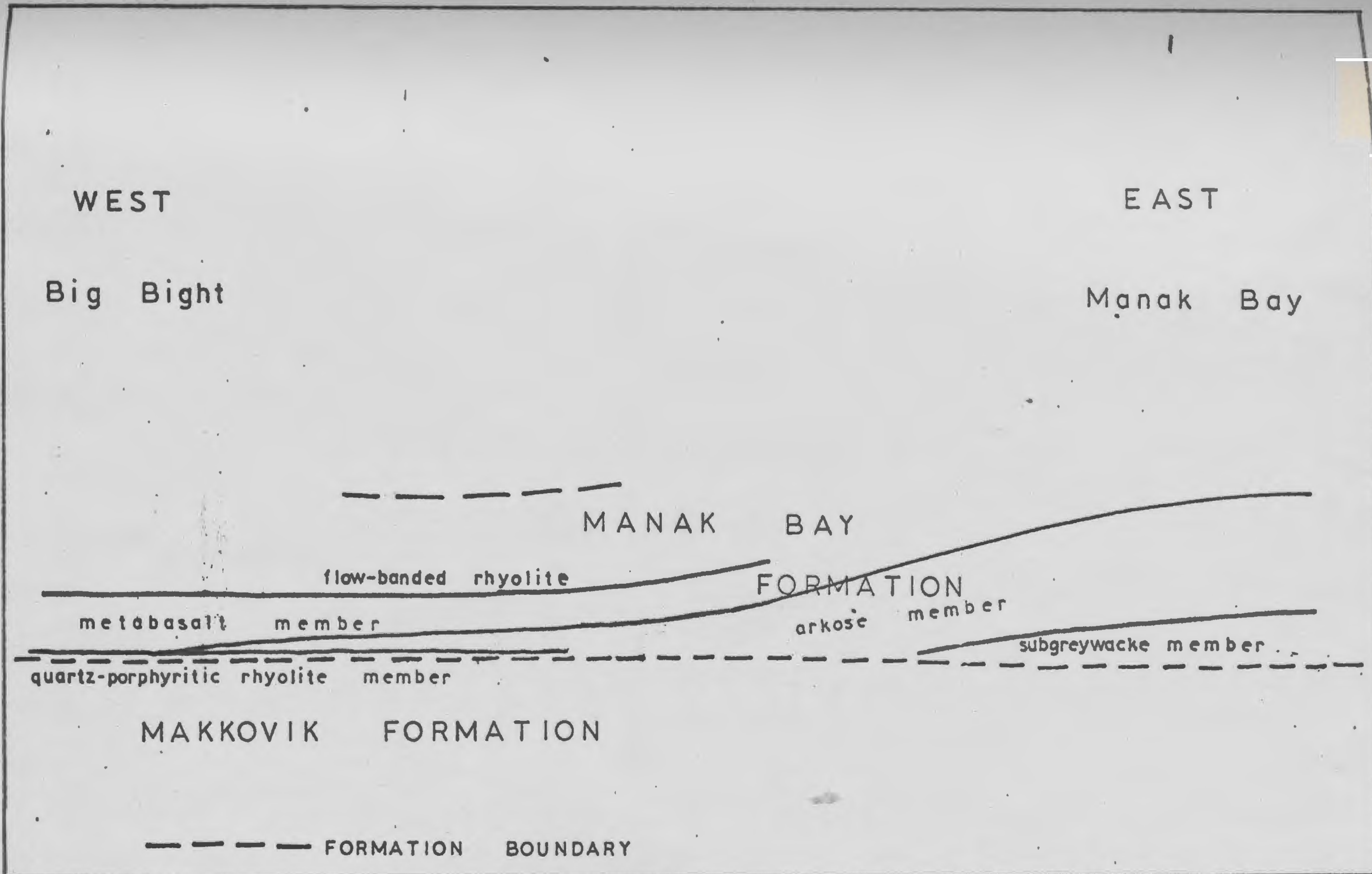


Fig. 2-25. Stratigraphic Diagram of the Manak Bay Formation from Manak Bay to Big Bight.

derived from a predominantly acid volcanic terrane, during a period of both acid and basic volcanism.

Both the Manak Bay Formation and the Pomiadluk Point Formation overlie the Makkovik Formation. However, the Manak Bay Formation disconformably overlies the Makkovik Formation whereas the Pomiadluk Point Formation conformably overlies it. Furthermore, the Manak Bay and Pomiadluk Point Formations cannot be correlated with one another in any way and they are, therefore, defined as two separate formations.

i. Subgreywacke Member: The subgreywacke member occurs southeast of Manak Bay. The member has a maximum strike-length of approximately 1 km. (1/2 mi.) and is approximately 10 m. (30 ft.) thick on the coast of Manak Bay where it is most clearly exposed. The member is very variable in appearance and composition but is mainly composed of dark grey-green, medium-grained, poorly bedded (on a 1 to 10 cm. (1/2 to 5 in.) scale) rock interlayered with pale grey and green fine- to medium-grained horizons, commonly showing incipient boudinage. The dark, grey-green horizons are composed of angular quartz, plagioclase and microcline grains up to 1/2 mm. (1/40 in.) in diameter set in a very fine-grained matrix of quartz, plagioclase, microcline and dark minerals (Fig. 2-26) arranged in 1 cm. (1/2 in.) beds of varying microcline/plagioclase ratio and quartz content.

Fig. 2-26. Photomicrograph of subgreywacke member of Manak Bay Formation, showing quartz and feldspar grains in a fine-grained feldspar-quartz-mica groundmass. Crossed nicols, X 10.

Fig. 2-27. Trough cross-bedding in arkose member of the Manak Bay Formation. South coast of Manak Bay.

The modal composition of this horizon is:

clasts: 30% of total rock

quartz	20%
microcline	5%
plagioclase	5%

matrix: 70% of total rock

plagioclase	35%
microcline	15%
quartz	5%
phlogopite	5%
epidote	4%
opaque	3%
hornblende	3%

The green horizons are diopside-rich beds containing up to 60% diopside, and the other horizons are composed of very fine-grained quartz and feldspar with 1/2 cm. (1/4 in.) phlogopite-rich bands. Some bands contain quartz phenoclasts indicating derivation from a porphyritic rhyolite.

The unusual distinctly bimodal grain-size distribution of the rock gives it the physical appearance in thin-section of a greywacke, though its mineralogy is distinctly different. The coarser fraction is of normal arkose composition, but the abundance of the fine-fraction precludes use of this term, and subgreywacke (Pettijohn, 1957) is considered the best descriptive term.

ii. Arkose Member: The arkose member occurs primarily in the vicinity of Manak Bay, where it has a 3 km. (1 1/2 mi.) strike length and a 2 km. (1 mi.) apparent thickness. Two narrow bands of cross-bedded arkose trend south from the head of Big Bight which, because of their structural setting and their lithologic similarity to some of the Manak Bay arkose, are considered to be part of that arkose. At Manak Bay the unit appears to conformably overlie the subgreywacke member, though the contact is not visible, and south of Big Bight the unit is conformably overlain by amphibolite.

Along the north and south coast of Manak Bay the arkose is a yellow-green to olive-green, fine grained meta-sediment with 1 cm. (1/2 in.) to 10 cm. (6 in.) bedding in which 2 to 4 mm. (1/8 to 1/4 in.) lamination is common. The lamination is defined by slight variation in grain size, and the bedding by erosional truncation of lamellae in cross-beds. Cross-bedding and ripple-marks are common (Fig. 2-27, 2-28 and 2-29), and parting lineations (Pettijohn, et al., 1964), flute-marks, mud-cracks (Fig. 2-29) and rain-pits were seen. Deformation of bedding in the form of both boudinage and brittle fracture is common (Figs. 4-15, 4-16, 4-50 and 4-51).

Higher up in the sequence, on the hill immediately west of Manak Bay, the bedding increases in thickness to approximately 1 m. (3 ft.), and the grain size also increases slightly. Interbedded with the arkose on the south coast

Fig. 2-28. Climbing current ripples (in vicinity of knife), overlain by oscillation ripples. Arkose member on south shore of Manak Bay.

Fig. 2-29. Current-ripples (left) and mud-cracks (right) in arkose on northwest coast of Manak Bay.

of Manak Bay are very fine-grained grey and yellow-green, thin-bedded (2 mm. to 4 mm. (1/4 to 1/2 in.) width) indurated units, which increase in number towards the base. Within these indurated units rare flame-structures and contorted bedding occur. Towards the north the unit changes through a pink and green, bedded (4 cm. (2 in.) width) arkose to a very fine-grained, dark and light grey arkose with 2 to 4 mm. (1/4 to 1/2 in.) lamination in which cross-bedding was not seen. The arkose members south of Big Bight are similar to the grey thin-bedded arkose north of Manak Bay, except that cross-bedding does occur in a few places.

The arkose is composed of:

quartz	25%
albite	25%
orthoclase & microcline	40%

with variable amounts of diopside (in some cases partially altered to hornblende), garnet, calcite, phlogopite, sphene, wollastonite and opaque minerals. Some of the dark green beds contain up to 40% albite and only 25% K-feldspar. For the most part, the K-feldspar is orthoclase, but where recrystallisation has occurred, such as in boudin necks or tectonically strained crystals, microcline has developed. In the fine-grained members the lamination is due to a variation in K-feldspar/plagioclase ratios and in dark mineral/feldspar ratios. The rock has equigranular polygonal texture of the major components, with generally decussate acicular and platy minerals.

However, a weak tectonite fabric is developed, especially in the more micaceous parts of the unit. Garnet and sphene, and to a lesser extent diopside consist of aggregates of small equigranular grains.

Current directions have been determined from cross-bedding, ripple marks and parting lineation in Manak Bay, (Fig. 2-30). The results of stereographically "unfolding" the folds to determine the original current directions are very approximate due to innate difficulties in the method of "unfolding", (Phillips, 1971; Ramsay 1967) which assumes cylindrical concentric folds, a condition not met here, Current directions appear to have been very variable (Fig. 2-30B) and there is not enough consistancy in the results to determine the overall current direction during deposition of the unit.

On the south side of Manak Bay six flattened ovoid structures were found. These were approximately 4 by 3 by 1 cm. (2 by 1 1/2 by 1/2 in.) in size, and lie parallel to the plane of the bedding and oriented with their lengths parallel to current lineations (Figs. 2-31, 2-32 and 2-33). The orientation of the structures is primary as they are not parallel to the local S- or L-tectonite fabrics. The ovoids are slightly raised up above the bedding plane due to weathering, and are formed of a finer grained material of the same composition and appearance as the beds themselves. Within the center of the structures is a smaller ovoid of

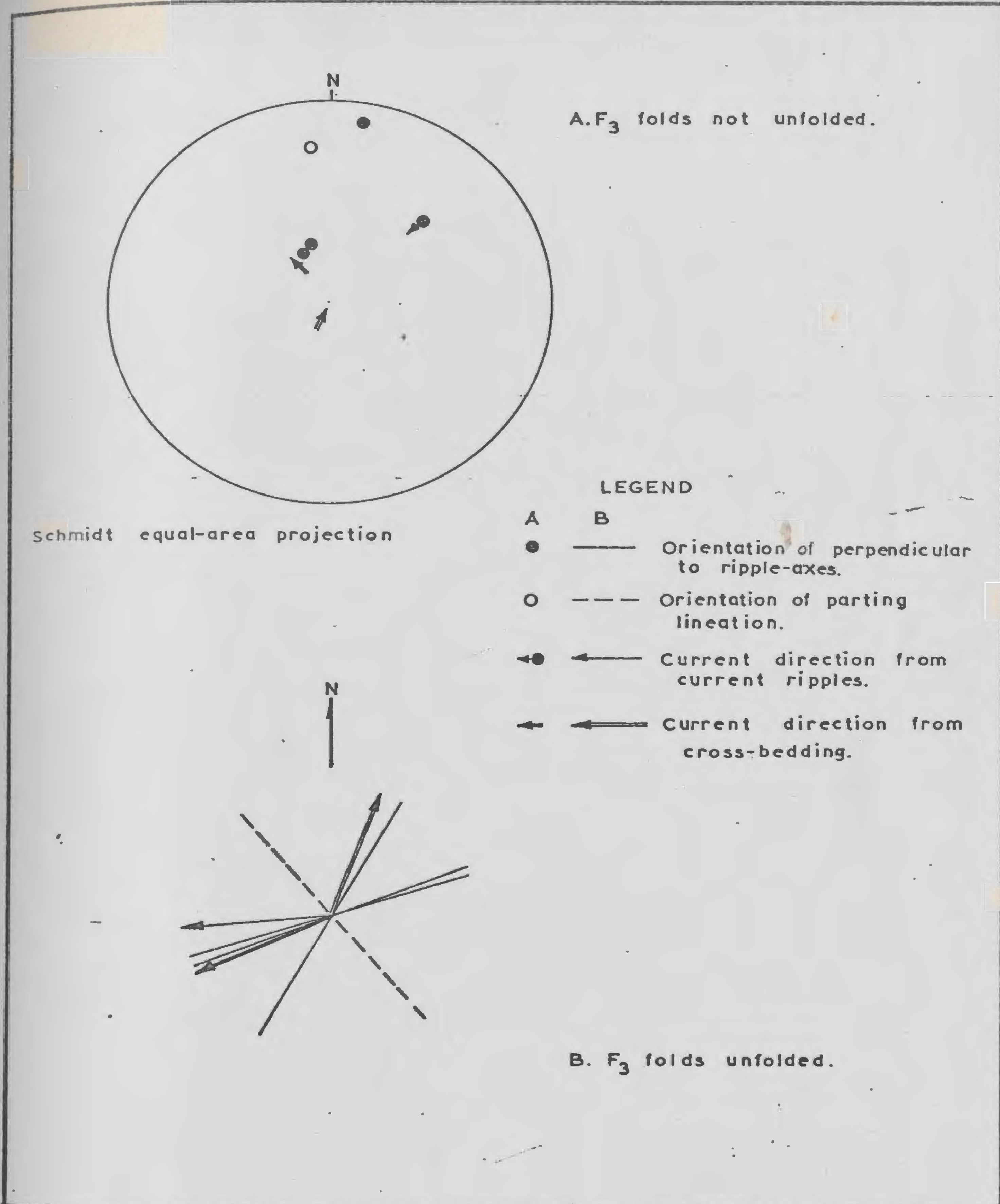


Fig. 2-30. Plot of current directions in the arkose member of the Manak Bay Formation. A: Stereographic projection before unfolding F_3 folds. B: Azimuthal plot after unfolding F_3 folds.

Fig. 2-31. Unusual concretions (?) (marked c) and parting lineation (marked l) in arkose on south shore of Manak Bay. Note elongate shape and core to the concretions.

Fig. 2-32. Concretion (?) from arkose on south shore of Manak Bay. Note coarser grained core and asymmetrical shape. Scale in millimeters.

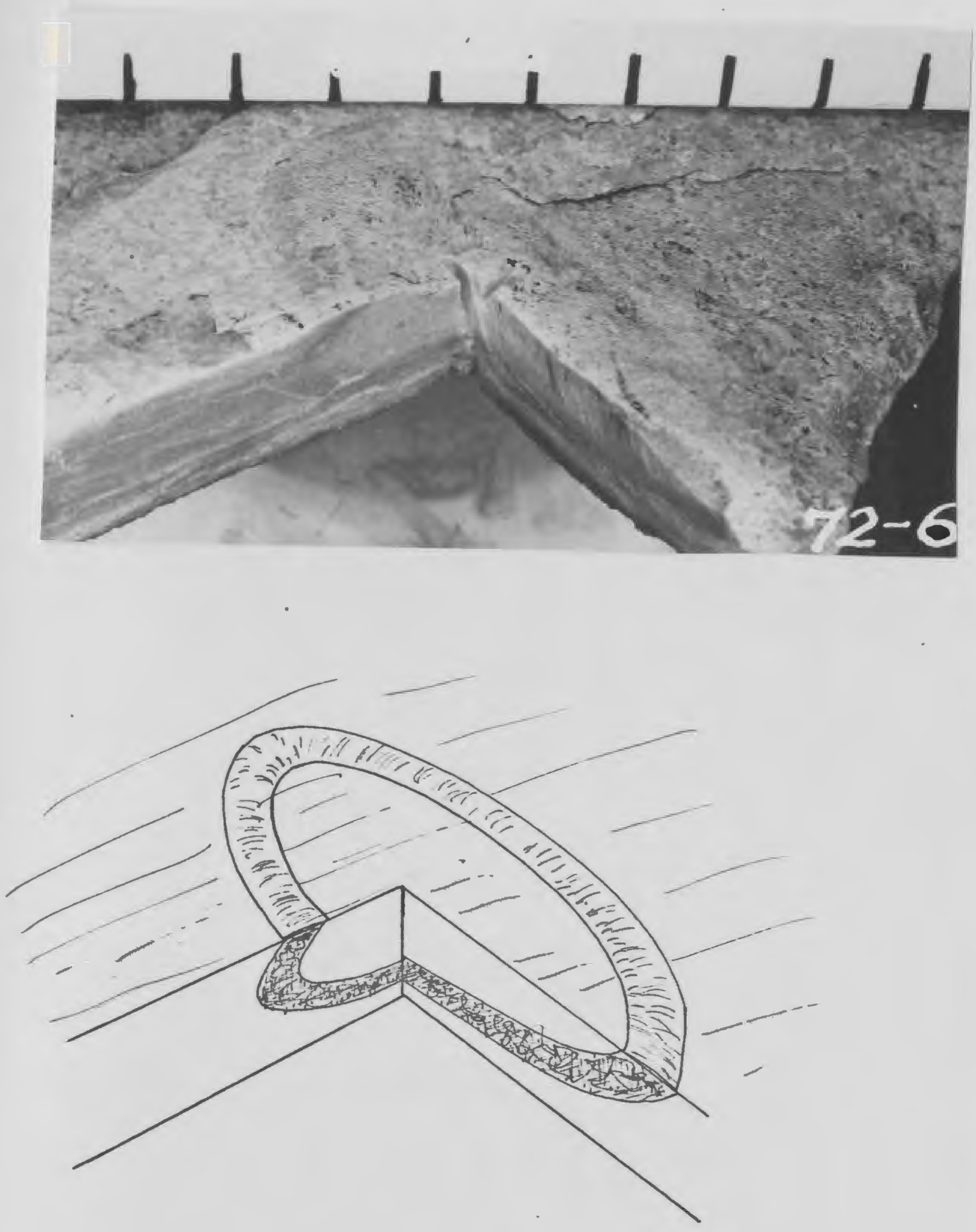


Fig. 2-33. Section of flattened concretion (?) from arkose on south shore of Manak Bay. Scale in centimeters.

slightly coarser material with the same appearance and weathering characteristics as the beds. Complete polygonal recrystallisation has destroyed any microscopic indications of origin there may have been, however, due to the age of the beds, (early Proterozoic, Gandhi, et al., 1969) these structures are not thought to be biogenic in origin. They are tentatively considered to be concretions (Pettijohn, 1957) which have undergone primary flattening and elongation. A loose boulder in the vicinity shows similar but smaller 2 by 1 by 1/2 cm. (1 by 1/2 by 1/4 in.) structures defined by an increase in mica in the outer section. These are thought to have been formed under more muddy conditions--the clay minerals having been converted to mica on metamorphism (Fig. 2-34).

iii. Metabasalt Member: The metabasalt member occurs as two 4 km. (2 mi.) long, 200 m. (600 ft.) thick parallel outcrops on either side of an anticline south of Big Bight, and conformably overlies the arkose member. The metabasalt is similar to that of the Nesbit Harbour Formation.

iv. Other Units: A flow-banded rhyolite similar to that in the Makkovik Formation has been outlined by Gandhi, et al., (1969) south of the eastern bay at the head of Big Bight, though no outcrop was seen by the present writer.

Summary of the Eastern Area

All the members and formations in the eastern area are in conformable stratigraphic contact and no disconformities

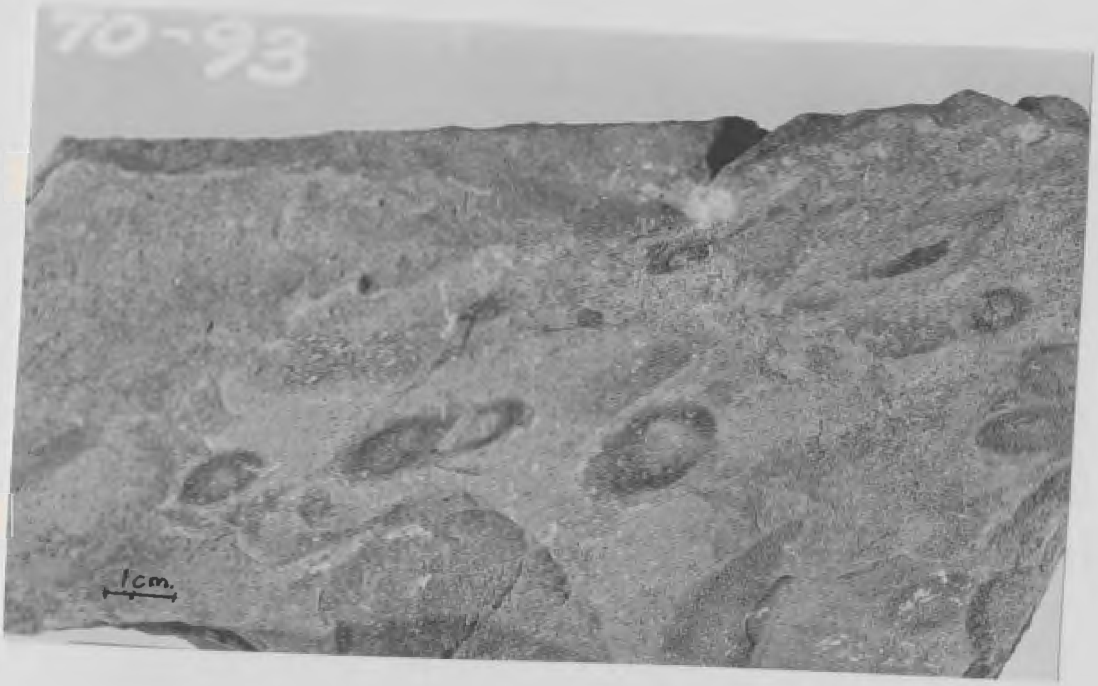


Fig. 2-34. Small micaceous mudballs from arkose on south shore of Manak Bay.

have been recognised. Also, there has been little change in rock type deposited either in the different parts of the area or at different times. The subdivision of the lithologies into formations has, therefore, been based partly on mapping convenience rather than any inherent characteristics in the rocks themselves. The Nesbit Harbour Formation metabasalt member is a widely occurring, distinctive rock-unit which resists erosion forming low hills. These factors make it easy to trace in the field and it is therefore a convenient marker unit for the upper boundary of the Nesbit Harbour Formation. Similarly, the porphyritic rhyolite member of the Makkovik Formation is distinctive in appearance and is therefore also useful as a marker. The two areas of rock which directly overlie the rhyolite and are therefore stratigraphically equivalent, the Pomiadluk Point and Manak Bay areas, are far removed from one another and the rocks cannot be directly correlated. They are therefore considered as separate formations.

The lithologies recognised in the area indicate a depositional history starting with sedimentary arkoses and conglomerates derived from acid volcanic rocks and overlain by a widespread basalt lava sequence, all comprising the Nesbit Harbour Formation. This was then followed by massive outpouring of an up to 10 km. (5 mi.) thick mass of acid extrusive rocks of three distinct types, flow-banded rhyolite, quartz-feldspar porphyritic rhyolite and microcline porphyritic rhyolite which comprise the Makkovik Formation (Plate 2). This was followed by further acid extrusives and associated conglomerates and tuffs in the northeast (the Pomiadluk Point Formation) and minor basic extrusive rocks and arkoses in the southeast (the Manak Bay Formation), which both form the highest recognised sequences in the area east of the Ranger Bight slide. Only one medium-sized occurrence of a plagioclase-porphyritic rhyolite has been recognised though the large tuff member in the Makkovik Formation contains mainly plagioclase phenoclasts, suggesting the plagioclase-porphyritic rhyolite must have been more extensive, or that other similar rhyolites occur in the Makkovik region.

The Nesbit Harbour Formation conglomerate member contains pebbles and boulders of an identical quartz-porphyritic rhyolite set in a quartzo-feldspathic matrix in which there is a high proportion of quartz phenoclasts. The very wide time-span of porphyritic rhyolite (from pre-Nesbit Harbour conglomerate to post-Manak Bay arkose) and its very wide distribution, uniform character and large volume, indicate either a single, very large magma chamber or many smaller chambers all developing near-identical products at different times, which suggests overall tectonic stability. (The conglomerates could be developed by local tectonism resulting from igneous activity i.e., caldera collapse, local crustal updoming, etc.). The abundant ripple-marks, cross-bedding and occurrences of conglomerate suggest shallow-water conditions throughout the depositional period of these rocks, with the occurrence of mud-cracks and rain-pits at Manak Bay indicating local intermittent subaerial conditions.

AREA WEST OF RANGER BIGHT SLIDE

Only one formation, the Big Island Formation, occurs west of the Ranger Bight slide.

Big Island Formation

The name Big Island Formation is proposed for a sequence of many small arkose, conglomerate, rhyolite and amphibolite units situated northeast of Tilt Cove (the location of the type-section is plotted on Plate 1). Few of these units could be delineated on the scale used for mapping the area (1 : 24,000). Neither the base nor the top of the formation have been recognised (Table 2-VI and Fig. 2-35).

The formation has a strike length of approximately 24 kms. (12 mi.) oriented from northeast to north, and a maximum true thickness of approximately 3000 m. (9000 ft.), though folding has increased the apparent thickness.

The Big Island Formation is a shallow-water sequence deposited during a period of both acid and basic volcanism and is partly correlated, tentatively, with the sequence west of Makkovik Bay (Clark, 1970 and 1971).

i. Lower Rhyolite Member: The lower rhyolite member crops out over a strike length of 5 kms. (2 1/2 mi.) north of Monkey Hill. It dips near-vertically and has a maximum true

Table 2-VI

Table showing approximate thickness of members of the Big Island Formation.

	m.	ft.
Upper rhyolite member	500-1000	1500-3000
Metabasalt member	300-500	900-1500
Cross-bedded arkose member	300-700	900-2000
Lower conglomerate member	0-1500	0-4500
Lower rhyolite member	0-1500	0-4500
Big Island Formation	2000-3000	6000-9000

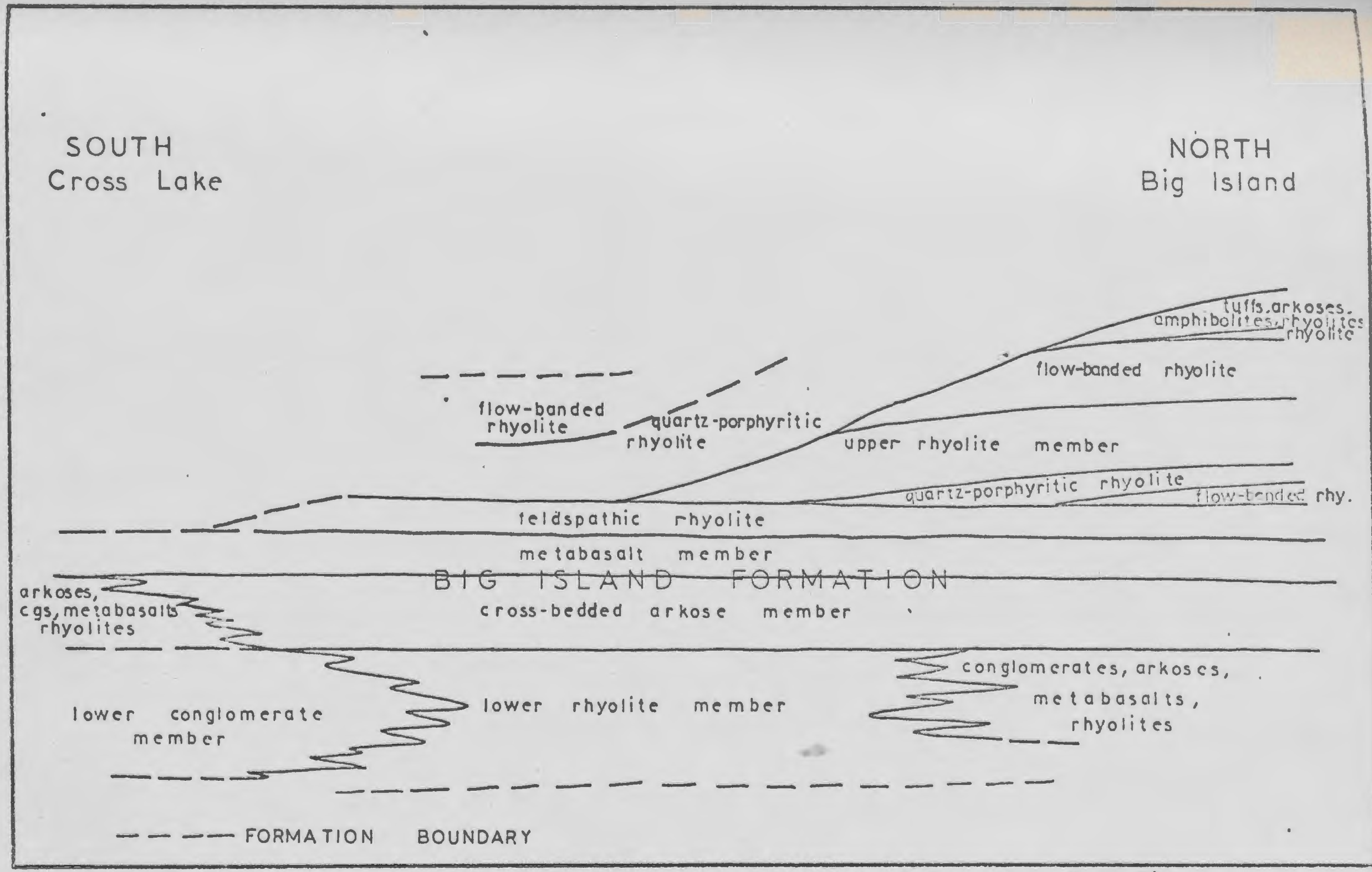


Fig. 2-35. Stratigraphic diagram of the Big Island Formation from Big Island to Cross Lake.

thickness of about 1 1/2 kms. (3/4 mi.) but wedges out northwards. The member is similar to the flow-banded rhyolite in the Makkovik Formation though flow-banding is less common. It is interbedded with various minor monomictic (rhyolite clasts) conglomerate beds, similar to the lower conglomerate of the Makkovik Formation, and is intruded on the eastern (lower) boundary by an amphibolite intrusion related to the Big Island slide zone (part of the Ranger Bight Complex). The rhyolite is interbedded in the north with numerous conglomerate, arkose, tuff, rhyolite and amphibolite beds which have not been stratigraphically distinguished from one another, and is interbedded in the south with the lower conglomerate member.

ii. Lower Conglomerate Member: The lower conglomerate member occurs between Cross Lake and the major granitic intrusion north of the lake. The conglomerate has a total strike length of 4 km. (2 mi.), but extends farther south of the area mapped, and a thickness of up to 1 1/2 km. (3/4 mi.). It is a polymictic conglomerate similar to the Nesbit Harbour conglomerate but contains epidote-rich pebbles as well as quartzo-feldspathic and rhyolitic clasts. Minor sandy beds occur, but no cross-bedding was seen.

In thin-section the matrix is seen to consist mainly of quartz and feldspar phenoclasts in a fine-grained quartzo-feldspathic matrix, with accessory amounts of biotite and opaque minerals. The majority of pebbles are rhyolitic in appearance and composed of quartz, plagioclase and microcline, though plagioclase-rich or microcline-rich psammitic (?) pebbles are also common.

iii. Cross-bedded Arkose Member: The cross-bedded arkose member occurs southeast of Tilt Cove. The arkose has a maximum thickness of 700 m. (2000 ft.), and a strike length of up to 12 kms. (6 mi.) of which the southern 4 km. (2 mi.) show cross-bedding. It conformably overlies the lower rhyolite member and the conglomerates, arkoses, etc., north and south of the rhyolite. The arkose is a fine to medium-grained, dark grey rock with 2 to 10 mm. (1/8 to 1/2 in.) laminations defined by variations in both dark mineral content and grain size. The laminations form cross-bedding up to 60 cm. (2 ft.) thick (Fig. 2-36) which indicate the unit youngs westward, but no other sedimentary structures were seen.

The arkose consists of:

albite	40%
oligoclase/andesine	20%
orthoclase	30%
dark green biotite	5%

and accessory amounts of opaque minerals and orthite. No quartz was seen. Large (up to 2 cm. (1 in.) diameter) garnet porphyroblasts occur in the arkose immediately west of Island Lake.

Fig. 2-36. Trough cross-bedding (top to right) in cross-bedded arkose member of the Big Island Formation. Arkose is cut by Monkey Hill Granite to top right of photograph.

Fig. 2-37. Tectonites [illegible] n about relict feldspathic blebs. Meta-basalt member of the Big Island Formation.

The arkose is similar in appearance to the cross-bedded psammite west of Makkovik Bay (Clark, 1970 and 1971), with which it is tentatively correlated.

iv. Metabasalt Member: The metabasalt member occurs from Big Island, south to the southern boundary of the area. The member is generally steeply dipping to vertical and has a total strike length of about 25 km. (12 mi.) of which 4 to 8 km. (2 to 4 mi.) in the south consists of sporadic xenoliths in the Monkey Hill Granite. It has a maximum thickness of approximately half a kilometer (1/4 mi.), and is best exposed west of Ranger Bight, where lack of vegetation and a transverse gully ensure excellent outcrop. The metabasalt is conformably underlain by the cross-bedded arkose. No evidence of chill margins has been recognised.

Pillow-lavas occur west of the head of Makkovik Bay and apparently dip eastward at a small to moderate angle. The pillows are well preserved, though metamorphism and deformation have destroyed the original glassy margins and radial fractures. They crop out over approximately 1 sq. km. (1/4 sq. mi.), and are conformably underlain and partly interbedded with arkose. There is no overlying unit in the area. The member here is a fine-grained amphibolite consisting of 20 to 50 cm. (1 to 2 ft.) diameter pillows with epidotic interpillow metasediment. The amphibolite shows an imperfect tectonic fabric and is identical in appearance to the pillow-lavas at Big Head on the west coast of Makkovik Bay (Clark, 1970 and 1971), with which the member is tentatively correlated.

The member generally is very similar to the Nesbit Harbour Formation metabasalt and is a fine- to medium-grained salt-and-pepper (feldspar and amphibole) textured amphibolite in which tectonite fabrics are imperfectly developed. In places the tectonic fabric forms augen up to 2 cm. (1 in.) in diameter of relict pre- and syn-tectonic feldspathic blebs (Fig. 2-37 and 2-38).

The metabasalt is generally composed of:

plagioclase	50 - 70%
hornblende, biotite	30 - 60%

and accessory amounts of sphene and opaque minerals. The plagioclase is usually labradoritic in composition suggesting the rock was originally a basalt rather than an andesite. However, northwest of Cross Lake metamorphic recrystallisation has resulted in the occurrence of both albite and andesine-labradorite. The hornblende and biotite are complementary, i.e., where hornblende is abundant biotite is rare, and vice versa, though hornblende is always present. On Big Island, the biotite has mainly been altered to chlorite, and epidote and calcite form up to 10% of the rock. Anthophyllite is locally developed as a contact metamorphic mineral.

v. Upper Rhyolite Member: The upper rhyolite member occurs over a strike length of approximately 8 km. (4 mi.) from Big

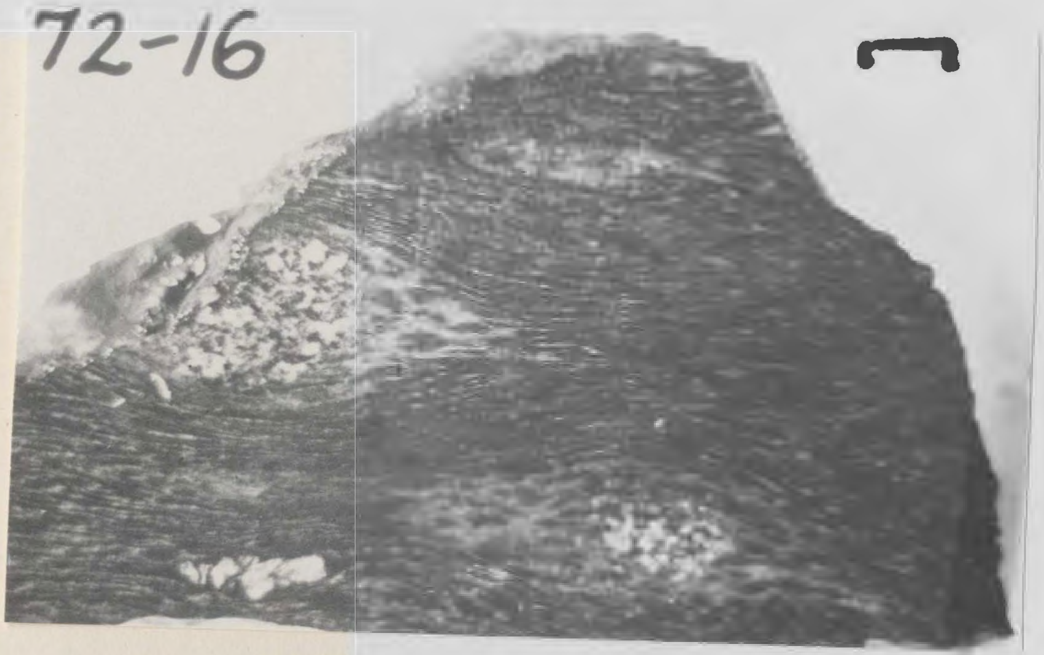


Fig. 2-38. Tectonite fabric (S_2) forming augen about relict feldspathic blebs. Metabasalt member of the Big Island Formation.

Fig. 2-39. Boudins developed from flow-banding in upper rhyolite member of the Big Island Formation.

Island, southwards to Tilt Cove. The member has a vertical to near-vertical dip and a maximum true thickness of approximately 1 km. (1/2 mi.) on Big Island. The rhyolite is most clearly exposed on Big Island. The member is a flow-banded rhyolite similar to the Makkovik Formation flow-banded rhyolite, but also contains some unbanded parts. On Big Island, and on the coast south of Big Island the rhyolite is coarsely banded on a 1/2 to 1 cm. (1/4 to 1/2 in.) scale. The bands are dark and light brown in colour and are interspersed with 1/2 by 1 cm. grey blebs. The whole has undergone tectonic flattening to produce small-scale boudinage and the appearance of a flattened pebble-conglomerate (Fig. 2-39). The rock is composed of microcline (mainly in the dark bands), plagioclase (mainly in the light bands) and quartz (the grey blebs), with minor amounts of carbonate and opaque minerals.

On the north coast of Big Island, deformed, ovoid concentric structures up to 1 by 4 by 6 cm. (1/2 by 2 by 3 in.) in size, but usually considerably smaller, occur in bands about 1 to 2 cm. (1/2 to 1 in.) apart (Fig. 2-40). These structures are similar in appearance to both lithophysae and accretionary lapilli (Moore and Peck, 1962). In hand-specimen the concentric layering appears as a colour variation, (pink-orange, grey, white and colourless), which in thin-section is seen to be due to variation in proportions of microcline, plagioclase and quartz respectively, and finely divided opaque minerals. This concentric mineralogical-layering could be due

Fig. 2-40. Deformed lithophysae in the upper rhyolite member of the Big Island Formation. North coast of Big Island.

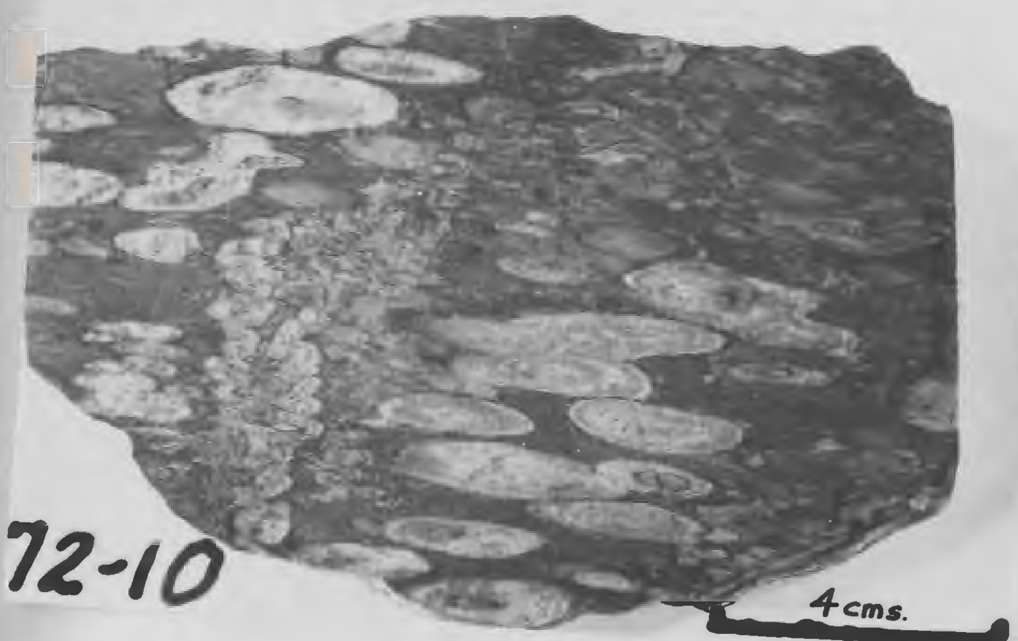


Fig. 2-41. Deformed lithophysae and early veinlet in the upper rhyolite member of the Big Island Formation. North coast of Big Island.

to either lithophysic growth (Cole, 1885) or to differential recrystallization of accretionary lapilli and consequent ionic partitioning between phases as is thought to have occurred in the development of the banding of the rhyolite (see section later). A thin, highly folded, originally planar (?) structure which bifurcates near the center of the specimen (Fig. 2-41) is considered to be an early veinlet. A similar mineralogical layering within hemi-ovoid structures occurs about the sides of the vein and is thought to be due to spherulitic crystallization which nucleated at points on the boundary of the vein. Similarly, the imperfect association of the structures with the banding in the rock (Fig. 2-40) is probably due to the initiation of spherulitic crystallization at the contacts of flow bands. This factor, and the lack of fragments of the structures in the rock (Moore, <u>et al</u>., 1962) indicate the structures are probably lithophysae and not accretionary lapilli.

vi. <u>Other Units</u>: A large number of small (1 to 20 m. (3 to 60 ft.) thick) arkose, tuff, conglomerate, rhyolite and amphibolite horizons occur below the cross-bedded arkose and are interbedded with one another and with the lower rhyolite member. No attempt has been made to accurately subdivide these, though the main beds have been noted (Fig. 4-19). The arkoses are all fine- to medium-grained, pink, white, grey or green coloured, banded rocks (2 mm. to 10 cm. (1/8 to 5 in.)

scale lamellae) with no recognisable sedimentary features. Many of the units have gradational contacts with the conglomerates, though other contacts are sharply defined. The lamellae may have been sedimentary in origin, but are now commonly transposed by tight folding and development of slides (Fig. 4-21 and 4-22). A banded (10 to 20 cm. (5 to 10 in.) scale), fine-grained, white arkose with a high proportion of muscovite, in which dark grey, ovoid (5 mm. by 2 cm. by 5 cm. (1/4 by 1 by 2 1/2 in.)) biotite blebs resembling flattened mudballs occurs on Big Island. In the northeast of Big Island, arkoses containing a high carbonate content are interbedded with salmon-pink carbonate bands on a 2 to 4 cm. (1 to 2 in) scale.

The arkoses are very variable in mineralogy and are composed of:

quartz	20 - 40%
albite	40 - 60%
microcline	10 - 40%

with minor amounts of diopside, hornblende, actinolite, wollastonite, biotite, phlogopite, muscovite, calcite, epidote and garnet. Accessory amounts of opaque minerals, sphene, chlorite, fluorite, tourmaline and zircon are usually present. The banding in the arkoses is usually defined by a change in grain size and dark mineral content, but a slight variation in ratio of microcline to plagioclase also defines a banding in some samples. The arkoses show polygonal texture, and clast outlines cannot be determined except in a few of the samples where quartz and feldspar phenoclasts and fine-grained lithic fragments are set in a fine-grained quartzo-feldspathic or biotite-rich matrix. Aslo, a few dark mineral-rich or opaque mineral-rich clots occur, indicating recrystallisation of a clast of appropriately different mineralogy to the matrix. These arkoses commonly grade into adjacent conglomerates. Some very fine-grained horizons, identical in appearance to the tuffs but usually lacking phenoclasts, form bands in the arkose. A similar, though less extensive interbedded lithology occurs above the lower conglomerate member in the south.

The microcline-porphyritic rhyolites are very similar to those of the Makkovik Formation and are massive, fine-grained, grey, porphyritic rocks with large (up to 1 cm. (1/2 in.) diameter) pink, perthitic, microcline phenocrysts (which in places mantle plagioclase phenocrysts, and which may show tectonic flattening to microcline streaks) distributed evenly throughout the rock.

The rhyolite is composed of:

quartz	25 - 40%
microcline	20 - 30%
oligoclase/ andesine	30 - 50%

with accessory amounts of opaque minerals and biotite. The matrix shows polygonal texture as do the phenocrysts, which are now composed of a medium-grained aggregate of microcline.

On the coast opposite Big Island, various conglomerates, similar to the Makkovik Formation conglomerates, are interbedded with the rhyolites, the most important of which are mentioned in the relevant sections.

A thin unit on Big Island was previously thought to be a possible remnant or sliver of basement (Sutton, et al., 1971). However, the irregular compositional "banding" has since been recognised as due to intense flattening and boudinage of pebbles of several different quartzo-feldspathic and rhyolitic (?) lithologies, and formation of small augen around quartz and feldspar porphyroclasts, indicating the unit is a polymictic conglomerate similar to others in the region.

On the north coast of Big Island there is a folded 2 m. (6 ft.) wide mafic conglomerate composed of dark grey-green to grey-brown, rounded, tectonically flattened, amphibolitic clasts up to 50 cm. by 10 cm. (1 1/2 ft. by 4 in.) in size, set in a fine-grained, black, amphibolitic matrix (Fig. 2-42).

Summary of Western Area

The Big Island Formation is the only formation west of the Ranger Bight slide, and is composed of many small units and a few large members of arkose, rhyolite, conglomerate and amphibolite which are very similar to those forming the members east of the slide. However, they are much thinner and more

Fig. 2-42. Folded (F_3) mafic conglomerate on north coast of Big Island.

Fig. 2-43. Ford's Bight conglomerate.

interbedded with one another, and far less homogeneous overall for the Big Island Formation, or any part of it, to be directly correlated with any sequence east of the slide. At present, it can only be stated that the lithologies are very similar and the formations on either side probably, therefore, approximately contemporaneous.

RANGER BIGHT COMPLEX

The name Ranger Bight complex is used informally in this thesis for those rocks associated spacially with the major tectonic slide situated along the western coast of Ranger Bight (Plate 1), which cannot be shown to be part of either the Makkovik or the Big Island Formations, or to be igneous intrusions. It is envisaged that with further, more detailed mapping in the area the various rock-types of the complex will be delineated and the formational name may be discarded.

The rocks of the complex are more deformed in the north than in the south of the thesis area. To the south they consist of medium-grained plagioclase-hornblende amphibolite which, because of its similarity in composition and texture to the mafic lavas and dykes in the area, is considered to be a deformed and metamorphosed pre- or syn-D_2 basaltic dyke. In the north, coarse- and medium-grained plagioclase-hornblende-biotite amphibolites are common, but biotite-hornblende-plagioclase schists, mafic feldspar porphyries and epidote-actinolite schists also occur (Figs. 4-12 and 4-38). The variation in amount of deformation and the difference in internal tectonic fabrics indicate these rocks are probably of different ages and many are probably, therefore, only fortuitously associated with the slide-zone. The composition of these rocks suggests they are also mainly metamorphosed mafic dykes or lavas which have either acted as a locus for the slide (lava) or have been intruded up it (dykes). The textures and metamorphic mineral growth features are discussed more fully in Chapters 4 (Structure) and 5 (Metamorphism).

The term Ranger Bight complex is, therefore, used informally in this thesis to describe a sequence of enigmatic mafic rocks spacially associated with the Ranger Bight slide-zone and thought to be metamorphosed basaltic dykes or lavas.

ORIGIN OF THE FLOW-BANDED RHYOLITES

The flow-banded rhyolites show the following features characteristic of acid lava flows:

1. Flow-banding (Boyd, 1961; Hughes and Malpas, 1971).

2. Autobrecciation (The rare possibility of autobrecciation of a "rheoignimbrite"--Rittmann, 1962--is considered unlikely).

The following features of ignimbrites were not seen:

1. Variation in grain-size or other parameter which might reflect original differential welding.

2. Lithic fragments.

3. Flattened pumice fragments (Ross and Smith, 1961).

The flow-banded rhyolites are, therefore, considered to be mainly lavas, but the following features suggest ignimbrites may form part of the sequence.

1. Conformable tuff horizons with apparently gradational boundaries with the associated flow-banded rhyolites.

2. Large area of coverage of the rhyolites. Acid lava flows form short, thick tongues (Holmes, 1966), which have not been recognised here.

Origin of the Flow-banding

Superficially the banding of the flow-banded rhyolites appears to be normal igneous flow-banding. However, in thin section the banding is seen to be due to a mineralogical differentiation.

This unusual banding consists of fine quartz bands, with rims of microcline, set in a matrix of plagioclase and quartz and minor microcline (Fig. 2-17). Opaque minerals and hornblende are concentrated in the microcline-rich bands. Compositional banding reported from the Katmai region, Alaska (Fenner, 1920) and from the Taupo area, New Zealand (Ewart, 1963) is thought, by the authors concerned, to be due to assimilation of basic material by the rhyolitic magmas. The present case is not considered to be of this type as the mafic minerals are concentrated in the microcline-rich rather than the plagioclase-rich bands. The most likely origin of the banding appears to be the segregation of (Na_2O + CaO) from K_2O during spherulitic crystallisation (Kesler and Weiblen, 1968; Ewart, 1971) along original flow-bands, possibly followed by leaching of Na_2O from the glassy portions of the rock (Lipman, 1965; Scott, 1971) by percolating groundwater. However, metamorphic differentiation as a result of clay-mineral and zeolite growth resulting in preferential concentration of some elements (Scott, 1971; Mumpton and Sheppard, 1972) may also have occured.

ORIGIN OF THE QUARTZ- AND MICROCLINE-PORPHYRITIC RHYOLITES

The quartz- and microcline-porphyritic rhyolites show a complete gradation from the quartz-predominating type to the microcline-predominating type. This gradation in mineralogy, as well as the apparently gradational contacts between some of them indicate a similar origin. These are considered to be lava-flows and not intrusive plutons for the following reasons:

1. The rhyolites are essentially conformable with the overlying and underlying units.

2. The lithology is extremely uniform in mineralogy and texture throughout the area.

3. On the east coast of Wild Bight, the porphyritic rhyolite contains angular xenoliths of a rock-type almost identical to itself.

4. No xenoliths of other lithologies have been seen.

5. Within the porphyry a few fine-grained, pale green, grey and yellow laminated beds up to 2 m. (6 ft.) thick occur, some of which show micro-cross-bedding. These are assumed to be tuff members as they generally appear to be of similar mineralogy to the rhyolite.

6. No pumiceous or definite crystal-fragments have been seen, though the latter would be extremely difficult to recognise due to recrystallisation of the phenocrysts.

There is no strong evidence against an ignimbritic origin for the rhyolites, and they are probably of mixed lava-flow and ignimbritic genesis. However, apart from some definite veins and dykes which do not commect with the main bodies of the rhyolite, the only non-conformable, transecting apophyses of the rhyolite into the country-rock occur south of Big Bight where small veinlets up to 2 cm. (1 in.) wide and 10 cm. (5 in.) long intrude the overlying Manak Bay Formation arkose and show chill-margins against the arkose. The rhyolites are therefore most likely not intrusive plutons, and these veins are probably related to local unrecognised later dykes in the rhyolite.

Ford's Bight Conglomerate (Mesozoic)

The Ford's Bight conglomerate (King, 1963a) crops out on the southeast side of Ford's Bight over an area of approximately 2 sq. m. (18 sq. ft.). It forms a sub-horizontal erosional shelf below the beach-sands and appears to extend under the sand from a small amphibolite cliff at the high-tide mark out towards the sea. The lower contact with the amphibolite bed-rock was not seen, but a 30 cm. (1 ft.) wide monchiquite (Moorhouse, 1959) dyke could be traced from the conglomerate (which it cuts) to the amphibolite (which it also cuts) indicating the conglomerate is in situ.

The conglomerate consists of 0.5 cm. to 20 cm. diameter pebbles and cobbles which show slight rounding of edges and corners, set in a fine-grained, dark grey-brown, carbonate matrix. The difference in amount of westhering between the resistant pebbles and the matrix gives the rock a very distinctive knobbly field appearance (Fig. 2-43). Many of the finer pebbles show well developed LS-tectonite fabrics, at different orientations in different pebbles, but no tectonite fabric is developed in the matrix indicating post-tectonic deposition.

The clasts of the conglomerate are identical in appearance to the amphibolites, acid volcanics and arkoses of the Aillik Group. Many of the clasts are separated entirely from surrounding clasts by the matrix (i.e., are "floating" in the matrix), indicating a syn-depositional origin for the matrix. Rock-types recognised in thin-section are: medium- to coarse-grained diorite, fine- to medium-grained amphibolite, fine-grained rhyolite, and monchiquite. The monchiquite clasts contain phenocrysts of nepheline, pyroxene and talc-carbonate replacement of serpentine, (Naldrett, 1966), set in a dark brown, partly carbonate matrix. The monchiquite clasts are similar to the dyke, but do not show any chill margin (Fig. 2-44).

The clasts show spalling of chips along the edges and intrusion of matrix into fractures and cracks in the pebbles, and some of the clasts show hazy boundaries with the matrix as a result of intense infiltration of grain-boundaries in these clasts by the matrix (Fig. 2-45).

The matrix contains angular fragments of quartz, feldspar, nepheline (?), amphibole, pyroxene, chlorite, talc, serpentine, opaque minerals and carbonate, representing all mineral species present in the clasts. The mineral fragments of the matrix are set in a cryptocrystalline, green-brown matt of generally unknown composition but in parts almost entirely composed of carbonate.

The conglomerate is unusual in that it has a carbonate matrix, some of the clasts "float" in the matrix, the matrix infiltrates the grain boundaries in some of the clasts, and angular fragments of clast-minerals occur in the matrix. These facts suggest the conglomerate was formed at the interface between a high-energy environment (such as an outwash fan) and a low-energy environment (such as a lake or sheltered part of the sea). The high energy environment would have supplied the clasts and clast-fragments, and the low energy environment the carbonate matrix. The occurrence of "floating" clasts and the infiltration of matrix into grain boundaries would be due to previously deposited carbonate being "overrun" by the high energy environment as in the case of a prograding fan.

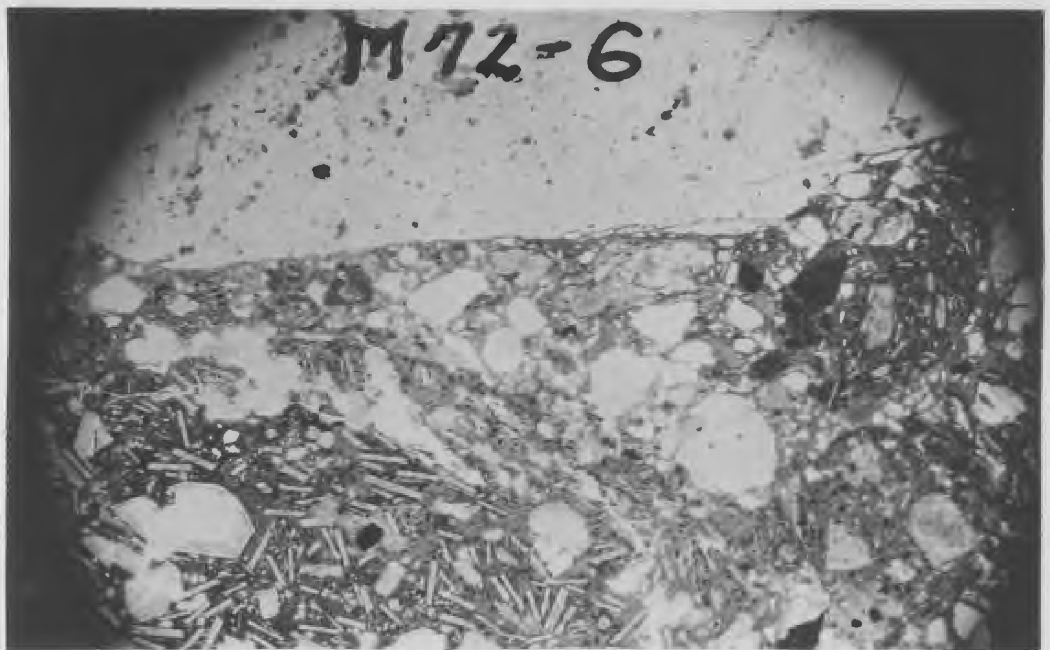


Fig. 2-44. Chill margin in monchiquite dyke cutting the Ford's Bight conglomerate. Plane polarised light, X20.

Fig. 2-45. Infiltration of clast-margins by matrix in Ford's Bight conglomerate. Plane polarised light, X150.

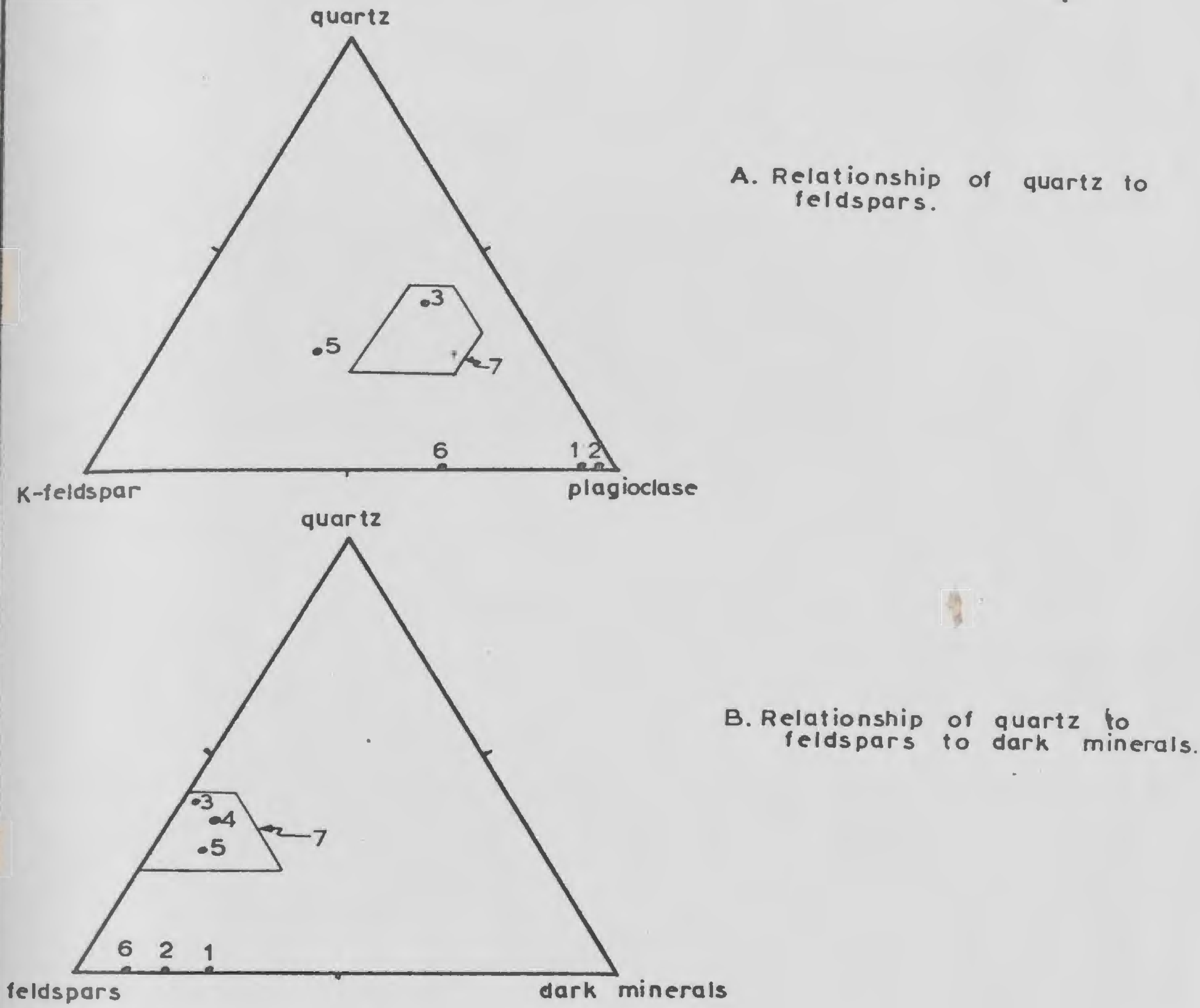
The conglomerate has been dated as Jurassic (probably Liassic) from coccoliths in the matrix, by geologists of Tenneco (Dr. N. J. McMillan, 1972, pers. comm.).

Summary and Discussion

The Aillik Group in the area under discussion has been shown to consist of a vast accumulation of acid volcanic rocks, and associated quartz-poor arkoses, (Fig. 2-46), all of which had previously been described as sedimentary quartzites (Gandhi, et al., 1969). Polymictic and monomictic (rhyolitic) conglomerates derived from an acid volcanic terrane, and widespread but relatively thin basaltic lavas complete the Aillik lithologies.

The basement to the Aillik Group does not occur in the area, but the oldest unit (the Nesbit Harbour Formation) consists of quartz-poor arkoses and a major polymictic conglomerate woth a large porportion of rhyolitic and rhyolite-like pebbles, indicating earlier acid volcanism which has not been recognised in the area (Fig. 2-47). This was followed by a widespread basalt-lava sequence, which is, however, relatively thin when compared to the following immense volume of acid volcanic rocks. The basalts are overlain by the Makkovik Formation of rhyolite lavas, rhyolitic conglomerates and rhyolitic tuffs, which form the thickest and most extensive formation in the area, reaching up to a maximum of 6 km. (3 mi.).

The Makkovik Formation is overlain in the northeast by the mixed acid volcanic/conglomeratic sequence of the



NESBIT HARBOUR FORMATION

1. Lower arkose member.
2. Upper arkose member.

MAKKOVIK FORMATION

3. Lower arkose member.
4. Transgressive arkose member.

MANAK BAY FORMATION

5. Arkose member.

BIG ISLAND FORMATION

6. Cross-bedded arkose member.
7. Arkoses of the Big Island-Ranger Bight area.

Fig. 2-46. Triangular diagrams showing the ratio of quartz to feldspars to dark minerals in some arkoses of the Aillik Group.

Recognised in Thesis Area

Not Recognised in Thesis Area

Pomiadluk Point Formation

rhyolite lavas, tuffs and polymictic conglomerates

Manak Bay Formation

arkoses and mafic volcanics

Big Island Formation

rhyolite lavas, tuffs, mafic lavas, conglomerates and arkoses

Ranger Bight Complex (amphibolites)

Makkovik Formation

rhyolite lavas, tuffs and conglomerates

Nesbit Harbour Formation

quartz-poor arkoses, polymictic conglomerate, basaltic lavas

acid volcanics and other lithologies which form pebbles in the Nesbit Harbour Formation conglomerate

basement (Hopedale Complex)

Fig. 2-47. Diagrammatic representation of relationship of Aillik formations to one another and to the basement.

Pomiadluk Point Formation, and in the southeast by the stratigraphically equivalent arkoses and mafic volcanics of the Manak Bay Formation, both of which indicate a waning of acid volcanicity and onset of normal sedimentary deposition, with the sediments being primarily derived from the local acid volcanic terrane.

The mixed sedimentary and acid volcanic lithology of the Big Island Formation suggests it is more likely to be related to the Nesbit Harbour Formation or the Pomiadluk Point Formation than to the adjoining Makkovik Formation, though direct correlation cannot be made due to separation by the Ranger Bight complex.

The early Proterozoic Aillik Group is unconformably overlain by the Ford's Bight conglomerate of Jurassic age.

It is noted that there is an almost complete lack of pelitic sediments and limestones (marbles), though carbonate is a common, and in some cases abundant, accessory in many of the arkoses, rhyolites, tuffs and conglomerates, and thin carbonate bands do occur in an arkose on the north coast of Big Island. It is, therefore, possible that some of the amphibolites, and especially those containing abundant epidote, scapolite and/or biotite that are associated with biotite schist, may have been derived by metamorphism of closely associated pelite and carbonate-rich sediment bands (Orville, 1969; Blank, 1972), though mapping was not carried out on small enough scale to permit correlation of such associations.

CHAPTER 3

IGNEOUS INTRUSIONS

Introduction

Approximately one-quarter of the thesis area is underlain by igneous intrusive rocks varying in composition from basic (gabbro, hornblendite) to acidic (granite, acid pegmatite) with most intermediate types being represented. The intrusions vary in age from pre-tectonic[1], through syn-tectonic to post-tectonic, but with no associated regular variation in petrology, most types being represented at all stages. Specific age relationships are given in Chapter 6.

The names of the various intrusions have been left unchanged from previous publications where possible, but most of the names are new (Table 3-I). The major intrusions are named as formations (American Commission on Stratigraphic Nomenclature, 1970, Articles 6h and 10i), but the minor intrusions are given purely descriptive petrographic names.

Those minor intrusions that are definitely related to a specific major intrusion are included in the description of that intrusion only.

Major Igneous Intrusions

The major igneous intrusions into the Aillik Group range in composition from the gabbroic Adlavik Igneous Complex to the Strawberry Granite, with intermediate dioritic

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[1]See p. 169 for definition.

Table 3-I

Table of major igneous intrusions in the thesis area, in order of increasing age.

Strawberry Granite (name informally used by Kranck, 1953, and Gandhi, *et al*., 1969).

Monkey Hill Granite (name informally used by Gandhi, *et al*., *op. cit*.; "Makkovik granite" of Kranck, 1939).

October Harbour Granite (new name; considered part of the Strawberry Granite by Gandhi, *et al*., *op. cit*.).

Adlavik Igneous Complex (new name; "Adlavik gabbro" of Gandhi, *et al*., *op. cit*.).

Round Pond Granodiorite (new name; "granoblastic feldspathic quartzite" of Gandhi, *et al*., *op. cit*.).

Kennedy's Cove Gneiss (new name; "granite gneiss" of Gandhi, *et al*., *op. cit*.).

Long Island Gneiss (Gandhi, *et al*., *op. cit*.).

Grampus Cove Gneiss (new name).

and quartz monzonitic intrusions also represented (Table 3-II). The intrusions are discussed in order of decreasing age, the evidence for which is discussed in Chapter 6.

None of the major intrusions in the area is pre-tectonic in age, though many of the syntectonic bodies were intruded during tectonically inactive periods between the various phases of deformation. The earliest intrusions are granitic gneisses and gabbroic bodies that were intruded prior to or during the main deformation in the area, as indicated by tectonic fabrics within and at the margins of the intrusions. Two granitic intrusions were emplaced after the main deformation, but before the end of tectonism. There are no major post-tectonic intrusions. There is no recognisable pattern to the distribution of any of the major intrusions in the area.

Minor igneous intrusions not directly related to major intrusions are discussed under a separate heading.

INTRUSIONS EMPLACED BEFORE OR DURING THE MAIN DEFORMATION

Grampus Cove Gneiss

The Grampus Cove Gneiss outcrops for approximately 1 km. (1/2 mi.) along the coast south of Manak Bay. The inland extension of the formation is not known. The unit is intrusive into the Manak Bay Formation and appears to be intruded by the diabase facies of the Adlavik Igneous Complex,

Table 3-II

Table listing the main features of the major igneous intrusions

Intrusion	Composition	Age relative to tectonism[1]	Intrusive relationships
Strawberry Granite	granitic	post main deformation (post-D_2) but syn-tectonism.	—
Monkey Hill Granite	quartz-monzonitic		—
October Harbour Granite	granitic	pre- or syn-D_2 but syn-tectonism	—
Adlavik Igneous Complex	hornblende-gabbroic to hornblende-dioritic		Intruded by Monkey Hill Granite
Round Pond Granodiorite	granodioritic		
Kennedy's Cove Gneiss	granitic to quartz-monzonitic		
Long Island Gneiss	monzonitic to dioritic		Intruded by Kennedy's Cove Gneiss
Grampus Cove Gneiss	granitic		Intruded by Adlavik Igneous Complex

[1]See p. 169 for definition.

as well as by various associated complex net-vein dykes.

The gneiss is a medium-grained, greyish-green to pink, feldspar-porphyritic granitic gneiss with a poorly developed foliation of elongate mafic clots. In thin-section the gneiss is seen to be composed of phenocrysts of andesine set in a groundmass of quartz, albite, andesine, perthitic microcline and dark minerals. The groundmass has undergone imperfect polygonal recrystallisation, and the edges of the phenocrysts have also been recrystallised. The modal composition of the gneiss is:

quartz	30 - 35%
plagioclase phenocrysts	10 - 15%
plagioclase in groundmass	10 - 20%
microcline	30 - 40%
dark minerals	10 - 20%

The dark minerals are mainly green hornblende, brown and green biotite and opaque minerals, with accessory and minor amounts of sphene, zircon, apatite, epidote, chlorite and saussurite.

Long Island Gneiss

The Long Island Gneiss crops out over 4 sq. kms. (1 sq. mi.) on the eastern coast of Makkovik Bay. It also crops out over 2 sq. kms. (1/2 sq. mi.) on the north face of Monkey Hill. Contacts were only seen at Makkovik Bay, where the gneiss is intruded by Monkey Hill Granite.

The Gneiss is usually a fine- to medium-grained pale cream, pink or grey rock with 2 cm. (1 in.) mafic xenolithic clots, feldspar phenocrysts and a well defined tectonic-fabric of hornblende needles and aggregates. However, away from the margins of the intrusion, the grain size increases and the porphyritic and tectonic fabrics die out so that the gneiss looks like a normal coarse-grained leucocratic intrusion.

The gneiss is composed of:

quartz	5 - 10%
albite and andesine	50 - 60%
K-feldspar	10 - 30%
hornblende, biotite	10 - 15%

with accessory amounts of chlorite, sphene, and pyrite with sparse chalcopyrite and magnetite. Epidote (5%) and actinolite (2%) occur in the undeformed coarse-grained variety, but only occur as very minor constituents along with chlorite and sericite elsewhere.[1] Albite is the only plagioclase in the margins of the intrusion, but andesine also occurs away from the margins. The gneiss in the area is monzonitic to dioritic in composition, whereas across Makkovik Bay in the vicinity of Long Point Cove and Swell Lake, and on Long Island in Kaipokok Bay, it is quartz-monzonitic to granodioritic in composition (Gandhi, *et al*., 1969; Clark, 1970 and 1971).

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[1] See page 273.

Kennedy's Cove Gneiss

The Kennedy's Cove Gneiss occurs over an area of approximately 15 sq. kms. (2 1/2 sq. mi.) between Cross Lake and Makkovik Bay. It is an extension of the granite gneiss of Clark (1970 and 1971) and Gandhi, et al., (1969) which occurs on the northwest side of Makkovik Bay, and is termed the Kennedy's Cove Gneiss because of its excellent exposure at Kennedy's Cove north of Big Island. In the thesis area the formation intrudes the Aillik Group on its southern margin, and is intruded by the Monkey Hill Granite on its northern margin.

The gneiss is an orange to buff coloured, coarse-grained, granitic to quartz-monzonitic gneiss and is composed of:

quartz	30 - 40%
albite/ oligioclase	15 - 25%
perthitic microcline	30 - 40%
biotite } hornblende }	10 - 20%

with accessory amounts of opaque minerals, zircon, chlorite and actinolite. The rock shows equigranular texture which has undergone tectonic flattening and polygonal recrystallization, with the development of a poorly defined gneissic foliation of dark minerals.

Round Pond Granodiorite

The Round Pond Granodiorite is sporadically exposed over an area of approximately 8 by 3 km. (4 by 1 1/2 mi.) in the core of the Round Pond dome. The contact with the overlying arkose was not seen, but has a map pattern essentially parallel to banding and bedding in the overlying units on the west side of the dome, where the two rock-types could be traced to within a few tens of yards of one another. Elsewhere in the dome no outcrop was seen and the contacts are, therefore, purely arbitary. The granodiorite may, therefore, be either intrusive into the Aillik sediments, or unconformably overlain by them. A very weak gneissic foliation parallel to the schistosity in the Aillik Group indicates the rock is pre- or syn-tectonic in age. It may be an anatexite of the Hopedale Complex basement, to some of which (west of Kaipokok Bay) it is closely similar in appearance (Sutton, 1972, and personal communication 1972), but the lack of deformation of the overlying arkoses within a few tens of meters of the contact suggests this is not the case as throughout the basement/cover contact (Fig. 1-2), intense localised deformation occurs (Sutton, et al., 1971). Also, the granodiorite is dissimilar to anatectic rocks of the basement in the vicinity of the contact (Marten, 1972, personal communication).

The Round Pond Granodiorite is intruded by the Monkey Hill Granite.

The granodiorite is a fine- to medium-grained, grey to pink, equigranular rock with a poorly developed gneissic foliation defined by dark minerals. The rock is composed of:

albite and oligoclase/ andesine	60%
quartz	30%

and accessory amounts of opaque minerals, hornblende, diopside, chlorite, sphene and microcline. In thin-section the granodiorite has a polygonal fabric, but quartz tends to cluster, as though derived by recrystallisation of single interstitial grains, indicating derivation from an igneous rock. Also, due west of Round Pond, glomero-porphyritic aggregates of plagioclase have resisted polygonal recrystallisation, but show antiperthitic intergrowth. Garnet occurs as aggregates of small grains associated with the larger plagioclases.

Adlavik Igneous Complex

The Adlavik Igneous Complex occurs in the southeast of the area and extends from east of Falls Lake to the coast south of Manak Bay for a distance of approximately 16 kms. (8 mi.), northwards and along the west coast of Big Bight and southeastward out of the thesis area (Stevenson, 1970). The complex underlies an area of at least 200 sq. km. (50 sq. mi.) (Stevenson, 1970) and consists of hornblende-gabbroic and dioritic rocks related to a large differentiated (appinitic?) basic intrusion of lopolithic shape (Gandhi, *et al*., 1969). It was not seen by Kranck (1953), but is described by Gandhi, *et al*., (*op. cit.*) who informally named it the Adlavik gabbro and separately described the surrounding diorites (called syenites by him). However, the spatial association of the diorites to the main rock-types of the Adlavik Igneous Complex and the later cross-cutting hornblende porphyritic and hornblende dykes strongly suggest the diorite is directly related to the complex.

Several igneous facies have been recognised in the intrusion and range from a basal chill-margin in the west, up the intrusion through a rhythmic layered facies to a predominantly massive plagioclase-cumulate (gabbro) facies on the coast. At Big Bight similar plagioclase-cumulates occur. On the northern boundary of the intrusion, and extending northwards and across to the west coast of Big Bight is a medium-grained diabasic facies. At Big Bight, and along the coast of Manak Bay, complex diorite and hornblendite intrusions are presumed to be related to the gabbro. The contact of the Adlavik Igneous Complex with the Aillik Group appears to be generally sub-vertical, except in the extreme west where it dips eastwards at 50° - 60°.

i. Chilled Margin Facies: A 3 m. to 50 m. (10 to 250 ft.) wide fine- to medium-grained chilled margin is preserved on the western contact with the Aillik Group. This chilled margin varies from a fine-grained amphibolite with well developed tectonite fabric at the contact to a medium-grained meta-diabase away from the contact, where it grades into the overlying rhythmic layered facies.

In hand-specimens, the meta-diabasic portion of the chilled margin away from the contact shows subhedral to euhedral inter-locking plagioclase laths up to 5 mm. (1/4 in.) long ophitically enclosed in dark green hornblende. In thin-section however, the plagioclase laths and "ophitic" hornblende are seen to have undergone tectonic stress which has produced irregular, commonly bifurcating, trains of small grains in a polygonally recrystallised mineral (Fig. 3-1).

The metadiabase is composed of:

plagioclase	50%
green hornblende	30%
opaque minerals	10%
biotite	5%

with minor amounts of sphene, chlorite, epidote and sericite. The last three minerals tend to be most concentrated in the vicinity of the irregular fine-grained zones of higher strain.

The amount of deformation in the rock increases, the grain-size decreases and alteration to chlorite, epidote, clinozoizite and sericite decreases towards the contact.

ii. Rhythmic Layered Facies: The rhythmic layered facies occur immediately above the chilled margin facies at the

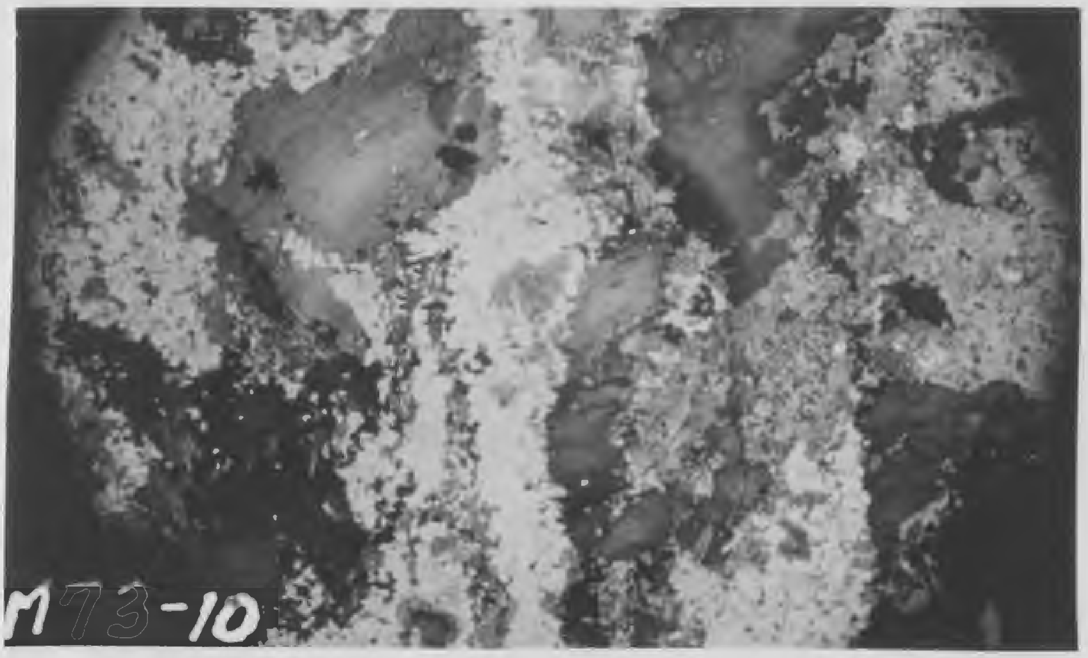


Fig. 3-1. Photomicrograph of fine-grained blasto-mylonitic fabric developed in micro-shear-zone in the chilled margin facies of the Adlavik Igneous Complex. Plane polarised light, X10.

Fig. 3-2. Rhythmic layering in the Adlavik Igneous Complex. Top towards right.

western contact of the complex and also on the west coast of Big Bight. Its thickness was not investigated, but is greater than 200 m. (600 ft.), though the upper border was not seen. The facies has a very variable dip, presumably due to folding, but a fairly constant northwesterly strike, and is in gradational contact with the underlying chilled margin facies.

The rhythmic layered facies shows very well developed 5 cm. to 50 cm. (2 1/2 in. to 1 1/2 ft.) thick, dark and light hornblende-cumulate and plagioclase-cumulate layers in which graded-bedding and scour structures are common and indicate the layering is upward facing towards the east (Fig. 3-2 and 3-3). The hornblende-cumulate consists of 2 cm. (1 in.) diameter, roughly spherical, dark green hornblende aggregates between which is an intercumulate of plagioclase and other minerals. This grades upwards to an accumulation of large (up to 1 cm. (1/2 in.) diameter) euhedral hornblende crystals (Fig. 3-4) which then gives way to the plagioclase-cumulate consisting of 2 to 3 cm. (1 to 1 1/2 in.) blocky plagioclase grains with an intercumulate of hornblende and other minerals. In thin-section the hornblendes may show indistinct zoning and commonly have a core of chlorite, pale green actinolite (?), and epidote or clinozoizite, which suggests the hornblende enveloped a pyroxene core which later altered to chlorite, actinolite, etc.. Rational boundaries on the subhedral to euhedral

Fig. 3-3. Rhythmic layering in the Adlavik Igneous Complex. Top towards left.

Fig. 3-4. Gradational boundary between layers in rhythmic layered facies of the Adlavik Igneous Complex. Top towards right.

hornblende indicate it is primary.

The rock consists of:

hornblende	30 - 40%
hypersthene	10 - 15%
labradorite-bytownite	15 - 20%
pigeonite	0 - 10%
serpentine (after olivine), opaque minerals, sphene, chlorite, calcite, epidote, phlogopite, actinolite	20 - 25%

The plagioclase is generally highly saussuritised. The hypersthene occurs as distinct intercumulus and small cumulus grains as well as randomly distributed inclusions in the hornblende, and commonly has olivine (now altered to serpentine) cores indicating reaction with the magma. The hornblende is zoned and has an irregular rim of pyroxene just in from the edge of the grains, over which is hornblende of a slightly different composition indicating initial crystallization of hornblende followed by reaction with the liquid, followed again by further hornblende crystallisation (Fig. 3-5). Phlogopite is abundant in the hornblendes where reaction has occured. The olivine (now

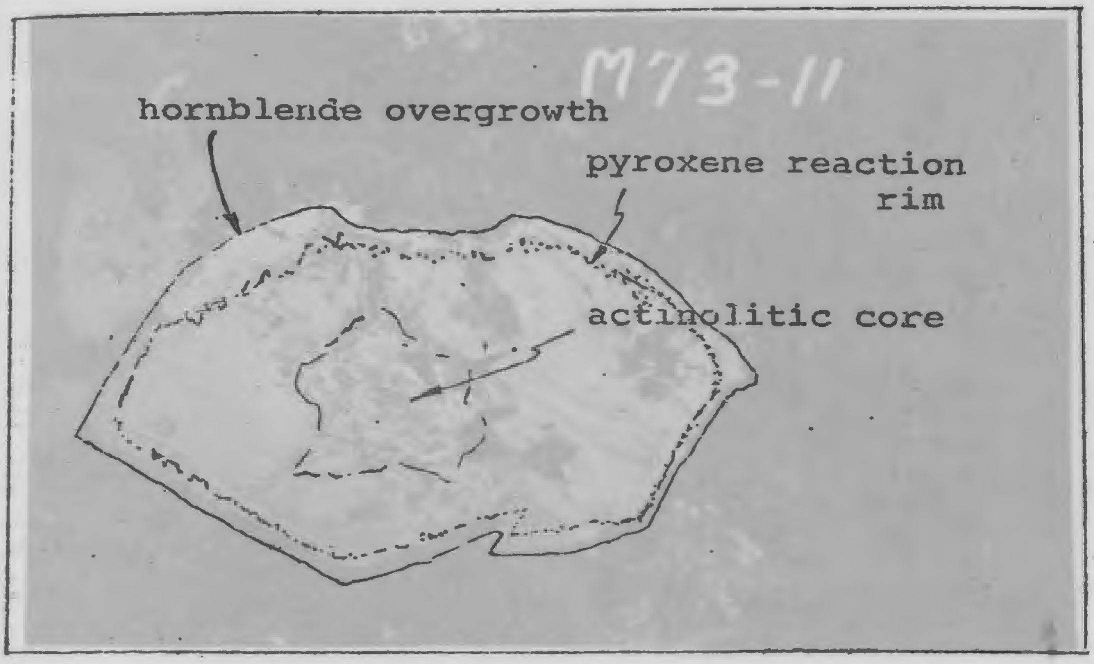


Fig. 3-5. Photomicrograph of hornblende grain in the rhythmic layered facies of the Adlavik Igneous Complex showing pyroxene reaction rim and later hornblende overgrowth. Note alteration of core of grain to actinolite. Crossed-nicols, X10.

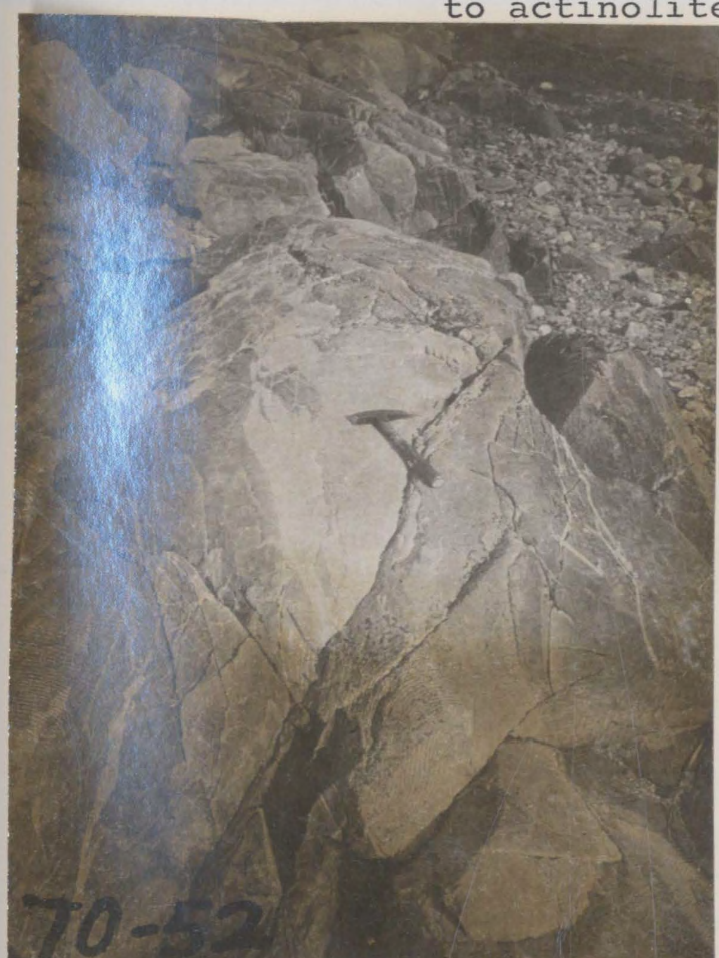


Fig. 3-6. Veins of diorite facies of the Adlavik Igneous Complex in the wacke member of the Manak Bay Formation. North of Grampus Cove.

Fig. 3-5. Photomicrograph of hornblende grain in the rhythmic layered facies of the Adlavik Igneous Complex showing pyroxene reaction rim and later hornblende overgrowth. Note alteration of core of grain to actinolite. Crossed-nicols, X10.

Fig. 3-6. Veins of diorite facies of the Adlavik Igneous Complex in the wacke member of the Manak Bay Formation. North of Grampus Cove.

The chill margin consists of 5 mm. (1/4 in.) long quartz and feldspar grains oriented perpendicular to the contact (Fig. 3-7).

Shadow extinction of quartz, minor kinking of plagioclase grains and minor intragranular crushing indicate slight post-solidification tectonic deformation has taken place.

Strawberry Granite

The Strawberry Granite was first named and described by Kranck (1953) and outcrops over an area of approximately 6 km. by 6 km. (3 by 3 mi.) on the northwest half of Cape Strawberry, and intrudes the Makkovik Formation. In the vicinity of the granite, associated pegmatite and quartz veins are common.

The granite is a coarse grained (up to 1 cm. size), pink to orange hornblendic granite containing:

microcline	50%
plagioclase	15%
quartz	30%

and various amounts of hornblende, biotite, fluorite and opaque minerals as minor accessory minerals. Graphic intergrowth of quartz and feldspar is common in the southeast of the intrusion.

A thorough hand-specimen, thin-section and mineralogical description of the granite is given by Cooper (1951).

Fig. 4-47. F_4 refolded F_3 folds (F_3/F_4 eye-pattern) in arkoses of the Pomiadluk Point Formation. North coast of Big Bight.

Fig. 4-47. F_4 refolded F_3 folds (F_3/F_4 eye-pattern) in arkoses of the Pomiadluk Point Formation. North coast of Big Bight.

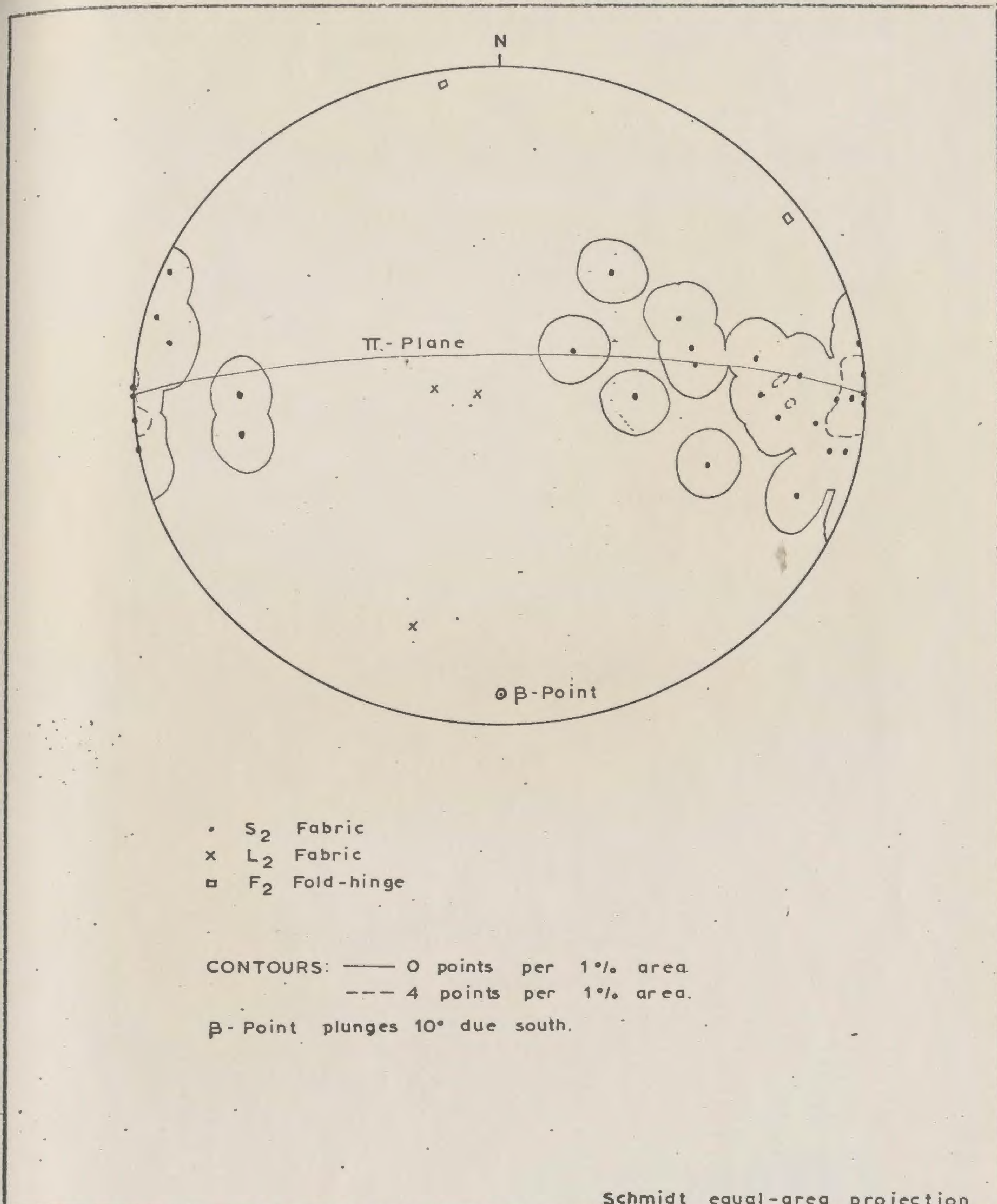


Fig. 4-48. Stereographic projection of S_2 and L_2 fabrics from the eastern part of Monkey Hill.

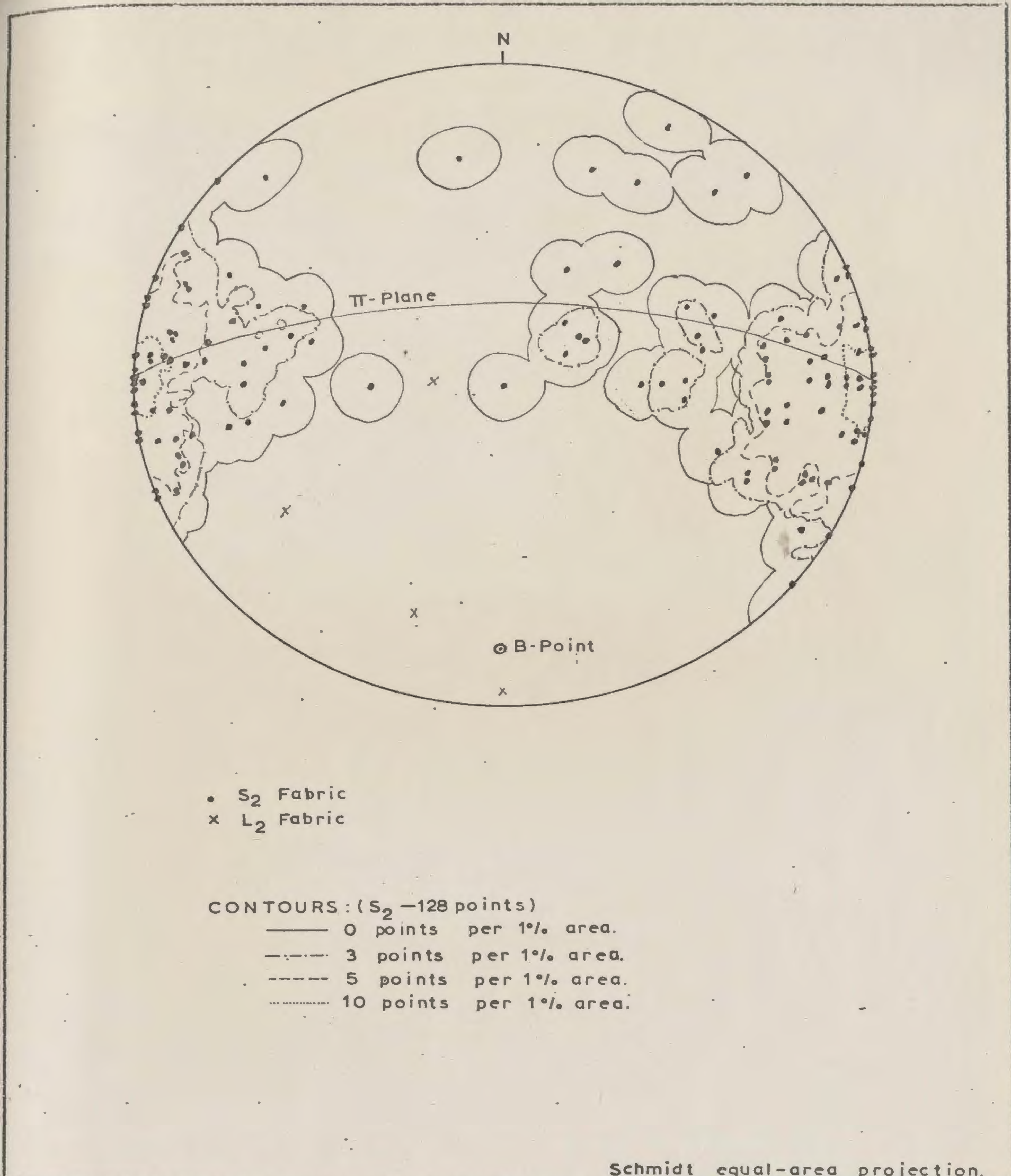


Fig. 4-49. Stereographic projection of S_2 and L_2 fabrics in the vicinity of the southern closure of the Round Pond dome.

those showing cross-bedding, are fractured at approximately 45^{o} to the axial-plane of the F_2 folds, and the fragments slid and "rotated" slightly relative to one another (Fig. 4-50). This fracturing and "rotation" is considered to be due to tectonic flattening under brittle conditions on axes oriented at an angle to bedding, producing boudins which have been "rotated" as deformation progressed (Ramsay, 1967, p. 103). No tectonic fabrics were recognised in the blocks. The surrounding finer grained beds have taken up the resulting differential strain in a plastic manner. Further north the rocks have similar fractures, but have behaved slightly more plastically resulting in curved fracture-planes and bedding (Fig. 4-51).

Although these fractures could have developed during either D_2 or D_3 they are thought to be related to the later (D_3) deformation as D_2 boudinage in the area (Figs. 4-15 and 4-16) indicates less brittle conditions than necessary for such fracturing. The hinge of the Manak Bay syncline is also fractured, and is thought to be due to tightening of the fold during D_3 (Fig. 4-7).

v. Simple-shear Belts: Small D_3 simple-shear belts (Ramsay, and Graham, 1970) occur in the quartz-porphyritic rhyolite on the east coast of Wild Bight. These belts deform and reorient the earlier S_2 fabric, but are not considered to have contributed much to the total deformation of the area.

Fig. 4-50. D_3 boudinage fractures in arkose member of Manak Bay Formation. Northwest coast of Manak Bay.

Fig. 4-51. D_3 boudinage fractures in arkose member of Manak Bay Formation. Coast north of Manak Bay. Note curved fractures in most beds, but straight (more brittle) fracture in thick more competent bed below.

The belts are, in some cases, localised by lithologic variations as shown by a shear belt at the contact between the flow-banded and the quartz-porphyritic rhyolites (Fig. 4-52).

The shear-belts are discussed further in Appendix II.

THE FOURTH DEFORMATION (D_4)

Several sets of folds trending approximately northeast and varying from small-scale crenulations of S_2 (Figs. 4-53, 4-54, 4-8 and 5-2) through larger open folds (Fig. 4-47) to possible large-scale open refolding of the Round Pond F_2 anticline to gove the Round Pond dome are all considered, because of their similar trends, to have developed during the same deformation (D_4). These F_4 folds are generally poorly developed and are uncommon in this area.

Measurement of fold orientations was difficult, but the results indicate a roughly east/west trending, steeply rather than shallowly dipping axial-plane, and variable plunge (Fig: 4-55). Vergence of F_4 folds of bedding and S_2 is variable.

POST-D_4 TECTONIC EVENTS

Several post-D_4 tectonic events are recognised which are too localised or have too small an effect to be considered as separate deformations. In all cases the events resulted in brittle fracturing. It has not been possible to determine the relative ages of the events due to a lack of cross-cutting relationships.

Fig. 4-52. D_3 simple-shear belt at tuffaceous(?) contact between the quartz-porphyritic rhyolite member of the Makkovik Formation (right) and the overlying lower rhyolite member of the Pomiadluk Point Formation (left). Note refraction of fabric into and out of the shear-belt (fabric parallel to pencil and hammer handle outside belt). East coast of Wild Bight.

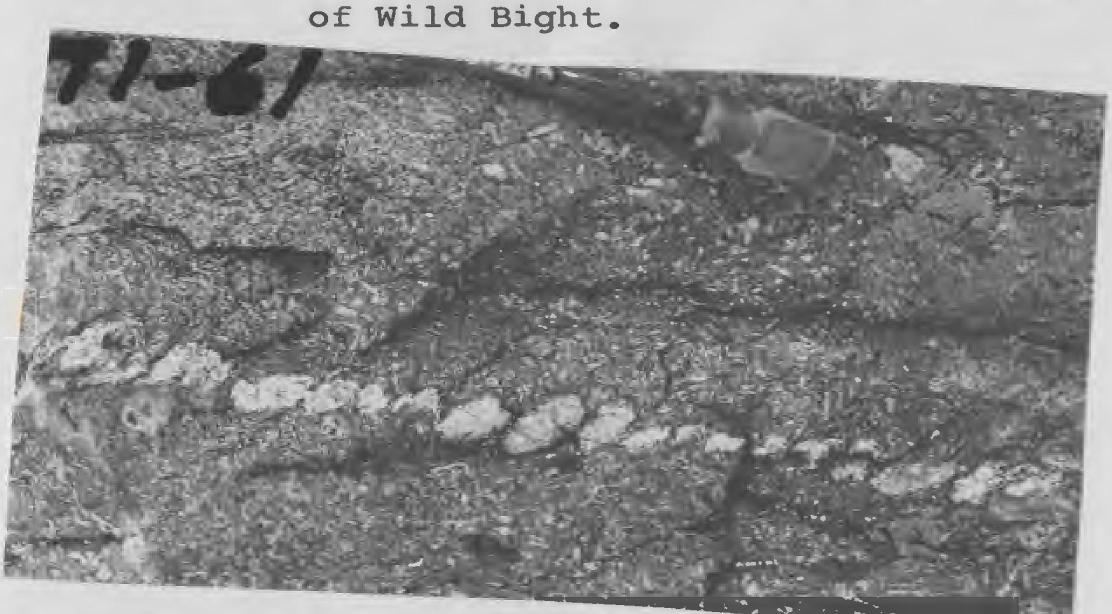


Fig. 4-53.. F_4 crenulation of S_2. Note the D_2 boudins of an early pegmatite veinlet, which are rod-like structures showing no signs of folding, have been rotated and have acted as loci for development of the F_4 folds.

Fig. 4-54. F_4 folds of bedding in the upper tuffite member of the Pomiadluk Point Formation. On hill west of October Harbour. Note: these are not chevion folds due to the interference of kink-bands.

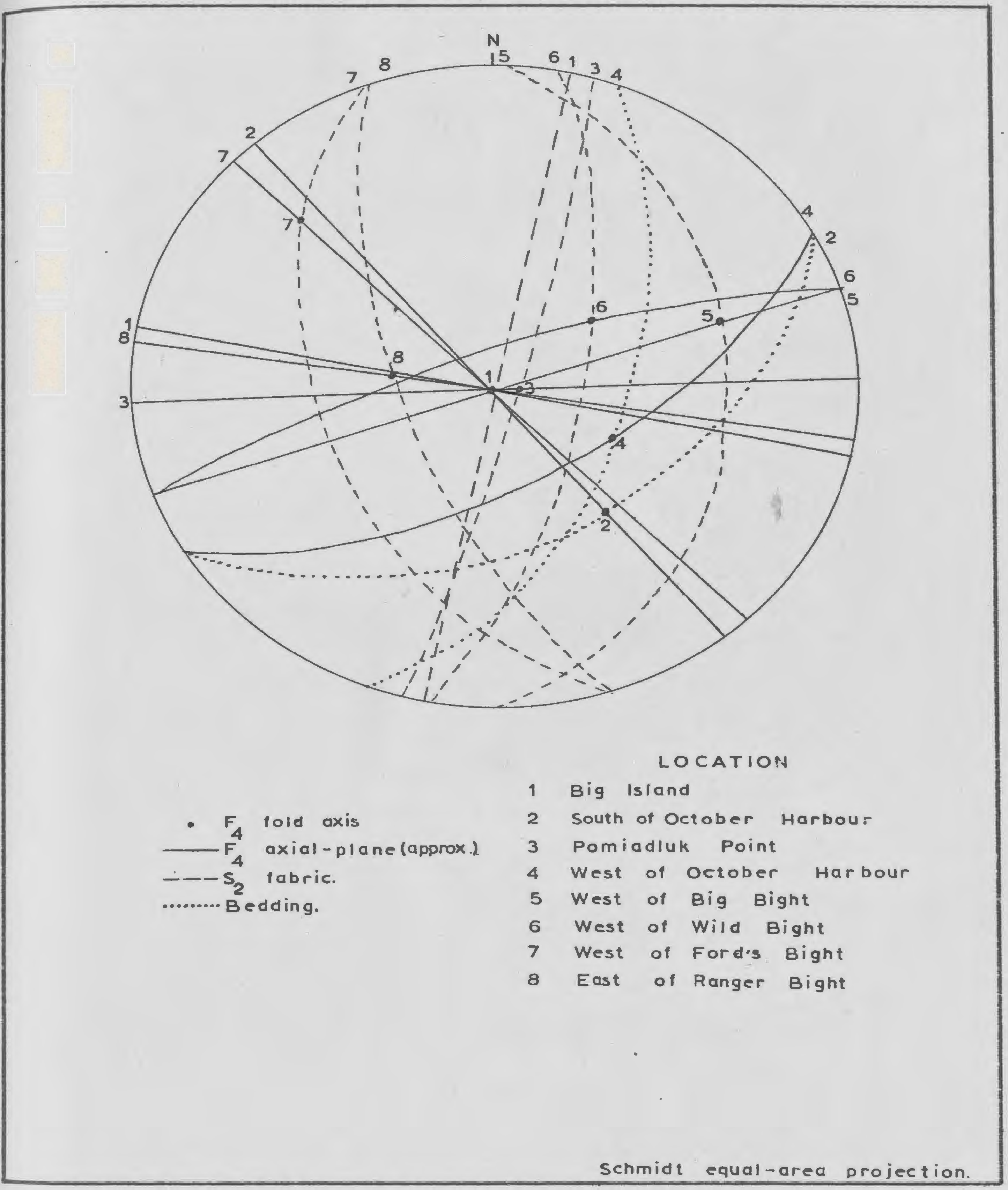


Fig. 4-55. Stereographic projection of F_4 folds and the attitude of the surface folded (S_2 or bedding) adjacent to the folds.

Fig. 3-7. Photomicrograph of the chill margin of the Monkey Hill Granite (left) in the flow-banded rhyolite of the Makkovik Formation (right) showing development of elongate quartz and feldspar grains. Crossed-nicols, X6.

Faults and Fracture-zones

Most of the faults in the area are northwesterly or northeasterly trending and appear to form a conjugate set. Their attitude is the same as the large, late faults recognised west of Makkovik Bay (Gandhi, et al., 1969; Clark, 1970 and 1971).

Several vertically dipping faults (joints) up to 2 km. (1 mi.) long but of small to no displacement (of the order of a few meters or less) occur in the area. As the displacement is small none of the faults were traced out in the field, and their length may be greater than presently recognised.

Slight displacement of the lithologies of Big Island with respect to west of Ranger Bight suggest a fault passing between the island and the point. The relatively straight nature of the northwest coastline between Cape Strawberry and west of Ranger Bight suggests the fault extends along this entire coastline and defines its location. Furthermore, a gully at the head of Long Point Cove on the western side of Makkovik Bay, and a diabase ("diorite") dyke extending south-west of Long Point Hill (Clark, 1970 and 1971), are also in alignment. If these features are related to the fault, then its total recognised length is of the order of 20 km. (10 mi.). The non-porphyritic diabase dyke which extends from Kaipokok Bay to the Adlavik Igneous Complex is probably intruded along one of this set of faults, and both the faults and the dykes appear related to similar faults and dykes in the Cape Makkovik area (King, 1963).

A large number of 20 to 50 cm. (8 to 18 in.) wide fracture zones occur throughout the area. These zones vary from en echelon trains of sigmoid tension gashes (Fig. 4-56) to faults of small displacement (displacement of the order of 30 cm. (1 ft.)). Conjugate sets indicate a northwest/ southeast maximum-stress orientation during deformation (Clark, 1970 and 1971) for some, but other less common sets also occur. The fractures of the most common set are, in all cases, open and unfilled and apparently have never been filled, though some of the other sets have quartz infilling and are, therefore, probably earlier and related to higher temperature and pressure (Fig. 4-57).

Sigmoid Joints

On the east coast of Wild Bight, an unusual sigmoid joint pattern (Figs. 4-58 and 4-59) is developed in grey diorite dykes. In cross-section the joints have the appearance of a twisted rope approximately 2 to 4 cm. (1 to 2 in.) wide and extending for a distance of 2 to 5 m. (6 to 15 ft.) before decreasing in width and finally dying out. The joint plane consists of an interconnecting set of rods with sigmoidal cross-section which gives the feature its characteristic pattern. Similar joints have been noted and described in Aillik Bay by Kranck (1961) and King (1963).

Fig. 4-56. Open, late-stage sigmoid tension gashes. In the tuff on southwest coast of Big Island.

Fig. 4-57. Quartz-filled conjugate joint-set. In conglomerate on east flank of Monkey Hill.

Fig. 4-58. Two planes of sigmoid joints in grey diabase dyke on east coast of Wild Bight.

Fig. 4-59. Sigmoid joint-set in grey diabase dyke on east coast of Wild Bight.

Possible Syn-depositional (Compactional) Folds

Complex refolded folds, producing eye- and mushroom-structures (Types 1 and 2 refolded folds; Ramsay, 1967), occur in the tuff member on the south coast of Ranger Bight (Fig. 4-60) and also 3 km. (1 mi.) south of the bight (Fig. 4-61 and 4-62). These structures are pre-D_2 as they are cut obliquely by S_2 and folded by open F_2 folds south of Ranger Bight. The structures may be due to F_1/F_2 fold interference or to interference of F_1 with pre-compactional folds, but the complete lack of any remnant pre-D_2 fabrics, and also the lack of similar structures elsewhere or in other lithologies in the area suggest they are directly related to the lithology they occur in, and are therefore assumed to be complex syn-depositional or syn-compactional slump (?) folds.

Summary of Structure

Previous work in the area has done little to define the structural history, though work outside the area (Clark, 1970 and 1971; Sutton, 1972b; Sutton *et al.*, 1971) has indicated the possible complexity of structures to be encountered. The total understanding of the complex structural evidence was confined to recognition of a few folds and faults and a penetrative S-fabric, and was probably best summarised by Gandhi, *et al.*, (1969) who stated that "the structural features point to a single cycle of orogenic deformations." However, the Aillik Group has been shown to have undergone at least four recognisable deformations, of which the second was the most

Fig. 4-60. Syn-depositional(?) interference folds in tuff on south coast of Ranger Bight. Bedding outlined on outcrop with pen.

Fig. 4-61. Sketch of complex syn-depositional (?) interference folds of bedding in Makkovik Formation tuff. Refolded about S_2.

Fig. 4-62. Syn-depositional (?) interference folds on same outcrop as Fig. 4-61.

pervasive producing many major folds and a regionally penetrative S-tectonite fabric. The S_2 fabric is subvertical throughout most of the area and the L_2 lineation is subhorizontal. The F_2 folds are upward facing and plunge from moderately southward at Manak Bay to gently northward at Ranger Bight, the variation in plunge being due to a major northwest trending F_4 fold through the center of the area, which also formed the Round Pond dome. A major D_2 slide of unknown displacement occurs to the west of the Round Pond dome, through Big Island and southwards, and separates two F_2 synclines. The amount and type of D_2 deformation has been calculated for a few locations, and is seen to be very variable. It has been shown that the entire thesis area may occur within a large simple-shear zone, but definitive evidence is, at present, lacking. One locally occurring pre-D_2 deformation is recognised, and is most extensively developed in amphibolite dykes in the vicinity of Ranger Bight and Big Island. D_1 fabrics, banding and minor boudinage are the main products of the first deformation, and their occurrence is apparently lithologically controlled. There is evidence of pre-D_1 fabrics, but the nature and extent of development is not known.

Two post-D_2 deformations are recognised: the third deformation which produced a few major folds of S_2 near Round Pond and on Big Island, but no fabrics, and the fourth deformation which caused the variable plunge of the F_2 folds, but also developed smaller folds on all scales down to microscopic crenulations of S_2 biotites.

Post-D_4 tectonic events have resulted in faults and fractures of minor displacement.

Possible depositional (sedimentary) folds have been recognised in the tuff member south of Ranger Bight.

Despite the amount of tectonism in the area, the Aillik Group remains upward facing throughout the area, except for very slight overturn (about 85^{o} dip) south of Tilt Cove and at Manak Bay.

Minor Igneous Intrusions

This section will deal with the different types of dykes and sills that have been investigated. It should be realised, however, that by no means all of the vast variety of these bodies have been recognised, especially as regards pre- and syn-tectonic amphibolites. Excluded from this section are dykes and sills definitely related to extrusive and large intrusive formations which are described in the relevant sections. The intrusions will be dealt with in order of emplacement, (or, where doubt exists, the assumed order). This order itself is discussed in Chapter 6. The complex age relationships and dyke compositions vary from simple single-composition dykes with normal intrusive contacts to complex net-veined and back-veined (Hughes, 1960) occurrences.

PRE-TECTONIC INTRUSIONS

Possible Aillik Group Intrusions

Many intrusions occur in the area which, though lithologically similar, and in many cases almost identical, to members of the Aillik Group, or of close spatial and chronological relationship to the Group, cannot be shown to be off-shoots from or feeders to them, and in many cases actually cross-cut and show chill-margins against them.

Many amphibolite dykes and sills (?) occur in the

CHAPTER 5

METAMORPHISM

Introduction

The regional metamorphism of the area is shown to have occured in the greenschist and amphibolite facies (Turner, 1968) with only local variations in grade due, for the most part, to contact metamorphic effects. However, due to a lack of facies- or grade-specific metamorphic minerals associated with the tectonite fabrics it has been difficult to determine the sequence of metamorphic mineral growth with respect to the fabrics, with any degree of confidence.

Regional Metamorphism

SYN-D_2 METAMORPHISM

The S_2 and L_2 fabrics throughout the area are defined by biotite and hornblende, and to a lesser extent by phlogopite and muscovite. Where these fabrics have developed as the result of crenulation cleavage and transposition of S_1 and L_1 fabrics, annealing (Voll, 1960) has occured resulting in unstrained grains. Evidence of syntectonic nucleation and growth of D_2 mineral species is rare but has been observed in some biotite- and amphibolite-rich dykes where S_2 biotite has developed from earlier hornblende (Fig. 4-33) or has nucleated separately (Fig. 5-1). It is therefore generally difficult to be certain that the minerals defining S_2 and L_2 were in fact in metamorphic equilibrium.

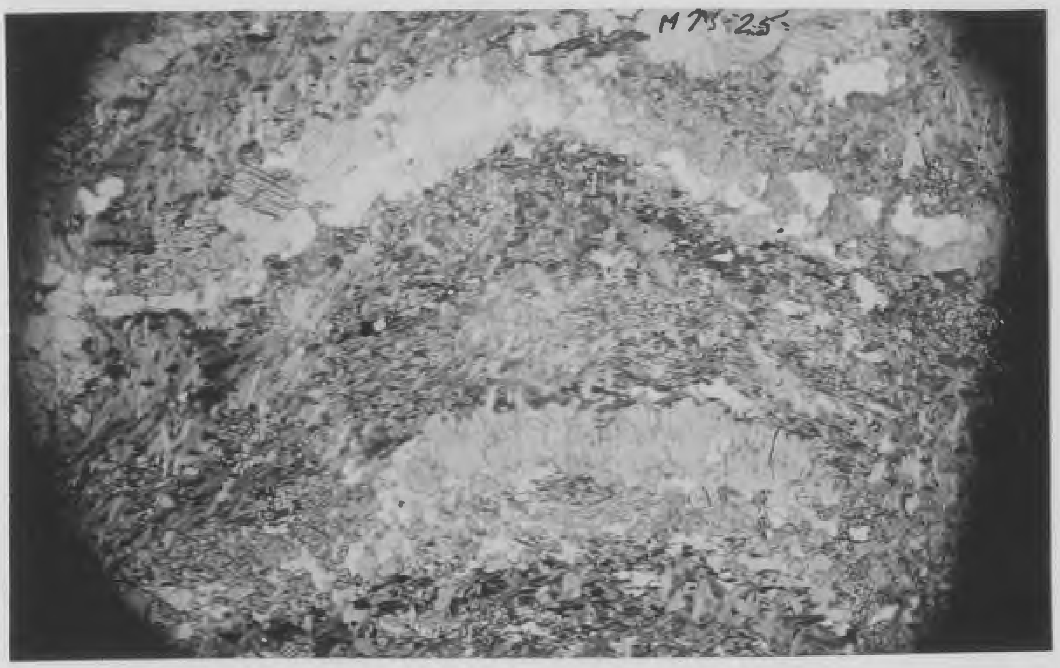


Fig. 5-1. F_2 fold of D_1 banding showing partial transposition of S_1 biotite towards S_2 orientations and overgrowth by newly nucleated S_2 biotite in S_2 orientation (vertical in photograph). Amphibolite on south coast of Big Island (Fig. 4-30). Plane polarised light, X10.

Nevertheless, as there is no evidence of crystallisation of minerals of considerably different metamorphic grade, metamorphism during D_2 is considered to have occured in the range of the greenschist and amphibolite facies. Minerals of S_2 and L_2 fabrics not obviously derived from earlier fabrics cannot be confidently assigned to either D_1 or D_2.

PRE-D_2 METAMORPHISM

Pre-D_2 metamorphism may be subdivided on mineral-growth and fabric cross-cutting relationships into post-D_1/pre-D_2 mineral growth, and syn-D_1 growth.

Post-D_1/Pre-D_2 Metamorphism

Scapolite and hornblende porphyroblasts in the Nesbit Harbour Formation metabasalt form augen within the S_2 biotite fabric and are therefore pre- or early syn-D_2 in age. Small included feldspar, biotite and hornblende grains define an imperfect earlier fabric. This earlier (S_1) fabric is generally straight, but in some porphyroblasts may show very slight sigmoidal curvature, indicating essentially post-D_1/pre-D_2 growth, but with some minor growth during the early stages of D_2 (Figs. 4-35 and 5-2). The scapolite is meionitic (Me_{75-100}), as determined from the birefringence (Deer, et al., 1970). If it is assumed chlorine metasomatism did not occur this suggests a temperature of formation of about 650°C (Hietanen, 1967), i.e., in the upper amphibolite or pyroxene hornfels facies. (It was not possible to use the plagioclase composition in the above determinations as it has undergone post-D_2 recrystallisation and separation into albitic and andesitic components).

Pre- to syn-D_2 hornblende porphyroblasts in a biotite-calcite dyke northeast of Ranger Bight have a disoriented internal fabric which develops into an oriented D_2 fabric at the edges of the porphyroblasts (Fig. 5-2), suggesting post-D_1 intrusion of the dyke (or post-D_1 annealing of an earlier dyke), and therefore post-D_1/pre-D_2 hornblende growth.

Syn-D_1 Metamorphism

Although biotite and hornblende define S_1 and L_1 fabrics in the Big Island/Ranger Bight area and at Pomiadluk Point, for reasons discussed above (the lack of evidence of syn-tectonic nucleation), these fabrics cannot confidently be assumed to have been stable at the time of formation, and therefore, although they suggest syn-D_1 metamorphism in the greenschist to amphibolite facies range, the actual range of metamorphism may have been higher or lower.

POST-D_2 METAMORPHISM

Post-D_2/Pre-D_3 Metamorphism

Quartz and feldspar throughout the Makkovik region show little or no sign of strain, and are generally thorouughly recrystallised to a polygonal fabric. This is

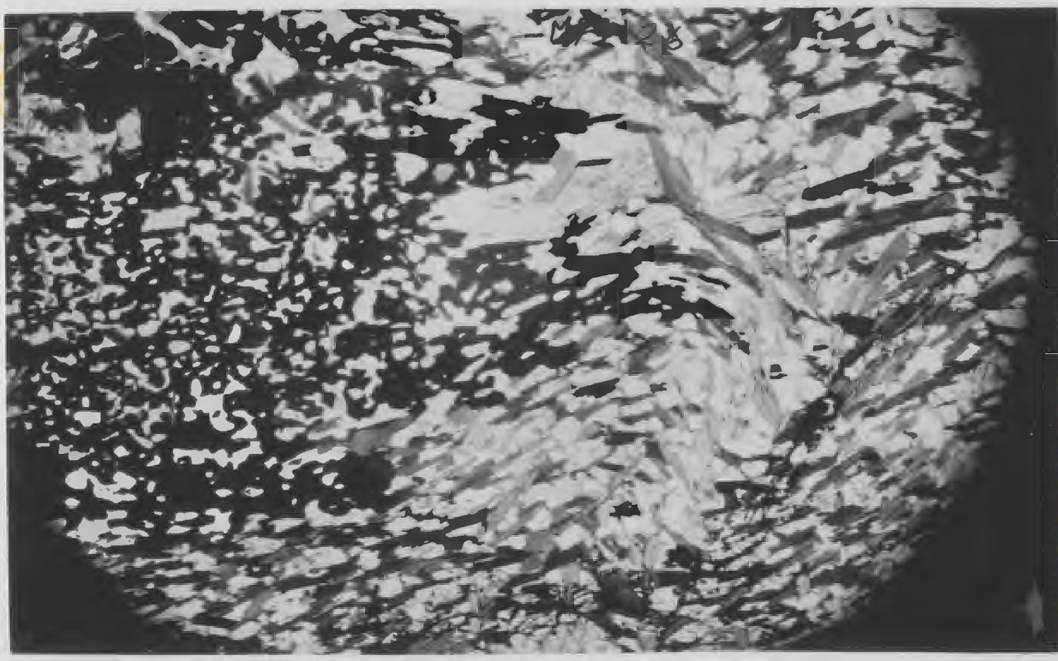


Fig. 5-2. Photomicrograph of large pre-D_2 hornblende porphyroblast (black, to left of photograph) with abundant disoriented inclusions showing development of S_2 orientation of inclusions at edges of porphyroblast. Porphyroblast and external S_2 biotite have been involved in F_4 crenulation in center of photograph. Biotite-carbonate dyke on coast north east of Ranger Bight. Plane polarised light, X10.

especially noticeable in the case of D_2 flattened quartz and feldspar phenocrysts (Figs. III-3, III-4, III-5 and III-8). The plagioclase is usually albite, and is accompanied by andesine, epidote or calcite, indicating metamorphic stabilisation and release of tectonic stresses by development of a polygonal fabric under greenschist or lower amphibolite facies conditions.

The Monkey Hill Granite, which is post-D_2/pre-D_3 in age (c.f., Chapter 6) is intruded into the flow-banded rhyolite of the Makkovik Formation. The rhyolite shows excellent post-D_2 polygonal fabrics, both adjacent to the granite, and in inclusions within it, whereas the intrusion shows only slight shadow extinction in the quartz grains (Fig. 3-7), indicating development of polygonal fabric prior to intrusion and therefore prior to D_3.

Syn-D_3 Metamorphism

Biotite defines S_3 fabrics in the metabasalt in the Nesbit Harbour Formation. However, it was not possible to determine whether the biotite was metamorphically stable during D_3. No other S_3 fabrics have been recognised.

Syn-D_4 Metamorphism

S_2 biotites have been crenulated by small F_4 folds in many biotite-rich lithologies, and have since undergone grain-boundary migration with consequent release of strain (Fig. 5-2). No minerals that nucleated during D_4 have been recognised, therefore, as with the F_2 folds, whether these

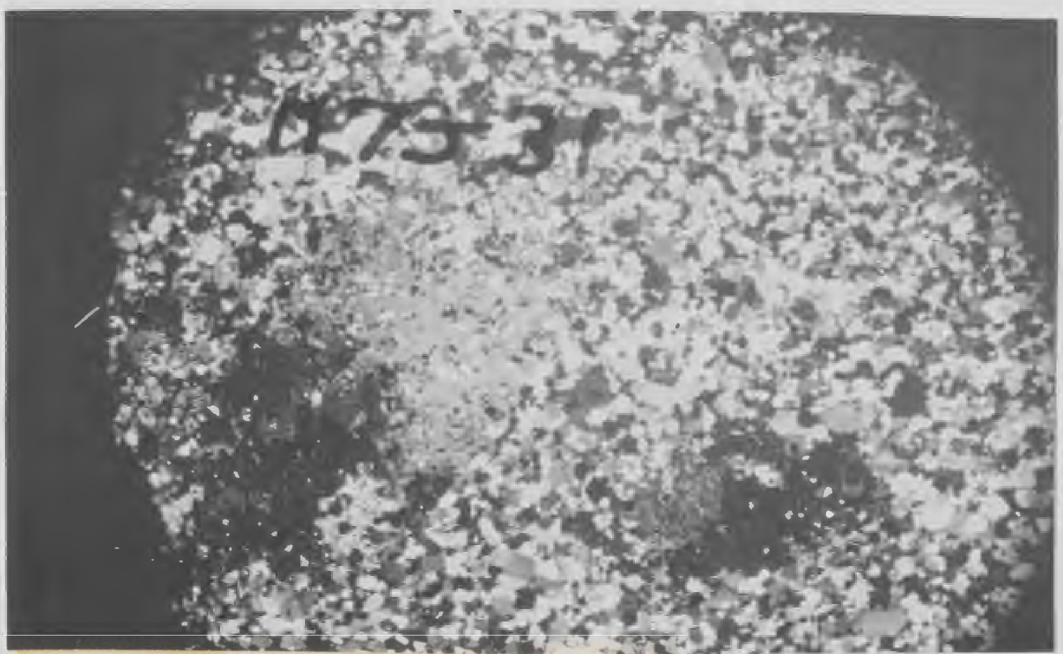


Fig. 5-3. Photomicrograph of microcline porphyroblast in the flow-banded rhyolite member of the Makkovik Formation, adjacent to the Monkey Hill Granite. Crossed nicols, X15.

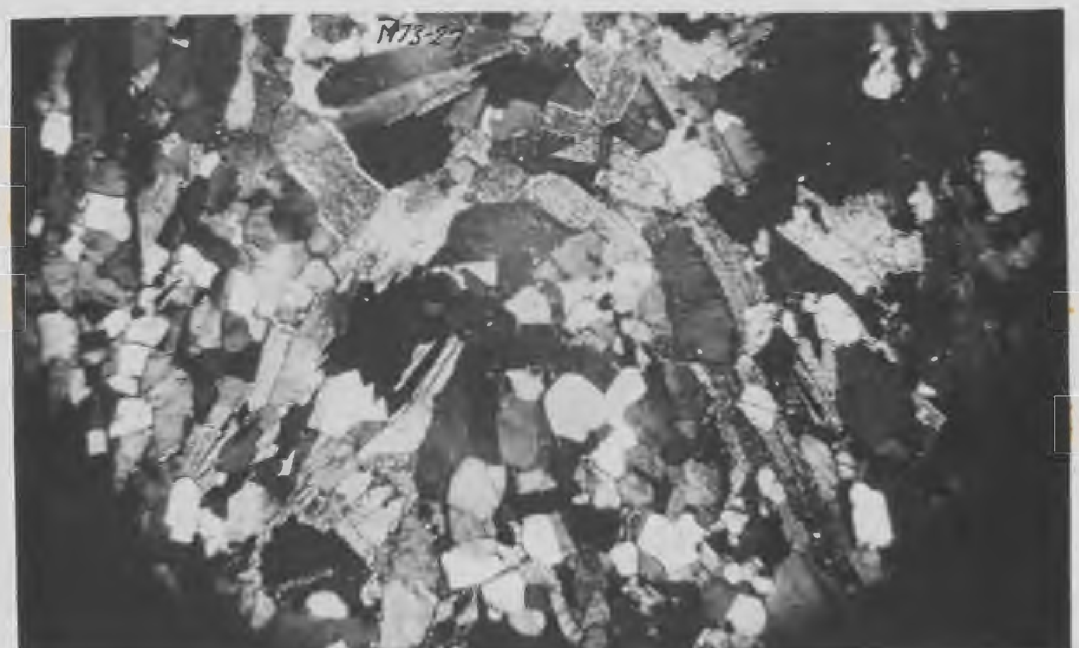


Fig. 5-4. Photomicrograph of F_4 fold showing lack of strain in biotite and feldspar grains, and imperfect polygonal fabric of the feldspar. Amphibolite dyke on west coast of Wild Bight. Crossed nicols, X40.

unstrained biotites are representative of the prevailing D_4 metamorphic conditions or not cannot be determined.

<u>Post-D_4 Metamorphism</u>

Quartz and feldspar associated with F_4 crenulations are unstrained and show polygonal fabric indicating post-D_4 recrystallisation (Fig. 5-4).

<u>Other Post-D_2 Metamorphism</u>

A few green, fine- to medium-grained, epidotitic masses varying in size from 5 mm. (1/4 in.) lenses and bands (Fig. 5-5) to 50 cm. (1 1/2 ft.) diameter spherical bodies (Fig. 5-6) occur in the pre-D_2 mafic (amphibolitic) lava and dykes. These pods have a pale, bleached border-zone surrounding a core of epidote with minor calcite, in which D_2 fabrics are imperfectly preserved. An outer rim of scapolite around the bleached zone is developed in some occurrences. The bleached zone consists of highly saussuritized plagioclase, minor epidote, and chlorite. Hornblende (and associated chlorite) define the weak S_2 fabric which passes directly through the pods with no development of augen. This lack of augen development and the unflattened shape of the pods indicate a post-D_2 age. Where bands or lenses of epidote occur the shape is apparently due to structural control, such as fractures (Fig. 5-5). A similar occurrence in Australia (Smith, 1968) is interpreted to be due to ionic migration during burial metamorphism in the prehnite-pumpellyite facies. A similar origin under higher

Fig. 5-5. Pre-D_2 amphibolite dyke with S_2 oriented feldspar phenocrysts (parallel to pencil) and post-D_2 epidote blebs and stringers. East coast of Wild Bight.

metamorphic conditions (greenschist-amphibolite facies) is the probable cause of these structures. Albite porphyroblasts have been recognised in tuffs in the porphyritic rhyolite between Ford's Bight and Wild Bight, and in the tuff south of Makkovik (Fig. 5-9) and indicate local greenschist facies conditions. The albite is unstrained and contains inclusions of the polygonal fabric indicating growth after development of the polygonal fabric.

Post-D_2 actinolite and anthophyllite occur in amphibolite rocks of the Big Island Formation. Late development of chlorite from muscovite and biotite is ubiquitous.

Contact Metamorphism

SYN-D_2 METAMORPHISM

The occurrence of wollastonite and anthophyllite in an imperfect S_2 orientation in arkoses and amphibolites, respectively, of the Big Island Formation, suggests local high temperatures and may be related to intrusion of the nearby Long Island Gneiss.

POST-D_2 METAMORPHISM

Post-D_2/Pre-D_3 Metamorphism

Microcline porphyroblasts occur in the vicinity of the Monkey Hill Granite (Fig. 5-3). Inclusions in the porphyroblasts show an earlier polygonal fabric (the post-D_2 polygonal fabric) indicating the porphyroblasts grew after vicinity of the Nesbit Harbour Formation metabasalt, and are especially notable on Monkey Hill. Most of these are probably directly related, genetically, to the metabasalt.

Rhyolite dykes and sills also occur. These dykes and sills are most commonly seen in the vicinity of Wild Bight, but also occur elsewhere. The dykes closely resemble the microcline-porphyritic rhyolites in being grey coloured with large (up to 1 cm. (1/2 in.) diameter) pink microcline phenocrysts evenly distributed throughout. However, quartz-porphyritic and intermediate types are also represented, though these are usually pink to orange in colour and may be so crowded with phenocrysts as to appear granitic in hand-specimen. The dykes are usually subvertical and therefore sub-parallel to the local bedding, and are in places intruded along the center of amphibolite dykes (Fig. 3-8), commonly accurately bisecting it for several hundred meters before a trangression occurs. There is no evidence of mixing of magmas at the contact though tectonically produced lobate structures are common (Fig. 4-24), and xenoliths of amphibolite from the basic wall-rock in the acid dyke show no reaction rims or other evidence of not being solid at the time of incorporation. They are also commonly seen to offset the contact with the amphibolite dyke (Fig. 3-9). The basic dyke is, therefore, considered to be a much earlier and genetically unrelated body, and the association of the two types of dyke to be due to preferential development of

Fig. 5-6. Epidote-rich pods (green) with scapolitic rim (white) in mafic lava. South of Monkey Hill.

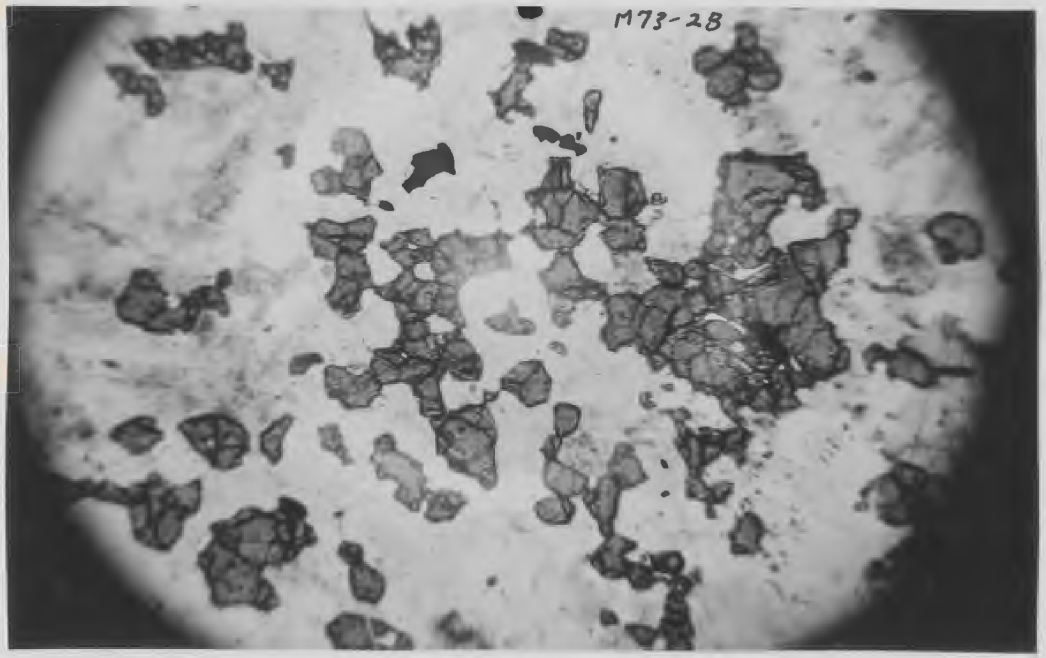


Fig. 5-7. Photomicrograph of garnet which has nucleated and grown on quartz and feldspar grain boundaries after polygonisation (post-D_2). Flow-banded rhyolite member of the Makkovik Formation. North of Monkey Hill. Nicols half-crossed, X40.

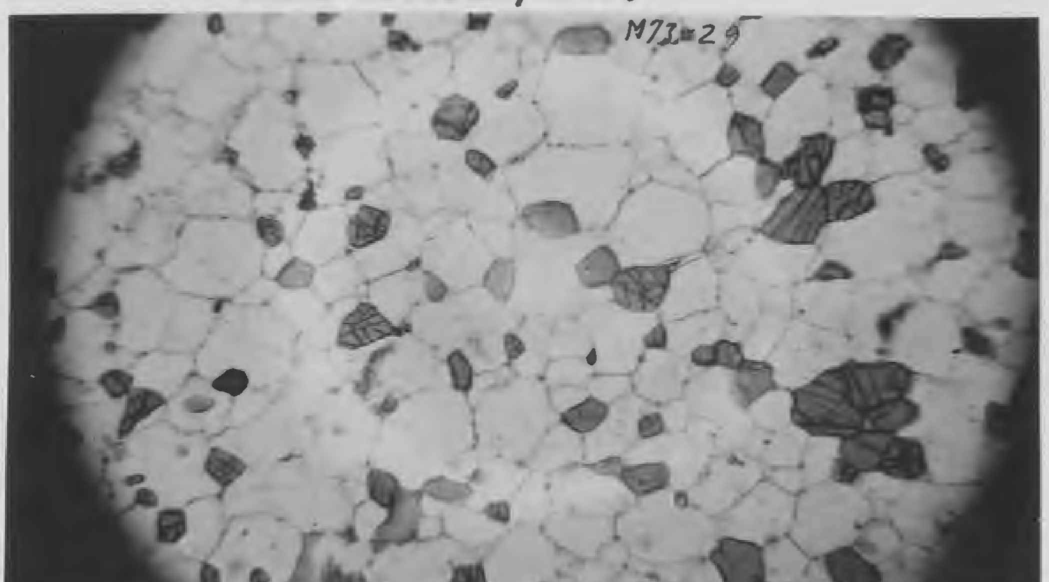


Fig. 5-8. Photomicrograph of post-D_2 diopside grains nucleated and grown at quartz and feldspar triple junctions. Lower arkose member of the Nesbit Harbour Formation. Northeast of Monkey Hill. Plane polarised light, X40.

development of this fabric, probably as a result of the rise in temperature and fluid pressure on intrusion of the granite.

Unoriented post-D_2 wollastonite in the Manak Bay arkose probably grew during the high prevailing temperatures in the vicinity of the syn- to post-D_2 Adlavik Igneous Complex (Chapter 6).

Garnet and diopside occur in the southern part of the region, and have grown after development of the polygonal fabric as indicated by their location and growth from triple-point junctions (Figs. 5-7 and 5-8). Together they indicate amphibolite facies conditions, and probably locally high temperatures related to intrusion of the Monkey Hill Granite and Adlavik Igneous Complex. Andalusite with polygonal fabric occurs in a rhyolite at Pomiadluk Point and may also indicate local post-D_2 attainment of pyroxene hornfels facies conditions.

Hypersthene in the Grampus Cove Gneiss at the border of the diabase net-vein dyke is presumably the result of late- and post-D_2 heating of the area by the Adlavik Igneous Complex, enhanced by the intrusion of the diabase and melting of the gneiss by the dyke.

Summary of Metamorphism

The regional metamorphic conditions during tectonism have been shown to be in the greenschist and amphibolite facies, with only local variations in grade (Tables 5-I and 5-II). No evidence has been recognised for major

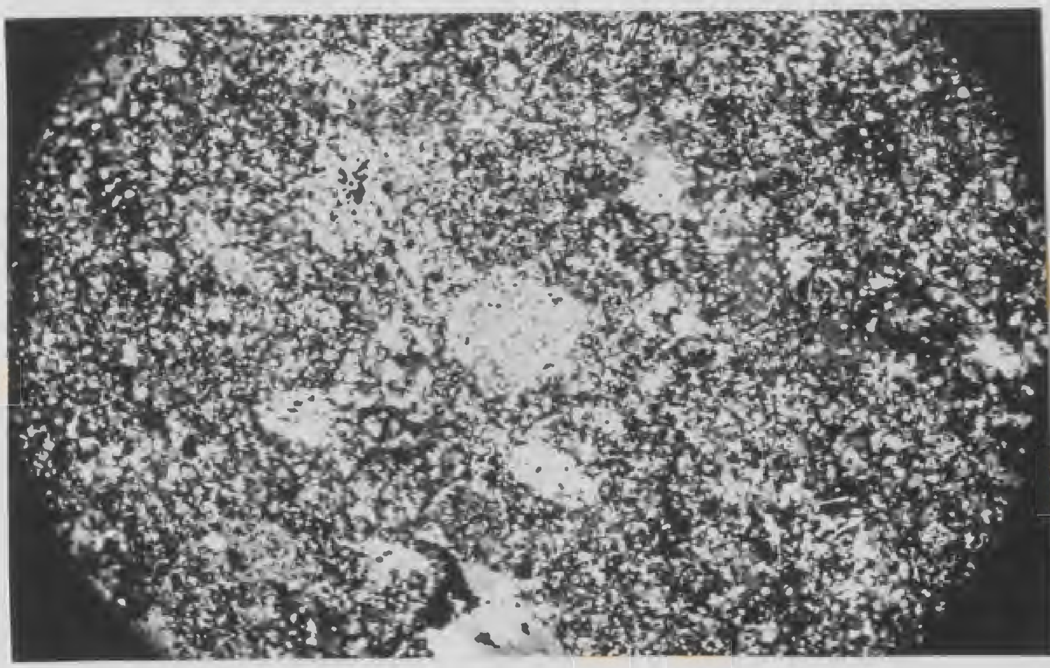


Fig. 5-9. Photomicrograph of albite porphyroblasts in tuff member of the Makkovik Formation. Southwest of Makkovik. Crossed nicols, X10.

metasomatism or for complete gradation from sediments to granite as suggested by Stevenson (1970). In fact, all granite (s.l.) contacts are clearly intrusive, usually with well developed chill margins, and only very minor growth of microcline, garnet(?) and diopside(?) in the enclosing sediments, indicating a lack of significant metasomatism. The presence of minor scapolite possibly indicates local chlorine metasomatism.

Contact metamorphic effects resulting in the growth of higher temperature mineral assemblages have been recognised in the vicinity of the Long Island Gneiss, the Adlavik Igneous Complex and the Monkey Hill Granite.

Table 5-I

Table showing stages of metamorphic mineral growth

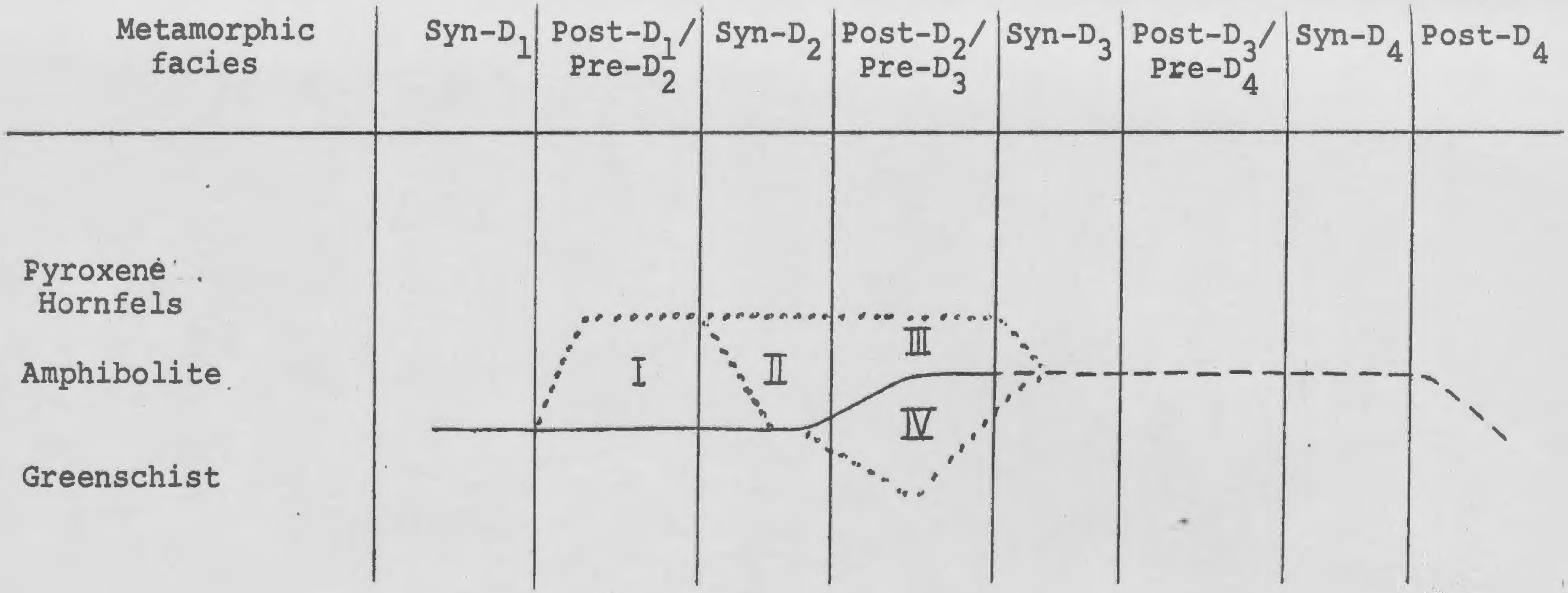
•• regional mineral growth or stability
? possibly stable
□ local growth only
--- range of uncertainty of growth (symbol in position of most likely growth).

Mineral Species	Syn-D_1	Post-D_1/ Pre-D_2	Syn-D_2	Post-D_2/ Pre-D_3	Syn-D_3	Post-D_3/ Pre-D_4	Syn-D_4	Post-D_4
Microcline				□ GrM				
Albite and Andesine (or epidote) together				•				•
Albite porphyroblasts				□ ---	---	---	---	---
Hornblende	?	•—	—•					
Biotite or phlogopite	?	?	•		?		?	
Scapolite		□—	—□	•---	---	---	---	---•
Wollastonite			□ LIgn	□ AIC				
Garnet				□ GrM AIC ---	---	---	---	---
Diopside				•---	---	---	---	---
Actinolite	---	---□		□---	---	---	---	---
Anthophyllite				□---	---	---	---	---
Chlorite				□---	---	---	---	---
Static Annealing (Polygonisation)				•				•

GrM Mineral growth in vicinity of Monkey Hill Granite
AIC Mineral growth in vicinity of Adlavik Igneous Complex
LIgn Mineral growth in vicinity of Long Island Gneiss

Table 5-11

Table showing grade of metamorphism
reached during and after tectonism



——— Approximate grade of metamorphism--known

------ Approximate grade of metamorphism--assumed

······· Local conditions indicated by:

- I Scapolite
- II Wollastonite
- III Wollastonite, garnet and/or anthopyllite
- IV Actinolite

CHAPTER 6

RELATIVE AGES OF IGNEOUS INTRUSIONS

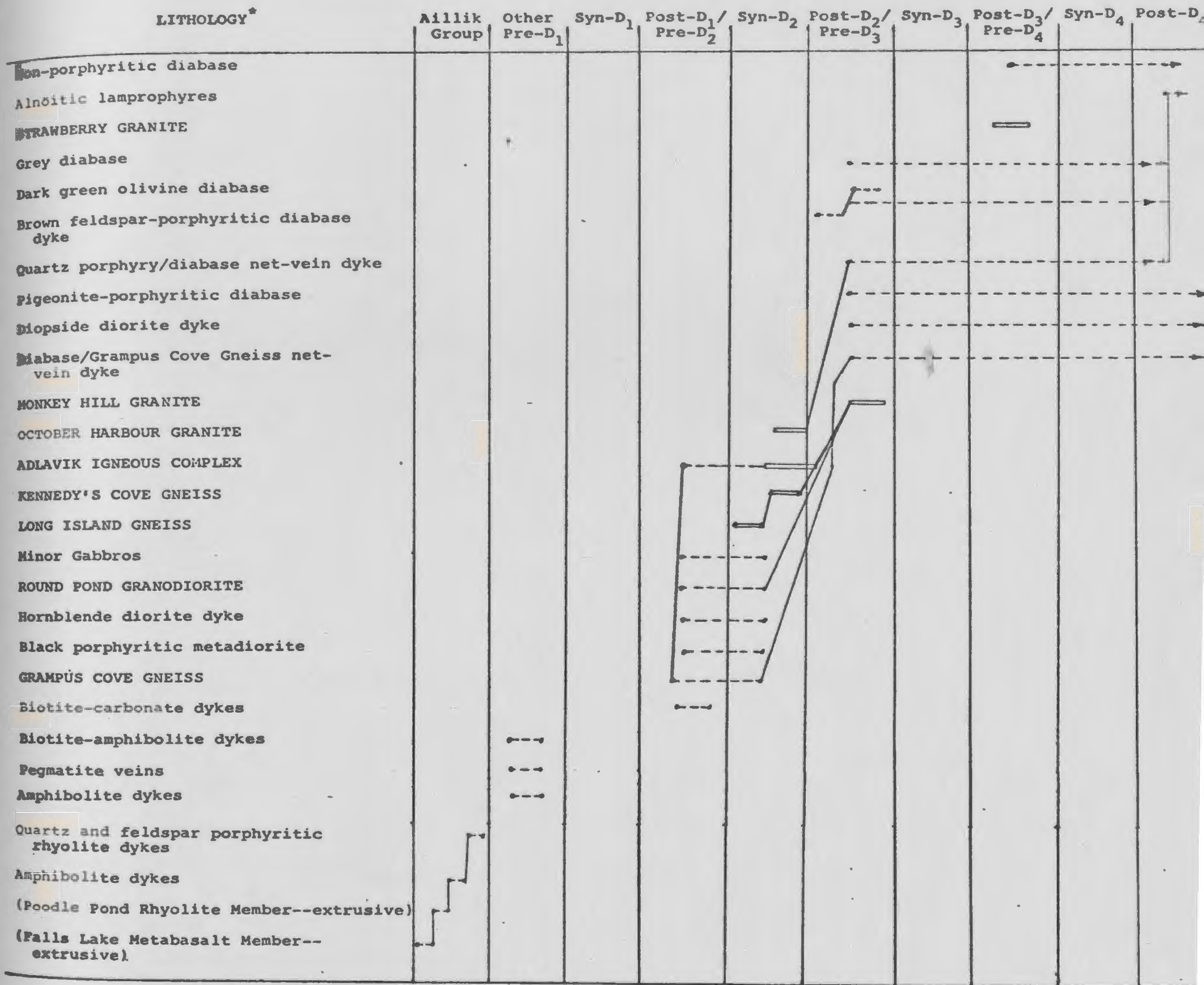
Introduction

The Aillik Group has been intruded by numerous dykes and sills related to the acid and basic extrusive formations within it, as well as by a few other pre-D_1 amphibolite dykes and pegmatite veins, and by many other pre-D_2 biotite-carbonate, amphibolite and pegmatite dykes and veins (Table 6-I). All the major granitic intrusions except the Monkey Hill Granite and the Strawberry Granite are pre- or syn-D_2 in age, as is the Adlavik Igneous Complex. The Monkey Hill Granite is post-D_2/pre-D_3 and the Strawberry Gramite is post-D_2/pre-D_4. Many post-D_2 diabase dykes of different relative ages occur. The youngest intrusions in the area are the lamprophyre dykes.

The relative ages of the vaiious major and minor igneous intrusions have been determined from both direct intrusive relationships and from comparisons with the tectonic deformational sequence. However, only the second deformation (the main penetrative deformation) is of broad enough occurrence to enable relative age correlations to be made throughout the area. The other deformations are only locally developed, though commonly at several different localities, and are consequently only locally relevant. For this reason most of the intrusive bodies can only be

Table 6-I

Table showing the relative age of the igneous intrusions in the thesis area



relevant intrusive (cross-cutting) relationships

uncertainty in intrusive age

range of intrusive age

*Major intrusives in Capitals, minor intrusives in lower-case confirmed as pre- or post-D_2 on the structural evidence, and finer definition of relative ages has to be made on the basis of intrusive relationships, where available.

Pre-D_2 Intrusions

PRE-D_1 INTRUSIONS

Aillik Group Intrusions

Several of the volcanic members of the Aillik Group have associated dykes, sills and veins. This is especially marked in the case of the mafic volcanic rocks, where many of the spatially and temporally associated amphibolite dykes are presumably feeder-conduits. Most of the pre-D_2 amphibolite dykes in the area are probably also related to the Aillik extrusives but have been completely recrystallised during the second deformation so that earlier fabrics are no longer seen. As regards the acid volcanic rocks, no dykes that could be related to the flow-banded rhyolites have been recognised, though quartz-porphyritic dykes are common especially in the vicinity of Wild Bight, and microcline-porphyritic dykes also occur, though are far less common. No dykes of plagioclase-porphyritic rhyolite have been recognised. The relative age relationships of the various Aillik Group igneous rocks is extremely complicated due to the broadly synchronous intrusion and extrusion of acid and basic volcanics throughout the

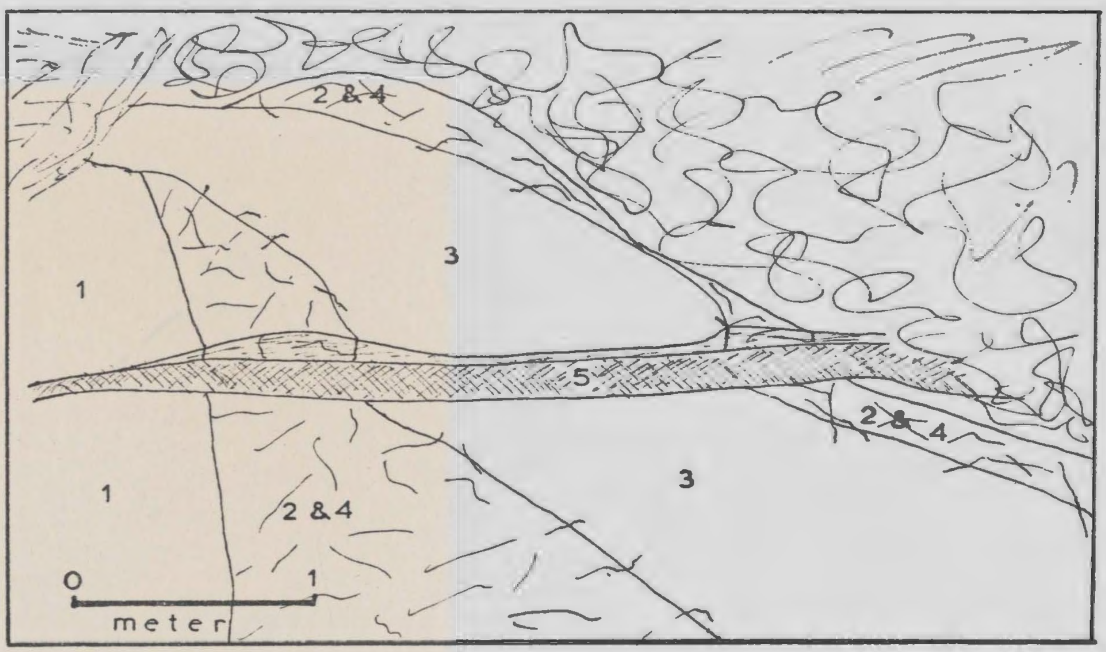


Fig. 3-8. Amphibolite (metabasalt?) dyke (2) in quartz-porphyritic rhyolite member (1) of the Makkovik Formation. The amphibolite dyke is cut longitudinally by a quartz-porphyritic rhyolite dyke (3) which is itself cut by molybdenite-bearing diopside veins (4). The entire complex is cut by a grey diabase dyke (5). West coast of Wild Bight.

period of deposition of the Aillik Group. This complexity is shown by the occurrence of the rhyolite member below the metabasalt of the Nesbit Harbour Formation, which is in turn overlain by the main acid volcanic lithologies of the Makkovik Formation. These are intruded by mafic (amphibolite) dykes, which are themselves cut by quartz- and feldspar-porphyritic dykes (Figs. 3-8 and 3-9) thought to be genetically related to the acid volcanics.

Other Pre-D_1 Intrusions

The oldest recognised igneous intrusion in the area is a single pre-D_1 amphibolite dyke on the south coast of Big Island (Figs. 4-30 and 4-42), which contains a pre-D_1 tectonite fabric. The Ranger Bight Complex probably contains many unrecognised pre-D_1 amphibolite dykes as the nature of the different amphibolites comprising the complex has not been determined. A few pre-D_1 pegmatite dykes and veins (Fig. 4-31), with D_1 boudins folded by F_4 folds (Fig. 4-40), have been recognised. Several biotite-rich amphibolite dykes, some with diopside bands, show S_1 fabrics and are therefore pre-D_1 (Figs. 4-31 and 4-32).

POST-D_1/PRE-D_2 INTRUSIONS

The biotite-carbonate dykes show well developed S_2 biotite fabrics. Pre- to early syn-D_2 hornblende porphyroblasts, the included fabric of which is random in the center tending to S_2-oriented at the edges of the porphyroblasts (Fig. 5-2) have been recognised and some of the smaller hornblende grains still have remnant cores of (original igneous?) pyroxene. The dykes are therefore probably post-D_1/pre-D_2 in age.

OTHER PRE-D_2 INTRUSIONS

Many pre-D_2 amphibolite dykes occur throughout the area, but their similarity to one another and lack of diagnostic mineralogy or textures precludes useful relative separation. Where intrusive relationships are recognised, they often cannot be extended to other intrusions in the vicinity. Pre-D_2 pegmatites with no recognisable relationship to D_1 are also common.

The Round Pond Granodiorite has imperfectly developed S_2 gneissic foliation and is therefore pre- or syn-D_2 in age. The contact with the Aillik Group was not seen, but the lack of complex deformational structures in the vicinity of the contact suggests it is not a basement lithology as such complex structures are ubiquitous where the basement/cover boundary is seen (Sutton, et al., 1971). Also, the simplicity of the S_2 fabric suggests no earlier deformational fabrics were developed, and the granodiorite is therefore considered to be post-D_1. Similarly, the Grampus Cove Gneiss, hornblende diorite dyke and black porphyritic metadiorite all show S_2 fabrics, but no earlier fabrics, and are therefore considered post-D_1/pre- or syn-D_2 in age.

Syn-D_2 Intrusions

The Long Island Gneiss generally shows well developed S_2 gneissic foliation throughout, though in some places, away from the contact, the rock has an undeformed igneous texture. The gneissic fabric indicates the intrusion is pre- or syn-D_2 in age. There is no indication of an earlier (D_1) fabric, suggesting a post-D_1 age for the gneiss. Wollastonite occurs in S_2 orientation in the arkoses of the Big Island Formation, and as wollastonite is indicative of fairly high temperatures, (Turner, 1968), it may indicate a syn-D_2 age for the nearby Long Island Gneiss.

The Kennedy's Cove Gneiss cuts the Long Island Gneiss, but also shows a thoroughly penetrative S_2 gneissic foliation, and is therefore also syn-D_2 in age.

The October Harbour Granite has S_2 fabrics in the margins, and weakly developed syn-intrusive S_2 fabrics (see below) elsewhere in the intrusion indicating emplacement during the second deformation.

The chilled margin facies of the Adlavik Igneous Complex shows well developed S_2 fabrics indicating a pre- or syn-D_2 emplacement of the intrusion though the fabric in the country-rock also tends to form an augen around the body (c.f. Mason, 1971). The diorite facies shows poorly developed D_2 gneissic fabric north-west of Big Bight, and very weak D_2 (?) fracturing and initial development of gneissic fabric west of Big Bight.

The host rhyolite shows well developed S_2 fabric of flattened quartz phenocrysts, but undeformed contacts, suggesting a late- to post-D_2 emplacement of the diorite. The Adlavik Igneous Complex as a whole, therefore, appears to have been a pre- or syn- to post-D_2 igneous event.

The late crystallisation of parts of the complex, and the concommitant retention of heat in the surrounding rocks after the end of D_2, is probably the reason for the poor development or lack of S_2 fabrics, and the widespread occurrence of decussate texture in the Manak Bay arkose, and also the growth of unoriented wollastonite in the member.

Both the small gabbro body at Pomiadluk Point and the gabbro between Wild Bight and Ford's Bight have S_2 fabrics, indicating a pre- or syn-D_2 age.

THE PROBLEM OF THE SYN-PLUTONIC FABRIC

Some of the intrusions show different types of tectonite fabrics, though they are all syn-deformational intrusions. The origin of these different fabric-types is discussed and is related to emplacement at different stages during the deformation.

The tectonite fabrics developed in the intrusions emplaced during deformation are of two types:

1. A preferred orientation of acicular or platy minerals combined with flattening and recrystallisation of the larger grains. This is developed in the margins of all the syn-tectonic intrusions, and is also developed on a coarser scale throughout the Kennedy's Cove Gneiss. This texture is identical to that in the enclosing sediments and rhyolites.

2. A preferred orientation of larger (especially

feldspar) crystals in an apparently undeformed intrusion in which the feldspars form part of the overall igneous texture of the rock. The orientation is parallel to an external country-rock S-tectonite fabric and is commonly at high angles to the contacts of the intrusion. This is identical in appearance to normal "flow-foliation" except that it is not related to the attitude of the contact of the intrusion, but is related to the attitude of external tectonic fabrics (Berger and Pitcher, 1970). This type of fabric has been recognised in the October Harbour Granite, where it grades into a "type 1" fabric at the margins of the intrusion.

Although both fabrics are obviously related to tectonism (in this case the second deformation) the factors leading to development of these different fabrics is not fully understood.

It would appear that the "type 2" fabric was developed as a result of deformation of the granite while it was in a "crystal-mush" state (Berger and Pitcher, 1970). However, the deformation must have ended by the time the granite was solid enough to transmit the prevailing differential stresses (except for the margins which cooled earlier and therefore were still being deformed after they were solid--hence the "type 1" fabric). The October Harbour Granite is therefore probably late syn-tectonic.

The "type 1" fabric is analogous to the fabrics in the enclosing sediments and is apparently due to the recrystallisation of deformed minerals (under stress) resulting in many smaller grains in place of one larger grain, and the concommitant release of internal (crystal-lattice) stress (Spry, 1969). This fabric-type, therefore, implies effective solidity of the rock to the extent that differential stress could be applied and maintained, i.e., the Kennedy's Cove Gneiss is probably early or middle syn-tectonic.

The fabrics of the Long Island Gneiss are the most difficult to explain. The gneiss was intruded prior to the Kennedy's Cove Gneiss (Gandhi, et al., 1969; Clark, 1970 and 1971), shows D_2 tectonite fabrics, and apparently caused wollastonite-growth in S_2 orientation in the country-rocks, all of which indicate a syn-D_2 or pre- and syn-D_2 intrusive age. However, although a gneissic foliation ("type 1") is developed in the margins, in some parts away from the margins no tectonic fabrics are developed at all, suggesting post-tectonic solidification--possibly due to fluid-rich "wet" spots in the magma or to continuing intrusion over a longer period of time than the Kennedy's Cove Gneiss. However, the occurrence of fluid-rich parts to the magma should be shown by granophyric intergrowth and pegmatitic phases (Hughes, 1971) which have not been seen. However, epidote is in greater abundance (up to 5%) in the undeformed parts of the intrusion than in the more gneissic parts, possibly as a result of retention of lower grade metamorphic conditions as a result of increased fluid pressure, particularly as metamorphic conditions began to rise at the end of D_2 (Chapter 5 and Clark, 1970).

The interpretation of a longer period of intrusion than the Kennedy's Cove Gneiss suggests the possibility of the Long Island Gneiss cutting the Kennedy's Cove Gneiss in places, which has also not been recognised.

Post-D_2 Intrusions

POST-D_2/PRE-D_3 INTRUSIONS

The Monkey Hill Granite shows no S_2 fabric, and chill margins cross-cut S_2 fabrics in the surrounding rocks. However, granitic and pegmatitic veins related to the intrusion in the vicinity of the contact south of Monkey Hill have been ptygmatically folded by a maximum strain with an approximately east/west orientation. Also the plagioclase grains of the veins show crush boundaries and slight bending and kinking of twin lamellae, and quartz has strain-extinction, all indicating post-emplacement strain. The relative age and orientation of the stresses indicate they were D_3, and the Monkey Hill Granite is therefore a post-D_2/pre-D_3 intrusion.

POST-D_3/PRE-D_4 INTRUSIONS

The Strawberry Granite has no tectonic fabrics, and there is no evidence of either F_2 or F_3 folding of associated pegmatite and granite veins. The granite is therefore post-D_3 in age. However, on the west coast of

Wild Bight, small quartz and pegmatite veinlets, which increase in number and size towards the granite and are therefore assumed to be derived from it, are intruded into pre-D_2 amphibolites, and both the amphibolites (S_2 fabric) and the veinlets show F_4 folding (Fig. 6-1) indicating a pre-D_4 age for the granite.

OTHER POST-D_2/PRE-D_4 INTRUSIONS

The metamorphic recrystallisation of the feldspar in the pigeonite-porphyritic diabase, but lack of any D_2 tectonite fabric, suggests it was intruded after D_2 but before D_4.

POST-D_4 INTRUSIONS

The non-porphyritic diabase dyke (Plate 1) is assumed to be post-D_4 as it shows no folding or deviation in orientation throughout its 20 km. (10 mi.) length.

The lamprophyres are post-D_4 as shown by cross-cutting relationships with F_4 (Fig. 6-2). A small lamprophyre dyke occurs in the Ford's Bight Conglomerate (Jurassic) on the east coast of Ford's Bight. It could not be determined from the field evidence whether the dyke intrudes the conglomerate or was an erosional remnant (a wall-like protruberance--King, 1972, pers comm.) around which the conglomerate was deposited. The occurrence of numerous lamprophyre pebbles in the conglomerate of identical appearance to the dyke, and the fact that all the isotopically age-dated lamprophyres in the region are Cambrian or

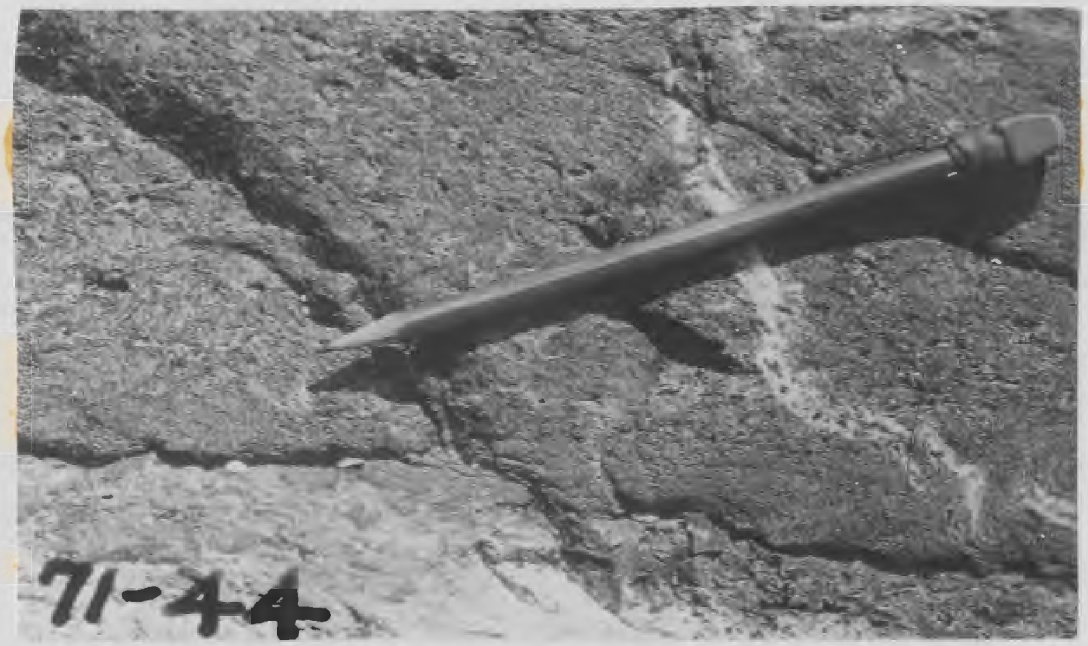


Fig. 6-1. Small F_4 crenulations of a pegmatite veinlet related to the Strawberry Granite. West coast of Wild Bight.

Fig. 6-2. F_4 folds of S_2 fabric in an amphibolite dyke cross-cut by a late carbonatitic lamprophyre dyke. West coast of Wild Bight. Plane polarised light, X15.

earlier in age (Gandhi, et al., 1968; Grasty, et al., 1969), suggest the dyke is older than the enclosing conglomerate. However, though none of the lamprophyre pebbles in the conglomerate show chill margins, the dyke does show an excellent chill margin (Fig. 2-44), with no evidence of it being weathered or eroded. Furthermore, the edge of the dyke is highly convoluted and commonly surrounds and encloses pebbles of the conglomerate (Fig. 6-3). These factors suggest the lamprophyre may well be later than the conglomerate and intrusive into it. It would then be the only young (Jurassic or younger) lamprophyre recognised on the Coast of Labrador.

OTHER POST-D_2 INTRUSIONS

The diabase/Grampus Cove Gneiss net-vein dyke is post-D_2 as the S_2 gneissic foliation in the host-rock in the vicinity of the dyke has been obliterated due to melting by the dyke. Its relationship to other deformations and intrusions is unknown but it is probably early post-D_2 as discussed in Chapter 5, (Metamorphism). The quartz-porphyry/diabase net-vein dyke is also post-D_2 as it has also truncated S_2 fabrics, but is cut by a later lamprophyre dyke.

A brown feldspar-porphyritic diabase dyke transects the S_2-fabric in the quartz-porphyritic rhyolite at Wild Bight, and is therefore post-D_2, and is cut by a dark-green olivine diabase dyke. Another dark-green olivine diabase

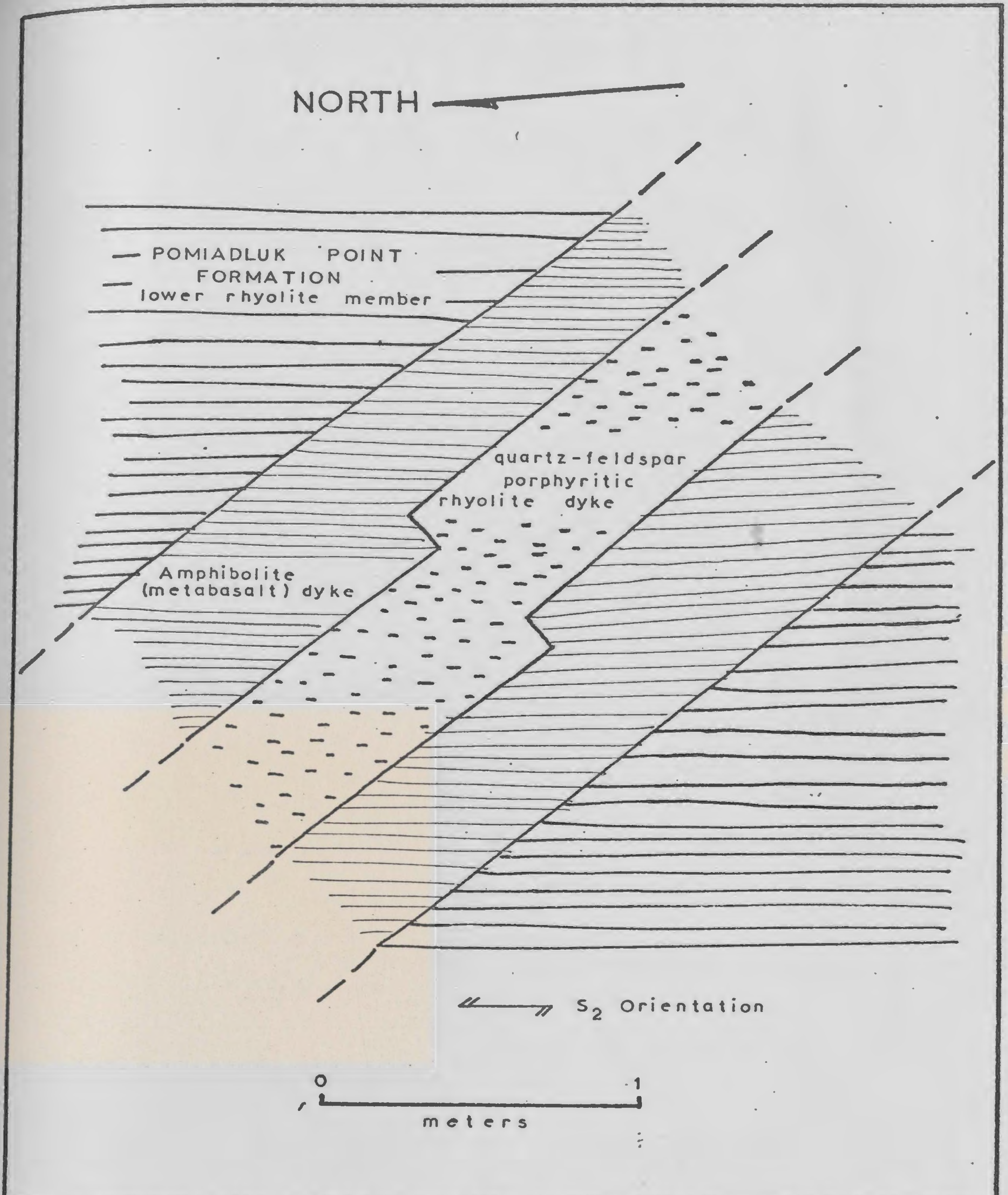


Fig. 3-9. Sketch showing off-set walls of later dyke of quartz-porphyritic rhyolite situated longitudinally in an earlier amphibolite (metabasalt?) dyke. East coast of Wild Bight.

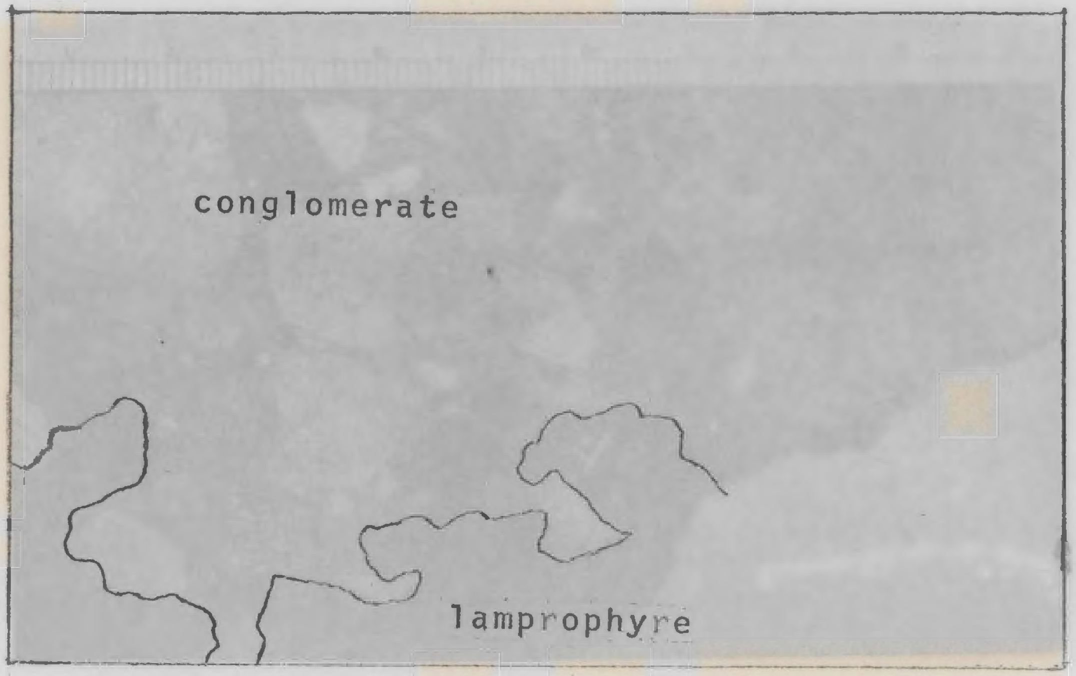


Fig. 6-3. Relationship of the contact of the lamprophyre dyke to the Ford's Bight Conglomerate. Scale in centimeters.

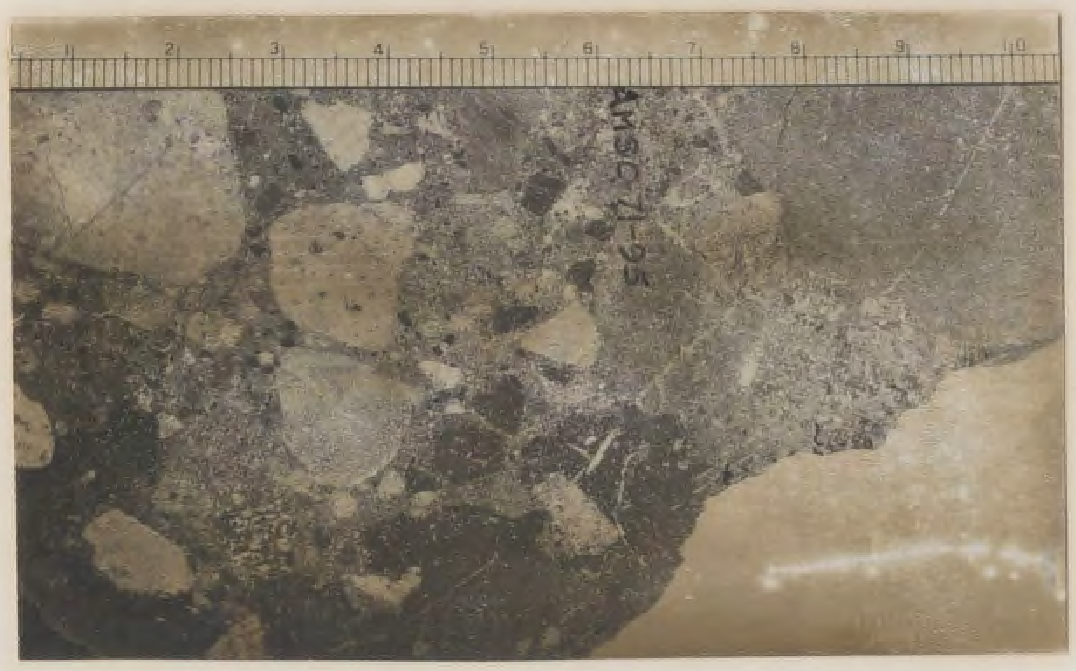


Fig. 6-3. Relationship of the contact of the lamprophyre dyke to the Ford's Bight Conglomerate. Scale in centimeters.

dyke is cut by a lamprophyre elsewhere in the area.

A grey diabase dyke cuts molybdenum-bearing diopside veins on the west coast of Wild Bight, and another grey diabase dyke is cut by lamprophyre on the coast east of Ranger Bight.

The very fresh nature of the constituent minerals of the non-porphyritic diabase, and the unfolded nature of the dyke, despite its length, indicates it is most likely post-D_4 in age.

Other post-D_2 intrusions are the diopside diorite and the pigeonite-porphyritic diabase.

Summary of Relative Ages of Intrusions

Dykes and sills of identical mineralogy and appearance to the acid and basic volcanic members of the Aillik Group are presumed to be related to those volcanics and are therefore the oldest recognised intrusions. Other pre-D_1 intrusions include several biotite-amphibolite and pegmatite dykes and veins in the Ranger Bight/Big Island area.

Biotite-carbonate dykes are the only recognised post-D_1/pre-D_2 intrusions, though there are many other pre-D_2 intrusions including biotite-amphibolite and metadiorite dykes, and pegmatite veins. Both the Round Pond Granodiorite and the Grampus Cove Gneiss have D_2 fabrics and are therefore either pre- or syn-D_2. The Long Island Gneiss and Kennedy's Cove Gneiss are both syn-D_2 intrusions, and the October Harbour Granite and Adlavik Igneous Complex are syn- to post-D_2 in age.

Both the Monkey Hill Granite and the Strawberry Granite are post-D_2 as indicated by their intrusive cross-cutting relationships with respect to S_2 fabrics in the host rocks. However, the pegmatite dykes and veins associated with the Monkey Hill body show F_3 folding, and the intrusion is therefore pre-D_3. Similar pegmatite veins associated with the Strawberry Granite show F_4 folding, but no F_3 folding, indicating a post-D_3/pre-D_4 age. Feldspar recrystallisation indicates the pigeonite-porphyritic diabase is also post-D_2/pre-D_4 in age. There are several other post-D_2 dykes, mainly of diabasic composition, which lack intrusive contacts with one another. The latest recognised dykes are lamprophyres.

CHAPTER 7

SUMMARY AND DISCUSSION

Geology of the Aillik Group

STRATIGRAPHY OF THE AILLIK GROUP IN THE AREA

The Aillik Group has been shown to consist of 8500 m. (25,500 ft.) of rhyolitic volcanics and genetically related arkoses and conglomerates, with minor mafic lavas.[1] The formations have undergone generally close to open folding and the stratigraphic sequence was readily determined using the principle of stratigraphic superposition modified slightly by a simple structural interpretation of the folding. The Aillik Group has been sub-divided into six formations, four of which are conformable on one another, and from which a fifth, to the west is separated by a major structural break, the Ranger Bight slide, in the slide zone the sixth formation, the structurally defined Ranger Bight complex,occurs.

The oldest rocks recognised in the main part of the area, east of the Ranger Bight slide, comprise the 800 m. to 2000 m. (2400 to 6000 ft.) thick Nesbit Harbour Formation. The formation is primarily a polymictic conglomerate unit derived by erosion of an acid volcanic terrane, but both below and above the conglomerate are quartz-poor arkoses, also probably derived from an acid

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[1]This thickness is measured perpendicular to sedimentary bedding and therefore cannot be an apparent thickness due to en echelon stacking as at Great Bear Lake (Hoffman and Cecile, 1974), but must be a true thickness.

volcanic terrain. The base of the formation is defined as the base of the lower arkose. The upper arkose is overlain by a mafic lava unit, the top of which is defined as the top of the formation. The Nesbit Harbour Formation is overlain by the 2000 m. to 7000 m. (6000 to 21,000 ft.) thick Makkovik Formation which is almost entirely composed of rhyolitic lavas with minor arkose and tuff interbeds. The rhyolites are approximately evenly divided into a lower flow-banded member and an upper non-banded quartz-feldspar-porphyritic member.

The Makkovik Formation is overlain in the northeast by the 2000 m. (6000 ft.) thick Pomiadluk Point Formation, and in the southeast by the 2000 m. (6000 ft.) thick Manak Bay Formation. Although these formations are stratigraphically equivalent to one another in that they both overlie the porphyritic rhyolite member of the Makkovik Formation, their lithology is too dissimilar for direct correlation with each other. The Pomiadluk Point Formation consists primarily of polymictic conglomerate and flow-banded rhyolite members with some associated tuffs, whereas the Manak Bay Formation consists primarily of quartz-poor arkose with minor mafic lavas at the top of the formation.

To the west of the Ranger Bight slide is the Big Island Formation which cannot be directly correlated with the formations east of the slide. The Big Island Formation is 3000 m. (9000 ft.) thick and is composed of many thin interbedded arkoses, conglomerates and rhyolites, with minor amphibolites.

Between the Big Island Formation and the formations to the east of the Ranger Bight slide is the structurally defined Ranger Bight complex of amphibolites which occupies the slide-zone.

RELATIONSHIP TO THE AILLIK GROUP ELSEWHERE

The most southerly part of the thesis area and the entire Aillik Group to the south were mapped by Stevenson (1970) on a reconnaissance basis (1 : 250, 000 scale). Although he recognised rhyolites in the group, he considered quartzites to constitute the dominant lithology. However, his "feldspathic quartzites of variable lithologies" occur in the southern border of the thesis area and south of Present Lake and Monkey Hill, where they were briefly investigated by the present author. These quartzites are equivalent to identically named quartzites of Gandhi, et al., (1969), and are flow-banded rhyolites, specifically the flow-banded rhyolite member of the Makkovik Formation south of Monkey Hill. Similarly Stevenson's feldspathic porphyroblastic quartzite presumably the feldspar porphyroblastic arkosic quartzite of Gandhi, et al., op. cit.,) was seen by the present author south of Shoal Lake and is an extension of the quartz-porphyritic rhyolite member. It therefore appears that there is considerably more acid volcanic material in the Aillik Group than recognised by Stevenson, and that the group elsewhere is similar to that part described in this thesis. This is bourne out by mapping to the north by Taylor and Baer (Taylor, 1972c), who also recognise acidic volcanic material in the Aillik Group.

No iron-formation has been found in the thesis area, though Stevenson (1970) reports iron-formation boulders in drift north of Present Lake. A half-mile long, gossan-encrusted, topographically sharp ridge immediately south of Present Lake has been mapped by Brinex geologists as iron-formation, but is recognised by the present author as a highly deformed, pre-D_2 pyroxenite dyke (?).

An area immediately west of Makkovik Bay which was previously mapped by the author (Clark, 1970 and 1971), may be directly correlated in part with the Big Island Formation. The cross-bedded psammite (ibid.) is equivalent to the cross-bedded arkose member of the Big Island Formation with the overlying pillowed amphibolite equivalent to the metabasalt member (which is pillowed north of Present Lake). However, the conglomerate, variable psammite (now recognised as flow-banded rhyolite) and porphyroclastic psammite (quartz-porphyritic rhyolite) cannot be directly correlated, presumably due to facies changes or removal of representative beds by erosion.

Barua (1969) considered some of the units east of Aillik Bay to the north, to be "acid spilites", based on a high Na_2O/K_2O ratio and overall chemical character. However, Collerson (personnal communication, 1972) considers the rocks may have undergone local soda-metasomatism, possibly related to intrusion of some of the lamprophyres.

The units in the vicinity of the contact between the Aillik Group and the Hopedale Complex (Sutton, et al., 1971: Sutton, 1972b) may not be basal members of the group, but representatives of units higher in the sequence that have been folded down by the recumbent F_2 anticline and F_3 antiform west of Makkovik Bay (Plate 2 and Clark, 1971).

DEPOSITIONAL ENVIRONMENT OF THE AILLIK GROUP

The Aillik Group is composed almost exclusively of acid volcanic rocks, and shallow-water arkoses and conglomerates derived from the volcanic rocks. Basic lavas and a few minor argillite and limestone units (Kranck, 1953) also occur, but form a very small proportion of the whole group. The scarcity or absence of greywackes, limestones, argillites, or orthoquartzites and dominance of acid volcanic rocks mitigates against a stable marine environment of deposition (ocean deep, rise or shelf). The occurrence of acid volcanic material to the almost complete exclusion of other rock-types and the occurrence of primarily potassic acid intrusive rocks suggest the entire sequence overlies and is derived from continental crust. The most likely types of environment for the development of this sequence are: Andean-type orogenic belts (Mitchell and Reading, 1969 and Jenks, 1956); aulacogenes (Fraser, et al., 1972); Basin and

Range -type tectonic zones (Gilluly, 1963, King, 1969 and Atwater, 1970); or New Zealand-type major continental wrench-fault environments (Wellman, 1955 and Thompson, et al., 1965).

None of the suggested models produces a depositional assemblage that is an exact analogy with the Aillik Group. The Andean-type orogenic belt results in much thinner acidic volcanic accumulations, with a greater proportion of basic volcanic and clastic sedimentary rocks. Aulacogenes develop large sequences of basic volcanic material and quartzite-carbonate-shale assemblages, with only minor acidic volcanics. Basin and Range tectonic zones develop a depositional sequence that is primarily acidic volcanic material, but that is not very thick. The model that results in a depositional sequence most closely analogous to the Aillik Group is the New Zealand-type tectonic zone, which results in very large thicknesses of acidic volcanic material, (up to 15,000 ft, but still only just over half the thickness of the Aillik Group volcanics), with little other material.

The depositional environment of the Aillik Group therefore appears most likely to have been one of deep faulting in continental crust, where the faulting was not so deep as to tap the upper mantle (thereby producing relatively large accumulations of basic volcanic rocks). The faulting may have had a major strike-slip component, as in the New Zealand model, but any form of deep crustal faulting, which is in existence long enough (i.e., longer than the fractures in the basic rock as a result of tectonic conditions. Development of perthitic structures in the microcline, mantling of plagioclase by microcline, and alteration zoning of the microcline are seen in some specimens.

Other Pre-tectonic Intrusions

A large number of other pre-tectonic dykes occur throughout the area. They are generally 10 cm. to 1 m. (4 in. to 3 ft.) wide and show well developed LS-tectonite fabrics, commonly with complex folds. The dykes are normally composed of hornblende and biotite in varying proportions, with albite and andesine, and minor amounts of diopside, chlorite, sphene, pyrite and magnetite, and sparse chalcopyrite, haematite, molybdenite and pyrrhotite. However, porphyroblastic aggregates of epidote are frequently developed, and scapolite, albite, or hornblende porphyroblastic aggregates and poikiloblasts are developed in a few dykes. The immense complexity and variation of these dykes precludes individual descriptions, but relevant textures and mineralogy will be described where necessary for the determination of structure and metamorphism of the area.

SYN-TECTONIC INTRUSIONS

Intrusions Emplaced Before or During the Main Deformation

Only one set of intrusions, the biotite-carbonate dykes, can be shown to have been intruded entirely prior to present-day models) would produce the immense thickness of acidic volcanic material.

STRUCTURE OF THE AILLIK GROUP

The Aillik Group in the thesis area has undergone four recognisable deformations, all within greenschist-amphibolite facies conditions, of which the second is regionally penetrative and produced the major structural features in the area.

The Makkovik region has undergone large-scale F_2 folding about northeast trending, steeply dipping axial planes (Fig. 4-2). The fold axes have low to moderate northward and southward plunges, the variation presumably being due to F_4 cross-folding. A major D_2 slide occurs down the western part of the area, and appears to be situated in the vicinity of pre-D_2 tensional fracturing, but it does not appear to be a rejuvenated D_1 slide. The other three deformations have produced only locally developed fabrics and structures.

The structural sequence determined within the thesis area may be directly correlated with that determined west of Makkovik Bay (Table 7-1; Clark, 1970 and 1971), as well as with that determined for the Hopedale Complex/ Aillik Group basement/cover contact area to the southwest (Sutton, *et al.*, 1971) and the post-Aillik deformation of the Hopedale Complex west of Kaipokok Bay (Sutton, 1972b).

Table 7-I

Table showing correlation of deformational events between the thesis area and other areas to the west and southwest

This thesis	Clark 1970 and 1971	Sutton, et al., 1971	Sutton, 1972b
			Complex pre-Aillik deformation of the Hopedale Complex
Pre-D_1 event?			
D_1	D_1	D_1	Zonally developed deformation
		D_2	
The Main Penetrative Deformation			Folding and development of LS-tectonite fabric
D_2	D_2	D_3	
D_3	D_3	Locally developed strain-slip fabrics	
D_4			
Post-D_4 faulting	Post-D_3 faults and kink-bands		

Igneous Intrusions of the Makkovik Area

Approximately a quarter of the thesis area is underlain by igneous intrusions. All the main intrusive plutons are syn-tectonic, and most are syn-D_2. The major intrusives vary from several syn-D_2 granitic orthogneisses and weakly deformed granites (s.l.) (Grampus Cove Gneiss, Long Island Gneiss, Kennedy's Cove Gneiss, Round Pond Granodiorite and October Harbour Granite) to two post-D_2 granites (the Monkey Hill Granite and the Strawberry Granite). Minor intrusions range from pre-tectonic amphibolites, to syn-tectonic pegmatites, amphibolites, diorites, gabbros and biotite-carbonate lamprophyres(?), and post-tectonic diabase, diorites, and carbonatitic lamprophyres. No regular variation in igneous petrology with geographic location or relative age has been recognised.

Isotopic Age Dates Relative to the Aillik Group

The Aillik Group has been involved in a major tectonic event about 1600 m.y. ago which has reset the K/Ar isotopic ages of members of the Group and pre-tectonic intrusions within the Group (Table 7-II and Fig. 7-1), as well as the age of the Hopedale Complex basement in the vicinity (Gandhi, et al., 1969; Taylor, 1971). The event is accepted as the Hudsonian orogeny (Gandhi, et al., 1969; Taylor, 1971, 1972a and b), though the ages are generally younger than the average age in the type area (1735 m. y., Churchill Province--Douglas, 1970) and there

Table 7-II

Table showing relationship of lithology, stratigraphy and isotopic (K/Ar) age-dates.

Palaeozoic (Cambrian)	Lamprophyre dykes (535 m.y.[1]; 585 m.y.[1]; 590 m.y.[1])
	±600 m.y.
Hadrynian	Various other diabase dykes (729 ± 55 m.y.[4]; 685 ± 32 m.y.[5]; 935 ± 64 m.y.[5]; 871 ± 64 m.y.[5])
Grenvillian Orogeny	955 m.y.
Helikian	Brown feldspar-porphyritic diabase dyke (956 ± 16 m.y.[2])
	"Mafic diorite" dyke (not recognised in this thesis; 992 ± 13 m.y.[2])
	Non-porphyritic diabase dyke (995 ± 13 m.y.[2])
	Strawberry Granite (1600 ± 34 m.y.[2]; 1565 ± 50 m.y.[3])
Hudsonian Orogeny 1735 m.y.	Monkey Hill Granite (1645 m.y.[1]; 1620 ± 60 m.y.[3])
	Kennedy's Cove Gneiss (1531 ± 38 m.y.[2])
	Long Island Gneiss (1832 ± 58 m.y.[2])
Aphebian	Hornblende lamprophyre (1550 ± 55 m.y.[4])
	Amphibolite dykes (1538 ± 20 m.y.[2])

Aphebian	Big Island Formation	Ranger Bight Complex (1497 ± 22 m.y.[2])	Pomiadluk Point Formation	Manak Bay Formation
			Makkovik Formation (Quartz-porphyritic rhyolite member--1545 ± 14 m.y.[2])	
	2480 m.y.		Nesbit Harbour Formation	
Archaean	Hopedale Complex (2430 m.y.[1]; 1728 ± 32 m.y.[2]).			

[1]Leech, *et al*., 1963. $\lambda_e = 0.585 \times 10^{-10}\ yr^{-1}$; $\lambda_{total} = 5.30 \times 10^{-10}\ yr^{-1}$

[2]Gandhi, et al., 1969. $\lambda_\beta = 4.72 \times 10^{-10}\ yr^{-1}$; $\lambda_e = 0.584 \times 10^{-10}\ yr^{-1}$; $^{40}K/K = 1.19 \times 10^{-4}$ g/g.

[3]Wanless, *et al*., 1970. $\lambda_\beta = 4.72 \times 10^{-10}\ yr^{-1}$; $\lambda_e = 0.585 \times 10^{-10}\ yr^{-1}$; $^{40}K/K = 1.19 \times 10^{-4}$ g/g.

[4]Wanless, *et al*., 1972. Constants as for Wanless, *et al*., 1970.

[5]Fahrig, *et al*., 1972. No constants given.

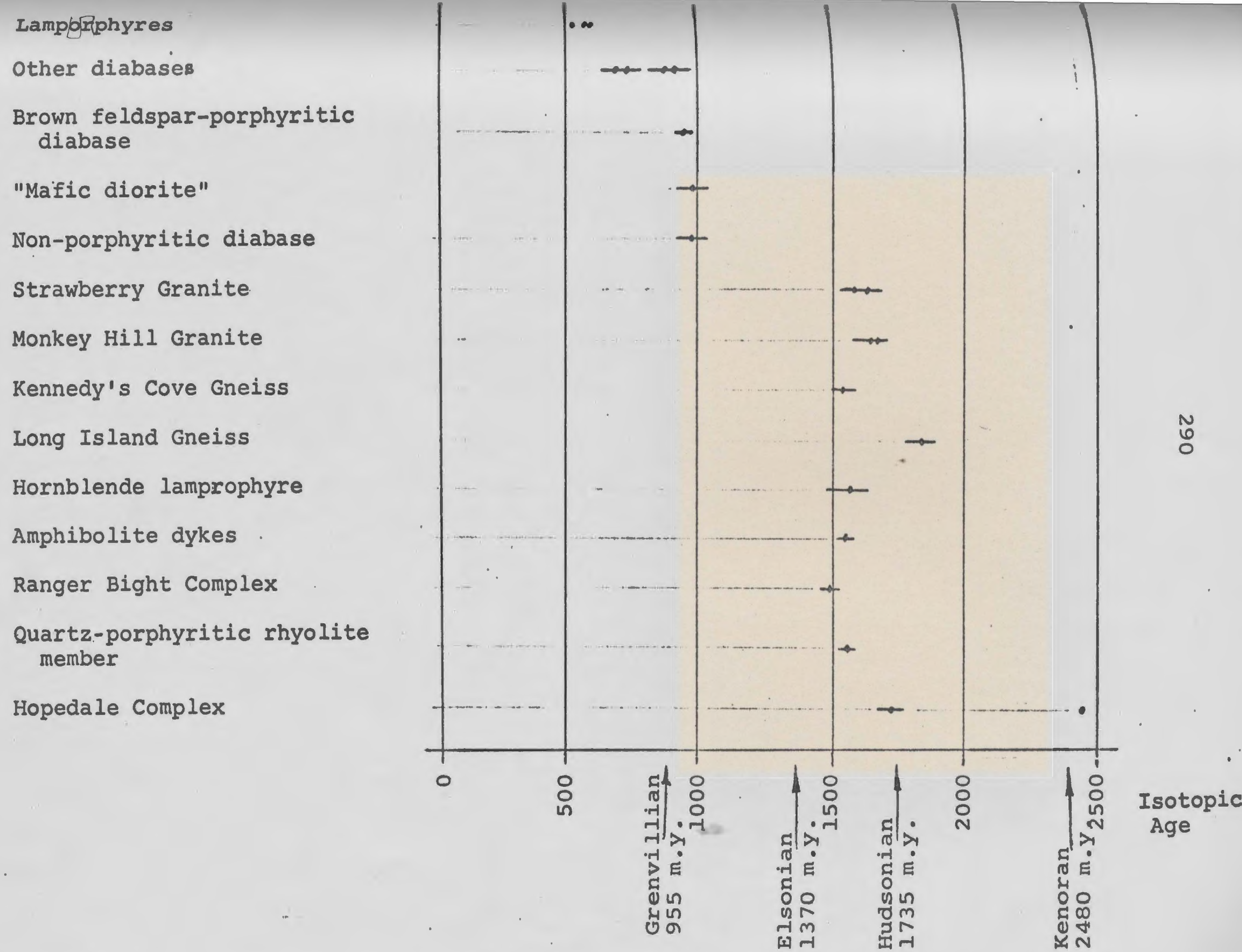

Fig. 7-1. Diagram showing the relationship of K/Ar isotopic age-dates to major Precambrian orogenies.

is no evidence that later thermal events have reset the isotopic ratios.

Syn-tectonic intrusions also have an approximate 1600 m.y. age, except for the Long Island Gneiss which has an anomalously old age of 1832 m.y., (Gandhi, et al., 1969), possibly indicating there are two different intrusions involved (Clark, 1971) or that it is a pre-tectonic (basement?) intrusion which was remobilised in part during tectonism.

Dykes emplaced during Grenvillian time are common (Gandhi, et al., 1969; Fahrig, et al., 1972), and emplacement may have continued sporadically (Wanless, et al., 1972; Fahrig, et al., 1972) through to the Cambrian when a lamprophyric suite was emplaced (Leech, et al., 1963).[1]

Tectonic Setting of the Aillik Group

The Aillik Group is situated in the Makkovik Subprovince (Taylor, 1971, 1972a and b) which is provisionally considered part of the Nain Province (Taylor, 1973a; Nutak

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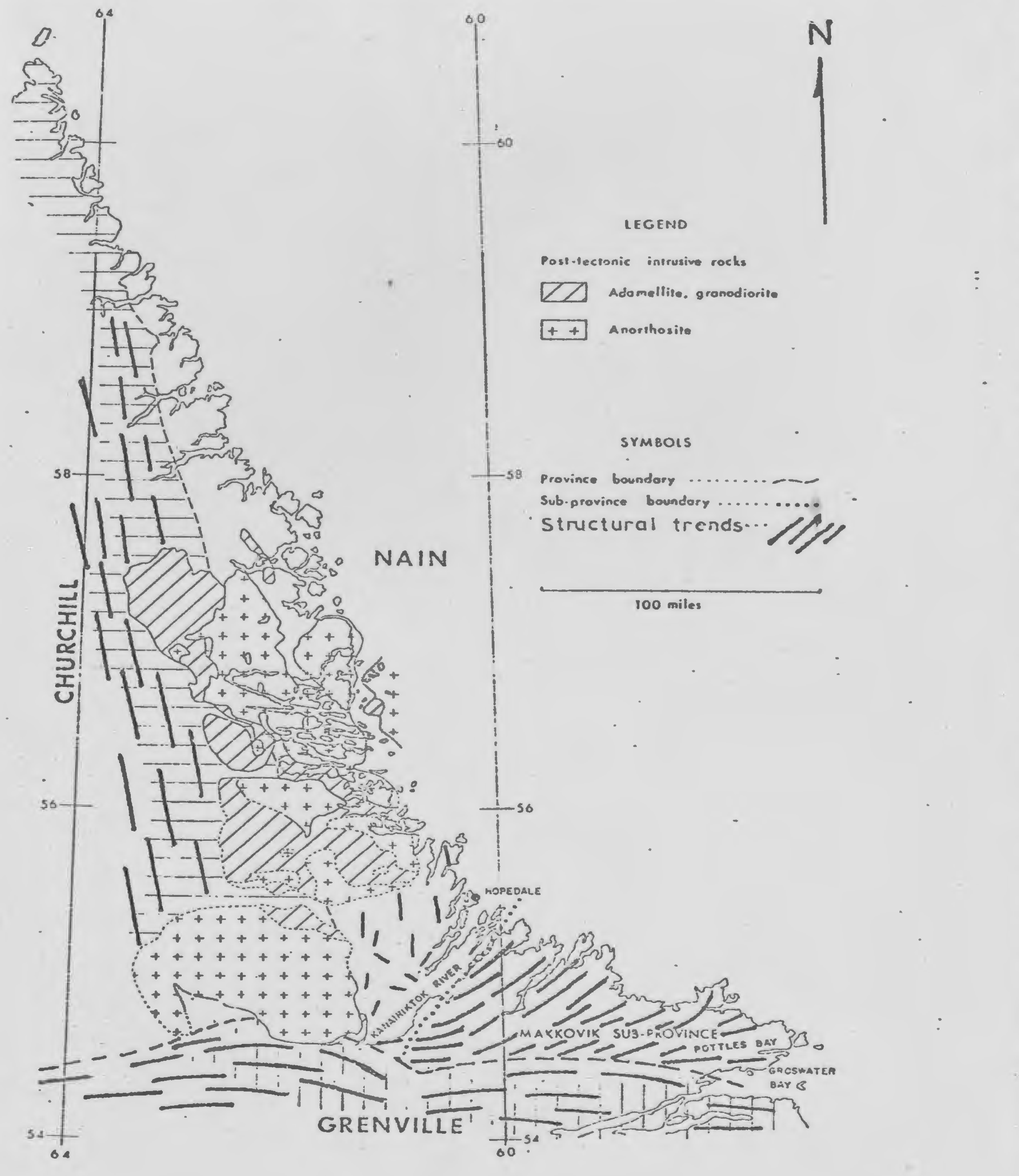
[1]It should however, be noted that diabase and lamprophyre dykes in southwest Greenland are considered to be Jurassic in age with the lamprophyres slightly younger than the diabases, and probably related to the opening of the Labrador Sea (Watt, 1969; Walton, and Arnold, 1970; Andrews, and Emeleus, 1971). Certainly, a kimberlitic dyke and sheets thought to be the same age gave K/Ar whole-rock isotopic age-dates of 609 ± 36 m.y., and 202 ± 6 m.y. respectively (Andrews, and Emeleus 1971). The older date, on further investigation, was shown to be in error, Rb/Sr whole-rock isotopic age-dates of each being a little over 200 m.y., (Andrews, and Emeleus, 1971). A similar situation has also arisen in the dating of kimberlites from Siberia (Bridgwater, 1971), indicating the need for caution in interpreting the Labrador samples.

Province of Douglas, 1972 and Douglas and Price, 1972). The region is included in the Grenville Foreland Belt by Wynne-Edwards (1972), though the only evidence of the Grenvillian orogeny is the occurrence of some diabase dykes of approximately Grenville Age. The entire character of the province was determined by the tectonism that took place at about 1600 m.y. B.P. This "orogeny" developed the general upright, northeast to east trending folds which become recumbent to the northwest at Kaipokok Bay (Fig. 4-2 and Plate 2), and the preservation of shallower stratigraphic and structural levels to the east. The trend of the "orogenic" belt appears to intersect the Churchill Province to the west (Fig. 7-2; Sutton, 1972a).

The sigmoidally oriented F_2 fold hinges in the Makkovik region suggest the Aillik Group may have undergone strike-slip (simple-shear) movement (Fig. 7-3) (Kennedy, personnal communication, 1973).

Comparison with the Croteau and Seal Groups

The Croteau Group is situated approximately 100 km. (50 mi.) southwest of the thesis area and along strike with the Aillik Group. It is divided into Lower, Middle and Upper parts (Greene, 1972) of which the Upper is Aphebian in age (Wanless and Loveridge, 1972). The Lower and Middle Croteau Group consist primarily of argillite, quartzite, greywacke, dolomite, conglomerate and basalt, and are therefore generally unlike the lithologies of the Aillik Group.



Modified after Taylor (1971 and 1972c) and Sutton (1972a) with data from Stevenson (1970), Greene (1972) and Fahrig et al. (1972).

Fig. 7-2. Map of eastern Labrador showing the orientation of structural trends in the Churchill and Grenville Provinces and the Makkovik Sub-province.

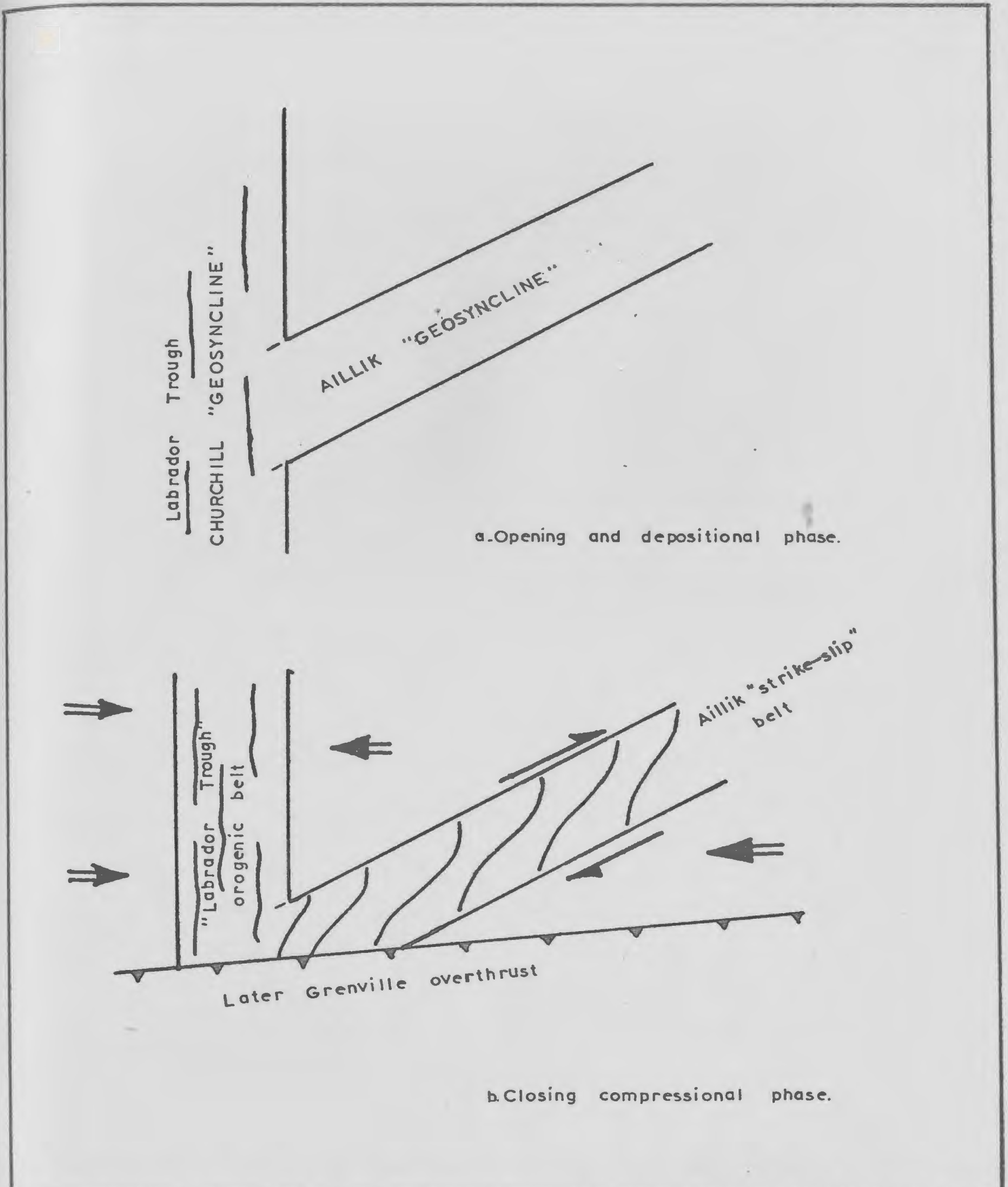


Fig. 7-3. Diagrammatic representation of the development of the Aillik "strike-slip" belt.

However, the Upper Croteau Group consisting predominantly of intermediate to acidic volcanics and feldspathic quartzites is, therefore, both very similar to the Aillik Group, and of the same age, suggesting it is a direct extension of the Group. The Neohelikian Seal Group of basic volcanics, quartzites, limestones and shales (Greene, 1972) occurs west of the Croteau Group.

Comparison with Southwestern Greenland

The Ketilidian Mobile Belt (Escher and Bridgwater, 1972) of southernmost Greenland is situated in a position that is thought to have been along strike from the thesis area prior to opening of the Labrador Sea (Fig. 7-4; Bullard, *et al*., 1965; Sutton, *et al*., 1972).

The Ketilidian supracrustal rocks are somewhat different in lithology to the Aillik Group, being composed of quartzites, conglomerates, shales, greywackes and pillow-lavas, but no acid volcanic rocks (Henriksen, 1969). The lithology is thought by Henriksen (*op. cit*.,) to represent deposition in a geosyncline close to the border of the pre-Ketilidian platform to the north.

The structural setting is similar, with the Ketilidian supracrustals situated unconformably on the underlying older gneissic basement with, in places, a tectonic (thrust) junction and associated "gneissic schist" (*ibid*.). The Ketilidian supracrustals become progressively more deformed and metamorphosed towards the south. These the main deformation (c.f. Chapter 6), all the others may have been intruded either prior to or during that deformation. No cross-cutting relationships between these and other intrusions were seen.

i. Biotite-carbonate Dykes: Several dark green, medium-grained biotite carbonate dykes, which commonly contain K-feldspar, hornblende or diopside as additional major mineral species, occur in the area. Plagioclase, fluorite, opaque minerals and sphene are common accessories. The biotite defines a good S-tectonite fabric, and metamorphic regrowth of mineral species is commonly shown by included tectonite fabrics in both hornblende and diopside. The original mineralogy of the dykes is unknown but carbonate forms up to 30% of the total mineral content and is therefore probably primary igneous carbonate, and the hornblende may, in some cases, have a core of primary(?) clinopyroxene (augite?) from which it has apparently developed by metamorphic recrystallisation. The original dykes were therefore probably lamprophyric carbonatites.

ii. Black Porphyritic Metadiorite: A 100 to 200 m. (300 to 600 ft.) diameter dark green, medium-grained, feldspar-porphyritic metadiorite intrusion occurs on Big Island (Fig. 4-19). Slightly finer grained, but otherwise identical 1 m. (3 ft.) dykes occur elsewhere on Big Island and near Ranger Bight. The feldspar phenocrysts are up to 1/2 cm. (1/4 in.) long, and are andesine in composition,

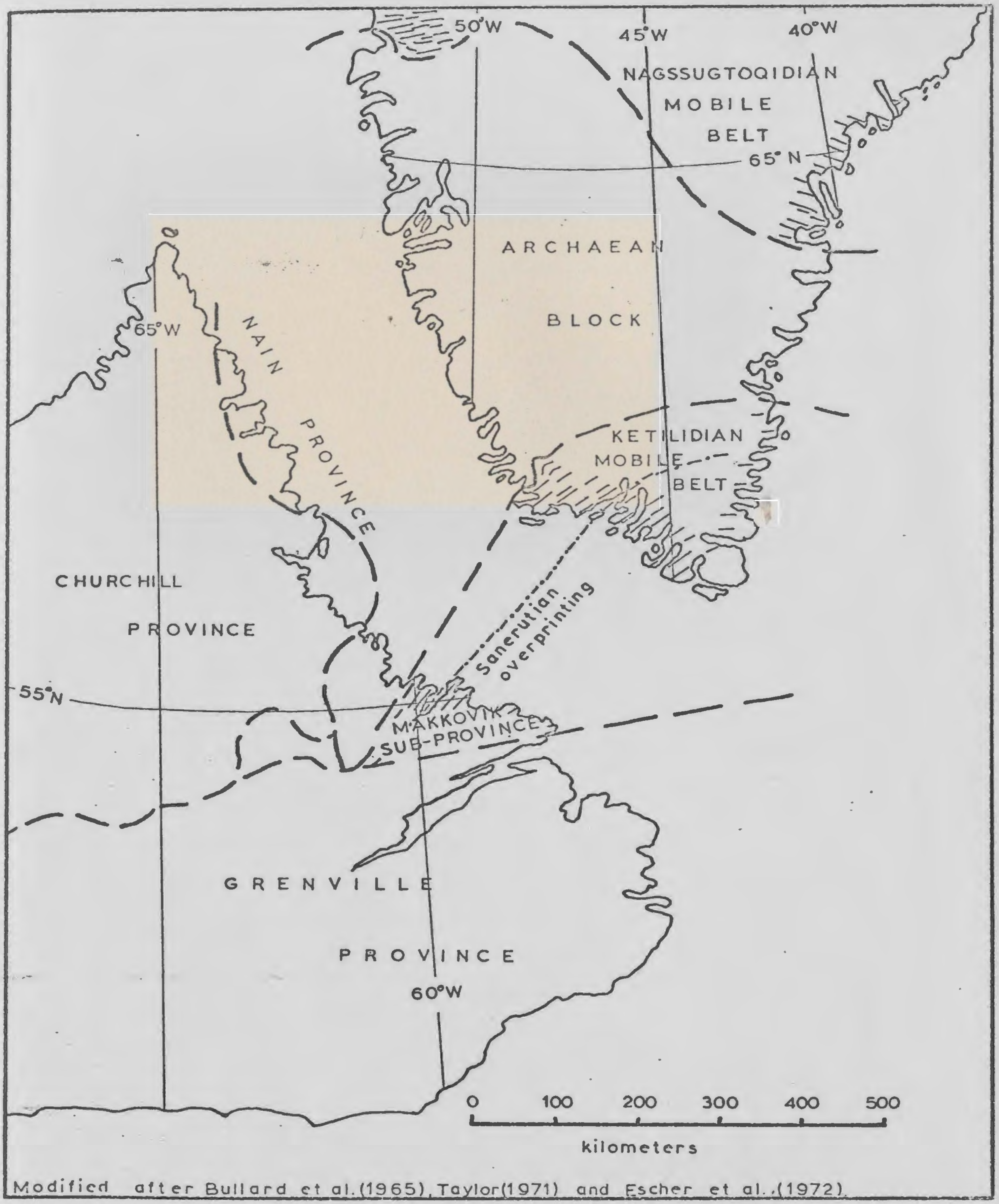


Fig. 7-4. Map showing comparison of tectonic provinces between parts of Labrador and Greenland prior to opening of the Labrador Sea.

relationships are closely comparable to those of the Aillik Group/Hopedale Gneiss junction (Sutton, et al., 1971).

Ketilidian tectonism is similar to that of the Makkovik region and has resulted in two main phases of deformation, both in the greenschist and amphibolite facies, which developed upright to overturned, northeast trending folds. Plutonic activity is also similar, with syn-tectonic granite emplacement followed by intrusion of an appinitic suite, then followed by emplacement of the younger granites (Bridgwater and Walton, 1964). However, the syn-tectonic granite intrusions of southwest Greenland were emplaced and deformed about 1800 m.y., ago (Bridgwater, 1965; Allaart, et al., 1969) whereas the only comparable intrusion from the Makkovik region is the Long Island Gneiss on Long Island (1830 m.y.). The other syn-tectonic granitic intrusions are much younger in age (1530 m.y.,) suggesting a greater development of tectonic and plutonic events of younger (Sanerutian equivalent: Bridgwater, 1965) age rather than of older (Ketilidian equivalent) age (Table 7-III and Fig. 7-4).

Conclusion

The thesis area is underlain by a thick sequence of Aphebian acid volcanic rocks and related sediments, with minor basic volcanic rocks (Table 7-IV). This sequence has been sub-divided into six formations which comprise the Aillik Group in the area. The Aillik Group is locally

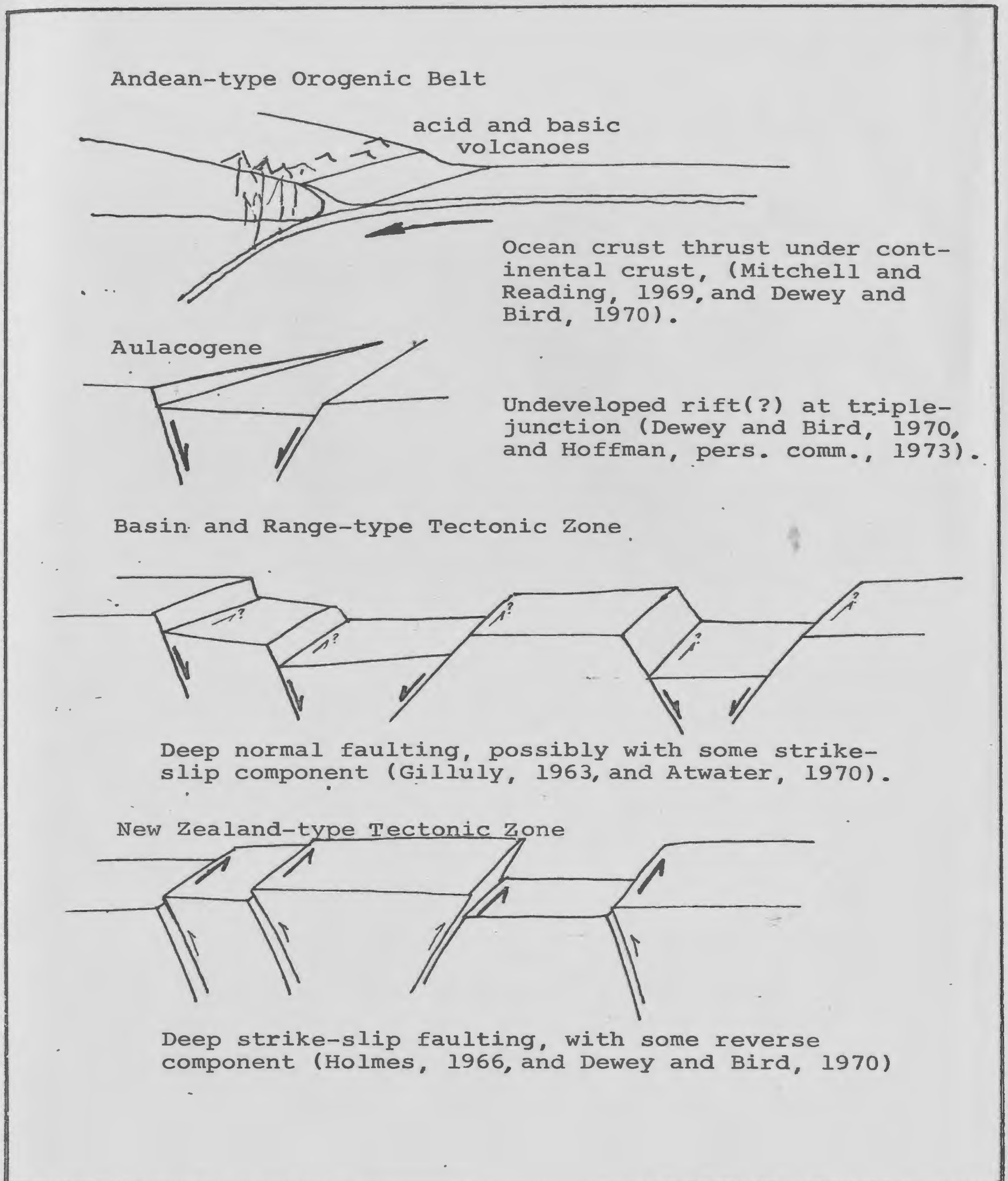


Fig. 7-5. Diagramatic representation of possible tectonic environments resulting in formation of the Aillik Group.

unconformably overlain by the Jurassic Ford's Bight Conglomerate. The area has undergone polyphase deformation of Sanerutian age and is situated on the northwestern boundary of a large orogenic belt extending from near the Labrador Trough in the west to the east coast of Greenland in the east. This orogenic belt may have undergone strike-slip (simple-shear) movement.

It has been suggested by Dr. A. J. Baer (pers. comm., 1974) that the Aillik Group is the syn-plutonic "blanket" of ejecta (and flows?) that accompanied the emplacement of (Elsonian) adamellites in Labrador. The Elsonian isotopic age-dates cluster around 1370 m. y. (Stockwell, 1964) and are therefore considerably younger than the age-dates from the Makkovik region or Greenland (1500 - 1600 m. y.). This suggestion implies intrusion (and associated extrusion) during tectonism at 1500 - 1600 m. y. followed by slow cooling for 100 to 200 m. y. before the isotopic ages of the intrusions were set. It seems unlikely that any plutonic rock, particularly a granitic (s.l.) rock which is molten at a relatively low temperature, would take such a long period of time to cool.

Table 7-III

Table showing comparison between depositional and intrusional events in the thesis area and Southern Greenland.

	Thesis Area		Southern Greenland
		Ketilidian-Sanerutian plutonic and tectonic episodes.	lamprophyres (≈ 200 m.y.)
"Hudsonian" orogeny	lamprophyres (550-600 m.y.??)		
	diabase emplacement (700-1000 m.y.)		diabase dykes (≈ 1100 m.y.)
	late-tectonic granites (Monkey Hill and Strawberry types) (1550-1650 m.y.)		younger granites (1500-1600 m.y.)
	Adlavik Igneous Complex		appinitic intrusive complexes (≈ 1650 m.y.)
	Syn-tectonic granitic intrusions (1500-1600 m.y.)		syn-tectonic granites (≈ 1800 m.y.)
	Aillik Group deposition		Ketilidian supracrustals (≈ 2000 m.y.)
		major u/c	
	Hopedale Complex (≈ 2500 m.y.)		pre-Ketilidian gneisses (≈ 2700 m.y.)

Bibliography

Allaart, J. H., Bridgwater, D and Henriksen, N., 1969. The Pre-Quaternary Geology of S. W. Greenland and its bearing on problems of correlation in the N. Atlantic. Am. Assoc. Petrol. Geol., Mem. 12, pp. 859 - 882.

American Commission on Stratigraphic Nomenclature, 1970. Code of Stratigraphic Nomenclature. Am. Assoc. Petrol. Geol., Tulsa, Oklahoma.

American Geological Institute, 1962. Dictionary of Geology. Doubleday, New York.

Andrews, J. R. and Emeleus, C. H., 1971. Preliminary account of Kimberlite intrusions from Frederikshåb district, S. W. Greenland. Grøn. Geol. Under., Rapport Nr. 31.

Atwater, T., 1970. Implications of Plate Tectonics for the Cenozoic Tectonic Evolution of Western North America. Geol. Soc. Am. Bull., v. 81, pp. 3513-3535.

Bailey, E. H. and Stevens, R. E., 1960. Selective staining of K-feldspar and plagioclase on rock-slabs and thin-sections. Am. Min., v. 45, pp. 1020-1025.

Barua, M. C., 1969. Geology of Uranium-Molybdenum-Bearing rocks of the Aillik-Makkovik Bay Area, Labrador. Unpublished M.Sc. Thesis, Queen's University, Kinston, Ontario.

Beavan, A. P., 1958. The Labrador Uranium Area. Proc. Geol. Assoc. Canada, v. 10, pp. 137 - 145.

Berger, A. R. and Pitcher, W. S., 1970. Structures in Granitic Rocks: A Commentary and a Critique on Granite Tectonics. Geol. Assoc. Proc., v. 81, pt. 3, pp. 441 - 461.

Blake, D. H., Elwell, R. W. D., Gibson, I. L., Skelhorn, R. R. and Walker, G. P. L., 1965. Some relationships resulting from the intimate association of acid and basic magmas. Quart. J. Geol. Soc. Lond., v. 121, pp. 31 - 49.

Blank, H. R., 1972. Hornblende Schists in the Manhattan Formation, in the Bronx, New York. Geol. Soc. Am., Bull., v. 83, pp. 1397 - 1412.

Boyd, F. R., 1961. Welded Tuffs and Flows in the Rhyolite Plateau of Yellowstone Park, Wyoming. Geol. Soc. Am. Bull., v. 72, pp. 387 - 426.

Bridgwater, D., 1965. Isotopic age determinations from S. Greenland and their geologic setting. Meddel. om Grøn., Bd. 179, Nr. 4.

Bridgwater, D., 1970. Observations on the Precambrian rocks of Scandinavia and Labrador and their implications for the interpretation of the Precambrian of Greenland. Grøn. Geol. Under., Rapport Nr. 28, pp. 43 - 47.

__________, 1971. Routine K/Ar Age Determinations on rocks from Greenland Carried out for G. G. U. in 1970. Grøn. Geol. Under., Rapport Nr. 35, pp. 52 - 60.

__________ and Walton, B. J., 1964. The tectonic-magmatic evolution of the Svecofennid chelogenic cycle in S. Greenland. Nature, Lond., v. 203, pp. 278 - 281.

Brown, W. L., 1962. Peristerite unmixing in the plagioclases and metamorphic facies series. Norsk. geol. tidsskr., v. 42 (Feldspar Volume), pp. 354-382.

Bullard, E., Everett, J. E. and Smith, G. A., 1965. The fit of the continents around the Atlantic. In: A symposium on Continental Drift, Roy. Soc. Lond. Phil. Trans. A, v. 258, pp. 41 - 51.

Cameron, E. N., 1961. Ore Microscopy. J. Wiley and Sons, N. Y.

Chapman, C. A., 1962. Diabase-granite composit dykes, with pillow-like structure, Mount Desert Island, Maine, J. Geol., v. 70, pp. 539 - 564.

Christie, A. M., Roscoe, S. M. and Fahrig, W. F., 1953. Preliminary Map, Central Labrador Coast. Geol. Surv. Pap. Can., 53 - 1.

Clark, A. M. S., 1970. A Structural Reinterpretation of the Aillik Series, Labrador. Unpublished M.Sc. Thesis, Memorial University of Newfoundland, St. John's, Newfoundland.

__________, 1971. Structure and Lithology of Part of the Aillik Series, Labrador. Geol. Assoc. Canada Proc., v. 24, no. 1, pp. 107 - 117.

Cole, G. A. T., 1885. On hollow spherulites and their occurrence in Ancient British lavas. Quart. J. Geol. Soc. Lond., v. 41, pp. 162 - 169.

Cooper, G. E., 1951. The petrology of some syenites and granites in Labrador. Unpublished M.Sc. Thesis, McGill University, Montreal, Quebec.

Deer, W. A., Howie, R. A. and Zussman, J., 1970. An introduction to the Rock Forming Minerals. Longman, London.

Delabarre, E. B., 1902. Report of the Brown-Harvard Expedition to Nachvak, Labrador, 1900. Preston and Rounds, Rhode Island.

Department of Energy, Mines and Resources, (Federal Government of Canada), Geographical Branch, 1968. Gazetteer of Canada: Newfoundland and Labrador Volume. Queen's Printer, Ottawa.

Dewey, J. F., 1965. Nature and Origin of Kink-bands. Tectonophys., v. 1, no. 6, pp. 459 - 494.

__________, 1969. The Origin and Development of Kink-bands in a Foliated Body. Geol. J., v. 6, pt. 2, pp. 193 - 216.

Dewey, J. F. and Bird, J. M., 1970. Mountain Belts and the New Global Tectonics. J. Geophys. Res., v. 75, pp. 2625 - 2647.

Douglas, G. V., 1953. Notes on localities visited on the Labrador Coast in 1946 and 1947. Geol. Surv. Canada, Paper 53 - 1, pp. 20 - 30.

Douglas, R. J. W. (editor), 1970. Geology and Economic Minerals of Canada. Geol. Surv. Canada, Economic Geol. Rept. 1, 1838p.

__________, 1972. A Revision of Precambrian Structural Provinces in Northeastern Quebec and Northern Labrador Discussion. Can. J. Earth Sci., v. 9, pp. 925 - 930.

__________ and Price, R. A., 1972. Variations in Tectonic Styles in Canada. Geol. Assoc. Canada, Spec. Paper No. 11. 688p.

Escher, A. and Bridgwater, D., 1972. Research sponsored by the Geological Survey of Greenland as a contribution to the International Geodynamics Project. Geol. Surv. Greenland, Copenhagen, 5p.

Ewart, A., 1963. Petrology and Petrogenesis of the Quaternary Pumice Ash in the Taupo Area, New Zealand. J. Pet., v. 4, pp. 392 - 431.

__________, 1971. Chemical changes accompanying spherulitic crystallisation in rhyolite lavas, Central Volcanic Region, New Zealand. Min. Mag., v. 38, no. 296, pp. 424 - 434.

Fahrig, W. F. and Larochelle, A., 1972. Paleomagnetism of the Michael Gabbro and Possible Evidence of the Rotation of Makkovik Subprovince. Can. J. Earth Sci., v. 9, no. 10, pp. 1287 - 1296.

Fairbairn, H. W., 1949. Structural Petrology of Deformed Rocks. Addison-Wesley, Mass., 344p.

Fenner, C. N., 1920. The Katmai Region, Alaska, and the great eruption of 1912. J. Geol., v. 28, pp. 569 - 606.

Fleuty, M. J., 1964a. The Description of Folds. Geol. Assoc. Proc., v. 75, pp. 461 -492.

__________, 1964b. Tectonic Slides. Geol. Mag., v. 101, pp. 452 - 456.

Flinn, D., 1958a. On the nappe structure of N. E. Shetland. Quart. J. Geol. Soc. Lond., v. 114, pp. 107 - 136.

__________, 1958b. On Tests of Significance of Preferred Orientation in Three-Dimensional Fabric Diagrams. J. Geol., v. 66, pp. 526 - 539.

__________, 1962. On Folding During Three-Dimensional Progressive Deformation. Quart. J. Geol. Soc. Lond., v. 118, pp. 385 - 433.

__________, 1965. Deformation in Metamorphism. In: Controls of Metamorphism, (Pitcher, W. S. and Flinn, G. W., editors), Oliver and Boyd, Edinburgh.

Fraser, J. A., Hoffman, P. F., Irvine, T. N. and Mursky, G., 1972. The Bear Province. In: Variations in Tectonic Styles in Canada, (Price, R. A. and Douglas, R. W. J., editors), Geol. Assoc. Canada, Spec. Paper no. 11, pp. 453 - 504.

Friedman, M., 1964. Petrofabric Techniques for the determination of Principle Stress directions in Rocks. In: State of Stress in the Earth's Crust, (Judd, W. R., editor), Elsevier, New York.

Gandhi, S. S., Grasty, R. L. and Grieve, R. A. F., 1969. The Geology and Geochronology of the Makkovik Bay Area, Labrador. Can. J. Earth Sci., v. 6, no. 5, pp. 1019 - 1035.

Gibson, I. L. and Walker, G. P. L., 1963. Some Composite rhyolite/basalt lavas and related composite dykes in Eastern Greenland. Geol. Assoc. Proc., v. 74, pt. 3, pp. 301 - 318.

Gill, F. D., 1966. Petrography of Molybdenite-bearing gneisses, Makkovik area, Labrador. Unpublished M. A. Thesis, University of Toronto, Toronto, Ontario.

Gilluly, J, 1963. The Tectonic Evolution of the Western United States. Quart. J. Geol. Soc. Lond., v. 119, pp. 133 - 174.

Grasty, R. L., Rucklidge, J. C. and Elders, W. A., 1969. New K/Ar Age determinations on rocks from the east coast of Labrador. Can. J. Earth Sci., v. 6, pp. 340 - 344.

Greene, B. A., 1972. Geological Map of Labrador. Mineral Resources Division, Department of Energy, Mines and Resources, Province of Newfoundland and Labrador.

__________ and McKillop, J. H., 1972. The first geological map of Labrador. Can. Mining J., v. 93, pp. 89 - 97.

Hancock, P. L., 1972. The analysis of en-échelon veins. Geol. Mag., v. 109, no. 3, pp. 269 - 276.

Henriksen, N., 1969. Boundary relations between Precambrian fold belts in the Ivigtut area, southwest Greenland. In: Age Relations in High-Grade Metamorphic Terrains, (H. R. Wynne-Edwards, editor), Geol. Assoc. Canada, Special Publ. No. 5, pp. 143 - 154.

Hietanen, A., 1967. Scapolite in the Belt Series in the St. Joe-Clearwater Region, Idaho. Geol. Soc. Am., Sp. Paper 86.

Hoffman, P. F. and Cecile, M. P., 1974. Volcanism and Plutonism, Sloan River Map-Area (86K), Great Bear Lake, District of Mackenzie. Geol. Surv. Canada. Paper 74-1, Part A.

Holmes, A., 1966. Principles of Physical Geology. Nelson, London.

Hughes, C. J., 1960. The Southern Mountains Igneous Complex, Isle of Rhum. Quart. J. Geol. Soc. Lond., v. 116, pp. 111 - 138.

__________, 1971. Anatomy of a Granophyre Intrusion. Lithos, v. 4, pp. 403 - 415.

__________ and Malpas, J. G., 1971. Metasomatism in the Late Precambrian Bull Arm Formation in Southeastern Newfoundland: Recognition and Implications. Geol. Assoc. Canada Proc., v. 24., pp. 85 - 93.

Jenks, W. F. (ed.), 1956. Handbook of South American Geology. Geol. Soc. Am., Memoir 65.

though now are highly saussuritised. The groundmass has recrystallised and shows a tectonic fabric of hornblende aggregates in a polygonal albite, oligoclase/andesine and microcline sub-fabric.

The metadiorite is composed of:

andesine phenocrysts	15 - 20%
groundmass albite and oligoclase/andesine	30 - 40%
green hornblende	20 - 30%
microcline	15 - 20%

and minor amounts of epidote, opaque minerals, chlorite, sphene, apatite, and calcite.

iii. <u>Hornblende Diorite Dyke</u>: A 1 1/2 km. (3/4 mi.) long sub-horizontal outcrop of hornblende diorite occurs south of Wild Bight. The rock is a medium-grained, salt-and-pepper textured, dark grey and white rock which in thin-section is seen to contain a matt of acicular green hornblende grains and feldspar laths with minor interstitial quartz, opaque minerals and sericitised K-feldspar (orthoclase?). The margins of the body show development of a schistose texture.

iv. <u>Gabbroic Intrusions</u>: A small 150 to 200 m. (450 to 600 ft.) diameter, 10 to 15 m. (30 to 45 ft.) thick sheet-like metamorphosed gabbroic body dipping approximately 20^{o} westward occurs between Wild Bight and Ford's Bight. The

Johnson, M. W. R., 1967. Mylonite Zones and Mylonite Banding. Nature, v. 213, pp. 246 - 247.

Kesler, S. E. and Weiblen, P. W., 1968. Distribution of elements in spherulitic andesite. Am. Min., v. 53, pp. 2025 - 2035.

__________, 1963b. Geology of the Cape Makkovik Peninsula, Aillik, Labrador. Unpublished M.Sc. Thesis, Memorial University of Newfoundland, St. John's, Newfoundland.

King, A. F., 1963a. Report on the Geology of the Makkovik Area, Labrador. Unpublished BRINEX Technical Report.

King, P. B., 1969. The Tectonics of North America--A discussion to accompany the Tectonic Map of North America Scale 1 : 5,000,000. U. S. Geol. Surv. Prof. Paper 628.

Kranck, E. H., 1939. Bedrock Geology of seaboard region of Newfoundland, Labrador. Geol. Surv. Newfoundland, Bull. no. 19.

__________, 1953. Bedrock Geology of the Seaboard of Labrador between Domino Run and Hopedale. Geol. Surv. Canada, Bull. no. 26.

__________, 1961. An unusual Type of Deformation in a Basic Sill. Bull. Geol. Insts. Univ. Uppsala, v. XL.

Leech, G. B., Lowdon, J. A., Stockwell, C. H. and Wanless, R. K., 1963. Age Determinations and Geological Studies. Geol. Surv. Canada, Paper 63 - 17.

Lieber, O. M., 1860. Notes on the Geology of the Coast of Labrador. Report of the U. S. Coast Survey.

Lipman, P. W., 1965. Chemical comparison of glassy and crystalline volcanic rock. U. S. Geol. Surv. Bull., 1201-D, pp. 1 - 24.

Lowden, J. A., 1961. Age Determinations by the G. S. C.. Dept. of Mines and Technical Surveys, Canada, Paper 61 - 17.

Mason, R., 1971. The structure and chemistry of the Sulitjelma gabbro. Norges. Geol. Unders., no. 269, pp. 108 - 142.

Mitchell, A. H. and Reading, H. G., 1969. Continental Margins, Geosynclines and Ocean-Floor Spreading. J. Geol., v. 77, no. 6, pp. 629 - 646.

Moore, J. G. and Peck, D. L., 1962. Accretionary lapilli in volcanic rocks of the western continental United States. J. Geol., v. 70, pp. 182 - 193.

Moore, T. H., 1951. Igneous dyke rocks of the Aillik-Makkovik area, Labrador. Unpublished M.Sc. Thesis, McGill University, Montreal, Quebec.

Moorhouse, W. W., 1959. The Study of Rocks in Thin-Section. Harper, New York.

Morris, P. G., 1959. Geology of Shoal Lake, Labrador. Unpublished BRINEX Technical Report.

Mumpton, F. A. and Sheppard, R. A., 1972 (March). Zeolites. Geotimes, v. 17, pp. 16 - 17.

Naldrett, A. J., 1966. Talc-Carbonate Alteration of some Serpentinized Ultramafic Rocks South of Timmins, Ontario. J. Pet., v. 7, pt. 3, pp. 489 - 499.

Nicholson, R., 1966. Metamorphic differentiation in crenulated schists. Nature, v. 209, no. 5018, p. 68.

Oertel, G., 1970. Deformation of a slaty lapillar tuff in the Lake District, England. Bull. Geol. Soc. Am., v. 81, pp. 1173 - 1188.

O'Hara, 1872. Map of Labrador. An unpublished, hand-draughted map of the Coast of Labrador attributed to O'Hara and in the possession of the Superintendent of Moravian Missions in Labrador.

Orville, P. M., 1969. A model for metamorphic differentiation origin of thin-layered amphibolites. Am. J. Sci., v. 267, pp. 64 - 86.

Packard, A. S., Jr., 1891. The Labrador Coast. N. D. C. Hodges, New York.

Pettijohn, F. J., 1957. Sedimentary Rocks. Harper and Row, New York.

__________ and Potter, P. E., 1964. Atlas and Glossary of Primary Sedimentary Structures. Springer-Verlag, N. Y., 370p.

Phillips, F. C., 1971. The Use of Stereographic Projection in Structural Geology. Edward Arnold, London.

Pitcher, W. S. and Berger, A. R., 1972. The Geology of Donegal: A Study of Granite Emplacement and Unroofing. Wiley-Interscience, New York, 448p.

Powell, D. and Treagus, J. E., 1967. On the geometry of S-shaped inclusion trails in garnet porphyroblasts. Min. Mag., v. 36, no. 279, pp. 453 - 456.

__________ and __________, 1970. Rotational fabrics in metamorphic minerals. Min. Mag., v. 37, no. 291, pp. 801 - 814.

Ramsauer, H., 1941. Achsensverteilungsanalysen an Quartz-tectoniten. Dissert. Univ. Innsbrück, Biblio. no. 304.

Ramsay, J. G., 1967. Folding and Fracturing of Rocks. McGraw-Hill, New York.

__________ and Graham, R. H., 1970. Strain variation in shear-belts. Can. J. Earth Sci., v. 7, no. 3, pp. 786 - 813.

__________ and Wood, D. S., 1973. The geometric effects of volume change during deformational processes. Tectonophys., v. 16, pp. 263 - 277.

Riley, G. C., 1951. The bedrock geology of Makkovik and its relation to the Aillik and Kaipokok Series. Unpublished M.Sc. Thesis, McGill University, Montreal, Quebec.

Rittmann, A., 1962. Volcanoes and their Activity. (Translator: E. Vincent), Wiley-Interscience, New York.

Ross, C. S. and Smith, R. L., 1961. Ash-flow tuffs: The origin, geologic relations and identification. U. S. Geol. Surv., Prof. Pap. 366.

Sander, B., 1950. Einführung in die Gefügekunde der Geologischen Körper. Springer, Berlin.

Schouten, C., 1962. Determination Tables for Ore Microscopy. Elsevier, New York.

Scott, R. B., 1971. Alkali exchange during devitrification and hydration of glasses in ignimbrite cooling units. J. Geol., v. 79, pp. 100 - 110.

Shackleton, R. M., 1957. Downward facing structures of the Highland Border. Quart. J. Geol. Soc. Lond., v. 113, pp. 361 - 392.

Shelley, D., 1971. The Origin of Cross-Girdle Fabrics of Quartz. Tectonophys., v. 11, pp. 61 - 68.

Slemmons, D. B., 1962. Determination of Volcanic and Plutonic Plagioclases Using a 3- or 4-Axis Universal Stage-Revision of Turner Method. Geol. Soc. Am., Spec. Pap. 69.

Smith, R. E., 1968. Redistribution of Major Elements in the Alteration of some Basic Lavas during Burial Metamorphism. J. Pet., v. 9, pt. 2, pp. 191 - 219.

Spry, A., 1969. Metamorphic Textures. Pergamon Press, New York.

Steinhauer, H., 1814. Notes on the Geology of the Labrador Coast. Trans. Geol. Soc., II, pp. 488 - 491.

Stevenson, I. M., 1970. Rigolet and Groswater Bay map-areas, Newfoundland. Geol. Surv. Canada, Paper 69 - 48.

Stockwell, C. H., 1964. Age Determinations and Geological Studies. Geol. Surv. Canada, Paper 64 - 17, pt. II.

__________, 1968. Geochronology of Stratified Rocks of the Canadian Sheild. Can. J. Earth Sci., v. 5, no. 3, pt. 2, pp. 693 - 698.

Stoeterau, W. W., 1970. Geology of the Round Pond East Area, Makkovik, Labrador. Unpublished B.Sc (Hons) Dissertation, Memorial University of Newfoundland, St. John's, Newfoundland.

Sutton, J. S., 1972a. Notes on the Geology of Labrador. Memorial Univ. of Newfoundland, Geological Report No. 5.

__________, 1972b. The Precambrian Gneisses and Supracrustal Rocks of the Western Shore of Kaipokok Bay, Labrador, Newfoundland. Can. J. Earth Sci., v. 9, no. 12, pp. 1677 - 1692.

__________, Marten, B. E. and Clark, A. M. S., 1971. Structural History of the Kaipokok Bay Area, Labrador, Newfoundland. Geol. Assoc. Canada Proc., v. 24, pp. 103 - 106.

__________, __________, __________and Knight, I., 1972. A correlation of the Precambrian supracrustal rocks of coastal Labrador and southwestern Greenland. Nature Phys. Sci., v. 238, pp. 122 - 123.

Sylvester, A. G. and Christie, J. M., 1968. The origin of cross-girdle orientations of optic axes in deformed quartzites. J. Geol., v. 76, pp. 571 - 580.

Taylor, F. C., 1971. A Revision of Precambrian Structural Provinces in Northeastern Quebec and Northern Labrador. Can. J. Earth Sci., v. 8, no. 5, pp. 579 - 584.

__________, 1972a. A Revision of Precambrian Structural Provinces in Northeastern Quebec and Northern Labrador: Reply. Can. J. Earth Sci., v. 9, no. 7, pp. 930 - 932.

__________, 1972b. The Nain Province. In: Variations in Tectonic Styles in Canada, (Price, R. A. and Douglas, R. J. W., editors), Geol. Assoc. Canada, Spec. Paper no. 11, pp. 435 - 452.

__________, 1972c. Reconnaissance Geology of a part of the Precambrian shield, northeastern Quebec and northern Labrador; part III. Geol. Surv. Canada, Paper 71-48.

Thompson, B. N., Kermode, L. O. and Ewart, A., 1965. New Zealand Geol. Surv. Handbook, Information Series 50, New Zealand Dept. of Scientific and Industrial Research.

Turner, F. J., 1968. Metamorphic Petrology. McGraw-Hill, New York.

__________ and Weiss, L. E., 1963. Structural Analysis of Metamorphic Tectonites. McGraw-Hill, New York, 545p.

University of Leeds, Research Institute of African Geology, 1962. 6th Ann. Rept. on Scientific Results, 1960 - 1961 Session.

Voll, G., 1960. New Work in Petrofabrics. Liv. and Manchr. Geol. J., v. 2, pt. 3, pp. 503 - 567.

Walker, G. P. L. and Skelhorn, R. R., 1966. Some associations of acid and basic igneous rocks. Earth Sci. Rev., v. 2, pp. 93 - 109.

Walton, B. J. and Arnold, A. R., 1970. Plutonic Nodules in Lamprophyric Carbonatite Dykes near Frederikshåb, S. W. Greenland. Meddel. om Grøn., Bd. 190, nr. 5.

Wanless, R. K. and Loveridge, W. D., 1972. Rubidium-Strontium Isochron Age Studies, Report 1. Geol. Surv. Canada, Paper 72-23.

Wellman, H. W., 1955. New Zealand Quarternary Tectonics. Geol. Rund., v. 43, pp. 248 - 257.

Wanless, R. K., Stevens, R. D., Lachance, G. R. and Delabio, R. N., 1970. Age Determinations and Geological Studies. K/Ar Isotopic Ages, Report 9. Geol. Surv. Canada, Paper 69 - 2A.

__________, __________, __________and__________, 1972. Age Determinations and Geological Studies. K/Ar Isotopic Ages, Report 10. Geol. Surv. Canada, Paper 71 - 2.

Watt, W. S., 1969. The coast parallel dyke swarm of southwest Greenland in relationship to the opening of the Labrador Sea. Can. J. Earth Sci., v. 6, no. 5, pp. 1320 - 1321.

Watterson, J., 1968. Homogeneous deformation of the gneisses of Vesterland, S. W. Greenland. Meddel. om Grøn., Bd. 175, nr. 6.

Wheeler, E. P., 1933. A Study of Some Diabase Dikes on the Labrador Coast. J. Geol., v. 41, no. 4, pp. 418 - 431.

__________, 1935. An Amazonite Aplite Dike from Labrador. Am. Miner., v. 20, no. 1, pp. 44 - 49.

Williams, F. M. G., 1970. Map. Geol. Surv. Canada, Open File 42.

Williams, H., Turner, F. J. and Gilbert, C. M., 1958. Petrography. Freeman, San Francisco.

Windley, B., 1965. The composite net-veined diorite intrusives of the Julianehåb district, S. Greenland. Meddel. om Grøn., Bd. 172, nr. 8.

Wynne-Edwards, H. R., 1972. The Grenville Province. In: Variations in Tectonic Styles in Canada, (Price, R. A. and Douglas, R. J. W., editors), Geol. Assoc. Canada, Spec. Paper no. 11, pp. 263 - 334.

APPENDIX I

CALCULATION OF PURE-SHEAR TECTONIC STRAIN

Two methods of calculating the total pure-shear strain have been used. The methods are adapted from methods described in Ramsay (1967--page 193, method 1; and page 195, method 3).

Direct Measurement Method(Ramsay Method 1)

Several photographs were taken of appropriate conglomerate outcrops so that the plane of the short and intermediate pebble axes (XY plane) was in the plane of one photograph and the ZX or ZY plane in the plane of another photograph (Fig. I-1). Scales of the photographs were made equal and the lengths of the pebble axes measured directly from the photographs in arbitrary units to give mX, mY and mZ, the lengths of the short, intermediate and long axes. These were then averaged and the mean used to calculate the following (see Figs I-2, 4-25 and 4-26):-

$$a = \frac{mZ}{mY}$$

$$b = \frac{mY}{mX}$$

$$k = \frac{a-1}{b-1}$$

$$r = a + b - 1$$

(After Flinn, 1962 and Watterson, 1968)

Fig. I-1. Conglomerate member east of Round Pond showing tectonic elongation of pebbles and boulders. (Plane of photograph intermediate between XZ and YZ planes).

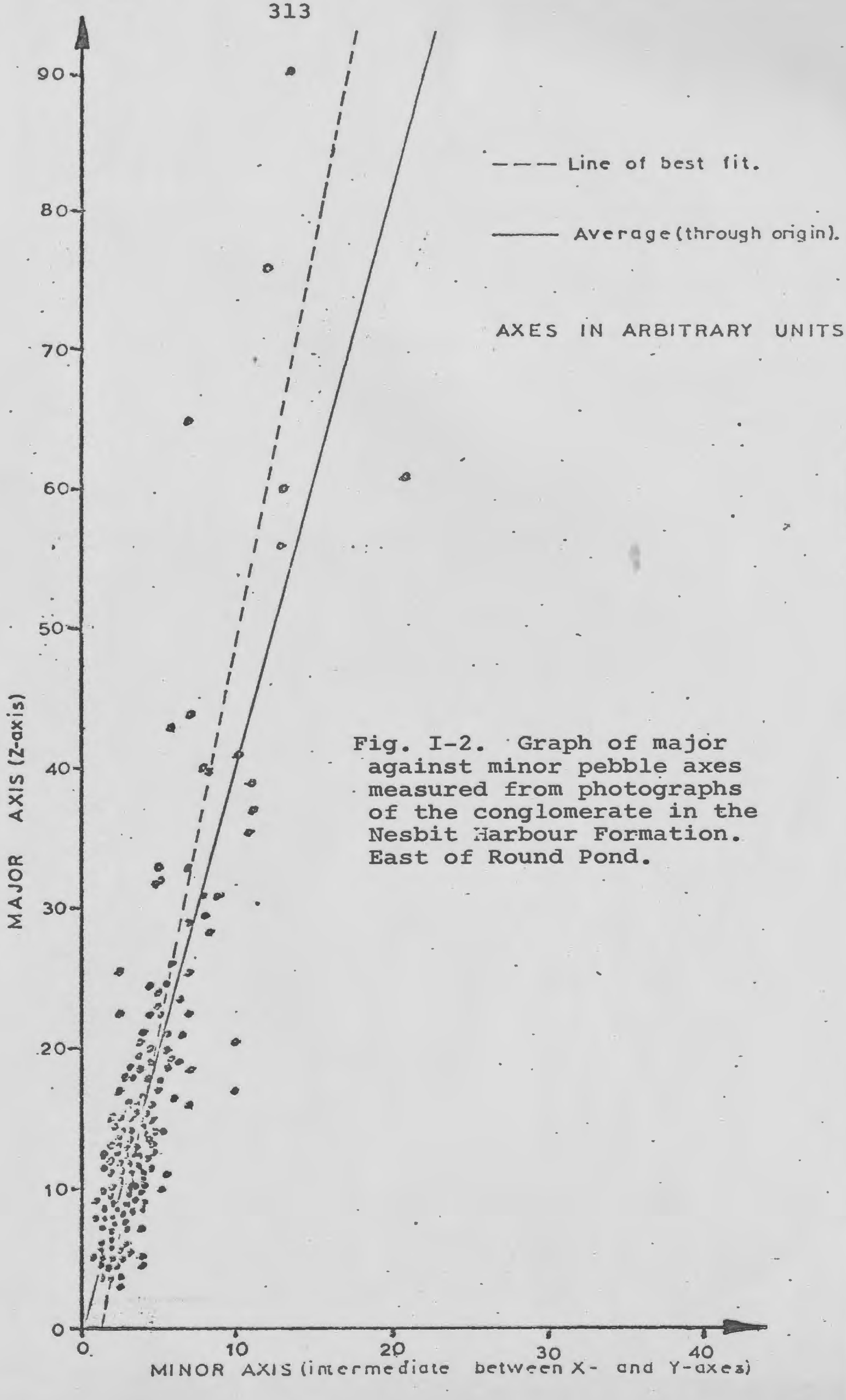


Fig. I-2. Graph of major against minor pebble axes measured from photographs of the conglomerate in the Nesbit Harbour Formation. East of Round Pond.

Note that this method will give a minimum value for the strain ellipsoid as some of the strain may have been taken up in the matrix, causing the matrix to flow around the pebbles. Also, some pebbles will be more competent than others and the pebbles are unlikely to have been initially spherical so that a straight-line plot of one axis against the other is not obtained (Fig. I-2). However, the average will be a straight-line plot provided there is no lithological relationship to pebble size. Note that the best-fit line through the points does not go through the origin and suggests an initial preferred orientation perpendicular to the Z-axis, possibly due to sedimentary layering or pre-D_2 deformation. The former interpretation is preferred as the initial pebble orientation is roughly parallel to the bedding, and also no evidence of earlier deformational effects has been seen in this location. The total effect of this preferred orientation on the strain ellipsoid, after averaging for all results, is negligible.

The method was also used in direct measurement of deformed lithophysae in the Doter Cove Rhyolite Member from Big Island (Fig. 2-41). As lithophysae very rarely show primary flattening or distortion (Ross and Smith, 1961) and also are rarely non-spherical except where in contact with other bodies (e.g. phenocrysts or other lithophysae), they are assumed to have been spherical prior to deformation.

A mafic dyke with vertically plunging F_2 folds which were mainly ptygmatic with only a small element of flow folding, occurs on the west coast of Wild Bight. The overall gabbro is a dark green to black medium-grained cumulate-textured rock in the center of the body, but is a dark green, fine-grained schistose amphibolite with elongate mafic clots up to 1/2 cm. (1/4 in.) long at the edges of the body. The border zone of the gabbro shows well developed tectonite fabrics, and a chill-margin, if originally present here, can no longer be discerned.

In thin-section the border zone is seen to consist of fine-grained polygonally recrystallised hornblende and feldspar grains, the hornblende generally occurring in strings or elongate aggregates. The coarser-grained central portions of the gabbro show relict cumulus textures of hornblende, which may have been derived from pyroxene, with an intercumulus of plagioclase. The plagioclase is zoned and is highly saussuritised in the cores, but fresh around the edges. Slight straining and grain-boundary migration has occurred around the edges and contacts with other plagioclase grains resulting in sutured boundaries of bytownite(?). The hornblende is largely altered to radiating masses of cummingtonite. Small grains of biotite which appear to have developed from some of the hornblende show further alteration to chlorite. Minor opaque minerals and sphene also occur.

At Pomiadluk Point a small 100 to 200 m. (300 to 600 ft.) long 20 m. (60 ft.) wide steeply tilted layered gabbroic body is intruded into the lower rhyolite member length of the dyke between two chosen points was measured, as was the direct distance between the points (which orientation was perpendicular to S_2 (Fig. I-3). The measurements indicate the rocks have been shortened in the X-direction to 1/3 or less of their original length.

Particle Distribution Method(Ramsay Method 3)

DESCRIPTION OF METHOD

In a rock in which particles have an initial random or known distribution, the post-tectonic distribution can be used to calculate the strain ellipsoid. In this thesis all calculations were made on the phenocrysts of quartz-feldspar porphyries (Fig. 2-8).

If the particles had an initial random distribution then the average distribution from any particle to its nearest neighbour will be the same for any direction, i.e., this distance may be thought of as the radius of an imaginary sphere and is inversely proportional to the concentration of particles. On deformation, this imaginary sphere will undergo distortion to an ellipsoid, the amount of distortion being related to the amount of strain, so that the calculated ellipsoid is a deformation ellipsoid. If the particles have size (e.g. phenocrysts) then the centers of the particles should be used for measurement. As this method does not rely on size or shape of the particles but only on the relative positions of their center-points it is independent of matrix

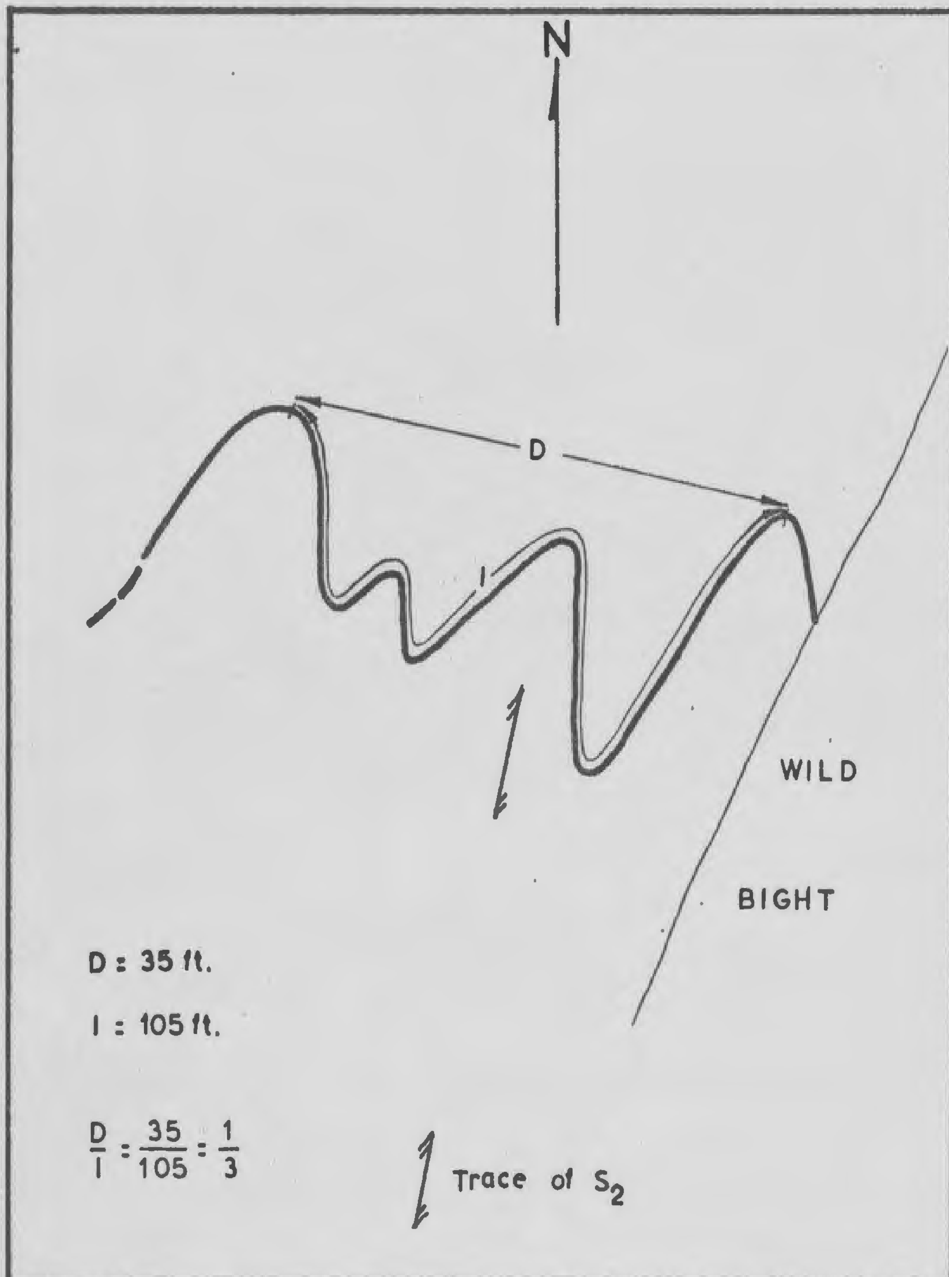


Fig. I-3. Sketch of folded post-D_1/pre-D_2 amphibolite dyke and calculation of amount of tectonic shortening. West coast of Wild Bight.

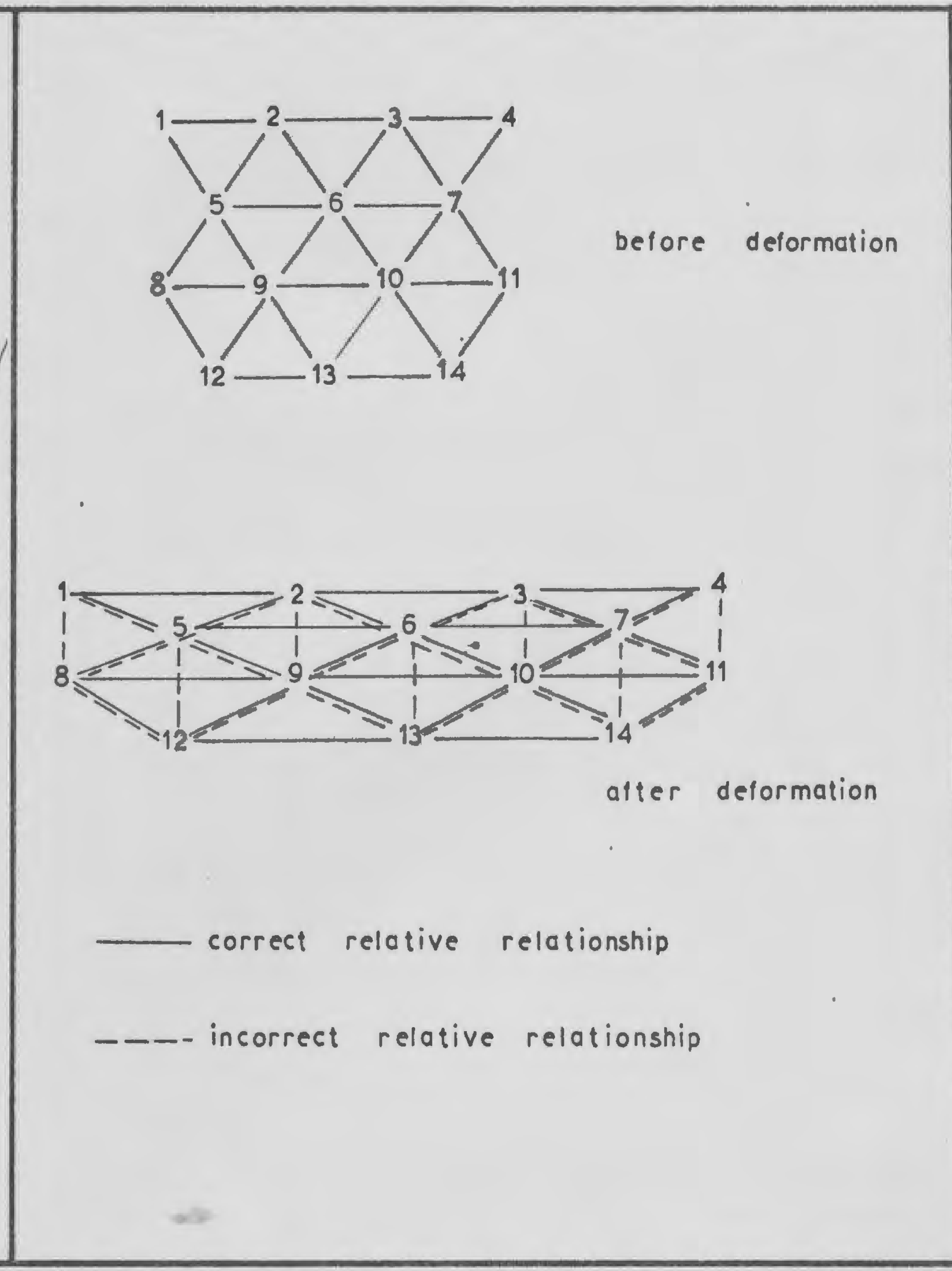


Fig. I-4. Schematic diagram showing origin of personal bias in measuring the relative positions of particles in a deformed rock.

flowage and will give a true value for the deformation of the specimen as a whole.

The choosing of pairs of points to be measured is largely subjective when no other deformational indicators are apparent, (though there is no mention of this in Ramsay (1967) as he deals with the simpler case of an aggregate of touching spheres.), as there is no way of deciding categorically which points should be considered as "nearest neighbours". However, the method appears to be more accurate than most when due care is taken. The author was unable to make a direct statistical evaluation of accuracy, but a personnal bias in measuring the relative positions of the particles was recognised. This bias is due to the difficulty in recognising which particles were nearest neighbours prior to deformation (Fig. I-4) and tends to make the values of both a and b smaller than their true value for large and small k-values, thereby increasing the error in k and decreasing the value of r.

The method employed by Ramsay involves cutting the rock parallel to one of the principal planes and plotting the distance between each pair of points chosen on that plane against the angle of the point relative to some reference direction on the plane. The average maximum and average minimum distances between points determined from the graph are then chosen as the long and short semi-axes of the ellipse. this is done for at least two faces of the rock, and if possible for a third face to act as a check. Values obtained will be for mX and mY, mY and mZ and mX and mZ, and a, b, k and r (Flinn, 1962; Watterson, 1968) may be calculated from these.

The above method, however, only uses the points about the maximum and minimum on the graph for the calculation of the ellipsoid. In order to improve the accuracy of the method by using all the points measured, a computer program has been developed and is included at the end of this section.

EXPLANATION OF COMPUTER PROGRAM

Note: For mathematical and computing reasons the terminology in this section--A, B, C, K, X, Y, etc.--bears no relation to the structural terminology as defined in the text).

The computer program is written on Fortran IV. The program is designed to calculate the ellipse constants A, B, C and K (see Fig. I-5 for definitions) by determining K, the ellipse constant, for an initial C, the foci coordinate, comparing this K with the average K obtained for the data-points, and changing C accordingly. This process is repeated until two successive values of C differ by less than a certain prechosen amount (ERAB).

Initial input consists of J, the number of data-points (each to its own data card) and ERAB, the maximum allowable difference between the final and semi-final values of C (Fig. I-7). This is followed by the data-points themselves, defined by an angle (ALFA) and a distance (D), (Fig. I-6).

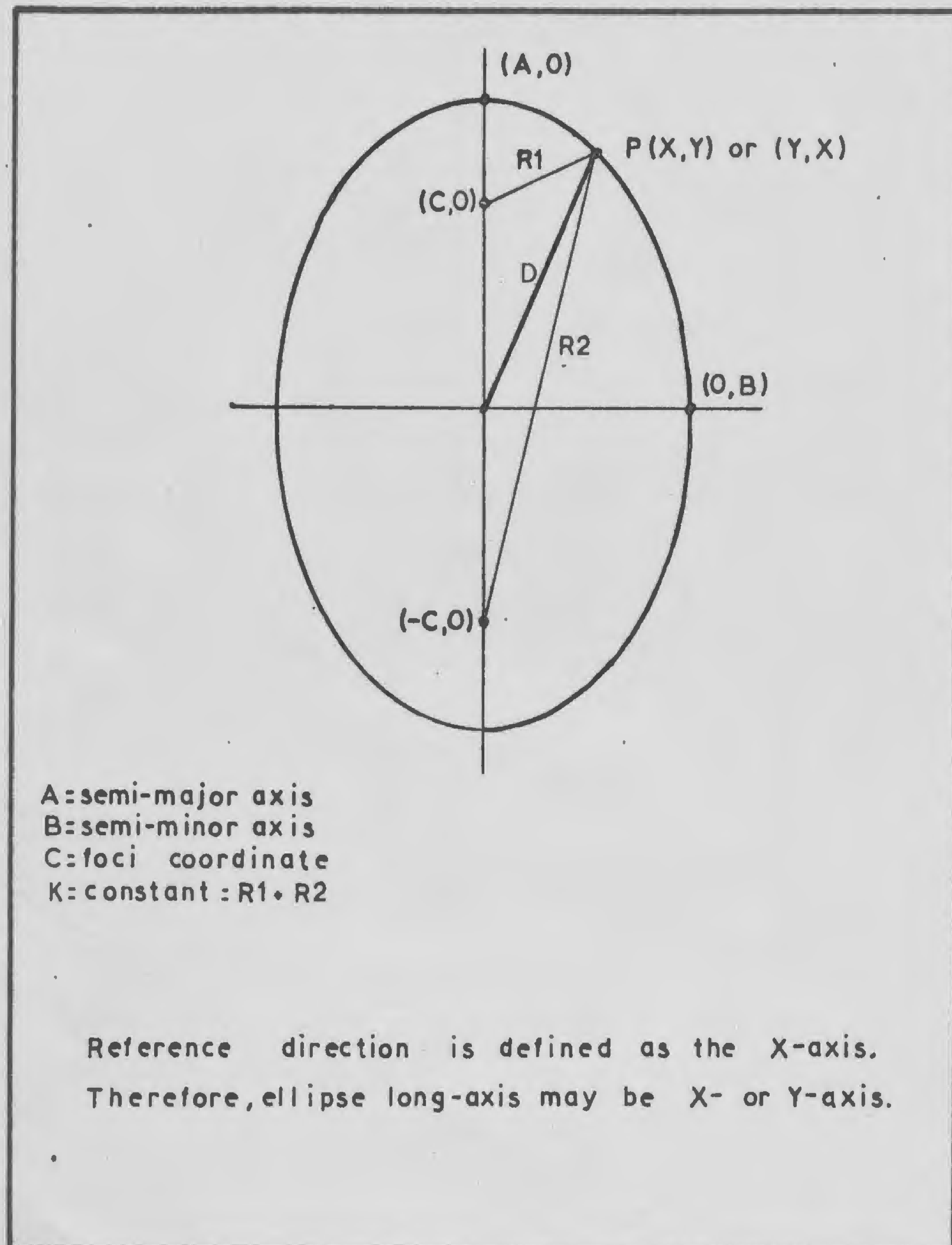


Fig. I-5. Diagram showing relationship of ellipse-constants A, B, C and K to the ellipse.

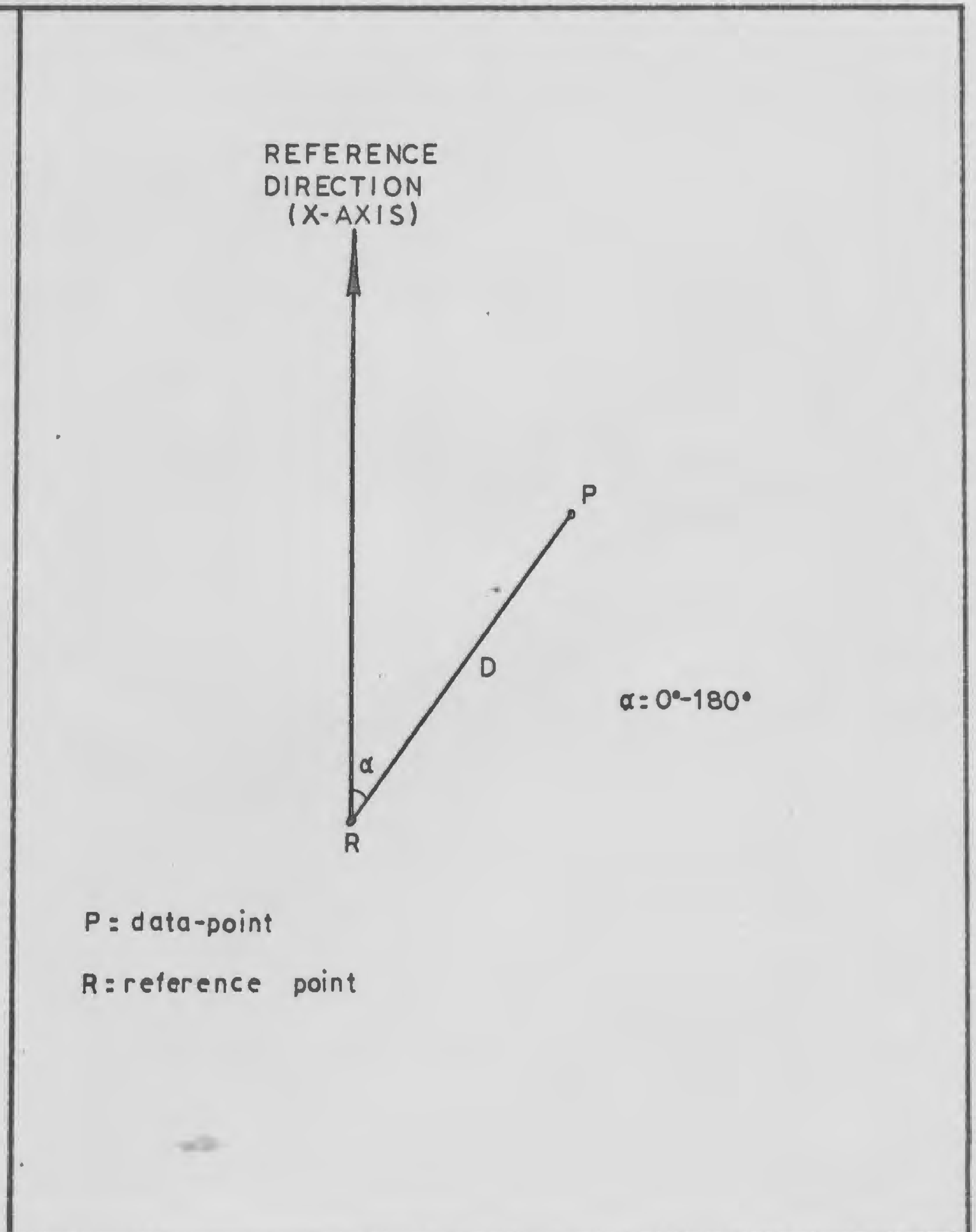


Fig. I-6. Diagram showing relationship of ALFA (α) and D to the reference and data points.

The ten largest D's are averaged to give a preliminary value for A (AVA) and the ten smallest to give a preliminary value for B (AVB). The X and Y coordinate values for each data-point P are calculated from D and ALFA. An initial value for C is calculated using the preliminary values A and B above.

Using the initial value of C above R1, R2 (Fig. I-5) and K (CONST) are calculated for each data-point. The various values of K are summed according to whether they lie about the X- or Y-axis, and the average value of K and a new value of A and B are determined.

The difference between the average of the sum of the value of K about the X-axis and those about the Y-axis is used to determine whether the ellipse foci should be separated further, or brought closer together. This new value for C is then used to recalculate K, and also A and B, which in turn are used to recalculate C. When the value of C is different from the previous value by an amount less than ERAB, then that value is taken as the final result. However if the value of C is less than ERAB it is assumed the long axis is perpendicular to the reference direction (X-axis).

A new preliminary value of C is set (equal to AVB, for convenience) and the previous calculations and iterations repeated using equations in which X and Y are interchanged, (the interchange of X and Y in the equations is necessary because the long axis is now assumed to be the Y-axis, i.e., perpendicular to the reference direction).

Final output consists of A, B, C and K, though prior to final output intermediate non-essential output, as marked, may be recorded for checking and cross-referencing.

Fig. 2-7

Simplified Computer Flow-diagram

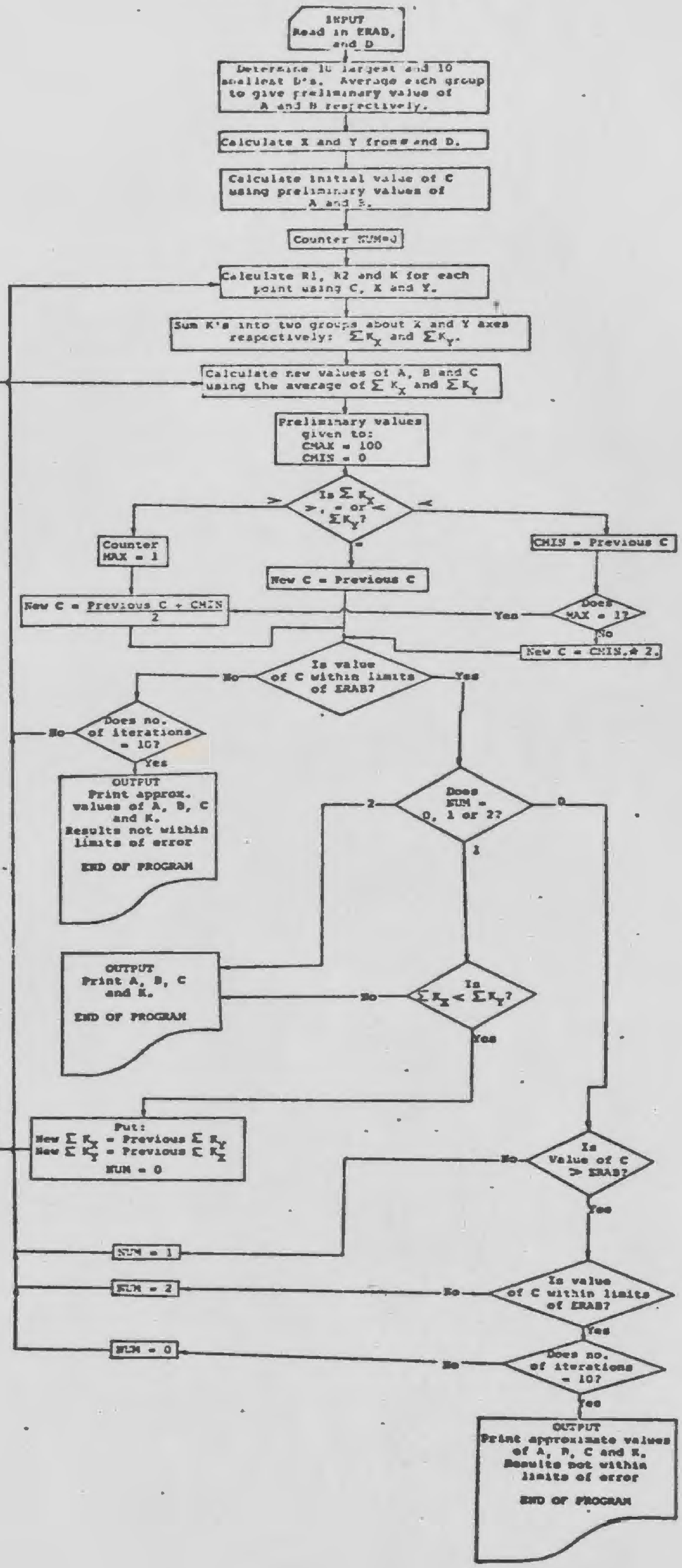


Fig.I-8. COMPUTER PROGRAM FOR DETERMINATION OF STRAIN ELLIPSE BY PARTICLE DISTRIBUTION METHOD

```
C           NOTE----------X-AXIS IS REFERENCE DIRECTION
C           All non-essential output marked ★
C INPUT, NO, OF DATA CARDS AND LIMITS OF ERROR FOR A AND B,
      DIMENSION DMN(100),DM(100),XE(100),YE(100),C(10),
     CALFA(100),DL(100)
      INTEGER CHCMX,CHCMN
★     PRINT 5700
 5700 FORMAT(1H , ' SAMPLE NUMBER                        ',//)
      READ 1102, J,ERAB
 1102 FORMAT (I3,F4,2)
★     PRINT 6102,J,ERAB
 6102 FORMAT (10X,'J=',I3,5X,'ERAB=',F4,2,//)
C
C READ IN ANGLE AND DIST, FOR EACH PAIR OF POINTS,
      N=0
 1201 N=N+1
      READ 1203, IALFA,D
 1203 FORMAT (I4,F6,2)
      ALFA(N)=IALFA
      DL(N)=D
      DM(N)=D
      DMN(N)=D
      IF(N,LT,J) GO TO 1201
C
C 10 LARGEST DM'S FOR AVA, & 10 SMALLEST FOR AVB
      M=0
      TOTDX=0,
      TOTDN=0,
 5303 M=M+1
      N=0
 5305 N=N+1
      IF(N,EQ,1) GO TO 5308
      IF(DM(N),LE,DMAX) GO TO 5311
 5308 DMAX=DM(N)
```

```
          CHCMX=N
          IF(N.EQ.1) GO TO 5312
     5311 IF(DMN(N).GE.DMIN) GO TO 5314
     5312 DMIN=DMN(N)
          CHCMN=N
     5314 IF(N.LT.J) GO TO  5305
          TOTDX=TOTDX+DMAX
          TOTDN=TOTDN+DMIN
          DM(CHCMX)=0.
          DMN(CHCMN)=TOTDX
          IF(M.LT.10) GO TO 5303
          AVA   =TOTDX/10.
          AVB   =TOTDN/10.
   C
   C CALCULATE X AND Y FOR EACH POINT
          N=0
     7201 N=N+1
          ANGL=ALFA(N)*3.1415927/180.
          YE(N)=DL(N)*SIN(ANGL)
          XE(N)=DL(N)*COS(ANGL)
     1212 PRINT 6200,ALFA(N),DL(N),ANGL,XE(N),YE(N)
     6200 FORMAT(2X,'ALFA=',F5.1,5X,'D=',F6.2,5X,'ANGL=',F6.2,5X,'X=',F6.2,
         C5X,'Y=',F6.2)
          IF(N.LT.J) GO TO 7201
   C
   C USE AVE A AND B TO FIND C (ELLIPSE FOCI COORD)
          PRINT 6800
     6800 FORMAT(/,10X,'ASSUME LONG AXIS PARALLEL TO REFERENCE DIRECTION',/)
          I=0
          C(1)=SQRT((AVA**2)-(AVB**2))
   C
   C USE C TO CALC R1 ,R2 & K FOR EACH P, AND AV. K'S
          MAX=0
          CMAX=100.
          CMIN=0.
          NO=0
     6900 NUM=0
```

serpentine and opaque minerals) is a distinct minor cumulate as are most of the opaque minerals.

The plagioclase-cumulate layers have plagioclase predominating with abundant small subhedral pigeonite prisms with hornblende cores poikilitically enclosed in them. Large euhedral hornblendes, which show all the features of dissolution and reaction mentioned earlier, decrease in quantity upwards from the hornblende-cumulate layer. Where pigeonite prisms are poikilitically enclosed in hornblende they show dissolution and reaction to form hornblende.

iii. Massive Gabbro Facies: The massive gabbro facies is a monotonous, coarse grained gabbro occurring along the coast south of Manak Bay beyond the diabase facies and complex marginal intrusions. It also occurs in the vicinity of Big Bight, and everywhere grades into the massive diabase facies which appears to be a finer grained marginal variation of the massive gabbro facies.

The rock is a plagioclase cumulate composed of 1 cm. to 5 cm. long, dark grey plagioclase laths with a black mafic-mineral intercumulus. The plagioclase is andesine-labradorite, and the intercumulus phase is pigeonite, basaltic hornblende, minor green hornblende, opaque minerals, olivine (now mainly antigorite?) and dark brown biotite. Apatite prisms are abundant. The hornblende is partly replacing the pigeonite.

(Fig. 3-10). The layering is only imperfectly developed and it could not be determined in the field whether it was due to crystal settling or to multiple magma intrusion. The gabbro consists of highly saussuritised plagioclase grains set in a tectonically deformed and metamorphosed aggregate of oriented hornblende, biotite and muscovite grains.

v. Other Early Syn-tectonic Intrusions: The gabbro at Pomiadluk Point is cut on its eastern side by a 30 cm. (1 ft.) wide quartz-arsenopyrite vein which shows imperfectly developed S_2 fabrics. The Long Island Gneiss east of Makkovik Bay is cut by amphibolite dykes with weakly developed S_2 fabric, which shows green copper-(malachite-) staining on the surface and contain accessory amounts of pyrite, chalcopyrite, chalcocite(?), magnetite and haematite (after pyrite and magnetite).

Intrusions Emplaced After The Main Deformation

None of the minor intrusions emplaced after the main deformation can be shown to have been emplaced prior to the end of tectonism. All intrusions emplaced after the main deformation will, therefore, be discussed in the section on post-tectonic intrusions.

POST-TECTONIC INTRUSIONS

Due to a lack of structural evidence most of the intrusions discussed in this section cannot be proven to be

```
 6503 NO=NO+1
      I=I+1
      N=0
      SUMKX=0.
      SUMKY=0.
      SKX=0.
      SKY=0.
      ICNTR=0
      JCNTR=0
 5502 N=N+1
      R1=SQRT(((XE(N)-C(NO))**2)+(YE(N)**2))
      R2=SQRT(((XE(N)+C(NO))**2)+(YE(N)**2))
      CONST=R1+R2
      IF((ALFA(N).GT.45).AND.(ALFA(N).LT.135)) GO TO 6501
      SUMKX=SUMKX+CONST
      ICNTR=ICNTR+1
      GO TO 6502
 6501 SUMKY=SUMKY+CONST
      JCNTR=JCNTR+1
 6502 IF(N.LT.J) GO TO 5502
      CONST=(SUMKX+SUMKY)/J
      SKX=SUMKX/ICNTR
      SKY=SUMKY/JCNTR
      AVA=CONST/2
      AVB=SQRT(((CONST/2)**2)-(C(NO)  **2))
      PRINT 6001 , I,AVA,AVB,C(NO)  ,CONST,SUMKX,SUMKY,SKX,SKY,CMIN,CMAX
 6001 FORMAT(/,8X,'I=',I4,5X,'AVA=',F8.3,5X,'AVB=',F8.3,5X,/,8X,
     C'C=',F8.3,5X,'CONST K =',F8.3,/,8X,'SUMKX=',F10.3,5X,'SUMKY=',
     CF10.3,/,8X,'AVE. SUMKX(SKX)=',F10.3,5X,'AVE. SUMKY(SKY)=',
     CF10.3,/,8X,'CMIN=',F8.3,5X,'CMAX=',F8.3,//)
      IF(NUM-1) 5333,5400,1705
C
C TO DETERMINE HOW C SHOULD BE CHANGED TO BRING IT NEARER TO BEST-FIT VALUE
 5333 L=0
      IF(  SKX-  SKY) 6507,6512,6505
 6507 CMAX=C(NO)
      MAX=1
 6510 C(NO+1)=(CMAX +CMIN__ )/2
```

```
      GO TO 6509
 6505 CMIN=C(NO)
      IF (MAX.EQ.1) GO TO 6510
      C(NO+1) = CMIN*2
      GO TO 6509
 6512 C(NO+1)=C(NO)
 6509 IF(C(NO+1).GT.ERAB) GO TO 5331
      NUM=1
      GO TO 6503
 5331 IF(ABS(C(NO+1)-C(NO)).GT.ERAB) GO TO 5332
      NUM=2
      GO TO 6503
 5332 IF(NO.LT.9) GO TO 6900
      GO TO 5326
C
C CALCS TO CHECK IF LONG AXIS IS POSSIBLY PERPENDICULAR TO REF. DIR.,
C DESPITE INDICATIONS FROM 10 LARGEST AND 10 SMALLEST QM'S
 5400 IF(  SKX-  SKY) 6600,5401,5401
 6600 C(1)=AVB
      PRINT 6613
 6613 FORMAT(/,1H ,'LONG AXIS IS PERPENDICULAR TO REFERENCE DIRECTION')
      PRINT 6801, C(1)
 6801 FORMAT(/,5X,'PRELIMINARY VALUE OF C CHOSEN AS ',F8.3,/)
      MAX=0
      CMAX=100.
      CMIN=0.
      NO=0
 6603 NUM=0
 5603 N=0
      NO=NO+1
      I=I+1
      SUMKX=0.
      SUMKY=0.
      SKX=0.
      SKY=0.
      ICNTR=0
      JCNTR=0
```

```
 5602 N=N+1
      R1=SQRT(((YE(N)-C(NO))**2)+(XE(N)**2))
      R2=SQRT(((YE(N)+C(NO))**2)+(XE(N)**2))
      CONST=R1+R2
      IF((ALFA(N).GT.45).AND.(ALFA(N).LT.135)) GO TO 6601
      SUMKX=SUMKX+CONST
      ICNTR=ICNTR+1
      GO TO 6602
 6601 SUMKY=SUMKY+CONST
      JCNTR=JCNTR+1
 6602 IF(N.LT.J) GO TO 5602
      CONST=(SUMKX+SUMKY)/J
      SKX=SUMKX/ICNTR
      SKY=SUMKY/JCNTR
      AVA=CONST/2
      AVB=SQRT(((CONST/2)**2)-(C(NO)  **2))
      PRINT 6001 , I,AVA,AVB,C(NO)  ,CONST,SUMKX,SUMKY,SKX,SKY,CMIN,CMAX
      IF(NUM.EQ.1) GO TO 6705
C
C TO CHANGE C TO BEST FIT
      IF(  SKY-  SKX) 6607,6612,6605
 6607 CMAX=C(NO)
      MAX=1
 6610 C(NO+1)=(CMAX +CMIN   )/2
      GO TO 6609
 6605 CMIN=C(NO)
      IF(MAX.EQ.1) GO TO 6610
      C(NO+1)=CMIN*2
      GO TO 6609
 6612 C(NO+1)=C(NO)
 6609 IF(ABS(C(NO+1)-C(NO)).GT.ERAB) GO TO 5330
      NUM=1
      GO TO 5603
 5330 IF(NO.LT.9) GO TO 6603
      GO TO 5326
 6705 C(NO+1)=C(NO)
      IF(C(NO+1).LT.ERAB) GO TO 1211
      L=1
```

```
      GO TO 1705
 1211 C(NO+1)=0.
      L=2
      GO TO 1705
C
C OUTPUT.
 5401 PRINT 5402
 5402 FORMAT (5X,'STRAIN-FIGURE IS A CIRCLE')
      GO TO 1705
 5326 PRINT 5700
      PRINT 1701
 1701 FORMAT(1H ,'RESULTS NOT WITHIN LIMITS OF ERROR')
      PRINT 5702, ERAB
 5702 FORMAT(1H ,'LIMITS OF ERROR=',F6,2)
      PRINT 1703, AVA   , AVB   ,C(NO+1),CONST
 1703 FORMAT(1H ,'LONG SEMI-AXIS APPROX.=',F8.3/1H ,
     C'SHORT SEMI-AXIS APPROX.=',F8.3/1H ,'FOCI COORD C APPROX.=',
     CF8.3/1H ,'CONST K APPROX.=',F8,3)
      GO TO 1707
 1705 PRINT 5700
      PRINT 1706, AVA   , AVB   ,C(NO+1),CONST
 1706 FORMAT(1H ,'LONG SEMI-AXIS =',F8.3/1H ,'SHORT SEMI-AXIS =',F8.3/
     C1H ,'FOCI COORD C =',F8.3/1H ,'CONST K =',F8.3)
      PRINT 5702, ERAB
 1707 IF(L-1) 5324,5322,5328
 5322 PRINT 5323
 5323 FORMAT(//,1H ,'LONG AXIS PERPENDICULAR TO REFERENCE DIRECTION',//)
      GO TO 5327
 5324 PRINT 5325
 5325 FORMAT (//,1H ,'LONG AXIS PARALLEL TO REFERENCE DIRECTION',//)
      GO TO 5327
 5328 PRINT 5402
 5327 STOP
      END
```

APPENDIX II

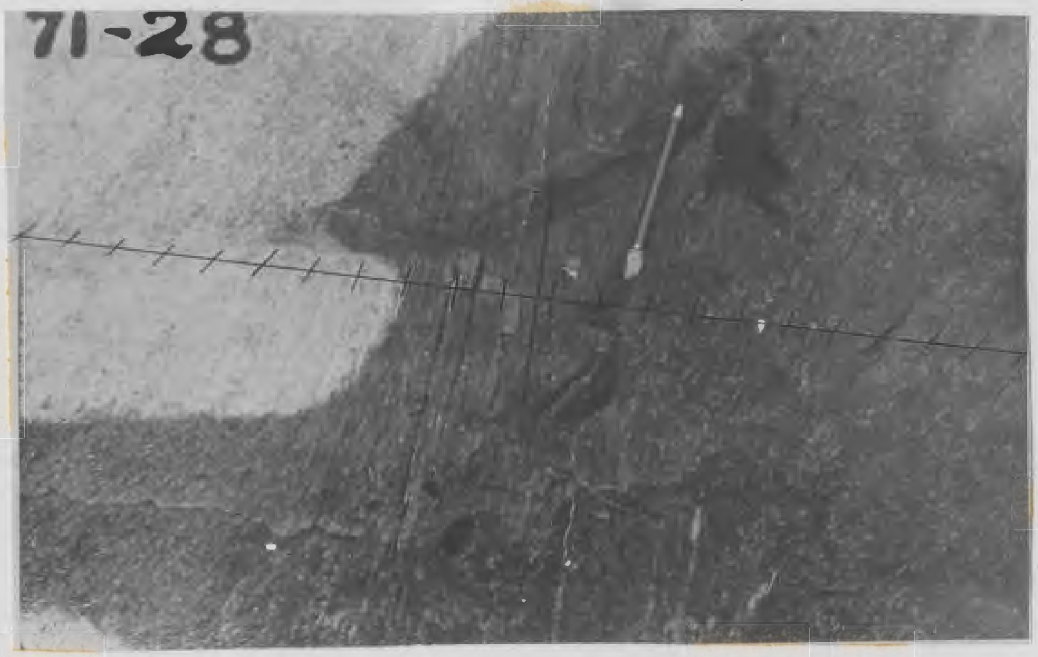
SIMPLE-SHEAR BELTS

Small 1/2 to 1 m. (1 1/2 to 3 ft.) wide shear belts (Ramsay et al., 1970) were recognised in the quartz-porphyritic rhyolite on the east coast of Wild Bight. All the shear belts recognised are oriented in the same direction and no complementary belts were seen. The shear belts were photographed and their orientation, and the orientation of the S-fabric outside the shear belts in the host-rock (the "external" fabric), were measured. As there is no evidence of shearing parallel to the S-fabric in the belts, they are considered not to be normal (positive) kink-bands (Dewey, 1965 and 1969) but simple-shear structures.

The external S-fabric is parallel to the S_2 fabric throughout the area and is therefore considered to be S_2. It is defined by flattened quartz and feldspar phenocrysts in which there is no evidence of rotation, shearing (as opposed to pure flattening), or transposition to a later fabric. This weak external S_2-fabric increases in intensity towards the shear belt, and its orientation changes towards parallelism with the shear belt (Fig. II-1).

The change in orientation and intensity of the S-fabric could be due to either development at the same time as the shear belts, or rotation and enhancement of the S_2 fabric by the shear belts at a later (post-D_2) stage.

The S_2 fabric in the vicinity of the shear belts,



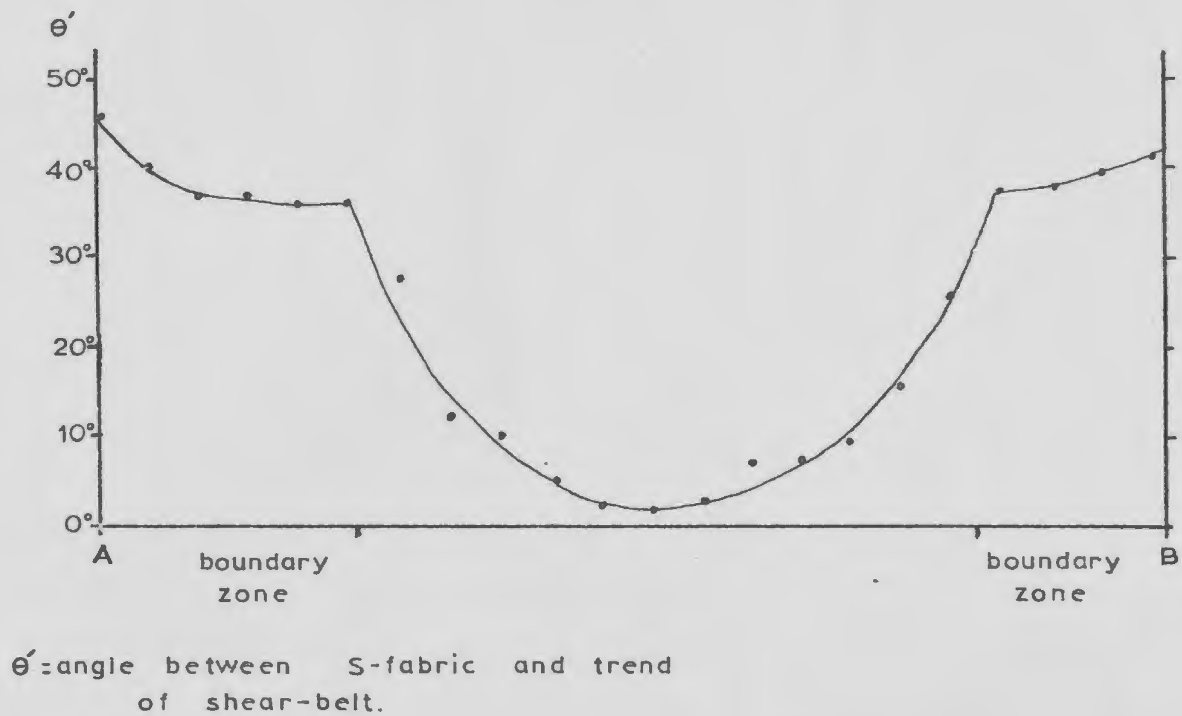


Fig. II-1. Photograph of small simple-shear belt on west coast of Wild Bight, and graph of θ′ (Ramsay and Graham, 1970). Pencil is 15 cm. long.

but outside them, strikes 205° and dips 65°E. The shear belts strike 140° and dip vertically, and there is therefore a 68° angle between the two (Fig. II-2). The external maximum stress that formed the S_2 fabric is therefore oriented at 22° to the shear belt (Fig. II-3). Although it is commonly assumed that shear belts (as opposed to the internal fabric) must form at 45° to the maximum stresses, this is seldom so in practice where angles of 25° to 35° are most likely under brittle-fracture conditions (Friedman, 1964). This factor does not, therefore, itself indicate a post-D_2 age for the belts.

If the external S_2 fabric developed contemporaneously with the fabric within the shear-belt, then the two fabrics would be expected to be continuous across the boundary of the belt. However, if the shear-belt is later, and reoriented and possibly enhanced the earlier S_2 fabric, one would expect an inflection or discontinuity of curvature at the shear-belt boundary. A plot of the orientation of the fabric across the shear-belt on a line perpendicular to the center-line of the belt has been made (Fig. II-1) and a significant break in curvature of the graph is seen to occur at approximately $\theta' = 35^\circ$. This approximately 10 cm. (4 in.) wide boundary zone (Fig. II-3) is thought to be a zone of recrystallisation and reorientation of the S_2 fabric to the internal S-fabric of the shear-belt. (A second possible inflection at $\theta' = 10^\circ$ is less than the error of measurement

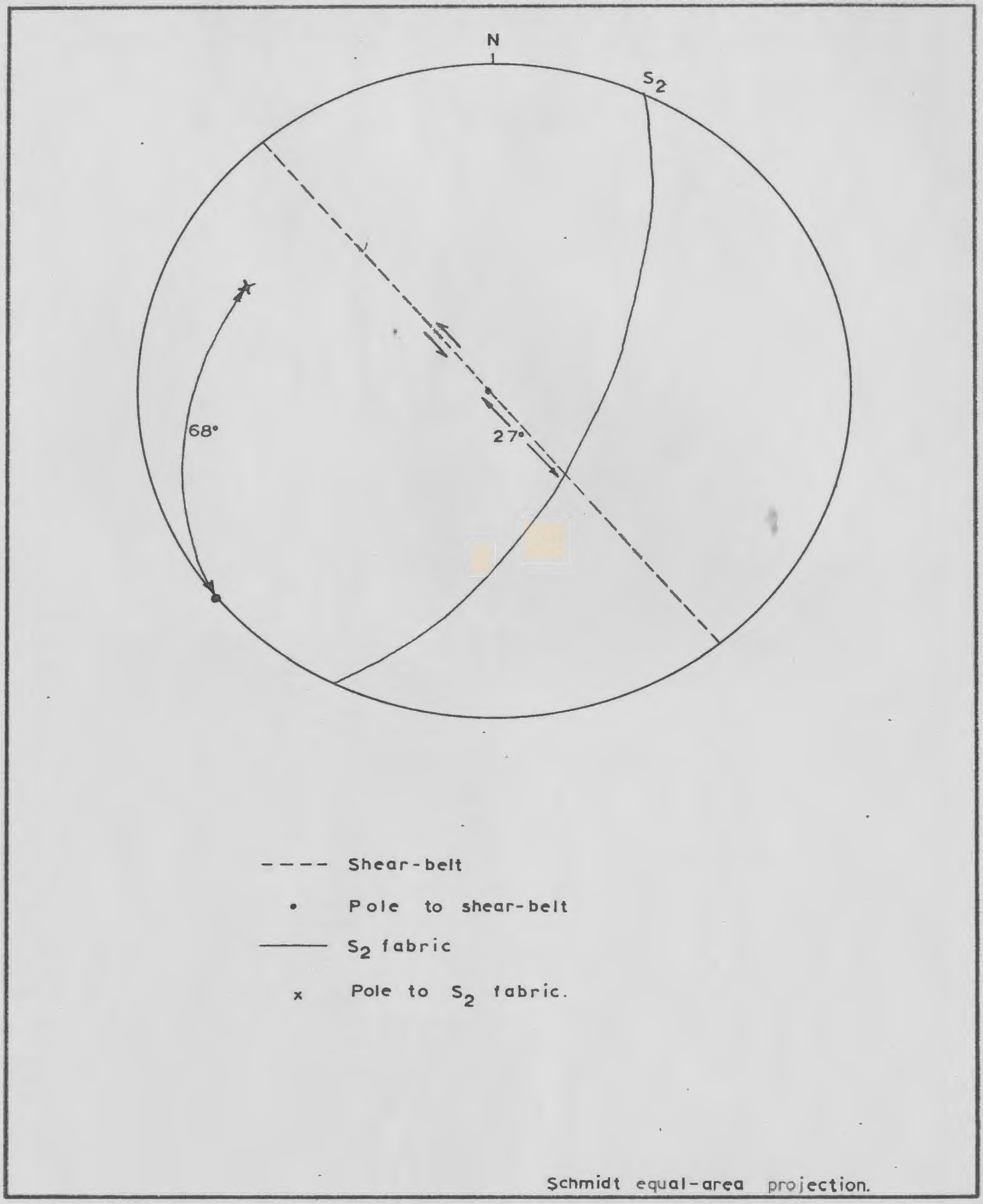


Fig. II-2. Stereographic projection of relationship of the S_2 fabric to the shear-belts on the east coast of Wild Bight.

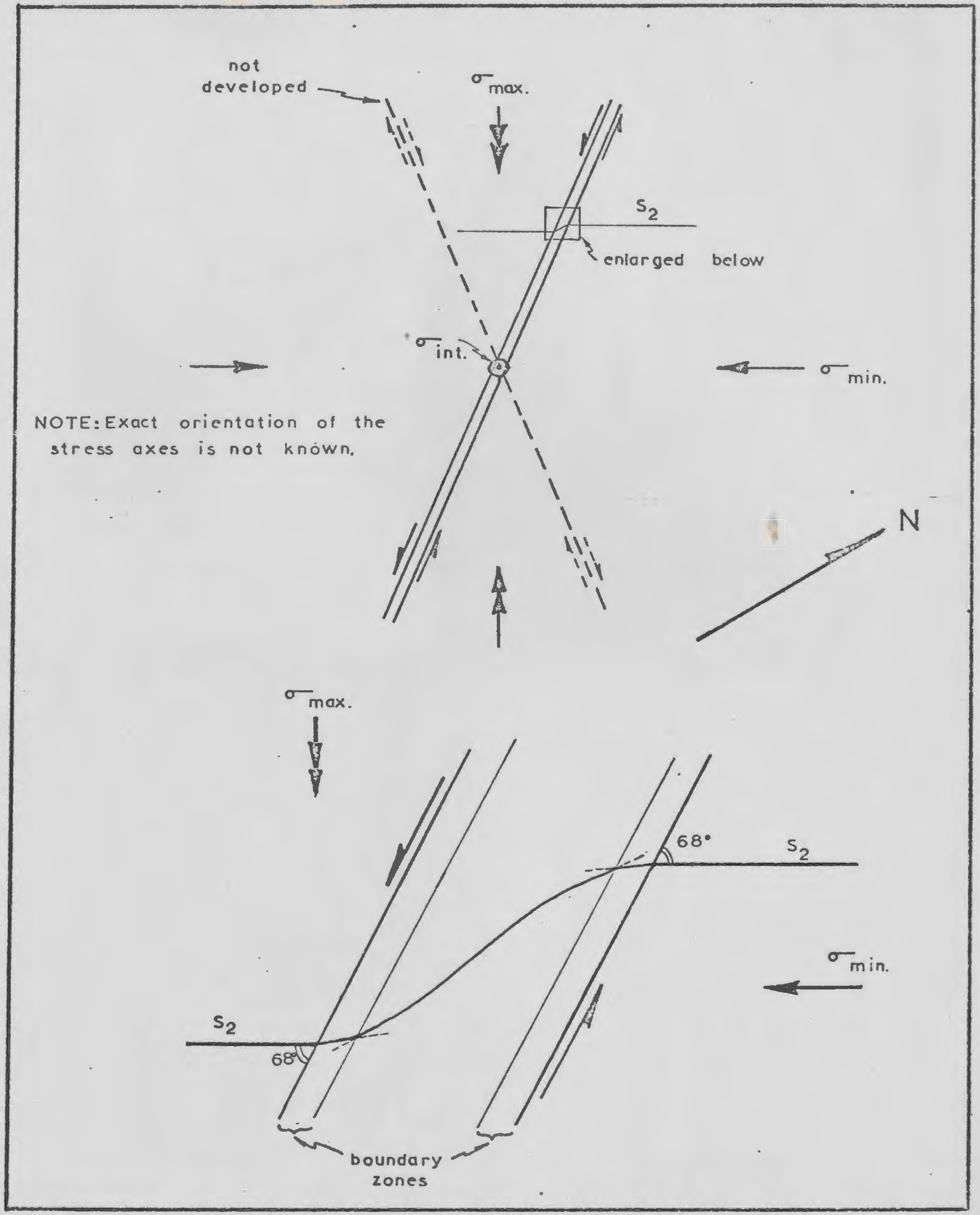


Fig. II-3. Schematic diagram showing the relationship of the shear-belts to the strain-axes of S_2 fabric.

and will therefore be ignored).

The amount of simple-shear deformation cannot be calculated as the internal fabric is the result of recrystallisation of phenocrysts and will therefore have a composite orientation (Ramsay, 1967, and Oertel, 1970). Even if the internal plane of flattening could be determined, its initial orientation of development with respect to the shear-belt is not necessarily known, and may be anywhere from 40° to 80° (the complements of the angles for tension gashes; Hancock, 1972) though it is theoretically determined to be at 45° (Ramsay, 1967; Ramsay and Graham, 1970). As the initial orientation of the internal plane of flattening is unknown and shear-belts are not necessarily propagated at 45° to the maximum stress, the orientation of the D_3 stress axes cannot be determined, though they must have been oriented approximately as shown in Fig. II-3.

Fig. 3-10. Layering in small gabbro body at Pomiadluk Point.

Fig. 3-11. Photomicrograph showing indistinctly bounded feldspar grains (dark and light blotches) crowded with small poikilitically enclosed augite grains. Pigeonite-porphyritic diabase dyke on south coast of Manak Bay. Crossed-nicols, X10.

APPENDIX III

AXIAL-DISTRIBUTION ANALYSIS (A.V.A.) DIAGRAMS FOR QUARTZ

Axial-distribution analysis (Achsensverteilungs-analysen--A.V.A: Ramsauer, 1941; Sander, 1950; as described in Turner and Weiss, 1963) diagrams have been made of the c-axis orientations of quartz sub-grains in recrystallised quartz phenocrysts which have undergone D_2 deformation. The phenocrysts occur in a sample of the quartz-porphyritic rhyolite of the Makkovik Formation (Fig. III-1) in which the S_2 fabric strikes 025° and dips 60° eastward, and the L_2 fabric plunges 5° northwards. The c-axis orientations were measured in ten phenocrysts, of which the four largest are shown (Fig. III-3, III-4, III-5 and III-8). Stereonets were compiled of c-axis orientations of all ten phenocrysts, of which four are shown (Figs. III-6 and III-9). Although the initial orientation of the phenocryst c-axes is not known, they are assumed not to have been parallel to one another. The stereonet patterns produced from each phenocryst aggregate show preferred orientations with one major concentration near the Y-axis of the strain ellipsoid, a lesser concentration between the Y- and Z-axes and a third possible weak concentration about the X-axis. There is a possible weak small-circle (conical) distribution symmetrical about the concentration near the Y-axis, and passing through the other two concentrations. These concentrations are not symmetrically orientated with respect to the strain ellipsoid, but are similarly

Fig. III-1. Sample of quartz-porphyritic rhyolite from which thin-sections were made for AVA diagrams. (XY-face shown). Scale bar = 1 cm.

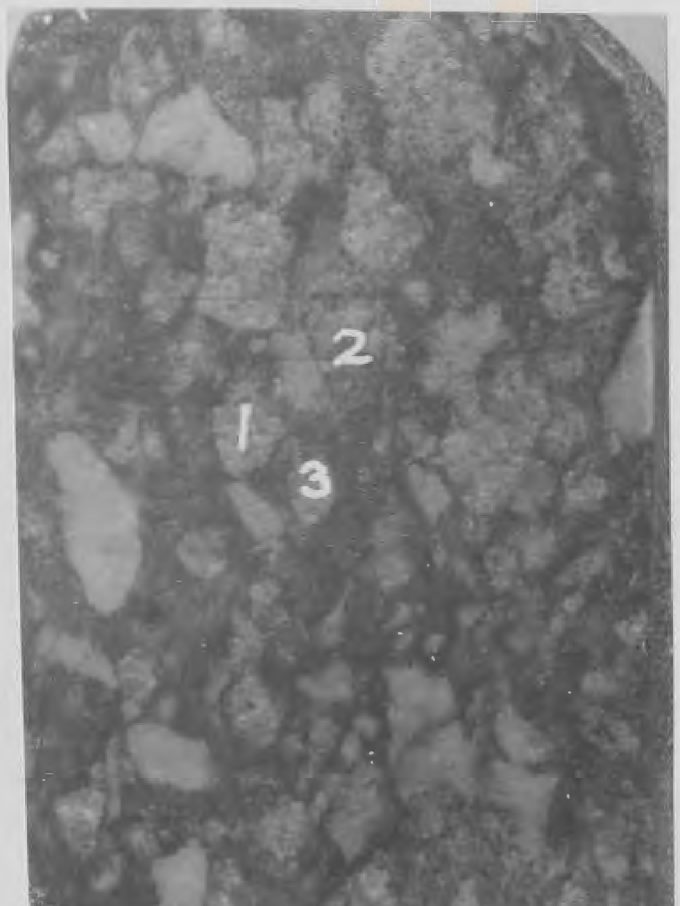


Fig. III-2. Thin-section (in ZY plane) showing quartz-phenocrysts numbers 1, 2 and 3.

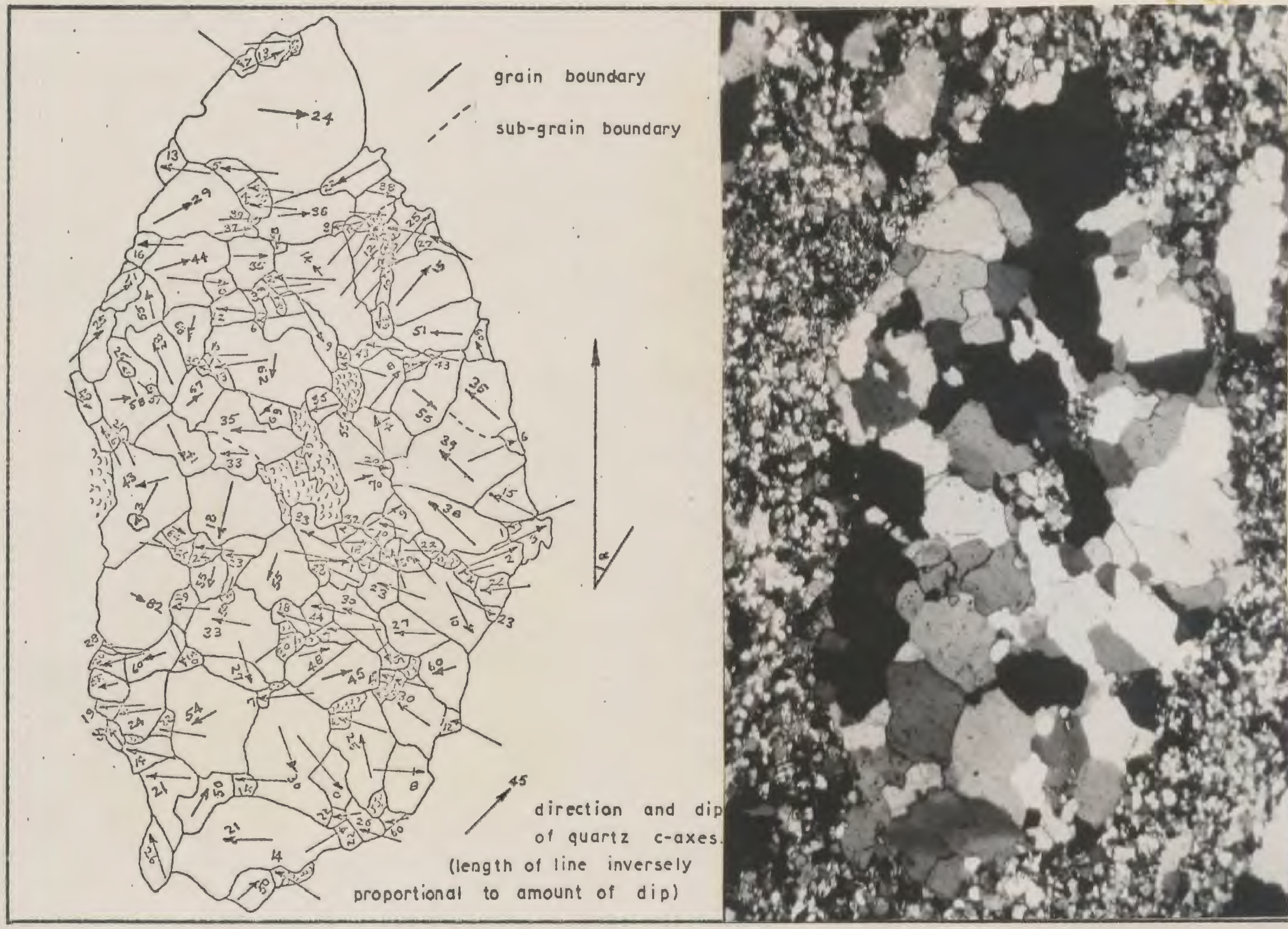


Fig. III-3. A.V.A. diagram and photomicrograph of quartz phenocryst 1. From quartz-porphyritic rhyolite between Ford's Bight and Wild Bight. Crossed nicols, X36.

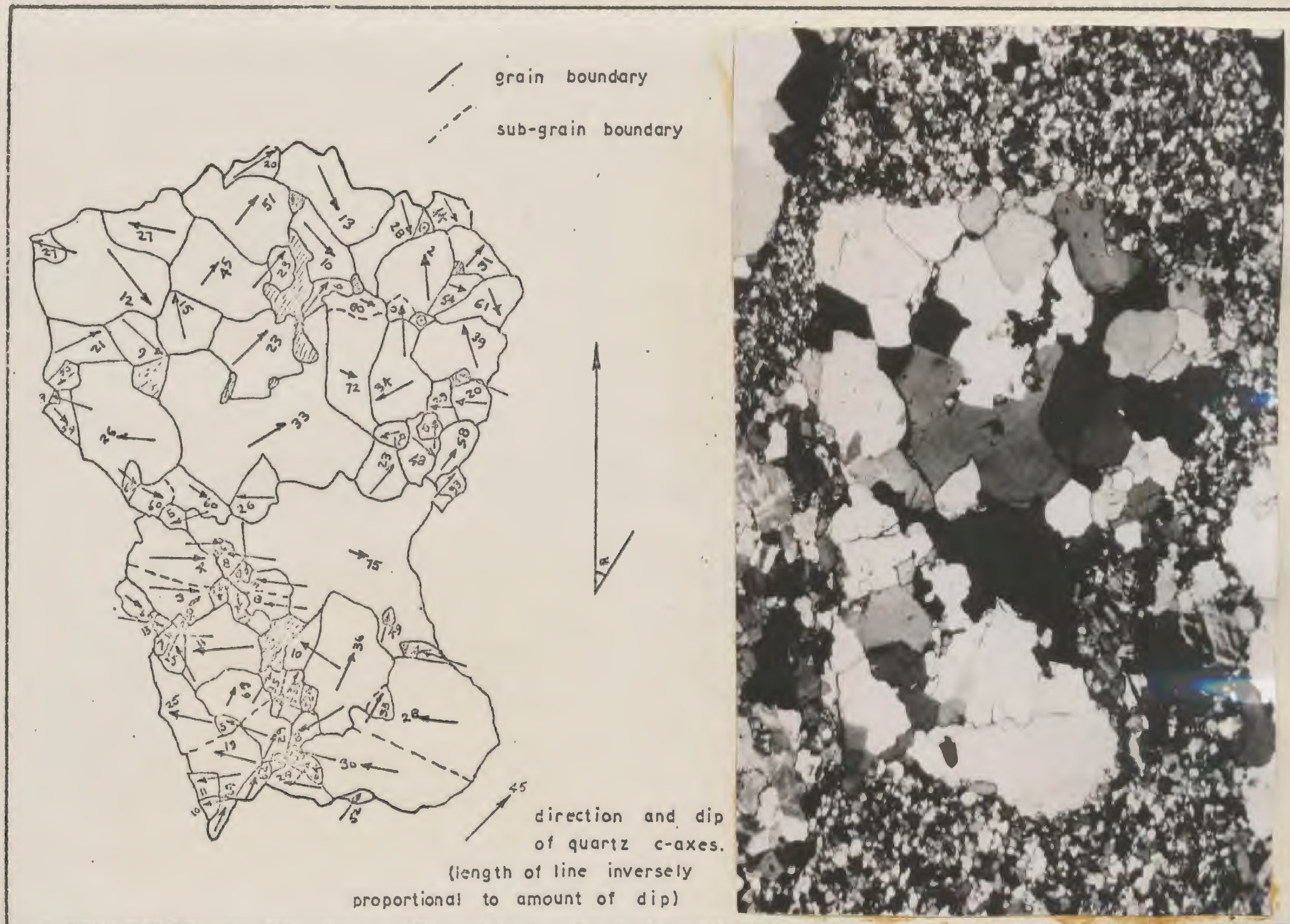


Fig. III-4. A.V.A. diagram and photomicrograph of quartz phenocryst 2. From quartz-porphyritic rhyolite between Ford's Bight and Wild Bight. Crossed nicols, X36.

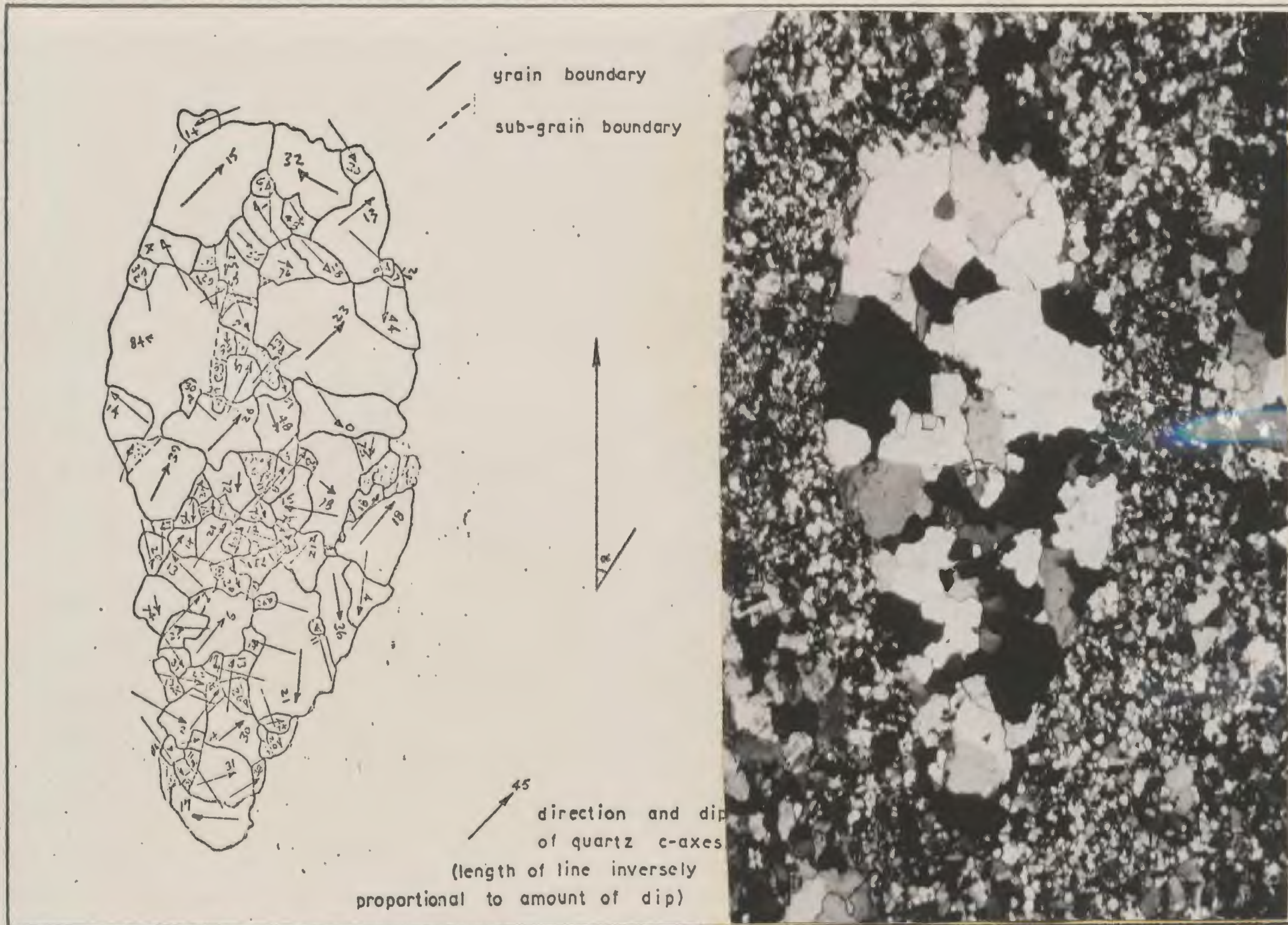


Fig. III-5. A.V.A diagram and photomicrograph of quartz phenocryst 3. From quartz-porphyritic rhyolite between Ford's Bight and Wild Bight. Crossed nicols, X36.

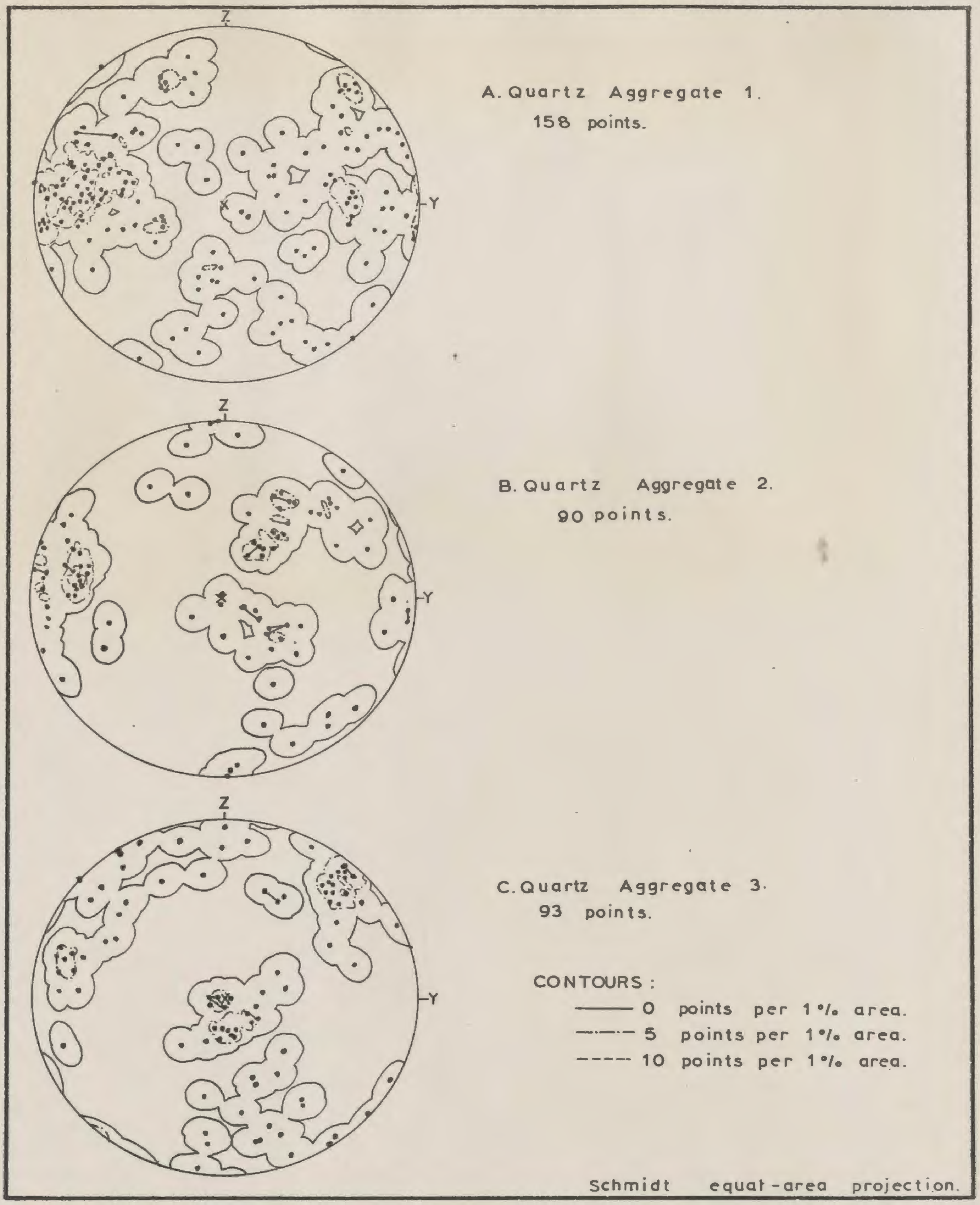


Fig. III-6. Stereographic projection of quartz c-axes from recrystallised phenocrysts in rhyolite. (Points representing subgrains of a single large grain are joined by lines.)

oriented relative to one another, indicating the c-axis orientations of the sub-grains. A similar conclusion was reached by Shelley (1971) from investigation of a deformed quartz porphyry from Scotland.

No statistical tests of significance of the preferred orientations have been made, and the number of points plotted for each phenocryst is too small to be statistically acceptable on its own. However, when the stereonets are compared to each other and to the composit stereonet, concentrations of points are seen to occur in the same relative positions in most of the plots. In the field, there is no evidence of folding of the porphyry in the vicinity, nor is there any evidence of local variation in strain orientation, or type or amount of strain. In hand-specimen (Fig. III-1) and thin-section (Figs. III-2 and III-7) the rock has an apparently homogeneous fabric, and since all stereonets with the same relative orientation of the X-, Y- and Z- axes to one another were made from phenocrysts in the same thin-section, the possibility of inhomogeneity of fabric is considered negligible. The comparison of concentrations in one stereonet with those in another is therefore considered a valid test of significance (Flinn, 1958b). Furthermore, the same concentrations are seen in a stereonet plot from a phenocryst in a different thin-section perpendicularly oriented with respect to the previous ones (Figs. III-9a and b). A combined plot of four of the ten phenocrysts was therefore

Fig. III-7. Thin-section (in ZX plane) showing quartz-phenocryst number 4.

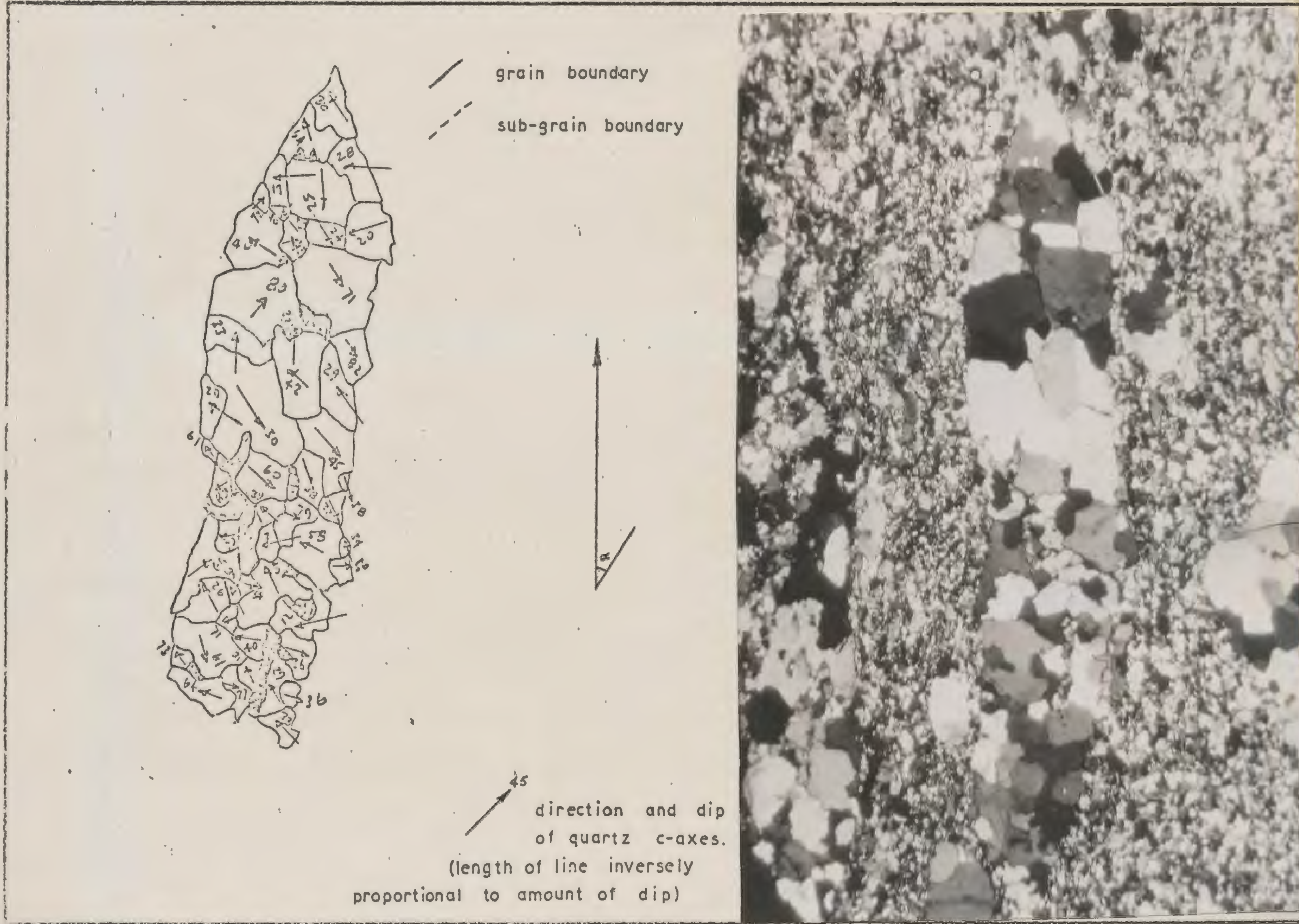


Fig. III-8. A.V.A. diagram and photomicrograph of quartz phenocryst 4. From quartz-porphyritic rhyolite between Ford's Bight and Wild Bight. Crossed nicols,

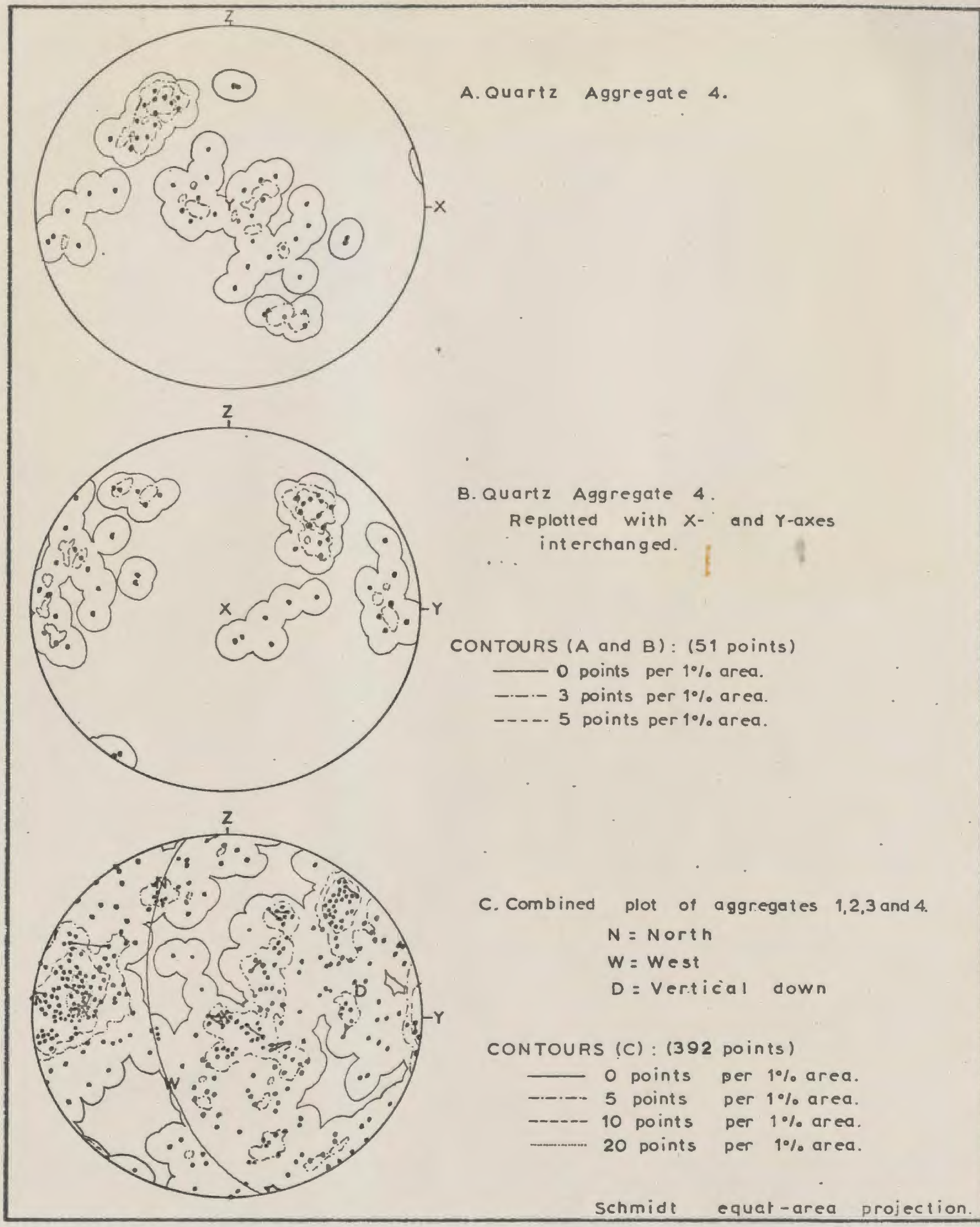


Fig. III-9. Stereographic projections of quartz c-axes from sub-grains in recrystallised quartz phenocrysts. (X, Y and Z are the D_2 strain axes. Sub-grains of a single strained grain are joined by a line).

post-tectonic (c.f. Chapters 4 and 6), but are included here for lack of contrary evidence.

Diabase/Grampus Cove Gneiss Net-vein Dyke:

A net-vein diabase/granite dyke occurs in the Grampus Cove Gneiss. The dyke consists of 1 cm. to 10 cm. (1/2 to 5 in.) diameter fine-grained diabase blebs and blocks set in a medium-grained salt-and-pepper textured biotite and hornblende granite (s.l.). The granite vein/host material grades into the enclosing gneiss. In thin-section the diabase is seen to consist of brown biotite, green hornblende and plagioclase, whereas the granitic portion is a hypersthene quartz monzonite composed of:

quartz	20 - 30%
K-feldspar	30 - 40%
plagioclase	30 - 40%
hypersthene, brown biotite, green hornblende	5 - 10%

with accessory amounts of opaque minerals and euhedral zircon. The K-feldspar is orthoclase which has partially inverted to perthitic microcline. Myrmekitic intergrowth is common on the borders of the K-feldspar. The plagioclase is highly saussuritised in the cores of the larger grains, but the rims are generally unaltered suggesting the cores are probably andesine/labradorite in composition which made (Fig. III-9c), the other six stereonets not being included as they are oriented in different principal planes with respect to the D_2 strain-axes.

The major c-axis concentration is oriented approximately 25° from the Y- direction in the XY-plane and approximately 10° from the Y- direction in the YZ-plane (Fig. III-9c). Although the orientation of the reference axes in the rock, and therefore in the thin-sections, could easily have been 5° off the true direction, and possibly even 10° off, due to the coarseness of the fabric, they could not be as much as 25° off the true directions as determined from the fabric of flattened and elongated phenocrysts (Fig. III-1). Furthermore, there appears to be a very weak concentration similarly oriented with respect to the X-axis, and the concentration between Y and Z is also not on the YZ-plane.

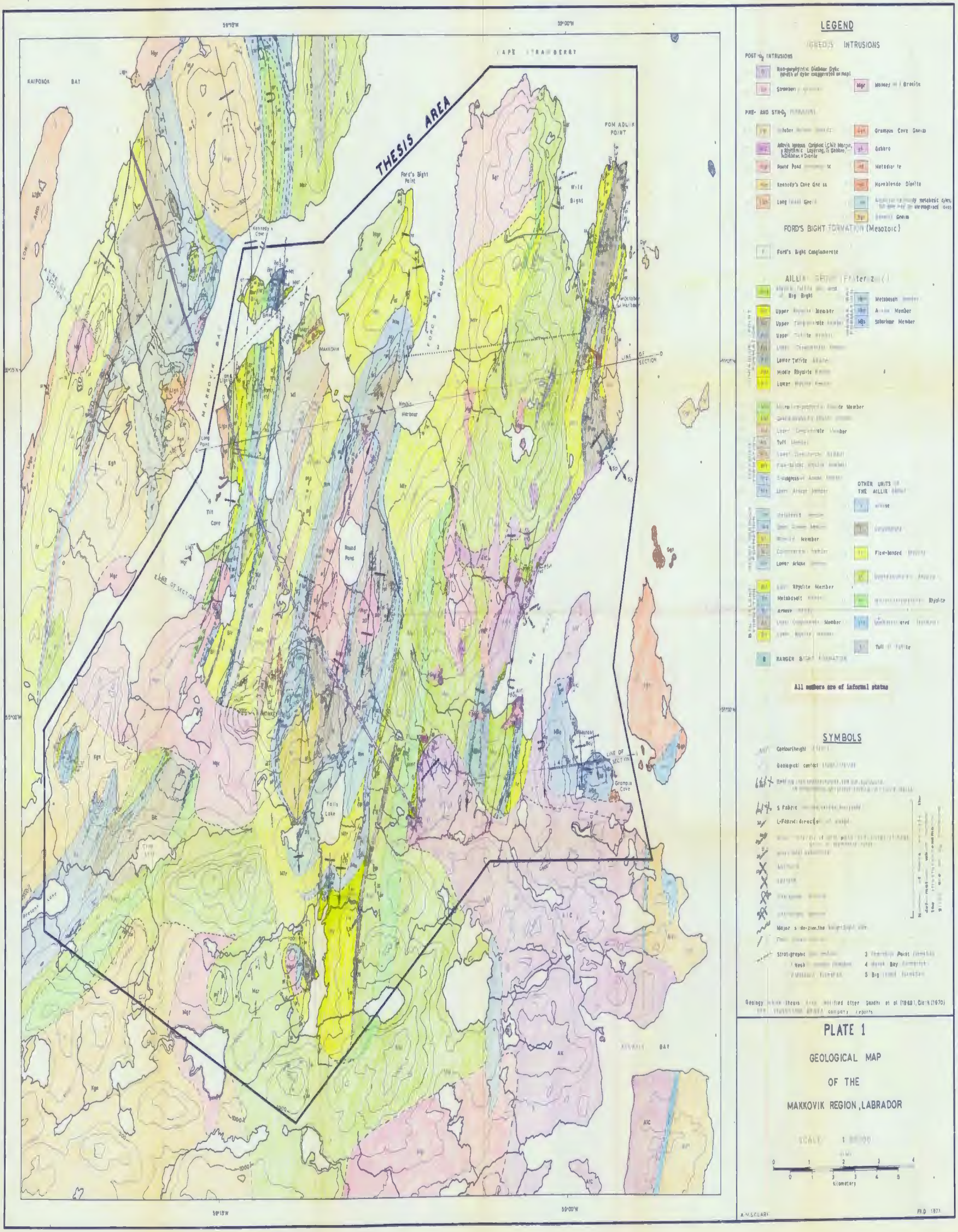
The lack of c-axis maxima correspondence or symmetry with respect to the D_2 strain axes indicates the c-axis orientations are probably not of D_2 origin, i.e., are not due to annealing recrystallisation. As quartz recrystallises extremely easily, at temperatures as low as 100°C to 150°C (Voll, 1960), (Zeolite facies, Turner, 1968) and D_3 or D_4 deformations could well have produced this fabric. There is no suggestion that the fabric pattern is similar to crossed girdles (Sylvester and Christie, 1968; Shelley, 1971), and a comparison of the maxima with theoretically determined quartz maxima positions based on a fracture hypothesis (Fairbairn,

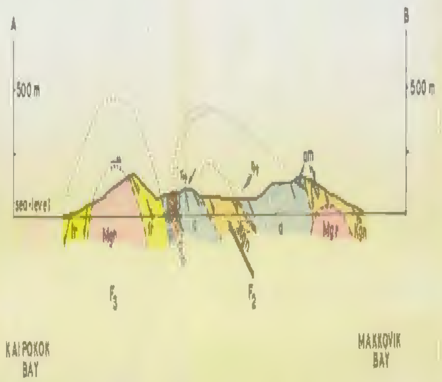
1949) leads to innumerable similarities, amongst which none can at present be chosen as more likely correct. No further interpretation can, therefore, be made on these few measurements.

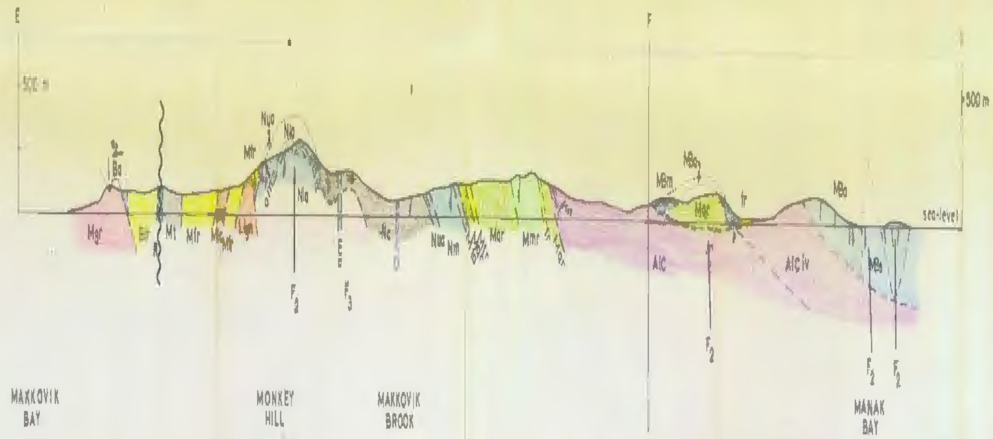
TABLE 7 -IV

Summary of the Geology of the Makkovik Region

	Sedimentary Sequence	Sedimentary Lithology	Deformation	Metamorphism	Major Intrusions	Important Minor Intrusions	Remarks
20.						Lamprophyre Dykes	Possibly Intruded Fords Bight Conglomerate
19.	Ford's Bight Conglomerate	Polymictic conglomerate					Contains Jurassic Coccoliths
18.						Lamprophyre Dykes	
17.						Diabase dykes	Hadrinian
16.						Diabase dykes	Helikian
15.			D_4	?			
14.					Strawberry Granite		
13.			D_3	Amphibolite facies?			
12.				Amphibolite facies	Monkey Hill Granite		
11.			D_2	Greenschist-amphiobolite transition facies	Adlavik Igneous Complex, October Harbour Granite, Kennedy's Cove Gneiss, Long Island Gneiss		
10.					Grampus Cove Gneiss Round Pond Granodiorite		Post-D_1/Pre.-D_2 or Syn.-D_2
9.			D_1	Greenschist-amphibolite transition facies			
8.						Amphibolite & Biotite-carbonate dykes	
7.						Rhyolite dykes	
6.						Amphibolite Dykes	
5.	Big Island Formation.	Arkoses, conglomerates, rhyolites, tuffs and mafic lavas.					This formation cannot be correlated with the other formations of the Aillik Group, but is Pre-D_1 in age.
4.	Manak Bay Formation	Arkoses and mafic lava.					Overlies the Makkovik Formation.
3.	Pomiadluk Point Formation	Extrusive flow-banded rhyolites, tuffs, conglomerate.					Overlies Makkovik Formation.
2.	Makkovik Formation	Extrusive flow-banded & quartz- & microcline-porphyritic rhyolites, tuffs arkoses.					Overlies the Nesbit Harbour Formation.
1.	Nesbit Harbour Formation	Arkoses, conglomerates, plagioclase-porphyritic rhyolite & mafic (basaltic) Lava.					Base of formation not known


THESIS AREA
CAPE STRAWBERRY
KAIPOKOK BAY
MAKKOVIK BAY
Ford's Bight Point
POMIADLUK POINT
Wild Bight
Kennedy's Cove
MAKKOVIK
Nesbit Harbour
Long Point
Tilt Cove
Round Pond
Falls Lake
Grampus Cove
AIVIKTOK BAY
LINE OF SECTION
LEGEND
INTRUSIONS
POST-D_2 INTRUSIONS
Non-porphyritic Diabase Dyke
(width of dyke exaggerated on map)
Granite
PRE- AND SYN-D_2
Grampus Cove Gneiss
Aillik Igneous Complex
Gabbro
Round Pond
Metadiorite
Kennedy's Cove Gneiss
Hornblende Diorite
Long Point Gneiss
Gneiss
FORD'S BIGHT FORMATION (Mesozoic)
Ford's Bight Conglomerate
AILLIK GROUP
Big Bight
Upper Rhyolite Member
Upper Tuffite Member
Lower Tuffite Member
Middle Rhyolite Member
Lower Rhyolite Member
Metabasalt Member
Arkose Member
Conglomerate Member
Tuff Member
Rhyolite Member
Lower Arkose Member
Flow-banded Rhyolite
Tuff or Tuffite
RANGER BIGHT FORMATION
OTHER UNITS OF THE AILLIK GROUP
All members are of informal status
SYMBOLS
Contour height
Geological contact
S Fabric
L-Fabric
Stratigraphic
PLATE 1
GEOLOGICAL MAP
OF THE
MAKKOVIK REGION, LABRADOR
Kilometers
A.M.S. CLARK
Ph.D 1973





LEGEND

As for Plate 1

SYMBOLS

Apparent dip of bedding in plane of section

Facing direction of bedding

Apparent dip of S_2 in plane of section

Attitude of axial-plane of fold.

Major slide-zone

all cross-sections looking northwards

PLATE 2

CROSS-SECTIONS SHOWING THE GEOLOGY IN THE MAKKOVIK REGION, LABRADOR

VERTICAL AND HORIZONTAL SCALE 1: 50,000

miles 0 1 2 3

kilometers 0 1 2 3 4

A.M.S.CLARK

Ph.D. 1973

grades out to rims of oligoclase/andesine (determined from refractive indices). The hypersthene is subhedral and partially altered to chlorite (?) in cleavage traces. A few hornblende grains have a core of clinopyroxene suggesting formation by alteration or reaction of clinopyroxene. A sample taken from the host Grampus Cove Gneiss immediately outside the trace of the dyke contains hypersthene with rims of clinopyroxene (augite?). No pyroxene has been recognised in the main body of the gneiss, and the occurrence of orthopyroxene here indicates very localised high temperature (pyroxene hornfels) metamorphic conditions.

The evidence of very localised high temperature conditions in the immediate vicinity of the dyke, and the gradational contact of the acid phase of the dyke with the host gneiss indicates derivation of that phase by partial melting of the gneiss by the intruding diabase. The occurrence of the gneissic texture and its destruction by the intrusion indicates the gneiss was effectively solid prior to the intrusion. Therefore, the more common origin of net-vein dykes by association of two magmas (Gibson, et al., 1963; Blake, et al., 1965; and Walker, et al., 1966), cannot have occurred here. However, a slight modification of the ideas of Blake, et al., (1965) in which basic magma passes through a previously solidified gneiss, rather than their viscous acid magma, would explain the observations. Similarly, the relationship of the acid phase with the gneiss is also conclusive evidence against Chapman's (1962) contraction-crack theory and Windley's (1965) shear-plane extension of that theory having occurred here, in both of which the acidic portion is a later introduction into a previously solidified mafic dyke.

Diopside Diorite Dykes

Diopside diorite dykes have been recognised at Manak Bay and elsewhere in the area and are composed of subhedral to euhedral elongate grains of diopside and green hornblende in interlocking blocky oligoclase grains and a matt of altered K-feldspar. The diopside has alteration rims of hornblende. The composition of the rock is:

oligoclase	50%
K-feldspar	20%
diopside	10%
hornblende	15%

with minor amounts of opaque minerals, chlorite, saussurite, clay minerals and sphene.

Pigeonite-porphyritic Diabase Dykes

The pigeonite-porphyritic diabase dykes are very fine-grained, grey-green dykes with 1/2 to 2 mm. long lath-shaped phenocrysts of a mafic mineral. On the south coast of Manak Bay one cuts an F_4 fold of amphibolite but is itself unfolded. In thin-section this rock is seen to contain pigeonite phenocrysts which have been partially altered to green hornblende, biotite and a large concentration of small hornblende grains, poikilitically enclosed and forming sieve-texture in relatively large (3 to 4 mm.) indistinctly bounded plagioclase grains. (Fig. 3-11). The plagioclase has undergone recrystallisation and has developed a polygonal aggregate of albite and oligoclase/andesine grains which have generally retained their original relative extinction orientations, so that they appear, under low magnification, to form a single grain. Epidote occurs as diffuse clots within the plagioclase in some parts of the rock.

The composition of the diabase is:

albite	10%
andesine	40%
pigeonite	10%
hornblende	20%
biotite	10%
epidote	10%

and accessory sphene.

Quartz Porphyry/Diabase Net-vein Dykes

A quartz porphyry/diabase net-vein dyke occurs at Pomiadluk Point. Many similar net-vein dykes occur between Pomiadluk Point and Big Bight, and possibly elsewhere in the area, though they are difficult to distinguish from other net-vein dykes. They consist of rounded, medium grey, very fine-grained, mafic blebs up to 30 cm. (1 ft.) in diameter with irregular dentate margins (Windley, 1965) set in a pale grey to white, fine- to medium-grained quartzo-feldspathic matrix of variable composition. The dykes are intruded with sharp contact into the various host lithologies, and show no development of chill-margins.

In thin-section the mafic material is seen to consist of zoned plagioclase laths and acicular to blocky green hornblende grains with interstitial biotite, plagioclase, hornblende and sphene. There is a regular gradation in lithologic composition from the mafic phase to the most acidic of the vein material. The most quartzo-feldspathic phase is composed of a few 2 mm. (1/10 in.) recrystallised quartz phenocrysts set in a fine-grained matrix composed of inter-locking quartz, microcline, zoned plagioclase is commonly mantled by microcline. Granophyric intergrowth is common in the interstices and also on the margins of the quartz phenocrysts. Myrmekitic intergrowth also occurs, most commonly on plagioclase grains, but also as embayments in microcline. This acidic member forms small veinlets and hazy-bounded "interstitial" areas to the diabasic blebs and to the lithology of intermediate composition. No fine-grained chill-margins have been recognised between any of the lithologies of the dyke.

The lack of chill margins between any of the phases (Fig. 3-12) and between the dyke and the wall-rock, and the occurrence of dentate margins to the mafic blebs,

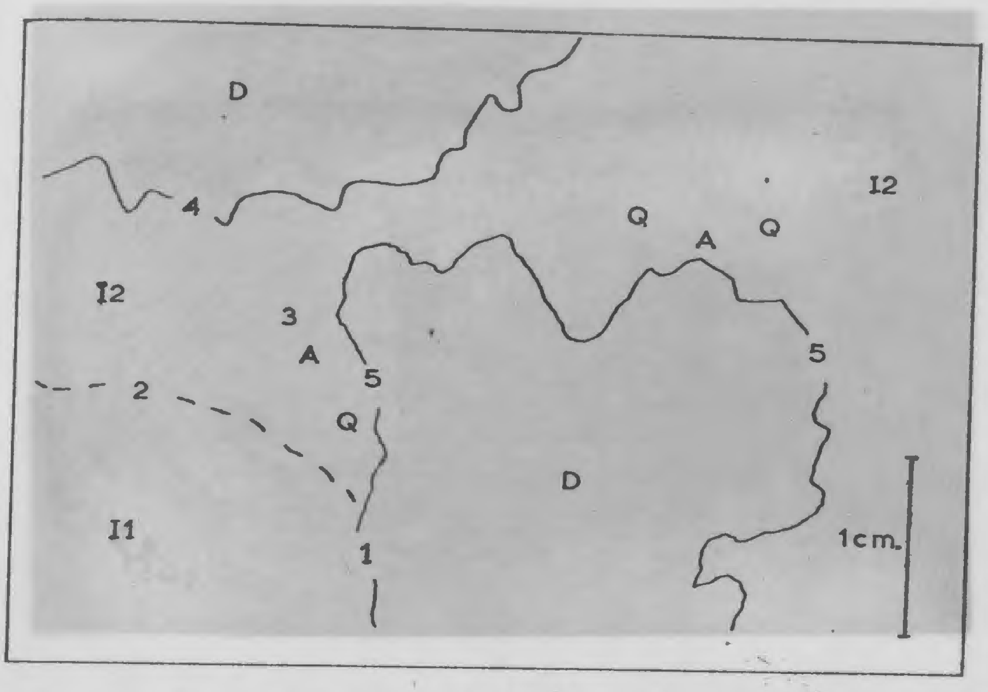


Fig. 3-12. Photograph of thin-section of quartz-porphyry/diabase net-vein dyke from Pomiadluk Point. Plane polarised light. See text for description.

A Acidic veinlets with quartz phenocrysts

Q Quartz-phenocrysts

D Diabase blebs

I_1 Intermediate lithology similar to diabase.

I_2 Intermediate lithology similar to acidic veinlets.

1 Hazy contact between D and I_1

2 Hazy contact between I_1 and I_2

3 Hazy contact between I_2 and A

4 Sharp dentate contact between D and I_2

5 Sharp dentate contact between D and A

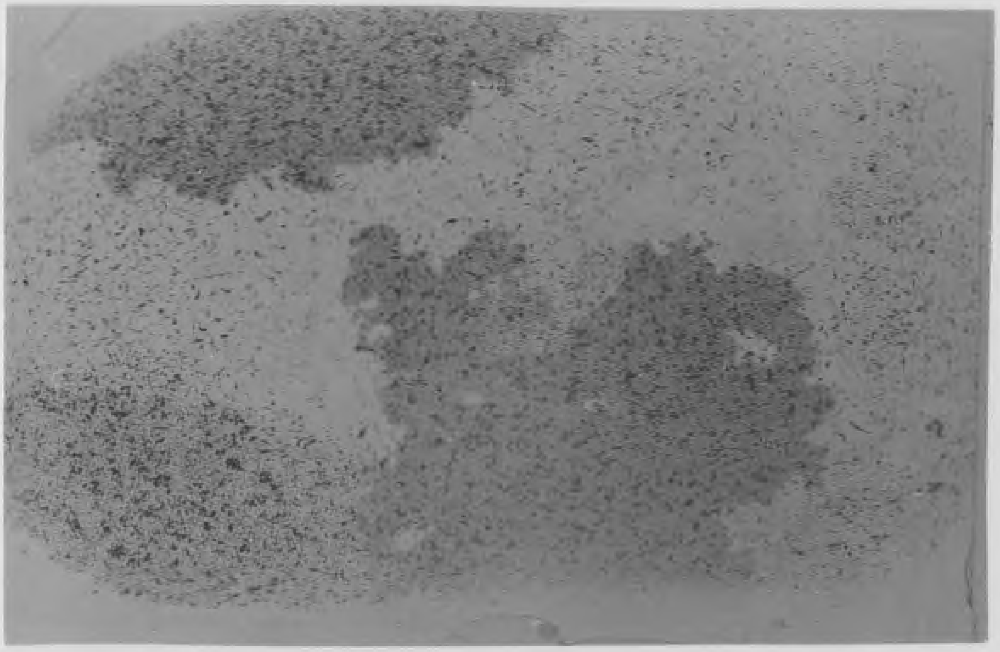


Fig. 3-12. Photograph of thin-section of quartz-porphyry/diabase net-vein dyke from Pomiadluk Point. Plane polarised light. See text for description.

A Acidic veinlets with quartz phenocrysts

Q Quartz-phenocrysts

D Diabase blebs

I_1 Intermediate lithology similar to diabase.

I_2 Intermediate lithology similar to acidic veinlets.

1 Hazy contact between D and I_1

2 Hazy contact between I_1 and I_2

3 Hazy contact between I_2 and A

4 Sharp dentate contact between D and I_2

5 Sharp dentate contact between D and A

suggest the dyke was intruded into hot country-rock as a composite magma of plastic mafic blebs in a more fluid acid phase. Although both phases may have been primary, the acid phase may, instead, have been the result of partial melting of an earlier quartz porphyry, which on re-solidification developed the microcline mantling to plagioclase and the granophyric intergrowth textures.

Brown Feldspar-porphyritic Diabase Dykes

Two brown feldspar-porphyritic diabase dykes have been recognised in the Wild Bight Pomiadluk Point area. Both dykes have steep to vertical dips. The dykes are medium- to fine-grained black to dark grey rocks, with a distinctive brown-coloured weathered surface, containing up to 10% of 1 cm. (1/2 in.) to 5 cm. (2 1/2 in.) sized pale green plagioclase phenocrysts.

In thin-section the rock is seen to consist of highly saussuritised plagioclase phenocrysts of indeterminate composition set in a groundmass of interlocking plagioclase laths and intersertal plagioclase making up to 50% of the rock. The rock is composed of:

plagioclase	40 - 50%
titanaugite	10 - 15%
opaque minerals	5 - 15%
green and brown biotite	0 - 5%
chlorite, serpentine	5 - 15%

The gabbro is composed of:

andesine/labradorite	50 - 65%
basaltic hornblende	10 - 15%
pigeonite	10 - 15%
opaque minerals	5 - 10%
dark brown biotite	0 - 5%

with minor amounts of green hornblende, green biotite, olivine and alteration products. Small 1 to 2 cm. (1/2 to 1 in.) diameter quartzo-feldspathic xenoliths showing incipient melting and development of spherulitic structures are common.

Plagioclase-pegmatite veins up to 10 cm. (5 in.) wide and several meters long containing dark grey plagioclase crystals up to 10 cm. (5 in.) long occur within this facies south of Manak Bay.

iv. Massive Diabase Facies: The massive diabase facies occurs between the massive gabbro facies and the country rock along the northeast contact of the main body of the intrusion. It also occurs about Big Bight. The rock is completely gradational with the massive gabbro and is apparently a finer grained border variation of it.

The rock is a homogeneous medium-grained (up to 1/2 cm. (1/4 in.) diameter), black and rusty-white, salt-and-pepper textured diabase. In thin-section it is seen to consist of stubby, commonly ophitic, plagioclase laths interspersed with mafic clots composed of pigeonite, basaltic

The chlorite, serpentine and minor epidote appear to have developed by the alteration of olivine, hornblende(?) and plagioclase.

These dykes appear to be equivalent to King's (1963) feldspar porphyritic diabase and to the porphyritic diabase of Gandhi, et al., (1969).

Dark Green Olivine Diabase Dykes

These dykes are common in the area and usually form 5 to 20 cm. (2 1/2 to 10 in.) thick extended sheets dipping at various angles. The dykes consist of a very hard, fine-grained, dark-green rock with conchoidal to brittle fracture and composed of:

phenocrysts 15%:	
plagioclase	10%
serpentine after olivine	5%
groundmass 85%:	
plagioclase	25%
opaque minerals	20%
glass	40%

and accessory apatite.

The serpentine occurs as small pseudomorphs after olivine phenocrysts (six-sided outline). One eight-sided bastite pseudomorph after pyroxene was seen. Serpentine occurs in accessory amounts in the groundmass and is probably also derived from olivine. The plagioclase in both the phenocrysts and the groundmass is saussuritised and partly altered to chlorite and its composition could not be determined. The groundmass consists of interstitial "glass" which does not go completely dark under cross nicols indicating sub-microscopic crystallisation. The "glass" forms the host of small randomly oriented plagioclase and opaque mineral laths.

Grey Diabase Dykes

The grey diabase is a ubiquitous very fine-grained, grey rock which forms dykes up to 1 to 2 m. (3 to 6 ft.) wide, but more commonly 10 to 20 cm. (5 to 10 in.) wide and of extreme length. In thin-section the rock is seen to be composed of very small, highly saussuritised plagioclase laths which form an interlocking network with zoned green hornblende laths and needles, in the interstices of which is plagioclase, sphene and minor pyrite.

Non-porphyritic Diabase Dykes

A non-porphyritic diabase dyke occurs as a vertically oriented 3 m. (9 ft.) wide intrusion extending across the entire area from Makkovik Bay to the Adlavik Igneous Complex (Plate 1). Another small sub-horizontal sheet of the diabase occurs on Big Island. The diabase is fine- to medium-grained and dark grey to grey-brown in colour, and is composed of ophitic plagioclase laths (35%), slightly saussuritised intersertal plagioclase (15%), pale pink non-pleochroic pigeonite (30%), 5% intersertal serpentine (after olivine) and olivine, and 5% opaque minerals.

At Makkovik Bay, the dyke is vertical and strikes south-easterly, and is the extension of a diabase dyke occurring across the bay (Clark, 1971). The mineralogy and field appearance of the dyke suggests it is the same as the olivine diabase of Gandhi, et al., (1969) and the diabase-olivine diabase of King (1963). However, the dykes mentioned by both Gandhi and King strike north-easterly, and this one therefore appears to have intruded a complementary joint.

Lamprophyre Dykes

Lamprophyre dykes are common and occur throughout the area. They are apparently the latest intrusions in the area, (though intrusive relationships with the non-porphyritic diabase dykes were not seen), and are usually 30 cm. to 1 m. (1 ft. to 3 ft.) in width and sub-vertical to sub-horozontal in attitude. The lamprophyres are dark brown, green or black coloured rocks and show small 1 to 2 mm. (1/16 in. to 1/8 in.) phenocrysts in a very fine-grained matrix. Zoning of the dykes is common, with the central portions being more phenocryst-rich than the borders (Fig. 3-13).

The rocks are variable in composition, but are generally composed of phenocrysts of olivine (now altered to serpentine), augite, biotite or phlogopite, and more

Fig. 3-13. Lamprophyre dyke showing concentration of phenocrysts towards the center of the dyke. East coast of Wild Bight.

Fig. 3-14. Net-vein dyke with mafic blocks of different compositions. West coast of Big Bight.

rarely hornblende, nepheline or muscovite, in various combinations. These phenocrysts are set in a very fine-grained to glassy matrix of carbonate, opaque minerals, plagioclase, chlorite, serpentine or talc. Although several different types of lamprophyre are represented, they are, for the most part, alnöitic or monchiquitic in composition (Moorhouse, 1959). A more detailed description of the lamprophyres, and chemical analyses are given in Kranck (1953), University of Leeds (Leeds, 1962, p. 55) and King (1963).

Other Minor Intrusions

Many net-vein dykes of various types occur in the area apart from those already described. The majority have mafic blocks of mixed parentage suggesting multiple mafic intrusions with the acid material being either locally derived by melting of country-rock, or a later intrusion (Fig. 3-14). On the coast east of Ranger Bight a 1 to 2 m. (3 to 6 ft.) wide string of 30 to 60 cm. (1 to 2 ft.) diameter pillow-shaped mafic blocks occurs in the porphyritic rhyolite (Fig. 3-15). The acid veining appears to be directly related to the enclosing rhyolite, and the elongate disposition of the "pillows" suggests a mafic dyke intrusion into the rhyolite and concomittent melting of the rhyolite (Blake, et al., 1965). Later tectonic deformation has resulted in development of lobate margins to the "pillows", probably initiated on the dentate margins (Windley, 1965).

Fig. 3-15. Pillow-shaped structures developed in mafic dyke intruded into quartz-porphyritic rhyolite. Coast east of Ranger Bight.

Fig. 3-16. Circular orientation of feldspar phenocrysts about flow-cells C1 and C2 in basaltic dyke on the south coast of Big Bight. Photograph in horizontal plane.

Minor pre- or syn-tectonic quartz veins occur in the vicinity of Round Pond and a few contain molybdenite. A small pre-tectonic quartz-microcline dykelet cuts the eastern side of the gabbroic body at Pomiadluk Point and contains minor phlogopite, plagioclase and arsenopyrite. A post-tectonic microcline-epidote veinlet containing minor plagioclase, sphene and hornblende was seen in the lower Big Island Formation conglomerate, and quartz-fluorite veinlets occur in the Nesbit Harbour Formation metabasalt and tuff members. An excellent example of cellular flow of magma during dyke intrusion has been noted on the point at the head of Big Bight (Fig. 3-16). The dyke is a 20 cm. (8 in.) wide basic dyke with tabular feldspar phenocrysts distributed throughout. Locally these phenocrysts are oriented in an oval or circular pattern (centered on C1 and C2 respectively in Fig. 3-16) and indicate vertical or sub-vertical flow.

On the west coast of Wild Bight, 1/2 cm. (1/4 in.) wide diopside veins containing small amounts of molybdenite fill post-D_2 joints in a pre-D_2 amphibolite dyke and are in turn cut by a grey diorite dyke (Fig. 3-8). The diopside occurs as long (up to 2 cm. (1 in.)) acicular crystals oriented in a radiating (sun-burst) pattern in the plane of the vein, and separated from the host amphibolite by a veneer of highly saussuritised feldspar which appears to be an alteration and replacement product of the amphibolite.

Summary

The Aillik Group has been intruded by igneous rocks of widely varying composition from granite to gabbro, which underly up to one-quarter of the thesis area. The main rock-type is granite/quartz monzonite/granodiorite which forms most of the early gneisses and the later undeformed granites, but hornblende-gabbro and associated diorites of appinitic affinities occurring at the border of the thesis area form a very large igneous complex, most of which is outside the area. The minor intrusions show a far greater range of compositions, including hornblendite and alnöite, with almost all types being represented in each of the pre-, syn- and post-tectonic periods.

CHAPTER 4

STRUCTURE

Introduction

Previous work in the Makkovik region (Kranck, 1939 and 1953; Gandhi, et al., 1969, and Stevenson, 1970) suggested a fairly simple structural style and history. However, work by the author (Clark, 1970 and 1971) in an adjacent area indicated that a complex sequence of both regional and localised deformational episodes had occurred. The apparent similarity between the two areas suggested that a similar sequence of tectonic events had probably occurred here. The recognition of the importance of a thorough understanding of structure to the determination of the stratigraphy was one of the main factors leading to the investigation of this area.

The area is shown to have undergone one major phase of folding with concomittent development of a penetrative axial planar schistosity during the second deformation, and minor, lithologically localised folding and schistosity development during three other deformations. The major folding of the second deformation produced upward facing structures (Shackleton, 1957) and northeast trends, but later folding about an easterly axis has developed a major interference dome in the center of the area. Other structures such as boudins, slides and faults are only locally common, though one large-scale slide of unknown displacement has been recognised.

As the second deformation schistosity (S_2) is the only fabric that may be recognised throughout the area, it has been used as a reference datum for determining the relative ages of the other, more locally developed, fabrics and structures. This schistosity is primarily defined by orientation of flattened quartz and feldspar phenocrysts and conglomerate pebbles, but S_2 schistosity is well developed in biotite- or amphibole-rich lithologies where it is generally folded by later structures.

The fabrics and structures developed in the area and described below are considered to be consistent, except where otherwise stated, with development under conditions of three-dimensional progressive deformation as described by Flinn (1962 and 1965) and further explained and developed by Ramsay (1967) and Watterson (1968). The textural analysis is based on the theories of metamorphic mineral growth described by Spry (1969) as modified by Powell and Treagus (1967 and 1970).

Structural Terminology

The structural history of the area has been subdivided into several phases of deformation which are referred to, in order of development, as D_1, D_2, D_3, and D_4. These phases of deformation are considered to have formed during one major period of tectonism. The terms pre-, syn- or post-D_1 (or D_2, etc.) are used to denote stages related to individual phases of deformation, whereas pre-, syn- or post-tectonism refers to stages of development with respect to the entire period of tectonism (which may be equivalent to hornblende, opaque minerals, biotite, olivine, and alteration products. The rock is composed of:

andesine-labradorite	50 - 65%
pigeonite	5 - 10%
basaltic hornblende	10 - 20%
opaque minerals	5 - 10%
biotite	0 - 5%

with minor amounts of prismatic apatite, olivine, green hornblende, calcite and serpentine.

v. Diorite Facies: The diorite facies occurs along the western shore of Big Bight and up to 6 km. (3 mi.) inland from the coast. Several small intrusions occur south of Wild Bight, west of Big Bight, and on the small peninsula east of Big Bight. The diorite was mapped as syenite but not formally named by Gandhi, *et al*., (1969). The area was plotted as Domino Gneiss on Kranck's (1953) map and is outside the limits of Stevenson's (1970) area. The diorite is intrusive into the Aillik Group and also into the massive diabase facies and is intruded by the Monkey Hill Granite and also by hornblende porphyrite and hornblendite dykes.

The diorite is very variable in appearance from a coarse-grained, pink and green gneiss with up to 20% mafic minerals north of Big Bight, through a fine- to medium-grained, pink, cream and black, speckled, very slightly gneissic, granitic rock west of Big Bight, to the more common blotchy an orogney).

The strain (deformation) ellipsoid axes are termed the X, Y and Z axes, where $X \leq Y \leq Z$, (Flinn, 1962).

The type of deformation (i.e., k-value) and amount of strain (r) are defined by:

$$k= \frac{a-1}{b-1} \quad \text{(Flinn, 1962 and 1965),}$$

and $$r= a+b-1 \quad \text{(Watterson, 1968),}$$

where $$a= \frac{Z}{Y} \text{ and } b= \frac{Y}{X} \quad (X \leq Y \leq Z).$$

Elements that define the plane of flattening (YZ-plane) of the deformation ellipsoid, such as schistosity or the plane perpendicular to the short axis of flattened pebbles or phenocrysts, are referred to as S-fabrics or S-tectonite fabrics (Flinn, 1958a). Similarly, elements that define the axis of elongation (Z-axis) of the ellipsoid, such as mineral lineation or the long axis of deformed pebbles or phenocrysts, are referred to as L-fabrics or L-tectonite fabrics. Note that the terms S- and L-fabric are used to describe the mutual relationship of the short or long axes of deformed pebbles or other bodies, as well as the relationship of oriented planar or acicular minerals.

S- and L-fabrics and folds (F) are related to their respective deformations by the terminology:

S_1, L_1 and F_1 for features developed during the first deformation (D_1);

S_2, L_2 and F_2 for features developed during the second deformation (D_2);

and similarly for the other deformations.

Fleuty's (1964b) use of the term "slide" to refer to a fault formed in close connection to folding, and which is broadly parallel to the axial plane of the fold, is adhered to here, as is his (1964a) terminology in describing the tightness (interlimb angle) of folds.

The term "band" (American Geological Institute, 1962) is used to denote any colour, compositional or textural layering that cannot be shown to be bedding. Where possible the mode of origin of the banding is stated in the text, but in many cases the origin is unknown. The types of banding that may occur in the area are:

1. Bedding, within which sedimentary features have not been recognised, either because they were not developed or because of later destruction (usually by tectonic flattening).

2. Flow-banding in rhyolites, which is more fully discussed in Chapter 2.

3. Banding due to repetition by isoclinal folding of beds or earlier bands.

4. Banding due to folding and sliding out of fold-limbs of a bedded or banded unit.

5. "Mylonite" banding due to tectonic flattening of pebbles, phenocrysts, etc., (Johnson, 1967).

6. Mineral segregation banding due to development of crenulation cleavages (Nicholson, 1966).

The Main Penetrative Deformation (D_2)

INTRODUCTION

The second deformation is the most widely developed and most thoroughly penetrative deformation in the region. The effects of second deformation flattening are seen throughout the area, though these effects are more pronounced in some lithologies that in others, and the amount of deformation is greater in some parts of the area than in others. D_2 structures and fabrics are used as a reference in determining the age of other structures and fabrics and will therefore be described first. The basic map-pattern in the area is the result of F_2 folding and is only slightly modified by other deformations. In all areas where tops could be determined, the beds were seen to face upwards on S_2.

D_2 TECTONITE FABRICS

The S_2 and L_2 fabric elements are defined in two ways in the area:

1. By the orientation of platy and acicular minerals grown during the deformation.

2. By the orientation of flattened and stretched (pre-D_2) bodies such as phenocrysts and clasts.

The S_2 fabric can be traced throughout the area from one lithology to another. It is also parallel to the axial-planes of F_2 folds of an earlier locally developed schistosity. The earlier fabric has therefore had a negligible effect on the S_2 orientation. The S_2 plane is therefore presumed to accurately represent the D_2 plane of flattening. The L_2 fabrics which represent the D_2 axis of elongation show similar distribution and behaviour. The S_2 fabric strikes northeastward and is subvertical throughout the area, except for locally, immediately east of Round Pond, where it is reoriented by later folds. The L_2 fabric is poorly developed, strikes northeastward and is predominantly subhorizontal.

Fabrics Defined by Mineral Orientation

Mineral orientation fabrics occur in almost all pre-D_2 amphibole and biotite bearing rocks, but are less well developed in the majority of rocks in the area due to their predominately quartzo-feldspathic composition and lack of acicular and platy minerals.

The D_2 schistosity and mineral lineation are defined by biotite, phlogopite, hornblende, and, very locally, by graphite. The fabrics are simple in the quartzo-feldspathic lithologies, but are usually derived from an earlier fabric and related to crenulation-cleavage in the biotite- and hornblende-rich lithologies. All the gneisses and the Round Pond Granodiorite show D_2 fabrics of mafic minerals in flattened and elongate mafic mineral aggregates, and have moderately developed S_2 schistosity in their margins. The Adlavik Igneous Complex and October Harbour Granite also show moderately schistose margins, though the former is undeformed internally, whereas the latter shows a "primary" orientation of feldspar parallel to S_2 which is considered to be tectonic in origin (c.f. Chapter 6).

Fabrics defined by Deformed Bodies

In some of the rhyolites and conglomerates of the Aillik Group the L_2 and S_2 fabrics are defined by the attitude, shape and distribution of deformed bodies such as conglomerate pebbles. Where the pre-tectonic attitude, shape or distribution is known, the strain-orientation, and amount and type of deformation may be determined (Appendix I). There is no evidence of reorientation of an earlier fabric in these rocks, even in the vicinity of F_2 fold hinges (Fig. 4-32), and it is therefore assumed the earlier fabrics were so weakly developed in the porphyritic rhyolites, conglomerates and lithophysic flow-banded rhyolite that they were completely masked by the development of D_2 fabrics.

i. Deformed Phenocrysts: Relict quartz and feldspar phenocrysts are abundantly distributed throughout the quartz and feldspar porphyritic rhyolites, and usually define D_2 tectonite fabrics. Acicular and platy minerals are rare or nonexistent in these lithologies and these rocks underlie up to half of the entire thesis area. There is no evidence of primary (flow) foliation or of any lamination or 'bedding' within the porphyritic rhyolites, apart from the few tuff horizons, and the flattening and extension of the phenocrysts is, therefore assumed to be parallel to the strain axes in the rock (Fig. 2-8). (Analysis of the quartz c-axis orientations of the subgrains of these phenocrysts is given in Appendix III. These c-axis orientations appear to be related to a later deformation). Where acicular or platy minerals occur in the rhyolite, or in a near-by lithology, they are oriented parallel to the D_2 fabrics defined by the relict phenocrysts in the rhyolite. Some of the basic dykes have phenocrysts or pre-tectonic porphyroblasts which also define L_2 and S_2 fabric orientations.

ii. Deformed Conglomerate Pebbles: The pebbles in the conglomerates have undergone tectonic flattening and extension parallel to the D_2 fabrics defined by mineral orientations in associated lithologies (Fig. I-1). As with the phenocrysts in the porphyritic rhyolites, this orientation is assumed to be directly related to the strain ellipsoid, though, in fact, a minor amount of pre-D_2 (depositional?) preferred orientation has been recognised in at least one of the conglomerates (see Appendix I). Small pebbles and phenocrysts in tuffs also show D_2 flattening and extension.

iii. Deformed Lithophysae: Lithophysae in a rhyolite on the north coast of Big Island have undergone D_2 flattening and elongation (Figs. 2-40 and 4-1). As with the conglomerate pebbles, there is no recognisable difference between

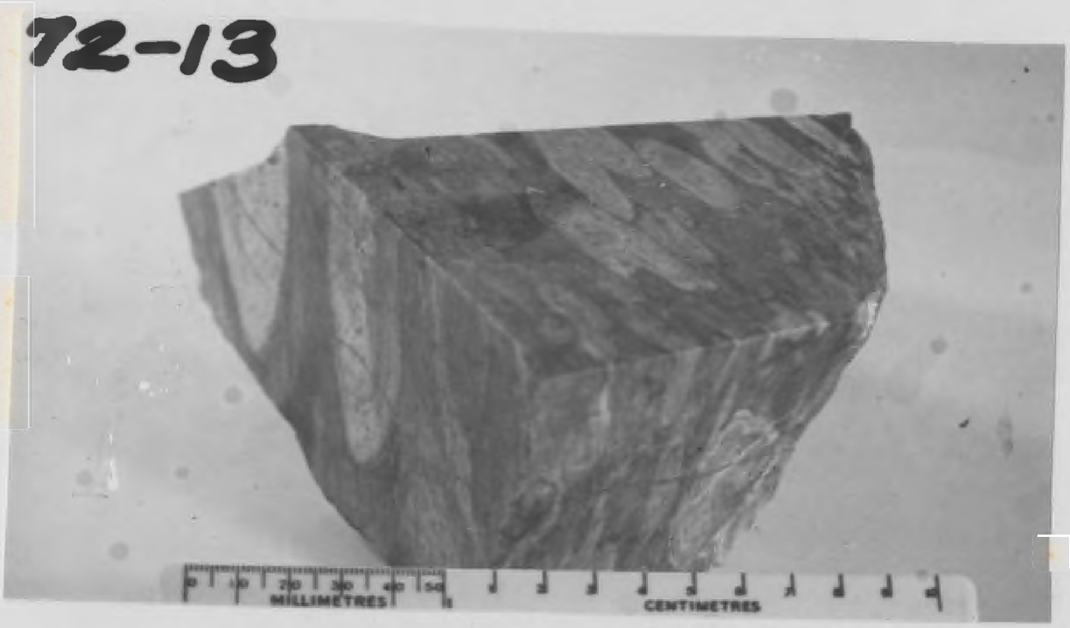


Fig. 4-1. Deformed lithophysae in the upper rhyolite member of the Big Island Formation. Sample cut parallel to the three principal tectonic planes.

the lithophysae ellipsoid orientations and the L_2 and S_2 fabric orientations in other lithologies in the vicinity, so that pre-D_2 tectonic strain is considered to be negligible (Oertel, 1970).

D_2 BANDING

Banding resulting from sliding developed in folds is common in the Big Island/Ranger Bight area, but also occurs elsewhere. The banding varies from 10 to 30 cm. (4 to 12 in.) to 1 to 2 cm. (1/2 to 1 in.) in width (Figs. 4-23, 4-21, 4-22 and 4-11). Transposition of earlier (pre-D_2) fabric (crenulation-cleavage) is common in the amphibole- and biotite-rich lithologies and has formed banding that is usually 2 to 4 mm. (1/10 to 1/5 in.) in width (Fig. 4-33).

D_2 STRUCTURES

The structures developed during the second deformation have produced the basic map pattern in the area, and are the only large scale, easily recognised structures. Generally the most conspicuous of the structures are the folds, but the Ranger Bight slide is equally as important. Many examples of minor folding, sliding and boudinage may be found in the area.

Major F_2 Folds

The major F_2 folds (Fig. 4-2 and Plates 1 and 2) have been determined primarily from outcrop patterns, but

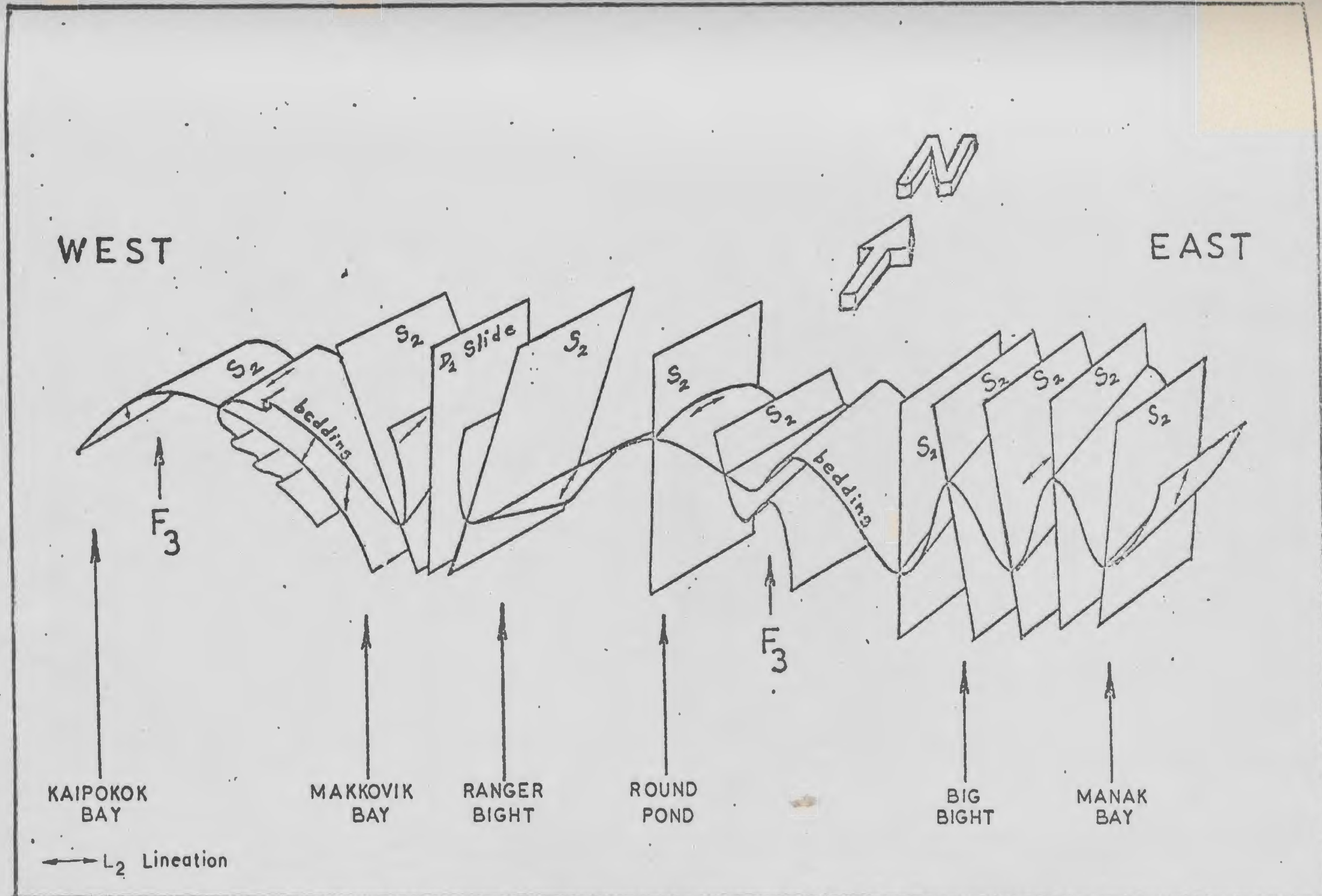


Fig. 4-2. Simplified three dimensional structural surface representation from Kaipokok Bay to Manak Bay, (partly after Clark, 1970 and 1971).

bedding/fabric and banding/fabric intersections (Fig. 4-3) have been used in conjunction with bedding attitudes to confirm closing directions. Apart from a slight overturn of bedding (85^{o} dip) between Manak Bay and Big Bight and east of Makkovik Bay, all bedding is upward facing on S_2.

The most distinctive F_2 fold in the area is the large 20 by 8 km. (10 by 4 mi.), elongate, anticlinal dome about Round Pond in the center of the area, the map-pattern of which is markedly modified in the south by the extreme topography of Monkey Hill (Fig. 4-4). As the L_2 lineation is sub-horizontal the D_2 stretching direction must also have been sub-horizontal, and the dome cannot, therefore, be due to a single deformation of prolate ellipsoid ($1 \leq K \leq \infty$) type (Flinn, 1962).

An attempt was made to determine the beta-point and pi-plane for the southern closure of the dome from a stereographic plot by first unfolding the F_3 folds about the F_3 fold axis, assuming they were open enough for this not to introduce too much error. However, the spread of points is too great and number of readings too small for the results to be meaningful (Figs. 4-5A and B).

Other, somewhat smaller but nevertheless still major F_2 folds occur to the east at Big Bight and Manak Bay. The latter are clearly exposed along the south shore of Manak Bay (Plate 2) and are upright, tight, southward plunging zig-zag folds (Ramsay, 1967; and Figs. 4-6 and 4-7).

pink and white, medium- to coarse-grained granitic rock with long (up to 4 cm. (2 in.)) acicular crystals of hornblende forming a very distinctive, randomly oriented, three-dimensional, criss-cross pattern. The diorite most commonly consists of:

quartz	0 - 10%
K-feldspar	0 - 15%
plagioclase	40 - 80%
hornblende	5 - 10%

with minor and accessory amounts of opaque minerals, sphene, apatite, biotite, and alteration products of chlorite, saussurite, epidote, antigorite (?) and calcite. However, the rock locally grades into quartz-monzonite. The gneissic variety in the north has the appearance of a plagioclase cumulate with minor quartz, microcline and dark minerals filling the interstices between the blocky plagioclase grains, and microcline rimming and replacing (?) some of the plagioclase which itself is highly saussuritised and partly replaced by (?) unaltered albite. The slightly gneissic diorite in the west shows similar features, but is finer grained (chill-margin) and has a higher percentage of microcline and perthitic orthoclase (which is partially inverted to microcline). The plagioclase is andesine and albite. At the head of Big Bight the diorite is composed of 80% andesine as interlocking grains with local pockets of microcline and perthitic orthoclase which has partly inverted to microcline. Hornblende occurs as subhedral acicular needles and antigorite (?)

Fig. 4-3. Intersection of tectonite fabric (S_2), defined by flattened conglomerate pebbles, and bedding (trending from lower right to upper left in the photograph) in the Makkovik Formation.

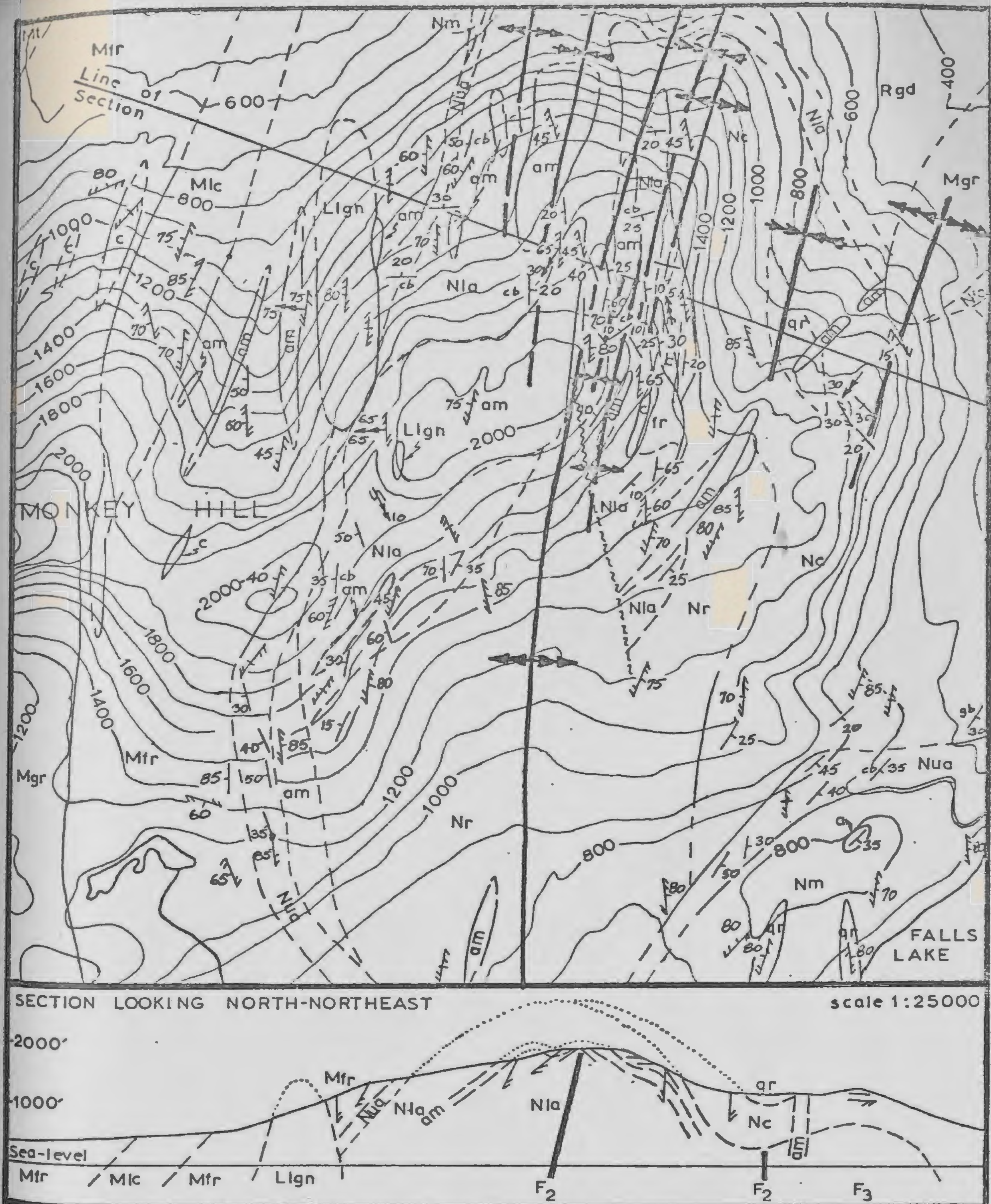


Fig. 4-4. Detailed geological map and cross-section of the Monkey Hill/Falls Lake area. See Plate 1 for legend.

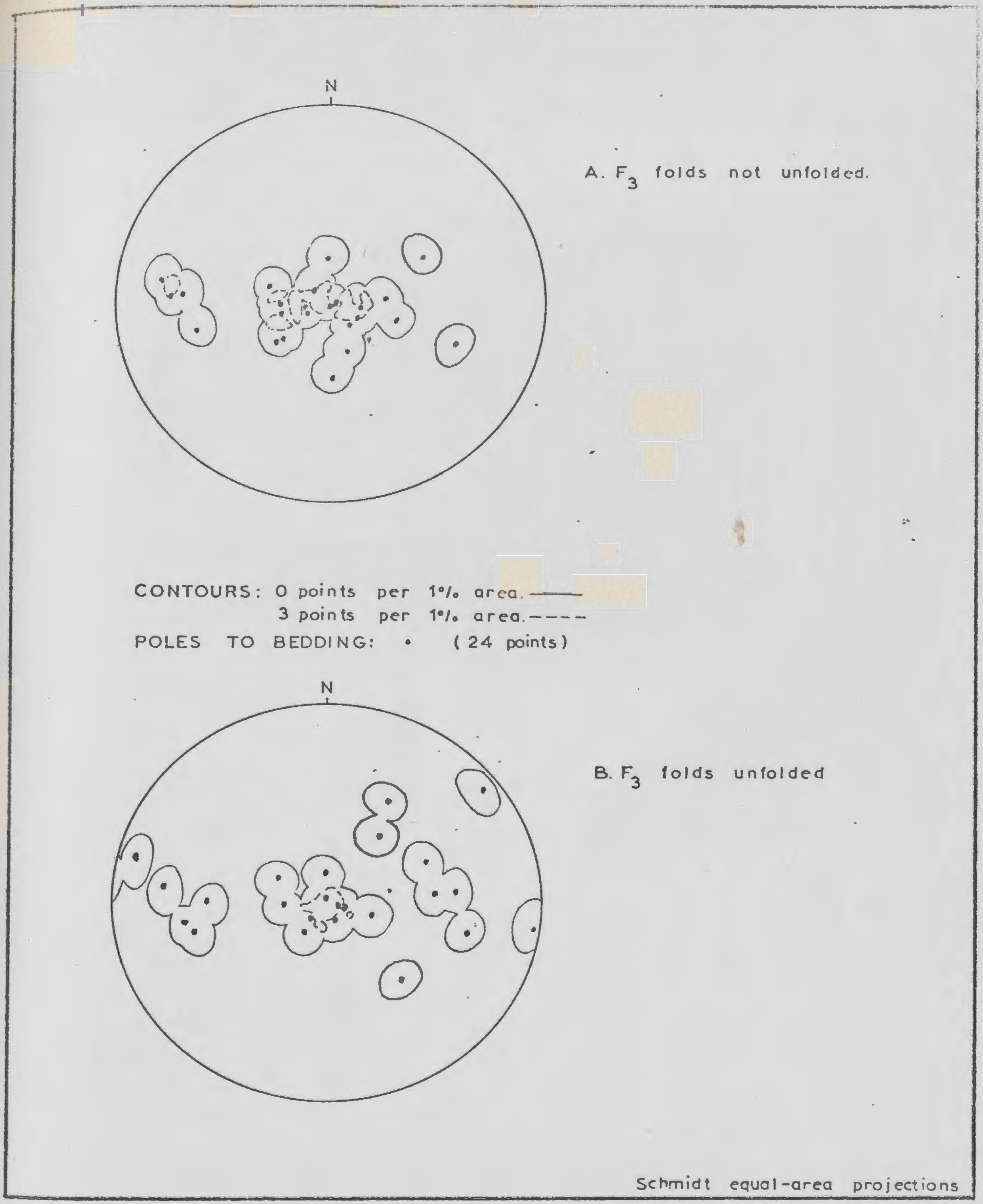


Fig. 4-5. Stereographic projections of bedding in the eastern part of Monkey Hill, with F_3 folds not unfolded and unfolded.

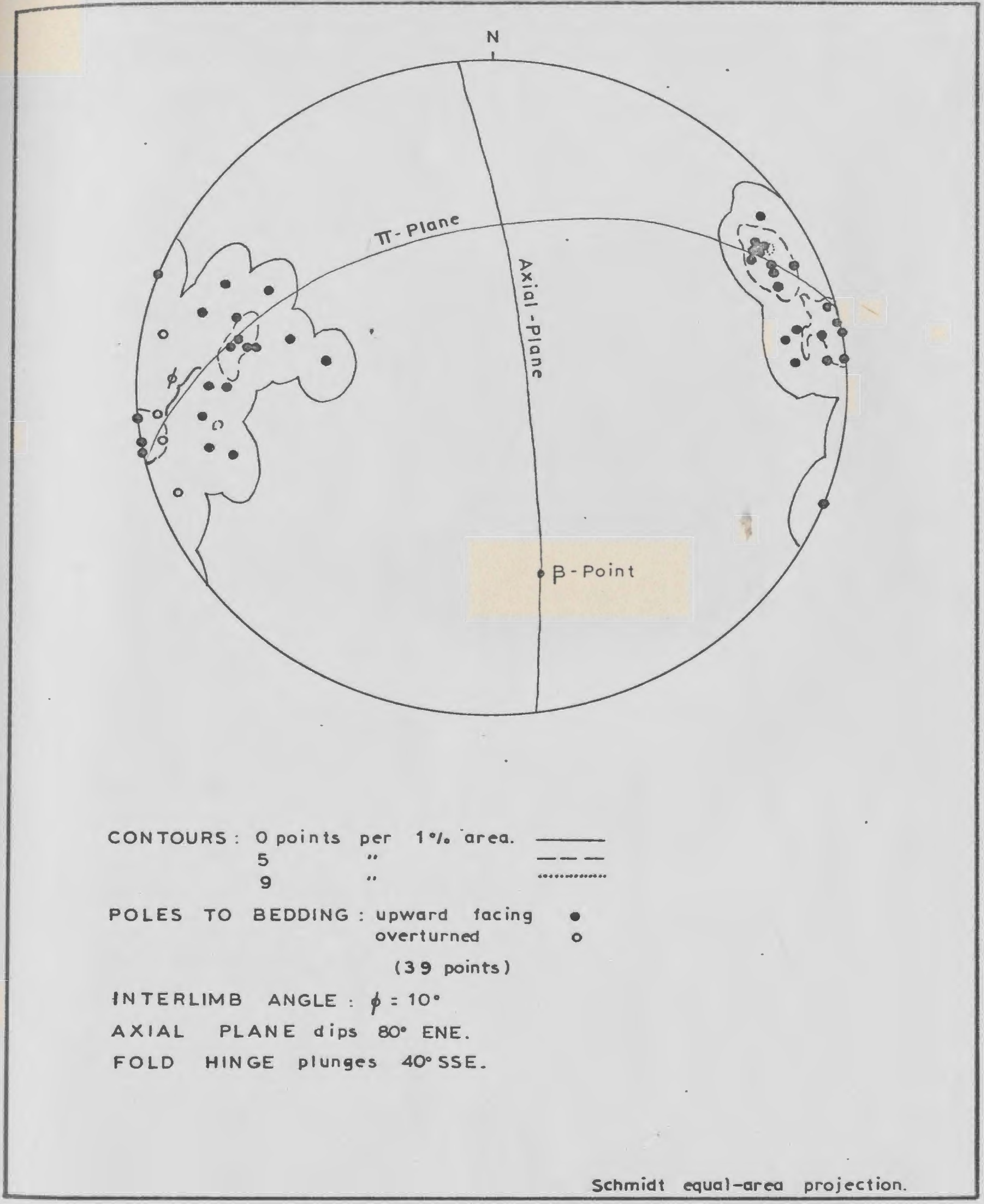


Fig. 4-6. Stereographic projection of bedding in the vicinity of Manak Bay.

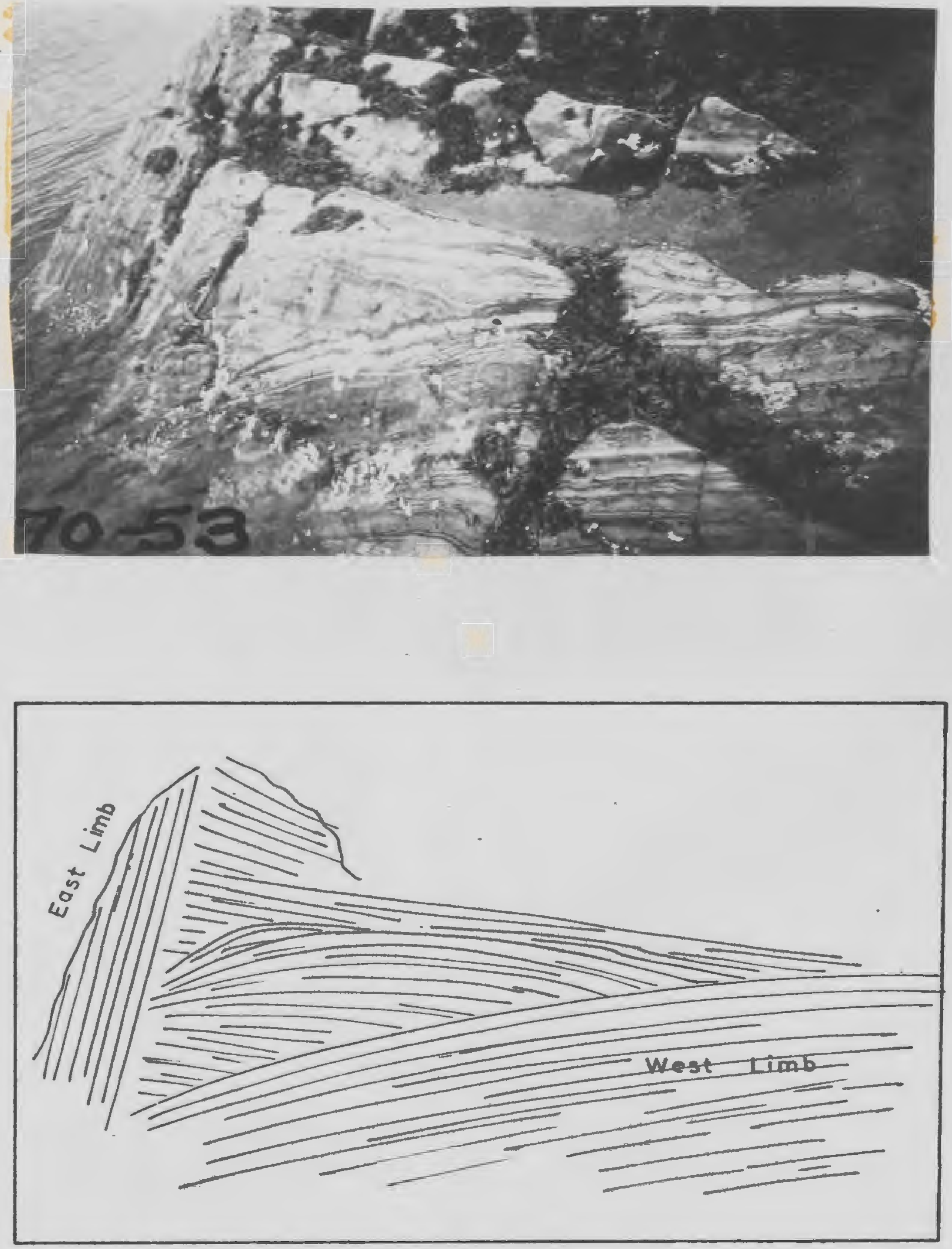


Fig. 4-7. Photograph of sketch of hinge-zone of F_2 syncline on the south shore of Manak Bay (looking south), showing brittle fracturing.

To the west of the Round Pond dome two major synclines occur, one northward plunging at a low angle through Retreat Lake, and one with near horizontal plunge through Tilt Cove. This latter appears to have a closure immediately south of Aillik Bay (Gandhi, et al., 1969) and is markedly modified by later folding on Big Island. These synclines are separated by a major slide (see below) and not by an anticline.

Minor F_2 Folds

Minor F_2 folds are generally common in amphibolite dykes but uncommon in bedded units and mafic lavas. Where they are well developed, as on Big Island and elsewhere, they are usually too tight, or have been affected by sliding, for their asymmetry to be recognised. However, where asymmetry is recognised it conforms to the major fold patterns described. The folds are usually small, irregular, asymmetrical, similar-type folds of earlier banding or schistosity (Figs. 4-8, 4-31, 4-32 and 4-37), but primary flow-folds in the flow-banded rhyolites tightened by D_2 (Fig. 4-9) are also common and folds of bedding (Figs 4-10 and 4-11) do occur.

Small tight folds in the Ranger Bight Complex at Ranger Bight occur in an auge around which the S_2 fabric is bent (Fig. 4-12), and are oriented at approximately 20° to the external S-fabric. The folds themselves show a well developed axial planar biotite S-fabric. Although the folds may be due to a pre-D_2 deformation, they are, for lack of

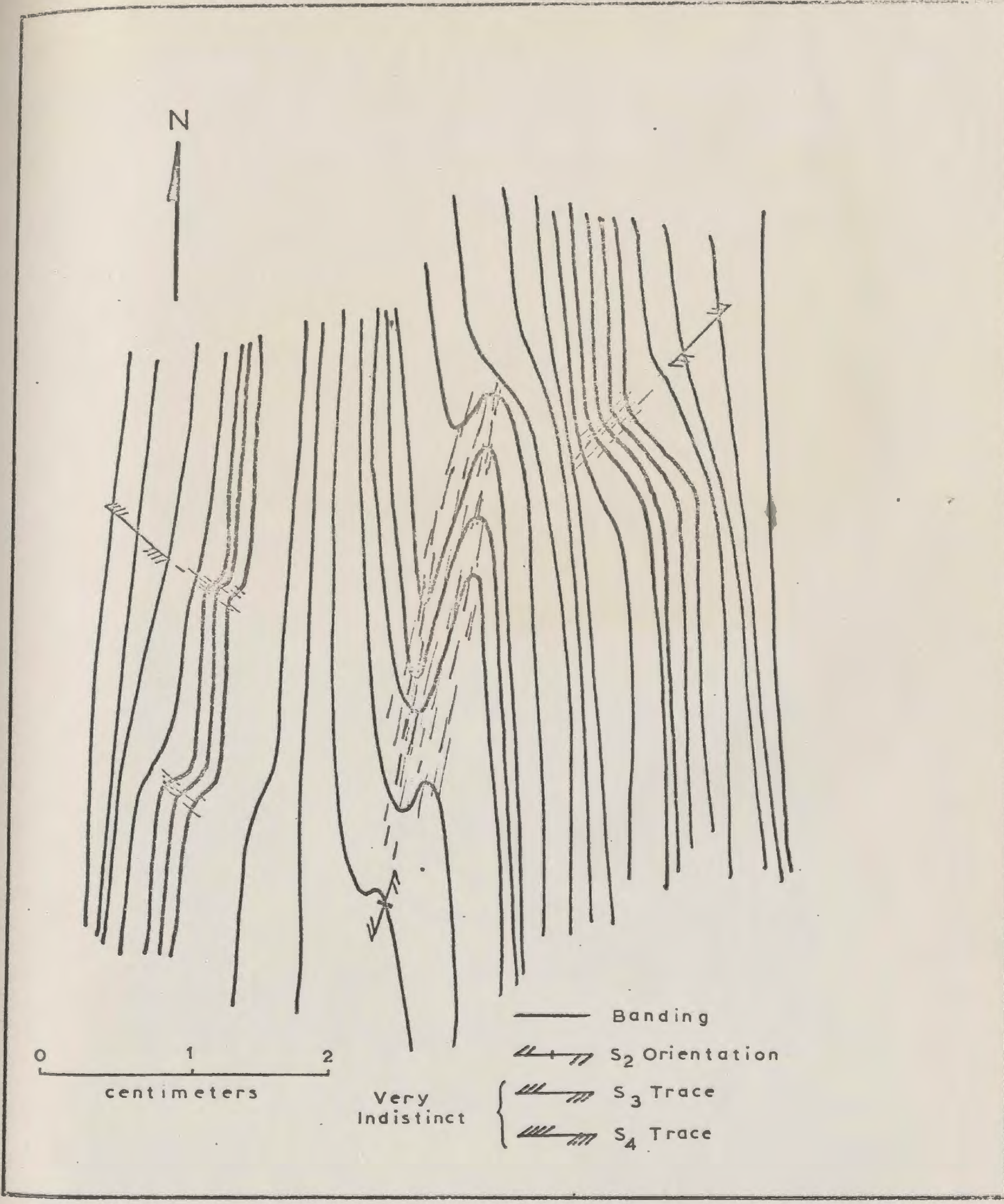


Fig. 4-8. Sketch of small folds in amphibolite on west coast of Ford's Bight.

Fig. 4-9. F_2 folds or D_2 tightened primary (flow) folds of flow-banding in lower rhyolite member of the Pomiadluk Point Formation. Pomiadluk Point.

Fig. 4-10. F_2 folds of bedding in the transgressive arkose member of the Makkovik Formation. Southeast of Falls Lake.

Fig. 4-11. Small tight F_2 folds of bedding in a tuffite unit in the quartz-porphyritic rhyolite member of the Makkovik Formation. Note extreme attenuation of limbs and development of isolated fold hinges giving a "pebbly" appearance to the unit.

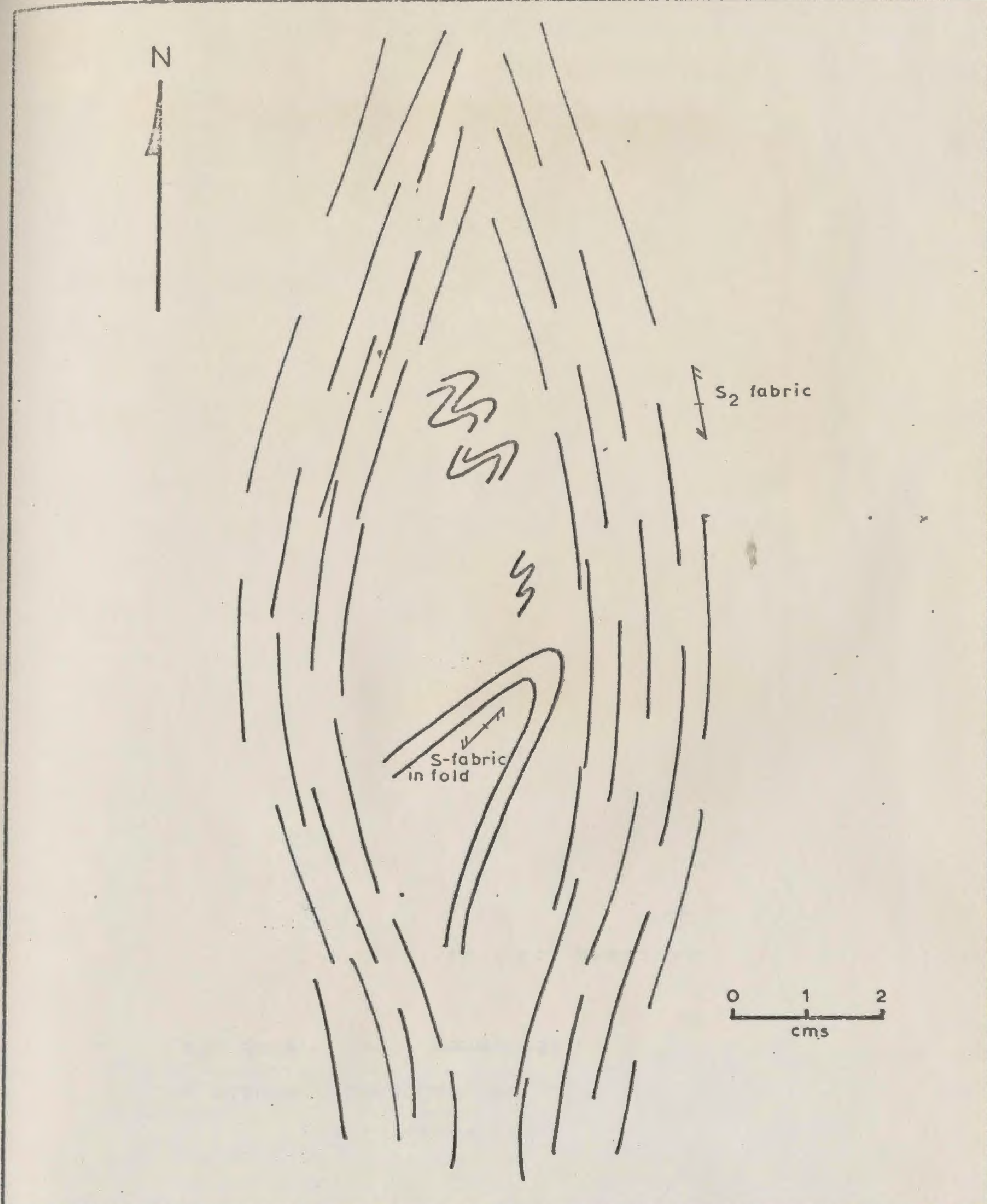


Fig. 4-12. Sketch of early formed F_2(?) folds forming an auge with later S_2 schistosity. Ranger Bight Complex west of Ranger Bight.

as interstitial radiating sheaves of fine acicular fibres.

vi. Associated Dykes and Veins: A 1 to 2 m. (3 to 6 ft.) wide hornblendite dyke intruded into the diorite facies west of Big Bight consists of 80% primary zoned euhedral to subhedral hornblende with interstitial plagioclase and opaque minerals, accessory euhedral apatite and secondary biotite (10 to 15%), epidote and saussurite. The small intrusion of diorite facies just south along the coast from Manak Bay has innumerable small veins and apophyses extending into the host wacke. These veins commonly give the dark grey wacke the appearance of a pillow-lava or net-vein intrusion (Fig. 3-6). Many true net-vein dykes (e.g. the diabase/Grampus Cove Gneiss net-vein dyke) occur in the vicinity of the Adlavik Igneous Complex and may well be associated with it. However, for lack of evidence, they will be treated separately in this thesis, as will other diabase and diorite dykes.

A pale pink, fine- to medium-grained, 1 to 2 m. (3 to 6 ft.) wide feldspar dyke is intruded into the amphibolites on the point at the head of Big Bight. The dyke appears to be an off-shoot of the diorite facies, but no direct evidence is available. The rock consists of:

K-feldspar	75%
andesine	20%
calcite	5%
dark minerals	10%

further evidence, assumed to be early F_2 folds which were later involved in D_2 boudinage, but not unfolded (Ramsay, 1967). Another occurrence of such early-formed F_2 folds is on the north coast of Big Island west of the Ranger Bight Complex, in a calcareous banded quartzo-feldspathic unit (Fig. 4-13). Both sets of folds indicate a southward-plunging syncline or northward-plunging anticline to the west, the first interpretation of which is in agreement with the location of the syncline through Tilt Cove.

D_2 Boudinage

The large clasts in the Pomiadluk Point Formation conglomerate have undergone D_2 boudinage (Figs. 4-14 and 4-36), with slight rotation of the boudins relative to one another indicating a maximum stress at an angle to the original boulder orientation. The long-axes of the boudins are steeply dipping and perpendicular to L_2. The sense of intersection of S_2 with the pre-D_2 (primary?) elongation of the boulder in conjunction with the upward and eastward facing direction of the beds indicates a southward plunging F_2 syncline to the east.

Well developed D_2 boudinage of the Manak Bay Formation arkose occurs throughout the Manak Bay area, (Fig. 4-15). The boudins are steeply dipping and are simple regular bodies up to 1 m. (3 ft.) in cross-sectional length and about 30 cm. (1 ft.) wide. The individual boudins tend to die out over

Fig. 4-13. Small early F_2(?) folds in a calcareous arkose unit of the Big Island Formation. North coast of Big Island.

Fig. 4-14. D_2 boudins developed from granitic boulders in the lower conglomerate, Pomiadluk Point Formation. Pomiadluk Point.

Fig. 4-15. D_2 boudins in the arkose member of the Manak Bay Formation. Northwest shore of Manak Bay.

Fig. 4-16. D_2 boudins in arkose member on northwest shore of Manak Bay. The boudins are shown as steeply dipping (to the left) warps on the vertical bedding surface.

a distance of several feet along the long axis directions indicating $k > 1$ (Fig. 4-16). Small scale D_2 boudinage of bedding, rhyolite banding and dykes is common (Figs. 4-17 and 2-39). The boudin necks in all cases are filled by quartz and microcline. The Ranger Bight Complex shows possible large-scale boudinage at Ranger Bight (Fig. 4-18).

The Ranger Bight Slide and Other D_2 Slides

On the coast northwest of Ranger Bight the adjacent limbs, as shown by bedding/S_2 intersections, of the syncline through Ranger Bight and that through Tilt Cove, are situated within 100 m. (300 ft.) of one another, with no intervening anticline (Fig. 4-19). The intervening rocks are highly deformed amphibolites of the Ranger Bight complex, and are postulated to be situated at, and partly intrusive into a major slide zone--the Ranger Bight slide. This slide zone extends southwards to Cross Lake and is occupied throughout by the Ranger Bight complex, which shows both D_2 and pre-D_2 tectonic fabrics. It is oriented at a small angle to both S_2 and to the syncline through Ranger Bight (Figs. 4-18 and 4-20). It has not been possible to determine the displacement on the slide as there is no distinctive lithology or sequence of lithologies on either side that can be correlated.

Small-scale D_2 slides are common in the vicinity of the major slide (Figs. 4-21, 4-22 and 4-23).

Fig. 4-17. D_2 boudins in a small tuffite layer in banded arkoses of the Big Island Formation. West of Ranger Bight.

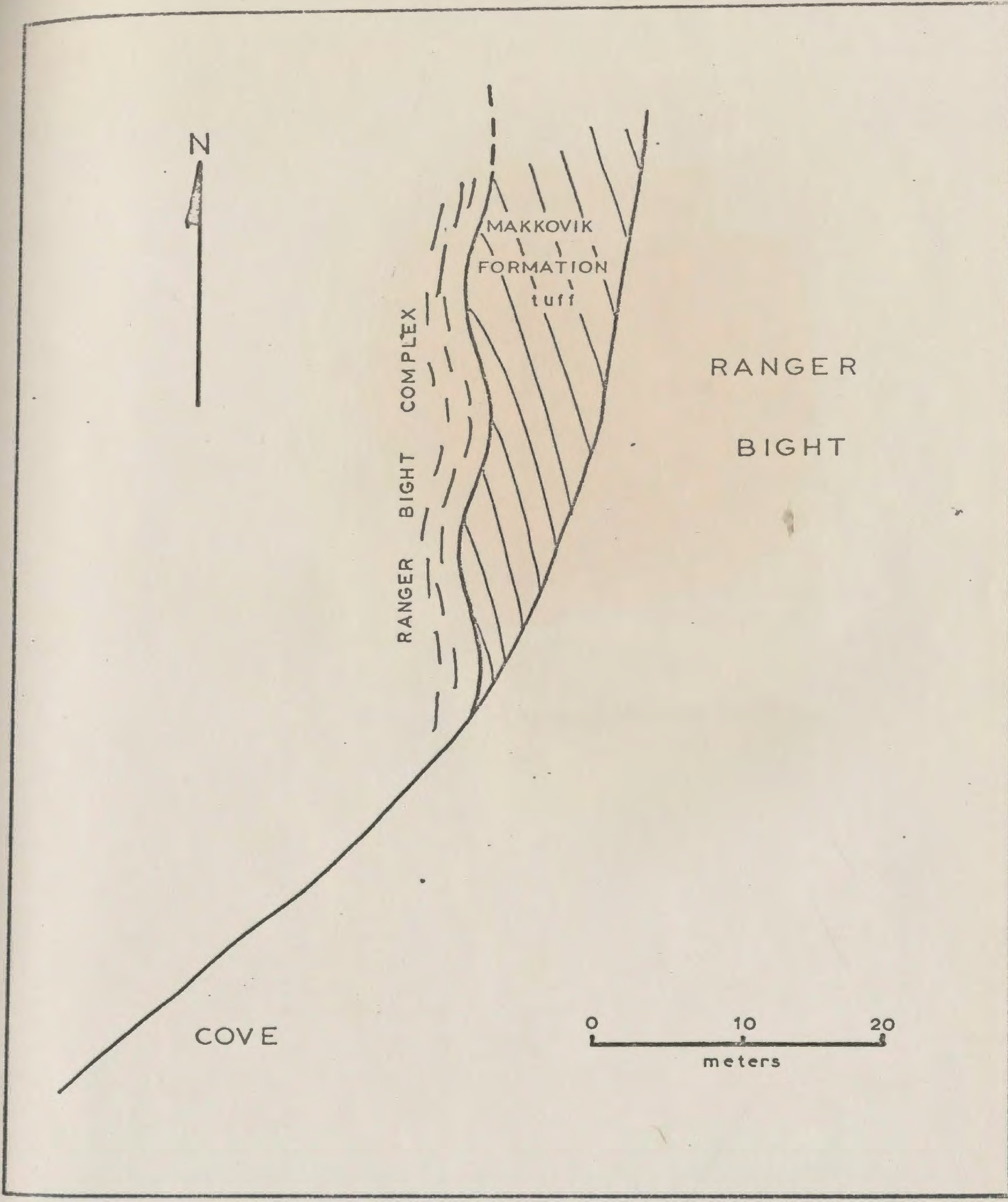


Fig. 4-18. Sketch-map of boudin-like boundary to the Ranger Bight Complex on west coast of Ranger Bight.

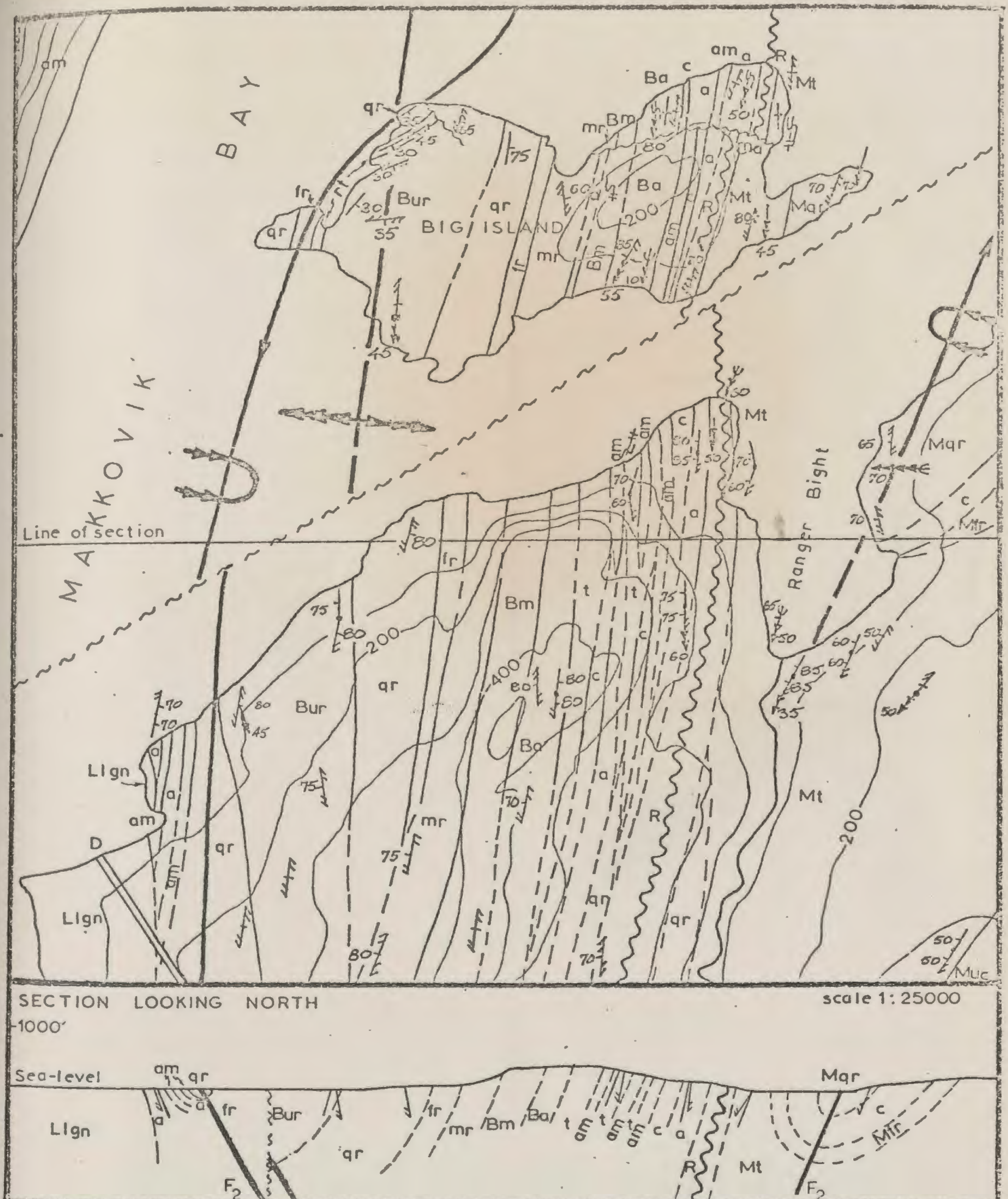


Fig. 4-19. Detailed geological map and cross-section of the Big Island/Ranger Bight area. See Plate 1 for legend.

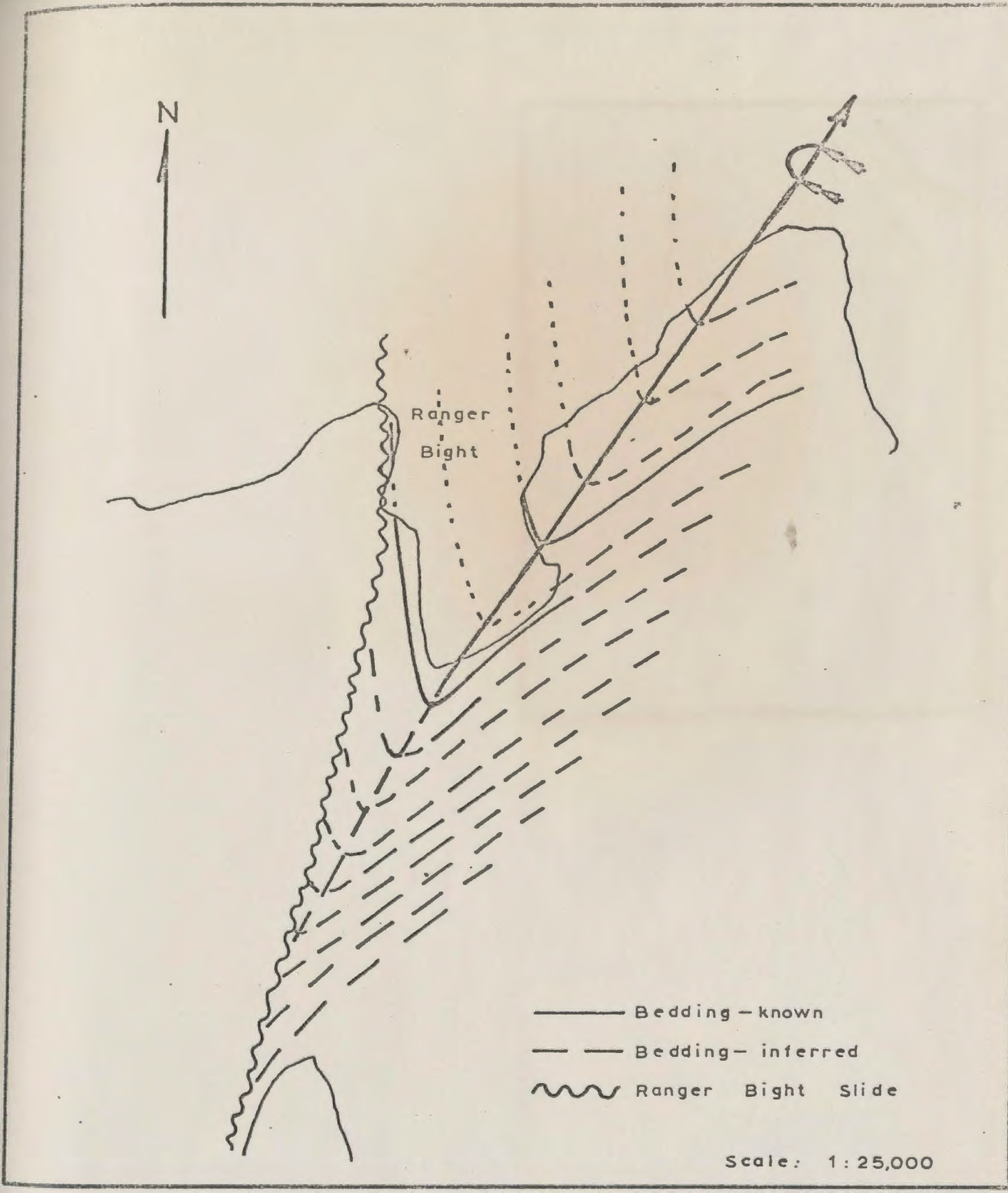


Fig. 4-20. Sketch map showing intersection of the Ranger Bight (D_2) slide and the Ranger Bight (F_2) syncline.

Fig. 4-21. D_2 slide in banded arkose of the Big Island Formation. South coast of Big Island.

Fig. 4-22. F_2 folds and associated D_2 slides in banded arkose of the Big Island Formation. South coast of Big Island.

Fig. 4-23. D_2 slides in interlayered arkose and conglomerates of the Big Island Formation. Coast west of Ranger Bight.

Fig. 4-24. D_2 lobate structures developed in an amphibolite dyke in the west coast of Wild Bight.

The K-feldspar is orthoclase, with allotriomorphic granular texture, which has largely inverted to microcline. The dark minerals consist of opaque minerals, serpentine and clay mineral alteration of the feldspars.

vii. Discussion of the Complex: The Adlavik Igneous Complex is a hornblende gabbroic intrusion which appears to have developed as the result of normal differentiation and crystal fractionation of a water-rich gabbroic magma to give primary hornblende-cumulate rocks instead of the more normal pyroxene cumulates. During the crystallisation of the rhythmic layered facies the water vapor pressure appears to have varied significantly as shown by reaction and partial dissolution of the early hornblende crystals followed by crystallisation of pyroxene rims, followed again by further hornblende crystallisation. Similarly, in the massive gabbro facies, hornblende is both a primary mineral and a reaction product of the pyroxene.

The complex appears to be a differentiated gabbroic intrusion of appinitic affinities (Pitcher and Berger, 1972) as shown by the occurrence of hornblende as a major primary mineral, and the development of many of the fabrics and microstructures of appinites (e.g. actinolite-cored hornblende, zoned early brown to later green hornblende grains, etc.). However, although net-vein dykes of various types are abundant in the vicinity of the complex, intrusive breccias, Other D_2 Structures

Lobate structures developed between rocks of different conpetencies (Ramsay, 1967), usually amphibolite and the quartz- or feldspar-porphyries, are common throughout the area (Figs. 3-15 and 3-24).

CALCULATION OF AMOUNT AND TYPE OF DEFORMATION

The fabrics and structures that developed during D_2 offer no evidence in themselves of simple-shear deformation. However, when taken as a whole, the large-scale sigmoidal attitude of F_2 fold-axes and S_2 fabric, and the sub-horizontal lineation suggest large-scale simple-shear deformation may have occurred (see p. 203). Nevertheless, because of the lack of any conclusive evidence for simple-shear, the calculations and discussion in this section and the following section are based on the assumption that D_2 was a pure-shear deformation. The case of simple-shear deformation will be discussed separately (p. 203).

The amount of deformation (r; Watterson, 1968) and type of pure-shear deformation (Flinn, 1962) vary throughout the Makkovik/Kaipokok region (Clark, 1970 and 1971; Sutton, et al., 1971; and Sutton, 1972b). It was therefore decided to try and calculate these factors where possible.

The amount of deformation and type of pure-shear deformation can be calculated if the post-tectonic shape, distribution or orientation can be determined for objects of known pre-tectonic shape, distribution or orientation. These methods calculate the total strain in the rocks, but as early flattening was at an angle to D_2 (see later) and the strain axes parallel the D_2 fabric in the enclosed or associated mafic rocks it is assumed that earlier deformational events contributed a negligible amount of strain to the total, which is therefore considered representative of the D_2 strain only. Of the various methods available the following were found useful in the area:

1. Direct measurement of shape of ellipsoids developed by deformation of originally spherical bodies (Ramsay, 1967). Pebbles in the Nesbit Harbour Formation conglomerate east of Round Pond were photographed in two planes containing the strain axis. The relative orientations of the photographs to the D_2 fabric elements was determined in the field, and measurements later made of the pebble axis-lengths on prints of the photographs.

A sample of the lithophysic rhyolite on Big Island was sectioned parallel to the three principle planes and the sections of the lithophysae measured.

A possible third group of measurements, of the shape of deformed phenocrysts in the porphyritic rhyolites, was not attempted as the subgrains are thought to be too coarse to allow measurements with enough accuracy to be made (Figs. III-3, III-4, III-5 and III-8).

2. The strain ellipsoid has been calculated from the relative distribution of phenocrysts in the porphyritic rhyolites according to Ramsay's Method 3 (Ramsay, 1967, p. 195), wherein random distribution of phenocrysts may be represented by an imaginary sphere of radius equal to the average distance between phenocrysts, which on deformation is transformed into

an imaginary ellipsoid directly related to the deformation ellipsoid. A simple computer program has been written to replace the graphical part of Ramsay's method and so to increase the accuracy of the method.

A more complete description and discussion of these methods is given in Appendix I.

It is assumed, for the purpose of calculating the k- and r-values in the area, that volume change of the rocks during deformation has been negligible. As there is no difference in density between deformed and undeformed rhyolites this assumption is valid for the calculations based on the rhyolites. However, although the conglomerate pebbles are unlikely to have undergone any volume change, the matrix may very well have decreased in volume due to dehydration, collapse of pores, recrystallisation (Ramsay and Wood, 1973). As only the pebble shapes were used to determine the amount and type of deformation the r-value calculated will be less than the true value, and the true k-value will be either more or less than the calculated value depending on whether pre-tectonic compaction or tectonic flattening caused the greater volume loss. However, as the pebbles are closely packed together in all directions, with little matrix between them, and show no sign of solution pitting, volume loss was probably negligible.

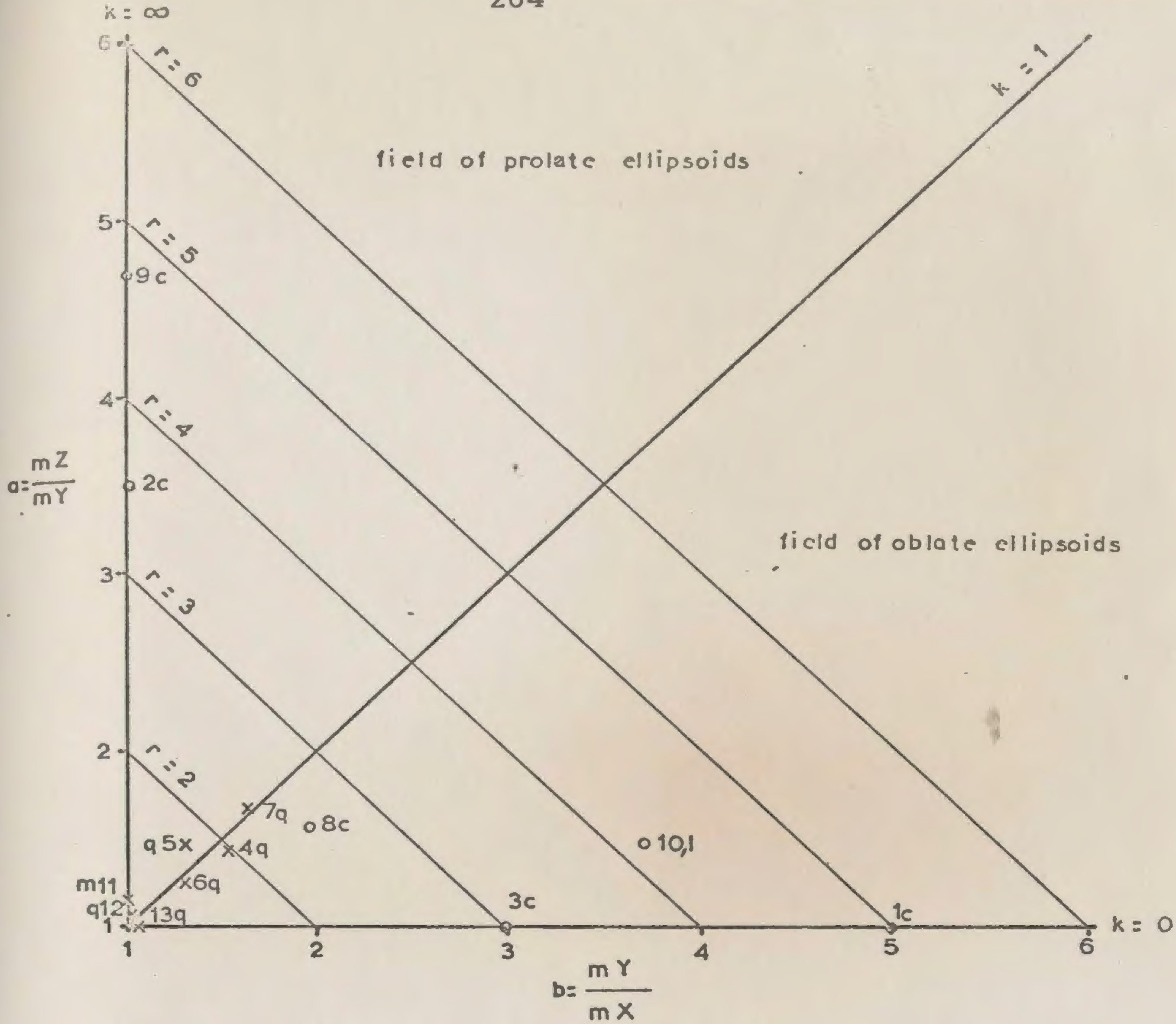
VARIATION IN AMOUNT AND TYPE OF DEFORMATION

A complete range of k-values, from zero to infinity has been recognised. However, in calculating the type of deformation a very small variation in a or b (for instance, in the second decimal place) can considerably alter the k-value, especially where a and b are close to 1 (as in points 12 and 13 in Fig. 4-25). As most of the rocks in the area apparently have a and b about2, and this low value is enhanced by a personnal bias which tends to make both a and b smaller than their true value in the particle distribution calculations (see Appendix I), the values for k for those points in the vicinity of the origin of the graph (Fig. 4-25) are likely to be unreliable. Nevertheless, the calculations do show that there is no consistency in the k-values throughout the area. There is no recognisable distribution or variation in k-values in the area (Fig. 4-26), though this is most likely due to the small number of calculations made, the wide dispersion of the samples and the inherent error in the calculations.

The calculations of the amount of strain are far less affected by errors in the values of a and b, though in the particle distribution calculations, the personal bias in the initial part of the calculations tends to reduce the value of r. As with the k-values, there is no recognisable variation in r-values in the area due to there being too few calculations made on samples distributed over too large an area, though a high range in values (from 1 to 5) is recognised. There is also no recognisable correlation between k, r, and the location of the sample.

Significance of Lineation Orientation and Style of Deformation

Under conditions of compressional tectonics where $k > 1$, and no rotation of stress axes occurs, mineral lineation which is nucleated and grown during the deformation develops



o Result determined by direct measurement of deformed bodies – c:conglomerate pebbles. l:lithophysae.

x Result determined by measurement of distribution of particles and calculation by computer – q:quartz-porphyritic rhyolite m:microcline-porphyritic rhyolite.

$k: \dfrac{a-1}{b-1}$ (Flinn, 1962 and 1965).

r: a + b - 1 (Watterson, 1968).

Fig. 4-25. Graph of a against b (Flinn, 1962) showing variation in amount (r) and type (k) of strain in the thesis area.

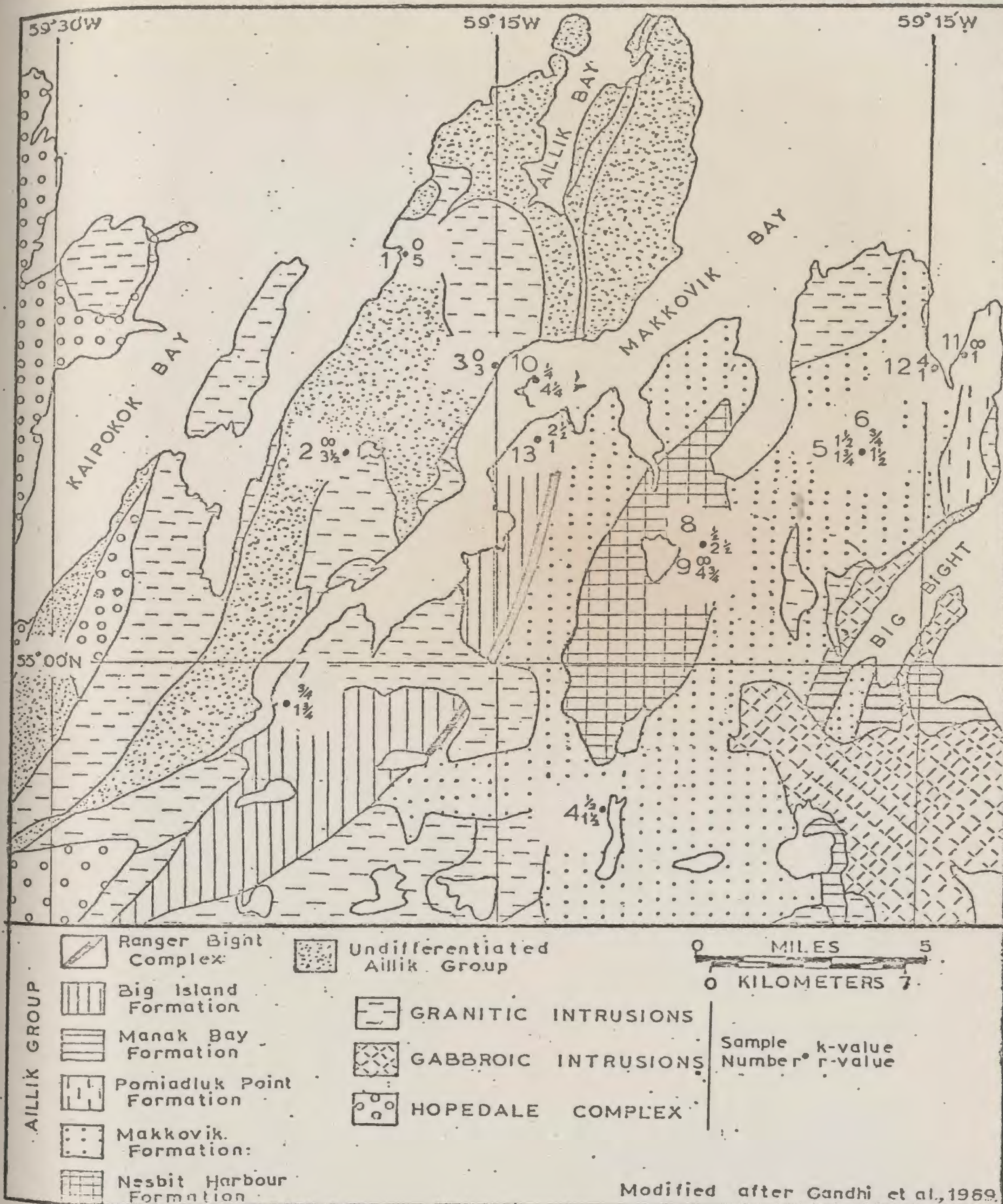


Fig. 4-26. Map showing type (k) and amount (r) of deformation in different parts of the area.

perpendicular to the axis of maximum stress and previously formed assymmetric minerals, whether forming a lineation or not, will be rotated towards the axis of maximum stress (Flinn, 1962). Where large-scale folds (e.g. nappes) are developed in sub-horizontal strata, and the folding acts as a means of vertical tectonic transport, the mineral lineation will develop perpendicular to the fold axes (Fig. 4-27).

Under conditions of simple-shear (k = 1), mineral lineation develops perpendicular to the intermediate axis, and within the sigmoidally-shaped plane of flattening (Ramsay and Graham, 1970). If the pole to bedding is sub-parallel to the intermediate stress axis, folds with sigmoidal axes parallel to the mineral lineation will develop in the shear-zone (Fig. 4-28).

Within the thesis area bedding is sub-horizontal on a regional scale. The L_2 mineral lineation shows a wide range of orientation, but is predominantly sub-horizontal and parallel to the F_2 fold hinges. The major F_2 fold hinges are sigmoidally shaped on a regional scale, together with the sub-parallelism of the L_2 mineral lineation, suggests that the entire thesis area may occur within a large simple-shear zone, the northwestern side of which moved towards the northeast with respect to the southeastern side. The variation in k-values and L_2 orientation indicate that if a shear-zone exists, it has been modified by pure shear.

The shear-strain (γ) varies regularly in amount and attitude through a shear-belt (Ramsay and Graham, 1970) and it is possible to determine the limits of a shear-belt from

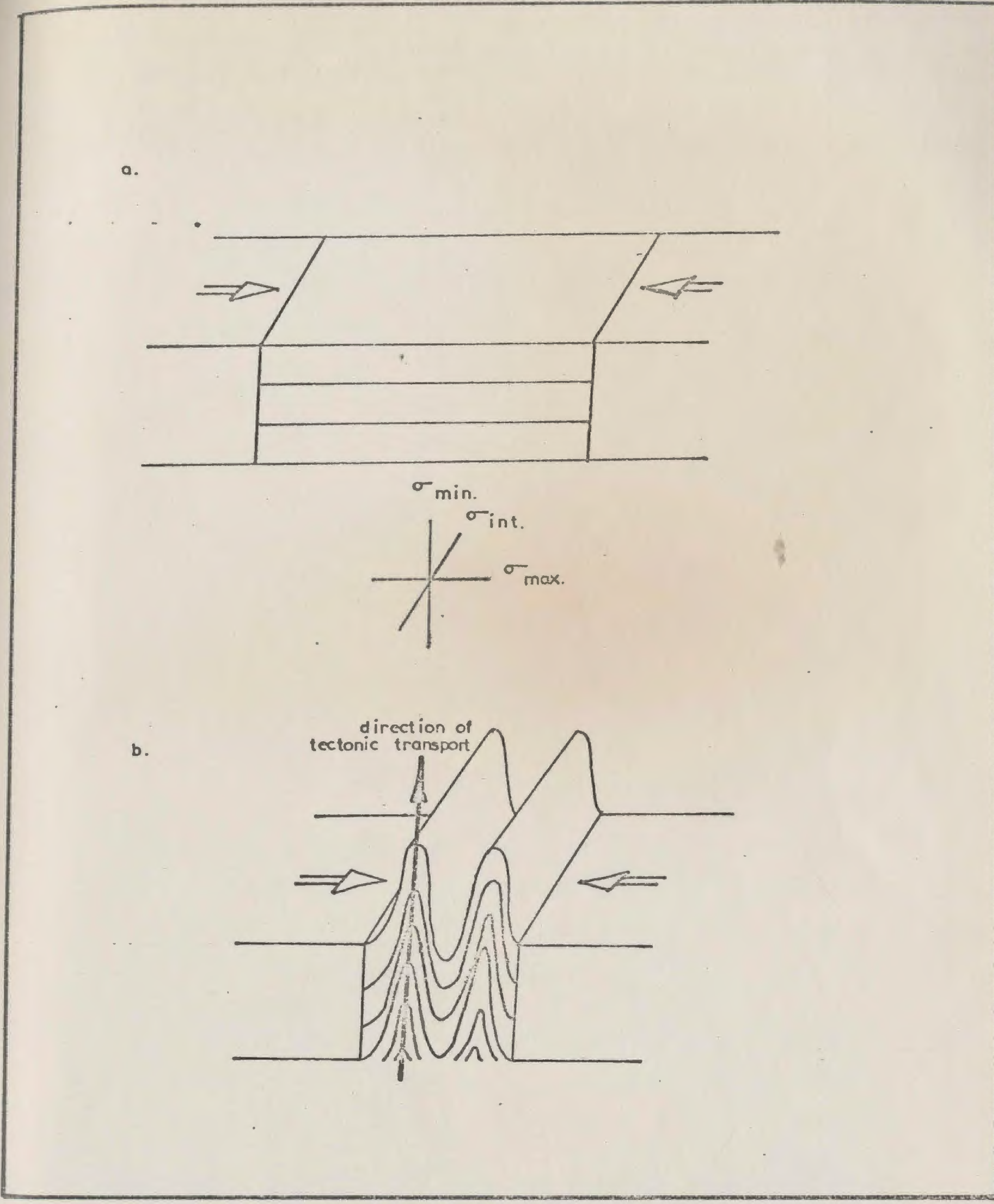


Fig. 4-27. Diagramatic representation of relationship of mineral lineation to fold hinges in development of nappe structures during compressional tectonics.

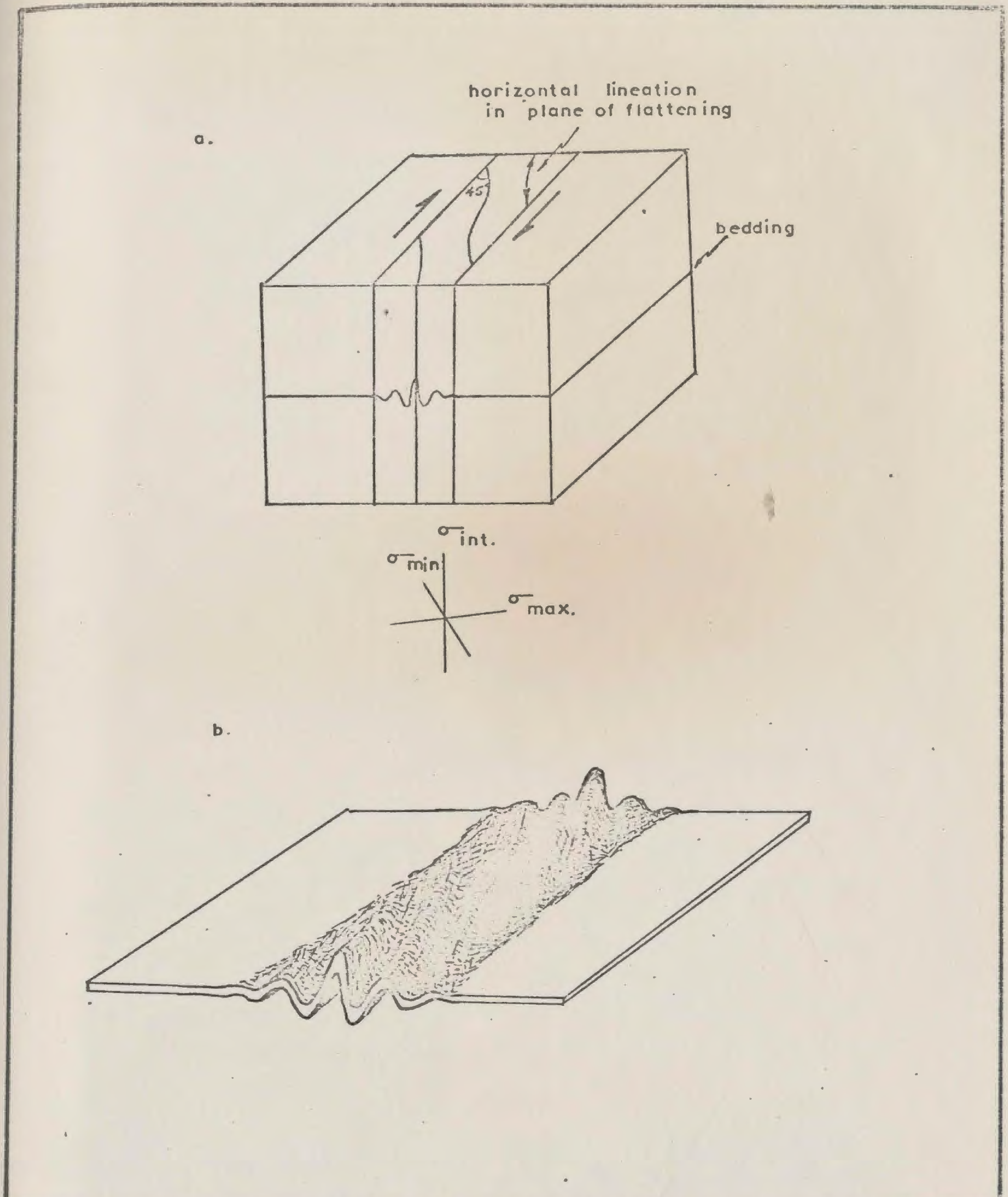


Fig. 4-28. Diagram showing relationship of mineral lineation and folds to simple-shear belt.

the relationship of the amount and attitude of γ. As the amount of deformation (r) at any location in a simple-shear belt is proportional to the shear strain (γ) (Ramsay, 1967, p. 68) and the strike of the schistosity (which is sub-vertical in the area) is parallel to the stretching direction, a graph of r against S_2-strike (Fig. 4-29) should be similar to the graph of θ' in Figure 8 of Ramsay and Graham, (1970).

The results as shown on the graph (Fig. 4-29) are inconclusive. This may indicate that a simple shear-zone does not exist, but does not disprove it as the number of readings is small and the calculation of r is open to significant error.

Pre-D_2 Tectonic Features

INTRODUCTION

One well defined, though very locally developed, pre-D_2 deformation is recognised and termed the first deformation (D_1). Other pre-D_2 tectonic features that cannot be confidently assigned to the first deformation have also been recognised.

THE FIRST DEFORMATION (D_1)

The effects of the first deformation have only been recognised in a few localities in the area. For the most part, these localities are in the Big Island/Ranger Bight area, but D_1 features have also been recognised from which appear to be characteristic of appinitic suites, have not, as yet, been recognised.

October Harbour Granite

The October Harbour Granite occurs along the eastern coast of Pomiadluk peninsula from 1 km. (1/2 mi.) north of October Harbour and extending towards the south. The greatest width seen was 1 1/2 km. (3/4 mi.). The granite was considered by Gandhi, et al., (1969) to be related to the Strawberry Granite, but, for reasons to be discussed later (Chapter 6), it appears to be somewhat older. The granite intrudes the Pomiadluk Point Formation, into which are also intruded many associated granitic and pegmatitic dykes and veins.

The granite is a coarse-grained, orange coloured homogeneous rock which shows a weak tectonite fabric in the vicinity of the margins that is parallel to a regular imperfect orientation of the larger primary feldspar laths within the granite (see p. 270). In thin-section the rock is seen to be composed of:

quartz	25 - 35%
microcline	50 - 60%
albite and andesine	15 - 20%

with minor and accessory amounts of opaque minerals, sphene, muscovite, schorlite, biotite and hornblende alteration rims to biotite, which all usually occur in small aggregates suggesting a xenolithic origin. The microcline is coarsely

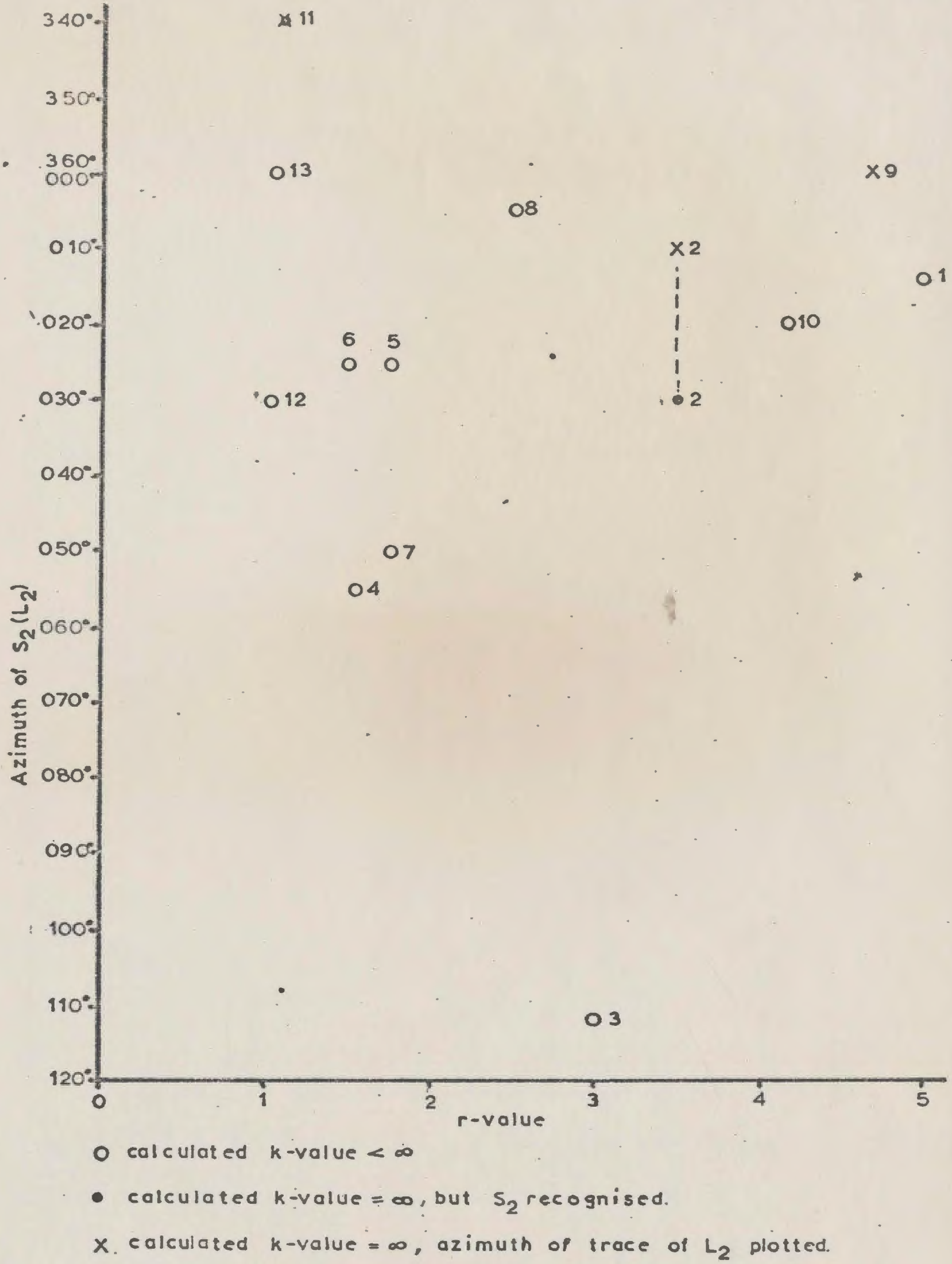


Fig. 4-29. Graph showing relationship of amount of deformation (r) to azimuth of S_2 or L_2 fabric.

rocks in the vicinity of Falls Lake and Pomiadluk Point. The distinct folding of these fabrics by F_2 and their recognition over a very large area indicates they are a regional deformational feature and not just local effects.

D_1 Tectonite Fabrics

The D_1 tectonite fabrics are defined by mineral orientation and have only been recognised in rocks of high biotite or hornblende content, usually metamorphosed basic dykes and lava flows, where they are folded by F_2 folds or partially transposed to the S_2 orientation in crenulation cleavage.

Well developed S_1 and L_1 biotite-hornblende fabrics and associated banding, and poorly developed L_1 fabric of diopside occur in some of the amphibolites of the Ranger Bight Complex, and also in some early basic dykes and the Big Island Formation metabasalt west of the complex. The L_1 and S_1 fabrics are commonly folded by small F_2 folds (Figs. 4-30, 4-31 and 4-32) and partially transposed to L_2 and S_2 (Figs. 4-33 and 4-34). Pre- to syn-D_2 scapolite and hornblende porphyroblasts with an included straight to sigmoidal S_1 fabric of feldspar, biotite or hornblende grains, occur in some of the amphibolites, particularly the metabasalt south and east of Round Pond (Fig. 4-35). Granite boulders in the conglomerate at Pomiadluk Point contain a pre-D_2 fabric of flattened xenoliths and mafic aggregates (Fig. 4-36) which may be an S_1 fabric, or a pre-depositional fabric.

D_1 Banding

Compositional pre-D_2 banding occurs in amphibolites

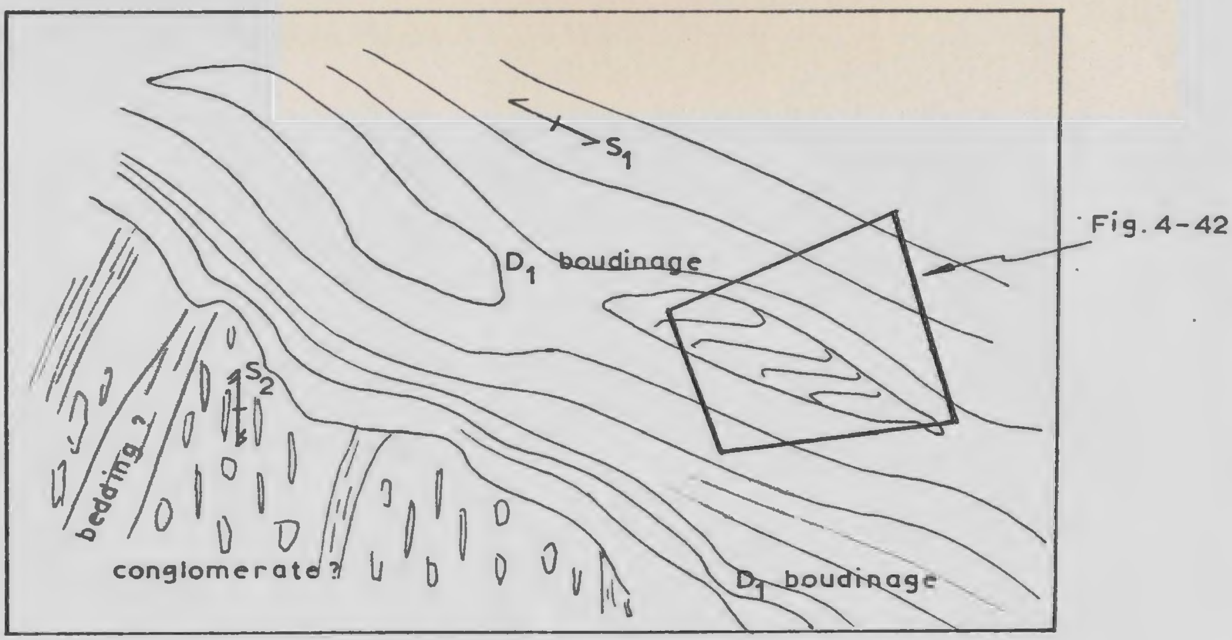


Fig. 4-30. Photograph and sketch showing relationship of amphibolitic bedded conglomerate(?), banded amphibolites and D_1-boudinised amphibolite dyke to D_1 and D_2 tectonic fabrics and structures. South coast of Big Island.

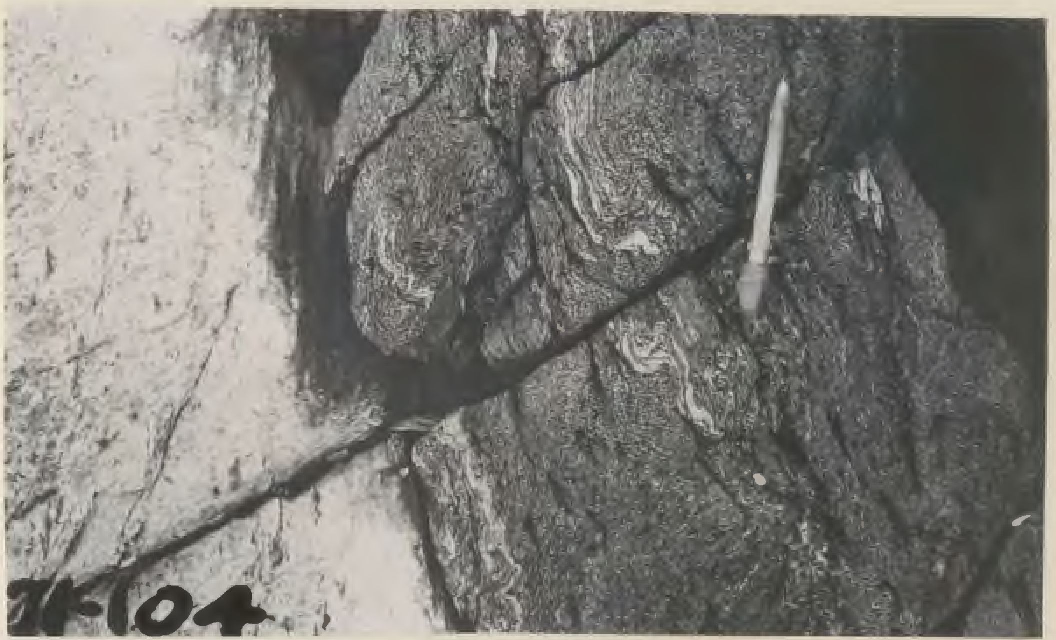


Fig. 4-31. Photograph showing S_1 fabric and D_1 flattened pegmatite veinlets folded by F_2 folds. Dyke on north coast of Big Island (see also Fig. 4-34). Pencil parallel to S_2.

Fig. 4-32. S_1 fabric folded by F_2 folds in amphibolite dyke on west coast of Ranger Bight. Pencil parallel to S_2.

Fig. 4-33. Photomicrograph of S_2 banding developed in margin of amphibolite dyke. East coast of Wild Bight. Plane polarised light, X10.

Fig. 4-34. Photomicrograph of S_1 biotite and diopside (d) fabric folded by F_2 folds and partly transposed to S_2 fabric in the hinges of the F_2 folds. Dyke on north shore of Big Island (Fig. 4-31).* Plane polarised light, X10.

*Note the diopside grains in the hinges of the folds appear to have been bodily rotated into their present position indicating pre-D_2 growth.

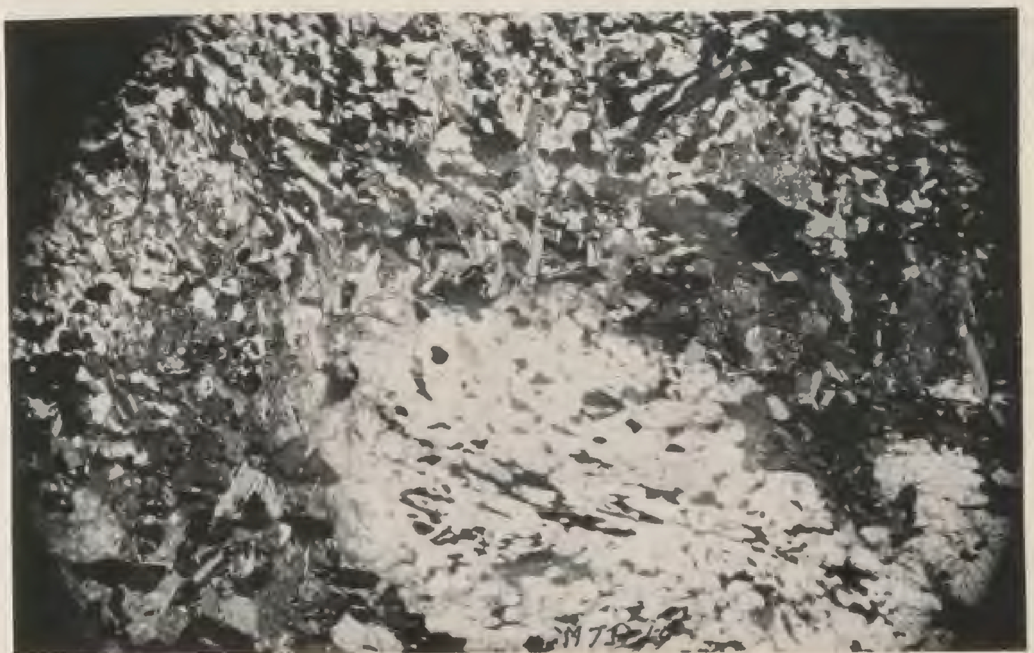


Fig. 4-35. Photomicrograph showing sigmoidal S_2 biotite inclusion trails in syn-D_2 scapolite (large white grain to lower right of photograph). Metabasalt member of the Nesbit Harbour Formation. East of Round Pond. Crossed-nicols, X10.

Fig. 4-36. D_2 boudins developed from granite boulder in lower conglomerate member of the Pomiadluk Point Formation. Note tectonite fabric in boulder is parallel to pencil. Pomiadluk Point.

at several localities in the Big Island/Ranger Bight area, and is defined by 1 to 2 mm. (1/20 to 1/10 in.) wide discontinuous monomineralic bands of quartz, epidote, diopside or hornblende and bimineralic quartz-feldspar bands (Figs. 4-37, 4-31, 4-32 and 4-38). The simple mineralogical nature of the bands suggests they may be derived from flattened phenocrysts or early veins, and some of the quartz-feldspar bands can be traced into small pegmatitic veins. Banding defined by variation in concentration of hornblende or biotite is common (Fig. 4-39) and may be due to transposition of a pre-D_1 fabric.

D_1 Structures

D_1 boudinage of early pegmatite veins and amphibolite dykes (Figs. 4-40 and 4-30) has been recognised and have been rotated into the D_2 stretching direction (Z-axis). Many of the flattened blebs which define an imperfect D_1 banding in the amphibolites may be due to F_1 folding and D_1 boudinage of earlier veins. No definite F_1 folds have been recognised.

The Ranger Bight slide is a D_2 feature and S_1 fabrics in the area are generally oriented at a high angle to S_2 and to the slide, suggesting it is not a rejuvenated D_1 slide. However, the larger number of pre-D_2 and pre-D_1 dykes in the vicinity suggest it may have been an area of local tension and weakness prior to D_2.

Fig. 4-37. F_2 fold of D_1 banding, with later (F_4) crenulation-folding. Amphibolite on north shore of Big Island. Coin is approximately 2 1/2 cm. in diameter.

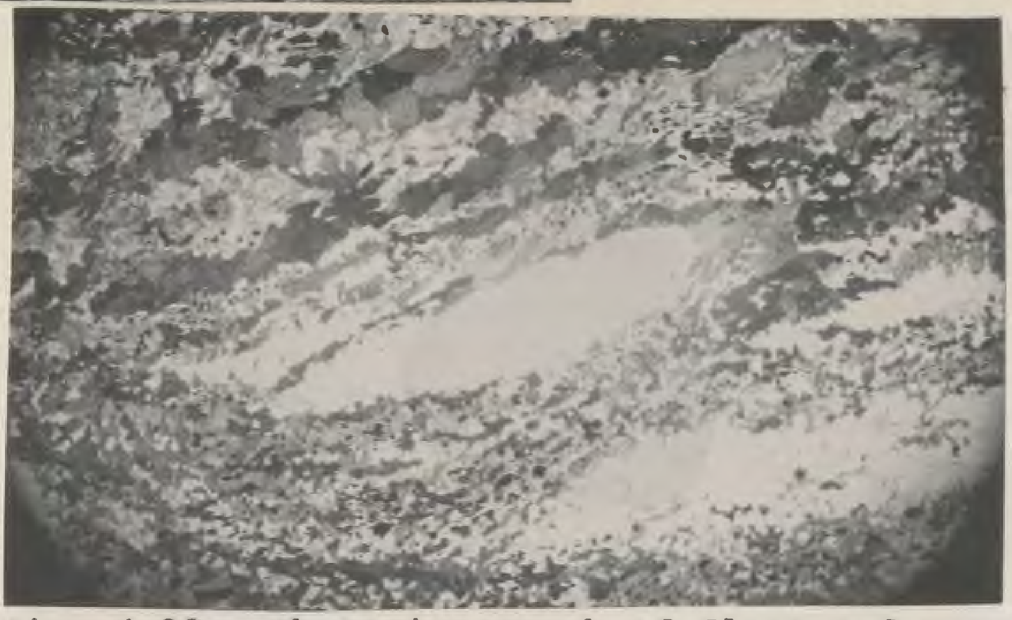


Fig. 4-38. Photomicrograph of flattened quartz blebs (boudins of early quartz vein?) in the Ranger Bight complex. West of Ranger Bight. Plane polarised light, X10.

Fig. 4-39. Photomicrograph of F_2 fold (closing to the left) of regular pre-D_2 biotite-rich banding. Biotite now defines S_2. Ranger Bight complex west of Ranger Bight. Plane polarised light, X10.

Fig. 4-40. D_1 boudins of pegmatite dyke folded by F_2 folds. Ranger Bight complex west of Ranger Bight.

perthitic.

INTRUSIONS EMPLACED AFTER THE MAIN DEFORMATION

Only two major intrusions were emplaced after the main deformation, the Monkey Hill Granite and the later Strawberry Granite. Both were emplaced before the last recognised phase of deformation in the area. A more thorough description, including chemical analyses, is given by Kranck (1953).

Monkey Hill Granite

The main outcrop of the Monkey Hill Granite forms a large 10 km. by 12 km. (5 by 6 mi.) body between Monkey Hill in the east and across Makkovik Bay in the west. Several smaller plutons of a similar rock outcrop south of Round Pond and elsewhere. The granite cuts the Aillik Group and has well developed chill margins. Numerous associated pegmatite and graphic granite dykes occur in the vicinity of the intrusions.

The Monkey Hill Granite is a homogenous, fine- to medium-grained (up to 2 mm. diameter), pink to grey equigranular rock. The rock is quartz monzonitic in composition and is composed of:

quartz	30 - 40%
microcline	20 - 30%
albite	30 - 40%

Accessory minerals include hornblende and sphene, and minor pyrite, arsenopyrite and molybdenite (?) in associated veins.

OTHER PRE-D_2 FABRICS AND STRUCTURES

Some of the discontinuous D_1 diopside bands in dykes in the Big Island/Ranger Bight area have pre-D_1 feldspathic inclusion trails (Fig. 4-41) which may represent a pre-D_1 tectonic event. Much of the D_1 biotite- and hornblende-rich banding in amphibolite dykes here is fine-scaled, 1/4 to 1/2 cm. (1/8 to 1/4 in.) in width, and regular (Fig. 4-39), which suggests it may have resulted by crenulation of an earlier fabric (Nicholson, 1966). An amphibolite dyke in the Ranger Bight Complex on Big Island has a pre-D_1 internal fabric which is deformed at the edges of the dyke and folded by F_1 folds (Fig. 4-42). Stoeterau (1970) has recognised a folded hornblende schistosity in scapolite porphyroblasts in the metabasalt east of Round Pond. These scapolites are considered to be pre- to syn-D_2 in age in this thesis, which implies the included fold is F_1 and the schistosity pre-D_1.

On the south coast of Big Island small, discontinuous, en echelon, green, microcline-rich, bands containing biotite occur in a plagioclase-rich, biotite-poor arkose (Fig. 4-43). The bands are arranged at approximately 45° to the main (D_1?) banding and are cross-cut by an S_2 biotite fabric. Because of their orientation and irregularly sigmoidal, lensoid shape, they are thought to be pre-D_2 tension-gashes.

Fig. 4-41. Pre-D_2 (pre-D_1?) diopside grain with sigmoidally shaped earlier (pre-D_1) included fabric of feldspar grains. External biotite fabric is S_1 partially transposed to S_2. Dyke on north coast of Big Island (Figs. 4-31 and 4-34), Plane polarised light, X40.

Fig. 4-41. Pre-D_2 (pre-D_1?) diopside grain with sigmoidally shaped earlier (pre-D_1) included fabric of feldspar grains. External biotite fabric is S_1 partially transposed to S_2. Dyke on north coast of Big Island (Figs. 4-31 and 4-34), Plane polarised light, X40.

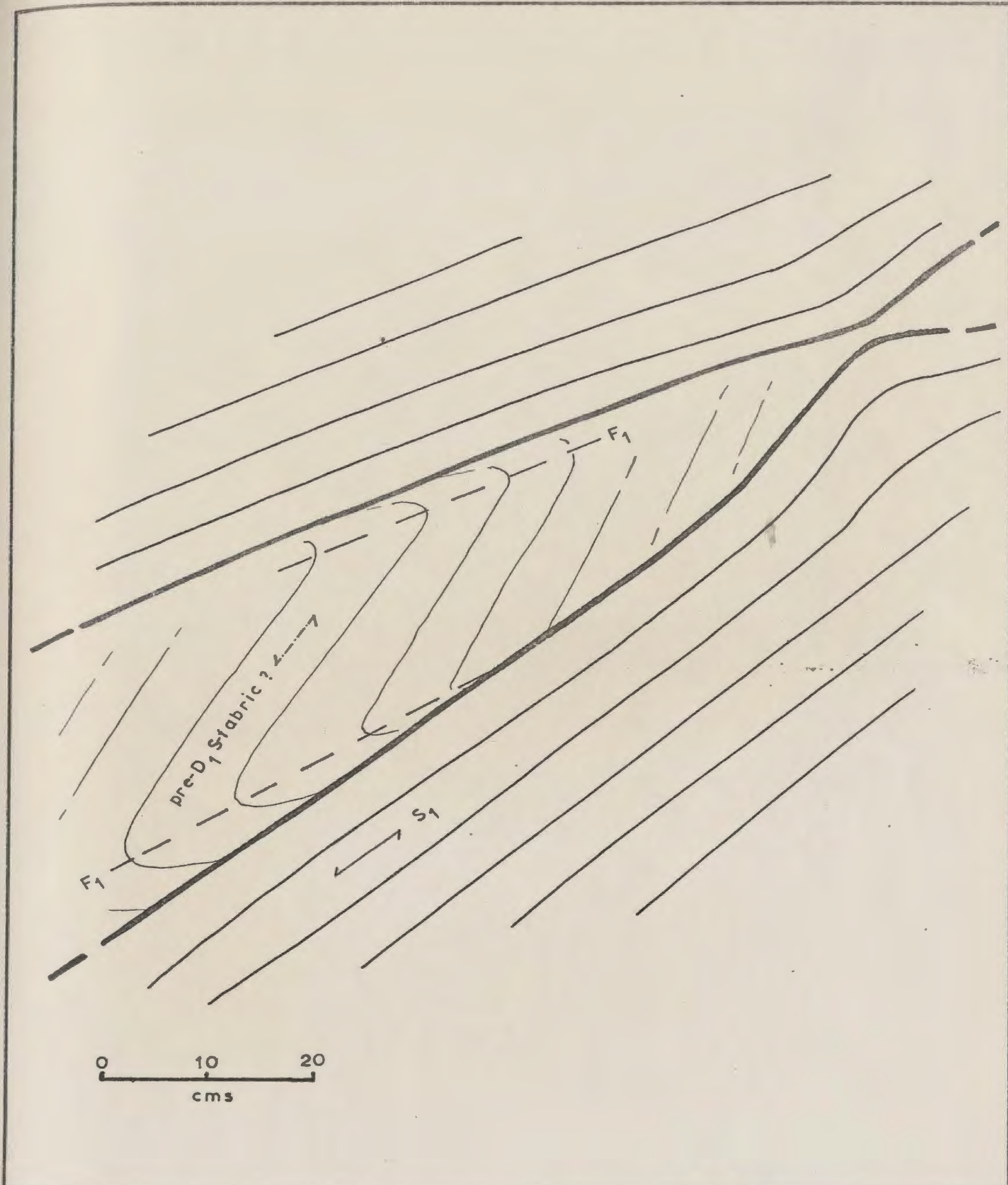


Fig. 4-42. Diagram of pre-D_1 amphibolite dyke in amphibolite (metalava?) on south coast of Big Island (see Fig. 4-30). Dyke shows D_1 boudins, and F_1 folds of pre-D_1(?) fabric.

Fig. 4-43. Pre-D_2 tension gashes in an arkose of the Big Island Formation. South coast of Big Island. Structures plunge perpendicular to the plane of the photograph.

Post-D_2 Tectonic Features

INTRODUCTION

One major post-D_2 deformation (the D_3 deformation) is recognised. Almost no fabrics have been produced by it, but several major open folds modify the D_2 map-pattern. Another deformation (D_4), which produced a few locally developed minor folds and associated fabrics, is tentatively considered as later than D_3, though proof of their relative ages is lacking. Minor faults and fractures are the latest signs of tectonism in the region.

THE THIRD DEFORMATION (D_3)

D_3 Tectonite Fabrics

S_3 fabrics have only been recognised in the metabasalt on the west coast of Nesbit Harbour (Fig. 4-8) where a very imperfect, roughly northerly oriented, biotite fabric, developed by recrystallisation of S_1 and S_2 biotite, is seen.

D_3 Structures

i. Major F_3 Folds: A southward plunging, westward facing F_3 antiform through Big Island (Figs. 4-19 and 4-44) refolds the F_2 Tilt Cove syncline. Southeast of Round Pond, a southward plunging F_3 antiform and synform refold the east limb of the Round Pond dome and associated S_2 fabrics. (Figs. 4-45 and 4-46). Another F_3 antiform occurs on the eastern

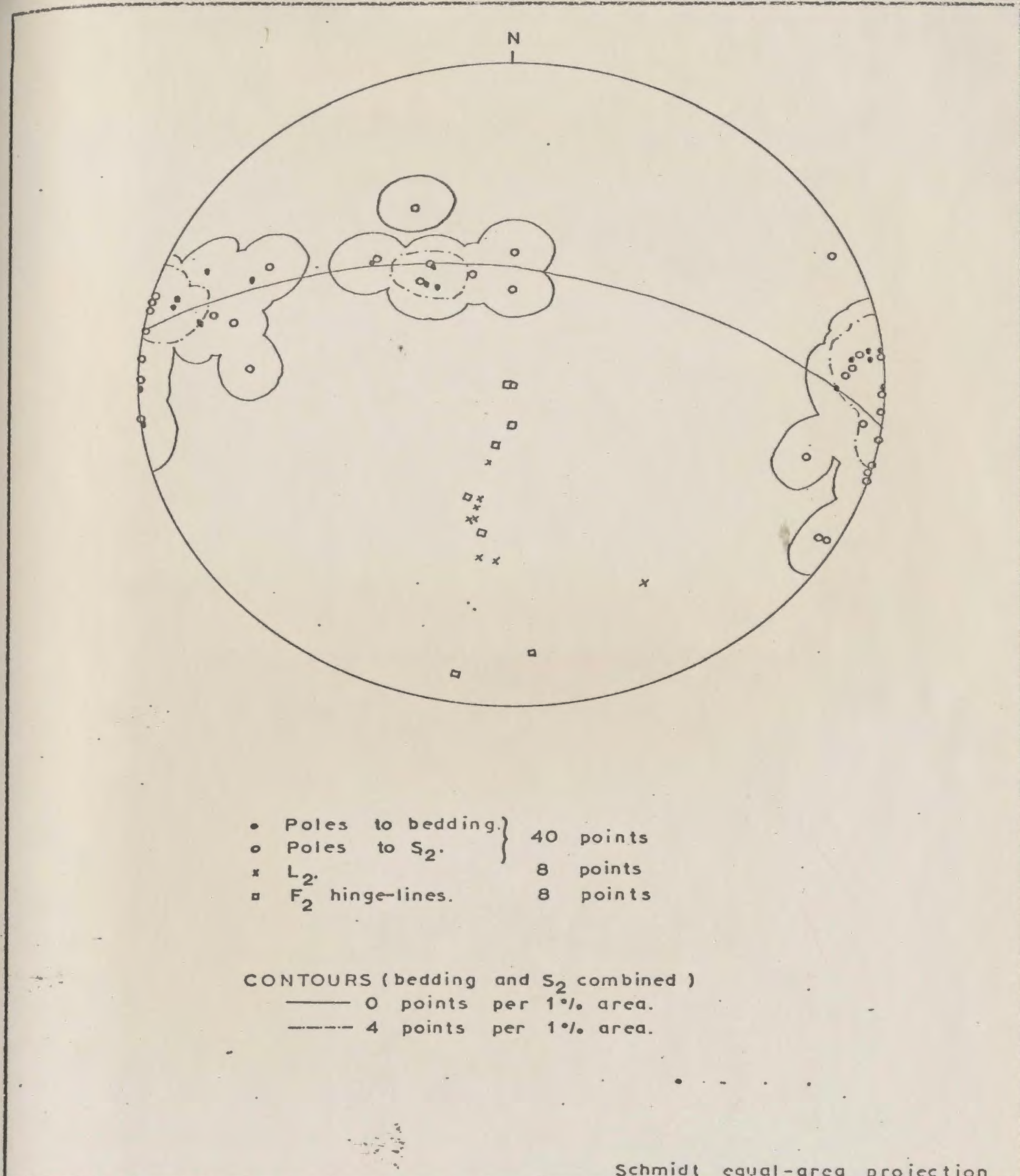


Fig. 4-44. Stereographic projection of bedding and D_2 features about the F_3 fold through the western part of Big Island.

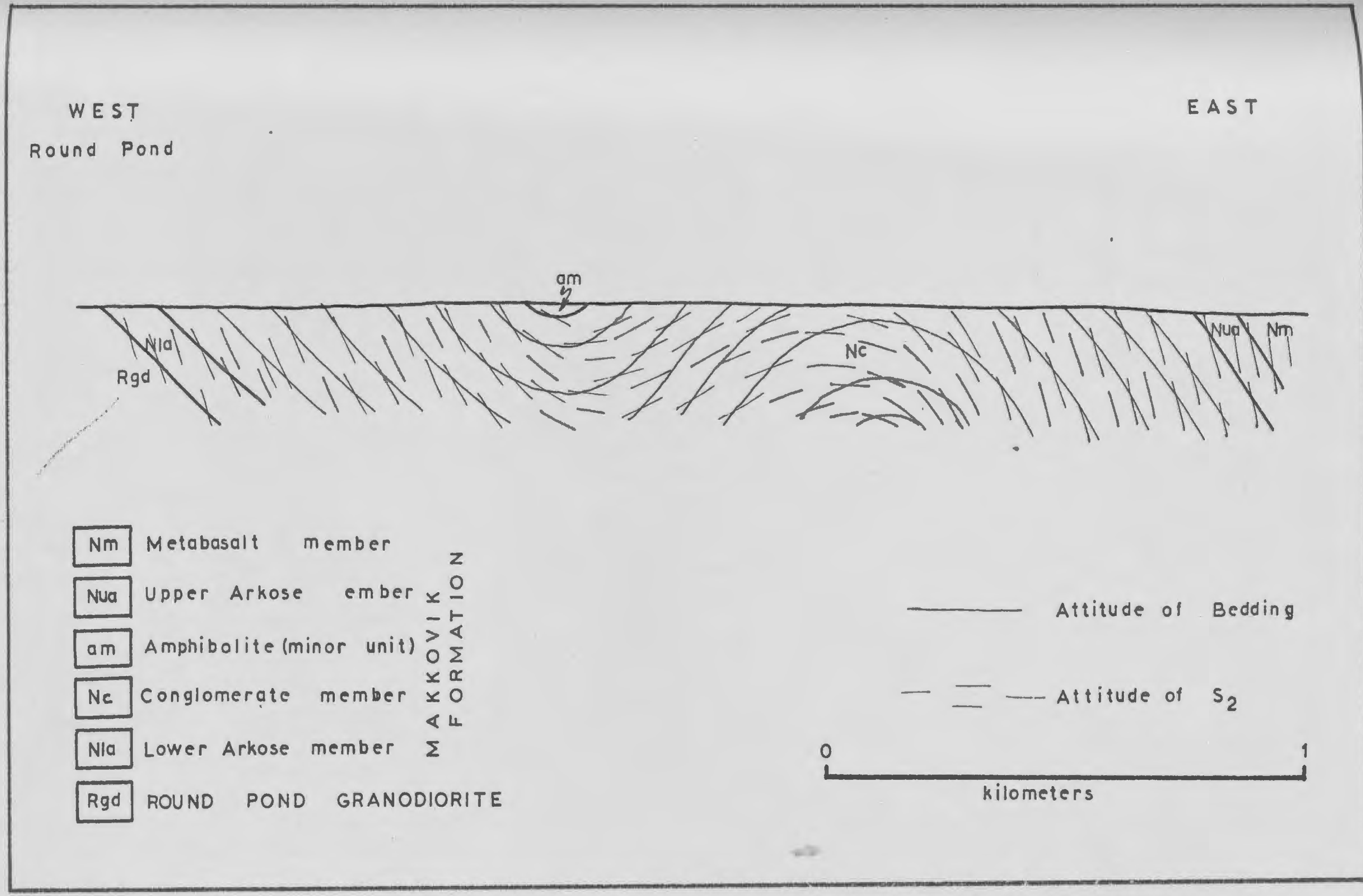


Fig. 4-45. Schematic cross-section through F_3 folds east of Round Pond, showing relationship of lithology, bedding and S_2 fabric.

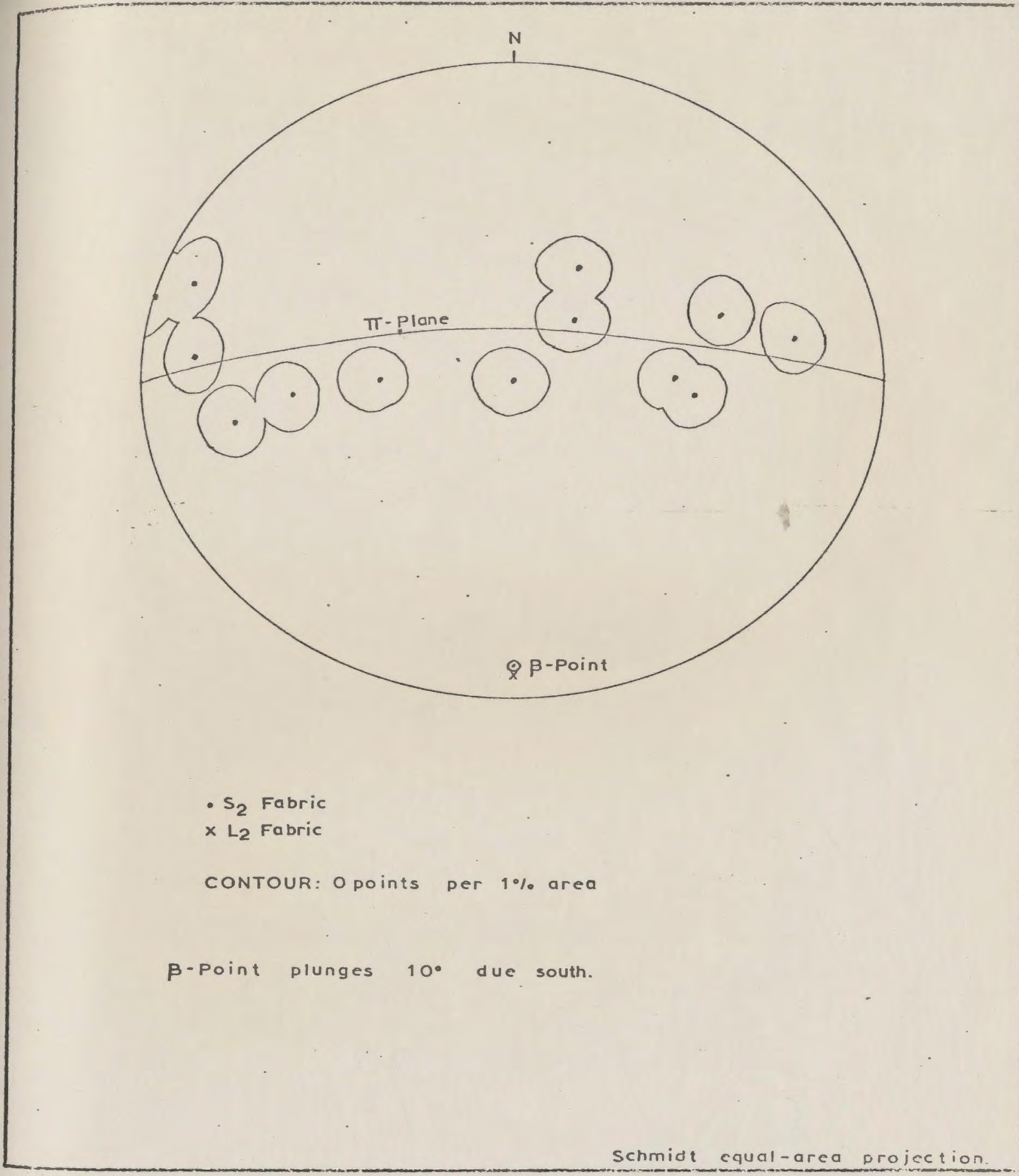


Fig. 4-46. Stereographic projection of S_2 and L_2 fabrics about F_3 folds due east of Round Pond.

flank of Monkey Hill (Fig. 4-4), and modifies the F_2 anticline through Round Pond.

The sense of closure of S_2 on these folds (Fig. 4-45 and Plate II) indicate a major F_3 antiform to the west, which is in agreement with the occurrence of an F_3 antiform down the east side of Kaipokok Bay (Clark, 1970 and 1971).

ii. <u>Minor F_3 Folds</u>: Only three occurrences of minor F_3 folds have been recognised, one forming a set of open F_3/F_4 dome and basin interference patterns north of Big Bight (Fig. 4-47), another occurring as small open asymmetrical folds of hornblende-rich bands in the metabasalt at Ford's Bight (Fig. 4-8) and the third as a single fold of D_2 flattened mafic conglomerate on Big Island (Fig. 2-42). Major F_3 closing-directions at Ford's Bight indicate an axial trace roughly north/south and F_3-closing on a horizontal surface in the same sense as the Round Pond dome northern closure, suggesting this part of the dome may have been modified by F_3 folding similar to that in the Monkey Hill area.

iii. <u>Analysis of F_3 Folds</u>: Stereonets of S_2 fabrics in the vicinity of Monkey Hill/Round Pond/Falls Lake (Figs. 4-46, 4-48 and 4-49) indicate the folds plunge due south at about 20°. However, it has not been possible to determine the orientation of the axial-plane.

iv. <u>D_3 Boudinage Fractures</u>: On the limbs of the folds at Manak Bay some of the more brittle, sandy beds, especially