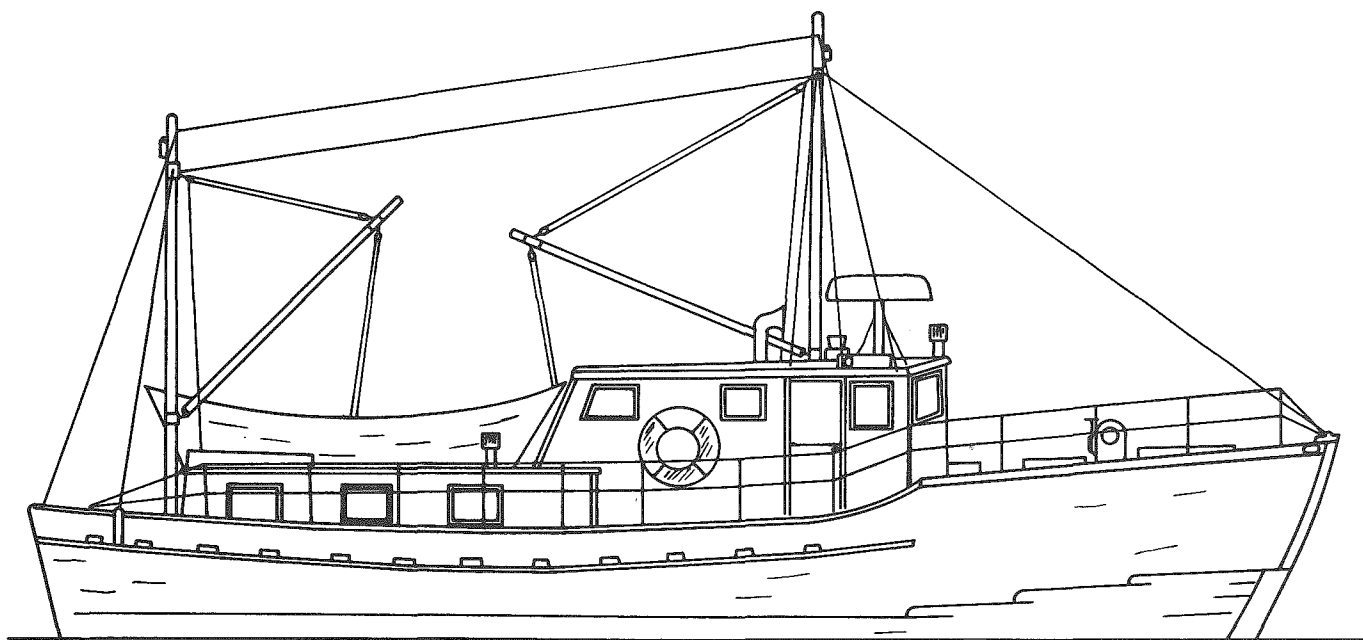


THE NAIN ANORTHOSITE PROJECT, LABRADOR: FIELD REPORT 1980

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THE NAIN ANORTHOSITE PROJECT, LABRADOR:

FIELD REPORT 1980

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"Mantle and Crustal Evolution as

Reflected in the Anorthositic Nain

Complex, Labrador."

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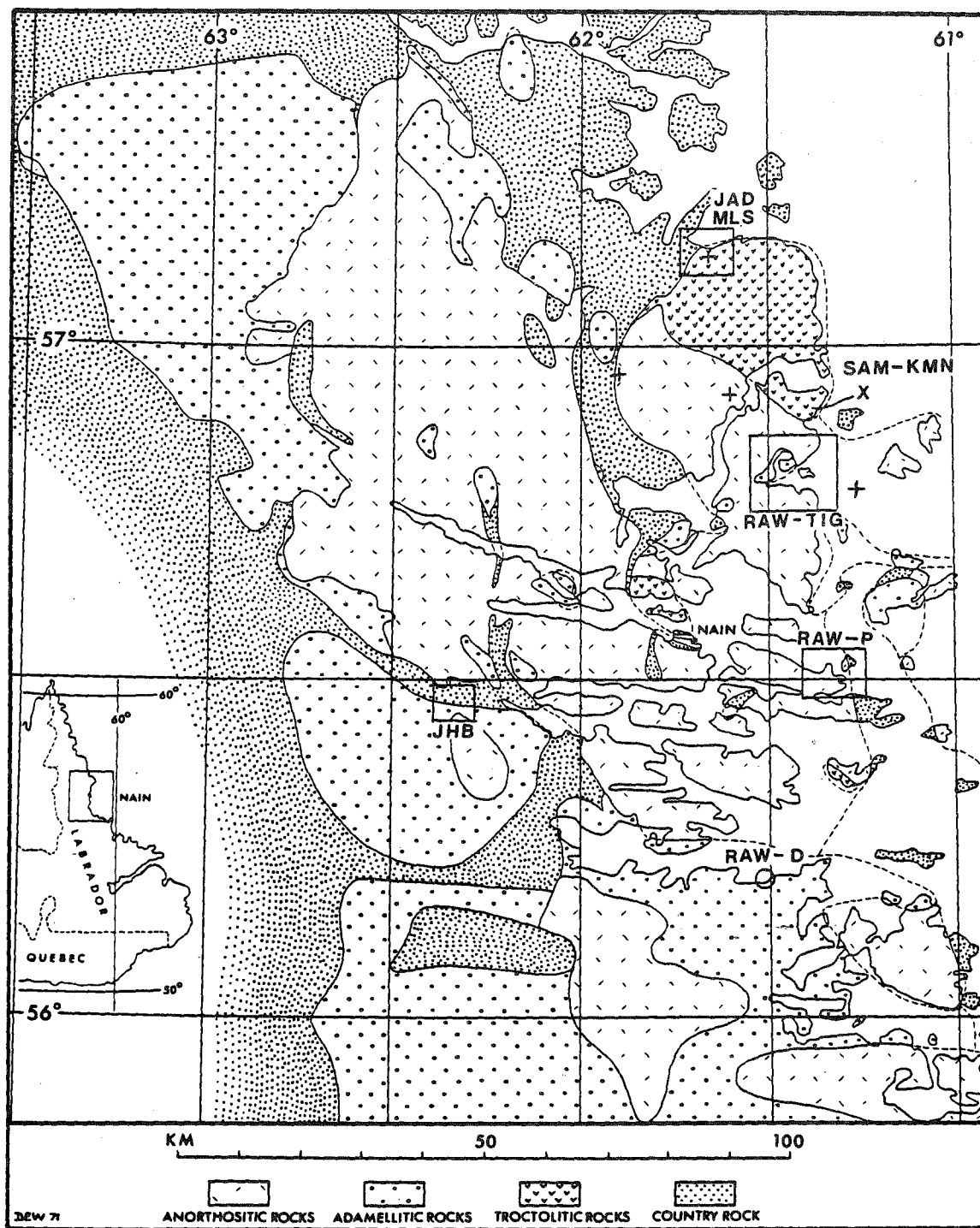


Fig. 1. Regional geology of the Nain area, after Wheeler (1968), showing locations of 1980 field areas. Key, north to south: JAD-MLS Docka and Schuh (Snyder Bay); SAM-KMN Morse and Nolan (Kiglapait); + Berg and Pencak (Troctolites); RAW-TIG Wiebe (Tigalak); RAW-P Wiebe (Paul I.- West Red I.); JHB Berg (Osumilite Lake); RAW-D Wiebe (Dikes).

THE NAIN ANORTHOSITE PROJECT, LABRADOR: FIELD REPORT 1980

INTRODUCTION AND REVIEW

As long emphasized in these pages, the anorthosite problem focuses a wide range of broader questions regarding the evolution of continental crust during the first three-quarters of earth history. As a result, investigators on this project have made important contributions in such diverse areas as the discovery of ancient Archean rocks, regional geochronology, the identification of Aphebian supracrustal rock sequences, the application of geobarometry, geothermometry, and volatile barometry to regional contact aureoles, the fractionation of basic magmas, and the documentation of anorthositic magmas and their ferro-dioritic residua. Moreover, many of the classical questions about anorthosite genesis, such as the role of volatiles, tectonic regime, pressure, and the spectrum of magma types, have now either been answered or satisfactorily resolved in a preliminary way, as summarized in the Introduction to our last Field Report (FR 1976). Large problems remain, among them the ultimate cause in the earth of an anorthosite episode which was at least broadly limited in space and time. At the same time, new opportunities arise from our increased understanding that the rocks of the anorthosite kindred, particularly in the Nain area of Labrador, contain many pristine igneous features of texture and chemistry that make them important potential recorders of the geochemical evolution of the mantle, and of possible crust-mantle interaction.

It is to these larger questions of mantle and crustal evolution that our new field research effort is ultimately addressed. Many old avenues of research are, of course, still being explored, as indicated by the published studies cited in the Bibliography. Such investigations still form an important thrust of our field research, now resumed after a three-year hiatus. But the new emphasis on a broader goal springs from many sources, rooted both in our increased understanding of the Nain complex and in recent progress in geochemistry, particularly isotopic geochemistry. It may be useful, therefore, to quote at length from the NSF proposal which led to the resumption of field studies, in order to remind ourselves and our readers of the increased scope of our questioning and the new evidence in favor of viewing the Nain complex as an important window on the earth's chemical evolution.

Extract from NSF proposal EAR-8011385, 1979.

INTRODUCTION

The earth's geochemical history, as presently understood, appears to contain three "crises" in the period 1-2 Gyr ago. (Whether these are crises for the earth or merely for geochemists remains to be seen.) Each such crisis represents a time when the geochemical evolution of the mantle appears to have changed direction or become closed. The crises are recognized by changes in the isotopic evolution of Sr, Pb, and Nd. A strontium crisis at about 1.5 Gyr emerges from the idea of mantle isochrons (Brooks, James, and Hart, 1976). A lead crisis near 1.5 Gyr is inferred by Tatsumoto (1978).¹ A neodymium crisis near 2 Gyr is suggested by Zindler and Hart (1979).¹

Nearly all the work suggesting these crises has been done on young volcanic rocks. In order to document or otherwise explicate the implied changes in the evolution of the earth's mantle, it will be essential to identify and characterize magmas and their mantle source regions that existed in the period 1-2 Gyr ago. Few fresh volcanic rocks are known from that stage of earth history. We propose that the most voluminous and ultimately the most valuable samples of mantle-derived magma of that age are the massifs anorthosites belonging to the anorthosite episode, about 1.7-0.6 Gyr ago. Of these, the most valuable geochemically are those in Labrador that have escaped later deformation and metamorphism (see Emslie *et al.*, 1972; Emslie, 1978; Berg, 1977; Wiebe, 1978). And of the Labrador anorthosite complexes, that in the Nain area offers the widest variety of bulk compositions and, very probably, the broadest spectrum of isotopic and minor element signatures to be found anywhere.

The Nain complex has yielded to date more well-documented candidates for magma compositions than any other such complex (Wiebe, 1978, 1979). The various parent magma compositions cannot be related by low-pressure fractionation, but must instead represent a variety of source regions or source processes, or possibly fractionation at depth. The enormous potential of the Nain complex as a geochemical window into the past is due not only to the great diversity and excellent preservation of the igneous rocks, but also to the clarity with which field relations are displayed in superb shoreline exposures.

The Nain complex is also a garden of exciting and informative metamorphic mineral assemblages. Several important minerals and mineral assemblages have first been recognized in the Nain contact aureoles (Berg and Wheeler, 1976; Wiebe and Berg, 1978; Berg and Wiebe, 1978) only to be found subsequently in many other places. With the possible exception of the Adirondack region being studied by E.J. Essene and his colleagues, there is probably no region in the world with as coherent a data base for quantitative regional geobarometry and geothermometry as the Nain region. With further examination of the metamorphic rocks, we will not only learn more about the assemblages, the local environment, and the local processes, but on a regional scale we will be able to specify more and more about the tectonic setting and physical evolution of this large magmatic complex.

THE NAIN COMPLEX

Recent research on the Nain complex is summarized in a series of six Field Reports of the NSF-supported Nain Anorthosite Project (Morse, editor, 1971-77). This project was conceived as a study in crustal evolution, now extended to mantle evolution in this proposal. The complex consists of about 9,000 km² of anorthosite and related basic rocks (Fig. 1) with a subequal area of granitic rocks described as an adamellite series by Wheeler (1955, 1968). The basic portion of the complex consists of a large number (certainly more than 25) of discrete plutons of anorthositic to gabbroic composition. The plagioclase composition of anorthositic rocks ranges at least from An₃₄ to An₅₀ (Morse, Field Report 1975), although it is much more restricted than this within individual plutons. Chilled margins occur against dehydrated Archean country rocks (Berg, 1974). Anorthosite dikes and homogeneous massive plutons show the prior existence of anorthositic magmas (Wiebe, 1979).

¹ The crises may simply represent the mean age of the mantle reservoir, according to Jacobsen and Wasserburg (1979 JGR 84, p.7411-7428; 7429-7445). Their second paper, on the Bay of Islands complex, illustrates the kind of inferences we hope eventually may be reached concerning the subcontinental mantle at Nain.

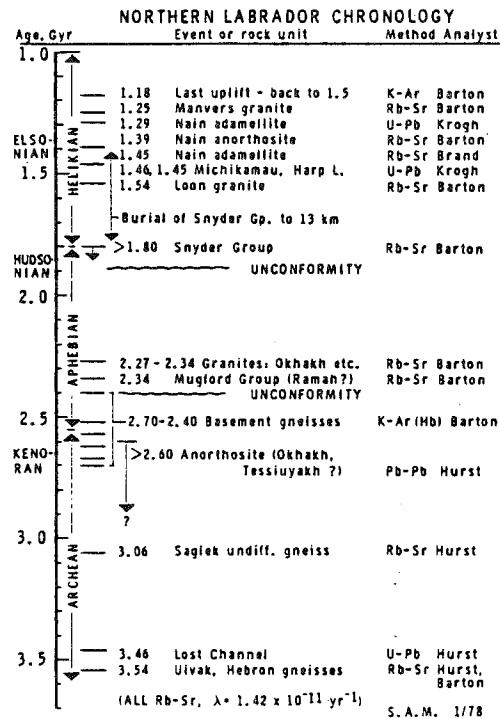


Fig. 2. Regional geologic history: the Kiglapait intrusion is considered slightly younger than 1.4 Gyr (10^9 yr). Data from Barton (1974, 1975a, 1975b, 1977), Barton and Barton (1975), Brand (1976), Hurst (1974), Hurst *et al.* (1975), and Krogh and Davis (1973). Note: all Rb-Sr ages adjusted to $\lambda^{87}\text{Rb} = 1.42 \times 10^{-11} \text{ yr}^{-1}$. All data originated with the Nain Anorthosite Project.

Tectonic setting

It is important to our argument to show that the Nain complex was emplaced anorogenically by the intrusion of many magmas, probably over a sustained period of time. Indirect evidence for the tectonic regime is supplied by Fig. 2, which shows that orogenic events were very rare in this part of the Archean craton. The major recrystallization episodes are restricted to the oldest rocks at 3.5 Gyr at Saeglekh, Hebron, and Lost Channel, an event at 3.1 Gyr at Saeglekh and Hebron, and the widespread Kenoran event near 2.6 Gyr. The next recorded events are the extrusion of the voluminous Mugford volcanics at 2.4 Gyr, deposition of the supra-crustal Ramah Group at a similar or later time, and deposition of the thin, shelf-type Snyder Group near the close of Aphebian time (1.8 Gyr). Then follows the Elsonian intrusive episode producing the Nain complex. There is no evidence for orogeny or regional recrystallization after 2.4 Gyr. Undeformed and unfoliated massive granites as old as 2.3 Gyr confirm the absence of any younger recrystallization.

Rifting tectonics were proposed as a setting for the anorthosite episode by Bridgwater and Windley (1973) and supported with Labrador evidence by Berg (1977, 1979) and Emslie (1978). Such a setting should maximize the opportunity for basic magmas to arise from the mantle, as they do now in rift and ridge systems, and minimize the opportunity for generating magmas *sui generis* in the crust by the thermal blanketing effects common to geosynclinal tectonics. We interpret the granitic magmas as crustal melts, but ones directly caused by introduction of magmatic heat sources from the mantle.

Only in Northern Labrador and in the Nain area in particular do we know the early geologic history with sufficient precision to make with confidence such statements about the tectonic setting of the anorthosite episode. Whether or not such a setting is implied for all other anorthosite terranes is immaterial, because the existence of such a setting at Nain suffices to show that orogenic tectonics are not required for the generation or emplacement of anorthosite. We assume, lacking persuasive evidence to the contrary, that a tensional setting applies to the emplacement of anorthosite generally, but this assumption is external to our quest in this proposal.

Our understanding of the tectonic history of coastal Labrador is nonetheless very incomplete, and its improvement forms, as before, an integral part of our research program.

Depth of emplacement

The work of Speer (1976) and Berg (1977) demonstrated for the first time that katazonal conditions are not a prerequisite to anorthosite genesis. Speer showed that the Snyder Group was contact-metamorphosed from greenschist facies in the hinterland to anatectic conditions at the Kiglapait contact, in the andalusite-sillimanite series. Kyanite has not yet been found in the contact aureole of the Nain complex. Berg showed that the anorthosite contacts record pressures of 4-6 kbar, and suggested (1979) that these could be reduced to 2-4 kbar. Emplacement depths were therefore quite certainly less than 20 km and perhaps as shallow as 7 km, with ambient temperatures of 200°-400°C in the country rocks. Such conditions are conducive to the production of chilled margins, which are being recognized with increasing frequency in the field.

Evidence for anhydrous magmas

The discovery by Berg and Wheeler (1976) of osumilite in a contact aureole opened the way to a demonstration that the surroundings of the Nain complex are characterized by dehydration, contrary to the case required by anorthosite crystallized from hydrous magma by expulsion of H_2O to the surroundings. Hensen (1977) confirmed that osumilite is unstable relative to cordierite in the presence of H_2O , and Goldman and Rossman (1978) showed that the Labrador osumilite contains less than 0.01% H_2O .

Other indications of dry surroundings occur in the common presence of basic pyroxene granulite at the margins of the anorthosite (de Waard, 1972; Wiebe, 1979; Ranson, 1979), and in the dehydration of muscovite + quartz to sillimanite + orthoclase without the production of melt (Speer, 1976).

Internal evidence for dry magmas resides in the high temperatures required by the occurrence of calcic mesoperthites (e.g. Morse, 1969), hypersolvus and narrow-solvus quadrilateral pyroxenes (Davies, 1974; Ranson, 1979), and the maximum H_2O content of 38 ppm calculated by Huntington (1979) for the Kiglapait intrusion.

The Nain complex is therefore characterized by dry, high-temperature magmas, consistent with their production in the mantle or, conceivably, the lower crust. A mantle source is suggested by the high temperatures implied (1150-1250°C at pressure), unless one invokes an abnormally steep geotherm or thick crust. These latter postulates are unlikely if the earth's geotherm at spreading centers is thermally buffered by melting at a cusp (Presnall *et al.*, 1979). The latter authors also argue persuasively that H_2O and CO_2 are of little importance in the generation of abyssal tholeiites, and it appears probable that at least some parent magmas of anorthosite have much in common with these modern products of mantle anatexis.

Direct sampling of plutonic magmas

The chilled margins of the Hettasch intrusion (Berg, 1974) and of the Michikamau intrusion (Emslie, 1978) furnish direct evidence for a high-Al, high-Fe, low LIL magma associated with anorthositic intrusions. The Hettasch composition is shown here in Table 1. A nearly identical composition was calculated for the Kiglapait intrusion by Morse (1974 and in preparation). Wiebe (1979) showed the existence of anorthosite dikes with uniformly high-Al compositions. The average composition of 17 massive rocks from the Tunungayualok Island leuconorite (Table 1, column 3) is equivalent to that of the dikes, and very similar to the Adirondack Whiteface facies composition of Buddington (1939). The other feldspathic compositions listed in Table 1 are considered by Wiebe to form a feldspathic magma series parental to anorthosites. These magmas cannot be related by low-pressure fractionation. We regard Table 1 as indicative of the spectrum of magma types that will be found responsible for anorthositic rocks, although much remains to be done to define this spectrum more closely. REE data for the rocks of Table 1 are plotted in Fig. 3.

TABLE 1. POSSIBLE PARENTAL MAGMAS OF THE NAIN COMPLEX

	1.	2.	3.	4.
	Hettasch Intrusion	Kheovik	Tun. Is. Ln	Uivakh type Ln
SiO ₂	47.68 %	53.30%	54.82 %	55.84 %
TiO ₂	1.23	0.28	0.94	0.78
Al ₂ O ₃	18.85	24.84	22.28	20.86
Fe ₂ O ₃	1.11			
	FeO _T	3.43	5.39	6.38
FeO	10.60			
MnO	0.14	0.05	0.08	0.08
MgO	7.63	2.88	2.05	1.37
CaO	8.55	10.46	9.13	8.27
Na ₂ O	3.19	4.33	4.31	4.86
K ₂ O	0.40	0.40	0.84	1.25
P ₂ O ₅	0.13	0.07	0.13	0.30
Rb	4.2 ppm	<10 ppm	11 ppm	24 ppm
Sr	409	640	579	590
Zr	61	123	155	183
Ni	~170	31	10	<d1
V		22	68	44
Ba	~140	236	490	696
FeO _T /MgO	1.45	1.19	2.63	4.66
Na ₂ O/CaO	.37	0.41	0.47	0.59

1. Ave. of 5 chilled samples (unpublished data)
2. Ave. of 16 massive rocks (similar to chilled margin)
3. Ave. of 17 massive rocks (equal to chilled dike; Wiebe, 1979 a)
4. Ave. of 2 chilled samples

Note: Ln = leuconorite; d1 = detection limit.

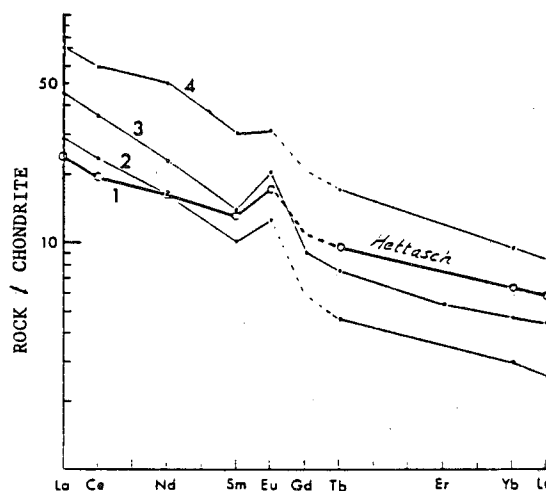


Fig. 3. Chondrite-normalized REE patterns for the rocks shown in Table 1. Hettasch curve from unpublished data of Haskin & Berg. Other data from Simmons and Hanson (1978) and new INAA analyses by R. A. Wiebe, made at NASA-JSC, Houston.

Other approaches to magma compositions

Berg is currently documenting evidence for a pulse of picrite-basalt magma injected into the partly crystallized Hettasch intrusion to form olivine and plagioclase crystals by constitutional supercooling. This magma composition extends the spectrum of Table 1 toward the basic end.

Morse (1979 d) has used mineral compositions and theory to evaluate oxygen and silica activities in basic plutonic magmas. These activities can be applied to set useful limits on the compositions of parental magmas of anorthositic rocks, particularly those containing olivine. Morse (1974 and in preparation) has also calculated the compositions of successive liquids in the Kiglapait intrusion, and from these and the mineral compositions he obtains useful mineral-melt distribution coefficients for K, Rb, Sr, Ba, REE, and other elements. The derived values tend to be closer to 1.0 than those commonly used in the literature, so the magma compositions calculated from them tend to be more like the rocks than would be the case using classical values. This approach to geochemical modelling of magma compositions furnishes a useful comparison (or contrast) with more direct evidence from chilled margins or dikes. A new formula for evaluating the X_{An} in parental magmas (Morse, 1979a) also has useful applications to the anorthosite prob-

NAIN STRONTIUM INITIALS

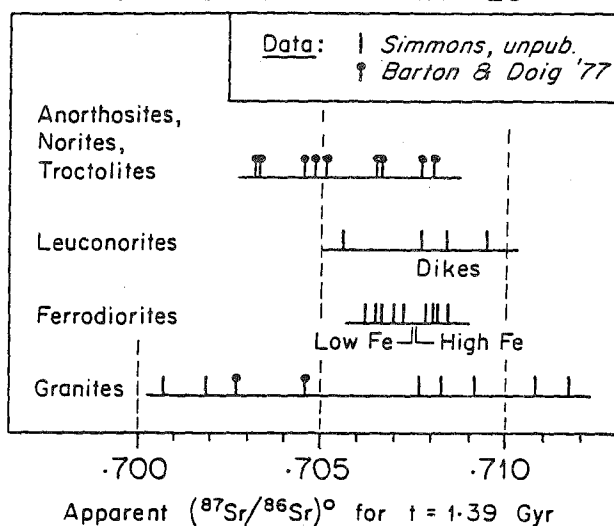


Fig. 4. Apparent initial strontium ratios in the Nain complex. The leuconorites, ferrodiorites, and the lowest granite are from rocks related to the Tunungayualok Island leuconorite intrusion being studied by Wiebe and Simmons. Unpublished data courtesy E. C. Simmons.

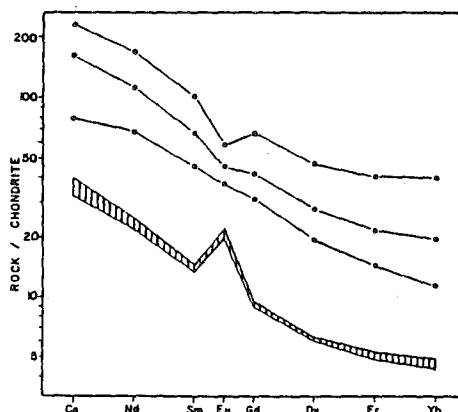


Fig. 5. Chondrite-normalized REE's for some rocks illustrating the liquid line of descent for the Tunungayualok Island leuconorite intrusion (determined by E. C. Simmons). Vertically ruled pattern shows the range of three fine-grained leuconorite dikes which are assumed to approximate the composition of the parental magma. The other REE's are from chilled ferrodiorite bodies closely associated with the leuconorite. These are presumed to represent late-stage residual magma.

lem, and provides a similar basis for comparison with direct evidence. We expect these various "synthetic" approaches to magma composition to converge with the direct estimates, or else lead to significant new understanding of the causes of disagreement.

Ranson (1979) has calculated the theoretical modal composition of a cotectic leuconorite magma, and found that it compares very closely with the average composition of leuconorite in the Puttualaak Lake area. He shows that plagioclase-rich anorthosites cannot have been cotectic with pyroxene at any crustal pressure, and must therefore represent either metastable or mechanical enrichment in plagioclase. Morse (1979b) suggests a mechanism for producing plagioclase-rich magmas metastably in the mantle.

Trace element and isotope chemistry

Minor element data on the Nain rocks are scanty. Gill and Murthy (1970) reported on K, Rb, Ba, Sr in selected rocks of the Nain complex, studied without detailed reference to field relations. Simmons and Hanson (1978) studied the REE distribution in a few widely scattered samples. Wiebe (1978, 1979) characterized several suites of anorthositic rocks, in field context,

with an informative group of minor elements, and the results of his current work are shown in Figs. 3-5. Heath and Fairbairn (1968) showed present $^{87}\text{Sr}/^{86}\text{Sr}$ ratios to range from 0.705 to 0.706 in a small suite of Nain samples. The range of apparent initial Sr isotope ratios is extended from 0.703 to 0.709 by the work of Simmons (in progress) and Barton and Doig (1977), as shown in Fig. 4. Perversely, the lowest initial ratio found in the Nain complex is the value of 0.7007 found for a granite (Fig. 4). There is an appalling lack of systematic minor-element and isotopic data on the now-abundant areas where field relations are well understood and well mapped.

Contamination

Abundant evidence has been found for the coexistence and comingling of granitic and ferrodioritic magmas in the eastern part of the Nain Complex (Wiebe, 1978 and 1979). The ferrodioritic magmas are interpreted as late-stage residua of the anorthosite parent magma, and the granitic magmas as crustal melts generated by heat from the anorthosites. Field, petrographic, and chemical evidence can be used reliably to identify such comingling and to identify the end members of the mixture, but further chemical data are needed to confirm the sources of the coexisting magmas. Whether or not the anorthositic rocks have been contaminated by crustal melts is less certain, but at least no profound contamination is indicated. The eastern anorthosites of the complex tend to carry minor quartz and K feldspar, whereas the western anorthosites appear to be devoid of these minerals or nearly so. Morse thinks the quartz and K feldspar signify some early contamination at the eastern margin; this proposition requires continued evaluation.

Summary

We are impressed with the clearly large range in parental magma compositions implied for the basic rocks of the Nain complex, and with the multitude of emplacement events representing separate draughts of magma. It is precisely this breadth of composition and diversity of intrusive events that makes the Nain complex such a valuable window on the Proterozoic mantle.

- - - - -

Our 1980 report incorporates the results of recent laboratory studies as well as field work, and three analytical discoveries are particularly noteworthy: the steady strontium isotope evolution found in the Kiglapait intrusion by Simmons and Lambert, the reversed rims on plagioclase of the Kiglapait Lower Zone found by Morse and Nolan, and the high-alumina orthopyroxene xenocrysts found in alkali basalt dikes by Wiebe. The strontium isotope data suggest an initial strontium isotope ratio for the source of the Kiglapait intrusion, at 1.4 Gyr ago, of 0.7040. Rocks of the Lower Zone show constant initial ratios near this value, but rocks of the Upper Zone show steadily increasing initial ratios to a value of 0.7066 at the top of the Layered Series. Contamination of the Upper Zone rocks can be ruled out because of their very low Rb/Sr ratio, and Simmons and Lambert ascribe the increase to magma mixing. This writer doubts that magma mixing could account for the elevated radiogenic strontium observed, and prefers a more heretical hypothesis involving fractionation of radiogenic strontium in the melt (1981 abstracts, AGU Spring Meeting, in press). In any event, the constraints on contamination and other conventional explanations of stratigraphically evolved strontium isotope initial ratios are much more severe in the Kiglapait intrusion than elsewhere found to date, and will require serious consideration of

thoughtful alternative hypotheses.

Reversed rims on plagioclase are abundant in the Kiglapait intrusion, and they predominate over normal rims. The reversals are locally large, up to nearly 30 mole percent An, with K/Na falling by a factor of six. These surprising results have important consequences for our ideas on the roles of kinetics, supersaturation, steady-state growth, and the adcumulus growth mechanism in the nucleation and solidification of feldspar-rich rocks.

High-alumina orthopyroxenes were reported earlier as megacrysts in anorthosite (Morse, 1975; Emslie, 1975), but Wiebe has now found them as xenocrysts in late, possibly Gardar-age, alkali basaltic dikes cutting anorthosite. The steep initial increase (with falling Mg ratio) of Al_2O_3 to 6 wt % may signify fractionation of a parent liquid from a meso-alumina to high-alumina variety, caused by removal of aluminous orthopyroxene. Study of these objects should therefore illuminate the origin of aluminous magmas possibly parental to anorthosite. A two-pyroxene assemblage reported by Wiebe from the dikes gives a calculated temperature of 1320°C , and a magnetite-ilmenite pair gives a directly calculated temperature of 1100°C ; these are uncommonly high temperatures for hypabyssal rocks, and they testify to the rapid quench imposed on the mineral assemblages during and after emplacement.

I turn now to the reports of field work in sequence as they appear in the following articles. Berg has at last been able to map an apparent osumilite isograd in the contact zone where plutonic osumilite was first discovered by Berg and Wheeler (1976). New sampling in this nearly inaccessible region will illuminate the reactions responsible for the growth of osumilite as a contact metamorphic mineral in aluminous paragneisses. This mineral is particularly important for its intolerance to H_2O , hence its ability to show that the contacts of anorthosite are not characterized by fugitive H_2O .

The Aphebian Snyder Group and its overlying supracrustal sequences preserve an important record of tectonic and sedimentological events prior to the emplacement of anorthosite. In a detailed re-examination of the Snyder Bay area, Docka reviews the problems associated with these metasedimentary rocks and their interstratified pyroxene granulites, and describes probable lateral facies changes that could eventually form the basis for a limited basin analysis. Schuh continues the review with a detailed description of stratigraphy and structure, including a new report of mildly overturned folds, followed by a separate study of the detailed stratigraphy of the banded and silicate

iron formation contained in the sequence overlying the Snyder Group.

The contact zone on the eastern end of Paul Island was one of the first critical areas studied by de Waard (FR 1971), who here named the Ford Harbour Formation and described the anorthosite contact. A revisit by Wiebe in 1980 resulted in new observations and reinterpretations in the light of current understanding of the role of dioritic rocks near anorthosite margins. A remarkable feature of the contact zone is a unit of basic granulite containing isoclinally folded anorthosite dikes which may represent the earliest stages of anorthosite emplacement, subsequently deformed by continued emplacement and expansion of the anorthosite body. The entire contact zone was intensively sampled by Wiebe and Morse as a target of high interest for geochemical and isotopic studies.

The internal relations of the Nain complex received further study by Wiebe in the Tigalak layered intrusion and by Berg and Pencak in an examination of supersaturation textures in melatroctolites occurring at various localities. The sequence of events in the Tigalak intrusion, as presently understood, includes emplacement and solidification of troctolite, followed by emplacement of a noritic body which fractionated to ferrodiorite near the top, and injection, perhaps laterally, of ferrodioritic liquid into a western lobe where it mingled with granitic liquid to form a large area of hybrid rocks. This sequence illuminates once again the ferrodioritic nature of residual liquids from anorthosite-forming leuconoritic magma, and the common phenomenon of magma mixing involving granite of presumed anatectic origin.

Berg and Pencak followed up on Berg's (1980) study of snowflake troctolite by visiting all known localities where macrospherulites (snowflakes) of plagioclase occur along with other textures indicative of supersaturation in the plutonic environment. Comb-layered olivine ranks high among the new discoveries. The remarkable textures in these rocks carry important information on the kinetics of crystal growth in rocks related to anorthosite, and Berg's recognition that the host rocks invariably have a high color index leads to the conclusion that picritic or melatroctolitic liquids formed an important component of the late-stage magmatism associated with the Nain complex.

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The broad temporal picture that begins to emerge from an examination of old and new data is one of early, sodic noritic anorthosite (Tikkoatokhakh region; Morse, FR 1976), followed by more calcic noritic anorthosite (both fractionating to ferrodiorite), followed by troctolite, followed by melatroc-tolite or picrite, all overlapping in time to some degree. High temperature granitic melts were probably generated in deeper parts of the crustal edifice and emplaced through the same plumbing system at such times as to mingle with ferrodioritic residua, and also at later times (Fig. 2). After 10^5 yr or less of cooling and uplift, basaltic dikes were emplaced during a renewal or continuation of tensional stress, and their magmas sampled lower-crustal and possibly upper mantle rocks left over from the anorthosite-making process. The overall petrologic picture is one of mantle magmatism in an aborted rift zone, in which a "crustal" component of perhaps some antiquity may have contributed isotopic signatures but could not, judging from the relatively low LIL content of the suite in general, have played a dominant role in determining the feldspar richness of the magmas. The origin of this anorthositic characteristic must be sought in process rather than source chemistry, and the mafic counterpart must be considered to have remained in the subjacent mantle and lower crust.

-- S. A. Morse

CONTACT RELATIONS

GEOLOGY OF OSUMILITE-BEARING PARAGNEISSES
IN THE REGION WEST OF ANAKTALAK BAYJ.H. Berg
Northern Illinois University¹Introduction

In the summer of 1973, E.P. Wheeler II collected a highly weathered sulfide- and graphite-rich gneiss in rugged terrain directly west of Anaktalak Bay. Laboratory work by the author resulted in the identification of the mineral osumilite in the sample. This was the first reported occurrence of osumilite in a metamorphic rock (Berg and Wheeler, 1976). Subsequently it was found in similar environments in Norway (Maijer, et al., 1977), the USSR (Grew, 1980), and Antarctica (Ellis et al., 1980).

In the summer of 1975, I attempted a one-day trip into the locality to collect more material. However, the traverse was so difficult and time-consuming that there was only time to collect one large sample, and field relations were left unexplored. The large sample was subdivided and distributed to several colleagues and institutions. Research on this material includes studies by Goldman and Rossman (1978) and Jarosewich et al. (1979).

During the summer of 1980, with much physical effort and many logistical problems, a base camp was established at "Osumilite Lake" in order to investigate the extent and field relations of the osumilite gneisses.

Geology

The generalized field relations in the Osumilite Lake area are shown in Fig. 6. The oldest rocks are aluminous paragneisses that pinch out in the southern part of the map area, but open up to the north. The gneisses are well-layered, having aluminous layers alternating with feldspathic layers on a scale of 2-15 cm. The aluminous layers contain cordierite + spinel + orthopyroxene + feldspar + quartz + osumilite + biotite + graphite + sulfides as the most obvious minerals, whereas the feldspathic layers contain for the most part just feldspar + quartz + hypersthene. The inten-

¹Authors' full addresses are given at the back of this volume.

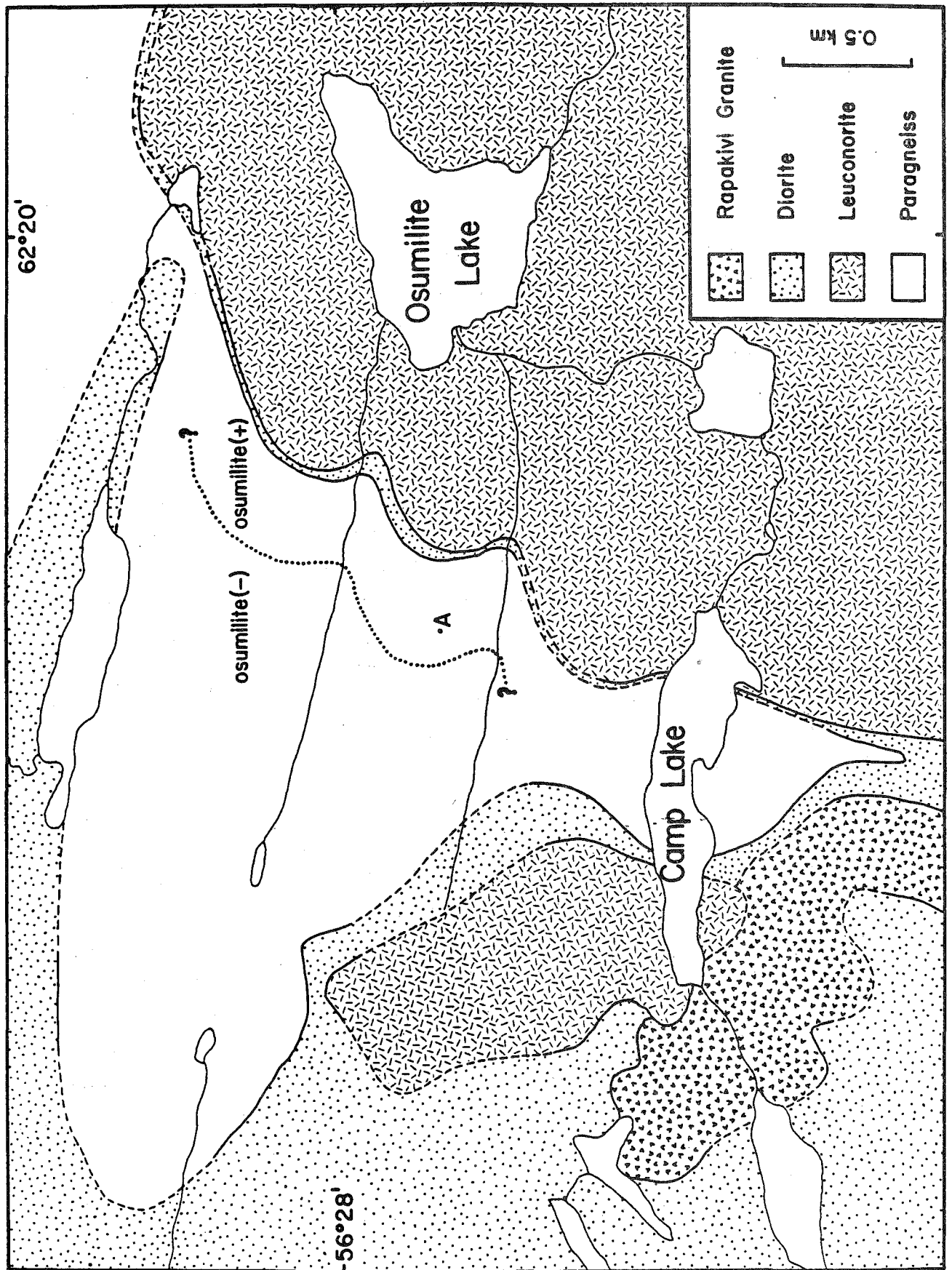


Fig. 6. Geologic map of Osumilite Lake area. Compiled from Wheeler (FR 1973) and 1980 study by Berg.

sity of deformation in this unit is variable. In some areas the layering is quite regular and continuous, and well defined folds are visible. In other areas the layering is quite incoherent and the rock has a chaotic appearance.

Although the aluminous paragneiss is heterogeneous on an outcrop scale (e.g., layering), on a broader scale the unit appears quite homogeneous. The two lithologies described above dominate the entire unit. The principal discernible variations involve concentrations of graphite + sulfide and osumilite (see below). One slight exception to this asserted homogeneity occurs at point A in Fig. 6, where garnet-bearing feldspathic veins cut across the layering of the aluminous paragneiss.

The zones enriched in graphite and sulfide minerals are easily recognized in the field by their rusty weathering and punky rock. The most prominent zone is along the eastern contact directly west of Osumilite Lake. In places, lenses of graphite are up to 2 cm thick.

The next oldest rocks are leuconorites and norites which occur principally to the east of the paragneisses, but also to the southwest. The leuconorite is coarse-grained (2-4 cm average grain-size) and is dominated by plagioclase and poikilitic orthopyroxene. I have not investigated the body to the southwest, but the leuconorite to the east typically becomes finer-grained and more mafic (CI~30-35) near the contact.

Seemingly present everywhere at the contact between leuconorite and paragneiss is fine- to medium-grained diorite. This unit is distinguishable from the leuconorite by its generally finer grain size, higher mafic and opaque mineral contents, and higher sulfide mineral concentration giving the weathered outcrops a rusty appearance. Locally the diorite is as rusty and punky as the paragneiss and likewise contains lenses of graphite.

To the east of the paragneiss the diorite occurs only as a thin unit at the contact between leuconorite and paragneiss, but to the west and north of the paragneiss, Wheeler (FR 1973) has mapped broad areas where the diorite is not restricted to the leuconorite contact (Fig. 6).

The youngest rock unit in the area is ovoidal rapakivi in the southwestern part of this area. The rapakivi was mapped by Wheeler (FR 1973), and he reported that it cut across both diorite and leuconorite.

Fig. 6. Geologic map of Osumilite Lake area. Compiled from Wheeler (FR 1973) and 1980 study by Berg.

Discussion

Based on the attitudes of contacts, best determined by the V-shaped map pattern in the stream valleys (Fig. 6), the aluminous paragneiss apparently underlies the igneous rocks in the area. The metamorphism of the paragneiss was undoubtedly effected by the intrusion of the leuconorite-diorite bodies. The rapakivi does not appear to have had any effect, at least in the most carefully studied area directly west of Osumilite Lake.

Whether the intrusion of leuconorite and diorite was one continuous event or two distinct events is difficult to ascertain. In other field areas Wiebe (e.g. FR 1975) has suggested that diorite represents late-stage interstitial liquid squeezed out of the leuconorite. The paragneiss has been metamorphosed at such high temperatures (probably greater than 850°C; Berg and Wheeler, 1976; Grew, 1980) that it would seem necessary that the heat source be a magma which would crystallize leuconorite and diorite in one cooling event, rather than a magma which would crystallize just diorite as the last discrete thermal event. Nevertheless, there are possible retrograde effects in some of the paragneiss (osumilite recrystallizing to very-fine-grained intergrowths of other unidentified minerals), and two events cannot be ruled out.

The presence of osumilite visible with the aid of a hand lens is restricted to an area of the paragneiss less than 0.4 km from the leuconorite-diorite contact in the area directly west of Osumilite Lake. At greater distances from the contact osumilite has not been recognized in the field, and in these areas biotite is quite common. The line drawn on Fig. 1 between the osumilite-bearing and osumilite free paragneiss is assumed to represent an osumilite "isograd."

I originally thought that the presence of osumilite was simply related to the concentrations of sulfide + graphite. The abundance of sulfur in the rock would have robbed the silicates of iron in order to produce the sulfides, leaving a Mg-enriched silicate assemblage which is necessary for the production of osumilite parageneses. This undoubtedly was a factor in the development of the osumilite-bearing assemblages; however, there is paragneiss, enriched in sulfides but lacking osumilite, located outside of the osumilite isograd. Therefore, based solely on field evidence, this postulated "isograd" does not appear to be merely a function of bulk composition. These problems will be explored with both petrographic and analytical studies.

GEOLOGY OF THE SNYDER BAY AREA: PART I

J.A. Docka
Harvard University¹

Introduction

The nature and origin of the metasedimentary rocks described by Berg (FR 1974, FR 1975), found overlying the Snyder Group northwest of the contact of the Kiglapait intrusion, are at present controversial. While Berg considers these metasedimentary rocks to be continuous formations which run parallel to the strike of the Snyder Group and hence would constitute an integral part of the Aphebian stratigraphy, Speer (1978) has proposed rather that they are large xenoliths "meters thick and tens of meters long."

Hand in hand with the problem of the metasedimentary rocks (hereinafter called the unnamed sequence) goes the origin of the enigmatic mafic granulites which are intercalated with them. These have variously been considered plutonic igneous (Outer Border Zone of the intrusion; Morse, 1969; Berg, 1971), auto-metamorphosed igneous precursors to the intrusion (Berg, FR 1974) and metabasites unrelated to the intrusion (Berg, FR 1974, FR 1975; Docka, 1979). Bulk chemical analyses (Morse, 1981b) as well as oxygen isotope data (R.I. Kalamarides, personal communication, 1980) preclude a simple petrogenetic relation with the intrusion, while field relations in the Avakutakh River region (Berg, FR 1975) suggest that the OBZ mafic granulites are older than the Kiglapait intrusion. Thus the suggestion that the mafic granulites are unrelated to the intrusion appears well-founded. However, this does not eliminate the possibility that the metasedimentary rocks of the unnamed sequence are xenoliths enclosed in mafic granulite, but simply that such an event is unrelated to the intrusion. The field relations of the mafic granulites then play a key role in unravelling the origins of the unnamed sequence and its relation to the Snyder Group, as well as being a puzzle in themselves. As described in this and the two following reports, more detailed mapping was performed to

¹Authors' full addresses are given at the back of this volume.

1980 in the vicinity of Snyder Bay, where the unnamed sequence and associated mafic granulites are best exposed, to define further their nature and relation to the Snyder Group as presently defined (Speer, 1978).

In addition, a comparative geothermometry study has recently been initiated in this portion of the contact aureole (Docka, 1979; Berg and Docka, 1980). The variety of protoliths and, consequently, metamorphic assemblages amenable to thermometry found within the aureole provides an unparalleled opportunity for the evaluation and calibration of geothermometers in a controlled setting. To date five geothermometers (opx-cpx, ol-opx, cpx-ilmenite, opx-ilmenite, ilmenite-magnetite) have been used extensively while two others (garnet-biotite, garnet-cordierite) have been applied to a limited extent. However, this study has been hampered by a lack of samples from the low-grade outer part of the aureole. Sampling was undertaken in Cold Comfort Bay and on Snyder Island specifically to obtain assemblages useful for geothermometry.

Middle Bay to Wendy Bay Area (Fig. 7)

Unnamed Sequence. These units have been previously described by Berg (FR 1974, FR 1975). More detailed stratigraphic sections and formation descriptions are given by Schuh in the following report. As noted by Berg, these units typically run parallel to the Snyder Group formations regardless of intercalated mafic granulites. Individual formations are never found at high angles to each other nor are small inclusions of these formations found in the enclosing mafic granulites. Cross-cutting dikes of mafic granulites are not commonly observed. Whether the few such dikes as do occur are equivalent to the mafic granulites is indeterminate in the field, and small apophyses of mafic granulite into metasediment are not observed. The above relations suggest that if the metasedimentary rocks are large xenoliths, they became hot and plastic rapidly enough to retain a coherent form, rather than fracturing to allow small xenoliths and apophyses to form. Because of the above observations and because rotated beds do not occur, one could also suggest that little significant movement has occurred. Indeed, it becomes difficult to entertain seriously the notion that these beds are xenoliths, since the bulk of evidence suggests otherwise.

Mafic Granulites. The mafic granulites exhibit several interesting textures. They have in general a well-defined foliation which parallels the contact of the intrusion. However, in several outcrops small isoclinal folds on the scale of 5-10 cm are ubiquitous. These folds are typically defined mineralogically by feldspar + amphibole + quartz + biotite. A second texture noted consists of interconnecting cracks, approximately 10-15 cm in length, filled with feldspar. These do not appear to resemble veinlets and seem to be limited to horizons parallel to strike. The texture is best viewed in exposures along Falls Brook, but can be seen on the point between Lakes Brook and Cold Comfort Bay as well. It is possible that this texture could represent brecciated flow tops. If so, at least some of the mafic granulites would have originated as flows.

The pillow textures first described by Berg (FR 1974) were again found. These were sampled in the hope that chemical analyses might give further evidence for their origin.

Distinct from the intercalated mafic granulites are various cross-cutting mafic dikes. These appear to have undergone contact metamorphism, and are mineralogically indistinguishable from the intercalated mafic granulites. Chemical analyses of these dikes would be of interest because, if similar to the intercalated mafic granulites, they would provide some evidence for the xenolithic origin of the unnamed sequence. If dissimilar, they might represent the igneous precursors of the Kiglapait as originally suggested for the intercalated mafic granulites (Berg, FR 1974).

Snyder Group. Perhaps the most important observation made during the recent field season involves the variation in lithology of the formations of the Snyder Group on the point between Wendy Bay and Middle Bay, Wendy Bay Hill (Hill 900 of Speer, FR 1974).

Approximately half way up the hill on the east side of Middle Bay, the iron-rich silicate minerals in the Silicate Iron Formation disappear and the beds of this formation tend to grade into a silty sandstone + sulfide minerals (see Schuh, this report). This trend continues over the top of the hill and into Wendy Bay. Because of the weathering properties of the Quartzite-Marble formation, it could not be traced continuously up the hill. However in the saddle at the top of the hill, an outcrop contains clinopyroxene + feldspar + quartz + amphibole. This rock is found stratigraphically above the equivalent of the Silicate Iron Formation but below

the Graphite-Sulfide Siltstone, and macroscopically resembles the Quartzite-Marble formation. This exposure seems to have been previously considered to be a variety of migmatite (Speer, FR 1974; Berg, FR 1975). However, the extent of exposure observed suggests it is a continuous unit rather than a block enclosed in migmatitic matrix. The apparent differences in mineralogy may be attributable to a difference in grade, fluid phase or bulk chemistry. This rock is suggested to be a more Si-Al rich, less carbonate rich equivalent of the Quartzite-Marble formation derived from a more clastic protolith. The Graphite-Sulfide Siltstone varies laterally as well, grain size becoming somewhat coarser and sulfides becoming less prominent as the crest of the hill is approached from the west (see Schuh, this report).

At the top of the hill the nature of the contacts between these three formations changes as well. In Cold Comfort Bay and along the exposures at the mouth of Falls Brook the contacts between these formations (with the exception of intercalated lenses) are relatively sharp. However the contacts at the top of Wendy Bay Hill are continuously gradational between the formations and no intercalated lenses are observed. Absolute contacts are difficult to define and all formations tend to thin as well. While on the west side of the hill thinning may be attributable to stoping by mafic granulites, at the top of Wendy Bay Hill no mafic granulites are found between these three formations.

All these formations appear to grade into silty sandstone intercalated with mafic granulites on the west side of Wendy Bay at the base of Wendy Bay Hill. This sandstone bears no striking resemblance to any of the three formations but it lies stratigraphically above the Lower Quartzite and given the gradational nature of the contacts on top of Wendy Bay Hill it is suggested that this unit is the time depositional equivalent of the Silicate Iron Formation, Quartzite-Marble and Graphite-Sulfide Siltstone found in Middle Bay. This would suggest a change in the depositional environment of the basin in which the Snyder Group was deposited. A simple model explaining the more clastic nature and thinning of the formations would call for a minor topographic high either in or to the west of Wendy Bay.

Cold Comfort Bay to Middle Bay Area

The area around and to the south of Cold Comfort Bay was mapped and sampled both for assemblages amenable to geothermometry and to try and reconcile discrepancies between the maps presented by Berg (FR 1975) and Speer (1978). One key area is the point on the coast between Lakes Brook and Cold Comfort Bay. This hill has been portrayed by Speer as mafic granulite considered to be part of the OBZ of the Kiglapait intrusion. Berg, however, notes the presence of Snyder Group Graphite-Sulfide Siltstone and iron formation and calc-silicate unit of the unnamed sequence. This summer, Quartzite-Marble formation of the Snyder Group was found below the Siltstone. In addition, it was found that the Siltstone as well as what may be a lens of Upper Quartzite crop out on the top of the hill. The units on this hill are thinner than Speer's measured sections (1978), a fact which might be attributed to stoping by mafic granulite. However, south of Cold Comfort Bay where no intercalated mafic granulites are observed these formations also thin, and it seems likely that depositional thinning was as important or more important than stoping in thinning the Snyder Group formations on the hill between Lakes Brook and Cold Comfort Bay. Juxtaposition of these formations with the Upper Quartzite on the east side of Cold Comfort Bay requires the presence of a NE-SW fault. The changing geometry of the metasedimentary rocks and intercalated mafic granulites make the reverse or normal nature of the fault indeterminate. In addition, the fault is obfuscated by an igneous breccia which intrudes between the hill and Cold Comfort Bay as mapped by both Berg and Speer. If the determination of a fault is correct, it does not seem unreasonable for the breccia to have intruded along it as a major zone of weakness.

Summary

The bulk of the evidence suggests that the unnamed sequence is a legitimate depositional sequence, rather than being xenolithic. Characterization of the different formations of this sequence becomes most important for possible correlations and reconstruction of paleo-environments. The mafic granulites remain a knotty problem. Whether they are a product of early igneous activity related to the Nain complex or are a relict of a much earlier period of volcanism has not been determined. However, it does seem that they are unrelated directly to the Kiglapait intrusion.

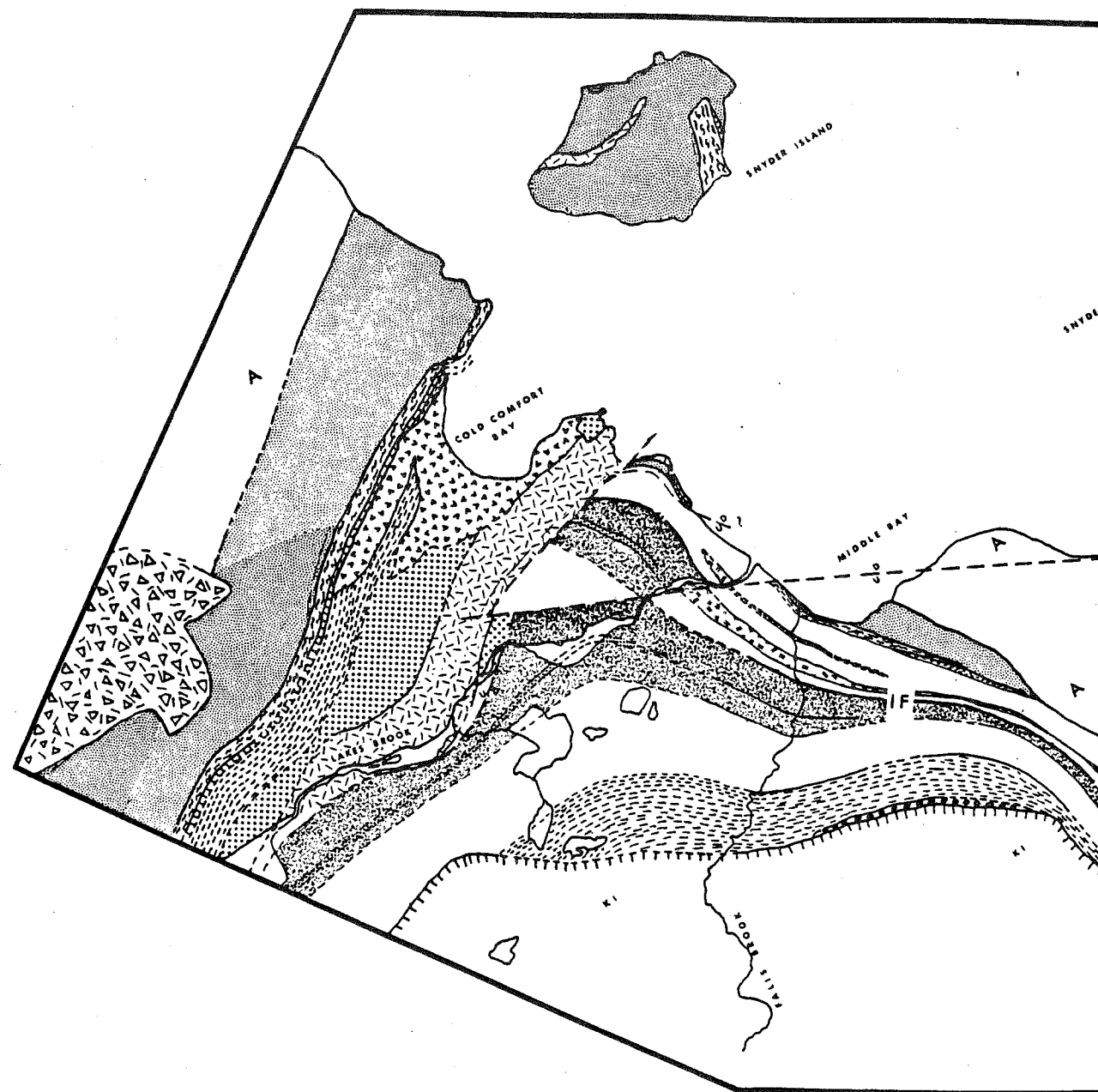


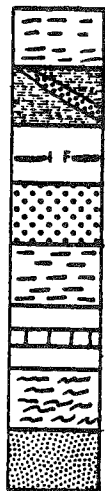
Fig. 7. Geologic map of the Snyder Bay area. Modified by M. L. Schuh from Speer (FR 1972) and Berg (FR 1975).

KEY

FAULT — — —

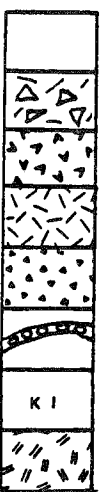
META-SEDIMENTARY UNITS

- Pyroxene paragonulite
- Calc-silicate unit
- Iron formation
- Upper Quartzite
- Graphite-Sulfide Siltstone
- Quartzite-Marble
- Silicate Iron Formation
- Lower Quartzite



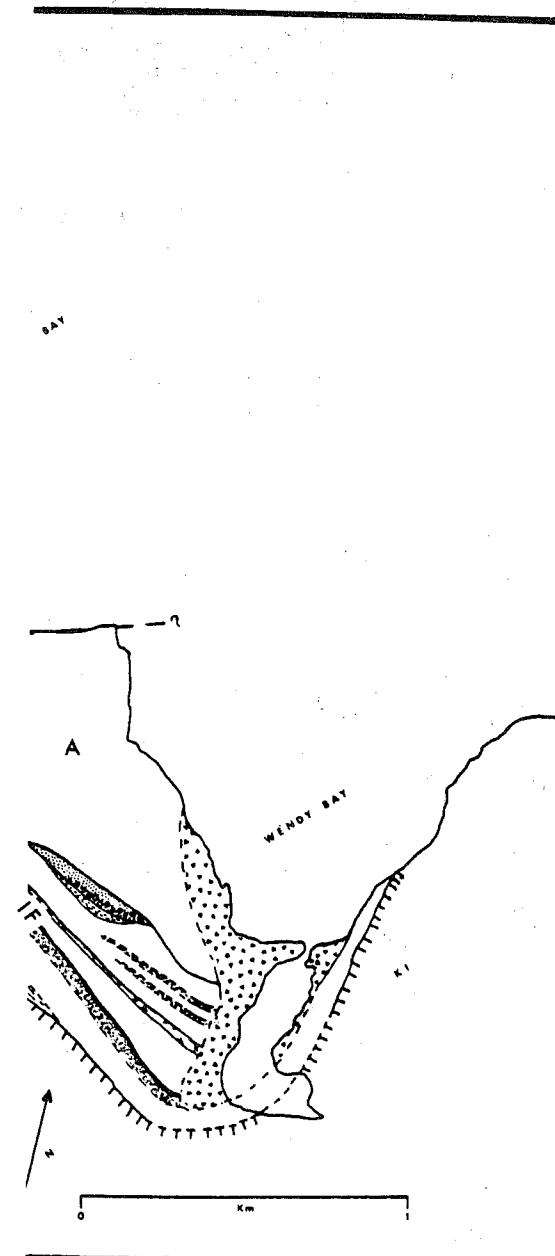
IGNEOUS UNITS

- Mafic granulite
- Amphibolite
- Ultramafics & metabasites
- Igneous breccia
- Agmatite
- Melatroctolite
- Kiglapait intrusion
- Spinel peridotite



ARCHEAN

- Gneisses



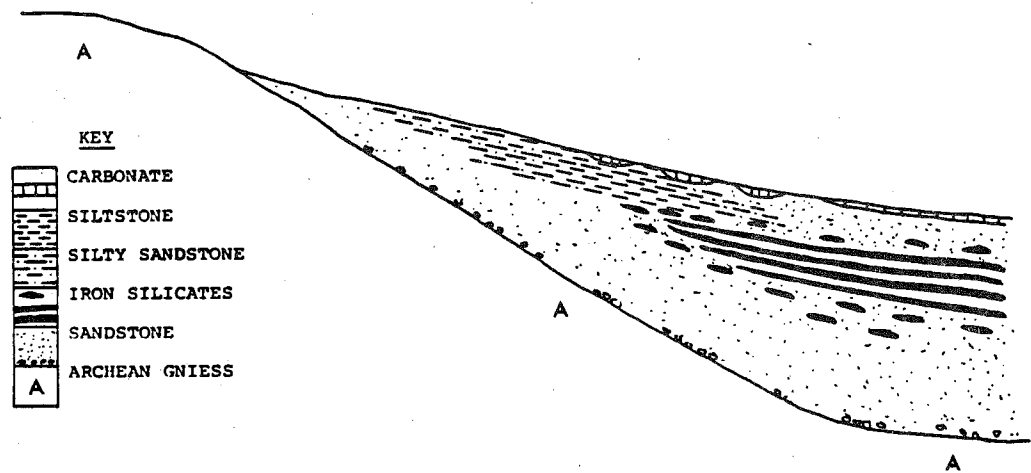


Fig. 8. Schematic cross section representative of the sedimentary facies changes observed on the western side of the hill between Middle Bay and Wendy Bay.

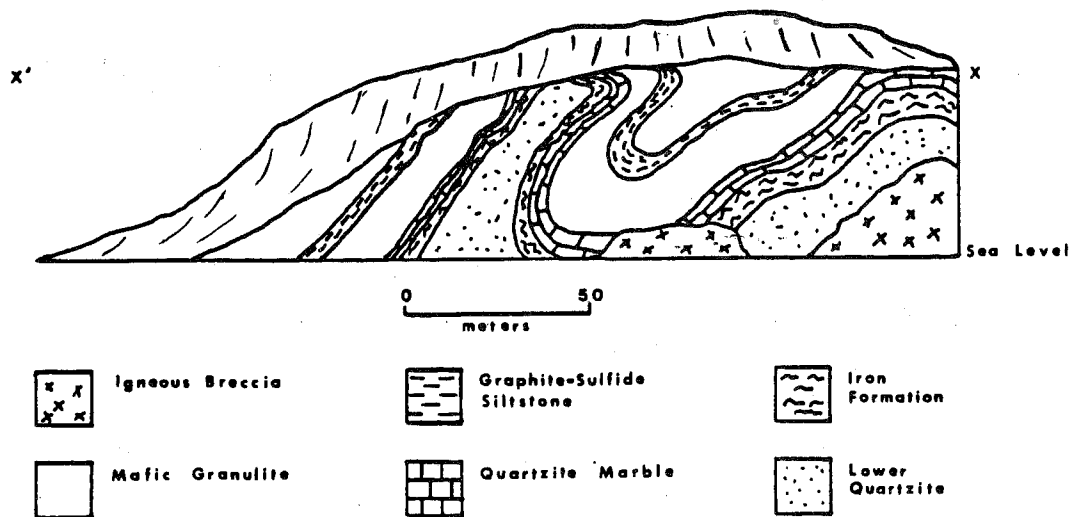


Fig. 9. Schematic cross section through the point between Cold Comfort Bay and Middle Bay. For location see Fig. 10.

GEOLOGY OF THE SNYDER BAY AREA: PART II

M. L. Schuh

Northern Illinois University¹Introduction

During the summer of 1980, an area previously mapped by Speer (FR 1972) and Berg (FR 1975) was revisited in hopes of clarifying and resolving certain aspects of the two maps. It was found that in the Cold Comfort Bay area, Speer's map is essentially correct, whereas in the Middle Bay area, Berg's map better represents that region.

Further study of this area (Fig. 7) resulted in a better definition of facies relationships of the Snyder Group on the hill between Wendy Bay and Middle Bay and the discovery of a small overturned anticline and syncline, within the Snyder Group, on the point between Middle Bay and Cold Comfort Bay. Evidence for three more faults was also recognized.

Snyder Group on "Wendy Bay Hill"

Stratigraphic units representative of the Snyder Group are present on the hill between Middle Bay and Wendy Bay. The stratigraphic thicknesses of all units are markedly diminished relative to the measured thicknesses at Cold Comfort Bay. All units thin or pinch out toward the top of the hill from either side.

The Archean rock which forms the depositional floor for the Snyder Group is a white to pinkish colored quartzo-feldspathic rock with migmatitic zones. Lenses of amphibole-rich rock with discrete pods of brilliant green diopside \pm other calc-silicate minerals are common. The contact between the Archean and the basal quartzite (Lower Quartzite) of the Aphebian Snyder Group is marked by a pebble conglomerate. The quartz pebbles are well-rounded and ≤ 1 cm in diameter. On this hill, the Lower Quartzite lacks the Al-silicate and phengitic porphyroblasts that characterize the unit in the Cold Comfort Bay region. Pelitic layers and lenses are also less common and, where they do occur, are quite thin. However, cross-bedding and pebble conglomerates are common to both areas.

¹Authors' full addresses are given at the back of this volume.

The contact between the Silicate Iron Formation and the Lower Quartzite is gradational. Sporadic pods of grunerite mark the start of the transition from Lower Quartzite to Silicate Iron Formation. These pods become more and more common near the base of the Silicate Iron Formation which is marked by the first continuous layers of iron silicates. The Silicate Iron Formation becomes thinner and more quartz-rich as one goes uphill. Approximately half-way up the hill, it is transitional into the Lower Quartzite.

The rock type overlying the Silicate Iron Formation is variable. At Falls Brook, a highly sulfidic and graphitic quartzite with a calcareous upper zone (Quartzite-Marble) conformably overlies the Silicate Iron Formation. Again, the contact is gradational. Further up the hill, mafic granulite overlies the Silicate Iron Formation. Beyond that, and shortly before the Silicate Iron Formation disappears, a siltstone unit occurs between the mafic granulite and the Silicate Iron Formation.

The siltstone generally exhibits good layering, and some cross-bedding can be discerned. Locally, pods of diopsidic and calcareous material can be found at the top of this unit. Where the siltstone first appears, it is composed of fine-grained quartz and plagioclase with discontinuous layers and pods of biotite + amphibole + plagioclase. Further uphill, bedding becomes more distinct and sulfides (pyrite and chalcopyrite) are present. This unit becomes progressively richer in quartz, coarser grained, less sulfidic, and more massive until it is indistinguishable from the Lower Quartzite. Shortly after the disappearance of the recognizable siltstone unit, the Lower Quartzite (which at this point is the sole representative of the Snyder Group) completely pinches out between the Archean gneiss and the mafic granulite.

At Falls Brook, the top of this mafic granulite is defined by the presence of a second siltstone unit which is highly sulfidic and very similar to the Graphite-Sulfide Siltstone at Cold Comfort Bay. However, it is only several meters thick at the Falls Brook exposure. This unit also thins out near the top of the hill. Above the siltstone is a second mafic granulite.

On the east side of "Wendy Bay Hill", the stratigraphy, where the Snyder Group reappears, is quite different. (Poor exposures of rock on this side of the hill prevent as detailed a survey as on the western side.)

The Snyder Group is represented by a thin, clean, white quartzite, an overlying, thicker conglomerate which contains boulders of quartzite and Archean migmatite (up to .5 meters in diameter) in a "dirty" quartz matrix, and a siltstone with sporadic occurrences of sulfides. Again, all units are gradational into the quartzite which, in turn, pinches out between the Archean basement rock and mafic granulite.

At the base of the hill, two sulfidic siltstone units occur between mafic granulites. These also appear to pinch out uphill.

All observations on "Wendy Bay Hill" seem to support Berg's (FR 1975) hypothesis that the hill represented a topographical high at the time of deposition of the Archean metasedimentary rocks. The disappearance of the different rock units can be accounted for by sedimentary facies changes (Fig. 8) within the depositional sequence. As to the mafic granulites' relationship to all this, they remain as enigmatic as before. The lower mafic granulite either dilates the Snyder Group between the sulfidic siltstones and Quartzite-Marble, or it is a flow upon which the siltstone was subsequently deposited.

Structural Interpretation of the Hill between Cold Comfort Bay and Middle Bay

The area between Middle Bay and Cold Comfort Bay is structurally complex. Features which are often diagnostic of faulting seem to indicate that several faults occur in this locality. Evidence presented below suggests the presence of a low angle reverse fault and two nearly vertical faults which may be related, and Berg (FR 1975) mapped a high angle normal fault which also cuts through this region. The igneous breccia obscures the relationships which may exist between any faults in the region.

On the hill between the two bays, two structural styles are displayed. At the point, on the shoreline between Middle Bay and Cold Comfort Bay, an overturned anticline and syncline can easily be traced on the steep faces of the promontory. Only Snyder Group units and interlayered mafic granulite sills comprise this structure. A schematic cross-section through this structure is shown in Fig. 9. The Snyder Group assemblage displayed on this point more closely resembles the Snyder Group exposed on "Wendy Bay Hill" than that cropping out in Cold Comfort Bay. However, the individual units, specifically the Silicate Iron Formation, Quartzite-Marble and Graphite-Sulfide Siltstone, are dramatically thinner. Mafic granulite and unnamed

group units, which comprise the bulk of the exposed part of the hill, overlie the overturned anticlinal and synclinal structure. Nowhere were folds of any type discerned in the unnamed group on this hill.

A nearly vertical scarp bounds the west side of this hill. On this scarp, there are exposures where mafic granulite overlies Upper Quartzite. The Upper Quartzite appears to be striking NNE and has an apparent thickness of 3 meters. The base of the overlying mafic granulite is characterized by abundant slickensides and sulfides. Nearer the shore, the Upper Quartzite is pinched out and the mafic granulite overlies a second basic, meta-igneous unit (possibly the igneous breccia). A clean, chloritic contact exists between these two units. This contact has an apparent dip at a low angle to the SSW.

The aforementioned observations suggest the presence of a low angle reverse fault on this hill. What happens to the southern trace of this fault is unknown as overburden and the igneous breccia obscure the field relations.

A second fault seems to be indicated by the nearly vertical scarp which bounds the west side of the hill between the two bays. The occurrence of the igneous breccia parallel to this scarp also suggests that this direction is a plane of weakness and reinforces the possibility of a fault in this location. The western side of the fault appears to have moved up with respect to the eastern side. However, there is not enough stratigraphic control to determine conclusively the relative movement.

STRATIGRAPHY OF THE UNNAMED IRON FORMATION
AND CALC-SILICATE UNITS IN THE SNYDER BAY AREA

M.L. Schuh

Northern Illinois University¹

Introduction

In the summer of 1974, J.H. Berg discovered a stratigraphic sequence of rocks overlying the Snyder Group (FR 1974). He identified a banded iron formation (~5 m thick) and a calc-silicate unit (~65 m thick) in addition to several other units. This summer, a detailed study of the iron formation was initiated. The overlying calc-silicate unit was also closely inspected in the hope of finding more iron-rich units.

Iron Formation

Throughout most of the study area (Fig.10) the iron formation is a continuous unit varying from ~1 m to a maximum of 4 m thick. On the top of the hill between Wendy Bay and Middle Bay, the iron formation becomes discontinuous and occurs as pods overlying a mafic granulite. On the Wendy Bay side of the hill, it attains a maximum thickness of 1 m near the bottom of the hill. It is represented by a medium-grained quartzite containing plagioclase and minor interstitial iron silicates and magnetite. Magnetite bands appear near the base of the hill and correlate with the disappearance of plagioclase in the unit.

Four detailed stratigraphic sections (Fig. 11) of the iron formation in the Middle Bay area and at hill 1700 show that sub-divisions within the iron formation can be correlated. In the Middle Bay region, the base of the iron formation is marked by a 2-4 cm thick layer of iron silicates and magnetite. The rest of the unit is predominantly quartzite with thinly to widely spaced magnetite bands (1-10 mm) and Fe-silicate-rich layers (4-20 mm). The quartzite varies from being white and relatively "clean" to greenish with interstitial fayalite, grunerite, and minor pyroxene and magnetite. Pods of iron silicates occur locally in the quartzite. In

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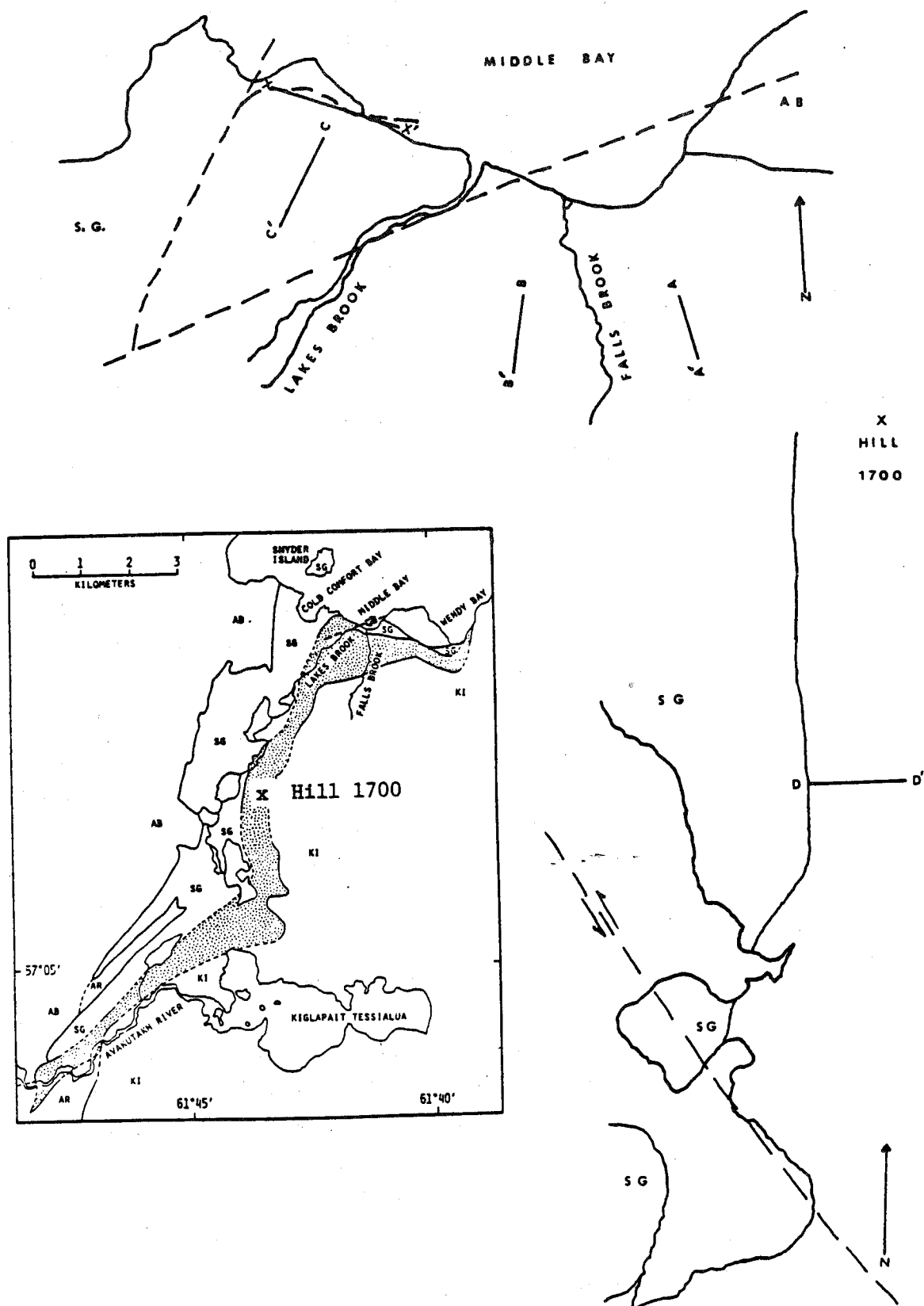


Fig. 10. Map showing locations of stratigraphic sections. Traverse A-A' corresponds to stratigraphic column A-A' and iron formation column A, etc.

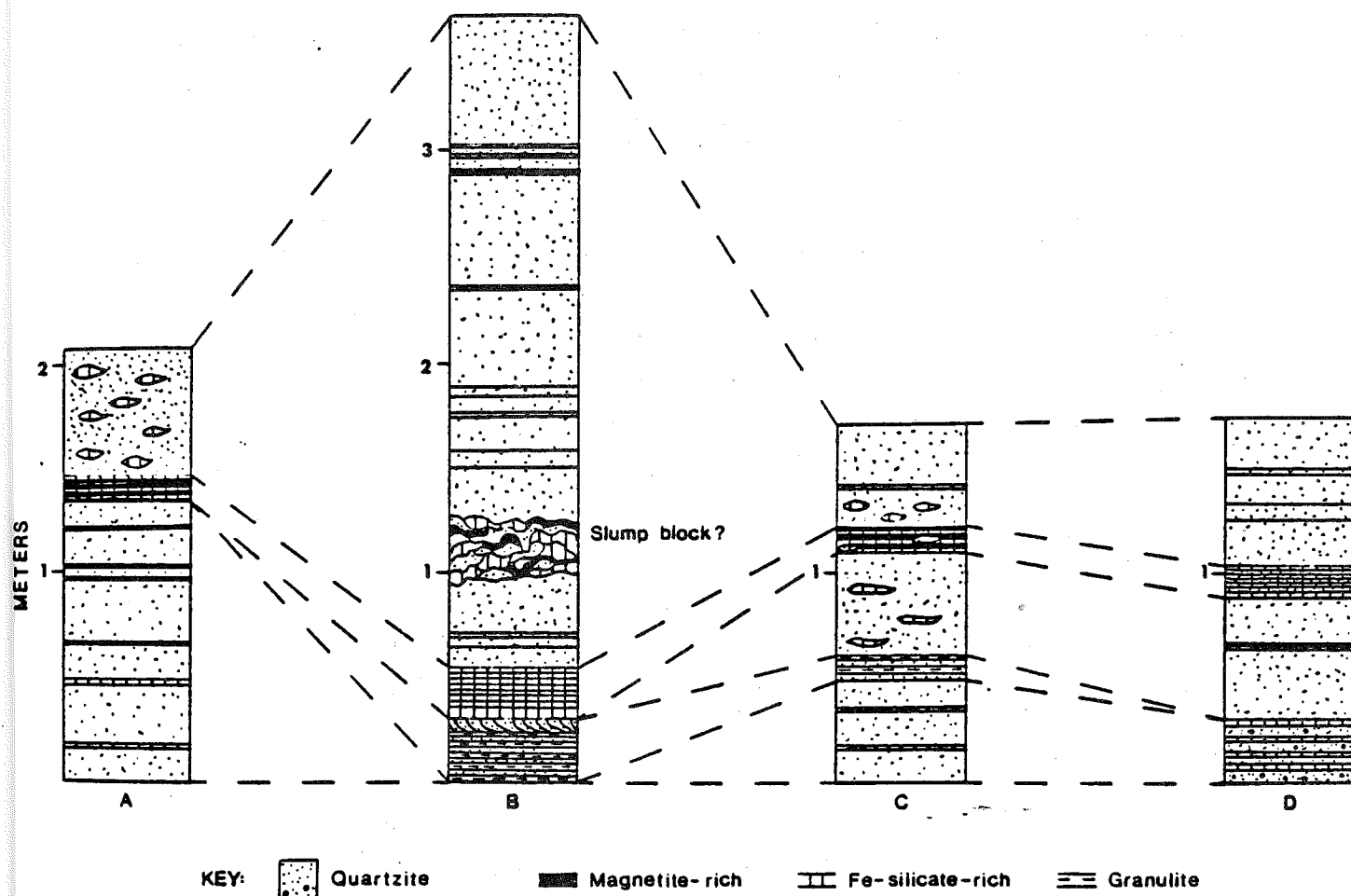


Fig. 11. Stratigraphic columns through the iron formation showing lithological changes and possible correlations of sub-units. See Fig. 10 for locations.

two of the sections (B and C), a thinly laminated granulite unit associated with fine-grained quartzite is present. In section B, what may be normal cross-bedding is observed in the top 4 cm of this unit. A sub-unit which is continuous throughout the iron formation consists of thinly laminated iron-silicate-rich and magnetite-rich or quartz-rich layers. The iron silicates are represented by fayalite, ferrohypersthene, hedenbergite and locally, grunerite. Locally, mafic granulite sills and dikes dilate the sequence, and in Fig. 11 they have been deleted from the sections in order to indicate true thickness and lithological variation.

Calc-Silicate Unit

The unit overlying the iron formation, termed the calc-silicate unit by Berg (FR 1975), was found to be composed of varying lithologies (Fig. 12), and ranges in thickness from a minimum of ~ 20 m on the hill between Wendy Bay and Middle Bay (Fig. 12, A-A') to at least 50 m elsewhere in the area. The major rock type in this unit is granulite (predominantly amphibole + plagioclase) which is variably magnetic depending on its proximity to the Kiglapait intrusion. It varies from being massive to thinly laminated (or to having a very well-developed foliation). The "laminations" are formed by alternating plagioclase-rich and amphibole-rich layers. In some of the layered granulites, structures reminiscent of soft sediment deformation occur and it is not uncommon to find small boudinage-type lenses of calc-silicate minerals, such as wollastonite, diopside, tremolite and olivine. Bands of thinly laminated granulite with magnetite-rich and iron silicate-rich layers, which are similar to the "granulite sub-unit" of the iron formation, are also found. These bands commonly have pods of magnetite or iron silicates at their tops. Lenses of other laminated rocks also occur. These include a banded magnetite and olivine lens and a thinly laminated lens consisting of magnetite-rich, pyroxene-rich, cordierite-rich, biotite-rich, and amphibole-rich layers. Berg (FR 1974) lists other lithologies which also occur.

A startling discovery was the recognition of another iron formation unit, within the calc-silicate unit, which attains a maximum thickness of 15 m. It is traceable from Falls Brook to near Lakes Brook, on the hill between Middle Bay and Cold Comfort Bay, and in outcrops in various areas on the east side of Lakes Brook up to the southern end of Hill 1700. Between

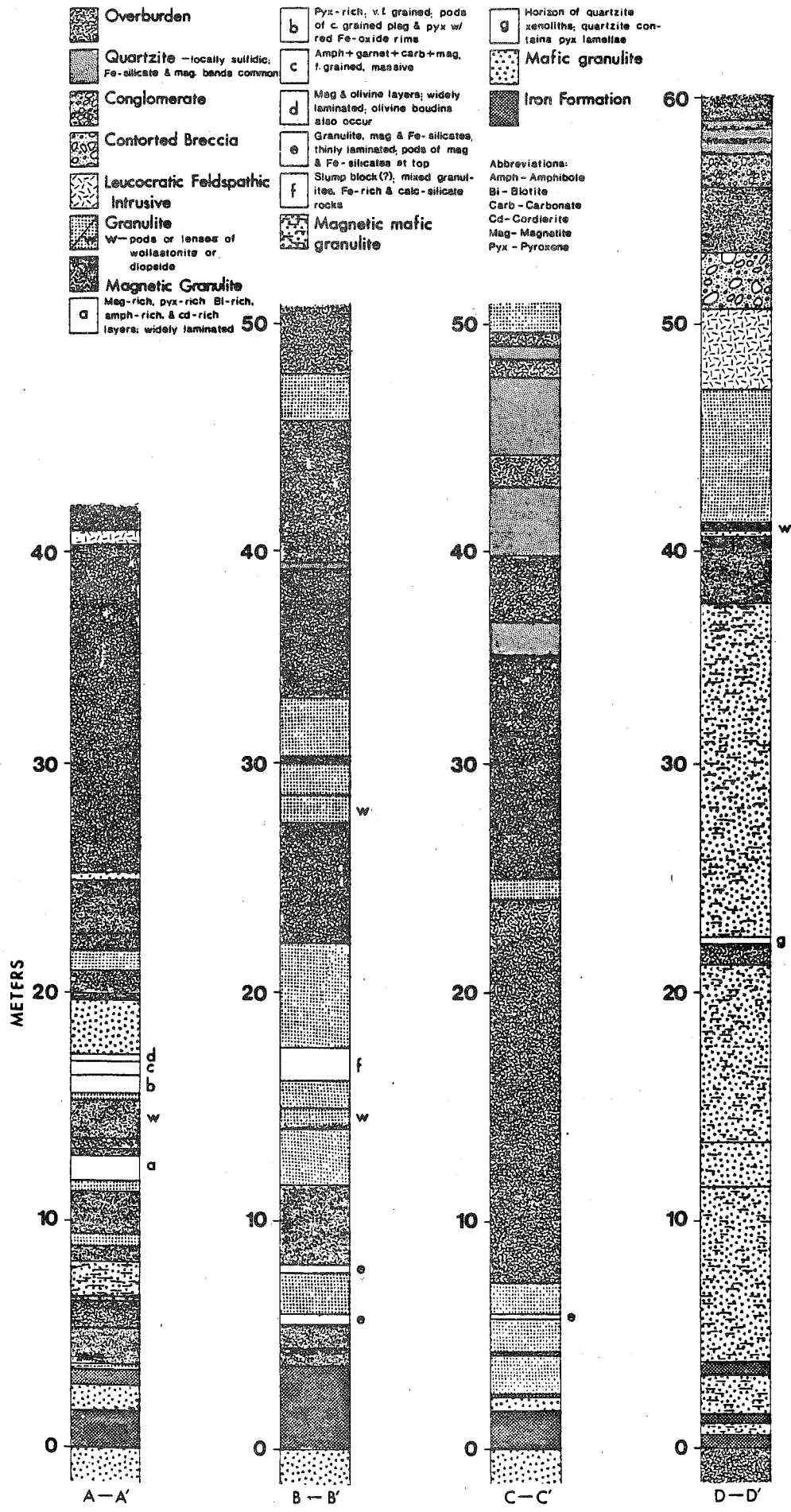


Fig. 12. Stratigraphic columns for the unnamed group from the base of the iron formation to the highest exposed portions of the calc-silicate unit.

Falls Brook and Lakes Brook, this unit is represented by a thinly laminated quartz-rich rock (~30 cm thick) with bands or lenses of quartz + fayalite + hedenbergite + ferrohypersthene + plagioclase + magnetite, quartz + fayalite + pigeonite + hedenbergite + plagioclase + magnetite, and quartz + hypersthene + fayalite + plagioclase + magnetite. This is transitional into a 60 cm thick layer of fine- to medium-grained rock composed of quartz and minor plagioclase, which appears to be a recrystallized arkosic siltstone. On the hill between Middle Bay and Cold Comfort Bay, the iron formation occurs as a 10 m thick quartzite which varies from being relatively pure, massive quartzite with minor interstitial magnetite, fayalite, and grunerite to thinly laminated quartzite with magnetite-rich, fayalite-rich, and quartz-rich layers to a highly sulfidic, massive quartzite, with local laminae of grunerite. Along the east side of Lakes Brook, exposures of this unit indicate a thickness of ~15 m. This unit is also exposed on the west side of Hill 1700 and is transitional into a conglomerate. The conglomerate is composed of well-rounded to sub-angular boulders and cobbles of mafic granulite and calc-silicate unit lithologies in a matrix of quartz and lithic fragments. The presence of mafic granulite cobbles implies that the mafic granulites represent supracrustal flows or near-surface intrusives. The conglomerate grades upward into thinly to thickly laminated quartzite and granulite rock. In all areas, calc-silicate type lithologies overlie this quartzite unit.

Mafic Granulites

Mafic granulites underlie the lower unnamed iron formation in all areas. Local dikes and sills dilate the iron formation, but the dilatory mafic granulites were never seen to emanate from the underlying mafic granulite. Nowhere were inclusions of iron formation found within the underlying mafic granulite.

In the Snyder Bay area, mafic granulite dikes and sills are rare in the calc-silicate unit. However, at Hill 1700, magnetic mafic granulite occurs between the lower iron formation and calc-silicate unit. It is approximately 35 m thick and varies from being massive with clots of coarse-grained pyroxene and plagioclase to thickly banded (consisting of alternating mafic granulite and plagioclase-rich layers) to massive as described

before. Chloritized zones are common, and xenoliths of calc-silicate and quartzite occur. A curious feature is the occurrence of a traceable horizon of quartzite xenoliths approximately in the middle of the mafic granulite sequence. The contact between the calc-silicate unit and the mafic granulite is clean and distinct with a wavy interface.

Feldspathic Intrusive Rock

At Hill 1700, a 5 m thick, light-yellow-colored feldspathic intrusive rock is exposed. It varies from medium to very coarse grained and contains 60-70% plagioclase, 30% amphibole and 0-10% pyroxene (?). It contains very well-rounded xenoliths of mafic granulite which suggest in situ fracturing and subsequent invasion by the feldspathic magma.

The lower contact is abrupt. The upper contact consists of a contorted three-meter-thick zone of angular to sub-angular blocks of mafic granulite, amphibolite, and quartzite in a quartz-rich matrix. The blocks are warped and bent and appear to have undergone plastic deformation.

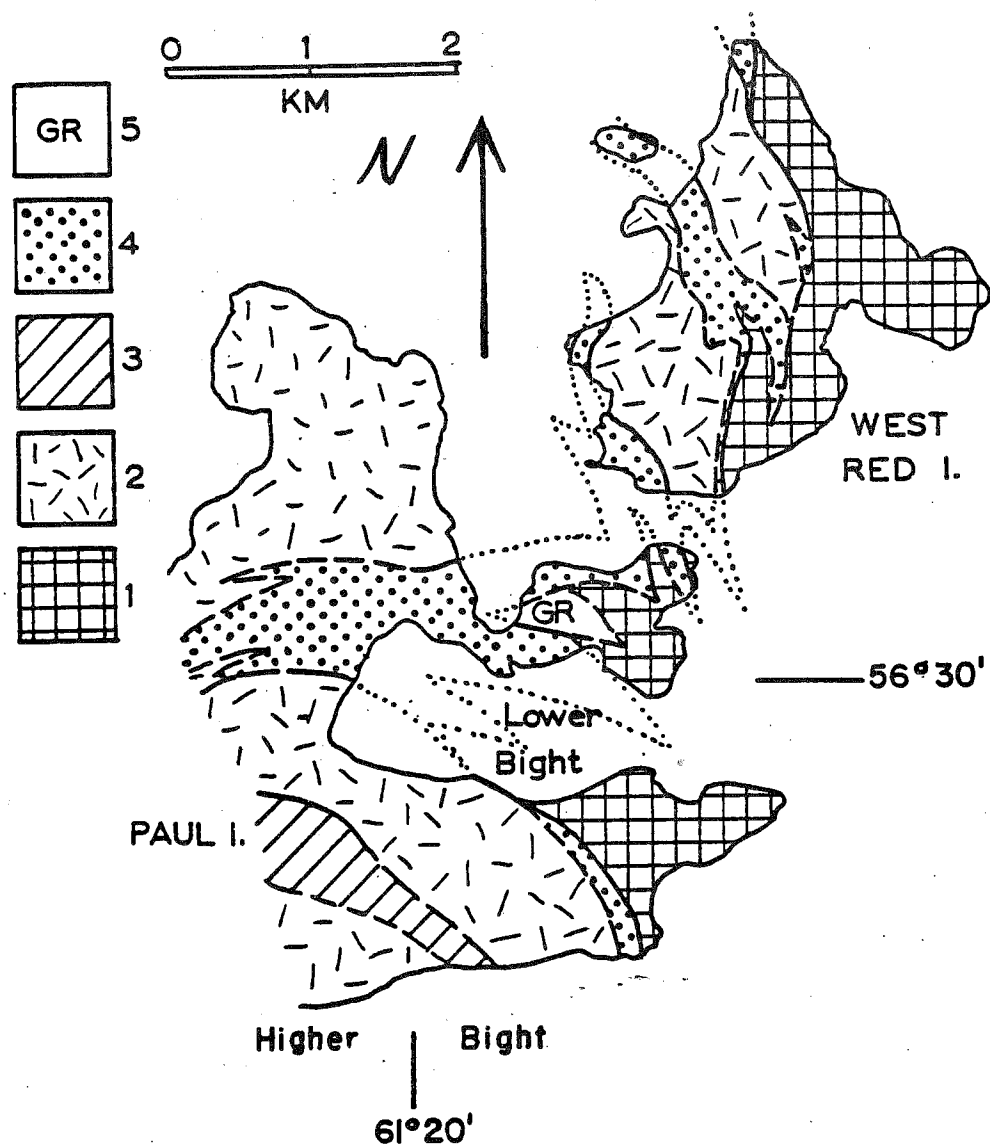


Fig. 13. Geologic map of West Red Island and the northeastern tip of Paul Island. Ford Harbour Formation (1); Anorthosite (2); Olivine-bearing anorthosite (3); Norite and diorite (4); Granite (5).

EASTERN MARGIN OF AN ANORTHOSITE PLUTON ON PAUL ISLAND

Robert A. Wiebe

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This field study was undertaken to determine (1) the nature of the compositional and textural variation of anorthositic rocks along an intrusive contact with the Ford Harbour Formation, (2) the relations between the West Red Island Layered Intrusion and adjacent anorthositic rocks, and (3) the relations between emplacement of the anorthositic rocks and deformation of the Ford Harbour Formation. Earlier field study by de Waard (FR 1971) provided a valuable framework for detailed remapping. His interpretations of this area also provided a useful focus for detailed re-examination of the West Red Island intrusion and the well-exposed anorthosite contact on the outer north shore of Higher Bight. Mineral data on these rocks are given by de Waard and Hancock (1976, 1977) and de Waard et al. (1977).

Figure 13 is a revised map of the area. The major difference between this map and de Waard's earlier one is the different distribution of dioritic rocks previously referred to as the West Red Island Layered Intrusion. The new map clearly indicates that there are several different zones of diorite, not a single arcuate body, and that each dioritic zone is related to a lens of basement rocks projecting into the anorthosite intrusion.

Main Features of the Contact Zone

West Red Island and eastern Paul Island provide superb exposures of the intrusive margin of a large anorthositic pluton. Relatively mafic rocks (diorite and norite, Unit 4, Fig. 13) occur prominently near the contact with the Ford Harbour Formation. These mafic rocks are locally gradational to typical coarse-grained leuconorite and anorthosite, and almost certainly comagmatic with them. The contacts between the intrusive and metamorphic rocks are consistently steep and sharp. On a small scale, structures within

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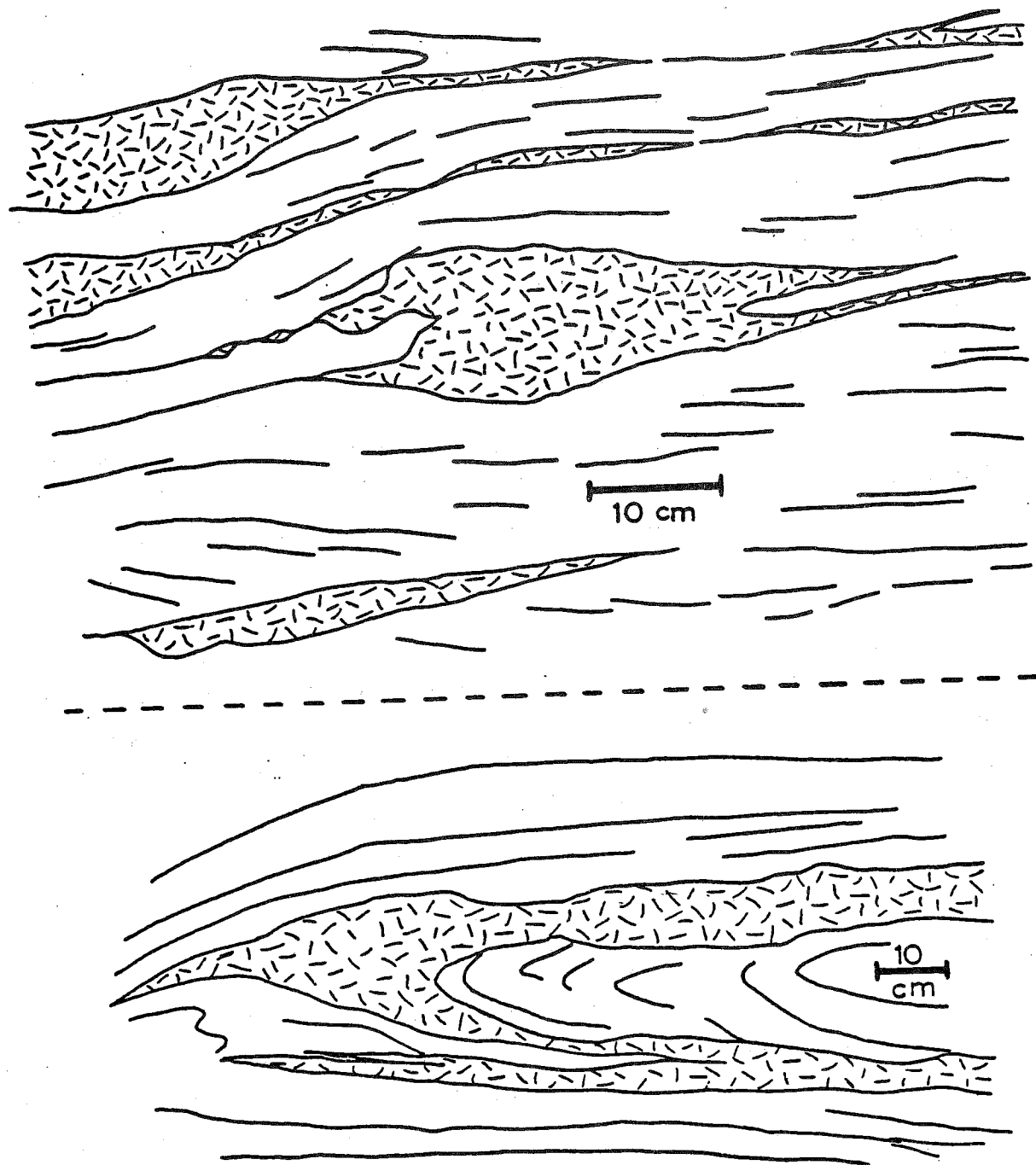


Fig. 14. Folded and boudinaged dikes of fine- to medium-grained anorthositic rocks in basic granulite of the Ford Harbour Formation. Color index (CI) of the anorthositic rocks ranges from less than 10 to approximately 40. Dominant CI is about 15 to 25.

the metamorphic rocks are commonly truncated by the intrusive contact. At the map scale the intrusive rocks appear to be broadly concordant. Nonetheless, the existence of metamorphic septa extending into the anorthosite pluton demonstrate that the anorthosite has cut across the metamorphic structures on a large scale.

Map Units

Ford Harbour Formation (Unit 1 of Fig. 13). The metamorphic rocks (Ford Harbour Formation) of this area have been thoroughly described by de Waard (FR 1971; de Waard and Hancock, 1976). The dominant rock types are basic granulite and quartzofeldspathic gneiss. Minor lenses and layers of calc-silicate, pelitic and ultramafic rocks occur sparsely. Diopside, grossular, and wollastonite were noted in the calc-silicate rocks. Pelitic rocks contain the assemblages cordierite-hypersthene and cordierite-sillimanite.

Strongly deformed dikes of anorthositic composition occur in gneissic granulite at several locations within 20 meters of the anorthosite intrusion (Fig. 14). These dikes are medium-grained and have compositions ranging from norite to nearly pure anorthosite. Except in pegmatitic patches, plagioclase is mostly between 2 and 5 mm in average dimension. Similar dikes were noted in one outcrop about 1 km from the contact.

Anorthosite (Unit 2 of Fig. 13). The area mapped as anorthosite includes both anorthosite (CI = 0-10) and leuconorite (CI = 10-35). Most rocks have CI between 5 and 25. The rocks are typically coarse-grained and display a wide variety of textures. On West Red Island the dominant variety consists of leuconorite (CI = 15) with blocky 1 cm plagioclase and scarce megacrysts up to 10 cm. Patches of norite and oxide-rich norite occur locally. Anorthositic rocks on Paul Island display many textural varieties of anorthosite and leuconorite with varying proportions, sizes and shapes of poikilitic patches of orthopyroxene. Some of these rocks have even-grained and equant plagioclase, 1-2 cm in diameter. These grade to rocks with seriate texture having variable proportions of phenocrysts, 1 to 20 cm in length.

Steeply dipping layering is apparent in many outcrops. Layers are defined by generally subtle changes in grain-size, mode or texture and have thicknesses from 10 cm to more than 10 meters. Most layers are homogeneous

internally. Structures commonly taken to indicate gravitational accumulation are lacking.

More pervasive than layering is a foliation commonly defined by preferred orientation of feldspar and/or by elongate dimensions of patches of poikilitic pyroxene. Although this foliation generally appears to be parallel with layering, the two structures were seen to be non-parallel in several outcrops. Uncommonly the foliation is axial to minor asymmetric folds in layering. Scarce dioritic dikes trend across the orientation of layering. Foliation within the anorthosite is also apparent within the diorite and cuts obliquely across the anorthosite-diorite contacts.

The anorthosite unit grades over short distances to the norite and diorite units. In some places the gradation occurs by a gradual decrease in the percentage and the size of the plagioclase. In other areas, the gradation is marked by well-defined steeply dipping interlayering of leuconorite, norite and diorite. Where the anorthosite unit is directly in contact with the Ford Harbour Formation, it has blocky plagioclase averaging between 0.5 and 1.0 cm and a CI of about 15-20.

Norite and Diorite (Unit 4 of Fig.13). Both of these units are gradational to the anorthosite. Nearly all rocks in the mapped areas have CI between 25 and 50. The only field distinction between these units is grain size: norite is characteristically coarse-grained with plagioclase and orthopyroxene >5 mm; diorite is relatively fine-grained with average grain size generally less than 2 mm. Areas mapped as norite grade to diorite along most contacts with the Ford Harbour Formation. Both rocks are composed dominantly of plagioclase and pyroxene with ilmenite and magnetite as important accessory phases.

Norite occurs along portions of the intrusive contact with the Ford Harbour Formation and as irregular areas of variable size within the anorthosite unit. Norite at the northern end of West Red Island displays prominent layering and foliation approximately parallel with the steeply dipping intrusive contact. Also present is a steeply plunging lineation, defined by elongate plagioclase and stretched clots of interstitial pyroxene. A few layers contain well-defined coarse perpendicular pyroxene.

Diorite occurs abundantly along the contact with the Ford Harbour Formation and as thick rinds apparently enclosing septa of metamorphic rocks which

project into the anorthosite. The dioritic rocks range from massive to layered on the scale of centimeters. All of the layering is steeply dipping. Dips between 60° and 90° are most common. In some sections the dip of layering varies continuously from 60° in one direction to 60° in the opposite direction with intermediate rocks having vertical dips. The color index of individual layers 1 cm to 10 meters thick varies from about 20 to 90. The most mafic layers consist of oxide-rich pyroxenite. These are prominent in dioritic septa on West Red Island. Many different textural varieties are present in these diorite septa, and abundant cross-cutting relations between different diorite bodies indicate that the septa have developed by multiple injections of dioritic magma. Many features in the layered rocks (e.g. boudinage, asymmetric drag folds, discontinuous faulting) suggest that they underwent deformation while still in a plastic state.

The diorite septum north of Lower Bight contains, in addition to the features described above, some prominent zones in which granitic rocks occur as a matrix to very fine-grained pillow-like masses of diorite.

Granite. A small body of massive granite occurs north of Lower Bight between diorite and the Ford Harbour formation. It clearly intrudes rocks of the Ford Harbour Formation and appears to include some rounded masses of diorite at one location.

Olivine-bearing Anorthosite (Unit 3 of Fig. 13). This small body has been described by de Waard and Hancock (1977). It is intrusive into the main anorthosite unit. A chilled margin and very fine-grained dikes of olivine-bearing anorthosite occur along the northern boundary of the intrusion.

Discussion

The layering within diorite and norite is consistently steep and either developed parallel to the contact or along zones extending from the ends of metamorphic septa which project into the anorthosite body. These relations, the character of the layering, and evidence for multiple injection strongly suggest that all layering is related to flow along steeply-oriented liquid-solid interfaces.

Some very fine-grained and massive dioritic rocks at the contact zone

probably represent chilled magma. The fine-grained dioritic rocks and coarser norite rocks do not appear to be reasonable parents to the anorthosite because they appear to be rich in oxide minerals and will probably prove to be relatively rich in apatite. Instead, these rocks appear to be differentiates from the same parent magma which produced the anorthosites.

The deformed anorthositic dikes in the Ford Harbour Formation near the contact south of Lower Bight appear to represent the oldest intrusive rocks related to this intrusion. I think it most probable that these dikes were intensely deformed during the continued intrusion and expansion of the main anorthositic pluton. The norites and dioritic rocks which form the present margin must therefore have been emplaced at the waning stages of pluton expansion and accordingly were deformed much less than the earlier anorthositic dikes. These anorthositic dikes have been thoroughly sampled and may provide important clues to the nature of the parental magma for the anorthosite pluton.

Most layering within the anorthosite probably originated during differential flow of the solidifying anorthosite. The foliation which locally cross-cuts the layering may reflect reoriented stresses which would have developed if a fluid interior of the anorthosite pluton continued to rise upward or expand outward deforming the more nearly solid exteriors.

This hypothesized expansion may also explain the localization of late stage mafic rocks along the contact. Expansion could result in stretching and tearing of the relatively solid exterior, permitting multiple injections of residual liquid into these zones.

INTERNAL RELATIONS

PRELIMINARY Sr-ISOTOPE DATA FOR THE KIGLAPAIT INTRUSION

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Colorado School of Mines¹

The Nain complex of Labrador is a plutonic complex of Proterozoic age (Krogh and Davis, 1973; Barton, 1974) which underlies approximately 15,000 km² of the Labrador peninsula (Wheeler, 1942; 1960). The complex is dominated by rocks of anorthositic and granitic composition (Wheeler, 1942; 1960), but minor volumes of gabbro, troctolite and diorite are also present (Fig. 1; Morse, 1969; Wiebe, 1979; Wheeler, 1968). These rocks represent the crystallization products of a wide variety of magmas, which were relatively anhydrous (e.g. Huntington, 1979) and emplaced at relatively shallow depths in the crust (e.g. Berg, 1977). A rifting environment has been proposed for the formation of the Nain complex (Berg, 1977, 1979; Emslie, 1978a, 1978b). The fact that the rocks of the Nain complex are almost completely undeformed and unmetamorphosed make this locality an ideal one for studying rocks of these types and ages.

One point which makes the Nain complex important is the fact that isotopic studies on recent volcanic rocks have indicated that an important event may have taken place in the earth's mantle between 1 and 2 Gyr ago. Sr (e.g. Brooks et al., 1976), Pb (e.g. Tatsumoto, 1978), and Nd (Zindler and Hart, 1979) isotopic data all suggest such an event. Unfortunately, unmetamorphosed, mantle-derived rocks of this age are rare. Thus, the Nain complex is also an important place to study the history of the earth's mantle during this interval. At present, isotopic data for the Nain complex are sparse, and in some cases imprecise. However, the available Sr-isotope data are tantalizing--

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Table 2. Rb, Sr, and $^{87}\text{Sr}/^{86}\text{Sr}$ data for Kiglapait samples.

<u>Sample # (PC)</u> ¹	<u>$^{87}\text{Rb}/^{86}\text{Sr}$</u>	<u>$^{87}\text{Sr}/^{86}\text{Sr}$</u>	<u>$(^{87}\text{Sr}/^{86}\text{Sr})_i$</u> ²
1. KI3109 (9)	0.007744	0.70412	0.70397 ± 8^3
2. KI3224 (21)	0.007546	0.70410	0.70395 ± 8
3. KI3034 (72)	0.01273	0.70422	0.70396 ± 8
4. KI3369 (89.3)	0.02066	0.70432	0.70390 ± 10
5. KI4142 (92.8)	0.01552	0.70477	0.70445 ± 9
6. KI4145 (95.5)	0.01440	0.70541	0.70512 ± 9
7. KI3379 (98.6)	0.02200	0.70600	0.70556 ± 10
8. KI4123 (99.4)	0.04654	0.70665	0.70572 ± 11
9. KI4119 (99.6)	0.02763	0.70638	0.70582 ± 9
10. KI4079 (99.90)	0.06297	0.70745	0.70618 ± 12
11. KI4077 (99.985)	0.3038	0.71272	0.70662 ± 46
12a. KI4086 (UBZ) ⁴	0.03791	0.70480	0.70405 ± 11
12b. KI4086-pl (UBZ)	0.004815	0.70416	0.70406 ± 8

Footnotes

¹ PC = "percent crystallized", from Morse (1980, written communication).

² Initial ratio calculated using:
 $t = 1.4 \times 10^9$ years
 $\lambda = 1.42 \times 10^{-11}$ years⁻¹

³ Uncertainties refer to the last significant figures of the initial ratios, and were calculated assuming no covariance, and using maximum uncertainties of: 0.01%, 2.0%, and 100 my. for $^{87}\text{Sr}/^{86}\text{Sr}$, $^{87}\text{Rb}/^{86}\text{Sr}$ and t , respectively. All errors are 2σ .

⁴ UBZ = Upper Border Zone.

a range of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.703–0.709 for the mafic rocks of the complex is apparent (Heath and Fairbairn, 1968; Barton, 1974; Barton and Doig, 1977; Simmons, unpublished data). Note that almost all of these numbers are higher than one would predict for the Proterozoic mantle. In addition, the large range of values suggests that Sr-isotopic data may be useful for petrogenetic studies.

One of the most important igneous bodies in the Nain Complex is the Kiglapait intrusion, a layered mafic intrusion, which contains an estimated volume of 3520 km^3 of rocks ranging in composition from troctolite to ferrosyenite (Morse, 1969; 1979b; 1979c). The superb exposures and remarkably coherent stratigraphy of the Kiglapait intrusion make it ideal for the geochemical study of magmatic fractionation processes. Previous studies have provided evidence which has been interpreted to support fractionation of a single magma (Morse, 1968; 1969; 1979b; 1979c), which was anhydrous (Huntington, 1979). Using the single magma hypothesis, a bulk composition and fractionation trends for the intrusion have been calculated, both for major elements (Morse, 1981b) and trace elements (e.g. Haskin and Morse, 1969; Morse, 1981a). However, preliminary Sr-isotope data (Barton, 1974) and the data presented here suggest that a closer look at this model is required.

Results

Table 2 gives the results for twelve wholerock samples and one plagioclase separate from the Kiglapait intrusion, analyzed by isotope dilution for Rb, Sr and $^{87}\text{Sr}/^{86}\text{Sr}$. Also given in Table 2 are initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios calculated using an age of 1.4 Gyr. The calculated initial ratios exhibit a surprisingly large range: 0.7039 to 0.7066; this range is well outside of analytical uncertainty. Only the Upper Zone samples show a range of calculated initial ratios, however; troctolites and gabbros from the Lower Zone and the Upper Border Zone have very similar calculated initial ratios: 0.70395 ± 0.00010 . The calculated initial ratios for the wholerock and plagioclase separate from Upper Border Zone sample KI 4086 agree very well.

Discussion

Figure 15 shows the data plotted on a $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ diagram. The data do not form a linear array, but rather a crude trend of increasing initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios with increasing $^{87}\text{Rb}/^{86}\text{Sr}$ ratio. Even though the

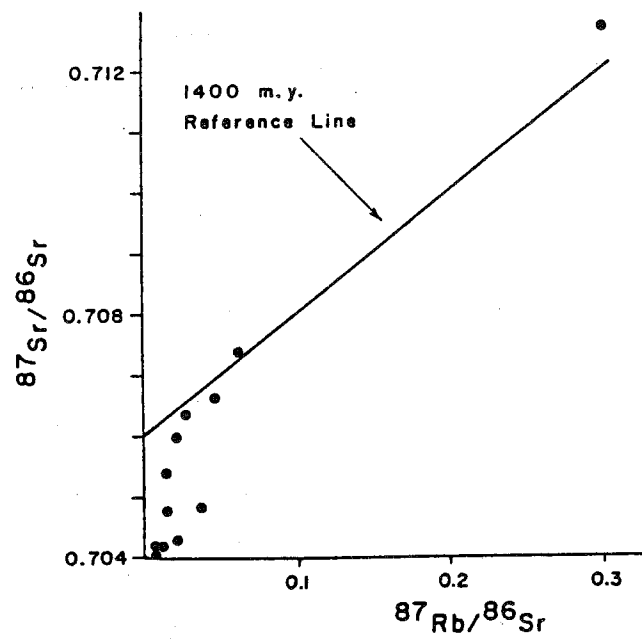


Fig. 15. $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ for whole-rock samples of the Kiglapait intrusion.

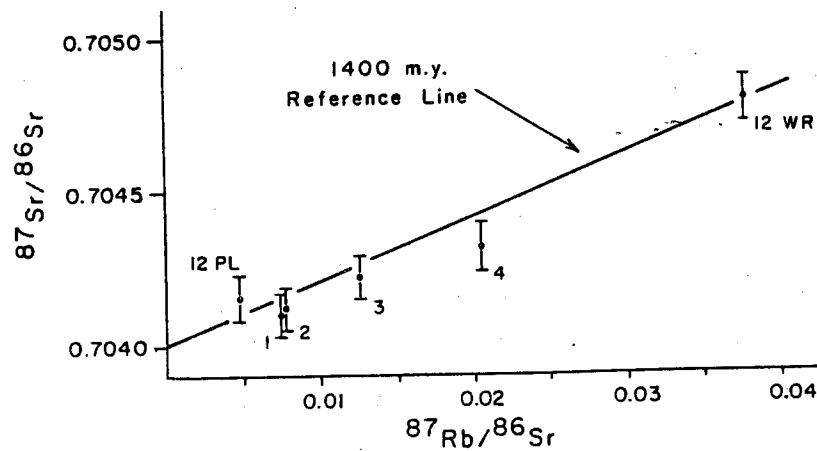


Fig. 16. $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ for samples of the Lower Zone and Upper Border Zone of the Kiglapait intrusion.

Lower Zone and Upper Border Zone samples have similar calculated initial ratios, these samples do not fall within analytical uncertainty of a 1400 m.y. reference line (Fig. 16). This suggests either that Rb and/or Sr have been severely remobilized within these samples, or that the Kiglapait intrusion crystallized from magma(s) having distinctly different $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. In light of the pristine condition of these samples and the lack of any evidence for a subsequent metamorphic event, we favor the latter interpretation.

Figure 17 shows the calculated initial ratios plotted versus the Percent Solidified (PCS) index of Morse. The data exhibit consistent initial ratios (0.70395) from the base of the intrusion to well within the Upper Zone. For PCS indices greater than 90%, there is a steady and progressive increase in the initial ratio within increasing PCS. It is interesting that this increase does not begin at the Lower-Upper Zone Boundary.

Clearly, the Sr-isotope data presented here do not support a single magma, closed-system model for the Kiglapait Intrusion. These results are best explained in one of two possible ways. First, the data could reflect progressive contamination of the Kiglapait magma by upper crustal material. If this is the case, then this process affected only the Upper Zone and more importantly, did not produce silica-oversaturated rocks. Alternatively, these results could mean that the Kiglapait Intrusion does not represent the crystallization products of a single magma, but rather that multiple injections of geochemically distinct magmas were involved. If the increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the Upper Zone is due to injection of a different magma, then the progressive nature of the increase suggests that the mixing process between the residual and new magmas was a gradual one.

While additional work is needed to distinguish the above two possibilities, data from elsewhere in the Nain Complex suggest that this is not an isolated phenomenon. Figure 18 is a histogram showing the Sr-isotope data presented above, as well as data from two other intrusive complexes: The Barth Intrusion (de Waard, 1976) and the Goodnews Complex (Wiebe, 1979b). In all cases, the highly differentiated, iron-rich rocks associated with these complexes have distinctly higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than do the troctolites or anorthositic rocks which make up the bulk of these complexes.

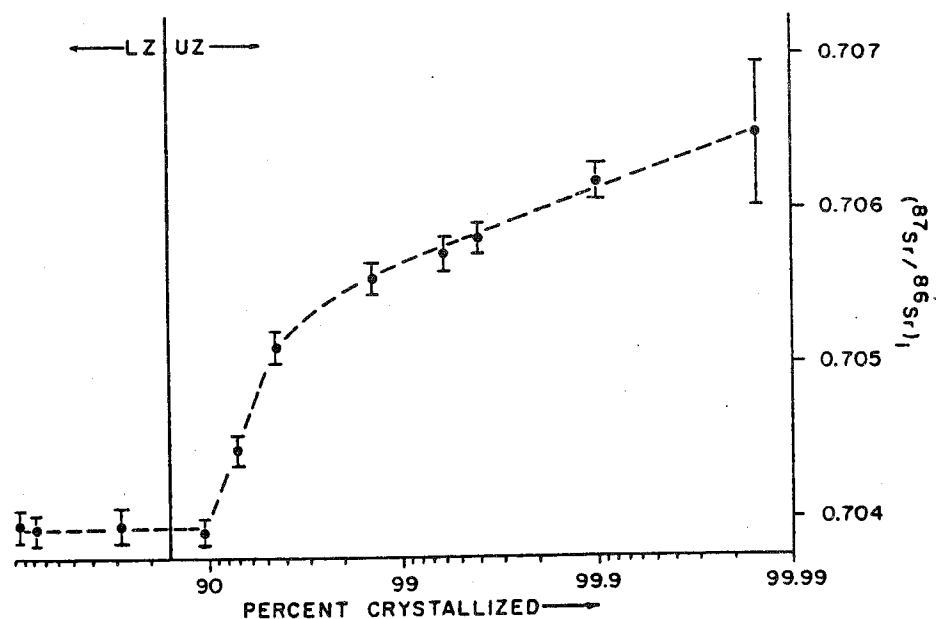


Fig. 17. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ versus the Percent Crystallized index of Morse, for whole-rock samples of the Kiglapait intrusion.

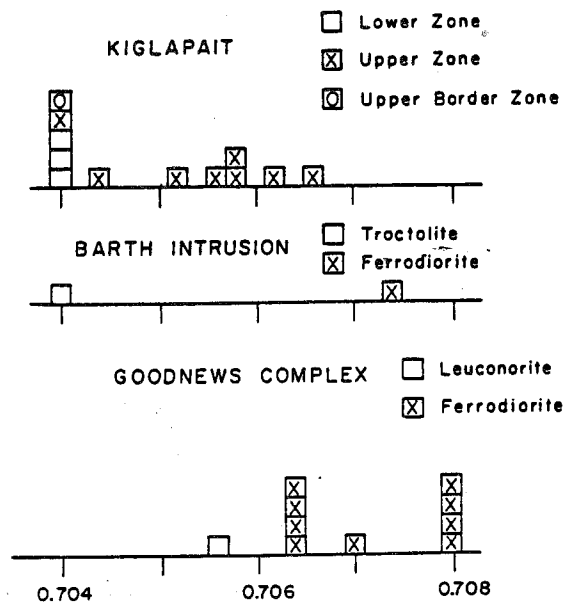


Fig. 18. Histogram showing initial $^{87}\text{Sr}/^{86}\text{Sr}$ data for three intrusive complexes of the Nain complex.

STRONG REVERSE ZONING IN PLAGIOCLASE OF
THE KIGLAPAIT INTRUSION

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University of Massachusetts¹

"If you have a rock, you have a problem."

H.S. Yoder, Jr.

Introduction

The Kiglapait intrusion is a large Precambrian layered intrusion on the coast of Labrador (Morse, 1969, 1979b, 1979c). It is notable for the unaltered nature and the completeness of its stratigraphy. Troctolites of the Lower Zone account for about 78% of the volume of the intrusion. These troctolites consist of medium-grained plagioclase and olivine showing nearly homogeneous mineral compositions as determined in grain mounts (Morse, 1978, 1979b); the rocks are thus adcumulates in the sense of Wager *et al.* (1960).

In this study, we examined polished thin sections from a suite of short drill cores collected from the 9-11 percent solidified (PCS) levels of the Lower Zone north of Slambang Bay, UTM coordinates 20V 984075, David Island Sheet of the 1:50,000 Canadian Topographic Series. The cores were taken from mafic layers (dunites) and intervening average rocks (leucotroctolites). The sections typically consist of laminated tabular plagioclase about 1 x 2 to 2 x 5 mm in cross section, having average composition An_{57} , and subhedral to granular olivine near Fo_{66} in composition. All mineral compositions were determined by electron microprobe analysis.

Sample KI 3109, from the 9 PCS level, shows a characteristic range of plagioclase texture and zoning. A compositional profile taken along the length of the largest cumulus plagioclase grain is shown in Fig. 19. The grain is nearly homogeneous, but there is a suggestion of a skeletal core flanked by a slightly more calcic zone, and a single normal-reverse oscillation occurs at either end of the grain. Similar features, much more strongly developed, were described by Maaløe (1976) from the Skaergaard

¹Authors' full addresses are given at the back of this volume.

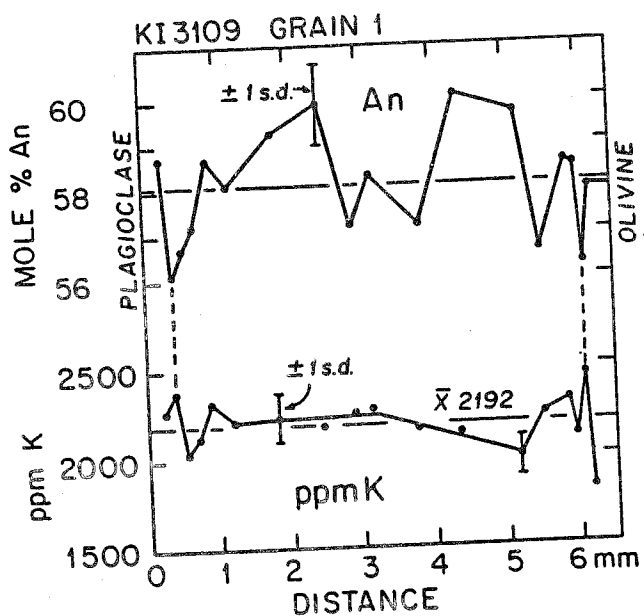


Fig. 19. Electron microprobe traverse along a large cumulus plagioclase crystal, 3109/1.

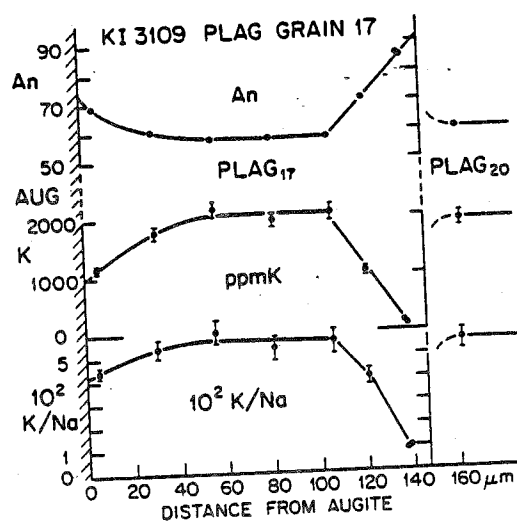


Fig. 20. Electron microprobe traverse along small zoned plagioclase grain 3109/17.

intrusion, East Greenland. Similar oscillations ending in reversals at the rims are nearly ubiquitous among the grains we have examined in six polished thin sections.

Strong Zoning

We searched the sections for areas showing minor augite rims or oxysymplectites of magnetite and hypersthene caused by the oxidation of olivine (Presnall, 1966; Morse, 1969, Plate 34; Morse, 1980), because there we expected to find normal zoning indicative of postcumulus growth from the last dregs of trapped liquid. Instead, in every case examined, we found reverse zoning near augite patches, and little if any zoning near oxysymplectites. Commonly the reverse zoning occurs in small grains or at the ends of larger grains, and shows a moderate gradation to the mid An 60's or low An 70's. Our analyzed grain 6 is characteristic. The An content rises to An 69 at one edge of the grain, where it is in sharp contact with a grain of An 57. A strong Becke line at the grain contact signals the sharp contrast in grain compositions. Although grain 6 is in contact with the calcic mineral augite, the most calcic parts of the plagioclase rim lie against other plagioclase grains, not augite. In grain 6 as elsewhere, the potassium concentration is inversely correlated with An.

The most striking example of reverse zoning at plagioclase rims is shown by grain 17, profiled in Fig. 20. The figure speaks for itself; the grain is zoned over an interval of 30 μm from An 57 to An 86 and from $K = 2200$ ppm to $K = 120$ ppm. Again, the most calcic part of the rim is in contact with another plagioclase, An 60. The K/Na ratio in grain 17 varies by a factor of 6, from 6 to 1×10^{-2} .

Other Examples

In a traverse of 38 analyses along the length of a polished thin section (sample KI 3113) paired data were obtained for plagioclase cores and rims. In 19 pairs, 14 showed reversed rims, with average $\Delta \text{An} = +1.1$ mole %, and only 5 showed normal rims, with average $\Delta \text{An} = -1.5$ mole %. Strongly reversed rims (An 56 core, An 71 rim) were also found in this section.

Speer and Ribbe (1973) reported narrow reversed rims on plagioclase, with ΔAn up to +20 mole %, from the 75 percent solidified level of the

Kiglapait intrusion. Taken together with their data, our petrographic examination of numerous thin sections from the Lower Zone and some from the Upper Zone suggests that reversed rims are ubiquitous in the intrusion.

Significance of Reverse Zoning and Inverse K versus An

A sense of the thermal history of the Kiglapait intrusion is essential to an appreciation of reversed zoning. The estimated crystallization time is about 10^6 yr (Morse, 1979b). Heat was extracted dominantly through the roof, as shown by the stratigraphy of the intrusion and the fine-grained, inverted Upper Border Zone, and so on the average, basal rocks of the Lower Zone took perhaps 10^6 yr to cool through the interval 1240 to 950°C. Because the samples examined in this study come from near a basal contact with warm anorthosite, it is likely that they experienced a long cooling history and were not quenched by heat loss at the contact. Certainly the higher-level samples in which zoning has been found required on the order of half a million years to cool below 1000°C. The characteristic cation diffusion coefficient measured for feldspars in the laboratory at 1000°C is 10^{-11} cm²/sec, and the blocking temperature for diffusion in feldspar is about 460°C (Hart, 1981). Equilibration on the scale of meters is therefore to be expected, at least for Na-K interdiffusion, in the geological time scale of the Kiglapait intrusion's cooling history. Instead, we find that it did not go to completion and probably did not occur at all.

These plagioclases do not coexist with K feldspar and are, therefore, not saturated with K. Abundant evidence exists in the literature and our collections for plagioclases with greater K content than these. The zoning in K cannot be ascribed to the constraints of the ternary feldspar solvus.

The principal immediate conclusions to be drawn, in the light of the long cooling history, are as follows:

1. Equilibrium crystallization of plagioclase did not occur despite ample opportunity.
2. Potassium diffusion did not occur in response to K concentration gradients.
3. The plagioclase compositions show no evidence for an evolved Na + K-rich residual liquid trapped in the final pore spaces of these cumulates.

These conclusions are especially significant because they bear heavily on the mode of nucleation and growth of crystals in the intrusion, and hence on the interpretation of mineral/melt distribution coefficients for the intrusion.

Origin of Strong Reverse Zoning

The reversed rims cannot be due to a subsolidus gain of Ca from augite because Al would be required also and augite is not an adequate source of Al. Moreover, the strongest reversals occur against other plagioclase grains, not against augite. No metamorphic or metasomatic origin is feasible because the features seen are local, and no source-sink minerals are locally available for $\text{CaAl} \rightleftharpoons \text{NaSi}$ exchange. The observed effects are therefore igneous.

Shimizu (1978) postulated steady-state crystallization at high supersaturation to explain the high K content of Skaergaard plagioclase. Morse (1979b) postulated delayed nucleation of plagioclase near the floor of the Kiglapait intrusion to account for accumulation and retention of plagioclase at the floor. Because the Lower Zone liquid was cosaturated with plagioclase and olivine, supersaturation and delayed nucleation of plagioclase are possible only if olivine nucleates first and drives the liquid well into the plagioclase field. Such compositional supersaturation must narrow the apparent plagioclase loop as the liquid is driven toward the rising plagioclase solidus (Morse, 1979b, Fig. 19). By the same argument, supersaturation will lead to plagioclase nuclei enriched in both Na and K.

We now suggest that the large cumulus grains typified by that shown in Fig. 19 nucleated at large supersaturation and grew isothermally by steady-state crystallization and by the adcumulus growth mechanism. Adcumulus, steady-state growth persisted as long as an interconnecting network of pore liquid could remove the latent heat of crystallization (along with NaSi in exchange for CaAl). When this network became closed, the latent heat could no longer escape and the plagioclase composition relaxed to or toward the equilibrium composition, causing reverse zoning. We therefore have the remarkable case of a cumulate completing its solidification at rising temperatures. Because the rim is more An-rich (e.g. An₇₇) than the plagioclase component of the liquid (<An₅₇), its excess An must come from the augite component of the residual liquid by such a reaction as $\text{CaAl}_2\text{SiO}_6 + \text{Mg}_2\text{Si}_2\text{O}_6 \rightleftharpoons \text{CaAl}_2\text{Si}_2\text{O}_8 + \text{Mg}_2\text{SiO}_4$ in liquid

plag ol

Conclusions

Narrow reversed rims, locally having large composition differences, are probably ubiquitous in the layered rocks of the Kiglapait intrusion. They show a strong inverse correlation of K with An, and prove that neither reactive equilibrium growth nor subsolidus cation diffusion occurred to a major degree during the late-stage solidification of these rocks, despite high temperatures that were maintained over long time periods. These observations demolish the "aquifer recharge" hypothesis considered by Morse (1981a) as a possible explanation for the high K content of Kiglapait plagioclases. That hypothesis required the influx of K to the cumulates from late-stage fractionated liquids that were postulated to have recharged the underlying cumulates from above. The residual traces of such recharging liquids would be expected to have more K than the bulk plagioclase rather than less, as observed.

The reversed rims are clearly not metamorphic in origin because no subsolidus reactants are present or accounted for. They appear instead to require steady-state crystal growth at large supersaturations, followed by relaxation of the local system toward a stable equilibrium state. The reversed rims therefore support the hypothesis of oscillatory nucleation as an explanation for the accumulation of plagioclase at the floor of the Kiglapait intrusion (Morse, 1979b).

FURTHER STUDY OF THE TIGALAK LAYERED INTRUSION

Robert A. Wiebe

Franklin and Marshall College¹Introduction

The Tigalak intrusion was first studied in detail during the 1976 field season and preliminary field results were presented in FR 1976. Laboratory work on samples collected during that season focused on problems in the northern half of the intrusion. The north central and northeastern portions appeared to consist of a relatively uncomplicated layered sequence displaying fractionation from leucocratic norites at the base to oxide-rich diorites at the highest observed levels. In contrast, the northwestern area was dominated by massive to layered diorite with many areas of hybrid rocks consisting of mixtures of diorite and granite. Because layering appeared to pass smoothly across all sections, the hybrid rocks were thought to have formed during all stages of fractionation recorded by the coherent layered sequence. Petrographic and chemical studies demonstrated that the dioritic rocks in the northwestern area show little variation and are all similar to the most fractionated rocks of the coherent layered sequence.

This field study was undertaken in order to clarify the structural relations between the layered diorite of the north and northeast and the hybrid diorites in the northwestern part of the Tigalak intrusion. Field relations observed on several new traverses suggest that the northwestern diorites may have formed from a late pulse of fractionated diorite which was injected laterally out from the main layered body while it was undergoing fractionation. The new mapping also indicates that the leucotroctolites north of the Tigalak intrusion lie above the layered diorites, not below as previously thought. A major remnant of the roof rocks to the Tigalak layered intrusion also occurs on the Tigalak peninsula. Fig. 21 is the revised map of the northern part of the Tigalak intrusion. A general description of all the units is given in FR 1976.

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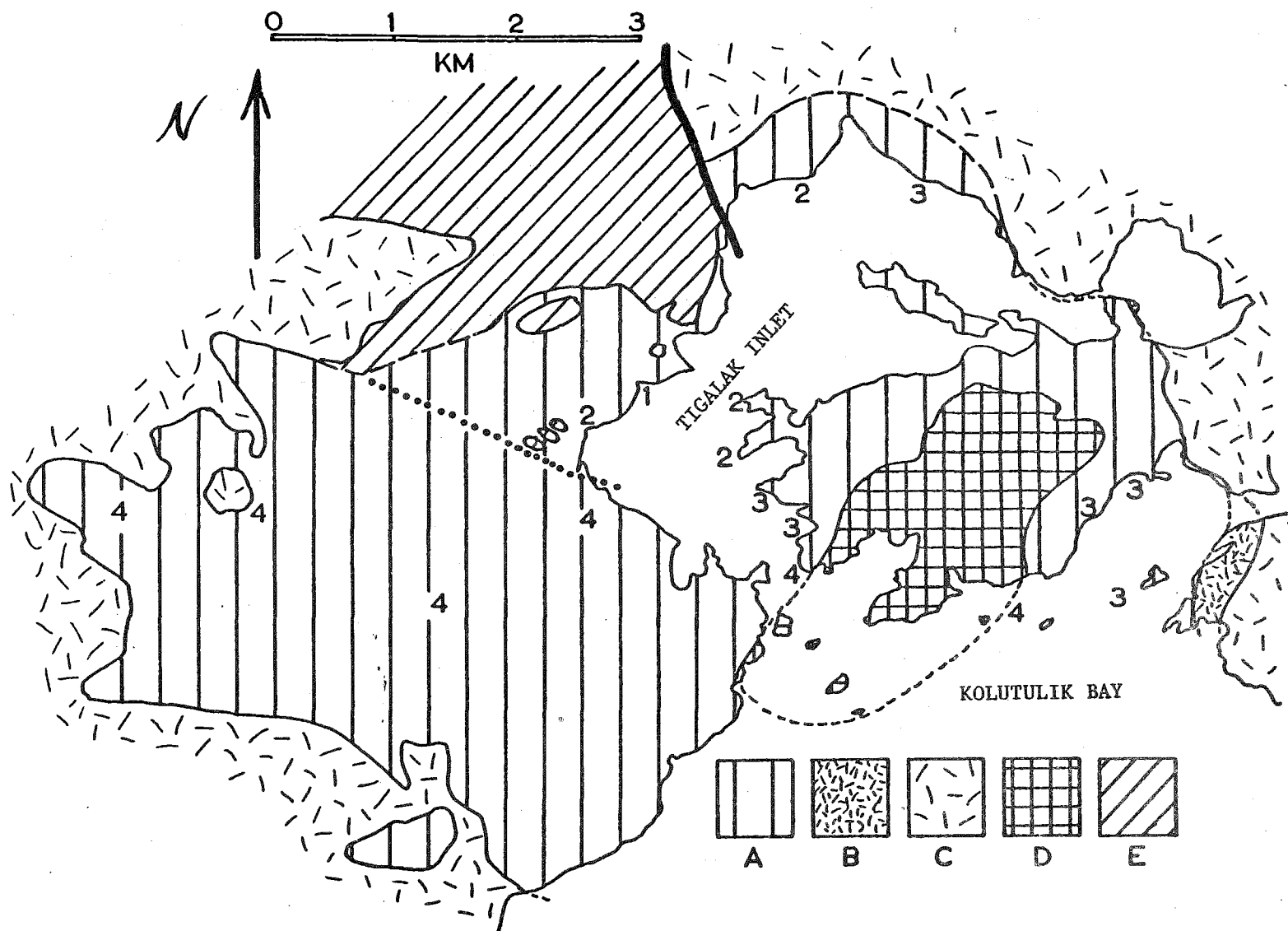


Fig. 21. Revised geologic map of the Tigalak layered intrusion (A). Norite (B); Leuconorite (C); Layered troctolite (D), Leucotroctolite (E). The Tigalak intrusion is a sheet-like mass of norite to ferrodiorite resting on the leuconorite and lying beneath a roof of leucotroctolite and layered troctolite. Mineral zones: 1. Orthopyroxene, 2. Pigeonite, 3. Augite, 4. Apatite. Dotted line

Tigalak Layered Intrusion

New data presented here relate to the nature of the external contacts of the intrusion. The hillside northwest of Tigalak Inlet appears to approximate the contact between the diorite and leucotroctolite. Topographic irregularities clearly indicate that the leucotroctolite lies above the Tigalak intrusion and the contact dips gently to the southeast, apparently truncating steeper layering in the overlying leucotroctolite. Leucotroctolite forms caps on many small hills north of the dotted line in Fig. 21. The diorite immediately below these cappings of leucotroctolite is medium-grained and leucocratic (CI = 25) with orthopyroxene as the major mafic phase and with oxide minerals essentially absent. South of the dotted line both leucotroctolite and medium-grained leucocratic diorite appear to be absent. On one traverse, the first outcrops south of the dotted line consists of very fine-grained ferrodiorite. Just north of the dotted line similar fine-grained ferrodiorite occurs in sharp contact with and beneath leucotroctolite. This oxide-rich ferrodiorite appears to be compositionally similar to the medium-grained ferrodiorites which dominate the western area.

Layered Troctolite

Strongly layered troctolite occurs as a partial roof to the Tigalak intrusion and similar troctolite occurs as rotated inclusions in the underlying diorite. Layering in the troctolite roof trends roughly N70E on average and dips between 20° and 50° SE. This attitude is comparable to that in the leucotroctolite unit which lies above the northern exposed edge of the Tigalak body, and it appears likely that both troctolites are structurally related to each other. The contact between the troctolite and underlying diorite appears to be irregular but approximately horizontal. Layering in the troctolite must be truncated by the underlying diorite. The area mapped as troctolite includes a central coherent slab of troctolite and an outer margin which is dominated by large rotated angular blocks of troctolite in a diorite matrix.

Layers in the troctolite unit vary widely in texture, mode and thickness. Channel scours are well developed locally and modally graded layers are common. Plagioclase in many of these rocks occurs as thin tablets about 1 cm in length; it defines a strong lamination in some layers, while in others it is randomly

oriented. Olivine is uniformly fine-grained and equant. CI ranges between about 20 and 80.

Pyroxene and oxide minerals are locally abundant in the southernmost (stratigraphically highest) exposures. Comb-layering of plagioclase is prominent in some layers. Some curved, branching crystals are 1-2 mm thick and up to 10-15 cm long.

Compositional Variation in the Tigalak Intrusion

Petrographic studies of material collected in 1976 provided a basis for understanding the relations between the layered sequence and the hybrid rocks of the Tigalak intrusion.

A suite of samples from the northern layered sequence defines a typical sequence of fractionation ranging from leucocratic norite at the base to oxide-rich ferrodiorite at the top. (Contrary to earlier field descriptions, olivine is absent from the lower layered rocks and orthopyroxene, not augite, is the dominant mafic mineral). Plagioclase is a cumulus phase throughout the section. Orthopyroxene is replaced upward by inverted pigeonite and augite. Ilmenite and magnetite appear to become cumulus phases at approximately the same level. Apatite first appears as a cumulus phase at higher levels. This mineralogical variation can be simplified by defining four numbered zones. The first, or least fractionated zone, is characterized by cumulus orthopyroxene, the second zone by cumulus inverted pigeonite, and the third zone by a dominance of cumulus augite. The fourth, or most fractionated zone, is characterized by the presence of cumulus apatite. All samples of apparent cumulus origin collected in 1976 were classified according to these zones. The zone number of each sample is plotted on Fig. 21. A consistent sense of inward and upward fractionation is apparent in the northern and eastern portions of the intrusion. The western lobe, however, contains only the most fractionated rocks, even at levels near the outer basal contact.

Discussion

The compositional data strongly suggest that the western lobe of the intrusion developed as a separate injection of more fractionated dioritic liquid. Granitic magma must have gained access at the same time to the

newly established magma chamber in order to effect the observed hybridization. The source of the granite and the reason for the timing of its emplacement are unknown.

The composition of diorite in the western lobe is similar to that of diorite at the highest levels of the layered sections. It is possible that the western lobe was formed by lateral emplacement of residual magma from the layered section. The shift of this magma may have been brought about by further injections of magma (dioritic or granitic) into the layered intrusion or possibly by partial collapse of the roof of the intrusion.

Acknowledgments

Petrographic studies of samples from the 1976 field season were carried out by Allan Kolker and Tom Wild at Franklin and Marshall College.

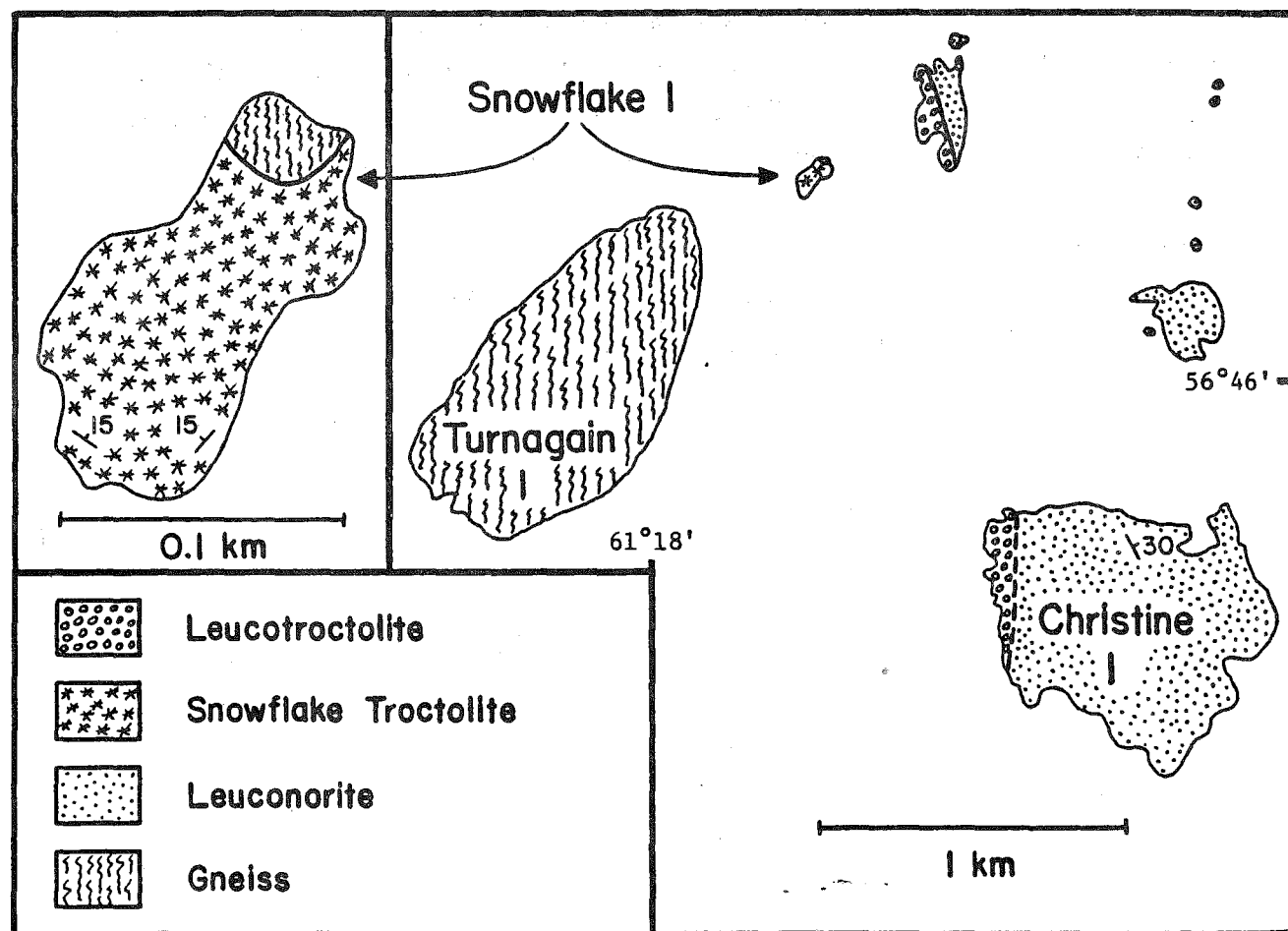


Fig. 22. Geologic map of Snowflake Island and vicinity, located east of Black Island harbour.

TROCTOLITIC ROCKS IN THE NAIN COMPLEX

J.H. Berg and M.S. Pencak
Northern Illinois University¹

Introduction

Recent work on the Snowflake Troctolite Zone (SFT) in the Hettasch intrusion (Berg, 1980), showing that it probably crystallized from a magma more mafic than basalt, has resulted in the search for other examples of this rock type and further elucidation of those previously known. Rocks that appear to have at least some similarity to the SFT have now been found as a dike cutting anorthosite just west of Second Rattle in Port Manvers Run, as xenoliths in and roof-rock to the Tigalak diorite intrusion, as part of a small pluton cutting basement gneisses on Snowflake Island (the small, westernmost island of the Vernon Island group), along the contact of the Hettasch intrusion with basement gneisses near Hettasch Lake, and possibly in several other places including a dike cutting anorthosite west of Port Manvers Run near Webb Bay and a dike along the northwest contact of the Kiglapait layered intrusion.

The features that seem to be characteristic of these occurrences are a mineralogy strongly dominated by olivine and plagioclase, a somewhat melanocratic nature (commonly 50-70% olivine), a typically well-developed layering, and supersaturation textures such as extremely flattened plagioclase crystals, comb layering, skeletal crystals, and large spherulites (snowflakes). In contrast, most olivine-plagioclase rocks in the Nain complex are leucotroctolites, rarely containing more than 35% olivine except in occasional mafic layers.

Snowflake Troctolite Zone (SFT) of the Hettasch Intrusion

The SFT was revisited for the purpose of resampling and closer study. Although much of it was covered by unusually large snowbanks, a number of new observations were made. Elegant comb-layered olivine was discovered in one place sandwiched between two plagioclase comb layers. The comb-

¹Authors' full addresses are given at the back of this volume.

layered olivine crystals are 4-6 cm in length and exhibit rather delicate branching. Unlike comb-layered plagioclase crystals, they do not appear to be curved.

Observations on comb-layered plagioclase showed it not to be curved always in the same direction, as had previously been thought. Layers that vaguely resemble comb layers display domains of parallel laths of plagioclase, the various domains showing almost every possible orientation. These domains of parallel plagioclase laths apparently did not nucleate at a magma-cumulate interface and therefore are quite different from comb layers. A tentative conclusion is that the texture of these layers is more analogous to the spinifex texture of komatiites.

In terms of relative age relations, it was observed that the SFT is clearly intrusive into the leucotroctolite below it and is intruded by the leucotroctolite above it.

Second Rattle Di

The Second Rattle dike is a sub-horizontal and tabular intrusive body of troctolite that cuts massive, almost-pure anorthosite in the Anorthosite Block Zone west of Second Rattle (FR 1973, p. 110). Both rock units in turn appear to be intruded by small dikes or veins of massive leucotroctolite that is believed to be related to the Hettasch intrusion. The dike was discussed in FR 1973, p. 115, and consists of extremely well-layered to nearly massive troctolite. A pervasive but very incomplete serpentinization of the olivine gives the mafic portions of the rock a very dark, blackish appearance. Portions of the dike show 1-2 cm layering that is very regular both laterally and vertically. A series of en echelon channel scours is visible (FR 1973) and plagioclase spherulites (snowflakes) up to 5 cm in diameter are present, especially toward the top. Some of the plagioclase shows spinifex-like texture (domains of parallel-oriented laths), and most of the plagioclase exhibits rather exaggerated flattening. Comb layering, however, has not been observed.

Previously it was believed that this dike was associated with the Hettasch intrusion (Berg, FR 1973). Discovery of the minor amounts of intrusive leucotroctolite suggests that the troctolite dike predated the Hettasch intrusion and was intruded by it along fractures during what may

have been a catastrophic collapse of the roof of the Hettasch intrusion (Berg, FR 1973).

Kolutulik Troctolite

The troctolite at the head of Kolutulik Bay, near Tigalak Inlet, was first described by Wiebe (FR 1976) as an area of abundant layered-troctolite xenoliths in the upper part of the dioritic Tigalak intrusion. As delineated this summer, the areal extent of the troctolite and the integrity of its layering are so great that it seems highly probable that this unit is actually a coherent section of the roof of the Tigalak intrusion (Wiebe, this Report).

Spectacular layering and channel scours are present in the troctolite. Plagioclase shows extreme flattening and is typically oriented parallel to the layering. However, it is not uncommon to find comb-layering on a scale ranging up to 8-10 cm in plagioclase length. Plagioclase spherulites (snowflakes), although present, are very rare. Comb-layering, clinopyroxene, and coarse, pegmatitic patches become much more abundant upward in the troctolite. The uppermost exposed rocks of the troctolite are quite rich in clinopyroxene and pegmatitic patches and the pegmatitic patches themselves are especially rich in clinopyroxene.

One feature of these rocks which remains particularly enigmatic involves shingles of anorthosite or leucotroctolite, typically 2 x 5 cm in section. These shingles seemed all too commonly to be oriented either perpendicular to layering or to be concentrated (radially??) around the margins of pegmatitic patches. In some outcrops one could see a progression from layers of anorthosite or leucotroctolite (2 cm thick) to broken-up en echelon shingles of anorthosite or leucotroctolite (2 x 15 cm), but whether this progression affords a real opportunity to explain the shingles escapes us for the present.

Snowflake Island Troctolite

On the smallest and westernmost island of the Vernon group is exposed a contact between a troctolitic pluton and basement gneisses (Fig. 22). This troctolite, with its spectacular snowflake textures, was briefly mentioned by Wheeler (FR 1972, p. 34).

At the contact a dark, coarse hybrid zone about 10 cm thick separates country rock from the troctolite. Inside this there is a zone of variable thickness (0-15 cm) which is fine-grained and massive. Locally within this zone minor comb layering or coarse-grained patches with snowflake texture are present. Stratigraphically above this are 2-5 repeated layers, the thickest as much as 15 cm thick. Each layer consists of a massive base from which coarse comb-layered olivine and plagioclase extend upward. Both minerals branch and become more acicular upward; however the olivine is particularly branched and feathery. At the top of each layer there is locally a coarser, snowflake-textured portion which is commonly enriched in pyroxene, oxide minerals and biotite. Above the layered zone, coarse-grained and more massive snowflake-textured troctolite is present. At least one xenolith of country rock was found with comb layers rimming it, much like an orbicular structure.

The snowflake troctolite has some vague layering, and although much of the troctolite is massive, a weak orientation of plagioclase tablets was observed locally. In some cases the plagioclase orientation appeared to be parallel to the rare layering, but more commonly it appeared to be perpendicular to it.

For the most part, the snowflake troctolite is massive and quite homogeneous. The snowflakes are very radial and spherical, and show little variation in size, being typically 3-4 cm in diameter. Where hydrothermal alteration along veins has accentuated color differences and weathering has etched the surface, delicate oscillatory zoning of plagioclase in the snowflakes can be observed.

Western Contact of Hettasch Intrusion

Along the western contact of the Hettasch intrusion, just south of Hettasch Lake (FR 1972, p. 53), troctolitic rocks are intruded by leucotroctolite which is apparently part of the Hettasch intrusion. The troctolites are characterized by their high color index (50-70) and the unusual habits of the olivine. The olivine ranges from perfectly euhedral to slightly skeletal to harrisitic. Although comb layering of plagioclase and olivine is present, it is not completely clear whether it is part of the troctolite or leucotroctolite. Even though the leucotroctolite

invades the troctolite, sub-spherical blobs of leucotroctolite occur in the troctolite. It is hoped that petrographic work on these rocks will help to remove some of the riddles.

Miscellaneous Occurrences

West of Port Manvers Run, Berg (FR 1973, p. 110) has mapped an isolated mass of troctolite within foliated anorthosite. This mass was originally thought to be part of the Hettasch intrusion, but we now conclude, based on its high color index (60-80) and the flattened habit of its plagioclase, that it is more akin to the snowflake troctolites described above.

Finally, along the northwest contact of the Kiglapait intrusion (Fig. 7), a narrow zone of troctolite occurs between the porphyritic chilled margin of the intrusion and cordierite-bearing pyroxene granulites of the country rock. Unusual textures have not been documented here, and this troctolite may or may not be related to the snowflake troctolites.

Discussion

A feature common to all occurrences of snowflake troctolite in the Nain area is their high color index, which commonly exceeds 50. Olivine invariably precedes plagioclase in the crystallization sequence observed in the field, and the magmas responsible for these units must have been decidedly olivine rich with respect to the normal olivine-plagioclase cotectic. A closer characterization of this high-temperature melatrolitic magma type will be permitted by the petrographic and geochemical studies planned for these rocks. In particular, the homogeneous snowflake troctolite on Snowflake Island is expected to be an excellent candidate for reflecting the original magma composition--if indeed there was a common parent magma to these rocks.

Table 3. Comparison of basaltic dikes from Nain with other Proterozoic dikes.

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>
SiO ₂	44.58	45.65	48.7	46.00
TiO ₂	3.34	2.64	2.1	2.62
Al ₂ O ₃	16.64	16.75	16.4	17.04
FeO*	15.29	15.13	13.6	13.60
MnO	0.20	0.20	0.21	0.20
MgO	5.76	6.14	6.5	7.65
CaO	8.20	8.93	8.7	7.81
Na ₂ O	3.48	3.37	2.9	3.25
K ₂ O	1.28	0.83	0.6	1.06
P ₂ O ₅	<u>1.23</u>	<u>0.36</u>	<u>0.29</u>	<u>0.77</u>
Total	100.00	100.00	100.00	100.00

All analyses recalculated volatile-free.
FeO* is total Fe as FeO*.

1. Average of 7 analyses of Group 1 basaltic dikes from Nain. This type tends mainly NNW. Nodules and xenocrysts are known to occur within these dikes.
2. Average of 7 analyses of Group 2 basaltic dikes from Nain. This type trends mainly NE. No nodules or xenocrysts have been located in these dikes.
3. Average of 9 Harp dike samples. (Meyers and Emslie, 1977, Table 3, Column 4).
4. Chill zone of giant gabbro dikes (Larsen, 1977, Table 4, Column 4).

Table 4. Comparison of high-Al orthopyroxene xenocrysts with megacrysts from anorthosite plutons.

	<u>1</u>	<u>2</u>	<u>3</u>
SiO ₂	50.94	50.36	49.19
TiO ₂	0.52	0.18	0.62
Al ₂ O ₃	6.06	8.38	7.28
Cr ₂ O ₃	0.12	0.23	0.08
FeO _T	16.00	11.63	16.45
MnO	0.24	0.21	0.28
MgO	24.37	25.56	22.48
CaO	1.76	1.99	2.15
Na ₂ O	<u>0.08</u>	<u>0.22</u>	<u>0.29</u>
Total	100.10	98.76	98.82
Fs	25.9	19.5	27.8
En	70.4	76.3	67.6
Wo	3.7	4.2	4.6

1. Typical high-Al orthopyroxene xenocryst
2. Emslie (1975) Table 2, Anal. 1
3. Emslie (1975) Table 2, Anal. 3

PROTEROZOIC BASALTIC DIKES AND THEIR INCLUSIONS

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The Nain complex is transected by many basaltic dikes which are similar chemically and petrographically to Proterozoic dikes that cut the Harp Lake anorthosite (Meyers and Emslie, 1977). These dikes have relatively high Fe/(Fe+Mg) and are alkaline in character. They are characterized by groundmass olivine and titanite, and most are slightly Ne-normative (see Table 3 for chemical compositions of typical Nain dikes and comparisons with other Proterozoic dikes). Meyers and Emslie (1977) noted the similarities between the Harp Lake dikes and the Nain dikes as described by Upton (FR 1973). It is likely that the Nain dikes are late Proterozoic and similar in age to the Harp dikes (probably about 1350 m.y.)

The dikes in Nain have received little attention except for a reconnaissance petrographic and petrochemical study by Upton (FR 1973). I have focused on two distinct groups of dikes: (1) a NNW- to NNE- trending set relatively rich in TiO_2 , K_2O and P_2O_5 and (2) a NE-trending set with distinctly lower values for these elements--particularly P_2O_5 . Dikes of the NE set are larger and more abundant than those of the first set.

At the end of the 1976 field season one dike belonging to the first group was found to contain abundant nodules and xenocrysts. Scarce xenocrysts and xenoliths have been noted this year in a few other dikes with similar orientation. All of the inclusions studied to date appear to be of crustal origin and none of the mineral assemblages is capable of providing reliable estimates of pressure. None of the inclusions resemble rocks or have textures characteristic of the exposed level of the Nain complex. Mineral compositions suggest that some of the xenocrysts and nodules with cumulus textures are genetically related to the anorthosites.

The most common xenocryst is homogeneous high-Al orthopyroxene. The xenocrysts are irregularly shaped, angular, and up to 5 cm in diameter. They appear to represent fragments of larger crystals. Orthopyroxene of

¹ Authors' full addresses are given at the back of this volume.

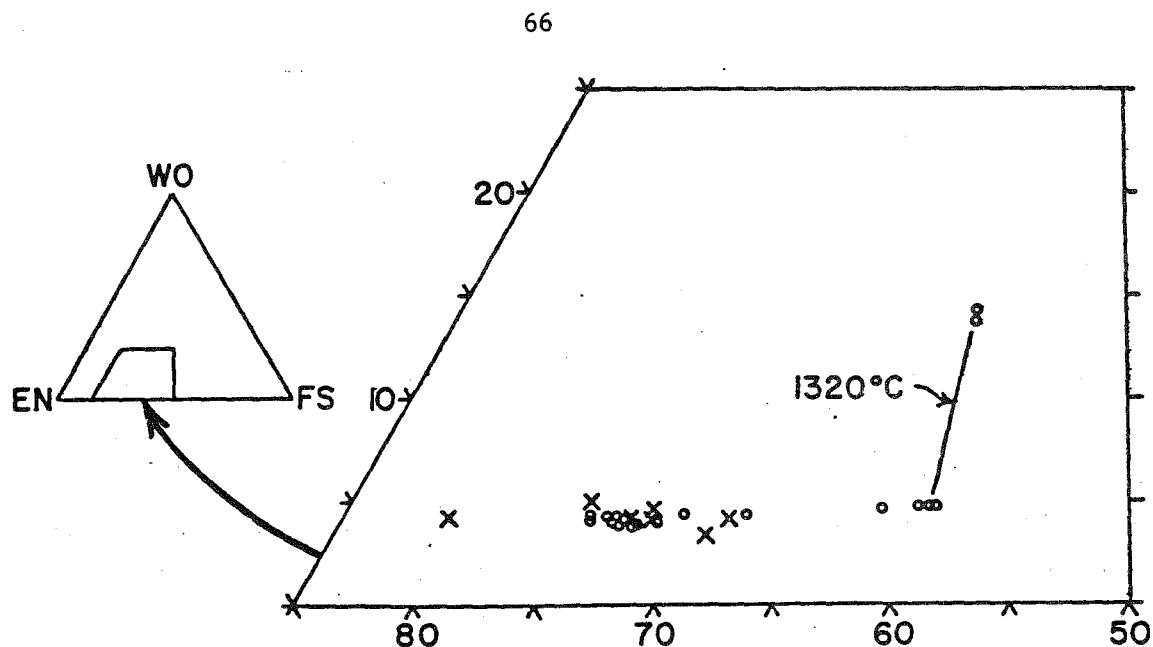


Fig. 23. Compositions of homogeneous orthopyroxene from xenocrysts and noritic nodules in basalt dikes (circles) compared with bulk analyses of partially exsolved pyroxene megacrysts (crosses) from different anorthosite complexes (Emslie, 1975). Temperature from Wells's (1977) 2-pyroxene geothermometer.

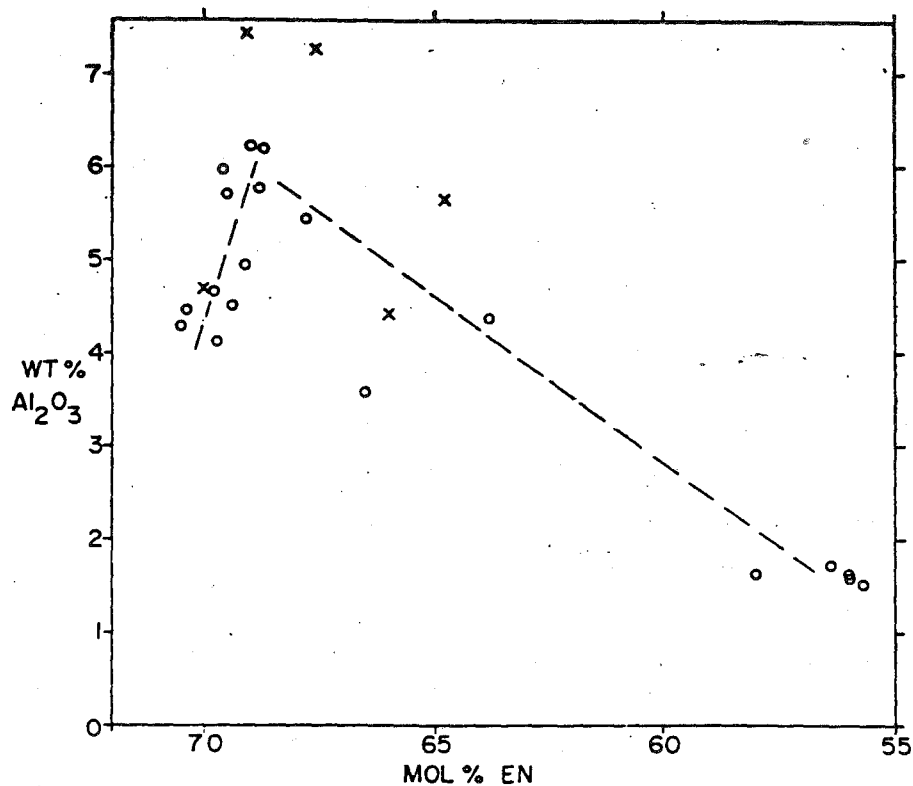


Fig. 24. Variation of Al_2O_3 with En of orthopyroxene from nodules and xenocrysts. Symbols as in Fig. 23.

similar composition also occurs in equilibrium with plagioclase in nodules up to 20 cm in diameter. Although most of this orthopyroxene is interstitial to euhedral plagioclase, irregular patches of orthopyroxene up to 6 cm wide do occur irregularly within the nodules. The most Fe-rich orthopyroxene (En_{56}) occurs in equilibrium with coarse equant crystals of low-Ca clinopyroxene and plagioclase and in association with interstitial Ti-rich oxide minerals. Application of Wells' (1977) 2-pyroxene geothermometer indicates a temperature of 1320°C (Fig. 23). The abundant Al-rich orthopyroxenes with higher $\text{Mg}/(\text{Mg}+\text{Fe})$ presumably crystallized at higher temperatures. The interstitial oxide minerals in this nodule consist of alternating coarse lamellae of titanomagnetite and ilmenite. The texture suggests oxyexsolution of an originally homogeneous titanomagnetite. The titanomagnetite lamellae are Usp_{79} . The coexisting lamellae indicate a subsolidus equilibration temperature of 1100°C and f_{O_2} somewhat lower than the QFM buffer (Lindsley, 1977). This equilibration temperature is comparable to that for interstitial oxide minerals in the matrix of the basalt host, and is remarkably high for Fe-Ti oxide minerals in an intrusive igneous rock.

Pyroxene compositions are compared with megacrysts from anorthositic plutons in Fig. 23 and Table 4. There is a considerable range in En and variation in Al_2O_3 (Fig. 24). For one suite of pyroxenes, there is a steep increase in Al_2O_3 as En decreases from 71 to 69. At lower En, Al_2O_3 appears to decrease gradually from over 6% at En_{69} to less than 2% at En_{56} . The Al-rich orthopyroxenes are similar in bulk composition to orthopyroxene megacrysts with plagioclase lamellae which have been reported from many anorthosite complexes (Emslie, 1975).

Some tentative conclusions can be drawn from the available data. Mineral assemblages suggest that crustal temperatures were unusually high following the emplacement of the anorthosites. Two-pyroxene equilibration temperatures of roughly 1300°C are recorded for assemblages which could hardly have been stable at pressures greater than 20 kb. If the basaltic dikes in the Nain complex are analogous to the Harp dikes (Meyers & Emslie, 1977), they are probably less than 100 m.y. younger than the anorthosites they intrude.

The lower crust may contain cumulate material related to the generation of anorthositic magmas. High-Al orthopyroxene may be the liquidus phase of a basaltic magma undergoing fractional crystallization in the lower crust or uppermost mantle. Early separation of this mineral could be responsible for

causing a "normal" basaltic magma to fractionate towards a high-Al composition. The steep initial increase in Al_2O_3 of the pyroxene may therefore reflect an increase of Al_2O_3 in the related magma. Nodules of high-Al orthopyroxene and intermediate plagioclase (An_{60-50}) may record a cotectic similar to that determined experimentally by Emslie (1971) for the system $\text{An}_{60}\text{-En-Di}$ at 15 Kb. The high magmatic temperatures indicated by coexisting pyroxenes approach the high temperatures needed for the existence of a feldspar-rich liquid.

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NAIN BIBLIOGRAPHY

We reprint in the following pages the complete bibliography of the Nain Anorthosite Project, bringing it up to date through 1980. Included are papers which make extensive use of samples or data gathered in the course of the Project. The Bibliography is organized in two sections: Papers, theses, reports; and Abstracts. The count at the close of 1980 is 14 theses, 43 papers, 6 reports and 31 abstracts.

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OPERATIONS

S. A. Morse

NARRATIVE REPORT

This Project is supported by a field vehicle, R/V *Pitsiulak*, which provides safe transport of personnel and supplies and at the same time offers laboratory facilities and such support services as outbound motor repairs and custom diamond drilling (provided the outcrop to be sampled is pleasingly located in the eyes of the drill crew). From 1971 to 1976, it was our practice to store the vessel in Labrador so as to have access to Nain early each season, inshore of the spring ice pack. With deteriorating conditions for storage in Labrador and the decision of the Newfoundland Fisheries Department to concentrate its Marine Centers on the island of Newfoundland, we joined the fishing fleet in moving out of Labrador to Flowers Cove, Strait of Belle Isle, in 1978. The need for enclosed working space over the vessel prompted a further move to Port Saunders, south of the Strait of Belle Isle, in the spring of 1980. This move was kindly attended to by Max H. Tiller, a Labrador contractor who is a long-time friend of this Project and who rarely overlooks a chance to go to sea. The advantages of the new location include ready accessibility by ground transportation, a superbly maintained storage lot and repair shed, and the proximity of shops and services. The potential drawback is, of course, the possibility that access to the field area will sooner or later be delayed by pack ice; but this is a risk worth taking for the security of the vessel and the chance that we may be lucky in working along shore inside the ice if it does arrive in force after breakup.

We were fortunate in 1980 in securing the expert services of D.C. Nutt, Jr., and Matthew Cole, assisted by T.A. McKean, for major repairs to the vessel. These were extensive, as indicated below. We were also fortunate to have the services of an experienced transit crew consisting of C.W. Williamson, A.J. Morse, and S.F. Morse in running north to the field area and breaking in new personnel, which included the accomplished S.P. Cobb as our new Executive Officer.

Our operations began on 1 July with the arrival of field personnel by

scheduled air service to the Nain area, and simultaneous departure of the vessel from Port Saunders. While the field crew began work near Nain, the vessel proceeded northward, running day and night to Postville, where ship-board lockers and dory were recovered from storage. From Postville north we were able to use our own previous sounding tracks to run inside. A damaged hydraulic steering line gave way on 4 July and necessitated manual steering to a temporary anchorage, where repairs were effected. As inevitably seems to happen, the difficult run through Porcupine Rattle near Davis Inlet came at moonless midnight. The vessel arrived at Khaukh Harbour, the site of the Project warehouse, at 0650 h. on 5 July, and soon thereafter began operations. The vessel worked almost steadily in the direct support of research and conferences until 24 July, when she began the normal cycle of resupplying and relocating field crews.

On 6 August air transport was secured for Berg's party to move to Hettasch Lake; the party was successfully retrieved after some remarkable piloting among squalls on 12 August.

Our first codfish in many years of dearth were caught on 13 August, at the proper insistence of the Cook. Formerly a staple in Labrador, codfish are just now moving in to fill the niche abandoned by the previous stock, which may be extinct.

Field crews were retrieved on 15 August and they departed by scheduled air service from Nain on 16 August, along with a chartered Beaver aircraft with a load of 1275 lb. of rock samples. *Pitsiulak* left at 2025 h. from Nain after a splendid farewell dinner ashore, and arrived at Port Saunders on 20 August after holding twice for weather, at Makkovik Bay and Grady.

TOPICAL SUMMARIES

Ice

No pack ice was encountered, for the second time in the history of this project. The 1976 season was also essentially ice-free.

Weather

As in Great Britain and elsewhere, the summer of 1980 was wet and cool in the Nain area. These words mean rain and drizzle with temperatures near 5°C. In 43 days of record, 10.5 were bad enough to inhibit work, for an average of 1 bad day in every 4 days; this is worse than the "normal" ratio of 1 in 5, but not as bad as the 1 in 2.5 or 3 that afflicted some months in prior seasons. There were no great gales, and only one northeaster. Most bad days had one or more stretches lacking heavy rain, and a great deal of work was done in such periods despite discomfort. The weather improved all during August, and was at its peak after we had left the field area to return to early classes.

Vessel Maintenance

Time, weather, and lay-up had brought about severe leakage through the cabin top of R/V *Pitsiulak* by the spring of 1980. The entire roof and overhead insulation were found to be saturated with water, and rot-promoting fresh water was prevalent in the interior of the vessel. A repair crew led by an experienced shipbuilder, D.C. Nutt, Jr., worked on her for more than two weeks in June. She was uncovered, dried out in the Marine Center repair shed at Port Saunders, and her roof was replaced and fiberglassed. Pieces of her deck plank were replaced, life rail stanchions refastened, the cabin top strengthened below the mast step, the rusted-out cookstove was replaced, the main engine injectors rebuilt, and many other major and minor repairs or improvements were effected. Subsequent use proved the vessel to be drier than ever inside, a condition giving her a new lease on life. Vigorous attention to maintenance during the summer also contributed greatly to her seaworthiness and potential longevity.

Communication

Radio conditions were generally good throughout the summer.

Health

One field assistant had the misfortune to aggravate an old ankle injury, and hence was unable to complete the season. No other unusual problems related to health or fitness arose.

Wintering

R/V *Pitsiulak* is now stored on land at the Fisheries Marine Center, Port Saunders, Newfoundland, where she was refitted in June, 1980. The new quarters are superbly suited to rapid (15-minute) hauling and launching, and provide much-needed protection, electric power, and work spaces for repairs.

SUMMARY OF OPERATIONS

The 1980 working season lasted 43 days, beginning on 4 July and ending on 16 August in time to resume college classes. A field conference was held 10-13 July, and there were several periods of collaborative research involving two or more investigators. Thirty-two drums of samples weighing about 800 kg were processed by the staff and shipped by air freight to destination in the U.S. The calendar below summarizes the main events.

CALENDAR

1977-78	Vessel used by Smithsonian Institution for Archaeological Survey of Northern Labrador and <u>Ungava</u> (W. Fitzhugh in charge).
1978-80	Vessel stored at Flowers Cove, Nfld., on parking lot.
June, 1980	Vessel moved to Port Saunders, Nfld., Marine Center.
July 1	Vessel left Port Saunders for Labrador.
3	Arrived Postville.
4	Left Postville in afternoon. Field crew at work.
5	Arrived Khaukh Hr. 0650. 20 hr. from Postville.
7	Refitting. Field crews to Kolutulik B.
9	Transit crew out to U.S.
10-13	Field conference, Tigalak to Slambang B.
14	Snyder Bay crews installed.
17	Resupply Snyder B., took Smithsonian speedboat to Kiglapait Hr.
18-19	Field work -- Vernon I. area.
20-21	Field work -- Uivakh - Tunungayualok area
22	Coordinator departed field area.
24	Berg crew to Osumilite L. area.
25	Resupply Snyder B.
31	Berg crew out to salt water again.

August 1 NE gale.
2-3 Field work, Port Manvers Run to Snyder B.
5 Coordinator returned to field area. Nolan, Wild out.
6 Berg crew to Hettasch Lake via Beaver aircraft.
7-9 Field work, Higher Bight.
10 Wiebe crew ashore Paul I. with Richards, asst.
12 Berg crew returned from H. Lake via Otter aircraft.
13 Retrieved Snyder B. crew.
14 Kiglapait Field Conference -- UBZ.
16 Field crew out. Vessel underway south 2025 hr.
17 Sheltered at Big I. near Makkovik.
18 Big I. to Grady.
19-20 Grady to Pt. Saunders.
21 Vessel laid up for winter on parking lot.

PERSONNEL

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