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THE GEOLOGY OF THE MICHIKAMAU INTRUSION, LABRADOR (13L, 23I)

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R. F. Emslie

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ABSTRACT

The Michikamau Intrusion underlies about 2,000 square kilometres in west-central Labrador and is one of a group of large anorthositic plutons in the Nain Province. Description of the intrusion and its setting is based chiefly on field observations and is meant to form a framework for future petrological investigations.

The original form of the intrusion was a funnel or lopolith that has since been modified by post-consolidation faulting. The marginal zone consists chiefly of olivine gabbro. The layered series is mainly troctolite and leucotroctolite which is overlain in turn by a thick anorthosite zone and an upper zone of leucogabbro and gabbro. Transgressive late differentiates are ferrodiorite, ferromonzonite and ferroadamellite. Rhythmic layering and planar plagioclase orientation are widespread and form the basis for structural interpretation. Layering in both the marginal zone and the layered series is subparallel to the walls and is believed to have formed largely through the action of convection currents. Faults complicate the interpretation but the maximum thickness of the intrusion is estimated to have been in excess of 15 km.

The Michikamau Intrusion appears to have been emplaced at and near the contact between a granulitic gneiss basement complex and a supracrustal sequence (the Petscapiskau Group). The magma that filled the Michikamau chamber is believed to have been an aluminous basaltic liquid carrying a suspension of plagioclase crystals. A model is outlined in which, shortly after emplacement, the suspended plagioclase rose to accumulate in the upper part of the chamber and thereafter solidification took place by bottom accumulation of crystals.

Figure 1. Location map showing the major anorthositic intrusions in Labrador (the Kiglapait intrusion is a layered troctolite, gabbro, syenite body probably related to the anorthositic suite).

THE GEOLOGY OF THE MICHIKAMAU INTRUSION, LABRADOR

INTRODUCTION

The Michikamau Intrusion of west-central Labrador is one of a group of large anorthositic intrusions (Fig. 1) that underlies a substantial part of the Nain Province (Stockwell, 1964; 1965) and extends far into the adjoining Grenville Province. There are advantages to studying these intrusions in the Nain Province, perhaps the most significant of which is that they are largely unaffected by metamorphic overprint that seriously hampers investigations of many anorthositic intrusions in the Grenville Province.

The Michikamau Intrusion is about 2,000 square kilometres in area. The main rock units are: a marginal zone of olivine gabbro; a thick, layered series of mainly troctolite and leucotroctolite; an anorthosite zone; an upper zone of leucogabbro and gabbro; a transgressive group of iron-rich differentiates. Textures and structures of the rocks provide evidence that accumulation of crystals played a major role in development of nearly all of the units. Post-solidification deformation of the intrusion is limited to faulting, most of which is of the nature of vertical displacements on vertical or near-vertical faults.

PREVIOUS WORK

The earliest recorded geological observations made in the region around Lake Michikamau were by A. P. Low (1897) during his epic explorations of Labrador and Quebec in the late nineteenth century. He was much impressed by the coarse, irridescent plagioclase crystals in anorthosite exposed on the shores of the lake. Low also described flat-lying red conglomerate and sandstone that he regarded as being of Cambrian age.

Since Low's work, there has been some prospecting activity in the area but no further published geological studies have appeared.

PRESENT INVESTIGATION

Preliminary maps and accounts of the geology of the region have been published (Emslie, 1963; 1964a), based on field work done in 1962 and 1963. Much of the mapping of the Michikamau Intrusion was completed in

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those two years and in 1966 another field season was devoted to more detailed examination and sampling of critical parts of the intrusion. Also in 1966, R. Wares, an M.Sc. candidate at Queen's University, mapped a section across the layered series on the south lobe of the intrusion. Some further sampling of the intrusion was done during a brief visit in 1967.

A number of aspects of the investigation of the Michikamau Intrusion have been started and are continuing. These include petrography, rock and mineral analyses, and mineral composition determination by physical methods. Some of the preliminary results of these investigations have already been reported (Emslie, 1964b; 1965; in press).

The aim of this contribution is to provide a description, based chiefly on field observation, of the intrusion together with an interpretation of the structure and certain features of the lithology. This is intended to form a framework into which subsequent studies can be fitted as they are completed.

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GEOLOGICAL SETTING

REGIONAL GEOLOGY

The geology of the area surrounding the Michikamau Intrusion is illustrated in Figure 2 which is based on published maps of the Geological Survey of Canada (Emslie, 1963; 1964a) amended where necessary by more recent data. Mapping in adjacent areas to the west has been done by Wynne-Edwards (1960), to the south by Eade (1952), and the east by Roscoe (unpublished). Systematic mapping of the adjoining area to the north has not yet been undertaken.

By comparison with the Michikamau Intrusion much of the surrounding areas are low-lying. The area lies on the divide separating northerly-flowing drainage into the George River system, from south- and east-flowing drainage into the Naskaupi and Churchill rivers and it is not surprising therefore that large low areas and extensive muskegs are common. Most parts of the intrusion are clearly defined topographically in relation to the surrounding country rocks. The highest hills stand over 600 feet above the level of Lake Michikamau but local relief exceeding 300 feet is rare. The topography of the intrusion is subdued to varying degrees by a mantle of drift. Larger exposures occur on hilltops and there is commonly a dearth of outcrops in the

LEGEND

8	Red conglomerate and arkose, sandstone, tuff, (equivalent to Seal Lake Group?)
7	Pink to red coarse-grained granite and syenite, typically massive
6	Coarse, even-grained to porphyritic granitic rocks; typically with a green or brown cast and bearing pyroxene, hornblende and biotite
5	Monzonite, syenite, and related varieties; pyroxene-bearing but may also contain variable amounts of hornblende and biotite
4	Michikamau and Harp Lake intrusions; chiefly leucotroctolite, leucogabbro and anorthosite
3	Gabbro and diorite
2	Metasedimentary and metavolcanic rocks (probably of more than one age); 2a, Petscapiskau Group
1	Quartzofeldspathic gneiss and granulite complex; local abundant lenses and layers of amphibolite, mafic granulite and paragneiss. Small granitic injections have produced some areas of migmatite

Heavy overburden . . . . . .

Foliation trends . . . . . .

Inferred fault . . . . . .

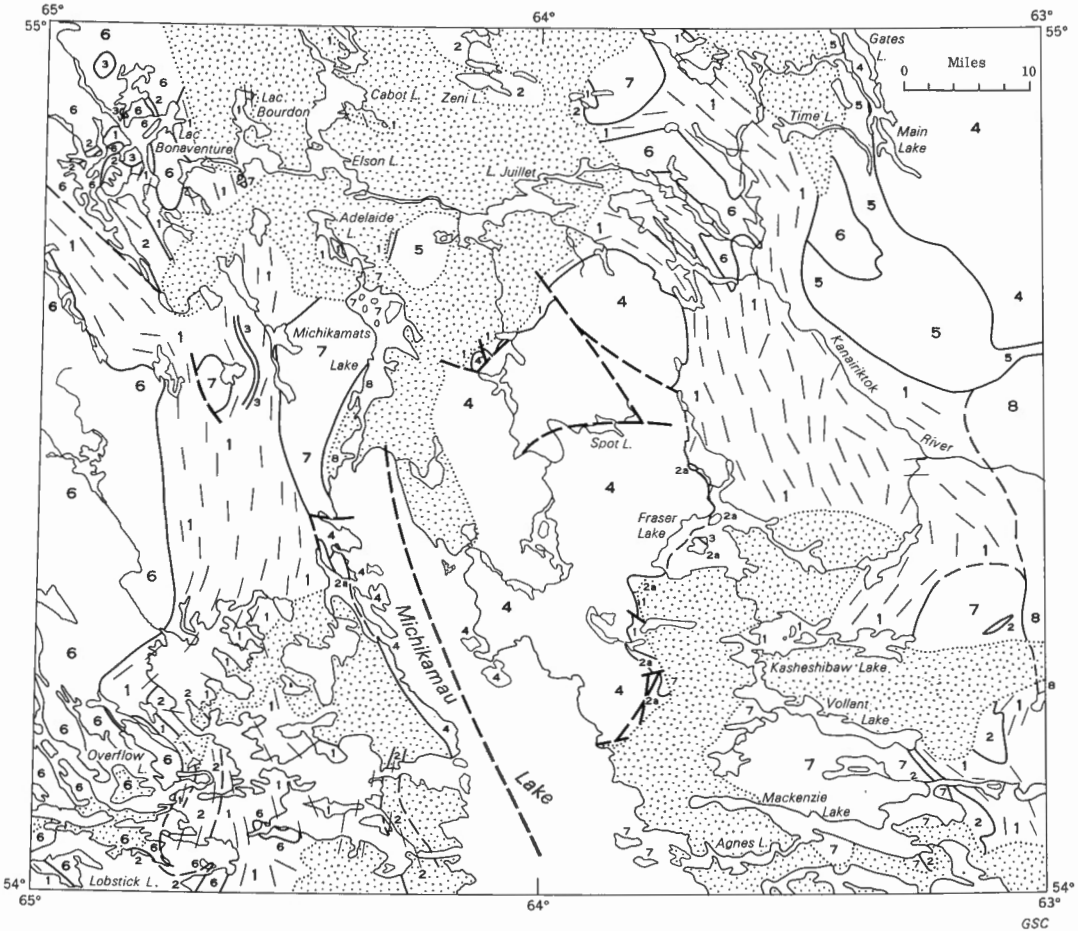


Figure 2. Regional geology in the vicinity of the Michikamau intrusion.

intervening valleys. Several large eskers cross the intrusion and some excellent exposures are found at intervals along their courses where the bedrock has been swept clean.

Within the area of Figure 2 the most widespread rock-unit is a quartzofeldspathic gneiss complex (1). Exposures range from uniform, nearly massive varieties to migmatitic rocks with bands and lenses of biotite gneiss and amphibolite, all of which are commonly injected with irregular small dykes and stockworks of pink or grey granite. The more homogeneous parts of this gneiss complex clearly show characteristics of granulite facies metamorphism. Feldspar has a greenish cast and orthopyroxene or clinopyroxene or both are commonly present in small amounts together with a little deep red-brown biotite and/or brown amphibole. At the scale of mapping it was not possible to separate clearly those gneisses with well-defined granulite facies mineralogy from more questionable varieties. However, even in rocks lacking pyroxenes the common occurrence of microperthite and antiperthite and the absence of muscovite suggest a high grade of regional metamorphism. The metamorphism represented in the bulk of the gneisses is believed to be at least upper almandine amphibolite and probably in large part, granulite facies. Apart from remnants of amphibolite, paragneiss, and mafic granulite too small to show on the map, the gneiss complex as a whole is regarded as the oldest rock-unit although it must be admitted that the evidence supporting this interpretation is indirect. Known igneous intrusions all appear to cut the gneiss complex. Contacts between the gneiss and significant bodies of paragneiss and/or amphibolite have not been directly observed but close to such contacts there is no evidence of granitic injection or 'granitization'. In the vicinity of the Michikamau Intrusion where detailed examination has been made, all contacts of the Petscapiskau Group dip away from the gneiss and there is little reason to doubt that it overlies the gneiss.

Because of its proximity to the Michikamau Intrusion, the Petscapiskau Group (2a) has been examined in greater detail than any of the other paragneisses and amphibolites (2) indicated in Figure 2. Most of the exposures lie within the thermal aureole of the intrusion and the widespread and abundant occurrence of cordierite and sillimanite indicates that pelitic and semipelitic compositions are common. Green spinel occurs in silica-undersaturated compositions but more commonly quartz is present both as an integral part of the mineral assemblage and as lensy segregations, veinlets and pods. In less aluminous layers hypersthene is common and garnet occurs sporadically. Amphibolite layers contain deep brown hornblende with orthopyroxene and clinopyroxene. Farther from the contact, and separated by faulting from the rocks just described, are upper greenschist facies schists (some bearing kyanite), paragneisses, quartzite, and meta-tuff. These are interpreted as Petscapiskau Group rocks at the grade of regional metamorphism they exhibited when the Michikamau Intrusion was emplaced. Whether the rocks attained this grade just prior to intrusion or long previously is not known.

Much less is known about the belts of metamorphic rocks of presumed sedimentary and/or volcanic origin (2) in the remainder of the area of Figure 2. The belt north of Zeni Lake is chiefly amphibolite gneiss, garnetiferous in part, as is the band south of Lac Bonaventure. Garnet-biotite

quartz-feldspar gneiss occurs in the belt just east of Overflow Lake. The northerly-trending belt farther to the east is poorly exposed. It contains impure quartzitic metasediments, fine-grained schist and meta-tuff. In the southeast part of Figure 2 fine-grained mafic schists and amphibolite have interbedded quartzite and paragneiss. On the basis of lithology, the rocks at these latter two localities are the most likely equivalents of the Petscapiskau Group.

The several small areas shown as gabbro and diorite (3) in the northwest part of the area are in part, and perhaps largely, contaminated granitic rock and/or strongly recrystallized amphibolite.

Medium- and coarse-grained pyroxene-bearing monzonite and syenite (5) occur in a large body adjoining the Harp Lake Intrusion and in a smaller mass northwest of the Michikamau Intrusion. These rocks are so similar in character and mineralogy to late differentiates of the Michikamau Intrusion that the author has no hesitation in considering them to be part of the same igneous suite. However, because less detailed study has been made of the Harp Lake Intrusion and of the Signal Hill body it was considered wise simply to show them as separate intrusions.

The coarse-grained and porphyritic granitic rocks (6) are not greatly dissimilar in appearance from the previously described monzonites and syenites, but the granites have more quartzose compositions and greater abundance of biotite and hornblende relative to pyroxene. The possibility must be entertained that these granitic rocks are also part of the troctolite-anorthosite-monzonite suite in this region. Wheeler (1955; 1960) considers that similar granitic rocks in the Nain district are related to the anorthosite suite.

The pink and red granites of unit 7 are biotite- and/or hornblende-bearing and are typically coarse to very coarse grained. As a group they are believed to be distinctly younger than the Michikamau Intrusion. Large and small pink granite dykes cut the Michikamau Intrusion on the west shore of Michikamau Lake near the granite contact and also southeast of Michikamau Lake.

On the west shore near the north end of Michikamau Lake a few outcrops of flat-lying red conglomerate, arkose and sandstone (8) have been found; the conglomerate was observed to rest unconformably on granite at one locality. These rocks are assumed to correlate with arkose, quartz sandstone and tuff occurring along the east side of the map-area (Fig. 2) and it is believed probable that all are stratigraphically equivalent to the Seal Lake Group (Brummer and Mann, 1961).

Only a very few diabase dykes were noted in the area. A swarm of small basic dykes dipping at low angles to the northeast was found filling joints in the Michikamau Intrusion near the contact east of Lac Montor. These are fine-grained, brown-weathering rocks unlike typical diabase dykes.

Potassium-argon dates, chiefly on biotite, have been determined on samples from a number of the rock-units. These data are summarized in the following table:

Table 1

Potassium-Argon Age Determinations

GSC Reference No.	Sample Description	K-Ar date (m. y.)	Reference
GSC 65-151	tuff of intermediate composition; flat-lying beds north of Kanairiktok River (Unit 8, Fig. 2).	843 whole rock	Wanless, et al., 1967, pp. 118-119.
GSC 63-165	pink, coarse, biotite hornblende microperthite granite, Michikamats Lake (Unit 7, Fig. 2)	1360 biotite	Wanless, et al., 1965, pp. 114-115.
GSC 61-197	porphyritic biotite granite; between MacKenzie and Windbound Lakes (Unit 7, Fig. 2).	1395 biotite	Lowdon, et al., 1963, p. 116.
GSC 63-164	leucotroctolite from Michikamau Intrusion; west shore of Michikamau Lake (Unit 4, Fig. 2).	1400 biotite	Wanless, et al., 1965, pp. 113-114.
GSC 63-148	paragneiss (probably inclusion in main gneiss complex); Prong Lake (Unit 1, Fig. 2)	1455 biotite	Wanless, et al., 1965, p. 103.
GSC 60-129	biotite pyroxene hornblende gneiss; Crossroads Lake (Unit 1, Fig. 2)	1500 biotite	Lowdon, 1961, pp. 74-75.
GSC 63-163	garnet biotite schist, Petscapiskau Group; Contact Lake (Unit 2a, Fig. 2).	1520 biotite	Wanless, et al., 1965, p. 113.

The latter six dates in the table fall within a range of only 160 m. y. which is not very much greater than the experimental error. All may be regarded as valid minimum ages for the rock units represented. Biotite from the Michikamau Intrusion (GSC 63-164) is a primary igneous mineral, not an alteration product. Specimen GSC 65-151 is a whole rock K-Ar date and as such it has a rather larger error than the others but still provides a useful minimum age for the rock.

COUNTRY ROCKS IN THE VICINITY OF THE MICHIKAMAU INTRUSION

In the course of mapping the Michikamau Intrusion a good deal of information was obtained on the country rocks adjacent to the intrusion. In the following paragraphs some of these data are presented and discussed in more detail than was provided in the preceding section. The principal rock-units to be considered are the quartzofeldspathic gneiss complex and the Petscapiskau Group but some comment is also devoted to certain other less extensive rock-units.

Rocks Older than the Michikamau Intrusion

Quartzofeldspathic gneiss complex

Most of the rocks in contact with the Michikamau Intrusion belong to a widespread gneiss complex. On the scale of an outcrop the gneiss is commonly heterogeneous, containing well-banded parts, more uniform parts, perhaps one or more dark ghosty schlieren, and sometimes well-defined inclusions. Small, irregular veins and pods of grey or pink granitic material may be seen in many outcrops. The general aspect of outcrops of the gneiss is typically a buff or light rusty colour. This is clearly a weathering effect for where fresh samples can be obtained they are light to medium greenish grey.

The grain size of the gneiss is commonly 1 to 3 millimetres. The content of mafic minerals is commonly less than 10 per cent but may reach as high as 20 per cent. Red-brown biotite, brown hornblende, clinopyroxene and orthopyroxene are the principal ferromagnesian silicates. Garnet is present in some specimens but is not common. Chlorite pseudomorphs after pyroxene have been recognized in a few specimens. Quartz and plagioclase comprise the bulk of most samples and occur in proportions of about 1:2. Plagioclase is generally moderately to strongly antiperthitic. Potash feldspar can be as abundant as 20 per cent of the total feldspar but generally it amounts to 10 per cent or less; it is commonly microperthite.

The rocks are generally equigranular or nearly so. Where biotite is abundant it may exhibit slight preferred orientation. Streaks and lenticles of quartz also contribute to the formation of a foliation. Most commonly, however, the chief foliation that is measured is a compositional variation (banding) due to minor fluctuation in concentrations of quartz, feldspar and mafic minerals. The bands are between 1 and 10 centimetres thick. In many outcrops, although an overall foliation attitude can be measured, the local structure is highly complex. Folds with amplitudes of a few centimetres to a metre are common and fold axes seem to be nonparallel more often than not. Streaky schlieren are elongate parallel to the foliation as are more definite inclusions of amphibolite, mafic granulite and siliceous paragneiss. Inclusions are rarely more than a metre in maximum dimension but a few larger lenses have been seen.

The regional trend of foliation in the gneiss complex strikes north to northwest. Dips in most places are steep to vertical. From the foliation attitudes concentric to the margin of the north lobe of the Michikamau Intrusion, it is clear that they have been deflected around the intrusion, presumably by stresses related to emplacement.

The Petscapiskau Group

Rocks of the Petscapiskau Group occur in contact with the Michikamau Intrusion for about one quarter the total length of its periphery - chiefly from Fraser Lake south and west to Lake Michikamau. A smaller area of Petscapiskau Group rocks occurs at the contact of the intrusion on the west shore of Lake Michikamau. Only those metasedimentary and metavolcanic rocks in contact with, and adjacent to, the Michikamau Intrusion are included in the Petscapiskau Group (Fig. 2).

TABLE OF FORMATIONS

EON	ERA	Formation	K/Ar biotite age	Lithology
Proterozoic	Helikian			Flat-lying red conglomerate, arkose, sandstone
		Unconformity		
			1360 m. y. (granite)	Granite, dykes of gabbro and diabase
		Intrusive Contact		
		Michikamau Intrusion	1400 m. y.	Olivine gabbro, troctolite, leucotroctolite, anorthosite, leucogabbro, gabbro, ferrodiorite, ferroadamellite, ferromonzonite
		Intrusive Contact		
	Aphebian (?)	Petscapiskau Group S	1520 m. y.	Paragneiss, schist, quartzite, meta-tuff, amphibolite
		Unconformity (?)		
Archean (?)				Quartzofeldspathic gneiss, granulite, migmatite

Rocks placed in the Petscapiskau Group comprise two markedly different metamorphic mineral facies. The most widespread assemblages are products of contact metamorphism by the Michikamau Intrusion. Less abundant, but well-exposed in the area between Contact Lake and Magnetite Lake, are regional metamorphic schists and gneisses. These two facies are everywhere separated by faults and it cannot be directly demonstrated that they are equivalent. However, there are sufficient lithological similarities so that there is little reason to doubt a correlation.

Contact metamorphic rocks. All of the rocks are too coarse to be adequately described by the term hornfels. Textures are equigranular and macroscopically isotropic. Banding on a scale of a centimetre to 30 centimetres or more is common so that the term gneiss is used to describe most of these rocks although it is not entirely appropriate. The compositional banding is most readily explained as relict bedding and is so interpreted.

The most common and distinctive rocks are sillimanite-cordierite gneisses. These are strikingly exposed on and around Petscapiskau Hill where

the cordierite content in some places is so high that entire rocky hillsides have a bluish cast - an effect that is especially pronounced when the rocks are wet. Layers containing abundant prismatic sillimanite crystals up to 0.6 by 2 centimetres have been observed. Porphyroblasts of pale pink to dark red garnet are not uncommon but never abundant. A universal feature of these gneisses is the presence of quartz lenses and pods which commonly comprise 10 to 25 per cent of an outcrop (Pl. IA). They range in size from thin partings to fat pods and, although most are concordant with the layering and folded with it, examples of transgressive quartz veining are usually also present. Small amounts of disseminated pyrite and pyrrhotite are commonly present and in a few places, such as Rusty Island in Fraser Lake, disseminated sulphide is sufficiently abundant to cause a marked rusty oxide weathering product over much of the outcrop.

Pyroxene-plagioclase-quartz rock, impure quartzite, quartzite, and pyroxene amphibolite occur interbedded with the main sillimanite-cordierite gneiss. A broad band of amphibolite occurs north of Contact Lake. The mineral assemblages of most of these contact rocks are indicative of pyroxene-hornfels facies. Examples of some of the observed mineral assemblages are:

Cordierite-sillimanite-microperthite-quartz
Cordierite-hypersthene-plagioclase-quartz
Cordierite-garnet-brown biotite-plagioclase
Cordierite-brown biotite-plagioclase-green spinel
Cordierite-sillimanite-green spinel
Hypersthene-plagioclase-brown biotite-quartz
Hypersthene-garnet-brown biotite-plagioclase-quartz
Clinopyroxene-brown hornblende-plagioclase-quartz

Wrinkling of layers in the gneisses is common and small scale folds with amplitudes up to 3 metres occur in some places (Pl. IB). For the most part, however, attitudes of layering are remarkably regular and persistent.

Regional metamorphic rocks. At the eastern end of Contact Lake there are biotite-rich schists and gneisses with some garnetiferous layers. These have abundant quartz lenses and stringers very similar to those seen in the contact metamorphic rocks. Farther east, and presumably underlying the biotitic rocks, is a thick section of finely banded, fine-grained amphibolite with small blebs and larger pods of quartz-epidote. The delicate layering is well preserved and these rocks are interpreted as bedded mafic tuffs. Beyond these to the east are biotite- and muscovite-bearing schists (one sample carries kyanite). Adjacent to the granite contact there is a coarse wollastonite-quartz rock - presumably the result of contact metamorphism of carbonate-rich beds by the granite. A few very pure white quartzite layers also occur in the sequence.

East and northeast of Magnetite Lake are numerous well-preserved outcrops of fine-grained amphibolites banded on a scale of 2 millimetres to 1 centimetre. These rocks are undoubtedly derived from mafic tuffs. They carry readily recognizable fragments up to 12 by 5 centimetres but many are less than 1 centimetre. Most of the fragments are somewhat elongate parallel to the banding. The fragments are slightly to markedly more felsic than the matrix, many have plagioclase phenocrysts, and one was found with amygdules of

carbonate and epidote. Epidote-rich pods and streaks are not uncommon. These rocks are strongly schistose adjacent to the fault passing through the pond east of Magnetite Lake.

Some examples of the mineral assemblages in these regional metamorphic rocks are:

Red-brown biotite-muscovite-kyanite-plagioclase-quartz
Garnet-deep olive biotite-plagioclase-quartz
Green hornblende-brown biotite-epidote-plagioclase-quartz
Olive brown hornblende-epidote-plagioclase-quartz
Diopside-brown hornblende-epidote-plagioclase-quartz

All of these regional metamorphic rocks are assigned to the Petscapiskau Group and show deformation on a small scale. Small folds and crenulations in schistosity are common, but the general attitude of foliation is more regular on the scale of an outcrop and over larger distances.

Rocks Younger than the Michikamau Intrusion

Granite

The granite on the west shore of Michikamau Lake is part of a larger body that extends north of Michikamats Lake. This is a massive, coarse to porphyritic granite bearing hornblende and biotite. It ranges in colour from pale pink to deep salmon red. For the most part the granite is fresh but on the north shore of Fault Bay, where it is faulted against rocks of the Michikamau Intrusion, shear zones are present and the mafic minerals are strongly chloritized. Granite has invaded the Michikamau Intrusion at the west end of Fault Bay where large blocks of leucogabbro have been stoped from the intrusion and engulfed by granite. Dykes of granite also cut rocks of the intrusion and of the Michikamau Group in the area south of Fault Bay. A few granite dykes cut anorthosite and leucogabbro near the edge of the low-lying muskeg area east of Michikamau Lake.

The small area of granite that outcrops east and northeast of Contact Lake is of unknown age relative to the Michikamau Intrusion although it is considered here with the post-intrusion rocks. The granite is pale pink to pinkish grey and medium grained with a finer facies near contacts with the Petscapiskau Group. It intrudes regionally metamorphosed facies of the Petscapiskau Group and an excellent wollastonite-quartz reaction zone is present at the contact in the northeast bay of Contact Lake. Other evidence of contact metamorphic effects on the Petscapiskau Group are not readily apparent but may be revealed by closer investigation.

Sedimentary rocks

Only a few exposures of unmetamorphosed sedimentary rocks are present in the area. Those on the northwest shore of Lake Michikamau are red arkosic pebble and boulder conglomerate, red arkose and sandstone, and grey siltstone. A granite wash was observed to grade up into arkosic conglomerate so that there can be little doubt that these sedimentary rocks unconformably overlie the granite. All observed bedding was flat-lying or nearly so. Pebbles in the conglomerate consist of quartz, granite and

porphyritic felsic volcanic rocks. The nearest known source of such volcanic pebbles is the Leticia Lake Group (Brummer and Mann, 1961) about 70 miles east.

Low (1897) described bluish grey limestone resting upon granite on a large hill on the east shore of Lake Michikamau, about 12 kilometres south of the southern end of the Michikamau Intrusion. He also described stromatolitic structures in limestone and believed that this rock was a stratigraphically higher part of the red conglomerate-arkose-sandstone sequence. Erratics of similar limestones were found during the present investigation but none was observed in place and no attempt was made to confirm Low's observations at the locality described above.

Inasmuch as these sedimentary rocks unconformably overlie granite dated at 1360 m. y. (see Table 1) which in turn cuts the Michikamau Intrusion dated at 1400 m. y. (see Table 1), it is concluded that they are most likely Upper Proterozoic in age.

Dykes

It has already been noted that granite dykes cut the Michikamau Intrusion in a few places, notably on the west shore of Lake Michikamau south of Fault Bay. Several small white granite dykes cut leucotroctolite at the head of Windy Bay and on the north shore of the bay near its mouth. A few small stringers of granite have been noted in other parts of the intrusion but they are rare.

An interesting swarm of mafic dykes occurs in the north lobe of the Michikamau Intrusion near its margin between Wolf Ridge and Castle Hill. The dykes are clearly controlled by a joint system that dips northeast at 20 to 40 degrees. The dykes are 2 centimetres to 2 metres in thickness. They are fine grained and weather to a brownish grey. Whether or not they are related to the intrusion is unknown but they have not been found outside the limits of the intrusion.

Only two small diabase dykes were observed. One on the south shore of Contact Lake in marginal rocks of the intrusion and another in leucotroctolite about one mile east of Muskeg Bay.

THE MICHIKAMAU INTRUSION

FORM AND INTERNAL STRUCTURE OF THE INTRUSION

Form of the Intrusion

The form of the Michikamau Intrusion is more complex than is implied by application of one of the common descriptive terms used for large mafic intrusions, such as lopolith or funnel. The intrusion probably can be most closely described as funnel-shaped although, as the sequel will show, there are significant differences from known funnel intrusions. Adequate characterization of the shape of the intrusion is difficult because its large

size makes extrapolation to great depths a necessary, but always hazardous, procedure. In addition, post-intrusion faulting is a complicating factor in the structural interpretation.

To facilitate discussion of form and geology the intrusion can be divided into four parts:

1. the north lobe - that part of the intrusion lying mainly, but not entirely, north and east of the Esker Fault and the Shallow Lake Fault.
2. the south lobe - that part of the intrusion south of a line drawn from the Buttress west to Lake Michikamau.
3. the central section - the main portion of the intrusion lying between the north and south lobes.
4. the west shore section - that part of the intrusion exposed on the west shore of Lake Michikamau.

The north lobe

Several lines of evidence indicate that the north lobe was probably one of the high parts of the intrusion at the time it crystallized. The main rock-units present there are anorthosite, leucogabbro and a suite of iron-rich rocks. These rocks are only found in quantity stratigraphically above the layered series which comprises about 70 per cent of the exposed rocks of the intrusion. Foliation in the quartzofeldspathic gneisses that surround the north lobe strikes parallel to the contact and dips nearly vertically in most places. Assuming that the dip of the contact also parallels foliation in the gneisses, we may conclude that the contact is nearly vertically-dipping around much of the north lobe. If so, this is the only known part of the intrusion where such is the case. Elsewhere, contacts dip beneath the intrusion, usually at low to moderate angles. Structural elements, which in the north lobe consist chiefly of planar-oriented plagioclase crystals, suggest that the gross structure is a shallow dome that initially comprised a shell of leucogabbro passing downward into anorthosite. Ferromonzonite has been intruded approximately along the northeast contact and its position was likely controlled, at least in part, by the contact. Truncation of the dome by erosion has produced the present distribution of rock-units. Large vertical movements are inferred along the Shallow Lake Fault and part of the Esker Fault - the north lobe has moved downward relative to the remainder of the intrusion adjoining it to the southwest.

The south lobe

The south lobe of the Michikamau Intrusion clearly has the form of a segment of a funnel. Although faulting has modified the original configuration to some degree, the inward-dipping contacts and internal layered structures indicate a funnel-shaped form centred in the vicinity of the Bay of Islands. The layered gabbros of the Bay of Islands apparently form a sequence continuous with the layered series, this being the only place in the intrusion where this relationship has been found.

The central section

In areal extent the central section comprises the largest part of the intrusion. Faulting has rotated and tilted blocks relative to one another so that a straightforward interpretation of the form is not readily apparent.

That part of the contact passing through Fraser Lake clearly dips inward beneath the intrusion. Northwest of Loon Lake faulting is present close to the contact so that the original form of the contact is doubtful. Dips of layering in the layered series and in the marginal gabbros are opposed - the only place in the intrusion where this has been observed. Either faulting, yet unrecognized, has played a role or the margin of the intrusion was steep and perhaps nearly vertical in this vicinity.

Northward to Shallow Lake the intrusion is bounded by a fault. West of Shallow Lake the movement on the Shallow Lake Fault must have been very large bringing lowermost troctolites of the layered series into contact with upper border leucogabbro.

Southwest of Boulder Lake part of the margin of the intrusion is preserved adjacent to a fault. Only a small area of marginal rocks remains but again the layering dips inward at a moderate angle.

Faulting that has dislocated and tilted large blocks in the central part of the intrusion makes impossible any attempt at extrapolation of dips to make inferences on the shape of the intrusion.

The west shore section

The part of the intrusion on the west shore of Lake Michikamau is believed to have been rotated counter-clockwise on a fault plane passing through the lake. This interpretation will be discussed in more detail in a later section on faulting. The presently exposed rocks along the west shore of Lake Michikamau are thought to represent a once steeply-dipping, perhaps nearly vertical section through the main part of the intrusion (see Fig. 4). If this interpretation is correct and it is assumed that the anorthosite zone was subhorizontal before faulting, then it appears that the western margin of the intrusion dipped inward at an angle of about 30 degrees.

Discussion

Everywhere that the contact of the Michikamau Intrusion is exposed and is uncomplicated by faulting, except for the northern lobe or dome, the walls and the layering in the marginal rocks dip inward beneath the intrusion at low to moderate angles. The walls surrounding the northern lobe are presumed to dip steeply at the present erosional level but at depths of a few kilometres they may be expected to change attitude and dip beneath the intrusion.

The central section of the intrusion is by far the largest and it may reasonably be supposed that it is underlain by the main feeder. Whether the northern and southern lobes were served by separate feeders and then coalesced with the main body or whether they are simply lateral bulges in the central magma chamber is open to question. There are no remnants of country rocks present where the lobes are joined to the central mass and internal structures do not indicate discontinuity. On the basis of this negative

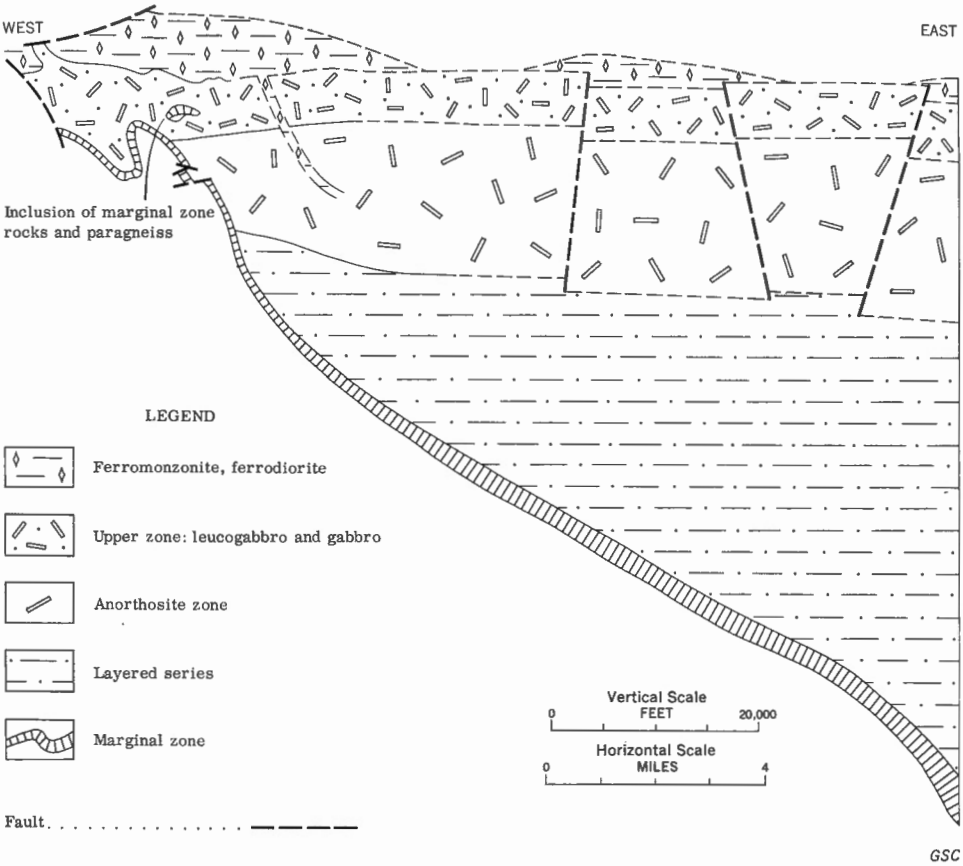


Figure 4. Reconstruction of a partial cross-section of the Michikamau intrusion based on the geology and structure on the west shore of Michikamau Lake. It is assumed that the initially subhorizontal anorthosite zone was brought into its present vertical position on the west shore by counterclockwise rotation on an east-dipping fault that underlies Michikamau Lake.

evidence it is assumed that the intrusion formed within one magma chamber served by a single feeder beneath the central part of the intrusion.

The original overall form of the intrusion is interpreted as having been a somewhat distorted funnel – elongated in a north-south direction with walls dipping inward at low to moderate angles. On evidence from the northern lobe, the roof of the intrusion was probably slightly domed with outward dips of 10 to 20 degrees. The isometric block diagram in Figure 3 is an attempt to portray some of the structural features of the intrusion.

Internal Structures

The most widespread mappable structures found in the Michikamau Intrusion are rhythmic layering and planar orientation of tabular plagioclase crystals. Where these structures were seen in proximity to one another they were sensibly parallel except in local areas disturbed by preconsolidation deformation. Mapping of these regularly disposed planar structures has revealed a reasonably clear picture of the structural and stratigraphic arrangement of the various units of the intrusion. In addition, some insight into the processes of crystallization and solidification of the intrusion can be gained through consideration of the reasons for the presence, absence and degree of development of these structures. Secondary foliations have been mapped in a few places only – always close to, and closely associated with, faulting. Macroscopic mineral lineation has not been observed in any of the rocks of the intrusion and microscopic fabrics have not yet been investigated.

Rhythmic layering

Rhythmic layering (Wager and Deer, 1939; Wager, 1968) is developed on vastly differing scales. Layers from a centimetre to several metres thick can be observed in individual outcrops. Layers over 100 metres in thickness are definable by mapping groups of outcrops.

Where marginal gabbro is chilled against the walls, the quenched rock is very fine grained, dense and massive. Passing inward, within 30 to 60 centimetres a transition begins, first to weakly and gradually to more strongly layered gabbro. Some layers, comprising concentrations of shiny black amphibole crystals, stand out strongly, but commonly, subtler variations in the concentrations of plagioclase, olivine and pyroxene cause a more subdued but nonetheless readily recognizable layering (Pl. IIA, IIB). The thickness of most of these layers is typically between 1 and 10 centimetres. Individual layers cannot be traced far as they commonly pinch out or are truncated by other layers within a metre or a few metres. The lateral persistence of a layer commonly seems to be roughly proportional to its thickness. Passing inward through the marginal gabbros layering becomes weaker and then virtually absent in the inner part of the marginal zone. Wherever the contact of the intrusion was observed, layering in the marginal gabbro was parallel to it.

The best-developed and most widespread rhythmic layering occurs in the layered series although it is by no means ubiquitous. Variation in proportions of olivine and plagioclase, the essential minerals of the layered series, is the chief cause of layering. Boundaries of layers may be sharp or either the lower or the upper boundary or both may be gradational. Most commonly the upper boundary is gradational and the lower boundary is sharp or sharper than the upper. Sections with abundant layering are separated by large thicknesses in which layering is rare or absent.

The most striking layers are those of dunitic and anorthositic composition. Of these extreme compositions the former is rare and forms layers of small size but the latter is widespread and commonly forms thick layers. Most dunitic layers are found in the lower part of the layered series - in the troctolites and lower leucotroctolites. Anorthosite layers are present throughout the layered series.

Layers caused by more moderate compositional variation are more common than either of the above types. In fact, a good deal of the layered structure can be overlooked when only a cursory examination is made. Sometimes this is because the layers are thick relative to the size of the outcrop but generally it is because of the subtle differences in composition between adjacent layers.

In the troctolites of the lower part of the layered series, layers typically are about 2 to 10 centimetres thick (Pl. IIIB). They are not sharply defined but are readily apparent on outcrop surfaces. Layers are more abundant per unit thickness in the troctolites than in any other part of the layered series. Well-developed anorthosite layers are rare in the troctolite. In several places horizons of elongate, irregular anorthosite clots define incipient anorthosite layers. Thin dunitic seams occur but comprise only a small part of the total troctolite section. They are commonly moderately to strongly serpentinized.

On the west shore of Lake Michikamau near the top of the ridge on the northeast side of Camp Island there is a zone about 12 metres thick of strongly layered rock. Layers range from 1 to 20 centimetres thick. The layered zone is bounded on both sides by uniform massive leucotroctolite. Much of the unlayered leucotroctolite has grain sizes larger than the layered rocks by a factor of up to 10. Numerous layers wedge out or are truncated at a low angle within a few metres or less though a group of layers may persist with nearly constant thickness across the outcrop Plates IXB, XA.

Strongly graded layers have been observed at several localities within the intrusion. Some of the most outstanding examples occur on wave-washed islands on the west shore of Lake Michikamau about 5 to 7 kilometres south of Camp Island (Pl. VIA, VIB, VIIA, VIIB). The layering dips about 50 degrees northeast. Individual layers range from a few centimetres to over 1 metre. The lower parts are dunitic and these grade upward regularly into leucotroctolite. The dunitic base of the next layer above is sharply bounded against leucotroctolite. In a few places, bulges suggestive of sole marks (Potter and Pettijohn, 1963; pp. 114-115) occur on the underside of dunitic layers. The ratio of dunitic material to leucotroctolite in layers is highly variable. The leucotroctolite is similar in all respects to massive leucotroctolite in nearby unlayered rock so it seems reasonable to infer that the layering was caused by a succession of pulses of deposition of olivine. Several small fragments of leucotroctolite in dunite and one fragment of dunite in leucotroctolite were observed. Plate VB shows a leucotroctolite fragment with a thin dunitic layer draped over it.

Lateral extensions of well-exposed layers commonly show thickening or thinning in one direction. On the basis of observations it is considered that lens-like layering of limited lateral extent is more characteristic of the intrusion than widespread tabular sheets. This is supported by the fact that it is commonly not possible to trace anorthosite layers, a hundred metres or more thick, for any great lateral distance.

Planar plagioclase orientation (igneous lamination)

In the Michikamau Intrusion planar orientation of plagioclase crystals (igneous lamination Wager and Deer, 1939; Wager, 1968) is at least as common as rhythmic layering, and perhaps more so. The degree of preferred

orientation varies from crude to very strong. It is often present but difficult to detect in rocks of low colour index. With increasing colour index plagioclase tablets are more clearly outlined by mafic mineral grains and orientation, if present, is readily visible.

Because plagioclase forms the framework of most of the rocks of the layered series, it is most obvious in determination of igneous lamination. However, it has been noted, in at least some troctolites, that flattened olivine grains produce a lamination in rocks in which plagioclase is equidimensional.

Plagioclase orientation is subparallel to rhythmic layering. Apart from this observation, there is no other obvious relationship between the two. Well-layered sections may show strong planar plagioclase orientation or it may be weak or absent. Strongly laminated rocks may have little or no layering. Even individual layers may be well-laminated yet neighbouring layers have no obvious preferred fabric although they contain abundant tabular plagioclase. This suggests that the plagioclase orientation is a primary depositional feature and not one formed during postdepositional compaction of a crystal mush.

Other structures

In addition to the structures described above there are several others that are less common and although their origin is not entirely clear, they are confidently regarded as primary.

On a low hill about 2 kilometres south of Boot Lake several anorthosite layers occur in leucotroctolite. In one place several separated anorthosite masses are present across the outcrop. They are up to one metre thick and two to three metres long as shown in Plate XVIA. The base of each anorthosite mass is sharply planar and they are subparallel. The tops are very irregular. This may be an example of an anorthosite layer that has been strongly eroded by magmatic currents after deposition. The fact that the planar bases have attitudes parallel to nearby layering would seem to rule out disruption of a layer by slumping.

Some anorthosite layers are moderately and even strongly folded. Other examples have been noted of layers that appear to have been abruptly thickened as if squeezed like toothpaste. Still other layers show incipient boudinage structure. In all of these structures the layers and the host rock are quite fresh and show no effects of secondary strain of recrystallization. The only conceivable explanation for such features is deformation of a crystal mush - in all likelihood by slumping on a depositional slope.

One curious, enigmatic structure that is widespread in the leucotroctolite subunit is anorthosite 'blobs' (see Pl. XVA, XVB). It is difficult to find a more suitable descriptive term because many of the structures are highly irregular, amoeboid masses. They range in size from perhaps 5 centimetres up to several metres across. In cross-section some may display one or more straight bounding faces and rarely, they appear nearly tabular. Individual bodies may be present in the midst of homogeneous rock or they may make up a substantial part of an outcrop. One example has been seen of an anorthosite layer that had partly broken up at one end producing irregular fragments. It is possible that slumping of anorthosite layers could produce a jumble of irregular fragments and if the layers had been incompletely

consolidated perhaps even subrounded shapes could result. Apart from this suggestion, the writer is at a loss to explain these peculiar structures.

DESCRIPTION OF ROCK-UNITS

In the following paragraphs the various rock-units of the Michikamau Intrusion are described by chiefly citing characteristics observable in the field. Here and there, additional petrographic and mineralogical data are included where they reinforce, or make possible, certain interpretations. However, there is no attempt to provide comprehensive information of this sort in this contribution.

The rock-units will be described in approximate chronological order from older to younger. It will become apparent from the ensuing discussion, however, that the order of succession is not clearly established for certain units at the present time. The succession has a critical bearing on interpretation of the manner in which the intrusion crystallized and solidified and efforts are currently being made to arrive at a solution.

The Marginal Zone

Many parts of the contact of the Michikamau Intrusion are not exposed and some exposed contacts are complicated by faulting. However, a sufficiently distinctive suite of rocks is present at many places along the contact that postulation of a marginal zone 'skin' enclosing the intrusion is a reasonable interpretation. The marginal zone is variable in thickness. It is thick adjacent to the lower (deeper) walls of the intrusion, much thinner along the upper walls and probably was thinnest over the roof of the intrusion. This is consistent with the interpretation that, apart from the initial thin 'chill', the bulk of the marginal zone rocks formed by relatively rapid accumulation of settled crystals on the steeply-dipping upper walls and more gently inward-dipping lower walls. The marginal zone is divided into two subunits: fine- to medium-grained banded border gabbro and medium- to coarse-grained massive olivine gabbro and pyroxene troctolite.

Fine- to medium-grained banded border gabbro

This rock has been found at all exposed and unfaulted contacts of the Michikamau Intrusion with older rocks. The rock is typically, though not invariably, banded on a scale of 1 to 10 centimetres. The banding (layering) reflects variation in proportions of mafic silicates, plagioclase and oxide minerals. Weathering accentuates the banding in these rocks and often reveals compositional variations that are not readily apparent on the fresh surface. The weathered surface typically appears as light to medium rusty brown or brownish grey. On glaciated outcrops drilling shows that the weathered zone is commonly only 1 to 2 centimetres deep. The fresh surface of hand specimens is medium to dark grey and has a definite granular appearance. A photomicrograph illustrating the texture of this rock has been published (Emslie, 1965, Pl. III, Fig. 7).

Where samples of the contact were obtained with a portable diamond drill it was found that about 30 centimetres of the intrusion adjacent to the contact was uniform, very fine grained to aphanitic brittle microgabbro. This is regarded as being a chilled contact and will subsequently be referred to as the 'chill' or 'chill zone'. Passing from this rock into the intrusion thin streaky bands begin to appear and there is a regular rapid increase in grain size until at 1 metre from the contact the average grain size is about 1 millimetre with some grains and thin seams of coarser size.

The principal minerals are plagioclase, clinopyroxene, orthopyroxene and olivine. Variation in concentration of these minerals is the chief cause of banding. In addition, thin layers enriched in magnetite and ilmenite are not uncommon and layers rich in dark brown amphibole are abundant at some, but not all, localities close to the contact.

The maximum observed thickness of the border gabbro is about 250 metres. In several places it is markedly less and commonly appears to vary in thickness along strike.

Medium-grained olivine gabbro and pyroxene troctolite

The inner part of the marginal zone is lithologically distinct from the outer, finer, banded gabbros. The boundary between the two is drawn where small scale layering ceases to be a regular feature of the rocks. This normally coincides with a gradual but marked increase in grain size and a less granular appearance of the textures. The boundary between fine-grained banded olivine gabbro and medium-grained uniform gabbro is, however, more or less gradational.

Outcrops weather to a medium or dark rusty brown. In many places weathering is deep and the outcrops present a rotten, rubbly appearance (Pl. IIIA). In spite of the appearance, the amount of chemical weathering, even in deeply weathered outcrops, is remarkably small. Although a heavy oxide stain may obscure the minerals it is apparent that, even in markedly friable material, much of the initial mineralogy is preserved. Oxidation of ferromagnesian silicates is the chief alteration effect.

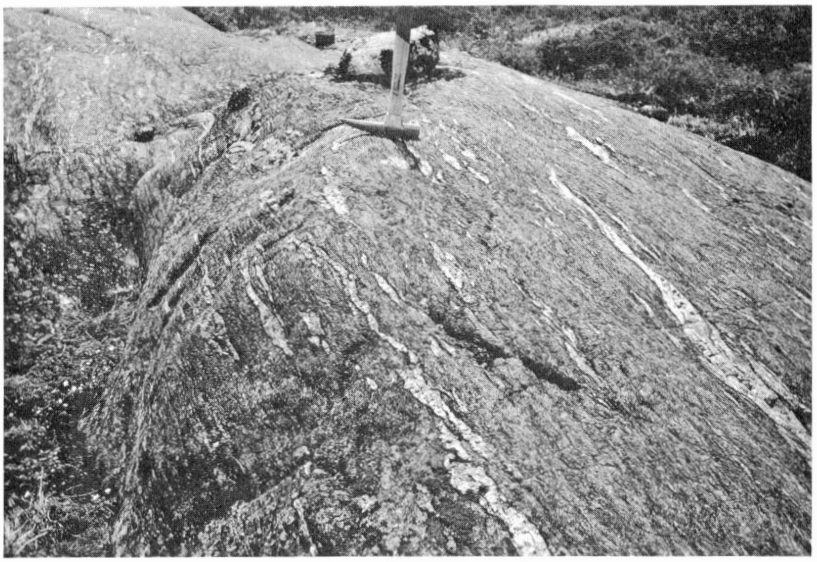
The fresh rock is medium to dark grey or greenish grey. It is normally equigranular with grain sizes of 3 to 4 millimetres although grain sizes larger than 5 millimetres are not uncommon. Plagioclase tends to be equidimensional or nearly so. This is in contrast to the more common tabular habit of plagioclase in the layered series. A photomicrograph showing the texture of this rock has been published (Emslie, 1965, Pl. III, Fig. 8).

Most outcrops are essentially structureless but rarely a thin streaky layer occurs and in the upper part of this unit, approaching the layered series, weak rhythmic layers seem to become more common.

The bulk of the rocks in this inner marginal zone may be described adequately as olivine gabbro. In some rocks the amount of olivine exceeds pyroxene and these are referred to as pyroxene troctolite. There is a gradual change from olivine gabbro to pyroxene troctolite in the upper part of the marginal zone which is itself transitional into the layered series. The lower boundary of the layered series is placed where significant amounts of pyroxene ($>10\%$ of the total mafic minerals) are no longer a regular feature of the rocks.

Plate IA Typical exposure of sillimanite-cordierite gneiss, Petscapiskau Group. Note abundant white quartz lenses and stringers, many of which show small scale deformation. Near the contact of the Michikamau Intrusion southwest of Hook Lake.

Plate IB Small shear folds in sillimanite-cordierite gneiss. South of Hook Lake. Photo by R. Wares.



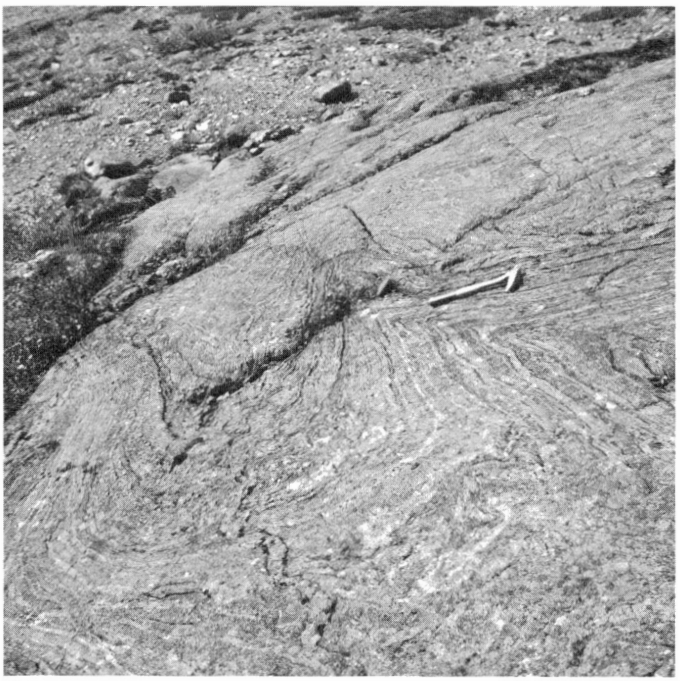


Plate IIA Small scale layering in marginal gabbro. Note what appears to be a primary trough structure in the middle of the photograph. South of Hook Lake.

Plate IIB Small scale layering in marginal gabbro. The darker layers contain concentrations of brown hornblende. South of Hook Lake.

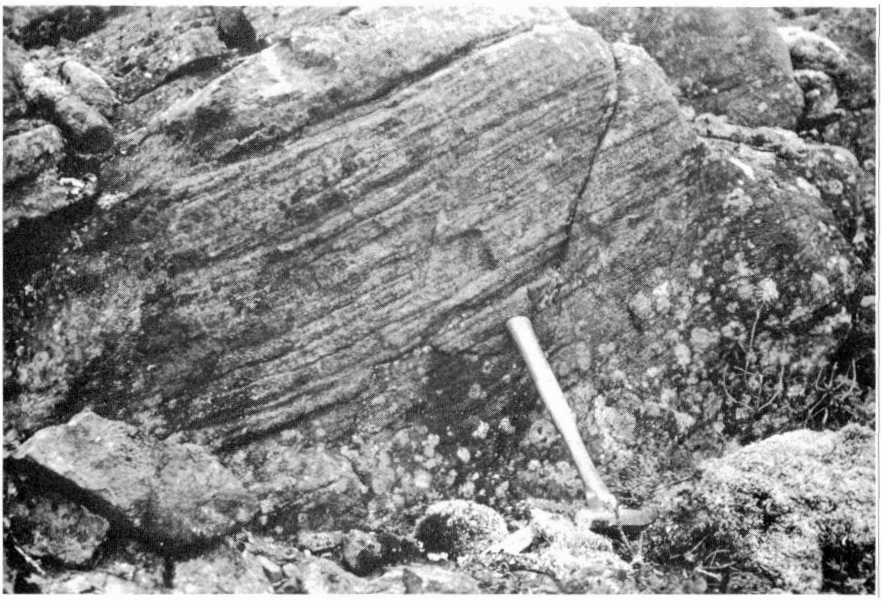


Plate IIIA Common rubbly-weathering appearance of marginal gabbros. West of Loon Lake, looking southeast.

Plate IIIB Small scale rhythmic layering in troctolite of the lower part of the layered series. West of Shallow Lake.

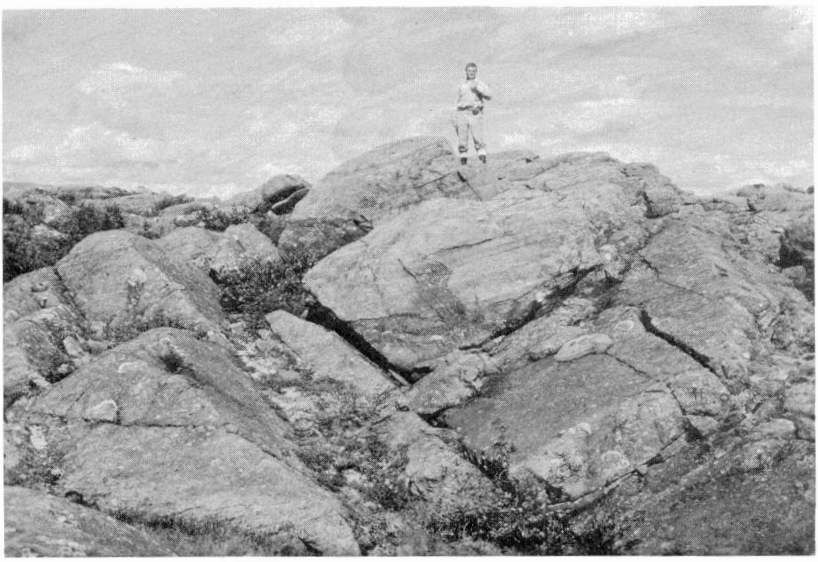


Plate IVA Thin anorthosite layers in leucotroctolite. The anorthosite layers weather in relief. About half way between Soaking Lake and Narrow Lake. Photo by R. Wares.

Plate IVB Rhythmically-layered section in leucotroctolite dips 50 degrees to the right. West shore of Michikamau Lake about 6 kilometres south of Camp Island.

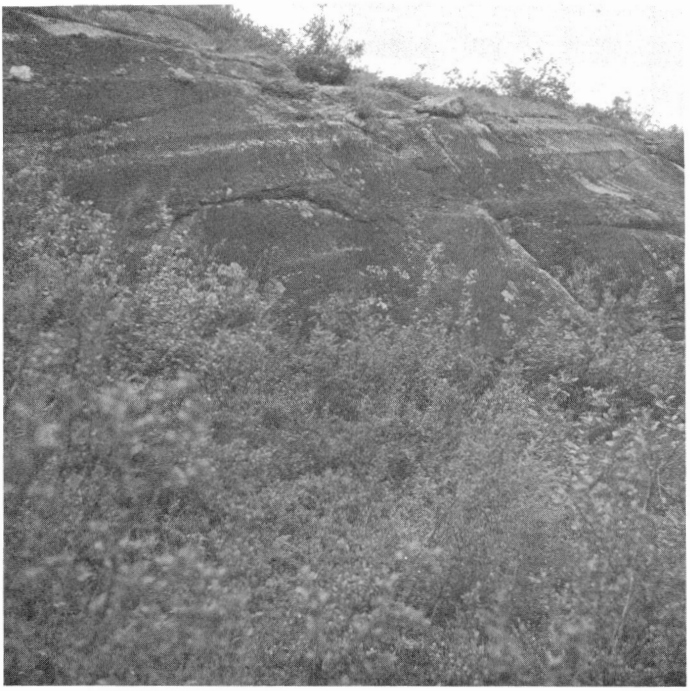


Plate VA Dunitic layers in leucotroctolite dip about 50 degrees to the right. About 9 kilometres south of Camp Island.

Plate VB Dunitic layers and fragments of dunite and anorthosite in leucotroctolite. Layering dips to the right about 50 degrees. At least three fragments are visible; a dunite fragment lies in leucotroctolite at left, an anorthosite fragment lies near the top of the main dunite layer and a thin graded dunitic layer is draped over an anorthosite fragment on the right. The fragments were presumably derived by slumping and/or erosion of layers some distance away and transported to the present site. Same locality as VA.

Plate VIA Sequence of rhythmic units with dunitic bases and anorthositic tops. Note the anorthosite fragment at lower right. West shore of Lake Michikamau about 6 kilometres south of Camp Island.

Plate VIB Closer view of mineral-graded layers shown in Plate VIA. Note the sharp boundaries between units but perfectly developed gradation within them.

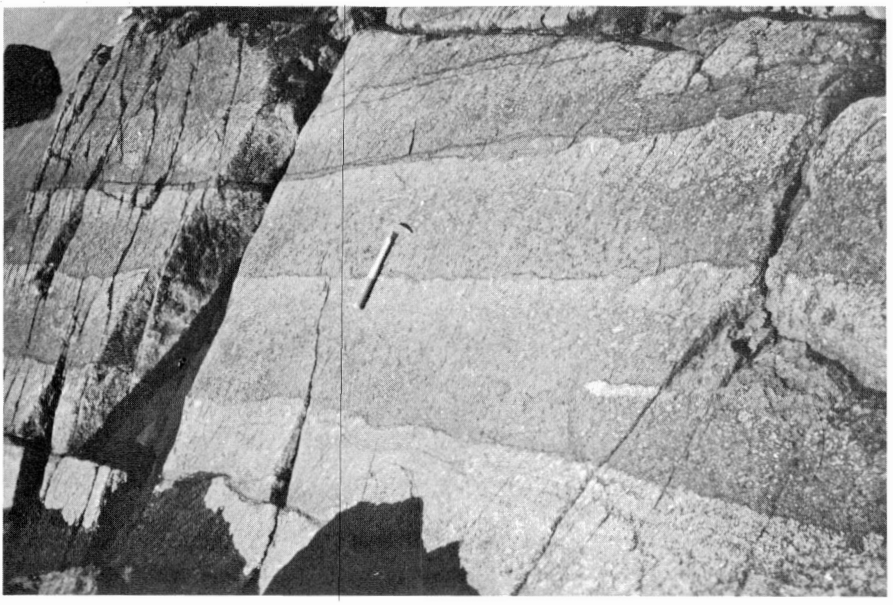


Plate VIIA Thick dunite layers in leucotroctolite. The layer upon which the hammer lies contains several anorthositic fragments and grades upward into a very thin feldspathic top. Same locality as VIB.

Plate VIIB Mineral-graded layers with olivine-enriched lower parts and plagioclase-rich tops. Near locality of VIA.

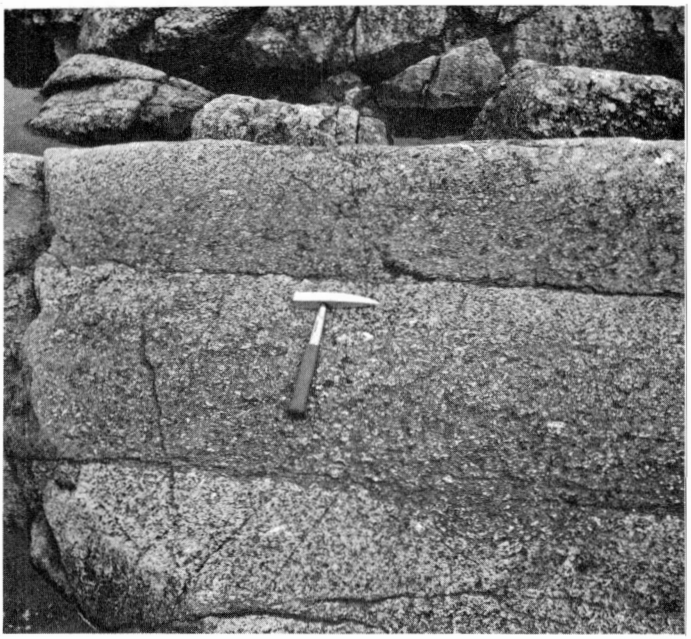


Plate VIIIA Thin dunitic layer has an irregular but sharp base and grades upward (to the right) into leucotroctolite. Near locality of VIA.

Plate VIIIB Thin diffuse rhythmic layers in leucotroctolite. Camp Island.

Plate IXA Closer view of diffuse rhythmic layers. Note large tabular plagioclase crystals subparallel to layering. Camp Island.

Plate IXB Fine scale layering in leucotroctolite. Note the convergence of groups of layers. Camp Island.

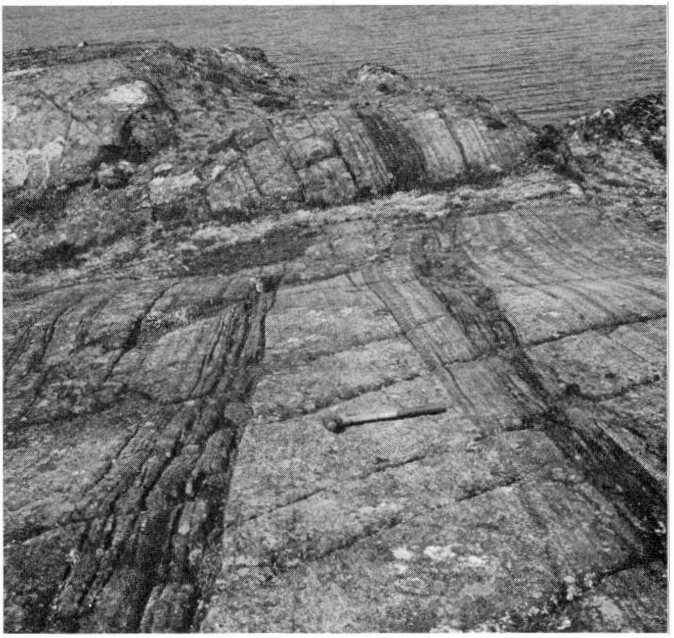


Plate XA Fine scale layering in leucotroctolite. Note the lensy nature of the layers. Same locality as IXB.

Plate XB Mottled zone of anorthosite in leucotroctolite. May represent slight disruption of a pre-existing incompletely consolidated anorthosite layer. Southwest of Hook Lake. Photo by R. Wares.

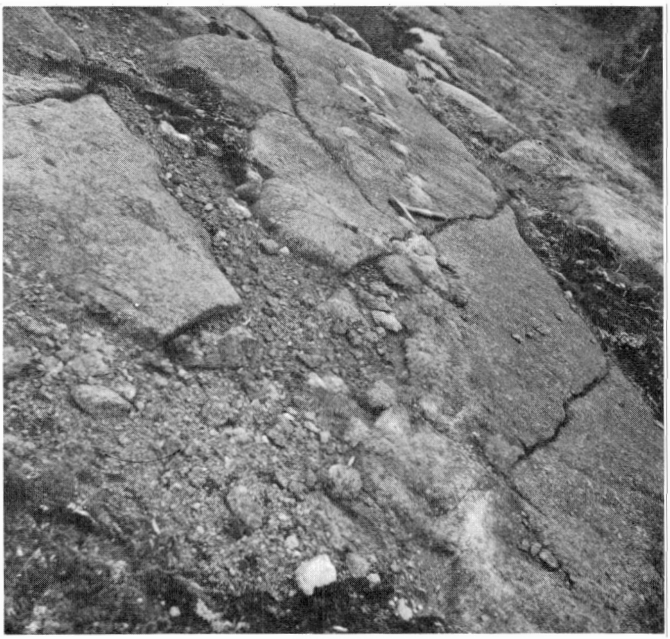


Plate XIA Sinuous anorthosite layers in leucotroctolite. They are inferred to be caused by minor deformation of a crystal mush, possibly due to slumping. Southwest of south end of Narrow Lake. Photo by R. Wares.

Plate XIB Anorthosite layers in leucotroctolite. Note the convergence of layers and wedging out of a thin leucotroctolite layer beneath butt of hammer handle. Southwest of Mosquito Lake. Photo by R. Wares.

Plate XIIA Two anorthosite layers in leucotroctolite coverage to form part of a trough-like structure. Near locality XIB. Photo by R. Wares.

Plate XIIB Trough-like structure formed by anorthosite layers in leucotroctolite plunges down dip to left. The anorthosite layers were more resistant to glacial scour than the leucotroctolite host. Southwest of Mosquito Lake.

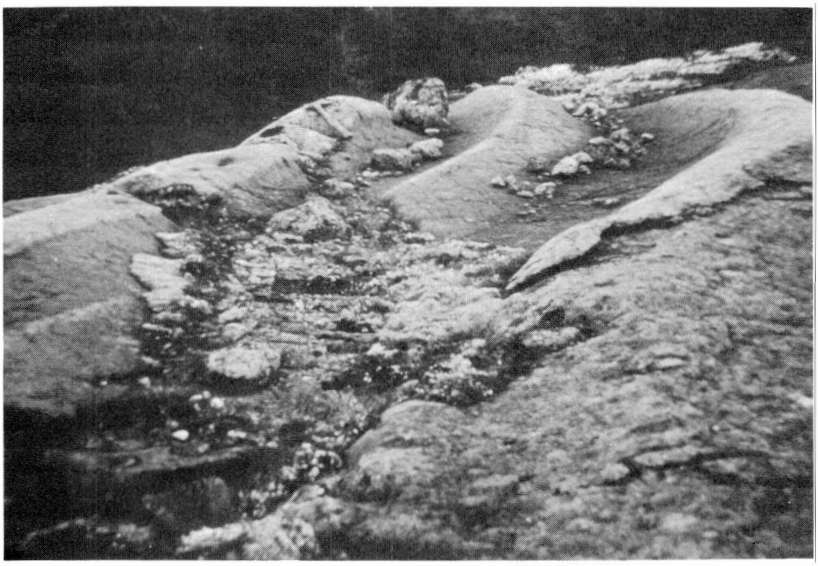
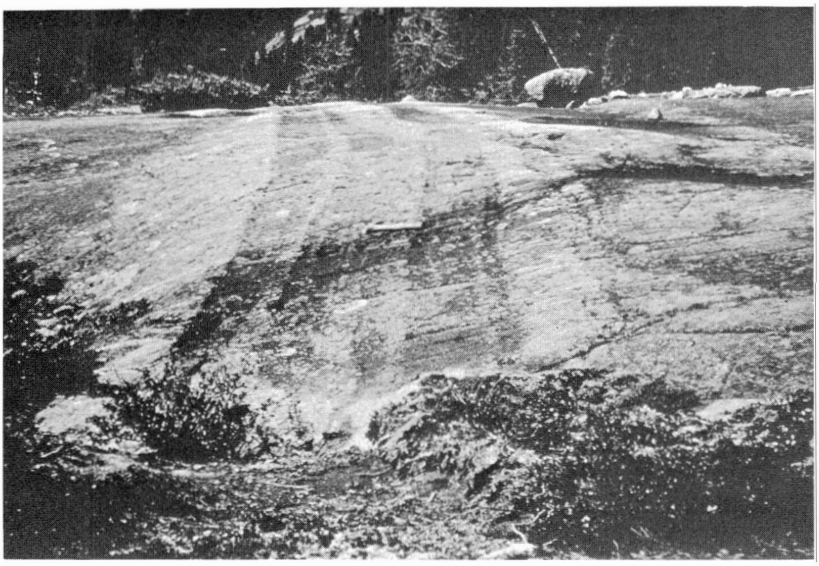


Plate XIIIA Trough-shaped interlayers of leucotroctolite and anorthosite. The upper layer of anorthosite (on the left) has a sharp base and a gradational top. Near locality XIB.

Plate XIIIB Weakly-developed layering in leucotroctolite. The layers dip gently to the left and away from the observer. About 2 kilometres north of Mosquito Lake.



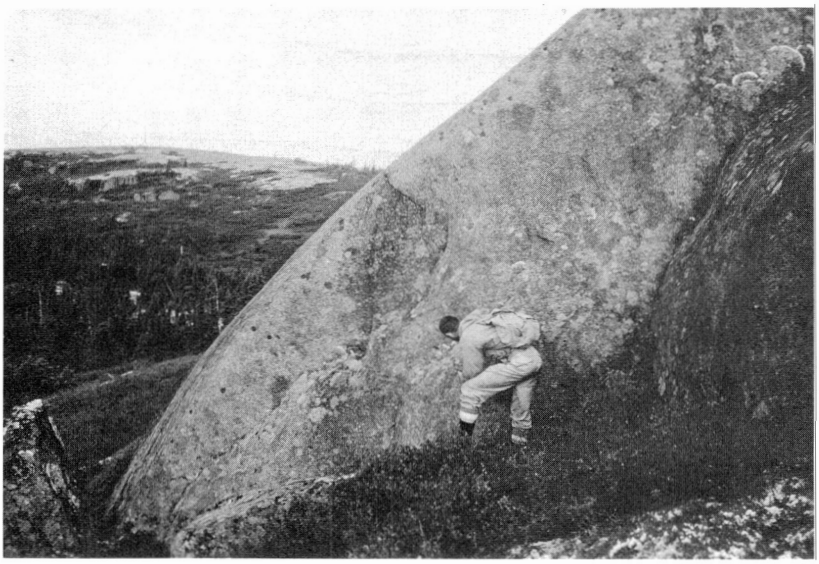
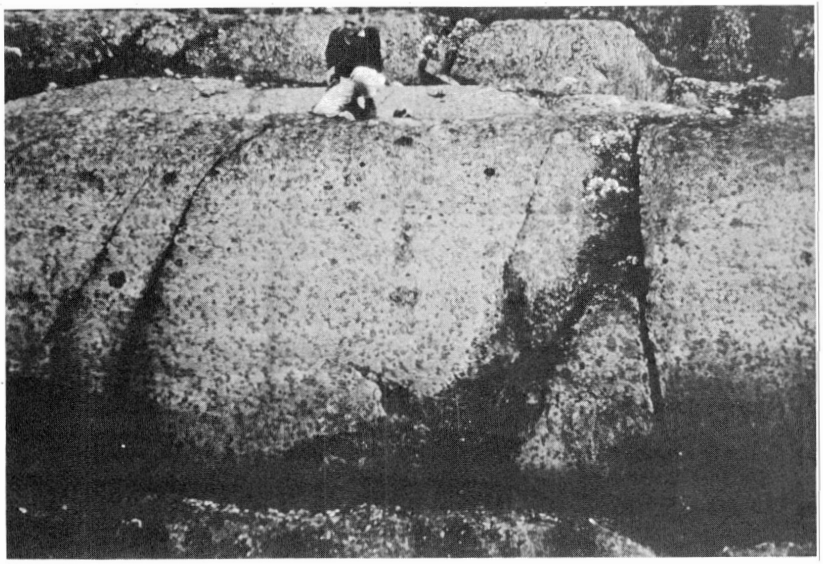


Plate XIVA Ledge of spotted leucotroctolite. Spots contain olivine, pyroxene and oxide minerals and intervening areas are essentially pure plagioclase. North of Bay of Islands.

Plate XIVB Closer view of ledge shown in XIVA. Note the uniform distribution of spots.



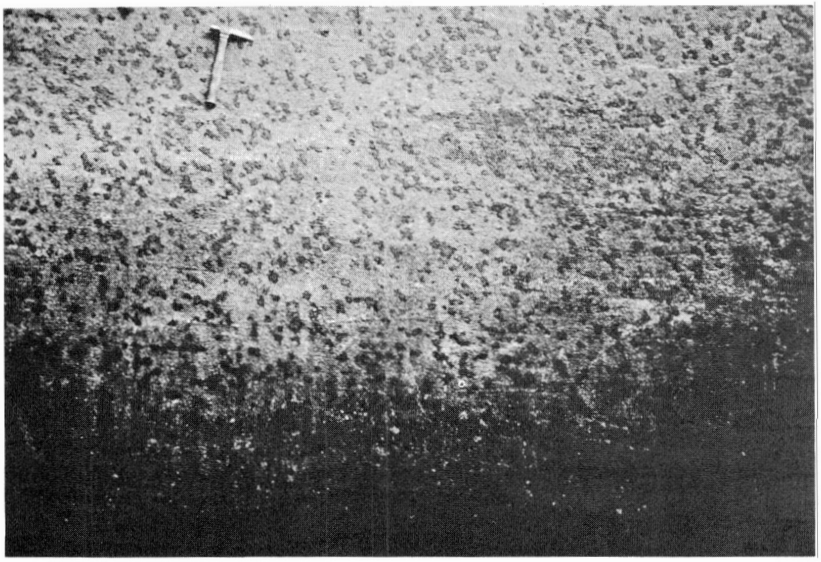


Plate XVA Highly irregular anorthositic mass in leucotroctolite. The origin of these masses is problematical; a possible explanation is that they were derived from pre-existing layers that were disrupted while still only partly consolidated. North of Mosquito Lake. Photo by R. Wares.

Plate XVB Abundant irregular "blobs" of anorthosite in leucotroctolite. Southwest of the Buttress.

Plate XVIA Isolated anorthosite mass with planar base and irregular top in leucotroctolite. South of Boot Lake.

Plate XVIB Resistant anorthosite layer in deeply weathered leucotroctolite. Northwest of Fraser Lake.

Plate XVIIA Cleavage face of large orthopyroxene oikocryst in coarse anorthosite. Distortion is caused by water over the surface of the outcrop. West shore of Shelter Bay.

Plate XVIIB Diffuse layers in the anorthosite zone. The hammer rests on a layer that is very slightly richer in plagioclase than the adjacent layers. West shore of Shelter Bay.

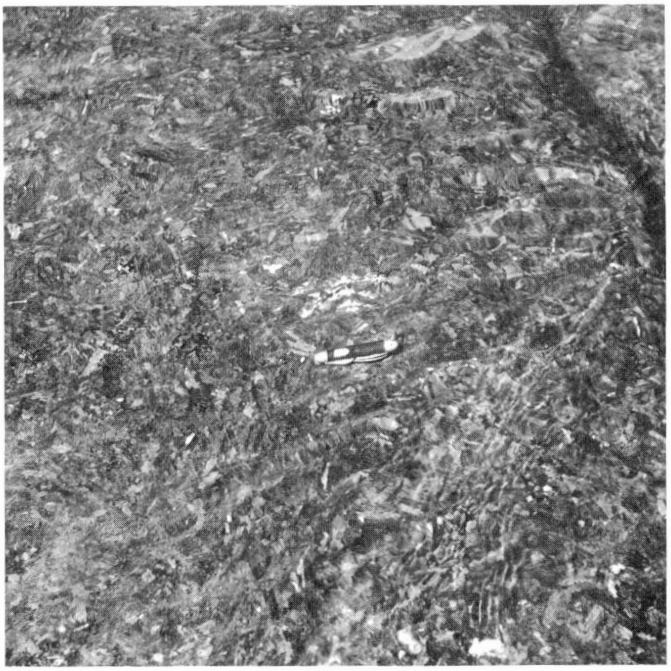
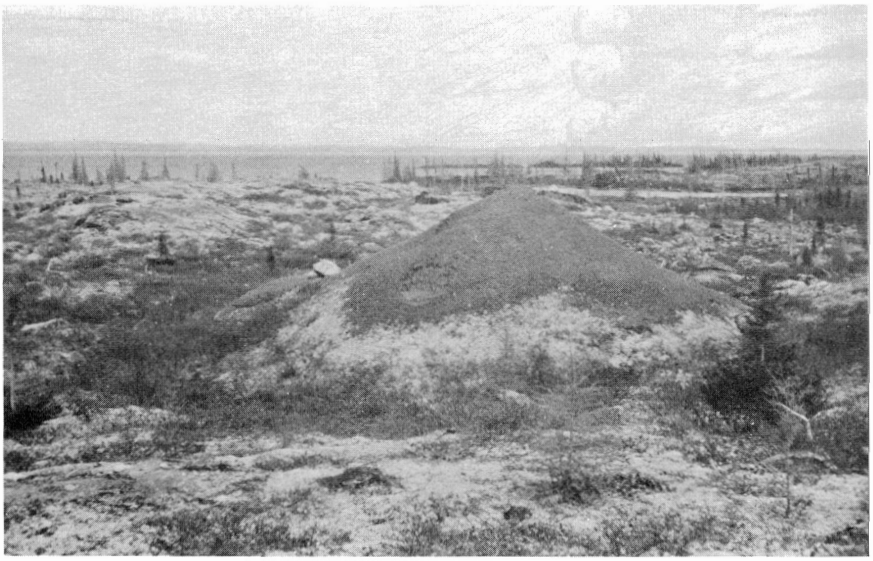


Plate XVIIIA Subophitic pyroxene patches in leucogabbro. West of Mosquito Lake. Photo by R. Wares.

Plate XVIIIB Layers due mainly to textural differences in the anorthosite zone. One thin layer has strongly planar-oriented plagioclase. Northwest of Last Lake.

Plate XIXA Strongly planar-oriented plagioclase crystals in the anorthosite zone. West shore of Shelter Bay.

Plate XIXB Large erratic of leucogabbro disintegrated into a cone of crystal gravel by weathering. South of Fault Bay.



The chief minerals are plagioclase, clinopyroxene, orthopyroxene (inverted pigeonite, at least in part) and olivine. Magnetite and ilmenite are present in variable amount. At some localities they are very abundant and occur concentrated in oxide-rich layers and heavy disseminations. A few flakes of bright biotite can be seen in some specimens and a small amount of brown amphibole is sometimes present. Scattered small sulphide grains, usually pyrrhotite and chalcopyrite, are not uncommon.

The boundary relations are of considerable importance in understanding the mechanism of intrusion and the early stages of crystallization of the magma. Brief descriptions are therefore given below of the principal localities where the geology of the marginal parts of the intrusion has been examined.

Description of localities at the margin of the intrusion

Magnetite Lake. At the west end of Magnetite Lake the marginal gabbros are notably rich in oxide minerals. The intrusion here is faulted against metatuffs and possibly also against sillimanite-cordierite gneisses and white quartzites; exposures close to the contact are sparse. None of the fine-grained banded gabbro was observed in the marginal rocks at this locality. Layers and lenses up to 10 centimetres thick very rich in oxides are present as well as zones up to a metre across that contain 30 to 50 per cent magnetite plus ilmenite. Analyses of magnetites from the oxide-rich rocks here show that they contain as much as 1.5 per cent V_2O_3. The outcrop width of the zone of oxide-bearing rocks is 300 metres or more and it bears further investigation as a possible commercially interesting source of vanadium.

Contact Lake. In the vicinity of Contact Lake the margin of the intrusion is faulted against paragneisses and amphibolite for a distance of about 2 kilometres. Most of the good outcrops are on islands in the lake and much of the lakeshore is low and boggy so that direct evidence of shearing is sparse. The abrupt truncation of units, however, makes faulting the most plausible explanation of relationships. South of the lake two northeast-trending faults are shown on the map. Abundant evidence of shearing is present along the easternmost fault about 1 kilometre south of Contact Lake.

The fine-grained border gabbro is streakily banded dipping northwest at 50 to 60 degrees. Most of the layering is caused by variations in plagioclase and ferromagnesian silicates but some thin layers rich in oxide minerals are also present. Individual layers are commonly 1 centimetre or less thick and lens out within a metre or so along strike.

Medium-grained olivine gabbro is gradational by increase in grain size from fine-grained border gabbro. It is generally massive although some irregular zones or ill-defined layers rich in oxide minerals are present.

Hook Lake. Between Contact Lake and Hook Lake several faults occur at and near the contact of the intrusion. Some of the best exposures of marginal gabbros are south of Hook Lake and it is unfortunate that faulting has obscured the relationships somewhat. In the two main outcrop areas a little more than a kilometre apart, two gabbro sills parallel the contact. In the northern outcrop area the one closest to the intrusion is about 15 metres thick and a

5-metre thick sill occurs a short distance farther from the contact. In the southern outcrop area a 6-metre sill is followed by a 3-metre sill away from the contact. These sills are identical in appearance to the banded marginal gabbro of the Michikamau Intrusion. Although the sills are generally concordant with banding in the enclosing sillimanite-cordierite gneiss, in detail small apophyses of gabbro can be seen injecting the gneiss and a few gneiss inclusions are found in the gabbro. Close to the contact of the large outcrop area about 1 1/2 kilometres southeast of Hook Lake hornblende is disseminated in the rock and concentrated in layers a centimetre or so thick. Passing inward from the contact hornblende decreases markedly in amount. Core taken with a portable diamond drill passed through the contact and it was found that the 25 centimetres of very fine grained 'chill' contains only anhydrous minerals adjacent to the sillimanite-cordierite gneiss. Marginal gabbro in the outcrop area just south of Hook Lake contains very little hornblende.

Fraser Lake. A wide fringe of marginal zone gabbro is present at the edge of the intrusion where it passes through Fraser Lake. Many of the outcrops are on islands, however, and it is not possible to trace variations in thickness in detail. Judging by the width of outcrop and recorded dips, the thickness of marginal gabbro in this part of the intrusion varies from about 450 to 900 metres. Fine-grained streakily banded gabbro is present at the contact where it passes through a group of islands at the west end of Fraser Lake. Passage into medium-grained massive gabbro occurs on the largest island. Although there is a gradual increase in grain size inward from the contact, there is also some interlayering of fine- and medium-grained gabbro. On two small islands northeast of the largest island the fine gabbro contains abundant thin lenses and streaks rich in shiny black hornblende. Fine-grained gabbro is exposed at only two other localities - both in the northeast part of Fraser Lake. Most of the remainder of the marginal zone is medium-grained to locally coarse-grained olivine gabbro. Scattered flakes of bright brown biotite can be seen in some specimens and up to several per cent of oxide minerals may be present. Rarely, a weakly developed layer can be seen but for the most part the rock is essentially massive.

North of Fraser Lake marginal gabbros continue to occur at the border of the intrusion. Both fine-banded and medium massive gabbros are present but due to poor exposure it was not possible to trace them as separate units at the scale of mapping.

Shallow Lake - Snow Lake. South of Shallow Lake there are several large and small dykes of fine- to medium-grained banded gabbro near the contact. They are remarkably similar to the marginal gabbros and it seems likely that they are related to them.

From Otter Lake north to Snow Lake and on toward Wolf Ridge sporadic outcrops of fine- and medium-grained gabbro are present at the contact. These are interpreted as forming a continuous, thin marginal zone. They are similar to the marginal gabbros elsewhere but here they mark the contact between leucogabbro and gabbro of the roof zone and wallrock gneisses rather than the boundary between the layered series and the wallrocks.

Boulder Lake. Southwest of Boulder Lake a fault slice contains an excellent succession from fine-grained, well-banded gabbro through medium-grained massive gabbro to medium- and coarse-grained pyroxene troctolites. Occasional thin (up to about 1 cm) layers rich in oxide minerals occur in the banded gabbros. It is clear from this succession that the small scale banding of the marginal gabbros gradually disappears as the grain size increases inward from the contact.

West shore of Lake Michikamau. On the west shore of Lake Michikamau marginal gabbros are exposed at the contact northwest of Petscapiskau Hill. Here, as around the northern lobe of the intrusion, the marginal zone is narrow and separates leucogabbro from wallrock paragneisses. Along the east side of Petscapiskau Hill only a few rusty weathered outcrops are present at intervals along the contact with sillimanite-cordierite gneisses. Passing southward, the contact is covered by water and low-lying boggy ground. Southeast of Turtleback Hill deeply weathered marginal gabbro is exposed here and there for about 1 kilometre along the contact.

Discussion

The thin 'chill' formed against sillimanite-cordierite gneisses suggests that the wallrocks, although cooler than the magma, were not cold. They may have been close to the maximum temperature indicated by the regional metamorphic mineral assemblages of the Petscapiskau Group (upper greenschist - lower almandine amphibolite facies) - say 400° C±.

The thin rhythmic layers commonly wedging out over short distances and sometimes truncated, presumably by scouring, suggest that strong currents were active in the magma. This is readily understood for in the initial period following intrusion strong temperature gradients near the walls must have been present. Due to these the degree of crystallinity of the magma near the walls may have been highly variable resulting in density currents that would move down dip under the influence of gravity. It is under such conditions that the fine-grained banded marginal gabbro is believed to have formed.

As the temperature gradients near the walls gradually became less steep due to heating of the wallrocks and deposition of an insulating layer of marginal gabbro, the intensity and variability of the currents probably waned. As steady-state cooling conditions gradually became dominant the more uniform medium-grained gabbro was deposited. It is not inferred that currents were absent at this stage but it is perhaps more likely that the rate of accumulation was sufficiently rapid that the ambient currents were not effective in providing a sorting mechanism.

In a funnel-shaped body with the dimensions of the Michikamau Intrusion it can confidently be inferred that the main direction of heat flow during cooling is upward through the roof. This being accepted, it follows that the likeliest site of crystal nucleation is the upper part of the intrusion. Following early deposition of fine-grained marginal gabbro which formed from crystals that may have nucleated largely in proximity to the walls of the intrusion, the main source of crystals was probably the upper levels of the intrusion. These crystals would then settle and travel down the depositional

slope (probably by a density current mechanism in which crystal-laden magma was able to displace lighter, crystal-poor or liquid magma) to be deposited.

Variations in thickness of marginal zone gabbros is not readily explained. They may be due to irregularities i. e. ridges and hollows in the contact, to changes in slope of the contact, or to lateral variations in current velocity or suspended load which resulted in irregular deposition or perhaps even erosion.

The erratic presence of hornblende in layers and disseminated through the rocks suggests that local conditions caused its formation. The magma seems to have been initially insufficiently hydrous to precipitate hornblende because it is absent from the 'chill'. Ingress of water to the magma from the wallrocks may have taken place locally, however, where fracturing took place after the initial 'chill' formed.

The Layered Series

The bulk of the exposed rocks of the Michikamau Intrusion have been assigned to the layered series. They are dominantly troctolitic in character although most are markedly feldspathic. Layers and zones within the sequence have sufficient pyroxene to be called gabbro but in proportion to the troctolitic rocks, these are not abundant.

Rocks of the layered series have a wide range of grain sizes. Grain sizes from 2 or 3 millimetres up to 2 or 3 centimetres are not uncommon but the most common size is in the range 5 to 10 millimetres. Plagioclase crystals up to 5 centimetres and even larger occur singly and in layers but these are exceptional.

On the basis of lithology, the layered series is divided into several subunits: troctolite, leucotroctolite, anorthosite layers in leucotroctolite, and layered gabbro. Troctolite forms a thick, layered sequence only in one area on the eastern margin of the intrusion, west of Shallow Lake. Leucotroctolite is by far the most abundant rock type of the layered series. Layered gabbro forms part of the layered series only in the Bay of Islands. Anorthosite layers are present in all of the above subunits but their maximum development is in the leucotroctolite where layers over 100 metres thick are not uncommon.

Troctolite

Troctolite is present in the lower part of the layered series, adjacent to the marginal zone, at several places in the intrusion. The main development of this subunit, however, is west of Shallow Lake. A rusty brown weathering habit is characteristic of these rocks and at this locality the broad north to northwest-trending ridge of troctolite with sparse vegetation is particularly striking. Besides the distinctive weathering, several other characteristics serve to distinguish the troctolite. Troctolite has a typically higher olivine content of 20 to 30 per cent in contrast to the 5 to 20 per cent that is more general in the leucotroctolite. Although coarser varieties occur, the grain size is most commonly about 2 to 3 millimetres - smaller than the

usual leucotroctolite mean grain diameter by a factor of 2 or more. Rhythmic layering on a scale of 2 to 10 centimetres is common and often well-developed (Pl. IIIB).

The northern and eastern boundaries of the Shallow Lake troctolite lie on strong lineaments that are interpreted as faults, chiefly on evidence along their extensions. The boundary on the southwest, however, appears to be gradational upward into leucotroctolite by increase in grain size and decrease in olivine content. The estimated thickness of the troctolite section here is about 1,600 metres.

The essential minerals are plagioclase and olivine accompanied by small amounts of clinopyroxene, orthopyroxene, and opaque oxides. Olivine is commonly relatively fresh but may be serpentinized to some degree producing a radial fracture pattern in the surrounding plagioclase. Olivine is more magnesian (Fo_{70}-Fo_{80}) and plagioclase more calcic (An_{65}-An_{72}) in troctolite than in leucotroctolite.

Troctolite is present in smaller amounts elsewhere in the intrusion. South of the Buttress several exposures indicate the presence of a lens of troctolite near the margin. About 3 kilometres south of Shallow Lake a small remnant of troctolite partly bounded by faults is present at the contact. About 3 kilometres southwest of Boulder Lake a sequence grades up from fine-grained banded marginal gabbro through medium and coarse uniform gabbro into troctolite. In addition, 5 kilometres east of Big Island Lake a wedge-shaped area of troctolite bounded by faults occurs within leucotroctolite. The relationship of this wedge to the leucotroctolite has not been established but a reasonable interpretation may be that the troctolite wedge is up-faulted with respect to its surroundings.

Leucotroctolite

This subunit of the layered series comprises about 70 per cent of the exposed rocks of the intrusion. Although layers with the extreme compositions of dunite and anorthosite, as well as some gabbroic layers, occur within the subunit, the great bulk of rock is conveniently described as leucotroctolite - that is, essentially a plagioclase-olivine rock with olivine typically present in amounts of 5 to 20 per cent. Other minerals are normally present in small amounts or are absent. They include orthopyroxene, clinopyroxene, Fe-Ti oxides, biotite, and hornblende.

Actual contacts of leucotroctolite with underlying troctolite have not been closely observed. Near the southwestern boundary of the Shallow Lake troctolite, layers of coarse leucotroctolite appear several hundred metres east of the mapped contact. It is believed likely, therefore, that interlayering or gradation is the normal relationship between these two subunits.

The upper boundary of leucotroctolite is against the anorthosite zone. Here again it is difficult to define the contact in many places to within less than a few hundred metres. Interlayering of leucotroctolite and anorthosite is common below the anorthosite zone on the west shore of Lake Michikamau. It is not entirely clear, however, whether the anorthosite interlayers should be regarded as the same as those that occur throughout the leucotroctolite or if they are precursors of the anorthosite zone. In any event, an objective boundary between the anorthosite zone and the leucotroctolite is drawn at the limit of olivine-bearing rocks.

Outcrops of leucotroctolite are normally not deeply weathered although strongly jointed exposures, and sometimes coarse-grained varieties, may be covered by a rubbly veneer. Weathered exposures are commonly light to medium brownish grey. With increased olivine content the colour becomes more rusty brown and increased plagioclase produces a greyer colour. Freshly broken rock is usually medium to dark grey or greenish grey. Because of the dark colour of the fresh plagioclase, estimates of colour index are extremely difficult to make on fresh surfaces but are readily done on weathered surfaces.

Plagioclase megacrysts 5 to 10 times the mean diameter of matrix plagioclase are commonly present. Only one or two crystals may appear in an outcrop and rarely do they comprise more than a few per cent of the rock.

Much of the rhythmic layering is subtly developed (Pls. IVA, IXA, XIIIB) with only the slightest differences in modal composition between layers. In many cases, given hand specimens of adjacent layers, it is virtually impossible to make a distinction by eye. However, the boundary between the same layers on the outcrop may be well-defined and their attitude readily measured. On clean glaciated outcrops some examples have been seen of low-dipping layers which are barely discernible even under close inspection, yet on the weathered vertical face of the same outcrop the layering is clearly visible. The common thickness of such layers is from about 30 centimetres to 1 metre but they range up to tens of metres and, although limitations of exposure prevent clear proof, traverses across strike suggest that layers over 100 metres thick occur in many places.

Extreme compositional variation in layers is rare. Anorthosite layers are common but, as the average rock contains at most perhaps 15 to 20 per cent olivine, the decrease in olivine required to produce anorthosite is slight. Further description of anorthosite layers in leucotroctolite is presented in the following section. Markedly olivine-enriched layers, on the other hand, are scarce. A few dunitic layers up to 1 metre thick have been observed at only one locality on the west shore of Lake Michikamau about 6 kilometres south of Camp Island (Pls. IVB to VIIIA). Partings and seams of dunite from 1 to 4 centimetres are more widespread but even these make up an infinitesimal part of the layered series. It is clear that the dominant departures from 'average' rock in the layered series is toward layers of more anorthositic composition.

Pods and irregular zones or ill-defined layers of coarse to very coarse grain size occur here and there in the leucotroctolite. They are not intrusive into the host rock but appear to be more of the nature of segregations.

Planar orientation of platy plagioclase crystals is widespread. Commonly a marked orientation is visible in the host leucotroctolite and an anorthosite layer in the same outcrop will have completely anhedral, equidimensional plagioclase. The reverse relationship has been observed but is uncommon. Even a layered section with only slight fluctuations in composition of successive layers may show marked planar plagioclase orientation in some layers and none in others, even though the shape and size of plagioclase crystals is identical.

Chatoyant plagioclase typically occurs in the anorthosite zone but at a few localities in the leucotroctolite some iridescent crystals have been observed. The blue is paler and more washed out looking than the intense colour commonly seen in the anorthosite zone plagioclase.

Because plagioclase is the principal mineral, its grain size and habit dominates the textural aspect of the rocks. Plagioclase varies in grain size from 2 millimetres to 3 centimetres but 5 millimetres to 2 centimetres is most common. Although olivine grains up to 10 centimetres have been observed, olivine is usually smaller than plagioclase in the same rock - 2 to 5 millimetres being the common range. Plagioclase varies in habit from plates with euhedral cross-section and nearly equal dimensions in the (010) plane, to completely anhedral equigranular grains. Marked changes in grain size and habit may occur from layer to layer. Olivine is commonly more irregular in shape, presumably due to overgrowth on initially settled grains. The irregularity of olivine shapes is the more apparent because it occurs as separated grains in a plagioclase matrix. Examples of truly poikilitic olivine crystals are rare (Pl. XIVA, XIVB).

Anorthosite layers in leucotroctolite

Anorthosite layers occur in all subunits of the layered series. However, they are most widely and strikingly developed in the leucotroctolite so that an additional brief account of their occurrence in that setting is warranted.

Anorthosite layers are more resistant to weathering than the enclosing leucotroctolite (Pl. XVIB). Consequently, in many outcrops they are readily observed because they stand above the general level of the outcrop surface. In some deeply weathered outcrops, fresh, well-preserved anorthosite layers retain their attitudes even though the enclosing leucotroctolite has been reduced to a rotten, slumping, crystal gravel. The greater weathering resistance of anorthosite layers has been taken into account in interpreting their thickness on the map where contacts are unexposed - a minimum thickness has been indicated in all cases.

Layers range in thickness from a few centimetres up to 200 metres or more. Thinner layers can commonly be seen to be wedge-like and where all or most of a layer is visible, a trough-like or lens shape convex downward is sometimes apparent. Some of these layers are thick in the axial region and thin rapidly to zero on the limbs.

The best information on the abundance of mappable anorthosite layers is from the south lobe of the intrusion. An approximately east-west band from the east margin of the intrusion between the Buttress and Contact Lake extending west to Lake Michikamau has been mapped by R. Wares at a scale of 1:25,000. The relatively greater number of anorthosite layers shown on the map in this area is believed to be a direct result of more detailed mapping and is probably a fair indication of the abundance of the larger anorthosite layers in leucotroctolite elsewhere in the intrusion.

The degree of development of anorthosite layering varies from zones of irregular anorthosite 'mottles' (Pl. XB) to sharply defined layers. Anorthosite layers with sharp contacts and those with gradational boundaries are about equally abundant. Layers may also have sharp lower boundaries and gradational tops and some examples with the reverse relationship have also been found. Examples of layering are shown in Plates XIA to XIIIA.

In general, there are significant textural differences between the anorthosite layers and anorthosite of the anorthosite zone. Grain sizes in the layers tend to be smaller - commonly in the range 0.5 to 1 centimetre. They are also more uniform in grain size as opposed to the generally poorly-sorted

sizes in the anorthosite zone. Planar orientation of plagioclase tablets is apparent in some layers but very commonly the plagioclase is equidimensional and the texture is xenomorphic-granular.

Many of the layers are pure, or nearly pure, plagioclase. Where mafic minerals are present, pyroxene is typical rather than olivine even though olivine is the chief ferromagnesian mineral of the host rock. The pyroxene normally occurs in small grains, clustered or widely dispersed. Poikilitic pyroxene 'spots' are present in some layers.

Layered gabbro

The rocks described under this section are found only in a limited area in the Bay of Islands. All of the outcrops are on islands in the outer bay. This unfortunately creates some difficulty in establishing the relationship of the rocks to those on the mainland.

Low-lying, wave-washed, glaciated outcrops provide some excellent fresh exposures. One of the distinctive features of the rocks is a light greenish brown cast seen over whole outcrop surfaces, a colour apparently largely due to plagioclase for it is evident even in very feldspathic varieties. Cores taken with a portable diamond drill show that fresh rock is medium to dark olive green or greenish grey.

Plagioclase and pyroxene are the chief minerals and are accompanied by lesser amounts of magnetite, ilmenite, and olivine. Although layers containing numerous plagioclase tablets up to 2 centimetres by 0.5 centimetre are not uncommon, the grain sizes in most of the rocks are consistently finer (2 to 5 mm) than is typical of the underlying leucotroctolites. Pyroxene has a generally more granular habit differing from its more common subophitic to poikilitic form characteristic of other rocks of the layered series. This suggests that it has become a primary precipitate rather than having grown interstitially. The colour index of many of the rocks is sufficiently low (20 and lower) that they can well be termed leucocratic.

Rhythmic layers occur at intervals and range from strongly contrasting compositions to others only slightly different from the host rock. There are some anorthosite layers over 1 metre thick but much layering is on a scale of 1 to 8 centimetres. Those layers enriched in pyroxene and oxides are most conspicuous and they commonly occur in groups, together comprising a thickness of 1 to 3 metres. The tops of some layers are graded but both top and bottom of many layers are equally sharply defined. Some thin layers can be seen to pinch out along strike but many layers are very regular and constant in thickness across the outcrop.

Some layers contain abundant euhedral plagioclase tablets. In some cases they are completely unoriented and in others they show a strong subparallelism to the enclosing layer. On the whole, igneous lamination is not a widespread feature of these rocks.

Discussion

Although the troctolite subunit has relatively limited exposure and its boundary relationships are complicated by faulting, there is little doubt about its position within the layered series. Where faulting has not obscured

the relations, it is apparent that troctolite underlies leucotroctolite and thus forms the lowermost exposed part of the layered series. This interpretation is further supported by the fact that olivine is more magnesian and plagioclase more calcic than higher members of the layered series. It is also known that troctolite is not everywhere present beneath leucotroctolite - that is, lying between leucotroctolite and the marginal zone. Stated another way, troctolite is not everywhere present at the base of the layered series but where it exists, its position is between leucotroctolite and the marginal zone. This suggests that troctolite deposition was confined to the central, deeper parts of the funnel-shaped chamber and that it was overlapped at higher levels by leucotroctolite deposition.

Leucotroctolite forms the bulk of the exposed rocks and considering the structure of the intrusion it must also have been the major unit of the original body. Rhythmic layering is by no means a universal feature of the rocks but it is widespread. This fact, together with consideration of the nature of the layers (which commonly wedge out over short distances, have lens and trough-like cross-sections, and graded tops) all point to the existence of currents in the magma from which crystals settled onto a depositional surface.

The most likely causes of the widespread plagioclase lamination are: (1) that subparallelism was produced by compaction of the crystal mush during burial, or (2) that the crystals settled initially with their short axes perpendicular to the depositional surface. The first explanation is unlikely to be universally applicable because it fails to explain orientation in some layers and its absence in adjoining layers of similar composition. Settling from a magma current of variable velocity, or alternatively, variable rates of accumulation from a current of constant velocity, are other possible ways of producing a succession of layers with variable degrees of plagioclase orientation. If a linear fabric could be demonstrated to exist, the argument for magma currents would be further strengthened. Unfortunately, plagioclase is typically platy rather than elongate and field observation did not discover a measurable lineation although it may be revealed by careful petrofabric studies.

Because all of the layered gabbro outcrops are on islands and there is faulting nearby, some uncertainty is attached to the interpretation of the relationship of these rocks to the layered series. Attitudes of layering in the layered gabbro sequence, although more steeply-dipping, are consistent in strike with the basin structure of the south lobe. Some of the section may be faulted out but there is sufficient information to judge with reasonable confidence that the layered gabbros are continuous with the underlying leucotroctolite. The existence of a continuous, conformable layered sequence from troctolite through layered gabbro has an important bearing on interpretation of the crystallization of the intrusion. Elsewhere in the intrusion layered gabbros do not appear between leucotroctolite and the anorthosite zone. The significance of these relations is further explored in a later section on interpretation.

The Anorthosite Zone

The rocks of the anorthosite zone are some of the most striking in appearance. The general coarse to very coarse grain sizes together with widespread chatoyant plagioclase contribute to the distinctive character of the rock.

Apart from essential plagioclase, other minerals present in the anorthosite zone are orthopyroxene, clinopyroxene, magnetite, ilmenite, apatite, and the odd sulphide grain. These minerals are invariably interstitial in habit toward plagioclase. In locally altered zones secondary minerals include green amphibole, chlorite, sericite, kaolinite(?), and zeolites. The great majority of rocks included in the anorthosite zone are properly designated as anorthosite - that is, they have a colour index of 10 or less. It is not uncommon however, to find imperfectly developed layers, irregular zones, and pods of leucogabbro here and there within the anorthosite zone.

Outcrops of anorthosite are usually remarkably fresh and unweathered with only a thin veneer of weathered products coating the feldspar. In very white-weathering outcrops the plagioclase is invariably strongly altered and opaque. Ferromagnesian minerals are oxidized to some degree and rusty patches may occur around local concentrations of these minerals. The colour of the fresh rock is dominated by plagioclase which is commonly bluish grey to dark grey. At some localities plagioclase is light grey and in a number of the westernmost exposures west of Last Lake, plagioclase has a definite pink colour.

Pyroxene is commonly erratically distributed in small or large subophitic nests and poikilitic crystals from the size of a quarter up to 20 centimetres or more across (Pl. XVIIA and XVIIIA).

The grain size of much of the rock is in the range 1 to 3 centimetres. However, numerous plagioclase crystals from 5 millimetres to 6 centimetres or larger are generally present in an outcrop so that, on the whole, size sorting is poorly developed. There are, on the other hand, examples of layers whose chief means of definition is that they are better sorted than the host rock that contains them. Layers can also be defined on the basis that they are coarser, finer, have different plagioclase habits, or a more strongly oriented plagioclase fabric than the enclosing rock (Pl. XVIIB, XVIIIB, XIXA). It should be emphasized that the layering that occurs sporadically in the anorthosite zone is not at all similar to the rhythmic layering of layered series. Most are rather irregular and ill-defined and might better be referred to as zones rather than layers.

Individual plagioclase crystals up to 50 centimetres across that occur singly or in groups of a few individuals may be encountered in almost any outcrop. In addition, segregations in the form of irregular zones of pods of extremely coarse plagioclase occur sporadically. These segregations sometimes also contain coarse pyroxene crystals up to 25 centimetres across.

One of the most distinctive characteristics of the anorthosite zone is the widespread presence of chatoyant plagioclase. A few individuals showing blue iridescence are readily visible in many outcrops and careful examination would probably reveal them in nearly all. Particularly noticeable in larger crystals is a colour-zoned iridescence in which a bronze core grades outward into a green rim followed by a blue rim.

Nearly equidimensional plagioclase crystals are most common but tabular crystal habits comprise a small to large proportion of many outcrops.

They may be quite unoriented but numerous measurements have been made on moderately to strongly developed igneous lamination.

As already noted, the nature of the lower boundary of the anorthosite zone is not clearly understood. On the west shore of Lake Michikamau, where we are apparently presented with a tilted cross-section of the intrusion, the gross structure of the anorthosite zone is discordant with the overall trend of layering in the layered series. However, near the boundary between the two units, layering in the leucotroctolite tends to become subparallel to the anorthosite zone. Faults separating anorthosite and leucotroctolite near the west end of Last Lake are largely inferred to explain the discordancy in attitudes between the two units. The faults may not exist and the discordance may be a real feature of the rocks (a possible explanation for the discordance is discussed in a later section on interpretation).

The upper boundary of the anorthosite zone is a graded transition into upper zone leucogabbros. This transition is most clearly evident in the north lobe where traverses inward from the margin to the anorthosite show the continuous change from gabbro at the margin, through leucogabbro with decreasing colour index, into anorthosite. The variation in colour index is imperceptible from outcrop to outcrop but is readily apparent in traversing the whole section.

Discussion

The anorthosite zone is one of the most remarkable units of the intrusion. On the west shore of Lake Michikamau where the anorthosite zone stands vertically, it is at least 3,500 metres, and may be as much as 4,500 metres thick. There is no way of being certain that the anorthosite zone once extended over the whole of the intrusion but because it outcrops in widely separated parts of the intrusion the assumption that it did is considered reliable. It is also not possible to establish whether the anorthosite zone was relatively uniform in thickness over the intrusion or if it was highly variable. Lacking better data, it is assumed that the thickness of the west shore anorthosite section represents a reasonable approximation of the thickness of a once widespread anorthosite zone.

It was pointed out that layering of a sort and igneous lamination are present in the anorthosite zone. The presence of these structures plus the extreme plagioclase-rich composition of the rocks is evidence supporting an origin by accumulation of plagioclase crystals. Plagioclase of the anorthosite zone overlaps in composition plagioclase from the upper part of the layered series and a possible conclusion is that the anorthosite zone formed by continuous plagioclase settling on top of the layered series (Emslie, in press). However, more recent investigations have brought new information to light. The reasons for believing that the Bay of Islands layered gabbros are a sequence continuous with the layered series have already been discussed. The existence of these layered gabbros, although limited to one locality, argues against an interpretation that transition from leucotroctolite to anorthosite took place simply by cessation of crystallization of olivine.

Upper Border Gabbros

The main area of exposure of upper border leucogabbro and gabbro is in the north lobe of the Michikamau Intrusion. Here they rim the central area of anorthosite and grade almost imperceptibly into it - the boundary being drawn where the colour index increases below 10. A smaller area of upper zone rocks is present on the west shore of Lake Michikamau where its relationship to the anorthosite is similar to that in the north lobe.

The leucogabbro and gabbro, except for increased colour index, are similar in appearance to rocks of the anorthosite zone. Grain size is commonly in the range 5 millimetres to 2 centimetres and plagioclase megacrysts of 4 to 6 centimetres and larger are scattered through the rock.

Outcrops of leucogabbro are somewhat more weathered, on the whole, than anorthosite - presumably due chiefly to the increased pyroxene content. A light rusty brown weathered surface is common. Some erratics of leucogabbro have been completely disintegrated to crystal gravel and remain as cones resembling giant ant hills (Pl. XIXB). Excellent examples occur on the peninsula south of Fault Bay.

The essential minerals of the upper zone rocks are plagioclase, orthopyroxene, and clinopyroxene. These are accompanied by small amounts of magnetite, ilmenite, and apatite. Orthopyroxene is generally dominant over clinopyroxene and in some rocks clinopyroxene is virtually absent so that they could properly be called norites or leuconorites. The colour index varies from 10 to about 35 but values greater than 20 are largely confined to the outer one-quarter to one-third of the upper zone gabbro rim in the north lobe.

The mafic minerals are invariably interstitial in habit and subophitic to poikilitic textures are common. Measurable planar plagioclase orientation is present in a number of places. Where it is most strongly developed the rocks are commonly more uniform in grain size and some plagioclase is so well-sorted and packed that it resembles brickwork.

Some rude composition layering has been observed locally and where comparison can be made it is consistent with attitudes of igneous lamination. Its use as a source of structural information is limited because of restricted development.

In the north lobe of the intrusion a number of traverses have crossed the upper zone gabbros from the anorthosite to the wallrocks. In all of these traverses, the decrease in colour index together with the increase in grain size is evident even though changes from outcrop to outcrop are virtually imperceptible.

Discussion

The anorthosite zone appears to grade laterally as well as vertically into upper zone leucogabbro. This is seen where the anorthosite zone approaches the wall of the intrusion east of Petscapiskau Hill on the west shore of Lake Michikamau. There is a gradual but marked increase in colour index of the rocks on the mainland toward the west and they could properly be designated leucogabbro. The main development of leucogabbro on the west shore does, however, overlie the anorthosite zone. The domed structure of

the north lobe of the intrusion also suggests that the gabbro may be interpreted as a shell roofing the anorthosite zone.

Where the contacts are exposed, a rind of fine-grained to medium-grained olivine gabbro separates the upper zone gabbros from quartzofeldspathic gneiss wallrock. Only a few such contacts have been observed but these are considered to be sufficient to infer that the marginal gabbro rind is a normal feature of the upper walls and probably extended over much, if not all, of the roof.

Transgressive Group

Strongly iron-enriched late differentiates of the Michikamau Intrusion occur in two main bodies widely separated in the intrusion. The larger body occurs along the northeast margin of the north lobe. A smaller mass is on the west shore of Lake Michikamau in the vicinity of Fault Bay. Both are intrusive into upper zone leucogabbros and occur at and close to the boundary between the intrusion and the country rocks. The rocks in these masses have a number of similarities but also some important differences in mineralogy.

North lobe

The large lens-shaped body in the north lobe consists mainly of ferrogranodiorite that grades in places into ferroadamellite. Dykes and irregular small intrusive masses of ferrogranodiorite are intrusive into leucogabbro south and east of Wolf Ridge.

Outcrops weather a rusty buff to medium rusty brown. Many outcrops are deeply weathered and it is nearly impossible to get a fresh specimen with a hammer. However, it is not difficult to find glaciated surfaces that are weathered to depths of only a few centimetres from which fresh samples can readily be obtained with a portable diamond drill. The fresh rock is markedly green or greenish grey.

The principal minerals are antiperthitic plagioclase, orthopyroxene, clinopyroxene, and variable amounts of mesoperthite, quartz, and Fe-Ti oxides. In addition, small amounts of brown biotite and shiny black hornblende are not uncommonly visible in hand specimens. The colour index of much of the rock is about 30 to 35.

Fairly uniform grain size and texture are characteristic of most outcrops. Grain sizes of 2 to 5 millimetres are common but a few crystals up to 1 centimetre or more of plagioclase and mesoperthite may be found here and there. In a few places, diffuse streaks enriched in pyroxene and oxide minerals that may represent incipient layering have been noted.

West shore of Lake Michikamau

The body of ferrous-rich rocks in the vicinity of Fault Bay contains rocks ranging from ferrodiorite to ferromonzonite and locally ferrosyenite. The bulk of the exposed rock is ferromonzonite. Ferrodiorite occurs as a marginal facies of ferromonzonite along the south shore of Fault Bay at the

contact against leucogabbro. It is finer (~0. 5mm) at the contact and increases to 2-3 millimetres away from the leucogabbro. Ferrodiorite also occurs at the margins of the large ferromonzonite dyke on Dyke Island.

The principal minerals are antiperthitic plagioclase, mesoperthite, clinopyroxene, olivine, and Fe-Ti oxides. A few shiny black poikilitic hornblende crystals occur here and there and a little biotite is present locally. Quartz occurs but is never present in amounts exceeding a few per cent. A colour index of about 20 is common in ferromonzonite and may reach 35 in ferrodiorite.

Many outcrops are very deeply weathered - sometimes to a deep red-brown, slumping, rotten heap. More commonly a buff or rusty buff colour characterizes the outcrop surface. Fresh rock has a strong green cast and, particularly in drill core, golden yellow fayalitic olivine grains set in the greenish matrix present a very striking appearance.

The feldspars commonly have grain sizes of 0. 5 to 1 centimetre but plagioclase crystals over 2 centimetres are found locally. The mafic minerals are typically smaller - about 1 to 2 millimetres.

Small dykes of ferromonzonite are present in leucogabbro near the south boundary of the main body. A number of dykes and irregular intrusive masses are present in the leucogabbro up to 400 metres south of the main body. Several large blocks of leucogabbro occur in the ferromonzonite within the main body.

Discussion

The rocks of the transgressive group are clearly the youngest members of the Michikamau Intrusion. They are the only rocks that show transgressive relations to other members of the intrusion and there is no reason to believe that crystal accumulation played a large role in their development.

The transgressive nature of the rocks indicates that some event took place prior to their crystallization which led to movement of the iron-enriched differentiated magma into parts of the already crystallized intrusion that it had not formerly occupied. A fine-grained ferrodiorite facies that is present against leucogabbro along the south shore of Fault Bay suggests a chill zone. If so, this implies that the leucogabbro, although undoubtedly not cold, was at a temperature sufficiently lower than the iron-rich magma to cause rapid crystallization close to the contact. It may be assumed, therefore, that the iron-rich magma was brought from a substantial distance away from its present site in the intrusion.

Inclusions

Inclusions within rocks of the Michikamau Intrusion are rare. Small inclusions of the wallrock are present in marginal gabbro adjacent to the contact with cordierite gneiss south of Hook Lake and northwest of Petscapiskau Hill. These are rarely more than 10 centimetres on a side and their parentage is readily recognizable.

A very large inclusion is contained in leucogabbro on the peninsula south of Fault Bay. It is composite, consisting of vitreous, purplish, cordierite-rich gneiss in sharp contact with fine- to medium-grained marginal

gabbro. Its position suggests that it was scaled off either the roof or upper walls of the intrusion, carried into the leucogabbro, and was trapped there when it solidified. If the inclusion was once part of the roof it would provide evidence that the cordierite gneiss did cover the intrusion and that marginal gabbros extended over at least part of the roof.

On the south shore of Contact Lake, just west of the marginal gabbro-leucotroctolite boundary, there is one large and several smaller inclusions of pyroxenite. They clearly have a recrystallized fabric and consist predominantly of orthopyroxene and clinopyroxene. The grain size of the pyroxenite is variable from 2 millimetres to 1 centimetre.

Five and one-half kilometres due west from the south end of Blackbear Lake, there is a large outcrop of dark brown to black orthopyroxenite. It consists chiefly of strongly pleochroic orthopyroxene with a little interstitial plagioclase. On the north-pointing peninsula in the lake adjoining the south end of Big Island Lake, another large mass of very similar dark brown to black orthopyroxenite occurs in leucotroctolite. It is not possible to be certain whether these orthopyroxenites are really inclusions or dykes.

For an intrusion of the size of Michikamau it is remarkable that inclusions are so scarce. Forceful injection and stoping cannot have played an important role in emplacement of the intrusion or one would expect to see more evidence of inclusions, particularly near the walls.

Dykes and Sills

Only those dykes and sills known or believed to be related to the Michikamau Intrusion are considered in this section.

Sills of fine- to medium-grained gabbro occur in sillimanite-cordierite gneiss close to the contact south of Hook Lake. The close association with the contact plus their similarity to the marginal gabbro of the intrusion leaves little doubt that their emplacement accompanied the main intrusion.

A dyke at least 15 metres wide cuts quartzofeldspathic gneiss close to the contact northwest of Loon Lake. It is composed of rusty weathering fine- to medium-grained gabbro. About 2 kilometres south of Shallow Lake a thick gabbro dyke occurs in quartzofeldspathic gneiss close to the contact. On the north side of the dyke two small fine gabbro dykes strike subparallel to the main dyke and dip about 90 degrees. About 1 1/2 kilometres south of this locality a steeply-dipping 2-metre, fine-grained gabbro dyke intrudes quartzofeldspathic gneiss. The gabbro in all of these dykes is remarkably like the marginal gabbro of the intrusion. The small dykes and the margins of the larger dykes have grain sizes of 0.5 to 1 millimetre whereas in the central part of the larger dykes it is 2 to 3 millimetres.

The relative scarcity of dykes in the wallrocks of the intrusion suggests that its emplacement was not a catastrophic event involving widespread fracturing of the crust. Rather, it implies that inflation of the magma chamber through one or more central feeders was accommodated by gradual adjustments in the surrounding crust. The paucity of open fractures also points to emplacement at a relatively deep crustal level.

A number of other dykes occur within the intrusion. Most are less than a metre thick and have grain sizes of 0.5 to 2 millimetres. In the layered series small, fine-grained olivine gabbro dykes occur 2 kilometres

south of Big Island Lake and 1 kilometre south of Blackbear Lake. An 8-metre gabbro dyke occurs 6 kilometres west of Blackbear Lake. All of these are steeply-dipping to vertical.

In the anorthosite zone, a large ferrodiorite-ferromonzonite dyke occurs on Dyke Island and at least one smaller dyke is subparallel to it. South of Trout Channel, three dykes of ferrodiorite, the largest 1 1/2 metres thick, cut anorthosite.

In upper zone leucogabbro two small dykes of brown-weathering ferroadamellite were found on the south and southwest margins of Wolf Ridge. Similar small dykes were found about 1 kilometre southwest of Otter Lake and about 6 kilometres northwest of Boot Lake.

It is difficult to draw significant conclusions from the small sampling of dykes. Dykes of gabbroic composition were found only in the layered series. A possible explanation is that during accumulation of these rocks, seismic activity opened local fractures in already solidified rock - or nearly solidified crystal mush - into which the superjacent magma was able to penetrate. Dykes similar in composition to the late differentiates are found in the anorthosite zone and the upper zone leucogabbro indicating that events causing their injection occurred late in the crystallization of the intrusion. Assuming that all of these dykes were formed by injection of essentially liquid magma at different times, their compositions provide a means of estimating the variation in magma composition during crystallization of the intrusion (Emslie, in press).

FAULTS

Faults have played an important role in developing the present surface expression of the Michikamau Intrusion. Their traces are reflected in the topography and this has been used to advantage in extrapolation. All of the inferred faults shown on the map, however, are based on evidence other than simply the presence of lineaments on air photographs. One or more of the following criteria are applicable to each of the faults indicated on the map:

1. presence of shearing or brecciation.
2. presence of localized strong alteration e. g. chloritization of mafic minerals, zeolitization of plagioclase, etc.
3. localized deformation of layering.
4. juxtaposition of units that are known to be not normally in contact.
5. abrupt and marked changes in attitude of layered sections.
6. horizontal displacement of contacts.
7. abrupt disappearance of layered sections along strike.

In addition to the faults indicated on the map, there are numerous other topographic lineaments for which no evidence is available. Some are likely to be faults but others are simply joint systems. Where evidence is lacking, no interpretation has been attempted. Some of the fault interpretations are discussed below.

DESCRIPTION OF MAJOR FAULTS

Michikamau Lake Fault

Evidence for the existence of this fault is indirect but nevertheless compelling. The surface trace of the fault is indicated as lying beneath the lake. Outcrops lying well out in the low ground 6 kilometres northeast of Last Lake are cataclastic, sheared and fractured. A number of outcrops in the group of small islands northwest of Stormy Island are strongly altered and fractured. On the west shore of Lake Michikamau a few outcrops at the eastern end of Trout Channel have patchy white alteration and a thin section examination shows that some plagioclase crystals are bent and slightly granulated at their margins. Farther south along the west shore, fracturing and severe alteration has been found on offshore islands.

Although the evidence cited above provides support for a fault zone underlying the lake, the strongest evidence is based on the layered structure of the intrusion. All of the layering on the west shore of Lake Michikamau is remarkably steep. Dips are vertical or nearly so over much of the northern part and gradually decrease to about 50 degrees northeast at the southern end. Nowhere else in the entire intrusion do such steep dips persist over so large an area. Layering in this west shore section is identical in character to that seen elsewhere in the intrusion and excellent examples of graded layers with tops facing northeast imply bottom accumulation of crystals. There can be no question of accounting for the steep attitudes of layering by flowage of a crystal mush or some other mechanism different from the rest of the intrusion. It is concluded, therefore, that major displacement on a fault passing through Lake Michikamau is the most satisfactory interpretation. If the anorthosite zone is assumed to have been subhorizontal, rotation of the west shore section relative to the remainder of the intrusion must have taken place. An easterly-dipping fault seems most likely on which the west shore section was rotated about an axis perpendicular to the fault plane. The south end of the west shore section is thus believed to have once underlain the central part of the intrusion at great depth. This interpretation views the west shore section as a vertical or steeply-dipping partial cross-section of the intrusion as indicated in Figure 4.

Esker Fault

The name derives from the large, discontinuous esker that follows the topographic lineament occupied by the fault for much of its length. Along the northwest part of the fault several contacts show horizontal displacements. A large outcrop area adjacent to the fault about midway between Boot Lake and Sandy Lake has much disturbed and deformed layering. The fault trace is remarkably straight as are most faults within the intrusion, indicating that steep to vertical fault planes are the rule. The horizontal displacement of contacts at the northwest end of Esker Fault show that the northeast block has moved up relative to the southwest block. The amount of vertical displacement at the northwest end is unlikely to be large because the dip of the contact between the upper zone gabbro and the anorthosite zone apparently is low.

Shallow Lake Fault

This fault extends from at least as far east as Shallow Lake and strikes northwest to join Esker Fault at Boot Lake. There is abundant disruption and deformation of anorthosite layers in leucotroctolite in a large outcrop area about 5 kilometres east-southeast of Boot Lake. This fault bounds much of the north lobe of the intrusion, separating it from the main body. The amount of vertical displacement on the fault must be exceptionally large because troctolites of the lower layered series have been brought to the same level as upper zone gabbros. A small wedge of upper zone gabbro is in fault contact with troctolite along the north edge of the main troctolite subunit.

Spot Lake Fault

There is evidence of vertical shearing and cataclasis in two outcrops about 2 kilometres apart at the eastern end of Spot Lake. Layering dips steeply south to the north of the lake; south of the lake dips are 30 to 35 degrees easterly. This strong discordance of attitudes on each side of the fault continues nearly to the eastern margin of the intrusion.

OTHER FAULTS

Other faults within the intrusion will not be discussed individually. Many of them occur entirely within the leucotroctolite which has no distinctive horizons that would be useful for estimating displacements. Most faults are remarkably straight or slightly curved indicating that they are steeply dipping or vertical. The faulting seems to have been mainly in the nature of vertical adjustments between adjacent blocks accompanied by more or less tilting which resulted in the present block mosaic pattern.

Examination of air photographs of the country surrounding the intrusion does not reveal a pattern of lineaments resembling that within the intrusion. The fault pattern over the intrusion as a whole appears almost 'random' rather than having well-defined persistent strikes such as might be produced by a regional stress field. These two facts may be interpreted to mean that faulting in the intrusion is largely the result of stresses related to its emplacement. This argument is expanded in a following section.

INTERPRETATION AND SYNTHESIS

OBSERVATIONS ON THE CHARACTER OF THE MAGMA

Considering the high proportion of plagioclase in the intrusion there can be little doubt that the magma was aluminous. This was not borne out by early analyses of fine-grained marginal gabbro that was originally thought to be chilled magma (Emslie, 1965; p. 394, analyses 1 and 2). These marginal gabbros are now known to have formed, in part at least, by crystal accumulation and thus are not representative of a liquid composition. A recent analysis of a sample of the true chill zone obtained by drilling has 18.33 per cent

Al_2O_3 and may be regarded as an aluminous basalt composition. Although the higher alumina content eases the problem, it is still too low to account for all of the plagioclase in the intrusion. If the chill zone rock is regarded as a sample of the liquid portion of the magma from which the Michikamau Intrusion crystallized, we are forced to conclude that a solid portion rich in alumina (probably plagioclase crystals) also existed contemporaneously. Whether an initially aluminous liquid existed at depth and crystallized a large amount of plagioclase during its rise to the magma chamber and consequent decompression, or whether a liquid-plagioclase crystal suspension existed even at depth, is not known. It is difficult, however, to escape the conclusion that the initial magma filling the chamber was a suspension of plagioclase crystals in a basaltic (probably aluminous) liquid.

EMPLACEMENT OF THE INTRUSION

Evidence has been presented which indicates that the walls of the Michikamau Intrusion dip inward everywhere except perhaps in the upper levels approaching the roof. The original form of the intrusion, although now much complicated by faulting, was therefore probably funnel-like and elongated in a north-south direction.

The walls of the intrusion are composed in part of the Petscapiskau Group and in part of the quartzofeldspathic gneiss complex. Not much more than about one kilometre of Petscapiskau Group rocks is exposed at any place around the intrusion. The Petscapiskau Group is presumed to have been deposited upon the gneiss complex. At all localities where the Petscapiskau Group is exposed but most notably between Fraser Lake and Lake Michikamau, attitudes in the Petscapiskau Group are subparallel to the walls of the intrusion. This relationship suggests that the Petscapiskau Group, or at least that part of it now remaining, was not deformed by large scale folding before the Michikamau Intrusion was emplaced. It is therefore concluded that the intrusion was emplaced at and near a surface (former unconformity) separating a supracrustal sequence (the Petscapiskau Group) and a basement gneiss complex (the quartzofeldspathic gneiss). The surface may have been folded slightly but major deformation seems unlikely.

Upon reaching this boundary surface, the rising magma spread laterally. In some places it penetrated the overlying Petscapiskau Group for distances of a few hundred metres up to a kilometre before spreading laterally. A gently domed roof is suggested by the structure of the north lobe of the intrusion and it is inferred that the roof rocks were lifted somewhat by the inflating magma chamber below them rather than being stoped and injected by the rising magma. However, considering the immense thickness of the intrusion, most of the room required for the magma chamber is likely to have been made available through downwarping of the floor that took place as the chamber was inflated.

The view that room for the intrusion was created largely by downward displacement of the floor requires some comment. There can be little doubt that the source of the Michikamau magma was within the mantle thus the enormous volume of magma injected into the crust to form a relatively localized intrusion can only have resulted in a concentrated loading of the crust. The stresses resulting from loading are expected to be most intense in the region beneath the thickest part of the intrusion. In the formation of a

centrally fed chamber such as the Michikamau Intrusion, the thickness from inception, would always be a maximum in the vicinity of the feeder. If the greatest crustal deflection takes place beneath the maximum vertical load, the expected result is a funnel-shaped cross-section. Such evidence as is available suggests that the source of basic magma is at depths of 60 to 100 kilometres, which places it at no great distance below the base of the crust. If the source region was localized beneath the intrusion, tapping of magma from it would cause a decrease in volume which may facilitate local crustal sagging. The process may be regarded as one of replacement in which the volume occupied by the intrusion within the crust was created by contemporaneous depression of the lower crust to replace the magma withdrawn from the upper mantle.

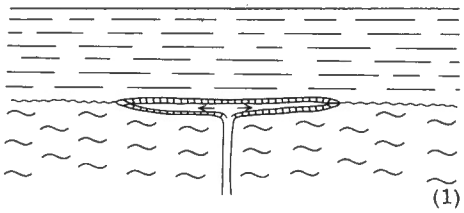
CRYSTALLIZATION OF THE INTRUSION

During and immediately after formation of the magma chamber, marginal gabbros formed a skin or rind around the intrusion. The outermost part of the marginal gabbro was formed by quenching against the cooler wall-rocks but the bulk of the marginal gabbros were deposited by accumulation of settled crystals on the walls and floor and to a lesser extent by welding of crystals on the roof. It is likely that fracturing and spalling of pieces of the marginal gabbro took place from time to time in response to adjustments of the magma chamber to its surroundings - one such fragment may be the gabbro-cordierite gneiss composite inclusion south of Fault Bay. The common presence of small scale rhythmic layering in the outer marginal zone suggests that magma currents were active at an early stage of crystallization.

One of the perplexing problems of the Michikamau Intrusion is to provide an explanation of the timing and the manner of development of the anorthosite zone. There is no reason to doubt that the anorthosite zone, like the bulk of the other rocks of the intrusion, formed by accumulation of crystals. The chief difficulty involved is to explain why 3,500 to 4,500 metres of plagioclase cumulates are present high in the intrusion, overlying a much greater thickness of layered rocks that formed by accumulation of olivine and plagioclase. Some of the possible ways in which the thick anorthosite zone may have developed high in the intrusion are:

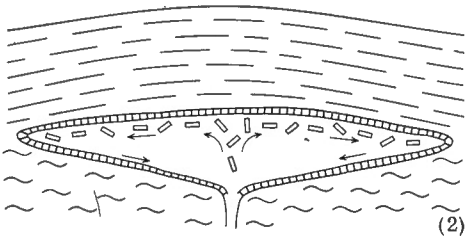
1. a mechanical process was responsible for accumulation of plagioclase by separating it from coprecipitating minerals, e.g. differential settling rates.
2. during the later stages of crystallization of the intrusion a condition was reached such that there was a long interval during which plagioclase alone crystallized from the magma and settled.
3. a mechanical process caused accumulation of plagioclase crystals at a time when no other crystalline phases were present in the magma chamber i.e. at a time soon after the magma chamber was filled.

Although there is some compositional overlap, plagioclase of the anorthosite zone (An_{44}-An_{55}) is more sodic than that of the layered series (An_{50}-An_{72}). If the anorthosite zone formed by accumulation of plagioclase due to differential settling rates, the process would seem to be restricted to a period following accumulation of most of the layered series. Differential rates of settling for plagioclase and whatever ferrogmagnesian minerals were



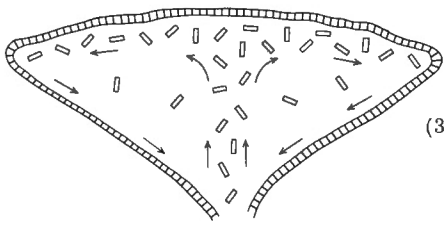
(1)

The magma rises to a relatively undeformed interface (unconformity?) between upper greenschist facies rocks of the Petscapiskau Group and a granulitic gneiss basement, spreads and chills against the walls.



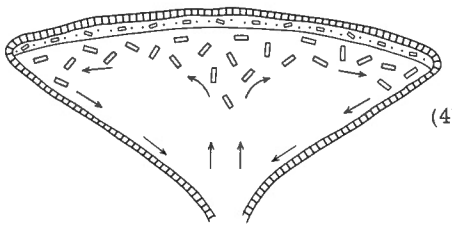
(2)

As the magma chamber is inflated, the roof is domed slightly and the floor is depressed.

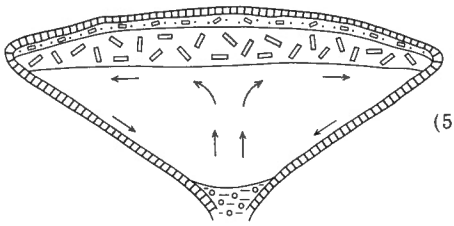


(3)

Further inflation accompanies continued depression of the floor and the chamber is filled with basaltic liquid carrying perhaps 25 to 30 per cent by volume of plagioclase crystals.

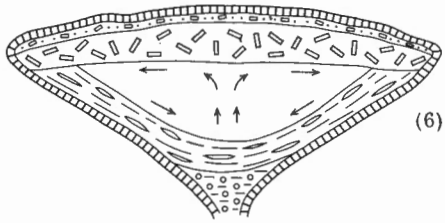


(4)



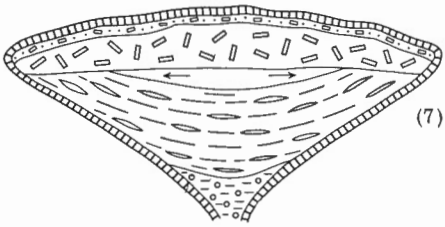
(5)

The density difference between crystals and liquid is sufficiently great that crystals rise to accumulate, at first rapidly, trapping much interstitial liquid to form upper border gabbro and leucogabbro, then more slowly (5) so that diffusion and adcumulus growth of plagioclase can expel the interstitial liquid and form the anorthosite zone.

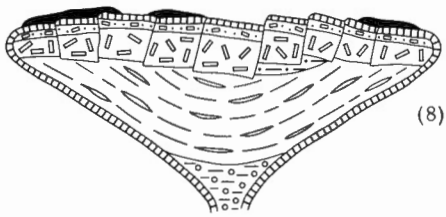


(6)

After most of the anorthosite zone has formed, an overall pattern of magma circulation is established due to convection and density currents caused by crystallization in the upper part of the chamber and bottom accumulation of the crystals takes place (7) until between 80 and 90 per cent of the intrusion has crystallized.



(7)



(8)

Roof collapse brings anorthosite and upper border gabbro and leucogabbro down upon the layered series. Much of the remaining liquid is squeezed to the upper part of the intrusion but local pockets are trapped between the anorthosite zone and leucotroctolite. After solidification, block faults with large vertical displacements break the intrusion into a mosaic (not shown).

LEGEND

- Transgressive group
- Upper zone: leucogabbro, gabbro
- Anorthosite zone
- Layered gabbro
- Leucotroctolite
- Troctolite
- Marginal zone
- Supracrustal rocks (Petscapiskau Group)
- Basement gneiss complex

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Figure 5. A sequence of diagrammatic sketches to illustrate a proposed model for the evolution of the Michikamau intrusion.

precipitating implies that a widespread zone enriched in mafic minerals underlies the anorthosite zone. No evidence to support the existence of such a mafic-enriched zone has been discovered. If typical conditions within the cooling intrusion included continual circulation of magma, as the widespread layered structures imply, it is difficult to imagine that any large volume markedly different in compositions, e. g. enriched in plagioclase crystals, could form and be maintained. Therefore, it is evident that severe difficulties are encountered in applying the first mechanism to explain the development of the anorthosite zone.

The second process suggested (see also Emslie, in press) is attractive for its apparent simplicity but as yet there are no known phase relations that allow equilibrium crystallization of large quantities of plagioclase alone at an advanced stage of crystallization of a basic magma. It is possible to appeal to nonequilibrium processes such as pressure variations on the magma, due for example to rupture of the roof of the intrusion, which could cause excursions of the magma into the field of plagioclase crystallization. However, it is difficult to accept such a process as capable of producing the entire anorthosite zone.

The last mechanism suggested to account for the anorthosite zone is that it formed soon after emplacement of the intrusion. Although, as pointed out earlier, lack of data at the present stage of investigation precludes drawing final conclusions, the interpretation to be presented is looked upon favourably by the author and is used in the model outlined in Figure 5. The conclusion has been reached that the initial magma was a suspension of plagioclase crystals in a basaltic liquid. Although plagioclase phenocrysts occur in the marginal gabbros, they are uncommon. Also, there is no evidence to support the existence of a thick layer of anorthosite in the lower part of the layered series which would be expected if plagioclase crystals suspended in the magma had settled soon after intrusion. Both these facts seem to argue against the presence of abundant plagioclase crystals in the initial magma. However, rejection of the hypothesis for these reasons assumes that the plagioclase crystals were capable of settling. Consideration must also be given to the possibility that the early plagioclase crystals were capable of floating or of remaining in suspension. Plagioclase crystals in basic magma are not expected to differ greatly in density from the magma. Hess (1960, p. 85) suggested that An_{85} is capable of being suspended or of slow settling. Pulvertaft (1965, pp. 28-29) considered that An_{60} may remain suspended or float. In these calculations, plagioclase compositions are such that they are considered by the above authors to have been in equilibrium with the enclosing magma. It is possible that the plagioclase crystals in the initial Michikamau magma were as sodic as An_{40} (An_{44} has been found) and thus significantly less dense than the usual plagioclase in basic magmas. This plagioclase would not have been in equilibrium with the liquid in the Michikamau magma chamber at the time of emplacement. Green (1966) has provided experimental data on a gabbroic anorthosite composition that crystallized plagioclase at the liquidus at pressures up to 22. 5 kilobars. His results indicate that plagioclase coexisting with liquid at 18 kilobars can be as much as 11. 5 mole per cent poorer in anorthite than at 9 kilobars. This suggests the possibility that sodic plagioclase crystallized at depth could be carried up into the crust in the magma providing that superheat due to decompression was dissipated so as to prevent resorption of the plagioclase crystals. If, after arrival in the crustal magma chamber, the crystals had

remained in suspension for a long period, it is expected that they would be resorbed or would react with the magma to become more calcic. However, if the crystals could accumulate sufficiently rapidly so that reaction with liquid was prevented or at least inhibited, they would remain more sodic than equilibrium compositions. It must be assumed that upward accumulation of plagioclase crystals could take place before a widespread connective circulation pattern was established which would tend to homogenize the crystal-liquid mixture.

Floating of plagioclase in some basic magmas is by no means a new idea, e. g. Elftmann (1893, p. 178), and Grout (1928, pp. 562-566). More recently, Bridgewater (1967) has postulated formation of an anorthosite capping by plagioclase flotation in a deep-seated magma chamber. He envisions a vertical, compositionally-graded, magma column in which plagioclase of An_{60} composition crystallizes in the upper part and floats at the same time as more calcic plagioclase crystallizes and sinks at lower levels in the chamber. Abundant evidence of current activity in the Michikamau magma would rule out development of any significant compositional gradients within the magma chamber.

The layered series began to form by bottom accumulation of olivine (Fo_{80}) and plagioclase (An_{72}) during, or soon after, formation of the anorthosite zone. During accumulation of the great thickness of layered series rocks, olivine changed composition from Fo_{80} to about Fo_{50} and plagioclase from An_{72} to about An_{50}. This change is not regular with height in the layered series as there is considerable local variation in mineral composition that presumably must be explained by different degrees of reaction between cumulus crystals and intercumulus liquid.

At some point in the later stages of solidification of the intrusion, crystal sedimentation built up the floor to where it began to approach the anorthosite zone. If crystallization had continued uninterrupted, the expected result would be that the final differentiates would be trapped beneath the anorthosite zone. This has clearly not occurred; the main bodies of iron-enriched differentiates are transgressive and lie in the uppermost parts of the intrusion. This is not to say that some differentiates may not be trapped below the anorthosite zone - the layered gabbros of the Bay of Islands may have resulted from such a process. However, the Bay of Islands rocks are the only such example; elsewhere the anorthosite zone rests directly upon leucotroctolites. An interpretation which satisfies these relationships involves collapse of the roof bringing the anorthosite zone (and the overlying upper zone gabbros) down upon the leucotroctolites. Where the anorthosite zone and the leucotroctolites are in contact, there is some indication of discordance but evidence of strong disturbance is lacking. It is likely, therefore, that collapse of the roof was not a catastrophic event. At the time of collapse the remaining liquid portion of the intrusion is unlikely to have exceeded 15 per cent of the original volume. The roof collapse is pictured as fracturing of the anorthosite zone and upper zone gabbros into large blocks that peeled off the roof and came to rest gently on the leucotroctolite floor. At the same time, remaining magma was displaced upward through fractures to find a resting place between the original roof and the upper zone gabbros. Not all of the residual magma was so displaced and part was trapped and completed crystallization beneath the anorthosite zone. An outline of the interpreted evolution of the Michikamau Intrusion is shown diagrammatically in Figure 5.

Rhythmic layering, some of which is graded, igneous lamination, and other structures indicating the presence of magma currents are widespread throughout the layered series and it is therefore concluded that currents were active during much of the crystallization interval of the intrusion. The existence of currents in any large basic intrusion is not surprising and perhaps less so in a funnel-shaped intrusion. Because heat flow from a large intrusion in the earth's crust must be dominantly upward, the large upper surface area of an inverted funnel must be an effective cooling surface. Combining this with the fact that heat is concentrated at the thick central part of the intrusion makes convection an appealing process. In fact, as indicated by Hess (1960, pp. 148-149) and Jackson (1961, p. 95), it is difficult to argue that convection did not occur in any large cooling body of basic magma. A funnel-shaped intrusion provides an excellent model for single cell convection and it may be assumed that this was the most likely configuration for the man circulation pattern although it is obviously a simplification. Rising currents in the centre would reach the roof and spread radially, following the roof to the margins and then turn downward to travel down a dip slope toward the lower centre of the intrusion before rising again. Because heat loss was dominantly through the roof, it is reasonable to expect that the main site of crystallization was also there. Formation of density currents with variable amounts of suspended crystals may be expected, therefore, to have reinforced magma circulation caused by simple convection.

A further reason for making arguments favouring convection more compelling is the structure of the layered series. In a stagnant magma it would be difficult to understand why simple settling of crystals would not rapidly fill up the axial region of the funnel and successive layers become more nearly horizontal and increasingly discordant to the walls at higher levels in the intrusion. This has not happened in the Michikamau Intrusion and layering, not only in the marginal gabbros but in the layered series as well, is subparallel to the walls (note one exception northwest of Loon Lake). In a funnel-shaped cross-section with single cell convection, velocities of magma currents must increase due to constriction as they pass down the dip slope toward the bottom of the intrusion. Other things being equal, e. g. size of suspended crystals, concentration of crystals in the magma current, etc., deposition at the lower levels may occur at no greater rate than at the upper levels thus helping to maintain layering subparallel to the walls.

POST-SOLIDIFICATION EVENTS

The major events that have affected the solidified Michikamau Intrusion are all connected with faulting. A certain amount of faulting is believed to have taken place in the later stages of crystallization as described above. This was probably confined to the upper parts of the intrusion and did not involve vertical movements exceeding perhaps 1 kilometre. Much larger, later movements must have taken place to bring, for example, troctolite of the lower layered series into contact with upper zone leucogabbro. There is some reason to believe that these major vertical movements are spatially related to the Michikamau Intrusion (see p. 53). There is no way of defining the age of this faulting except that it postdates consolidation of the intrusion. Because the displacements are nearly vertical, one is inclined to consider

that perhaps crustal uplift confined to the region of the intrusion was a response to the earlier crustal downwarp at the time the intrusion was emplaced.

Movement on the Lake Michikamau Fault appears to have been largely rotational. This is sufficiently different from the other faults that a different mode of origin may be involved. Emplacement of the large body of granite west of Lake Michikamats may have produced stresses that caused the fault.

ECONOMIC GEOLOGY

Concentrations of oxide minerals in certain parts of the Michikamau Intrusion are of potential economic interest. Most of these are layers, pods, and heavy disseminations in marginal zone gabbros. Magnetite and ilmenite are the common oxide minerals. A broad zone (up to 300 metres wide and several kilometres along strike) extends west from Magnetite Lake. Magnetite separated from these oxide concentrations contains up to 1.5 per cent V_2O_3. None of the observed pods or layers is extensive but a magnetic survey would readily define the extent of the oxide bearing zone and provide an indication of relative magnetite concentrations. The vanadium-bearing magnetite is of especial interest as a possible source of vanadium.

Iridescent plagioclase is a common constituent of the anorthosite zone. Although no large concentrations of coarsely crystalline material were observed, careful prospecting might reveal a source of gem quality crystals.

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