METAMORPHOSED TRONDHJEMITIC BASEMENT BENEATH THE ARCHEAN FAVOURABLE LAKE VOLCANIC COMPLEX, NORTHWESTERN ONTARIO

BY

ELIZABETH M. HILLARY

Submitted to the University of Manitoba

Faculty of Graduate Studies

in Partial Fulfilment of the Requirements

for the Degree

Master of Science

March, 1980

METAMORPHOSED TRONDHJEMITIC BASEMENT
BENEATH THE
ARCHEAN FAVOURABLE LAKE VOLCANIC COMPLEX,
NORTHWESTERN ONTARIO

BY

ELIZABETH MUSSEN HILLARY

A thesis submitted to the Faculty of Graduate Studies of
the University of Manitoba in partial fulfillment of the requirements
of the degree of

MASTER OF SCIENCE

Abstract

The Favourable Lake metavolcanic-metasedimentary sequence in Northwestern Ontario was intruded to the north by the North Trout Lake batholith approximately 2700 Ma ago. Within the batholith is a 15 km^2 remnant of older basement (2910 Ma) composed largely of gneissic hornblende-biotite trondhjemite. The basement is separated from the metavolcanic-metasedimentary sequence by about 1 km of younger batholithic rocks, and was uplifted into its present position during emplacement of the batholith. The trondhjemite was metamorphosed during and possible prior to the emplacement of the batholith. Metamorphism is variable in intensity but is mainly amphibolite facies. It has resulted in moderate to intense recrystallization, and development of gneissosity and lineation.

However, the primary plutonic nature of the trondhjemite can still be recognized. It is medium-grained and locally porphyritic, with an average of 11 percent biotite and 2 percent hornblende. The trondhjemite intruded an earlier dioritic unit and contains amphibolite xenoliths that are probably derived from the diorite. Melanocractic trondhjemite sills and dikes with slightly higher concentrations of biotite, hornblende and microcline comprise up to 10 percent of the unit and were intruded prior to metamorphism.

Chemical and modal data show that the trondhjemite is relatively homogenous, although some components, particularly K_2O, Na_2O, Al_2O_3 and SiO_2, were mobile during metamorphism. The degree

of metamorphism makes petrogenetic considerations based on chemistry less reliable than those based on field and petrographic data, but the trondhjemite probably was derived by shallow partial melting of amphibolite or basalt.

CONTENTS

LIST OF TABLES

LIST OF FIGURES

Page

Introduction

The nature of the crust prior to the Kenoran orogeny about 2600 Ma ago is currently being investigated in many shield areas of the world (see Young 1978, various papers in Windley 1976). Some workers speculate that the volcanoes and associated sediments of the Archean greenstone belts developed on a sialic basement of unknown thickness and extent (Anhaeusser and others 1969, Henderson and Easton 1977, Ayres 1974, Baragar and McGlynn 1976, 1978) whereas others propose an early simatic crust (Glikson 1971, 1972, 1978, Glikson and Lambert 1976, Viljoen and Viljoen 1969, Anhaeusser 1973, Wilson and others 1974). However, documentation of the character of basement material has been made in only a few areas (Bridgewater and Collerson 1976, Glikson 1972, Arth and Hanson 1975).

In light of this problem, the origin and metamorphic history of a small area of probable basement trondhjemite gneiss in the Favourable Lake area of northwestern Ontario is documented in this thesis. The trondhjemite has been dated at 2910 Ma using U-Pb zircon methods (Krogh and Davis 1971) and is therefore about 200 Ma older than most greenstone belts in the Superior Province (Krogh and Davis 1971). It is separated from the nearby deformed and metamorphosed Favourable Lake volcanic complex, which has not been isotopically dated, by a 1 km wide, younger granodiorite pluton that is part of the North Trout Lake batholith. This batholith intruded and now forms the north edge of the Favourable Lake volcanic complex. Spatially the trondhjemite forms part of the batholith (phase N-3, Ayres 1974), but it is not genetically related to the other phases of the batholith, all of which are younger. Field and laboratory studies indicate that the trondhjemite is a metamorphosed plutonic rock and is not remobilized supracrustal material as suggested elsewhere for many ancient crustal remnants (Anhaeusser and others 1969, Barker and Peterman 1974, Arth and Hanson 1975).

The basement trondhjemite is 0.5 km north of North Trout Lake, about 60 km east of the Ontario-Manitoba border (Fig. 1). It occupies an oval area of 15 km^2 and is 5 km long and 3 km wide. The area is accessible by float equipped aircraft from Red Lake, Ontario, 200 km south of North Trout Lake. The only nearby permanent settlements are a fishing camp that was operated by the late N. Kondradt on South Trout Lake, 1 km south of North Trout Lake, and Sandy Lake Indian Reserve, 27 km northeast of North Trout Lake. Two winter roads cross the trondhjemite (Fig. 37). The west road is overgrown and almost obliterated, but the east road is relatively clear and is still travelled during the winter months.

Early reconnaissance mapping was done in the area by Low (1887), Douglas (1926) and Hurst (1930). Part of the Favourable Lake greenstone belt and surrounding granitic bodies was mapped at a scale of 2:31,680 by the Ontario Division of Mines (Ayres 1974).

The trondhjemite is covered by boreal forest and muskeg. Relief is generally less than 10 m but locally is as much as 40 m. Outcrop is relatively good in the east half of the basement unit and along the western edge, but is poor in the centre of the area (Fig. 37).

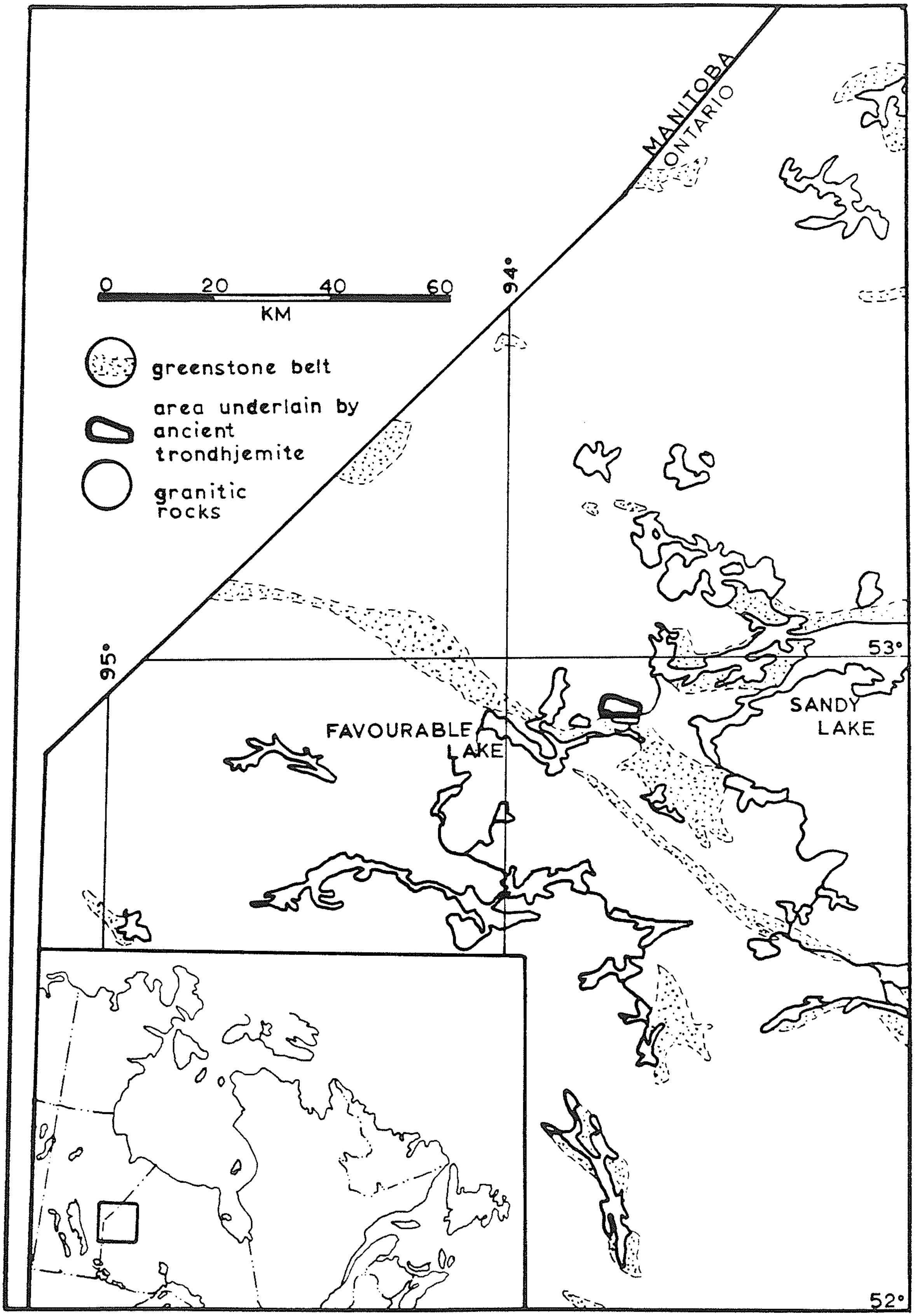


Figure 1 - Location map.

Method of Study

Field work was carried out during June and July of 1976. All possible outcrops of the basement trondhjemite were located on 1:15,840 aerial photographs, using the previous mapping of Ayres (1974) and stereoscopic examination. All outcrops were then examined in detail. The contacts of the trondhjemite were defined by traverses into adjacent units.

Three hundred hand samples were collected for petrographic and chemical work carried out during the winter of 1976-77. An effort was made to obtain a uniform sample distribution, but this was hampered by poor outcrop distribution and, in most of the central and northwestern parts of the unit, by almost complete obliteration of the basement remnant by crosscutting batholith phases. To investigate local structural and textural complexity, two large outcrops were sampled at 30 m intervals on a 300 m by 150 m grid. Structural measurements were made at each sample station.

One hundred and forty thin sections were made of the trondhjemite and nearby rock units. Seventy-three modal analyses were done (Appendix 1) and eighteen chemical analyses were made by K. Ramlal at the University of Manitoba. The trondhjemite was analyzed for ten major oxides as well as P_2O_5, CO_2, H_2O, Ni and Pb. Two samples were analyzed for rare earth elements by neutron activation by H.Y. Kuo at the University of Manitoba.

Acknowledgements

Several people have helped in the preparation and writing of this thesis. Gayle Schwab ably assisted me in the field; and at the University of Manitoba, K. Ramlal and H.Y. Kuo analyzed samples for major oxides and rare earths respectively. Thanks are especially due to Dr. L.D. Ayres, who suggested the topic, supervised the work and offered many helpful ideas and suggestions. The thesis has been reviewed by L.D. Ayres, G.S. Clark and J. Graham. I am very grateful to Judy Bradley for her careful preparation of the final manuscript.

The work has been supported by National Research Council Grant A8996 to L.D. Ayres.

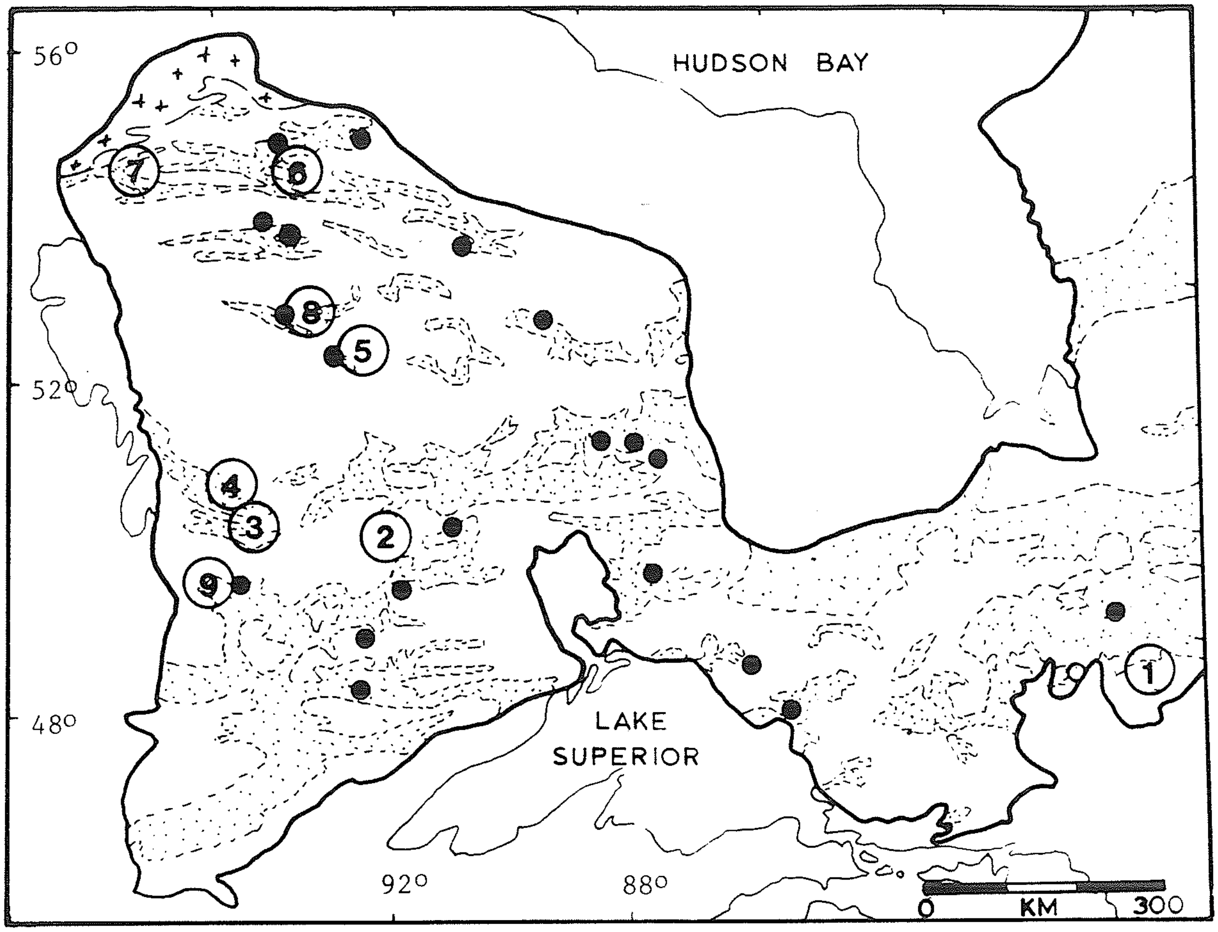
A. REGIONAL SETTING

(1) The Superior Province

During the Kenoran orogeny in the Superior Province, about 2600 Ma, volcanic and sedimentary supracrustal sequences were deformed, metamorphosed and intruded by large composite granitic batholiths. The remnants of these supracrustal sequences now form linear belts within a largely granitic terrain (Fig. 2).

Volcanic rocks predominate in most belts and the lower part of the sequence is commonly a thick series of mafic submarine flows. The upper part contains an increasing amount of intermediate to felsic volcanic rocks, and these grade, vertically and laterally, into greywacke and mudstone turbidites. Intrusion of the batholiths deformed and metamorphosed the supracrustal rocks. Metamorphic grade ranges from prehnite-pumpellyite to amphibolite, and locally to granulite facies.

Granitic plutons occupy about 60 percent of the province and range in size from narrow dikes to immense composite batholithic complexes, but they are poorly documented. The complexes range in composition from gabbro to granite, but quartz diorite, trondhjemite, grandodiorite and quartz monzonite make up most of the intrusive phases (Goodwin and others 1972). Multiphase batholiths usually have early, relatively melanocratic phases and later leucocratic phases.



location of recognized sialic basement

location of conglomerate containing granitoid clasts

metavolcanic-metasedimentary supracrustal belt surrounded by granitic rocks

Pikwitonei province (Bell 1971)

1 - *Abitibi belt, Quebec*
2 - *Lac Seul area, Ontario*
3 - *Winnipeg River area, Ontario-Manitoba*
4 - *Rice Lake area, Manitoba*
5 - *North Spirit Lake area, Ontario*
6 - *Oxford Lake area - Manitoba*
7 - *Cross Lake area, Manitoba*
8 - *North Trout Lake area, Ontario*
9 - *Gundy Township, Ontario*

Figure 2 - The western part of the Superior province showing locations of recognized sialic basement and conglomerate containing granitoid clasts, modified from Baragar and McGlynn (1976).

Age dating of various plutons using U-Pb zircon, Rb-Sr whole rock and K-Ar mica methods has outlined a major magmatic event between 2700 and 2500 Ma and most of the granitic material was emplaced at that time. However, a few ages indicating an earlier sialic event between about 2900 and 3100 Ma have been obtained from several small felsic to intermediate units within the large batholiths (Fig. 2, Krogh and Davis 1971, Krogh and others 1974, 1976).

(2) Sialic Basement in the Superior Province

In the Superior Province there are several known or suspected occurrences of sialic basement rocks (Fig. 2) but no documented areas of simatic basement. The evidence in support of older sialic basement includes isotopic ages, metamorphic features, granitoid clasts in early conglomerates, high quartz content in some sandstone units, and unconformities between basement and overlying supracrustal rocks. The occurrences are all relatively small and have been intruded by younger granitic plutons. The large area over which they occur (Fig. 2) indicates that at one time the sialic basement may have been extensive.

Nine areas of documented sialic basement are located on Figure 2 and pertinent data for eight of these is tabulated in Table 1. In addition to these specific examples, several other localities of possible sialic basement have been recognized on the basis of granitoid clasts in conglomerates (Fig. 2). Near the base of the Uchi Lake, Birch Lake, Fort Hope and other greenstone successions, conglomerates locally contain granitoid pebbles (Bateman 1940, Goodwin, 1967, Prest 1939, 1944)

TABLE 1 DESCRIPTION OF RECOGNIZED SIALIC BASEMENT IN THE WESTERN SUPERIOR PROVINCE[1]

	ABITIBI BELT, QUEBEC	LAC SEUL AREA, ONTARIO	WINNIPEG RIVER AREA, ONT-MAN	RICE LAKE AREA, MANITOBA
EVIDENCE FOR BASEMENT	unconformity between "granitic mass" and overlying sediments containing pebbles of identical granitic material	isotopic age	extension of Lac Seul geology	isotopic age; quartzo-feldspathic clasts in conglomerates are same as in nearby orthogneiss
COMPOSITION	granitic	tonalite to trondhjemite	quartz diorite to granite	quartz monzonite to granodiorite
STRUCTURE		layered orthogneiss	layered orthogneiss	layered orthogneiss; gneissosity truncated by Kenoran orogeny
AGE		3008-3040 Ma (U-Pb zircon)		2900-3000 Ma (U-Pb zircon)
ORIGIN OF BASEMENT UNIT	plutonic	plutonic	plutonic	plutonic
REFERENCE	Holubec 1972	Harris and Goodwin 1976; Krogh and others 1976	Beakhouse 1977	Ermanovics 1970; Krogh and others 1974

[1] locations of areas are given on Figure 2.

TABLE 1 (continued)

	NORTH SPIRIT LAKE AREA, ONTARIO	OXFORD LAKE AREA, MANITOBA	CROSS LAKE AREA, MANITOBA	GUNDY TOWNSHIP AREA, ONTARIO
EVIDENCE FOR BASEMENT	orthoquartzite and granitoid clasts in conglomerate	isotopic age; early pillowed basalts are underlain conformably by quartz-rich sediments and conglomerates of apparent plutonic derivation	boulders in a conglomerate are the same composition as nearby tonalitic gneiss; conglomerate unconformably overlying granodiorite and containing identical boulders	isotopic age; intense deformation and high metamorphic grade (middle amphibolite); intrusion of basement by less deformed plutons
COMPOSITION	orthoquartzite and granitoid	granodiorite to granite	tonalite to granodiorite	trondhjemite
STRUCTURE				
AGE				2950 Ma (whole-rock Rb-Sr)
ORIGIN OF BASEMENT UNIT	plutonic	plutonic	plutonic	plutonic
REFERENCE	Donaldson and Ojakangas 1977; Donaldson and Jackson 1965	Campbell and others, 1972	Horwood 1935; Rousell 1965	Bald 1977, personal communication

of tonalitic, trondhjemitic or granodioritic composition. Rare gneissic granitoid clasts (Donaldson and Jackson 1965, Bass 1961) necessitate an earlier tectonic event and subsequent deep erosion before deposition of the congolmerate.

The quartz content of some Archean sandstone units is higher than might be expected if they were derived from volcanic rocks (Pettijohn 1970, Donaldson and Jackson 1965). The most likely source is a plutonic terrain, although some orthoquartzite is locally present in the source (Donaldson and Ojakangas 1977). The presence of polycrystalline quartz and coarse-grained fragments of vein quartz could also indicate a plutonic source. Quartz-rich greywacke and conglomerate containing granitoid clasts occur in both Favourable Lake and North Spirit Lake areas (Donaldson and Jackson 1965, Ayres 1974, Gordanier 1976) fairly close to the sialic basement remnant, and may be derived from it.

These examples demonstrate that a pre-volcanism sialic basement was present in some parts of the Superior Province. Its original extent and range of composition and lithology remain to be determined, but present information supports a trondhjemite to granodioritic orthogneiss of plutonic origin.

(3) The Favourable Lake Area

(a) The Volcanic Complex

The Favourable Lake greenstone belt has a maximum width of about 13 km near Setting Net Lake. Here Ayres (1977) has mapped a 7.5 km thick, isoclinally folded sequence of metavolcanic and derived metasedimentary rocks, bordered on both sides by younger, composite granite batholiths. In the sequence, 15 formations are grouped into 5 cycles. Metamorphic grade ranges from mid-greenschist facies in the centre of the belt to amphibolite and hornblende facies at the margins (Ayres 1978).

Most cycles have a lower basaltic flow unit with some andesitic to dacitic flows and pyroclastic units in the upper part. Minor metasedimentary rocks, mainly volcanogenic greywacke, conglomerate and argillite, occur near the top of some cycles. The sequence represents a subaerial andesitic to dacitic stratovolcano (Cycle 1) with three successive, subaqueous basaltic shields and superimposed subaqueous to subaerial andesitic to dacitic stratovolcanoes, cones and calderas (Cycles 2, 3 and 4) developed on its northwestern flank. Cycle 5 is a complex subaerial to shallow water andesitic to dacitic unit that represents the final stages of volcano construction.

Formation K in Cycle 3 consists of a lower conglomerate member and an upper quartz-rich greywacke member. Volcanic clasts in the conglomerate and greywacke, and volcanic plagioclase in the greywacke indicate a volcanic source. However, the conglomerate also contains trondhjemite clasts with hypidiomorphic-granular texture, indicating a local plutonic source (Gordanier 1976). The framework fraction of the greywacke is 50 to 100 percent quartz, which may also indicate a partly plutonic source. The basement trondhjemite outcrops 3 km north of the quartz-rich greywacke and may be partly the source of the quartz fraction of the greywacke and the trondhjemite clasts in the conglomerate (Ayres 1977).

(b) The North Trout Lake Batholith

In the North Trout Lake batholith north of the Favourable Lake volcanic complex, twenty discrete intrusive phases forming stocks, sills, dikes and dike swarms were recognized by Ayres (1974). The three earliest phases represent pre-volcanism basement, but two of these were recognized in only one outcrop. The other early phase (N-3) is more widespread and is the subject of this thesis. The 17 later phases intruded both the volcanic complex and the basement, and now separate the basement from the volcanic complex. The general sequence of intrusion of the younger phases is from diorite to quartz monzonite, and the sequence shows increasing K_2O, Na_2O and SiO_2 and decreasing CaO, MgO and total iron, although local interruptions and reversals of the sequence are common. The average composition of the batholith is trondhjemite.

In addition to the 2910 Ma age of the trondhjemite, a time gap between the three early phases and the later phases is indicated by several factors. (1) The early trondhjemite (phase N-3) does not fit the mafic to felsic trend of the batholithic phases. It is more felsic and quartz-rich than the next oldest unit (N-4) which is a diorite and is the start of the trend toward mafic-poor, quartz- and microline-rich quartz monzonite. (2) The trondhjemite is the only gneissic unit in the batholith, and is also the only unit with relatively abundant amphibolite xenoliths. (3) Chilled contacts in the batholith are best developed between diorite (N-4) and the basement trondhjemite, and are rare between succeeding phases. This suggests that a long time interval separated emplacement of the trondhjemite and diorite, but the later phases were emplaced within a short time interval.

(c) Structure and Metamorphism

The supracrustal sequence was metamorphosed before and during the emplacement of the batholith. Metamorphic recrystallization, gneissosity and deformation are present in the early trondhjemite, but the later phases of the batholith show little evidence of metamorphism except for incipient recrystallization and minor metamorphic foliation.

B. THE BASEMENT COMPLEX

The basement remnant is a xenolithic screen in the North Trout Lake batholith. Within a 15 km^2 area, it forms about 40 percent of the outcrop, although its abundance varies from more than 75 percent to less than 5 percent (Fig. 3, 37). The remaining 60 percent is later batholithic phases. The basement is mainly a foliated to gneissic trondhjemite with local amphibolite xenoliths; up to 10 percent concordant melanocratic layers, centimeters to meters wide; and up to 5 percent discontinuous leucocratic layers less than 5 cm wide and several meters long. At the east end of the basement remnant, a 0.25 km^2 area of pyroxenite and biotite-hornblende diorite is intruded by trondhjemite dikes. Thin dikes of pre-trondhjemite syenodiorite also intruded the pyroxenite.

Later biotite-hornblende diorite (N-4) and leucocratic biotite quartz monzonite (N-17, 19) of the batholith intruded and are interlayered with the basement complex on the scale of centimeters to meters. Other batholith phases do not cut the basement remnant, although they partly surround the remnant on the south and west.

Average modal analyses of each rock type are given in Table 2. Modal analyses of 61 trondhjemite samples, 7 leucocratic layers and 5 melanocratic layers are given in Appendix 1. Sample locations are shown in Figure 4. The variation in the modes within rock units (see Appendix 1) probably reflects both primary compositional variation and the error inherent in analyzing inhomogenous rocks. To be representative, modal

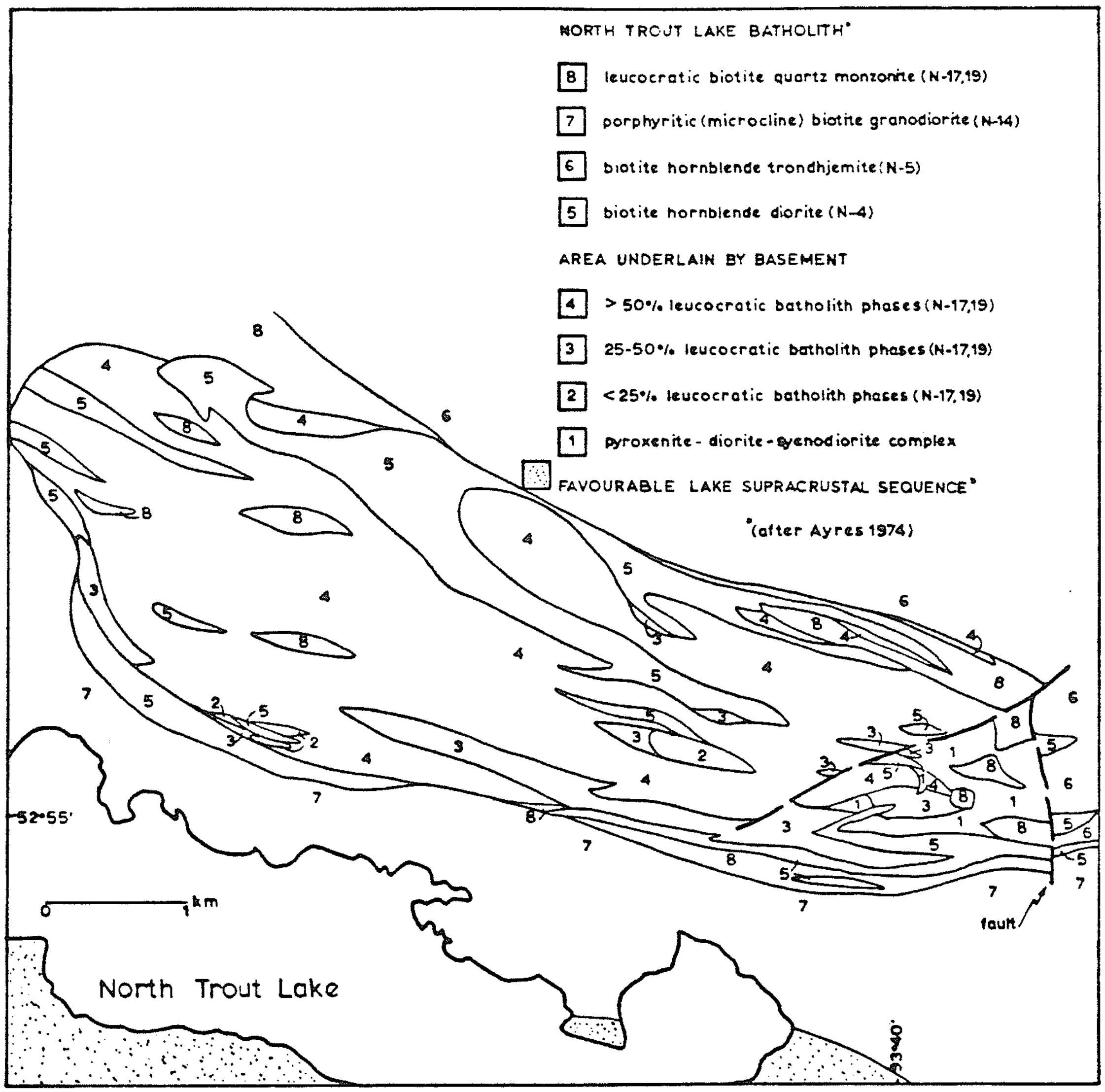


Figure 3 - The basement complex in relation to other rock types in the area.

TABLE 2 AVERAGE MODAL ANALYSES OF UNITS IN THE BASEMENT COMPLEX

	quartz	plagioclase	microcline	biotite	hornblende	others[1]
BASEMENT COMPLEX						
diorite - gabbro complex[2]						
melanocratic phase (3 samples)	2.7 (4.4)[3]	37.2 (2.2)	0	7.3 (3.4)	50.6 (3.2)	2.2 (0.05)
leucocratic phase (2 samples)	0.5	62.0	0	6.5	27.0	4.0
trondhjemite[4] (61 samples)	27.1 (3.8)	56.1 (5.5)	0.1 (0.2)	10.9 (3.4)	2.1 (2.5)	3.7
leucocratic layers[4] in trondhjemite (7 samples)	30.7 (7.2)	67.1 (8.1)	0.1 (0.3)	1.8 (1.7)	0	0.3
melanocratic layers[4] in trondhjemite (5 samples)	22.5 (1.9)	52.3 (7.5)	1.0 (2.2)	14.1 (3.9)	4.7 (2.7)	5.4
LATER PHASES						
biotite - hornblende diorite (N-4)[2] (6 samples)	10.5 (2.8)	51.3 (4.7)	1.1 (2.3)	10.9 (3.5)	21.4 (9.1)	4.8
leucocratic biotite quartz monzonite (N-17, N-19)[2] (1 sample)[5]	30.0	21.4	46.8	0.1	0	1.7

[1] muscovite, epidote, allanite, pyrite, apatite, sphene, zircon, iron-titanium oxide
[2] personal communication, Ayres, 1979
[3] numbers in brackets indicate 1 standard deviation
[4] indicates averages from modal analyses in Appendix 1
[5] modal data for this phase elsewhere in the batholith given in Ayres (1974)

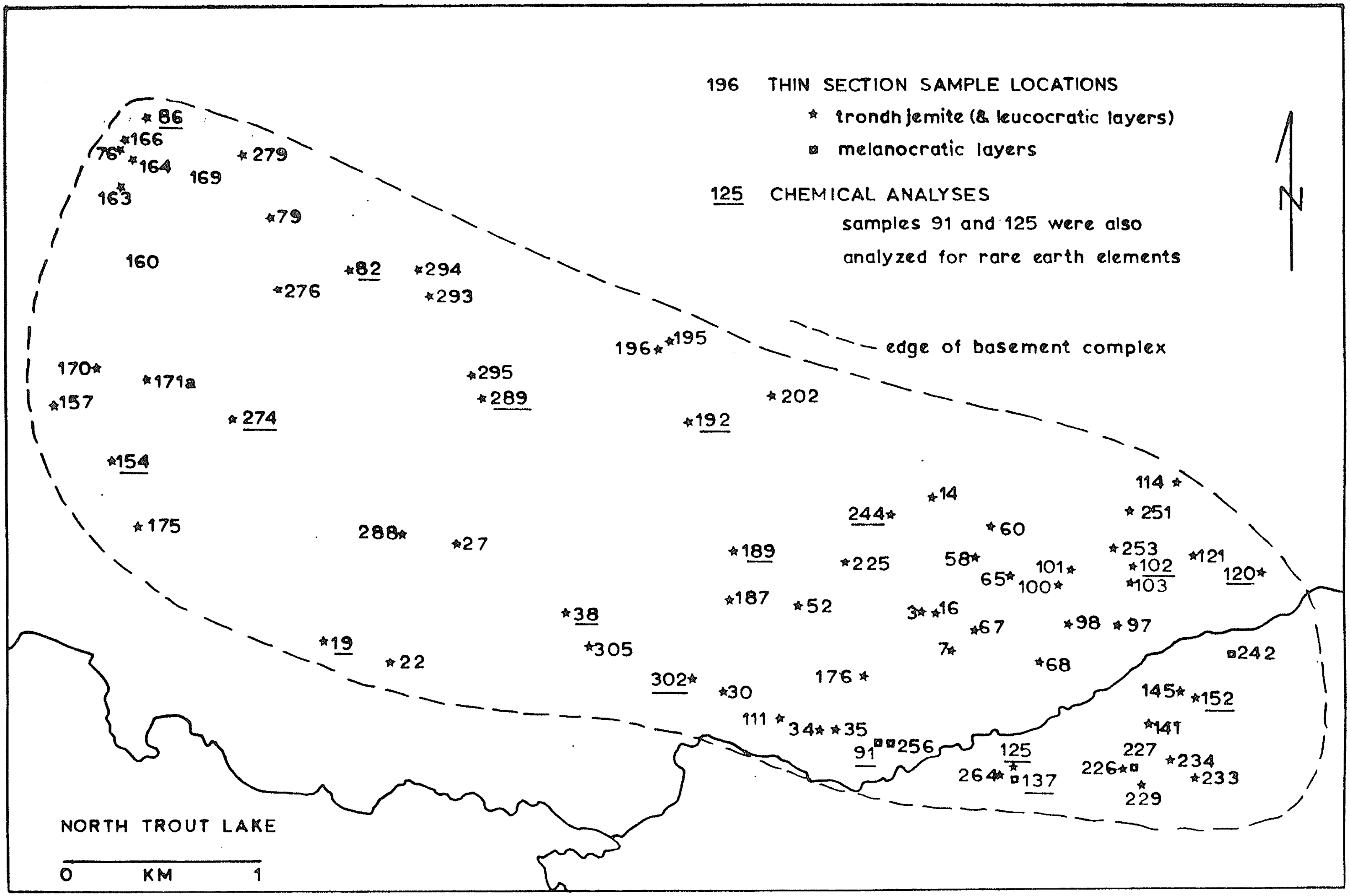


Figure 4 - Sample locations.

analysis of a rock with a coarse, foliated to gneissic fabric and local phenocrysts requires a much larger surface area than analysis of a fine-grained unfoliated rock. According to Chayes (1956) 3 to 4 thin sections of each trondhjemite would be required for a statistically representative analysis. Each modal analysis in Appendix 1 was done on only one thin section, but the possible error thus introduced is probably smaller than any regional variation. Because of the potential statistical errors in individual analyses, compositional variations based on differences between specific samples are probably less meaningful than regional variations shown by groups of samples.

(1) Trondhjemite

Trondhjemite is the most widespread unit in the basement complex, comprising about 90 percent of the remnant. It is a fine- to coarse-grained, locally porphyritic, homogeneous to gneissic, pale to dark grey rock. Despite the rather high hornblende content of some samples (Appendix 1) the name trondhjemite is used rather than tonalite, because the plagioclase is oligoclase except for the andesine cores of some zoned crystals, and the colour index is usually less than 20 (Streckeisen 1976).

In thin section the trondhjemite is remarkably consistent in mineralogy, but is texturally and structurally variable. It is a plutonic rock, but primary plutonic textures are partly destroyed in some areas and almost totally destroyed in others by metamorphic recrystallization. Metamorphic grade is difficult to document because the main minerals, quartz, plagioclase and biotite, are relatively stable with increasing temperature. They recrystallize but do not react. Recrystallization is ubiquitous but variable in intensity. Many samples show only minor recrystallization, and unrecrystallized grains commonly coexist with more recrystallized grains of the same mineral, reflecting energy variations on a local scale. In addition, different mineral species are affected to varying degrees by recrystallization.

The history of metamorphism is thus better documented by textural changes than by mineralogical changes, and these indicate three stages of recrystallization, with a general southeastward increase in recrystallization. In general the most reliable indicator of metamorphism in the trondhjemite is recrystallization of quartz and plagioclase, and this was the main factor used in determining the three stages of recrystallization in Table 3. In stage 1, which is the least recrystallized, primary minerals are strained and fractured, but there is only minor recrystallization. Stage 2 is characterized by about 50 percent recrystallization of quartz and minor recrystallization of plagioclase. In the most intense recrystallization of stage 3, quartz is almost completely recrystallized and plagioclase is about 50 percent recrystallized. There is a corresponding but less pronounced change in the colour and habit of biotite and hornblende with increasing recrystallization. Stage 1 is slightly more abundant than stage 2, while stage 3 is developed sporadically in less than 10 percent of the trondhjemite.

TABLE 3 STAGES OF RECRYSTALLIZATION IN THE TRONDHJEMITE

	STAGE 1 (WEAKLY RECRYSTALLIZED)	STAGE 2 (MODERATELY RECRYSTALLIZED)	STAGE 3 (STRONGLY RECRYSTALLIZED)
QUARTZ	interstitial crystals 1-4 mm (Fig. 9); some straining and suturing of grain boundaries; rare recrystallized grains	50 % or more of quartz recrystallized to grains $\lesssim$0.5 mm (Fig. 12); still grouped in shape of original 1-4 mm grain; some fine-grained intergranular quartz; unrecrystallized quartz is strongly strained	recrystallized mosaic of unsutured, unstrained crystals $\lesssim$0.5 mm (Fig. 13); a few original grain shapes remain in clusters of recrystallized quartz
PLAGIOCLASE			
GROUNDMASS	composition An 21-48; local normal zoning (Fig. 6); good magmatic habit; relatively good twinning albite, Carlsbad-albite (Fig. 5), rare pericline); crystals to 3 mm; some strain and fracturing	10-15% plagioclase recrystal- to grains 1 mm (Fig. 12); development of wedge (Fig. 8) and pericline (Fig. 7) twinning; some grains only partly twinned; partly recrystallized to aggregates with some optical continuity	up to 50% of grains are recrystallized and poorly twinned; no optical continuity in aggregates (Fig. 13); composition An 21-30; some original grain shapes can be seen in clusteres of recrystallized plagioclase grains; in hand sample the plagioclase is more translucent
PHENOCRYSTS	form up to 10% of the trondhjemite; 5-8 mm; same zoning and twinning as the groundmass grains but cores are slightly more calcic (Fig. 6); strained and fractured	recrystallized to 3-5 mm grains that retain some optical continuity (Fig. 12)	recrystallized to 1-5 mm grains that lack any optical continuity (Fig. 13)
MICROCLINE	interstitial anhedral grains 1-2 mm	no change	no change

TABLE 3 (continued)

	STAGE 1	STAGE 2	STAGE 3
BIOTITE	0.5-1 mm; phenocrysts to 8 mm; dark brown to olive green; subhedral, equant grains; strained	slightly ragged; light brown to olive green	ragged; pale olive green
HORNBLENDE	subhedral; green; 1-2 mm	slightly ragged; green	ragged; blue-green; anhedral
ACCESSORIES[1]	euhedral to subhedral	no change	no change
STRUCTURE	gneissic to foliated; locally lineated	gneissic to foliated (Fig. 19); partly recrystallized, particularly along grain boundaries	foliated; partial loss of gneissosity
TEXTURE	magmatic; equant, subhedral plagioclase; interstitial quartz; mineral types evenly distributed; relatively few plagioclase-plagioclase and quartz-quartz boundaries; quartz and plagioclase commonly elongated parallel to foliation; grain size is 1-8 mm; average grain size is 2 mm	minor intergranular quartz, plagioclase and microcline; quartz, plagioclase and microcline commonly elongated parallel to metamorphic foliation with local development of leucocratic lenses and layers (Fig. 19); grain size is 0.5-5 mm	metamorphic; a large number of plagioclase-plagioclase and quartz-quartz boundaries (Fig. 6); no preferred orientation of quartz or feldspar grains; grain size is 0.5-5 mm; average grain size is 1 mm; rock is finer grained, and appears darker because of translucent plagioclase (Fig. 8)

[1]muscovite, epidote, allanite, sphene, apatite, zircon, iron-titanium oxide, pyrite

(a) Petrography

The process of recrystallization is the key to the petrographic textures in the trondhjemite. Recrystallization rearranges the existing mineral phases in a rock, in the absence of mineral reactions, to produce a combination with lower free energy. In strictly thermal metamorphism, ionic mobility is high enough for recrystallization at about half the melting temperature of a mineral (Spry 1969). New crystals being to grow at isolated points, leaving the original texture unchanged. With continued application of heat, the crystals decrease in size and increase in number until the original texture is obliterated and the rock is almost totally recrystallized. Application of pressure lowers the temperature at which recrystallization will occur and affects the shape and orientation of the newly formed crystals.

(i) Recrystallization Stage 1

In stage 1 the trondhjemite was affected by relatively mild metamorphism and is slightly recrystallized, with minor defomation. Plagioclase forms unrecrystallized, subhedral, equant to slightly elongated and oriented grains 1 to 3 mm long (Table 3) with up to 10 percent equant, commonly idiomorphic phenocrysts 3 to 8 mm long. Both albite and pericline twins are present, as well as rare Carlsbad-albite twins (Fig. 5).

The plagioclase has continuous mornal zoning (Fig. 6) in the upper oligoclase range, with some cores as calcic as An_{48}. Local discon-

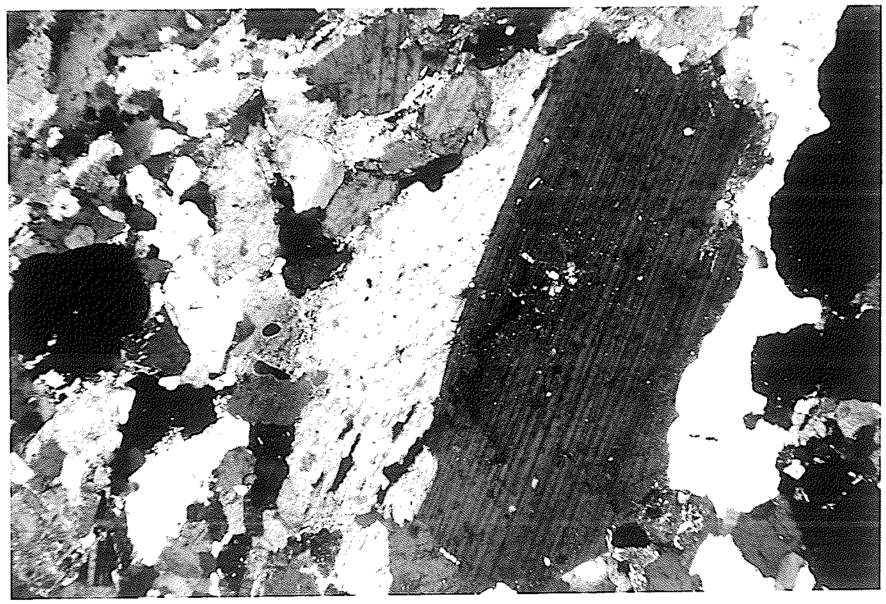


Figure 5 - Carlsbad-albite twinning in plagioclase phenocryst in recrystallization stage 1 (length of view is about 10.0 mm; crossed nicols).

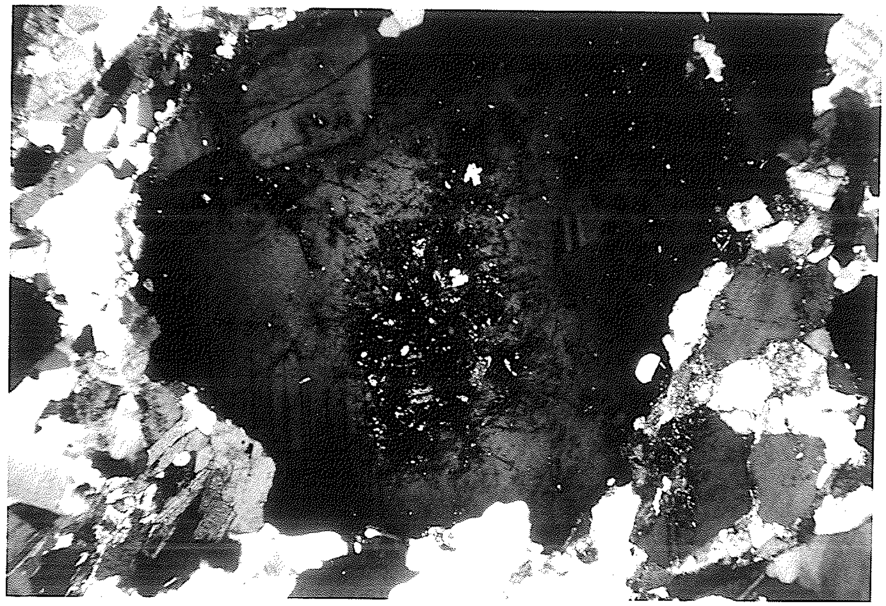


Figure 6 - Normal zoning in plagioclase phenocryst (recrystallization stage 1). Note strong alteration of calcic core, faint deformation twinning near the edges of the crystal, and incipient recrystallization of upper left edge of crystal (length of view is about 4.0 mm; crossed nicols).

tinuities are present in some grains, and a well-defined albite rim is locally common. In a few samples, gradational oscillatory and reverse zoning were observed. Some calcic cores are strongly altered to fine-grained epidote, sericite and iron-titanium oxides (Fig. 6).

The plagioclase phenocrysts have the same zoning and twinning characteristics as the groundmass grains but consistently have more calcic cores, which are locally corroded. Many phenocrysts contain sparse randomly oriented inclusions of euhedral apatite and biotite, and are mantled by smaller crystals of biotite, hornblende, epidote and sphene, producing a pseudo-augen texture. Quartz embayments and albite rims are common. The phenocrysts are definitely primary magmatic crystals and are not metamorphic porphyroblasts. This is shown by the presence of Carlsbad-albite twinning, a primary igneous twin law, the presence of discontinuous and oscillatory zoning, and the similarity in composition and morphology to the smaller plagioclase.

Although the minerals in stage 1 trondhjemite are only slightly recrystallized, the local development of intersecting, wedge-shaped, curved, discontinuous, polysynthetic twins in plagioclase (Fig. 7, Fig. 8, Spry 1969) indicates strain and deformation. Some plagioclase crystals, particularly the phenocrysts, show minor fracturing. In some samples discontinuous cracks extend across several crystals of unrecrystallized plagioclase and quartz, but are partly healed by small crystals of quartz or biotite that have grown across the fracture.

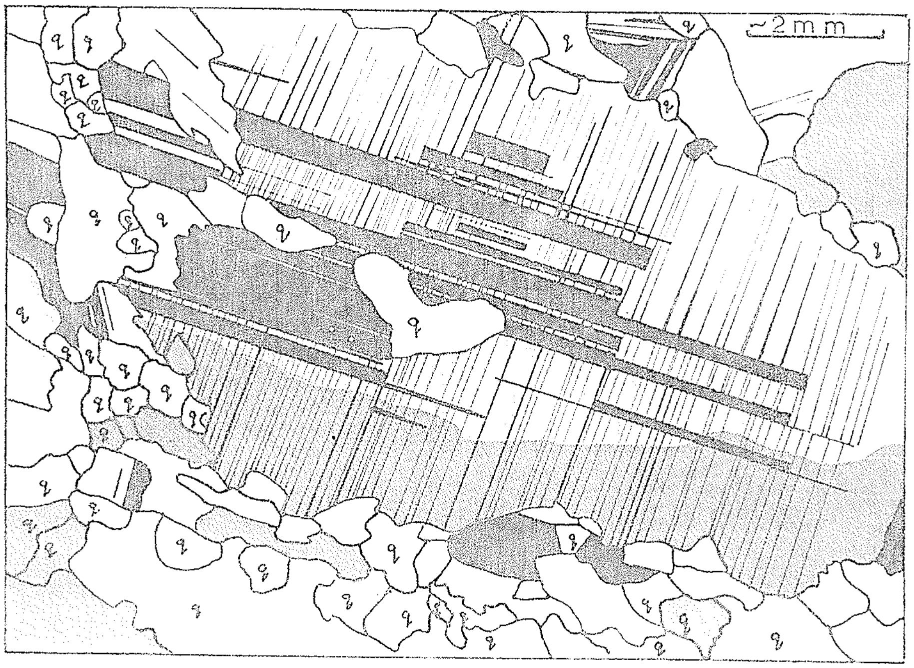


Figure 7 - In stage 2 recrystallization, discontinuous poorly developed pericline twinning occurs in a plagioclase crystal that is beginning to recrystallize at its boundaries. Note warping of grain, causing undulatory extinction. (q = quartz, all other grains are plagioclase; sketched from a photograph).

Figure 8 - Well developed wedge-shaped polysynthetic (pericline deformation) twinning in plagioclase in recrystallization stage 2 (crossed nicols; length of view is about 0.7 mm).

Minor anhedral microcline, with poorly developed grid twinning, forms 1 to 2 mm grains interstitial to, or replacing primary plagioclase. Myrmekite is locally found where plagioclase is adjacent to microcline.

Quartz form 3 to 4 mm, slightly elongated, oriented grains with sutured quartz-quartz boundaries and undulatory extinction (Fig. 9). It is rarely recrystallized to fine-grained intergranular quartz.

The biotite commonly forms dark brown, subhedral, equant grains up to 8 mm but averaging 0.5 to 1.0 mm long. Clots 1 to 2 mm across of small greenish-brown flakes are common throughout the remnant, but their relationship to the recrystallization process is unknown, because they occur in samples of all three recrystallization stages. Locally the biotite is slightly altered to chlorite. A few biotite crystals are bent or broken.

Subhedral to anhedral, green to blue-green hornblende occurs erratically throughout the unit and is not recrystallized. It is generally 1 to 2 mm long but locally forms phenocrysts up to 4 mm long. Biotite and hornblende commonly form clots with epidote, sphene and iron-titanium oxide, rather than being evenly distributed. Locally the biotite and hornblende are intimately interlayered, to form a 2 mm aggregate composed of 5 or 6 parallel plates of hornblende and biotite, each less than 0.5 mm thick. This texture rarely occurs in more than 10 percent of the hornblende and biotite in any one sample, and appears in all three recrystallization stages. Both hornblende and biotite contain

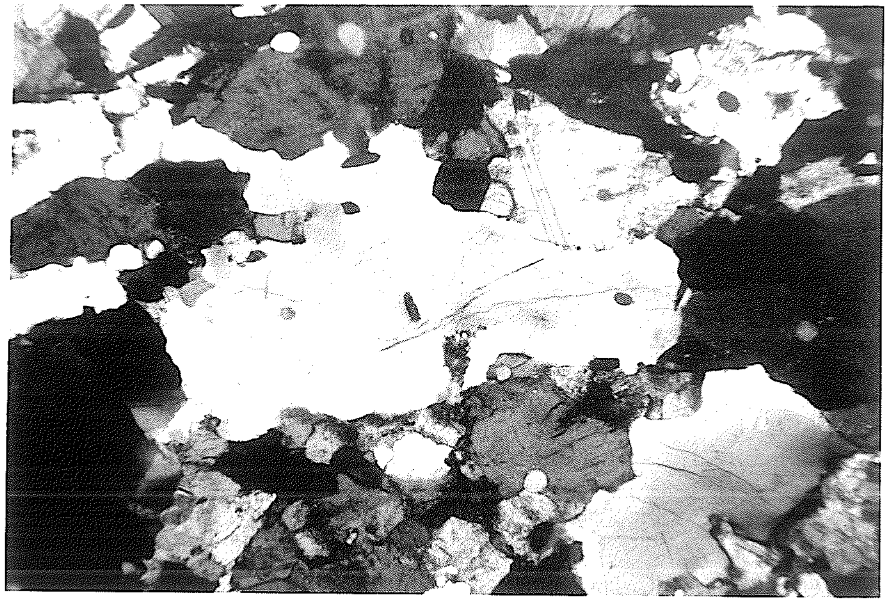


Figure 9 - Mainly unrecrystallized quartz in typical igneous habit, interstitial to plagioclase. Note undulatory extinction (length of view is about 4.0 mm; crossed nicols). Stage I recrystallization.

minor amounts of rutile and iron-titanium oxide exsolved along cleavages, and have local symplectic intergrowths of quartz or epidote.

Minor accessory minerals include apatite, zircon, sphene, allanite, iron-titanium oxide and pyrite, all of which are subhedral to euhedral and locally included in plagioclase, biotite and hornblende. Most of the accessory minerals are less than 1 mm in diameter. Allanite forms yellow-brown euhedral grains that are generally rimmed by euhedral to subhedral iron-rich epidote (Fig. 10). Epidote grains imbedded in biotite are generally euhedral where protected by the biotite but are strongly corroded where adjacent to quartz or plagioclase. Euhedral epidote rims on subhedral pyrite and sphene rims on iron-titanium oxide are common (Fig. 11).

(ii) Recrystallization Stage 2

In stage 2 the effects of deformation are progressively eliminated by recrystallization. Ten to fifteen percent of the groundmass plagioclase is recrystallized to grains less than 1 mm long that retain some optical continuity but are generally untwinned (Fig. 12). The phenocrysts are partly to completely recrystallized to aggregates of 1 to 5 mm grains that retain the shape of the phenocryst and have some optical continuity. Unrecrystallized plagioclase grains have poorly to well-developed pericline (Fig. 7) and wedge-shaped deformation twinning (Fig. 8), but albite and Carlsbad-albite twins are rare.

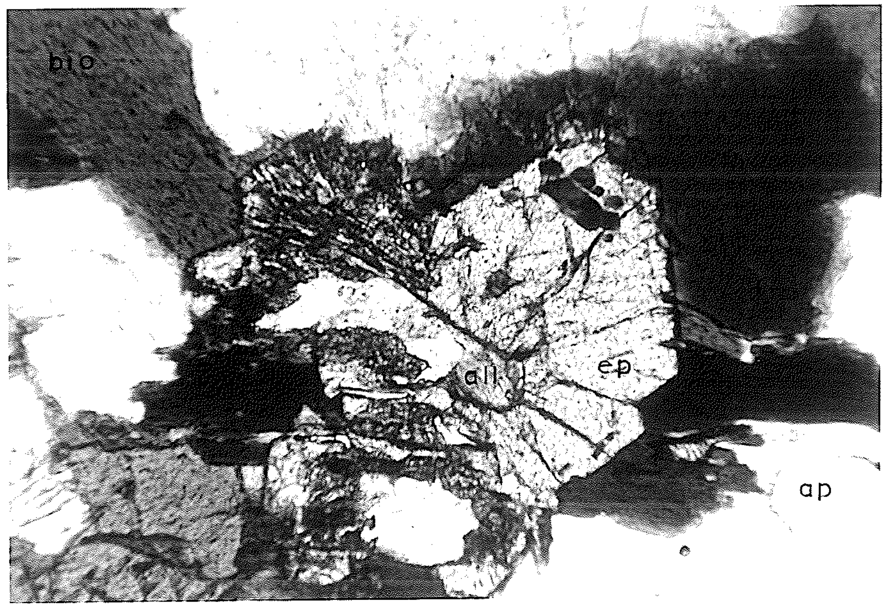


Figure 10 - Allanite (all) rimmed by euhedral epidote (ep) (one polar; length of view is about 0.7 mm; bio = biotite; ap = apatite). Recrystallization stage 1.

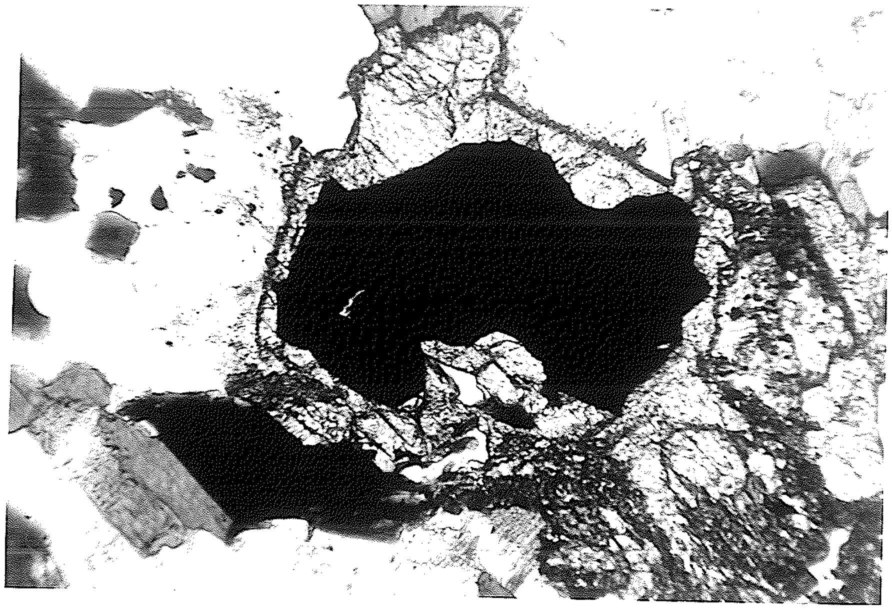


Figure 11 - Pyrite rimmed by late magmatic euhedral epidote (one polar; length of view is about 0.7 mm). Recrystallization stage 1.

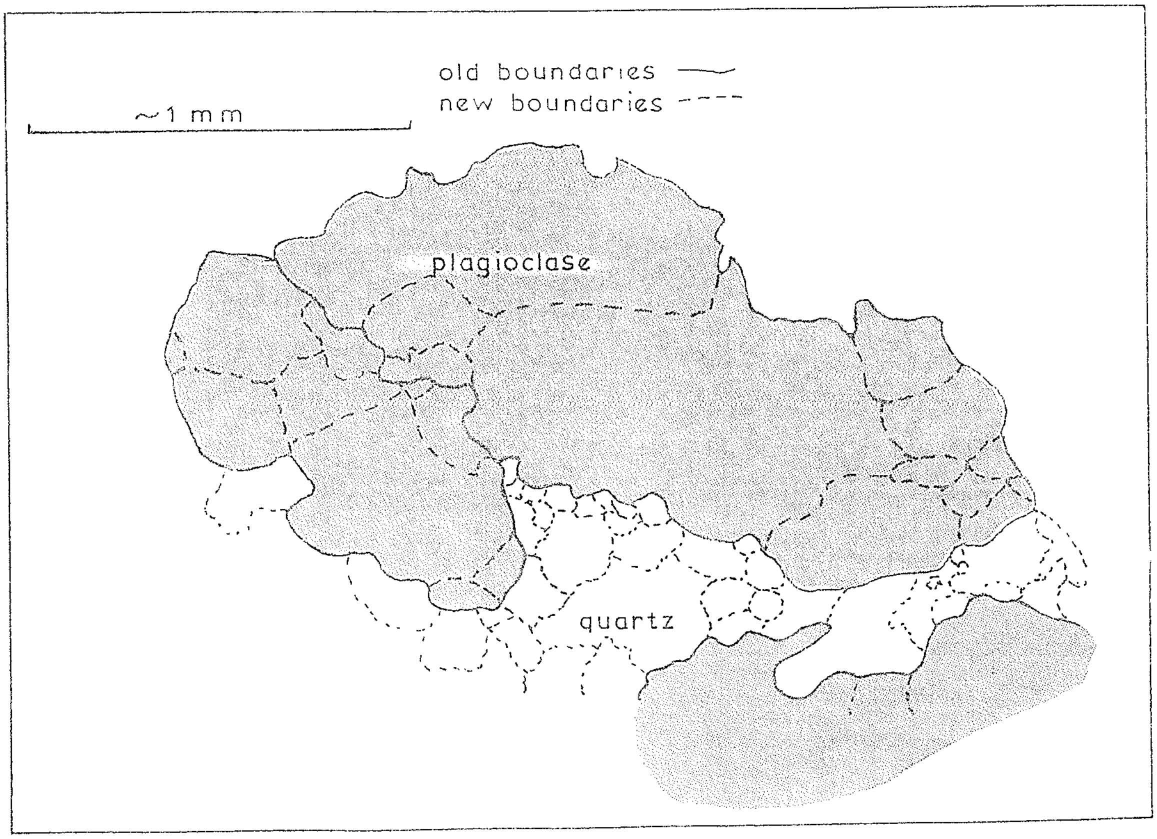


Figure 12 - Recrystallization stage 2. Large plagioclase grains with interstitial quartz. Both mineral species have recrystallized to mozaics of smaller grains which retain the shape of original crystals (drawn from a photograph).

Fifty percent or more of the quartz is recrystallized to polycrystalline aggregates of grains less than 0.5 mm in diameter. The remaining primary quartz is highly strained.

Biotite is generally pale olive green, anhedral and ragged. Trace amounts of muscovite occur as isolated euhedral crystals less than 2 mm long in hornblende-poor rocks.

These characteristics indicate that the stage 2 trondhjemite was more strongly recrystallized and slightly more deformed than the stage 1 trondhjemite by the continued application of heat and minor stress. As a result of the recrystallization, the preferred orientation in stage 1 is partly destroyed, and the number of plagioclase-plagioclase and quartz-quartz boundaries is increased.

(iii) Recrystallization Stage 3

In stage 3 the primary and deformation characteristics are almost totally erased by extensive recrystallization under static conditions, leaving a metamorphic texture (Fig. 13). Up to 50 percent of the plagioclase is recrystallized and poorly twinned, and the maximum anorthite content is only An_{30} rather than An_{48}. The recrystallized aggregates have no optical continuity. Quartz grains are completely recrystallized to unsutured, unstrained crystals less than 0.5 mm in diameter. Although biotite and hornblende are less consistent than quartz and plagioclase, they also tend to be recrystallized, to smaller pale olive green and blue green grains respectively. The number of

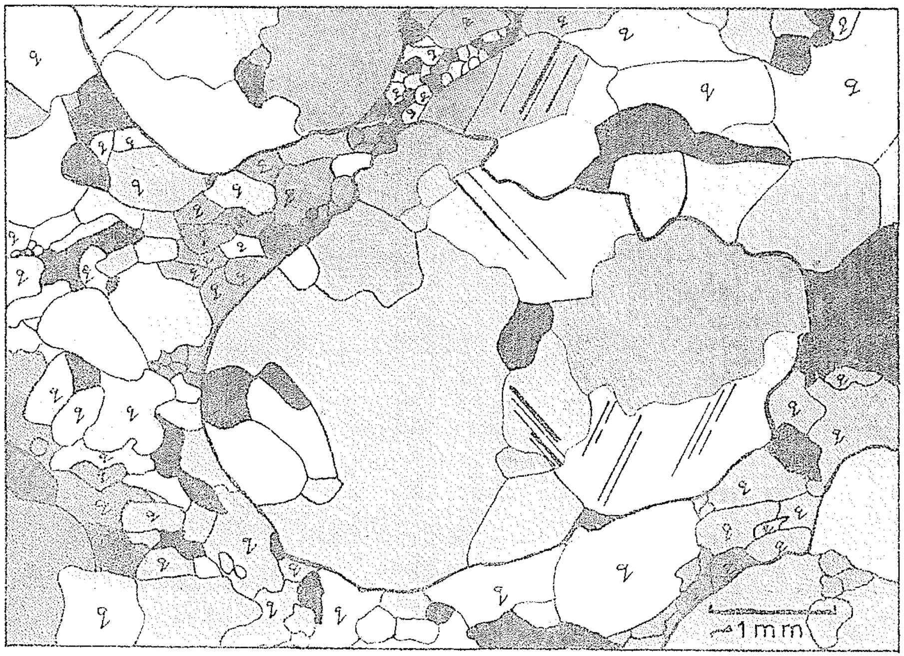


Figure 13 - Strongly recrystallized stage 3 trondhjemite. Plagioclase and quartz (q) are a mosaic of unsutured, unstrained crystals. Recrystallized aggregates of plagioclase, outlined by heavy line, after large primary phenocrysts have no optical continuity. Only very minor twinning remains. (Crossed nicols; sketched from a photograph).

plagioclase-plagioclase and quartz-quartz boundaries is markedly greater than in stage 1.

(iv) Discussion

A recrystallized grain can be readily recognized when the original shape of the primary grain is preserved, particularly when the primary grain was surrounded by different mineral species. A recrystallized grain is more difficult to recognize when the primary grain shape has been destroyed or where several primary grains of the same mineral were in juxtaposition. In the latter case a mixture of large and small crystals of the same species results. Such grain size variations could also be caused by differences in nucleation rates or supply of material during magmatic crystallization. However, in normal magmatic crystallizations, there should be numerous quartz-plagioclase grain boundaries, whereas in recrystallized material the number of plagioclase-plagioclase and quartz-quartz boundaries is drastically increased.

In some samples the initial stage of recrystallization results in an aggregate of several small grains (Fig. 12) showing optical continuity at one position of rotation under crossed nicols. This indicates that one large crystal is in the process of separating into several entitites which are now only slightly related in structure, and which will eventually be completely dissociated from each other. The end result, which is illustrated well only by the few isolated examples of stage 3 recrystallization (Fig. 13), is a mosaic of unrelated, undeformed crystals.

Deformation textures are also ubiquitous in the remnant, but they occur erratically and are not as well developed as the recrystallization textures. These include wedge-shaped albite and pericline twins in plagioclase (Fig. 13) that developed mechanically, in contrast to primary Carlsbad, Acline, Baveno or Manebach growth twins (Spry 1969). Such twins occur throughout the trondhjemite and were probably caused by external strain or by local strain produced by recrystallization. Most of the plagioclase twinning, even when it is not wedge-shaped, is discontinuous, poorly developed pericline twinning (Fig. 7) which may also be the result of deformation (Spry 1969, Vernon 1965). Increasing recrystallization and deformation results in untwinned plagioclase. Undulatory extinction in biotite, quartz and plagioclase (Fig. 7) is caused by the warping of grains during deformation. Other deformation effects include the local fracturing of plagioclase grains and the granulation of quartz and plagioclase borders. The common elongation of quartz and plagioclase crystals parallel to gneissosity may be a primary magmatic crystallization effect, or the result of dislocation creep. This type of deformation involves a series of gliding intracrystalline dislocations that operates under applied stress at upper greenschist and higher grades of metamorphism to change grain shapes (Kerrich and Allison 1978).

In many samples the deformed grains are partially recrystallized to undeformed new crystals. Broken plagioclase crystals with deformation twinning are recrystallized to untwinned, undeformed crystals. Quartz grains with undulatory extinction are recrystallized to smaller grains with even extinction. Plagioclase and quartz grains that were originally aligned parallel to foliation are recrystallized to aggregates of equidimensional grains that retain the alignment of the original crystal, suggesting that the original grains formed under directed stress whereas the recrystallization developed to its present state under relatively static conditions. These changes indicate that the thermal effects of metamorphism outlasted the deformational effects. The final stage of the thermal event was probably a cooling period after prograde events.

The amount of recrystallization is highly variable in individual outcrops but there is a general southeastward increase in intensity of recrystallization. The specific recrystallization stages were not recognized in the field, but in outcrop variations in degree of recrystallization are interlayered on the scale of centimetres to metres (Fig. 14), and can be recognized by variations in grain size, translucency of plagioclase, and development of gneissosity. This interlayering is best developed in the southeast, although its presence in the northwest may be masked by the abundance of later intrusive phases (Fig. 3). Contacts between layers are gradational and invariably concordant to gneissosity. However, in many cases contact relations are uncertain because the younger phases of the North Trout Lake batholith were commonly intruded along the contacts between these textural variations.

(b) Modal Variation

The trondhjemite is poor in microcline and relatively low in total biotite, hornblende and epidote, but it plots as a coherent igneous rock

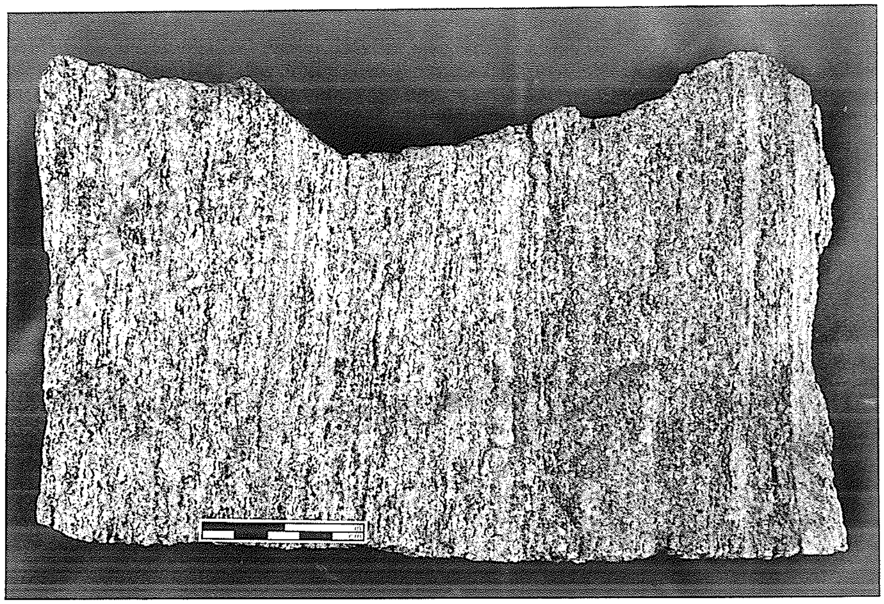


Figure 14 - Slight differences in texture of trondhjemite. Left two-thirds (stage 2) of sample is more gneissic but less crystallized, whereas the right third, which is more recrystallized (stage 3), is slightly finer-grained and felsic minerals are more translucent. Note that both zones have quartzofeldspathic segregations.

on the ternary modal diagrams (Fig. 15, 16). There are only minor modal variations, as shown by the low standard deviation (Table 2) and the clustering of sample points on modal variation diagrams (Fig. 15, 16). The only consistent variations are in biotite, which is slightly more abundant in the south part of the remnant than the north part (Fig. 17) and hornblende, which is less common in the eastern part than elsewhere (Fig. 18). These trends are not statistically significant, but are probably original plutonic variations. In individual outcrops hornblende content is erratic and varies from zero to several percent within an area of several square metres.

(c) Structure

The trondhjemite is better foliated than the various phases of the North Trout Lake batholith, but the structure varies from locally homogeneous to lineated to foliated to coarsely gneissic (Fig. 19). Grain size and the degree of development of foliation and gneissosity are highly variably throughout the remnant and within individual outcrops. The foliation is defined by indistinct, commonly monomineralic, 1 to 2 cm thick lenticular aggregates of quartz and plagioclase and by the concentration of biotite, hornblende, sphene, allanite, epidote, apatite and iron-titanium oxides in discontinuous layers. In stage 1 and stage 2 trondhjemite, individual plagioclase, quartz and potassium feldspar crystals commonly are slightly elongated parallel to this foliation. As the thickness and lateral extent of the quartzofeldspathic aggregates increase, the foliation grades into gneissosity and the grain size commonly increases. Locally the quartzofeldspathic aggregates form a

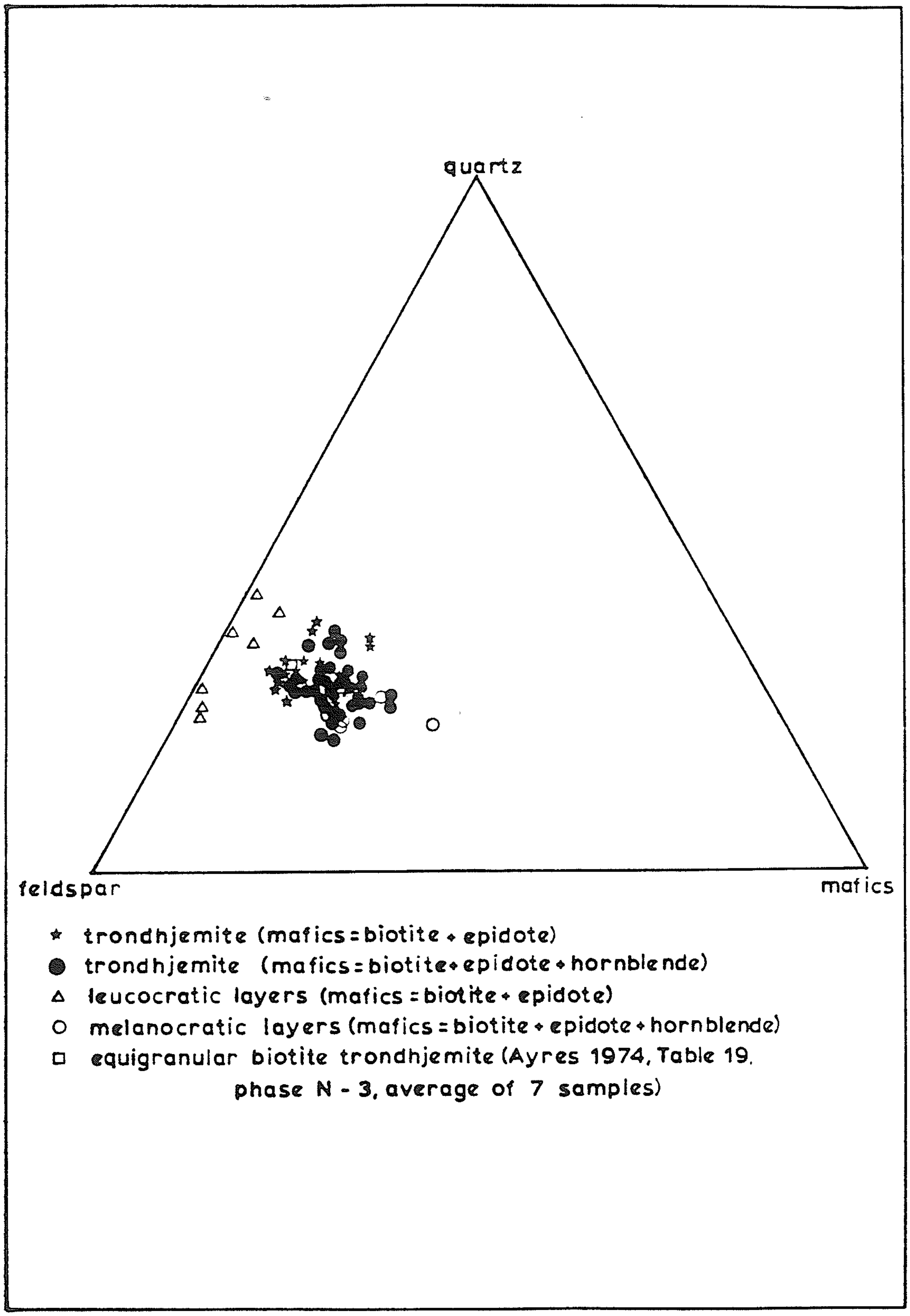


Figure 15 - Modal analysis - Quartz-feldspar-mafics.

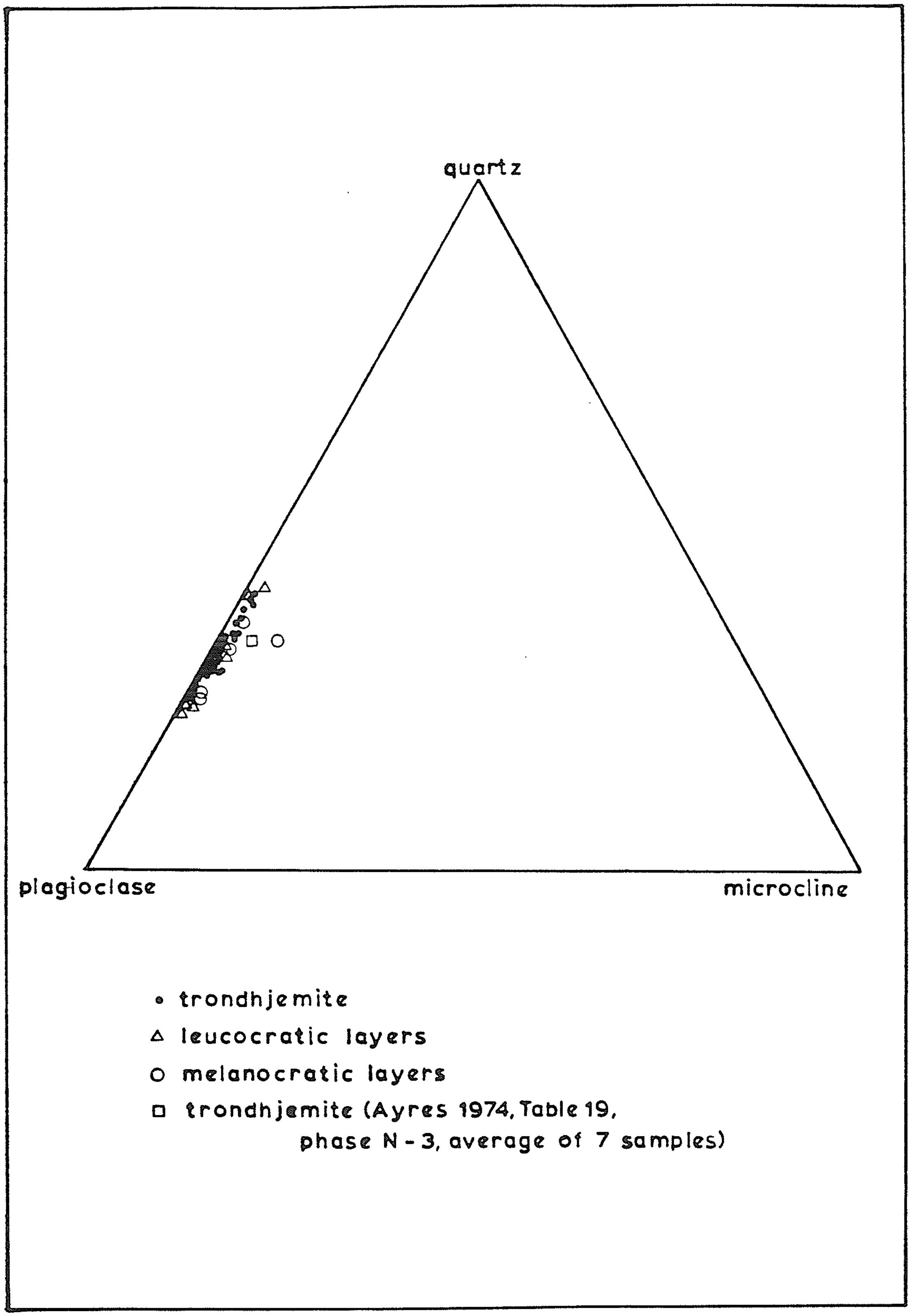


Figure 16 - Modal analysis - Quartz-plagioclase-microcline.

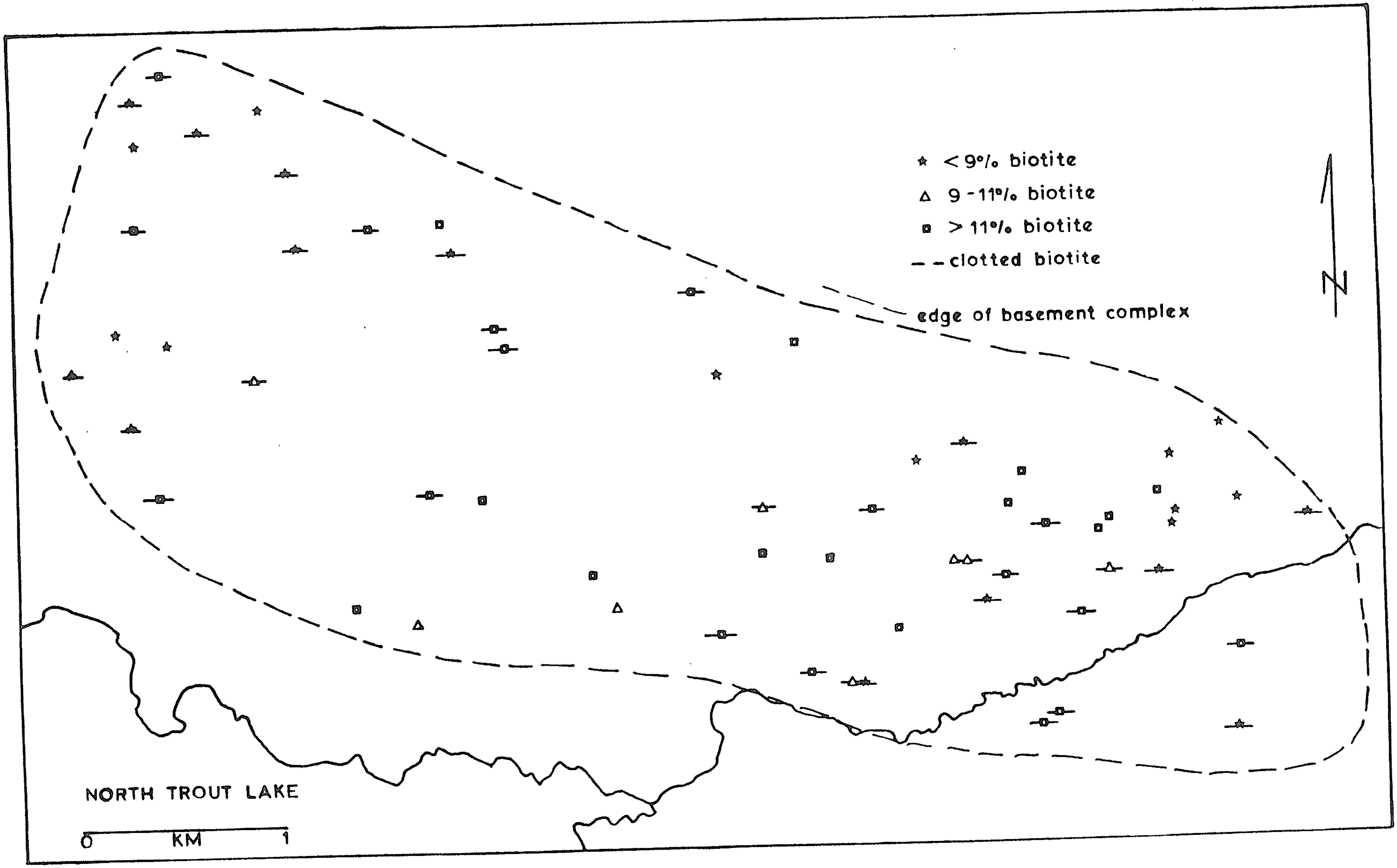


Figure 17 - Modal biotite content in trondhjemite.

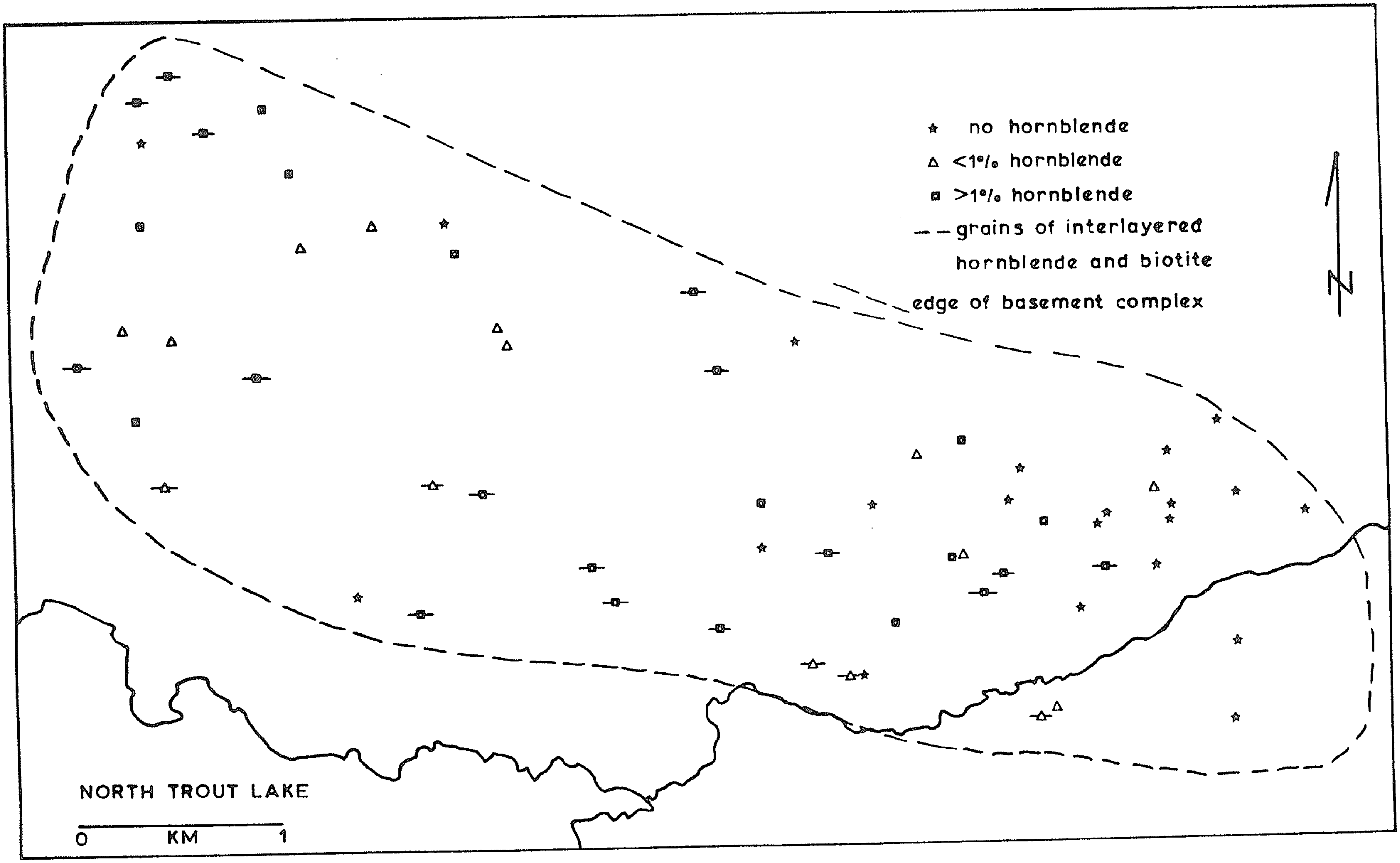


Figure 18 - Modal hornblende content in trondhjemite.

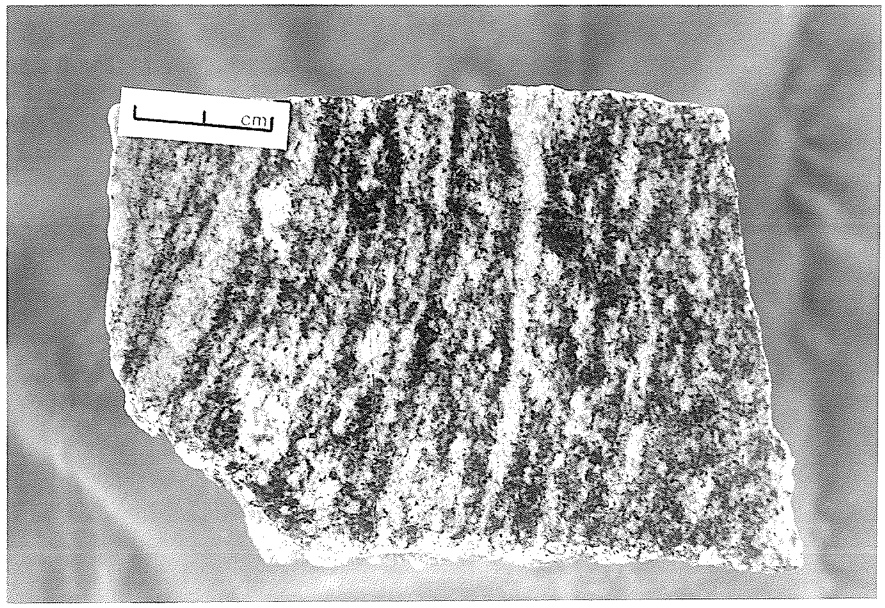


Figure 19 - Well developed gneissosity in trondhjemite. Note development of quartzofeldspathic lenses into continuous felsic segregations. Recrystallization stage 2.

lineation that plunges 30 to 50 degrees easterly (Fig. 37). The foliation and gneissosity commonly trend west-northwesterly parallel to the long axis of the remnant, but they are locally contorted into complex plastic folds on the scale of centimetres (Fig. 20).

In some outcrops, a slight change of foliation across a sharp boundary was observed, with one foliation parallel to the contact and terminating the other. This could be the result of an original magmatic foliation that has been cut off by a slightly later movement of magma, or due to a faulting of the rock in a plastic state.

In stage 3 the gneissosity is partly destroyed by recrystallization, leaving a remnant foliation or almost homogeneous rock. This indicates that an early deformational event preceded the main recrystallization.

(2) Minor Basement Phases

Six minor phases are associated with the trondhjemite in the basement complex. Four phases, pyroxenite, gabbro-diorite, syenodiorite and amphibolite, are older than the trondhjemite, whereas intercalated melanocratic and leucocratic layers are contemporaneous with and/or younger than the trondhjemite. The compositions of three of these phases are summarized in Table 2.

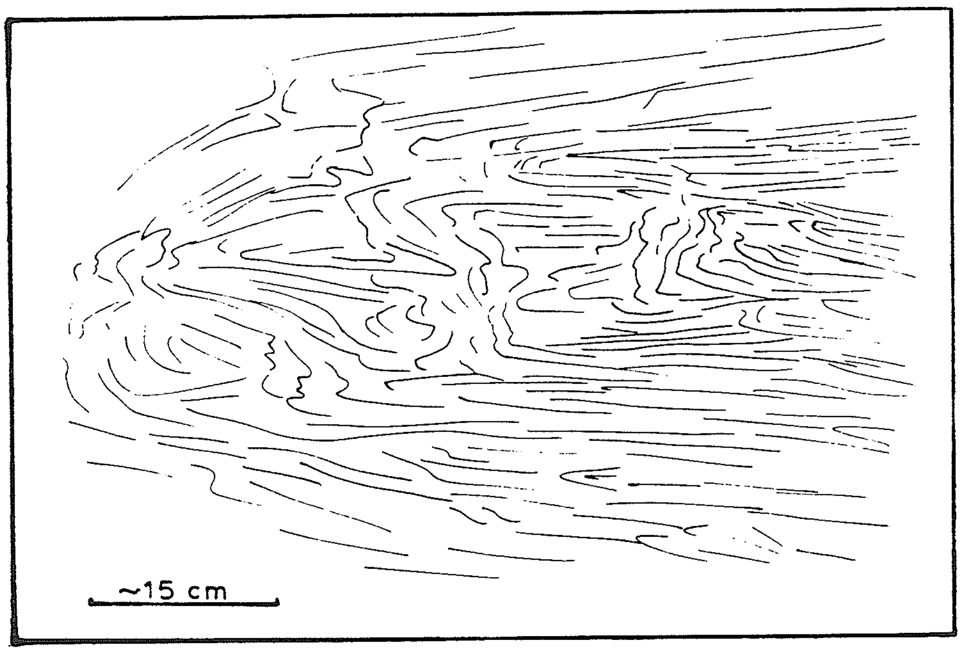


Figure 20 - Complex plastic fold patterns in trondhjemite defined by foliation (sketched from a photograph).

(a) Units Older Than the Trondhjemite

(i) Pyroxenite-Gabbro-Diorite Complex

Three of the four pre-trondhjemite phases are restricted to a melange of medium-grained dioritic and gabbroic intrusions in the southeast end of the basement complex (Fig. 3). Within this is a block of metapyroxenite, several metres square (Fig. 37), that has been intruded by rare dikes of fine-grained porphyritic syenodiorite. The diorite, which includes both phase N-2 of Ayres (1974) and areas previously mapped as mafic metavolcanics (Ayres 1974) is variable in composition, with more felsic diorite intruding more mafic diorite or gabbro (Table 2). The diorite is largely recrystallized to amphibolite but is considered to be intrusive rather than volcanic because of the absence of identifiable volcanic structures and the presence of intrusive relationships. The diorite is well foliated but lacks the gneissosity of the trondhjemite, although discontinuous concordant quartzofeldspathic segregations are locally present. The diorite is distinguished from post-trondhjemite diorite (phase N-4, Ayres 1974) by lower quartz and biotite contents (Table 2).

(ii) Amphibolite Xenoliths

The xenoliths are fine- to medium-grained, oval and rounded to elongated and ragged, and mineralogically similar to the early diorite (Table 2) from which they may have been derived, but are more completely recrystallized. They form up to 5 percent of the trondhjemite adjacent to the diorite in the southeast corner of the remnant, are commonly elongated parallel to the gneissosity of the trondhjemite and are up to several metres long (Fig. 21). Northwest of Adam's Creek (Fig. 37) they decrease in abundance to less than 1 percent, are generally less than 0.5 m long, and are elongated parallel to the gneissosity of the trondhjemite. The xenoliths have locally developed, discontinuous, concordant and discordant felsic segregations that are less than 5 cm thick and composed mainly of plagioclase with minor biotite and hornblende. In some xenoliths, more mafic, hornblende-rich selvages 1 to 5 cm wide occur adjacent to trondhjemite that is depleted in biotite and hornblende compared to trondhjemite elsewhere (Fig. 22). The foliation and gneissosity of the trondhjemite are deflected around the xenoliths. Some xenoliths are broken, with fragments slightly displaced relative to each other and surrounded by trondhjemite (Fig. 21). Hairline fractures in the trondhjemite developed locally where the gneissosity of the trondhjemite is severely deflected around a xenolith.

(b) Intercalated Layers in the Trondhjemite

(i) Leucocratic Layers

Discontinuous, concordant to slightly discordant quartzofeldspathic lenses and layers form up to 5 percent of the trondhjemite in the southeast part of the remnant but decrease in abundance northwestward. They are generally less than 5 cm thick but are up to several metres long. The layers have similar mineralogy (Fig. 15, 16), grain size and textures to the trondhjemite, but have a lower biotite content and lack

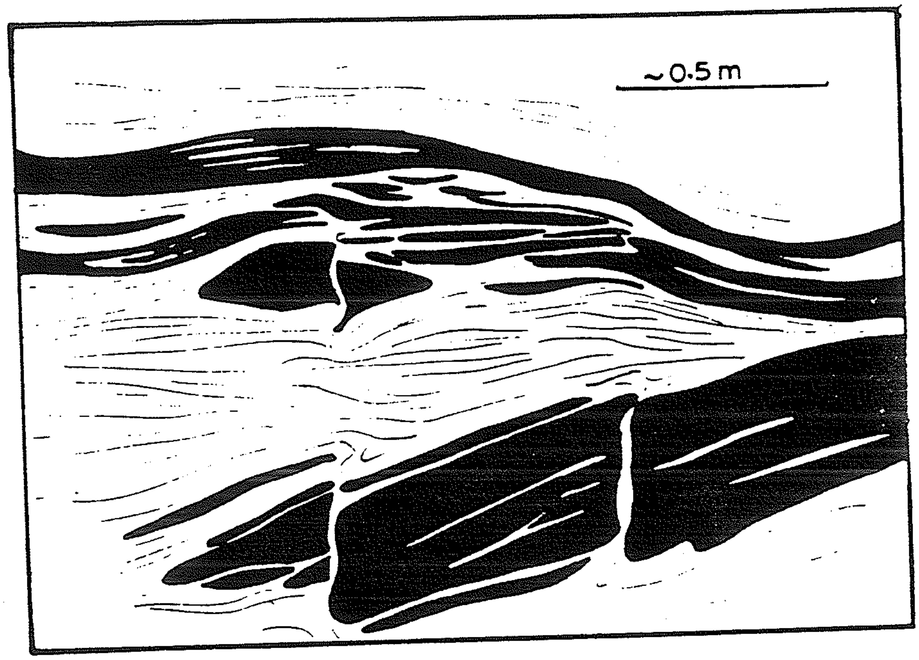


Figure 21 - Typical amphibolite xenoliths (black) in trondhjemite in southeast corner of the basement complex. Crosscutting leucocratic material is trondhjemite. Note that xenolith is fractured whereas trondhjemite host has both fractured and flowed.

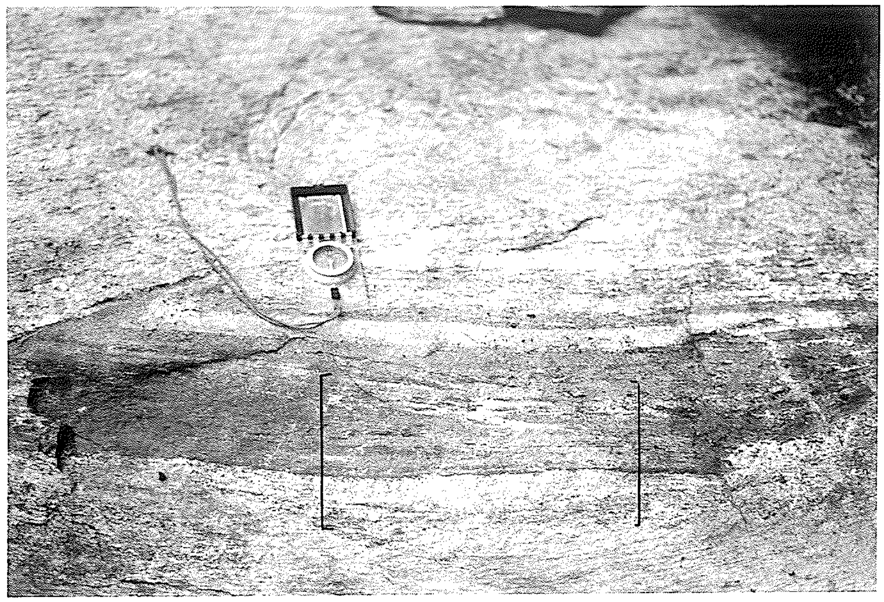
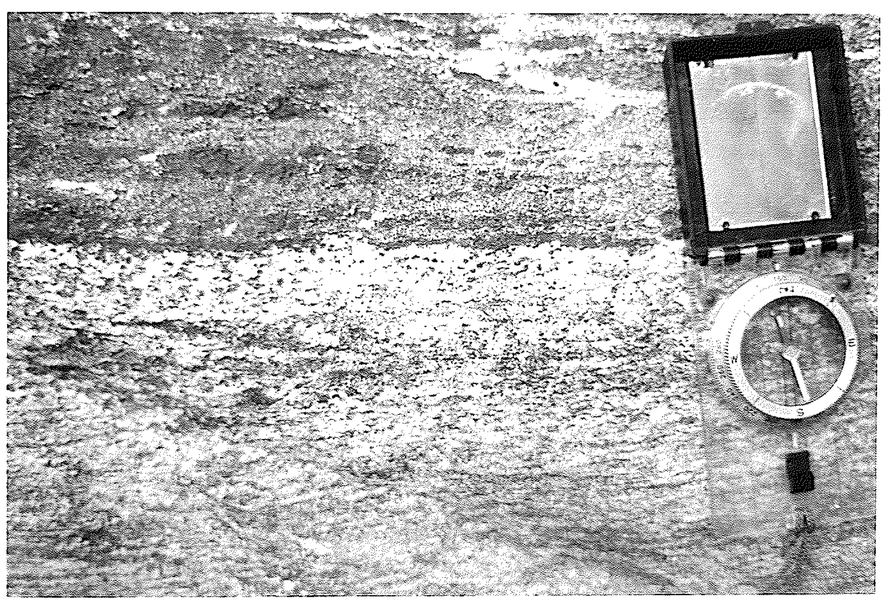


Figure 22 - (a) Thin mafic selvages and felsic segregations in amphibolite xenoliths in trondhjemite. Bracketed area is shown in Figure 22.b.



(b) Close-up of bracketed area in Figure 22.a. Note hornblende-rich selvages on xenolith and relatively leucocratic trondhjemite adjacent of xenolith (probably recrystallization stage 2). Compass is 6.5 cm wide.

hornblende (Table 2, Fig. 8). They are less recrystallized than the trondhjemite in that recrystallization phenomena are present in only some of the leucocratic layers. Contacts between layers and trondhjemite are gradational over a few millimetres and show no evidence of intrusion or movement.

These characteristics suggest that the layers are the result of metamorphic segregation from the trondhjemite rather than deformed dikes, sills or xenoliths. Similar quartzofeldspathic segregations occur in the older diorite complex.

(ii) Melanocratic Layers

Intercalated with the trondhjemite are slightly finer-grained more melanocratic layers. These are foliated and medium to dark grey, with a grain size of 1 to 2 mm. They are mineralogically similar to the trondhjemite (Fig. 15, 16) except for slightly higher biotite, hornblende and locally microcline contents (Table 2). Recrystallization textures are developed to the same extent as in the adjacent trondhjemite, and narrow concordant quartzofeldspathic segregations similar to those in the trondhjemite are common (Fig. 23).

The melanocratic layers are most abundant in the southeast part of the basement complex, where they form as much as 20 percent of the trondhjemite, but they decrease in abundance northwestward and are rare in the northwest part. They are usually less than 1 mm thick, and have sharp contacts with the trondhjemite. A few layers have chilled, cross-

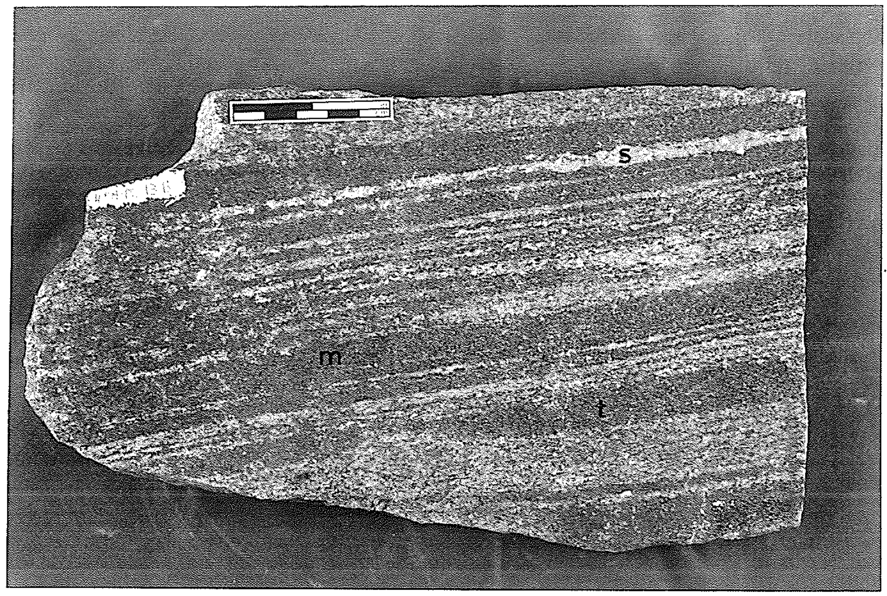


Figure 23 - Fine interlayering of melanocratic layers and trondhjemite (t). Note abundant quartzofeldspathic segregations (s) in melanocratic layers (m). Recrystallization stage 2.

cutting contacts with the trondhjemite. Where abundant, they occur in swarms, with approximately equal amounts of interlayered trondhjemite and melanocratic material (Fig. 23). Some layers consist of discontinuous slivers a few cm thick, separated by thin trondhjemite layers, the foliation of which follows the boundaries of the slivers (Fig. 24). No chilled contacts were found in these slivers.

The presence of chilled, crosscutting contacts indicate that at least some of the layers represent intrusion of a more melanocratic trondhjemite. However, the amount of recrystallization is identical to that in the trondhjemite, indicating that the layers were intruded prior to metamorphism. The discontinuous slivers are probably remnants of melanocratic intrusions that were brecciated and deformed during the metamorphism that caused the recrystallization.

(3) <u>Phases of the North Trout Lake Batholith that intruded the Basement Complex</u>

The younger phases of the North Trout Lake batholith that intruded the basement complex were examined only briefly in the present study, but they are discussed by Ayres (1974). Seventeen younger phases have been identified to date in the batholith, but only three of these, biotite-hornblende diorite (N-4) and leucocratic biotite quartz monzonite (N-17 and N-19) occur within the basement complex (Fig. 3, 37). Older phases border the basement complex but these will not be considered here. Modal compositions of the phases that intruded the basement complex are given in Table 2.

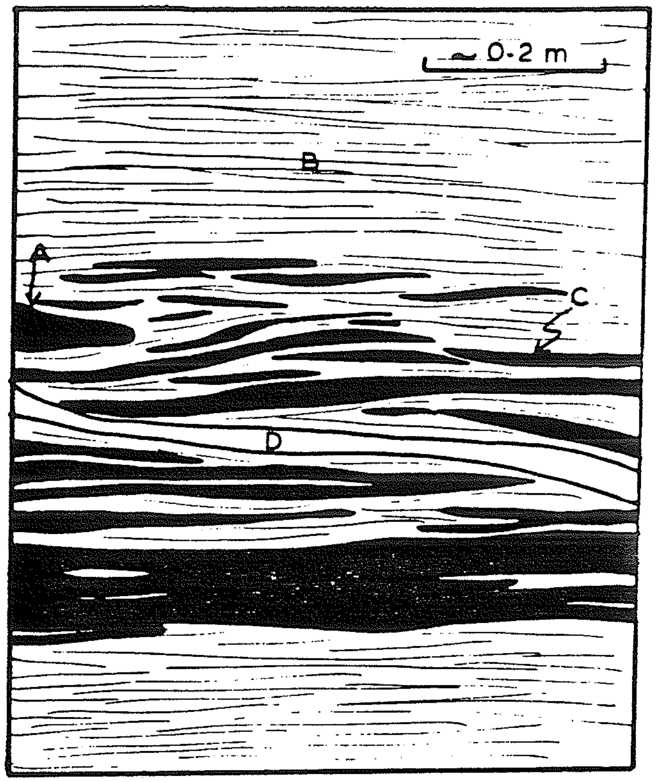


Figure 24 - Interlayered trondhjemite and melanocratic layers, enclosing an amphibolite xenolith and intruded by a late aplite dike (drawn from a photograph); a = amphibolite xenolith; b = trondhjemite; c = melanocratic layers; d = aplite dike.

(a) Biotite-hornblende Diorite (N-4)

Sills of fine- to medium-grained biotite-hornblende diorite up to 25 m wide were intruded passively parallel to the gneissosity of the trondhjemite, although some sills are slightly discordant. The sills are distinguished from earlier diorite of the basement complex by a lack of felsic segregations, higher quartz and lower hornblende contents (Table 2), more weakly developed foliation, and lack of extensive recrystallization. They do not contain trondhjemite xenoliths, but they are intruded by quartz monzonite (N-17, N-19) dikes.

(b) Leucocratic Biotite Quartz Monzonite (N-17, N-19)

Most of the remnant contains at least 50 percent dikes and sills, centimetres to metres wide (Fig. 3, 37) of biotite quartz monzonite, minor aplite, and pegmatite. Except for pegmatite, quartz monzonite is the youngest unit identified in the batholith, but there are at least two and possibly more ages of leucocratic quartz monzonite that can be distinguished only where they are in contact with intervening phases (Ayres 1974). They are distinguished from the basement phases by a slightly pink colour reflecting high microcline content, very sharp, locally crosscutting contacts, lack of any foliation or other metamorphic textures and lower mafic mineral contents.

These leucocratic phases occupy up to 95 percent of some outcrops. In the southeast part of the remnant, the leucocratic phases were intruded as thin sills, concordant to the foliation. Pegmatite occurs only in the northwest part in association with the leucocratic phases, where the trondhjemite is brecciated and commonly occurs as isolated, rotated xenoliths in the quartz monzonite (Fig. 37).

(4) Metamorphism

The stability of the mineral components of the basement remnant makes the metamorphic history difficult to decipher. However, the diorite mineral assemblage in the early pyroxenite-gabbro-diorite complex indicates that the remnant has undergone at least one episode of regional amphibolite facies metamorphism, possibly coincident with metamorphism of the nearby Favourable lake supracrustal sequence (Ayres 1974). The relationships among the foliation and gneissosity, recrystallization and layering of the trondhjemite indicate that the trondhjemite also underwent an earlier event that produced the foliation and gneissosity.

Although field evidence on the shape and size of the isolated areas where stage 3 is developed are inconclusive, they probably represent local zones of increased deformational strain in which recrystallization was accelerated by deformation (Spry 1969, Pitcher and Berger 1972), and continued to affect the rock long after the strain had ceased.

(a) Foliation and Gneissosity

The actual mechanism that concentrated the felsic minerals along concordant planes to produce the foliation and lenticular gneissosity in the trondhjemite is poorly understood. The even development of foliation and gneissosity throughout the remnant without correlation to stages of recrystallization, and partial destruction of gneissosity with the most intense recrystallization (stage 3) indicate that the gneissosity predated the recrystallization. It may have been initiated during the emplacement of the trondhjemite, with possible modification by later metamorphism. The presence of relict primary magmatic textures suggests that some structures could be the result of magmatic flow. However, the deformation twinning, bending and fracturing of crystals, and elongation and lenticular concentration of crystals to form foliation and gneissosity are not normal magmatic flow characteristics. Instead they are structural rearrangements of crystals that collectively allowed the rock to deform upon addition of heat and stress, when it was largely or even wholly crystalline.

Similarly, foliation in most of the North Trout Lake batholith phases is not a flow phenomenon but is due to submagmatic reactions in a stress field (Ayres 1978). In many respects the foliation in younger phases such as N-5 is similar to that in the trondhjemite but is not as strongly developed, and lacks gneissosity (Ayres 1979, personal communication).

(b) Leucocratic Layering and Recrystallization

Since the leucocratic lenses and layers are less recrystallized than the parent trondhjemite, and locally crosscut the early gneissosity, they were probably the result of metamorphic segregation that was initiated during the second metamorphic event. Similar origins have been cited for similar layering in other ancient gneiss complexes (Harris and Goodwin 1976, Archibald and Bettenay 1976, Bridgewater and Collerson 1976, Schmidt 1932, Kerrich and Allison 1978). Both leucocratic layers and recrystallization are more strongly developed in the southeast part of the remnant (Fig. 25), indicating a pressure and temperature gradient across the remnant in which the southeast part was hotter and under greater stress than the northwest part. The intensity of recrystallization is best shown by the distribution of recrystallized quartz and ragged or green biotite (Fig. 26).

This metamorphism was probably caused by the intrusion of the North Trout Lake batholith. The southeast part of the remnant was more strongly affected by the heat and strain accompanying the intrusion. The sill-like character of the quartz monzonite phases (N-17, 19) in the southeast part of the trondhjemite compared to the crosscutting and brecciating dikes in the northwest, also indicates hotter and more plastic conditions in the southeast than in the northwest. Pegmatite (N-20), which is commonly a low temperature phase that is emplaced in areas of tension late in the emplacement history of the batholiths, is concentrated in the northwest part of the remnant.

(c) Relationship Between Emplacement of the Batholith and Metamorphism of the Basement

The heat from the North Trout Lake batholith produced a zone of amphibolite facies metamorphism in the Favourable Lake supracrustal sequence. The amphibolite zone is 1 to 2 km wide and grades inward to

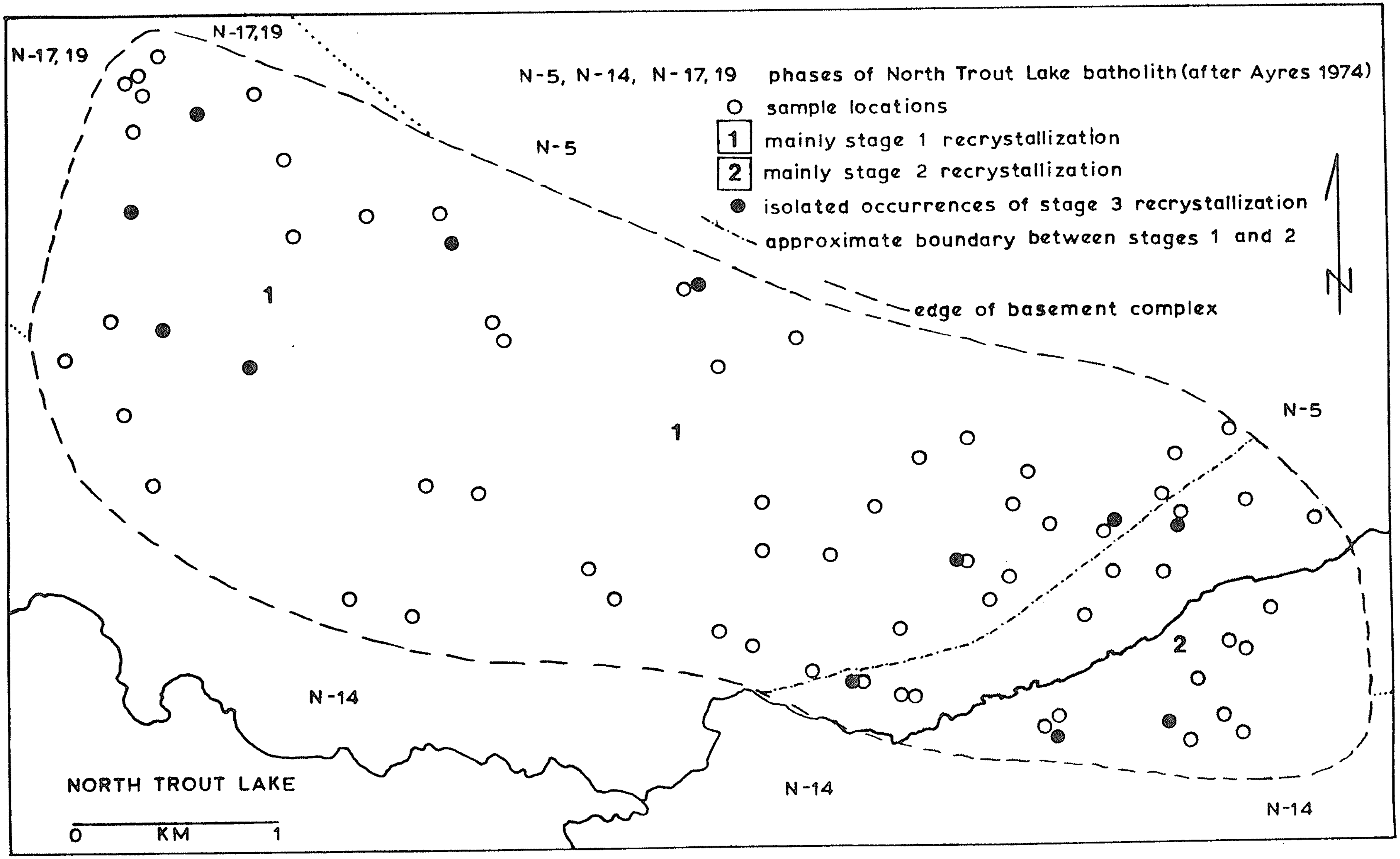


Figure 25 - Stages of recrystallization.

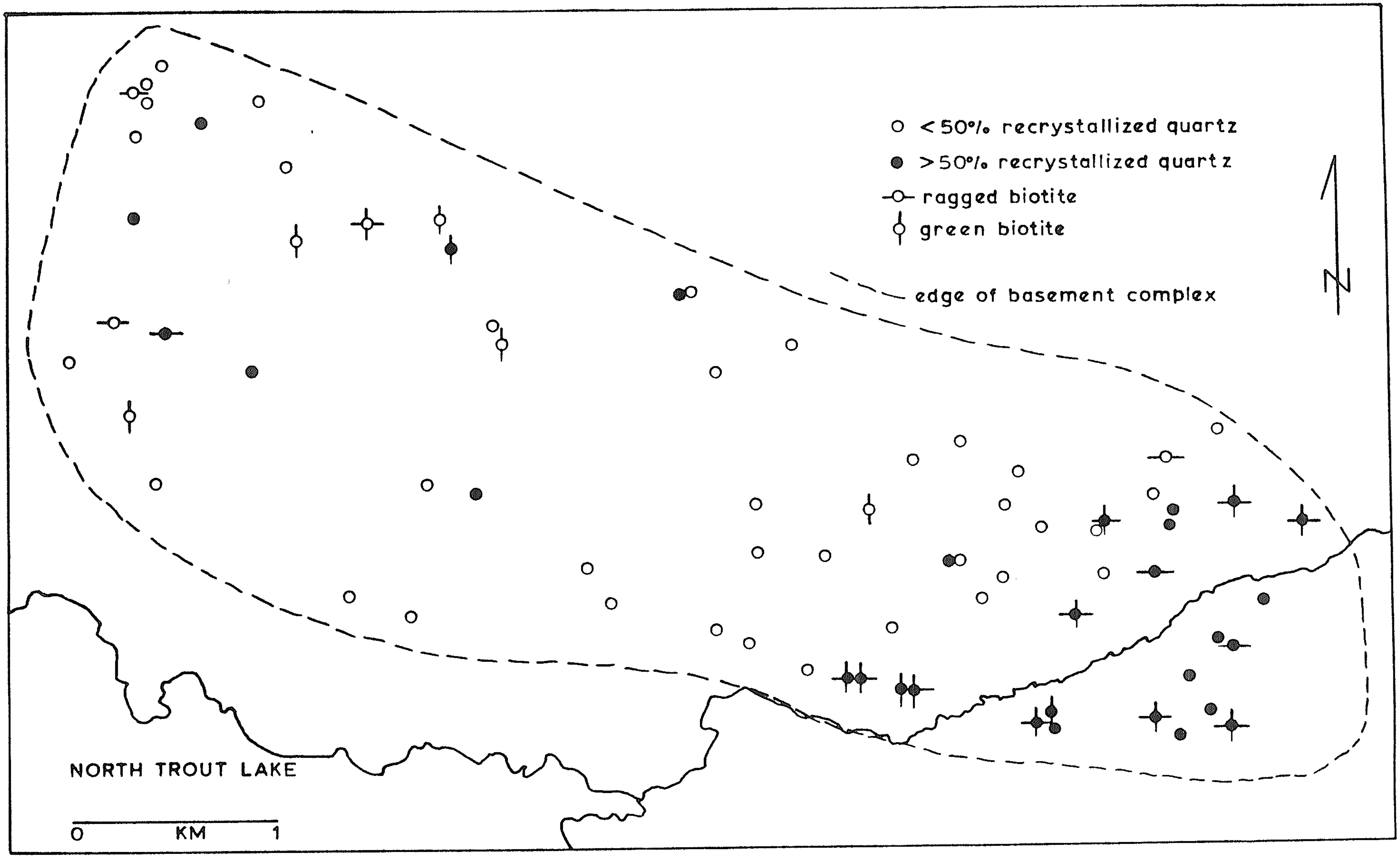


Figure 26 - Distribution of recrystallized quartz and ragged and green biotite.

greenschist facies (Ayres 1978). This large heat source must also have affected the basement rocks.

The stratigraphy of the Favourable Lake supracrustal sequence indicates that the basement trondhjemite was uplifted to its present position during the emplacement of the batholith. The trondhjemite was probably the provenance for the quartz-rich greywacke of Formation K in cycle 3 (Ayres 1977). There is only 1 km now separating the trondhjemite from exposures of cycle 3. However cycle 3 is at least 5 km above the base of the supracrustal section. Therefore subsequent intrusion of the batholith must have uplifted the basement remnant at least 5 km (Ayres 1977, Fig. 6) to its present position near cycle 3 (Table 4).

The time hiatus between the intrusion of the basement trondhjemite and the emplacement of the batholith is indicated by the fact that the first intrusive phase (diorite N-4) of the batholith is the only phase with chilled contacts. This implies that the trondhjemite was relatively cool when the batholithic emplacement began, and that the diorite (N-4) was still warm when the later batholithic phases were intruded.

In the basement, the main metamorphic effect of the emplacement of the batholith was recrystallization. However, the main recrystallization did not occur until relatively late in the intrusive sequence, because the distribution of the recrystallization stages shows no relationship to the distribution of the early batholith phases. The chilled contacts of the phase N-4 diorite indicate that it intruded relatively cold wall rocks. It forms relatively small intrusions that would have

TABLE 4 SEQUENCE OF EVENTS IN THE HISTORY OF THE REMNANT

1. The trondhjemite was emplaced and crystallized at some unknown depth about 2910 Ma. Foliation and gneissosity were probably initiated then but may have been modified by later events. The nature of the country rocks is unknown.

2. Unknown events prior to erosion.

3. The trondhjemite was eroded and exposed to form basement for the supracrustal sequence. The first evidence of the trondhjemite as a subaerial source is the quartz-rich Formation K in cycle 3.

4. The basement was at least partly buried by subsequent eruption of the volcanic sequence.

5. Burial metamorphism affected the basement, but probably had little effect.

6. The diorite phase (N-4) of the batholith intruded the trondhjemite, initiating the warming trend but causing little metamorphism.

7. Each successive intrusive phase of the batholith warmed the trondhjemite to the point where, with the intrusion of Phase N-14, the trondhjemite was hot enough for widespread recrystallization to occur. The recrystallization was strongest at the south-east edge of the remnant where Phase N-14 was widest and would have the greatest effect. A temperature and pressure gradient was in existence across the remnant.

8. During the succession of intrusions the basement was uplifted to its present position opposite cycle 3 of the supracrustal sequence.

9. Phases N-17, N-19 intruded the trondhjemite in the framework of the temperature and pressure gradient, resulting in their variable mode of emplacement across the remnant.

had insufficient heat to produce extensive recrystallization in the trondhjemite. Phase N-5 was intruded mainly on the northeast side of the trondhjemite (Fig. 3), while the main pressure and temperature gradient indicated by the recrystallization pattern trends northwest-southeast.

Phase N-14, a porphyritic biotite granodiorite that occurs between the basement remnant and the Favourable Lake supracrustal sequence (Fig. 3), was the eleventh batholith phase to be emplaced in the area of the remnant. The granodiorite, being more felsic, was probably cooler during emplacement than the more mafic N-5 diorite, which is of similar size. However, the recrystallization appears to be related to the emplacement of the granodiorite (Fig. 25). The heating effect of the early phases was therefore used only to increase the temperature of the trondhjemite, with possible minor incipient recrystallization. By the time the later phases were intruded, their heat and strain energy would cause extensive recrystallization in the already warmed trondhjemite. The heat and strain caused by its emplacement would have been greatest where the granodiorite is widest, at the more highly recrystallized southeast corner of the remnant. Local areas throughout the remnant that have been recrystallized to stage 3 have probably undergone increased deformational strain which would accelerate and prolong the recrystallization process (Spry 1969, Pitcher and Berger 1972).

There is therefore no evidence for a pressure and temperature gradient in the remnant before the intrusion of phase N-14. However, the habit and abundance of the late quartz monzonite phase (N-17, N-19) indicate that the pressure and temperature gradient lasted until the late intrusive phases of the batholith.

It is noteworthy that the early batholith phases did not experience the same recrystallization that affected the trondhjemite (Ayres 1974) during intrusion of later phases. It is possible that this type of extensive recrystallization of minerals into finer-grained mosaics requires a cold, rigid starting material like the trondhjemite, whereas the early batholith phases would have remained hot and plastic during the intrusion of later phases. A type of high temperature submagmatic recrystallization in the early phases of the batholith produced a foliation, but the textures are still magmatic (Ayres 1978).

(5) Chemistry

(a) Major and Trace Elements

Chemical and normative analyses of 16 trondhjemite samples and two melanocratic layers are tabulated in Table 5 and rare earth analyses are given in Table 6. Of the two melanocratic layers, both less than 1 m thick, sample 137 is from a well defined sill and sample 91 is from a layer with diffuse borders. An average chemical analysis of the Favourable Lake trondhjemite with standard deviation and range is compared to other trondhjemites and early gneisses in Table 7.

The most obvious chemical features of the Favourable Lake trondhjemite are its similarity to other plutonic trondhjemites, both metamor-

TABLE 5 CHEMICAL AND NORMATIVE ANALYSES[1]

Chemical Analyses[2]

Sample #	16	19	38	82	86	102	120	125[3]	152	154	189	192	244	274	289	302	91[4]	137[4]	AVG[5]
SiO_2	66.50	69.00	63.90	68.05	65.50	68.00	68.80	65.35	66.85	67.70	64.10	66.55	71.95	62.60	65.95	66.90	65.50	64.95	66.73
Al_2O_3	15.58	15.58	16.41	15.64	16.21	16.84	16.90	16.06	16.94	16.10	16.68	16.20	14.40	16.68	16.28	15.82	15.78	15.56	16.14
Fe_2O_3	2.21	1.82	2.43	2.11	1.87	1.58	1.37	2.36	1.16	1.87	2.45	1.86	1.70	2.62	2.53	2.29	2.39	2.50	2.01
FeO	2.30	1.76	2.88	1.72	2.36	1.04	0.92	2.64	2.22	1.90	2.52	1.94	1.36	2.98	1.80	1.92	2.30	2.62	2.02
MgO	1.70	1.33	1.90	1.37	1.70	0.87	0.77	1.73	1.70	1.37	1.80	1.60	0.80	2.23	1.63	1.47	1.63	1.70	1.50
CaO	3.90	3.30	4.27	3.67	4.00	2.87	3.53	4.20	3.10	3.40	4.40	4.00	2.90	4.93	3.93	4.15	3.80	4.13	3.78
Na_2O	5.12	4.76	4.88	4.88	5.06	5.42	5.78	4.40	4.52	4.96	4.94	5.16	4.62	5.06	5.08	4.94	4.14	4.64	4.97
K_2O	1.35	1.41	1.38	1.16	1.19	2.04	0.88	1.46	1.78	1.36	1.36	1.06	1.01	1.27	1.16	1.10	2.52	2.04	1.31
TiO_2	0.52	0.40	0.62	0.43	0.58	0.26	0.21	0.60	0.64	0.43	0.59	0.48	0.31	0.47	0.52	0.44	0.68	0.70	0.47
P_2O_5	0.13	0.10	0.15	0.11	0.14	0.06	0.10	0.19	0.20	0.09	0.13	0.14	0.10	0.18	0.15	0.12	0.22	0.24	0.13
MnO	0.09	0.08	0.10	0.07	0.08	0.05	0.04	0.08	0.06	0.07	0.09	0.08	0.04	0.12	0.10	0.08	0.08	0.08	0.08
CO_2	0.07	0.09	0.05	0.07	0.08	0.08	0.10	0.13	0.09	0.03	0.12	0.08	0.07	0.10	0.08	0.07	0.12	0.08	0.08
H_2O	0.77	0.63	0.82	0.60	0.69	0.66	0.53	1.04	0.80	0.60	0.65	0.74	0.48	0.85	0.67	0.63	0.83	0.71	0.70
ppm Rb	88	77	72	90	75	92	43	64	97	103	97	52	57	64	92	66	82	64	78
ppm Ni	16	16	18	12	20	11	10	15	10	10	19	21	10	26	13	15	19	15	16
K/Rb	127	152	160	107	132	184	170	189	153	110	116	169	147	164	104	138	254	264	140
Fe_2O_3/FeO	.96	1.03	0.84	1.22	0.79	1.51	1.48	0.89	0.52	0.98	0.97	0.95	1.25	0.87	1.40	1.19	1.41	0.95	0.99
*	1	1	1	1	1	2	2	2	2	1	1	1	1	3	1	1	2	2	

Cation Equivalent Norms[6]

Sample #	16	19	38	82	86	102	120	125[3]	152	154	189	192	244	274	289	302	91[4]	137[4]	AVG[5]
Q	19.25	24.80	16.40	23.46	18.83	19.74	21.52	20.58	22.53	22.34	16.39	20.05	31.08	12.93	19.40	21.29	19.83	17.83	20.66
C	0	0.51	0	0	0	0.70	0.29	0.07	2.66	0.55	0	0	0.75	0	0	0	0	0	0.69
Or	8.03	8.40	8.25	6.93	7.12	12.10	5.20	8.76	10.61	8.10	8.11	6.31	6.07	7.56	6.92	6.57	15.13	12.21	7.81
Ab	46.22	43.04	44.30	44.25	45.94	48.79	51.85	40.08	40.90	44.86	44.73	46.64	42.15	45.71	45.97	44.76	37.74	42.15	45.01
An	15.63	15.83	19.01	17.52	18.21	13.89	16.85	19.88	14.18	16.40	19.49	18.04	13.96	19.06	18.34	17.91	17.29	15.79	17.18
Di	1.71	0	0.75	0.09	0.55	0	0	0	0	0	0.91	0.65	0	2.19	0.19	1.19	0.23	1.86	0.51
He	0.67	0	0.39	0.03	0.20	0	0	0	0	0	0.42	0.17	0	1.31	0.07	0.47	0.08	0.83	0.23
En	3.86	3.70	4.93	3.78	4.47	2.41	2.12	4.85	4.73	3.81	4.55	4.12	2.25	5.10	4.44	3.50	4.45	3.82	3.91
Fs	1.51	1.03	2.54	1.08	1.59	0.23	0.24	2.31	1.85	1.16	2.07	1.09	0.56	3.04	1.50	1.40	1.50	1.70	1.45
Mt	2.12	1.92	2.24	2.04	1.98	1.66	1.43	2.23	1.22	1.97	2.20	1.96	1.81	2.07	2.13	2.05	2.31	2.33	1.93
Il	0.73	0.56	0.87	0.61	0.82	0.36	0.29	0.85	0.90	0.60	0.83	0.67	0.44	0.66	0.73	0.62	0.96	0.99	0.65
Ap	0.27	0.21	0.32	0.23	0.30	0.13	0.21	0.40	0.42	0.19	0.28	0.30	0.21	0.38	0.32	0.25	0.47	0.51	0.26
DI[7]	73.49	76.24	68.95	74.64	71.89	80.63	78.57	69.42	74.04	75.31	69.24	73.01	80.03	66.19	72.29	72.61	72.70	72.19	73.53
CI[8]	10.60	7.21	11.72	7.61	9.60	4.65	4.09	10.23	8.70	7.54	10.99	8.66	5.05	14.37	9.05	9.21	9.54	11.52	8.70
An%	25	27	30	28	28	22	25	33	26	27	30	28	25	29	29	29	31	27	27.5

1 for sample locations, refer to Figures 4 + 37.
2 analyst = K. Ramlal, University of Manitoba (for analytical methods see Appendix 2).
3 location of age determination.
4 melanocratic layer in trondhjemite.
5 average chemical analysis and cation equivalent norm for 16 trondhjemite samples (excluding melanocratic layers).
* = stage of recrystallization.
6 norms were calculated using the computer program "Renorm," which was adapted in 1973 by the GSC Regional and Economic Geology Division from Irvine and Baragar 1971. CO_2 was removed from the analysis and the remainder was normalized to 100%. Ferric iron was adjusted to be TiO_2 + 1.5, and the excess was converted to ferrous iron.
7 DI = differentiation index = normative quartz + albite + orthoclase (weight percent).
8 CI = colour index = normative diopside + hedenbergite + enstatite + ferrosilite + magnetite + ilmenite.

TABLE 6 RARE EARTH ELEMENT ANALYSES[1]

Sample	La	Ce	Sm	Eu	Tb	Yb	Lu
91	53[2]	111	5.7	1.30	0.50	1.73	0.26
125	24	45.0	3.9	0.98	0.49	1.50	0.25

[1]analyzed by H. Y. Kuo, University of Manitoba, by neutron activation

[2]see Figure 34 for plot of chondrite-normalized values

TABLE 7 COMPARISON OF FAVOURABLE LAKE TRONDHJEMITE WITH OTHER TRONDHJEMITIC UNITS

	A			B	C	D	E	F	G	H	I	J	K
wt %	range	X	σ										
SiO_2	62.60–71.95	66.73	2.26	64.41	72.3	59.53	69.10	70.4	68.6	70.9	69.30	67.17	67.60
Al_2O_3	14.40–16.94	16.14	0.74	15.95	14.9	18.04	17.0	14.6	17.1	15.1	16.81	15.79	15.19
Fe_2O_3	1.16–2.62	2.01	0.42	1.46	0.6	6.00	0.69	3.92	0.53	0.15	0.28	1.68	1.75
FeO	0.92–2.98	2.02	0.59	3.81	1.3		1.12		2.36	2.36	1.26	1.41	1.28
MgO	0.77–2.23	1.50	0.33	2.45	0.65	2.86	0.69	1.44	1.30	0.84	1.08	1.09	1.28
CaO	2.87–4.93	3.78	0.51	5.36	3.04	5.79	3.60	3.55	1.83	2.60	3.34	3.54	3.22
Na_2O	4.40–5.78	4.97	0.21	3.39	5.40	4.98	5.69	4.45	3.92	4.37	6.00	4.74	4.15
K_2O	0.88–2.04	1.31	0.29	1.45	1.16	1.55	1.31	1.32	2.72	1.50	1.39	2.10	2.58
H_2O	0.48–1.04	0.70	0.13	0.8	–	0.81	–	–	0.63	0.66	0.50	–	–
CO_2	0.03–0.13	0.08	0.02	–	–	–	–	–	0.29	0.39	0.15	–	–
TiO_2	0.21–0.64	0.47	0.13	0.62	0.31	0.66	0.22	0.3	0.32	0.32	0.23	0.31	0.33
P_2O_5	0.06–0.20	0.13	0.04	–	0.10	–	0.21	–	0.18	0.11	0.03	0.12	0.11
MnO	0.04–0.12	0.08	0.02	0.1	0.02	0.11	0.03	–	0.05	0.06	tr	0.04	0.05
total	–	99.92	–	99.80	99.78	100.33	99.64	99.98	99.7	99.3	100.37	97.99	97.54
ppm Ni	10–26	16	5.07	–	23	–	–	–	15	10	–	13	18
ppm Rb	43–103	78	18.07	–	70	–	267	68	–	–	–	44	77
K/Rb	126–222	140	–	–	163	–	49	194	–	–	–	395	278
FeO/Fe_2O_3	–	0.99	–	0.38	0.46	–	0.62	–	0.22	0.06	0.22	1.19	1.37

A – Favourable Lake trondhjemite (from Table 5, 16 samples)
B – Nockolds' average hornblende-biotite trondhjemite (Nockolds 1954, 22 samples)
C – Best original composition of Uivak I gneiss (Bridgewater and Collerson 1976)
D – Tonalite gneiss, Lac Seul area, Ontario (Harris and Goodwin 1976, 3 samples)
E – Northern Light gneiss (Hanson and Goldich 1971)
F – Ancient gneiss complex, South Africa (Condie and Hunter 1976)
G – North Trout Lake batholith, phase N-14 (Ayres 1974, sample 105, biotite granodiorite)
H – North Trout Lake batholith, phase N-16 (Ayres 1974, sample 352, biotite trondhjemite)
I – Trondhjemite, Trondhjem, Norway (Goldschmidt 1916)
J – Average of 41 grey gneisses, Angmagssalik, east Greenland (Sheraton and others 1973)
K – Average of 22 Lewisian grey gneisses, Harris and Lewis, Outer Hebrides (Sheraton and others 1973)
σ – one standard deviation
X – average

phosed and unmetamorphosed, and its relatively restricted compositional range. Although metamorphic effects partly obscure trends on Larsen variation diagrams (Fig. 27), the general trends of decreasing CaO, MgO, TiO_2, P_2O_5, MnO and total FeO (FeO^T), increasing SiO_2, and constant K_2O and Na_2O are normal magmatic trends. This supports the petrographic and field evidence that the trondhjemite has a magmatic origin.

Some of the oxides show consistent areal variations. These do not coincide with metamorphic zones defined by recrystallization, and probably reflect minor primary chemical variations within the trondhjemite. In general, K_2O, total iron as FeO and Rb are slightly higher and Na_2O slightly lower in the south part of the remnant than the north part (Fig. 28).

Total iron content is similar to that of other trondhjemites, but the Fe_2O_3/FeO ratio is approximately 1, which is considerably higher than most other trondhjemites (Table 5). To check for possible analytical error, samples that covered the range of Fe_2O_3/FeO values were reanalyzed. The new analyses had slightly higher FeO values, but the difference is minor (Appendix 4). In samples with relatively high Fe_2O_3, epidote is slightly more abundant (Fig. 29), which may reflect oxidation during metamorphism. However, there is no correlation among the epidote content and the Fe_2O_3/FeO ratio or the degree of recrystallization. More specific correlations among Fe_2O_3/FeO, Fe_2O_3, epidote contents and metamorphism are not possible because of the small number of analyses.

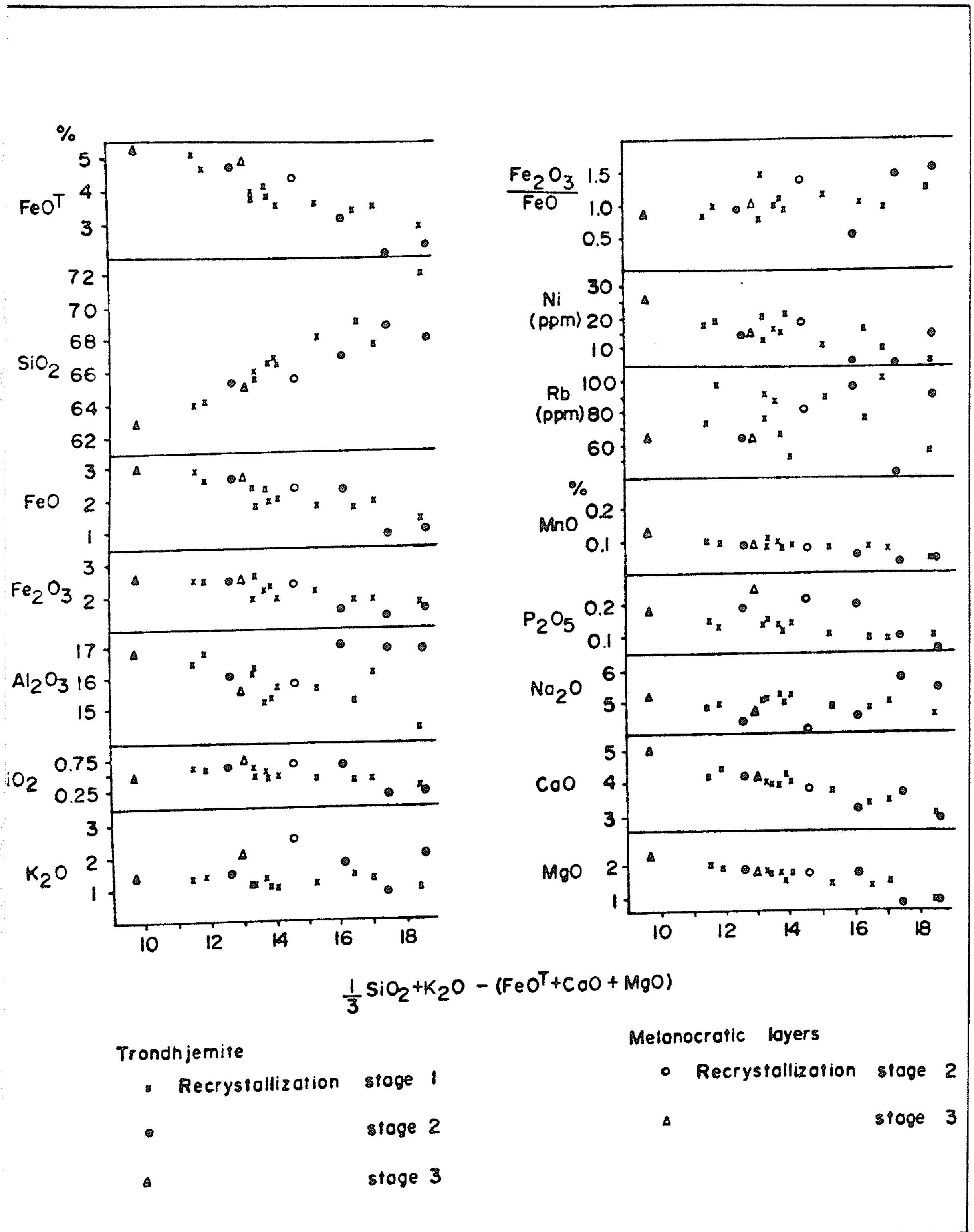


Figure 27 - Larsen Variation Diagrams.

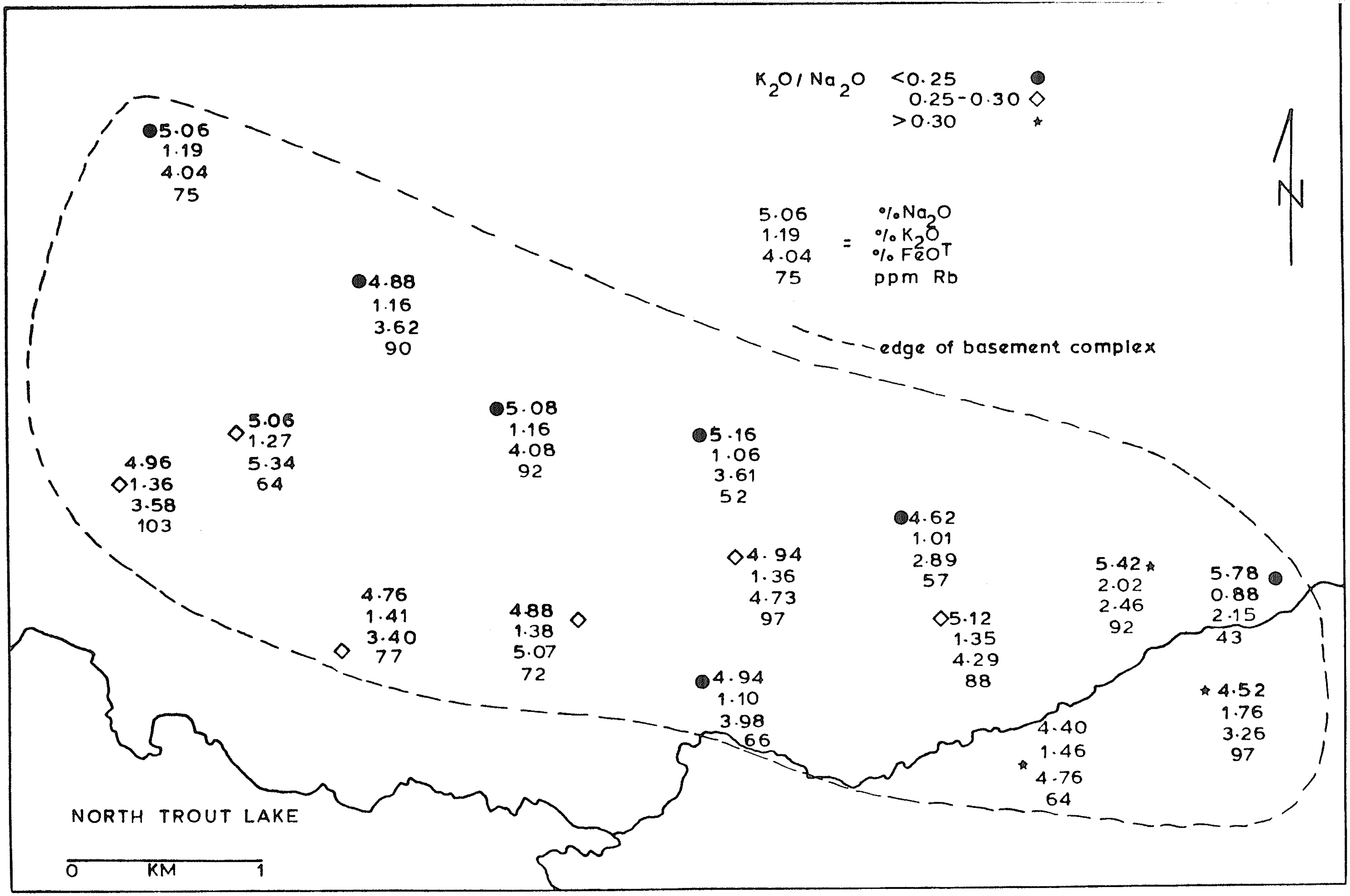


Figure 28 - Variations in concentration of: percent Na_2O, percent K_2O, percent total FeO, and ppm Rb.

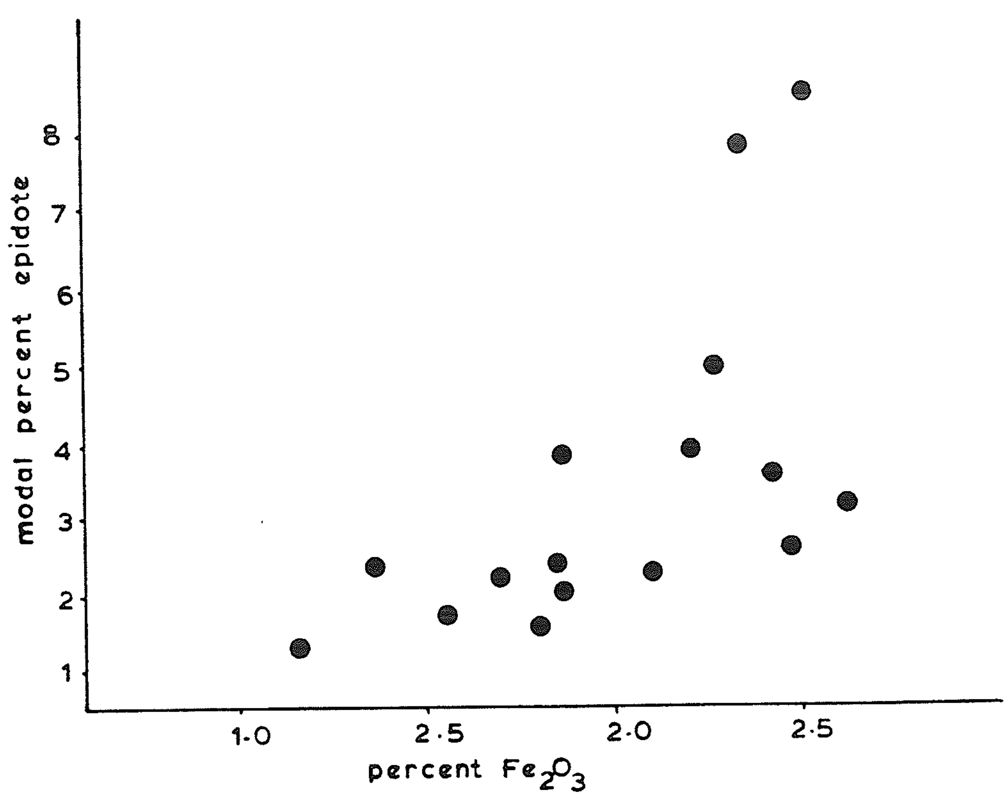


Figure 29 - Correlation between Fe_2O_3 and modal epidote content.

The K/Rb ratios are in the same range as other ancient gneisses (Fig. 30), but define a field of lower K/Rb ratios than representative Archean metavolcanics and greywackes. The less recrystallized samples are well clustered and give an average K/Rb ratio of 133. This is well below 232, suggested by Shaw (1968) as the lowest K/Rb ratio for rocks of recycled anatectic origin, and supports a primary igneous origin. The K/Rb ratio for the more recrystallized samples is slightly higher.

Barth-Niggli norms can also be used (Table 5) to confirm the igneous origin of the trondhjemite. On a plot of Niggli al-alk versus Niggli c (Fig. 31), which Leake (1964) used to discriminate between igneous and sedimentary rocks, the trondhjemite samples plot in a small compact group in the igneous field. This discriminant would be affected by any change in the major element composition produced by metamorphism, but to minimize this effect, only rocks from recrystallization stage 1 were used.

The Ni values do not define a good trend (Fig. 27). In part this reflects the correlation between Ni and modal hornblende, which is very erratic in the trondhjemite (Fig. 32).

Although the trondhjemite illustrates good magmatic trends, there is considerable scatter on most of the oxide plots (Fig. 27). Some of the scatter may be caused by problems inherent in analyzing a small population of medium-grained, gneissic, locally porphyritic samples. However, some of the variation can be directly correlated to the stages of recrystallization. Values for Al_2O_3, total iron as FeO, Na_2O, K_2O,

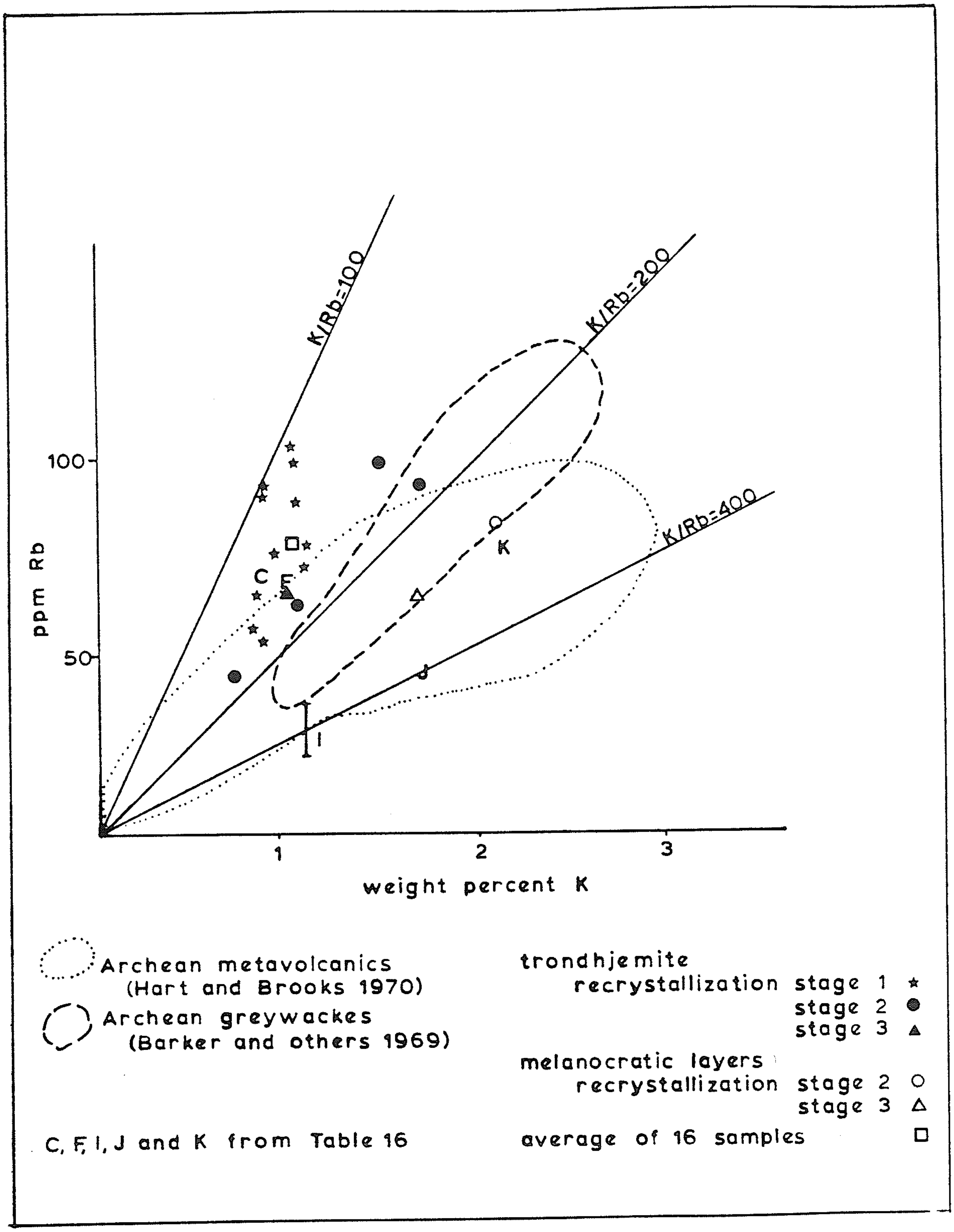


Figure 30 - K/Rb Ratios.

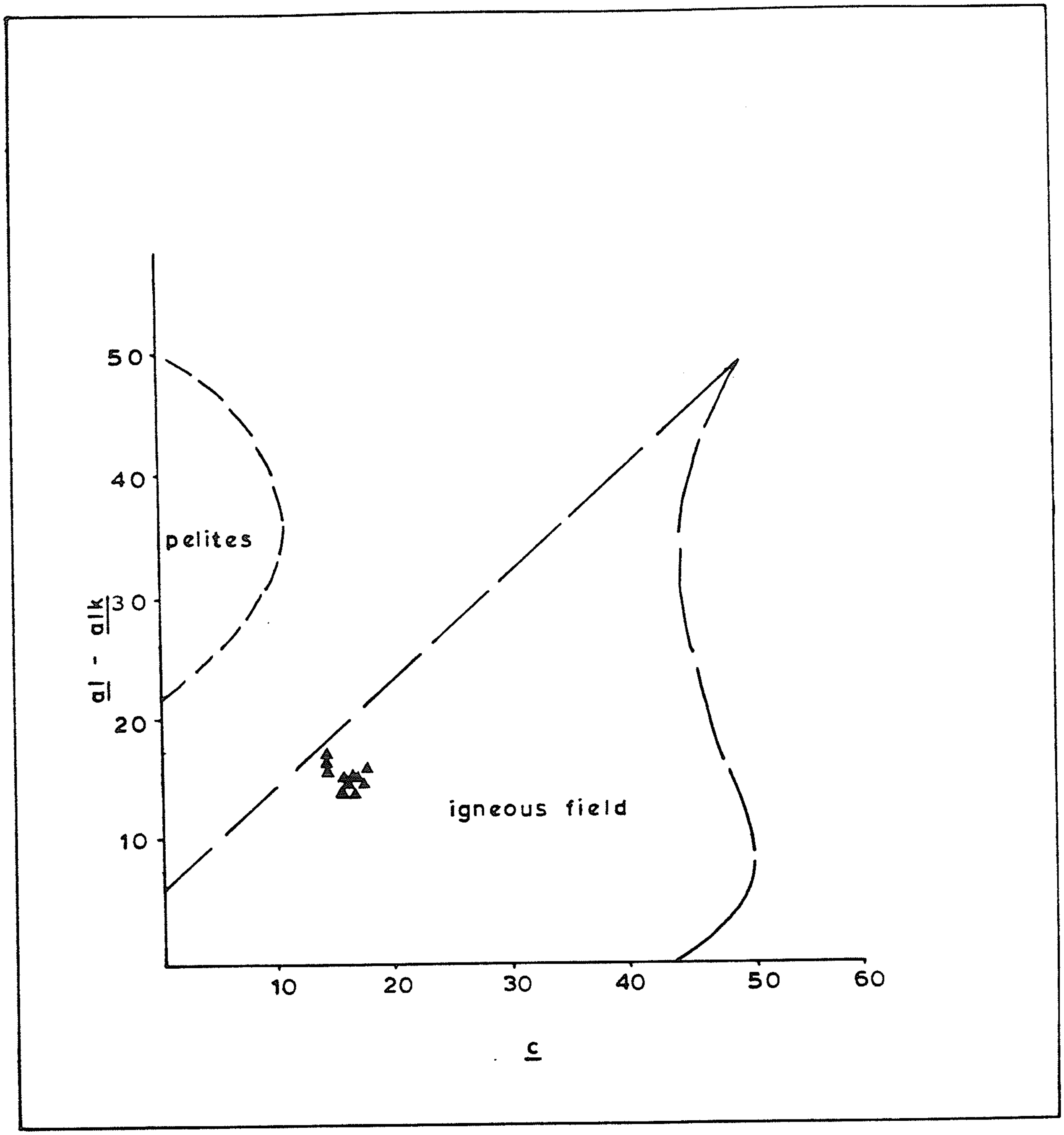


Figure 31 - Niggli al-alk versus c discriminant for igneous rocks (after Leake 1964).

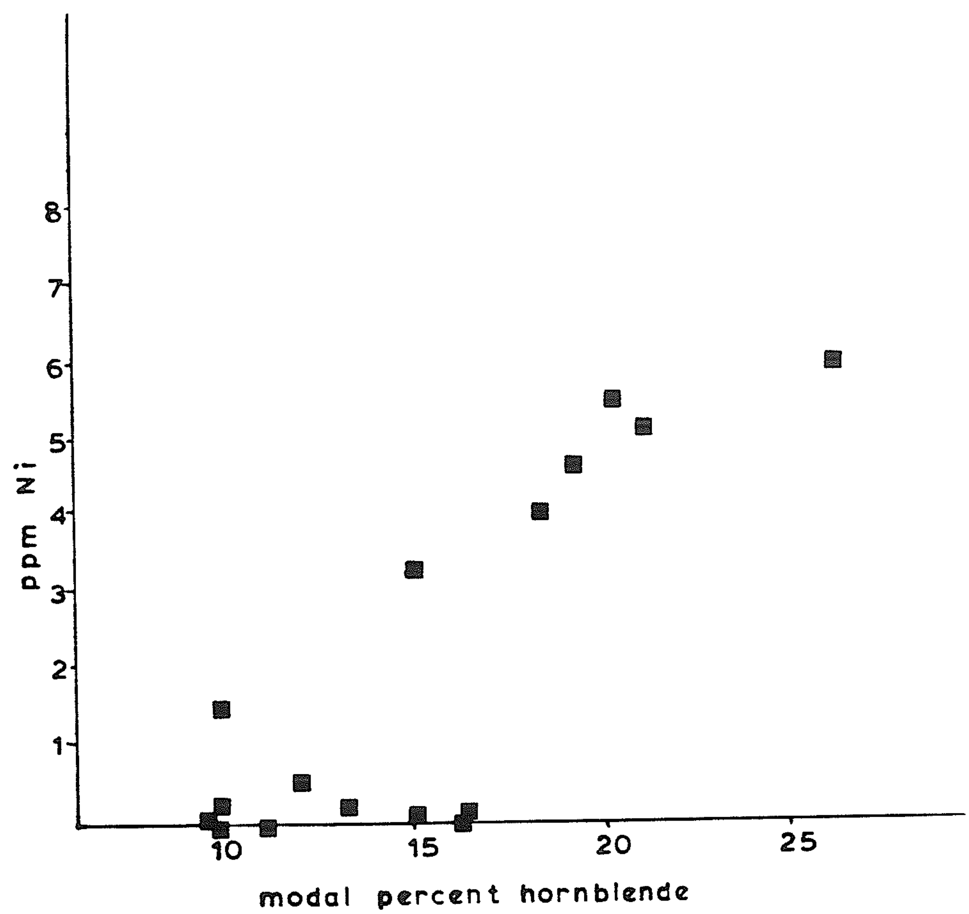


Figure 32 - Correlation between Ni concentration and modal hornblende content.

P_2O_5, K/Rb ratio and Fe_2O_3 (Fig. 30) ratio are all more scattered in stage 2 samples than in stage 1 (Fig. 27, 30). This scatter probably reflects the remobilization of elements to form leucocratic lenses and layers, thereby increasing the inhomogeneity of the rock. The single sample in recrystallization stage 3 has a very low Larsen index but plots on the trondhjemite trend (Fig. 27). The lack of additional analyses of stage 3 samples hampers any correlation between stage 3 metamorphism and chemical trends.

On a normative quartz-albite-orthoclase plot (Fig. 33) the trondhjemite samples plot on the sodic trend defined by the younger members of the North Trout Lake batholith. The more recrystallized samples are much more scattered, with three of the samples displaced toward the orthoclase corner. This demonstrates the scattering effect of the metamorphism.

The two samples from melanocratic layers (samples 91, 137 on Table 4) have more K_2O, TiO_2, total iron as FeO and P_2O_5, a higher K/Rb ratio, and slightly less SiO_2 and Na_2O than the main trondhjemite. Both samples plot closer to the orthoclase corner on the quartz-albite-orthoclase diagram (Fig. 33) than the main trondhjemite. The higher K_2O, TiO_2 and P_2O_5 relative to the trondhjemite are reflected by increases in modal biotite, sphene and apatite.

Rare earth element (REE) abundances (Table 6) of one trondhjemite sample (125) and one sample of a melanocratic layers (91) normalized to chondritic values are plotted on Figure 34, with trends from other rock

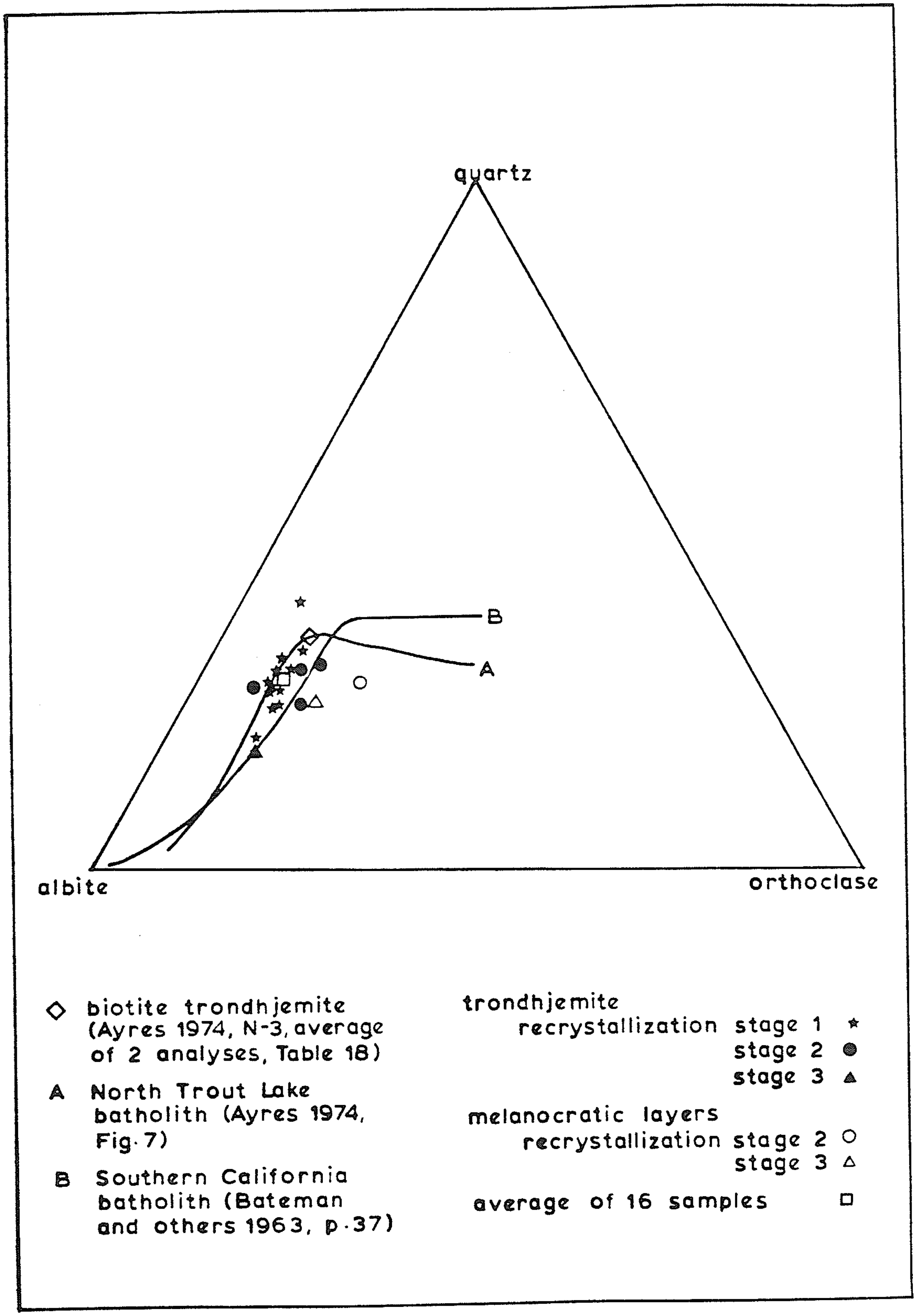


Figure 33 - Quartz-albite-orthoclase normative ternary diagram.

suites for comparison. Sample 91 is more melanocratic than sample 125 and has higher K_2O, Fe_2O_3/FeO and K/Rb ratios. The two samples have similar trends, although total abundances in the melanocratic layer are slightly higher than in the trondhjemite. The trend is light REE enriched, has a flat heavy REE trend, and does not have a Eu anomaly. The other Archean trondhjemites and tonalites on Figure 34 show a range of total abundances that bracket the Favourable Lake samples, and with the exception of the Saganaga tonalite they also tend to be enriched in light REE's with flat heavy REE patterns.

Rare earth element abundances have generally been considered to be immobile during hydrothermal and low grade metamorphism (Frey and others 1968, 1975) although some recent investigations (Ludden 1978, Mason 1978) have shown that they are locally mobile. The amphibolite facies metamorphism has caused remobilization of some elements, particularly in the more highly recrystallized samples, but there is no direct evidence for remobilization of REE's. The two samples analyzed for REE's both have stage 2 recrystallization but the major elements of the trondhjemite sample (125) plot very close to the trend defined by stage 1 samples (Fig. 27). In the absence of direct evidence of metamorphic remobilization of REE's, the REE data are considered to be primary magmatic trends.

(b) <u>Genesis of the Trondhjemite Magma</u>

Chemical, field and petrographic evidence have shown that the basement remnant beneath the Favourable Lake volcanic complex was

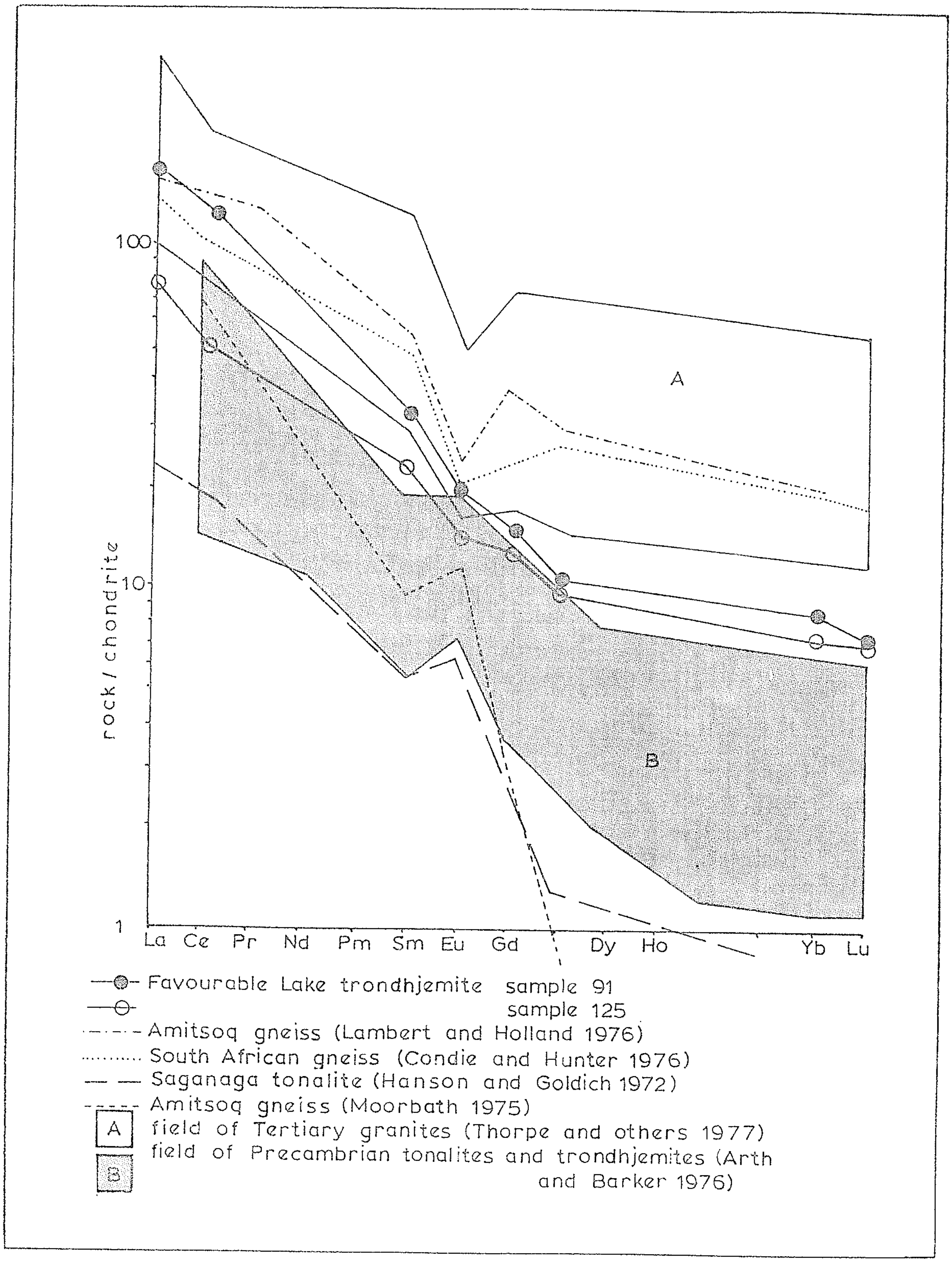


Figure 34 - Chondrite normalized REE plot for some leucocratic plutonic rocks.

originally a plutonic trondhjemite (Table 8). However, as Bridgewater and Collerson (1977) have pointed out, the petrogenetic history of Archean gneissic rocks is difficult to document because of the effects of approximately 3000 Ma of metamorphism and tectonism.

Physical conditions and environment of plutonism during the Archean were very different from present conditions. There was probably an almost continuous, thin (5 to 10 km, Phinney and others 1978) sialic surface layer (Hargraves 1976, West 1978, Baragar and McGlynn 1976). Studies of radioactive heat gradients and komatiites indicate much steeper geothermal gradients (Fig. 35) and higher heat flow (Green 1975, Brooks and Hart 1974) in the Archean. This would preclude subduction, at least in the Cenozoic and Mesozoic mode of plate tectonics, because the plates would deform plastically rather than act as solid entities. With a higher geothermal gradient, melting would occur at shallower crustal levels.

The REE abundance trends of the Favourable Lake trondhjemite are very similar to those of the 3000 Ma Lac Seul trondhjemite gneiss (Fig. 36, Chou 1978). Their similarity in age, geographic location, and geology (Harris and Goodwin 1976) suggests similar origins for the two ancient gneisses. Their REE trends and major element compositions generally fit the high Al_2O_3 trondhjemite suite of Barker and others (1976). Barker and Arth (1976) and Arth and others (1978) suggested that such magmas may have been produced by up to 40 percent partial melting of an amphibolite parent at depths of less than 50 or 60 km.

TABLE 8 SUMMARY OF EVIDENCE SUPPORTING A PLUTONIC IGNEOUS ORIGIN

FIELD DATA

1. Presence of xenoliths
2. Dikes of ancient trondhjemite cutting older pyroxenite-gabbro-diorite complex
3. Lack of any sedimentary or volcanic structures

PETROGRAPHIC DATA

1. Consistent mineralogy and composition over 15 km^2 area
2. Presence of relict plagioclase phenocrysts with Carlsbad-albite twinning, and discontinuous and oscillatory zoning
3. Groundmass plagioclase with Carlsbad-albite twinning and oscillatory zoning
4. Groundmass textures are typical of felsic plutonic rocks, although they are partly obscured by recrystallization; medium-grained quartz is interstitial between subhedral, equant twinned grains of plagioclase
5. Modal data shows typical igneous composition with only small variations from the mean

CHEMICAL DATA

1. Similarity to other trondhjemites in major element and trace element data
2. Igneous compositional trends in less recrystallized samples on variation diagrams
3. Average K/Rb of less recrystallized samples is similar to that of other plutonic rocks but lower than that of volcanic or metasedimentary rocks
4. Plots in igneous field on Niggli *al*-*alk* versus *c* diagram

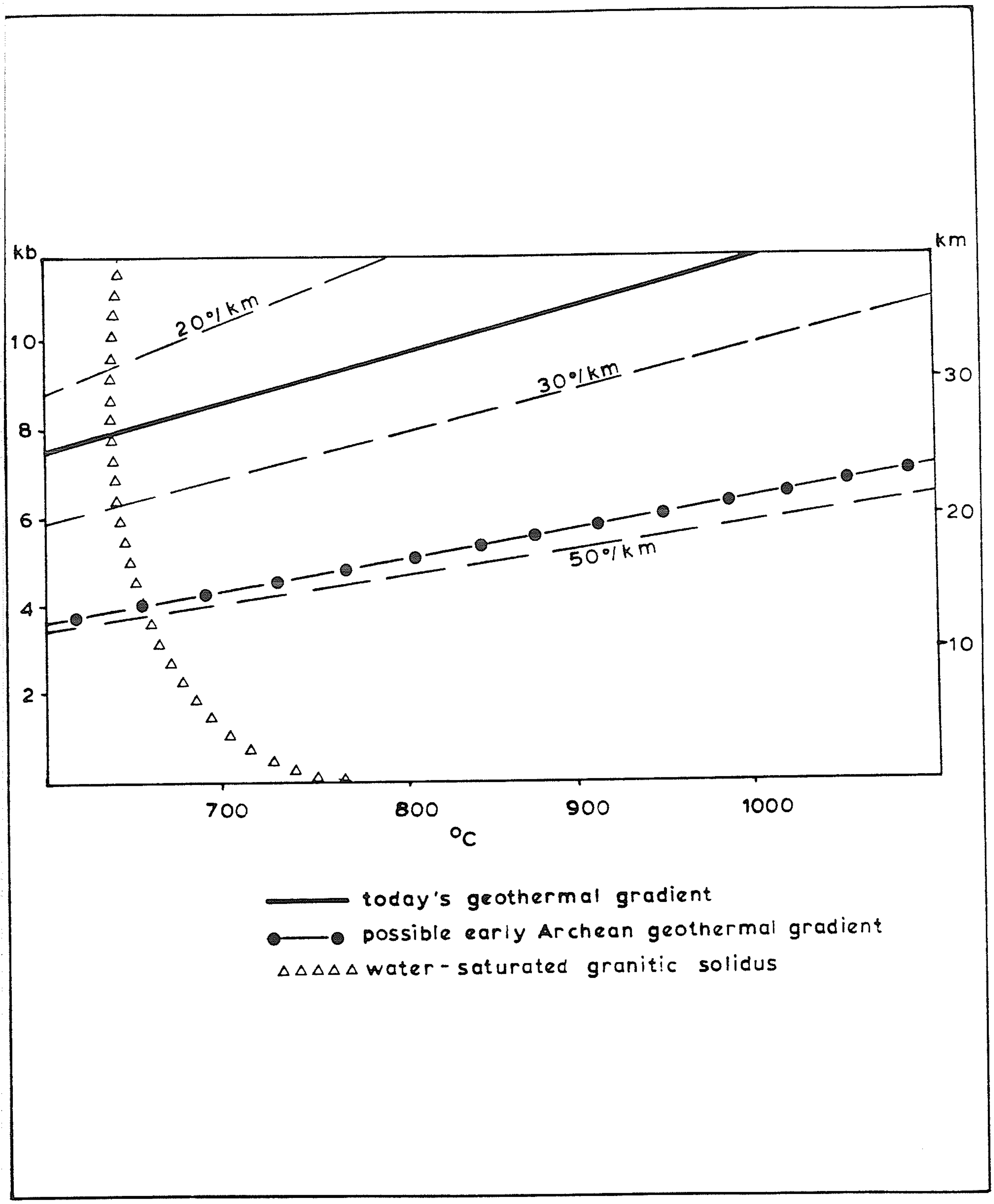


Figure 35 - Geothermal gradients (after Brown and Fyfe 1972).

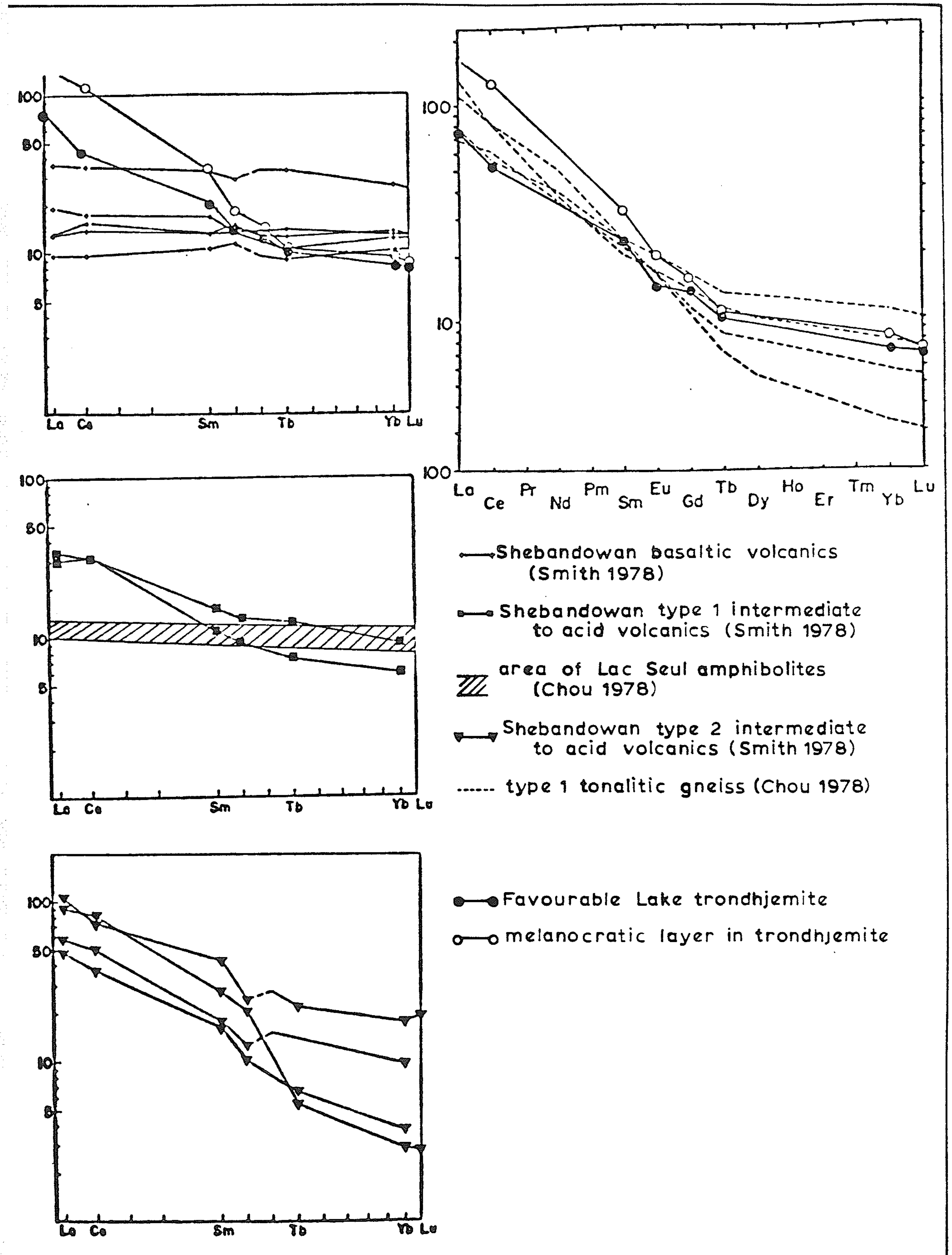


Figure 36 - Rare earth element abundance trends.

The REE trends of the Favourable Lake trondhjemite are compared to those of the Lac Seul amphibolites (Chou 1978) and the Shebandowan volcanic suite (Smith 1978) in Figure 36. The trends of the type 2 intermediate to acid volcanics and the trondhjemite are very similar, indicating that the trondhjemite could have been produced by almost the same process that produced the volcanic magma. The volcanic magma was probably derived by partial melting of basalt or amphibolite.

The similarity among the heavy REE patterns of the Shebandowan basalts, the Lac Seul amphibolites and tonalite and the Favourable Lake trondhjemite (Fig. 36) also suggests that partial melting of an amphibolite parent could have produced the trondhjemite, if hornblende or garnet were residual mineral species. Residual hornblende and garnet both produce a light REE enrichment and heavy REE depletion in the magma relative to the parent, and the effects of the two minerals on REE abundances are very similar (Arth and Barker 1976). However, at the depth level where amphibolite would begin to melt, hornblende would probably be the residual phase. Garnet is more likely to be residual in higher pressure zones. The heavy REE depletion caused by hornblende could be offset by minor associated plagioclase fractionation, which would enrich the magma in heavy REE's (Arth and Barker 1976).

In contrast to Arth and Barker (1976), Chou (1978) stated that residual hornblende will produce a flat heavy REE pattern rather than a depleted trend. In this case simple hornblende fractionation from a basalt parent magma similar to the Lac Seul amphibolite could produce the Favourable Lake trend.

C. CONCLUSIONS

The trondhjemite gneiss at North Trout Lake, near the Favourable Lake supracrustal complex, was dated at 2910 Ma by Krogh and Davis (1971). It is part of an old pluton of unknown extent, which intruded and contains xenoliths of an even older pyroxenite-gabbro-diorite complex. The trondhjemite is intruded in turn by thin melanocratic trondhjemite sills that are commonly brecciated. Prior to the intrusion of the younger North Trout Lake batholith, a strongly foliated to gneissic fabric was produced in the trondhjemite. During the formation of the Favourable Lake supracrustal complex, the trondhjemite probably was a provenance for the quartz-rich sediments in cycle 3. It was uplifted at least 5 km by the intrusion of the North Trout Lake batholith.

During emplacement of the later batholith phases, the trondhjemite was heated, strained and recrystallized to varying degrees by successive intrusions. Early phases progressively heated the remnant and recrystallization culminated with the intrusion of phase N-14. Because phase N-14 was intruded to the south and east of the remnant, that part of the trondhjemite was more strongly heated, deformed and recrystallized than the rest of the trondhjemite. This is reflected in the variable mode of emplacement of the late leucocratic phases (N-17, N-19) across the remnant. The leucocratic phases were intruded concordantly into the hotter, more strained southeast area, but they brecciated the trondhjemite and intruded it in greater volumes in the cooler, less strained northwest part of the basement remnant.

The metamorphism has masked some of the primary features of the trondhjemite. However, the presence of xenoliths and plagioclase phenocrysts, and typical igneous trends on chemical variation diagrams indicate that the trondhjemite is a plutonic igneous rock. Chondrite-normalized rare earth element abundances suggest that it had a shallow depth of origin, and is probably a partial melt of amphibolite or basalt.

REFERENCES

Anhaeusser, C.R., 1973. The evolution of the early Precambrian crust of Southern Africa; Phil. Trans. R.Soc.Lon.A. 273, p 359-388.

Anhaeusser, C.R., Mason, R., Viljoen, M.J. and Viljoen, R.P., 1969. A reappraisal of some aspects of Precambrian Shield geology; Geol. Soc.Am.Bull 80, p 2175-2200.

Archibald, N.J. and Bettenay, L.F., 1976. Indirect evidence for tectonic reactivation of pre-greenstone sialic basement in Western Australia; Earth Plant. Sci.Lett., 33, p 370-378.

Arth, J.F. and Barker, F., 1976. Rare earth partitioning between hornblende and dacitic liquid and implications for the genesis of trondhjemite-tonalitic magmas; Geology, 4, p 534-536.

Arth, J.G., Barker, F., Peterman, Z.E. and Friedman, I., 1978. Geochemistry of the gabbro-diorite-tonalite-trondhjemite suite of southwest Finland and its implications for the origin of tonalitic and trondhjemitic magmas; Jour. Petrology, 19, p 289-316.

Arth, J.F. and Hanson, G.N., 1975. Geochemistry and origin of the early Precambrian crust of northeastern Minnesota; Geochim et Cosmochim. Acta, 39, p 325-362.

Ayres, L.D. 1974. Geology of the Trout Lakes Area, District of Kenora (Patricia Portion); Ont. Dep. Mines, G.R. 13, 197 p.

Ayres, L.D. 1977. Importance of stratigraphy in early Precambrian volcanic terranes; cyclic volcanism at Setting Net Lake, northwestern Ontario; Geol.Ass.Can.Sp.Paper 16, p 243-264.

Ayres, L.D. 1978. Metamorphism in the Superior Province of northwestern Ontario and its relationship to crustal development in Metamorphism in the Canadian Shield, Geol.Surv.Can., Paper 78-10, p 25-36.

Bald, R.C. 1977. Pre-volcanism Archean sialic crust Gundy-Tannis Lake area, northwestern Ontario; Cen.Precambrian Stud. 1977 Annual Report, p 98-105.

Baragar, W.R.A. and McGlynn, J.C. 1976. Early Archean basement in the Canadian Shield: A review of the evidence; Geol.Surv.Canada Paper 76-14, 20 p.

Baragar, W.R.A. and McGlynn, J.C. 1978. On the basement of Canadian greenstone belts: Discussion; Geoscience Canada S, p 13-15.

Barker, F., Peterman, Z.E. and Hildreth, R.A. 1969. A rubidum-strontium study of the Twilight Gneiss: West Needle Mts, Colorado; Contr. Mineral. Petrol., 23, p 271-282.

Barker, F. and Peterman, Z.E., 1974. Bimodal tholeiitic-dacitic magmatism and the early Precambrian crust; Precambrian Research, 1, p 1-12.

Barker, F., Peterman, Z., Millard, H.T., 1975. Geochemistry and Rb-Sr age of Goldschmidt's type trondhjemite and associated rocks, Norway; Geol.Soc.Am.Abstracts with Programs 7, no 7, p 988-989.

Barker, F., Arth, J.G., Peterman, Z.E., and Friedman, I., 1976. The 1.7 to 1.8 BY old trondhjemites of southwestern Colorado and New Mexico: geochemistry and depths of genesis; Geol.Soc. of Am.Bull., 87, p 189-198.

Barth, T.F.W., 1952. Theoretical Petrology; Wiley and Sons, New York, p 387.

Bass, M.N., 1961. Regional tectonics of part of the Canadian Shielf; Jour.Geol. 69, p 668-702.

Bastin, E.S., 1909. Chemical composition as a criteria for identifying metamorphosed sediments, Jour.Geol. 17, p 445-472.

Bateman, P.C., Clark, L.D., Luber, N.K., Moore, J.D. and Rinehart, C.D., 1963. The Sierra Nevada batholith, a synthesis of recent work across the central part; U.S.G.S. Prof.Paper 414-D, 46 p.

Bateman, J.D., 1940. Geology and gold deposits of the Uchi-Slate Lakes area; Ont.Dept.Mines Ann.Rep., 49, pt 8, p 1-43.

Beakhouse, G.P., 1977. A subdivision of the western English River subprovince; Can.J. Earch Sci., 14, p 1481-1489.

Bell, C.K., 1971. Boundary geology, Upper Nelson River area, Manitoba and Northwestern Ontario; Geol.Assoc.Can., Sp.Paper 11, p 11-39.

Bridgewater, D. and Collerson, K.D., 1976. The major petrological and geochemical characters of the 3600 MY Uivak Gneisses from Labrador; Contr.Mineral.Pertol., 54, p 43-59.

Bridgewater, D. and Collerson, K.D., 1977. On the origin of early Archean gneisses: A reply: Mineral.Petrol, 62, p 179-191.

Bridgewater, D., Keto, L., McGregor, V.R. and Myers, J.S., 1976. Archean Gneiss Complex of Greenland, in Geology of Greenland, Escher, A. and Watt, W.S. ed; Andelsbogtrykkeriet in Okense, Denmark.

Brooks, C., and Hart, S.R., 1974. On the significance of komatiite; Geology 2, p 107-110.

Brown, G.C. and Fyfe, W.S., 1972. The transition from metamorphism to melting: Status of the granulite and eclogite facies; 24th Int. Geol. Conf., Section 2, p 27-34.

Campbell, F.H.A., Elbers, F.J. and Gilbert, H.P., 1972. The stratigraphy of the Hayes River Group in Manitoba; Man. Mines Br., Geol. Paper 2/72.

Chayes, F., 1956. Petrographic Modal Analysis; Wiley and Sons, London.

Chou, L.L., 1978. Trace element studies in the Gestraverse region-Eastern Lac Seul area; in Smith and Williams (ed) Proceedings of the 1978 Archean Geochemistry Conference, Univ. of Toronto Press, p 151-159.

Clark, G., 1974. Isotopic age studies in progress; Cent. Precambrian Stud. 1974 Annual Report, p 56-57.

Condie, K.C. and Hunter, D.R., 1976. Trace element geochemistry of Archean granitic rocks from the Barberton region, South Africa; Earth Planet. Sci.Lett., 29, p 389-400.

Donaldson, J.A. and Jackson, G.D., 1965. Archean sedimentary rocks of the North Spirit Lake Area, Northwestern Ontario; Can.J. Earth Sci., 2, p 622-647.

Donaldson, J.A. and Ojakanagas, R.W., 1977. Orthoquartzite pebbles in an Archean conglomerate, North Spirit Lake, Northwestern Ontario; Can.J.Earth Sci., 14, p 1980-1990.

Douglas, G.V., 1926. Reconnaissance from Red Lake to Favourable Lake, District of Patricia; Ont.Dep.Mines, 35, pt 4, 1925, 28 p.

Ermanovics, I.F., 1970. Geology of Beren's River-Deer Lake Map Area, Manitoba and Ontario and a preliminary analysis of tectonic variation in the area; Geol.Surv.Can., Paper 70-29, 23 p.

Frey, F.A., Bryan, W.B. and Thompson, G., 1975. Atlantic Ocean Floor: Geochemistry and Petrology of basalts from legs 2 and 3 of The Deep Sea Drilling Project; Jour.Geophys.Res., 79, p 5507.

Frey, F.A., Haskin, M.A., Poetz, J.A. and Haskin, L.A., 1968. Rare earth abundance in some basic rocks; Jour.Geophys.Res., 73, p 6085.

Glikson, A.Y., 1971. Primitive Archean element distribution patterns; Chemical Evidence and Geotectonic Significance; Earth Planet, Sci. Let., 12, p 309-320.

Glikson, A.Y., 1972. Early Precambrian evidence of a primitive ocean crust and island nuclei of sodic granite; Geol.Soc.Am.Bull., 83, p 3323-3344.

Glikson, A.Y., 1977. On the origin of early Archean gneisses; Cont. Mineral. Petrol. 62, p 171-178.

Glikson, A.Y., 1978. On the basement of Canadian greenstone belts; Geoscience Canada, 5, p 3-12.

Glikson, A.Y. and Lambert, I.B., 1976. Vertical zonation and petrogenesis of the early Precambrian crust in western Australia; Tectonophysics, 30, p 55-89.

Goldschmidt, V.M., 1916. Kristiania Vidensk-selsk. Skr, Math.Naturv. Kl.2, p 75.

Goodwin, A.M., 1967. Volcanic studies in the Birch-Uchi Lakes area of Ontario; Ont.Dep.Mines, Misc. Paper 6.

Goodwin, A.M., Ambrose, J.W., Ayres, L.D., Clifford, P.M., Currie, K.L., Ermanovics, I.M., Fahrig, W.F., Gibb, R.A., Hall, D.H., Innes, M.J.S., Irvine, T.N., MacLaren, A.S., Norris, A.W. and Pettijohn, F.J., The Superior Province, in Variations in Tectonic Styles in Canada, Price, R.A. and Douglas, R.J.W., 1972, Geol.Ass.Can. Sp. Paper 11, 688 pp.

Gordanier, W., 1976. The depositional environment of formation K, Favourable Lake volcanic complex, northwestern Ontario; Cent. Precambrian Stud. 1976 Ann. Report, p 73-84.

Green, S.H., 1975. Genesis of Archean peridotitic magmas and constraints on Archean gradients and tectonics; Geology 3, p 15-18.

Hanson, G.N. and Goldich, S.S., 1971. Early Precambrian rocks in the Saganaga Lake-Northern Light Lake area, Minnesota-Ontario, Part 2, Petrogenesis; Geol.Soc.Am.Mem. 135, p 179-192.

Hargraves, R.B., 1976. Precambrian geologic history; Science, 193, p 363-371.

Harris, N.B.W. and Goodwin, A.M., 1976. Archean rocks from the eastern Lac Seul region of the English River gneiss belt, Northwestern Ontario, Part 1. Petrology, Chemistry and Metamorphism; Can.J. Earth Sci., 13, p 1201-1211.

Hart, S.R. and Brooks, C., 1970. Rb-Sr mantle evolution models; Carn. Inst. Wash Yearbook 68, p 426-429.

Henderson, J.B. and Easton, R.M., 1977. Archean supracrustal basement rock relationships in the Keskarrah Bay map area, Slave Structural province, District of Mackenzie: Geol.Surv.Can. Report of Activities, Part A., Paper 77-1A, p 217-221.

Holubec, J., 1972. Lithostratigraphy, structure and deep crustal relations of Archean rocks of the Canadian Shield, Rouyn-Noranda Area, Quebec, Krystalinikum 9, p 63-88.

Horwood, H.C., 1935. A pre-Keewatin (?) tonalite; Roy.Soc.Can.Trnas.29, p 139-147.

Hurst, M.E., 1930. Geology of the area between Favourable Lake and Sandy Lake, District of Kenora (Patricia Portion); Ont.Dept. Mines 38, pt 2, 1929, p 49-84.

Irvine, T.N. and Baragar, W.R.A., 1971. A guide to the chemical classification of the common volcanic rocks; Can.Jour.Earth Sci. 8, p 523-548.

Kay, R.W. and Gast, P.W., 1973. The rare earth content and origin of alkali-rich basalts; Jour.Geol. 81, p 653-682.

Kerrich, R. and Allison, I., 1978. Flow mechanisms in rocks; Geoscience Canada, 5, no 3, p 109-118.

Krogh, T.E. and Davis, G.L., 1971. Zircon U-Pb ages of Archean metavolcanic rocks in the Canadian Shield; Carn.Inst.Wash. Yearbook 70, p 241-242.

Krogh, T.E., Ermanovics, I.E. and Davis, G.L., 1974. Two episodes of metamorphism and deformation in the Archean rocks of the Canadian Shield; Carn.Inst.Wash. Yearbook 73, p 573-575.

Krogh, T.E., Harris, N.B.W. and Davis, G.L., 1976. Archean rocks from the eastern Lac Seul region of the English River gneiss belt, Northwestern Ontario, Part 2, Geochronology; Can.J.Earth Sci, 13, p 1212-1215.

Lambert, R.St.J. and Holland, J.G., 1976. Amitsoq gneiss geochemistry - preliminary observations, in Windley, B.G. (ed) The Early History of the Earth, John Wiley and Sons, London, p 191-201.

Leake, B.E., 1964. The chemical distinction between orth- and para-amphibolites; Jour.Petrol 5, p 238-254.

Low, A.P., 1887. Preliminary report on an exploration of country between Lake Winnipeg and Hudson Bay; Geol.Surv.Can., Ann.Rep. 2, Part F, 24 p.

Ludden, J.N., 1978. Are the REE's mobile during alteration processes? Geol.Soc.Am.Abstracts with Programs, 19, p 447.

Mason, D.R., 1978. Geochemistry of Archean tonalitic gneisses from the Superior Province (Wabigoon subprovince), Northwestern Ontario; Geol.Soc.Am.Abstracts with Programs, 10, p 451.

Moorbath, S., 1975. The geologic significance of early Precambrian rocks; Proc. Geol. Assoc. 86, p 259-279.

Nockolds, S.R., 1954. Average chemical composition of some igneous rocks; Geol.Soc.Am.Bull, 65, p 1007-1032.

Pettijohn, F.J., 1970. The Canadian Shield, a status report in Symposium on basins and geosynclines of the Canadian Shield; Geol.Surv. Can.Paper 70-40, p 239-255.

Phinney, W.C., Warner, J.L. and Morrison, D.A., 1978. Proto-Archean rocks: do they exist on other planets or Earth? Geol.Soc.Am. Abstracts with Programs, 10, p 472.

Pitcher, W.S. and Berger, A.R., 1972. The geology of Donegal; Wiley Interscience, New York.

Prest, V.K., 1939, Geology of the Keezhik, Miminiska Lakes area; Ont. Dep. Mines, Ann.Rep 49, Part 6.

Price, R.A. and Douglas, J.W. (eds), 1972. Variations in tectonic styles in Canada; Geol. Ass. Can. Sp. Paper 11, 688 pp.

Rousell, D.H., 1965. Geology of the Cross Lake area; Man. Mines Br. Pub. 62-4.

Schmidt, W., 1932. Tektonik and Verformungslehre; Barnstraeger, Berlin.

Shaw, D.M., 1968. A review of K-Rb fractionation trends by covariant analysis, Geochim. et Cosmochim. Acta, 32, p 573-602.

Sheraton, J.W., Skinner, A.C. and Tarney, J., 1973. The geochemistry of the Scourian gneisses of the Assynt district, in Park, R.G. and Tarney, J. (eds.) The early Precambrian of Scotland and related rocks of Greenland; Univ. Keele, p 13-30.

Smith, I.E.M., 1978. Volcanic and plutonic rocks of the Lake Shebandowan area in Smith and Williams (ed) Proceedings of the 1978 Archean Geochemistry Conference, University of Toronto Press, p 185-192.

Spry, A., 1969. Metamorphic textures; Pergamon Press, Oxford, 350 pp.

Streckeisen, A., 1976. To each plutonic rock its proper name; Earth Science Reviews, 12, p 1-33.

Thorpe, R.S., Potts, P.J. and Sarre, M.B., 1977. Rare earth evidence concerning the origina of granites of the Isle of Skye, Northwestern Scotland; Earth Planet.Sci. Lett. 36, p 111-120.

Vernon, R., 1965. Plagioclase twins in some mafic gneisses from Broken Hill, Australia, Mineral.Mag. 35, p 488.

Viljoen, M.J. and Viljoen, R.P., 1969. A reappraisal of granite-greenstone terrains of shield areas using the Barberton model in Upper Mantle Project; Geol. Soc. South Africa, Spec. Publ. 2, p 245-274.

West, G.F., 1978. A geotectonic model for the Superior province; Geol. Soc.Am.Abstracts with Programs, 10, p 514.

Wilson, H.D.B., Morrice, M.G. and Ziehlke, D.V., 1974. Archean continents; Geoscience Canada, 1, p 12-20.

Windley, B.F. (ed), 1976. The early history of the earth; John Wiley and Sons, London, 609 pp.

Young, G.M., 1978. Some aspects of the evolution of the Archean crust; Geoscience Canada, 5 p, 140-149.

APPENDIX 1 MODAL ANALYSES

Sample #	Micro-cline	Sphene	Fe-Ti Oxide, Pyrite	Horn-blende	Epi-dote(1)	Biotite	Quartz	Plagio-clase	Musco-vite	An%
Trondhjemite										
3	tr	0.8	0.4	4.0	3.5	9.7	27.2	54.0	0	24-43
7	tr	0.7	0.3	5.7	2.7	6.3	28.8	56.5	0	22-31
14	tr	tr	0.3	5.4	2.9	8.1	26.8	56.5	0	26-40
16	tr	0.6	0.6	tr	3.9	11.2	32.0	51.3	0	23-27
19	tr	tr	tr	0	1.5	12.5	26.7	58.5	tr	22-28
22	tr	0.6	tr	6.8	2.9	9.0	25.6	54.5	0	24-27
27	tr	0.7	tr	5.0	3.9	12.0	23.3	54.9	0	24-26
34	0.8	0.4	0.3	0.7	4.0	9.4	26.2	58.2	0	ND(2)
38	tr	1.0	tr	4.2	3.5	9.8	24.6	53.8	0	23-27
52	0	0.8	0.4	1.6	3.2	15.6	27.2	51.2	0	ND
58	tr	tr	0.6	0	1.4	12.0	27.4	58.4	tr	22-26
60	tr	1.4	tr	4.6	2.4	14.9	21.8	54.8	tr	25
65	tr	0.6	1.4	3.6	4.2	12.8	24.0	53.5	tr	23-25
67	0	0.7	0.9	3.3	4.1	10.4	25.1	55.6	tr	24-27
68	tr	0.2	tr	tr	5.8	20.5	23.5	50.0	0	24-31
76	0	0.6	0.2	4.4	1.0	4.6	31.4	57.8	0	ND
79	tr	0.2	0.9	3.2	2.3	11.4	24.7	57.3	0	25-27
82	0	0.3	0.5	0.6	2.3	12.8	28.0	55.2	0	23-27
86	tr	0.6	0.3	5.6	2.3	11.4	22.5	57.3	0	23-27
97	tr	0.1	0.3	0	1.8	8.2	35.0	53.1	1.5	24

Sample #	Micro-cline	Sphene	Fe-Ti Oxide, Pyrite	Horn-blende	Epi-dote(1)	Biotite	Quartz	Plagio-clase	Musco-vite	An%
Trondhjemite (cont'd.)										
98	0	0.2	0.4	2.0	2.2	9.8	33.4	52.0	0	ND
100	tr	0.2	tr	0	4.4	14.4	32.8	47.6	0.2	ND
101	0.6	tr	0.2	0	3.2	14.4	24.3	55.5	1.6	ND
102	0.4	tr	0.1	0	1.8	9.0	30.9	56.4	1.8	21-28
103	0.2	tr	tr	0	2.2	7.6	27.8	60.6	1.6	ND
111	tr	0.5	0.3	0.4	1.6	15.5	25.6	56.1	0	22-31
114	tr	tr	0.6	0	0.5	7.0	29.5	62.4	0	23-30
120	tr	tr	tr	0	2.4	8.8	26.0	62.5	1.0	22-25
121	0.2	0.3	tr	0	1.3	8.0	26.7	63.1	1.2	22-27
125	tr	0.8	0.7	tr	7.8	12.5	21.5	56.7	0	23-27
152	0.3	0.3	tr	0	1.3	17.2	33.4	47.0	0.5	24-31
154	tr	0.2	0.2	1.5	1.9	7.7	28.0	60.5	0	25
157	0	0.8	0.6	4.9	1.3	6.4	27.6	57.9	0	24-30
160	0	0.4	0.2	3.9	1.7	12.4	27.4	54.0	0	26
163	tr	0.6	tr	5.4	2.2	6.9	26.4	58.4	0	26
169	tr	0.4	0.4	9.2	1.0	8.2	19.6	61.2	0	ND
170	0	tr	0.5	0.2	4.0	8.1	25.3	61.8	0	25
171a	0.3	0.1	0.2	0.6	0.9	6.9	31.5	59.6	tr	26
175	0	0.3	0.2	0.3	3.1	14.8	28.9	52.4	0	26
176	tr	0.9	0.5	2.8	5.6	11.2	22.5	56.5	0	26

Sample #	Micro-cline	Sphene	Fe-Ti Oxide, Pyrite	Horn-blende	Epi-dote(1)	Biotite	Quartz	Plagio-clase	Musco-vite	An%
Trondhjemite (cont'd.)										
187	tr	0.2	0.4	0	1.1	15.9	28.2	54.2	0	22-30
189	tr	0.8	0.4	5.2	2.6	10.3	23.8	56.7	0.2	24
192	0	0.6	0.5	5.8	3.8	8.8	26.1	54.0	0	22-36
195	0	0.4	0.3	2.7	2.2	13.7	25.1	55.6	0	25
202	tr	0.4	0.4	0	2.1	11.5	29.9	55.4	0.4	24-29
225	tr	0.8	0.6	0	3.0	15.2	23.6	56.8	tr	27
233	tr	0.4	0.5	0	1.2	8.0	36.1	52.3	1.3	26
244	tr	0.4	0.5	0.1	2.2	7.9	32.9	55.1	0.9	23-32
251	0.8	tr	0.2	0	1.2	6.8	30.8	59.6	0.6	ND
253	tr	0.2	tr	0.4	1.6	12.2	32.4	52.6	0.6	ND
264	0.3	0.5	0.4	tr	3.9	15.9	27.8	51.2	tr	23-28
274	0.3	0.9	0.2	6.2	3.1	10.0	20.3	59.1	0	26
276	0	0.2	0.4	tr	0.7	9.0	28.5	61.2	0	24
279	0	0.3	0.8	2.8	1.2	6.1	27.6	61.1	0	25
288	0	0.1	0.9	0.7	1.7	12.8	28.7	55.4	0	26
289	0	0.5	0.8	0.3	8.5	16.8	24.9	48.4	0	26
293	0.2	0.3	0.2	5.6	2.7	4.9	34.7	51.4	0	25
294	0	0.1	0.5	0	0.8	4.5	24.3	62.4	0.4	22-26
295	0	0.8	1.0	0.1	2.8	14.3	24.4	56.6	0	24-28
302	0	0.3	0.2	3.5	4.1	15.1	24.4	51.7	0	22-28
305	0	0.8	0.4	7.7	2.1	11.0	18.7	59.3	0	22-28

Sample #	Micro-cline	Sphene	Fe-Ti Oxide, Pyrite	Horn-blende	Epi-dote(1)	Biotite	Quartz	Plagio-clase	Musco-vite	An%
Melanocratic Layers										
22	tr	tr	tr	0	tr	5.0	37.1	57.6	0	24-27
86	0	0	0.6	0	0	1.0	23.4	76.0	0	23-27
160	tr	0	0	0	0	1.2	23.3	75.5	0	26
170	0	tr	0	0	0.8	3.2	32.8	63.2	0	25
244	tr	0	0	0	0	1.0	34.7	64.3	0	23-32
264	tr	0	0	0	tr	tr	40.1	58.3	0	23-28
Leucocratic Layers										
91(3)	5.0	1.8	0.8	tr	8.2	20.4	23.4	40.4	0	25-38
137	0.2	2.0	0.8	5.2	3.2	13.6	24.8	50.2	0	24-47
227	tr	0.2	0.2	5.8	0.9	13.4	20.6	58.8	0	24-30
242	0	0.8	0.9	5.9	2.9	13.6	22.3	53.6	0	23-30
256	tr	0.9	0.2	6.8	3.0	9.4	21.3	58.4	tr	25-35

(1) Most samples contain trace amounts of apatite, chlorite, zircon and allanite. Allanite content is included with epidote.

(2) ND = not determined.

(3) Samples of leucocratic layers are narrow and occur in three thin sections of the main trondhjemite, resulting in duplicated sample numbers.

APPENDIX 2 SUMMARY OF ANALYTICAL METHODS FOR CHEMICAL ANALYSES
(as done at the University of Manitoba, Department of Earth Sciences)

Si, Al, Fe(total), Mg(high), Ca, K, Ti, Mn
X-ray fluorescence spectrometry - weighed sample plus $Li_2B_4O_7$ and La_2O_3 is heated in a graphite crucible at about 1100°C for 0.5 hr; resulting glass bead is combined with H_3BO_3 (to a total weight of 2.1000 grams) and ground to -200 mesh and then compressed to 50,000 p.s.i.; elements are then simultaneously analyzed in multi-channel ARL X-ray fluorescence spectrometer.

Na_2O, MgO(low) and trace metals
Atomic absorption spectrophotometry - rock is dissolved with HF, H_2SO_4, HNO_3 in platinum crucibles; the Perkin Elmer 303 atomic absorption spectrophotometer is used for determinations.

P_2O_5 Colorimetry - solution as for Na_2O; the absorption at 430 m of molydivanadophosphoric acid complex is determined with Unicam sp 500 spectrophotometer.

FeO Rock is decomposed with HF and 1:4 H_2SO_4 solution is titrated with $K_2Cr_2O_7$ using sodium diphenylamine sulfonate as indicator.

H_2O^- Determined by heating the sample to constant weight at 110°C.

H_2O total Determined by heating sample in a stream of dry oxygen in an induction furnace (temp 1100°C); H_2O is collected on Anhydrone and weighed.

H_2O^+ H_2O total minus H_2O^-

CO_2 (low sulfur samples)
Determined simultaneously with H_2O total; CO_2 is collected on Ascarite, with small amounts of SO_2 removed on MnO_2(act).

APPENDIX 3 NIGGLI al-alk and c VALUES

Sample #	16	19	38	82	86	154	189	192	244	289	302	av.
$Fe + 2Fe_2 + Mg + Mn$	114	80	119	85	101	85	111	92	62	99	93	94
Al	153	153	161	153	159	158	163	159	141	160	155	158
Na + K	97	92	94	91	95	94	94	94	85	94	92	94
Ca	70	59	76	66	71	61	79	71	52	70	74	68
Total	434	384	450	395	426	398	447	416	340	423	414	412
fm	26	21	26	22	24	21	25	22	18	23	22	22
al	36	40	36	38	37	40	36	38	42	38	38	38
alk	22	24	21	23	22	24	21	23	25	22	22	23
c	16	15	17	17	17	15	18	17	15	17	18	17
al-alk	14	16	15	15	15	16	15	15	17	16	16	15

where $Ca = \frac{\text{wt\%60}}{\text{approx.ml.wt.}} \times 1000$ (after Barth 1952, p 76)

$Al = \frac{\text{wt\%}Al_2O_3}{\text{approx.mol.wt.}} \times 1000$

$Fe + Fe_2 + Mg + Mn$ from Appendix 3(a)

$Na + K = \frac{\text{wt\%Na20}}{\text{approx.mol.wt.}} \times 1000 + \frac{\text{wt\%}K_2O}{\text{approx.mol.wt.}} \times 1000$

and al = Al_2O_3 (including Cr_2O_3 and rare earths)

fm = FeO + MgO + (MnO + CoO + NiO)

c = CaO (including BaO + SrO)

alk = $K_2O + Na_2O$ (including rare alkali earths)

100 = al + fm + c + alk

APPENDIX 4 RESULTS OF REANALYSIS OF FOUR FeO VALUES TO CHECK Fe_2O_3/FeO RATIO

SAMPLE	FeO - Original Analysis	FeO - Reanalysis
38	2.88	2.94
102	1.04	1.06
120	0.92	1.02
244	1.36	1.40

The second analyses for FeO are all higher than initial values and would result in slightly lower Fe_2O_3 values.

LEGEND

SYMBOLS

area of bedrock outcrop

gneissosity: inclined, vertical

foliation: inclined vertical

lineation, with plunge

geological boundary: observed, interpreted

fault, assumed

muskeg or swamp

winter road

sample location

- age dated sample
- chemical analysis
- rare earth analysis

NORTH TROUT LAKE BATHOLITH PHASES

10 Late biotite-hornblende trondhjemite (phase N-5') and porphyritic biotite granodiorite (phase N-14')

9 Late leucocratic phases forming an intrusive complex with the basement: biotite quartz monzonite (phases N-17, 19'), porphyritic biotite granodiorite (phase N-14'), minor pegmatite and aplite

8 leucocratic phases containing basement phases as rotated xenoliths, > 90 % leucocratic phases

7 > 75% leucocratic phases

6 50-75% leucocratic phases

5 25-50% leucocratic phases

4 Late mesocratic phases: biotite hornblende trondhjemite (phase N-5') and biotite hornblende diorite (phase N-4')

BASEMENT PHASES

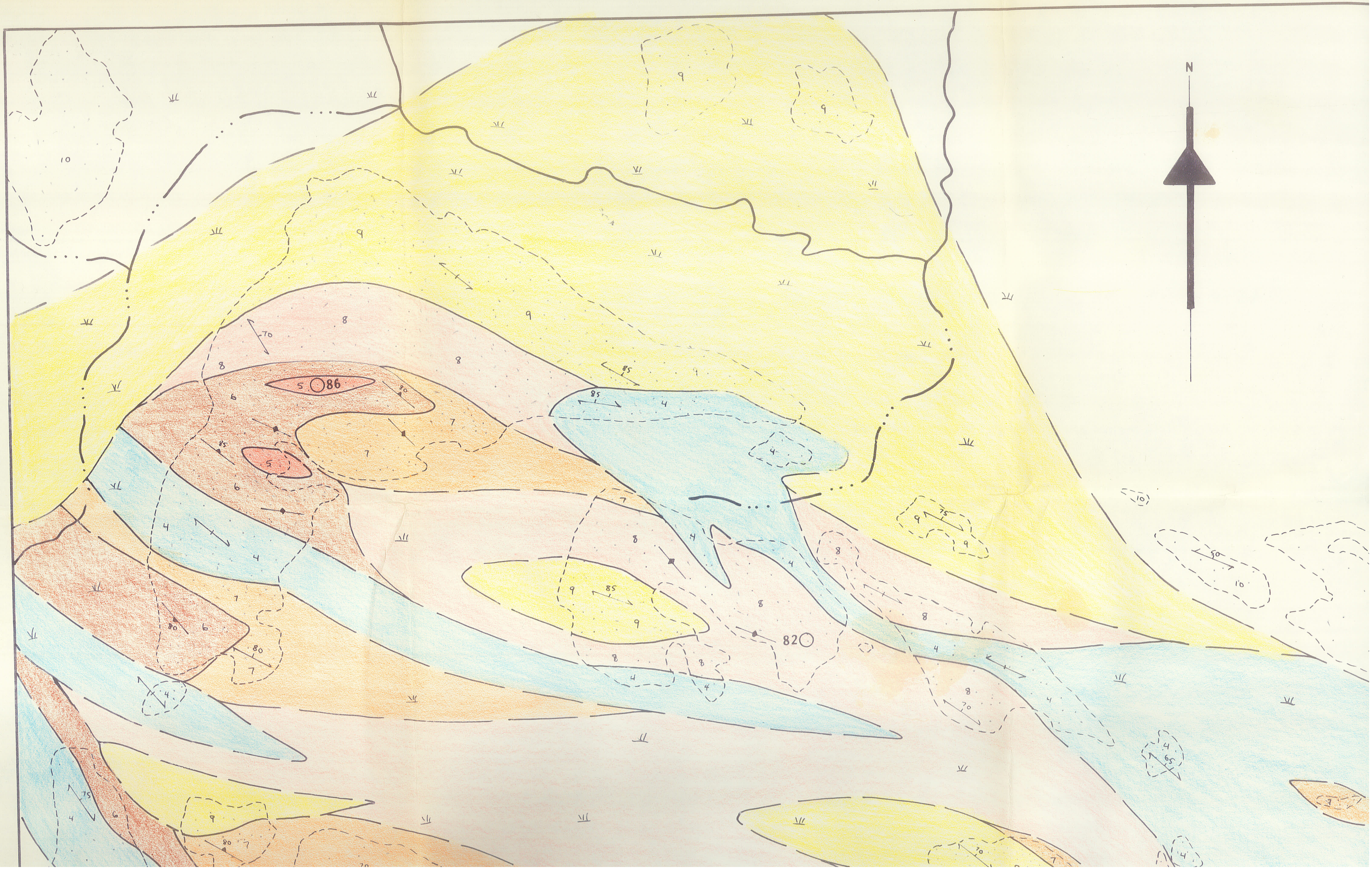
3 equigranular to porphyritic hornblende biotite trondhjemite (phase N-3'): < 25% late leucocratic phases

early melanocratic phases

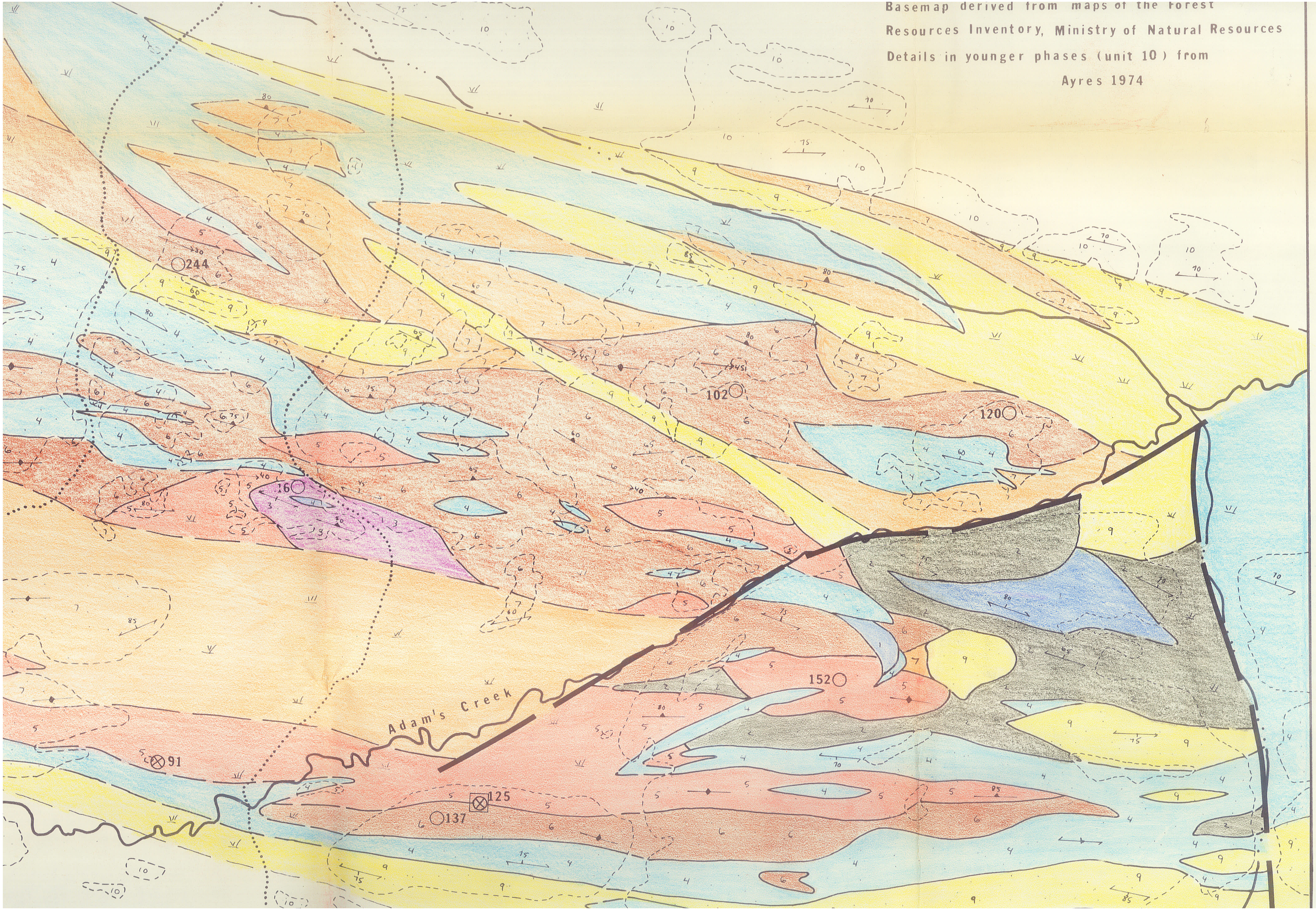
2 gabbro-diorite assemblage

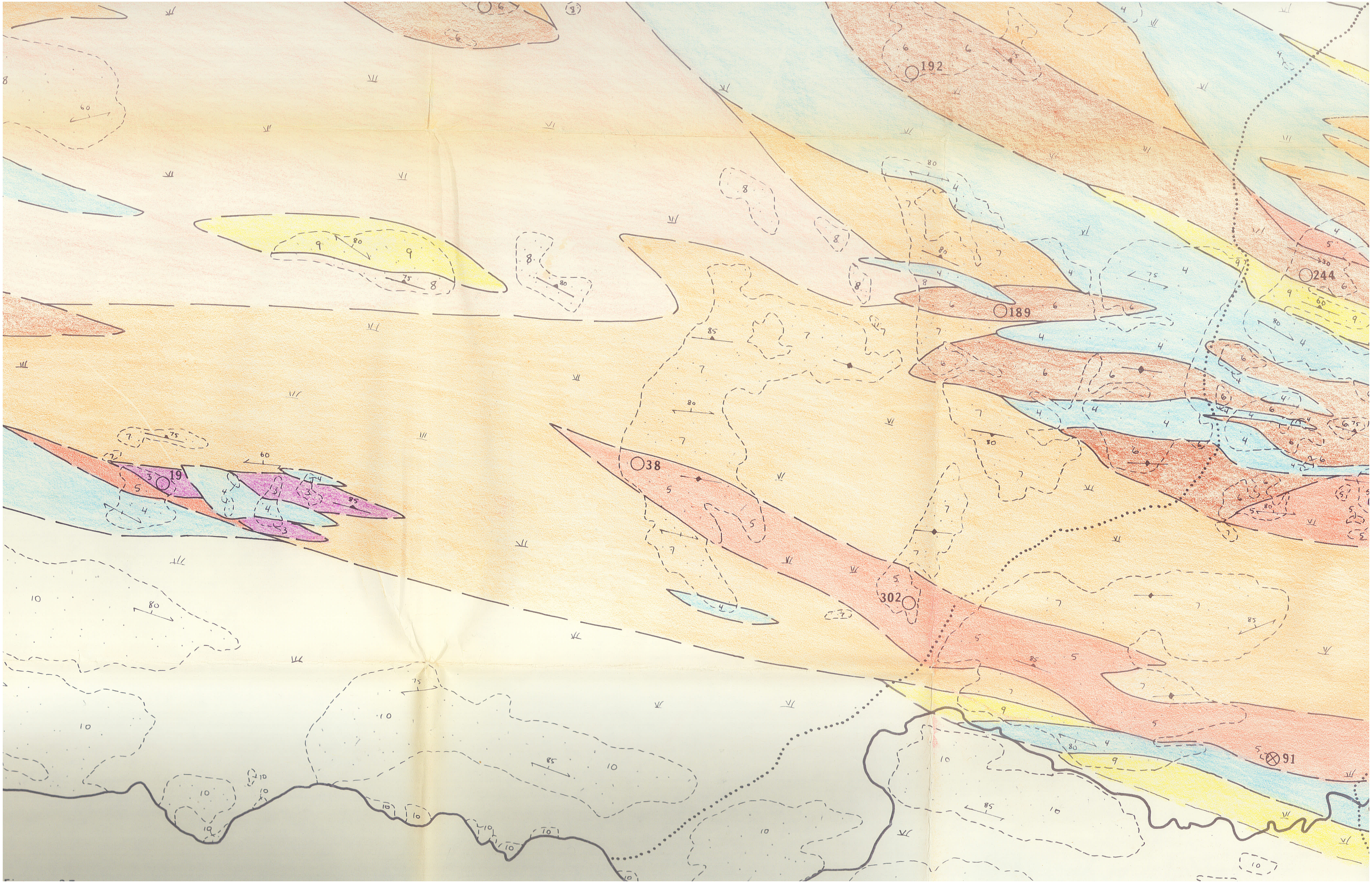
1 pyroxenite

(' after Ayres 1974)


N
5
86
82


N
82
9
8
7
4
10

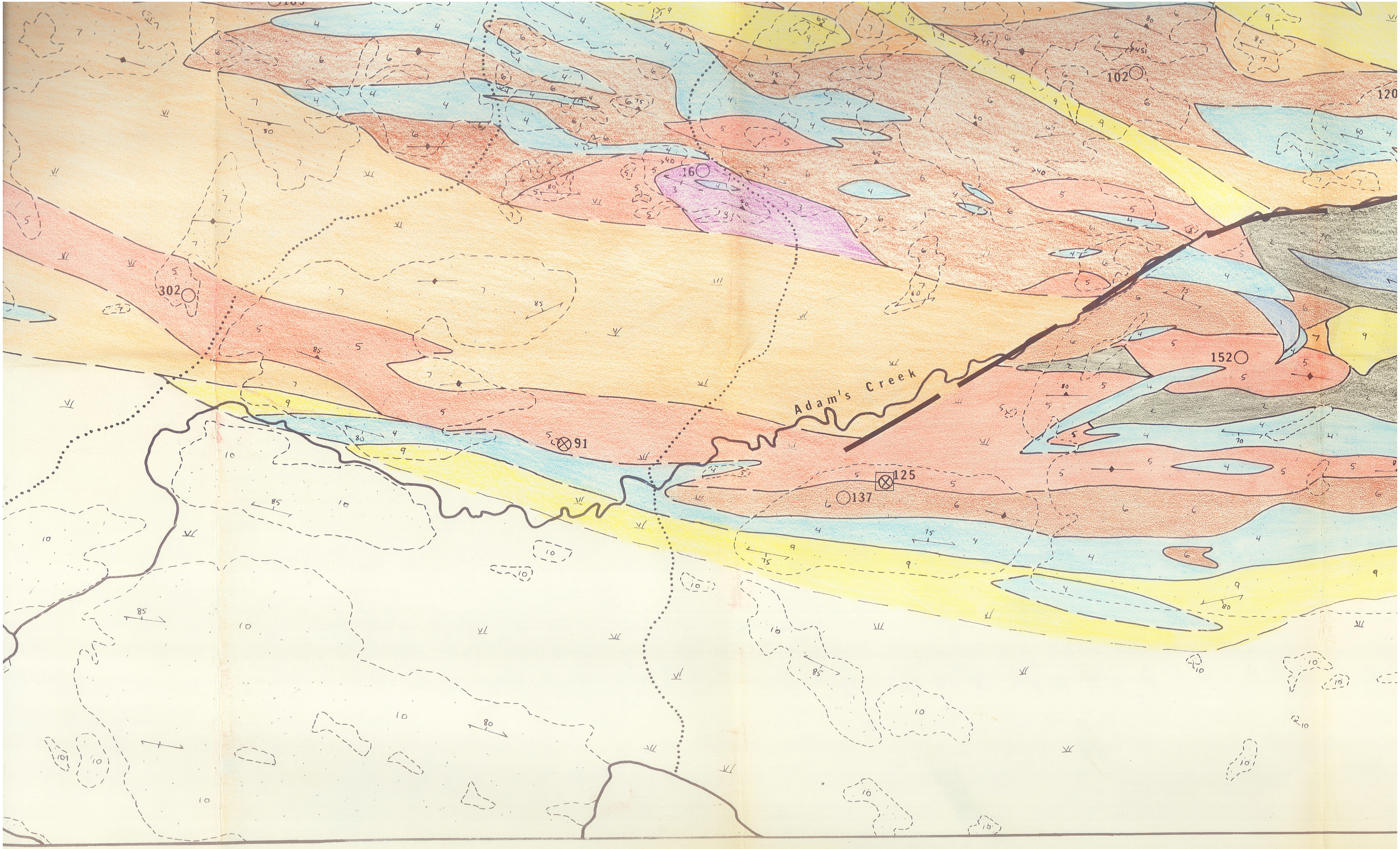

Basemap derived from maps of the Forest
Resources Inventory, Ministry of Natural Resources
Details in younger phases (unit 10) from
Ayres 1974
Adam's Creek
244
102
120
16
152
91
125
137

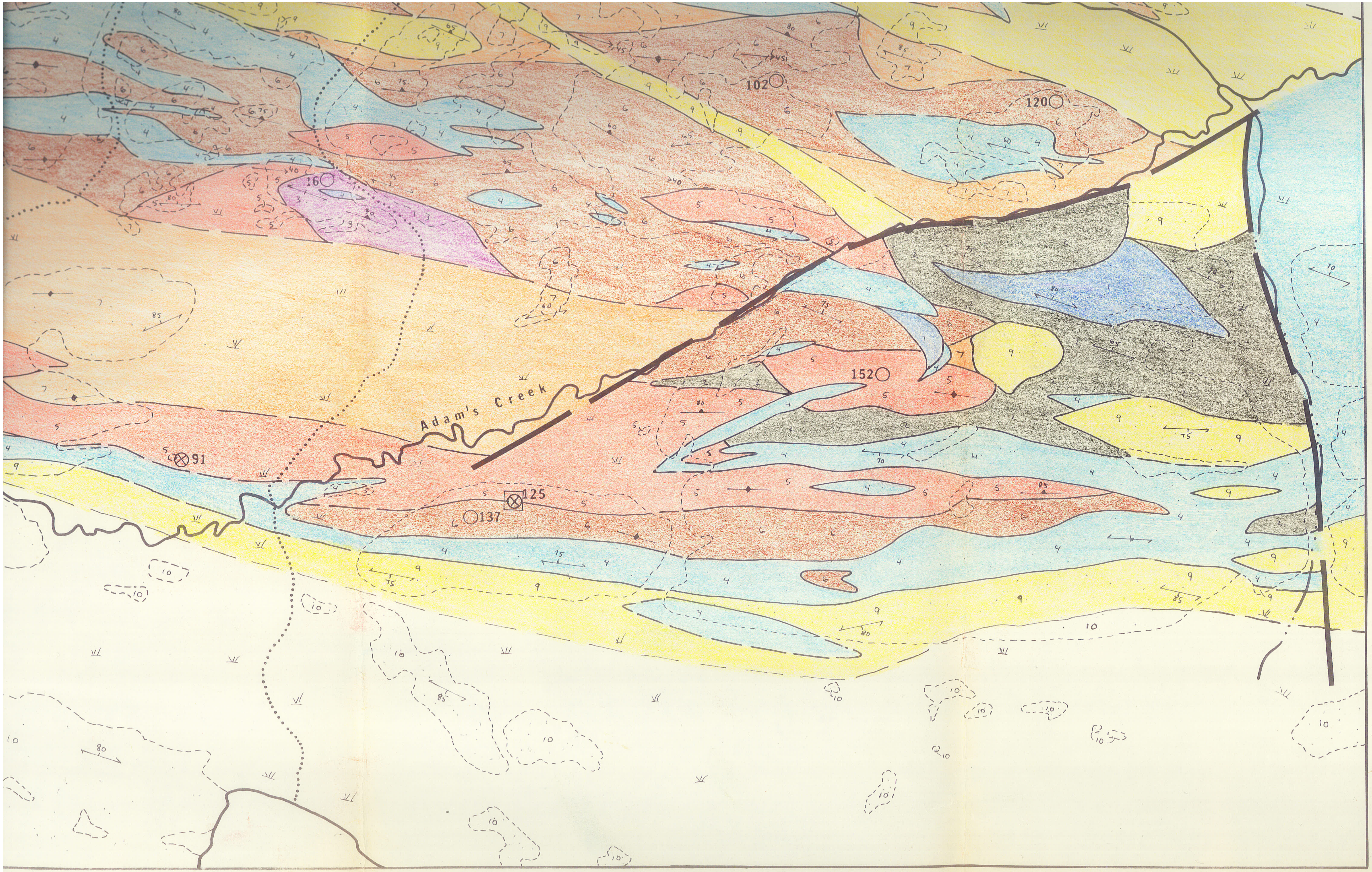

192
244
189
38
19
302
91


NORTH TROUT LAKE
274
289
154
19
38

Figure 37

GEOLOGY OF THE NORTH TROUT LAKE
ANCIENT CRUSTAL REMNANT


Adam's Creek
302
91
125
137
152
16
102
120


Adam's Creek
102
120
16
152
91
125
137