GRENVILLE PROVINCE-SOUTHERN PROVINCE BOUNDARY, SUDBURY DISTRICT

STRUCTURAL AND PETROLOGIC RELATIONS ACROSS THE GRENVILLE PROVINCE-SOUTHERN PROVINCE BOUNDARY, SUDBURY DISTRICT, ONTARIO

By

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A Thesis

Submitted to the Faculty of Graduate Studies

in Partial Fulfilment of the Requirements

for the Degree

Doctor of Philosophy

McMaster University

October 1967

DOCTOR OF PHILOSOPHY (1967) (Geology)

McMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Structural and Petrologic Relations across the Grenville Province-Southern Province Boundary, Sudbury District, Ontario

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NUMBER OF PAGES: xii, 119

SCOPE AND CONTENTS:

A region across the Grenville province-Southern province boundary was mapped, described and analysed. Structural fabric analyses and relationships between deformation and metamorphism were emphasized; especially in regard to Southern province rocks that were strained and recrystallized during the final Grenville orogenic episode.

Critical review of the literature on the region showed that metasedimentary rocks in the research area belong to the Huronian System, and that a granitic body of batholithic size was emplaced more than 600 million years before it was recrystallized last.

Many tectonite fabrics were shown to be products of homogeneous strain, and, wherever possible, principal strain axes were determined in relation to fabric symmetry axes. Comparison of internal fabrics of salt domes with structural fabrics in the research area indicated that the principal flowage occurred in the direction of maximum elongation. Finally, the Grenville-Southern province boundary was defined genetically as a dynamo-thermal metamorphic gradient.

ABSTRACT

South of Sudbury, Huronian sedimentary rocks, Sudbury-type gabbro and Killarnean granite of the Southern structural province were deformed and metamorphosed during the orogenic episode in which Grenville province gneisses attained their present structural and metamorphic state. A granitic batholith parallels the boundary between the provinces, but pre-dates the final orogeny in the Grenville province by as much as 600 million years. A zone of blastomylonitization and green-schist grade of metamorphism extends 2 to 5 miles west of completely recrystallized almandine-amphibolite grade gneisses of the Grenville province.

Mineral lineation and foliation are typical structures formed during the final deformation, but mesoscopic folds are absent from many penetratively deformed rocks. The strain recorded by unfolded, lineated tectonites is considered to be homogeneous, and many structural fabrics are shown to have orthorhombic symmetry. Mineral lineations, passive fold axes and quartz c_v girdle axes parallel the direction of maximum strain; mineral foliations, and passive surfaces are shown to lie in a plane containing the maximum strain direction.

ACKNOWLEDGMENTS

Professor P. M. Clifford suggested the topic and supervised the research. Excellent field assistance was provided by Mr. Robert Kier in the summer of 1964, and by Mr. Peter Waldo in the summer of 1965. My wife, Timea, who has been a constant source of encouragement throughout my graduate studies, typed the manuscript and colored the maps.

Financial assistance was provided by the Ontario Research Foundation (1963-66), the National Research Council (1963-67) and the Geological Survey of Canada (1963-67).

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CHAPTER I

INTRODUCTION

The Canadian Precambrian Shield contains rocks which were formed over a period of more than two billion years. The major internal geological features of the Canadian Shield are large structural provinces each characterized by a common period of most recent orogenesis. The Grenville[1] province of Ontario, Quebec and Labrador is the youngest structural province of the Canadian Shield. The northwest border of this province, the Grenville "front", extends for more than twelve hundred miles from the coast of Labrador southwest to Georgian Bay on Lake Huron where it disappears under the cover of Paleozoic rocks. The Grenville province apparently truncates four older structural provinces along its northwest boundary.

The research area is underlain by rocks which previously have been assigned to both the Grenville and Southern structural provinces of the Canadian Shield. The Southern province is composed largely of metasedimentary rocks which were deformed during the Hudsonian orogeny to form the Penokean fold belt. The Grenville province in this region is composed of gneissic rocks, largely of unknown primary age and origin, that were metamorphosed during the Grenville orogeny. The nature of the Grenville province-Southern province boundary, the origins and primary ages of the Grenville province gneisses have been subject to speculation and controversy since the region was first

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[1]The names of structural provinces, orogenies and the time-classification of rocks of the Canadian Precambrian Shield used here are those adopted by Stockwell (1964) and used on the Tectonic Map of the Canadian Shield (Stockwell, 1965) unless stated in a historical context.

investigated in the nineteenth century.

The particular area was chosen for investigation because of its location across the Grenville front in an easily accessible region where previous workers had described the "front" as a transitional boundary uncomplicated by younger tectonic events. The major objectives of the field work were to define the boundary between the two structural provinces and to map sufficiently large areas within both the Grenville and Southern provinces to provide an understanding of the regional geology.

I. LOCATION, ACCESS AND MAP COVERAGE

The mapped area is ten miles south of Sudbury, Ontario (Fig. 1), and includes the townships of Eden, Tilton and the southern part of Broder--encompassing an area of approximately seventy-five square miles. The northern and western parts of the area are accessible by gravel roads which branch from the terminus of Route 543 (See township index map on Fig. 2, pocket); the southern part of the area is accessible by foot and canoe.

Topographic maps covering the area are the Copper Cliff (41 I/6 East) and Coniston (41 I/7 West) 1:50,000 scale sheets of the National Topographic System. Areal photographs (approximate scale 4 inches to 1 mile) were used to map contacts between rock units, and to locate observation stations and sample localities.

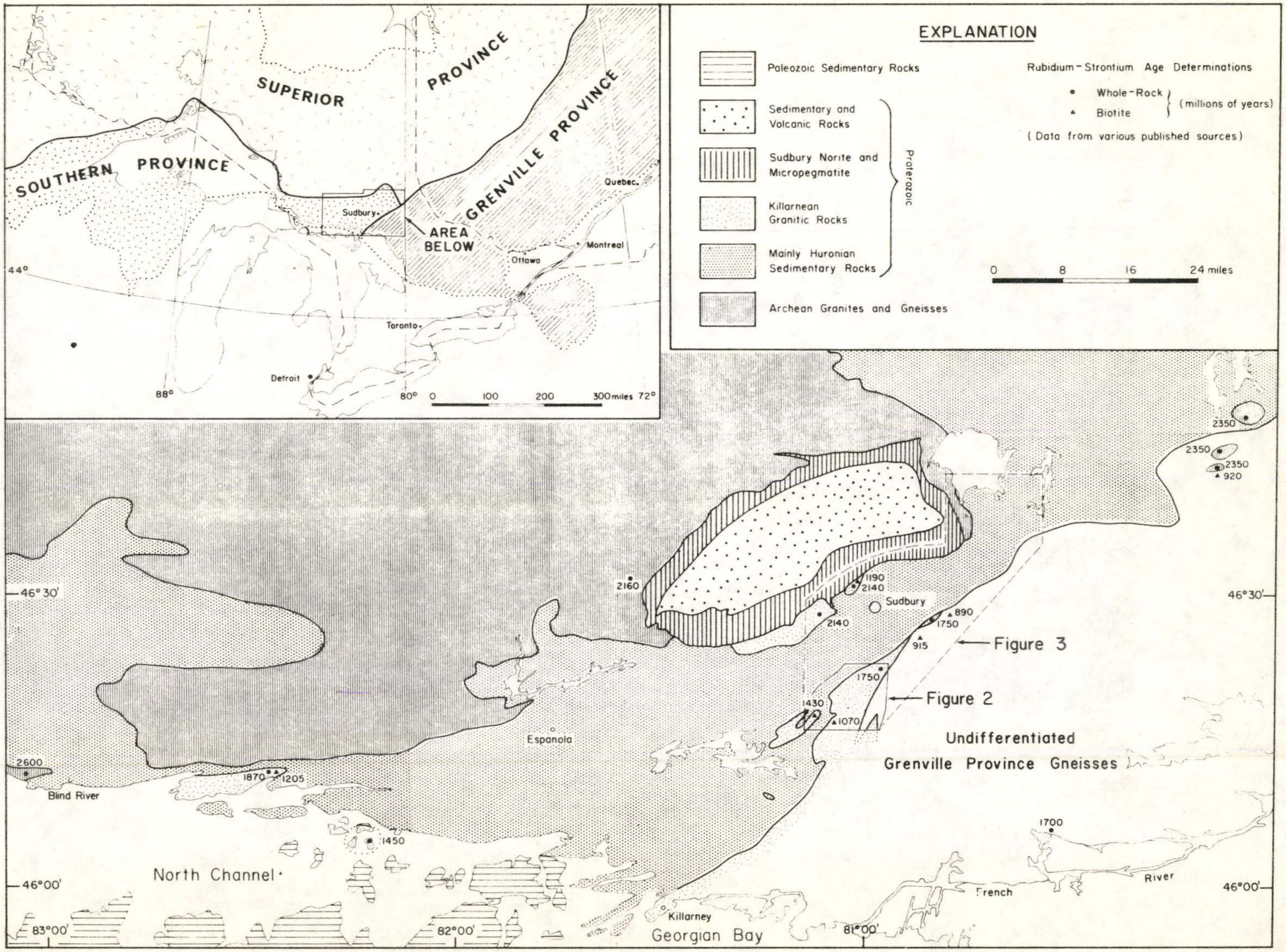
EXPLANATION
Paleozoic Sedimentary Rocks
Sedimentary and Volcanic Rocks
Sudbury Norite and Micropegmatite
Killarnean Granitic Rocks
Mainly Huronian Sedimentary Rocks
Proterozoic
Archean Granites and Gneisses
Rubidium-Strontium Age Determinations
Whole-Rock
Biotite
(millions of years)
(Data from various published sources)
0 8 16 24 miles
SUPERIOR PROVINCE
SOUTHERN PROVINCE
GRENVILLE PROVINCE
AREA BELOW
Sudbury
Quebec
Montreal
Ottawa
Toronto
Detroit
44°
88°
80°
72°
0 100 200 300 miles
Figure 3
Figure 2
Undifferentiated
Grenville Province Gneisses
Espanola
Blind River
North Channel
Killarney
Georgian Bay
French
River
46°30'
46°00'
83°00'
82°00'
81°00'
2350
2350
2350
920
2160
1190
2140
2140
890
1750
915
1750
1430
1070
2600
1870
1205
1450
1700

Figure 1

II. CRITICAL REVIEW OF PREVIOUS WORK

Because of its great mineral wealth, the Sudbury district has been investigated by more geologists than any other area of comparable size in the Canadian Shield. The belt of sedimentary and volcanic rocks extending along the north shore of Lake Huron between Bruce Mines and Sudbury has been extensively mapped--especially since the discovery of uranium-bearing conglomerate at the base of the Mississagi Formation. The mineral-poor region southeast of Sudbury in the Grenville province, on the other hand, has received scant attention, and remains practically unmapped except for a narrow strip adjacent to the Southern province.

The map area has been investigated by geologists working for both the federal (Bell, 1891; Collins, 1925 and 1938), and provincial (Coleman, 1914; Baker, 1917; Phemister, 1960, 1961; Grant, Pearson, Phemister, and Thomson, 1962) surveys. With the exception of Baker's work, which lies wholly within the research area, all previous workers were concerned mainly with mapping much larger areas lying west and north of the research area. H. R. Spaven (1966) conducted a detailed geological survey of a small portion of the map area.

Within the past ten years, radiometric age determinations have been obtained from rocks and minerals within the research area (Fairbairn, Hurley, and Pinson, 1960; Tilton, Wetherill, Davis, and Bass 1960; Krogh, 1966) and from surrounding areas (See especially Hart, 1961; Fairbairn, Hurley, and Pinson, 1965; Van Schmus, 1965.). The determination of absolute times of intrusion and metamorphism in the Sudbury region has contributed greatly to the understanding of the regional geologic history.

Southern Province

No generally acceptable correlation of the sedimentary rocks in the Southern province south of Sudbury has yet been published. Recent field work by the Ontario Department of Mines in Waters Township (Card, 1964) and in Broder Township (Grant, et al., 1962) suggests that, due to structural complications, even a local stratigraphic sequence cannot be worked out.

Until 1953, the time-stratigraphic correlation of rocks in the Sudbury region that had evolved largely through the work of the Geological Survey of Canada under the direction of W. H. Collins was accepted as correct. On the Copper Cliff Sheet, Collins (1938) placed the sedimentary rocks in the thesis area within the Huronian succession, and classified the granitic rocks and gneisses as Killarnean.

In 1953, J. E. Thomson published a summary of his field work in Baldwin Township (Thomson, 1952a) southwest of Sudbury. Thomson (1953) found a conformable succession upwards from what Collins (1938) had considered to be Keewatin lavas, through the Pre-Huronian Sudbury Series and into the Huronian Bruce Series (Ramsay Lake Conglomerate and Mississagi Formation). The absence of the great Pre-Huronian unconformity at the base of the Ramsay Lake Conglomerate in this region, according to Thomson (1957, p. 52), made it impossible to establish a satisfactory division of the formations in the area south and west of sudbury into an Archean (i.e., Pre-Huronian) and Proterozoic sequence. Subsequently, working along the south shore of Lake Wanapitei, Thomson (1961a) discovered what he considered to be the "great unconformity" at the base of the Huronian. According to Thomson, all rocks south of the unconformity are Pre-Huronian and belong to the Sudbury Group. These Pre-Huronian rocks had been mapped previously as Huronian by Collins and Cooke (1946a). Thus Thomson cast serious doubt on both the accuracy of mapping and the stratigraphic succession established in the Sudbury region by the Geological Survey of Canada.

Rather than attempt to reconcile his findings with the results of previous workers, Thomson (1953 and 1957)... "advocated a moratorium on current time-stratigraphic terminology throughout the Sudbury-Espanola area [which includes the research area] until detailed restudies have been made over a considerable part of the country." In 1961, Thomson (1961b) published a compilation map of the area south and west of Sudbury. On this map, Thomson arbitrarily placed all of the sedimentary rocks south of the Murray Fault into either the Sudbury Group or left them as unclassified sedimentary rocks. In a footnote, Thomson stated that "Field relationships favour the interpretation that the unclassified sedimentary rocks are largely Pre-Huronian in age."

Thomson (in Grant <u>et al</u>., 1962) placed the sedimentary rocks in Broder, Neelon and Dill townships in the Sudbury Group, and Card (1964) called the sedimentary rocks overlying a basal metapelite in Waters Township the Ramsay Lake Conglomerate and Wanapitei Quartzite. The stratigraphic succession of the formations within Thomson's Sudbury Group has never been defined. Thomson (1961a, p. 5), in his table of formations, listed the Sudbury Group as: Conglomerate, quartzite, graywacke and biotite paragneiss, limestone. In his description of the Sudbury Group, Thomson (1961a pp. 7-9) implied that all of the rock types are interbedded with no regular stratigraphic succession. Phemister (1961, p. 17), said that "in Broder, Dill, Neelon and Drydon townships there is no sharp field distinction between Wanapitei Quartzite, Ramsay Lake Conglomerate and McKim Graywacke, because there are many bands of quartzite within McKim Graywacke, and because argillaceous phases are common in Wanapitei Quartzite beds."

Recent work southwest of Sudbury, where the rocks are less intensely deformed, shows a regular succession of lithologic units within the sedimentary sequence. Frarey (1966), working southwest of the thesis area in the Lake Panache region, stated that "the normal Huronian sedimentary succession can be followed in all of the ground covered, and that with the exception of minor revision and some subdivision the formations previously mapped by Quirke and Collins are essentially correct." Similarly, Young and Church (1966), in a review of the stratigraphy in the area between Bruce Mines and Cobalt, concluded that the rocks throughout the area are Huronian in age--including the rocks formerly known as the Sudbury Series, and the underlying volcanic rocks.

A compilation map of the area east and south of Sudbury (Fig. 3, pocket) was made in order to determine if the so-called Pre-Huronian Sudbury Group can be divided into regional stratigraphic units. The interpretation of the regional stratigraphy will be discussed in a later section (p.14), but it should be pointed out here that a consistent stratigraphic sequence within the sedimentary rocks can be traced around the east end of the Sudbury Basin, and, as nearly as can be determined without field checking, the succession is similar to that defined by Collins (1938) and Young and Church (1966). It is concluded, therefore, that the Southern province sedimentary rocks in the research area belong to the Huronian System.

Collins (1916) defined two periods of granitic intrusion in the area north of Lake Huron: an older series (Archean) on which the basal Huronian was deposited unconformably, and a younger series intrusive into the Huronian sedimentary rocks. This younger granite series Collins correlated with the granite in the Grenville province at Killarney on Lake Huron (Fig. 1), and proposed the time-stratigraphic term "Killarney Granite" to include <u>all</u> of the granitic rocks intrusive into the Huronian succession. To Collins, the term "Killarnean" had an orogenic sense and apparently was equivalent to the Hudsonian orogeny. Radiometric dates of intrusion and metamorphism in the Southern Province indicate that the Hudsonian orogeny occurred about 1600 million years ago (Stockwell, 1964).

Grenville Province

The Grenville front was first noted by Alexander Murray in 1856 in the vicinity of Ashagami Lake northeast of Sudbury. Bell (1891) considered the gneisses of both the Grenville and Superior provinces to be correlative and placed them in Logan's "Laurentian System" (Archean). The less deformed sediments of the Southern province were thought to be younger than the Laurentian gneisses and granites and to belong to the Huronian System. In his description of the boundary between the Southern and Grenville provinces, Bell noted that the strike of the adjacent Huronian rocks did not always parallel the course of the dividing line between the two systems, and suggested that a "considerable fault" coincided with this common boundary from Broder Township to beyond the Wanapitei River.

Coleman (1914) concurred with Bell's Laurentian correlation of the granites in the Grenville province, but, in order to explain their intrusive relationship to the Huronian System, he reclassified the sedimentary rocks as Pre-Laurentian and, therefore, Pre-Huronian. Based on the presence of quartzite bands, kyanite schists and, most significantly, an exposure of crystalline limestone, Coleman (1914) correlated the paragneisses in the Grenville province southeast of Sudbury with the Grenville Series in Eastern Ontario and Quebec.

Quirke and Collins (1930) thought that the granitic rocks and paragneisses of the Grenville province were the metamorphic and metasomatic equivalents of the Huronian sedimentary rocks to the west. In their mapping from Bruce Mines southeastward to Killarney, they noticed a progressive increase in the intensity of folding, granitic intrusion, and metamorphism of the Huronian strata. In the vicinity of Killarney and northward for a distance of about twenty miles, they noted that the contact between sedimentary rocks and granite is fairly abrupt and straight--suggesting a fault contact along which Killarney granite magma had intruded. However, north of T. B. 68 (now known as Sale Township) the contact appears to be transitional and is no longer straight, but becomes extremely dentate with interfingering lobes of granite and sedimentary rocks. Within the "Killarnean batholithic complex", Quirke and Collins carefully described what they considered to be recognizable Huronian formations in various stages of transformation into granite. They believed that these outliers of sedimentary rocks showed on a small scale the process of transformation of the Huronian sequence as a whole.

The gneisses in the Grenville province are now known to have been metamorphosed about 1000 million years ago (Stockwell, 1964). From Quirke and Collins' description of the relations between "Killarnean" granite and gneiss, it is clear that they had no conception of the 1000 million year (Grenville) orogeny that affected these rocks, and considered the Grenville province gneisses to be the deep-seated products of the Hudsonian orogeny. Phemister (1960) considered the Grenville province rocks in Broder and Dill townships also to be metamorphic and metasomatic equivalents of the Huronian rocks to the northwest. His evidence is the same as that of Quirke and Collins, i.e., "...field evidence of structural continuity and progressive transition from unmetamorphosed to metamorphosed material is fundamental and is indicative of a corresponding stratigraphic and petrogenetic continuity." (op. cit., p. 118).

Recent work along the Grenville-Superior and Grenville-Churchill (Labrador Trough) province boundaries suggests that, in part at least, rocks of older structural provinces have been incorporated into the Grenville province during the Grenville orogeny (See especially: Davis, Hart, Aldrich, Krogh, and Munizaga, 1967; Stockwell, 1965; Grant, 1964; Gastil and Knowles, 1960; Deland, 1956.)

In the Grenville province south of Lake Timagami, Grant (1964) demonstrated by rubidium-strontium istopic studies that granitic rocks of the Superior province with a primary age of approximately 2400 m.y. were metamorphosed by the Grenville orogeny at approximately 920 m.y. (See location of dated samples near the northeast corner of the larger-scale map on Figure 1.) In a similar type of study, Krogh (in Davis, et al., 1967 p. 383) determined a 1750 m.y. rubidium-strontium whole-rock isochron for "Killarnean granite" along the Grenville front near Coniston and in the northeast corner of the research area (See locations on Fig. 1). These granites are in contact with gneisses containing biotite dated at approximately 900 m.y. (Fig. 1).

The 1750 m.y. isochron for granitic rocks adjacent to the Grenville front suggests that these rocks are products of the Hudsonian orogeny metamorphosed during the Grenville orogeny, rather than Huronian sedimentary rocks transformed into granite during the Grenville orogeny. This conclusion differs fundamentally from that of Quirke and Collins (1930) and Phemister (1960).

CHAPTER II

STRATIGRAPHY AND PETROGRAPHY

Huronian sedimentary rocks are the oldest rocks found within the thesis area. This sedimentary sequence is intruded by sills and dikes of hornblende metagabbro (Nipissing-type diabase or Sudbury gabbro). The sedimentary rocks and metagabbro are intruded by the Eden Lake complex[1] (hornblende gabbro to hornblende diorite and trondhjemite) and the Chief Lake batholith[1] (quartz diorite to quartz monzonite). The Eden Lake complex and the Chief Lake batholith are separated by a screen of quartzite and cannot be given an age relative one to another on the basis of field relations. Portions of both plutonic bodies have undergone post-intrusion deformation which has produced penetrative foliation and lineation that are characteristic structural features of the gneisses in the eastern part of the area. The gneisses east of Chief Lake batholith attained their present structural and metamorphic state during the youngest "orogenic" event in the region, and are intruded by several small post-orogenic diabase dikes.

The relative sequence of geologic events can be correlated to a certain extent with radiometric dates from minerals and whole-rocks in the Sudbury area. Van Schmus (1965) determined that Huronian sediments were deposited between 2500 and 2100 million years ago (based on 2500 m.y. ages

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[1]The Eden Lake complex and the Chief Lake batholith are the names used in this thesis for two spatially separate igneous bodies classified as "Killarnean" by Collins (1938). The part of the Chief Lake batholith in Broder township was mapped as a zone of feldspathization and alteration of Wanapitei Quartzite and Sudbury Gabbro by Phemister (1961; map in Grant, et al., 1962).

of Archean basement and a 2155 $\pm$ 80 m.y. age for the Nipissing diabase). Stockwell (1964) dated the Hudsonian orogeny in the Southern Province at 1640 million years--which presumably was the time of formation of the Penokean fold belt.

One whole-rock and two biotite separates from the research area have been dated by the rubidium-strontium method (Locations shown on Fig. 1). A specimen of the Chief Lake batholith gave a 1750 m.y. date (Krogh, in Davis et al., 1967); biotite from the migmatitic western border of the batholith gave a 1070 m.y. date (Tilton, et al., 1960). The other biotite separate is from the northeast border of the Eden Lake trondhjemite and it gave a 1430 m.y. date (Fairbairn, et al., 1960).

The whole-rock determination probably indicates the primary age of the batholith, and the biotite age may indicate the final recrystallization during the Grenville orogeny. The biotite date from the Eden Lake trondhjemite, however, may not represent a specific event because the specimen was collected less than one mile west of the 1070 m.y. old biotite from the batholith. If this portion of the Eden Lake complex was affected by the Grenville orogeny, the 1430 m.y. date may be intermediate between the final metamorphic event in the area and the primary age of the complex.

Table 1 summarizes the major geologic events in the area. The Huronian sediments were deformed during the Hudsonian orogeny and possibly were intruded at about that time by the Eden Lake complex and Chief Lake batholith--although it is certain only that both of these intrusions pre-date the Grenville orogeny and post-date the intrusion of Sudbury-type gabbro. The gneiss and amphibolite in the eastern part of the area were metamorphosed during the Grenville orogeny. The migmatite around the borders of the Chief Lake batholith was deformed during the Grenville orogeny, and may have been originally a xenolithic border phase formed when the batholith was intruded.

Age Range	Geological Event	Related rocks
1100-900 m.y.*	Grenville orogeny	gneiss, amphibolite, migmatite
1700-1400 m.y.*	Hudsonian orogeny, intrusion of plutonic rocks	Chief Lake batholith, Eden Lake complex
2235-2075 m.y.	gabbroic intrusion	Hornblende metagabbro (Nipissing diabase or Sudbury gabbro)
2500-2075 m.y.	Huronian sedimentation	quartzite, graywacke, conglomeratic graywacke

*The age ranges 1100-900 m.y. and 1700-1400 m.y. for the Grenville and Hudsonian orogenys represent the range in metamorphic mineral radiometric dates from the Grenville and Southern provinces near Sudbury, and probably mark the last phase rather than the entire time-span of the orogenic events.

HISTORY OF MAJOR GEOLOGICAL EVENTS

Table 1

I. HURONIAN SEDIMENTARY SEQUENCE

Within the Huronian rocks in the map area, feldspathic quartzite and arkose are the dominant lithologies with subordinate amounts of graywacke and conglomeratic graywacke. These sedimentary rocks previously have been correlated with the Lower Huronian Bruce Group by Collins (1938) and the Pre-Huronian Sudbury Group by Thomson (in Grant et al., 1962).

Figure 3, a compilation map of the area south and east of the Sudbury Basin, shows the regional structure and stratigraphy within the sedimentary sequence between the Sudbury Basin and the Grenville gneisses. The area is structurally complex with both northeast and northwest trending folds, and numerous transverse and strike faults. However, a consistent stratigraphy can be determined from facing directions of beds near unfaulted contacts between quartzite and graywacke-conglomerate units. Regionally, the sedimentary sequence is a continuous framework of feldspathic quartzite within which discontinuous lenses of graywacke and conglomerate occur at certain stratigraphic horizons.

Table 2 compares the stratigraphic succession derived from Figure 3 with the successions given by Collins (1938) and Thomson (1961a). Grouping conglomerates, graywackes and limestones as single formational units enables the dominantly quartzitic sedimentary sequence to be broken into five (structure section E-G) or six (structure sections A-B and C-D) formational units defined by stratigraphic position and lithology. Grouping the McKim-Ramsay Lake formations and the Bruce-Esponola formations produces a perfect correlation on Table 2 with Collins' (1938) succession. Thomson's (1961a) definitions of the Sudbury, Bruce and Cobalt groups include too many lithologies to allow detailed correlations with the stratigraphic succession derived from Figure 3.

		Collins (Copper Cliff Sheet, 1938)	Structure Sections A-B and C-D (Fig. 3)	Structure Section E-G (Fig. 3)	Thomson (1961; in Grant et al., 1962)		
	Cobalt Series	banded cherty quartzite	(intrusive contact)				
		Lorrain Quartzite	upper vitreous quartzite	(erosion)			
		Gowganda Formation	upper graywacke and quartzite (minor conglomerate)	upper conglomerate, graywacke, quartzite and limestone	Gowganda Formation	Cobalt Group	Huronian
		(unconformity)					
Huronian		Serpent Quartzite	middle feldspathic quartzite	middle feldspathic quartzite	Mississagi Formation	Bruce Group	
	Bruce Series	Espanola Graywacke limestone	middle graywacke and conglomerate	middle conglomerate graywacke and limestone			
		Bruce Conglomerate					
		Mississagi Quartzite	lower feldspathic quartzite	lower felspathic quartzite	(great unconformity)		
		Ramsay Lake Conglomerate	lower graywacke and conglomerate	lower conglomerate			
		(unconformity?)					
Pre-Huronian	Sudbury Series	McKim Graywacke			conglomerate, quartzite, graywacke and biotite paragneiss, limestone	Sudbury Group	Pre-Huronian
		volcanic rocks with interbeds of graywacke	volcanic rocks with interbeds of graywacke	volcanic rocks	acid and basic volcanics	Volcanic Group	
		Copper Cliff Rhyolite					

CORRELATION OF METASEDIMENTARY ROCKS

Table 2

Collins believed that the Ramsay Lake Conglomerate marked the base of the Huronian sequence. However, Young and Church (1966) found a conformable succession from the volcanic rocks upwards through the lower part of the Bruce Group, and,therefore,included the McKim Formation and the underlying volcanic rocks in the Huronian sequence.

In the following sections, the Huronian sedimentary rocks within the area (Fig. 2) are described in detail. Wherever possible, the informal stratigraphic units used in the thesis are correlated with the formations established by Collins.

Lower graywacke (McKim-Ramsay Lake formations)

Chlorite schist exposed in the extreme north-central part of the map area appears to underlie the lower feldspathic quartzite and is the only observed occurrence of the lower graywacke in the area. The pelitic portions of the unit are well foliated, and thin quartzite interbeds commonly are transposed parallel to the foliation. Exposures of the metagraywacke are not abundant, and generally occur as low, narrow ridges aligned parallel to the foliation. Southwestward the metagraywacke appears to interfinger with feldspathic quartzite. Several pebble bands were seen in the northeast portion of the metagraywacke.

In hand specimen, the lower graywacke has the appearance of a green schist with rectangular to ovoid, gray porphyroblasts. The porphyroblasts, which resemble staurolite or andalusite in hand specimen, are fine grained aggregates of muscovite, chlorite and quartz. Mineral phases in the lower graywacke in order of decreasing abundance are muscovite, quartz, chlorite and plagioclase.

An accurate estimate of the sedimentary thickness of the lower graywacke can not be made because of the strongly foliated and sheared nature of the unit; however, the maximum exposed thickness is about 1700 feet.

Lower feldspathic quartzite (Mississagi Formation)

The quartzite occurring northwest of the shear zone along the western border of the Chief Lake batholith is correlated with the Mississagi Formation. The lower feldspathic quartzite is overlain by the middle conglomeratic graywacke along the southeast shore of Long Lake. The upper contact of the quartzite is sharp, possibly because of shearing along the contact between quartzite and the pelitic matrix of the overlying conglomerate.

The lower feldspathic quartzite is resistant to weathering and forms ridges parallel to the strike of bedding. The color of the weathered rock surface varies from dull white to red-brown.

The quartzite contains abundant cross-beds, generally of the torrential type (Plate 1A) so that it is difficult to determine facing directions. Top and bottom-set beds are generally semi-pelitic and are more readily weathered than the coarser grained cross-sets, thus giving outcrops a distinctive ribbed appearance. Some graded beds are present (Plate 1B). Bedding thickness is variable, averaging about one foot, but cross-bedded units more than five feet thick are present.

An average of three modal analyses of the lower feldspathic quartzite is: quartz (60 per cent), feldspar (35 per cent), and mica (5 per cent). The range in feldspar content is from 25 to 40 per cent. On the basis of these modal determinations, the lower feldspathic quartzite unit is identified as an arkose.

Stratigraphic duplication due to folding within the lower feldspathic quartzite makes a thickness estimate difficult; however, the unit appears to be at least 4500 feet thick.

Middle conglomeratic graywacke (Bruce Conglomerate)

The middle conglomeratic graywacke occurs in the core of a tight syncline along the southeast shore of Long Lake. The northeast end of the structure is modified by faulting, but in the southwest the conglomerate pinches out into feldspathic quartzite. The syncline is not basin-shaped, but opens to the south indicating that the southern termination of the conglomerate is probably due to restricted deposition. Collins (1938) mapped the middle conglomeratic graywacke as Bruce Conglomerate.

The conglomeratic graywacke is more resistant to erosion than the surrounding quartzite and forms a ridge with a maximum relief of about 150 feet. The rock weathers dark gray to black with a gritty, uneven surface.

Pebbles of granite and quartzite less than one foot in diameter are dispersed in the graywacke matrix. The rock is a graywacke microconglomerate in thin section. The coarser grained fraction is made up of clastic quartz and feldspar grains, whereas the fine grained matrix is muscovite, biotite and quartz.

Provided that a double thickness of the conglomerate is exposed in the syncline, and that the average dip is sixty degrees, the maximum thickness of the unit is 650 feet.

Middle (?) feldspathic quartzite (Serpent Quartzite?)

The middle feldspathic quartzite occurs south of the westward extension of the Chief Lake batholith border shear zone in Eden Township, and is tentatively correlated with the Serpent Quartzite. Collins (1938) mapped this feldspathic quartzite as Mississagi Formation. However, the middle feldspathic quartzite is overlain by an interbedded mixture of graywacke and quartzite with minor conglomerate. This overlying heterogenous lithologic sequence is thought to represent the Gowganda Formation; hence the correlation of the underlying feldspathic quartzite with the Serpent Formation.

On the basis of field appearance, the middle and lower feldspathic quartzites are indistinguishable.

An average of two modal analyses of the middle felspathic quartzite is: quartz (66 per cent), feldspar (26.5 per cent), and mica (7.5 per cent). The base of the middle quartzite unit is not present in the area, and a minimum thickness is 2000 feet.

Upper graywacke-quartzite (Gowganda Formation?)

The middle feldspathic quartzite grades upwards into a sequence of graywacke and feldspathic quartzite which is tentatively correlated with the Gowganda Formation. Collins (1938) included this heterogeneous lithologic sequence within the Mississagi Formation. The upper graywacke-quartzite can be traced as scattered outcrops trending northwest for about two miles from the north shore of Wavy Lake in Eden Township. A similar lithologic sequence also occurs as part of the large inclusion mass in the Chief Lake batholith at Chief Lake (See Fig. 2.).

Minerals present in the graywacke portion of the unit are quartz, microcline and plagioclase which may be of detrital origin, and chlorite, biotite, and muscovite which are probably secondary. Microscopic garnets are present in one thin section. An argillitic portion of the unit contains quartz, plagioclase, chlorite, biotite, amphibole, epidote and minor microcline.

In the exposures north of Wavy Lake, graywacke is more abundant than quartzite. At least one pebble bed is present, but could not be traced. In the inclusion mass at Chief Lake, however, conglomeratic graywacke about 50 feet thick is present. The top of the upper graywacke-quartzite is marked by a sharp contact with the vitreous quartzite unit. The upper graywacke-quartzite is about 2600 feet thick.

Vitreous quartzite (Lorrain Quartzite)

A sequence of vitreous quartzite beds occurs stratigraphically above the upper graywacke-quartzite north of Wavy Lake, and is correlated with the Lorrain Quartzite. Collins (1938) included these beds in the Mississagi Formation. The vitreous quartzite is less feldspathic and cross-bedded than the underlying quartzite units, and is easily distinguished in the field by its glassy appearance and light orange to pale green color. A narrow band of green vitreous quartzite on the western side of the inclusion mass at Chief Lake is thought to be correlative with the vitreous quartzite unit. In the area north of Wavy Lake, the quartzite is a ridge former.

The vitreous quartzite is divided into three subunits on the basis of color and relative bedding thickness. The color of the vitreous quartzite varies from orange to pale green to white as it contains microcline, muscovite or is composed of quartz alone. The lower subunit is composed of beds varying from one to three feet thick containing microcline and muscovite, with muscovite more concentrated along bedding planes. The middle subunit is the most orthoquartzitic, and contains a few per cent of muscovite, but very little feldspar. Bedding thickness in the middle subunit varies from one to three feet, and a few cross-beds are present. The upper subunit contains laminated beds one to six inches thick and is richer in feldspar and muscovite than the underlying vitreous quartzite.

The top of the vitreous quartzite has been assimilated by the Chief Lake batholith which contains numerous inclusions of the refractive quartzite. A minimum thickness for the vitreous quartzite is 3300 feet.

Uncorrelated Sedimentary Rocks

Isolated xenoliths of sedimentary rocks occur within the intrusive rocks in the area, but most cannot be correlated with a specific stratigraphic unit. Inclusions of both feldspathic and pure vitreous quartzite, metagraywacke, and conglomerate occur within the Eden Lake complex and the Chief Lake batholith. Generally, the metamorphic grade increases from west to east in the area and the protoliths of most inclusions within the Chief Lake batholith can be recognized as far as the eastern border migmatite zone. East of the migmatite zone the only rocks of recognizable sedimentary origin are a few discontinous bands of vitreous-type quartzite.

Summary of Huronian Stratigraphy

The sedimentary rocks within the research area appear to represent the Huronian stratigraphic succession from the McKim Graywacke upwards to the Lorrain Quartzite. No unconformities occur in the succession, but the base of the Serpent Quartzite appears to have been removed by faulting. Correlation of the Serpent Quartzite is the most doubtful, and is based on a correlation of the overlying graywacke-quartzite unit with the Gowganda Formation. The total thickness of Huronian sedimentary rocks exposed in the area is more than 14,750 feet.

II. HORNBLENDE GABBRO (NIPISSING DIABASE OR SUDBURY GABBRO)

The Huronian sedimentary rocks in the research area and throughout the region between Bruce Mines and Sudbury are intruded by bodies of gabbro and diabase, which have been termed Nipissing diabase or Sudbury gabbro. As noted previously, Van Schmus (1965) determined an intrusive age of 2155 ± 80 m.y. for the Nipissing diabase.

In the map area, hornblende gabbro is dark green to black, massive, and equigranular or sheared. It is more resistant to erosion than feldspathic quartzite and forms ridges. The general geometry of the larger gabbroic bodies is sill-like, with the borders parallel to regional bedding trends. The metagabbros west and east of Lohi Lake in Broder Township form the limbs of a south-plunging syncline within the lower feldspathic quartzite (See Fig. 3.). Another gabbroic body is intruded along the contact of the middle feldspathic quartzite and upper graywacke-quartzite northwest from Wavy Lake. Numerous gabbroic inclusions occur in the Eden Lake complex and the Chief Lake batholith.

Gabbroic rocks exhibit signs of retrograde metamorphism. Amphibole rather than pyroxene is the major mafic mineral, and plagioclase is altered to sericite and epidote. The amphibole commonly is actinolite, as pale green prismatic crystals and fibrous mattes. The more felsic portions of the gabbroic bodies contain interstitial quartz, and micropegmatitic intergrowths of quartz and plagioclase. Plagioclase is too highly altered for optical determinations.

III. EDEN LAKE COMPLEX

The igneous rocks that intrude the middle feldspathic quartzite in southwestern Eden township are collectively termed the Eden Lake complex. This plutonic complex ranges in composition from hornblende gabbro and hornblende diorite to trondhjemite. Collins (1938) correlated the basic portion of the complex with the Nippissing diabase and the acid portion with Killarnean granite. Previously, however, Collins (1925) noted that the basic portion of the complex differed appreciably from the Nippissing diabases in the North Shore district.

The Eden Lake complex is divided into two map units: (1) hornblende gabbro and hornblende diorite, and (2) biotite trondhjemite. The contact between the two field units is commonly a narrow interbanded zone of acid and basic layers, rather than a continuous gradation from acid to basic rocks or a sharp contact. The geometry of the complex is imperfectly known because only the northeast portion is exposed in Eden Township. Collins (1938) showed the complex to be about six miles long by two miles wide, trending northeast parallel to the regional bedding strike. The ratio of basic to acid portions is about 1:6, with the gabbro-diorite phase occurring only in the northwest part of the complex.

Basic Rocks

Hornblende gabbro and hornblende diorite are surrounded on three sides by trondhjemite; the northern side is in contact with the middle feldspathic quartzite. Inclusions of feldspathic quartzite and metagraywacke occur within the basic phase of the complex. Disseminated gold-bearing arsenopyrite was mined from one large feldspathic quartzite inclusion (Baker, 1917).

Generally, the gabbro-diorite is a medium grained, fresh-looking rock composed of about two-thirds light gray plagioclase and one-third black hornblende. A few outcrops of banded hornblende-rich and plagioclase-rich layered gabbro were found, which suggests that gravity settling of minerals occurred within the basic portion of the complex (Plate 1C). The present attitude of the layers is vertical, striking northeast subparallel to the regional trend of bedding in quartzite. If formed by primary gravitational settling, the layers would have been "sedimented" in a nearly horizontal position and subsequently tilted into the vertical position. Parallelism of the layering and the regional trend of bedding suggests that the Huronian rocks were in a near horizontal position during the intrusion, and, therefore, that the Eden Lake complex is pre-Hudsonian orogeny.

Four modal analyses of hornblende diorite and gabbro are listed in Table 3. Plagioclase (andesine to labradorite) is generally the dominant phase and occurs as subhedral laths which commonly exhibit normal zoning. Compositions were determined by maximum extinction on albite twins. However, the majority of the plagioclase crystals are untwinned, and the determinations may not be representative of the rock as a whole. Specimens 908 and K325 were collected from near the trondhjemite contact and specimens 774 and 775 are from the central portion of the basic phase of the complex. The central portion appears gabbroic and the borders dioritic.

Hornblende is dark green pleochroic, and occurs as subhedral crystals and as large poikilitic grains enclosing plagioclase laths. Biotite appears to be a minor primary phase and is commonly altered to chlorite. Magnetite is accessory. One dioritic specimen (specimen 908) from near the trondhjemite contains interstitial quartz and biotite.

Spec.	Qtz.	Plag.	Hbd.	Bio.	Ep.	Ap.	Cht.	Mte.	Total Counts
908	5.4	59 (An_{46})	20.4	5.8	2.0	3.0	1.2	3.2	500
775	_	39 (Alt.)	55.6	_	0.2	_	2.8	1.8	500
774	_	71 (An_{63})	23	1.8	3.2	_	_	1.0	500
K325	_	62.6 (An_{46})	24.8	_	9.8	_	2.4	0.4	500

MODAL ANALYSES OF HORNBLENDE DIORITE AND GABBRO
FROM THE EDEN LAKE COMPLEX

Table 3

Trondhjemite

Most of the Eden Lake Complex exposed in Eden Township is medium grained, equigranular, biotite quartz diorite or trondhjemite. The term trondhjemite is better suited for this rock than quartz diorite because of the virtual absence of alkali feldspar. Potassium feldspar was seen in only four of eleven thin sections examined, and ranged from 1.7 to 2.8 per cent. The minor amount of potassium feldspar is in contrast to that in the Chief Lake batholith, in which microcline constitutes from one-quarter to three-quarters of the total feldspar.

Metasedimentary and metagabbroic xenoliths are common within the trondhjemite and usually occur in clusters. The shapes of inclusions shown on the geologic map (Fig. 2) are generalized because many of the inclusions are too small to be depicted accurately at the scale of the mapping. The eastern and northern borders of the trondhjemite are sharp and non-migmatitic.

The trondhjemite south and east of Eden Lake is lineated and foliated, the result of post-intrusion deformation. The central part of the Eden Lake complex, however, is unfoliated and does not show any evidence of primary or secondary flowage. The trondhjemite west of the basic portion of the complex is not foliated near the northern contact with the middle feldspathic quartzite, but becomes foliated and migmatitic to the southwest. The origin and local development of this migmatitic area, which is composed of about equal proportions of trondhjemite and contorted feldspathic quartzite, is not understood.

The results of modal analyses of trondhjemite are listed in Table 4. Plagioclase (oligoclase to andesine), quartz, biotite and epidote are present in all the specimens examined. Muscovite (generally within plagioclase) is common, and microscopic garnets are present in four thin sections. Specimen 787, from the contact between hornblende

Spec.	Qtz.	Ksp.	Plag.	Bio.	Hbd.	Musc.	Ep.	Cte.	Cht.	Mte.	Gar.	Total
798	39	-	41.4 (An_{27})	10.2	-	4.2	4.6	0.2	-	0.4	-	500
787	11.4	-	73.0	7.4	4.0	-	0.2	-	2.0	-	-	500
K347	18.2	-	60.8 (An_{31})	19.5	-	-	1.3	-	-	0.2	-	400
2176	20.3	-	51.7 (An_{33})	22.3	-	-	4.0	-	-	-	1.7	300
907	16.6	-	52.4 (An_{29})	23.4	-	2.0	2.6	-	1.2	-	1.8	500
K275	42.4	-	53.4 (An_{38})	18.0	-	1.0	2.4	-	0.6	-	0.2	500
650	49.0	-	39.3	7.0	-	2.3	2.3	-	-	-	-	300
933	35.3	2.7	43.0	13.0	-	4.7	1.3	-	-	-	-	300
K286	33.5	1.7	53.2	7.7	-	2.0	0.7	-	-	-	1.2	400
834	43.8	2.8	39.2 (An_{43})	6.6	-	4.0	3.4	-	-	-	-	500
59	35.5	2.5	44.2	9.5	-	3.3	5.0	-	-	-	-	400

MODAL ANALYSES OF TRONDHJEMITE FROM EDEN LAKE COMPLEX

Table 4

diorite and trondhjemite, contains hornblende as well as biotite, and less quartz than any of the other specimens. Chlorite occurs as an alteration product of biotite, and epidote commonly is found as inclusions within plagioclase.

Petrogenesis of the Eden Lake Complex

Field evidence suggests that the Eden Lake complex was intruded into the Huronian sedimentary rocks early in their tectonic history, possibly when bedding was horizontal. The hornblende gabbro-diorite and trondhjemite phases seem to have been derived by differentiation of a single magma. The preponderance of trondhjemite over gabbro-diorite, and the presence of what appear to be primary hornblende and biotite suggest that the magma was dioritic in composition, with a rather high water content. Barth (1962, p. 222) related the lack of potassium feldspar in trondhjemite to a high water content in the magma; early formed biotite extracts so much potassium from the melt that none is left for the formation of potassium feldspar in the later stages.

The petrology of the Eden Lake complex is an interesting research topic that has not been thoroughly investigated. The Eden Lake complex probably was intruded before the Chief Lake batholith, although little field evidence exists to support this conclusion. It is certain only that the complex is younger than Sudbury gabbro and older than the Grenville orogeny.

IV. CHIEF LAKE BATHOLITH

The Chief Lake batholith is the major lithologic unit in the area, and occupies a geographic position between the Huronian sedimentary rocks on the west and the Grenville gneisses to the east. Within the map area, the batholith covers about thirty square miles, varying in width from five miles in southern Tilton Township to less than two miles in southern Broder Township. Batholithic rocks extend at least two miles north of the thesis area into Dill Township. (See Fig. 3.) The southern limit of the batholith is not known, but it is believed that the Chief Lake batholith represents the northern extension of the Killarney batholith mapped by Quirke and Collins (1930). A body of massive gneiss with microcline megacrysts in southeastern Tilton Township is thought to be a metamorphosed outlier of the Chief Lake batholith. The small plutons occurring west of the batholith also are thought to be genetically related to the main body.

Contact Relations

Migmatite zones occur along the southwestern and eastern borders of the batholith (Fig. 2). Both migmatites are foliated and lineated mixtures of about equal proportions of batholithic and country rocks.

The migmatite contacts are transitional. For example, the middle feldspathic quartzite on the southwestern shore of Wavy Lake grades eastward into migmatite by a progressive increase in granitic dikes and lenses. As the granitic component increases (e.g., on the southeastern shore of Wavy Lake), the feldspathic quartzite beds become isolated inclusions within the granitic matrix. Further eastward from Wavy Lake, the granitic component is dominant, and the migmatite grades into inclusion-free quartz diorite.

The migmatite zone between the batholith and the Grenville metamorphic rocks is a banded mixture of quartzo-feldspathic paragneiss with minor amphibolite, and gneissic quartz diorite. The migmatite east of the batholith is more thoroughly recrystallized than that along the southeast shore of Wavy Lake; it is a foliated mixture of banded paragneiss and orthogneiss which grades westward into blastoporphyritic quartz diorite gneiss, and grades eastward into porphyroblastic augen gneiss and massive amphibolite.

Most of the western border of the batholith is not migmatitic. From the north shore of Wavy Lake northward for a distance of about two miles, the contact between Huronian sedimentary rocks and the Chief Lake Batholith trends northwest. The rocks in this area are not as intensely deformed as at Wavy Lake or along the eastern border of the batholith, and it is possible to map sills and dikes of granitic rocks intrusive into the Huronian sequence as well as inclusions of vitreous quartzite (Plate 2C, Plate 2D) and metagabbro within the batholith.

The geometry of the contacts between rock types in this area is extremely complex, and was the subject of an M. Sc. thesis by H. R. Spaven (1966). Spaven concluded that field relationships strongly suggest that the Chief Lake batholith was intruded into brittle country rocks, and that quartzite xenoliths are apparently stoped remnants of what was once a continuous quartzite unit. Vitreous quartzite and metagabbro are the most common inclusions found in the batholith. The quartzite xenoliths generally show a preferred orientation, with their longest dimension parallel to the strike of penetrative foliation which transects both the batholith and the inclusions. The borders of the inclusions are sharp, and embayments of quartz monzonite occur parallel to bedding in the inclusions (Plate 2C).

North of the area of xenoliths the contact between the batholith and the envelope rocks swings abruptly northeast, parallel to a major shear zone. The sheared contact is well exposed east of Clearwater Lake where the rocks for about two hundred feet on either side of the contact possess a vertical foliation and mineral lineation. Batholithic rocks within the sheared zone are blastomylonitic with porphyroclasts of microcline surrounded by a fine grained matrix of comminuted feldspar and recrystallized quartz (Plates 3A, 3B). The shear zone passes into the batholith east of Clearwater Lake and joins with another shear zone to the north.

The contact between the batholith and the lower feldspathic quartzite at the northeast corner of Clearwater Lake is not sheared. In this area, the batholith is composed of coarse grained, unfoliated, equigranular quartz monzonite. The contact trends at a large angle to the strike of bedding in the quartzite, and on the scale of an exposure is very irregular. The quartz monzonite is free of quartzite inclusions, but embayments of igneous material parallel to bedding in the quartzite are common. Isolated lenses and dikes of coarse grained quartz monzonite are found within the feldspathic quartzite in the vicinity of the batholith contact. Contacts are sharp and show no obvious effects of metasomatism or contact metamorphism.

Agmatite

Along the northern border of the batholith a large body of hornblende metagabbro is intruded by quartz diorite forming an intrusion breccia. The term "agmatite" is used for this blocky mixture of two igneous rocks. Contacts are generally sharp, but, locally, metagabbro contains porphyroblasts of orange microcline, and quartz diorite is hornblende-bearing. The agmatite mapped along the northern edge of the batholith is similar in appearance to that shown in Plates 2A and 2B, although the agmatite in the photographs is near Wavy Lake.

Texture and Fabric

The batholith contains foliated, porphyritic and equigranular rocks. The general distribution of coarsely porphyritic rocks is indicated by the stippled pattern on the geologic map (Fig. 2). Foliation is the result of post-intrusion deformation which has modified the primary textures to form, most commonly, augen gneiss (Plate 3). Foliation planes generally contain a linear structure due to parallel alignment of elongate quartz and feldspar.

Textures of unfoliated rocks vary from hypidiomorphic granular to porphyritic (Plate 2D). Phenocrysts of perthitic microcline are common, and some anti-perthitic plagioclase phenocrysts are present, as well as plagioclase mantled by microcline. Generally, the matrix of porphyritic rocks is composed of plagioclase and quartz with a subordinate amount of microcline, whereas the phenocrysts are dominantly microcline. Microcline contains from zero to thirty per cent perthite, averaging between five and ten per cent. Myrmekite commonly has replaced microcline along contacts with plagioclase. Central areas of large plagioclase grains generally contain epidote and white mica. Some plagioclase grains exhibit normal zoning.

Foliation and lineation are produced by planar and linear alignment of recrystallized quartz and mica. Feldspar generally has undergone brittle deformation, and phenocrysts are bent, fractured and surrounded by a rim of fine grained feldspar (Plate 6).

Mineral Composition of the Batholith

The composition of the Chief Lake batholith varies from quartz diorite to quartz monzonite[1] with accessory

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[1]Quartz diorite has a ratio of Na, Ca feldspar to K, Na feldspar in excess of 5:3; quartz monzonite has a ratio of K, Na feldspar to Na, Ca feldspar between 5:3 and 3:5 (after Wahlstrom, 1955, p. 307).

biotite and epidote. Minor phases not present in all specimens are: muscovite, hornblende, chlorite, sphene, calcite, magnetite and zircon.

Table 5 contains modal analyses of 51 thin sections from the main body of the batholith, the small plutons west of the main body, and the outlier of massive orthogneiss in the southeastern part of the map area. The distribution of the rocks examined is presented on Figure 8. The more basic varieties contain up to 25 per cent hornblende, 20 per cent biotite, and less than 20 per cent quartz. Hornblende-bearing batholithic rocks generally are associated with xenoliths of metagabbro, and hornblende- and biotite-rich quartz diorite and quartz monzonite probably were formed by assimilation of basic rocks by the magma. Contact metasomatism of basic inclusions is shown by the presence of microcline megacrysts and quartz in metagabbro (Plate 2B); however, many basic inclusions show no metasomatic effects.

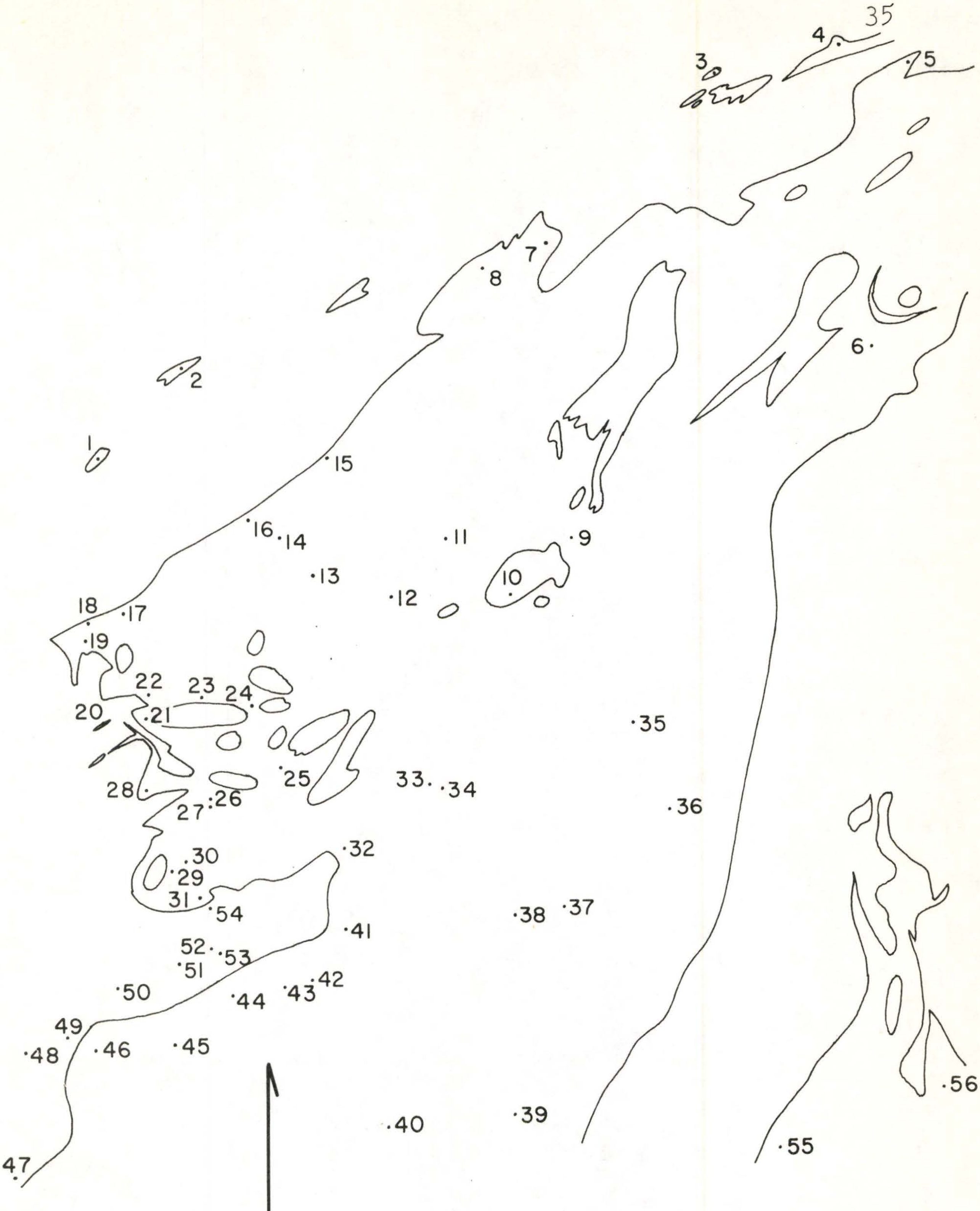
Modal compositions in volume per cent of 33 specimens containing more than 80 per cent quartz, plagioclase and microcline are plotted on Figure 9a. These specimens are the least altered and recrystallized of the batholithic rocks examined. Four specimens plotted on Figure 9a are from small plutons east of the main body of the batholith (numbers 1, 2, 3, and 4), but they plot with the majority of the rocks from the main body. Number 56 is from the orthogneiss in the southeastern part of the map area, and appears to be low in microcline content; however, the modal analysis may not be representative of the bulk composition as the microcline in the gneiss is contained in large megacrysts which are not present in the slide examined.

No systematic relationship between geographic position and modal composition of the rocks appears to exist on the scale sampled. Two-thirds of the specimens plotted on Figure 9a are quartz diorite, and one-third are quartz monzonite. Figure 9b is a diagram of the modal analyses from

No.	Sample	Quartz	K-spar	Plag.	Bio.	Epidote	Musc.	Hbd.	Cht.	Cte.	Sphene	Mte.	Total	%(Qtz.+ Ksp., + Plag.)	Ave. Plag. Comp.	Name of Rock
1	1081	34.3	14.4	39.9	6.0	2.4	8.2	–	2.0	–	–	–	1250	88.6		quartz diorite
2	1133	29.5	5.6	54.9	2.9	3.3	3.2	–	1.0	–	–	–	1250	90.0		quartz diorite
3	1916	24.3	23.1	45.5	3.4	1.1	2.6	–	–	–	–	–	1250	92.9		quartz diorite
4	2177	31.2	17.8	42.2	4.8	2.1	1.7	–	2.0	–	–	–	1250	91.2		quartz diorite
5	2178	24.8	23.3	39.0	7.8	3.5	5.0	–	–	–	0.5	0.6	1250	87.1		quartz diorite porphyry
7	2229	24.2	22.9	45.4	3.8	2.2	1.4	–	–	1.0	–	–	1250	92.6		quartz diorite
8	NEC	19.9	28.0	44.1	3.0	4.2	1.6	–	1.9	–	–	–	1250	92.0		quartz monzonite
9	W110	19.6	2.1	46.3	2.7	9.0	0.1	–	18.6	0.6	–	–	1250	68.9	An_{27}	quartz diorite
11	W139	20.3	15.5	52.6	7.4	1.7	0.9	–	–	–	1.0	0.6	1250	88.4		quartz diorite
12	W26	22.6	20.9	49.4	4.0	1.7	0.9	–	0.4	0.1	–	–	1256	92.9	An_{17}	quartz diorite
13	W20	19.0	9.2	59.0	–	2.2	0.4	–	10.0	0.2	–	–	1250	87.2	An_{18}	quartz diorite
14	2131	26.4	21.3	48.9	0.9	0.6	1.4	–	0.5	–	–	–	1250	96.7		quartz diorite
15	1042	34.4	10.8	33.4	0.4	11.0	8.0	–	–	–	–	–	500	78.6		quartz diorite porphyry
16	1035	27.4	16.2	47.9	0.6	5.7	2.2	–	–	–	–	–	1250	91.5		quartz diorite
17	1112	33.0	21.0	40.9	4.0	1.0	–	–	–	–	0.1	–	1260	94.9		quartz diorite porphyry
18	1107	22.8	19.3	50.5	2.4	3.4	1.4	–	–	–	–	–	1260	94.9		quartz diorite
19	1109	24.1	21.1	42.4	6.4	3.3	2.7	–	–	–	–	–	1500	87.6	An_{18}	quartz diorite
20	2231	34.9	8.7	39.4	7.4	3.0	6.3	–	–	0.3	–	–	1250	82.9	An_{16}	quartz diorite
21	534	9.6	12.6	64.6	8.7	3.5	–	1.0	–	–	–	–	1250	86.9		quartz diorite porphyry
22	H135	21.8	27.8	45.8	3.2	0.6	0.1	–	–	–	–	0.7	1250	95.5		quartz diorite
23	H184	2.0	14.7	56.7	7.0	5.0	–	13.3	–	–	1.3	–	300	73.3		hornblende diorite
24	K38	10.8	–	56.8	19.2	5.0	–	8.2	–	–	–	–	400	77.5		hornblende trondhjemite
25	K10	16.3	3.5	41.5	10.0	4.2	–	24.5	–	–	–	–	400	61.3		hornblende trondhjemite
26	327	18.0	7.7	46.8	13.0	5.5	2.5	6.5	–	–	–	–	400	74.5		hbd. quartz diorite porph.
27	326	8.3	34.7	39.3	15.3	6.4	–	–	–	–	–	–	300	82.3		monzonite porphyry
28	250	25.2	13.0	47.3	8.2	4.3	2.0	–	–	–	–	–	1250	85.4		quartz diorite
29	210	19.8	28.7	44.6	4.3	2.2	–	–	–	0.2	–	0.2	1250	93.0		quartz monzonite
30	H334	17.5	31.9	42.1	5.3	2.9	–	–	–	–	0.3	–	1250	91.5		quartz monzonite porphyry
31	95	20.7	27.5	31.8	15.7	3.5	0.5	–	–	0.3	–	–	400	80.0		quartz monzonite porphyry
32	W385	34.9	28.4	25.4	1.3	2.8	7.2	–	–	–	–	–	1250	88.6		quartz monzonite
33	1540	25.7	18.2	35.4	9.4	9.2	1.5	–	–	–	–	–	1250	85.9	An_{17}	quartz diorite porphyry
34	K407	25.3	17.7	29.1	20.0	6.7	1.3	–	–	–	–	–	300	72.0		quartz monzonite porphyry
36	1695	19.6	35.1	31.3	11.1	0.2	2.7	–	–	–	–	–	1250	85.9	An_{24}	quartz monzonite porphyry
37	1682	29.3	21.0	32.8	13.0	1.3	1.8	–	0.8	–	–	–	1250	83.0	An_{28}	quartz monzonite
38	K395	32.5	39.5	17.2	0.8	0.3	9.7	–	–	–	–	–	400	89.8		granite
40	950	30.5	21.4	39.0	5.4	1.1	2.6	–	–	–	–	–	1250	91.0		quartz diorite porphyry
41	437	34.3	29.3	24.1	2.0	2.0	8.3	–	–	–	–	–	300	87.7		quartz monzonite
42	940	48.1	21.3	14.3	0.3	2.3	13.7	–	–	–	–	–	300	83.7		quartz monzonite
44	K211	31.5	35.5	29.7	1.4	0.8	1.0	–	–	–	–	0.1	1250	96.7		quartz monzonite
45	911	30.4	24.2	36.9	6.3	0.5	1.7	–	–	–	–	–	1250	91.5		quartz monzonite
46	173	26.2	24.4	39.7	7.1	1.3	1.1	–	–	–	–	0.2	1250	90.3		quartz diorite
47	1000	31.8	19.8	38.1	6.1	2.2	2.0	–	–	–	–	–	1250	89.7		quartz diorite
48	103	39.3	22.6	21.7	9.7	3.0	4.7	–	–	–	–	–	300	83.5		quartz monzonite
49	K356	31.5	24.2	31.0	4.3	4.0	5.0	–	–	–	–	–	400	86.8		quartz monzonite
50	1013	24.2	20.0	44.6	3.2	3.4	4.6	–	–	–	–	–	500	88.8		quartz diorite porphyry
51	69	32.9	12.4	40.5	8.2	1.3	4.2	–	–	–	–	–	600	85.6		quartz diorite porphyry
52	72	31.4	22.5	31.7	2.2	–	12.2	–	–	–	–	–	1250	85.5		quartz monzonite
53	73	37.4	17.4	35.3	3.3	2.3	5.7	–	–	–	–	–	300	89.3		quartz diorite
54	84	35.1	16.2	36.2	10.2	0.8	1.5	–	–	–	–	–	1250	87.5		quartz diorite
55	1771	23.2	27.2	40.8	8.0	0.2	0.6	–	–	–	–	–	500	91.2		quartz monzonite porphyry
56	1849	36.0	1.5	57.7	4.7	0.1	–	–	–	–	–	–	1250	95.2		trondhjemite porphyry

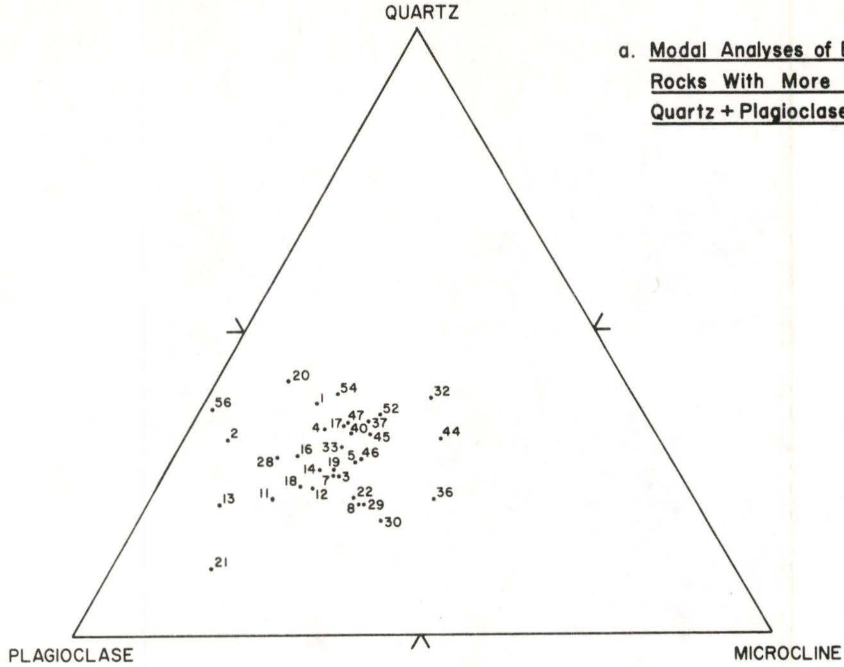
MODAL ANALYSES OF ROCKS FROM THE CHIEF LAKE BATHOLITH

Table 5

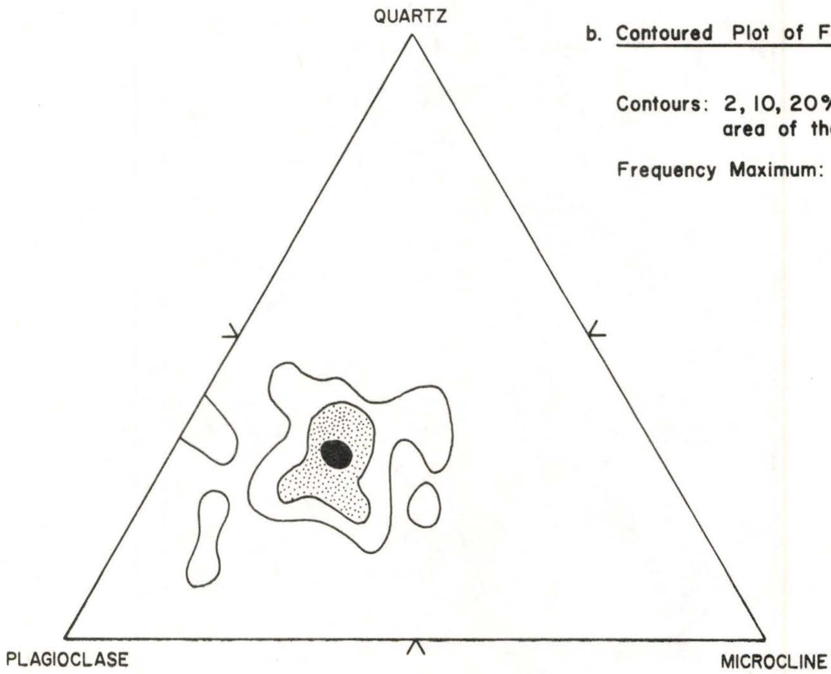


OUTLINE MAP OF CHIEF LAKE BATHOLITH SHOWING LOCATIONS OF ANALYSED THIN SECTIONS

Figure 8



a. <u>Modal Analyses of Batholithic Rocks With More Than 80% Quartz + Plagioclase + Microcline</u>



b. <u>Contoured Plot of Figure 9a</u>

Contours: 2, 10, 20% per unit area of the diagram

Frequency Maximum: 30% / 1% area

Figure 9

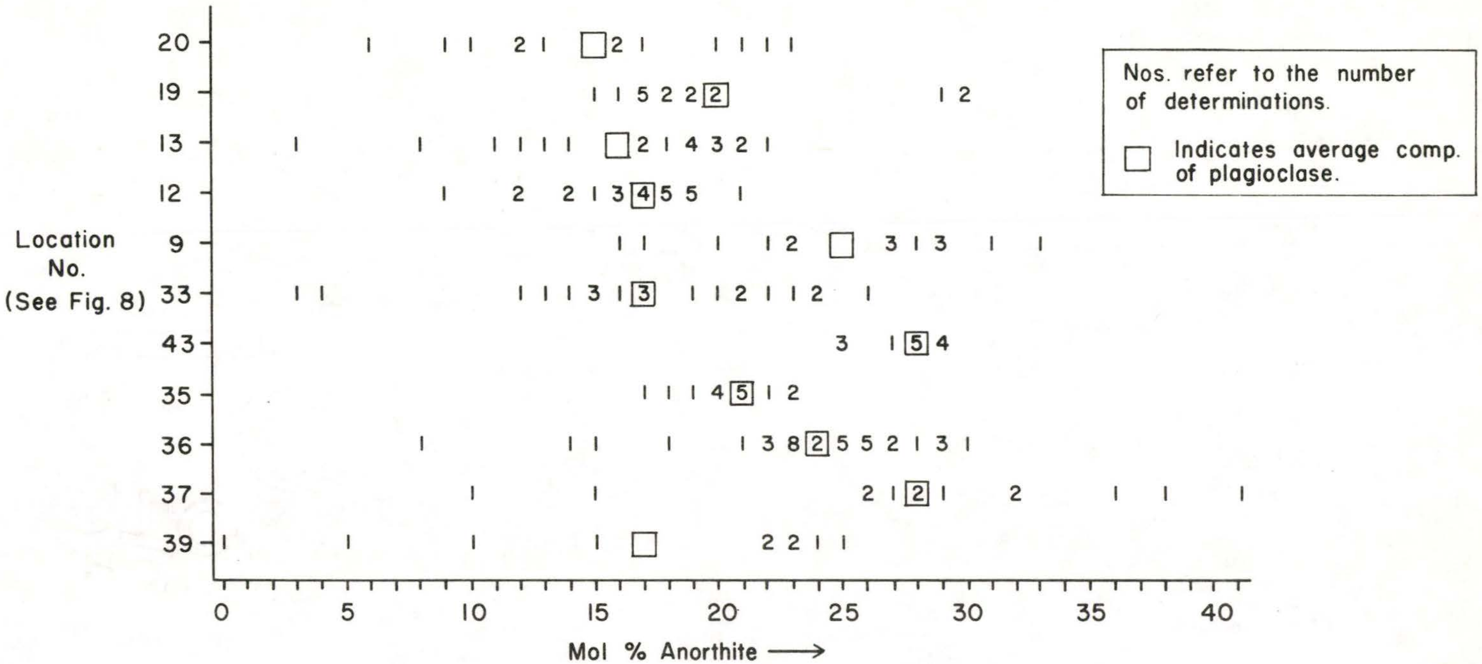
Figure 9a contoured according to the number of analyses within a unit area of the triangular diagram. The frequency maximum on Figure 9b contains thirty per cent of the data (i.e., ten of the thirty-three analyses fall within one per cent of the area of the diagram), and represents a modal composition of 30 per cent quartz, 45 per cent plagioclase, and 25 per cent microcline.

Plagioclase Composition

The anorthite content of plagioclase in the batholith was determined by means of the electron probe microanalyser. Plagioclase was analysed in polished thin sections by comparison of the intensity of Ca Kα X-radiation from the specimen with a calibration line drawn by plotting the intensity of Ca Kα X-radiation against anorthite mol per cent for standards of known anorthite content. Glasses of known plagioclase composition (An_0, An_{20}, An_{40}, An_{60}, An_{80}, An_{100}) were used as analytical standards.

Figure 10 shows the range and average anorthite content of plagioclase in thin sections. The geographic locations of the specimens are shown on Figure 8. The numbers shown on Figure 10 (e.g., 1, 2, 3, etc.) refer to the number of determinations at a particular composition, and may be either within a grain or from different grains.

The range in anorthite mol per cent within each specimen is mainly variation between fairly homogeneous grains. The largest compositional variation within a single grain was found in number 37, wherein a plagioclase phenocryst varies from An_{40} in the center to An_{10} at the edge. The maximum variation in anorthite content found in a single plagioclase grain in the ten other specimens is 5 per cent (number 36). The anorthite content of three grains of myrmekite replacement of microcline in number 12 ranged as follows: An_{12-16}, An_{16-17}, An_{16-19}. The composition of 11 non-myrmekitic



VARIATION OF MICROPROBE DETERMINATIONS OF PLAGIOCLASE FROM THE CHIEF LAKE BATHOLITH

Figure 10

plagioclase grains in the same slide ranged from An_9 to An_{21}, averaging An_{18}.

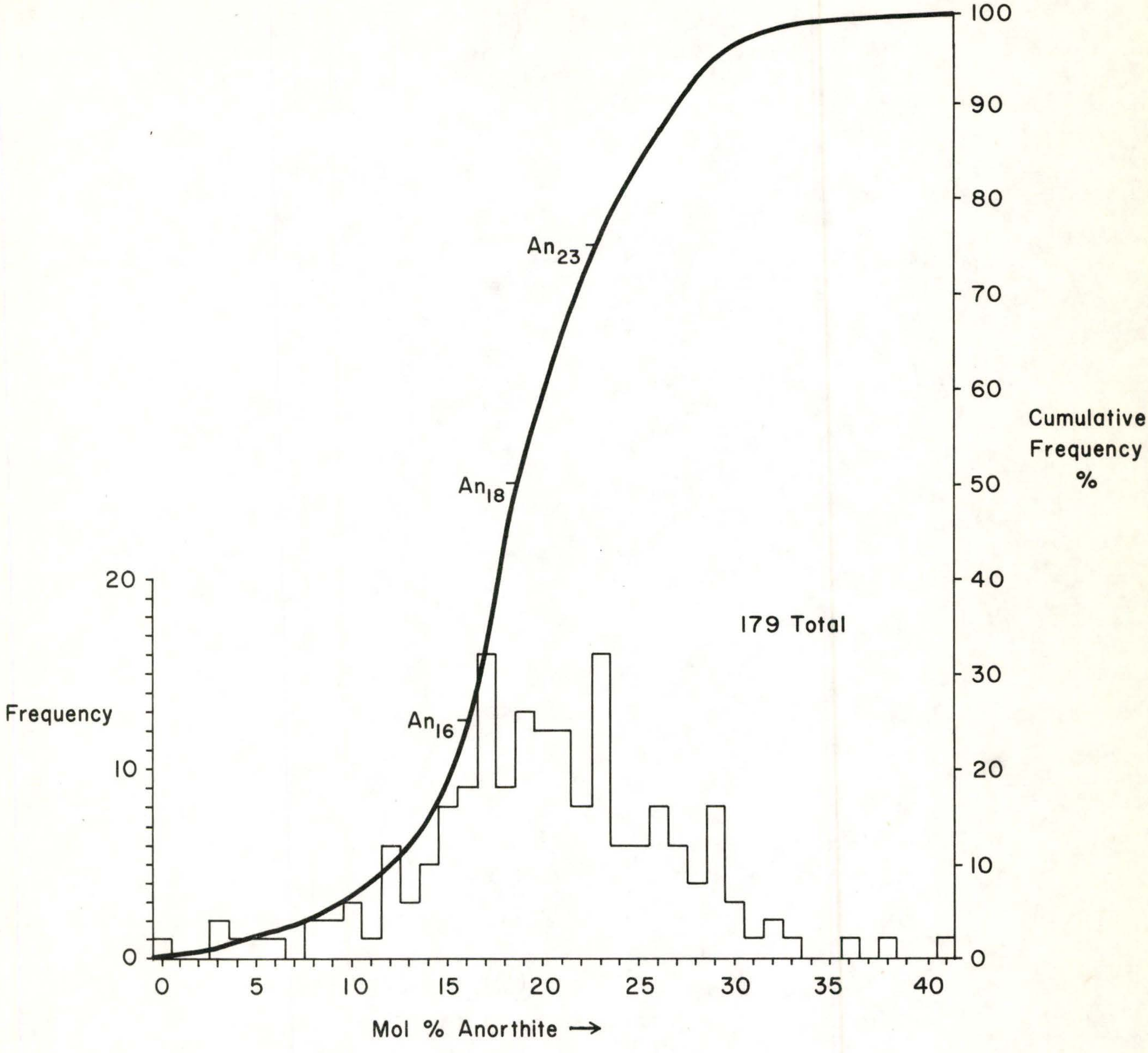
The frequency distribution of all microprobe determinations of plagioclase is presented on Figure 11 as both a histogram plot and a cumulative frequency curve. The median composition is An_{18}, and 50 per cent of the determinations are within the range An_{16} to An_{23}. The total range in composition is An_0 to An_{41}.

Petrogenesis of the Batholith

Phemister (1960, 1961) considered the rocks of the Chief Lake Batholith to be the product of alkali metasomatism of sedimentary rocks (largely quartzite) and gabbro. His interpretation is based entirely on textural and structural observations in the field. Phemister gave the following evidence in favor of a metasomatic origin for these rocks:

1. Porphyroblastic growth of alkali feldspar in Grenville province gneisses, metagabbro and quartzite in the vicinity of the Grenville-Southern province boundary.
2. Parallelism of planar bedding in "unreacted" rafts of quartzite, and foliation in surrounding "pseudogranite".
3. "...it is the problem presented in mapping where one is confronted with the impossibility of isolating clearly igneous material from country rocks that is the most convincing proof to a geologist of the metasomatic origin of the felspar in these rocks." (Phemister, 1960, p. 117).

A metasomatic origin of microcline megacrysts in gabbroic rocks associated with the batholith seems to be unquestionable. However, wholesale transformation by alkali metasomatism of vast amounts of rocks varying in composition from pure quartzite to gabbro to form a rock unit as restricted in composition as the Chief Lake batholith (See Fig. 9.)



FREQUENCY DISTRIBUTION OF MICROPROBE DETERMINATIONS OF PLAGIOCLASE FROM THE CHIEF LAKE BATHOLITH

Figure 11

seems improbable. The parallelism of planar structures in the batholith and enclosed metasediments was more likely caused by homogeneous deformation of both the batholith and inclusions rather than preservation of relic bedding in pseudogranite.

Field relations of the batholith presented in previous sections indicative of a magmatic origin are summarized as follows:

1. Unfaulted and unsheared contacts with the country rocks are sharp and commonly discordant.
2. Dikes and apophyses from the major body of the batholith cross the strike of the enclosing country rocks.
3. Intrusion breccias of granitic material into metagabbro are common.
4. The most common xenoliths in the batholith are composed of refractory material, e.g., pure quartzite and metagabbro. Metasedimentary inclusions with a lower melting temperature, e.g., arkose and graywacke, are conspicuously absent from the batholith.

The restricted modal composition of the batholith in terms of quartz-microcline-plagioclase-rich rocks is suggestive of an igneous origin; whereas the large compositional variation between plagioclase grains within a single thin section implies chemical disequilibrium, which is not proof of either magmatic or metasomatic origin.

The total range of plagioclase determinations from the batholith (An_{0-41}) is large; however, 50 per cent of the data fall within the much more restricted compositional range An_{16-23}. Data shown in a later section suggest that chemical equilibrum (as shown by constant plagioclase composition within a single thin section) has been achieved in the Grenville gneisses.

V. GRENVILLE PROVINCE ROCKS

The complex of metamorphic rocks east of the Chief Lake batholith is assigned to the Grenville structural province (See Fig. 2.). Although no radiometric ages are available from the eastern part of the research area, the rocks are similar in structure and metamorphic grade to rocks about ten miles northeast which were metamorphosed during the Grenville orogeny. Collins (1938) mapped as far east as the southwestern shore of White Oak Lake in Tilton Township; the portion of the map area north and east of White Oak Lake has not been mapped previously.

The Grenville province rocks within the research area are divided into two contrasting lithologies: massive amphibolite and dominantly quartzo-feldspathic gneiss. In general, the Grenville province rocks are much more heterogeneous on the scale of an outcrop than Southern province rocks. Migmatitic phases are common in the gneisses, but pegmatites are rare.

Coarse grained, foliated quartz dioritic gneiss in southeastern Tilton Township (Fig. 2) is probably a metamorphosed portion of the Chief Lake batholith, but which is spatially separate from the main body in the map area.

No stratigraphic succession is implied by the sequence of Grenville province map units either in the text or on Figure 2.

Massive Amphibolite

Massive amphibolite constitutes about 10 per cent of the metamorphic rocks mapped, and is the most homogeneous of the Grenville province lithologic units. Numerous unmapped bodies of amphibolite occur as narrow discontinuous bands within quartzo-feldspathic gneiss. The massive amphibolite between White Oak and Bluff lakes immediately east of the migmatite zone has an unusual texture consisting of ovoid segregations of white plagioclase up to one foot in diameter in hornblende-rich matrix (Fig. 2).

Generally, amphibolite is a massive black rock composed essentially of plagioclase and hornblende. Quartzo-feldspathic bands within amphibolite constitute less than 20 per cent of the total rock unit. Hornblende grains commonly possess a preferred orientation.

Modal analyses of three amphibolite specimens are listed in Table 6. Garnet is absent from the thin sections examined, and is absent in amphibolite seen in the field--although garnet commonly is present in surrounding quartzo-feldspathic gneiss. Optical determinations of plagioclase by the method of maximum extinction angles of albite twins range from An_{33} to An_{53}. A specimen of amphibolite (not listed on Table 6) analysed with the electron microprobe contains plagioclase ranging from An_{61} to An_{84}, with an average value of An_{77}.

Table 6

MODAL ANALYSES OF AMPHIBOLITE

Spec.	Qtz.	Plag.	Hbd.	Cpx.	Bio.	Ap.	Mte.	Total Counts
2092	3.0	46.5 (An_{33})	49.5	-	1.0	-	-	600
1823	9.6	31.8 (An_{47})	31.4	1.2	15.6	3.4	7.0	500
1646	-	38.8 (An_{75})	40.2	20.3	0.7	-	-	600

Quartzo-feldspathic Gneisses

Quartzo-feldspathic gneisses constitute the bulk of the Grenville province metamorphic rocks exposed within the research area. Four types of quartzo-feldspathic gneiss were recognized in the field and are shown on Figure 2. The types are distinguished on the basis of mineral composition, scale of gneissic banding (i.e., laminated or massive) and weathering characteristics.

The gneisses generally are heterogeneous and contacts between units are mostly approximate. Abundant folds and generally poor exposure contribute to the difficulty of tracing contacts between quartzo-feldspathic gneisses.

Biotite-garnet gneiss Migmatitic biotite-garnet gneiss occurs in northeast Tilton Township and along the east shore of White Oak Lake. Characteristically, this unit is a massive quartz dioritic gneiss with biotite and small red garnets dispersed in a white matrix of plagioclase, quartz and microcline. Small bodies of amphibolite, and discontinuous schistose layers rich in biotite and garnet are locally abundant.

Laminated biotite-garnet-sillimanite gneiss Three small bodies of schistose biotite-garnet-sillimanite gneiss occur along the eastern shore of White Oak Lake. The unit is characterized by abundant red garnet porphyroblasts, up to one inch in diameter. The matrix is composed of plagioclase, biotite, sillimanite and quartz. Plagioclase was determined with the electron microprobe; its composition ranges from An_5 to An_7 for sixteen determinations in one specimen, and from An_{29} to An_{32} for twelve determinations in a second.

Rusty-weathering banded biotite gneiss This unit occurs east of White Oak Lake, and east and north of Bluff Lake; it has the appearance of a "bedded feldspathic quartzite". The rock contains abundant quartz and feldspar, with minor biotite and rare dispersed porphyroblasts of hornblende and garnet, although one thin section contains scapolite, hornblende and calcite. The reason for the characteristic rusty weathering is not known.

Microcline-augen gneiss with felsic and mafic laminae Laminated augen gneiss is the most abundant quartzo-feldspathic gneiss in the map area. This unit is typified by alternating hornblende-biotite rich, and quartz-feldspar rich laminae. Hornblende and large microcline porphyroblasts also are typical and abundant. Folded laminae of the augen gneiss are shown on Plates 4A and 4B.

Modal analyses of three thin sections of laminated augen gneiss are given in Table 7. Plagioclase determinations were made with the electron microprobe. The modal analyses are variable, probably due to the characteristic segregation of minerals into felsic and mafic laminae.

Table 7

MODAL ANALYSES OF LAMINATED AUGEN GNEISS

Spec.	Qtz.	Plag.	K-spar	Bio.	Hbd.	Gar.	Ap.	Mte.	Total
1700	7.4	35.2	10.6	46.8	- (in hand spec.)		-	-	500
2208	21.8	13.8 (An_{17})	46.9	4.6	12.1	0.6	-	0.2	1250
W285	-	60.0 (An_{34})	2.5	21.5	15.5	-	0.4	-	1250

VI. DIABASE DIKES

Small dikes of diabase intrude both Southern and Grenville province rocks. The largest body of diabase is at White Oak Lake, and is about 100 yards wide and one mile long. Another basic dike cuts across the contact of the Chief Lake batholith and the vitreous quartzite north of Wavy Lake.

In hand specimen, diabase is dark olive green and of medium grain size. The rock commonly weathers reddish-brown. The red-brown weathering characteristic distinguishes diabase from hornblende gabbro. However, some of the smaller bodies of hornblende gabbro mapped may be diabase dikes without the distinctive weathered appearance.

Two thin sections of diabase from the large dike at White Oak Lake contain essential plagioclase and clinopyroxene, with accessory olivine and magnetite. Red-brown biotite surrounds magnetite, and the few olivine grains have reaction rims of pyroxene. Textures are intergranular with clinopyroxene and olivine contained in a felted matte of labradorite laths. Plagioclase laths are bent and broken in one thin section.

Diabase dikes are not recrystallized and, therefore, appear to post-date the Grenville orogeny. Dikes of similar composition in other parts of the Sudbury area commonly have been classified as Keewenawan or Late Precambrian.

CHAPTER III

METAMORPHIC GEOLOGY

The area is divided into two metamorphic zones as indicated by the regional development of certain metamorphic minerals: a chlorite-biotite zone and a garnet-sillimanite zone. The appearance of garnet porphyroblasts in quartzo-feldspathic rocks, the garnet isograd (Fig. 2), marks the boundary between the two zones. Mineral assemblages in the chlorite-biotite zone are typical of the greenschist facies of regional metamorphism; in the garnet-sillimanite zone, mineral assemblages appear to have been formed under the higher temperature conditions of the almandine-amphibolite facies (Turner and Verhoogen, 1960).

In the following discussion, the rocks are divided into two general categories for convenience of description of metamorphic mineral paragenesis: (1) semi-pelitic and quartzo-feldspathic rocks, and (2) mafic rocks. Except for electron microprobe plagioclase determinations, all data are the result of field observation and thin-section petrography.

The regional developement of minerals in the chlorite-biotite and garnet-sillimanite zones is shown in Table 8. Quartz, plagioclase and biotite are ubiquitous in semi-pelitic and quartzo-feldspathic rocks. Chlorite, muscovite, epidote and actinolite occur only in the chlorite-biotite zone. Hornblende and potassium feldspar in quartzo-feldspathic meta-igneous rocks in the chlorite-biotite zone probably are relic phases. Garnet is rare in the chlorite-biotite zone.

Porphyroblasts of garnet, hornblende and microcline are common in quartzo-feldspathic gneisses of the garnet-sillimanite zone. Sillimanite occurs in one map unit adjacent to the garnet isograd in the southeast portion of the area

ROCK TYPE	CHLORITE–BIOTITE ZONE	GARNET–SILLIMANITE ZONE
Semi-pelitic Meta-sedimentary Rocks and Quartzo-Feldspathic Rocks	—— Quartz ——	————
	—— Plagioclase ——	————
	—— Chlorite ——	
	—— Muscovite ——	
	—— Biotite ——	————
	—— Epidote ——	
	—— Actinolite ——	
	—— relic —— Hornblende ——	————
	— Garnet —	————
		—— Sillimanite ——
	—— relic —— Potassium	Feldspar ——
Mafic Rocks	—— Quartz ——	————
	—— Plagioclase ——	————
	—— Actinolite ——	
	—— relic —— Hornblende ——	————
	—— Epidote ——	
	—— Chlorite ——	
	—— Biotite ——	————
		—— Clinopyroxene ——

MINERAL ZONING

Table 8

(Figure 2).

All mafic rocks contain plagioclase and amphibole; some contain minor quartz and biotite. Amphibole is actinolite or hornblende in the chlorite-biotite zone, but only hornblende in the garnet-sillimanite zone. Chlorite and epidote are found only in the chlorite-biotite zone. Several amphibolites from the garnet-sillimanite zone contain clinopyroxene. Garnet is absent.

I. CHLORITE-BIOTITE ZONE

Rocks west of the garnet isograd are placed in the chlorite-biotite zone because of the regional development of these two minerals in quartzo-feldspathic and semi-pelitic rocks. Protoliths of rocks within the chlorite-biotite zone west of the Chief Lake batholith are easily recognized, and are described in Chapter 2. These rocks are feldspathic quartzite and arkose, with minor graywacke and orthoquartzite of Huronian age. Mafic rocks are metagabbro. Quartzo-feldspathic and migmatitic gneiss along the east border of the batholith and west of the garnet isograd are more intensely deformed and recrystallized than the Huronian rocks to the west.

Semi-pelitic and Quartzo-feldspathic Rocks

Mineral assemblages found in semi-pelitic metagraywacke are:

(1) quartz-plagioclase-muscovite-chlorite
(2) quartz-plagioclase-muscovite-biotite
(3) quartz-plagioclase-muscovite-biotite-chlorite
(4) quartz-plagioclase-muscovite-chlorite-(garnet)
(5) quartz-plagioclase-muscovite-biotite-(garnet)
(6) quartz-plagioclase-muscovite-chlorite-epidote
(7) quartz-plagioclase-biotite-epidote

(8) quartz-plagioclase-muscovite-biotite-actinolite-(garnet)
(9) quartz-plagioclase-chlorite-actinolite
(10) quartz-plagioclase-muscovite-biotite-chlorite-actinolite

The distribution of rocks containing these mineral assemblages are shown on Figure 4 (pocket). Chlorite, biotite and muscovite are widely distributed, but garnet, actinolite and epidote are more restricted. Garnet was not observed in the field within the chlorite-biotite zone, but is present in minor amount in four thin sections.

Rectangular glomeroporphyroblasts (See Plate 1D.) of muscovite and quartz, and muscovite and chlorite, which are found at three localities shown on Figure 4, are interpreted as pseudomorphic replacements of andalusite or staurolite. Staurolite-bearing metapelitic rocks occur about four miles northwest of the research area in Waters Township (Card, 1964).

The assemblages listed for the chlorite-biotite zone are typical of the greenschist facies of regional metamorphism, probably the lowest temperature quartz-albite-muscovite-chlorite subfacies, and the quartz-albite-epidote-biotite subfacies defined by Turner (1958).

Quartzo-feldspathic rocks include feldspathic quartzite, quartz monzonite, quartz diorite and trondhjemite, as well as migmatite and quartzo-feldspathic gneiss. Mineral assemblages are similar to those of the semi-pelitic rocks, but the relative proportions of minerals are not. Quartz, plagioclase and potassium feldspar generally constitute 80 to 90 per cent of the rock. Accessory minerals are chlorite, biotite, muscovite, epidote and hornblende. Micas commonly are aligned parallel to foliation, and are believed to be of metamorphic origin. Epidote occurs both in the groundmass of meta-igneous rocks and as inclusions within plagioclase grains. Microcline and plagioclase commonly are fractured and rimmed by fine grained feldspar (Plate 5). Quartz is recrystallized, but shows undulose extinction.

The anorthite content of plagioclase in the Chief Lake batholith and Eden Lake trondhjemite has been discussed previously. For comparison with analysed plagioclase in metamorphic rocks, the variation in anorthite content of all samples determined with the electron microprobe are shown on Figure 4. Some samples indicate an extremely wide range in plagioclase composition which is due partly to single determinations that are five to ten mol per cent anorthite higher or lower than the rest of the determinations.

The average variability in plagioclase composition for the 15 specimens on Figure 4 west of the garnet isograd is 18 mol per cent anorthite. The maximum range is 32 per cent (no. 37), and the minimum is 5 per cent (no. 43). A comparison of these data with similar studies of unmetamorphosed granitic intrusions would be interesting to evaluate whether the apparent chemical disequilibrium is due to crystallization from a silicate melt or partial re-equilibration of igneous plagioclase under the conditions of the greenschist facies of regional metamorphism.

Mafic Rocks

Mineral assemblages found in metagabbroic rocks and amphibolite in the chlorite-biotite zone are:

(1) plagioclase-actinolite
(2) plagioclase-actinolite-epidote
(3) plagioclase-actinolite-biotite
(4) plagioclase-actinolite-epidote-biotite
(5) plagioclase-actinolite-epidote-chlorite
(6) plagioclase-hornblende
(7) plagioclase-hornblende-biotite
(8) plagioclase-hornblende-biotite-epidote
(9) plagioclase-hornblende-biotite-chlorite-epidote
(10) plagioclase-hornblende-microcline-biotite-epidote
(11) plagioclase-hornblende-chlorite-epidote

Quartz is a possible additional phase in all assemblages. Actinolite and hornblende were not observed in the same thin section. Hornblende is dark green, and actinolite is pale green laths or fibers.

Chlorite and actinolite are rare in metagabbroic xenoliths within the batholith. The common assemblage in basic inclusions is plagioclase-hornblende-biotite. Potassium feldspar phenocrysts and quartz are common in basic inclusions, and probably were formed during metasomatism. Hornblende in metagabbro west of the batholith may be a relic phase because it was observed only in rocks with a gabbroic texture. Foliated mafic rocks contain actinolite.

Plagioclase composition in a basic inclusion within the Chief Lake batholith (number 10 on Fig. 4) varies from An_{34} to An_{46} and averages An_{42}. The composition of plagioclase in a specimen of massive amphibolite located immediately west of the garnet isograd varies from An_{61} to An_{84} (c on Fig. 4) and averages An_{77}.

Mineral assemblages in mafic rocks in the chlorite-biotite zone west of the Chief Lake batholith are typical of the greenschist facies. Hornblende-plagioclase-biotite and hornblende-plagioclase assemblages could have formed under the conditions of the upper greenschist facies or the almandine-amphibolite facies (Turner, 1958).

II. GARNET-SILLIMANITE ZONE

The rocks of the garnet-sillimanite zone are quartzo-feldspathic gneiss and amphibolite. Their protoliths are uncertain, but probably both quartzo-feldspathic paragneiss and orthogneiss are present. The massive amphibolite may have been derived from gabbroic rocks.

Quartzo-feldspathic rocks

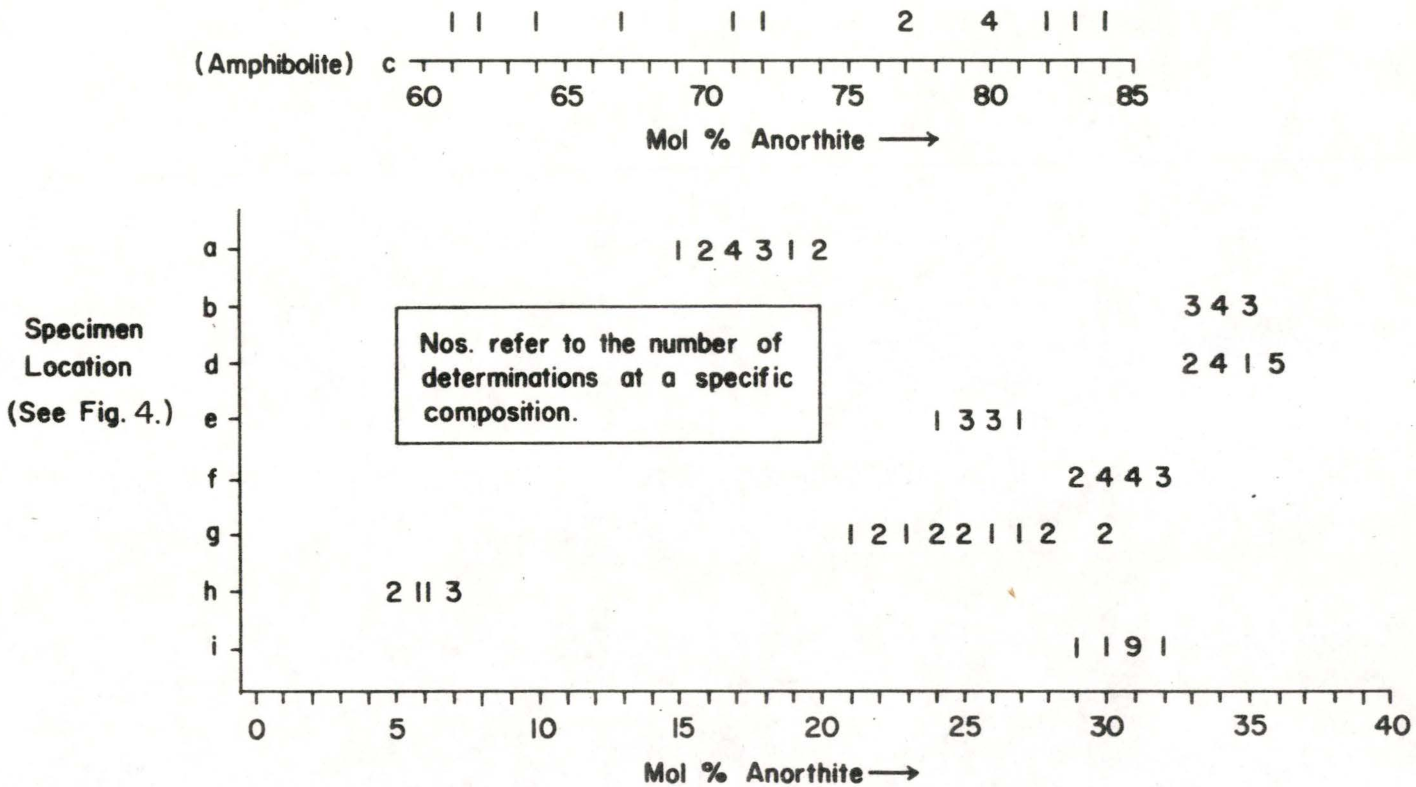
Observed mineral assemblages in quartzo-feldspathic rocks in the garnet-sillimanite zone are:

(1) quartz-plagioclase-potassium feldspar-biotite
(2) quartz-plagioclase-potassium feldspar-biotite-garnet
(3) quartz-plagioclase-potassium feldspar-biotite garnet-sillimanite
(4) quartz-plagioclase-potassium feldspar-biotite-garnet-hornblende
(5) quartz-plagioclase-potassium feldspar-biotite-hornblende

Quartz, plagioclase, potassium feldspar and biotite are ubiquitous. Potassium feldspar (commonly microcline), garnet and hornblende occur as porphyroblasts. Sillimanite was observed only in thin sections from biotite-garnet-rich laminated gneiss adjacent to the garnet isograd in the southeast part of the area (Fig. 4).

The mineral assemblages are typical of the almandine-amphibolite facies of regional metamorphism, probably the sillimanite-almandine subfacies (Turner, 1958).

Compositions of plagioclase in eight specimens of quartzo-feldspathic gneiss from the garnet-sillimanite zone were determined with the electron microprobe. The plagioclase determinations are plotted on Figure 12, and the specimen locations are shown on Figure 4. In contrast to plagioclase determinations from the batholith, the range of plagioclase composition in specimens of the higher grade rocks varies from 3 to 10 mol per cent anorthite--excepting the amphibolite specimen (c on Fig. 4) which is from west of the garnet isograd in the chlorite-biotite zone. The variation in anorthite content determined in specimens a to i are shown on the figure. The range in anorthite content in the samples of quartzo-feldspathic gneiss is from An_5 to An_{37}.



VARIATION OF MICROPROBE DETERMINATIONS OF PLAGIOCLASE FROM QUARTZO-FELDSPATHIC GNEISS

Figure 12

Mafic Rocks

The mafic rocks in the garnet-sillimanite zone are massive amphibolite. Observed mineral assemblages are:

(1) plagioclase-hornblende-(quartz)

(2) plagioclase-hornblende-biotite-(quartz)

(3) plagioclase-hornblende-clinopyroxene-biotite

Clinopyroxene is colorless and is probably diopside. No plagioclase determinations were made with the microprobe. Pyroxene-bearing amphibolite occuring between the garnet isograd and sillimanite-bearing gneiss contains plagioclase of composition An_{75} (optical determination by method of maximum extinction on albite twins). Garnet was not observed in amphibolite in the area.

Pyroxene-bearing amphibolite is typical of the almandine-amphibolite facies (Turner, 1958).

III. PETROLOGIC INTERPRETATION

Observed mineral assemblages in rocks in the chlorite-biotite zone generally are typical of the greenschist facies of regional metamorphism. Correlation of the rocks with the quartz-epidote-biotite subfacies is more speculative because of the lack of chemical analyses of minerals and rocks.

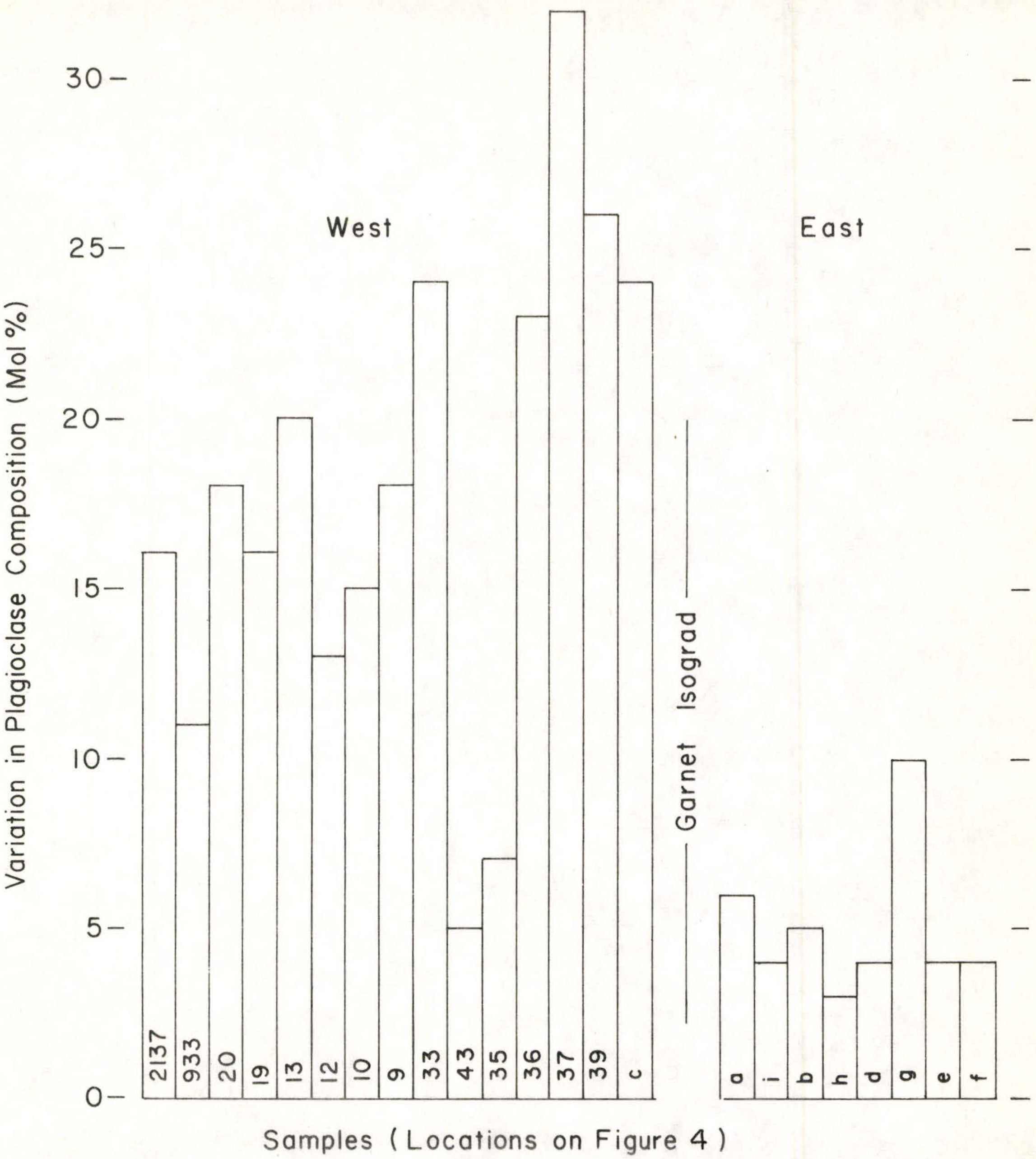
The rocks east of the garnet isograd contain mineral assemblages typical of the almandine-amphibolite facies. The absence of muscovite and the presence of microcline porphyroblasts and sillimanite suggest that the rocks belong to the sillimanite-almandine-orthoclase subfacies.

Regional zoning of minerals (Table 8) indicates that the garnet isograd represents a metamorphic discontinuity. For example, within one-half mile on either side of the isograd chlorite, muscovite, epidote and actinolite disappear as mineral phases in the rocks, and sillimanite, garnet, clinopyroxene and microcline (porphyroblastic) appear.

The microprobe data do not directly reflect the metamorphic grade of the rocks, but the much narrower range in plagioclase composition in the gneisses east of the garnet isograd suggests that chemical equilibrium of plagioclase was attained during metamorphism. The ranges in plagioclase composition for the specimens shown on Figure 4 are plotted on Figure 13. The rocks are not on a straight traverse across the area, so it is difficult to place the data on a linear scale. However, the specimens from east of the isograd are so much lower in plagioclase compositional variation than those to the west that the data strongly suggest a discontinuity of some sort rather than a transition near the garnet isograd.

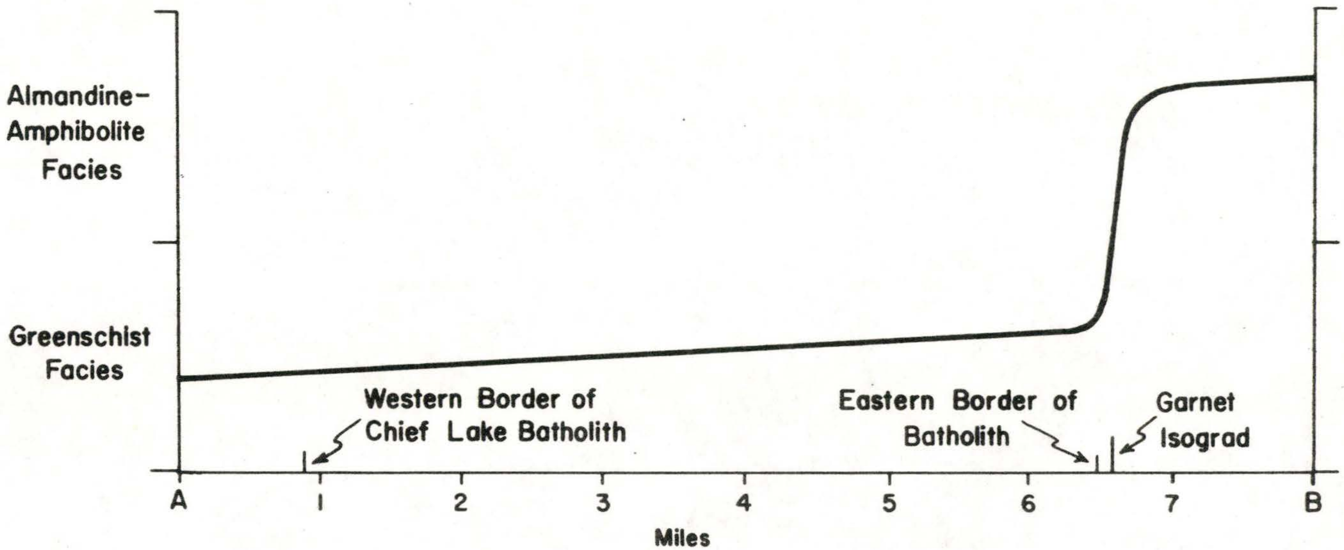
Turner and Verhoogen (1960) suggested possible ranges in temperature and pressure for the greenschist facies of 300-500°C. and 3,000 to 8,000 bars P_{H_2O}. For the almandine-amphibolite facies they suggested that the upper limit may be near 700° to 750° C. and about 8,000 bars pressure.

Figure 14 is a schematic interpretation of the metamorphic profile across the area along A-B on Figure 4. The metamorphic intensity rises gradually across the Huronian metasedimentary rocks and the Chief Lake batholith; the gradient steepens considerably less than a mile to the west of the garnet isograd, and passes through the upper part of the greenschist facies and the lower part of the almandine-amphibolite facies. The proximity of sillimanite-clinopyroxene-bearing rocks to the garnet isograd indicates that the isograd lies on a very steep metamorphic gradient. The gradient probably decreases east of the garnet isograd, rather than continuing to rise into the granulite facies, because biotite and hornblende are present in the gneiss at least six miles east of the isograd.



VARIATION IN PLAGIOCLASE COMPOSITION WITHIN SAMPLES FROM WEST AND EAST OF THE GARNET ISOGRAD

Figure 13



INTERPRETIVE METAMORPHIC PROFILE
ALONG LINE A-B ON FIGURE 4

Figure 14

CHAPTER IV

STRUCTURAL ANALYSIS

The structure of the Huronian metasedimentary rocks is characterized by large-scale folds in bedding which can be observed only indirectly by measurement of facing reversals in cross-bedded quartzite units. The southern and eastern portion of the Chief Lake batholith, and the Grenville Province gneisses, however, are characterized by a penetrative deformation which has produced a tectonite whose dominant structures are a foliation and lineation. This foliated and lineated structural fabric has been superimposed on the large-scale structures of the Huronian metasedimentary rocks southwest of the batholith.

The major objectives of structural analysis are: (1) to describe the form, orientation and mutual relationships of various structural fabric elements, and (2) to interpret the fabric data in terms of more theoretical considerations such as possible principal strain and stress directions and deformation mechanisms responsible for the development of fabric. The descriptive phase of structural analysis is referred to as geometric analysis; correlation of observed geometries of fabric elements with inferred stress trajectories and deformation mechanisms constitutes kinematic and dynamic analysis of structures.

The total fabric of a deformed rock includes both active fabric elements formed in response to a particular stress field acting upon the rock (e.g., preferred crystallographic orientation of quartz in a quartzite), and passive fabric elements inherited from a previous period in the history of the rock body (e.g., the line of intersection of normal and cross-beds in a folded sandstone).

I. STRUCTURAL GEOMETRY

Proceedure and Terminology

For convenience, the following three scales are used in the description of structural data:

(1) Structures observable in thin section are referred to the microscopic scale.

(2) Structures directly observable in hand specimen or a single outcrop are referred to the mesoscopic scale.

(3) Structures indirectly observable on a scale larger than a single outcrop are referred to the macroscopic scale.

The lower hemisphere equal-area (Schmidt) projection is used to illustrate the three-dimensional orientation of structural fabric elements. The orientation of a plane can be represented on the equal-area net as a great circle (meridian) trace or as the impingment point of the line drawn normal to the plane (i.e., the pole to the plane). The orientation of a line is plotted as a point on the petrofabric diagram. In this study, diagrams containing more than fifty points generally are contoured according to the percentage of points per unit area of the diagram.

Orientations of the following structural fabric elements were recorded in the field:

(1) Bedding (S_b) was measured in the Huronian metasedimentary rocks. The orientations of crossbeds were not recorded.

(2) Gneissic banding (S_g) was measured in the Grenville Province gneisses east of the Chief Lake batholith. In some places, S_g may represent original sedimentary bedding, but in general the origin of S_g is not clear.

(3) Foliation (S_1) was recorded throughout the map area as the plane of preferred mineral orientation. In general, S_1 is parallel to either S_b (in metasedimentary rocks) or S_g (in gneissic rocks) except in fold hinges.

(4) Axial planes of mesoscopic folds in S_b and S_g were measured, and were found to be mesoscopically parallel to foliation, S_1.

(5) Mineral lineation (L_1) was measured in the southern and eastern parts of the area (where it is developed). L_1 is defined by elongate aggregates of quartz (rodding) and feldspar. In rocks containing a foliation, S_1, mineral lineations lie on the foliation plane. Some rocks are lineated, but not foliated.

(6) Fold axes (L) were measured in the metasedimentary rocks and gneisses.

Several faults and shear zones were mapped. Exposed fault zones contain brecciated rocks, whereas rocks within shear zones are mylonitic.

On Figure 5 (pocket), the map area is divided into nine structural subareas based on lithology and internal structural homogeneity. The lithologic subdivision are:

(1) rocks west of the Chief Lake batholith (areas 1, 2, 3, 4, 5, 7,)

(2) the main body of the batholith (areas 6, 8)

(3) rocks east of the batholith (area 9).

Within the three lithologic divisions, the smaller areas are outlined on the basis of apparent structural homogeneity. For example, in area 7 the foliation, S_1, has a dominant east-west strike, whereas S_1 in area 4 generally strikes northwest. In areas 1 and 2, however, S_1 is subparallel, but the two regions are separated by a shear zone.

Geometry of areas 1 and 2

The contact between the Chief Lake batholith and the rocks to the west is sheared along the border with area 1 and 2 except for a short segment east of Clearwater Lake in southwestern Broder Township (See Fig. 2 .). The boundary between areas 1 and 2 is a shear zone which branches from the the shear zone parallel to the western border of the batholith. The southern boundary of area 2 is marked by the extension of the batholith border shear zone westward through the Huronian rocks. (The shear zones and their associated structures will be discussed in another section.) The northern and western boundaries of area 1, and the western boundary of area 2 are marked by the limits of mapping. Within these two structural divisions, the rocks appear to be the least recrystallized.

Foliation, S_1, generally is developed in the more pelitic beds, and is absent or is present as fracture-type cleavage in coarser grained quartzite beds. Rarely is foliation developed in metagabbro in areas 1 and 2, but the best developed fracture direction strikes subparallel to the conformable contacts with the country rocks. Mesoscopic folds in quartzite are rare, and, therefore, few L lineations were measured. The observed mesoscopic folds are intrafolial folds in thin quartzite beds within more pelitic sequences. Mineral lineations, L_1, are rare in areas 1 and 2, but are common in most of the other subareas. Where observed, L_1 is formed by elongate quartz grains in the plane of foliation.

Composite equal-area diagrams of foliation and mineral lineations, and bedding and fold axes for areas 1 and 2 are shown on Figure 5. The contoured plots of S_1 poles for both areas indicate a dominant northeast strike and steep southeast dip. Mineral lineations plunge southward at more than 45 degrees. Contoured plots of bedding poles show a dominant northeast strike and southeast dip, and fold axes generally plunge southward at more than 45 degrees.

Geometry of areas 4, 5 and 6

Areas 4 and 5 are south of the west extension of the shear zone which parallels the western border of the batholith (Fig. 5). Area 6 lies within the batholith east of areas 4 and 5. The common boundary of areas 4 and 5 with area 6 is an irregular, unfaulted contact between the country rocks and the batholith. The boundary between areas 4 and 5 is the contact between the upper graywacke-quartzite and the vitreous quartzite. This contact is conformable and the boundary is drawn for convenience of structural description. The boundaries between areas 3 and 4, and areas 4 and 7 are defined by changes in the dominant trend of S_1.

The rocks in areas 4 and 5 are more recrystallized than rocks of similar composition in areas 1 and 2, but primary sedimentary structures are common. Both normal and cross-beds are abundant and are well defined in feldspathic quartzite in area 4. Feldspathic quartzite beds are abundant in the upper graywacke-quartzite in area 4 and provide good structural control on the trend of S. Bedding generally is poorly defined in the massive vitreous quartzite in area 5, and in the quartzite xenoliths in area 6, but S can be defined in places by thin concentrations of pale green muscovite flakes.

Foliation is found in both quartzite and metagraywacke in areas 4 and 5, and is defined by planar orientation of lenticular quartz grains parallel to which quartzite develops finely spaced cracks. Quartzite, especially the vitreous quartzite, contains several other more widely spaced fracture directions which are mistaken easily for S_1. One direction of fracturing is commonly more intensely developed, generally normal to S_1.

Although most of the rocks in areas 4 and 5 are foliated, mesoscopic folds are rare, and S and S_1 generally are planar and nearly parallel within a single exposure. Changes in trend of bedding between exposures are followed by similar changes in foliation trend.

The trend of both bedding and foliation in area 4 is predominantly northwest (Fig. 5), whereas foliation in area 5 has both northwest and northeast trending maxima. On the macroscopic scale, two foliations appear to exist in area 5, one with a dominant northeast strike and another with a dominant northwest strike. However, only one foliation was recognized in the field.

The Chief Lake batholith in area 6 contains abundant xenoliths of vitreous quartzite and metagabbro. Foliation in the batholith is defined by planar orientation of microcline porphyroclasts and narrow lenses of quartz. The longer dimensions of small xenoliths and foliation within xenoliths parallel S_1 in subjacent batholithic rocks (Plate 2C).

Mineral lineations are more common in areas 4, 5 and 6 than in areas 1 and 2, and rodding of quartz on the plane of foliation is visible in all rock types except metagabbro. L_1 is best developed in vitreous quartzite (Plate 5D) and in quartzite xenoliths, but quartz and feldspar lineations are found also in the batholithic rocks. However, L_1 and S_1 are not ubiquitous. Some rocks, especially within the batholith, contain no visible planar or linear structures.

On Figure 4, S_1 and L_1 in area 6 are plotted on separate diagrams for the batholith and the quartzite xenoliths. Both diagrams show a dominant east-west strike and steep south dip for S_1. Mineral lineations plunge steeply southward in the batholith and inclusions. Bedding in the xenoliths is vertical and generally strikes northwest. Fold axes in the xenoliths are too rare to indicate any significant trend, although all plunge steeply.

Generally, bedding and foliation are more variable in orientation in areas 4, 5, and 6 than in areas 1 and 2. The orientation patterns shown by S_b and S_1 poles in these three subareas do not define a strong point maximum (as does S_1 in area 1) or a complete dispersion along a great circle of the equal-area net. L and L_1 generally plunge steeply southward in areas 4, 5, and 6.

Geometry of Area 3

Area 3 includes most of the mapped portion of the Eden Lake complex. The southeastern boundary is the western limit of typically well foliated rocks of area 7. The boundary between areas 3 and 4 is through a region of poor exposure where the dominant northwest strike of S_b and S_1 in area 4 changes to a northeast strike in area 3.

The structural geometry of area 3 is unlike that of any other portion of the map area. The igneous rocks of the Eden Lake complex rarely are foliated or lineated. Abundant xenoliths of metasedimentary rocks in the trondhjemite generally lack a preferred strike direction--even within a small area. Mesoscopic folds are almost non-existent in both the xenoliths and in the country rocks north of the complex. Northwest striking mineral layering is present in the basic portion of the complex. On the macroscopic scale (Fig. 5), foliation and bedding are dominantly vertical, but no dominant strike direction exists; the poles of S_1 and S_b define complete girdles about the perimeter of the petrofabric diagrams. Linear structures in area 3 are insignificant.

Geometry of areas 7, 8 and 9

Areas 7, 8 and 9 comprise more than 50 per cent of the research area and are characterized by tectonites containing a foliation, S_1, and a mineral lineation, L_1. The boundary between areas 7 and 8 parallels the gradational contact of migmatitic and non-migmatitic rocks of the Chief Lake batholith. Area 8 comprises the foliated and lineated portion of the batholith. The large area of the batholith northwest of area 8 generally does not contain any penetrative structures. Area 9 includes the migmatite along the southwest border of the batholith as well as the gneissic rocks assigned to the Grenville province.

L_1 and S_1 are the dominant structures in area 7.

Mineral lineations and foliation are marked by flattened and elongated quartz and feldspar grains. Both normal and cross-bedding were observed in the middle feldspathic quartzite and in feldspathic quartzite xenoliths in the migmatite. Mesoscopic folds in bedding are rare, and S_1 and L_1 are developed in all components of the migmatite.

S_b and S_1 in area 7 have dominant east-west trends and steep dips to the south (Fig. 5). L_1 generally plunges south at about 60 degrees. Only 13 fold axes were measured, and they show considerable dispersion.

In area 8, the Chief Lake batholith is foliated and lineated. Generally, steeply dipping foliation and down-dip lineation occur together, but S_1 and L_1 may occur separately. The texture of the batholith in area 8 commonly is blasto-mylonitic with microcline porphyroclasts surrounded by a rim of finer grained plagioclase and microcline; quartz and biotite form a foliated and lineated matrix (Plates 3C, 3D).

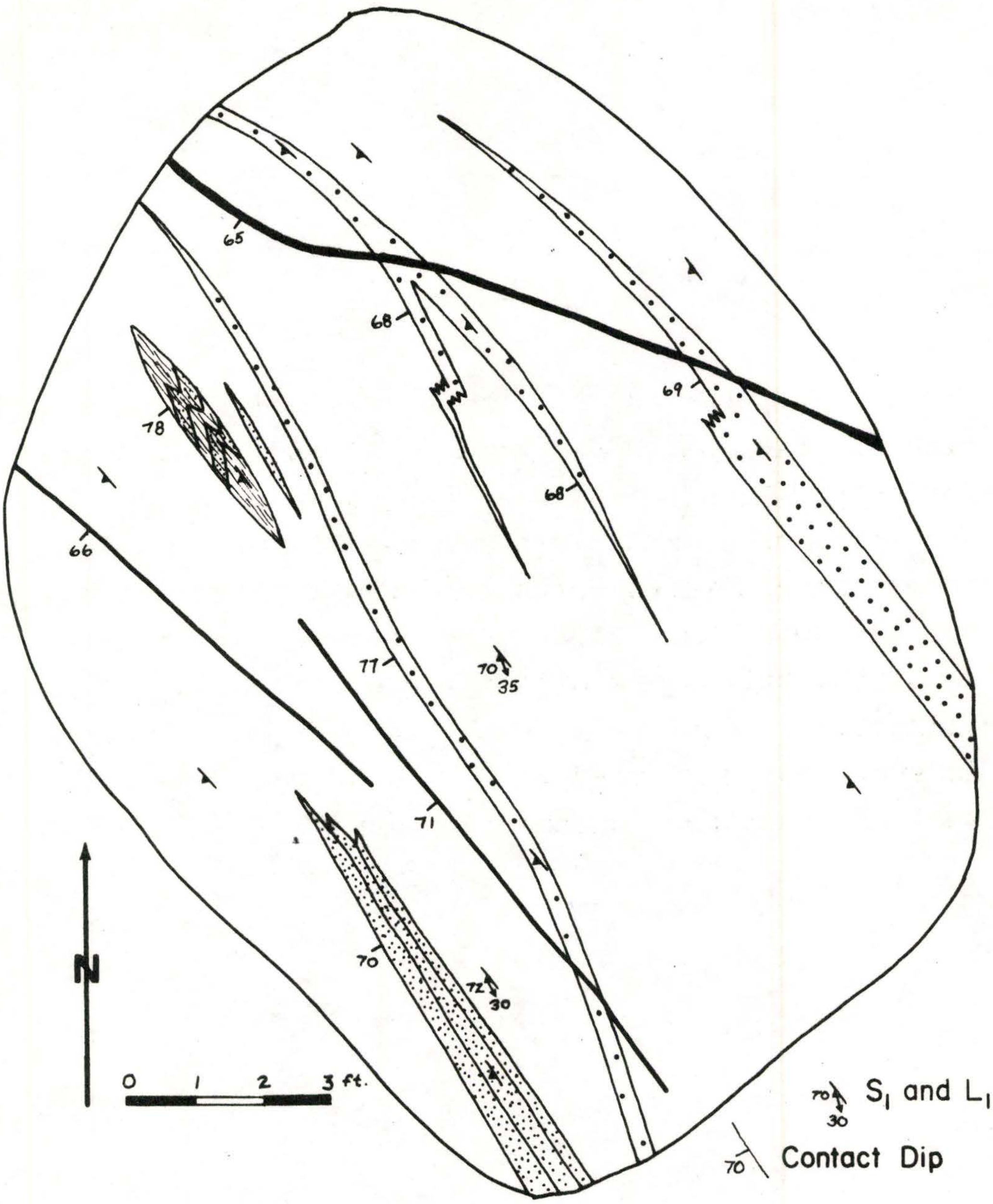
In the northern part of area 8, numerous foliated and lineated xenoliths and dikes of coarse grained quartz diorite occur in the batholith. Whereas the xenoliths generally are aligned parallel to S_1, the dikes commonly strike obliquely to the trend of foliation. The relationship between S_1 and quartz diorite dikes in the batholith is sketched on Figure 15. Although the dikes and xenoliths predate the formation of S_1, their planar form has been maintained.

The macroscopic geometries of S_1 and L_1 in area 8 are shown on Fig. 5. S_1 has a dominant northeast strike and steep southeast dip; L_1 generally plunges down the dip of S_1.

The structures in area 9 differ from those in areas 7 and 8 in that mesoscopic folds are common. The style of the folds in S_g is similar, and the limbs are symmetrically inclined with respect to the axial surface, although one limb commonly is the longer (Plate 4). Axial planes parallel

Figure 15

Sketch of an outcrop of the Chief Lake batholith along Brodill Lake road east of Linton Lake, Broder Township. Unpatterned area is quartz monzonite, large dots are foliated pegmatitic dikes, small dots and dashes are metasedimentary xenoliths, and black lines are quartz veins. Note that folds are preserved in one metasedimentary xenolith, but the xenoliths and dikes, though foliated (lineation is not well developed), are not folded. Folds in surfaces that pre-date S_1 and L_1 are rare in the region west of structural subarea 9.



SKETCH OF FOLIATED DIKES AND XENOLITHS IN THE BATHOLITH

Figure 15

S_1 and fold axes parallel L_1 (Plates 5A, 5B). Foliation is marked by planar orientation of micas and preferred orientation of microcline augen. Mineral lineation (quartz rods, microcline augen and hornblende prisms) is common, and is the dominant structure in some parts of area 9 (Plate 5D).

The macroscopic geometry of structural fabric elements in area 9 are shown on Figure 5. Petrofabric diagrams of S_1 and L_1, and S_g and fold axes are identical. S_1 and S_g poles are distributed parallel to great circles normal to the densest concentrations of L_1 and L.

Shear Zones

Three shear zones are shown on the geologic map (Fig. 2): (1) along the northern boundary of the agmatite, (2) along the southern boundary of the agmatite and the western border of the inclusion mass at Chief Lake, and (3) parallel to the western border of the Chief Lake batholith. None are well exposed, but they generally are marked by linear topographic depressions.

The shear zones are characterized by a zone of variable width containing fine grained, foliated blastomylonites, commonly with mineral lineations on the foliation surfaces (Plates 3A, 3B). The rocks within the shear zones generally are foliated much more intensely than surrounding rocks, although in the northeast corner of the map area (Fig. 5, area 8) shear zones merge into a wide area of penetrative foliation and lineation.

Faults

Two east-west striking, apparently vertical faults displace northwest-trending Huronian rocks north of Wavy Lake (Fig. 2). The apparent displacement on both faults is right-lateral. Exposed rocks along the fault traces are brecciated.

Regional Trends of S_1 and L_1

Figure 6 (pocket) shows the regional variation of foliation and mineral lineation trends. S_1 is drawn as continuous traces through areas of subparallel foliation surfaces on the figure. Where S_1 and L_1 occur together, the horizontal projection of the mineral lineation is connected to the S_1 trace. Some mapped areas do not contain any penetrative structures (e.g., part of the batholith).

West of the Chief Lake batholith (areas 1 and 2) mineral lineations are rare, and foliation strikes northeast and dips steeply southeast. Foliation in the Huronian metasedimentary rocks and the Eden Lake complex south of the batholith border shear zone (areas 3, 4, 5, 7) trends in various directions, but, except for the quartzite in area 5, changes in trend of S_1 are progressive and can be accounted for by regional (rather than local) variability.

S_1 forms an arc in the quartzite around the northern border of the Eden Lake complex, and cuts across the eastern border, but does not penetrate the entire body. The foliation trends within the complex show locally variable strike directions, but dip steeply.

The Chief Lake batholith generally contains northeast trending foliation with a southeast plunging lineation. Both foliation and lineation tend to maintain constant orientation over large areas, but in the southwest portion of the batholith S_1 progressively assumes a more westerly trend and L_1 plunges southward. Foliation in the Grenville province gneisses east of the migmatite zone is folded on a macroscopic scale about axes parallel to the regional trends of mineral lineations and axes of mesoscopic folds in S_g.

Regional Bedding Trends

Macroscopic folds in bedding were recognized by observation of reversals in facing direction of subparallel bedding surfaces, and because mesoscopic folds are rare bedding trends can be projected parallel to strike with certainty. Figure 7 (pocket) shows bedding trend lines and facing directions for the Huronian metasedimentary rocks in the western part of the research area and in the southern half of Waters Township. The trend lines of bedding in Waters Township are derived from Card's (1964) map.

Figure 7 is drawn in a manner similar to that for S_1 trends (Fig. 6), but with the additional information of facing directions of beds. Macroscopic fold axes were determined on the stereographic projection by the orientation of the line of common intersection of bedding traces from opposite limbs of a particular fold. The horizontal traces of axial surfaces are loci of points of maximum curvature between bedding traces. It is assumed that sufficient data exist regarding strike, dip and facing direction of beds to trace a folded horizon within the lithologically homogeneous sequence. The trend lines may locally transgress bedding, but the tendency of a particular fold to appear at several stratigraphic horizons indicates that the bedding geometry shown on the figure is generally correct.

The form of the bedding trend lines on Figure 7 shows that the macroscopic folds are large-scale plications in the trough of a northeast-trending regional synclinorium in the Huronian sedimentary sequence. (This synclinorium can be traced for more than 30 miles to the northeast as indicated by Figure 3.) Axial surfaces of macroscopic folds are nearly vertical throughout, but appear to dip steeply northwest in the northwestern part of Figure 7, and dip steeply southeast less than two miles from the batholith border. Fold plunges vary from 10 to 35 degrees in the region northwest of the batholith. South of the batholith border shear zone, in the region between the trondhjemite and the Chief Lake batholith, the axial surface of a large, partially overturned anticline strikes northwest, and the fold axis plunges south at 70 degrees. Northeast of the axial surface trace of this anticline, beds are nearly vertical and face consistently toward the border of the batholith. North of the batholith border shear, however, bedding trends "concertina"-like northwestward across Waters Township, and the sedimentary sequence as a whole faces southwest, normal to the plunge of the macroscopic folds. The stratiform, folded shape of several of the large gabbroic bodies suggests that they were intruded before the folding.

Discussion of Macroscopic Geometry

Areas 7, 8, and 9 comprise the portion of the area in which mineral lineation with foliation constitutes the dominant penetrative structure. Linear structures are marked by a variety of minerals in rocks of different metamorphic grade distributed over a wide area. The most intense development of L_1 is east of the western boundaries of structural subareas 7 and 8 (Fig. 5).

Steeply plunging mineral lineations are typical of rocks deformed during the Grenville orogeny; mineral foliation, however, may be of several ages. Although cross-cutting foliations were not observed in the field, diverse trends of S_1 on the macroscopic scale, especially in area 5 (Figs. 5, 6) suggest that more than one generation of foliation surfaces is present. Mineral lineations on foliation surfaces within sheared zones west of the region of penetrative L_1 (Fig. 6) indicate that shearing was concurrent with the regional development of L_1 further east.

Macroscopic folds in the Huronian metasedimentary sequence northwest of the Chief Lake batholith probably were formed during the Hudsonian orogeny and modified in shape and orientation during the Grenville orogeny. Evidence for this statement is largely based on the difference in orientation of macroscopic and mesoscopic fold axes. Macroscopic folds in the Huronian rocks in Waters Township (Fig. 7) have northeast striking, nearly vertical axial surfaces and shallow, southwest plunging axes, whereas mesoscopic folds associated with mineral lineations in the research area plunge steeply south or southeast parallel to L_1. The manner in which the older folds were modified will be discussed in a later section on kinematic analysis.

II. QUARTZ MICROFABRIC GEOMETRY

Orientations of quartz c-axes were determined with a universal stage for 21 specimens of quartzite and quartzo-feldspathic rocks to determine the geometric relationship between mesoscopic structures and quartz crystallographic fabrics. Lower hemisphere equal-area plots of quartz c-axis orientations are shown on Figure 5. Series A to D are petrofabric diagrams of quartzitic rocks; series P shows the orientations of quartz c_v in four plutonic rocks, three from the Chief Lake batholith and one (P4) from the Eden Lake trondhjemite. Within a particular series (e.g., B1 to B7) the diagrams are arranged in a sequence of increasing preferred orientation of quartz c-axes.

Diagrams A2, A3, B1, B3, B4 and B5 are oriented with respect to geographic coordinates. The primitive circle of these diagrams is the horizontal plane, and the small arrow at the top of the circle indicates north. The geographic orientation of the other diagrams is not known, but the thin sections generally were cut normal to the foliation in the hand specimen. The location of each specimen is indicated on the structural subarea outline map on Figure 5.

Microfabrics of Quartzitic Rocks

Quartzitic rocks contain more than 70 per cent quartz with accessory feldspar and mica. The shape of grains and the nature of quartz-quartz boundaries vary systematically. Generally, quartz in areas 1 to 7 (Fig. 5) is highly strained, showing undulose extinction in polarized light, and grain boundaries are irregular and sutured. Grain shapes vary from irregular with no preferred orientation to flattened and extremely elongate parallel to S_1 and L_1. In contrast, quartz in area 9 is generally strain-free, and shows mosaic texture with straight, unsutured boundaries and little preferred shape orientation. Foliation and lineation in the gneisses of area 9 commonly are defined by planar and linear alignment of biotite and feldspar rather than quartz morphology.

The petrofabric diagrams of quartzitic rocks shown on Figure 4 are divided into four series based upon the morphology of quartz grains.

Series A: Quartz grains show undulose extinction, but have little or no preferred shape orientation. Quartz-quartz boundaries are irregular and sutured. (See Plates 7A, 7B.)

Series B: Quartz grains are flattened in the plane of S_1 and elongate parallel to L_1; all grains show undulose extinction. Quartz-quartz boundaries are irregular and sutured. (See Plates 7C, 7D.)

Series C: Quartz grains generally are strain-free (no undulose extinction), but are slightly flattened and elongate. Many grains have polygonal outlines and the texture approaches mosaic. (See Plates 8A, 8B.)

Series D: Quartz grains exhibit no preferred shape orientation, and are strain-free. The texture is a mosaic of polygonal grains. (See Plate 8C.)

<u>Series A</u> Specimen 1091 (diagram A1) is the least deformed rock collected; it is from an outcrop of middle feldspathic quartzite which contains no penetrative foliation or lineation. Plate 7A is a photomicrograph showing the sutured, strained quartz grains in this specimen. Specimens OS5 and OS3 (diagrams A2 and A3) are from a vitreous quartzite xenolith in area 6, and the vitreous quartzite in area 5, respectively. Foliation in specimen OS5 is poorly defined in the thin section, and

quartz grains appear to be aligned parallel to more than one direction in the slide. In OS3, the easterly trending foliation is defined by dispersed muscovite flakes aligned in the same general trend.

Quartz c_v have a preferred (i.e., non-random) orientation, but the scatter of points is large. The geometric relationships between quartz c_v, S_b, S_1 and the line of intersection of S_b and S_1 in diagram A2 are not symmetrical. Quartz c-axes on diagram A3 are scattered along a northeast trending great circle normal to the trend of L_1 and normal to the plane of S_1. The other specimens in series A contain no penetrative structures to which the quartz c_v can be related.

Specimen 1950 (diagram A4) is from an inclusion of vitreous-type quartzite in the gabbroic portion of the agmatite along the northern border of the batholith. No mesoscopic structures are visible in the hand specimen. Plate 7B shows the typical sutured boundaries between quartz grains in this rock.

Series B In series B, quartz c_v characteristically lie in the plane normal to L_1. This relationship is shown best by specimen 104 in diagram B7n (Section cut parallel to L_1) and diagram B7p (cut normal to L_1). Plate 7C, shows the linear form of quartz grains in specimen 104. Quartz grains in specimen 104 have no preferred shape orientation in the plane normal to the rods. Mica flakes also are oriented randomly in the plane normal to L_1, but are "folded" about axes parallel to L_1. Specimen 104 is from a vitreous quartzite xenolith in the Chief Lake batholith along the north shore of Wavy Lake.

In series B, all specimens except OS4 and 1736 (diagrams B1 and B2) contain mineral lineations. The quartz c_v fabric shown in diagram B1 forms a diffuse partial girdle along a great circle normal to the line of intersection of S_b and S_1, indicating a relationship similar to that of quartz c_v and L_1 in diagram A3. The 10 per cent maximum concentration of quartz c_v in diagram B1 is inclined obliquely to both S_1 and S_b.

Specimen OS4 is an oriented sample of vitreous quartzite from area 5. Specimen 1736, a quartzo-feldspathic gneiss, is the only rock in series B from area 9. Foliation is defined by biotite and tabular grains of quartz. The geometry of quartz c_v in specimen 1736 is a diffuse concentration obliquely inclined to the plane of foliation.

Specimens 1003, 1037 and 1012 (diagrams B3, B4 and B5) are oriented samples of feldspathic quartzite. L_1 is subparallel in the three specimens, but S_1 strikes northwest in diagram B3 and northeast in diagrams B4 and B5. Quartz c_v lie along a northeast trending great circle normal to L_1 in all three diagrams, whereas S_1 bears no consistent geometrical relationship to the preferred orientation of quartz c-axes. Specimens 1003 and 1012 are from structural subareas 4 and 7 in which L_1 is regionally developed. However, specimen 1037 (diagram B4) is from a portion of area 2 where L_1 is locally developed in a zone parallel to the sheared contact of the Chief Lake batholith.

Specimen 2126 (diagram B6) is vitreous-type quartzite from a xenolith in the foliated and lineated portion of the Chief Lake batholith in area 8. Foliation and lineation are defined in the hand specimen by flattened rods of quartz, and in thin section (Plate 7D) by lensoid aggregates of quartz grains with similar crystallographic orientations. The two quartz c_v maxima on diagram B6 lie on a great circle normal to the trend of L_1, and are symmetrical with respect to S_1.

Series C Quartz crystallographic fabrics of the rocks in series C, with the exception of specimen 1930 (diagram C1), indicate a tendency towards a single maximum concentration dispersed in a girdle normal to L_1, and slightly oblique to S_1. Specimen 1930 is feldspathic quartzite from the shear zone which parallels the agmatite border north of the Chief Lake batholith, the three remaining specimens in series C are quartzo-feldspathic gneisses from structural subarea 9.

Specimen 1930 is unusual because the texture of quartz grains in the rock is more typical of rocks from subarea 9, whereas the location of the specimen is structurally similar to that of specimen 1037 (diagram B4). Specimen 1930 was chosen for petrofabric analysis because the rock contains a well defined foliation, but no lineation. However, the orientation of quartz c_v in diagram C1 has no symmetrical relationship with respect to S_1.

Specimen 2083 (diagram C2) is a banded gneiss in which S_1 and L_1 are defined in hand specimen by microcline augen flattened and elongated in the plane of foliation. In thin section, S_1 is defined by narrow quartz-rich bands bounded by plane-parallel biotite grains. Quartz c_v are concentrated normal to L_1 and oblique to S_1.

Specimen 1635 (diagram C3) is foliated, but not lineated, quartzo-feldspathic gneiss. The textures of quartz and biotite in this specimen are shown on Plate 8A. Quartz grains flattened parallel to S_1 contain inclusions of biotite which parallel S_1 also. Quartz c_v on diagram C3 indicate a maximum concentration slighly oblique to the plane of S_1, and a dispersion along a great circle normal to S_1.

Specimen 2180 (diagram C4) is vitreous-type quartzite with narrow feldspar-rich bands parallel to S_1 and S_g. Lineation, L_1, is defined by mullion-like grooves on the foliation surface. Plate 8B shows the slightly flattened polygonal quartz grains in specimen 2180. Quartz c_v in diagram C4 are concentrated at a point slightly oblique to the plane of S_1 and normal to L_1.

Series D Specimen 1961 (diagram D1) is from a quartzite xenolith in the northeast portion of the batholith. The thin section is cut normal to the axial surface and subparallel to the axis of an isoclinal fold (approximately three inches across the limbs) in the hand specimen. The specimen does not contain a mineral lineation or a foliation. The preferred orientation of quartz c_v in diagram D1 appears to consist of a single broad maximum concentration which forms a partial girdle parallel to a great circle oriented approximately 45° from the trend of the fold axis, L. A secondary girdle of c_v parallel to a small circle centered on L is present also.

Specimen 1829 (diagram D2) is quartzo-feldspathic gneiss from area 9. The hand specimen is characterized by rod-like aggregates of quartz and feldspar grains parallel to L_1. Foliation is poorly developed in the hand specimen, and the trace of S_1 shown on diagram D2 is the average orientation defined by dispersed biotite flakes in the thin section. The polygonal outlines of quartz grains in specimen 1829 is illustrated in Plate 8C. Quartz c_v in diagram D2 are concentrated normal to S_1 and L_1, and dispersed partially along a great circle normal to L_1 and S_1.

Quartz Microfabrics of Plutonic Rocks (Series P)

Quartz c_v were measured in three specimens from the Chief Lake batholith and one specimen from the Eden Lake trondhjemite in order to determine the effect of post-intrusion deformation on the crystallographic orientation of quartz in rocks in which quartz is not the dominant mineral phase. The four rocks in series P contain 20 to 30 per cent quartz; the bulk of the rocks are composed of feldspar, and mica and epidote are minor contituents.

Specimen NEC (diagram P1) is quartz monzonite from a portion of the batholith that contains no penetrative structures. Undulose extinction of quartz grains in the thin section indicates that sample NEC has undergone some post-intrusion deformation, but no penetrative structures are visible. Diagram P1 indicates a random distribution of quartz c_v in the rock.

Specimen 911 (diagram P2) is quartz monzonite from the southwest portion of structural subarea 8. No penetrative structures are visible in the hand specimen or the thin section of specimen 911, but the feldspars in the thin section are fractured, and quartz occurs as irregularly oriented lenses of strained grains. However, petrofabric diagram P2 does not indicate any obvious pattern of preferred quartz c_v orientation in the rock.

Specimen 1118 (diagram P3) is foliated and lineated quartz diorite from the sheared northwest border zone of the batholith. S_1 and L_1 are defined by elongated and flattened feldspar rods in hand specimen. Plate 6C shows the texture and fabric of specimen 1118. Quartz occurs as sub-parallel lens-shaped aggregates of strained grains; feldspars are sub-round in the plane of the thin section, and commonly are offset along fractures. The petrofabric diagram of quartz c_v indicates a partial girdle of c-axes along a great circle normal to L_1 and S_1. The degree of preferred quartz c_v orientation in diagram P3 is surprising--considering that quartz constitutes less than thirty per cent of the rock, the bulk of which is composed of much larger feldspar grains that apparently were deformed in a brittle manner. Intuitively, it might be expected that local stress differences adjacent to the feldspar porphyroclasts would determine a less regular quartz c_v pattern than that shown.

Specimen 650 (diagram P4) is foliated trondhjemite from the portion of the Eden Lake complex in area 7. The texture of specimen 650 is similar to specimen 1118 in that quartz is recrystallized into lens-shaped aggregates, whereas plagioclase has been deformed in a brittle manner. Quartz in specimen 650, however, commonly occurs as large, strained, lens-shaped single grains with smaller grains along the borders (Plate 8D). Quartz c_v form a partial girdle in a plane normal to S_1 on diagram P4.

<u>Summary of Quartz Microfabrics</u> Petrofabric diagrams of quartz c_v are divided into four series based upon morphology of quartz grains in thin section. The shape, nature of quartz-quartz boundaries, and amount of internal strain shown by

quartz grains appear to vary systematically in the area. Huronian quartzites containing no penetrative structures (e.g., specimens 1091 and 1950 illustrated in Plates 7A and 7B) are composed of quartz grains with no preferred shape orientation and sutured quartz-quartz boundaries. Quartzites containing prominant mesoscopic structures are composed of quartz grains whose shape reflects the type of mesoscopic structure present in the rock (e.g., specimen 104, Plate 7C).

The nature of quartz-quartz boundaries varies independently of the general shape of the grains, as both sutured grains and grains with slightly curved or straight boundaries may have an overall preferred shape orientation (Compare Plate 7C and 7D.). In general, however, quartz in greenschist facies rocks (structural subareas 1-7) has sutured boundaries, and quartz in higher grade rocks (subareas 8 and 9) has mosaic texture with slightly curved or straight-line-segment boundaries. The amount of internal strain in quartz (as indicated by undulose extinction) generally is less in rocks from the almandine-amphibolite facies than from rocks of the greenschist facies.

The common pattern of preferred quartz c_V orientation in specimens containing a lineation, a foliation, or both S_1 and L_1 is a complete or partial girdle of c-axes in a plane normal to the mesoscopic structural fabric elements. The petrofabric diagrams of series A show more random distribution patterns of c_V than the other diagrams of quartzitic rocks. Preferred morphological orientation of quartz is greatest in series B, and tends to be less obvious or absent in series C and D; whereas, quartz c_V show a high degree of preferred orientation regardless of the shape of grains in the specimens of these series. Quartz c_V in rocks from area 9 (diagrams B2, C2, C3, C4, D2) indicate a tendency towards a partial girdle or point maximum concentrated normal to L_1. The position of the maximum concentration with respect of the trace of S_1 varies from slightly oblique to the plane of foliation (diagrams

B2, C2, C3, C4) to normal to S_1 (diagram D2). Oriented diagrams A3, B3, B4, and B5 are significant because L_1 and the c_v girdles have approximately the same geographic orientation in each of the petrofabric diagrams, but S_1 strikes in various directions.

Discussion of Texture and Crystallographic Fabric of Quartz

Although it has never been demonstrated experimentally, it is generally assumed that a pattern of preferred crystallographic orientation of quartz is due to recrystallization under the influence of a non-hydrostatic stress system. Stress induced preferred crystallographic orientation of calcite aggregates and metals has been demonstrated in the laboratory (See Turner and Weiss, 1963, p. 327-333 and p. 345-355.).

By analogy with deformed metals, marble has been deformed experimentally by both cold-working and hot-working processes (Turner, Griggs, and Heard, 1954). The principal difference between the two processes is that cold-working occurs at temperatures low enough to permit appreciable work hardening and storage of strain energy within the deformed grains, whereas hot-working takes place above a transition temperature where strain energy is dissipated and work hardening does not occur. Prolonged heating of cold-worked materials in a hydrostatic stress field causes annealing recrystallization. Annealing of marble tends to randomize pre-existing patterns of preferred crystallographic orientation developed by cold and hot-working (Griggs, Paterson, Heard and Turner, 1960).

Texturally, cold-worked marble resembles the quartz of series B, whereas hot-worked or annealed marble more closely resembles the mosaic quartz texture of series C and D. The high degree of preferred c_v orientation of the majority of rocks in series C and D is indicative of syntectonic rather than annealing recrystallization if, like calcite, quartz c_v

patterns are de-oriented by annealing.

Conclusions Comparison of quartz microfabrics with the fabric of experimentally deformed marble suggests that cold-working of quartz has occurred in greenschist facies rocks, and that hot-working of quartz has occurred in the almandine-amphibolite facies. The general absence of randomly oriented quartz c_v patterns in foliated or lineated rocks may indicate that static recrystallization of quartz (annealing recrystallization) did not occur after deformation had ceased. Absence of strained and sutured quartz in the almandine-amphibolite facies rocks of area 9 suggests also that deformation did not continue below the transition temperature between hot-working and cold-working of quartz.

Discussion of recrystallization mechanisms of quartz is, at the present time, speculative and will not be elaborated upon. Experimental study of recrystallization mechanisms of quartz is an active field of research, but none of the results appear to explain the origin of quartz fabrics in highly strained rocks. Regardless of the actual deformation mechanisms, however, quartz crystallographic fabrics show consistent symmetry patterns which can be compared to symmetries of larger scale structures.

III. STRUCTURAL FABRIC SYMMETRY

> A principal aim of structural analysis of tectonites is to deduce something of the nature, and especially the symmetry, of strain from the tectonite fabric. (Paterson and Weiss, 1961 p. 872.)

The "argument" of symmetry is useful to locate the principal directions of strain in a rock--although the symmetry of the tectonite fabric may not define which fabric symmetry axes parallel specific principal directions of strain. Generally, the symmetry of the rock fabric eliminates certain types of strain and orientations of principal strain axes from consideration, but does not give a unique solution to the problem.

The symmetry class of a structural fabric element is determined by the number of symmetry axes or planes (mirror planes) that can be drawn on a petrofabric diagram of the geometry of the structural element. The symmetry of the total fabric is given by the symmetry elements that are common when the symmetries of the subfabrics of the various kinds of fabric elements are superimposed (Paterson and Weiss, 1961, p. 868 and Fig. 9).

Primary subfabrics (e.g., bedding) provide a record of the total strain imposed on a rock, although, if the strain results from several superimposed deformations, inherited subfabrics may have a lower symmetry than the individual strains. Imposed subfabrics (e.g., mineral lineations, foliations, crystallographic fabrics), on the other hand, do not indicate the total strain, but commonly reflect the symmetry of the final deformation more perfectly than inherited subfabrics.

With respect to the symmetry of data plotted on equal-area diagrams, Paterson and Weiss (1961, p. 862) stated that:

> The symmetry of the preferred orientation of the fabric elements is determined by the

> symmetry elements that can be observed in such a projection. In this connection, only the form of the main features of the patterns (such as well defined girdles or strong maxima) should be considered, since the diagrams usually lack all symmetry if every detail is taken into account. The main features are reproducible in different sets of adequate data from the same homogeneous body, whereas the detail of the contour patterns is not.

The symmetry of quartz c_v subfabrics in foliated and lineated rocks depends upon the significance attached to the orientation of c_v maxima within the great circle girdles. As only the main features of the patterns are considered important (Paterson and Weiss, 1961, p. 862), quartz c_v girdles are interpreted as having an axial symmetry pattern (i.e., an infinite number of symmetry planes intersecting parallel to L_1). A point concentration of quartz c_v is axial also (A single mirror plane is normal to the symmetry axis.), whereas a single concentration dispersed parallel to a segment of a great circle is an orthorhombic symmetry pattern. Petrofabric diagrams without mirror planes are triclinic. Orthorhombic and axial symmetries are indicated by several other geometries of structural fabric elements, but they were not recognized in this research.

In the following sections, the symmetry of tectonite fabrics will be discussed in an effort to determine principal directions of strain. The basic assumptions involved are that structural fabric elements sharing common symmetry elements are genetically related, and that the symmetry of the tectonite fabric reflects the symmetry of the strains responsible for the evolution of the fabrics.

Fabric Symmetry of Lineated Rocks

Lineated rocks with, or without, a foliation are the typical tectonites in most of the area. These rocks include the Grenville province gneisses, a large portion of the Chief Lake batholith (and associated migmatites), the Huronian metasedimentary rocks in the southwest portion of the map area, and the eastern border of the Eden Lake trondhjemite. Penetrative L_1 occurs throughout structural subareas 7, 8 and 9, and in portions of subareas 4, 5 and 6 (Figs. 5 and 6). Foliation occurs as a penetrative structure throughout the area, but is developed most intensely in areas 4 to 9.

Fabric Symmetry of Areas 7 and 8 Mesoscopic folds in bedding are rare in these subareas, and S_b generally parallels S_1 in feldspathic quartzite and in metasedimentary xenoliths in the batholith. The mesoscopic combination of plane-parallel foliation with homoaxial mineral lineations on the S_1 surface defines an orthorhombic symmetry pattern (Fig. 16a). One symmetry plane (m_1) is normal to L_1; the other two planes intersect parallel to L_1, and one (m_2) is the foliation surface. The third plane of symmetry (m_3) is normal to S_1.

Petrofabric diagrams of quartz c_v in rocks from areas 7 and 8 show various symmetry patterns (Fig. 5, diagrams B5, B6, D1, P2, P4). Rocks containing mesoscopic foliation and lineation have orthorhombic fabrics. For example, diagram B5 (Fig. 16b) has three symmetry planes: the plane bisecting the quartz c_v girdle (m_1), the plane of foliation (m_2), and a third plane (m_3) containing L_1 and normal to the other two symmetry planes. Diagram B6 (Fig. 5) also shows an orthorhombic combination of S_1, L_1, and c_v, and diagram P4 shows an orthorhombic combination of S_1 and c_v. Rocks without penetrative lineation or foliation (Fig. 5, diagrams D1 and P2) have triclinic quartz c_v subfabrics.

The macroscopic fabric diagram of S_1 and L_1 in area 7 shows orthorhombic symmetry (Fig. 16c). m_1 is normal to L_1 and bisects the partial girdle of S_1 poles; m_2 passes through the dominant orientation of S_1; m_3 is normal to m_1 and m_2. The symmetry of S_b in area 7 (Fig. 16d) is orthorhombic also, and the symmetry planes are coincident with those of S_1 and L_1. On Figure 16d, m_2 represents the dominant attitude of bedding. Fold axes are not considered significant because only thirteen

Figure 16

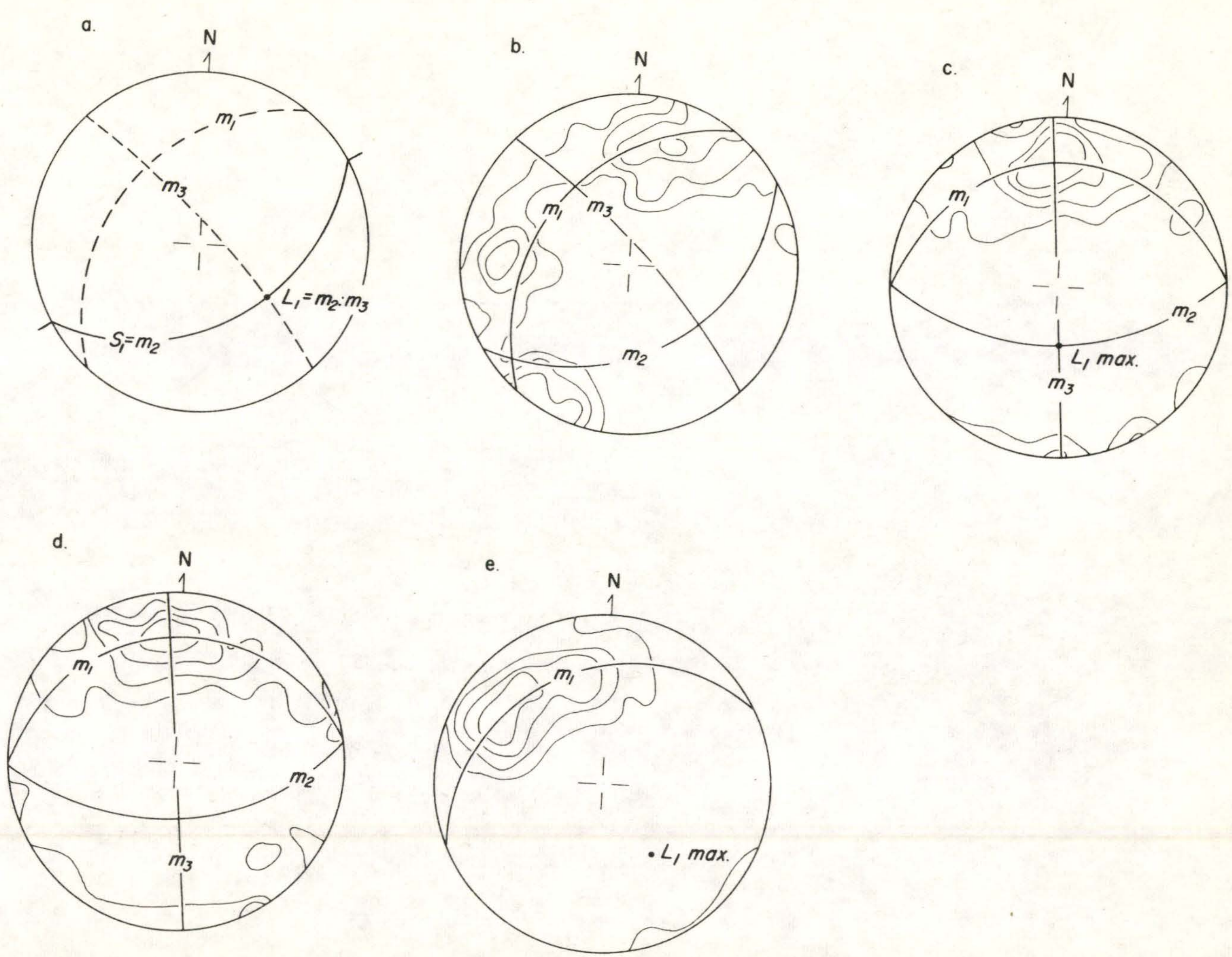
a. Orthorhombic mesoscopic symmetry of S_1 and L_1 in areas 7 and 8. The orientations of S_1 and L_1 in petrofabric diagram B5 (Fig. 5) are used as a typical example.

b. Symmetry of quartz c_v in diagram B5 (Fig. 5). The symmetry of the total microscopic fabric is orthorhombic ($m_2=S_1$, and $m_2:m_3=L_1$ in the thin section), but the symmetry of the quartz c_v subfabric is axial with the symmetry axis parallel $m_2:m_3$. (200 c_v, contours 1, 2, 4, 7% per 1% area)

c. Orthorhombic macroscopic symmetry of S_1 and L_1 in area 7. (S_1 contours 1, 3, 5, 7, % per 1% area, 251 poles)

d. Orthorhombic macroscopic symmetry of S_b in area 7. (S_b contours 1, 3, 5, 7% per 1% area, 169 poles)

e. Monoclinic (nearly orthorhombic) macroscopic symmetry of S_1 and L_1 in area 8. (S_1 contours 1, 3, 5, 7% per 1% area, 433 poles)



FABRIC SYMMETRY OF AREAS 7 AND 8

Figure 16

folds in bedding were observed in area 7.

In area 8, the macroscopic fabric diagram of S_1 and L_1 shows nearly orthorhombic symmetry (Fig. 16e). Perfect orthorhombic symmetry is not achieved because the maximum concentration is not in the center of the partial girdle of S_1 poles. The macroscopic geometry of area 8 is monoclinic, and the symmetry plane (m_1) parallels the partial girdle of S_1 poles. The reason for the off-center spread of S_1 poles is that in the southern part of the batholith the regional trend of S_1 progressively changes from northeast to east (See Fig. 6.). The regional orientation of L_1 remains constant so that a maximum trend can be designated on Figure 16e.

Fabric Symmetry of Area 9 Individual mesoscopic folds in Grenville province gneisses exhibit monoclinic symmetry (Fig. 17a). The symmetry plane (m_1) is normal to the fold axis; no other symmetry planes can be drawn because one limb generally is longer than the other, although individual fold limbs tend to be symmetrically oriented with respect to the axial surface. Mesoscopically planar S_1 and homoaxial L_1 define an orthorhombic symmetry in the gneisses. The total mesoscopic fabric defined by S_g, S_1, L and L_1 is monoclinic because fold axes parallel mineral lineations and axial surfaces parallel S_1.

Microfabric studies of quartzitic rocks from area 9 (Fig. 5, diagrams B2, C2, C3, C4, D2) indicate that quartz c_v tend to form a partial girdle parallel to the great circle normal to S_1 and L_1 (in rocks containing mesoscopic foliation and mineral lineation), but no consistent symmetrical relationship exists between the orientation of S_1 and quartz c_v maxima. The fabrics of quartz c_v in the specimens examined indicate a tendency towards a single maximum concentration, rather than a complete great circle spread. Both of these patterns (i.e., point maximum and complete girdle) represent axial symmetry, whereas a partial girdle of c_v is orthorhombic.

Figure 17

a. Monoclinic mesoscopic symmetry of folds in S_g. S_1 parallels axial planes and L_1 parallels fold axes.

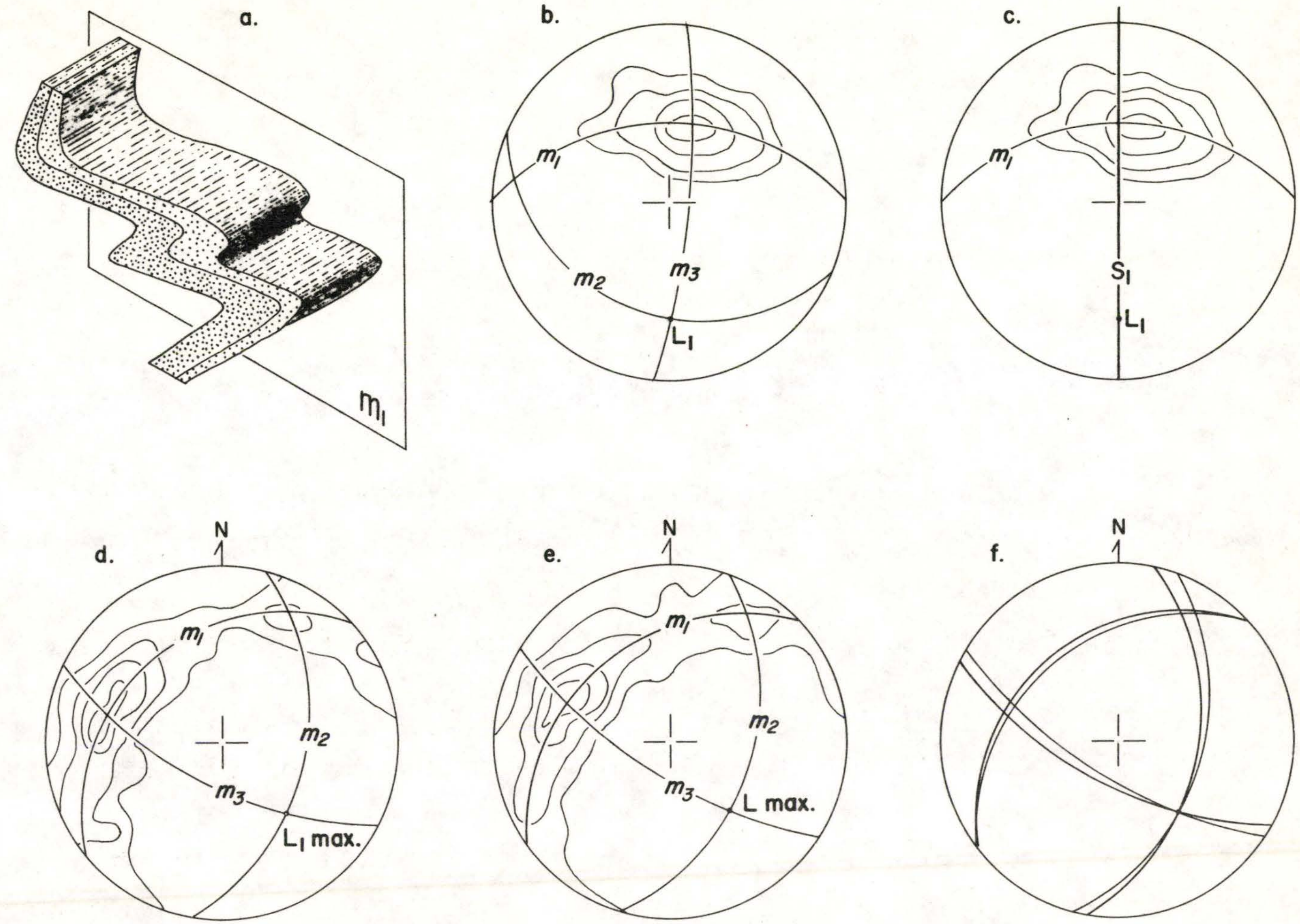
b. Orthorhombic microscopic symmetry of quartz c_v and L_1 in diagram C4 (Fig. 5). (200 c_v, contours 1, 3, 10, 20% per 1% area)

c. Monoclinic microscopic symmetry of quartz c_v, L_1 and S_1 in diagram C4.

d. Orthorhombic macroscopic symmetry of S_1 and L_1 in area 9. (S_1 contours 1, 3, 5, 7% per 1% area, 307 poles)

e. Orthorhombic macroscopic symmetry of S_g and L in area 9. (S_g contours 1, 3, 5, 7% per 1% area, 349 poles)

f. Symmetry planes from Figure 17d and 17e superimposed.



FABRIC SYMMETRY OF GRENVILLE PROVINCE GNEISSES

Figure 17

Using diagram C4 as an example, quartz c_v and L_1 define an orthorhombic symmetry (Fig. 17b), but S_1 is inclined obliquely to two of the symmetry planes, and the symmetry of the total microfabric is monoclinic (Fig. 17c).

The macroscopic fabric symmetry of S_1 and L_1 in the Grenville province gneisses is orthorhombic (Fig. 17d). One symmetry plane (m_1) is normal to the dominant trend of L_1 and parallels the girdle of S_1 poles; the other symmetry planes intersect parallel to the maximum concentration of L_1, and one (m_2) is normal to the maximum concentration of S_1 and the other (m_3) passes through the S_1 pole maximum. The symmetry of S_g and L (Fig. 17e) is orthorhombic, and is almost exactly the same as S_1 and L_1. The total macroscopic fabric symmetry is orthorhombic as shown by the coincidence of the symmetry planes for S_1 and L_1, and S_g and L (Fig. 17f).

<u>Fabric Symmetry of Areas 4, 5 and 6</u> Penetrative mineral lineations occur in the southern parts of areas 4, 5 and 6, and occur locally in the northern parts of these structural subareas. On the mesoscopic scale, S_1 and L_1 define an orthorhombic symmetry similar to that described previously in area 7 (Fig. 16a). Mesoscopic folds in bedding are monoclinic with the symmetry plane normal to the fold axis. Fold axes do not always parallel mineral lineations in the same exposure, but insufficient folds were observed to deduce any consistent geometric relationship between the two linear elements. Where folds are present, the total mesoscopic symmetry may be triclinic, but generally folds are absent and the mesoscopic fabric is orthorhombic.

Petrofabric diagrams of quartz c_v from areas 4, 5 and 6 are triclinic for rocks that do not contain L_1 (Fig 5, diagrams A2, B1). Lineated rocks show a girdle of quartz c_v normal to L_1 and exhibit axial symmetry (Fig. 5, diagrams A3, B3, B7). The microscopic combination of S_1, L_1 and c_v

is orthorhombic, but foliated rocks without a mineral lineation in which S_b and S_1 are not parallel (diagrams A2, B1) show triclinic symmetry. The symmetry patterns of microscopic fabrics in rocks from area 4, 5 and 6, however, are not as well defined as for areas 7, 8 and 9.

On the macroscopic scale, linear structures (L_1 and L) tend to plunge southward at more than 45 degrees in these areas and indicate an axial (point maximum) symmetry pattern. S_1 and S_b poles show peripheral girdles or partial girdles which approach axial or orthorhombic symmetry. The total macroscopic symmetry is difficult to define, however, as the points on the petrofabric diagrams are dispersed more irregularly than in areas 7, 8 and 9. The total fabric is considered to be triclinic, but could be interpreted also as monoclinic with m_1 parallel to the girdles of S_1 and S_b poles.

Fabric Symmetry of Non-lineated Huronian Rocks

The significant structures in non-lineated meta-sedimentary rocks are macroscopic folds in bedding (Fig. 7). Mesoscopic folds are too rare to be important in determining the structural fabric symmetry of these rocks. Quartz c_v fabric of a mesoscopically undeformed specimen of quartzite (i.e., no penetrative foliation or lineation) from area 2 has no well defined symmetry planes (Fig. 5, diagram A1), suggesting that deformation was not penetrative to the microscopic scale.

The macroscopic symmetry of fabric diagrams for areas 1 and 2 are shown on Figure 18. The composite diagram for area 1 (Fig. 18a) shows the dominant orientation of S_1, a partial girdle of bedding poles, and the distribution of mesoscopic fold axes. The macroscopic symmetry of area 1 is monoclinic (m_1 parallels the partial girdle of S_b poles), although the mesoscopic fold axes indicate a variance from the pole to the symmetry plane, β.

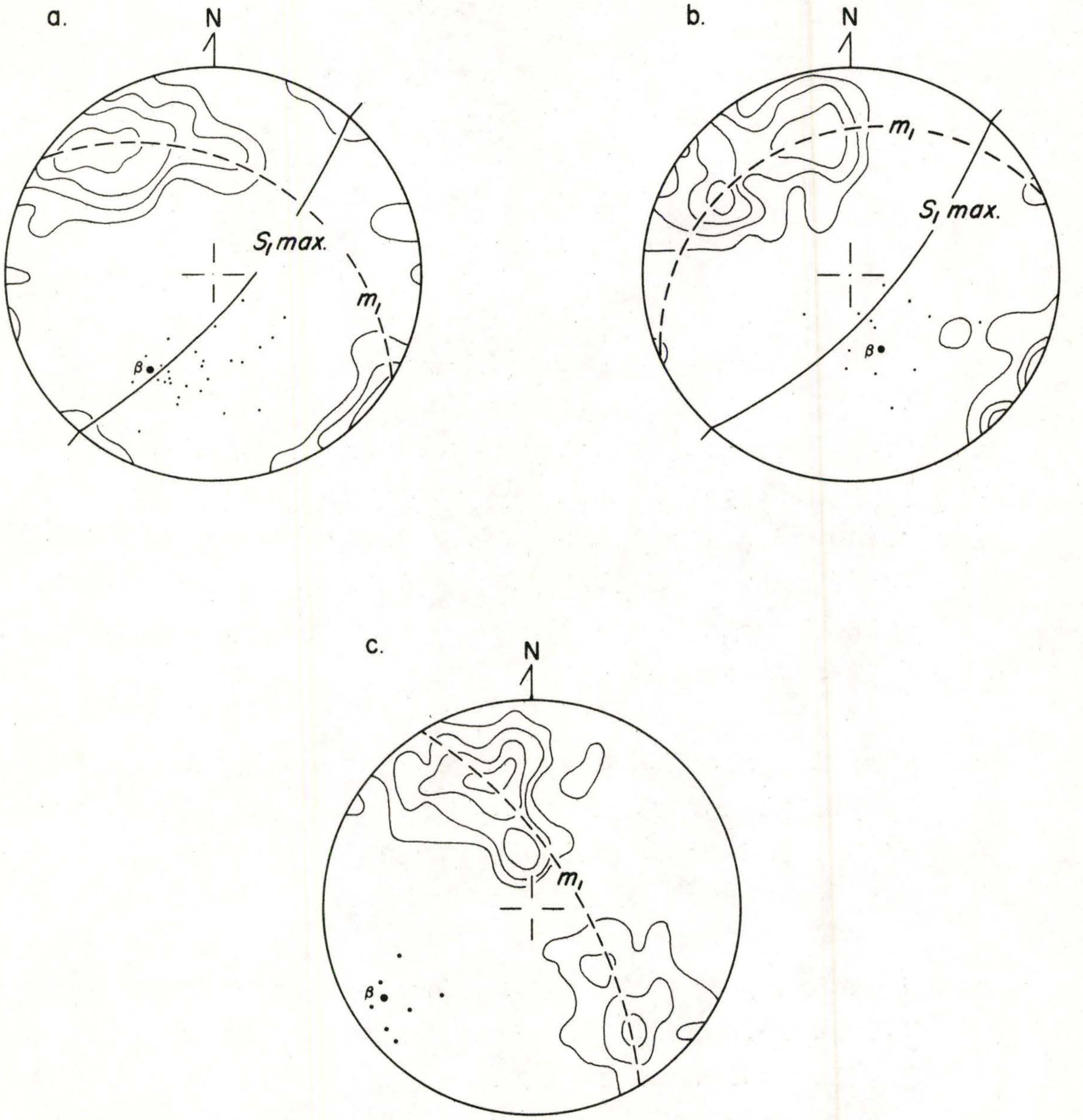
The macroscopic symmetry of area 2 is shown on Figure 18b, and is less well defined than in area 1 (Fig. 18a).

Figure 18

a. Area 1 composite diagram showing S_b (contours 1, 3, 5, 10% per 1% area, 75 poles), the dominant trend of S_1, and mesoscopic fold axes (small dots). The symmetry of the diagram is monoclinic (nearly orthorhombic), and the pole to the symmetry plane, β , is indicated.

b. Area 2 composite diagram showing S_b (contours 1, 3, 5, 10% per 1% area, 110 poles), the dominant trend of S_1, and mesoscopic fold axes (small dots). The pole to the symmetry plane, β , is indicated.

c. Monoclinic (nearly orthorhombic) symmetry of S_b in Waters Township. The pole to the symmetry plane β , is indicated as well as the orientation of macroscopic fold axes (small dots) from Figure 7. (113 poles, contours 1, 3, 6, 9% per 1% area)



MACROSCOPIC FABRIC SYMMETRY
OF NON-LINEATED HURONIAN ROCKS

Figure 18

The dominant trend of S_1 is northeast, parallel to S_1 in area 1. Bedding poles appear to lie along a great circle (m_1 on Fig. 18b), but the symmetry may be triclinic.

Figure 18c is a petrofabric diagram of S_b from Waters Township mapped by Card (1964). The data are taken from Card's map and cover the area in Waters Township in which the bedding trend lines are drawn on Figure 7. Bedding poles on Figure 18c lie along a great circle (m_1) and define a monoclinic (nearly orthorhombic) symmetry pattern. The pole to the symmetry plane (β) indicates the general orientation of fold axes in the area, and is consistent with the orientations obtained for individual macroscopic folds from Figure 7.

Summary of Fabric Symmetries

Orthorhombic fabrics are common in lineated and foliated tectonites. These include Huronian metasedimentary rocks, the Chief Lake batholith and Grenville province gneisses. Homotactic orthorhombic fabrics occur at all scales investigated in unfolded lineated rocks, but only the macroscopic symmetry of the Grenville province gneisses is orthorhombic.

Northwest of the batholith, penetrative lineations are not common, and the structural fabric is characterized by macroscopic folds in bedding. In Waters Township, macroscopic folds are nearly orthorhombic, but the macroscopic fabric becomes monoclinic or triclinic near the western border of the batholith.

IV. PRINCIPAL STRAIN DIRECTIONS IN LINEATED ROCKS

The significant aspect of the strain observed in lineated rocks west of the Grenville province gneisses is that mesoscopic folds are rare. For example, lineated and foliated dikes in the batholith rarely are folded, and, similarly, mesoscopic folds generally are absent from lineated and foliated quartzite beds west of the batholith. The strain in these rocks is finite as indicated by strong preferred morphological and crystallographic quartz fabrics, and cataclastic deformation of feldspars. Absence of mesoscopic folds indicates that the strain is irrotational. Rotational strain can appear irrotational if: (1) inherited fabric elements are not present to record rotational components of strain, (2) rotation was external to the field observed, or (3) all inherited planar fabric elements had a preferred orientation normal to the rotation axis prior to deformation. Because the primary array of layers in cross-bedded quartzite southwest of the batholith (Fig. 5, areas 4 and 7) could not have been parallel originally, it is thought that the strain in unfolded lineated rocks is irrotational, and that the rocks were structurally isotropic during deformation. Absence of folds in inherited planar elements of rocks containing penetrative foliation or lineation indicates that within the field considered deformation was homogeneous.[1]

Rock deformation has been analysed in terms of finite homogeneous strain by Flinn (1962, 1965a, 1965b), Brace (1961), Turner and Weiss (1963) and Ramsay (1967). Flinn's analyses are general and relate common tectonite fabrics to principal directions of homogeneous strain.

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[1] A stricly homogeneous deformation can be specified in terms of a displacement field, such that all points initially on any straight line remain on a straight line. (Turner and Weiss, 1963, p. 367)

Flinn (1965a) related morphological and crystallographic fabric symmetries of quartz to the shape of the deformation ellipsoid. The deformation ellipsoid is the ellipsoid resulting from the finite homogeneous strain of an original sphere (Flinn, 1962, p. 386). The major axes of the ellipsoid are designated $Z \geqslant Y \geqslant X$, and refer to the maximum, intermediate and least principal strain directions. The ellipsoid varies in shape from a prolate axial ellipsoid ($Z>Y=X$), through a series of orthorhombic triaxial ellipsoids ($Z>Y>X$), to an oblate axial ellipsoid ($Z=Y>X$) (Flinn, 1965a, p. 44).

Relationship Between Quartz Microfabric Symmetry and the Deformation Ellipsoid

Quartz grains in series B and C (Fig. 5) commonly are elongated and flattened parallel to L_1 and S_1, showing an orthorhombic symmetry and morphology similar to a triaxial deformation ellipsoid with principal directions $Z>Y>X$. The Z-direction parallels L_1, and the X-direction is normal to S_1. The tendency of quartz grains to show a strong morphological orientation decreases with increasing metamorphic grade, and in higher grade rocks L_1 and S_1 commonly are defined by feldspar augen and biotite.

The symmetry of quartz c_v subfabrics also can be related to the deformation ellipsoid (Flinn, 1965a). The complete girdle of c-axes parallel to a great circle is an axial symmetry pattern corresponding to the prolate deformation ellipsoid with the girdle axis parallel to Z. A single concentration of c_v is axial also, but corresponds to the oblate deformation ellipsoid with the fabric symmetry axis parallel to X. Orthorhombic partial girdles are compatible with triaxial deformation ellipsoids of the type $Z>Y>X$.

In the western part of the area, quartz c_v in lineated rocks form complete girdles in the plane normal to L_1 and S_1. Quartz c_v in Grenville province gneisses, however, form partial girdles in the plane normal to L_1 and S_1, but commonly the maximum c_v concentration is inclined obliquely to S_1. Assuming that quartz crystallographic fabrics were imposed late in the deformational history of the rocks, homotactic relationships between L_1 and quartz c_v indicate that L_1 was an active structure until the final recrystallization of quartz, whereas the heterotactic relationship commonly observed between S_1 and quartz c_v fabrics indicates that in some areas foliation was a passive structure when the crystallographic fabric was formed.

The general type of strain indicated by rocks containing mineral lineations is elongation parallel to L_1 with flattening normal to S_1, or essentially equal flattening in the plane normal to L_1 (where S_1 is absent). In rocks containing an active lineation and passive foliation, the maximum principal strain direction, Z, parallels L_1 and is normal to the plane containing the quartz c_v girdle or partial girdle.

In summary, morphological and crystallographic microfabrics of quartz are related to triaxial deformation ellipsoids in the manner described by Flinn (1965a). The shapes of quartz grains in lower grade rocks commonly define triaxial deformation ellipsoids of the type Z>Y>X, wherein the direction of maximum elongation (Z) parallels L_1 and the direction of maximum shortening (X) is normal to S_1. In the higher grade rocks, quartz grains tend towards subhedral or equant shapes, and L_1 and S_1 are defined by preferred orientations of feldspar and mica. Quartz c_v in lineated rocks consistently show strong tendencies to lie in the plane normal to L_1 and S_1, but, because S_1 is not consistently oriented normal to quartz c_v maxima in partial girdles, it is thought that foliation in some areas was passive at the time of final recrystallization of quartz. Unfortunately, few oriented specimens were collected, and it is not possible to define principal strain directions based on quartz crystallographic fabrics alone.

Macroscopic Principal Strain Directions in Unfolded Lineated Rocks

Orthorhombic mesoscopic fabrics in areas 7 and 8 are related to a triaxial orthorhombic deformation ellipsoid of the type Z>Y>X with L_1 parallel to Z and S_1 normal to X. The macroscopic symmetry of area 7 is orthorhombic as well, and at this large scale it is convenient to assign geographic coordinates to the orientations of principal strain directions. Figure 19a shows the petrofabric diagram of S_1 and L_1 for area 7 (Fig. 5) with the principal directions Z, Y and X parallel to the appropriate fabric symmetry axes. Z in area 7 trends 176 degrees and plunges 64 degrees parallel to the intersection of $m_2:m_3$ and the regional trend of L_1; X is normal to the average orientation of S_1 and parallels the line of intersection of $m_1:m_3$; Y is horizontal and strikes at 090 degrees parallel to the intersection of $m_1:m_2$.

Principal strain directions are less well defined in area 8 than in area 7 due to to the macroscopic fold affecting the regional trend of S_1 and L_1 (Fig. 6). This broad fold may be the youngest structure of regional significance in the area, or it may be coeval with the formation of L_1. If the latter interpretation is correct, it implies a progressive regional change in the orientations of the principal strain directions. The pole of the macroscopic symmetry plane for area 8 (Fig. 19b) parallels the dominant trend of L_1, and indicates the orientation of Z. The macroscopic orientations of X and Y are not defined because of the monoclinic symmetry.

Macroscopic Principal Strain Directions in Grenville Province Rocks

The Grenville province gneisses (area 9) have orthorhombic macroscopic symmetry. The structural geometry, however, is markedly different from the rocks in areas 7 and

8 because mesoscopic folds are abundant in the gneisses. Mesoscopic folds in S_g and macroscopic folds in S_1 have axes parallel to L_1, indicating the total strain is not homogeneous. It cannot be determined whether the mesoscopic folds in S_g are the result of an initial inhomogeneous strain or whether they originated prior to the final deformation. Regardless of their origin, folds in S_g were passive structures during the final penetrative deformation. The similar style of folds, and penetrative foliation and lineation indicate that gneissic banding exerted little or no control on principal strain directions as recorded in the final fabric geometry.

Differential (inhomogeneous) slip on S_1 is a possible mechanism for generation of passive folds in pre-existing planar surfaces. However, a slip mechanism is two dimensional (no deformation parallel to Y), and cannot account for any final parallelism of fold axes except in the unrealistic case where the passive layers were parallel prior to slip folding. Considering that fold axes are parallel to L_1 (which is parallel to Z in areas 7 and 8, and is not a visible rotation axis), differential slip on S_1 does not provide an adequate interpretation of the symmetry of the gneisses or a consistent interpretation of the structure of all lineated rocks.

The presence of folded layers in the gneisses is not incompatible with a homogeneous final strain imposed on previously folded rocks . For example, Flinn (1962) demonstrated that homogeneous strain accentuates irregularities in passive surfaces, and pre-existing folds may change shape and orientation during homogeneous deformation. In the general case of three-dimensional homogeneous strain (principal directions $Z>Y>X$), inherited linear structures rotate toward Z, and inherited planar elements rotate towards the Y-Z plane (Flinn, 1962).

Fold axes and axial surfaces in rocks with a homogeneous component of strain have no symmetrical relationship to principal strain directions--except in special cases, such as extreme flattenings ($Z=Y\gg X$) or extreme elongations

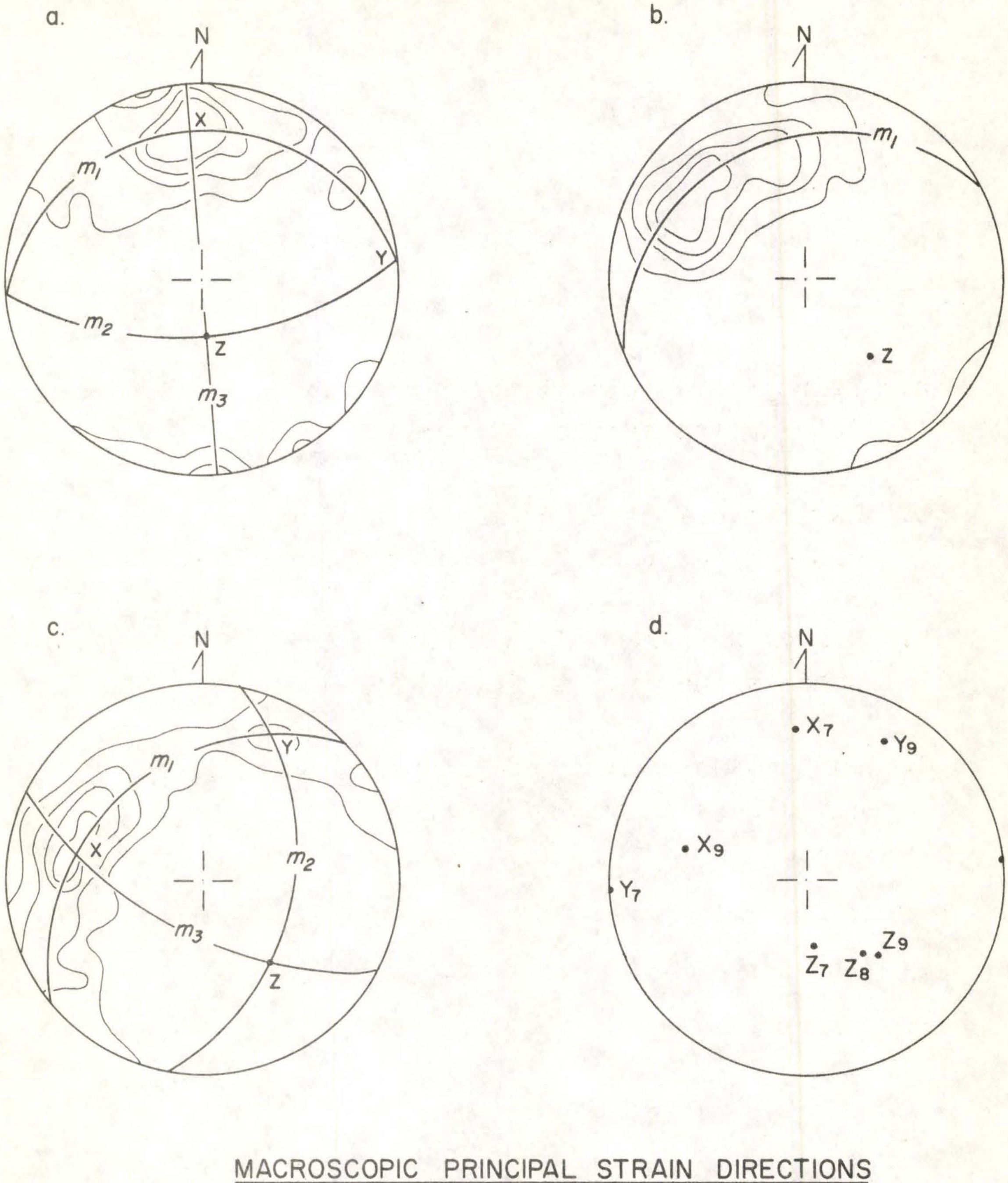
Figure 19

a. Principal strain directions in area 7. Symmetry data is the same as Figure 16d.

b. Maximum principal strain direction in area 8. Symmetry data is the same as Figure 16e.

c. Principal strain directions in area 9. Symmetry data is the same as Figure 17d.

d. Comparison of macroscopic principal strain directions in areas 7, 8 and 9.



MACROSCOPIC PRINCIPAL STRAIN DIRECTIONS

Figure 19

(Z>>Y=X). In the case of extreme uniaxial elongation, all passive linear structures (regardless of original orientation) rotate to positions subparallel to the maximum strain direction, and passive planar elements have complete freedom of orientation in the X-Y plane, but intersect parallel to Z (Flinn, 1962, Fig. 7). Extreme triaxial strain of the type Z>>Y>X causes passive linear elements to rotate parallel to Z, and passive planar elements to rotate normal to X (Flinn, 1962, Fig. 7).

The orthorhombic macroscopic symmetry of area 9 can be related to a triaxial deformation ellipsoid of the type Z>>Y>X, if it is assumed that L_1 was an active structure and L, S_g and S_1 were passive, and that strain is extreme. On Figure 19c, Z is drawn parallel to the dominant trend of L_1 and L (Parallel m_2:m_3) oriented at 136 degrees and plunging 42 degrees to the southeast. The dominant orientations of passive S_1 and S_g in the case of extreme triaxial strain are oriented normal to X (parallel m_1:m_3) which trends at 284 degrees and plunges 32 degrees to the west. Y is normal to X and Z, and trends parallel to the intersection of m_1:m_2 oriented at 028 degrees and plunging 22 degrees to the northeast.

Interpretation of the Regional Warp in S_1

The orientations of principal strain directions derived for areas 7, 8 and 9 are superimposed on Figure 19d. The Z-directions are closely grouped for the three structural subareas, but the Y- and X-directions in areas 7 and 9 lie approximately parallel to a great circle normal to the Z-directions. This regional strain geometry can be interpreted either as a progressive rotation of X and Y about Z, or as a subsequent warp in the foliation about an axis subparallel to Z. The first interpretation is more favorable because of the subparallelism of the rotation axis with Z. The latter interpretation suggests that it is only coincidental that the regional warp shares a symmetry axis with the dominant strain.

Summary of Macroscopic Symmetries

In summary, orthorhombic macroscopic symmetries in areas 7, 8 and 9 are interpreted as products of finite homogeneous strain. The macroscopic fabric symmetry is compared with extremely elongate orthorhombic deformation ellipsoids of the type Z>>Y>X, wherein Z parallels L_1 and X is normal to S_1. The direction of maximum elongation plunges steeply to the south or southeast, but the direction of maximum shortening is variable in the plane normal to Z, and depends upon the dominant macroscopic orientation of S_1.

V. KINEMATIC AND DYNAMIC ANALYSIS

Interpretation of axial and orthorhombic tectonite fabrics as products of extreme homogeneous strain indicates that the dominant penetrative movement occurred parallel to the direction of maximum elongation, and rotational components of the strain are either inherited or subsidiary. On a smaller scale, salt diapirs provide a geologically significant example of similar relationships between fabric symmetry and movement directions.

Studies by Balk (1949), Muelberger (1960), and Hoy, Foose and O'Neill (1964) showed that the internal fabric of salt domes has axial and orthorhombic symmetry. Passive layers (bedding) in the salt are folded about vertical axes, and halite and andydrite commonly are recrystallized into spindle-shaped aggregates aligned vertically. In the horizontal plane, axial surfaces of folds are folded irregularly about vertical axes. Near the border of the Winnfield Salt Dome, Hoy et al. (1964) showed that foliation parallels the vertical walls between salt and surrounding sediments, but that no consistent foliation strike exists inward from the walls.

Vertical linear elements are the dominant structures in salt domes, and the symmetry axes of both active (mineral lineations) and passive (fold axes) structures are parallel. In the example of salt diapirs, it is known that considerable vertical flowage has occurred within the salt, and, therefore, the fabric symmetry axis parallels the direction of flow. Rotational components of the strain, shown by passive folding of foliations, are subsidiary to the dominant strain which is vertical elongation.

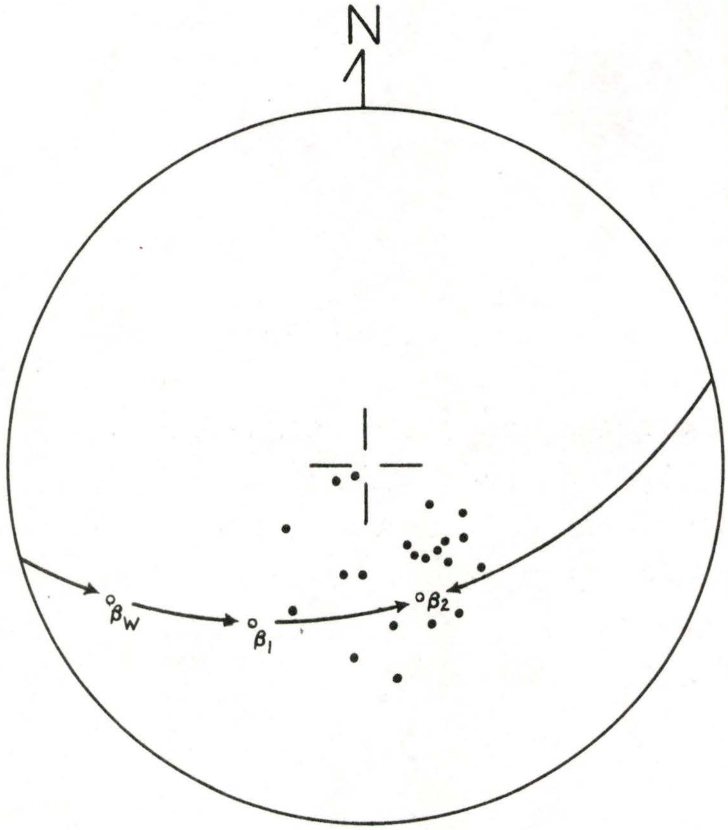
The similarity between the internal fabric symmetry of salt domes and the fabric symmetry of lineated rocks in the research area indicates that the dominant flow direction was parallel to the regional trend of L_1. Folds are either inherited inhomogenous strains or subsidiary rotational components of a dominantly homogeneous strain. Strong parallelism of passive fold axes and mineral lineations in Grenville province gneiss indicates that flow parallel to L_1 was "extreme", and spatial relations of high and low grade rocks indicate that flow was upwards to the northwest.

The macroscopic folds in Huronian rocks (Fig. 7) are thought to be Southern province structures that pre-date the flowage in the Grenville province. The kinematics of formation of Southern province folds is beyond the scope of this study, but analysis of the <u>deformation</u> of these structures shows that the relative amount of flow parallel to L_1 increased progressively from Waters Township to the western border of the Chief Lake batholith. Figure 20 shows the dominant trend of fold axes (β-directions) in Waters Township (Fig. 7), area 1 and area 2 as well as mineral lineations from area 2. Assuming that the folds in Waters Township were least affected by the deformation in the Grenville province, the arrows drawn along the great circle on which the three β-directions lie represent the rotation path of Southern province folds into parallelism with the flow direction during the final deformation in the area. The macroscopic folds are rotated into axial parallelism with the Grenville province structures in a distance of about one mile.

Very little is known about deformation mechanisms and the nature of stresses responsible for the evolution of silicate

Figure 20

Open circles are β-directions in Waters Township (β_W), area 1 (β_1) and area 2 (β_2). Dots are mineral lineations from area 2. Arrows indicate possible rotation path of fold axes from Waters Township to area 2.



MOVEMENT PATH OF FOLD AXES IN HURONIAN ROCKS

Figure 20

tectonites. Observed textures of lineated rocks in the research area show that quartz was the most ductile common mineral, and was deformed by a mechanism of recrystallization flow which produced fabrics similar to cold-worked metals under the conditions of the greenschist facies and hot-worked metals under higher grade conditions.

Considering that quartz has not been recrystallized experimentally--even at extreme pressures and temperatures--time-dependent flow (creep) appears to be a favorable recrystallization mechanism. Salt creeps under the influence of non-hydrostatic body forces, and, in view of the structural similarity with salt diapirs, time-dependent flow of quartz may have been the dominant deformation mechanism in the research area.

CHAPTER 5

TECTONIC SYNTHESIS

Southern province rock units in the research area consist of Huronian sedimentary rocks intruded by hornblende metagabbro, the Eden Lake complex and the Chief Lake batholith. The Grenville province consists of heterogeneous quartzo-feldspathic gneisses and massive amphibolite. According to Stockwell (1964), the Southern structural province was formed about 1600 m.y. ago during the Hudsonian orogeny. Macroscopic northeast-trending folds in Huronian rocks probably were formed at this time, and the rocks may have been matamorphosed to staurolite grade. The Chief Lake batholith was intruded about 1750 m.y. ago (Krogh, in Davis, et al., 1967), and was metamorphosed last about 1000 m.y. ago. The time of last metamorphism approximates that of the Grenville orogeny. During the final Grenville orogenic episode, deformation and metamorphism were essentially simultaneous, and extreme rock flowage was accompanied by metamorphism to the middle almandine-amphibolite facies.

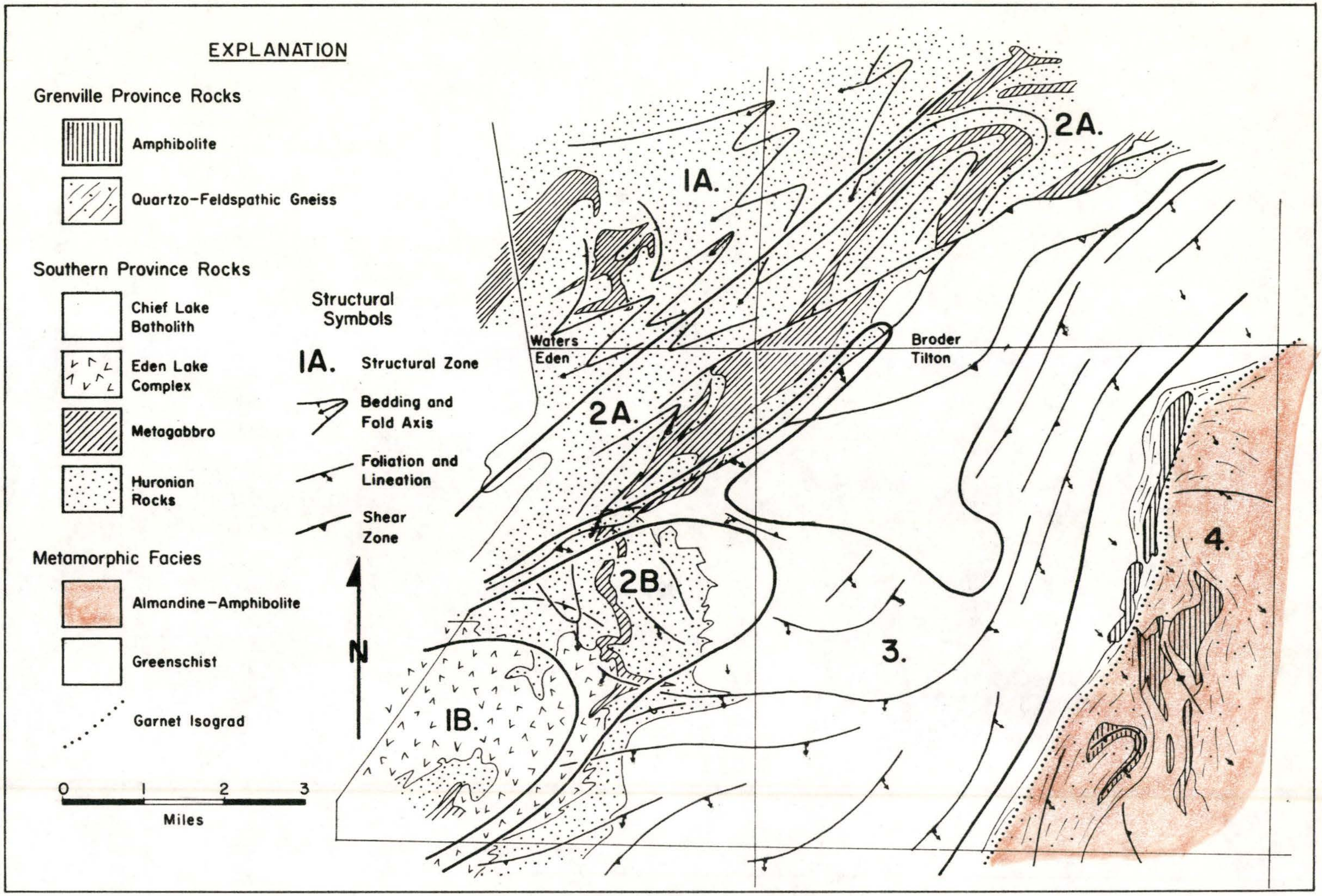
The Chief Lake batholith separates Huronian metasedimentary rocks from Grenville province gneisses and prevents direct correlation of paragneisses with Huronian rocks. However, the 1750 m.y. intrusive age of the batholith gives a minimum age for the protoliths of the gneisses, and considering that the batholith has intruded Huronian rocks along its west border, the gneisses east of the batholith also may be primarily of Huronian age.

I. THE NATURE OF THE GRENVILLE-SOUTHERN PROVINCE BOUNDARY

The Grenville-Southern province boundary has not been defined in previous chapters because a distinctive break between the provinces is not obvious. Distinction can be made easily between Southern province rocks west of the Chief Lake batholith and Grenville province gneisses to the east, but without radiometric ages batholithic rocks could not be assigned with certainty to either the Grenville or Southern provinces.

Figure 21, a tectonic map, shows the transitional nature of the boundary. Lower and middle greenschist facies rocks (chlorite-biotite zone) were recrystallized during the final Grenville metamorphism, but hornblende metagabbro and staurolite-bearing schist (e.g., in Waters Township) indicate that the chlorite-biotite zone contains higher grade assemblages (Hudsonian?) partly re-equilibrated to lower grade conditions. In the garnet-sillimanite zone, however, the rocks were completely recrystallized, and east of the garnet isograd they have equilibrated to the conditions of the middle almandine-amphibolite facies. The metamorphic grade near the garnet isograd changes in less than one mile from middle greenschist to middle almandine-amphibolite facies. The garnet isograd occurs in the region of greatest variation in metamorphic intensity, and provides a convenient metamorphic boundary between the provinces.

The area shown on Figure 21, is divided into four zones which show the progressive regional deformation that occurred during the final Grenville orogenic episode. Zones 1A and 1B were least affected by the final deformation. Macroscopic folds in Huronian rocks in zone 1A are upright, open, and plunge southwest at less than 40 degrees. Zone 1B contains



TECTONIC MAP

Figure 21

vertical foliation which may be related to the intrusion of the Eden Lake complex or the Hudsonian orogeny. In zones 2A and 2B, axes of macroscopic folds in Huronian rocks are rotated progressively parallel to the direction of maximum elongation and flow during the final deformation.

Deformation in zone 3 was dominantly homogeneous and flowage parallel to foliation and lineation transposed inherited planar elements into subparallelism with S_1 and L_1. The greatest flowage appears to have occurred in zone 4 because the abrupt increase in metamorphic grade in the vicinity of the garnet isograd indicates that higher grade rocks to the east flowed upwards from a deeper crustal level than rocks west of the isograd.

On the basis of structural data, the Grenville-Southern province boundary is defined as the western limit of penetrative mineral lineations (zone 3). Note that this places the "structural Grenville front" as much as 6 miles west of the metamorphic boundary (the garnet isograd), and includes part of the shear zone along the west border of the batholith with Grenville province structures.

II. CONCLUSIONS

The Grenville front is a dynamo-thermal metamorphic gradient, east of which Southern province rocks were transformed during the Grenville orogeny into gneiss and amphibolite with little or no addition of exogenous material, but with considerable chemical and mechanical re-working. The Chief Lake batholith is the only Southern province rock unit that was definitely traced across the boundary zone, but massive amphibolite east of the batholith may be metamorphosed Sudbury gabbro. Protoliths of quartzo-feldspathic gneisses in the Grenville province are indeterminate on the basis of field work.

Absence of mesoscopic folds in many penetratively deformed rocks indicates that deformation during the Grenville orogeny was dominantly homogeneous. Folds observed are thought to be inherited structures that were deformed passively by homogeneous strain. Strong axial parallelism of passive folds with mineral lineations in Grenville province rocks indicates that exteme elongation (probably viscous flow) occurred parallel to the linear structures. The abrupt increase in metamorphic grade across the garnet isograd suggests that differential flowage was greatest near the isograd, and that higher grade rocks to the east flowed upward from deeper crustal levels than those to the west.

Suggestions for Further Work

Rubidium-strontium isochron studies of mineral and whole-rocks might differentiate primary and metamorphic ages, and provide more positive correlations of rock units across the Grenville-Southern province boundary. Radiometric studies also might indicate more clearly the western extent of mineral recrystallization during the Grenville orogenic episode.

In regards to possible areal extensions of this research, the region to the southwest (Bevan Township) is most interesting because lineated and foliated Huronian metasedimentary rocks occur there. Also, the Chief Lake batholith should be mapped southward in order to determine its relationship to the Killarney batholith. If it is part of the Killarney batholith, the question arises as to why the largest igneous body in the Southern province parallels what is now the Grenville front.

The Grenville front is a major tectonic feature of the earth, and the conclusions of this study may be significant only locally. Many more detailed structural and petrologic studies (especially in conjunction with detailed radiometric studies) should be made across other segments of the front before any general conclusions are made concerning its nature and origin.

Plate 1

A. Torrential cross-bedding in lower feldspathic quartzite exposed east of Wavy Lake road at Tilton Lake.

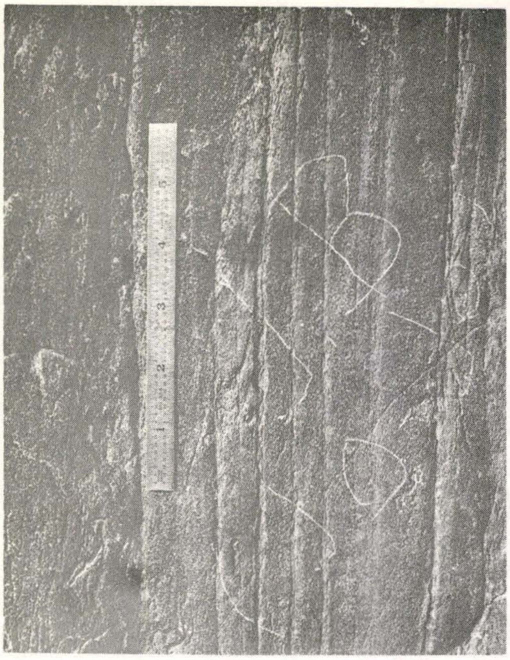
B. Graded semi-pelitic beds in lower feldspathic quartzite exposed west of Wavy Lake road on shore of Tilton Lake.

C. Primary igneous layering in the dioritic portion of the Eden Lake Complex. Upper layer contains glomeromegacrysts of hornblende disseminated in a plagioclase-rich matrix. Lower layer contains evenly distributed hornblende and plagioclase.

D. Metagraywacke portion of large xenolith at Chief Lake showing muscovite-quartz pseudomorphs of andalusite.

A

B



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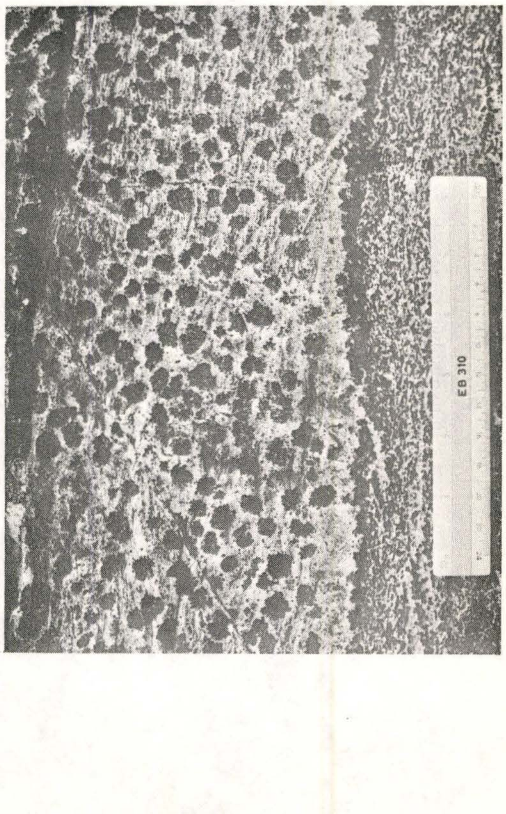


D

Plate 2

A. Agmatite or intrusion breccia of granitic veins and dikes into hornblende metagabbro. The locality is a small gabbroic mass in upper vitreous quartzite north of Wavy Lake near the west border of the Chief Lake batholith.

B. Same locality as Plate 2A. "Feldspathization" of metagabbro--note megacrysts of microcline in dark colored hornblende metagabbro.

C. Xenolith of vitreous quartzite in porphyritic quartz monzonite of the Chief Lake batholith north of Wavy Lake. Note embayments of quartz monzonite parallel to bedding in the inclusion. Foliation in the batholith and xenolith parallels the hammer handle.

D. Small xenolith of feldspathic quartzite in unfoliated porphyritic quartz monzonite north of Chief Lake.

A

B

C

D

Plate 3

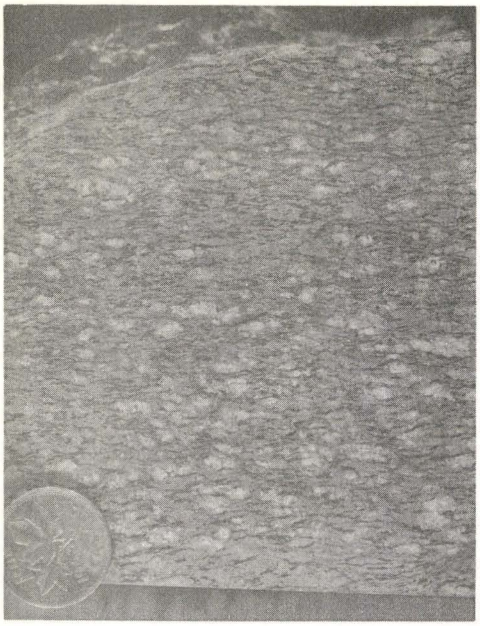
A. Specimen (2132) from mylonitic border of Chief Lake batholith east of Clearwater Lake. Surface is cut normal to foliation and parallel to lineation. Dark streaks on photo are recrystallized quartz surrounding augen-shaped feldspar (perthitic microcline) porphyroclasts.

B. Same as Plate 3A, but cut normal to both foliation and lineation.

C. Augen gneiss (specimen 2116) from Chief Lake batholith northeast of Chief Lake. Dark streaks are recrystallized quartz and plagioclase. Augen are perthitic microcline porphyroclasts.

D. Augen gneiss (specimen 1540) from Chief Lake batholith east of Chief Lake. Large porphyroclasts of perthitic microcline are in a matrix of recrystallized quartz, biotite and plagioclase. Foliation is irregular due to coarse-grained porphyroclasts.

A



B

C

D

Plate 4

A. Passive, similar-type folds in laminated microcline-augen gneiss, west shore of White Oake Lake.

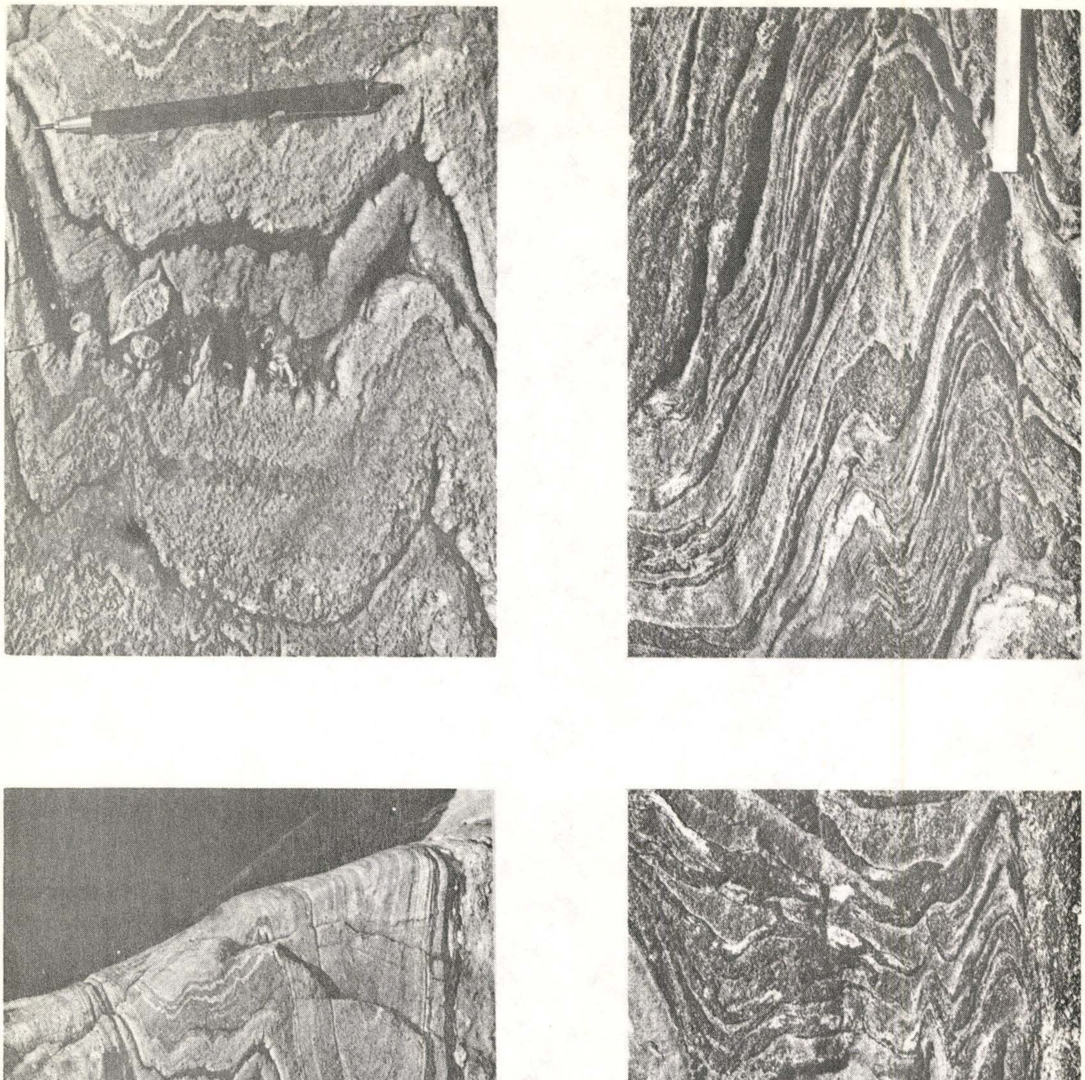
B. Close-up of Plate 4A showing porphyroblastic development of microcline-augen and parallelism of biotite (dark layer) with axial surface in hinge zone.

C. Passive, similar-type folds in laminated amphibolite (dark) and quartzo-feldspathic gneiss (light). Axial plane foliation is not present, but fold axes parallel mineral lineation. Location west of Brodill Lake in migmatite zone.

D. Same locality, compositional and structural relations as Plate 4C.

B

D



A

C

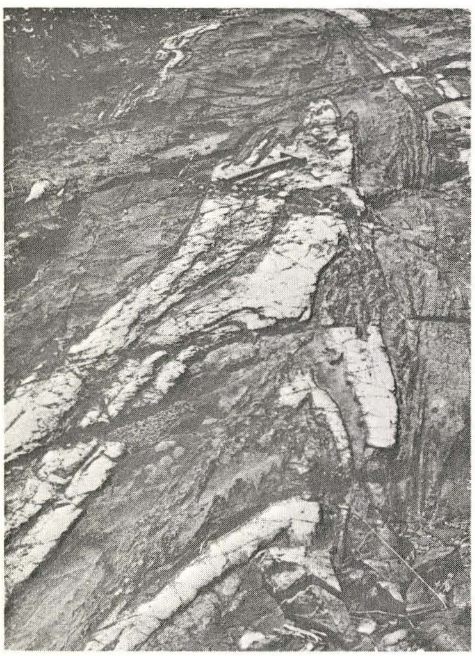
Plate 5

A. Steeply plunging, similar-type fold in quartzite band in migmatite zone along eastern border of Chief Lake batholith.

B. Close-up of hinge zone of fold shown in Plate 5A. Note quartz rods which parallel fold axis.

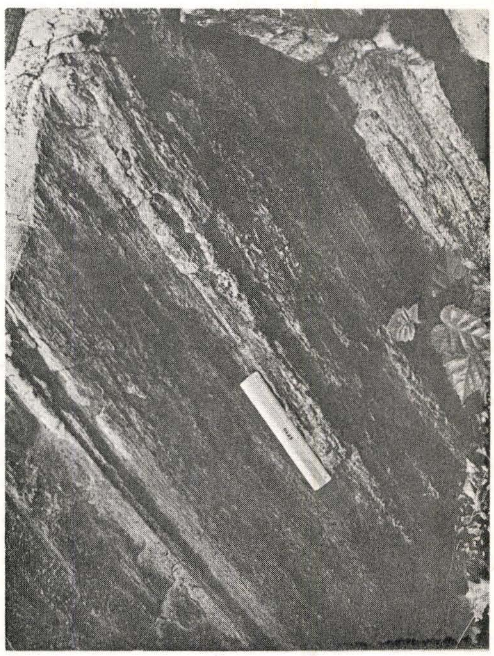
C. Flutes or mullions on gneissosity surface. Foliation is not developed, and the rock is a "lineated tectonite". Location is in the migmatite west of Brodill Lake.

D. Flutes or mullions on bedding surface parallel to mineral lineation of vitreous quartzite. Specimen collected near contact of upper vitreous quartzite and Chief Lake batholith north of Wavy Lake.



A

B



C

D

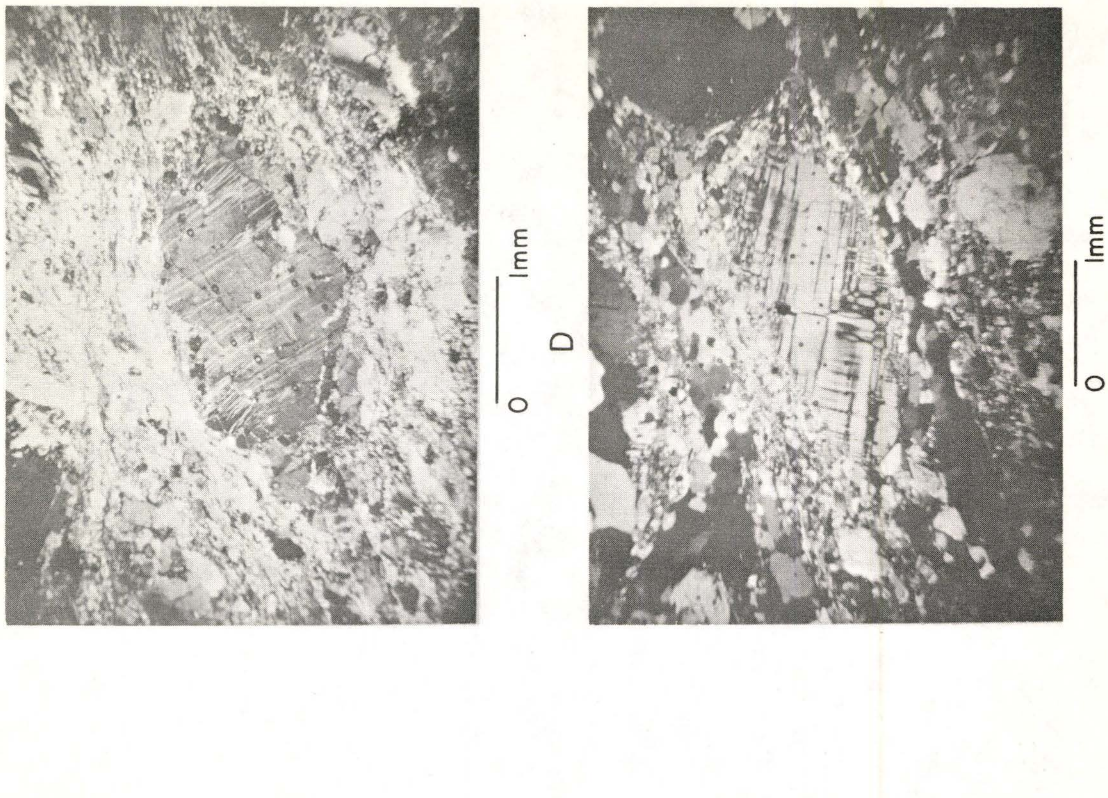
Plate 6

A. Fractured microcline porphyroclasts in mylonitic zone of Chief Lake batholith (specimen 2179) north-east of Crowley Lake along road to Brodill Lake. Fine grained matrix is a mixture of recrystallized quartz and comminuted feldspar. (Polarized light)

B. Fractured microcline porphyroclast in specimen (2132) from mylonitic western border of Chief Lake batholith east of Clearwater Lake. This is same specimen shown on mesoscopic scale in Plate 3A and 3B. (polarized light)

C. Recrystallized quartz, and fractured and comminuted feldspar in foliated and lineated quartz diorite (specimen 1118) from the sheared northwest border of the Chief Lake batholith southeast of Tilton Lake. (polarized light)

D. Microcline porphyroclast surrounded by rim of fine grained feldspar and coarser grained recrystallized quartz (specimen 1540). This is same specimen shown on mesoscopic scale in Plate 3D. (polarized light)


A
0
1mm
B
0
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C
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1mm
D
0
1mm

Plate 7

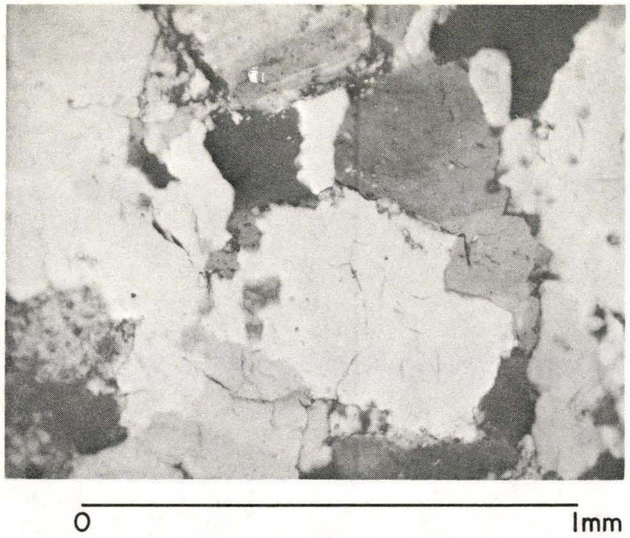
A. Sutured quartz grains showing undulose extinction in specimen (1091) of lower feldspathic quartzite from east shore of Long Lake near mouth of Wavy Creek. (polarized light)

B. Sutured quartz grains showing undulose extinction in specimen (1950) of vitreous quartzite from xenolith in agmatite along northern border of Chief Lake batholith. (polarized light)

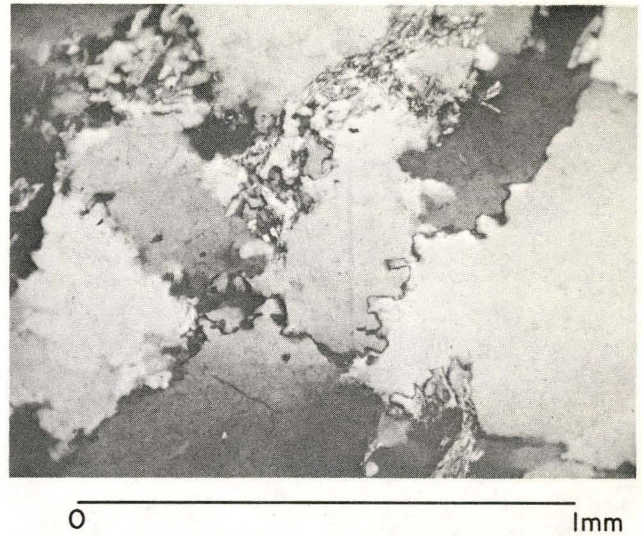
C. Quartz rods in lineated vitreous quartzite specimen (104) from xenolith in Chief Lake batholith along north shore of Wavy Lake. Large grains with undulose extinction bands are thought to be highly strained relic detrital grains (smaller grains are recrystallized quartz). Section is cut parallel to mineral lineation. (polarized light)

D. Recrystallized quartz grains in quartzo-feldspathic gneiss (specimen 2126). Grain boundaries are curved and irregular, and many grains show undulose extinction. Note linear groups of small grains with similar crystallographic orientation. (polarized light)

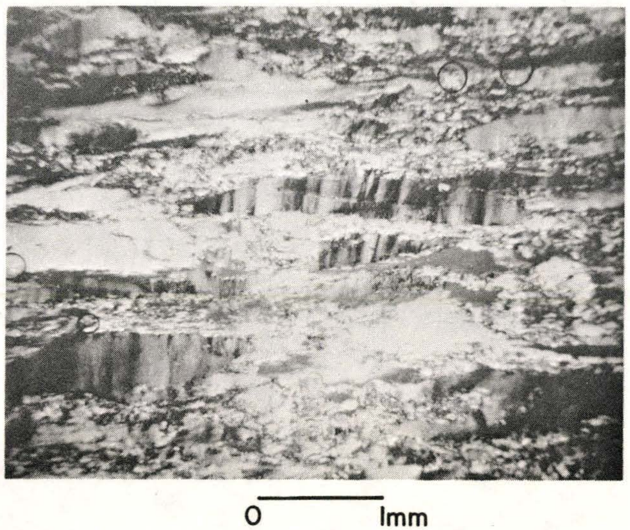
A



B



C



D

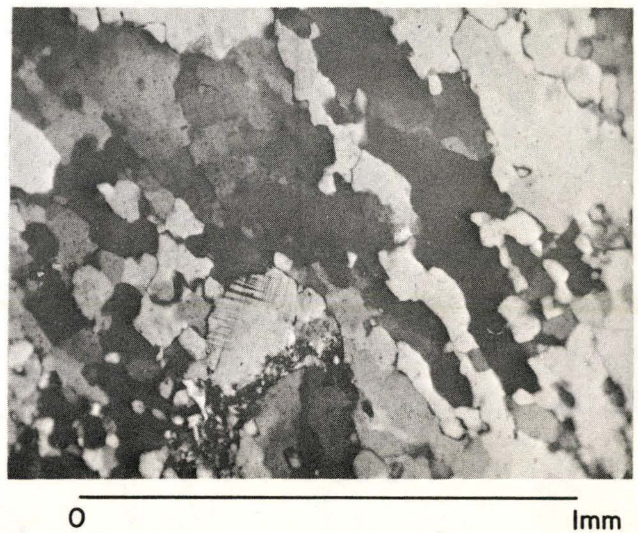


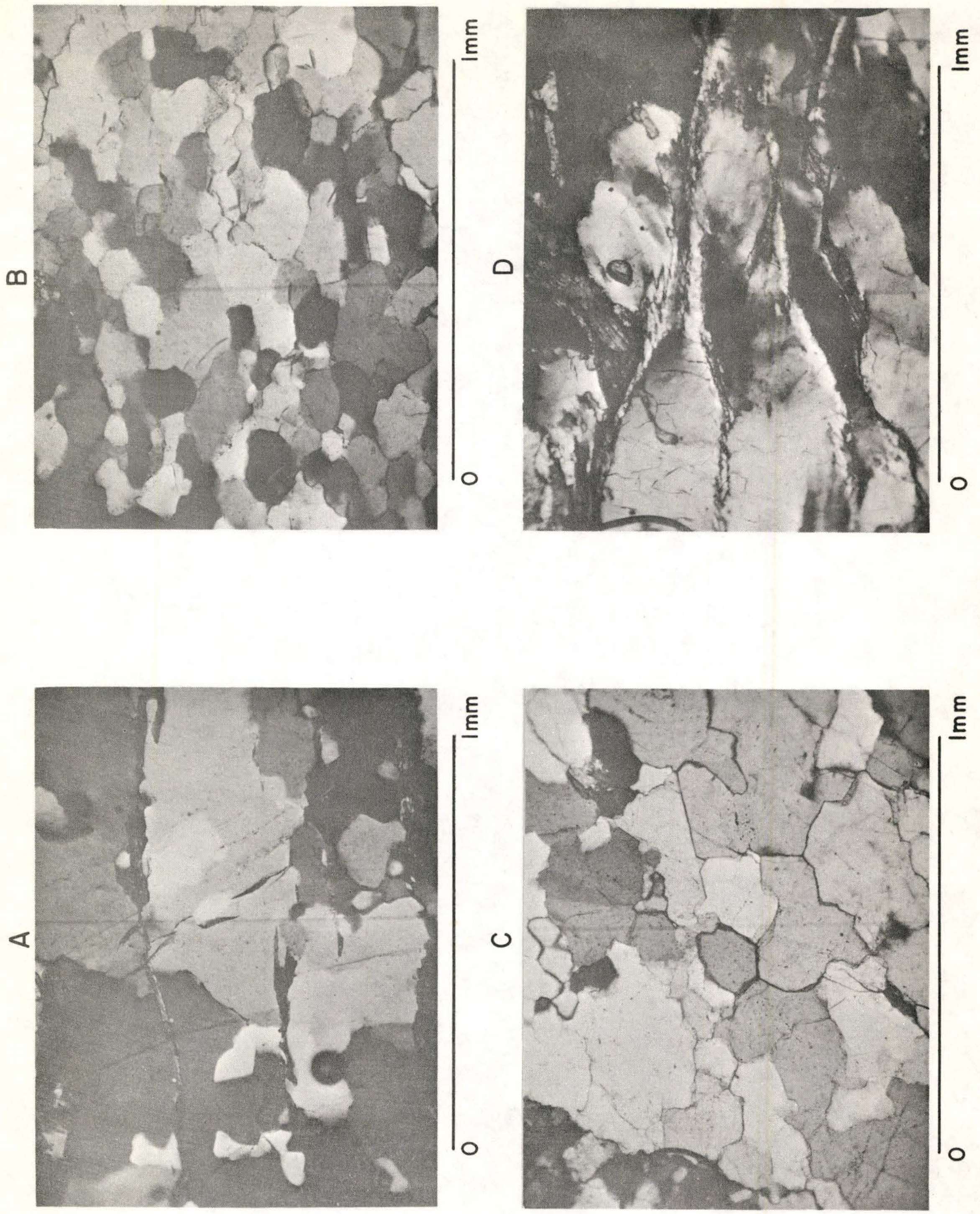
Plate 8

A. Foliated quartz and biotite (horizontal on photo) in quartzo-feldspathic gneiss (specimen 1635) from east shore of White Oak Lake. Note rectangular shape of quartz parallel foliation; biotite parallel foliation occurs as inclusions in quartz grains. Section is cut normal to foliation (no lineation present). (polarized light)

B. Polygonal quartz grains, slightly flattened in plane of foliation (horizontal on photo) in quartzo-feldspathic gneiss (specimen 2180) from contact of batholith and migmatite zone west of Brodill Lake. Section is cut normal to foliation, and oblique to lineation. (polarized light)

C. Polygonal (subhedral) quartz grains in quartzo-feldspathic gneiss (specimen 1829) from north of White Oak Lake. Foliation and lineation in specimen are defined by mica and feldspar, but not by shape of quartz grains. (polarized light)

D. Foliaform, strained relic igneous quartz from eastern border of Eden Lake trondhjemite (specimen 650). (polarized light)



A
B
C
D
0
1mm

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LEGEND
SOUTHERN PROVINCE ROCKS
12 CHIEF LAKE BATHOLITH (QUARTZ DIORITE, QUARTZ MONZONITE) (STIPPLED AREAS CONTAIN FELDSPAR MEGACRYSTS)
11 HORNBLENDE GABBRO-QUARTZ DIORITE AGMATITE
10 QUARTZITE MIGMATITE
EDEN LAKE COMPLEX
9 HORNBLENDE GABBRO-DIORITE
8 TRONDHJEMITE
7 HORNBLENDE GABBRO (SUDBURY GABBRO)
HURONIAN METASEDIMENTARY ROCKS
6 VITREOUS QUARTZITE (LORRAIN QUARTZITE) c. LAMINATED PALE GREEN AND ORANGE QUARTZITE b. MASSIVE PALE GREEN OR WHITE QUARTZITE a. PALE GREEN AND ORANGE QUARTZITE
5 UPPER GRAYWACKE-QUARTZITE a. CONGLOMERATIC GRAYWACKE (GOWGANDA FORMATION)
4 MIDDLE (?) FELDSPATHIC QUARTZITE (SERPENT QUARTZITE ?)
3 MIDDLE CONGLOMERATIC GRAYWACKE (BRUCE CONGLOMERATE)
2 LOWER FELDSPATHIC QUARTZITE a. GRAYWACKE (MISSISSAGI FORMATION)
1 LOWER GRAYWACKE a. CONGLOMERATIC GRAYWACKE (McKIM-RAMSAY LAKE FORMATIONS)
B UNCORRELATED QUARTZITE
A UNCORRELATED GRAYWACKE
GRENVILLE PROVINCE ROCKS
T MASSIVE AMPHIBOLITE (INCLUDES SOME PYROXENE AMPHIBOLITE)
QUARTZO-FELDSPATHIC GNEISSES
S MASSIVE BIOTITE, GARNET GNEISS MIGMATITIC IN PART
R LAMINATED BIOTITE, GARNET, SILLIMANITE GNEISS
Q RUSTY WEATHERING BANDED BIOTITE GNEISS
P MICROCLINE AUGEN GNEISS WITH FELSIC AND MAFIC LAMINAE
GARNET ISOGRAD
X DIABASE DIKES
GEOLOGIC MAP
Figure 2
MAGNETIC
TRUE NORTH
8°
0 1/2 1 2 miles
STRUCTURAL SYMBOLS
CONTACT, EXPOSED OR GOOD CONTROL
CONTACT, POOR CONTROL
SHEAR ZONE
FAULT
STRIKE AND DIP OF BEDDING (S_0), FACING IN DIRECTION OF DIP
STRIKE AND DIP OF OVERTURNED BEDDING
STRIKE AND DIP OF BEDDING, FACING DIRECTION UNKNOWN
STRIKE OF VERTICAL BEDDING
STRIKE AND DIP OF GNEISSOSITY (S_g)
STRIKE OF VERTICAL GNEISSOSITY
STRIKE AND DIP OF FOLIATION (S_1)
STRIKE OF VERTICAL FOLIATION
HORIZONTAL PROJECTION AND PLUNGE OF FOLD AXIS (L)
HORIZONTAL PROJECTION AND PLUNGE OF MINERAL LINEATION (L_1)
VERTICAL MINERAL LINEATION
LONG LAKE
Lohi Lake
CLEARWATER LAKE
TILTON LAKE
CHIEF LAKE
Crowley L.
Linton L.
BRODILL L.
Bluff Lake
WHITE OAK LAKE
WAVY LAKE
Wavy Creek
LUKE LAKE
EDEN LAKE
Bazi L.
Sudbury
545
69
to Parry Sound
WATERS
BRODER
DILL
TILTON
SECORD
EDEN
BEVAN
HALIFAX
LAURA
46° 25'
46° 20'
46° 15'
81° 10'
81° 05'
81° 00'
80° 55'
0 2 4 miles
TOWNSHIP LOCATIONS

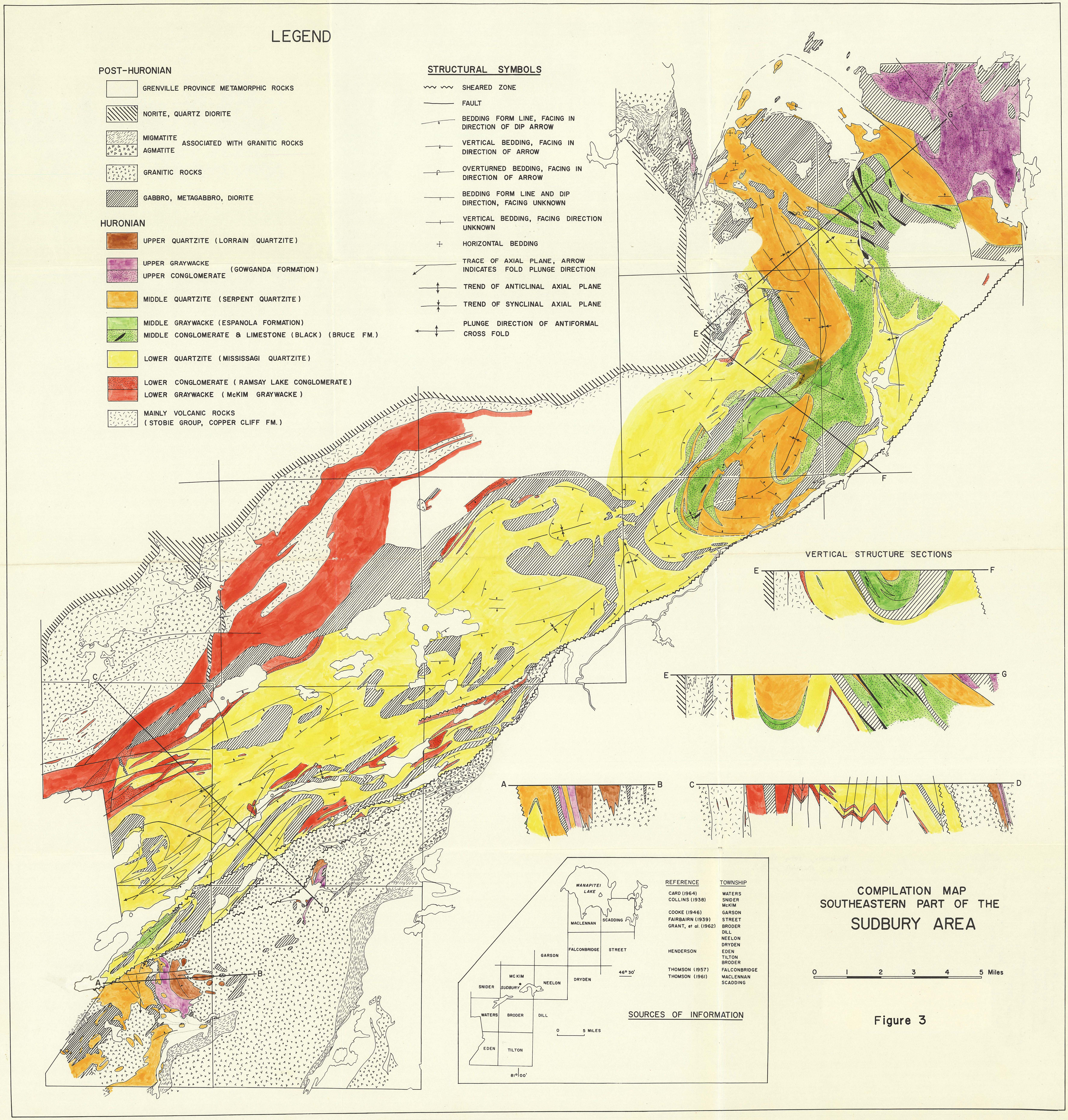
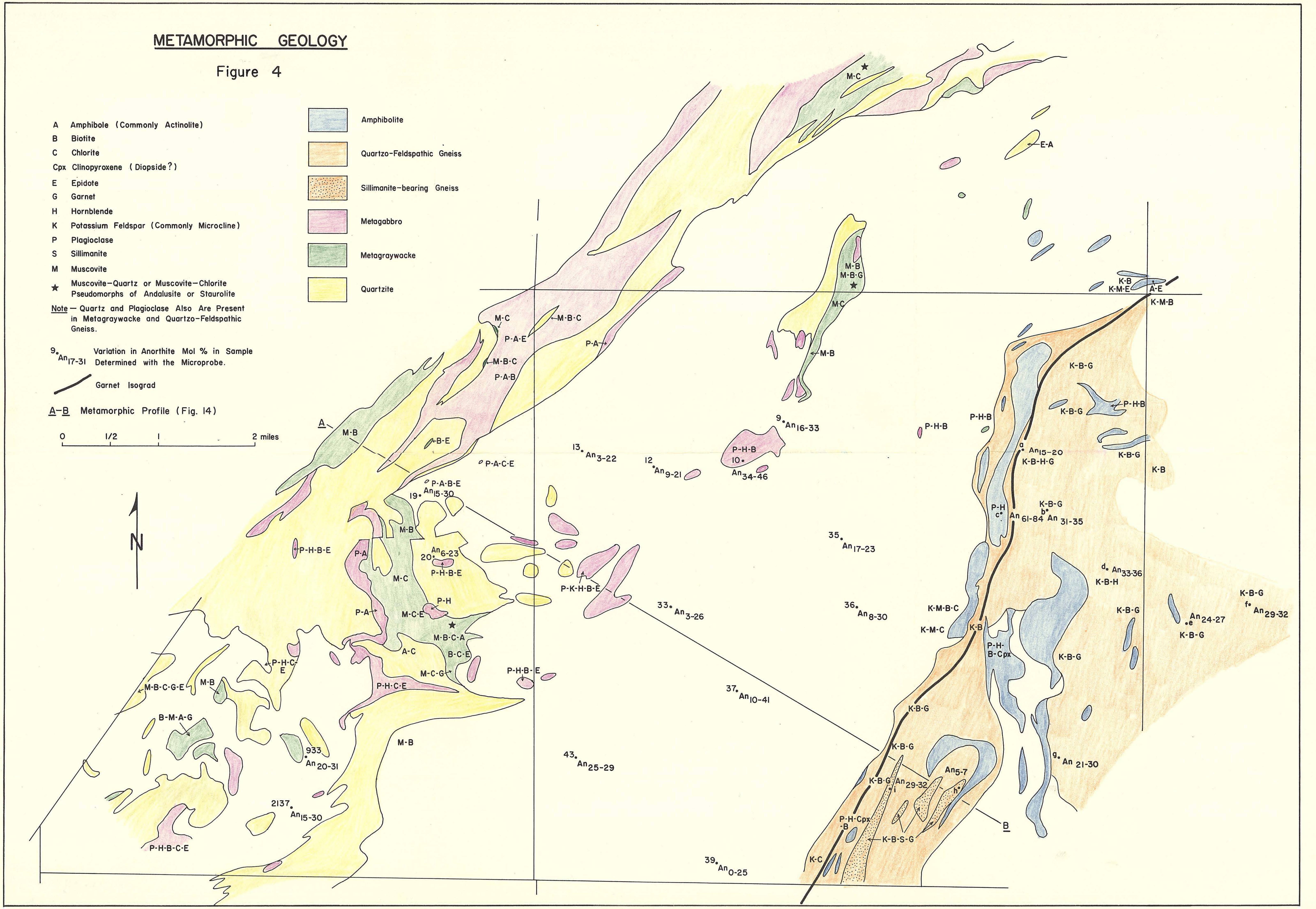
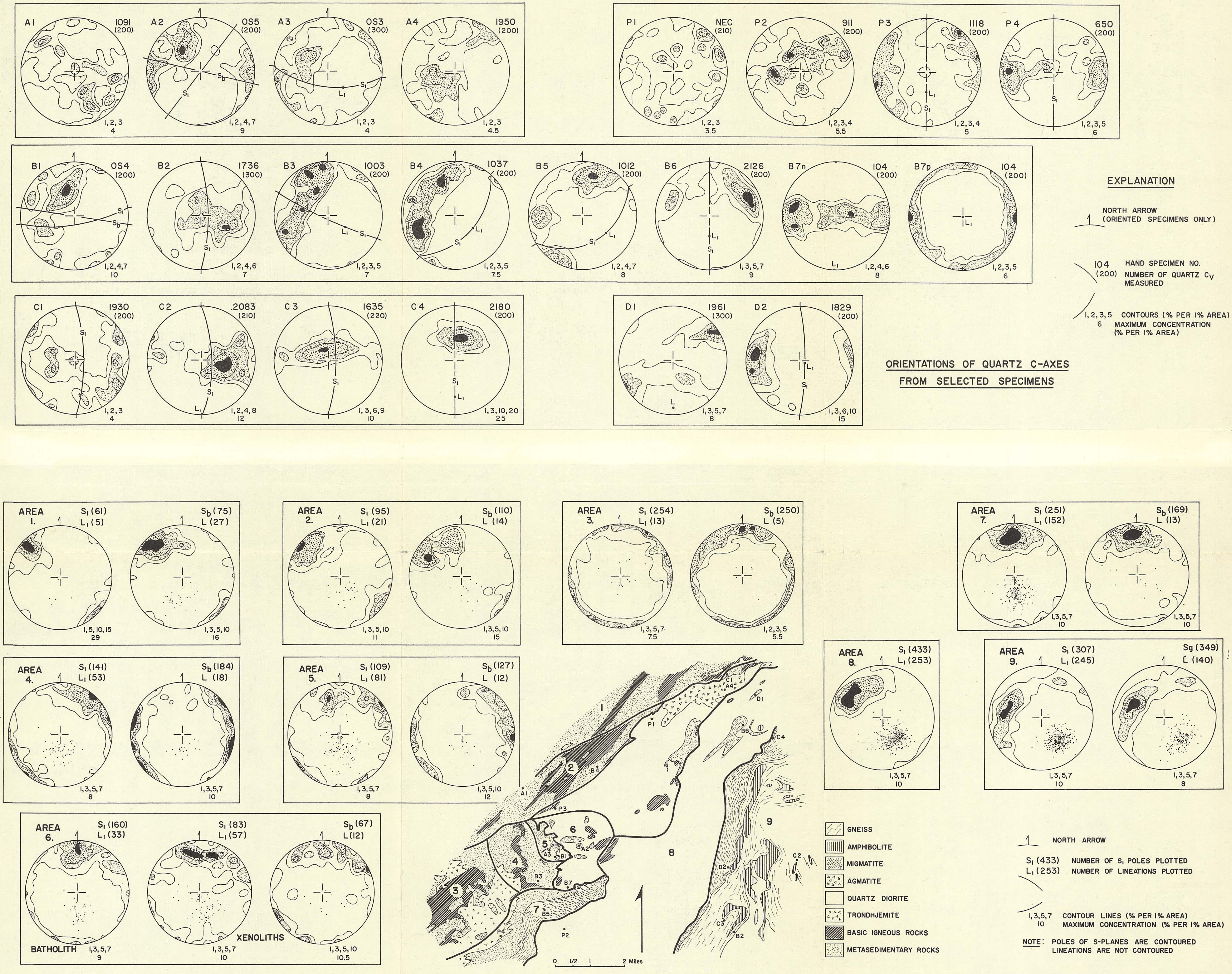


Figure 3

Fig 3


METAMORPHIC GEOLOGY
Figure 4
A Amphibole (Commonly Actinolite)
B Biotite
C Chlorite
Cpx Clinopyroxene (Diopside?)
E Epidote
G Garnet
H Hornblende
K Potassium Feldspar (Commonly Microcline)
P Plagioclase
S Sillimanite
M Muscovite
★ Muscovite-Quartz or Muscovite-Chlorite Pseudomorphs of Andalusite or Staurolite
Note — Quartz and Plagioclase Also Are Present in Metagraywacke and Quartzo-Feldspathic Gneiss.
9. An 17-31 Variation in Anorthite Mol % in Sample Determined with the Microprobe.
Garnet Isograd
A-B Metamorphic Profile (Fig. 14)
0 1/2 1 2 miles
N
Amphibolite
Quartzo-Feldspathic Gneiss
Sillimanite-bearing Gneiss
Metagabbro
Metagraywacke
Quartzite
A
B



PETROFABRIC DIAGRAMS

Figure 5

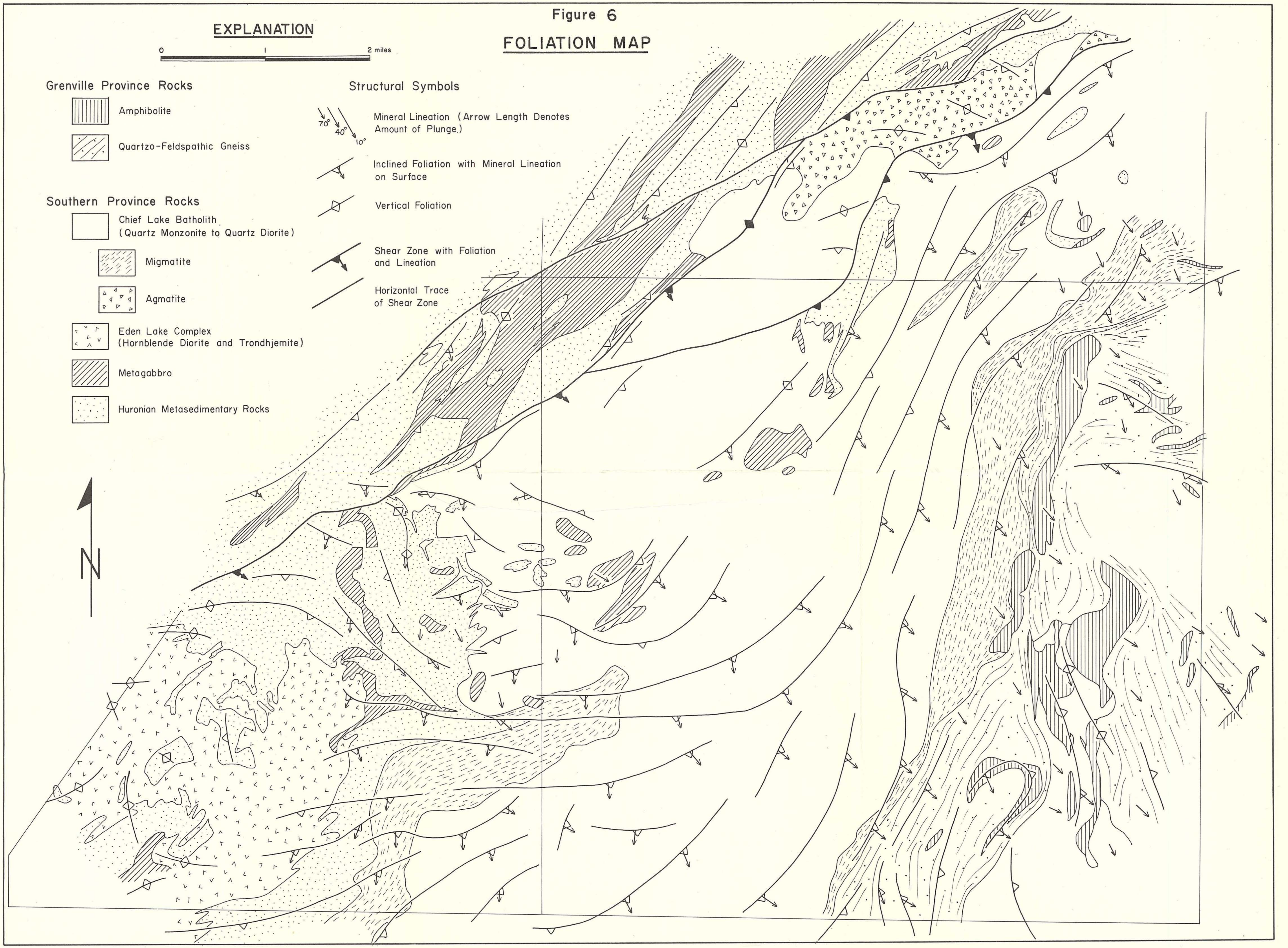
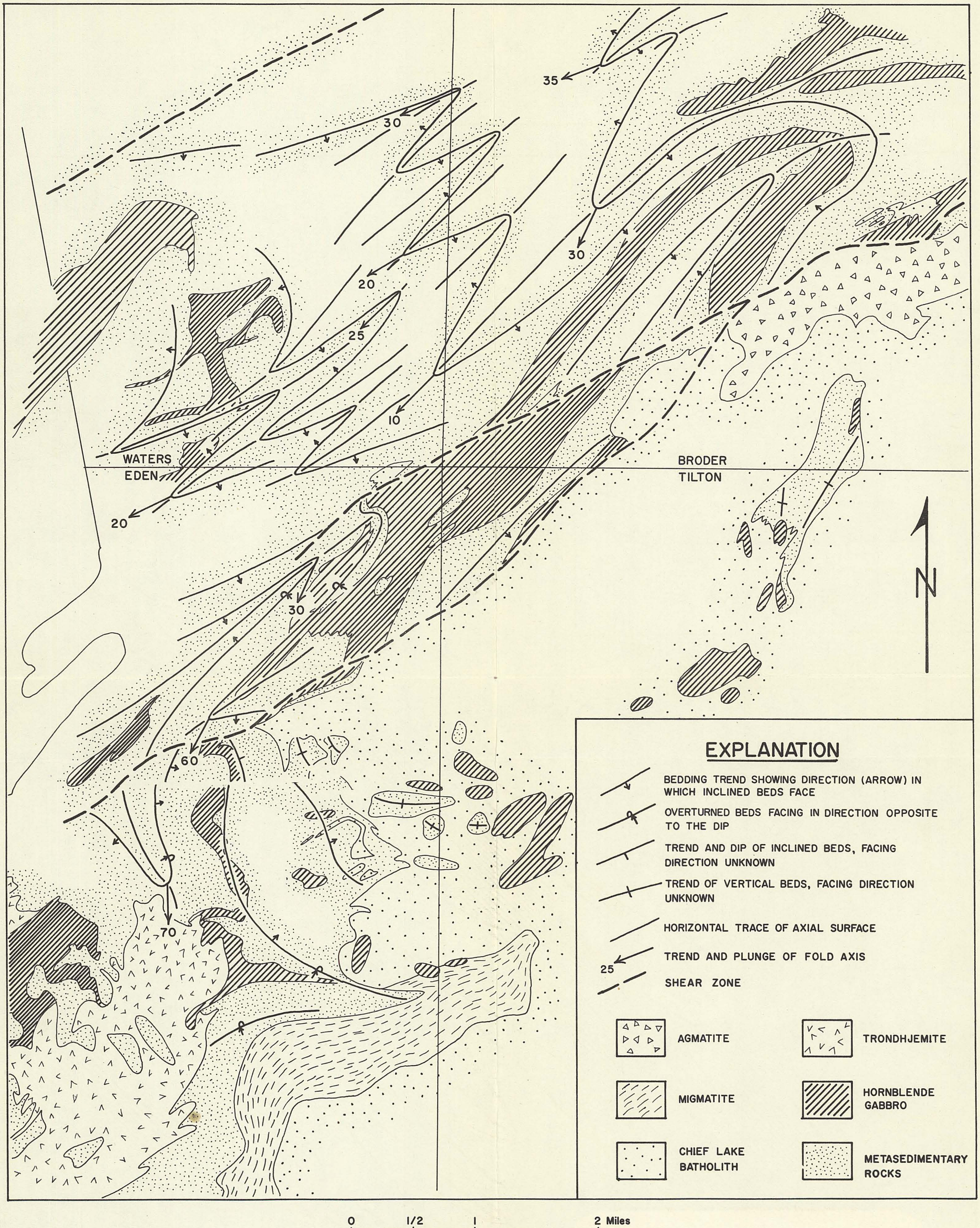


Figure 6

FOLIATION MAP



BEDDING MAP

Figure 7