STRUCTURAL STUDY OF KAKAGI LAKE AREA

NORTHWESTERN ONTARIO

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by

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TITLE: Structural Study of Kakagi Lake Area,
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SCOPE AND CONTENTS:

An area situated near Nestor Falls, Northwestern Ontario was geologically mapped, described, and structurally analyzed.

Rock specimens were sectioned to provide additional information on the petrology and metamorphic geology.

Six structural domains outlined in the area are examined using data obtained in the field. Many effects of both ductile and brittle deformation were recognized and are described. Mesoscopic and macroscopic structures were analyzed when sufficient data was available. At least three generations of folding were recognized and are described. Finally, an interpretation of the successive sequence of geologic events is given.

ABSTRACT

Volcanic rocks and associated intrusives in the Kakagi Lake area display both ductile and brittle deformation. Three different phases of deformation are recognized. The scale of folding associated with these phases decreases through time from large E-W trending macro-folds in the north to small scale kink bands in the south-west. These different phases have resulted from three different directions of maximum compressive stress.

The first phase of deformation probably affected the entire area. It is reflected in large E-W trending macro-folds plunging steeply to the east and has associated with it an axial plane schistosity. Mesoscopic structures (ptygmatic and boudinaged quartz veins) indicate one principal shortening normal to the axial plane and one principal extension in this plane. Once the shortening reached a critical value 'brittle deformation' (ductile faulting?) occurred on both a mesoscopic and macroscopic scale. The point where deformation shifted from a ductile regime to a brittle one varied with rock lithology.

A second phase of folding and schistosity development is evident in the south-west. Maximum compressive stress acted in a N.E. - S.W. direction and gave rise to mesoscopically folded quartz bodies and lamprophyre dykes as well as the second schistosity.

A change in the maximum principal stress direction to N.N.W. has resulted in the formation of kink bands on the south-west shore of Kakagi Lake. These bands have formed by kinking of the late

schistosity planes produced during the second phase of deformation.

Most of the area has undergone a low greenschist regional metamorphism but in the south-west a higher grade metamorphic zone (biotite zone) is evident. This zone is thought to represent the outer edge of a contact aureole lying north of the granitic batholith, the north edge of the batholith being less than two kilometres to the south. The structural geology, especially in the south and south-west of the map area, appears to be genetically related to this granitic body to the south.

ACKNOWLEDGMENTS

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CHAPTER 1

INTRODUCTION

Several volcanic-sedimentary belts of Keewatin age are found in Northwestern Ontario. A characteristic feature of many of these assemblages is that they are bordered by large batholiths ranging in composition from granite to diorite. A question that has been asked about these volcanic belts is whether or not their structural geology is genetically related to the adjacent granitic intrusives.

The Kakagi Lake area is only one of several areas in Northwestern Ontario where structurally deformed volcanic-sedimentary assemblages are found bordered by large granitic batholiths. During the summer of 1970 the author spent three months studying the geology and in particular the structural geology of an area situated in and around Kakagi Lake in Northwestern Ontario. The particular area was chosen for investigation because of its location adjacent to a large granitic batholith in an accessible region. The major objectives of the field work were to map the geology in as much detail as possible and to describe the structural geology of the area.

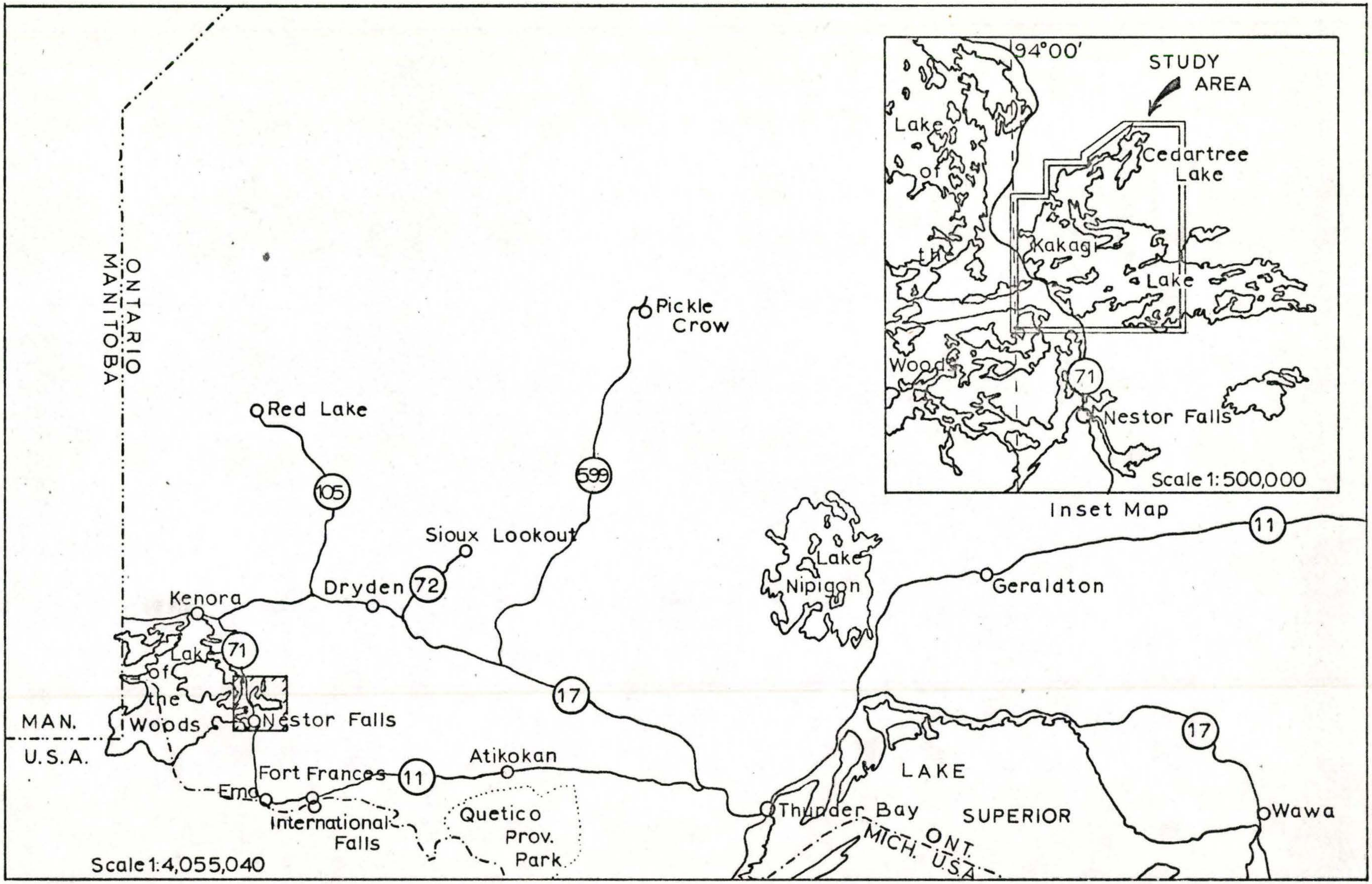
Location and Access

The study area (Figure 1) covers Cedartree Lake and the western part of Kakagi Lake. Kakagi Lake is situated in northwestern Ontario just east of Sabaskong Bay on Lake of the Woods at Lat. 49^{o} 15'N and Long. 93^{o} 55'W. Excellent roads provide access to the study area. Highway 17 crosses within 50 miles to the north and Highway 71, joining Kenora on the north and Ft. Frances on the south, passes along the west side of Kakagi Lake.

Previous Work

Although A.C. Lawson (1885) did a general reconnaissance of the area, the first detailed study was by E.M. Burwash in 1930 - 31 (Burwash, 1933). He furnished a geological map of the area at a scale of 1 inch to 1 mile. A.M. Goodwin did a regional study in 1963 - 64 which included Kakagi Lake area (Goodwin, 1965). The resulting map at a scale of 1 inch to 4 miles shows the general overall structure in the Kakagi Lake area. Map 2115, a compilation map of the Ontario Department of Mines published in 1967 at a scale of 1 inch to 4 miles, also outlines the general geology of the area. Ridler (1966) did a petrographic study of one of the ultrabasic sills within Kakagi Lake.

Unpublished data of numerous mineral exploration companies has been filed in the past with the Kenora office of the Ontario Department of Mines.



Location Map
Figure 1

Field Methods

Aerial photographs at a scale of approximately 4 inches to 1 mile were used extensively during the field work. Outcrops were outlined on the photographs both prior to entering the field and while in the field.

Several traverses were made inland where it was thought the most information could be obtained. However, detailed examination of shoreline exposures provided most of the data. Due to the overburden and lichen cover exposures away from the shoreline provide less information and of a lower quality.

The data was recorded for each outcrop outlined on the photographs. Some of it was then transposed to base maps at a scale of 4 inches to 1 mile.

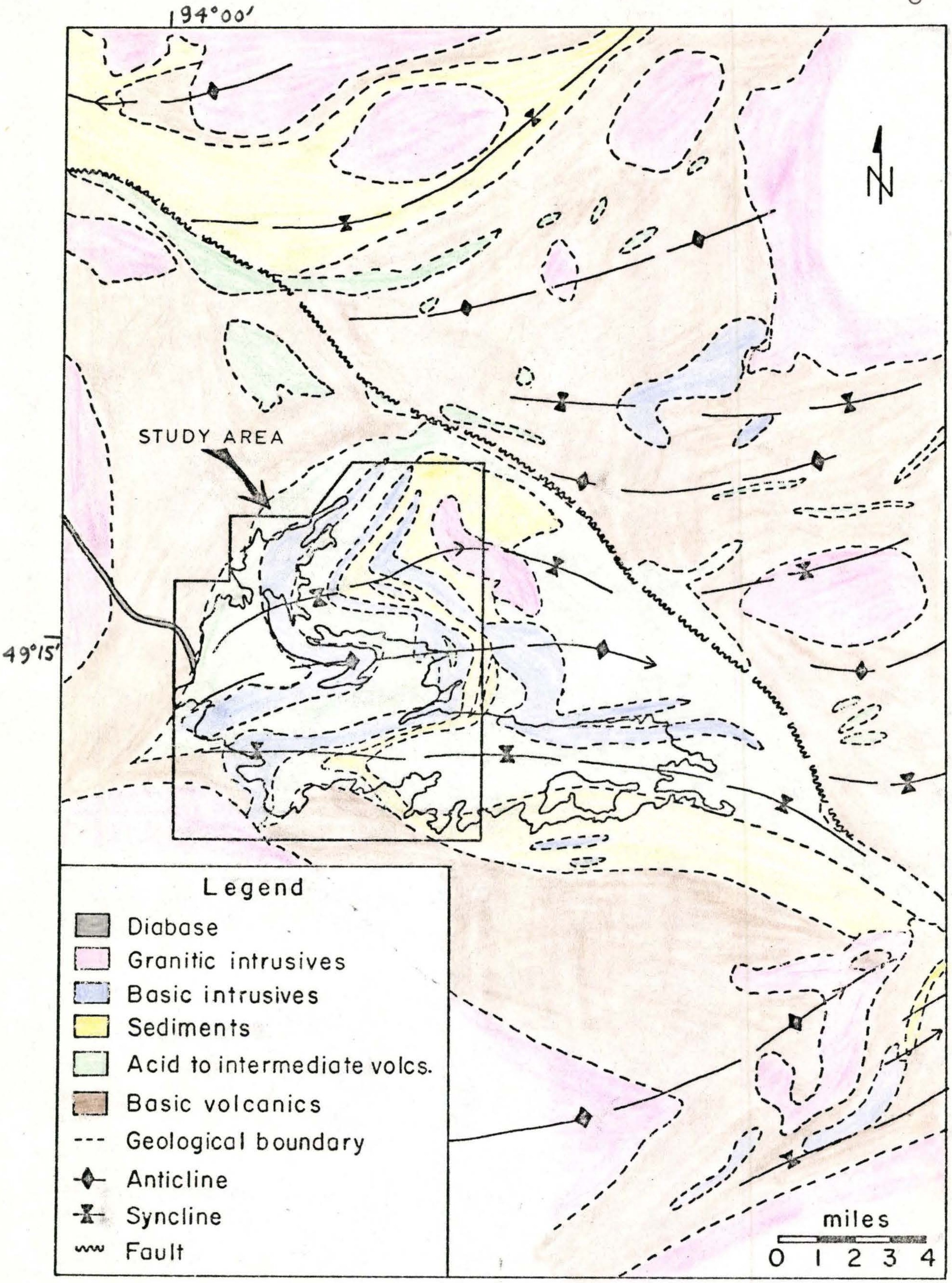
General Geology

The bedrock of the area is Precambrian. Basic and acid volcanic rocks and the overlying sediments comprise the Keewatin 'series' of the area. Younger Precambrian rocks are intrusive (Figure 2).

The Keewatin rocks are characterized by widespread and local overlapping, interchanging, and gradations between volcanics and sediments as well as between the different types of volcanics and of sediments. A picture of volcanic and sedimentary materials being deposited contemporaneously has been recognized by many workers starting with Lawson (1885).

Goodwin (1965) recognized a periodic or cyclic arrangement of lithologic types in the area. The lithologic parts of a cycle, arranged in ascending order, are: (1) basic volcanics, (2) acid volcanics, and (3) sediments (Goodwin, 1965).

As seen in Figure 2, many basic intrusive sills, dykes and irregular masses are present; the composition of which ranges from peridotite to grano-diorite. The basic sill in the west part of Kakagi Lake areawas described by Burwash (1933) as Haileyburian in age and 'subsequent to the shearing of the Keewatin ... and to the intrusion of the Laurentian gneisses'. Ridler (1966) considered this particular sill to be a differentiated stratiform body intruded in two pulses of magma contemporaneous with the host volcanic rocks and before any major structural deformation.



General Geology of the Area
(after Goodwin, 1965)

Figure 2

Large acid intrusions form regional batholithic masses marginal to the volcanic - sedimentary assemblages. Small restricted stocks and bosses are found enclosed within the Keewatin rocks. The batholithic masses are composed of a variety of rock types varying from granitic at their centre, to dioritic at their outer edge (Goodwin, 1965).

The intensity of metamorphism in the area is low (green schist), and original features of the rocks are usually well preserved.

Structure of the region according to Goodwin (1965) is characterized by the superposition of large intrusive masses, upon pre-existing east-west linear fold patterns. In the Kakagi Lake area major fold axes plunge steeply to the east. A south-east trending lineament (fault?) east of Kakagi Lake separates east-facing, open folded strata to the west (i.e. in the study area) from west-facing, isoclinally folded strata to the east.

CHAPTER 2

STRATIGRAPHY and PETROGRAPHY

General Statement

Geology of the study area as mapped by the author is given as Figure 3. The general stratigraphy of the area consists of a volcanic sequence that displays a progressive transition from basic to more acidic rocks. They are considered to be of Keewatin age (Burwash, 1933). Nearly concordant with these volcanic rocks are a number of basic intrusive units which show varying amounts of differentiation from peridotite to granophyre. These intrusive rocks are considered to have intruded the volcanic pile nearly contemporaneously with the volcanism (Ridler, 1966). They vary greatly in thickness from less than 50 metres to more than one kilometre. While the small intrusive bodies can only be traced for a short distance, two of the large ones traverse the entire map area (Figure 3). Both the volcanic rocks and the basic sills have been cut by numerous minor discordant intrusives, consisting both of black lamprophyre and quartz porphyry. Minor granitic bodies have also intruded into the volcanic sequence. A late north-west trending diabase dyke has been traced for over a kilometre in the most westerly part of the map area.

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Cuddy, Robert Graham.
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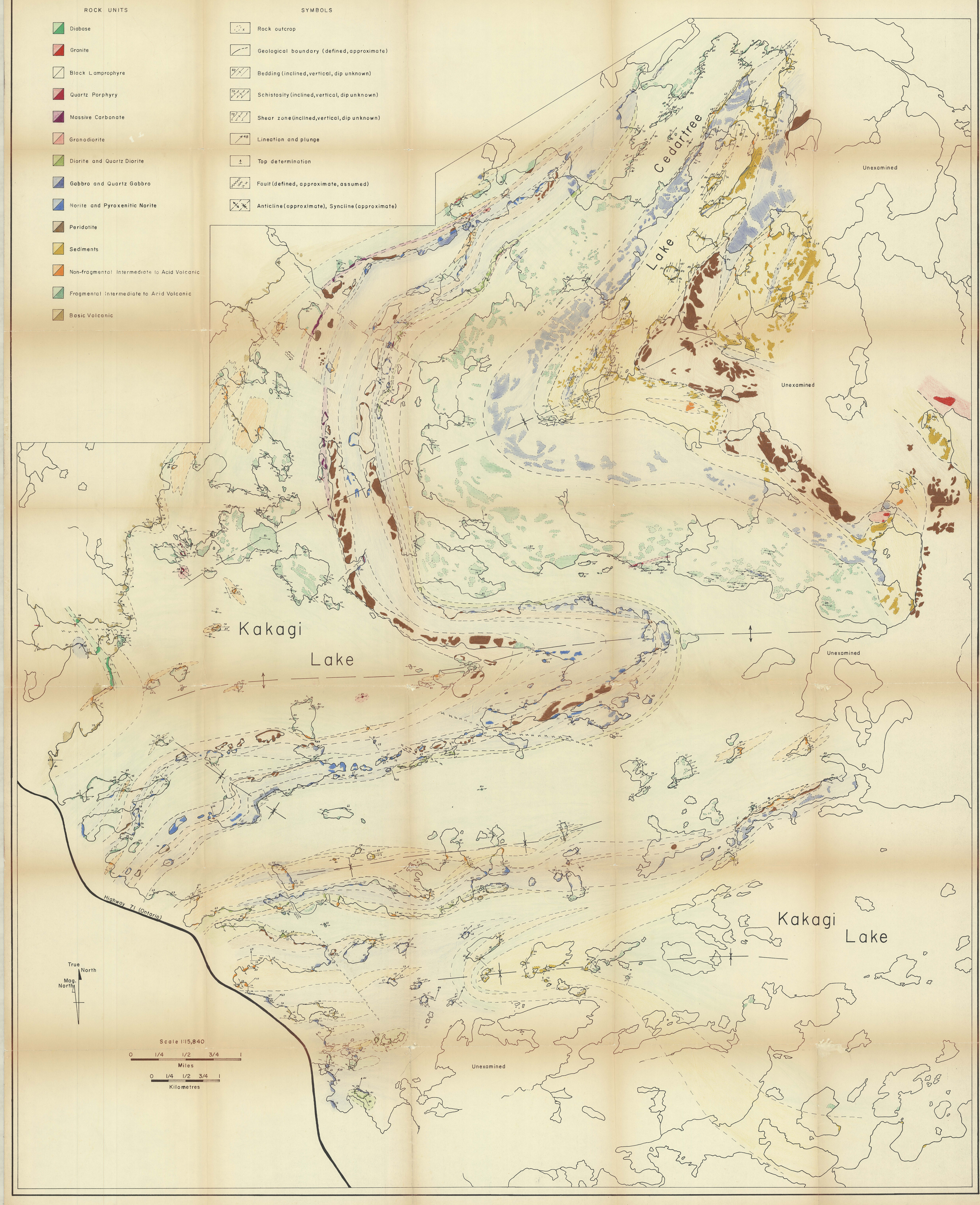
Figure 3
Geology of Kakagi Lake - Cedartree
Lake Area

GEOLOGY OF KAKAGI LAKE-CEDARTREE LAKE AREA

NORTHWESTERN ONTARIO



Basic Volcanics

Basic volcanics are confined mainly to the west shore of Kakagi Lake (Figure 3) and are thought to be the oldest rocks in the area. These rocks were considered by Burwash (1933) to belong to the Lower Keewatin as defined by Lawson (1885). The basic lava flows exhibit pillow structures (Plate 1) and vesicular tops. In places the lavas are amygdaloidal. The pillow structures indicate that at least some of the lava has been deposited in an aqueous environment. These structures, as well as the scoraceous and amygdaloidal parts of the flows (Plate 2), indicate direction of tops.

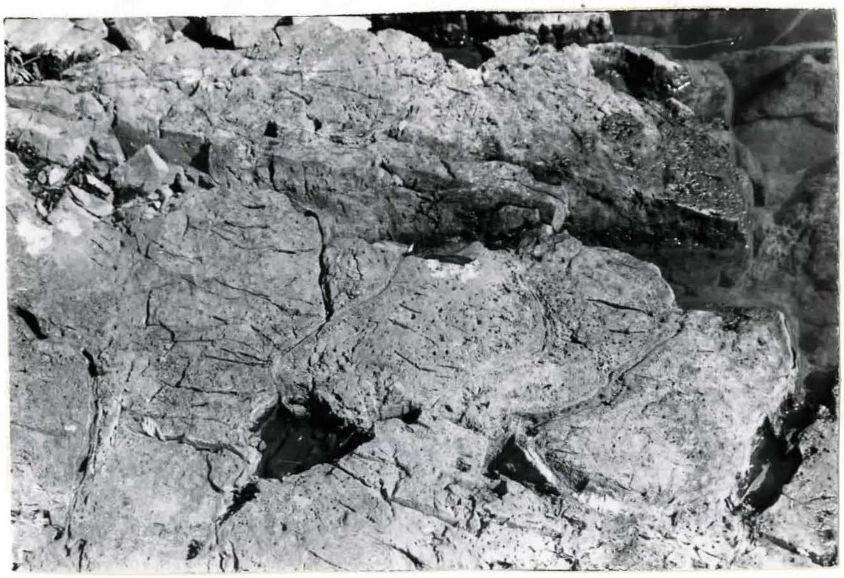
The basic lavas are, as recognized by Burwash (1933), mainly andesitic in composition. Few basaltic rocks were noted in the field. The feldspar is generally andesine and green hornblende is ubiquitous in all thin sections examined. In addition much chlorite, biotite and pyroxene, the composition of which was not determined, was often present.

PLATE 1.

Pillow basalt on the west shore of Kakagi Lake. Pillows are 10 - 12 inches in width and 18 - 24 inches in length on the outcrop surface.

PLATE 2.

Contact in the basic volcanics. Note numerous vesicles and amygdales near the top of the flow and the 'drag' near the contact caused by the flowage of the overlying lava. Stratigraphic top faces 015°. Compass points N.



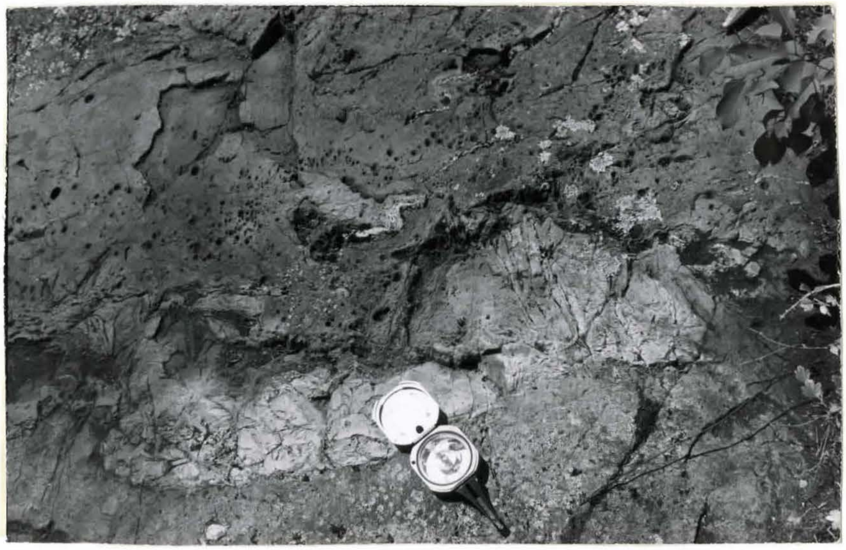
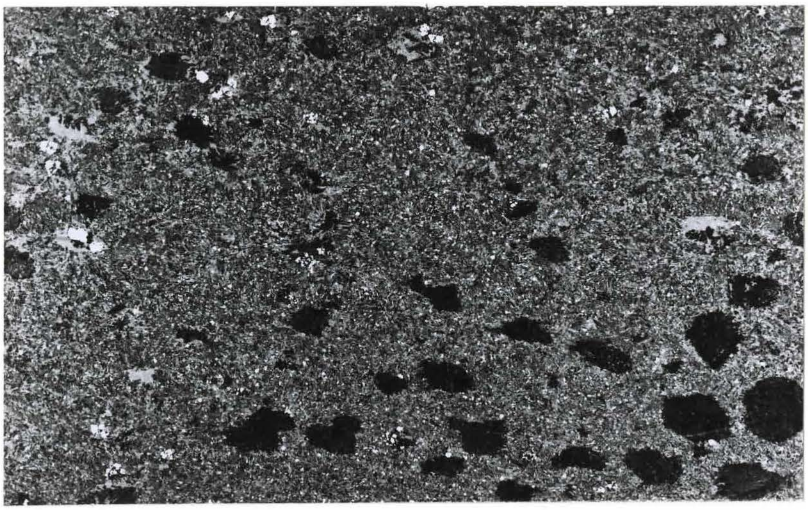


PLATE 3.

Calcite amygdales in pillow basalt on west side of Kakagi Lake. Photograph of thin section (Plane light).

PLATE 4.

Close-up of Plate 3 showing a single amygdale consisting of calcite displaying rhombohedral cleavage. (X-Nicols).



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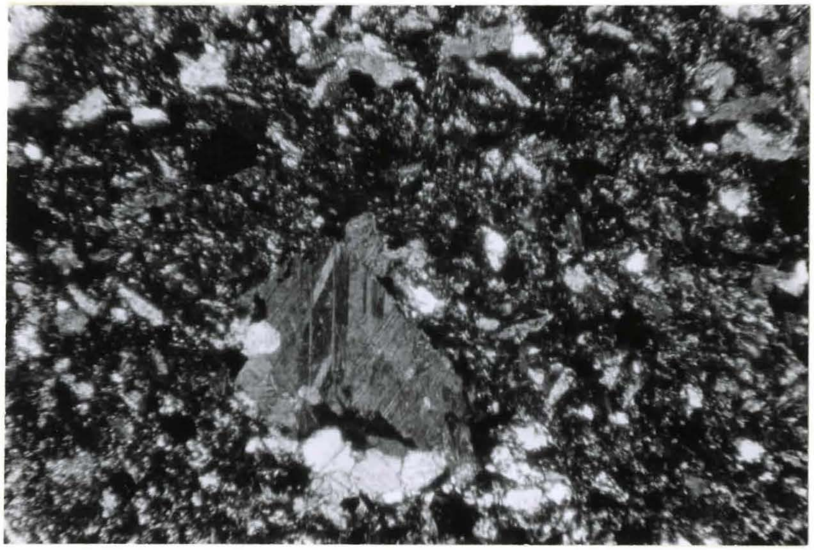

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Fragmental Intermediate to Acid Volcanics

Grouped under the broad term of fragmental volcanic rocks are all those volcanic rocks in which fragments are readily apparent. The fragmental rocks constitute more than 50 percent of all the volcanics exposed in the study area.

For the most part they show very little stratification or layering of any kind. In places, however, one is able to determine primary layering in the otherwise unstratified massive rocks. This layering is usually recognized by graded fragment size; a graded sequence from coarse fragments at or near the bottom of a unit to finer and in places more tuffaceous, well banded fragments near the top (Plate 5).

One of the best exposures of the massive fragmental unit is found near the north end of Cedartree Lake (Plates 7,8). In many places the fragmental nature of the rock is obscured by lichen cover.

The fragments generally show little sorting (Plate 7) and vary in size to greater than a metre in diameter (Plate 6). The two-dimensional shape of individual fragments varies from well rounded to extremely angular. Many of the fragments are elongate in one direction (Plate 10). Although in certain exposures the long axis of adjacent fragments varies considerably (Plates 6, 7); in most exposures the long axis of fragments in two dimensions does exhibit a definite trend. More will be said later concerning this general orientation of fragments.

PLATE 5.

Crude bedding observed in the fragmental acid to intermediate volcanic unit.

PLATE 6.

Typical shoreline exposure of fragmental volcanic rock. Note two large fragments with apparent long axis about normal to one another.

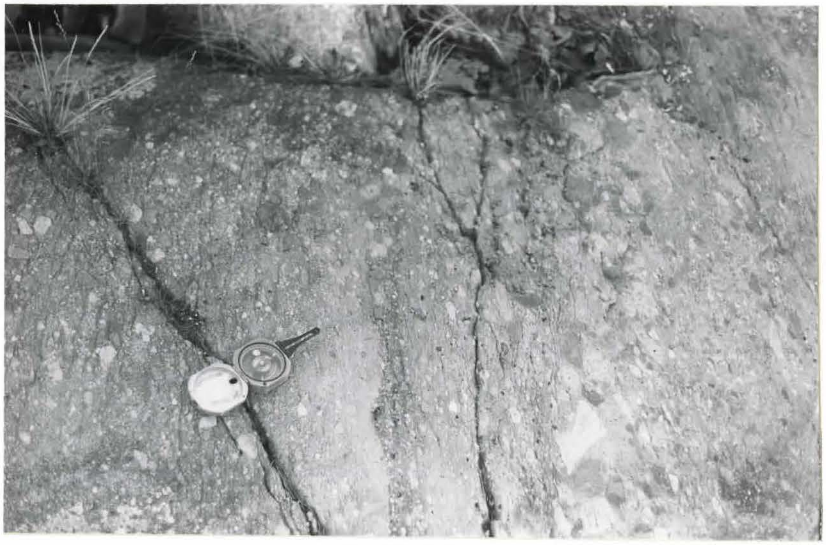
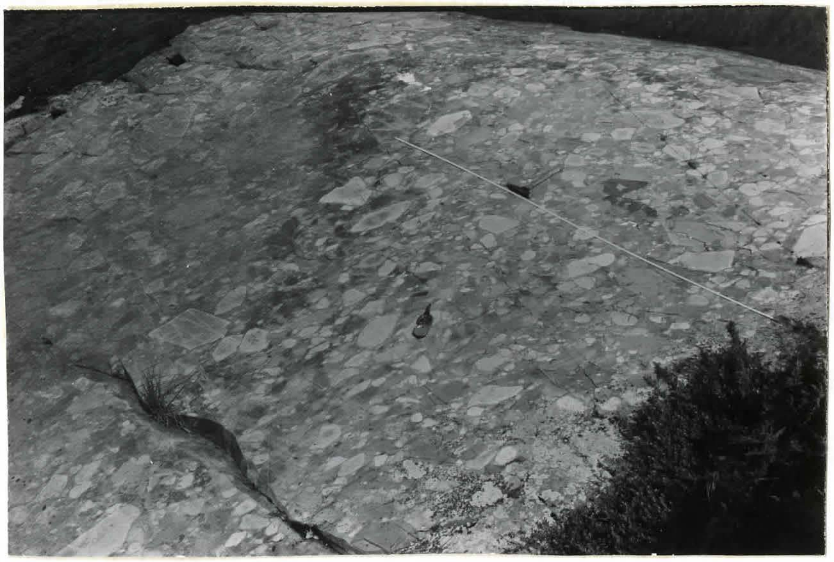


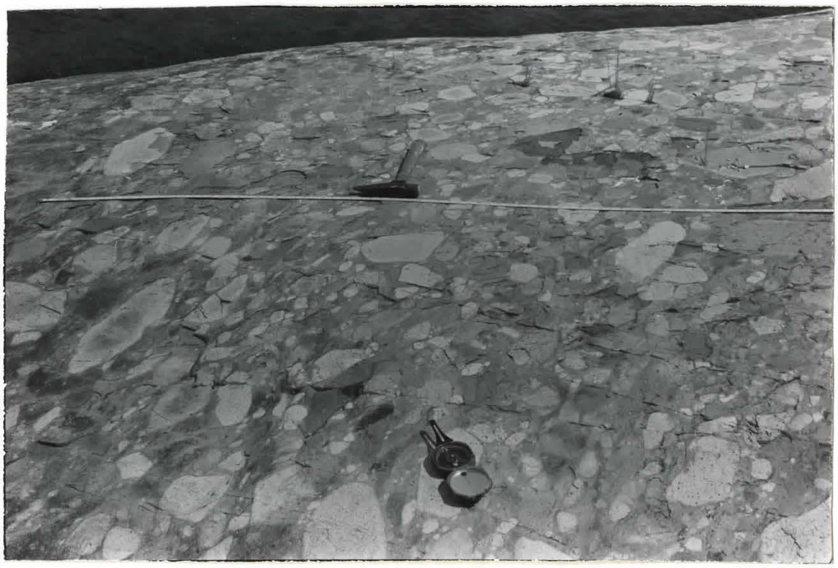
PLATE 7.

Excellent exposure of fragmental volcanics near north end of Cedartree Lake.

PLATE 8.

Close-up view of above exposure.





The density of fragments to matrix varies from outcrop to outcrop, as well as within a single exposure. Where the density of fragments to matrix is high, the finer (tuffaceous) material weathers more readily than the fragments leaving the latter in relief. Usually, however, there is no preferential weathering. The fragments show differing shades of colour and vary in texture; some are noticeably porphyritic while others appear tuffaceous. On first appearance all the fragments seem to be a lighter colour than the matrix but upon closer examination one finds many which blend in with the matrix as to colour and texture. In fact, numerous fragments are darker in colour than the adjacent matrix (Plate 8), and often have less well defined boundaries than those of a lighter colour. This is probably due to their difference in physical properties with respect to temperature and pressure. Some of the fragments display evidence of being quite brittle at the time of their deposition with matrix material subsequently flowing in along fractures and isolating small angular fragments (Plate 9).

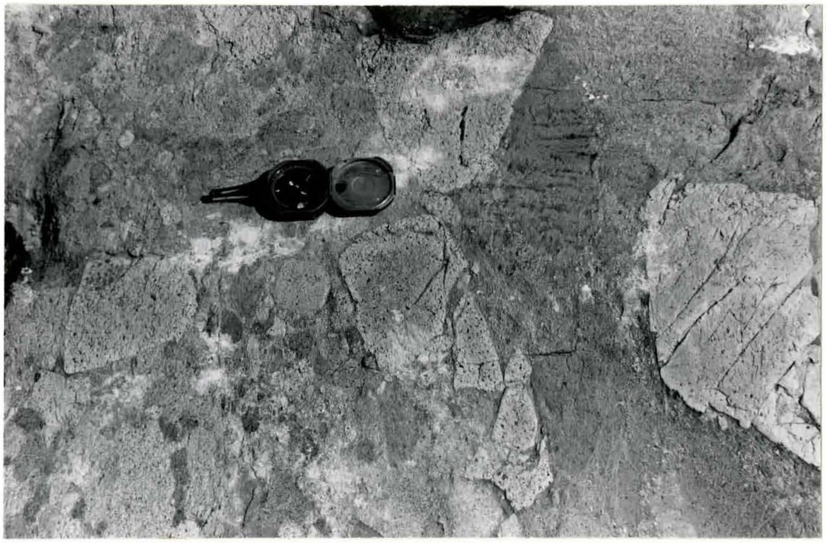
The composition of the fragments and matrix is variable. Macroscopically the light coloured fragments appear more acidic (rhyolitic) than the matrix. This is also reflected in places by their higher resistance to weathering. Many of the darker fragments, which appear more tuffaceous, are often much softer. Much of this difference might be due to alteration both during deposition and later green-schist metamorphism. Some of the lighter coloured fragments display a colour zonation from darker

PLATE 9.

Exposure of fragmental volcanics showing brecciation of fragments.

PLATE 10.

Fragments showing tensional joints. Note distortion is due to angle at which photograph was taken.



cores to lighter outer rims (Plate 8). This, no doubt, is an alteration effect. The composition of the fragments and matrix has been studied by Smith (1971); the present author did not investigate this aspect. In one exposure in the middle of Kakagi Lake fragments within fragments were noted. In this isolated example lighter coloured fragments were found within darker fragments.

A size distribution analysis of the fragments in the area with respect to stratigraphy was carried out by Smith (1971). The author was unable to recognize any significant areal fragment size distribution in the field.

Non-fragmental Intermediate to Acid Volcanics

Interbedded among all the volcanic rocks are some light coloured rocks which have been interpreted as volcanic in origin (Figure 3). Although minor in extent they are found throughout the area. They are usually fine grained siliceous units. Both their stratigraphic and lateral extent is limited. They have been defined on the lack of banding or bedding as well as their lack of recognizable fragments. For the most part they appear dacitic in composition and seem to be crystal tuffs.

Sediments

This unit consists of interbanded cherty and tuffaceous rocks which are well bedded (Plate 11 for microphotograph; Plates 22, 27 for field exposures). In general they are confined to two areas; one in the vicinity of Cedartree Lake and the other in the southern part of Kakagi Lake (Figure 3). On the south shore of Kakagi Lake the banded nature can be seen on aerial photographs. Stratigraphically these rocks rest above the thick fragmental unit.

Contrary to what Burwash (1933) considered, the contact between the sediments and the fragmental rocks is not an unconformity but rather a gradational and/or interfingering contact as recognized by Goodwin (1965).

These sediments are believed to be waterlain volcanic tuff and ash. Most appear to have been reworked.

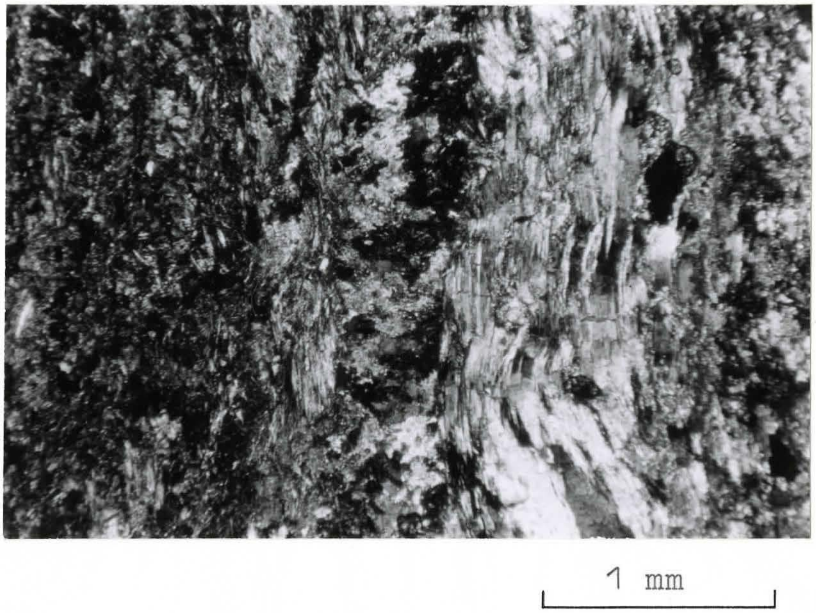
Basic Intrusive Rocks

The large, most westerly, intrusive body found in Kakagi Lake consists of a differentiated series of rocks from peridotite to granophyre (Ridler, 1966). The author has not examined in any detail the mineralogy and petrology of this sill. The lithologic types and their distribution, as shown in Figure 3, have been defined on examination of hand specimens with a minimum of thin section petrography. For detailed petrographic description of the various rock types comprising the large differentiated intrusive unit the reader is referred to Ridler (1966).

In addition to the intrusive studied by Ridler there are two other large intrusive bodies roughly concordant to the volcanic stratigraphy (Figure 3). These other intrusives, although not as thick as the most westerly one, do display some degree of differentiation. In the N.E. (Cedartree Lake area) however, the intrusives

PLATE 11.

Contorted banding within the sediments due to flowage. Lighter material is 'cherty' while darker band is 'tuffaceous'. (Plane light).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do not reveal on macroscopic scale nearly the degree of differentiation that is evident elsewhere.

Quartz Porphyry Intrusives

Quartz porphyry has intruded into both volcanic and ultrabasic intrusive rocks. The largest concentration of quartz porphyry is found in the north-west part of Kakagi Lake (Figure 3). In general, these intrusive bodies display very little schistosity or shearing. They appear very fresh and often are a pinkish colour in hand specimen.

Petrographically, three thin sections of the porphyry from different intrusives consisted of phenocrysts of both quartz and feldspar in a very fine - grained matrix of similar mineralogy. The feldspar, for the most part, is oligoclase (Ab_{88}; Michel - Lévy technique), and often exhibits both albite and Carlsbad twinning. Phenocrysts of quartz were not euhedral but rounded and embayed. They showed no evidence of strain (i.e. no undulatory extinction). Euhedral cubes of pyrite approximately twice the size of the matrix grains are often present throughout the matrix. The only alteration noted was selective sericitization of the plagioclase. Hornblende, although minor, was ubiquitous throughout the three sections examined. A modal analysis (1000 point count) of one section indicated the minerals as: plagioclase (oligoclase), 12.8%; quartz, 7.7%; hornblende, 3.4%; pyrite, 1.1%; and fine - grained matrix, 75.0%.

Lamprophyre Intrusives

Numerous small black lamprophyre dykes and sills are found throughout the study area. Several of the larger ones have been placed on the geological map (Figure 3). The general morphology of these intrusives will be described later. With the exception of the late north-west trending diabase dyke, the granitic intrusives and quartz porphyry, they are considered to be one of the youngest rock units in the area.

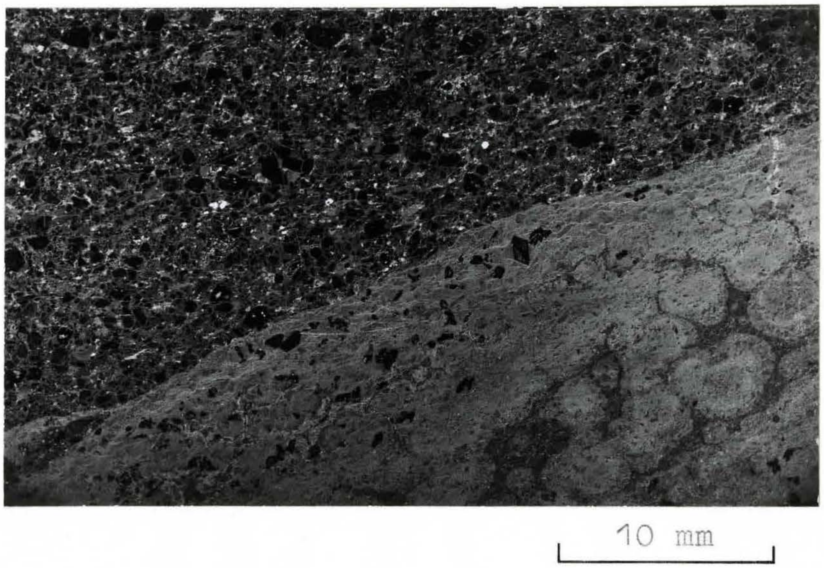
The lamprophyres are massive, dark grey to black, fine to medium - grained rocks with a distinct porphyritic texture due to biotite crystals approaching 5 mm in length. The rock is frequently too fine-grained to reveal biotite in hand specimen. In the outcrop these intrusives are often readily apparent as recessive rusty coloured bands.

Although no detailed chemical or petrographic examination was made, the lamprophyres appear to be variable mineralogically. In general, the rock consists of hornblende and biotite laths embedded in a feldspathic matrix. Some thin sections show much calcite as well as varying amounts of opaque mineral(s). A few sections displayed much pyroxene (mostly augite).

Contacts with host rock are sharp. The chilled margin of small intrusives (i.e. $<$ 25 cm in width) is often very limited (Plate 12).

PLATE 12.

Contact between lamprophyre dyke and fragmental volcanics. Note the growth of secondary quartz (dark mineral) within the volcanic rock near contact. Chilled margin is extremely narrow.



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Diabase Dyke

Possibly the youngest rock of the area is the diabase dyke that extends into the study area from the north-west (Figure 3). It cuts through both basic volcanic rocks and gabbro intrusives. Its relative age cannot be determined in the study area but from existing maps of the adjacent area it appears to continue for some distance in a north-west direction disregarding structure and is thus later than the major tectonic episodes of the area.

Where observed, the dyke is approximately 400 feet (120 metres) wide and appears vertical. The rock seems quite fresh in hand specimen. No petrographic examination was done.

This dyke may be one of those classified by Fahrig and Wanless (1963) as 'Aphebian and older (+1700 my.)' on the basis of (K/Ar) age determinations of other similar dykes in northwestern Ontario.

CHAPTER 3

METAMORPHIC GEOLOGY

The grade of metamorphism in the study area is low (greenschist) and thus the original features of the rocks are generally well preserved. The mineral assemblages found over most of the area are monotonously similar.

Because of the extent of the area and the various lithologies this discussion on metamorphism is only intended to describe the general metamorphic picture of the area.

In the field an increase in metamorphism associated with an increase in deformation was noted from north to south. The area can be roughly divided into two metamorphic zones - a chlorite zone and a biotite zone - as indicated by the corresponding mineral assemblages. The presence of biotite marks the boundary between the two zones.

A problem in discussing the metamorphic geology is the change in lithology across the area - from acid to intermediate fragmental volcanics over most of the area to more basic lavas in the south and south-west. Thus the bulk composition, a determining factor in the mineral assemblages, changes about where the isograd has been placed (Figure 4). The basic volcanics along the west shore to the north have aided in avoiding this problem since they allow for a comparison of mineral assemblages in the two zones. While the fragmental rocks show little change in metamorphic grade

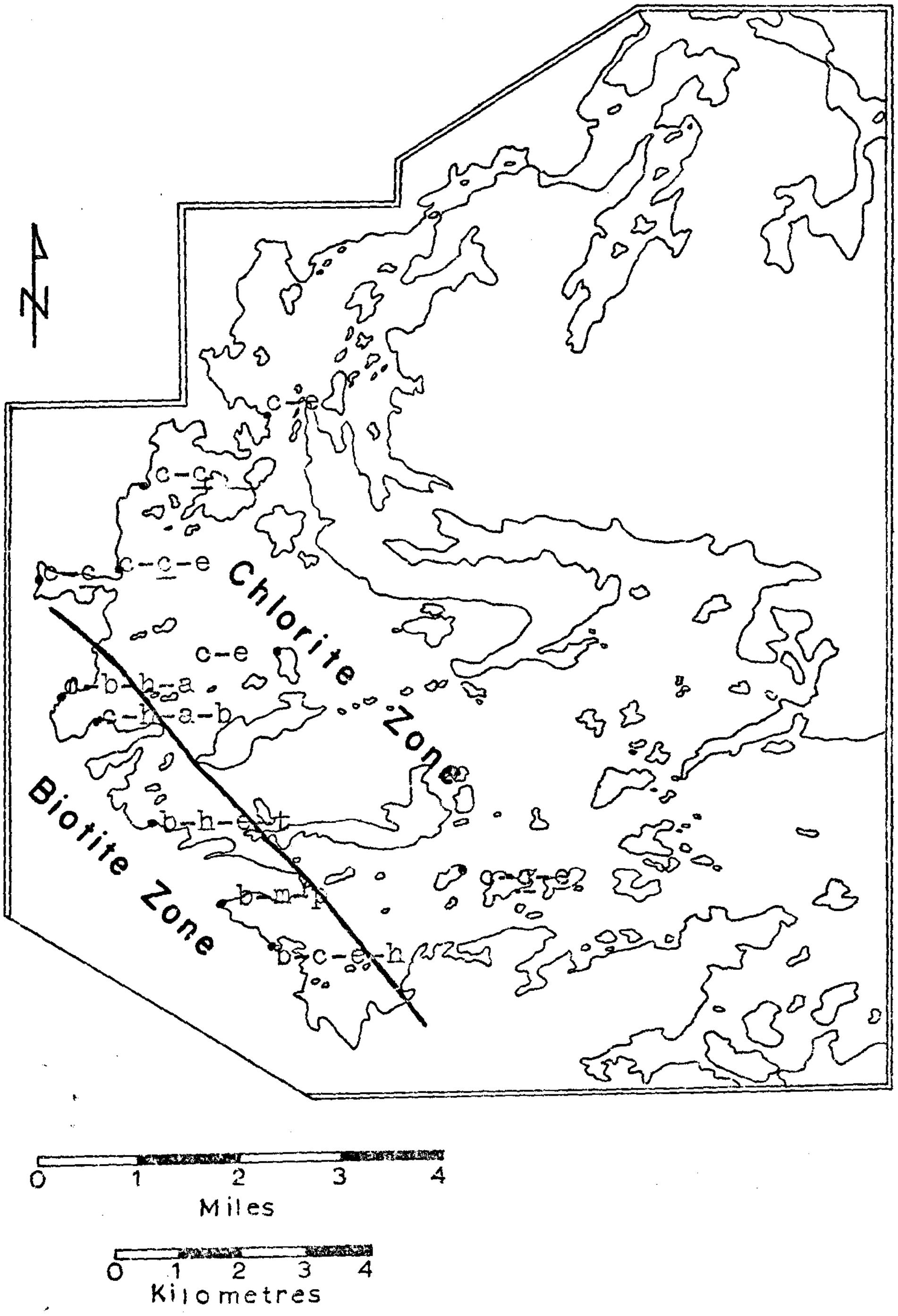
FIGURE 4.

Metamorphic Zones of the Area

Legend:

a - actinolite
b - biotite
c - chlorite
c - calcite
e - epidote
h - hornblende
t - tremolite
p - plagioclase
m - muscovite

Quartz and plagioclase are found in many of the mineral assemblages.


c-e
c-c
c-c
c-c-e
Chlorite Zone
c-e
e-b-h-a
c-h-a-b
Biotite Zone
b-h-e-t
c-g-e
b-m-p
b-c-e-h
0
1
2
3
4
Miles
0
1
2
3
4
Kilometres

over the area, the more basic volcanics do indicate an increase in metamorphic grade southward along the west shore of Kakagi Lake.

Chlorite Zone

Chlorite is ubiquitous in both the fragmental acid to intermediate rocks and the more basic lavas. Only minor amounts of chlorite were observed in the sediments on Cedartree Lake, whereas those on Kakagi Lake appeared to contain more. The mineral assemblages noted in thin sections vary with the lithology (Figure 5) but all evidence indicates the lower greenschist facies (Quartz - Albite - Muscovite - Chlorite subfacies of Turner and Verhoogen,(1960)).

Common mineral assemblages noted in sections of rocks within the chlorite zone include:

A. Acid to intermediate volcanic rocks:

1. Quartz - albite - muscovite - chlorite
2. Epidote - albite - chlorite - actinolite(tremolite)

B. Basic volcanic rocks:

3. Albite - epidote - chlorite - calcite

Biotite Zone

The biotite zone lies south of the isograd as shown in Figure 4 and is characterized by the first appearance of biotite in thin section. The position of the isograd is approximate for the reasons mentioned above.

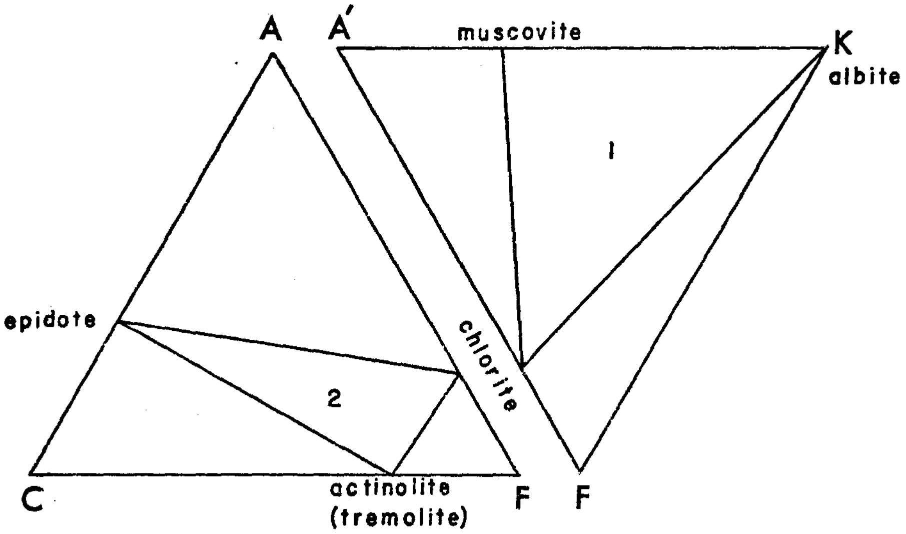
Associated with the growth of biotite in the basic lavas is

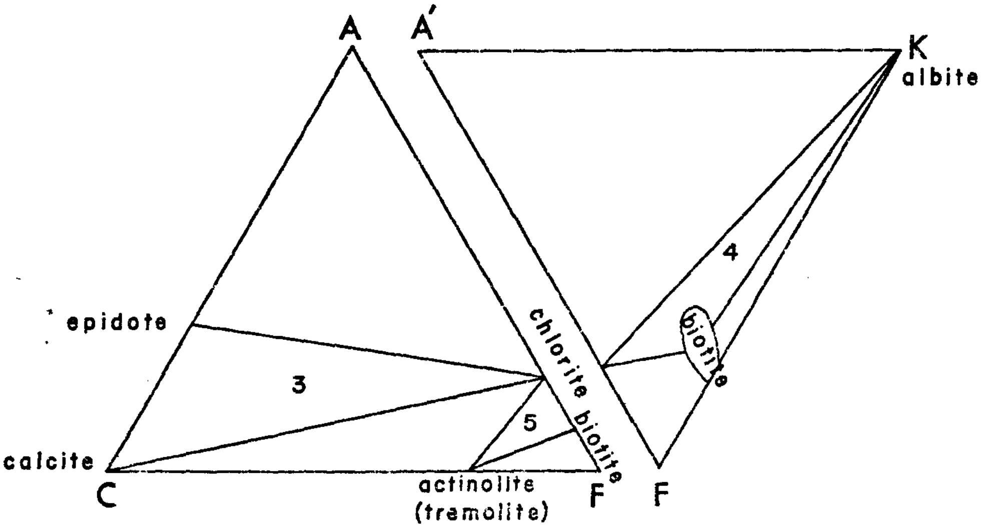
FIGURE 5.

ACF and A'KF Diagrams of Metamorphic Mineral Assemblages.

(a) Mineral assemblages found in the chlorite subfacies of the greenschist facies for the acid to intermediate volcanic rocks. Both quartz and plagioclase are possible additional phases.

(b) Mineral assemblages found in the basic volcanic rocks. Quartz and Plagioclase are possible additional phases. Assemblage 3 is found in the chlorite zone; assemblages 4 & 5 are found in the biotite zone.


A
A′
muscovite
K
albite
1
epidote
chlorite
2
C
actinolite
(tremolite)
F
F


A
A′
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albite
4
epidote
chlorite
biotite
3
5
calcite
C
chlorite biotite
actinolite
(tremolite)
F
F

the appearance of amphiboles, mostly hornblende and actinolite. Clusters of these minerals give the rock a 'knobby' weathered appearance which has been referred to as spotted lava (Burwash, 1933).

In thin section some of these aggregates of hornblende and biotite display a paracrystalline (syntectonic) nature, while most of the minerals show a pretectonic crystallization with biotite cleavages visibly bent and fractured. Late fine-grained biotite has developed along penetrative schistosity planes and can only be observed in thin section due to the still high percent of chlorite present in the rock.

Mineral assemblages found in the basic volcanic rocks and considered to be within the biotite zone are:

A. Basic volcanic rocks:

4. Chlorite - biotite - calcite
5. Chlorite - biotite - actinolite

Possible reactions that have occurred at the biotite isograd may be:

$$\text{Muscovite} + \text{chlorite} \rightleftharpoons \text{biotite} + \text{chlorite}$$

$$\text{Chlorite} + \text{calcite} + \text{quartz} \rightleftharpoons \text{actinolite} + H_2O + CO_2$$

In a few restricted areas within the zone evidence exists that the rocks might better be described as belonging to a hornblende - hornfels facies of contact metamorphism rather than the biotite zone of regional metamorphism. However, not enough data has been collected to justify outlining these small areas. The biotite zone is believed to be the northern edge of a contact aureole about the granitic batholith to the south.

The isograd shown in Figure 4 appears to coincide with the north-west striking fault over much of its length. This fault zone is on strike with the late diabase dyke to the north (Figure 3). It may be that the fault has a vertical as well as horizontal component of movement and the biotite zone represents an up-lifted block to the south-west.

CHAPTER 4

STRUCTURAL GEOLOGY

The main objectives of a structural analysis are two fold:

1. to describe the form, orientation and mutual relationships of various structural fabric elements, and
2. to interpret the data in terms of possible principal stress and strain directions and deformation mechanisms.

The former constitutes an analysis of the geometry, the latter an analysis of the kinematics and dynamics producing the geometry.

Total fabric of a deformed rock includes not only those elements formed in response to a particular present stress field but also those formed by previous stress fields differing from the present one.

Method and Terminology

The three scales of structure used in this study may be defined as follows:

1. microscopic scale: those structures observable in thin section.
2. mesoscopic scale: those structures readily observable in a hand specimen or a single rock outcrop.
3. macroscopic scale: those structures indirectly observ-

able on a scale larger than a single outcrop.

Three-dimensional orientation of structural data is illustrated by using the lower hemisphere equal - area (Schmidt) projection. Orientations are given by either a great circle trace or a pole plot. Diagrams are contoured where considered necessary, according to the percent poles or number of poles per unit area of the diagram.

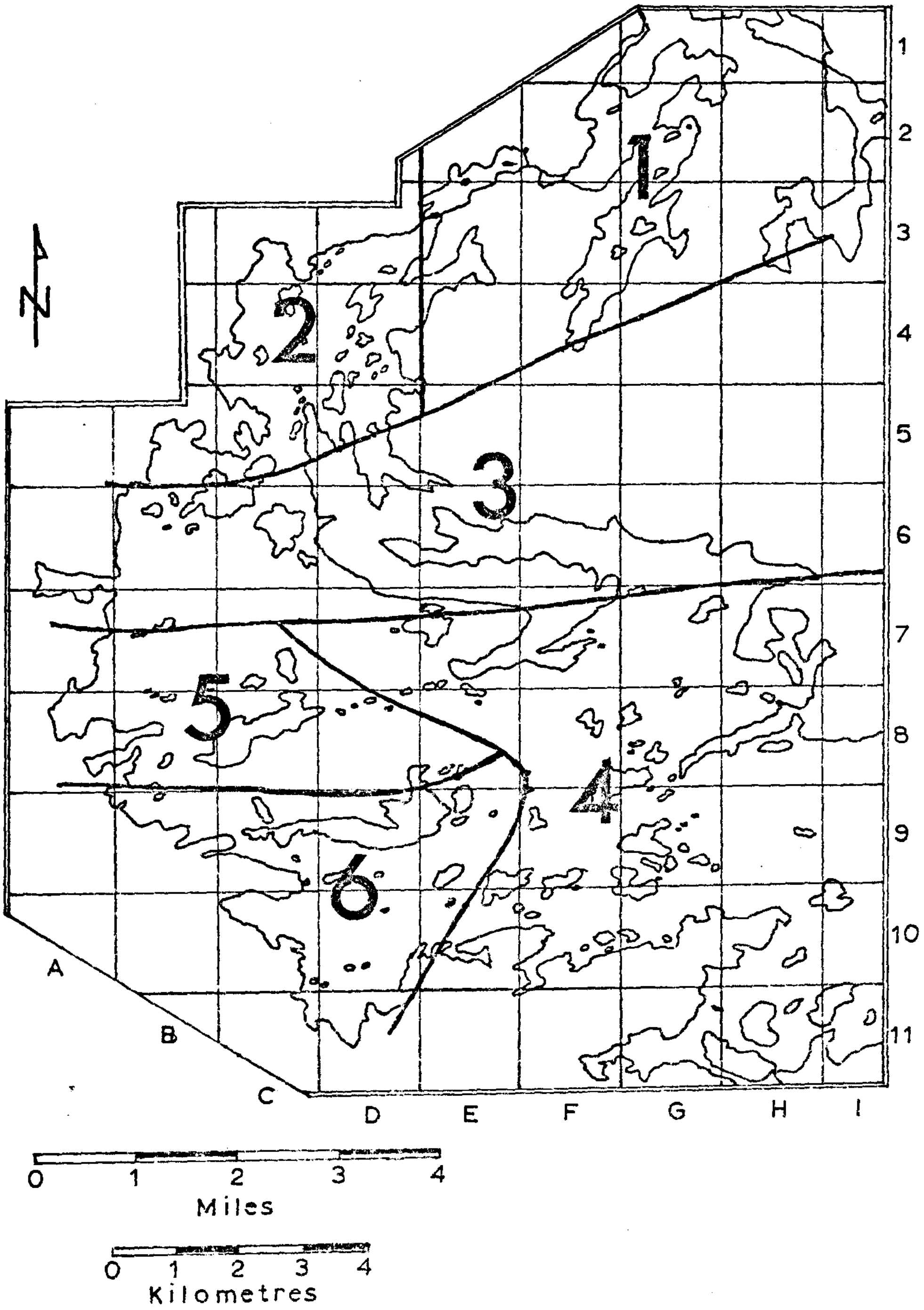
Orientations of the following structural elements were recorded in the field:

1. Bedding (S_0) was measured both in the fragmental volcanic rocks and the fine-grained cherts and tuffs. No bedding was found in the intrusive rocks.
2. Schistosity (S_1,S_2) was measured throughout the field area as the penetrative cleavage. It was generally approximately parallel to S_0 on the limbs of the macrofolds and to the axial plane near the crests of these folds.
3. Axial planes of mesoscopic folds in S_0 were measured and found to roughly parallel that of the macroscopic folds.
4. Fold axes were measured throughout the area where found.
5. Lineations, the result of intersecting s-surfaces; and crenulation cleavage axes were measured.
6. Joints, where prominent, were recorded.
7. Several faults and shear zones were mapped.

The map area was first divided into sections measuring one square mile (Figure 6) and much of the data were plotted in each of the sections. Upon inspecting the resulting plots the area was then broken into six large domains on the basis of apparent structural homogeneity (Figure 6). Structural data was then plotted for each of these large domains. For certain features (e.g. minor intrusives) a single plot was constructed for the entire area.

FIGURE 6.

Sketch showing Sections and Domains used in Analysis of Structural Data.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Kilometres

Mesoscopic Structure

Planar Structures

The rocks in the area were found to contain various primary stratification layers and cleavages which the author refers to as s-surfaces. Primary stratification layers have been designated as S_0, while the other s-surfaces have been numbered chronologically as S_1 and S_2. Their relative time of formation has been deduced from their mutual relationship in the field.

Primary Stratification Layers (S_0)

These S_0 - surfaces in nearly every case represent bedding found in the volcanic rocks. The majority of the primary stratification is confined to the waterlain, fine-grained acid volcanic rocks considered by previous workers (Goodwin, 1965) as sediments. Very few S_0 - surfaces were found in the volcanic fragmental rocks and where found, such surfaces were defined on the graded aspect of the fragments. In the sediments they were defined both on the colour and composition change from one layer to another. Movement has occurred along these S_0- surfaces in many places.

In the vicinity of Cedartree Lake the S_0- surfaces are the most prominent planar structures in the rocks. It is precisely these S_0- surfaces that reveal the style of deformation (folding) in the area. A plot of S_0- surfaces on an equal-area projection will give the geometry andorientation of the macroscopic structure in the area as will be shown later.

Secondary Planar Structures

These include the two schistosities observed within the study area (S_1 and S_2). One set of schistosities (S_1) consists of those planes generally developed parallel to the axial planes of the large macro-folds (F_1). In places, however, S_1- surfaces are parallel to subparallel to bedding planes on the limbs of these folds. The other set of schistosities (S_2) is roughly parallel to the axial planes of the later developed cross folds (F_2) found in the south and west parts of the area (Domains 5, 6).

In addition to planes of schistosity, joints are prominent in certain places throughout the field area. These constitute another type of secondary planar structure which will be discussed later under a separate heading.

S_1- Surfaces

Near the axial traces of the large E-W trending macro-folds, as shown in Figure 3, a parallel set of schistosity planes (S_1) has developed. These planes are marked by a weak orientation of minerals; as well they represent obvious planes of weakness. At the crest of the most northerly syncline in the vicinity of Cedartree Lake a slaty cleavage has developed which is considered a more local intense development of what has developed elsewhere as penetrative schistosity. S_1- surfaces as axial plane schistosity are best developed near the axial trace of the large folds. In this area where S_1 is oblique to S_0 the intersection of the two surfaces defines a lineation lying parallel to the axis of the fold. S_1- surfaces strike between E.N.E. and E.S.E.

and dip steeply both to north and to south. The S_1- surfaces on the limbs of the folds occur parallel or sub-parallel to the S_0- surfaces and appear to have formed as a consequence of flexural slip on S_0- surfaces during the F_1 folding. In places these S_1- surfaces show fanning about the axial plane as seen in the north-west part of the area. On the west side of Kakagi Lake S_1- surfaces are partially obscured by development of later schistosity planes (S_2).

S_2- surfaces

Evidence for a second phase of deformation (F_2) is seen on the west and south-west side of Kakagi Lake, in the form of minor superposed folds. The axial planes of these F_2 folds have associated with them in certain places a set of cleavage planes referred to as S_2- surfaces. These S_2- surfaces are oblique to S_1 and defined in some outcrops by closely spaced, discrete, parallel fractures varying from 1 to 3 mm apart and in other outcrops by preferred mineral orientation. In some of the intermediate to basic volcanic units preferred orientation of such minerals as hornblende help to accentuate the S_2- surfaces.

Meso-folding

Meso-folding may be defined as folding observed on the scale of a single outcrop or hand specimen. Meso-folding is often a sympathetic part of folding on a larger scale (macro-folding). Kink folds, a type of meso-fold, will be discussed later in this thesis.

Few meso-folds were found within the study area. Where present they are reflected by the deformation (folding) of a previously straight quartz vein or minor intrusive, with little observable deformation of the adjacent rock (Plates 13, 14). This is due to the lack of any other surface which might have displayed the folding, thus the orientation of the fold axis was the only measurable parameter to describe the fold. Few axial planes could be determined. Only the strike of the fold axis has any significant meaning since the plunge will depend entirely on the original unknown attitude of the folded body.

The orientation of the fold axes recorded in the field for each of the six domains is given in Figure 7, along with the orientation of slickensides recorded within each domain. It should be pointed out that these fold axes have not been recorded from the small buckles associated with ptygmatic folding of quartz veins (described later) but that they may in certain cases be of the same orientation and origin.

One very important difference noted in the meso-folding is the occurrence of folded quartz bodies throughout the study area while folded black lamprophyre dykes and sills occur only in the south-west of the study area. If one assumes that all the lamprophyre intrusives are of the same age then one is able to define more accurately the area affected by a later deformation.

More meso-folds are found in the south than in the north (Figure 7). In domains 1 and 2 the folded quartz bodies have their fold axes striking about normal to the S_0- surfaces. They have resulted from buckling of concordant quartz bodies during the early folding (F_1) that gave rise to the large north-east trending synclinal

PLATE 13.

Deformed (open-folded) lamprophyre dyke (25 cm wide) in section C-9. Note small offsets along transverse cutting fractures.

PLATE 14.

Deformed (folded) quartz vein in section A-7. Approx. fold axis 280;70W.

FIGURE 7.

Orientation of Mesoscopic Folds and Slickensides

(a) Domain 1; small dots are slickensides; triangles are fold axes.

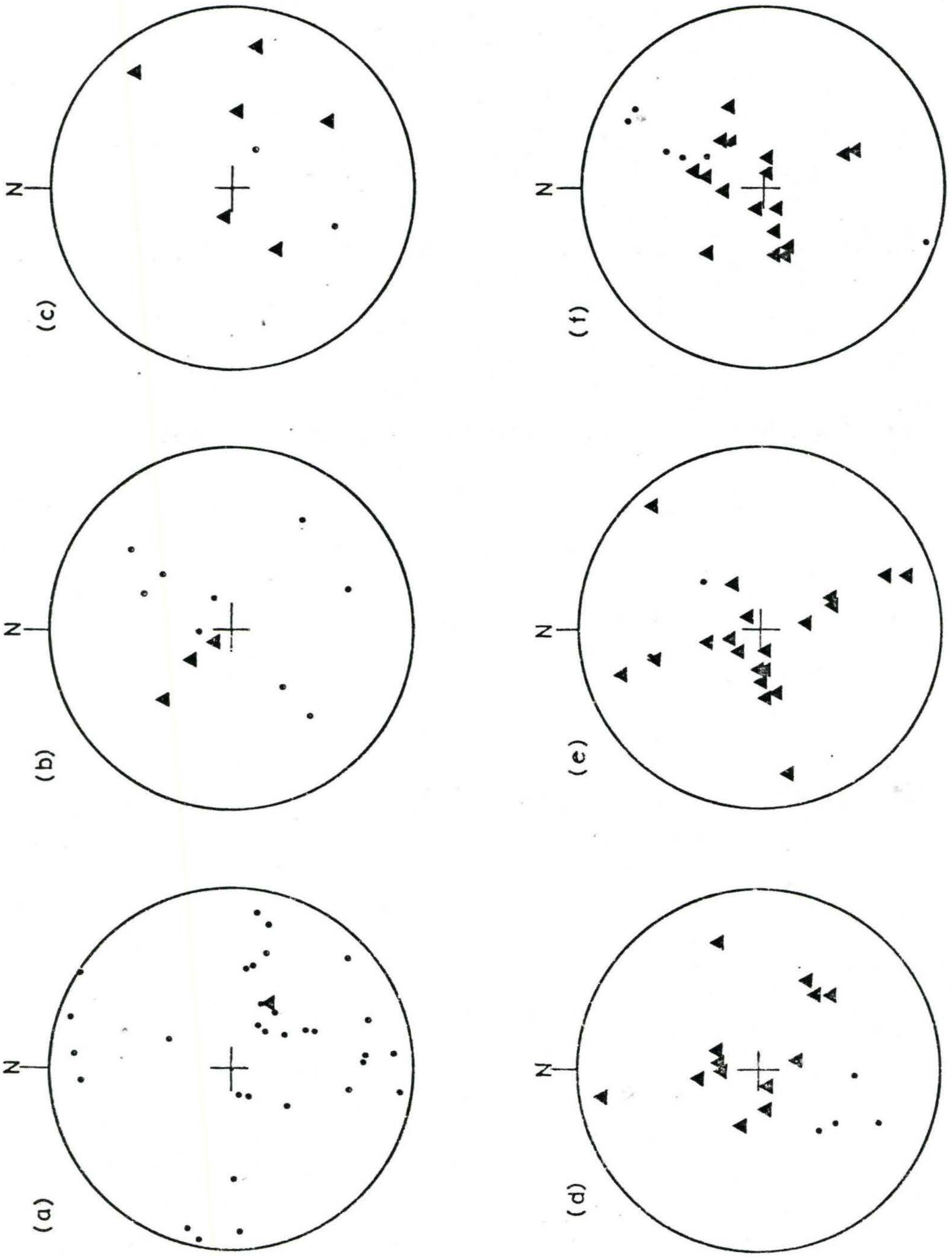
(b) Domain 2: Symbols same as (a).

(c) Domain 3: Symbols same as (a).

(d) Domain 4: Symbols same as (a).

(e) Domain 5: Symbols same as (a).

(f) Domain 6: Symbols same as (a).


(a)
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(b)
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(c)
N
(d)
N
(e)
N
(f)
N

fold. South of this fold axis, the orientation of the meso-fold axes becomes more random. In domain 5 (Figure 7e), however, two axes of folding appear evident; the one trending about N25W coincides with S_2 while the other (N75E) parallels the general S_1 trend. A less obvious but perhaps similar spatial relationship exists between meso-fold axes and local schistosity in domains 4 and 6 (Figures 7d,f).

Slickensides

Slickenside orientations plotted for each domain (Figure 7) show the lack of preferred orientation observed in the field. Most were found in the north part of the area and have resulted from both flexural-slip concordant with the S_0- surfaces and from movement along discordant faults.

It is concluded that they are of use only in an analysis of movement of a particular structure and were used as such where other evidence was lacking to determine the direction and sense of movement along faults.

Meso-faulting

Numerous meso-faults were noted in the field. A meso-fault may be defined as a dislocation observed within a single outcrop. The term 'meso-faulting' as used here includes the term 'strain-slip cleavage' as described by Turner and Weiss (Turner and Weiss, 1963).

Total displacement within a rock exposure through meso-faulting may occur along a single fracture (Plate 15) or several fractures. Displacement along single fractures ranges from a fraction of a centimetre to well over a metre. Where several faults occur they are usually systematic in orientation (Plate 16). The distance between systematic faults varies from less than a centimetre to over a metre and orientation of dislocations varies considerably from outcrop to outcrop. In similar lithologic units this is probably due to the distance between rock exposures and the resulting change in stress orientation. Also considered important is the change in stress concentrations that have developed as a result of fracturing. In many places the common geometry of two conjugate shear fractures is observable.

The two most common indicators of displacement are S_0-surfaces and mineral (quartz) veins. In most cases the markers show a very 'clean' break with little or no drag along the fracture; in others, however, friction effects are evident (Plate 15). Dislocations showing drag were usually single fractures found most often in the gabbroic and dioritic parts of the large sills or in the fragmental volcanic rocks. The sediments displayed negligible drag along dislocations.

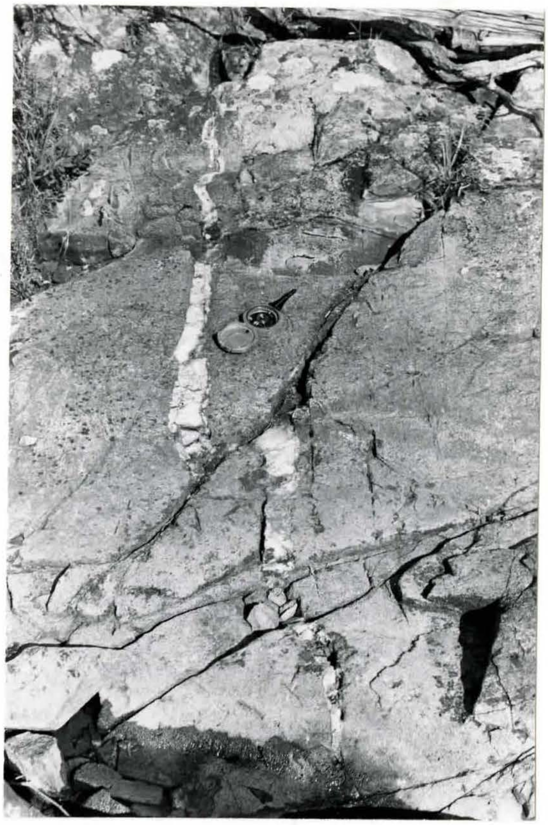
Although much meso-faulting was found in the sediments it was not limited to them. The scale and distribution density of this type of dislocation appears to depend not only on its location within the macro-structures but also on lithology. The fine-grained cherty and tuffaceous sediments display numerous dislocations of a small scale (Plate 17); whereas the thick fragmental volcanic units and the intrusive rocks display fewer dislocations and of a larger scale. This is, no doubt, a result of the various physical properties of the different

PLATE 15.

Deformed (faulted) quartz vein in acid fragmental volcanics.

PLATE 16.

Meso-faulting of quartz veins within gabbro intrusive.



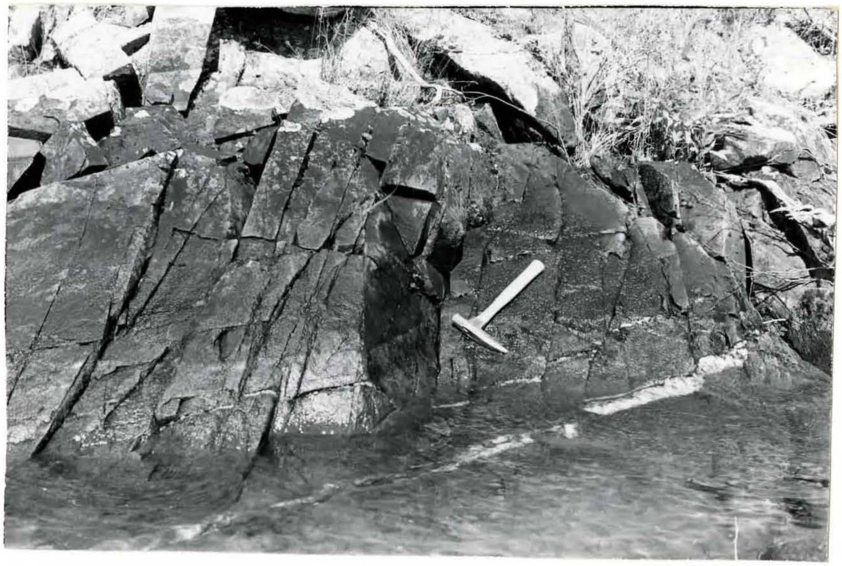
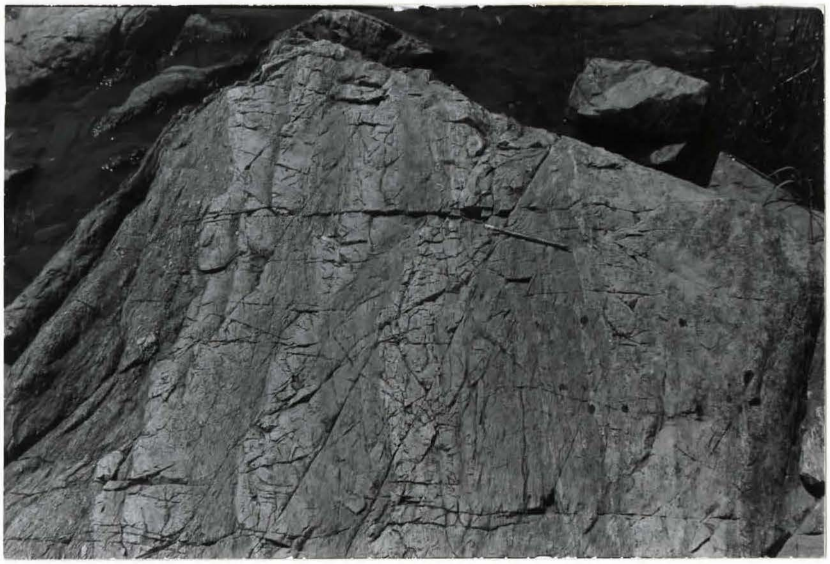


PLATE 17.

Meso-faulting of sediments near south end of Cedartree Lake.



rocks. The sediments being more brittle and thin bedded allowed numerous fractures to propagate across them with small amounts of movement occurring along each. All other parameters being equal, various workers have noted that thin beds contain more fractures than do thick beds (Harris et al, 1960).

Shearing and Shear Zones

Numerous shear zones as shown on the geological map (Figure 3) have been recognized as discrete bands of intense shearing. In places they may only be an extreme example of localized schistosity, however, where restricted to a definite band were considered as shearing rather than penetrative schistosity (Plates 18, 19).

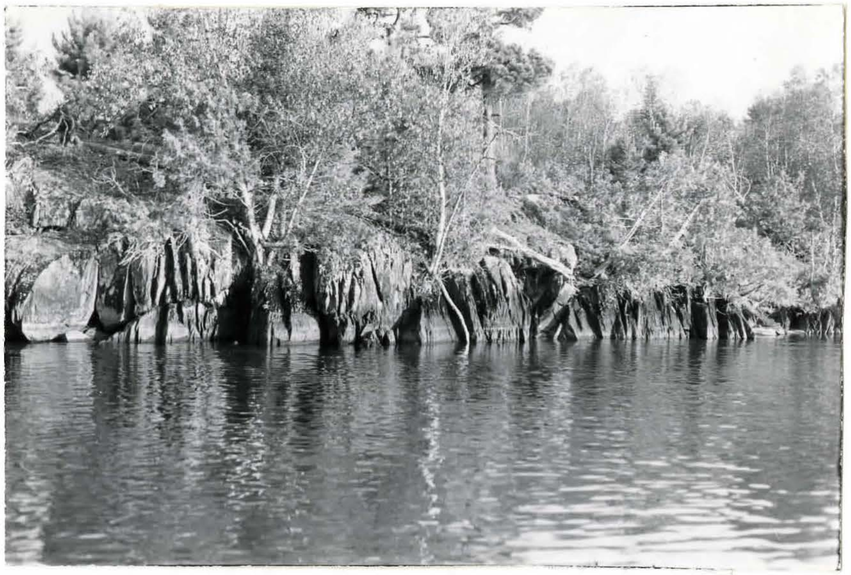
In a few places the shear zones are concordant with the local schistosity but in most places they cut across the schistosity, often at high angles. In the Cedartree Lake area two types of shear zones exist: one parallel to sub-parallel to S_0- surfaces and another cutting across S_0- surfaces. The former is a result of localized movement (flexural-slip) along S_0- surfaces while the latter represents conjugate shears cutting across the strata at angles up to 90 degrees. Well developed shear zones are often found at or near lithologic contacts such as between the large sills and the sediments or the sediments and the fragmental volcanic unit. In both these cases shearing is usually developed best in the sediments. At two places near the north end of Cedartree Lake intense shear bands concordant with S_0- planes were found to pass along strike into penetrative bedding plane schistosity.

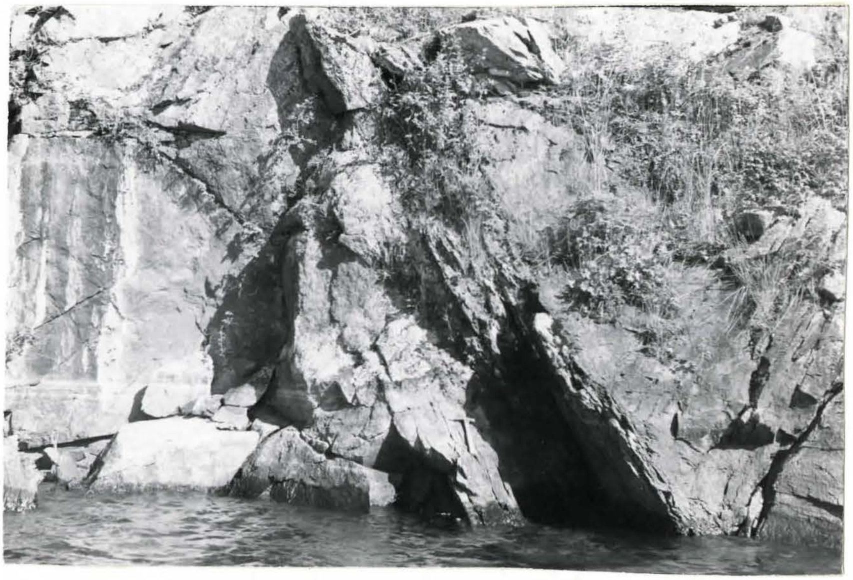
PLATE 18.

Wide N.W. striking shear zone within carbonate unit in section C-5.

PLATE 19.

Narrow shear zone in fragmental volcanic outcrop.





Within the gabbro rocks on Cedartree Lake the shear zones are characterized by weak but noticeable bands of fine-grained mylonites. These shear zones as well as those that cut through the acid fragmental rocks are often iron stained. Some display minor topographic depressions due to preferential weathering.

In contrast to shear zones found in the Cedartree Lake area those found on Kakagi Lake are usually discordant with respect to S_0 and S_1- surfaces. They generally show no preference to lithology and trend parallel or sub-parallel to the north-west striking faults of the area. They often dip steeply. Small drag folds were noted within two of the wider shear zones and these indicated sinistral sense of movement similar to the nearby faulting in the north of Kakagi Lake. The fold axes plunge rather steeply toward the south-east, possibly indicative of sub-horizontal movement within these two particular shear zones.

Some of these shear zones may actually represent faults with a considerable amount of displacement, however, due to lack of exposure this cannot be proven. A few were noted to cut across joints and are therefore younger than some of the joints. In no place was one able to observe a shear zone cutting a lamprophyre or quartz porphyry intrusive and thus their relative age remains unknown.

Tension Gashes

Tension gashes associated with small shear zones were noted at various places throughout the study area (Figure 8). They often display a sigmoidal shape and an en échelon character. These tension gashes are not confined to any particular lithology, and usually are filled with quartz. In the Cedartree Lake area they result from simple shear along bedding planes whereas in the south they form along shear zones that are discordant with the bedding and other s-surfaces.

An analysis of 12 sets of tension gashes at 10 stations gives the local principal stress orientations as shown in Figure 9 and Table I. These results are based on the assumption that the σ_1 σ_2 plane is the plane of tension and that the gashes resulted from simple shear that caused both their initiation and subsequent rotation. The important assumption that they did not result from movement along previous joints (Ramsay, 1967) has also been made. As shown in Figure 9 and Table I the local orientation of σ_1, the maximum compressive stress, varies from 299^0 to 064^0 in strike with a plunge ranging from 39^0S to 17^0N about the horizontal.

Figure 8

Location of Tension Gashes and Orientation of Responsible Stress Field

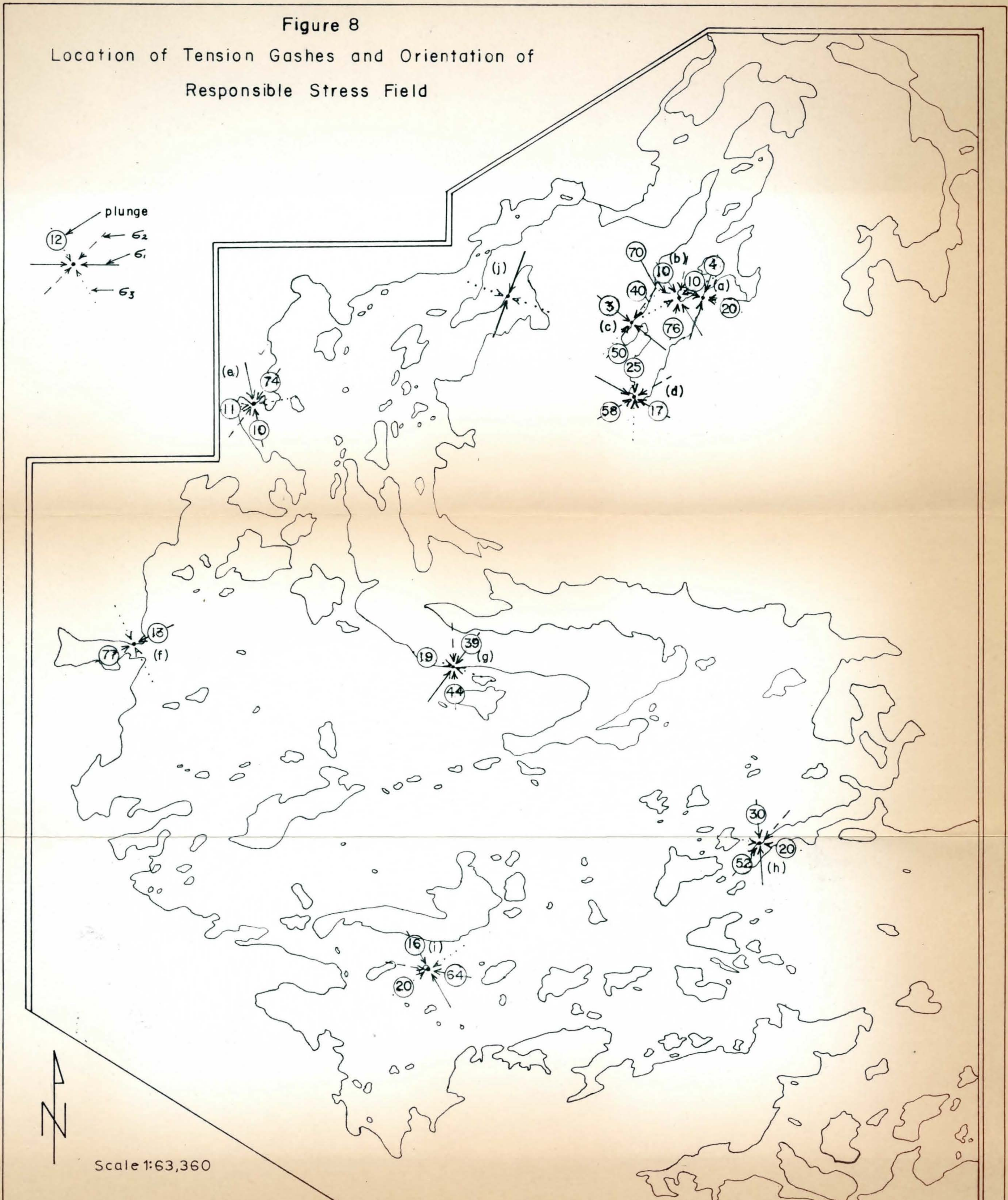


FIGURE 9.

Orientation of principal stress axes obtained from an analysis of 12 sets of tension gashes at 10 stations throughout area. Small dots are σ_1; crosses are σ_2; large dots are σ_3.

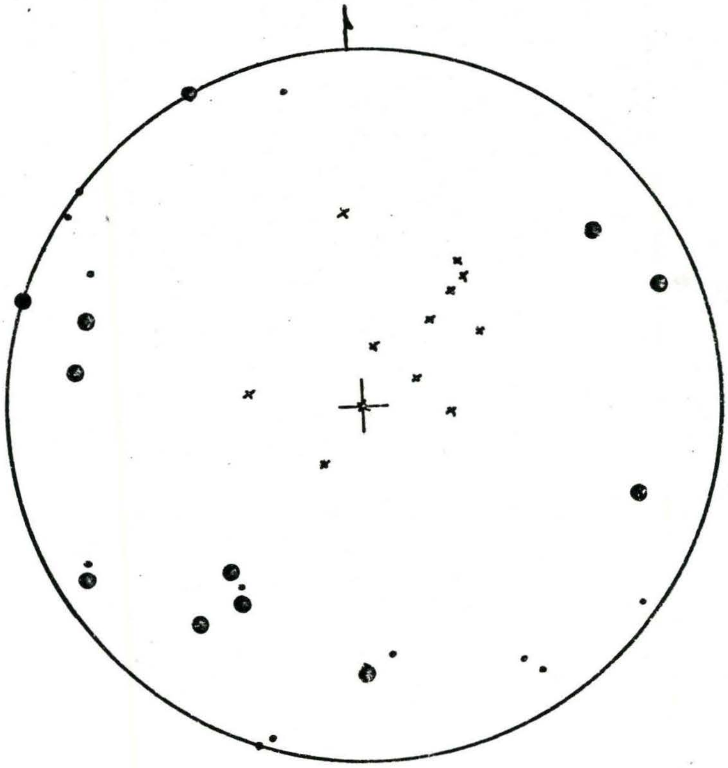


TABLE 1

Orientation of Principal Stress Responsible for Tension Gashes

(a)	1071	018;4S	275;70E	290;20W
(b)	2661	328;10SE	012;76E	061;10SW
(c)	2563a	310;Horiz.	040;65NE	040;25SW
	2563b	308;3SE	036;50NE	041;40SW
	2563c	306;2NW	040;57NE	035;33SW
(d)	5711	299;17NW	060;58NE	002;26S
(e)	2819	348;10NW	037;74SW	081;11E
(f)	27721	064;13SW	064;77NE	334;Horiz.
(g)	1810	037;39SW	357;44N	290;19SE
(h)	15824	356;30S	040;52NE	280;20W
(i)	2575	330;16SE	280;64W	056;20NE
(j)	30728	020;Horiz.	Vertical	290;Horiz.

Macroscopic Structure

Schistosity Trends

By far the most common structural element found in the field beside joints is the schistosity - both S_1 and S_2. The general trend of the schistosity over the study area can be seen on Figure 10. This plot shows the pronounced change in attitude of S_1 from the limbs of the folds to near the axial trace. In the north the attitude of the schistosity is quite uniform (unimodal) within a small area (eg. a section). In the south and south-west (Figure 10) the attitude of the schistosity is not unimodal but rather bimodal with the appearance of S_2 cutting obliquely across S_1.

No statistical treatment with respect to density of schistosity has been done so the data shown as poles or contoured poles in Figure 10 has no significance other than to indicate the general trend. The amount of data recorded in any one place or any single sector varies and is especially dependent on the amount of rock exposure which in turn depends on the relative amounts of land to water.

The attitude of the schistosity in each of the six domains is shown as equal area projections (Figure 11). The bimodal character resulting from the second schistosity (S_2) is again clearly indicated in Figure 11f, a plot of schistosity for Domain 6.

Figure 10

General Trend of Schistosity

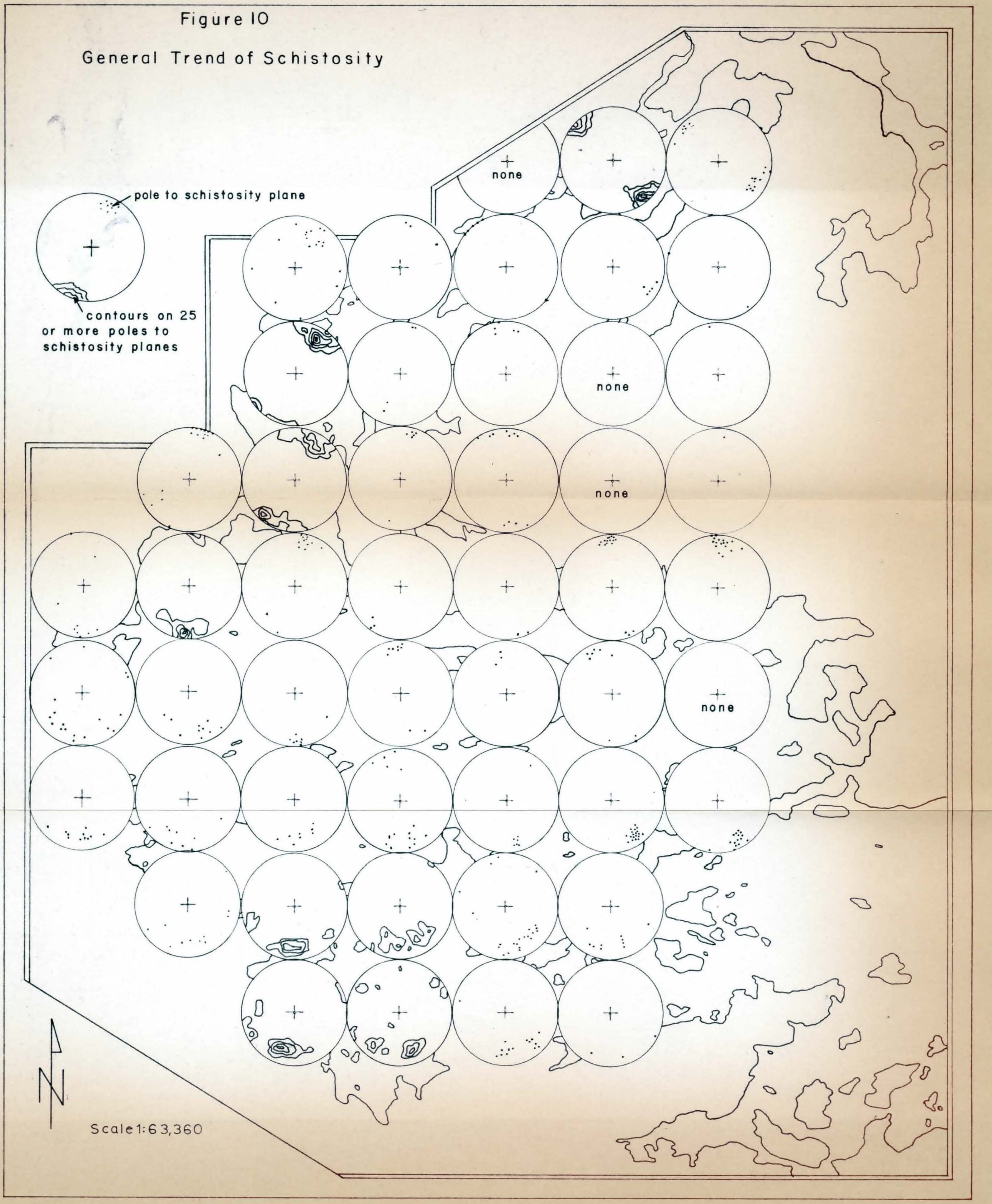


FIGURE 11.

Orientation of Schistosity (S_1 and S_2-surfaces)

(a) Domain 1: Poles to schistosity contoured.
n=71

(b) Domain 2: Poles to schistosity contoured.
n=142

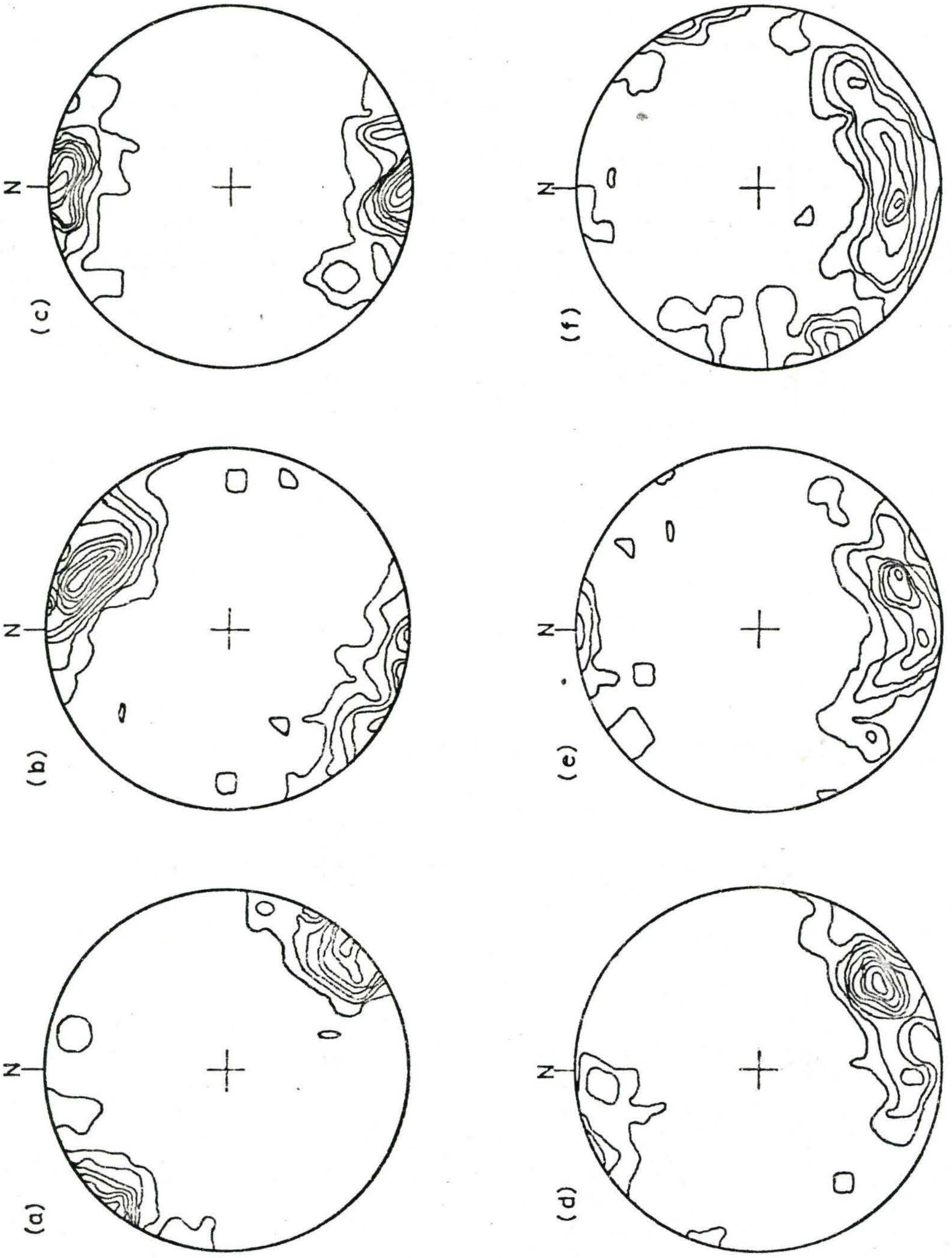
(c) Domain 3: Poles to schistosity contoured.
n=114

(d) Domain 4: Poles to schistosity contoured.
n=83

(e) Domain 5: Poles to schistosity contoured.
n=86

(f) Domain 6: Poles to schistosity contoured.
n=140

Note: all lower hemisphere, equal-area net projections; contours at
1,3,5,7,9,11,13,15,17,19 poles per 1% area.


(a)
(b)
(c)
(d)
(e)
(f)
N

Macro-folding

Two phases of macro-folding appear to be evident in the study area. The major east-west trending folds formed early, with subsequent development of cross-folds in the southwest part of the area. These cross-folds have axial traces roughly N.W. - S.E.

These macro-folds have their orientation and geometry revealed in part by one or more of the following:

1. distribution of lithology,
2. mesoscopic folds, and
3. a prominent s- surface.

Early Formed Macro-folds

Cedartree Lake Syncline

The prominent structure in the northern part of the area is the large macro-fold, a syncline, within which Cedartree Lake lies. It will thus be referred to as the Cedartree Lake Syncline. This fold is outlined by two bands of well bedded volcanic sediments consisting of interbedded cherts and tuffs (Figure 3). The gross geometry is also shown by the distribution of the adjacent ultrabasic sills and fragmental acid to intermediate volcanic rocks.

A total of 468 bedding plane (S_0) attitudes were recorded in the most westerly band of sediments and 93 in the easterly band. In this case, where crude mappable markers are present, structural analysis applied to S_0- surfaces serves to add to the picture of folding displayed in the geological map. As is usual most of the attitudes were measured along the limbs of the fold resulting in the familiar π -diagrams (Figures 12a,b) in which the majority of the poles to bedding lie near the equatorial plane of a Schmidt or equal-area projection. The axial trace of the fold is indicated by the lithology distribution (Figure 3). The β -pole obtained for the π - plot of 468 S_0- surfaces plunges 55°N.E. at 044° (Figure 12a) while that obtained for the most easterly folded sediments plunges 55°N.E. at 022° (Figure 12b). Folding appears to be nearly cylindrical and thus β = B (the fold axis). Figures 12a,b indicate that the syncline is approximately symmetrical with an average dihedral angle of 94 degrees and is thus an open fold (Fleuty, 1964). The strike of the axial plane of the fold seems to change from 030° in the vicinity of Cedartree Lake to 010° further east with a steep dip of about 80 degrees.

This syncline affects the youngest volcanic rocks of the area, but is cut by younger lamprophyre and quartz porphyry intrusives. The minor intrusives in this area show no evidence of deformation. No major faults within the Cedartree Lake area were noted to displace any part of the syncline. However, large faults offset the limbs of this fold in the Kakagi Lake area. These will be discussed elsewhere in the text.

One of the most striking features of the macro-folding in general and the Cedartree Lake syncline in particular is the almost total absence of any sympathetic (parasitic fold of Turner and Weiss 1963) meso-folds. This,in addition to other factors discussed later, is probably the result of a small competence difference between lithologic units.

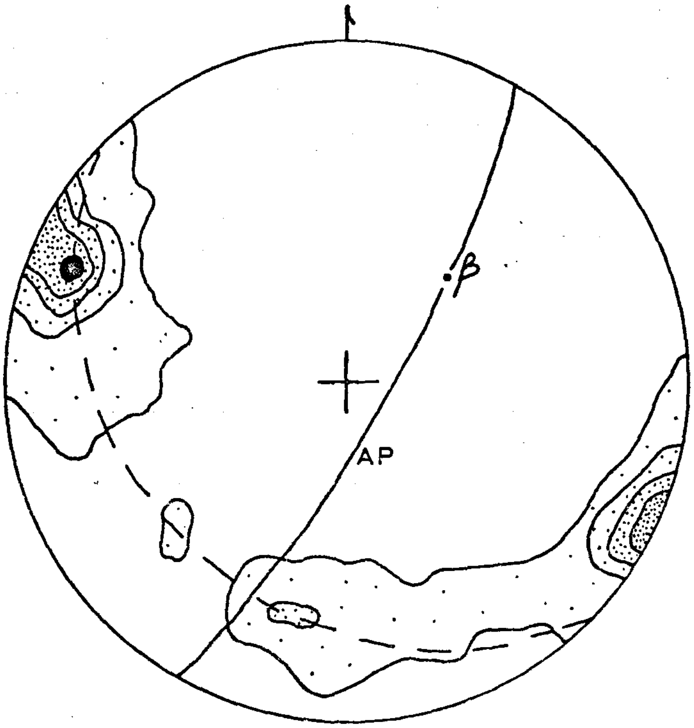
FIGURE 12.

Geometry of Cedartree Lake Syncline

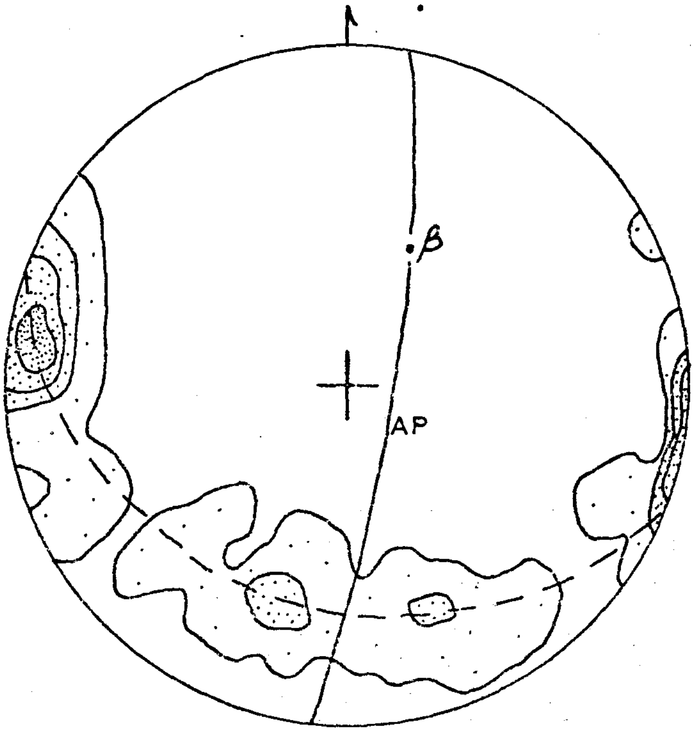
(a) π- projection of Cedartree Lake Syncline (most westerly band of sediments) illustrating macroscopic fold geometry. Attitudes of S_o (bedding) planes have been plotted as poles and then contoured. Contours at 1,5,10,15,20 % per 1 % area. n=468 β= fold axis. AP = axial plane.

(b) π- projection of Cedartree Lake Syncline (most easterly band of sediments). Contours at 1,5,10,15, % per 1% area. n=93 β= fold axis. AP = axial plane.

(a)



(b)



Kakagi Lake Anticline

The large anticlinal structure situated in the central part of the study area defies any detailed analysis other than on a microscopic scale. No s- surfaces other than schistosity and jointing were noted in the field. The form of the folding is illustrated by the gross lithologic distribution. The axial trace of the fold strikes slightly north of east. The few S_0- surfaces noted in the adjacent fragmental volcanics indicate that it is steeply dipping and thus would probably have a fold axis plunging steeply to the east. If the S_1- surfaces are axial plane schistosity near the axial trace, as the author suggests, then we have some additional evidence that the axial plane is nearly vertical. From Figure 3 we see that near the hinge area of the anticline the lithologic units are much thicker (measured parallel to S_1) than on the limbs. Not enough detail is available in the surrounding rocks to say whether or not this is a function of the folding and resulting shearing or a primary feature of the ultrabasic sill. Perhaps the fold hinge is merely situated near the thickest part of this particular sill. Furthermore, one might have reason to consider that the thickest part of the sill lies near its point of entrance at this stratigraphic level. Little evidence exists to support either suggestion mainly because the hinge area of the fold is almost entirely surrounded by water.

Kakagi Lake Syncline I

Folded cherts are found in the south-east part of the map area (Domain 4) and are thought to be a southern extension of those in the Cedartree Lake area. As was common to the north these sediments possess much prelithification and slump structure (to be described later). The general shape of the fold in plan view is shown in Figure 3 by the distribution of the sediments, as well as the recorded fluctuations in bedding attitudes. Folding is very tight (closed-Fleuty, 1964). The north limb is overturned in places while the south limb dips steeply to the north. This south limb can be traced both in the field and on airphotographs for a considerable distance to the east along the south shore of Kakagi Lake.

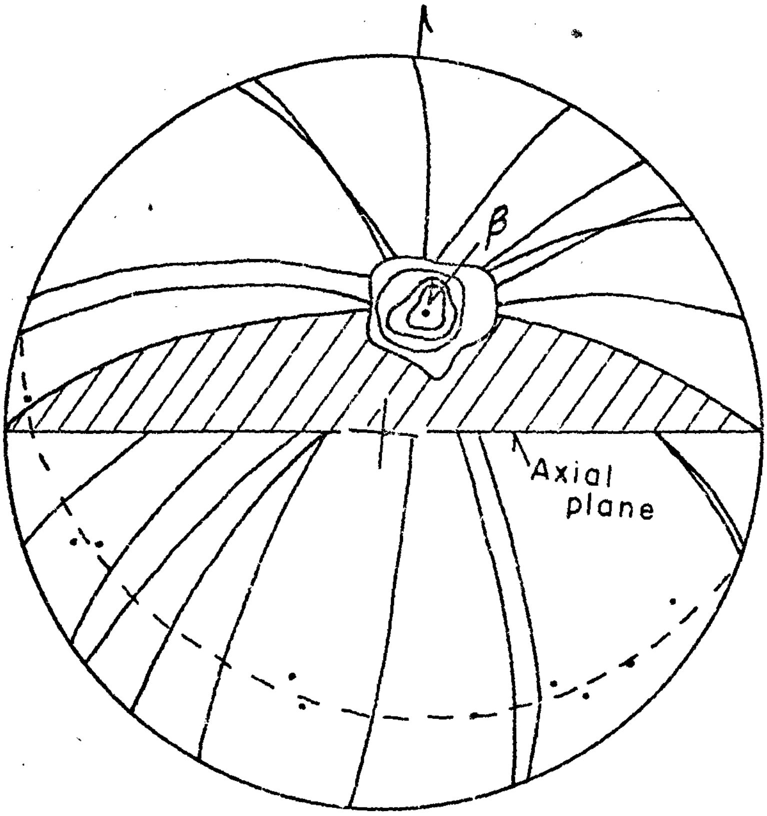
In exposures where S_0- surfaces were found to maintain a fairly uniform trend for several metres it was assumed that they represent true bedding within that area and were not affected by penecontemporaneous slumping. Using these S_0- surfaces the author attempted to analyze this fold although, due to lack of exposure, only 10 S_0- surfaces were recorded. Figure 13 shows the data as a combined π- pole and β- diagram. In addition to the inherent assumptions and errors as pointed out by various workers (Turner and Weiss, 1963; Ramsay, 1962, 1967) this analysis also suffers from a lack of data. It must therefore be viewed as an extremely crude approach to describing the fold geometry.

The β- diagram (Figure 13) would seem to support the conclusion reached in the field that the north limb of the fold is reclined since β plunges 58^{o}N.E. at 016^{o}. The axial plane strikes 080^{o} and dips 60^{o}N.

No major faults were noted to cut the syncline. Many small black lamprophrye dykes traverse the limbs of the fold and are post-

FIGURE 13.

Combined π- pole and β- diagram for Kakagi Lake Syncline. Attitudes of S_0 are plotted as great circles which intersect at a common point β. n=10. Number of β intersections = 45. Contours on β intersections at 5, 10, 25, and 50 % per 1 % area. Small dots are poles to S_0. AP = axial plane.


β
Axial plane

folding as indicated by their lack of deformation.

This synclinal fold is outlined only by the sediments and immediately adjacent fragmental and basic volcanic rocks. The continuance of the structure to the west (much beyond Domain 4) is not evident but instead the geology becomes extremely complex with the lithologic units, most of which are intrusive, seemingly striking discordant to the structure.

Kakagi Lake Syncline II

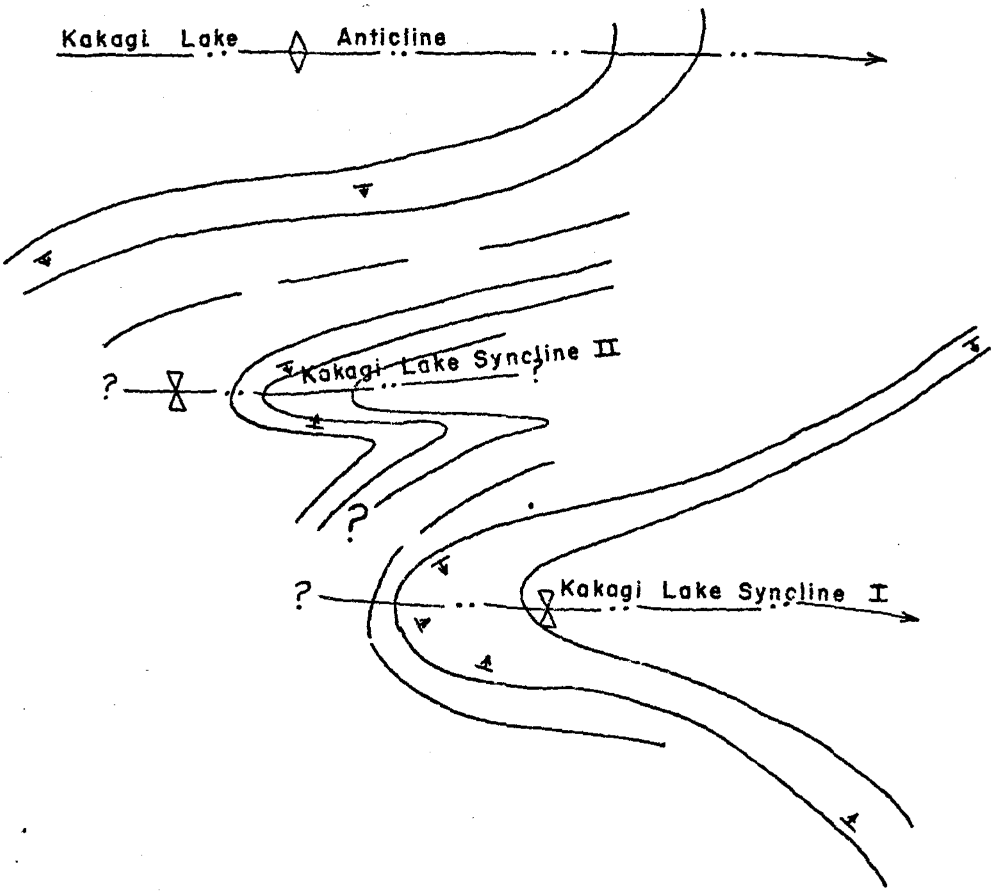
Evidence, although weak, exists for a second synclinal fold situated about three kilometres north of the above described syncline. Due to lack of exposure it is not well defined. The only significant evidence is provided by top determinations got from graded bedding and pillow structures found in the volcanic rocks (Figure 3). Its axial trace approximately parallels that of the southern fold striking slightly north of east. This axial trace lies within non-fragmental and fragmental volcanic rocks.

The bedding on the south limb of the fold dips quite steeply to the north while that on the north limb seems to dip steeply to the north (i.e. overturned).

A possible form of the fold and its relation to the adjacent folds may be sketched as shown in Figure 14. Little more can be said about this macro-structure.

FIGURE 14.

Sketch showing possible relationship of small syncline in domains 5 and 6 to adjacent macroscopic folds.


Kakagi Lake Anticline
Kakagi Lake Syncline II
?
?
?
?
Kakagi Lake Syncline I

Macro-faulting

Numerous large scale dislocations (macro-faults) displace the lithologic units in varied amounts as shown on the geological map (Figure 3). Most are inferred from apparent displacement of rock units. They generally pass between islands and are thus concealed entirely by a body of water.

No macro-faults were observed in the Cedartree Lake area. West and south, however, on Kakagi Lake there are a number of faults believed to be wrench faults with horizontal displacements of up to 500 metres. However, some display evidence of a vertical component of displacement as well as horizontal. This is inferred to explain different stratigraphic thicknesses of a particular unit across a fault. With the exception of the two most northern faults whose displacement is sinistral, all others have a dextral movement. The faults generally trend north-west, an exception being the inferred north-east trending fault found in section D6. The dip on all faults is unknown.

The actual fault trace was observed at only two places in the outcrop. The most northerly fault that displaces the large basic sill in Kakagi Lake cuts across the south-east tip of a large island (section D-4)., and at this point is characterized by a narrow (4 metres wide) band of highly disturbed gabbro that displays excellent fault breccia. The breccia consists of highly polished and slickensided 'horses' in a vein network of white often fibrous calcite.

The morphology and extent of the faults is unknown both in plan and section. No evidence of splay or secondary faulting was found, no doubt due to the extremely limited exposure. No relationship between the faults and the minor intrusives (lamprophyre and quartz porphyry) was observed leaving their relative age, as yet, unknown.

Kink Band Development

Kink bands (folds) observed in the area are confined to a single large rock exposure along the south-west shore of Kakagi Lake. They are described separately from the other structural features because of :

1. their confinement to a local area,
2. their ease of being distinguished from other suites of folds, and
3. their more obvious quantitative use in determining their responsible stress fields than other structures.

A total of 16 kink bands were examined in detail. They are found in laminated acid volcanic rock, the lamination being a cleavage laminae. The average attitude of the laminae is 335;70NE but there are several local departures from this mean value (Figure 18a). All kink bands may be regarded physically as singular since an opposite sense of external rotation is not found within a short distance along the strike of the same laminae.

The following observations and measurements (Figure 15) were made of each of the 16 kink bands examined:

1. sense of external rotation of the laminae across the band (d or s, where d=dextral and s=sinistral)
2. total length of the kink zone (L)

3. average attitude of the unrotated laminae immediately outside the band (A)
4. average attitude of the laminae within the band where the band measured the widest (B)
5. average attitude of the kink zone boundary (KZB) (average of the two sides of the band where it is the thickest) (C)
6. maximum width of the kink band measured along cleavage within the band (D)
7. maximum width of kink band measured perpendicular to its sides (G)
8. the angle the laminae outside the band makes with the KZB (α)
9. the angle the laminae within the band makes with the KZB (β)
10. the average thickness of the laminae within the kink band (t)
11. whether or not a joint plane marks the sides of the kink band, and
12. the distance (where possible) between adjacent kink bands measured along strike of laminae (B).

Morphology and Geometry

An idea of the size and morphology of some of the kink bands may be obtained from the accompanying photographs and line drawings (Plates 20, 21; Figures 16, 17). In each of the kink

FIGURE 15.

Right Cross-Section of Dextral Kink Bands illustrating Measured Angles and Distances.

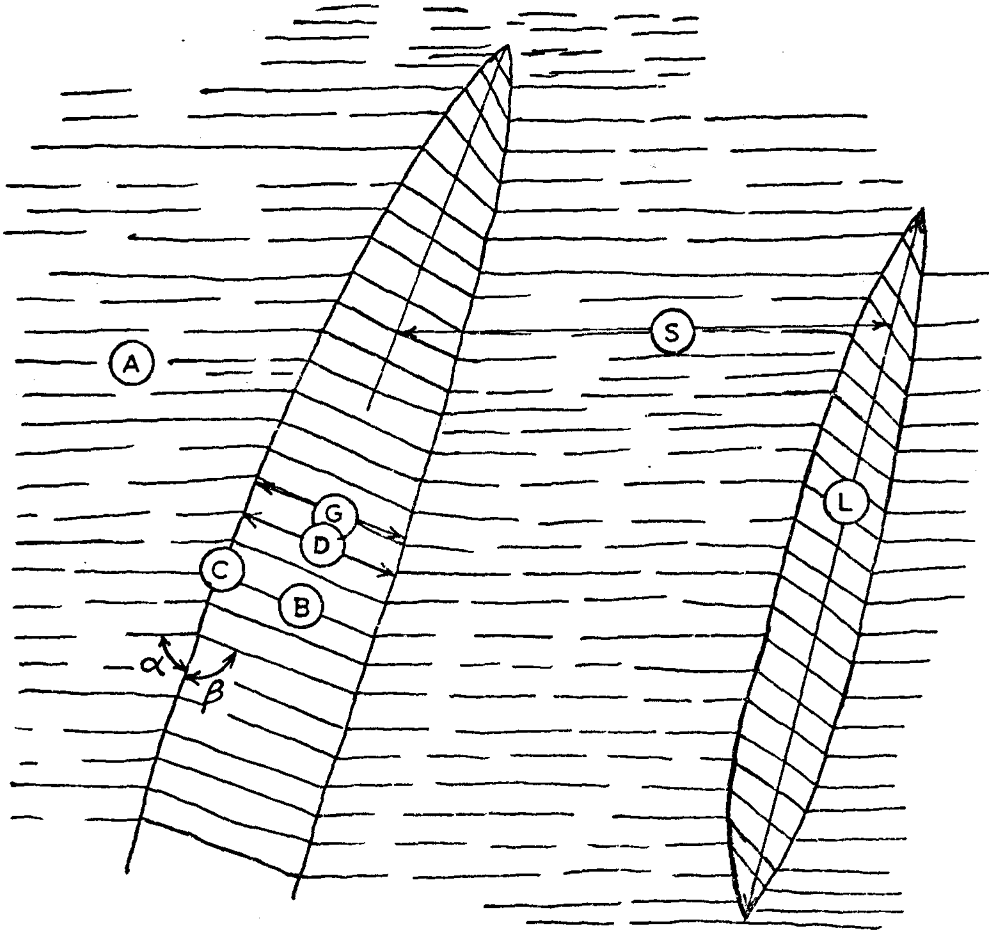
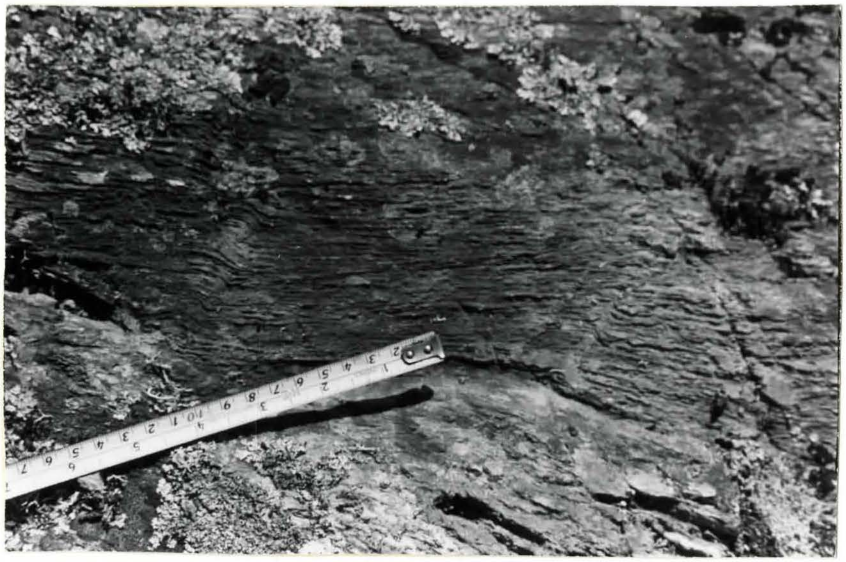

A
S
G
D
C
B
α
β
L

PLATE 20.

Singular dextrally-rotated kink bands in thin laminated meta-volcanics. Measure calibrated in both inches and centimetres.

PLATE 21.

Numerous small singular sinistrally-rotated kink bands. This is typical of much of the rock surface after thin overburden (club mosses) is removed. Measuring tape case is 2 inches (5 cm) in length.



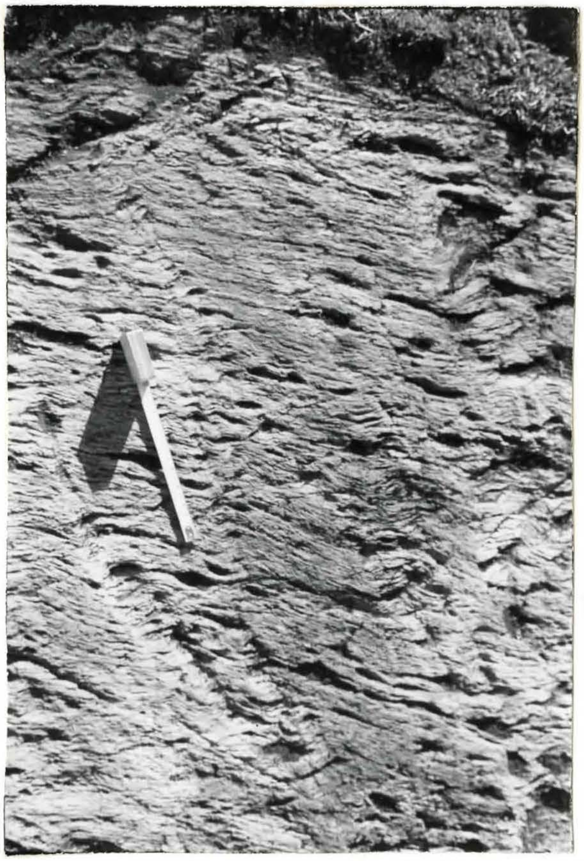
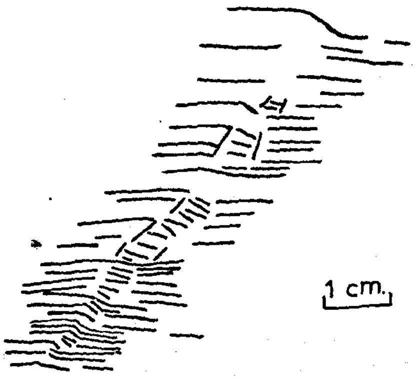
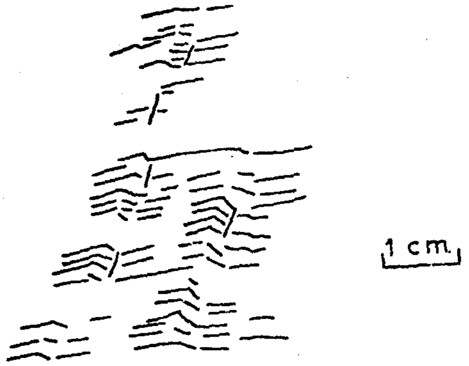


FIGURE 16.

Line drawings showing the morphology of three of the better developed dextral kink bands. (All drawings based on photographs).


1 cm.


1 cm.

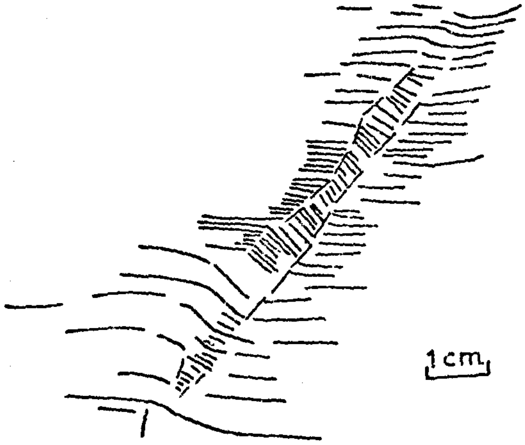
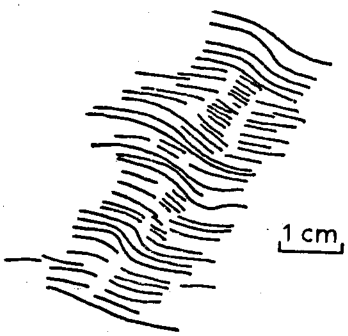
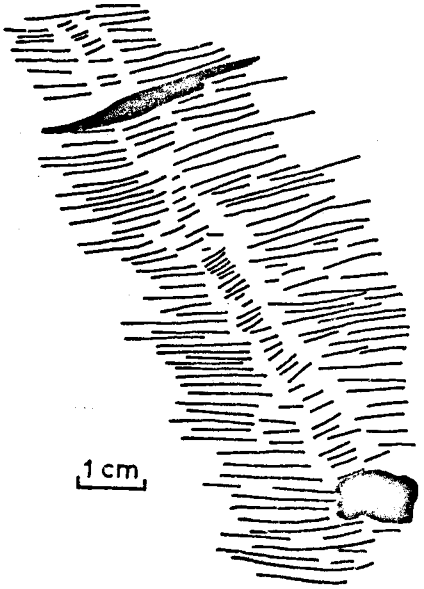
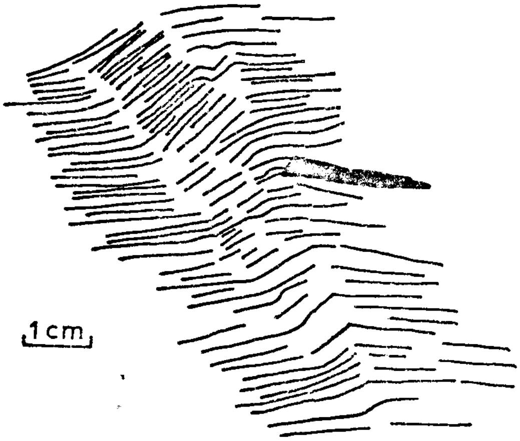

1 cm.

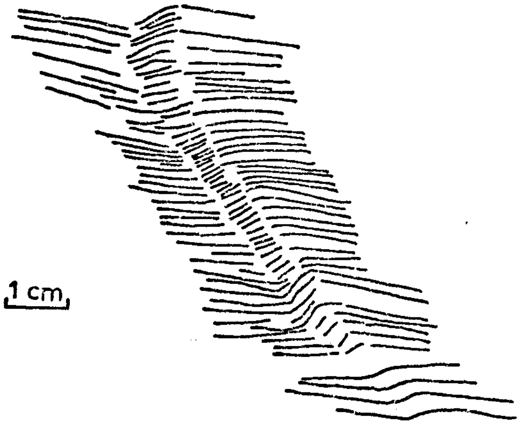
FIGURE 17.

Line drawings of kink bands, all based on photographs.


1 cm


1 cm


1 cm


1 cm

bands the laminae is twice kinked so that within the band it has an attitude significantly different from that outside the band. The sharpness of the bend varies from fold to fold and the domain boundaries are not always parallel. Although no general statement can be made due to the limited amount of data, the more rounded kink bands - those of the larger size (Plate 20), seem to occur in the less thinly laminated rocks. No joints were noted to mark the kink zone boundaries.

Where kink bands developed the cleavage laminae varied from 0.2 mm to 1.5 mm (mean of 0.4 mm) in thickness. α, the angle between the kink band and the cleavage outside the band, ranged from 35 to 85 degrees with a mean of 51 degrees. In general, there is a greater angle (β) (mean of 73 degrees) with the rotated laminae within the band. The width (D), measured parallel to laminae within the band, ranged from 0.5 cm to 9.0 cm with a mean of 1.7 cm. Of the 16 kink bands observed only two were greater than 20 cm in length, the largest being 92 cm. Where the entire length was visible all were noted to terminate in two directions by the convergence of the kink planes, thus giving a lenticular body of disoriented laminae on a flat surface. In nearly every case there was a gradual but marked increase in the amount of rotation of the laminae within the kink band near its extremities (Figure 17).

All the KZBs are nearly vertical which aided in their measurement. The sense of offset across the kink planes was either sinistral or dextral. Seven of the 16 kink bands displayed sinistral rotation while the others displayed dextral rotation. The sinistral kink bands seem, for some unknown reason, to be among the largest of the bands.

The kink bands exhibit an area of dilation which has allowed for more pronounced weathering. The laminae are more widely separated within the band than without and no filling of the dilational areas between laminae was found.

The mean orientation of the sinistral and dextral kink bands was determined using the statistical method outlined by Ramsay (1967) pp. 14-17, and is shown as great circle traces in Figure 18b.

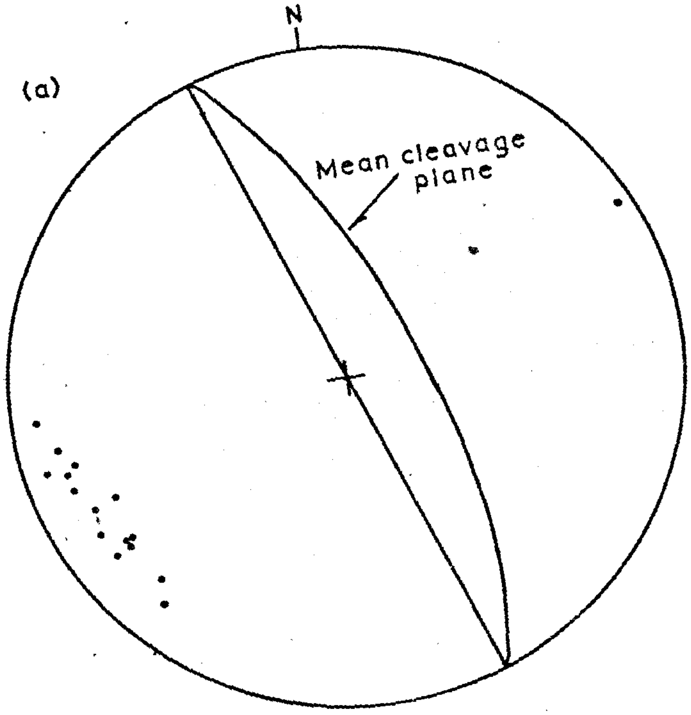
A plot of α against β (Figure 19) shows that in 14 out of the 16 kink bands β was larger than α . When kink bands have developed entirely by slip on the laminae and the slip is confined to the rock between the kink zone boundaries, then β should be larger than α in individual bands (Anderson, 1969). This condition seems to be satisfied in 14 of the kink bands. The two dextral kink bands which have $\beta < \alpha$ may be a result of erroneous measurement or a different deformation mechanism in which there has been slip outside the KZB. The author tends to favour the former considering the difficulty measuring parameters of these small kink bands.

With the exception of one, all the sinistral kink bands were found to have a smaller α (mean of 39 degrees) than the dextral kink bands (mean of 60 degrees) (Figure 19). This is in contrast to what Anderson (1969) found for an area in the Southern Uplands of Scotland. The deviation about a mean value of α is smaller for the sinistral kink bands than the dextral. The mean β for the two types of kink bands is very similar (73 degrees).

FIGURE 18.

(a) Poles to Unrotated Cleavage just outside Kink Band. Orientation of mean cleavage plane shown as a great circle trace.

(b) Poles to Kink Bands. Squares are dextrally - rotated kink bands ; dots are sinistrally - rotated kink bands. Mean orientations of kink bands shown as great circle traces.


(a)
N
Mean cleavage plane

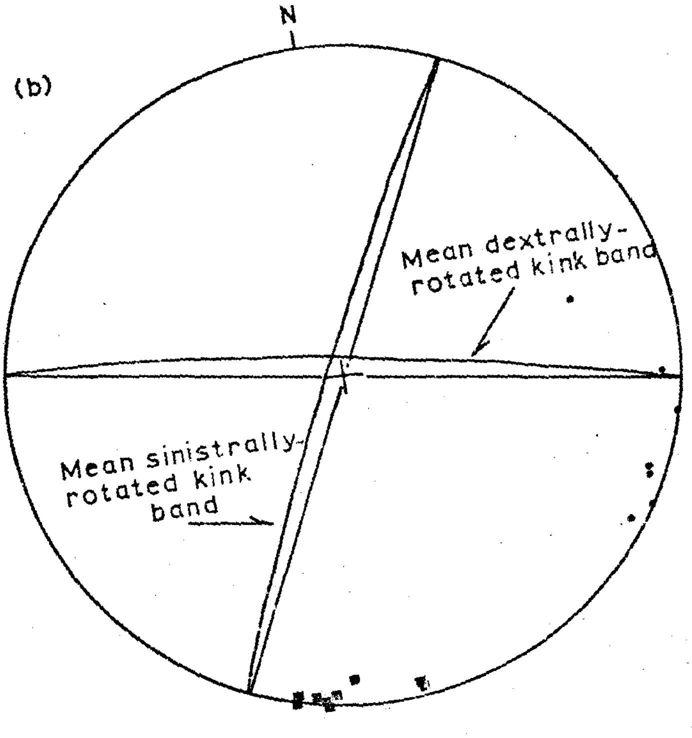
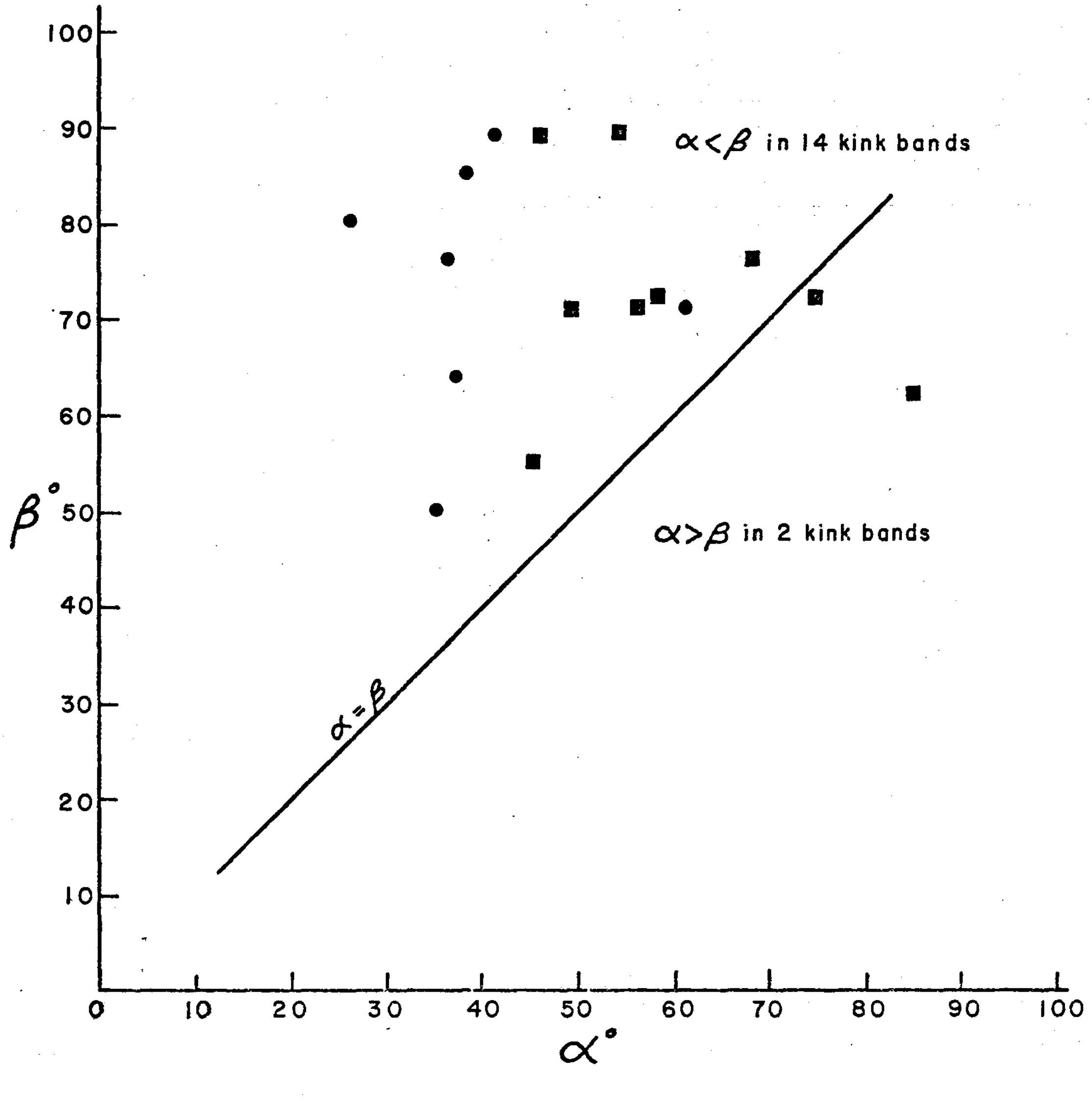

(b)
N
Mean dextrally-rotated kink band
Mean sinistrally-rotated kink band

FIGURE 19.

Plot of α vs. β for each of the 16 Kink Bands. (Small dots are sinistrally - rotated kink bands; squares are dextrally-rotated kink bands).


100
90
80
70
60
50
40
30
20
10
0
10
20
30
40
50
60
70
80
90
100
β°
α°
$\alpha<\beta$ in 14 kink bands
$\alpha>\beta$ in 2 kink bands
$\alpha=\beta$

Responsible Stress Field

Johnson (1956) used kink bands in conjugate array to determine the axis of maximum compressive stress (σ_1). Ramsay (1962) has formalized the use of conjugate kink bands to determine the orientation of the responsible stress field. He considered that the intersection of the conjugate kink zone boundaries coincides with the orientation of the intermediate principal stress axis (σ_2) and their two bisectrices to the maximum and minimum principal stress axes, σ_1 and σ_3 respectively. This use of kink bands to infer the responsible stress field has been shown experimentally to be valid (Paterson and Weiss, 1966).

Although all the kink bands have a single sense of external rotation it seems appropriate to consider the entire suite as if it were actually conjugate; as has been done by Clifford (1969), for the following reasons:

1. the limited areal extent (approx. 300 m^2) over which the kink bands are found, and
2. the proximity of the kink bands displaying sinistral and dextral external rotation.

Whereas conjugate arrays of kink bands are largely a result of σ_1 lying parallel or nearly parallel to the mean undeformed lamination; singular kink bands with a single sense of external rotation occur when σ_1 does not lie parallel to the lamination (Paterson and Weiss, 1966).

Using the assumption that singular kink bands displaying both sinistral and dextral rotation formed at the same time under a single stress system and that the entire suite of kink bands is really conjugate, one gets a mean orientation of the principal stress axes (Figure 20) as follows:

σ_1 = 332;4S

σ_2 = 340;86N

σ_3 = 062;Horiz.

Thus the orientation of σ_1 approximately lies in the plane of the undeformed laminae (335;70NE).

Discussion

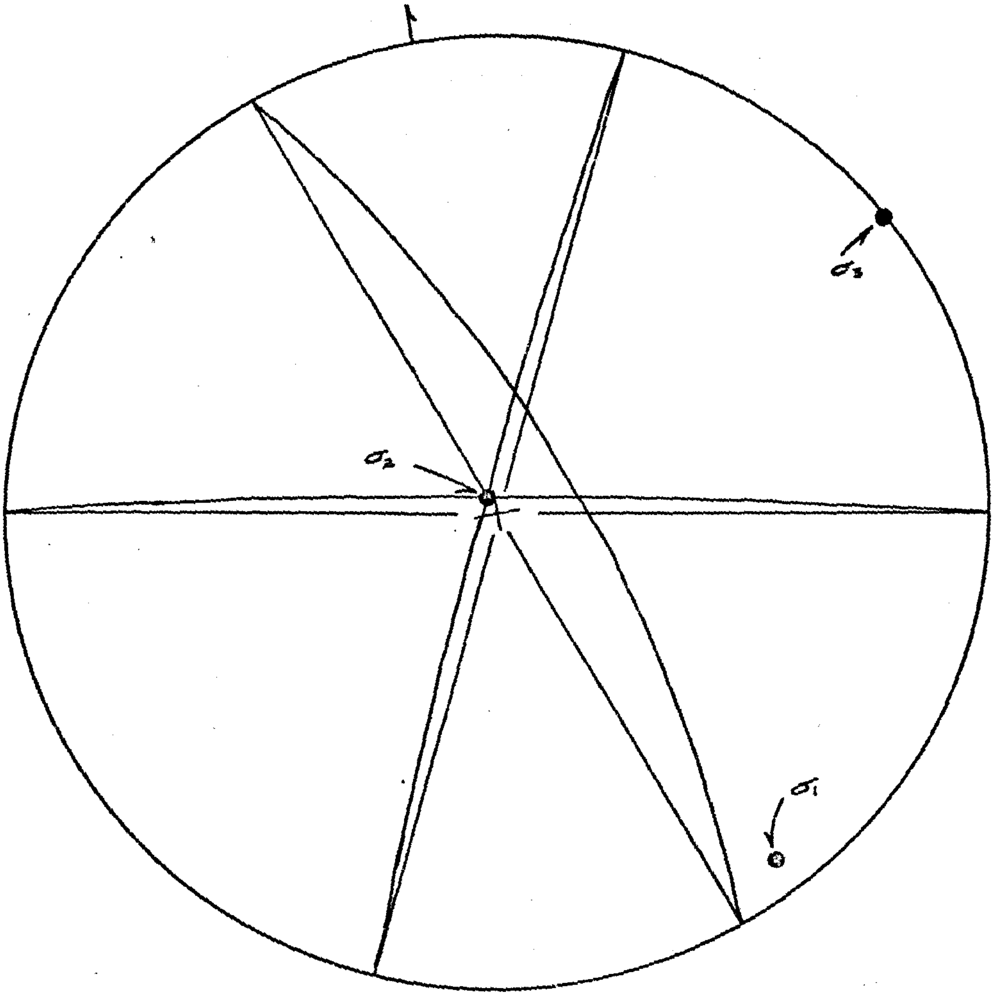
Kink band development appears to be the last deformation to occur in the area. Why do kink bands only occur in this particular area? The answer, no doubt, is a combination of several factors, some of which are:

1. the fulfilment of the requirement of a sufficiently thin rock lamination,
2. the fulfilment of the requirement of a maximum compressive stress oriented nearly parallel to the laminae, and
3. the proximity to the granitic batholith that lies to the south, and the probable effect of higher strain.

It should be noted that if the lamination (cleavage) has developed as a penetrative cleavage as a result of a compressive stress normal to it, as is suggested by the schistosity and other mesoscopic structural elements (meso-folds) in adjacent rocks, then there has been an approximate 90 degree change in the direction of compressive stress between phase F_2 and phase F_3 deformation.

FIGURE 20.

Orientation of stress field responsible for development of kink bands. Great circle traces same as in Figure 18.


σ_3
σ_2
σ_1

The above analysis of the stress orientation assumes that the planar anisotropy (laminae) has no effect on the orientation of the principal stress axes at the time of kink band nucleation. Both structural field studies (Norris, 1969) and experimental studies (Donath, 1963) have shown that planar anisotropy does have an effect in reorienting local stresses under conditions of faulting. Thus the above positioning of the 'responsible' stress system is at best only an approximate one.

Penecontemporaneous Structure

Within the cherty and tuffaceous members of the volcanic sequence are many pre-lithification (soft-sediment) and slump structures. In small outcrops it is difficult, even impossible, to differentiate them from post-lithification or tectonic structures. Since these structures define a study in themselves particular attention was not given to them and only a brief description is included here for completeness. These structures have been grouped for ease of discussion as:

1. primary sedimentary structures,
2. load-induced structures, and
3. gravity-induced structures.

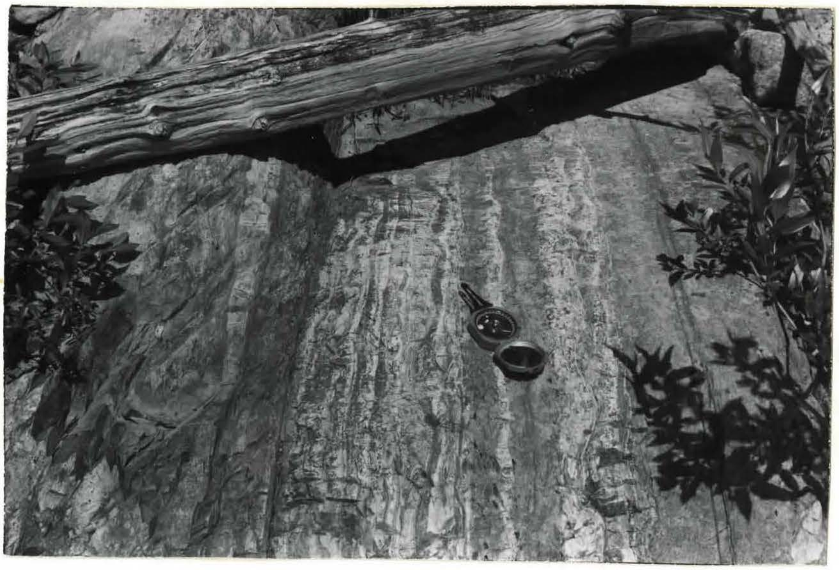
Primary sedimentary structures as used here include only those that have developed in situ or nearly so, at time of deposition of sediment. Such structures consist of ripple marks (Plate 22), graded bedding, and sole marks. Graded bedding was often found in the more tuffaceous members and served in conjunction with other

PLATE 22.

Bedding found in sediments at south end of Cedartree Lake. Brunton compass points N. Stratigraphic tops to the south-east.

PLATE 23.

Convolute bedding in sediments near hinge of Kakagi Lake Syncline (Section E-9). Brunton compass points N. Stratigraphic tops to the east.



evidence in stratigraphic top determinations.

Convolute bedding (Plate 23) is one example of structures caused by load of overlying strata prior to lithification.

Structures observed in the field and thought to have resulted from movement induced by gravity include slump folds (Plate 24), and thrust sheets (Plate 25).

Slumping is due to failure of sediments in shear. Several slump structures were recognized in the field. Such gravity-induced structures in contrast to those caused by load (mass) only, developed after original sedimentation and in certain cases a considerable time interval is evident. They are distinquished from load-induced structures in that their origin involves horizontal mass transport. The only definitive criterion for distinguishing slump structures from later tectonic structures is the truncation of the structures by overlying strata with a clear penecontemporaneous depositional contact. A few such contacts were recognized. A number of workers have described other features which tend to characterize slumping (Fairbridge, 1946; Maxwell, 1962;and Helwig, 1970). Some of those characteristics recognized in the field include:

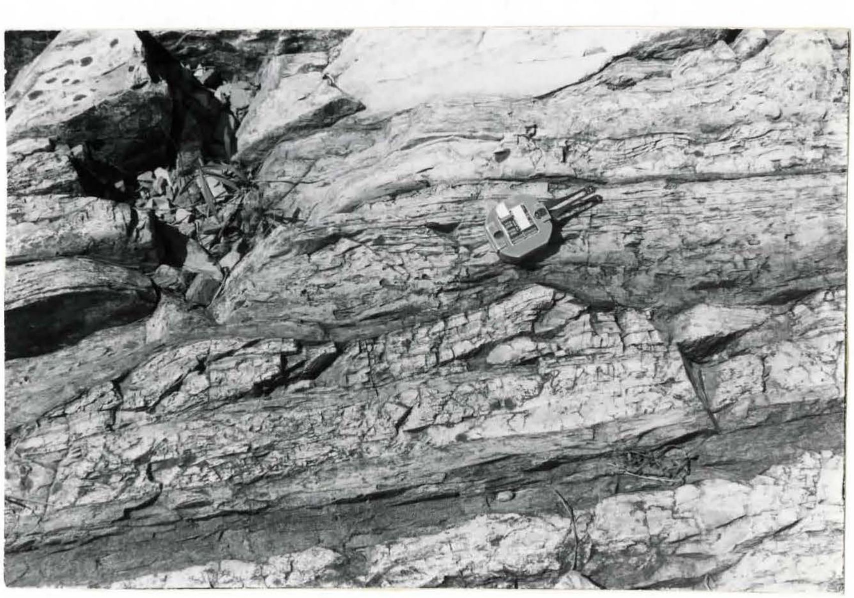
1. disrupted bedding sequence,
2. folds confined to a certain stratigraphic horizon,
3. extremely chaotic folding within a small area,
4, nappe-like folding on a small scale,
5. décollement, and
6. detached fold blobs.

PLATE 24.

Large slump fold on shore of Cedartree Lake (Section F-4). Brunton compass points N. Stratigraphic tops to east.

PLATE 25.

Penecontemporaneous thrusting of sediments. Stratigraphic tops toward top of photograph.



If coherence is maintained during slumping the sediments take the form of slump folds (Plate 26); if coherence is not maintained less defined structures develop (Plate 27).

Minor Dykes and Sills

Both lamprophyre and quartz porphyry intrude the major rock units occurring mainly as dykes. Approximately 80% of these intrusives are lamprophyre.

The lamprophyre dykes range in width from a fraction of a centimetre to over 10 metres with the majority averaging about 15 cm. Although the larger dykes extend for some distance, with one observed to maintain a constant width for 400 metres, many of the smaller dykes display a curved or sinuous plan. Abrupt changes in width are common, often locally pinching out and re-appearing further along strike. The intrusives weather a rusty brown colour. In the fragmental volcanic rocks they form recessive bands which are not well exposed, especially back from the water. Many were probably not observed in the lichen covered rocks.

Contacts of these dykes are extremely sharp. In two places a marginal selvage of quartz up to 2 cm wide was found but, in all other cases the intrusive had a fine-grained chilled margin 1 to 6 mm wide in sharp contact with the country rock. The smaller dykes are often lenticular in both plan and section (Plate 28). They display many different forms in the outcrop surface, ranging from bifurcation with small wedges of host rock separating the two parts, to anastomosing where the lamprophyre repeatedly joins again along strike.

PLATE 26.

Small-scale slump folding in sediments on Kakagi Lake (Section E-9). Brunton compass points N. Stratigraphic tops toward lower right-hand corner of photograph.

PLATE 27.

Sediments on island near south end of Cedartree Lake showing incoherent slumping. Bedding is approximately N-S. Stratigraphic tops to the east (right-side of photograph).

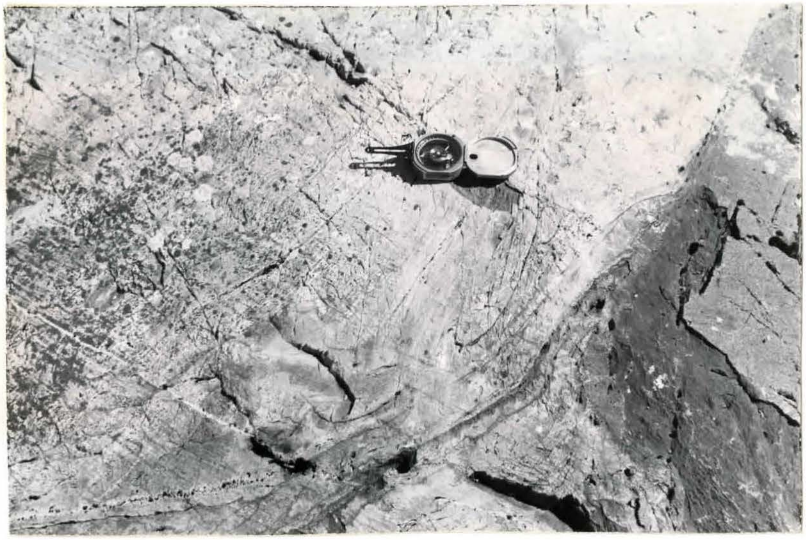


Plate 29 shows a common feature of some of the dykes. The dyke appears to have been offset 25 cm in a dextral sense. No shear or fault plane cuts through or extends beyond the dyke suggesting that this 'offset' occurred during its emplacement. The fact that in many cases where there is an offset of a dyke it is not a clean 'break' but rather has associated with it 'spikes' of the dyke rock protruding beyond on both sides, is further evidence that this offsetting is not caused by deformation (faulting) but is a result of some process of dyke emplacement.

In addition to the lamprophyre intrusives there exists several quartz porphyry and associated felsite dykes. These very coarse to extremely fine-grained intrusives seem more locally restricted with the majority found in the north-west part of Kakagi Lake. The contacts are very sharp and display narrow chill margins next to the host rock. They do not exhibit the variety of shapes like the lamprophyres but instead are generally very straight and continue for some distance. The longest dyke traced was 300 metres in length with its extremities hidden by overburden. Quartz porphyry, in addition to forming dykes and sills, is also found as large masses; an example of which is one of the islands (Figure 3) in the north-west part of Kakagi Lake which consists entirely of quartz porphyry.

Orientation of Intrusives

The orientation of the minor intrusives has been plotted for each of the domains (Figure 21) and the area as a whole (Figure 22).

PLATE 28.

Irregular, narrow, en échelon, north-trending lamprophyre dykes within exposure of fragmental volcanics on island in section C-5.

PLATE 29.

A 10 inch (25 cm) wide lamprophyre dyke strikes 005° through narrow gabbroic sill 250 m N.W. of south end of Cedartree Lake.

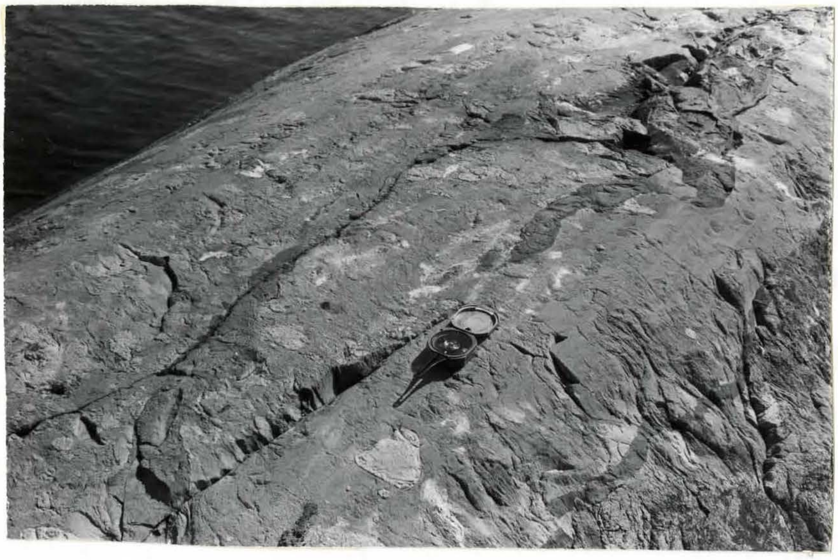


Figure 21 indicates little preferred orientation within the domains. This is probably due in part to the limited number of poles in each projection and in part to the lack of any strongly preferred trend to these intrusives.

When considering the entire area three weak trends appear (Figure 22) : (1) N.W.;80N, (2) N.N.E.;Vertical, and (3) N.-S.;50E. These trends are probably more apparent than real as there is a very high standard deviation in Figure 22. However, it is noted that all of the intrusives dip at angles greater than 45 degrees with a majority of near vertical dip.

Age of Intrusives

As observed by Burwash (1933), these minor intrusives especially in the north half of the area, lack any evidence of shearing or development of s- surfaces other than joints. Where joints are prevalent they do cut across the intrusives. The quartz porphyry bodies are often characterized by short, cross-cutting, quartz-filled tension joints - probably shrinkage cracks (Burwash, 1933).

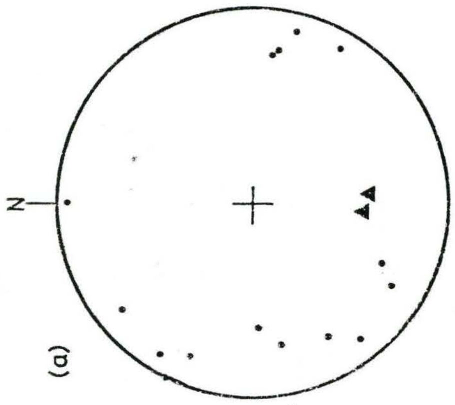
Both the lamprophyre and quartz porphyry intrusives are younger than the volcanic rocks and the large ultrabasic intrusives as shown by their cross-cutting relationships.

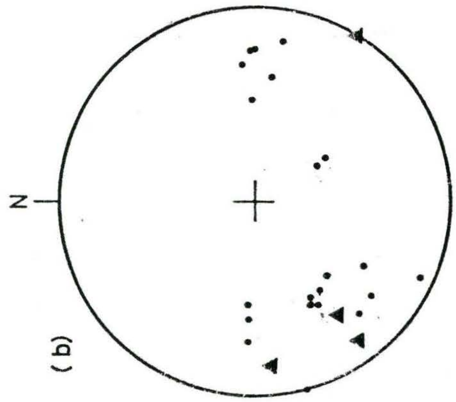
The quartz porphyry and felsite intrusives appear younger than the lamprophyres although the only available information would suggest the contrary. In the south-east corner of section C-5 a pink quartz porphyry dyke, at least 4 metres in width,

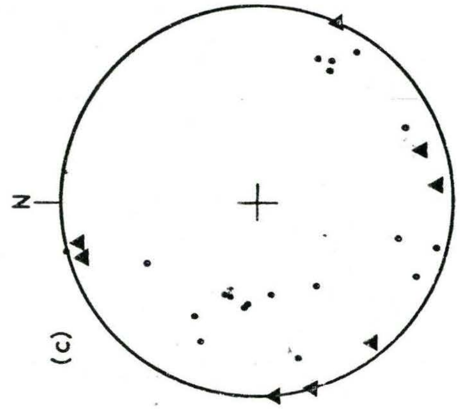
FIGURE 21.

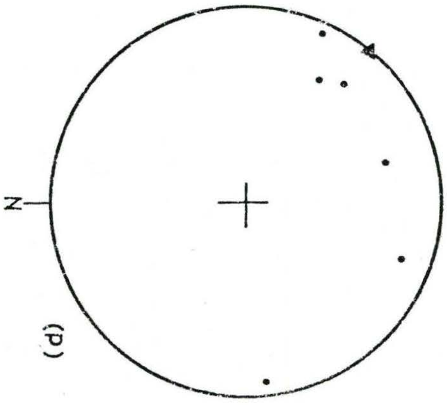
Orientation of Minor Intrusives

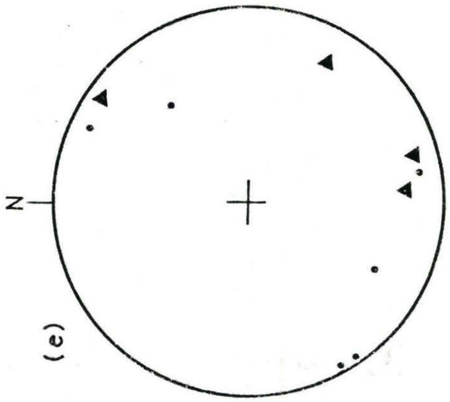
(a) Domain 1: Small dots are lamprophyre; triangles are quartz porphyry and felsite.
(b) Domain 2: Symbols same as (a).
(c) Domain 3: Symbols same as (a).
(d) Domain 4: Symbols same as (a).
(e) Domain 5: Symbols same as (a).
(f) Domain 6: Symbols same as (a).


(a)


(b)


(c)


(d)


(e)

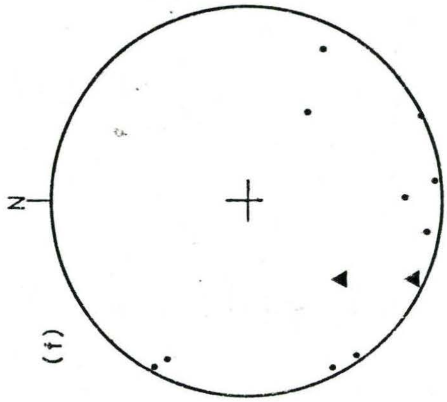

(f)

FIGURE 22.

Equal-Area Net Projection of Total Minor Intrusives (lamprophyre and quartz porphyry / felsite) over Entire Area. Total number = 99. Contours at 1,3,5 (6) % per 1 % area.

contains a metre wide and seemingly concordant, lamprophyre body (sill) about mid-way between its contacts. This lamprophyre appears to have intruded into the quartz porphyry since it seems to have a chilled margin against the porphyritic intrusive. Other than this isolated case no other information was found to give the relative age of these two types of intrusives.

In a few places black lamprophyre dykes were noted to cut small quartz veins (Plate 30) indicating that some of these intrusives are younger than some quartz veins.

No data was found that might give the relative age of these two types of minor intrusives and the small granitic bodies.

Quartz Veins

Quartz veins are prominent throughout the area. They are common in the volcanic rocks and in particular the more acid fragmental units. A few were found in the basic intrusive rocks, usually confined to the gabbroic and dioritic members.

These veins occupy two different structural environments - joints and shear zones. A milky white, coarse - grained quartz is the only common constituent of the veins. In places the quartz is accompanied by varying amounts of epidote, calcite and chlorite. The calcite generally occupies the centres of the veins and thus post-dates the deposition of the silicates. This calcite is often iron-stained resulting in a brown coloured mineral. The epidote and chlorite are found

PLATE 30.

A narrow lamprophyre dyke cuts through both fragments and small quartz veins. Extension of largest quartz vein in intrusive indicates sense of flowage within dyke (i.e. from east to west). Minor sinistral displacement across dyke. Compass points N.

both at the selvage of the quartz and included within it.

The quartz veins that occupy joints vary in width from less than a millimetre to several centimetres. These veins occupy tension joints and generally have a characteristic lenticular shape in both plan and section. Although some of the veins extend great distances laterally, most are short often being less than a metre. To be noted is the abundance of tension joints that do not have any mineral fillings. Usually the quartz completely fills the joint but occasionally open spaces remain in the centre into which comb quartz protrudes. Where there is a high density of tension joints they are generally not completely mineral-filled but gape in the centre.

Few veins are found in the shear zones. They do not display the sharp, straight contact with the host rock as do those occupying tension joints. Their width may vary considerably along strike but they extend for greater distances than the joint fillings. In places the brecciated aspect of the quartz indicates post depositional movement. Shear zone fillings have a much higher percentage of other minerals associated with the quartz, mainly chlorite, epidote and iron-stained calcite. Chlorite and calcite, in particular, form pods along the shear zone.

Although most veins are found in the two environments mentioned above, quartz-filled gash fractures often sigmoidalshaped are noted locally.

Orientation of Veins

The attitude of the veins within each domain are plotted on the lower hemisphere equal-areaprojection as shown in Figure 23. The results of the plot may be summarized as follows:

Domain 1: maxima at 302;48N.E.; sub-maxima at 302;64S.W.
Domain 2: maxima at 284;56N.; sub-maxima at N.-S.;60W.
Domain 3: maxima at 327;70N.E.;sub-maxima at 008;64W.
Domain 4: maxima at 309;54N.E.
Domain 5: maxima at 282;84S.
Domain 6: maxima at 346;72W.;sub-maxima at 082;59N.

Figure 23 shows a definite change from a more preferred orientation of quartz veins in the north to a more random orientation in the south, reflecting a more complex structure to the south. In the north most of the quartz veins represent fillings of tension joints that have developed on the limbs and hinge area of the large syncline, whereas in the south a higher percentage of the veins are found along shear zones. The latter helps to explain the more random orientation of the veins in the south since the shear zones tend to be more random in orientation than the tension joints.

Origin of Veins

The silica that gave rise to the quartz veins in the volcanic rocks is thought to have been secreted out of the host

FIGURE 23.

Orientation of Quartz Veins

(a) Domain 1: Poles to quartz veins contoured.

(Except those in carbonate unit.)

n= 131

(b) Domain 2: Poles to quartz veins contoured.

n= 66

(c) Domain 3: Poles to quartz veins contoured.

n= 98

(d) Domain 4: Poles to quartz veins contoured.

n= 53

(e) Domain 5: Poles to quartz veins contoured.

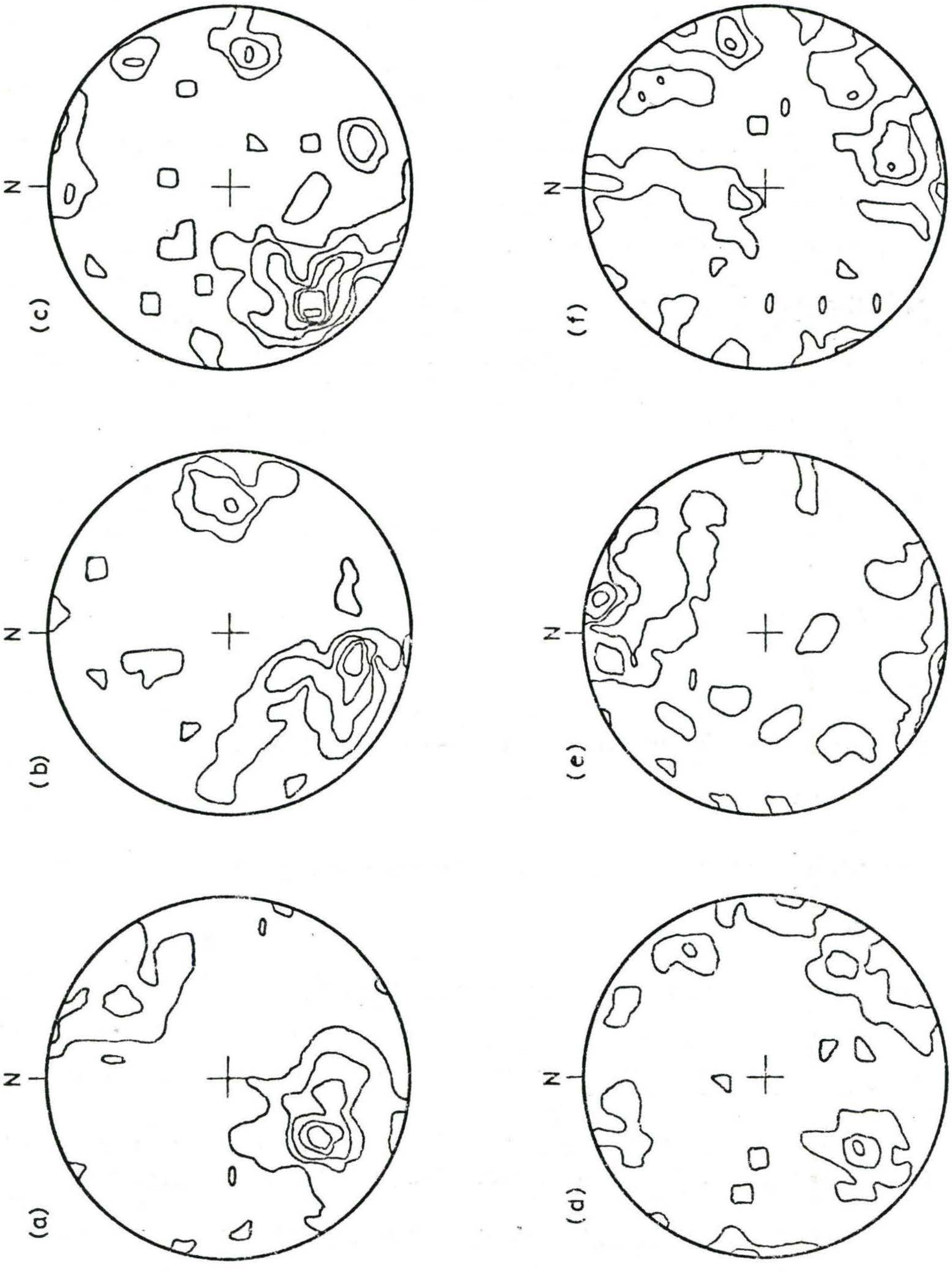
n= 61

(f) Domain 6: Poles to quartz veins contoured.

n= 77

Note: all lower hemisphere equal-area net projections;

contours at 1,2,5,7,9,11 poles per 1% area.


(a)
N
(b)
N
(c)
N
(d)
N
(e)
N
(f)
N

rock and migrated to the dilational (low pressure) area that resulted from the tensional nature of the joints. The following is given in support of such a hypothesis:

1. a comparison of the orientation of the quartz veins, especially in the north area, and that of the joints in the same section indicate that they do occupy tension joints.
2. the fact that most of the veins are found in the more acid rock types.
3. the completeness of the filling of the fractures depends on the number of fractures within a small area and the distance between adjacent fractures. The closer spaced the fractures the less complete their filling.
4. the fact that mineral-filled veins are rare in shear zones. The shear zones are probably compressional features in contrast to the joints which are tensional in nature.
5. the characteristic shape of most of the veins - lenticular, short, vuggy and straight-sided.
6. the fact that comb quartz often protrudes inward from the edge of the fracture.

Quartz Veins within Carbonate Unit

Numerous quartz veins are found in the carbonate unit situated at the base of the large ultrabasic sill in sections

C-4 and C-5 (Figure 3). They have been described separately from the rest of the quartz veins because of :

1. their extreme abundance relative to the area as a whole, and
2. their characteristic morphology - sigmoidal for the most part.

Plate 31 shows, in addition to their abundance, the general shape and orientation of these veins. They appear to occupy both sigmoidal gash fractures and near horizontal tension or release fractures. None extend continuously for a great distance; the longest noted being less than seven metres. In addition to quartz-filled fractures there are a few calcite veins. Very few fractures exist that are not filled with one of the two minerals.

Equal-area net plots were constructed from data obtained at four different stations (Figure 24a) along this carbonate unit (Figures 24b-f). The plots show the dominantly sub-horizontal attitude of the veins. The strike of the veins display a marked variation from station to station, a probable reflection of variation in direction of shear within the carbonate rock.

Origin of Veins

It has been suggested that the carbonate rock has resulted from replacement of tuff and breccia by volcanic emanations (Ridler, 1966). As described by Ridler, the rock in thin section (Plate 32) is composed mainly of massive anhedral grains of carbonate. Many of these grains display excellent rhombohedral cleavage

PLATE 31.

Quartz veins and quartz-filled sigmoidal gash fractures within carbonate unit (section C-5).

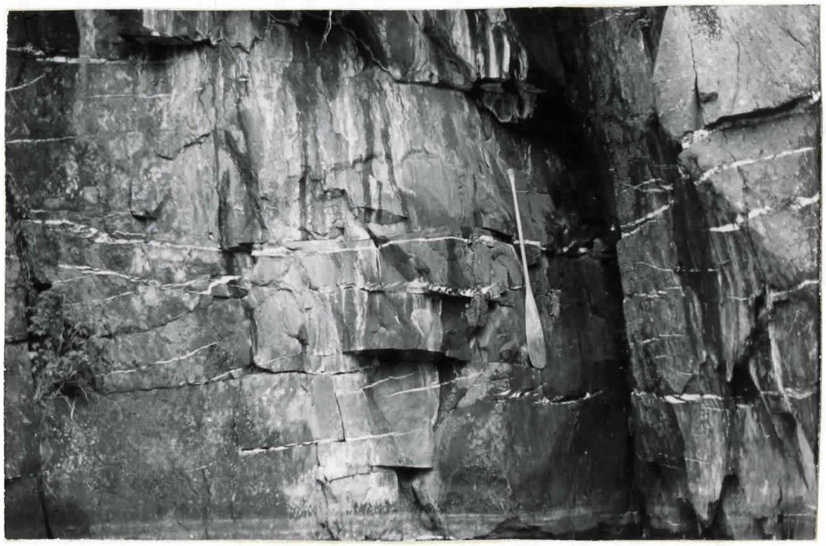


FIGURE 24.

Quartz Veins in Carbonate Unit

(a) Sketch map showing location of stations where quartz veins were observed.

Note: In (b) to (f) dots are poles to quartz veins; contours at 1,3,5,7 poles per 1% area.

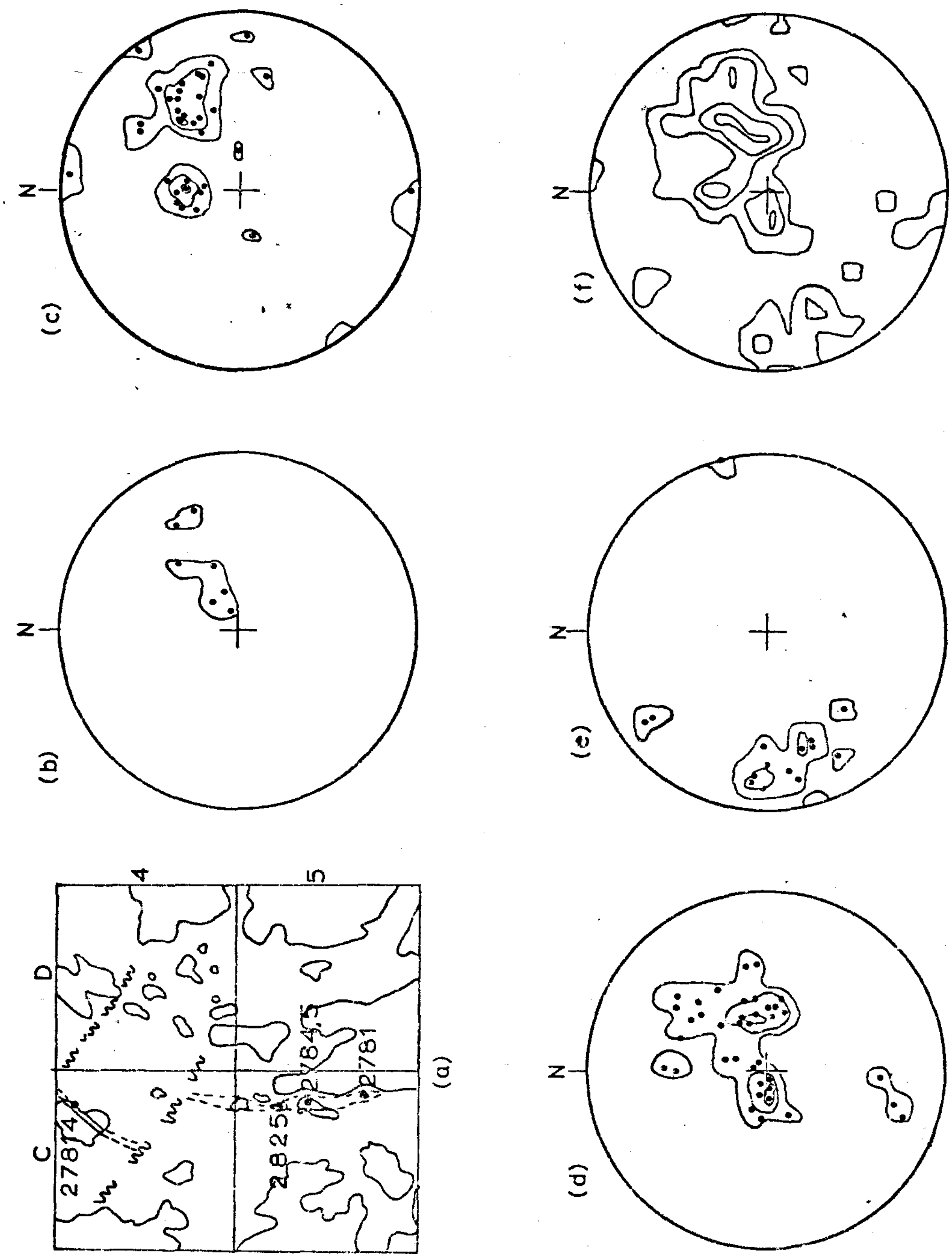
(b) Poles to quartz veins at station 2781. n=7.

(c) Poles to quartz veins at stations 2784 and 2785. n=34.

(d) Poles to quartz veins at station 2825. n=38.

(e) Poles to quartz veins at station 27814. n=15.

(f) Contoured plot of total poles to quartz veins found in carbonate unit. n=94.


(a)
C
D
4
5
27814
2825
2784.5
2781
(b)
(c)
(d)
(e)
(f)
N

with varying amounts of yellow to rusty coloured stain. In some sections numerous veinlets of quartzo-feldspathic material cut through the rock. A light green micaceous mineral is ubiquitous and thought to be a chrome-rich mineral (Ridler, 1966). Small pyrite cubes and 'pieces' of pyrite were present in some of the slides examined (Plate 33).

The carbonate unit due to its position between the two more brittle units - acid fragmental volcanics to the west and the ultrabasic sill to the east has acted as a band of intense shearing during the folding of the resistant intrusive units. Evidence of shearing is indicated by :

1. the elongation of some of the carbonate grains,
2. the texture outlined by the sub-parallel seams of the green mineral (chrome-rich mica),
3. the common undulatory extinction of quartz,
4. the pressure shadow zones (now filled with quartz) behind the pyrite grains, (Plate 33),
5. the fragmentation of the pyrite, and most important,
6. the tension gashes (Plate 31) observed in the outcrop.

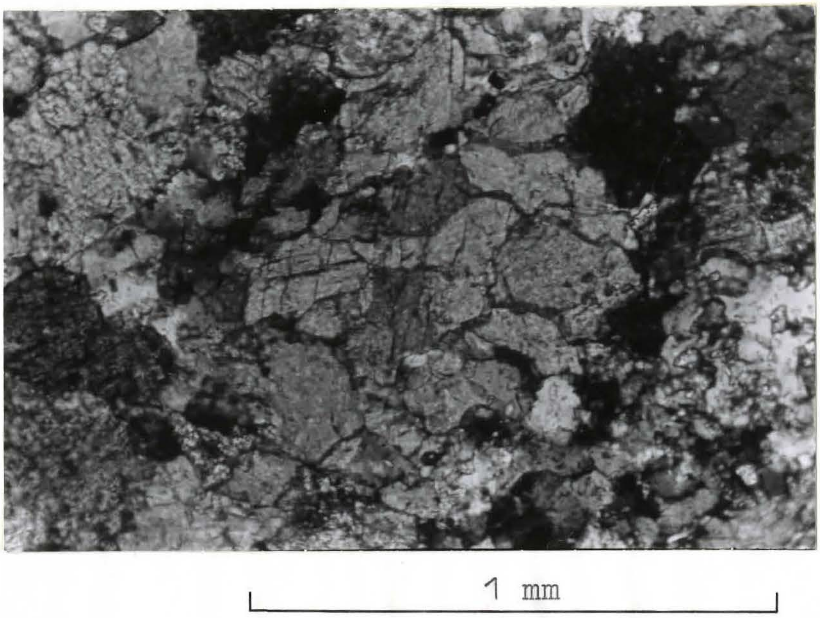
This shearing formed the tension gashes and other fractures that were later filled with quartz. The carbonate that fills a few of the fractures probably has been derrived locally from the host rock, while the silica probably originated from the original volcanic rocks or from the adjacent volcanic rocks stratigraphically

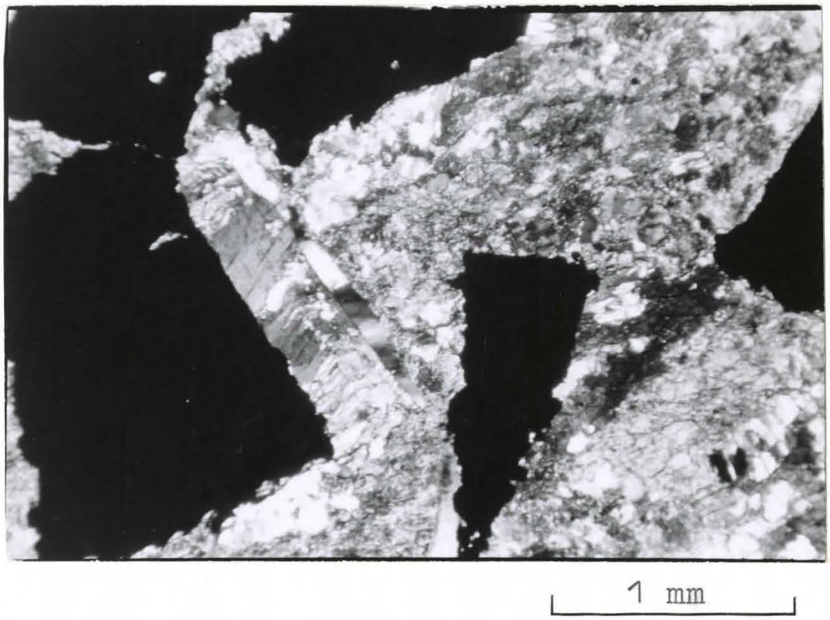
PLATE 32.

Anhedral carbonate grains showing rhombohedral cleavage. Dark mineral is chrome-rich mica. (X-Nicols).

PLATE 33.

Pressure shadow zones behind pyrite (black) filled with quartz. (Plane light).


1 mm


1 mm

below the carbonate unit.

Deformed Quartz Veins and their Relation to Local Schistosity

Although most of the quartz veins found in the area are straight and not deformed some show the effects of compression and extension along their length (Plates 34,35). Few deformed veins extend any great distance and are thus of little value in a quantitative assessment of the resulting strain. Both buckles and stretched (boudinage) veins were observed in the field. Where buckled veins could be traced for any lateral distance, the following measurements were recorded:

1. the wavelength of the buckle,
2. the thickness of the vein,
3. the length of a segment of the buckled vein measured along the central axis, and
4. the length of the same segment as in (3) above by measuring along a line joining the consecutive inflexion points of the buckles.

Unfortunately very few viens could be measured in the above manner due to their poor exposure, lenticular morphology, and limited horizontal extent. The location and mean attitude of the deformed quartz veins is given in Figure 25.

Biot, <u>et al</u> (1961) and Ramberg (1963) have shown experimentally that the wavelength to thickness ratio of a buckled competent sheet of rock is a measure of the contrast on competency between the enclosed sheet and the host rock. For the case of

PLATE 34

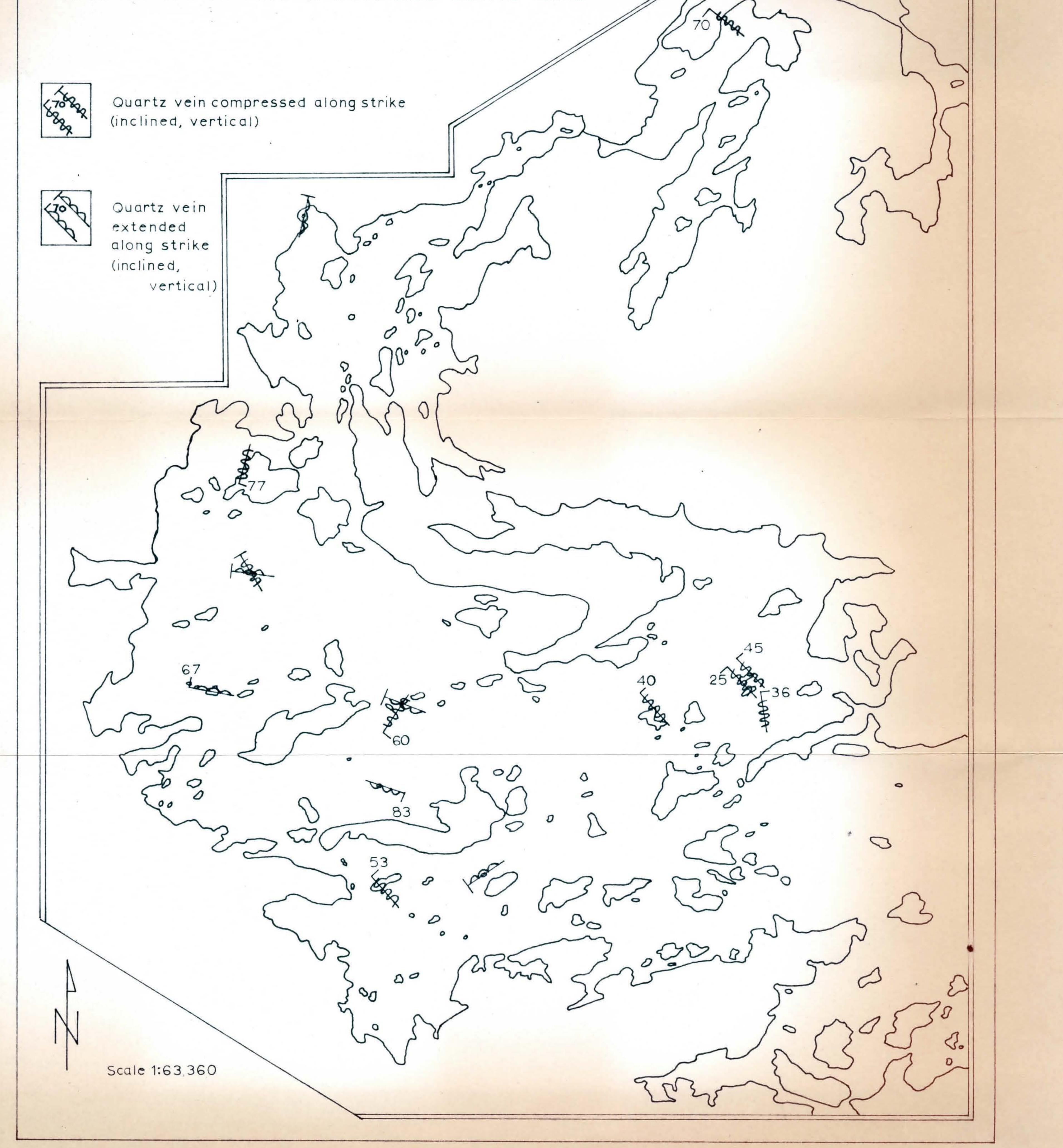
Ptygmatically folded quartz vein on the northwest shore of Cedartree Lake (section G-1), Note elongation of fragments normal to quartz vein.

PLATE 35

Buckled ribbon of quartz in fragmental volcanic rock on west shore of Kakagi Lake (section C-5).

Figure 25

Location and Attitude of Deformed Quartz Veins



elasticity we have: $\frac{\lambda}{t} = 2\pi\left(\frac{1}{6}\ \frac{G_1}{G_2}\right)^{\frac{1}{3}}$; and for materials displaying Newtonian viscosity: $\frac{\lambda}{t} = 2\pi\left(\frac{1}{6}\ \frac{\mu_1}{\mu_2}\right)^{\frac{1}{3}}$

where; λ = wavelength

t = thickness

G = rigidity, and Subscript 1 = buckling layer

μ = viscosity subscript 2 = host rock

Since rocks follow neither of the above cases but fall in between the two ideal cases, use of either of the above formulae is only an approximation to the real behaviour of the rocks.

If enough data had been available one might have found, as did Ramberg and Ghosh (1968), that the λ/t value was fairly constant within each outcrop although the dimensions themselves varied considerably. Since rocks do not behave in a strict Newtonian sense their λ/t value depends not only on the ductility contrast but also on the intensity of the stress field. This would, as pointed out by Ramberg and Ghosh (1968), result in a different λ/t value from place to place. One would expect that if numerous measurements could have been obtained in the Kakagi Lake area not only would areas of high stress concentrations within a single macro-fold be outlined but there would also be a noticeable increase in strain toward the granitic mass to the south. The effects of contact strain may also have shown up near the ultrabasic sills. The above concept is based on the assumption that we are dealing with materials of the same contrasting ductility throughout the area. This assumption would be approximated if we restricted our study to only those deformed quartz veins found in the fragmental volcanic rocks. Unfortunately, in this study area, data is lacking and the above has to remain as 'theoretical thinking'.

Measurement of the type 3 and 4 give a measure of the minimum shortening parallel to the length of the vein; five such measurements were made at three stations in the field and the results are given as Table 2. These measurements were done on veins that could be traced for a distance greater than two metres and were fairly uniform in thickness.

Examples of good boudinaged quartz veins were even more difficult to find than buckled veins. Generally, only one or two boudins could be recognized in an area thus making it difficult to distinguish them from small lense-shaped tension joint fillings. The few noted, however, were described and measured in as much detail as possible in a fashion outlined by Ramsay (1967). Some of the boudins found at the most southern deformed quartz vein showed evidence of having been rotated 10 degrees clockwise about their length. In only one exposure did both buckled and boudinaged veins exist together.

The deformed quartz veins appear to be genetically related to the local schistosity but not enough data is available to show this conclusively. If the schistosity is an axial plane

TABLE 2.

Minimum Shortening of Quartz Veins

Location	Attitude	Width	Host Rock	Shortening
(a) Section G-1	310;70S	1.2 cm	Fragl. volc.	10.2%
(b) Section G-8	350;36E	2.5 cm	Non-fragl. volc.	17.6%) 7.1%) 12.4%
(c) Section F-8	327;67N	0.5 cm	Fragl. volc.	13.8%
"	320;40N	0.8 cm	"	7.1%

Note: See Figure 25 for location of veins.

schistosity as suggested then it coincides with the plane of maximum extension in the strain ellipsoid. The fact that boudinaged veins lie parallel to sub-parallel to the trend of the schistosity and buckled veins are normal - or sub-normal to it tends to support the idea that the schistosity has formed normal to the local direction of maximum finite compressive strain.

Horizontal Shortening and Flattening

Shortening indicated by the outline of the large ultrabasic sill in Kakagi Lake is nearly horizontal and about perpendicular to the main schistosity (S_1).

Measurement along the sill in Figure 3 gives a shortening close to 70% assuming the sill was initially sub-horizontal. The theoretical maximum shortening permitted in perfect concentric folds is 36% (de Sitter, 1958). Ramsay (1962) has pointed out that flattening in the competent beds may take place before the concentric folding maximum is attained. As the sills show some thickening at the fold hinges, the above shortening is likely to be a minimum.

It follows that the associated volcanic rocks have undergone a similar amount of shortening. Due to lack of stratification one is unable to see their suspected similar geometry that has resulted from both brittle and ductile deformation. The volcanic material acquired an axial-plane cleavage (schistosity) probably at an early stage of the shortening. Associated with this schistosity is the flattening and orientation (or re-orientation) of the volcanic fragments. Continued shortening resulted in the development of the large and small scale dislocations and shear zones.

Fragment Orientation and its Relation to Schistosity

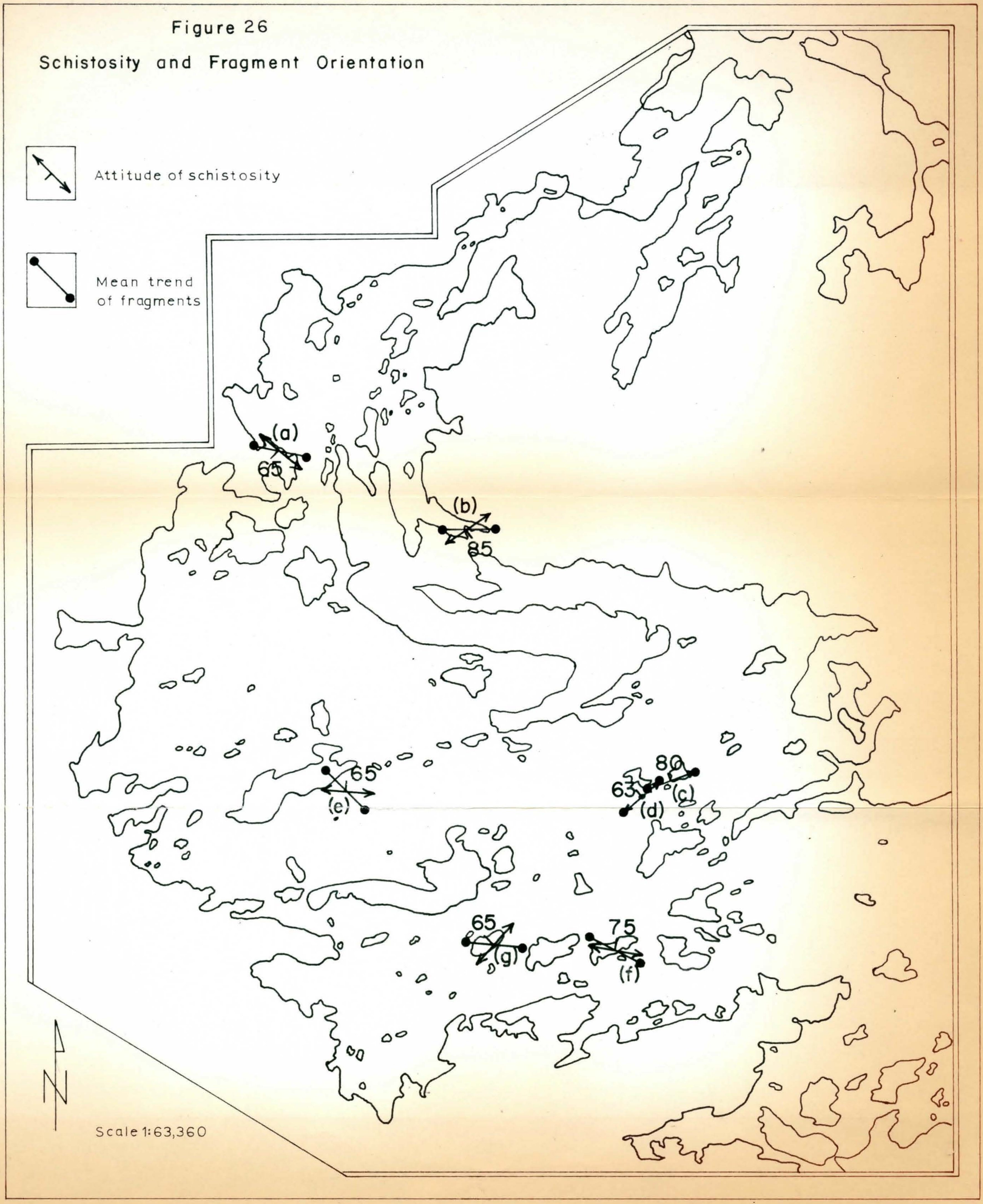
As mentioned earlier, many of the fragments throughout the study area display an orientation. This orientation is defined as the direction of the apparent long axis of the fragments.

The length of the major and minor axes of the fragments as displayed in the outcrop surface were recorded at a number of stations. In addition, the orientation of the major axis (apparent) and the slope of the surface on which the measurements were made was determined. In most of the exposures where measurements were made (Figure 26) a weak, but discernible schistosity was evident and thus recorded. Figure 26 shows the location of the stations where measurements were made and the mean orientation of the fragments as well as the trend of the schistosity.

No measurements were made on Cedartree Lake because of apparent random orientation of fragments and lack of any obvious penetrative deformation. On the north-west limb of Cedartree Lake syncline the orientation of fragments is usually random (Plate 8) however in a few places a preferred orientation exists and it appears to be parallel to subparallel to primary bedding. Some elongation of fragments has occurred locally near shear zones but for the most part the fragments in this area are thought to have been originally deposited with their long axis in the depositional

Figure 26

Schistosity and Fragment Orientation



plane and have suffered little deformation (elongation) or rotation.

In the field the fragments in the southern half of the map area appeared to have their long axis lie in the plane of schistosity. However, measurements of fragments at some stations indicate that such is not always the case. Seven graphs have been plotted for locations where good schistosity attitudes could be obtained (Figure 27). The number of fragments measured at any one station varied from 5 to 31 with an average of 20. Each of the graphs shows the apparent axial ratio (R) of the fragments as well as the angle (ϕ) between the apparent long axis and the schistosity (Figure 27). The axial ratio varies from 1.98 to 3.28 with a mean of 2.49. At some stations (Figure 27c) R displays only a small spread about a mean but in others (Figure 27g) the deviation about the mean is large. Some of the stations (Figure 27c) show a near symmetrical distribution of their long axis about the schistosity while others indicate angles greater than 50 degrees about the schistosity trend.

A rough correlation between ϕ and the distance from the axial plane of a fold might be suggested by the data; the closer to a fold axis the smaller ϕ seems to be. This may, however, be more apparent than real.

Deformation with respect to the Strain Ellipse

Ramsay (1967) has shown that two dimensional strain represented by an ellipse may fall into definite well defined

FIGURE 27.

Orientation of Fragments with respect to Schistosity

R = apparent axial ratio of fragments.

ξ = angle between apparent long axis and schistosity.

$-$ = dextral rotation (the long axis must be rotated in a clockwise direction to parallel the trend of the schistosity in the rock).

$+$ = sinistral rotation (the long axis must be rotated in a counter clockwise direction to parallel the trend of the schistosity in the rock).

Note: The location of (a) to (g) is shown in Figure 26.

fields (Figure 28). When strain occurs in the plane of a competent material such as a quartz vein embedded in a more ductile matrix, different types of minor structures will develop in the layer according to the deformation field in which the strain ellipse is positioned. The minor structures found in the study area include, as mentioned previously, buckled and stretched quartz veins. In general, these veins are at a high angle to one another within a small area (Figure 25). Near the hinge of Kakagi Lake Syncline (Domain 4; Section F-10), deformed pillows (Plate 36) indicate a general N-S compression and E-W extension. The ptygmatic veins and buckled pillow selvages were generally oriented about normal to the general schistosity trend; this is to be expected if both features have resulted from compression. The long axis of fragments, where present, was often approximately parallel to sub-parallel to the schistosity and the extended quartz veins.

In summary, we seem to have over the entire area, at any one location and during a single phase of deformation, one principal extension and one principal contraction. Thus deformation in the area falls within field 2 of Figure 28.

FIGURE 28.

The Three Fields of the Strain Ellipse and their respective Geological Structure (after Ramsay, 1967).

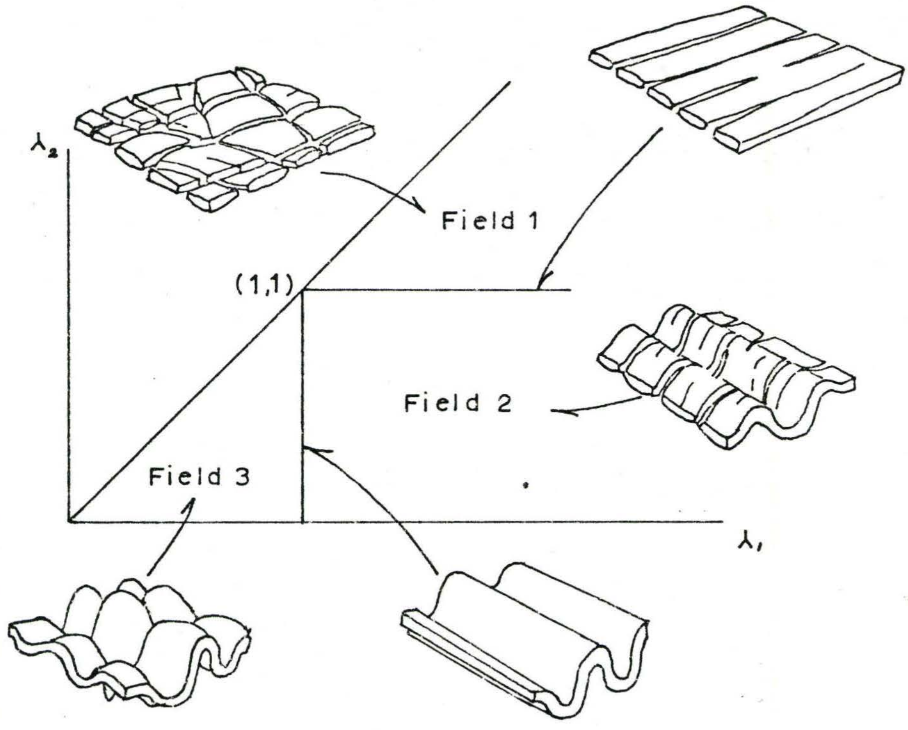
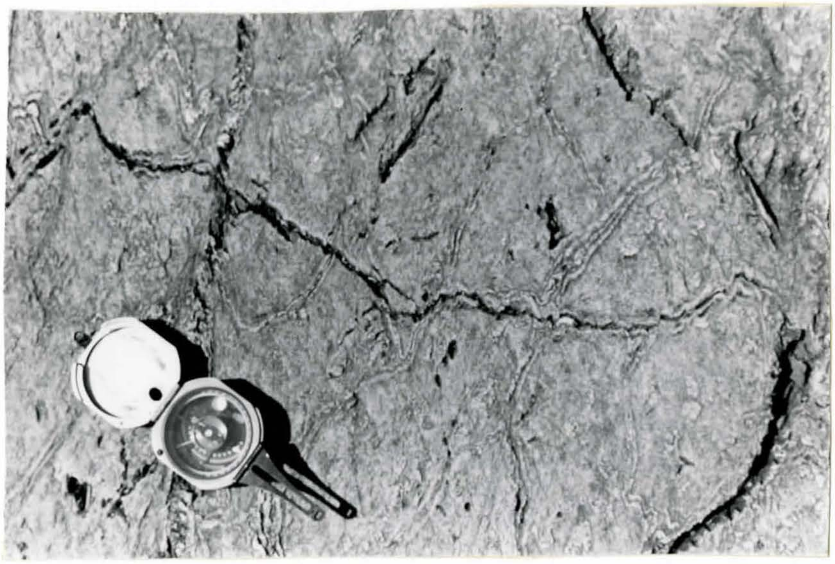

λ_2
Field 1
(1,1)
Field 2
Field 3
λ_1

PLATE 36.

Deformed pillow lava near hinge of Kakagi Lake Syncline I. Pillows have been compressed in a N-S direction and stretched in an E-W direction. Mean axis of buckles (27 measurements) was 066° with variable plunge.



Jointing in the Area

Well developed joints are the most prominent structural feature in much of the study area. Joints which are planar and parallel or sub-parallel so they form 'sets', are said to be systematic (Price 1966). A total of 5,132 systematic joints were observed and their attitudes recorded.

A joint is a fracture displaying negligible relative displacement of the walls and is a member of a group of fractures spatially extensive in three dimensions. In this study no differentiation was made between a joint and a fissure as defined by Mitcham (1963).

The joint frequency varies considerably within a small area. Although the data recorded is roughly proportional to the number of well developed joints in an outcrop they were not systematically recorded for the overall area. Depending on the character, size of the outcrop etc. the number of measurements at each station varied from zero to well over 100.

Joint development is particularly dependent on lithology. The more massive and finer grained rocks show more and better developed joints. The sediments, due to their bedding, display much jointing, as has been found by many workers. In the fragmental unit the higher density of fragments and the larger the fragments, the better developed the joints and less well developed the schistosity. In the large ultrabasic intrusives few joints are present; even fewer are found in the late small black lamprophyre and quartz porphyry dykes and sills.

The joints recorded in the field fall into two broad categories:

(1) Shear joints: characterized by long, straight, and clean fractures.

(2) Tension joints: characterized by short, lenticular, gapeing (when not mineral-filled), rough fractures. The attitude of the shear joints is much more consistent than that of the tension joints in any one outcrop. Whereas, the tension joints are often filled with such minerals as quartz, chlorite, and epidote, the shear joints are seldom filled with any minerals. As shown earlier, equal area plots of quartz veins in the different domains indicate that they are merely infillings of these tension joints.

All joints recorded in the field were plotted for each of the sections (Figure 29). Data was contoured when the net contained 50 or more poles to joint planes and three most prominent joint planes are shown as great circle traces in the contoured diagrams. These plots are at best of a quasi-quantitative nature. Equal numbers of joints were not recorded at each station (outcrop) or within equal areas. Many criticisms have been levelled at the use of Schmidt plots in the statistical treatment of joints (Rodgers, 1952). The author, however, feels that since the field data in this particular case does not warrant much statistical treatment, these plots are satisfactory for showing general trends. Nevertheless, it should be noted that since the plots have been contoured as percent poles per unit area therefore the greater the

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M.Sc., McMaster, ~~1971~~.
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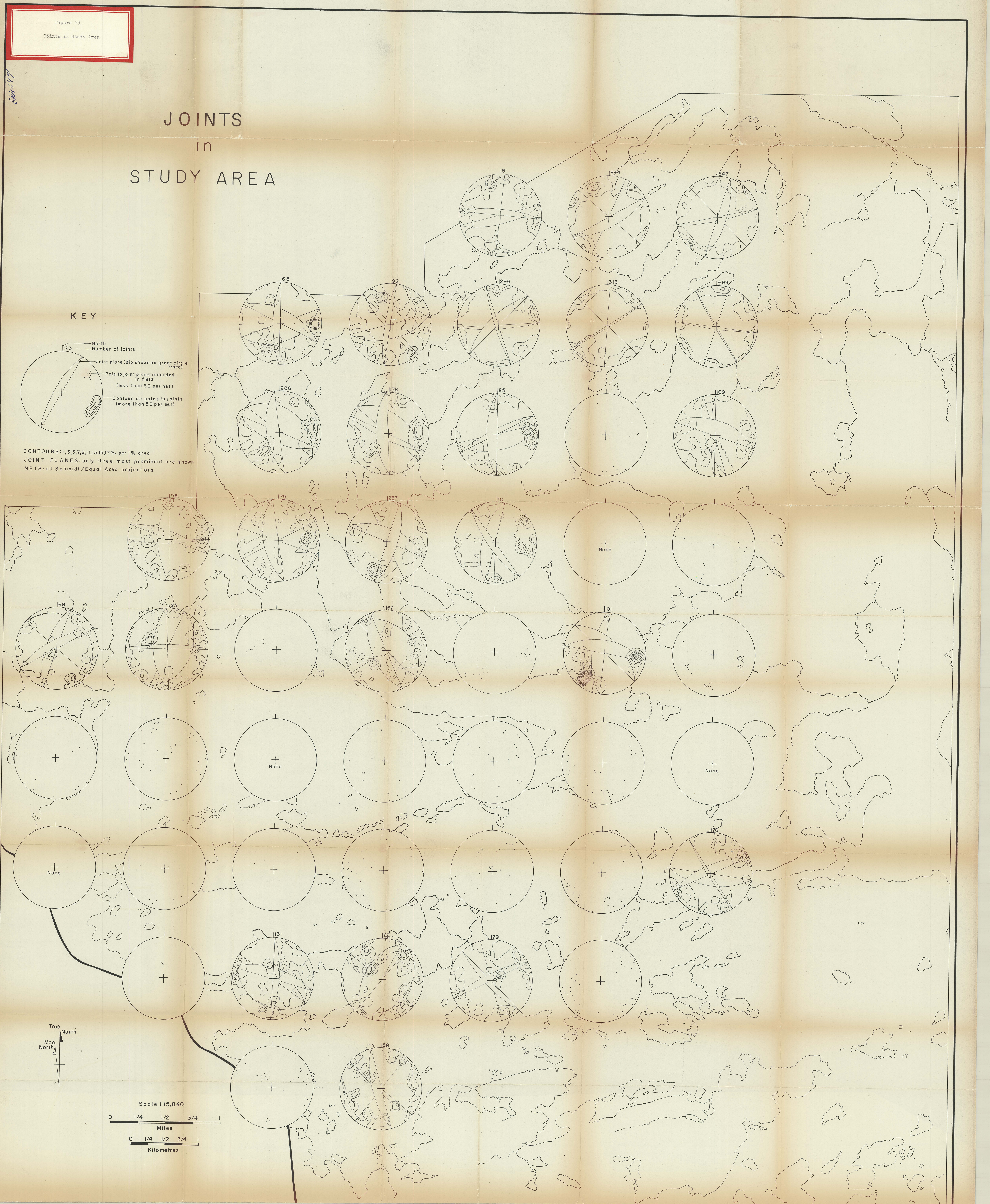
Figure 29

Joints in Study Area

6H4097

JOINTS
in
STUDY AREA



number of data the less the apparent standard deviation.

Interpretation of Jointing

It was noted both in the field and from Figure 29 that the joints were more systematic within an area (eg. a section) in the north than in the south. This probably reflects the more complex structure and varying lithology in the southern part of the study area.

Systematic joints have been found by various workers to show well-defined relationships to structures providing they developed during the same tectonic period.

The joint pattern found superimposed on the Cedartree Lake Syncline appears to be related to this structure. Those joints striking N.W. are tension joints, many of which have subsequently been mineral filled (Figure 23), while the predominantly north-east trending joints are shear joints and in some places (e.g. Section F-2) coincide with bedding planes and in others (e.g. Section E-4) are about parallel to the axial plane of this large fold.

Near the hinge of the fold (Domain 2) two tension joints appear - one N-S, and another N.W. - S.E.. Associated with these tension joints is a shear joint that in some places parallels the axial plane and in others the N.W. trending schistosity in this domain (Figure 11).

On the east side of Domain 3, three well developed joint sets exist;.two are tensional and are often mineral-filled while the third, striking about east-west, is a shear joint approximately parallel to the bedding. It was noted that the joints near the hinge of the large folds tend to show a more preferred orientation (e.g. in east part of Domain 3) than those at a distance from it (e.g. in west part of Domain 3).

In the southern part of the area (Domains 4,5,6) the joints cannot be related to the structure because of the small amount of data, and also the more complex structure. The apparent increase in the number of joint sets present and the random orientation of these sets from section to section is probably related to the more complicated structural deformation of this area in which joints formed during later periods of deformation are superimposed on those formed during a previous deformation.

CHAPTER 5

DISCUSSION AND SUMMARY

The steeply plunging macro-folds within the area suggest that more than one phase of deformation has occurred. It seems reasonable to suggest, although no supporting evidence was found, that the area experienced at least one period of deformation prior to F_1. This deformation resulted in the eastward tilting of the volcanic pile and its associated ultrabasic sills. Such a deformation may have resulted in part from a tectonic stress related to a nearby batholith and in part from the downward sinking of the centre of the volcanic pile contemporaneous with the volcanism. The volcanic pile may have continued to tilt through the early part of the F_1 period of deformation.

Subsequent to the tilting the rocks in the area experienced from one to three periods of deformation (folding) as indicated by the various macroscopic and mesoscopic structures present. These three periods of deformation appear to have been characterized by three different directions of maximum compression (σ_1) and have been summarized in Table 3. All three phases of deformation are not found over the entire map area but instead the number of phases increases to the south and south-west.

Table 3.

Summary of Tectonic Events

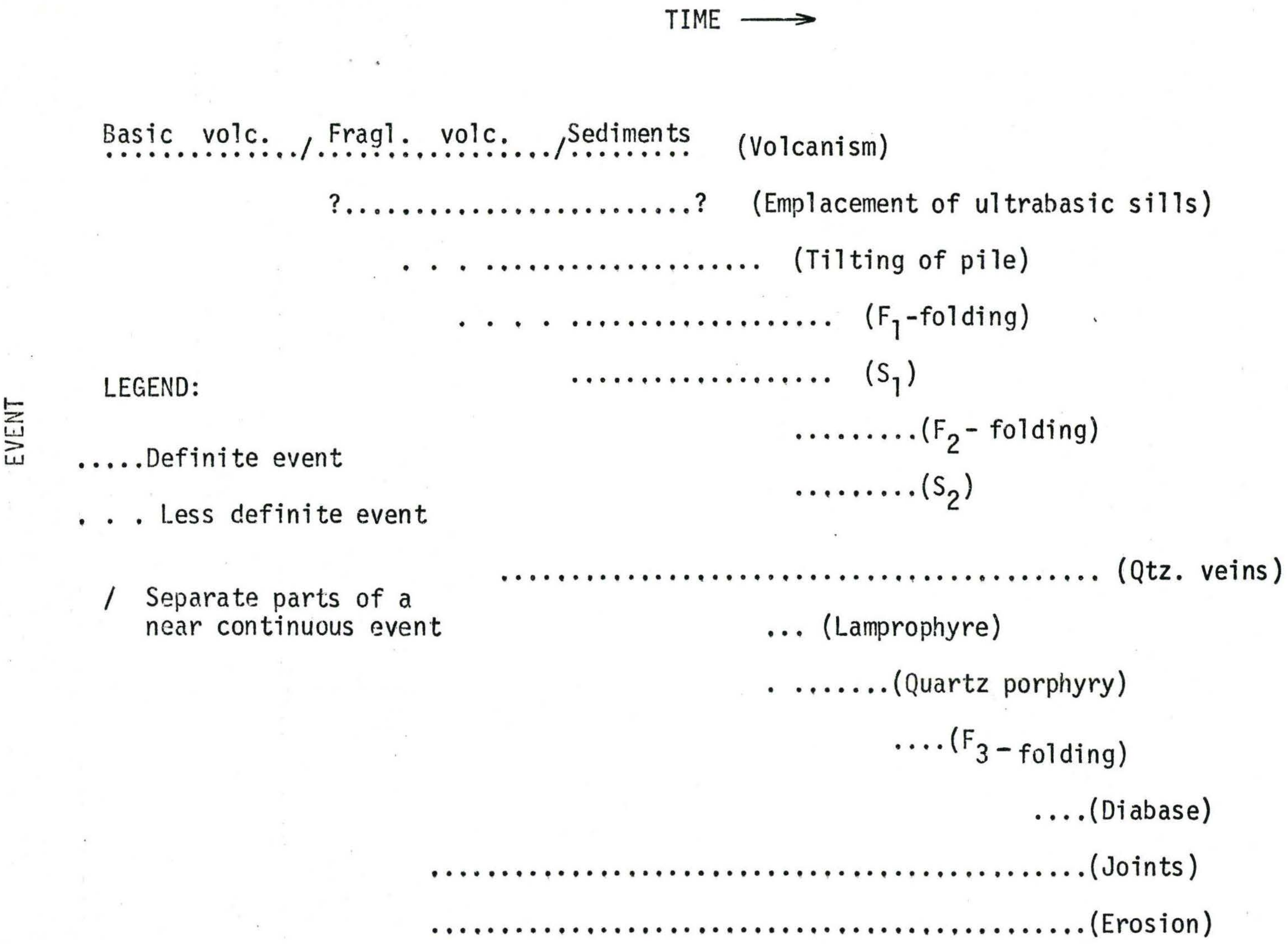
Phase of deformation	Area affected	Orientation of principal stress axis	Deformation shown in rocks
Pre-F_1	entire area?	σ_1 approximately E-W	- possible tilting of volcanic strata to east.
F_1	entire area (Domains 1-6)	$\sigma_1 \simeq$ N-S $\sigma_2 \simeq$ E-W $\sigma_3 \simeq$ Vert.	- large E-W trending macro-folds. - development of S_1. - macro-faults. - buckled and boudinaged quartz veins.
F_2	Domains 5&6	$\sigma_1 \simeq$ N.E.-S.W.	- development of cross-folds. - development of S_2 oblique to S_1.
F_3	Domain 6 (esp. Section C-10)	$\sigma_1 = 332;4S$ $\sigma_2 = 340;86N$ $\sigma_3 = 062;$Horiz.	- kink bands. - crenulation schistosity

Summary

1. Geological evolution of the Kakagi Lake area as interpretated by the author is shown in Figure 30.
2. The volcanic rocks of the area in ascending age consist of: 1. basic volcanics, 2. fragmental acid volcanics and 3. sediments.
3. The contact between the different volcanic units is both gradational and interfingering.
4. Emplacement of the ultrabasic sills within the volcanic pile has occurred nearly contemporaneous with the volcanism and prior to any major deformation as recognized by Ridler (1966).
5. The grade of metamorphism in the area is low (greenschist). An increase in metamorphism associated with an increase in deformation was noted from north to south. Two zones of metamorphism - a chlorite zone and a biotite zone were outlined. The biotite zone is thought to be the northern edge of a contact aureole about the granitic batholith to the south.
6. The general structure of the study area is revealed by large E.W. trending macro-folds developed during the F_1 phase of deformation. Folding appears to be nearly cylindrical with axes plunging steeply to the east. Superimposed on this E.W. structure in the south-west part of the map area are cross folds of a much smaller scale with axial traces roughly N.W.-S.E.

FIGURE 30.

Evolution of the Kakagi Lake Area.


TIME
EVENT
Basic volc.
Fragl. volc.
Sediments
(Volcanism)
?
?
(Emplacement of ultrabasic sills)
(Tilting of pile)
(F1-folding)
(S1)
(F2- folding)
(S2)
(Qtz. veins)
(Lamprophyre)
(Quartz porphyry)
(F3 - folding)
(Diabase)
(Joints)
(Erosion)
LEGEND:
.....Definite event
. . . Less definite event
/ Separate parts of a near continuous event

7. The minor intrusives (black lamprophyre and quartz porphyry) appear to be post F_1 deformation and prior F_2 deformation as indicated by their folding in the south of the area.
8. Few mesofolds were found in the study area. This factor is attributed to the absence of any S_0- surfaces in much of the rock. The few mesofolds observed were in the southern part of the area and have resulted from deformation (folding) of the quartz veins and minor intrusives.
9. Meso-faulting is prominent over much of the area. The scale and distribution density of this type of dislocation appears to depend on both the location within the macro-structures and the lithology. The fine-grained sediments display numerous small-scale dislocations whereas the thick fragmental volcanics and the intrusive rocks display fewer dislocations of a larger scale.
10. Tension joints - both those void of any filling and those filled in part or whole, appear to have resulted from the F_1 deformation (folding).
11. Numerous quartz veins are found in the fragmental acid to intermediate volcanic rocks.In most places these veins represent infillings of tension joints by silica migrating from adjacent rock.

12. A number of deformed quartz veins occur throughout the area. The orientation of the ptygmatically folded veins is about normal to the extended veins and the penetrative schistosity where such were found.
13. Associated with the folding (both F_1 and F_2) is the development of a penetrative schistosity over much of the study area. The early formed schistosity (S_1) is most prominent in the northern part of the area. Near the axial plane of Cedartree Lake Syncline this cleavage represents axial plane schistosity while on the limbs of this fold it lies parallel or sub-parallel to S_0. The later formed schistosity (S_2) cuts obliquely through S_1 in the southern part of the area. It appears to parallel the axial plane of F_2 folds and is the cleavage upon which kink bands have later developed.
14. Kink bands, developed in the S.W. corner of the area, appear to have formed entirely by slip on the laminae with the slip confined to the rock between the kink zone boundaries.
15. The sediments in the area reveal much slumping and penecontemporaneous structure. Slumping is of both a coherent and incoherent nature.
16. Shortening close to 70% is nearly horizontal and about perpendicular to the main schistosity S_1.

17. Deformation within the area as a whole, at any one location and during a single phase of deformation, reveals one principal extension and one principal contraction.

18. The carbonate unit situated between the volcanic rocks stratigraphically below and the intrusive above has experienced much shearing during the folding as a result of flexural-slip folding during the F_1 phase of deformation. This has resulted in the unit displaying numerous quartz veins.

19. Many small shear zones cut through the area and are believed to be genetically related to the macro-folding.

20. Jointing is particularly dependent on lithology with the fine-grained sediments displaying much systematic jointing. Both shear joints and tension joints are found in the study area; the attitude of shear joints being more consistent than that of the tension joints. In the fragmental acid to intermediate volcanic rocks, the higher density of fragments and the larger the fragments, the better developed the joints and the less well developed the schistosity.

21. Three phases of deformation have been recognized in the area.

(a) F_1 deformation:

Associated with the F_1 phase of deformation are features indicative of both ductile and brittle deformation. The large macrofolds appear to have formed at an early stage in a ductile fashion with subsequent brittle deformation having occurred as the folding became more tight. Different lithologic units changed from a ductile to a more brittle regime at different points in the folding. The ultrabasic sills appear to have flowed by intragranular mechanisms while the more brittle sediments behaved as a macroscopically ductile material and was able to flow cataclastically. Meso-fracturing and faulting have operated as a mechanism of cataclastic flow. The acid to intermediate fragmental volcanics seem to have been ductile at least during much of the shortening as indicated by ptygmatically folded and boudinaged quartz veins. Flexural slip has accounted for much of the folding in the north. The carbonate unit experienced much shearing during the development of Cedartree Lake Syncline. Shearing along S_0- surfaces especially near the contacts of lithologic units is common. With continued shortening, the rocks on the whole started to behave in a brittle fashion resulting in fractures (joints) and faulting both on a macroscopic and mesoscopic scale. The two large dislocations with sinistral movement that cut the north-west limb of the northern syncline in Kakagi Lake and the inferred one of dextral movement that cuts the sill south of the axial trace appear to be conjugate shear faults caused by continued shortening in a N.S. direction. A somewhat similar situation has occurred on the south limb of Kakagi Lake Anticline where numerous dislocations with a dextral sense of movement are suggested by apparent displacement of various lithologic units. These probably have formed at fairly low angles to the direction of maximum compression and have aided in the increased shortening noted in the south.

(b) F_2 deformation:

The second phase of deformation affected only the southern part of the area (Domains 4,5,6) and was most intense in the south-west (Domain 6). Here quartz veins and black lamprophyre dykes have been folded on a mesoscopic scale and a penetrative cleavage (S_2) developed oblique to S_1. The deformation is only weakly evident on a macroscopic scale. Much of the deformation was of a ductile nature in the early stages but appears to have become more brittle with time as indicated by meso-faulting of the limbs of meso-folds.

(c) F_3 deformation:

Deformation (F_3) of the late cleavage (S_2) has occurred locally. In one large exposure on the south-west shore of Kakagi Lake (Domain 6) it takes the form of kink band development and on one of the islands in the same domain a crenulation cleavage has resulted.

22. In going from F_1 to F_3 the area over which each phase of deformation is found decreases in size, the scale of deformation decreases, and the deformation appears to become more brittle with time.

Suggestions for Further Work

The author has only considered the structural geology in any detail thus leaving many important and interesting aspects of the geological evolution of the area. Suggestions for further work in the area include:

1. a detailed study of the sediments to ascertain their origin, nature of deposition and prelithification and slump structures which might lead to a determination of the paleoslope and thus the direction the volcanic vent was situated.
2. a study of the granitic body and surrounding volcanic rocks adjacent to the north-east side of the map area to see if this body (1) represents a volcanic plug and (2) has been folded as suggested in Figure 2.
3. a radiometric study might be done on the above mentioned intrusive and the age compared with that obtained for the granitic intrusive (batholith) to the south of the area, and
4. a metamorphic and structural study continuing south and west of the study area might be of great significance in relating genetically the structure described here to episodes of movement in the batholith. The author would liked to have examined this area but time did not allow it.

REFERENCES CITED

Anderson, T.B. (1969): The geometry of a natural orthorhombic system of kink bands, in A.J. Baer and D.K. Norris, eds., Proceedings, Conference on Research in Tectonics. Geol. Surv. Can. Paper 68-52, pp. 200-220.

Biot, M.A., et al. (1961): Experimental verification of the folding of stratified viscoelastic media. Bull. Geol. Soc. Amer., v. 72, pp. 1621-1630.

Burwash, E.M. (1933): Geology of the Kakagi Lake area. Ont. Dept. of Mines, v. 42, pt. 4, pp. 41-92.

Clifford, P.M. (1969): Kink band development in Lake St. Joseph area, northwestern Ontario, in A.J. Baer and D.K. Norris, eds., Proceedings, Conference on Research in Tectonics. Geol. Surv. Can. Paper 68-52, pp. 229-241.

de Sitter, L.U. (1958): Boudins and parasitic folds in relation to cleavage and folding. Geol. Mijnbouw. v. 20, pp. 277-286.

Donath, F.A. (1963): Fundamental problems in dynamic structural geology, in T.W. Donnelly, ed., Earth Sciences - Problems and Progress in Current Research. Univ. Chicago Press, pp. 83-103.

Fahrig, W.F., and Wanless, R.K. (1963): Age and significance of diabase dyke swarms of the Canadian Shield. Nature,. v. 200, pp. 934-937.

Fairbridge, R.W. (1946): Submarine slumping and location of oil bodies. Bull. A. Assoc. Petrol. Geol., v. 30, pp. 84-92.

Fleuty, M.J. (1964): The description of folds. Proc. Geol. Assoc., v. 75, pt. 4, pp. 461-492.

Goodwin, A.M. (1965): Preliminary report on volcanism and mineralization in the Lake of the Woods - Manitou Lake - Wabigoon region of Northwestern Ontario. Ont. Dept. of Mines Prelim. Rept. 1965-2, 63p.

Harris, S.F., et al. (1960): Relation of deformational fractures in sedimentary rocks to regional and local structure. Bull. Am. Assoc. Petrol. Geol., v. 44, pp. 1853-1873.

Helwig, J. (1970): Slump folds and early structures, northeastern Newfoundland appalachians. J. Geol., v.78, pp. 172-187.

Johnson, M.R.W. (1956): Conjugate fold systems in the Moine Thrust Zone in Lochcarron and Coulin Forest areas of Wester Ross. Geol. Mag., v. 93, pp. 345-350.

Lawson, A.C. (1885): Report on the geology of the Lake of the Woods region with special reference to the Keewatin (Huronian) Belt of the Archean rocks. Geol. Surv. Can. Annual Rept. (new series) v. 1, report CC.

Maxwell, J.C. (1962): Origin of slaty and fracture cleavage in the Delaware Water Gap area, New Jersey and Pennsylvania, in Petrologic Studies (Buddington volume). New York. Geol. Soc. Amer., pp. 281-311.

Mitcham, T.W. (1963): Fractures, joints, faults, and fissures. Econ. Geol., v. 58, no. 7, pp. 1157-1158.

Norris, D.K. (1969): Discussion (on kink bands) in A.J. Baer and D.K. Norris, eds., Proceedings, Conference on Research in Tectonics. Geol. Surv. Can. Paper 68-52, pp. 220-225.

Paterson, M.S., and Weiss, L.E. (1966): Experimental deformation and folding in phyllites. Bull. Geol. Soc. Amer., v. 77, pp. 343-374.

Price, N.J. (1966): Fault and joint development in brittle and semi-brittle rock. Pergamon, Oxford, 176p.

Ramberg, H. (1963): Fluid dynamics of viscous buckling applicable to folding of layered rocks. Bull. Am. Assoc. Petrol. Geol., v. 47, pp. 484-515.

........, and Ghosh, S.K. (1968): Deformation structures in the Hovin Group schists in the Hommelvik - Hell region (Norway). Tectonophysics, v. 6, no. 4, pp. 311-330.

Ramsay, J.G. (1962): The geometry and mechanics of formation of similar type folds. J. Geol., v. 70, pp.309-327.

........, (1967):—Folding and fracturing of rocks. New York, McGraw-Hill, 568.

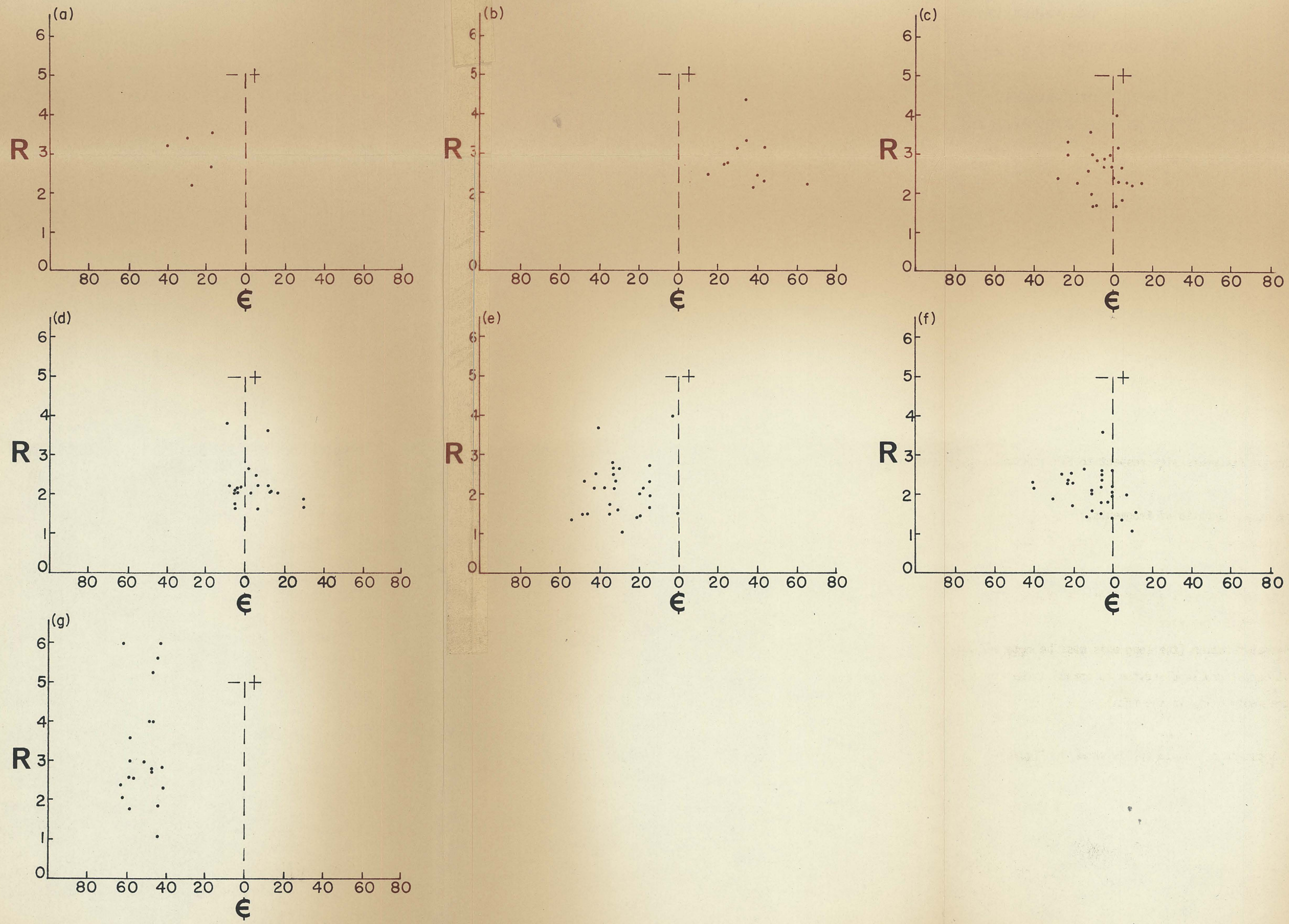
Ridler, R.H. (1966): Petrographic study of a Crow Lake ultrabasic sill, Keewatin volcanic belt, Northwestern Ontario. unpubl. M.A.Sc. thesis, Univ. of Toronto, 104p.

Rodgers, J. (1952): Use of equal-area or other projections in the statistical treatment of joints. Bull. Geol. Soc. Amer., v. 63, pt. 1, pp. 427-430.

Smith, D.R. (1971): Physical and chemical studies of the salic volcanic breccias, Kakagi Lake, Ontario. unpubl. M. Sc. thesis, McMaster Univ. 79p.

Turner, F.J., and Verhoogen, J. (1960): Igneous and metamorphic petrology, 2nd ed., New York, McGraw-Hill, 694p.

........., and Weiss, L.E. (1963): Structural analysis of metamorphic tectonites. New York, McGraw-Hill, 545p.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