

STRUCTURAL ANALYSIS OF GRENVILLE ROCKS
NEAR BANCROFT, ONTARIO, CANADA

A thesis
presented to the
Department of Geology
University of Ottawa

In partial fulfillment
of the requirements for the Degree
DOCTOR OF PHILOSOPHY
in Geology

by



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January, 1972

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ABSTRACT

Southeast of Bancroft, Ontario, Precambrian metasedimentary and metavolcanic rocks of the amphibolite facies lie tightly folded between a gneiss complex and granitic plutons underlying the Haliburton-Hastings highlands to the north, and the Wesslemkoon granitic batholith to the southeast. The metamorphic rocks were subjected to at least three discontinuous phases of deformation, with resultant folds F_1 , F_2 the dominant, and F_3 .

F_1 folds, which are only exposed in metasedimentary rocks at a few localities, are small-scale and isoclinal. The preservation within later garnet porphyroblasts of a fine-grained schistosity S_1 that is axial-planar to micro F_1 folds in relict layering suggests that F_1 formed under conditions of relatively low-grade metamorphism. The orientation of S_1 before overprinting by later folding and crystal growth is not clear, but in some areas the geometry of low-plunging F_2 folds suggests that it was subhorizontal.

F_2 folds affect both bedding and S_1 on all scales, vary in tightness, and are upright to overturned northwest. An axial-surface foliation S_2 , either a schistosity or gneissic layering, is regionally penetrative and extends into synkinematic or earlier plutons. Some sillimanite crystals are bent by F_2 folds, but most lie within S_2 , and a few are cross-cutting; thus high-grade metamorphism preceded, accompanied, and outlasted the folding.

Hinge lines of minor F_2 folds and axial lineations L_2 are curved and irregular in detail, but commonly plunge at low angles northeast or southward along the axial traces of large-scale F_2 folds. As demonstrated by rake isogons and moving averages, towards the northern complex of gneiss and plutons L_2 progressively changes direction and steepens until, as in the gneiss, it plunges southeast down the dip of S_2 . To the southeast near the large Weslemkoon pluton, F_2 axes and L_2 also plunge steeply down-dip.

Flattening of F_2 folds, as calculated from systematic changes in thickness of calc-silicate layers in fold profile, varies locally, but in general increases with the axial steepening towards the plutons. Thus it appears that the presence or emplacement of relatively rigid plutons during folding imposed an element of constrictional strain that influenced the orientation and shape of adjacent F_2 folds.

F_3 folds, only locally present, are open, recumbent structures bending S_1 and S_2 foliations; mineral grains disrupted in the axial regions indicate that folding was virtually post-metamorphic.

Narrow mylonite zones occur within the gneiss complex near its southern border. Structures within the mylonites are of similar form and orientation to those associated with F_1 , F_2 , and F_3 . The similarity of the small-scale structures and related fabrics within the gneisses, mylonites, and metasedimentary-metavolcanic group of rocks indicates that the mylonite zones were initiated at an early stage in a common deformational history.

K-Ar radiometric ages obtained from minerals forming S_2 or L_2 fabrics, nepheline from the gneiss complex, biotite and hornblende in the southern metasediments, and biotite from a foliated part of the Weslemkoon pluton, are compatible with a Grenvillian age (about 1000 m.y.) of deformation and metamorphism throughout the area.

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INTRODUCTION

Location and Rock Outcrop

The study area (Fig. 1) is within the Bancroft and Coe Hill map sheets (31 F/4 and 31 C/4) of the Canadian National Topographic 1 : 50,000 Series; it includes most of Dunganon and Mayo townships, and part of Faraday Township in Hastings County, southeastern Ontario. The area is easily accessible by Highway 62, which runs southward from Bancroft to Madoc, Highway 500, which runs eastward, and by Township and logging roads.

Glacial deposits and thick vegetation cover much of the bedrock. Bordering the Haliburton-Hastings Highlands in the north (Fig. 1), the local relief is up to 700 feet and small outcrops of fresh rock are common. In the south relief is less, and apart from ^{along} Highway 62, outcrops are scarce. Many temporary rock exposures were available during the summer of 1968, when Highway 62 was broadened and new county roads were laid.

Geologic Setting and Previous Work

The study area lies in the Grenville Province of the Canadian Shield (Fig. 1). Except near Madoc to the south, the province is largely a region of medium-to high-grade metamorphic rocks of sedimentary, volcanic and plutonic origin. Recognisable metasedimentary and metavolcanic rocks are prominent in southeastern Ontario (Baer, 1970,

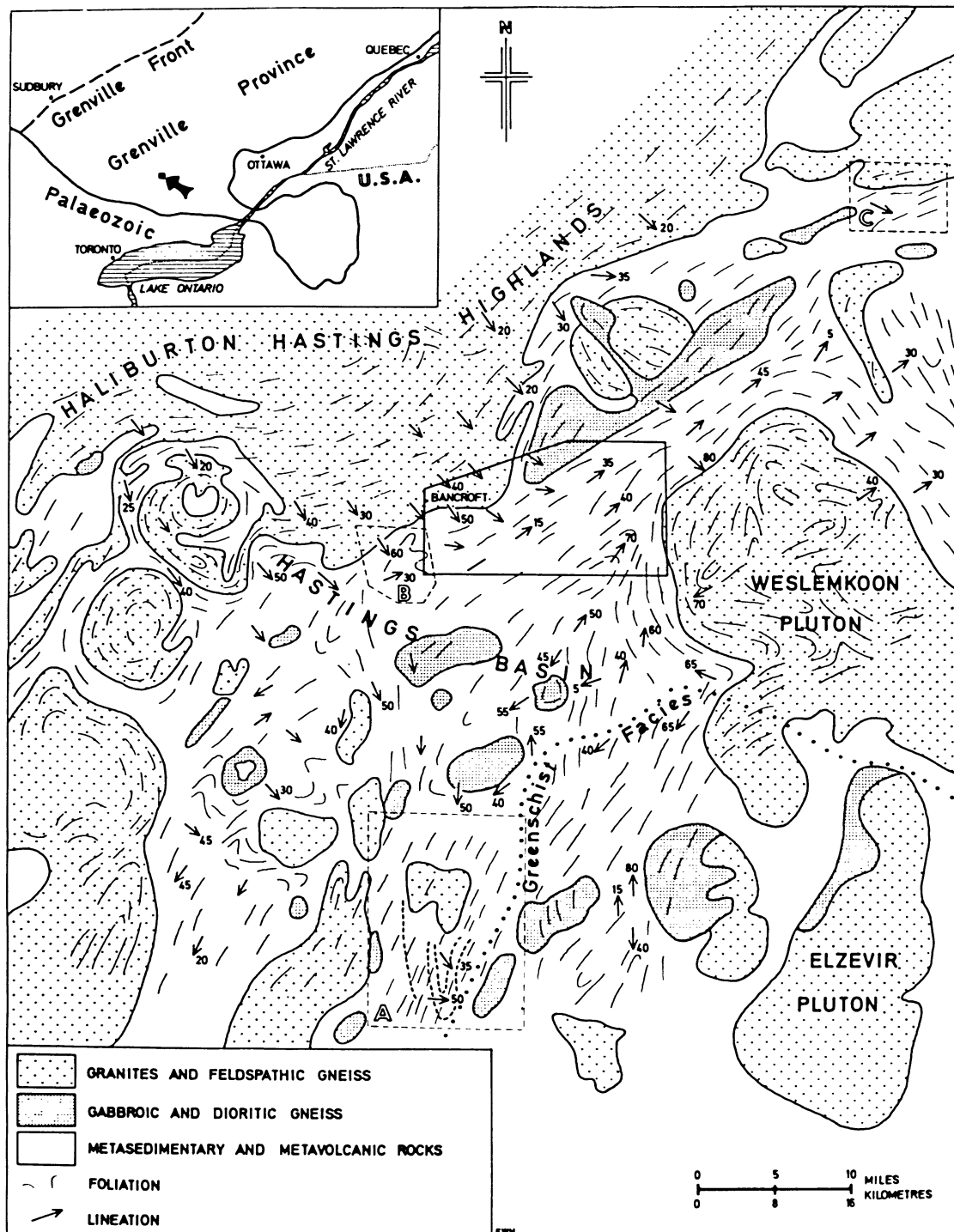


Fig.1 : Generalized geological-structural map of Bancroft region. Modified from Hewitt, 1957, 1962; Evans, 1964; Lumbers 1964, 1967b, 1968, 1969; Best, 1966 (Area B); Carmichael, 1967 (Area A); and Appleyard, 1967 (Area C). Study area outlined east of Bancroft. Layers oblique to foliation shown in Area A by dotted line.

Fig. 1; also see Map 2197 and 2198 of Ontario Department of Mines), including the Bancroft area south of the Haliburton Hastings highlands (Fig. 1) where they are known as the Grenville Group. Elsewhere within the province similar rocks occur as isolated lenses in a widespread terrain of plutonic rocks and gneisses whose original characteristics are obliterated as a result of reconstitution during medium-to high-grade regional metamorphism.

Rb-Sr isotopic studies on Grenville rocks in southeastern Ontario indicate that there was sedimentation 1300 million years ago (Lumbers, 1967b), and regional metamorphism and plutonism between 1300 and 1000 million years ago (Krogh and Hurley, 1968). A regional thermal event that affected the Grenville Province about 1000 million years ago is apparent from numerous K-Ar age determinations. A few older ages indicate Aphebian (Lower Proterozoic) metamorphism and plutonism for some rocks in the northwestern part of the Grenville Province (see Geological Survey of Canada's Isotopic Age Map of Canada - map 1256A; also Krogh and Davis, 1969; Krogh, 1971; and Dalziel and others, 1969).

Immediately south of the Grenville Front (Fig. 1), typical Grenville sediments are lacking, and metamorphosed equivalents of Superior, Churchill and Southern province rocks have been recognized (Quirke and Collins, 1930; Johnston, 1954; Déland, 1956; Knowles and Gastil, 1959; Clarke, 1960; Grant, et al., 1962; Wynne-Edwards et al., 1966; and Wynne-Edwards, 1969). Following this evidence, many geologists

consider the major part of the Grenville Province, except the Grenville Group, as 'remobilized basement' (Osborne and Morin, 1962; Walton and De Waard, 1963; Stockwell, 1964; Wynne-Edwards, 1964, 1969; and Wynne-Edwards and others, 1966). Although no radiometric evidence has so far proven older material to exist more than 15 km southeast of the Grenville Front, except at two localities (see Krogh and Davis, 1969), many rocks in this region extending into the Haliburton-Hastings highlands are shown in the Geological Survey of Canada Map 1250A as remobilized older rocks.

Geological map 770 (Adams and Barlow, 1910) published by the Geological Survey of Canada, and maps 1955-8, 1957b (Hewitt) and Fig. 7 of Preliminary Report 1964-4 (Lumbers), published by the Ontario Department of Mines, cover much of the study area, but no previous detailed structural work has been carried out. In a regional study, Hewitt (1962) noted that the Haliburton-Hastings highland gneisses characteristically display steep southeast 'A' lineations whereas the Hastings Basin metasedimentary and metavolcanic rocks to the south more commonly exhibit shallow northeast axial lineations; the rocks with contrasting structures were considered to be in fault contact, represented by mylonite zones.

Best (1966) recognized in a small area to the southwest of the study area (area B, Fig. 1) two generations of folds and axial lineations that trend northeast in the south and southeast in the north. He suggested that with increased compression, shallow northeast lineations and fold axes

rotated to become steep southeast lineations; but he offered little supporting evidence.

For a wide region to the southeast of Bancroft, Lumbers (1967b) outlined a long period of deformational, metamorphic and plutonic activity. However, detailed information about the different phases of deformation is not given.

Baer (1970) divided the Grenville Province of southeastern Ontario into two regions of contrasting lithology and structural trends. The 'line' separating the two regions passes through the Bancroft area northeastward along the highland-basin boundary. He suggested (1970, p. 120), that close to the line, the southeasterly trends in the gneisses to the north are earlier than the northeasterly trends in the metasedimentary rocks to the south.

In the studies mentioned, the time relations of structures within the highlands and the basin are not clearly defined.

Purpose of Present Work

The objectives of the present work are:

1. To determine the geometry of structures in the northern gneisses, the southern metasediments, and the mylonite zones.
2. To investigate the time relations of the structures in the various rock types; for example to find whether or not there is evidence that the northern gneisses represent remobilized basement.
3. To investigate the time relations between deformation,

metamorphism and plutonism in the study area.

4. To correlate the tectonic history of the area with that in adjacent areas in order to further define the regional tectonic history of the Grenville rocks.

Acknowledgments

The writer is indebted to Dr. W.K. Fyson for his supervision and constructive criticism of the work. Discussions with Dr. R. Kretz, Dr. A. Baer, Dr. E.C. Appleyard and my fellow graduate students, particularly Mr. A. Carrara and Mr. D.C. Kamineni, have been most valuable. Dr. R. Kretz and Dr. D.D. Hogarth gave valuable suggestions in improving the manuscript. The work was supported by a research grant by the National Research Council of Canada to Dr. W.K. Fyson. Thanks are also due to the Geology Department, University of Ottawa for providing necessary facilities for the research.

Dr. M. Shafiquallah, while at Carleton University, determined the radiometric ages of the samples. Mr. D.C. Kamineni helped in separating nepheline used in radiometric age dating. Dr. B. Tilak helped the writer in the computer work.

Mr. B. Munroe and Mr. M. Armstrong assisted the writer in the field during summer, 1968. The writer is also thankful to many residents of Bancroft who helped him in some way during the summers of 1967 to 1969. Photographs of samples and thin sections were taken by Mr. E.W. Hearn, who also drafted figs. 1 and 13. Mrs. S. Groulx typed the thesis. Finally, the writer is deeply indebted to his parents, who made it possible for the writer to come to the University of Ottawa.

ROCK UNITS

Introduction

The petrography, outcrop pattern and metamorphic grade of the rock units shown in Figs. 2 and 3 are described. The rock units in Fig. 2 are essentially similar to those of Map 1955-8 (Ontario Department of Mines) described by James (1957, p. 24). However, boundaries and designation of some rock units are modified in the light of the present study. Also, the metamorphic isograds in the area are modified from those shown by James (1957) (Fig. 2).

Primary structures which aid in determining the stratigraphic tops are nearly all obliterated in the study area, and the rock units in Fig. 2 are not necessarily in stratigraphic order. Within the southern metasediments marble (unit 3), the most widely distributed rock unit, is closely associated and locally interfingers with rock units 2, 4, 5, 6 and 7. The northern gneisses are comprised essentially of rock units 8, 10, and 11, and metasediments are rare in this area.

Petrography

Southern Metasedimentary and Metavolcanic Rocks

The mafic metavolcanic rocks (unit 1) are commonly medium grained and consist of dark green amphibolites with hornblende, plagioclase, biotite, and occasional garnet. The rocks possess a well-developed foliation in which lie the long dimensions of

hornblende and biotite crystals. Primary textures and structures are virtually obliterated as the rock has undergone strong deformation and recrystallization. However, deformed pillow-structures are present to the south of the area (Lumbers, 1967a, p. 36).

Rock unit 2 previously known as 'feather amphibolite' (Adams and Barlow, 1910), is an interbanded amphibolite and fine-grained grey to brown marble. The amphibolite bands consist of porphyroblastic laths of hornblende aggregated in feather-like shape in a granoblastic plagioclase-quartz matrix. The width of the banding varies from a few millimeters to more than ten meters. Some spectacular boudinage structures are seen in this unit.

Marble, rock unit 3, is the most widespread rock type within the southern metasediments. The grain size varies from fine to medium in the south to medium to coarse in the north. Generally the rock is grey, dark grey, bluish-grey or white, depending on the impurities and grain size. For example, though they contain the same amount of graphite the fine-grained graphite marbles in the south are various shades of grey whereas the coarse-grained rock in the north is white.

Both calcitic and dolomitic marbles occur, but the former predominate. Dolomitic marble, can be seen in some outcrops along Highway 62 south of L'Amable lake. In the silicated or impure varieties, the following minerals are common: diopside, tremolite, phlogopite, graphite, scapolite, quartz, pyrite and apatite.

Locally, metamorphic pyroxenite occurs within and grades into marble, as for example 4 km southeast of Bancroft. The pyroxenite is a medium to coarse-grained, dark green rock consisting primarily of diopside or augite, often accompanied by phlogopite, scapolite, calcite and tremolite.

At a single locality immediately to the southeast of Mayo lake, metaconglomerate outcrops within marble (Fig. 2). Calc-silicate, granitic and quartzitic pebbles lie in bands in a matrix of biotite, muscovite and quartz.

Unit 4 is essentially a fine-to medium-grained, grey to brownish-grey quartz-feldspar-biotite rock, which occasionally contains porphyroblasts of biotite and garnet. Plagioclase constitutes most of the feldspar, but potassic feldspar occurs in some zones. Where interbanded with carbonate metasediments, the rock commonly contains a few coarse-grained porphyroblasts of hornblende.

Locally in some outcrops the rock consists mainly of quartz with a little feldspar; gneissosity is poor and the rock is essentially a quartzite.

Rock unit 5, which mainly consists of quartz-biotite-feldspar, is similar in mineral content to the semi-pelitic schist unit (unit 7), but differs from it in commonly possessing 5-10% pyrite and in being more pelitic. Due to weathering, pyrite is oxidized, thereby giving the rock a rusty appearance. Narrow veinlets and lens-like bodies of massive pyrite and rarely, of massive pyrrhotite are common in many outcrops. Graphite, muscovite and diopside are present as minor consti-

tuments. Narrow interbands of marble occur in the rusty schist.

Rock unit 6 consists of fine-grained grey-white rhyodacite with porphyroblasts of altered albite. Primary flow structures are lost and there is a weak foliation defined by flattened quartz and feldspar grains and also by aligned biotite. Plagioclase is altered to scapolite.

The semi-pelitic schist, unit 7, contains mainly quartz, biotite and muscovite; garnet porphyroblasts are common. The highest grade of metamorphism in the area is indicated by the appearance in this unit of sillimanite which occurs as sheaves and needles within the foliation. A little plagioclase is commonly present. Accessory minerals are magnetite, zircon and apatite.

Interlayered within the pelitic schist is a band of coarse-grained scapolite-amphibole rock, which is well exposed immediately to the west and southwest of Upper Hermon. The rock is banded with alternating scapolite-rich and amphibole-rich layers.

Northern Gneisses

The Mallard Lake metagabbro, unit 8, is a metamorphosed basic intrusive occupying about 28 square miles in the north-eastern part of the area. The term 'metagabbro' is used to describe coarse-grained hornblende-plagioclase gabbroic gneiss in which clotted aggregates of hornblende crystals occur in a medium-grained granoblastic matrix of plagioclase.

The metagabbro grades from light-coloured scapolite-

plagioclase rock, with clots of black or dark-green hornblende crystals up to one centimeter in length that constitute 20-40% of the rock, to a coarse-grained hornblendite of interlocking crystals.

Other rock types present in this large body include fresh looking, fine-to medium-grained quartz diorite, massive gabbro, and hornblende schist. Remnants of olivine are present in the central part of the mass. A highly metamorphosed border facies on the west side of the metagabbro body is difficult to distinguish from the amphibolite of unit 1.

The nepheline gneisses of rock unit 9 at 2 km northeast of Bancroft consist essentially of nepheline, amphibole and plagioclase (usually oligoclase). Microcline and microcline microperthite are sometimes present. Biotite, hornblende and magnetite are important mafic constituents in some of the gneisses and may make up over 30% of the rock. Garnet, scapolite and corundum are characteristic accessories in some nepheline gneisses.

The nepheline rocks 9 km further northeast near Mallard Lake metagabbro are more gneissic and have higher percentages of mafic minerals than those near Bancroft.

Unit 10 consists of hybrid gneisses or migmatites of granitic composition, which contain many bands of amphibolite and granitized amphibolite.

Five plutons are composed of granitic rock (unit 11). Some are virtually structureless, others are well foliated. In the southeast corner of the area, about 10 square kilometers of

the Weslemkoon granitic pluton outcrops. It is the largest single intrusive body in this part of the Grenville province (Fig.1). A foliation is formed by lens-shaped aggregates of coarse-grained quartz and feldspar (microcline and usually oligoclase), and also by biotite crystals. The Bessemer granite 2 km to the northwest is a medium-grained, grey biotite-plagioclase-quartz rock with a pronounced gneissosity.

The Great Bend granite immediately to the west of the Mallard Lake metagabbro is also gneissic and has albite, quartz, and microcline as the major constituents. Large inclusions or bands of marble, amphibolite and nepheline gneiss sometimes occur in the pluton especially near its margin (Hewitt and James, 1956). Further northwest the Bronson granite is a poorly foliated pink leucocratic granite composed of microcline, albite and quartz.

The McArthurs Mills granitic pluton in the northeast corner of the map-area differs from the other granitic rocks. It is a coarse-grained, massive, pink, biotite-microcline-microperthite granite that varies to quartz syenite. The pluton does not exhibit any foliation.

Mylonites

East and west of Bancroft, discontinuously exposed mylonite zones, ranging in width from 3 to 60 meters, occur in the northern gneiss complex within 3 km of the boundary with the southern metasediments (Fig. 3). The term mylonite is used here following the definition given by Lapworth in 1885

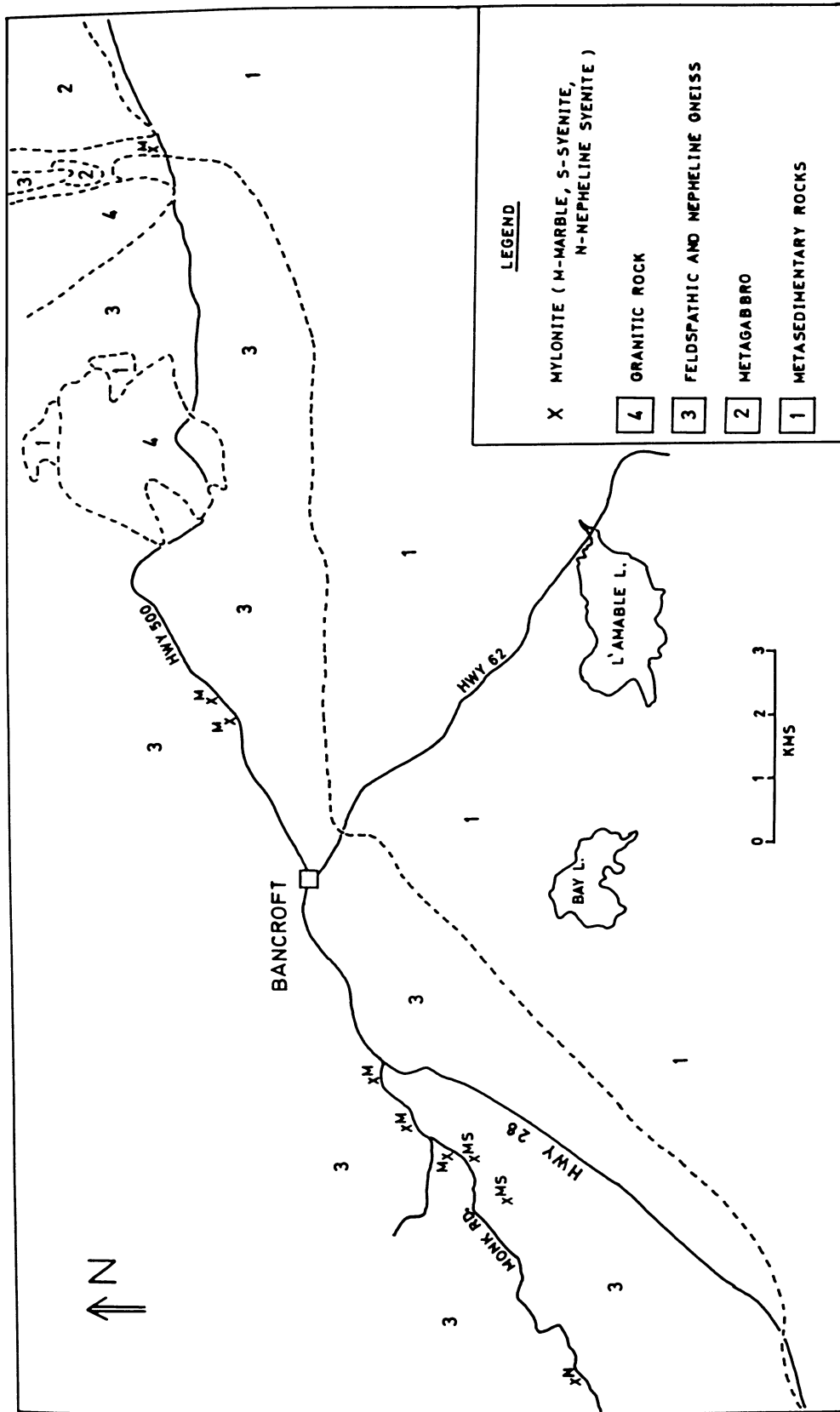


Fig. 3 : Distribution of Mylonites east and west of Bancroft.

(see review by Christie, 1960). The rock must be fine-grained, laminated, and the minerals show effects of cataclastic break down. But in the present area, some of the mylonitic rocks exhibit various stages of recrystallization, masking their cataclastic textures. To these rocks, the term "blastomylonite" (Sander, 1912) is used.

Marble mylonites occur east of Bancroft along Highway 500, and west of Bancroft along Monk Road (Fig. 3).

The marble mylonites are various shades of grey and fine grained. The laminated and foliated material is composed of flattened grains of calcite and dolomite and equidimensional grains of the same minerals in blastomylonites.

In this matrix porphyroclasts of different rock composition (which originated during the process of mylonitization by cataclasis) occur as "floating" bodies in the outcrop (Plate 1A). These porphyroclasts have been described as "foreign fragments" (Hewitt, 1957). Following the terminology used for such fragments in mylonites of the Moine Thrust zone (Christie, 1960; Johnson, 1961; and Barber, 1965), the term porphyroclast is preferred to emphasize the associational character of such fragments with mylonites.

The porphyroclasts range from submicroscopic to bodies several feet in diameter and consist of heterogeneous material similar to and probably originally part of the surrounding rocks outside the mylonite zone - amphibolites, quartz-feldspar-biotite schists, and syenitic gneisses with or without nepheline. Some of the porphyroclasts are sharply angular but

most of them are rounded or ovoid (Plate 1B) and lie in the mylonitic foliation.

Both crystalloblastic and typical cataclastic textures are common, but in blastomylonites the texture is granoblastic, in which case the grains do not show any post-crystallization strain.

Syenite mylonites occur mostly west of Bancroft along Monk Road (Fig. 3). The zones are discontinuous and vary in width from 3 to 40 meters. In some outcrops the syenite mylonites are interlayered with marble mylonites.

The syenite mylonites are dark pink, with matrix and porphyroclasts largely composed of pink potash feldspar and plagioclase; but some fine bands in the matrix are quartzose. The porphyroclasts have variable shape - sharp and angular, rounded, and ovoid .

Unlike marble mylonites, recrystallization of either the matrix or porphyroclasts is limited, and nowhere is the original cataclastic texture masked to form blastomylonites.

Near the western margin of Fig. 3 and 300 m north of Monk Road, grey nepheline syenite mylonites occur in a narrow zone up to 30 m wide, flanked on both sides by syenite gneiss. The nepheline syenite mylonites contain porphyroclasts of orthoclase, microcline, plagioclase and nepheline, and unlike the other two types, contain biotite; these are aligned and accentuate the mylonitic foliation. Porphyroclasts of nepheline are more often rounded or ovoid than those of feldspar, probably reflecting a relatively higher ductility for the nepheline.

Stratigraphic Relations

A reconstruction of the stratigraphic section for the metavolcanic-metasedimentary sequence is made difficult by the complex intertonguing between and gradation within original lithofacies, by recrystallization of the rocks into different mineral assemblages as a result of variation in the metamorphic grade, and by the complex folding which has affected the rocks.

Though primary structures which aid in determining the tops are nearly all obliterated in the study area, pillow structures occur in metavolcanic rocks to the south (Lumbers, 1967a, p. 36).

Hewitt and James (1956) mapped the southern part of the metavolcanic unit (unit 1) as gabbro, but this rock continues further south into Cashel township and into Tudor township where it becomes part of the Tudor metavolcanics (Lumbers, 1967a).

The stratigraphic sequence given by Lumbers (1967b, p. 18) (Fig. 4) for the Grenville metavolcanic and metasedimentary rocks of the Bancroft-Madoc region is tentatively accepted until more data are available. In the study area unit 1 is part of the Tudor metavolcanics. Contrary to the opinion expressed by Hewitt (1956) that these are in the upper part of the metavolcanic-metasedimentary sequence, they are assigned to the base (Lumbers, 1967b). That the Tudor metavolcanic rocks are basal is also supported from the fact that ^{in rhyolites} zircons give a radiometric (U-Pb) age of 1310 ± 15 million years,

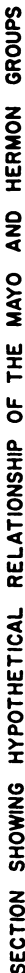


Fig. 4 : Stragigraphic Section showing hypothetical relationship of the Mayo and Hermon Groups, (Lumbers, 1967b) .

which is the oldest age obtained for any rock in the meta-volcanic-metasedimentary sequence (Lumbers, 1967b, p. 19). The base of the Tudor metavolcanics is nowhere found.

The layered Feather Amphibolite (unit 2) and the Marble (unit 3) are part of the Dungannon Formation of the Mayo Group (Hewitt and James, 1956). The position of the quartzitic rock (unit 4), the pyritic schist (unit 5) and the semi-pelitic schist (unit 7) in the stratigraphic sequence is not clear. Their lithological character indicates that they could belong to the Vansickle Formation of Lumbers. Similarly the albite-quartz rock (unit 6) is correlated with the Burnt Lake Formation.

Cross-cutting relations of the rock units indicate that the Mallard Lake metagabbro is the oldest unit among the northern intrusives (Hewitt and James, 1956). The Bronson and the Great Bend granitic plutons have intrusive relationships to the nepheline rocks (unit 9) and evidently are later. Similarly in the south the granitic plutons cross-cut the metasedimentary-metavolcanic rocks and thus are later.

The Bessemer and Weslemkoon plutons, foliated like parts of the Bronson and Great Bend plutons, are considered as syn-tectonic and earlier than the McArthurs Mills granite, which, lacking any foliation is late or post-tectonic.

Metamorphic Grade

Spatial variations in regional metamorphism are usually shown by isograds. In the study area, however, it is difficult

to locate isograd lines due to the discontinuous nature of and great compositional variation within the lithological units. Thus more detailed work indicates that some of the isograds (presumably drawn on the basis of the first appearance of an index mineral) delineated by James (1957) need revision.

James (1957) indicated green schist facies in the southeast and amphibolite facies in the northwest. The facies are separated by a line that runs northeast and passes immediately north of the Bessemer granite. Roughly parallel to this line and within the amphibolite facies zone, hornblende, garnet and sillimanite isograds are shown.

However, hornblende, garnet, and oligoclase occur in rocks of appropriate basic composition throughout the area. Therefore all rocks are considered to belong to the amphibolite facies, as shown on a regional map by Lumbers (1964, Fig. 2), and the garnet and hornblende isograds of James (1957, Fig. 2) are incorrect.

As recognized by James (1957), the highest grade within the amphibolite facies is indicated by the occurrence of sillimanite in the semi-pelitic schist (unit 7). An isograd, based on the first appearance of sillimanite and modified in position from that of James, trends northeast parallel to the edge of the gneiss complex, which lies 0.75 to 1.5 km to the north. The topographic relief is not prominent enough to infer the slope of the isograd surface.

Radiometric dating (potassium-argon method) of biotite

associating with prismatic sillimanite gave an age of 935 ± 50 million years (location in Fig. 2). As both the minerals are the product of the same reaction (see Appendix) it can be concluded that the highest grade of metamorphism in the area occurred at least 935 ± 50 million years ago (MacIntyre and York, 1967). Other age dates obtained (Fig. 2) will be referred to in conjunction with structures (p.103).

STRUCTURES

Introduction

In this chapter the structures in the northern gneisses and southern metasediments are considered. The structures in the mylonites are considered separately in a later chapter.

The northern gneisses and southern metasediments are affected by at least three phases of deformation, D_1 , D_2 , and D_3 , which produced various structures - folds F_1 , F_2 , and F_3 ; foliations S_1 , S_2 , and S_3 ; and lineations L_1 , L_2 , and L_3 respectively. The three generations of structures are distinguished from each other on the basis of their character, orientation, and relationship to one another at various scales of observation.

D_1 structures are widespread in the southern metasediments, but they are not clear in the northern gneisses. Except in the case of the McArthurs Mills granite, which does not possess a recognizable tectonic fabric, all the rock units exhibit well developed D_2 structures. Analysis of the areal variation in D_2 fold styles is one of the main aspects of the thesis. D_3 structures of lesser significance occur sporadically in both the northern gneisses and southern metasediments; they are readily identified in the field.

The main characteristics of the three generations of structures are summarized in Table 1, and the following is an amplification of this table and further interpretation.

Table 1: Character of D₁, D₂ and D₃ Structures

	D ₁	D ₂	D ₃
Geographic Distribution and Scale	Few small hinges in the southern metasediments but axial-surface foliation throughout; not clear in the northern gneisses	Dominant folds of all scales throughout the map-area	Mesoscopic folds, local but widely distributed throughout the map-area
Surface folded	S ₀ bedding (lithologic layering)	S ₀ and S ₁ stratiform foliation	S ₀ , S ₁ and S ₂ fracture cleavage/schistosity/gneissosity
Fold Profile	Isoclinal, interfolial, detached hinges	open to isoclinal and detached; flattened parallel	Gentle to open, parallel
Axial-surface Foliation	S ₁ stratiform foliation	S ₂ fracture cleavage, crenulation-cleavage (only in sillimanite schist) schistosity similar to subparallel S ₁ , gneissosity	S ₃ fracture cleavage, crenulation cleavage
Axial lineation	L ₁ axes of lens-like remnant fold hinges, colour streaks of S ₀ on S ₁ ; long axes of elongated pebbles	L ₂ intersection of S ₁ , and S ₂ , quartz rods, mica edge, amphibole/sillimanite/nepheline aggregate alignment	L ₃ intersection of S ₃ and S ₁ /S ₂ , crenulation axes
Orientation	Variable axial trend, not clear; F ₁ folds initially recumbent	S ₂ dip to SE steepens towards southeast; variable axial plunge to NE and SW; reclined axes in the northwest and southeast	S ₃ subhorizontal; F ₃ axes trend northeast

D₁ Structures

Distribution and Scale

In the southern metasedimentary rocks small-scale detached fold hinges (F₁) whose development is definitively attributable to the D₁ phase of deformation are found in only a few localities, where they occur in layers folded by open F₂ folds; for example, 6 km southwest of Mayo Lake, immediately to the southwest of Jamieson Lake and 2 km southwest of Detlor (Fig. 5). Elsewhere similar structures could be the detached hinges of F₂ folds. No large-scale F₁ folds are recognized.

Similarly, except in a few localities L₁ lineation is not clearly distinguished from L₂ lineation. The axial-surface foliation S₁ is widespread, but largely overprinted by a later parallel foliation S₂.

D₁ structures in the northern gneisses are obscure because of post-D₁ intrusion, recrystallization, and replacement of the rocks; also because of the high strain these rocks experienced during the D₂ phase of deformation.

Fold Form

The F₁ folds recognised in competent quartz layers in semi-pelitic schists (unit 7) and competent calc-silicate layers in interlayered marble/calc-silicate rocks, (unit 3) are isoclinal, interfolial, and with attenuated limbs along which layers thin and disappear (Plate 2A,B).

Profiles of F₁ folds always indicate layer thickening in

the hinge area, and they approach those of a "general similar type" (Ramsay, 1962) (Plate 3A,B). Hinges are both rounded and sharply pointed, with a very small hinge area depending upon the difference in competency (ductility contrast) of folded layers. Where the difference is low, as in the quartzite rock (unit 4), the folds have pointed hinges on both outer and inner surfaces of the folded layer (Plate 3 B). On the other hand, where the difference is high, as in some calc-silicate-marble rocks of rock units 2 and 3, the F_1 folds in more competent layers have rounded hinges on the outer surfaces and pointed hinges on the inner surfaces. Although complex profiles with alternating rounded and pointed hinges on outer surfaces of quartz layers also occur (Plate 4A) in folds with attenuated limbs, the hinges are pointed on both the surfaces.

Axial-Surface Foliation S_1

In the metasediments the axial-surface foliation S_1 is strongly penetrative. S_1 is discordant to S_0 only locally in the F_1 hinge areas (Plates 2, 3, 4A); elsewhere on the isoclinal limbs, it is parallel to the lithologic layering S_0 and it is known as 'stratiform foliation'. In most outcrops, due to later D_2 deformation S_1 is parallel and indistinguishable from S_2 . S_1 where not obscured by S_2 , is defined by the alignment of micas (Plate 4A) and by the plane of flattening in which elongated grains of quartz or pebbles lie.

In pelitic rocks near Mayo Lake (Fig. 5, unit 5) the S_1 foliation oblique to S_2 is a schistosity defined by aligned biotite and muscovite, and flattened quartz grains. In a narrow band of metaconglomerate near the same locality the two long axes of flattened pebbles lie in the S_1 schistosity, which is at an acute angle to S_0 defined by pebble-free bands (Plate 4B).

S_1 in the quartzitic rock (unit 4) is represented by the plane in which flattened quartz and feldspar grains lie and also by the alignment of biotite crystals in segregated zones of biotite (Plate 5A). The S_1 with aligned biotite crystals is warped around 'bulges' representing the detached and later flattened F_1 fold hinges in quartz rich layers (Plate 5A).

Axial Lineation L_1

An axial lineation L_1 is defined in various ways. Long axes of boudin-like 'bulges' of calc-silicate (Plate 5B) or quartz (Plate 5A) in S_1 foliation represent detached F_1 hinges. Rarely, fine colour striations on S_1 in calc-silicate rocks indicate intersections of bedding S_0 on S_1 . In the metaconglomerate band east of Mayo Lake the L_1 lineation is defined by the intersection of bedding S_0 and schistosity S_1 (Plate 4B) and by the long axes of the pebbles (Plate 6).

Pebble Strain: Axial determinations of 11 pebbles, both in the outcrop and from loosened pebbles, indicate that the shortest (X), intermediate (Y), and the longest (Z) axes are

in a ratio of 1 : 2-4 : 6. Too few pebbles were measured for further strain analysis. Nevertheless, it is apparent that calc-silicate pebbles are flattened to a greater degree than granitic and quartzitic pebbles, reflecting a greater ductility.

S_1 in the matrix is warped around the pebbles, apparently as a result of unequal flattening of S_1 during late D_1 or D_2 . Parallellism between the S_1 schistosity in the matrix and the plane of flattening of pebbles indicates that the former also represents a plane of flattening.

If the F_1 folds were shear folds with S_1 a shear plane and pebble deformation due to simple shear, the plane in which the long and intermediate axes of pebbles lie (YZ) should be at an angle to S_1 . On the contrary, YZ coincides exactly with S_1 as to be expected if S_1 was a plane of flattening during folding. The elongation of pebbles parallel to L_1 is too large to be accounted for by the 'primary shape factor' (Ramsay, 1967, p. 203), which is the influence of initial non-sphericity of pebbles before deformation, on the finite shape of pebbles after deformation. Thus, the strain was non-plane with significant component of stretching along the F_1 axes. Carmichael (1968) also noted similar stretching of pebbles parallel to F_1 axes in area A (Fig. 1) The amount of pebble strain during the later D_2 deformation is not clear.

Orientation

Attitudes of S_1 foliation and L_1 lineation before deformation by the D_2 phase are not clear. Too few axes of F_1 folds

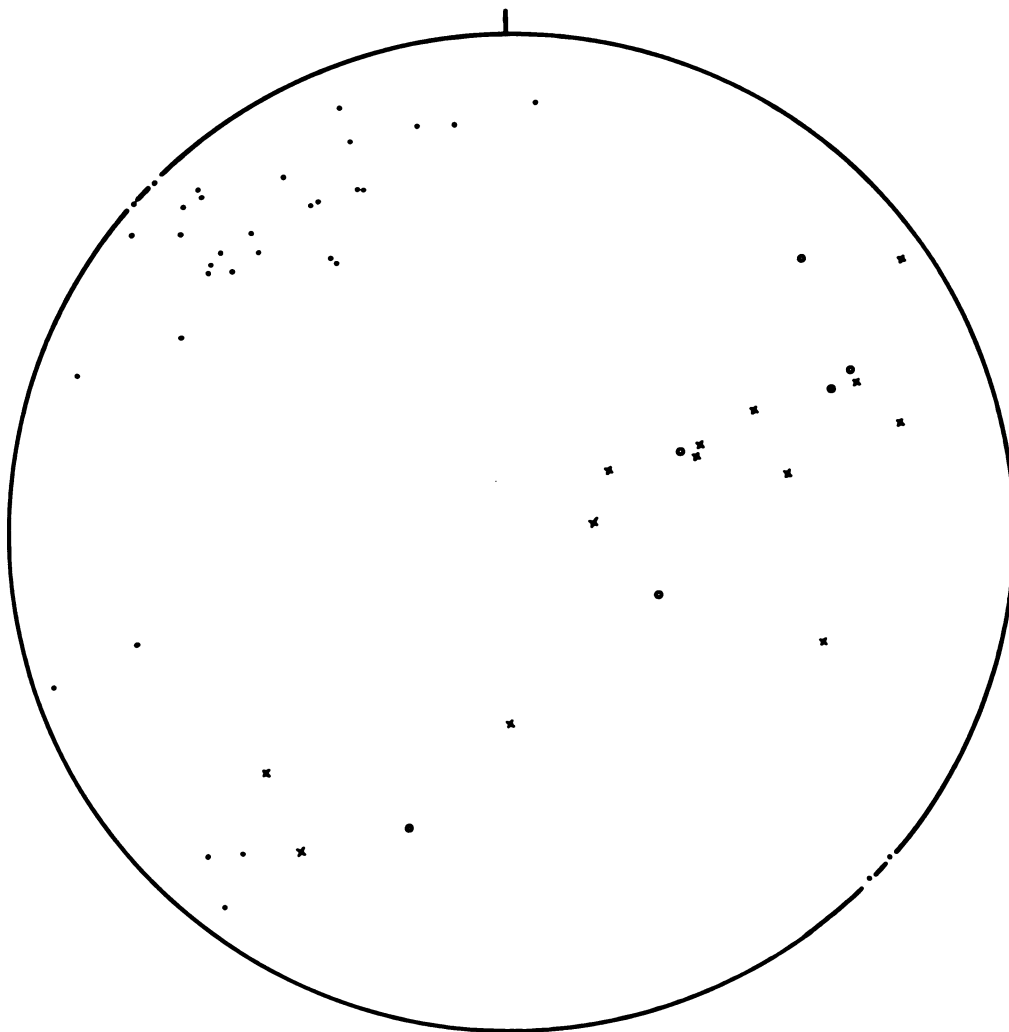


Fig.6 : Orientation of D₁ structures. Dots - Poles to S₁ foliation. Circles - F₁ fold axes. Crosses - L₁ lineation. This and other plots are on the lower hemisphere of Schmidt net.

and L_1 lineations have been measured (Fig. 6) to warrant many generalizations about reoriented trends, but a few are parallel to D_2 fold axes. In some localities as for example one kilometer east of Detlor (Fig. 5), S_1 is bent by open near upright symmetrical F_2 folds with shallow axes. This suggests that before F_2 folding, S_1 was horizontal and the F_1 folds were recumbent.

D_2 Structures

Distribution and Scale

The D_2 structures - F_2 folds, S_2 axial surface foliation, and L_2 lineation, are the most dominant and widespread structures on all scales, both in the northern gneisses and in the southern metasediments. Two major antiforms in the area at Hermon and Mayo Lake (Fig. 5) are F_2 folds. Mesoscopic F_2 folds are present in most of the outcrops in the southern metasediments, but in the northern gneisses where layering is less prominent they are represented by a few detached F_2 fold hinges in feldspathic layers. A penetrative S_2 foliation is present in all the rocks in the area except in the post- D_2 McArthur's Mills granite. In the southern metasediments the L_2 lineation in various forms is also penetrative; in the north it is most apparent in nepheline gneisses (unit 9) and less prominent in the other gneisses.

Fold Form

Distinction of F_2 from F_1 folds is possible when S_1 is bent and when an axial-surface foliation S_2 is oblique to S_1

in the hinges, for example in the hinge area of the major anti-form at Mayo Lake. However, when folds are isoclinal and hinges detached, they can be either F_1 or F_2 . In such cases study of the fabrics within and outside porphyroblasts aids in the identification (see p. 94).

In the south, F_2 folds are open to close (Fleuty, 1964) with inter-limb angles ranging between 100° to 50° and in the north, close to isoclinal with hinges detached. But local variations exist; for example, isoclinal folds west of the Weslemkoon pluton and open folds near Upper Hermon in the north.

The F_2 hinges are rounded to angular, the variation apparently depending upon the ductility contrast between layers, and upon the inter-limb angle. When the ductility contrast appears high, as in the case of layered marble and mica-quartz layers, the F_2 hinges in the less ductile mica-quartz layers are rounded, even when they are attenuated and the limbs detached from their limbs (Plate 7A). With small inter-limb angles the F_2 hinges in a sequence of layers with moderate ductility contrast, for example in scapolitic amphibolite, are detached, and pointed (Plate 7B).

Profiles of F_2 folds vary in shape; towards the north they approach those of ideal similar folds. The modification of the profile shapes of F_2 folds in response to the D_2 flattening strain is dealt with later (p. 61).

Small-scale F_2 folds are both symmetric and asymmetric. The asymmetry is not consistent with that expected for para-

sitic folds on opposing limbs of the major folds. Possibly folds of intermediate size, not apparent on the map-scale but larger than single outcrops, complicate the symmetry pattern.

Axial-Surface Foliation S_2

In the hinge areas of F_2 folds, an axial-surface foliation S_2 is discordant to the S_1 foliation. Away from the hinges, especially where folds are isoclinal, S_2 is closely parallel to S_1 .

The character of S_2 varies. In calc-silicate rocks it is commonly a fracture cleavage which in open to close folds in some feldspathic layers fans outward (Fig. 7). But more generally the fracture cleavage is closely parallel to the axial surfaces of F_2 folds (Plates 8A,B). Where the fracture cleavage crosses lithologic layers whose competency varies, it refracts from one layer to the other (Plate 9A,B). The spacing between the cleavage planes measured in 6 folds varies from 1 mm to 6 mm, and appears to be proportionally related to the thickness of the layer in much the same manner as joint sets elsewhere (Price, 1966). Sometimes calcite is recrystallized along the axial surface fracture cleavage S_2 of F_2 folds (Plate 10A).

In marbles and quartzite, S_2 is commonly a schistosity in which long axes of flattened grains of calcite and quartz lie. In pelitic rocks an S_2 schistosity is defined by the alignment of micas and in the northern pelites (unit 7) sillimanite is

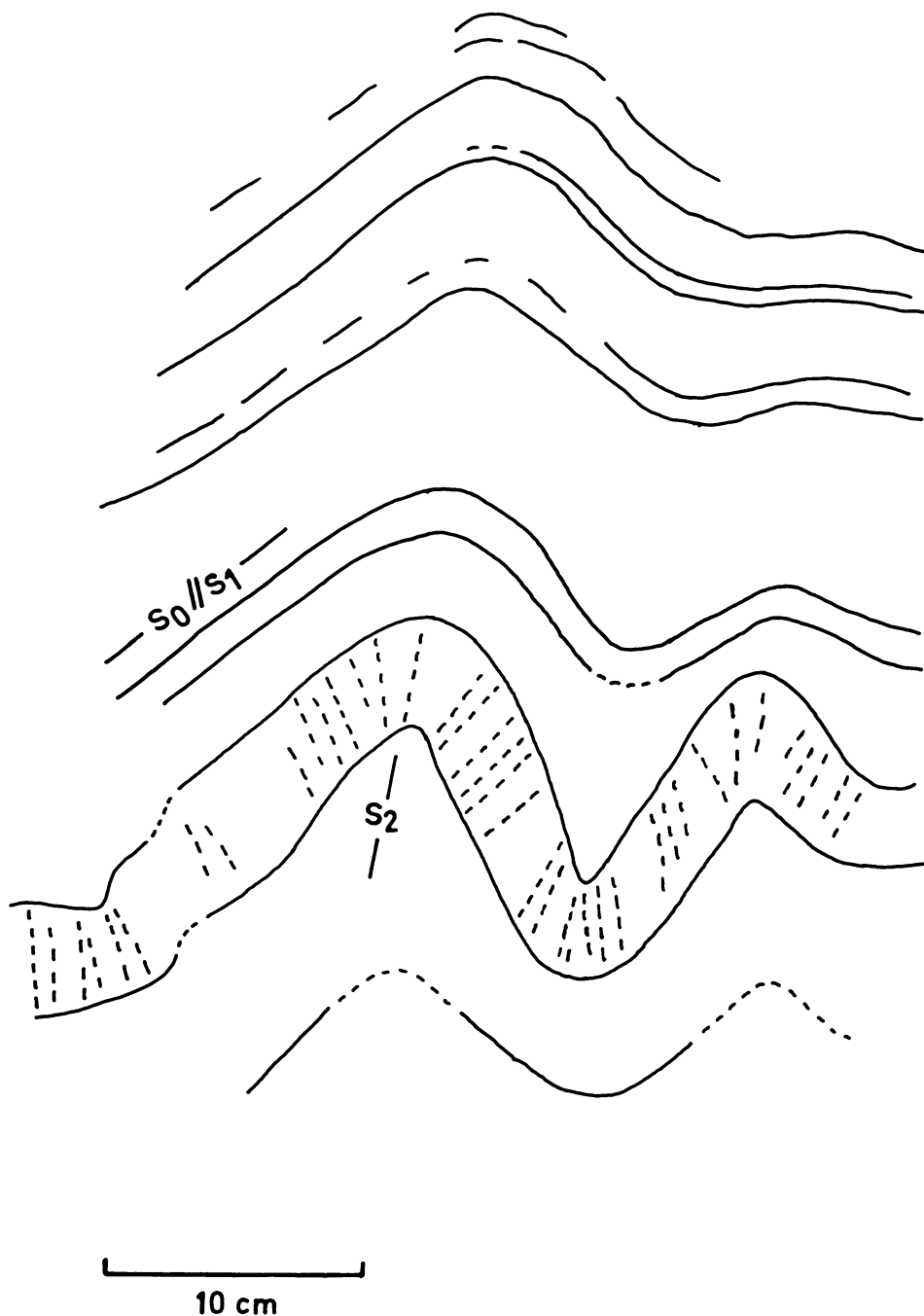


Fig.7 : Fanning of S_2 fracture cleavage in F_2 fold in feldspathic layer within calc-silicate rock (Unit 2).
Locality : One kilometer south of Green Corners.

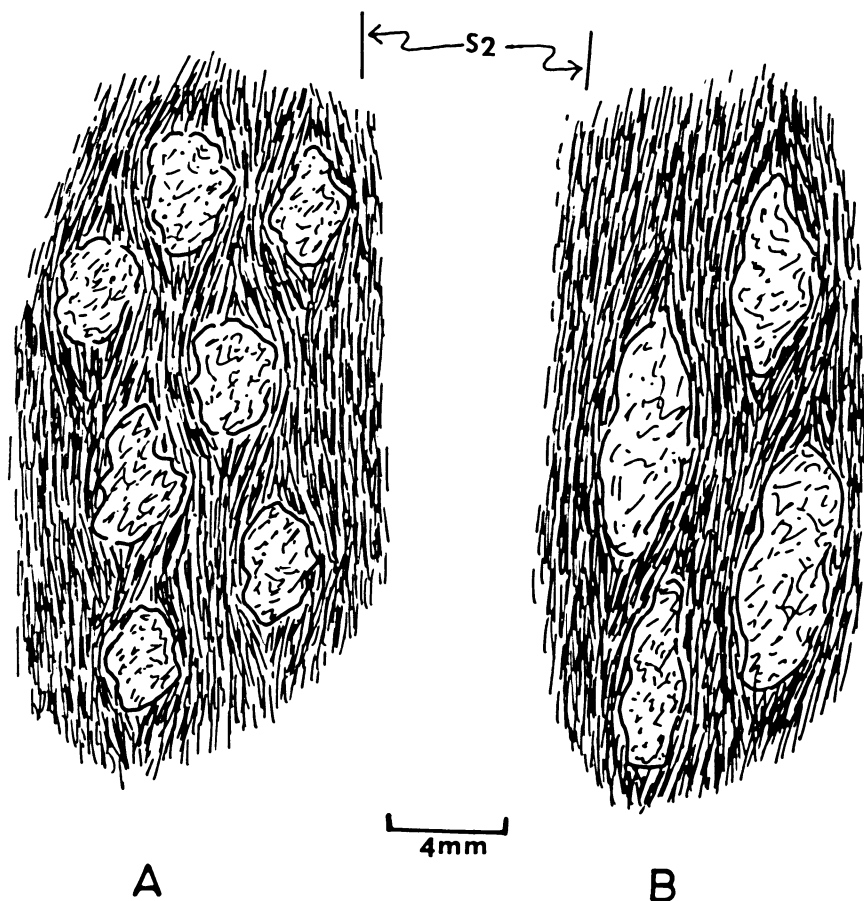


Fig.8 : Alignment of sillimanite pods parallel to L_2 in semi-pelitic schist (Unit 7). S_2 is defined by the alignment of micas and sillimanite.

A: Surface cut perpendicular to S_2 and L_2 .

B: Surface cut perpendicular to S_2 and parallel to L_2 .

Locality : One kilometer north northeast of Upper Hermon.

also aligned parallel to S_2 (Plate 10B). Flattened sillimanite 'pods' lie in and accentuate S_2 (Fig. 8). In some sillimanite schists, pre or early D_2 sillimanite needles are crenulated during the D_2 phase of deformation. On the limbs of these crenulations, the sillimanite needles are aligned subparallel to S_2 (Plate 11A).

Within the banded scapolite amphibole rock of unit 7 and amphibolite (unit 1), S_2 in which hornblende crystals are aligned is oblique to S_1 in some F_2 folds (Plate 11B) and locally in some places where such folds in S_1 are absent (Plate 12A). However, towards the north where F_2 are nearly isoclinal S_2 is essentially parallel to S_1 (Plate 12B).

Where S_1 and S_2 are parallel, the foliation may be considered as a "composite schistosity" first developed during D_1 as S_1 , and then rotated and accentuated by recrystallization to coarser grains parallel to S_2 . In the absence of mesoscopic F_2 folds, the composite character of S_2 can be seen in thin sections containing porphyroblasts. For example, in pelitic rocks garnet porphyroblasts include a fine quartz fabric S_1 (S_i) which is both oblique and parallel to the coarse external schistosity S_2 (S_e) (Plate 13). Apparently, the pre- D_2 garnet porphyroblasts preserved the fine S_1 fabric (see also p. 86 and Fig. 24C, D). The fine quartz fabric (S_1) could be interpreted as to have developed as a result of consumption of quartz during the growth of garnet. However, the nature of the fabric does not support this interpretation. Individual quartz grains in the fabric are uniformly fine grained and are

inequidimensional in all the garnet porphyroblasts; it is extremely unlikely that all garnets consumed quartz in preferential directions and to the same amount in every locality.

In the northern gneisses where layering is less pronounced, the gneissosity is considered to be S_2 as it is subparallel to the S_2 foliation in the metasedimentary and metavolcanic rocks to the south. Also the gneissosity is slightly discordant to some feldspathic layers or lenses (S_1 ?) and axial planar to a few disrupted and tight fold hinges (F_2 ?).

In the nepheline gneisses (unit 9) the S_2 gneissosity is defined by segregated mineral layers which are discontinuous. Inequidimensional minerals such as hornblende and biotite are aligned parallel to the gneissosity as are elongated nepheline and plagioclase grains. Similarly, hornblende is aligned parallel to S_2 gneissosity in the Mallard Lake metagabbro (unit 8). In the coarse grained mixed gneisses (unit 10), the S_2 gneissosity is less prominent.

Axial Lineation L_2

The character of the L_2 lineation varies according to the rock type and the intensity of the deformation. It is expressed in different forms as intersections of S-surfaces, oriented elongate minerals, axes of crenulations, rodding, and elongated mineral segregations.

L_2 is ubiquitous on the S_2 surfaces in the southern metasediments. In the northern gneisses it is not as prominent and widespread. The L_2 lineation in all cases is parallel

to the axis of the local mesoscopic F_2 folds.

In outcrops where mesoscopic F_2 folds are either absent or not readily distinguished from F_1 folds, the lineation could be either L_1 or L_2 . In such cases the character of the lineation helps in identification. For example, in many outcrops within the northern semi-pelitic unit (unit 7), no mesoscopic F_2 folds could be found, but because sillimanite growth was during D_2 not D_1 , the needle alignment defines L_2 .

The most commonly represented form of L_2 in the southern metasediments is the intersection trace of the stratiform foliation S_1 and S_2 fracture cleavage or schistosity (Plate 8A,B). Particularly in the calc-silicate-marble rocks (unit 3) L_2 is represented on S_2 by a series of parallel ribbons representing the intersections of darker calc-silicate layers and lighter marble layers on the S_2 fracture cleavage (Plate 14A).

Towards the north, where the F_2 folds are disrupted and transposed (Plate 12B), their detached hinges form rod-like lines on the S_2 foliation.

In some pelitic rocks a very characteristic type of lineation is present in the form of aligned edges of biotite and muscovite crystals on the S_2 surface. As this type of lineation can form in various ways that are not readily distinguishable, a general descriptive term "mica-edge lineation " (Fyson, 1970, p. 113) is preferred. In the semi-pelitic schist (unit 7) outcropping along Highway 500 east of Bancroft and near Mayo Lake the mica edge lineation can be identified as:

a) S_1 mica flakes intersecting S_2 foliation at low angles; b) S_2 mica warped around L_2 quartz lenses or rods; and c) S_2 mica flakes crinkled near the hinges of F_2 folds. Some individual mica flakes following the crinkles are bent; others are not, evidently due to post- D_2 mimetic recrystallization.

Axes of quartz and feldspar rods define L_2 in some pelitic rocks, as for example in sillimanite bearing schists about one mile northwest of Detlor. Unlike L_1 quartz 'bulges', the L_2 rods are closely and regularly spaced in the mica schist. Occasionally grooves resembling slickensides parallel the axes of the rods. These 'striae' probably represent the impingement marks of S_1 mica on the rods.

Alignment of individual crystals of hornblende in amphibolites, tremolite in calc-silicate rocks, and sillimanite in the northern semi-pelitic schist give another form of L_2 lineation. Hand specimens cut in three mutually perpendicular directions best exhibit this type of lineation. In the northern semi-pelitic schist (unit 7) L_2 is also represented by elongated pre- D_2 biotite porphyroblasts and "pods" of sillimanite (Fig. 8A,B). The longest dimension of the biotite porphyroblasts and the sillimanite pods is parallel to the F_2 hinge lines in quartz rich layers.

The nepheline gneisses (unit 9) exhibit a well developed lineation lying within the gneissosity plane. The lineation is defined by elongated lens-like and rod-like aggregates of nepheline, plagioclase, and hornblende (Plate 14B); it is parallel to sillimanite needles aligned along L_2 in the nearby

semi-pelitic rocks (unit 7) to the south, and hence it is taken to represent L_2 . As mentioned previously, there are a few detached fold hinges, presumably F_2 , to which the gneissosity S_2 is axial planar. The fold axes like F_2 in the metasediments are parallel to the lineation. Biotite flakes lying in the gneissosity are strongly bent around the linear aggregates, probably as a result of strong flattening of S_2 during late D_2 .

Within the L_2 aggregates, the long axes of hornblende crystals lie parallel to S_2 , but there is no preferred orientation to the elongation direction of the aggregates. This fact is clear for hornblende in thin sections cut perpendicular and parallel to L_2 . To investigate whether c-axes of nepheline crystals are aligned parallel to the L_2 aggregates, thin sections were made perpendicular to L_2 and the orientation of the c-axes of nepheline measured with a universal stage. Like hornblende some c-axes of nepheline are aligned (Figs. 9A,B) parallel to the S_2 gneissosity though others are random (Figs. 9C, D). There is no apparent preferred orientation parallel to L_2 .

In the Mallard Lake metagabbro (unit 8) a nonpenetrative L_2 lineation is defined by lenses of plagioclase and hornblende elongated within the gneissosity. Unlike the hornblendes within the nepheline gneisses, the c-axes are aligned along the L_2 lenses; this is evidenced from basal sections of hornblende in thin section perpendicular to L_2 . Another interesting feature is apparent in thin section perpendicular to L_2 ; the longer diagonals of the rhombohedral basal sections of the

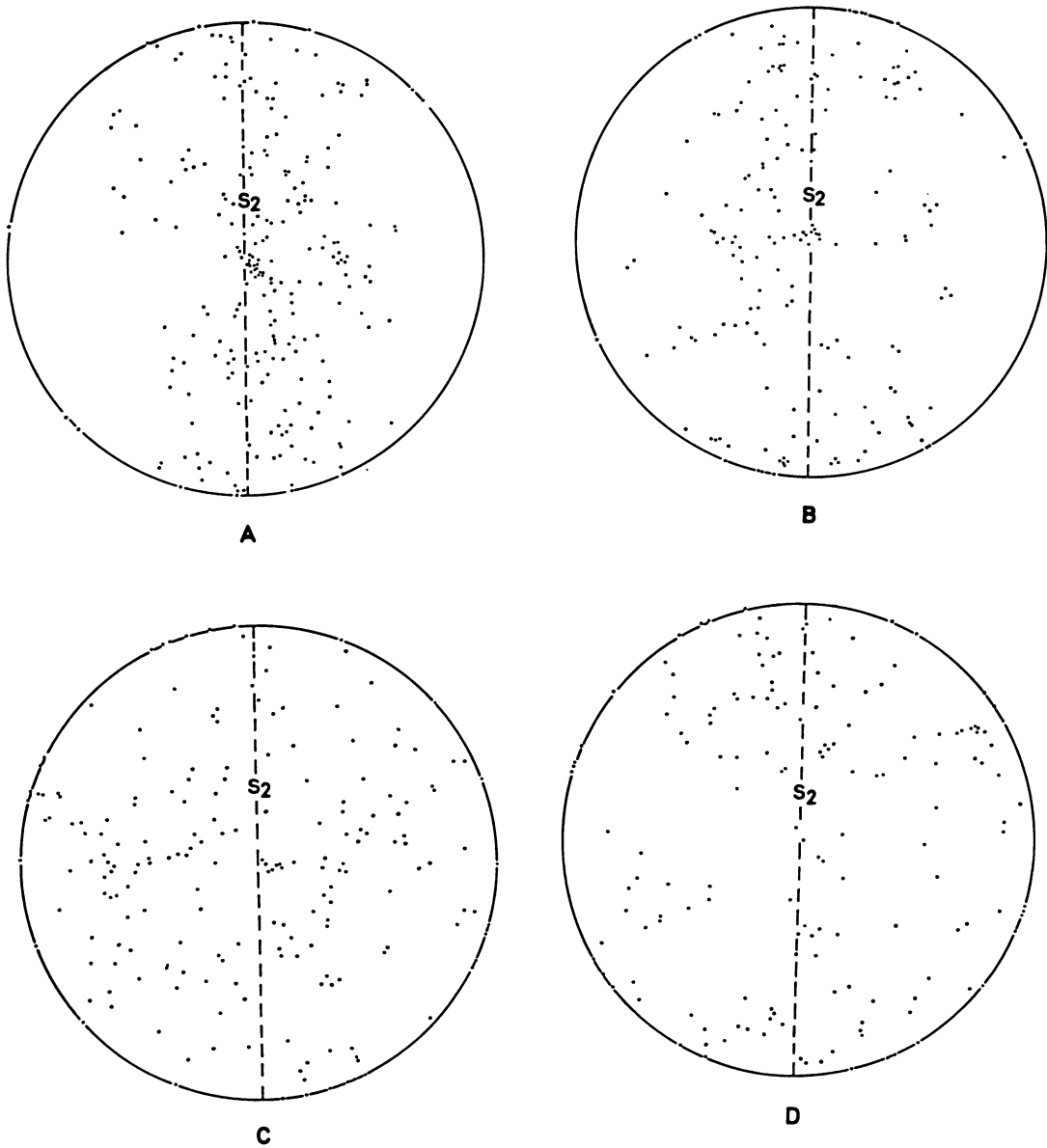


Fig. 9 : Orientation of c-axes of nepheline crystals.

hornblende crystals are aligned parallel to the S_2 gneissosity.

Boudinage is a common feature associated with F_2 folds (Plate 15). In the limb areas of tight and isoclinal folds, the more competent layers are boudinaged, presumably as a result of stretching parallel to the layer. This feature is displayed strikingly by the competent dark amphibolite layers which alternate with light coloured marble layers. Axes of boudins have variable geometric relations with associated and genetically related folds elsewhere (Fyson, 1962, p. 214; Jones, 1959, p. 118), but in the southern part of the study area they are statistically parallel to the axes. Towards the north, the axes of the boudins have no consistent relation to the fold axes. The probable cause of this variation is explained later (p.69).

Orientation

Areal variation of S_2 and L_2 : Orientation of D_2 structures varies from place to place within the area (Fig. 5). To illustrate the spatial variation, the area is divided on the basis of structural homogeneity of L_2 and lithology into 13 domains (Fig. 10). Following standard procedures, the fabric elements of D_2 for each domain are plotted on a lower hemisphere equal-area (Schmidt) projection (Fig. 11). Since the size of the domains is not large, the number of points on each projection except for a few cases is less than or near 100, and contouring is not necessary to indicate the relations.

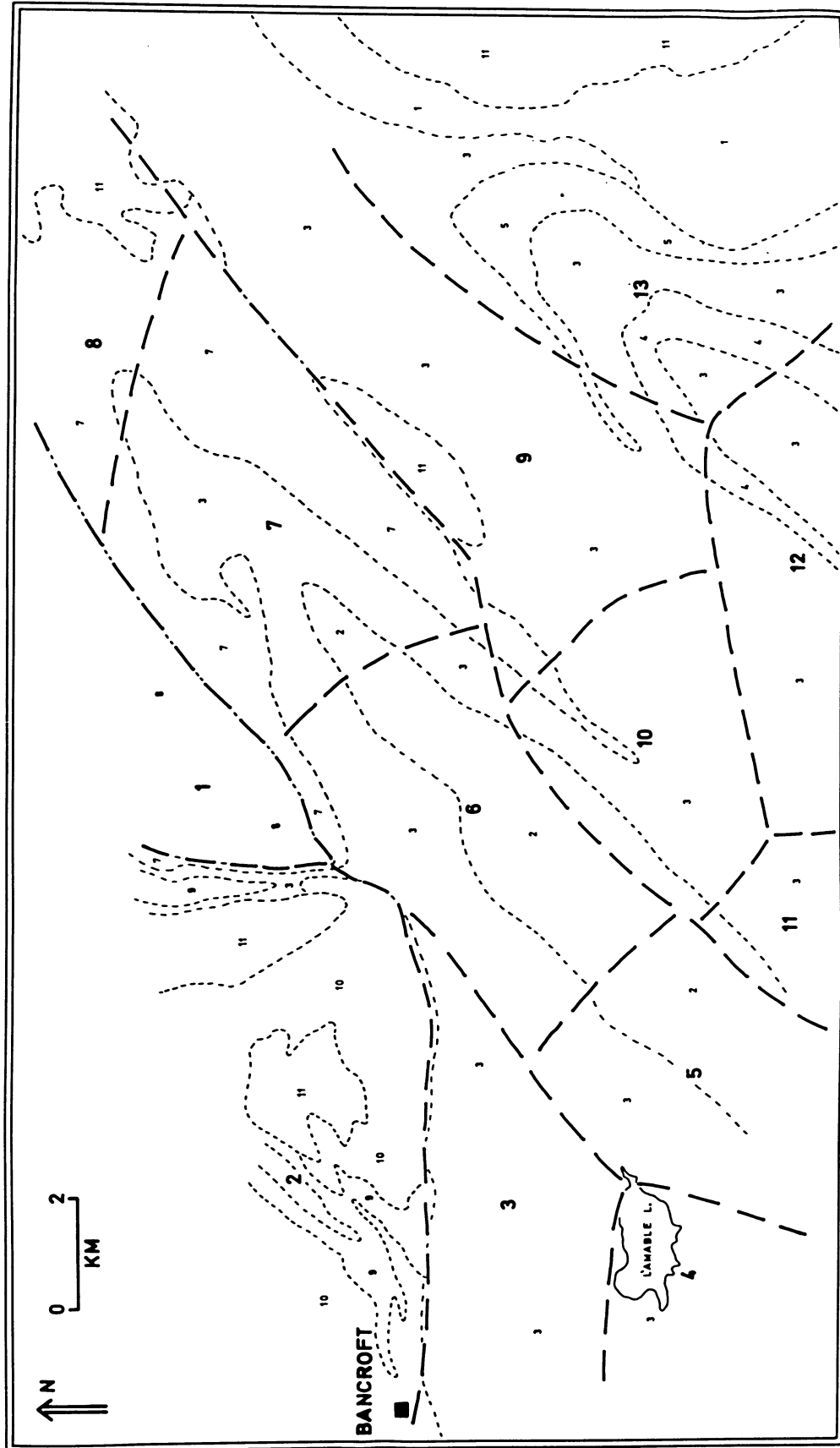


Fig. 10 : D2 structural domains in Bancroft area. Domain boundaries are drawn on the basis of homogeneity of S₂ and L₂, and lithology. Rock units numbered as in Fig.2.

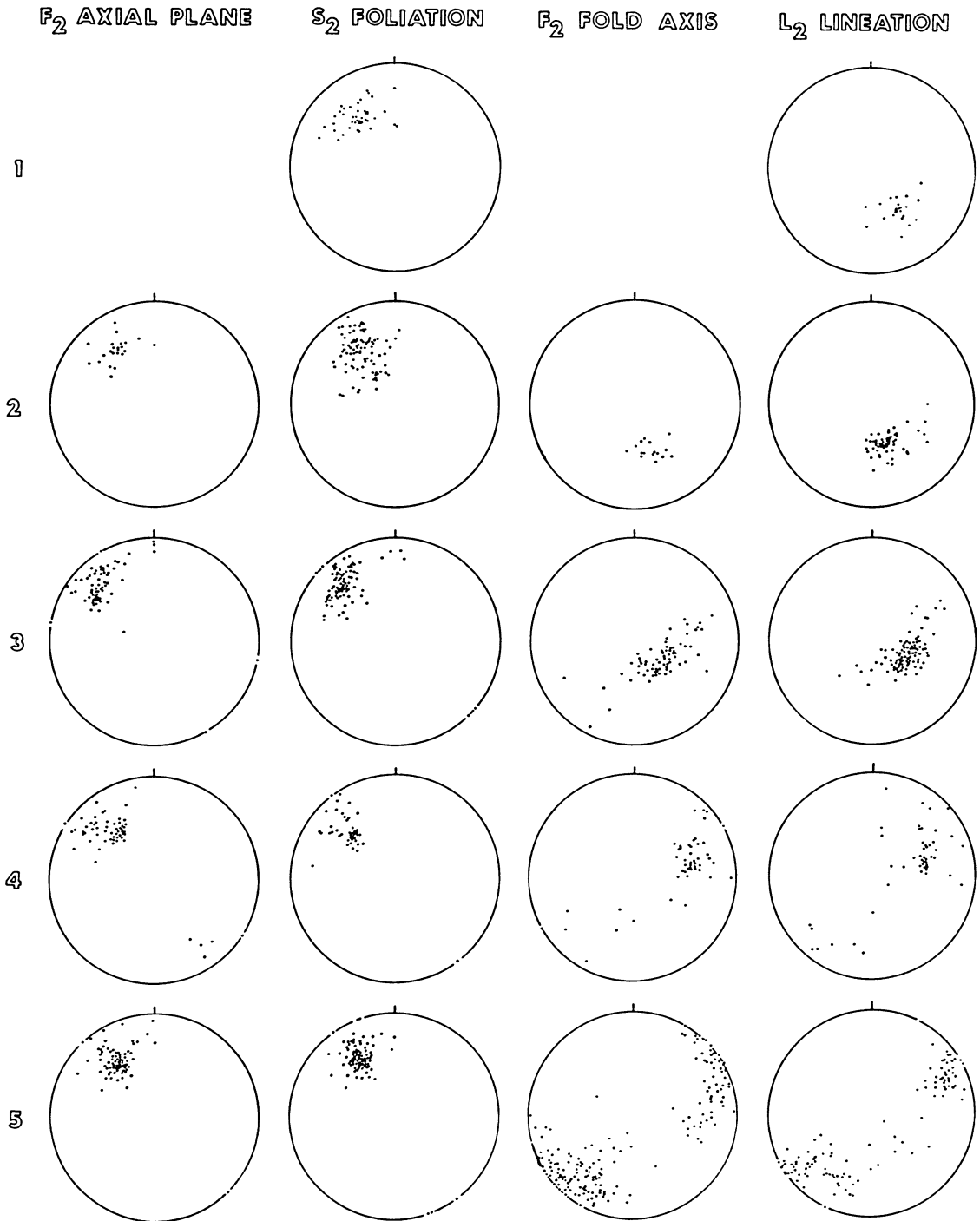
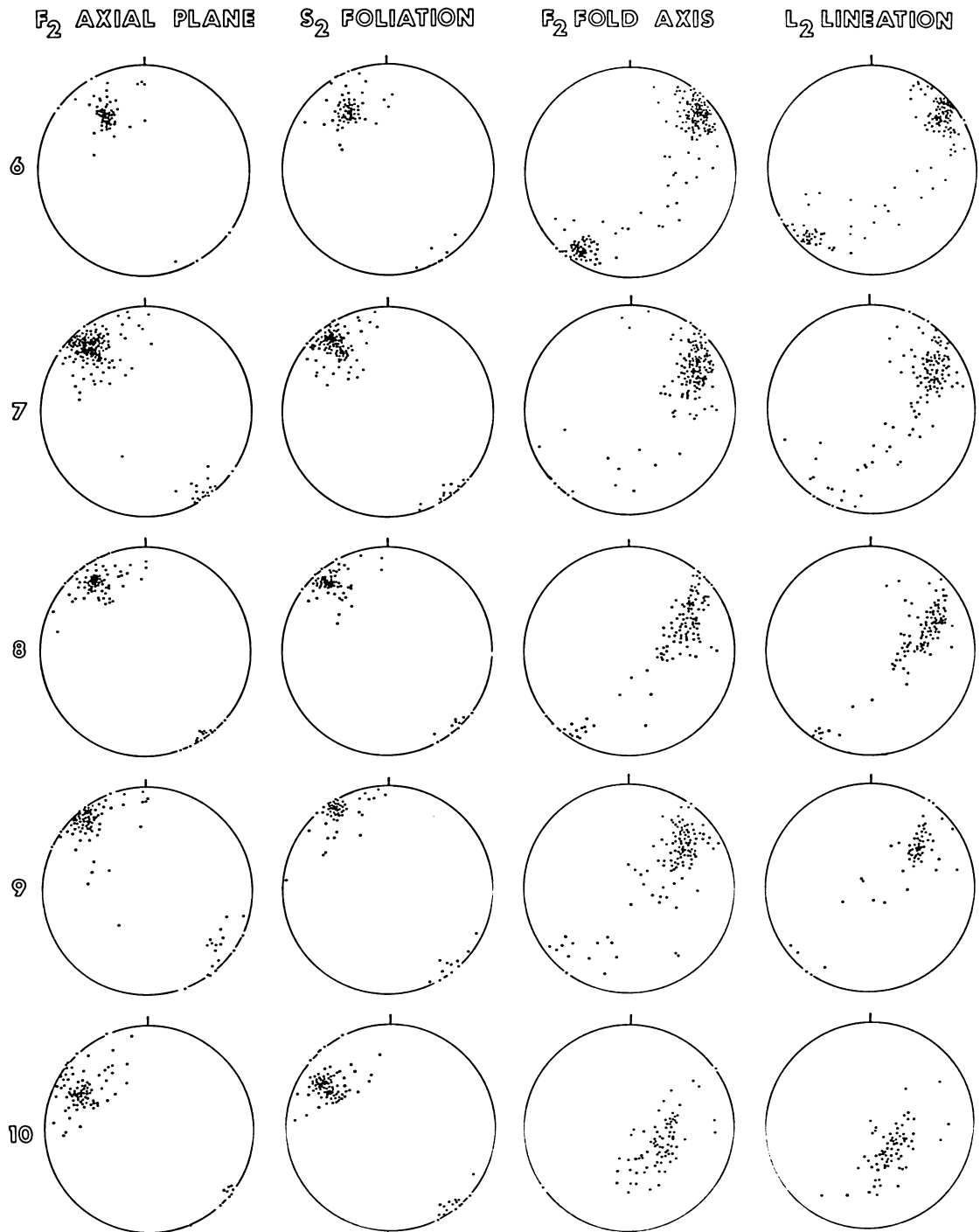
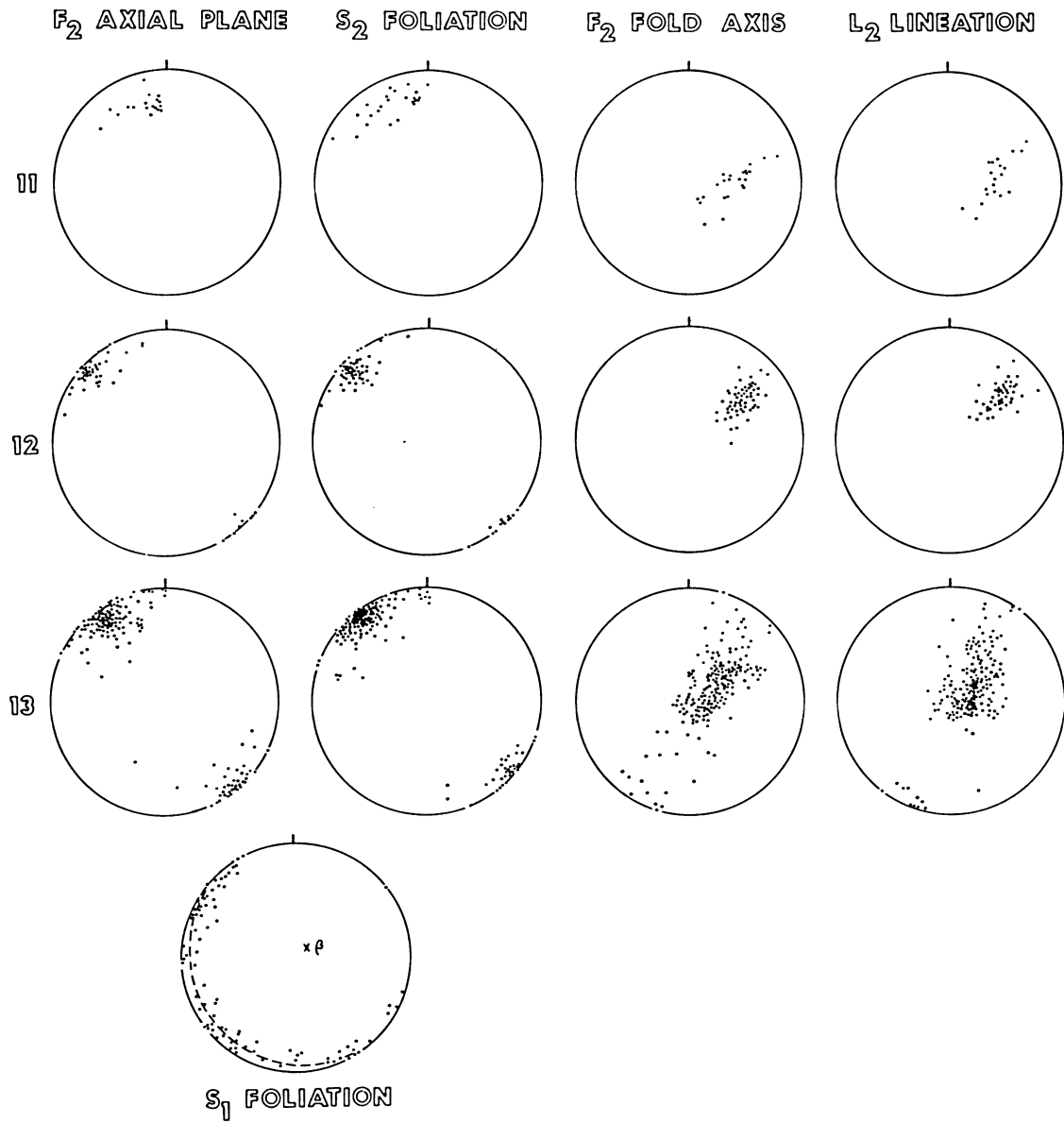


Fig.11 : Net projections of D_2 structural elements in domains 1 to 13. β axis of F_2 antiform at Mayo Lake in domain 13 is also represented.





Close parallelism of the S_2 foliation to the axial surfaces of the F_2 folds both in the northern gneisses and in the southern metasediments is evident from Figure 11. S_2 strikes fairly constantly northeastward, but locally, veers towards the east, for example in domain 11 (Fig. 11-11). In most cases of the area the S_2 foliation is moderate to steeply dipping to the southeast. S_2 dips moderately ($40^\circ - 55^\circ$) in the northern gneisses, but steepens in the metasediments towards the Weslemkoon pluton near which it is subvertical. Locally, away from the Weslemkoon pluton, S_2 is also steep to vertical (domains 3 and 8). These areal variations will be considered in more detail in a later part of this chapter.

Parallelism between axes of the F_2 folds and L_2 lineations in all domains (Fig. 11) reflects the invariant axial orientation of L_2 . In outcrops where mesoscopic F_2 folds are scarce, such as in the northern gneisses, L_2 thus reveals the axial direction.

In the southern metasediments, the F_2 axes and L_2 vary considerably in trend and plunge. For example, in domains 4 to 9, and 12, they trend northeast to east, and plunge at low angles both northeast and southwest (Fig. 5). To the northwest, in marble-calc-silicate and pelitic rocks in the northern part of domain 3, the axes of tight to isoclinal folds plunge at moderate to steep angles to the southeast, often down the dip of S_2 . Farther to the northwest, in the northern gneisses, L_2 and a few F_2 hinges also, plunge moderately to the southeast down the dip of S_2 . In the hinge

region of the antiform at Mayo Lake (Fig. 5; domain 13, Fig. 10) fold axes and L_2 plunge at steep to very steep angles to the northeast (Fig. 11-13). The S_1 foliation is bent around the hinge of the antiform and the β axis determined by poles to S_1 also plunges steeply to the northeast (Fig. 11-13).

S_2 dip Isogons: It is important to note whether or not the regional orientation of S_2 in the study area is reasonably constant; if too variable, lineation isogons which are considered later, could be misleading. In the study area, S_2 has a constant regional northeast-southwest strike. Therefore it is not necessary to prepare strike-line maps (Agterberg, 1964, Fig. 4).

But S_2 varies in dip and these areal variations are best illustrated by mapping dip isogons to S_2 foliation (Fig. 12). The isogons are the lines joining points where angles of dip are equal; they are the trace upon the ground surface of isogonic surfaces. To prepare the dip isogonic map for S_2 , control stations were selected so that the areal density reflected the degree of areal change. Selection of control stations in this manner yielded more significant results than the use of regular grid pattern. The 'modal' dip value for each control station was established visually from net projection of measurements made at the station and adjacent locations, and isogons were drawn at a regular interval of 10° .

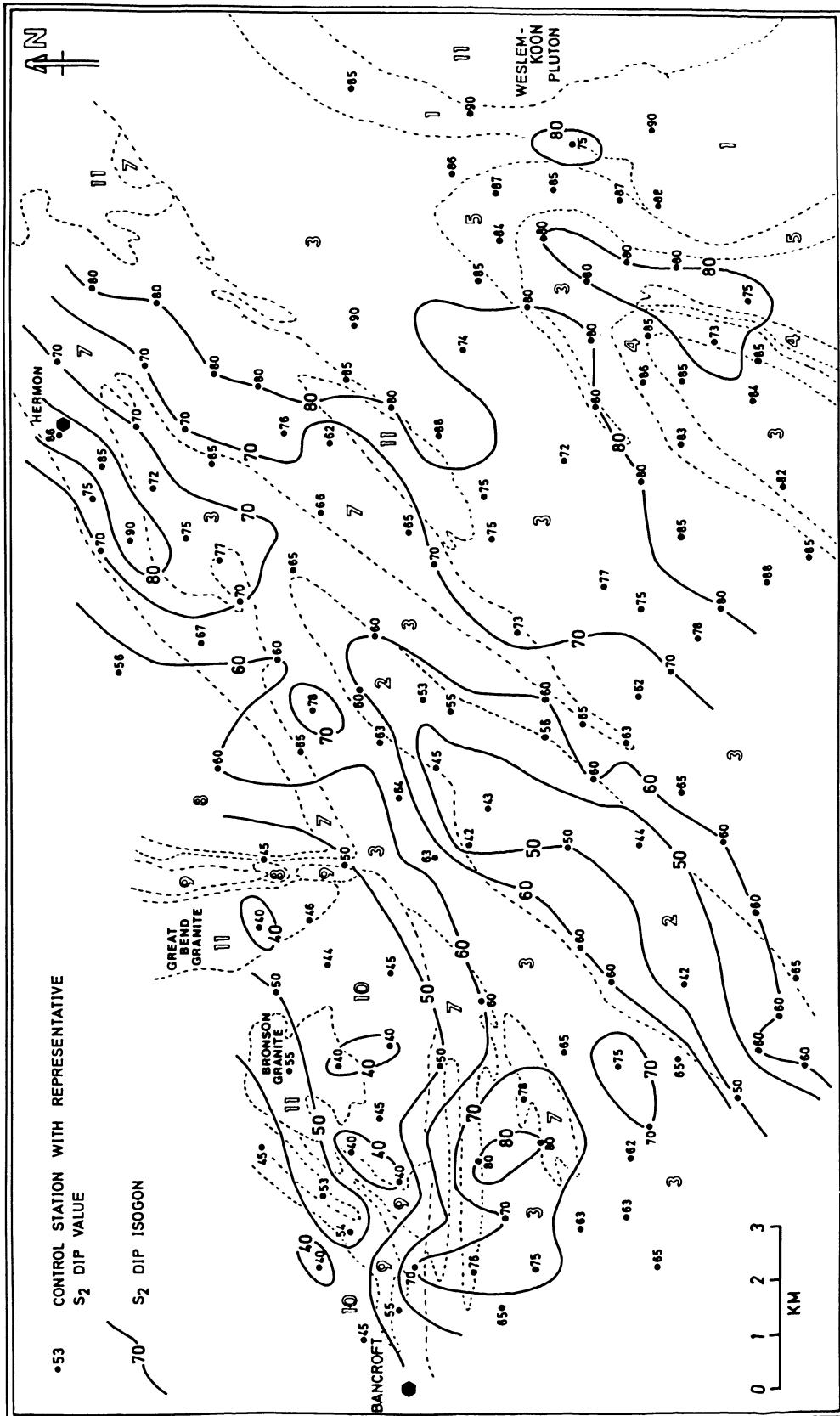


Fig.12 : S₂ dip Isogonic map. All dips to southeast. Rocks units numbered as in Fig.2.

Fig. 12 shows that in most of the area S_2 dips over 50° southeast. In a zone extending several km westward from the Weslemkoon pluton, S_2 is near vertical, and similarly it is steep in a discontinuous zone bordering the northern gneiss and granite complex. Within the complex S_2 is generally less steep, with most dips less than 50° southeast.

For the map-area as a whole, the dip isogons tend to parallel the northeasterly strike of S_2 and the axial traces of the major F_2 folds. S_2 is thus most variable in dip across the strike. However towards the northeast, for example along the antiform that extends past Hermon, S_2 steepens, and in the northeastern sector it is steep throughout. Of note, is that in this sector the Weslemkoon pluton is closest to the northern gneiss complex (see also Fig. 1).

The areal relationships suggest that S_2 may be steep marginally to the Weslemkoon pluton and the gneiss complex because these bodies horizontally constricted the intervening rocks during D_2 deformation. This suggestion is further amplified after considering the variation in plunge of F_2 folds and L_2 axial lineations.

L_2 Rake Isogons: In illustrating on a map areal variations in the orientations of L_2 , the rake angle in the S_2 plane, rather than the trend or plunge, was found to be most useful. For example, where L_2 plunges steeply on steep S_2 , slight

Variations in dip of S_2 drastically affect the trend. Also it can be demonstrated that in most cases, the rake of L_2 is more sensitive than plunge in revealing variations (Elliott, 1965).

An isogonic map for the rake of L_2 (Fig. 13) was prepared following the method given by Elliott (1965). A modal rake angle for each of the control stations was found from the net projections of L_2 (10-25 measurements) measured around the control station. As the standard deviations at the control stations rarely exceeded 12° , the statistically significant contour interval is 20° (see Elliott, 1965, Fig. 9). In drawing the isogons, the L_2 rake angle is treated as non-parametric in S_2 , that is, northeast and southwest rakes are not differentiated.

The L_2 rake isogonic map (Fig. 13) indicates that the L_2 rakes in the southern metasediments are in general low to moderate, except southeast of Detlor where locally the rake is high.

The most important feature in Fig. 13, is that L_2 rakes approach the possible maximum of 90° towards and within the northern gneisses. High rakes within the gneisses are readily inferred from the standard trend-plunge map (Fig. 5) where it is apparent that L_2 lies close to the dip direction of the S_2 gneissosity. However, not as apparent as in the rake map, is that in a northerly and northwesterly directions from Detlor and Green Corners, the L_2 rakes gradually increase (Fig. 13) from low through medium to high, and reach nearly

90° in the northern gneisses.

In the metasediments with subvertical S_2 close to the Weslemkoon pluton the L_2 rakes are high. Away from the Weslemkoon pluton in a northwesterly direction, the L_2 rakes become moderate again.

Locally, southeast of Detlor, L_2 rakes change from 0° to 90° in a distance of 2 kilometers. Unlike in the metasediments close to the Weslemkoon pluton, S_2 foliation here is not subvertical, but has moderate to steep dips.

The mechanical interpretation of the changes in rake will be given later (p. 69).

The L_2 rake isogonic map aided in delineating the D_2 structural domain boundaries (Fig. 10). The boundaries initially drawn on the basis of obvious structural homogenities were modified to follow where the isogons are close together and thus indicating rapid changes in L_2 orientation.

L_2 Rake Moving Averages: It is apparent from Fig. 5, and from Fig. 13, that locally, the attitude of L_2 lineation changes rather abruptly. This local effect can mask the regional trend in L_2 variation in orientation. Therefore, it is useful to analyse the L_2 orientation on a larger 'statistical' scale than the one in Fig. 13, so that local 'anomalies' are dominated and masked by the regional trend.

The method of analysis is similar to the one used widely by sedimentologists. Two-dimensional "moving averages" have

been constructed for maps that show the generalized trends in the areal variation of scalar properties such as grain size, as well as orientation of directional structures such as current direction (Potter, 1955; Schlee, 1957; Pelletier, 1958; and Fahrig, 1961).

The two dimensional regional variation of L_2 orientation in S_2 foliation (that is L_2 rake) is analysed by using the moving average method (Fig. 14). Though any arbitrary shape of the grid can be used, it is found out that if the distance between the grid lines is at least half of the width of the zones in which local variations occur, the effect of the local variations on the regional trend will be minimized. Obviously, when the grid width is too large, the 'strength' of the regional trend becomes weak and it may disappear. On the other hand if the grid width is too small, the local variations do not disappear, and the regional trend cannot be clearly shown.

A part of the grid system used is shown at the northwestern corner of Fig. 14. At each grid intersection, a mean value for all the observations in the four adjacent quadrants (1, 2, 3 and 4) is calculated. Because the variation of L_2 lineation in constantly oriented S_2 foliation is simply the change in azimuth of a line (L_2) in a plane (S_2), the arithmetical mean of these azimuths will not necessarily always indicate the true central tendency of the distribution. Instead, the vector mean (Potter and Pettijohn, 1963, p. 264) gives the true central tendency of the distribution, and should be computed for any azimuth data.

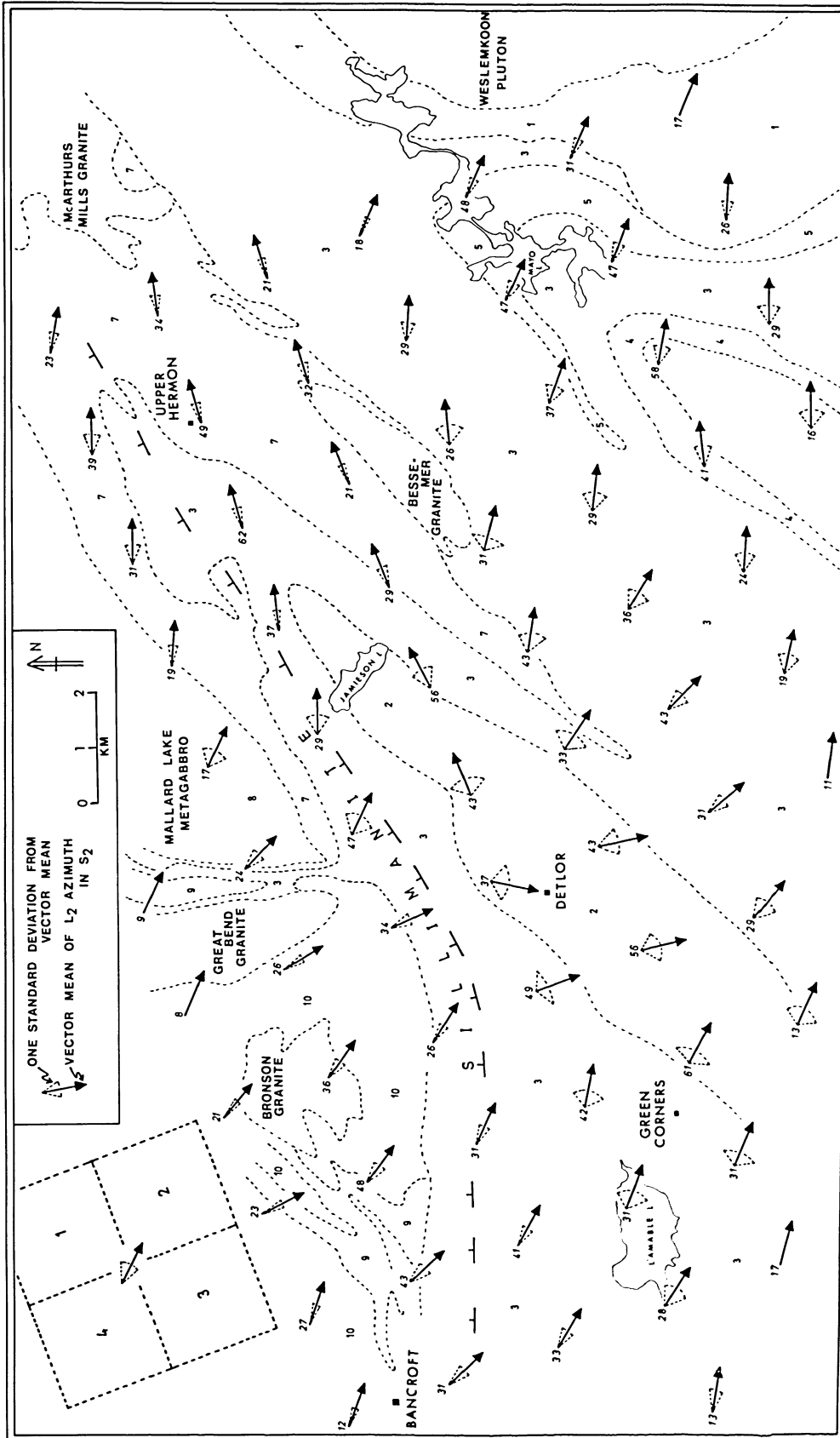


Fig.14: L₂ rake Moving Average map. S₂ with constant northeast strike and steep south-east dip has been rotated on its strike axis to horizontal map plane. Number of measurements shown at the tail end of arrow. Rock Units numbered as in Fig.2. Grid system for averaging as in northwest corner.

At each grid intersection in Fig. 14, the S_2 foliation dipping to the southeast is rotated on its northeast strike line to the horizontal, so that the plane of the map sheet is parallel to S_2 foliation. Therefore, what the map exhibits essentially is, the areal variation in orientation of L_2 in S_2 whose original 'dip line' is now horizontal and trending southeast in the map.

The direction pointed by the small arrow head at each grid intersection represents the vector mean of L_2 in S_2 . A circular arc representing one standard deviation from the vector mean of L_2 in S_2 is drawn on both sides of the line, at the other end of the arrow, and represents the amount of dispersion of L_2 about its mean value at each grid intersection.

Fig. 14 indicates certain features which are significant. Except in the vicinity of Detlor, the azimuth of L_2 in the metasediments lies between east northeast and east southeast. These azimuths gradually change to the southeast as we pass through the metasediments in a northwesterly direction and reach the northern gneisses. This means that L_2 becomes closer and eventually parallel the dip direction of S_2 as we pass through the metasediments north of L'Amable Lake (Fig. 14) and into the northern gneisses. Similar changes in L_2 azimuth in the metasediments can be seen also when we proceed from Jamieson Lake (Fig. 14) in a southeasterly direction towards the Weslemkoon pluton. The pattern of the change in L_2 indicated in the L_2 rake isogonic map (Fig. 13), is thus

supported by the moving average map.

Another interesting feature that can be observed in Fig. 14 is that, both northwestwards from Green Corners and Detlor, and southeast from Bessemer granite toward the Weslemkoon pluton, the rakes become larger and the standard deviations become smaller. That means, in these two regions L_2 is dispersed when it is away and clustered when it is close to S_2 dip direction. However, such inverse relationship between rake and standard deviation is not evident for the map as a whole, because locally, as near Upper Hermon, both rakes and standard deviations are low.

Profile Analysis of F_2 folds

Introduction: In many but not all cases, F_2 folds are open to close (Fleuty, 1964) in regions of low to moderate rake of L_2 and F_2 axes (Plates 10A, 14A), and close to isoclinal in regions of moderate to high values of rake (Plate 12B). Some relationship between the rake of F_2 fold axes in S_2 foliation and profile shapes of F_2 folds is thus suggested from direct observations. The following is a more complete analysis of the fold shapes using detailed measurements of layer thicknesses.

The F_2 fold profiles cannot be grouped in a simple way into parallel and similar types. In fact, folds belonging exactly to these classes are never seen in the study area. Rather, the methods given by Ramsay (1967, p. 359) are used to describe, classify, and analyse the F_2 profile shapes.

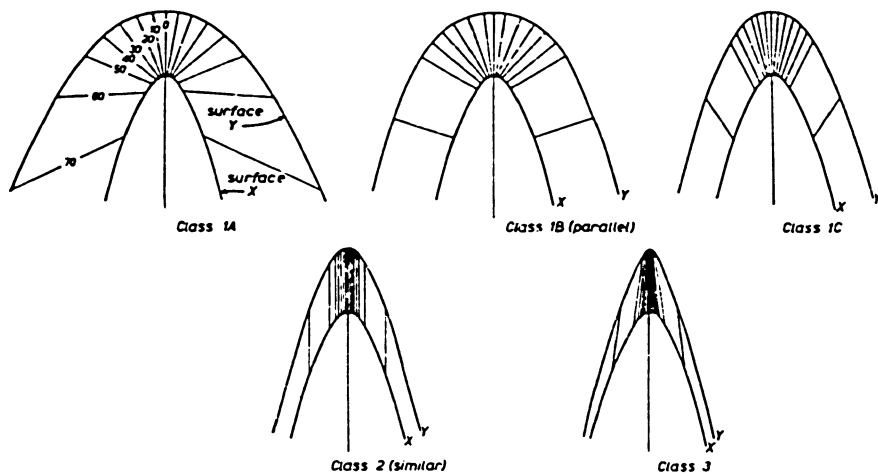
The methods are entirely geometrical, and any possible folds affecting two or more surfaces can be analysed by measuring one or all three geometrical parameters: the orthogonal thickness, the thickness parallel to the axial trace, and the dip isogons. These parameters, which are not independent of each other, are used to graphically indicate relative rates of change of the inclination of successive surfaces and hence to depict fold shapes.

In brief, Ramsay's methods for the three parameters are:

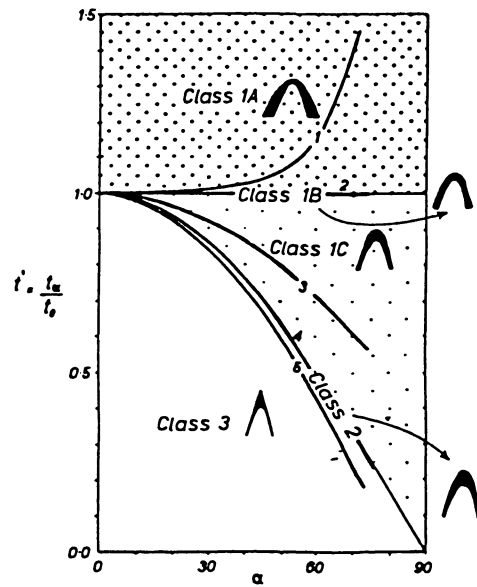
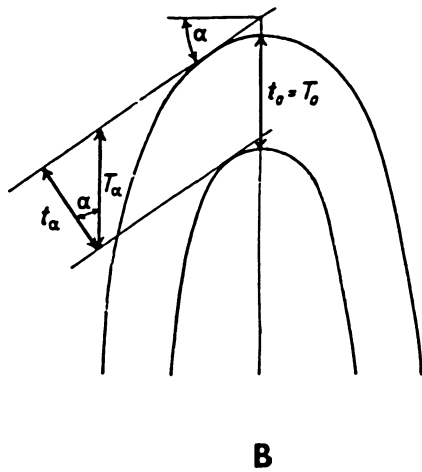
1. The orthogonal thickness t (Fig. 15B) of the material between two folded surfaces can be determined by constructing parallel tangents to each surface. At any inclination $\mathcal{L}$ of the tangents it is possible to express the ratio of the thickness $t_{\mathcal{L}}$ to t_0 measured at the hinge of the fold (Fig. 15B) where $\mathcal{L}$ is 0. In non-parallel folds this ratio $t_{\mathcal{L}}/t_0$ or t' varies with $\mathcal{L}$ and defines the proportional change in the orthogonal thickness. The variation in t' with $\mathcal{L}$ relates to the fold shape and is recorded graphically (Fig. 15C).

With every fold examined in this way two sets of values of t' can be established, one for the sector of the fold between the hinge and one inflection point, and the other for the other half of the fold. For symmetric folds the two graphs coincide, and for asymmetric folds they form two separate graphs.

2. The second parameter is the thickness T of the layer measured in a direction parallel to the axial trace of the fold (Fig. 15B). Geometrical and graphical constructions are similar



A Fundamental types of fold classes. Dip isogons have been drawn at 10° intervals from the lower to the upper surfaces X and Y.



C Position of t_α plots for the various fundamental fold classes.

Fig. 15 : Classification and graphical representation of fold shapes in profile (Ramsay, 1967).

in technique for t' ; but as $t_{\angle} = T_{\angle} \cos \angle$ (Fig. 15B) and $t_o = T_o$, it follows that $t' = T' \cos \angle$ and it is not necessary to measure T .

3. The third parameter is the 'dip isogon' (Fig. 15A; Elliott, 1965, p. 870). Parallel tangents as for $t_{\angle}$, and points where these tangents touch the bounding surfaces are joined to form dip isogons. The inclination of the isogons reflects the variations in the slope of adjacent surfaces (Fig. 15A).

Ramsay (1967, p. 365) divided all possible shapes of fold profiles into three fundamental classes (Fig. 15A), whose definition is based on the comparison of curvature changes on adjacent surfaces in the folded layer.

To analyse the shapes of the F_2 folds, photographs of outcrops were taken looking down the fold axes and the profiles were traced. Profiles were also traced from polished surfaces of hand specimens cut perpendicular to the axes.

Precise location of the axial trace of the fold is necessary for Ramsay's methods; the axial trace is normal to the tangents at the hinges, which have zero inclination. But, particularly in asymmetric folds with rounded hinge areas, the exact position of the hinge points on each surface is not obvious. In such cases the axial trace may be located by joining the centres of circles best fitting the curvatures of the inner and outer arcs at the hinge areas (Fig. 16A). Less time consuming is a trial and error method by use of set squares (Fig. 16B). A mark on the moving set

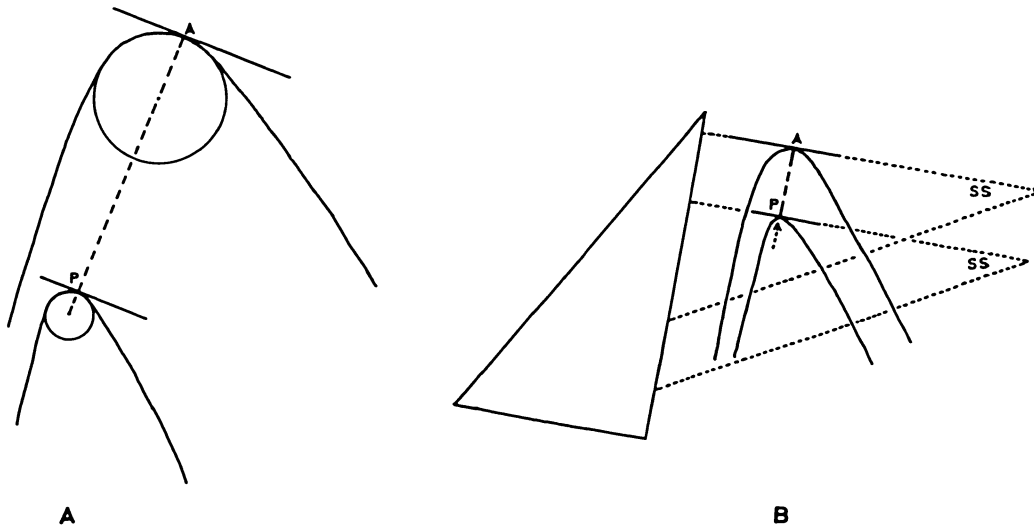


Fig. 16 : Location of axial trace in fold profile.
AP - Axial plane. SS - Set'square.
A. " Best Fit Circle " method.
B. " Set Square " method.

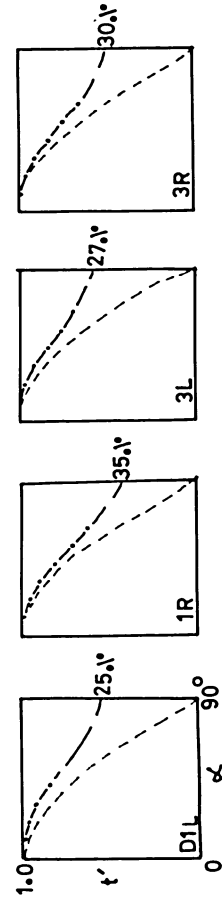
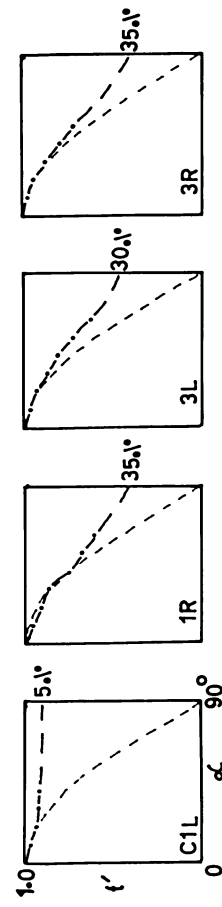
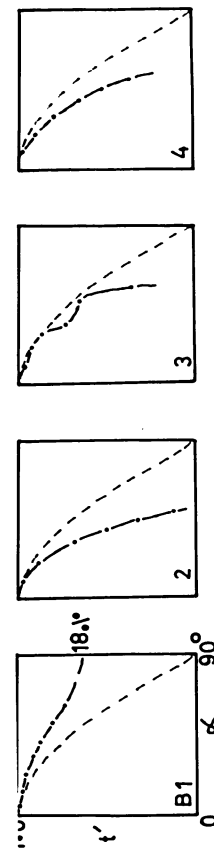
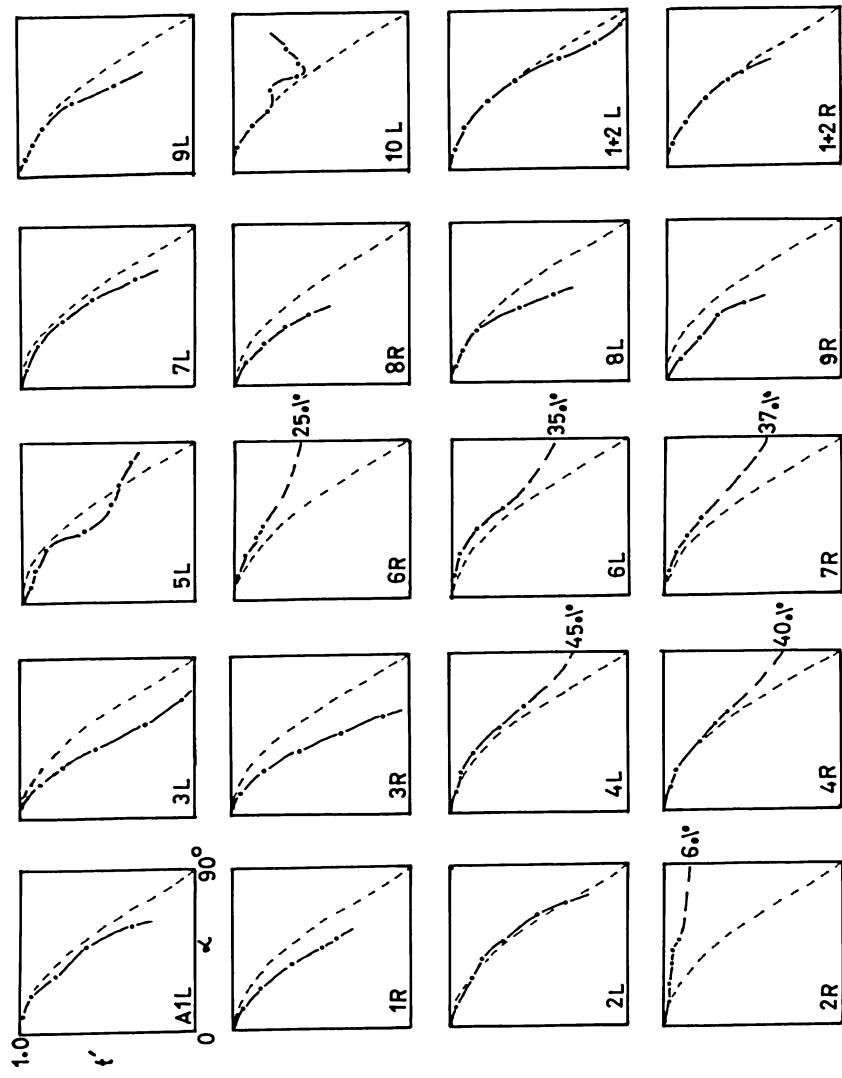
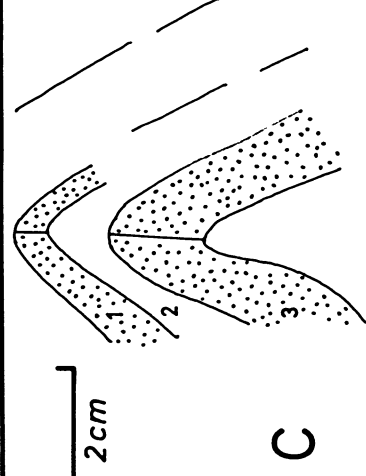
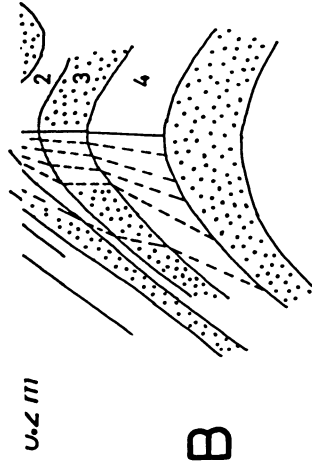
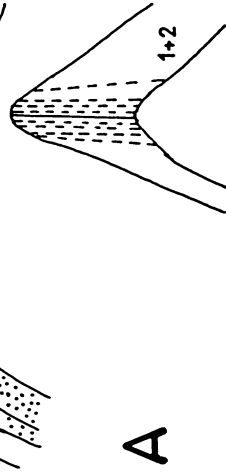
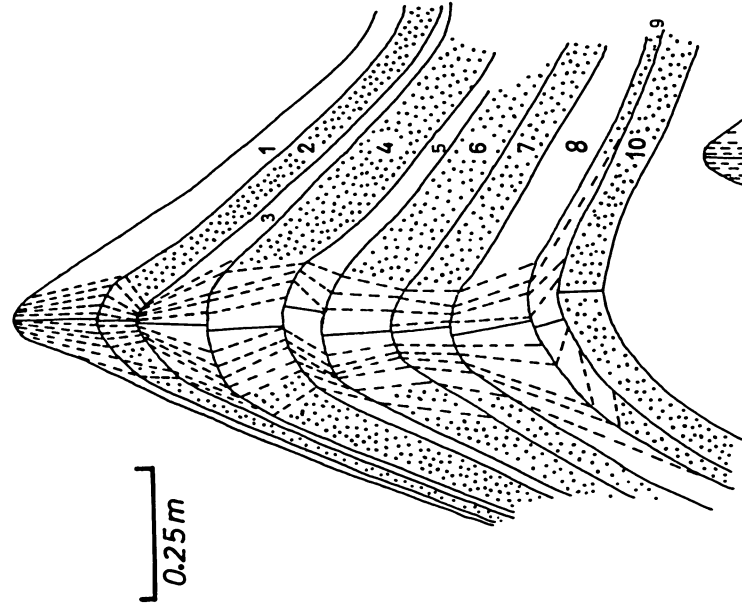
square that touches two or more fold surfaces at tangents moves along the axial trace.

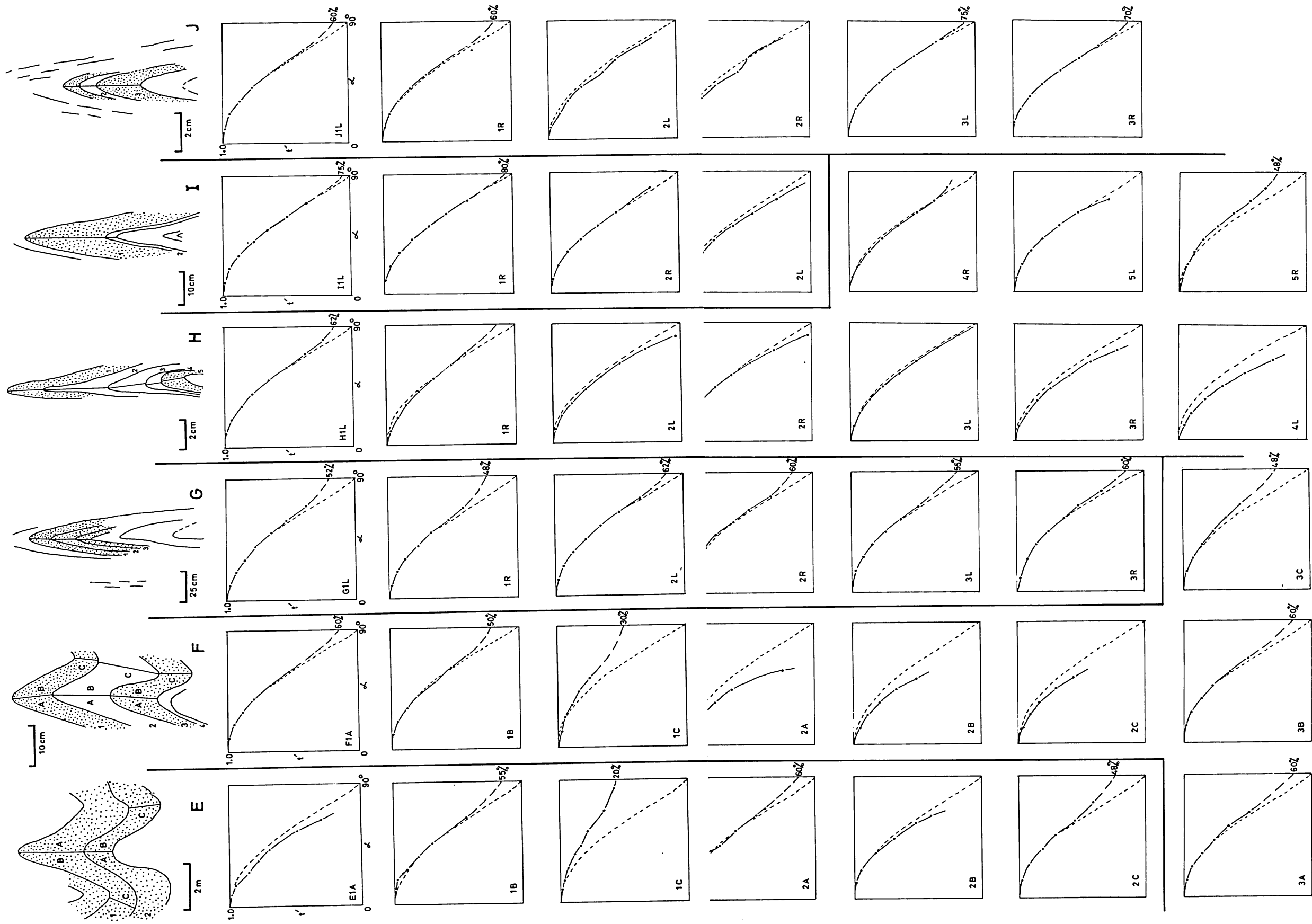
Isogons are drawn with reference to the axial trace of the folded layer as it changes direction from one layer to the next. In this way the mathematical and geometrical relations (Fig. 15B) can be strictly maintained.

Fold Shape and Lithology: The most common profiles of the F_2 folds in the metasediments in the central and southern part of the area are those of class 1C and 3 (Fig. 15A). Shapes closely resembling that of class 2 (similar fold) are common in the northern part, but none fit perfectly this class. Classes 1A and 1B have not been observed.

Lithology strongly controls the shape of the profiles. In a sequence containing layers differing in lithologic character, as for example in the area of marble and calc-silicate layers, the profiles of folds in the calc-silicates are consistently those of class 1C with isogons diverging outward (Fig. 17A, layers 2, 4, 6; Fig. 17B, layers 1, 3) whereas those in marble layers are consistently class 3 with isogons converging outward (Fig. 17A, layers 1, 3; Fig. 17B, layers 2, 4). Similarly, in scapolite amphibole rock (unit 7), profiles in amphibolite-rich layers are those of class 1C (Figs. 17C,D: layers 1, 3) whereas those in marble-scapolite layers belong to class 3 (Figs. 17C,D: layer 2). The calc-silicate and amphibolite layers forming class 1C folds are more competent (they change thickness less readily) than the marble layers forming class 3. It is apparent that

Fig 17: F₂ fold profiles in marble and calc-silicate layers in map units 2 and 3. In folds C and D, amphibolite layers are stippled, whereas in the other folds calc-silicate layers are stippled. t'α graph of each 'half hinge' in each layer is represented in the box by solid line joining the dots, which are actual measurements. Dashed line separates fold class 1C field (above) from fold class 3 field, and represents the t'α graph for an ideal similar fold. Numbering at the lower left hand corner in each box refers in sequence to - fold (A), layer (1), and half hinge (A, B, C, or L, left and R, right). Percentage of flattening is represented at the end of t'α graph.





competence controlled profile shapes.

For folds to affect a succession of layers indefinitely, the geometry requires that either the profile shapes in all layers remain those of class 2, or they change class from layer to layer so that increases in curvature (class 1) inward are compensated by decreases (class 3). In Fig. 17B, profiles in successive layers do not change sufficiently to be compensated, and the folds tend to die out downward. In Fig. 17A, profiles alternate between class 1C and class 3 in layers 1 to 4, but further down the profiles do not change enough from layer to layer between class 1C, class 3 and nearly class 1B, and the folds appear to die out.

If fold profiles of two or more adjacent layers completely 'compensate' each other, the profile of the combined sequence should be that of class 2. This is demonstrated by layers 1 and 2 of Fig. 17A. Though they individually from class 3 and class 1C their combined profile (Fig. 17A: 1 + 2 L, 1 + 2 R) is very close to that of class 2.

The above features, like the influence of competence on the shape and changes in the amplitude of folds that die out at depth, indicate that the F_2 folding involved an element of buckling, that is folding in response to compression along the layering (cf. Ramberg, 1963).

Fold Shape and Flattening Strain: The features of F_2 folds indicate that they formed at least in part by buckling. As the folds are not truly parallel (concentric) folds (class 1B), but rather 1C or 3, there is consistent relative thickening in

the hinge areas and thinning in the limbs.

De Sitter (1958, p. 283) has suggested that there is a value of maximum shortening of 36% of the competent layers by buckling alone. This value is attained when the limbs of parallel folds are isoclinal. He suggested that further compressive strain leads to the modification of the shape of the fold by a process of flattening which involves thinning of the limbs and thickening in the hinge area. Folds formed in this way have been termed 'flattened parallel folds' (Ramsay, 1962). By estimating total shortening and subtracting the maximum 36% due to initial buckling, De Sitter (1964) obtained a value for flattening strain. Ramsay (1962, p. 313) has shown that the flattening process can be initiated well before buckle folds become isoclinal. In such cases the calculations of De Sitter (1958, p. 283) for estimating the amount of flattening assuming maximum buckling cannot be applied.

Since it is not possible, to know when the flattening process started, Ramsay (1962, p. 314; 1967, p. 411) has established a method of estimation of the flattening which is independent of the amount of buckling. In Fig. 15A the orthogonal thickness t is constant throughout the parallel fold. Therefore the ratio t_{α}/t_0 or $t' = 1$ for any value of α , and the graph of $t'\alpha$ in Fig. 15C is horizontal from the origin. Homogeneous flattening of the fold perpendicular to its axial trace modifies the profile and the $t'\alpha$ line is displaced downward to indicate class 1C folds. For these

1C folds Ramsay has calculated t' graphs that would be produced by given ratios $\sqrt{\frac{\lambda_2}{\lambda_1}}$ where λ_1 and λ_2 are the principal 'quadratic extensions' parallel and perpendicular to the axial trace of the fold. The quadratic extension is defined as the square of the length of a line of originally unit length.

Easier to visualize are the curves he uses for percentages of flattening (shortening in strict sense; Fig. 18). The basic assumption is that the strain is homogeneous and that along the fold axis it is zero.

For the calculation of flattening strain, 50 examples were selected from the many F_2 folds classified according to profile shapes. The size of the folds ranged from that of a hand specimen to that of an outcrop. To test the effects of axial orientation the folds were selected from localities so that various rakes were represented (Fig. 13). Layers with viscosities that differ, flatten to different amounts under the same order of compression. To eliminate this variable, folds are selected from the calc-silicate rocks. However, two examples (Fig. 17C,D) from scapolite-amphibole rock (rock unit 7) are also used for comparison.

A few typical and representative profiles are shown in Figs. 17 E to J in addition to Figs. 17A,B. Percentages of flattening in competent layers of each fold are represented on the $t' \propto$ graphs for these folds. Some general features which are discussed below are apparent. For example, there is a general increase in the tightness of a

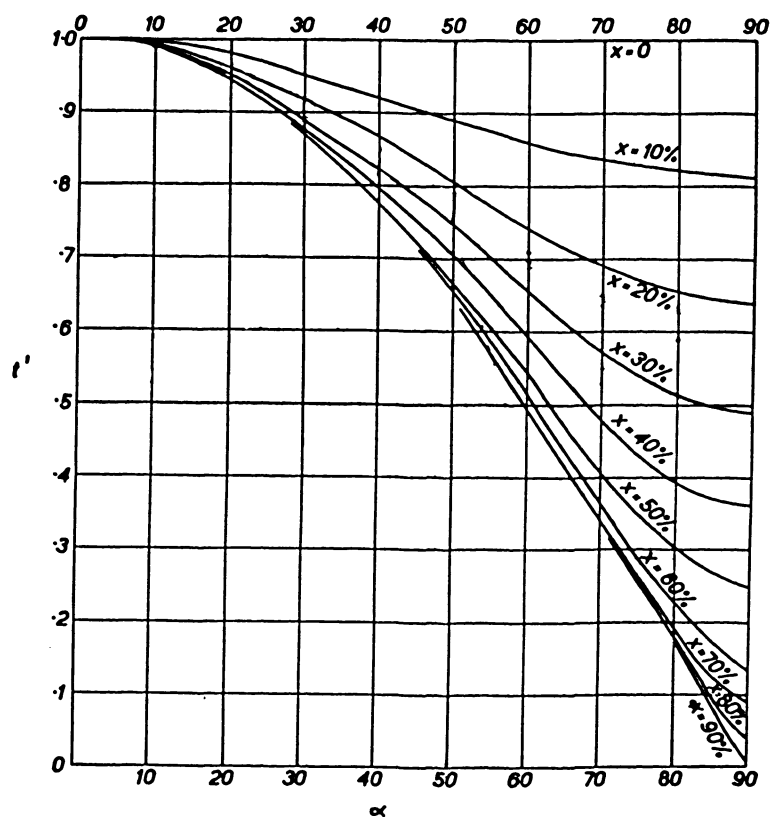


Fig.18 : Graph showing relationship between thickness variations and flattening strain in fold profiles. t' = ratio of thickness of layer at dip α (t_α) to thickness of layer at hinge (t_0). x = percentage flattening (Ramsay, 1962).

fold with increase in the flattening strain in the fold (compare Fig. 17A and Fig. 17G to J). Also, as the flattening strain in the competent layer increases the $t'\mathcal{L}$ graphs for both the competent and incompetent layers, become closer to the $t'\mathcal{L}$ graph for the profile of an ideal similar fold, class 2 (Fig. 17I: compare $t'\mathcal{L}$ graphs of layers 1 and 2)

When the flattening strain in the folds is small ($\approx 30\%$), profiles retain buckling features like fanning fracture cleavage in competent layers (Fig. 7) and a decrease in the amplitude of the fold in successive layers (Figs. 17A,B).

Where the flattening strain in the F_2 folds has moderate values ($\approx 50\%$), buckling features are less obvious (Figs. 17E,F). The fold profiles resemble those of class 2 (similar folds), but as shown in $t'\mathcal{L}$ graphs, shapes of competent layers belong to class 1C and those of incompetent layers to class 3. Folds in both competent and incompetent layers are nearer on the $t'\mathcal{L}$ graphs to class 2 than the folds with lower flattening strain (Figs. 17A,B).

When the flattening strain in the F_2 folds has high values ($> 60\%$), the profiles in both relatively competent and incompetent layers resemble very closely those of class 2 similar folds (Figs. 17G-J). But the $t'\mathcal{L}$ graphs still lie in their respective class 1C (Fig. 17J: 1L, 1R, 3L, 3R) and class 3 (Fig. 17J: 2L, 2R) fields, though very close to that of class 2. That means, that the geometry imposed by a buckling process is almost completely modified

by flattening strain. It is important to emphasize that profiles resemble and approach but are never identical to those of class 2, similar folds.

Though fold profiles may appear gradually transitional in curvature from one surface to the next, local irregularities in shapes are displayed in their $t'\mathcal{L}$ graphs. In particular when the graphs for the competent and incompetent layers are close to that of class 2, any local change in the curvature will make them cross that of class 2; as a result, "anomalous" class 3 (Fig. 17H: 1R) profiles of competent layers and class 1C (Fig. 17I: 2R) profiles of incompetent layers are produced. Such irregularities from the smooth curves of $t'\mathcal{L}$ graphs may, in part, be due to inhomogeneous flattening, which should affect a succession of layers.

Some irregularities in the shape of $t'\mathcal{L}$ graphs could also be due to initial variations in layer thickness. In this case, the irregularity in the $t'\mathcal{L}$ graph should be restricted to a single layer or possibly include the two adjacent layers (Fig. 17A: 5L, 10L; Fig. 17E: 1C, 2A).

A. Carrara (1971, personal communication) has demonstrated that such irregularities in $t'\mathcal{L}$ graphs are very common for folds in sedimentary and low-grade metamorphic rocks in which the bedding surface is found to be irregular. In folds where flattening strain is high, the initial variations in thickness should virtually be eliminated; but in such folds small irregularities still occur (Fig. 17J: 2L, 2R). These could be due to differential slip parallel to the axial surface,

generated in consequence of inhomogeneous flattening (Ramsay, 1962a, p. 321).

Flattening Strain and Axial Orientation: As shown by the examples given, the flattening strain calculated for a single locality varies from fold to fold, from one layer to the next in a single fold, and even from one side to the other of a single hinge. Thus the average flattening strain given for each of the localities shown on Fig. 13 is calculated from the total number of individual measurements for each "half-hinge" in all layers for all folds; 4-12 measurements are averaged, and in each case one standard deviation $\sigma < 8\%$ of flattening.

In Fig. 19 the average flattening strain at each locality is plotted against the rake of F_2 axes. Though the points are scattered, there is an increase in the rake of F_2 axis with the increase in the flattening strain.

Certain features in Fig. 19 are interesting to note. The trend line, does not pass through the origin, signifying that the orientation of F_2 axes in S_2 are not necessarily affected until the percentage of flattening strain in F_2 folds reaches at least 20. Also there is an inverse relationship between the spread in F_2 rake values and the flattening strain in F_2 folds. For example, in F_2 folds displaying 30% flattening, their rakes vary up to 40° ; at 60% flattening the spread of rake values is 28° ; and after 70% flattening strain the spread is only 10° . This means that F_2 axes initially of variable rake in S_2 , at later stages of increased flattening converge closer together. As shown by the trend towards a 90° rake, this convergence is towards the dip direc-

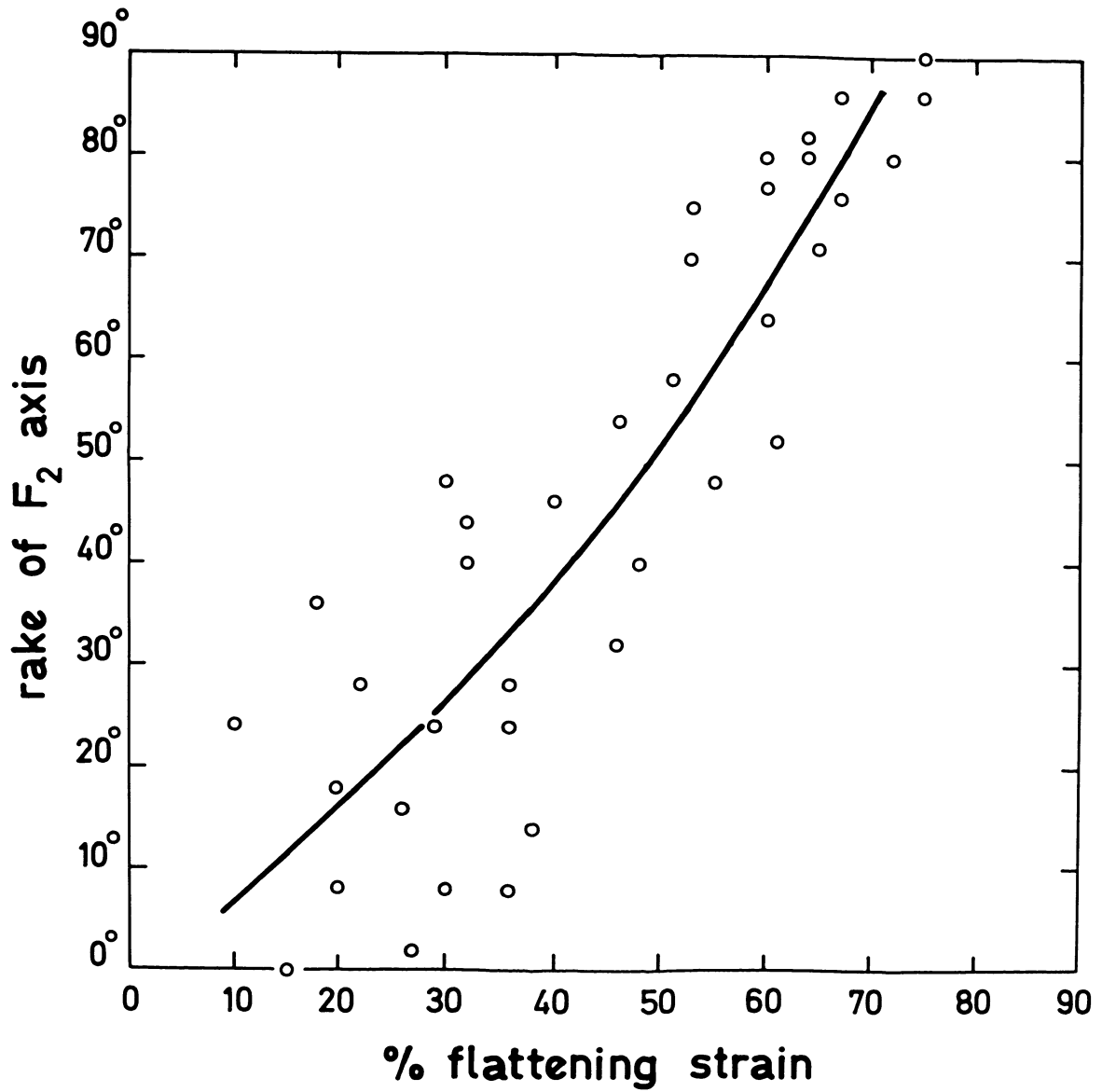


Fig.19 : Relationship between rake of F_2 axis and flattening strain in F_2 folds.

tion of S_2 .

Flinn (1962, p. 397) showed how lines rotate towards the maximum strain extension direction in the strain ellipsoid as a result of three dimensional homogeneous strain during progressive deformation. A similar mechanism by which F_2 axes are rotated is represented in Fig. 20 (see also Best, 1966).

In the general case of deformation, F_2 axes were initially inclined to the Y axis of the strain ellipsoid. As a necessity, these F_2 axes oblique to Y were rotated toward Z as a result of a constrictional type of deformation ($1 < K < \infty$). As F_2 axes were passively rotated, the profiles of F_2 folds changed their shapes to approach those of similar folds (Ramsay, 1967, p. 413). During this stage of deformation the compressive strain in the rock must have been greater than that inferred from the flattening of the F_2 profiles, because when the F_2 hinges approached the Z axes, they were elongated and such strain is not accounted for in the profile measurements.

It is important to test whether or not the local variations in flattening strain are truly less significant than the regional variation.

For example, in Fig. 19 flattening in localities where the rakes of F_2 axes are less than 30° is not immediately different from the flattening strain in localities where the rakes of F_2 axes are between 30° and 60° .

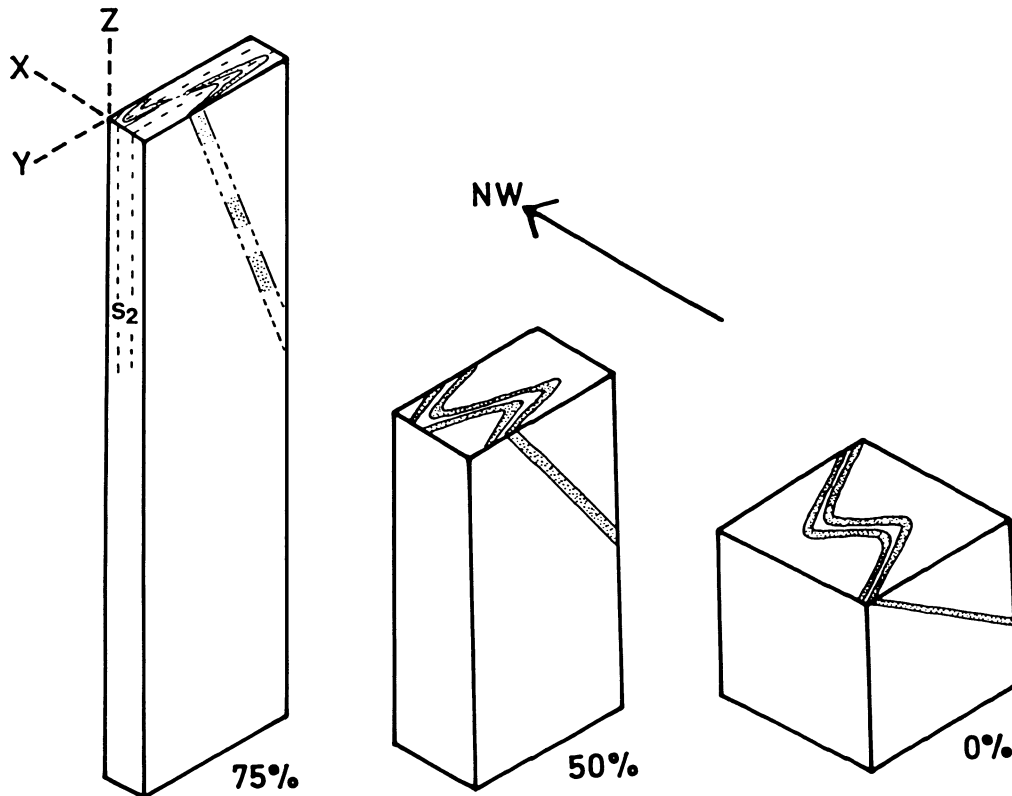


Fig.20 : Mechanism of rotation of F_2 axis. Percentage flattening strain is represented beside the blocks. X ; Y , Z are directions of minimum, intermediate and maximum strain extension.

The flattening strain in the three successive 'rake domains' (in a northwest direction from Green Corners and Upper Hermon) outlined by areas with rakes $0-29^{\circ}$ (A), $30^{\circ}-69^{\circ}$ (B) and $70^{\circ}-90^{\circ}$ (C), are plotted on a histogram in Fig. 21; the mean flattening strain and one standard deviation from the mean for each group are also shown. Though the means are distinctly different, the standard deviations of the domains are close to overlapping. It is considered therefore necessary to test statistically whether in fact the variance of flattening strain between the domains is significantly different and greater than the variance of flattening strain within the domains.

A common method used in such cases is the analysis of variance, which is applied in the present case. As A and C are widely separated and distinctly different (Fig. 21), it is enough to apply analysis of variance for A and B, and B and C only. The results are shown in Table 2.

Table 2 indicates that the variance between the domains is much greater than the one within the domains. Therefore, the three groups of flattening strains in the domains are considered as samples of three populations and the F-test is applied. The F values for both A and B, and B and C are higher than the value (4.6) for the 1% confidence level, so that the conclusion that the samples are from different populations has a probability of 99%. Thus the flattening strain in the rake domains differs significantly. In fact these domains are also "strain domains".

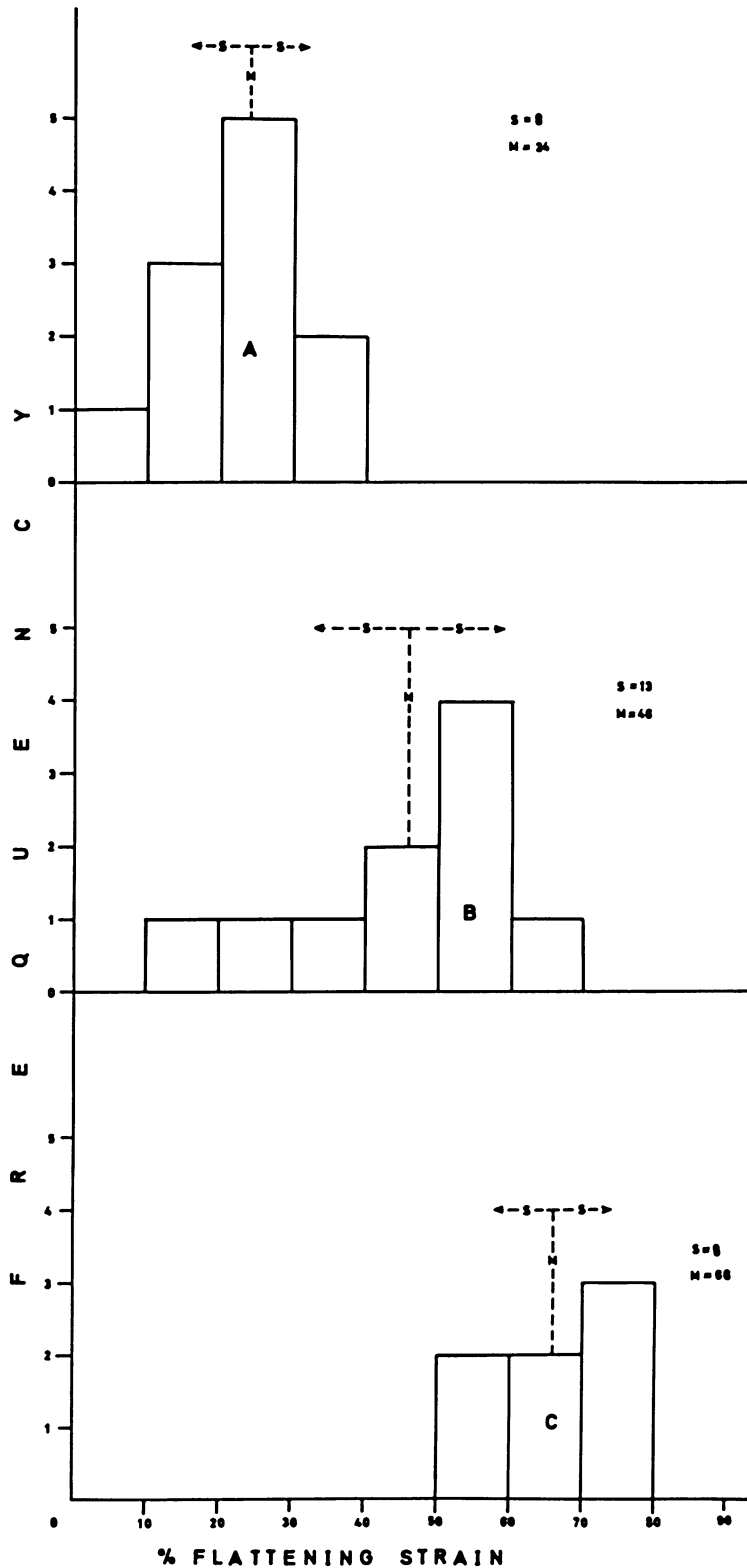


Fig.21 : Frequency distribution of flattening strain in three successive rake domains. A - Domain with rakes 0° to 29°. B - Domain with rakes 30° to 69°. C - Domain with rakes 70° to 90°. M - Mean. S - One standard deviation from the mean.

Table 2: Analysis of variance for flattening strain
in three successive rake domains

	Source of Variance	Sum of Squares	Degrees of Freedom	Variance Estimate	F
Domains A and B	Between Domains	23259.04	1	23259.04	23.89
	Within Domains	18497.23	19	973.53	
Domains B and C	Between Domains	1708.2	1	1708.2	11.85
	Within Domains	2161.3	15	144.08	

The significant increase in flattening strain accompanying the increase in rake from domain A to C is northwest toward the gneiss complex. There also appears to be a similar increase in flattening toward the area of high rakes near the Weslemkoon Pluton (Fig. 13). As suggested when considering the zones of steeply dipping S_2 marginal to the complex and pluton, these bodies may have horizontally constricted the intervening rocks during D_2 deformation and this could account for the flattening and axial steepening by the mechanism shown in Fig. 20.

Fold Shape and Metamorphic Grade: It is still not known whether an increase in the grade of metamorphism in an area necessarily increases the "mobility" of rocks. On the contrary, loss of vapour phases, especially H_2O , from the rocks with increase in temperature might have opposite effects on the mobility of the rocks (Holland and Lambert, 1969).

As a partial test, profiles of F_2 folds in rocks that have undergone a relatively high grade of syn- D_2 metamorphism, shown by the occurrence of sillimanite (Fig. 13), are compared to those of similar lithology in the southern part of the area where sillimanite does not occur.

For example, the characters of profiles in both Fig. 17I and Fig. 17J are very similar - both are nearly isoclinal, both have large amplitude/thickness ratio, and in both the profile shapes of individual layers belong to either class 1C or class 3.

The $t' \propto$ graphs for both folds are close to that of an ideal similar fold (class 2). Therefore, it is suggested from comparison of 9 examples that the F_2 folds are increasingly flattened both northwestward and southeastward, irrespective of the slight change in metamorphic grade.

D_3 Structures

Distribution

The D_3 structures are widely distributed, but they occur in isolated localities and are not penetrative. F_3 folds are best displayed in marble/calc-silicate rocks (unit 3) outcropping 4 km south-southwest of Mayo Lake and near Green Corners (Fig. 5). In feather amphibolitic rocks (unit 2) 2 km south of Green Corners, they are also well exposed on Highway 62.

In the northern gneisses, F_3 are seen only in the nepheline rocks (unit 9), for example those west of the Mallard Lake metagabbro.

Fold Form

F_3 are readily distinguished from F_1 and F_2 folds in that they affect not only S_0 and S_1 but also S_2 surfaces. The folds are gentle to open with rounded hinges (Plate 16A,B).

In marble/calc-silicate rocks, F_3 in the more competent calc-silicate layers are parallel folds (Plate 16B). But variations in layer thickness and irregular profile shapes do occur in some layers (Plate 17A).

F_3 folds affect a restricted number of layers. Away from competent layers, such as the calc-silicates in marbles, the amplitudes decrease and the folds die out. Evidently contrasts in competency are necessary for folding (Ramberg, 1964).

Rarely in nepheline gneisses, F_3 occur as kinks in biotite lying in S_2 gneissosity (Plate 17B).

Axial-Surface Foliation S_3

A non-penetrative axial-surface foliation S_3 is developed in some F_3 folds. In folded nepheline gneisses, S_3 is a crenulation cleavage affecting S_2 (Plate 18A). Amphibole crystals are slightly rotated towards S_3 , but there is no evidence of growth along S_3 .

In marble/calc-silicate rocks, the S_3 foliation is a fracture cleavage along which the rock weathers (Plate 16A, 17A). Often calcite and quartz are recrystallized in the form of veins parallel to S_3 (Plate 18B).

A subhorizontal fracture system is well developed in many outcrops in the marbles, especially those along Highway 62 south of Bancroft where mesoscopic F_3 are absent. It is possible that this fracture system is associated with recumbent F_3 folds.

Axial Lineation L_3

Intersection of S_3 fracture cleavage with folded S_1/S_2 gives rise to a lineation L_3 that is parallel to F_3 axes.

In the nepheline gneisses L_3 is defined by the axes of the crenulations of S_1 and S_2 . These F_3 crenulations are most readily distinguished from F_2 crenulations by the attitude of the axial surfaces S_3 (see next section). Where crenulations are too fine to clearly indicate the orientation of S_3 and where mesoscopic F_2 folds are absent, it is not clear whether they are F_2 or F_3 structures.

Orientation

The F_3 fold axes and L_3 trend northeast to east-northeast; most are subhorizontal and lie on subhorizontal S_3 axial surfaces which cross subvertical S_1 and S_2 (Fig. 22). Some of the S_3 surfaces dip gently to the northwest. In this case S_1 and S_2 are not vertical, but away from F_3 dip steeply to the southeast. As where subhorizontal, the S_3 axial planes thus tend to lie normal to S_1 and S_2 , and the folds are symmetrical. Possibly the F_3 folds are the result of compression parallel to the layers, rather than a simple vertical shortening.

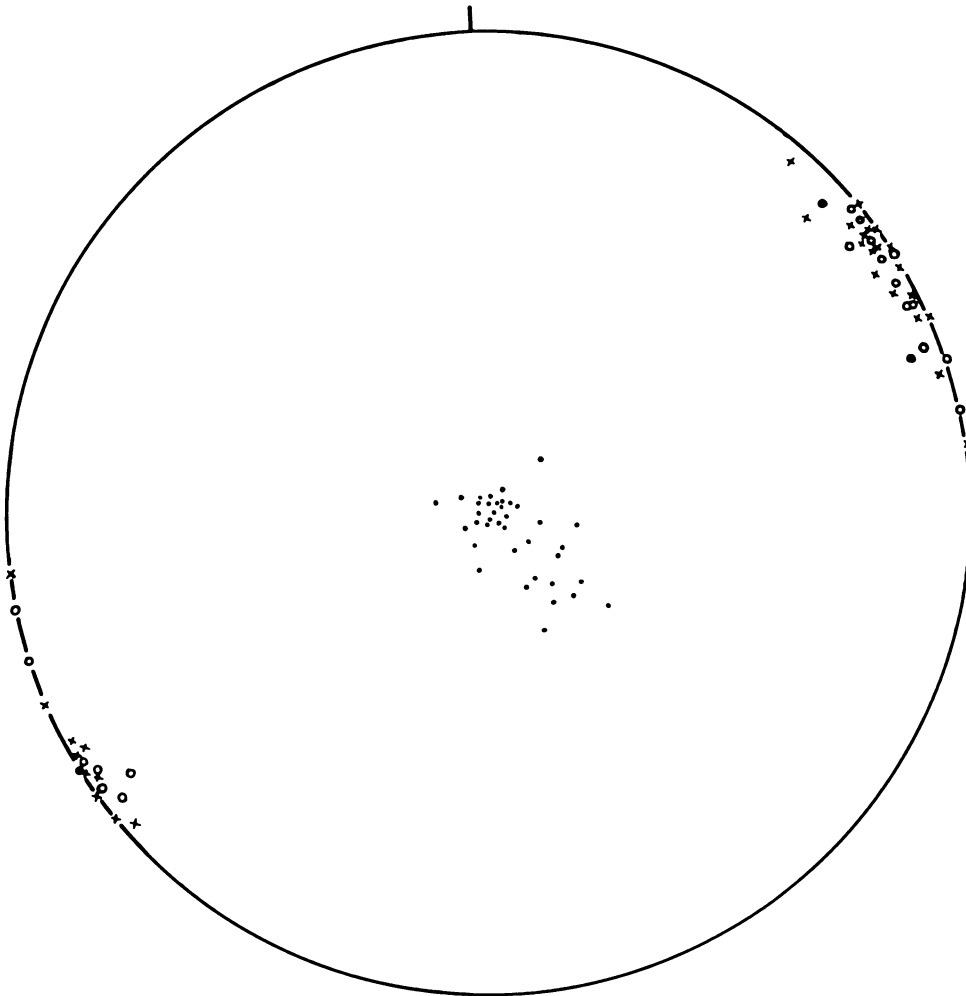


Fig.22 : Orientation of D_3 structures. Circles - F_3 fold axes. Crosses - L_3 lineation. Dots - poles to S_3 foliation.

DEFORMATION AND METAMORPHISM

Introduction

Chronological analysis of crystallization and deformation consists of a study of the relations between the various effects of crystallization (recrystallization of existing minerals or the growth of new minerals) and deformation (strain of existing minerals, generation of new S-surfaces and lineations, and folding of existing S-surface). The textural criteria used in the following description of the time of growth of the various porphyroblasts relative to the three phases of deformation D_1 , D_2 , and D_3 , are similar to those used by various authors, for example, Rast (1958, 1965) and Johnson (1963) in the Scottish Caledonides, Zwart (1960, 1963) in the Hercynian Pyrenees, and Chatterjee (1961) in the Swiss Alps, and by Fyson (1970) in the Shuswap terrain. Following these authors, several basic assumptions, as summarized by Fyson (1970, p. 117), are made in establishing the time of growth:

1. All preferred orientation fabrics (gneissic layering, schistosity, lineation) are formed during deformation, they are synkinematic, and the particular phase of deformation may be recognised by the relationship to folds and associated structures.
2. Unstrained mineral grains aligned in a fabric are either synkinematic and grew during the end of a deformation phase, or post-kinematic and mimic

earlier synkinematic minerals (mimetic crystallization).

3. Grains bent by folds grew earlier than the folding.
4. Grains that deflect an external foliation grew before that foliation formed in its present position. Conversely, if grains cross a foliation without deflection, growth was later than the foliation.
5. Grains, such as large porphyroblasts that include an internal foliation (S internal) grew after that foliation. The inclusion or non-inclusion of structures affecting the external foliation reveals lower or upper limits of times of growth.

A number of difficulties were encountered during the study when trying to establish the relative time of mineral growth.

1. The determination of whether matrix mineral grains are synkinematic for a particular deformation phase or post-kinematic and mimetic. In particular, the distinction between subparallel S_1 and S_2 foliation forming grains is not clear when earlier porphyroblasts with internal fabrics are absent in thin section.
2. Whereas the upper or lower limits of growth time are commonly apparent, rarely are both limits established in the same locality or thin section. Also the later structures, particularly those of D_3 , affect only a limited number of outcrops, so that the relationship to mineral growth elsewhere is not clear.

3. Where the foliation is weakly defined in thin sections, as in the northern gneisses with many equant grains, deflection or non-deflection by porphyroblastic minerals is not obvious.

Over 200 thin sections were made from a collection of more than 800 oriented hand specimens collected from widely distributed outcrops. To understand the relation between the mesoscopic structural features and microscopic textural features, thin sections were made from three mutually perpendicular cut surfaces - one perpendicular to foliation and parallel to lineation, one parallel to both foliation and lineation, and the other perpendicular to both foliation and lineation.

In the descriptions of mineral textures the term "rotation of porphyroblasts" is used here, but at present the mechanics of rotation are not clearly understood, and it seems that there may be either an actual rotation of porphyroblasts (Spry, 1963, p. 214) or a relative rotation, with porphyroblasts remaining stationary and the matrix being flattened around them (Ramsay, 1962, p. 322). The term will therefore be used here in only a relative sense without consideration of the rotation mechanics.

Textural evidence for the time relations of mineral growth to deformational phases is summarized in Fig. 23 which is similar in form to that of Fyson (1970, p. 118, Table 2). In addition, the relative amounts of growth of a mineral are also represented in Fig. 23.

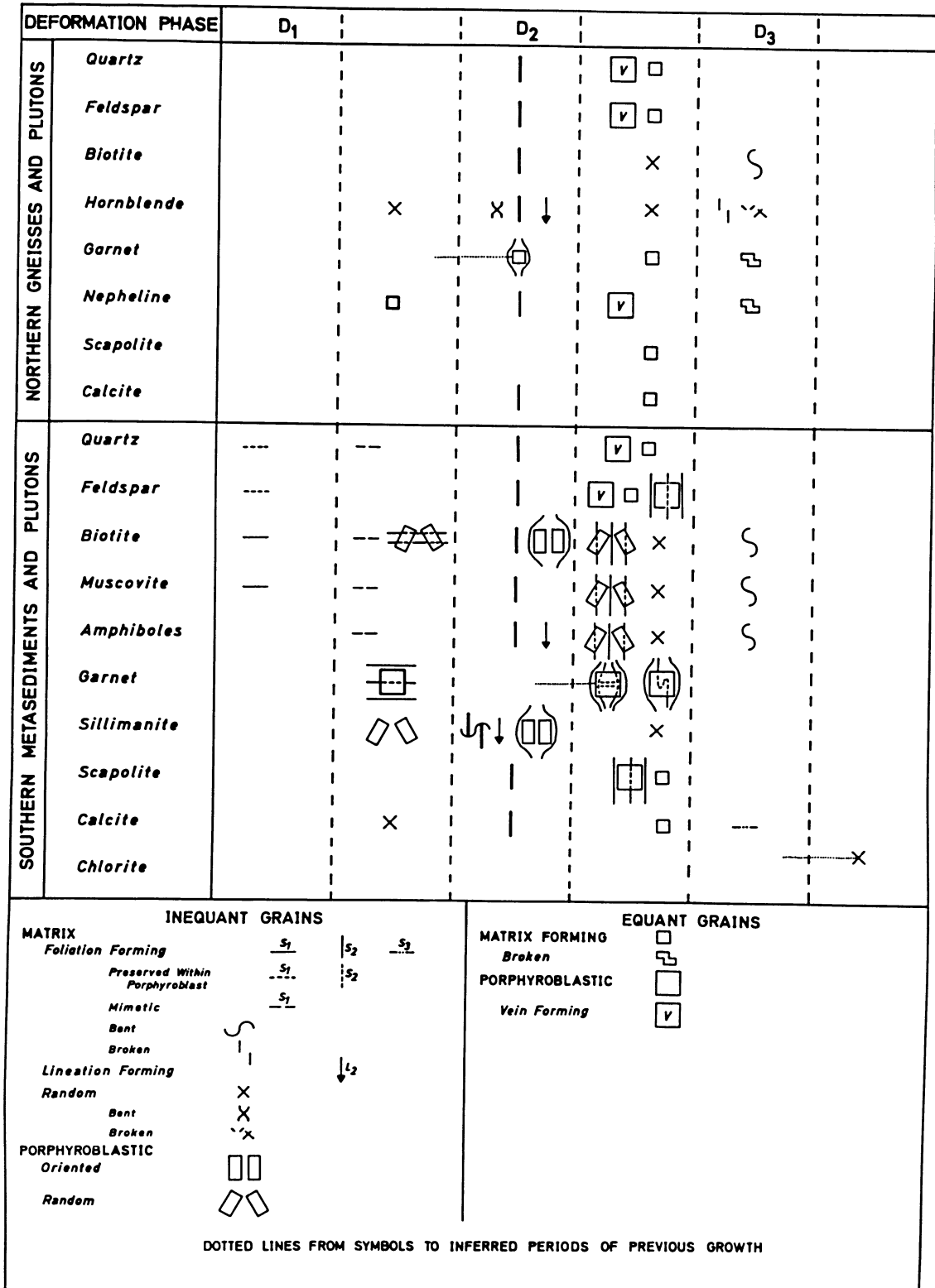


Fig.23 : Time relations between growth of minerals and deformational phases. Symbols outlined in thick lines indicate time of maximum crystal growth.

Crystallization of some quartz, feldspar, and mica took place during D₁ deformational phase, probably under low grade metamorphic conditions. But most of the minerals indicating amphibolite facies metamorphic conditions were crystallized before, during, and after the D₂ deformational phase (Fig. 23). Except for a little chlorite during (?) D₃, no other new minerals were produced either during or after the D₃ phase. The paragenesis of various minerals is elaborated in the following pages.

Mineral Growth

Quartz

The concept that the shape of mineral grains in metamorphic rocks can be used to deduce growth either under stress or non-stress conditions is well known and has been discussed by Voll (1960), Flinn (1965), Kretz (1966) and Rast (1965), together with experimental data by Griggs and others (1960). In the study area there is a marked contrast between the shape and size of quartz grains preserved as inclusions in porphyroblasts and the shape and size of matrix quartz grains.

In semi-pelitic schist (unit 7) and the pyritic schist (unit 5) the quartz grains forming inclusion trails in garnet porphyroblasts differ from the matrix quartz in showing a distinct tendency towards an elongate shape and in being appreciably fine grained (Plate 13; Fig. 24C,D).

Fig.24: Textures indicating growth features of minerals.

A. Pre D_2 biotite porphyroblast enclosing S_1 schistosity in the semi-pelitic rock (unit 7) one kilometer northeast of L'Amable Lake. S_1 is flattened outside during D_2 . B. Post D_2 hornblende porphyroblasts cutting across S_2 in calc-silicate rocks one kilometer south of Green Corners. Quartz grew epitaxially along cleavage in hornblende. C-D. Pre D_2 garnet porphyroblasts overgrowing S_1 quartz fabric. Outer 'rim' enclosing S_2 schistosity (mica and sillimanite alignment) grew during post D_2 times. Locality: One kilometer NNE of Upper Hermon in semi-pelitic schist. E. Pre D_2 biotite porphyroblast partially 'replaced' along its cleavage by sillimanite needles, which are bent 'into' S_2 schistosity during D_2 . Locality: One kilometer NNE of Upper Hermon in semi-pelitic schist.

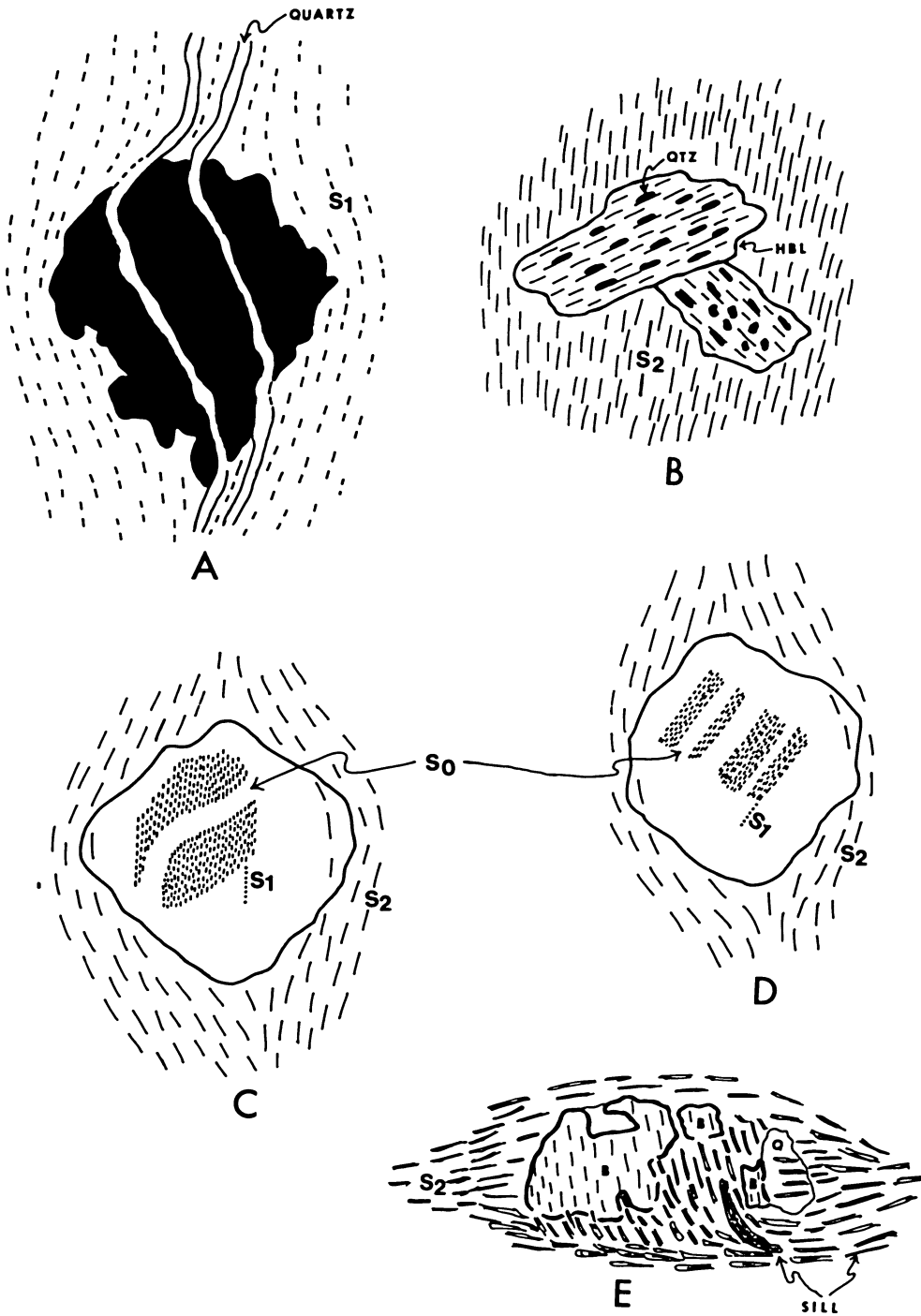


Figure 24

The elongate shape of the included quartz grains suggests formation under stress conditions during D_1 , and the trails of these inclusions represent S_1 foliation (see section on garnets). Some annealing recrystallization could have occurred immediately after D_1 , with the elongate habit being maintained, because the inter-facial energy of quartz against phyllosilicates is such that the former has limited growth in directions other than parallel to the schistosity (Voll, 1960, p. 531-534).

Along with many authors, it is assumed that the fabric of quartz Si (S-internal) within porphyroblasts is a relic of the rock texture existing before and during the growth of the porphyroblast. Some authors (Galwey and Jones, 1962, p. 472) have doubted this assumption. But the arrangement of inclusions is not controlled crystallographically by the enclosing mineral because the orientation of the Si is not at a consistent angle in different crystals of the mineral. Thus, biotite porphyroblasts include Si in which individual elongate quartz grains may be seen intersecting the (001) cleavage at a variety of angles (cf. Kretz, 1966, p. 82).

Whereas elongated quartz grains forming S_1 fabric within the garnet porphyroblasts in the pelitic schist are 0.1 mm in their average maximum dimension, quartz grains forming S_2 foliation outside these porphyroblasts often show a variation in size from 0.2 mm to more than 6.0 mm in their maximum

diameter. The approximate average grain size (maximum dimension) for quartz grains forming S_2 foliation is 1.0 mm. It is clear that the fine-grained quartz fabric of S_1 is preserved in the garnet porphyroblasts, but that outside these porphyroblasts, S_2 coarsened during and following D_2 . However, medium-grained inequidimensional quartz fabric S_1 , which is preserved within post D_1 and pre- D_2 biotite porphyroblasts and which is folded and flattened outside the porphyroblasts during D_2 , suggests some mimetic growth along S_1 before D_2 . The argument against the assumption, that the fine grained nature of quartz fabric in the garnet porphyroblasts may be due to quartz consumption by and during garnet growth, was given earlier (see p. 33).

An approach to equilibrium texture amongst these matrix quartz grains during post D_2 annealing is shown by a marked tendency in some quartz aggregates towards polygonal equidimensional grains with triple junctions approaching 120 degrees. But many grains do have curved and irregular grain boundaries, and undulose extinction and subgrain structure are common, particularly in the larger grains. The extent to which the latter features are a result of continuation of deformation D_2 after crystallization rather than failure to reach equilibrium during recrystallization is unknown.

Within the quartzitic rock (unit 4), quartz grains forming S_2 foliation evidently grew during the D_2 phase. The well-foliated Bessemer granitic pluton has elongated quartz grains lying in the foliation, which is parallel to S_2 in

the metasediments outside the pluton. This suggests growth of the quartz grains during the D_2 phase. Some quartz grains in the Weslemkoon pluton exhibit strain effects of continued D_2 deformation after syn-tectonic (D_2) crystallization of the quartz during the emplacement of the pluton. Quartz in the non-foliated McArthurs Mills granite is equidimensional and grew when the granite intruded the metasediments in post- D_2 times.

Growth of quartz prior to D_2 phase in the northern gneisses is difficult to establish because of the absence of porphyroblasts with an internal fabric and because these very coarse-grained gneisses exhibit only one foliation, S_2 . Most of the quartz in the northern gneisses lies in S_2 gneissosity and grew during D_2 . Additionally, there was some growth of quartz in the quartz veins which cut across S_2 gneissosity.

Feldspars

In the metasediments, feldspar, like quartz grew during and after D_1 and D_2 . The S_1 fabric within the garnet porphyroblasts is mainly composed of quartz, but some elongated plagioclase is also present which must have developed during D_1 . Flattened plagioclase grains showing strain effects and elongated plagioclase grains accentuate S_2 schistosity, which is mainly defined by the elongated quartz grains and alignment of micas. Apparently, this feldspar grew during early D_2 and was later flattened as a result of late D_2 strain. An S_2

schistosity containing aligned sillimanite needles passes through plagioclase grains in the semi-pelitic schist (unit 7); this texture indicates post D_2 growth of the feldspar. Definite textural evidence of feldspar growth during a "static" period between D_1 and D_2 is lacking.

Feldspar in nepheline gneisses, and that in mixed gneisses and gneissic granites where it is closely associated with quartz lies in S_2 gneissosity, and evidently grew during D_2 . Feldspar in the feldspathic veins which cut S_2 grew, as did part of the feldspar in S_2 , after D_2 .

Micas

Biotite in the metasediments, as with quartz and feldspar, probably grew during and after D_1 and D_2 phases of deformation. There is no indication of growth of biotite either during or after the D_3 phase of deformation.

Coarse biotite forming an S_2 fabric is abundant both in the semi-pelitic schist in the north and the pyritic schist in the south. But, unlike quartz, post D_1 coarsening of biotite is difficult to ascertain. This is because of lack of preservation of S_1 biotite fabrics in post- D_1 garnet and biotite porphyroblasts. Possibly S_1 biotite was involved in and consumed during the metamorphic reaction which produced these porphyroblasts.

However, rarely, some biotite in S_1 intersects S_2 foliation, a schistosity or fracture cleavage, to produce mica-edge lineation L_2 (see page 35) and suggests its growth during D_1 .

Additionally, syn- D_1 growth of biotite is also suggested from biotite aligned in and forming S_1 schistosity in the metaconglomerate band. However some of the medium-to coarse-grained biotite now lying in S_1 could have mimetically grown after D_1 and before D_2 .

Although S_2 biotite lies oblique to an S_1 quartz fabric at the hinge areas of D_2 folds, the earlier S_1 biotite is completely obliterated. S_1 and S_2 biotite schistositities away from the D_2 hinges are subparallel and indistinguishable from each other on the basis of shape and size of grains. Therefore, the regional schistosity S_2 containing biotite must be regarded as a 'composite schistosity' ($S_2//S_1$), first developed during D_1 and subsequently reinforced as a result of mimetic growth of biotite parallel to S_1 modified during D_2 deformation.

Biotite porphyroblasts in the semi-pelitic schist and pyritic schist grew during the 'static' periods following D_1 and D_2 . The porphyroblasts of biotite which grew prior to D_2 are deformed by D_2 (Fig. 24A) and are elongated parallel to L_2 (Plate 19A,B). These porphyroblasts show a form orientation, but their cleavages are random except in one specimen. From this specimen two thin sections perpendicular to S_2 were made - one perpendicular to L_2 and one parallel to L_2 . In the thin section perpendicular to L_2 and S_2 , the biotite porphyroblasts do not show significant elongation parallel to S_2 and the cleavage within the porphyroblast is at high angle (60 degrees to 90 degrees) to the S_2 outside.

In the thin section parallel to L_2 , the biotite porphyroblasts are greatly elongated parallel to S_2 and no trace of cleavage is apparent in the porphyroblast i.e. L_2 is parallel to the cleavage in the biotite porphyroblast.

A possible explanation of the fabric in this specimen is that the biotite porphyroblasts grew during late D_1 with their cleavages parallel to S_1 and are later elongated parallel to L_2 during folding of S_1 by D_2 . The high angle between the cleavage in the porphyroblast and S_2 outside may indicate that the specimen is taken from the hinge area of an F_2 fold.

One important feature of the pre- D_2 biotite porphyroblasts is that there is no internal kinking or folding within the porphyroblasts, inspite of the intense D_2 deformation which affected the rocks in which they are present. Apparently, the porphyroblasts were deformed by indirect componental movements according to the Nabarro-Herring principle (Spry, 1969, p. 159). An indication of this is the presence of embayed outlines parallel to S_2 .

Post D_2 biotite porphyroblasts enclose S_2 foliation. These porphyroblasts are distinguished from pre- D_2 biotite porphyroblasts, by the absence of either form or lattice orientation (Plate 20A).

Most of the biotite flakes in the northern gneisses are parallel to and accentuate S_2 gneissosity in the rock and grew during D_2 phase. Some biotite grew randomly across S_2 gneissosity after the D_2 phase. There was no growth of biotite during or after the D_3 phase, and earlier biotite was kinked during D_3 (Plate 17B).

Most of the muscovite in the semi-pelitic schist and pyritic schist is coarse-grained and the muscovite flakes are parallel to S_2 foliation represented by the plane of flattening of quartz and feldspar grains. This coarse-grained muscovite has apparently grown during D_2 deformation. Pre- D_2 growth of muscovite is indicated in a few specimens only. This is expected because most of pre- D_2 muscovite was involved in and consumed by the reaction which produced sillimanite (see Appendix). Within the metaconglomerate (Fig. 2), S_1 schistosity is represented by the plane of flattened pebbles, and the less coarse muscovite in this schistosity grew during D_1 , and some of it, like biotite, probably mimetically recrystallized after D_1 . Porphyroblasts of random muscovites enclose and a few small crystals cross S_2 foliation with sillimanite and evidently grew after D_2 . Pre D_3 muscovite crystals are crinkled during D_3 .

Amphiboles

Syn- D_2 and post D_2 growth of amphiboles is clearly indicated, and early growth during or prior to D_2 is suggested. The amphiboles consist of tremolite in the calc-silicate rocks and hornblende in amphibolites, metagabbro and nepheline gneisses.

Early growth is suggested by the following: In the scapolite-amphibole rock (unit 7) individual amphibole crystals and aggregates lie in S_1 bent around F_2 fold hinges. This suggests either Syn D_1 or post D_1 mimetic growth along S_1 . Some crystals in the fold intrados exhibit shear fractures

oblique to S_2 and probably formed during the folding.

In the Mallard Lake metagabbro (unit 8), most of the amphibole grains are elongated within the S_2 gneissosity, but some grains contain bent translation lamellae (Plate 20B). It is considered that they represent 'relic' grains, which grew before or during early D_2 and were subsequently deformed during D_2 without recrystallizing completely. Parts of these grains were annealed during post D_2 , and so, do not contain these slipbands (Voll, 1960, p. 513). Growth of hornblende prior to D_2 can also be inferred from 'igneous textures' still preserved in the Mallard Lake metagabbro. The time of emplacement of this plutonic body will be discussed in a later chapter (chapter - Discussion and Conclusions).

Syn- D_2 growth of hornblende and tremolite is indicated by the alignment of these minerals in S_2 -schistosity/gneissosity in various rock types both in the southern metasediments and northern gneisses. Lying within S_2 , these minerals are also sometimes aligned parallel to the L_2 lineation.

In the so-called 'feather amphibolite' (unit 2, Fig. 2), poikiloblastic hornblende crystals cut across S_2 foliation (Plate 21A) and must have grown in 'static' conditions after D_2 deformation. As will be mentioned later (p. 95), these random amphiboles have interlocking relations with post- D_2 gneiss and so these two minerals are coeval. In one outcrop, (at Jamieson Lake, Fig. 5), F_3 folds bend these random hornblende, which therefore, must have grown prior to D_3 .

One interesting feature, which is consistent in all post D_2 random hornblende crystals, is that the quartz and the feldspar inclusions in poikiloblastic hornblende crystals are aligned. These inclusions have their longest dimensions parallel to the trace of the $l10$ cleavage in hornblende (Fig. 24B). The oriented inclusions do not represent an older fabric existing at the time of growth of the hornblende crystals. On the contrary, the 'inclusions' and 'host' hornblende grew together, but hornblende imposed its ($l10$) form on the inclusions to attain the low energy requirements for satisfying equilibrium conditions (Kretz, 1966, p. 82).

Garnet

Porphyroblasts of garnet up to 2 cm across, for example in the pyritic schist (unit 5) are widely distributed in most of the metasedimentary rocks. Most of the garnet porphyroblasts in the semi-pelitic schist (unit 7) and the pyritic schist contain an internal fabric, which aids in understanding their times of growth.

The regional schistosity, S_2 (S_e , fabric external) is either deflected or truncated by the garnets in the pelitic schist. The inclusions in garnet are usually quartz and feldspar, but more rarely biotite and sillimanite. Most often the inclusions are elongated and arranged in trails (S_1) which are straight. As described when considering quartz, in all trails the quartz inclusions are appreciably smaller than grains of the same mineral in the matrix. Thus it is assumed

that the garnets grew in a schistose fabric that was finer grained than the present matrix of the rocks. There is no consistent relationship between the orientations of S_i and S_e . The angle between S_i and S_e varies from 0-50 degrees, and as had been mentioned earlier, it is difficult to determine whether rotation of garnets took place or whether the matrix was 'obliquely' flattened around them.

That the straight trails of fine quartz grains in the garnet porphyroblasts represent S_1 foliation is indicated by the following features: Within the garnet porphyroblasts, bands with parallel quartz inclusion trails alternate with bands without these trails (Fig. 24D). Probably the alternating bands represent bedding S_0 . Less frequently, the bands without trails traverse the inclusion trails at high angles (Fig. 24C). Here the garnets appear to have grown over F_1 hinges with S_0 , represented by the quartz-free bands, oblique to S_1 . Thus the first 'stage' of garnet growth was post- D_1 .

Most, if not all, the garnets in the semi-pelitic schist and in the pyritic schist, have internal zoning. The core of the garnet has a fine S_1 fabric preserved as S_i . The outer 'rim' is usually inclusion-free, but sometimes encloses S_2 foliation containing sillimanite which is bent around the garnet (Fig. 24C,D). This indicates that the outer rim had grown or recrystallized after S_2 , defined in part by sillimanite, had been flattened during late D_2 deformation; that is the growth was late or post D_2 . Some unzoned garnets enclose F_2

folds (Plate 21B). The grain size within the folds is greater than that of the S_1 fabric enclosed in pre D_2 garnets and though smaller is comparable to the grain size of S_2 outside. Also, grains in the layers of the enclosed folds are elongated parallel to the layering, and this is taken to represent S_1 parallel to S_0 .

Therefore, it is argued that the folds are F_2 structures and that the garnets enclosing them grew during late or post D_2 times; quartz in the matrix outside the porphyroblast was coarsened later. Post- D_2 growth of garnets is also indicated by the interlocking growth between random hornblende (feather amphibole whose growth is post- D_2) and garnet. In a narrow amphibolitic band southwest of Mayo Lake dendritic garnets (Rast, 1965, p. 94) enclose S_2 foliation, indicating that growth was post- D_2 (Plate 22A).

Garnets in the hornblende-plagioclase-nepheline gneiss (unit 9) do not contain any internal fabric and are wrapped around by S_2 gneissosity. Their exact time of growth is difficult to determine, but was probably either post- D_1 or early D_2 . Post D_2 growth of garnet is suggested from sub-hedral to euhedral garnets which abut against S_2 gneissosity.

Sillimanite

Sillimanite in the semi-pelitic schist (unit 7) is usually closely associated with biotite and needles are typically parallel to the biotite cleavage. Similar growth features are common elsewhere, for example in the Moine schists (Chinner, 1961;

Tozer, 1955). Chinner (1961, p. 323) has proposed that the nucleation of sillimanite on biotite is epitaxial and due to the fact that the arrangement of Al-O octahedra and Si-O tetrahedra in biotite and sillimanite is similar. Although this is possible, it is still not known whether biotite and sillimanite grow simultaneously or sillimanite nucleates later epitaxially on biotite.

In the Bancroft area, sillimanite occurs mainly in two forms - as individual needles and as pods (faserkiesel). The pods consist mostly of the fibrolite variety, but occasionally are composed of small prisms. The pods are about 0.5 to 1.5 cm long and are very prominent in the field; they are particularly well developed at a locality about one mile north-northeast of Upper Hermon (Fig. 2).

Sillimanite pods are morphologically inequidimensional and lie in the S_2 mica foliation, and are elongated parallel to the L_2 lineation. In a specimen cut in three mutually perpendicular directions, the elongation of pods along L_2 is clearly seen (Fig. 8A,B). Within these pods the fibrolitic sillimanite is mostly random, suggesting that growth was during 'static' conditions, either before or after D_2 .

That the fibrolite sillimanite in the pods grew before the end of D_2 deformation is indicated by the following features. Firstly, in some of the pods the needles are folded (by F_2) about axial planes parallel to S_2 to produce, in some cases, a crenulation-type cleavage (Plate 11A). As the pods are random and deformed, their growth was before D_2 .

Replacement of pre-D₂ biotite porphyroblasts by sillimanite is a very common feature, and in some cases nearly complete. The sillimanite needles intergrown along the cleavage planes in the biotite porphyroblasts are drawn out at the edges of the porphyroblasts, and pass into the S₂ foliation which warps around the porphyroblasts (Fig. 24E). Apparently, the needles grew before or during early D₂ and subsequently were deformed during late D₂.

Although sillimanite in the pods grew before or during early D₂, the shape of the pods indicates that they subsequently were flattened in S₂ and elongated parallel to L₂.

Individual sillimanite needles larger than the fibrolite variety lie in S₂ and also are aligned parallel to L₂; the alignment is well displayed at about 2½ km northwest of Detlor (Fig. 2). Obviously, the needles were crystallized during D₂. Some needles are warped around garnet porphyroblasts as a result of late D₂ flattening of the S₂ foliation in which they lie. A few random needles of sillimanite cross S₂ schistosity. This suggests they were crystallized in a 'static' period following D₂ deformation.

Nepheline

Most of the nepheline in the area, except that in nepheline syenite pegmatites and that along S_{2m} mylonitic foliation (described later, p. 111) lies in S₂ gneissosity (unit 9). Though the grains lie with the shortest dimensions across the S₂ gneissosity, the shapes vary, and it is not easy to distinguish

between the 'patch' and 'ribbon' nepheline as developed in the Wolfe Belt area to the northeast (Fig. 1, area C; Appleyard, 1969, p. 695). Here post D_1 metasomatic 'ribbon' nepheline with random c-axes was recrystallized during D_2 deformation to 'patch' nepheline with axes aligned parallel to an L_2 lineation.

To determine if nepheline crystals lying in the S_2 gneissosity, were recrystallized during D_2 , the orientation of c-axes was established, using thin sections cut perpendicular to the mesoscopic lineation. The c-axes measured are either random (Fig. 9C,D), suggesting little recrystallization (cf. Appleyard, 1969) or they tend to lie parallel to the S_2 -gneissosity (Fig. 9A,B), and thus indicate some crystal growth during D_2 . Alignment along L_2 , however, has not been demonstrated.

Nepheline pegmatites, measuring up to 30 m in width and several hundred feet in length, are discontinuous and occur cutting nepheline gneisses or as replacement patches in nepheline gneisses (Hewitt and James, 1956, p. 31). The pegmatite patches are enclosed by and include pods of nepheline gneisses and do not show any internal deformation. It is not certain how these patches formed, but probably by replacement of the nepheline gneisses, as suggested by Hewitt and James (1956, p. 31). If so, the timing of their formation should be post D_2 .

In summary, nepheline formed before D_2 and subsequently part of it was recrystallized during D_2 , and some formed later.

Scapolite

Scapolite occurs in most of the rock types and it is a major constituent of almost all the calc-silicate rocks (unit 2).

Elongated scapolite grains, sometimes with strain-shadows in them, lie in S_2 . These must have grown during early D_2 and subsequently been strained during late D_2 . Calc-silicate boudins lying in S_2 also contain flattened scapolite grains with long axes paralleling S_2 outside the boudin.

Where scapolite occurs in large interconnected patches, such as in the scapolite-amphibole rock (unit 7) west of Upper Hermon and in Mallard Lake metagabbro (Fig. 2), it encloses S_2 schistosity. Obviously, growth of this scapolite was post D_2 . Scapolite of this period has a widespread occurrence in the Wolfe Belt area (Fig. 1, area C); (Appleyard, 1969, p. 707).

Calcite and Chlorite

Calcite probably grew in the metasediments during D_1 , but textures exhibit only features indicating growth before, during and after D_2 . 'Knots' representing D_1 fold hinges in S_1 (Plate 5B) contain equidimensional calcite grains that must have recrystallized after D_1 . Syn- D_2 growth of calcite is shown by D_2 folds with S_2 defined by flattened calcite grains (Plate 22B). Deformation lamellae within these grains indicate late D_2 or D_3 strain.

Calcite within veins crossing S_2 grew during post D_2 times. Growth of calcite in relation to D_3 is not clear from

the limited study. However, some calcite grew during D_3 , in the form of veins parallel to S_3 fracture cleavage, as for example near L'Amable Lake.

Almost all chlorite in the area is present as a retro-grade mineral, derived, for example, from garnet, hornblende and biotite. Most of the prograde metamorphic minerals in the metasediments and gneisses ceased to grow before D_3 , and as the temperature started to fall, the retrograde chlorite may have grown. It is not known exactly when chlorite grew in relation to D_3 . However, chlorite replacing S_2 biotites bent by D_3 , suggests some growth after D_3 .

Summary

Especially in the northern gneisses, the nature of mineral growth during D_1 is difficult to unravel because of the absence of an S_1 fabric, but the fine grained S_1 fabric preserved in the post- D_1 biotite and garnet porphyroblasts in the southern metasediments indicates that these rocks were schistose and fine-grained during D_1 . The lack of minerals in S_1 that are indicative of the amphibolite facies, suggests that the rocks were deformed during D_1 under low-grade metamorphic conditions.

Increase in the grade of metamorphism, from low-grade to amphibolite facies conditions, and a widespread growth of minerals during post- D_1 and pre- D_2 times is indicated by the first appearance of porphyroblasts of biotite, garnet and hornblende (Fig. 23 and 25). The temperature continued to rise in the

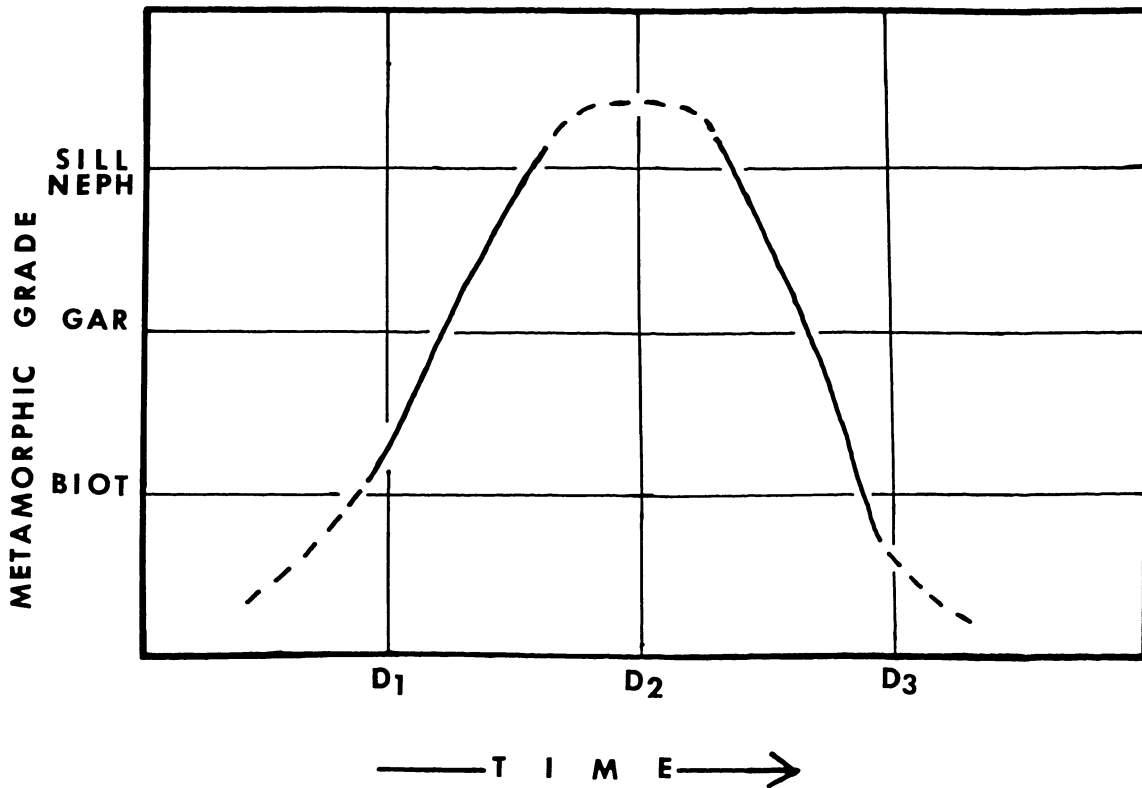


Fig.25 : Suggested time relations between metamorphic grade and deformational phases.

area containing the semi-pelitic schists when random sillimanite 'pods' crystallized before D_2 .

During D_2 the earlier formed sillimanite pods were elongated parallel to L_2 lineation, and outside the pods additional sillimanite grew in the form of needles parallel to L_2 . In the northern gneisses, the earlier metasomatic nepheline was recrystallized during D_2 . Most of the minerals in the northern gneisses lying in S_2 foliation were grown or recrystallized during D_2 .

Regional growth of scapolite both in the northern gneisses and southern metasediments was essentially syn- D_2 , but like in the Wolfe Belt area (Appleyard, 1969, p. 707) took place in post- D_2 times also. That the metamorphic grade was still in the amphibolite facies conditions during post- D_2 times was indicated by the growth of garnet, hornblende (feather variety), and random sillimanite. Retrogressive muscovite and chlorite growth during post- D_2 times is an indication of cooling of the rocks. Possibly the chlorite grew after D_3 deformation.

It is apparent from the above that there was a progressive metamorphism from possible greenschist conditions during D_1 to amphibolite facies conditions during D_2 deformation. Though the high-grade metamorphism (indicated by sillimanite growth) began prior to D_2 and outlasted it, it is essentially coeval with D_2 . There was no strong penetrative deformation after the onset of cooling during post- D_2 times. In the Whetstone Lake area (Fig. 1, area A) 40 km to the south-southwest, Carmichael (1968, p. 9) also records syntectonic (syn- D_2) high-grade metamorphism of the Grenville metasediments.

In the Bancroft area, like in many other metamorphic terrains, as in the Hercynian belt (Zwart, 1969), the Dalradians of Scotland (Johnson, 1963), the Alpine schists of New Zealand (Grindley, 1963) and the Western Alps (Bearth, 1962, 1966), high-grade metamorphism post-dates initial horizontal movements suggested by recumbent folding. In a general way, as shown in Fig. 25 the metamorphic climax and the period of re-folding (D_2) overlap in time (see also Sutton, 1965, p. 38; Zwart, 1969, p. 12).

Radiometric Ages

To determine the ^{minimum} age of formation of the structural fabric in the rocks in the study area, fabric forming minerals have been dated by the Potassium-Argon radiometric method.

Radiometric ages (K-Ar) of 2 minerals forming D_2 fabric and one mineral enclosing D_2 fabric, have been determined (refer to Fig. 2 for the location of samples).

1. Nepheline morphologically aligned along L_2 lineation in S_2 gneissosity in the nepheline-amphibole gneiss (unit 9) yielded an age of 1071 ± 59 million years.

2A. Biotite which is aligned in S_2 schistosity and which is associated with L_2 sillimanite in the semi-pelitic schist (unit 7) yielded an age of 935 ± 50 million years.

2B. Biotite aligned in S_2 gneissosity in the Weslenkoon batholith (sample's locality is about 20 km east of the study area) yielded an age of 928 ± 49 million years.

3. Post-D₂ random hornblende porphyroblast enclosing S₂ schistosity in the 'feather amphibolite' (unit 2) yielded an age of 919 ± 86 million years.

The above ages are compatible with the Grenvillian age of deformation and metamorphism of all the rocks in the study area.

STRUCTURES AND MINERAL GROWTH IN MYLONITIC ROCKS

Introduction

Structures in the mylonites occurring both within and to the west of the study area (Fig. 3), were investigated in an attempt to correlate them with the structures in the country rock. Such information is pertinent in attempting an explanation of the origin and tectonic significance of the mylonites.

Hewitt (1957, p. 36) considered that the mylonite zones mark a fault boundary between the Haliburton Hastings highland rocks and Hastings basin rocks. However, mylonites do not necessarily occur along the boundary, but occur far within the highland rocks (Fig. 3).

Hewitt (1962) maintained that there was synchronous development of the mylonite zones, the fault, and the down-dip lineations (L_2). On the other hand, Best (1966) suggested that the mylonite zones originated late in the metamorphic history, after the formation of gneisses and S_2 . These workers carried out no detailed work on the fabrics in the mylonites.

All the three types of mylonite (p.12-15), the marble mylonite exposed at 8 localities, the syenite mylonite at 2 localities, and the nepheline syenite mylonite at one locality, are complexly deformed on mesoscopic and microscopic scales, giving rise to different generations of foliations and lineations. Investigation of the textural and structural elements reveals the genetic relationships between these foliations and lineations, and forms the basis of the present interpretation, which is that mylonitization occurred early in the deformational history.

As in the surrounding rocks, the mylonitic rocks were affected by three phases of deformation; these are conveniently labelled as D_{1m} , D_{2m} , and D_{3m} and the resultant folds as F_{1m} , F_{2m} , and F_{3m} . In Table 3 the various structures produced during the three phases of deformation in the mylonitic rocks are compared with those in the surrounding rocks.

D_{1m} Structures

Fold Form

Mesoscopic and microscopic F_{1m} folds (Plate 23A,B) bend lithologic layering in marble mylonites. The well-defined layering is composed of alternate bands of marble-rich and extremely fine-grained graphite-rich material. Structures that possibly are stromatolites were observed in some marbles outside the mylonites, and it is possible that the graphite in these well-defined layers represents recrystallized organic matter in sedimentary layers (S_0 bedding?).

In the syenitic mylonites, no F_{1m} are apparent, possibly because of the lack of a pre D_{1m} layering S_0 . However, analogous F_1 folds appear to be absent in the enclosing feldspathic gneiss where S_0 is present, suggesting that mineral growth may have obscured both F_1 and F_{1m} .

The F_{1m} folds are tight to isoclinal, and detached (Plate 23A,B). They have pointed hinges and as for F_1 elsewhere, the profiles resemble those of a "general similar" type (Ramsay, 1962).

Axial-Surface Foliation S_{1m}

An axial-surface foliation S_{1m} is prominent in the marble

Table 3: Comparison of Folds in Mylonitic Rocks with those in Adjacent Rocks

Fold Generation	F _{1m}	F ₁	F _{2m}	F ₂	F _{3m}	F ₃
Scale	Microscopic Mesoscopic	Mesoscopic	Microscopic Mesoscopic	All scales	Microscopic Mesoscopic	Microscopic Mesoscopic
Surface folded	Lithologic layering (bedding S ₀ ?)	Lithologic layering/ bedding, S ₀	Lithologic layering and parallel S _{1m} mylonitic foliation	S ₀ and S ₁ schistosity	S ₀ , S _{1m} and S _{2m} schis- tosity/ transposed mylonitic foliation	S ₀ , S ₁ and S ₂ schistos- ity/gneissosi- ty
Profile	Tight, iso- clinal, attenuated limbs	isoclinal attenuated limbs, deta- ched hinges	close to isoclinal, detached hinges, flat- tened paral- lel	open to isoclinal, detached hinges, flat- tened paral- lel	Gentle to open	Gentle to open
Axial-surface Foliation	S _{1m} mylonitic foliation	S ₁ stratiform foliation	S _{2m} schisto- sity, transpo- sed mylonitic foliation	S ₂ fracture cleavage, crenulation cleavage (only in sillimanite schist), schis- tosity, gneis- sosity	Not clear	S ₃ fracture cleavage, crenulation cleavage (?)
Axial Lineation	L _{1m} colour streaks of lithologic layering on S _{1m} mylonitic foliation	L ₁ axes of lens-like remnant fold hinges, colour streaks of S ₀ on S ₁ , long axis of elon- gated pebbles	L _{2m} colour streaks of S _{1m} mylonitic foliation on S _{2m} schistos- ity/transposed mylonitic foliation	L ₂ intersec- tion of S ₁ and S ₂ , quartz rods, mica- edge, amphi- bole and silli- manite align- ment, seggre- gated trails of nepheline and amphibole	Not clear	Intersection of S ₃ with S ₁ /S ₂ , axes of crenula- tions
Orientation	Not clear	Initial atti- tude unknown probably recumbent, L ₁ variable in S ₁ steeply dipping to southeast	F ₂ axes and parallel L _{2m} variable in S _{2m} modera- tely dipping to southeast	F ₂ axes and parallel L ₂ variable in S ₂ steeply dipping to southeast	Not clear	L ₃ trends northeast in sub- horizontal S ₃

and syenite mylonites. The geometric relationship between S_{1m} and the F_{1m} folds like S_1 and F_1 elsewhere, can be seen in hinge areas where S_{1m} is discordant to lithologic layering (compare Plate 23A with Plate 4A). Elsewhere, the lithologic layering and S_{1m} are parallel.

In marble mylonites S_{1m} is defined by a surface containing elongated, streaked, and lens-shaped bodies of calcite (Plate 23A). These lenses or augen are bounded by streaks of fine-grained graphite. Individual crystals of calcite in the lenses were recrystallized completely during post- D_{1m} metamorphism of the rock. In fact where the streaks of graphite defining the foliation are absent, no foliation is apparent.

Unlike in syenitic rocks elsewhere (Fig. 3) metamorphism later than early folds has not obliterated the early foliation in the syenitic mylonites and the initial cataclastic textures can be seen clearly. The mylonitic foliation S_{1m} is defined by thin laminae of varying composition parallel to which lie long dimensions of ovoidal porphyroblasts of feldspars (Plate 24A) and the long dimensions of quartz grains in the quartz-rich matrix (Plate 24B). The matrix is warped around the porphyroclasts.

Axial Lineation L_{1m}

The intersection of S_{1m} with lithologic layering in the marble mylonites gives a very striking lineation L_{1m} which is axial to F_{1m} folds. L_{1m} is defined by well-defined, discontinuous, and alternating colour streaks or ribbons composed of graphite and calcite. It is clearly seen in outcrop and in hand specimen

(Plate 25A). Since calcite within these ribbons is completely recrystallized after the formation of S_{1m} it is not known whether initially there was any dimensional elongation of the minerals parallel to the ribbons.

In syenite mylonites where no F_{1m} mesoscopic folds are observed the L_{1m} intersection lineation is absent. The porphyroclasts lying in S_{1m} are ovoidal with no preferred direction for the long axes and similarly the long axes of quartz grains in the matrix exhibit no linear alignment. Thus L_{1m} appears to be completely absent.

Orientation

Since S_{1m} and the few L_{1m} lineations are folded during the D_{2m} phase of deformation, initial orientations are not known.

D_{2m} Structures

Fold Form

Both lithologic layering and S_{1m} are folded on mesoscopic and microscopic scales by F_{2m} (Plates 25B). Hence, F_{2m} folds are younger than the F_{1m} folds to which S_{1m} is axial planar. Unlike F_{1m} , which are only recognized in the marble mylonites, F_{2m} are well displayed in all three types of mylonites.

F_{2m} like F_2 in nearby rocks, are close to isoclinal (Plates 25B, 26B; Fig. 26A, B). Most have rounded hinges (Fig. 26A; Plate 26B) but in some the hinge areas are restricted and some pointed (Plate 26A, 27A). In many folds the limbs

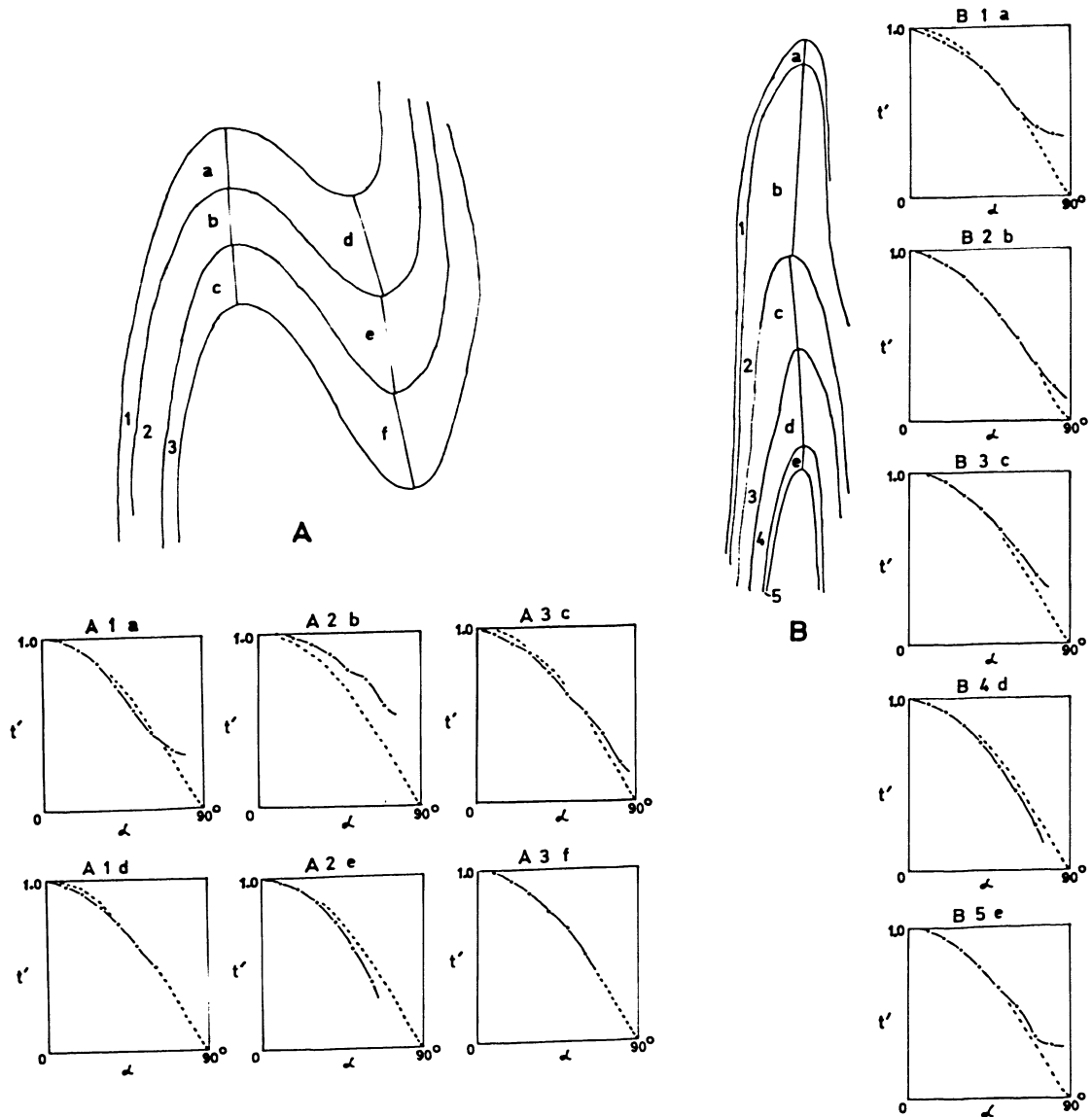


Fig.26 : F_{2m} fold profiles in syenitic mylonite outcropping near Monk road. In the t' - L graphs, the dashed line separates fold class 1C field from fold class 3 field. Numbering above each box refers in sequence to - fold (A), layer (1) and 'half hinge' (a).

are attenuated and the hinges are detached (Plate 27A,B).

Profiles of F_{2m} in syenitic mylonite layers in a matrix of marble mylonite were traced from polished hand specimens and photographs, and thickness variations studied by plotting t' against $\angle$ (as explained on p. 54).

Figure 26 shows typical profiles of F_{2m} in syenitic mylonite layers. $t'\angle$ graphs of the two folds indicate that as with nearby F_2 folds, although close to class 2 the profiles belong either to class 1C or to class 3.

As syenitic layers are more competent than marble, it might be expected, by analogy with F_2 and other folds, that all the syenite mylonite layers would have profiles of class 1C. The presence in some layers of class 3 profiles (Fig. 26: A2e, B4d) may be due to inhomogenities in lithology, such as variations in quartz content, or in the relative proportions of porphyroclasts and matrix.

Also individual layers exhibit combinations of class 1C and class 3 shapes even in a "half-hinge". For example, the $t'\angle$ graphs in Fig. 26: A3c, B1a, indicate that class 3 shapes in the hinge area change to class 1C in the limb area. For some layers, the $t'\angle$ graphs do not form smooth curves with a regular gradient change (Fig. 26: A2b, B5e). As explained earlier (p.66) this could be the result of inhomogeneous flattening of F_{2m} folds.

Axial Surface Foliation S_{2m}

A schistosity S_{2m} that is axial surface to F_{2m} folds is

well developed in layers rich in quartz and calcite (Plate 28A,B). In these layers, such as within sequences of marble and syenitic mylonite S_{2m} is defined by surfaces in which the long dimensions of quartz and calcite grains are aligned. Quartz grains are also crystallographically oriented as can easily be seen by using a gypsium plate to produce a common interference colour.

Within the syenitic mylonites, apart from the quartz-rich layers, the matrix and porphyroclasts, which are largely feldspars, do not exhibit a well developed S_{2m} schistosity. Though some of the porphyroclasts are flattened and their longer dimensions aligned parallel to the axial surfaces of F_{2m} folds (Plate 28A), others either are still dimensionally oriented in bent S_{1m} (Plate 26B) or are random (Plate 26A).

The orientation of c-axes of unstrained nepheline crystals which are flattened and lying in S_{2m} was determined with a universal stage to find out if there is any preferred orientation of c-axes. It is apparent from Fig. 27A,B that the c-axes are close to random. This suggests post- D_2 recrystallization of the nepheline grains (see section on metamorphism).

Axial Lineation L_{2m}

The intersection of S_{1m} with S_{2m} schistosity produces a streaky L_{2m} lineation which is parallel to F_{2m} axes. In marble mylonites L_{2m} on S_{2m} surfaces is represented by ribbons and streaks of carbonaceous and calcitic material (Plates 29A,B). The ribbons represent detached hinges of F_{2m} .

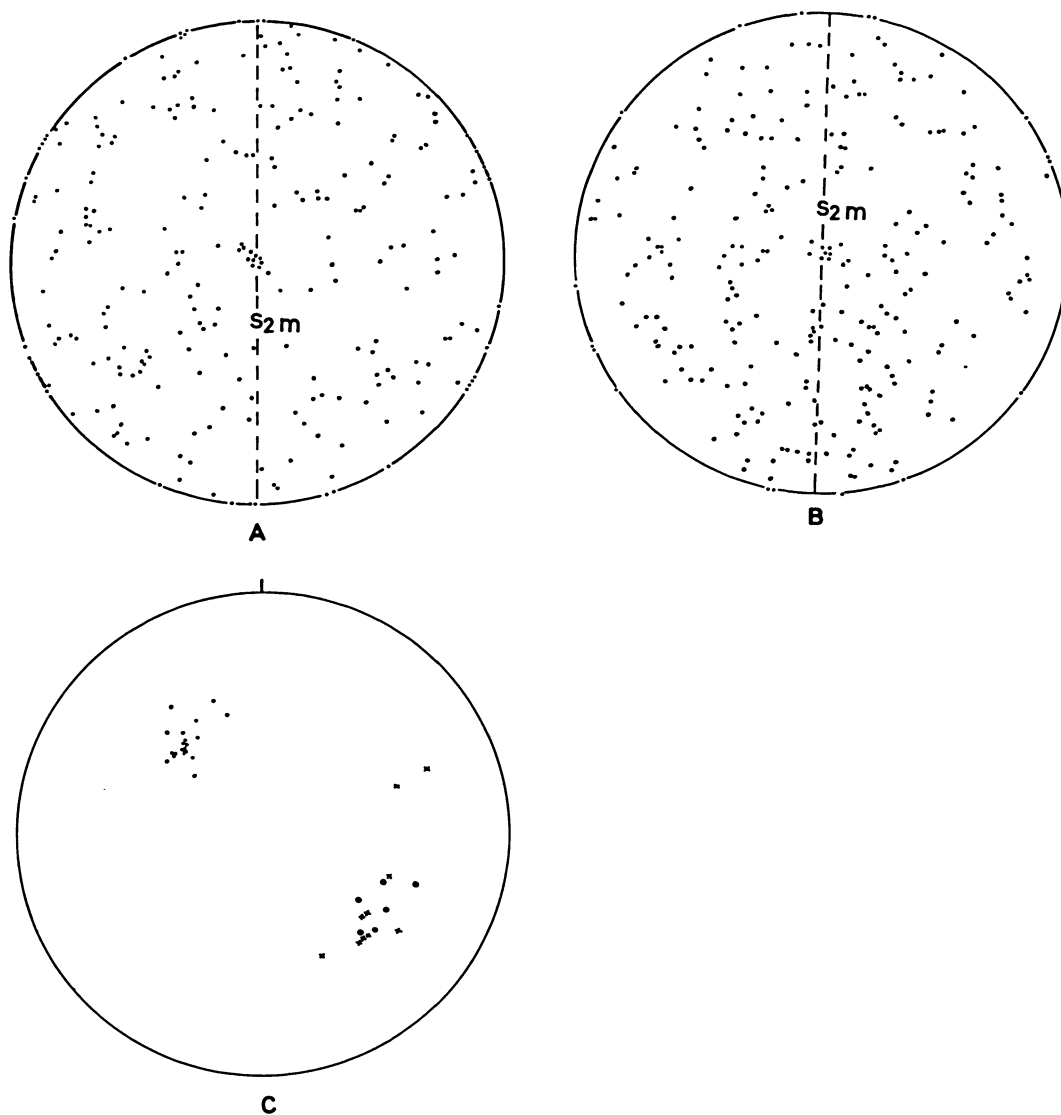


Fig.27 : Orientation of fabric elements in mylonites.

A and B - c-axes of nepheline crystals.

C - Circles are F_{2m} fold axes. Crosses are L_{2m} lineation. Dots are poles to S_{2m} foliation.

L_{1m} and L_{2m} in marble mylonites are closely similar in character (compare Plate 25A with Plate 28A,B). L_{2m} is distinguished in the field by the association with F_{2m} mesoscopic folds that bend both lithologic layering and parallel S_{1m} . Where mesoscopic folds are not present, L_{1m} and L_{2m} cannot easily be distinguished.

Axes of detached F_{2m} hinges in syenite mylonite layers in a matrix of marble mylonite form a distinct L_{2m} lineation. Small 'grooves and ridges' on the F_{2m} hinges probably represent the impingement of the edges of flattened grains of calcite and feldspar in S_{2m} .

F_{2m} fold axes are often highly curved within S_{2m} planes. Even at the scale of the hand specimen, axes of F_{2m} folds vary direction 90° or more. For example the axis of the F_{2m} fold in the syenitic layer in Plate 27B curves within 10 cm up to 100° .

Orientation

S_{2m} has a fairly constant orientation; it strikes northeastward, and dips at moderate angles to the southeast (Fig. 27C) parallel to adjacent S_2 . Though some F_2 axes and L_2 are variable in orientation, most, like L_2 in the gneisses outside the mylonites, plunge down the dip of S_{2m} (fig. 27C), so that the folds are reclined as are F_2 in adjacent rocks.

D_{3m} Structures

D_{3m} structures are not well represented in the mylonitic

rocks. Isoclinal F_{2m} folds and S_{2m} are bent around open folds with subhorizontal axial surfaces, which are similar in character and orientation to F_3 .

Correlation of Structures in Mylonites and Surrounding Rocks

It is apparent from the above description of F_{1m} , F_{2m} and F_{3m} folds and their associated planar and linear elements, that the characters and mutual relationships are similar to that of F_1 , F_2 , and F_3 folds in the rocks surrounding the mylonites (Table 3). However, both sets of structures are not necessarily coeval.

Perhaps the most convincing correlation is that of the style of the F_2 and F_{2m} folds. Both bend an earlier tectonic foliation, are tight to isoclinal and have axes plunging to the southeast down the dip of S_2 or S_{2m} foliation, and S_2 is parallel to S_{2m} . The simplest hypothesis is, therefore, to assume F_2 and F_{2m} formed at about the same time.

S_1 or S_{1m} are the only tectonic foliations present before the D_2 or D_{2m} phase of deformation. Therefore, it appears reasonable to correlate S_1 and S_{1m} , and to consider that they were developed during the same phase of deformation D_1 . The close similarity of F_3 and F_{3m} is apparent from the descriptions given in Table 3.

As the mylonitic foliation S_{1m} is an axial surface foliation to F_{1m} folds it developed at a very early stage in the deformational history of the area. Certainly the mylonite zones were not initiated, as suggested by Hewitt (1962) as a

result of the deformational phase (D_2) which produced the L_2 lineations. However, it is possible that during D_2 there was further granulation and mylonitization (See section on metamorphism).

The relationship between the mylonite zones and the ill-defined fault supposedly separating the northern gneisses and the southern metasediments (Hewitt, 1962) is not clear. Moreover, no displacements across either the "fault" boundary or the nearby mylonite zones are established, though the linear contact between the F_2 antiform at Hermon and the Mallard Lake metagabbro may, as indicated in earlier maps (1955-8 and 1957b by Hewitt of Ontario Department of Mines), represent a fault boundary. Any such faulting, possibly following an earlier weak mylonite zone, must have occurred after the D_2 phase of deformation. Similar relations between early mylonite zones and late fault zones are noted elsewhere (Dalziel and Bailey, 1968; Barber, 1965).

Formation of Mylonitic Foliation

According to some, a mylonitic foliation surface represents a plane of high resolved shear stress (Turner and Verhoogen, 1960, p. 631; Christie, 1963; and Eisbacher, 1970, p. 2017). These authors are of the opinion that the foliation surfaces represent surfaces of strong differential movement. On the other hand, Johnson (1967) and Dalziel and Bailey (1968, p. 559) suggest that mylonitic foliation, like slaty cleavage forms perpendicular to the maximum compressive strain in the rock.

In the marble mylonites of the study area, mylonitic foliation is parallel to the axial surface of the F_{1m} folds in lithologic layering. The mechanism of F_{1m} folding is not known and the initial fabric is lost due to later crystallization. Therefore, the mechanical significance of the relict S_{1m} is not clear.

However, in the syenite mylonites, though no F_{1m} folds are present, the grain fabric displays certain features suggesting that S_{1m} was a plane of flattening. Thus feldspar porphyroclasts, in addition to alignment of long axes along S_{1m} , have extension fractures perpendicular to the foliation. Often the fragments separated by these extension fractures are drawn into the mylonitic foliation. Also, some porphyroclasts exhibit shear fractures along with there is displacement (Plate 30). Though variable, the angles between the shear fractures and the mylonitic foliations are more commonly greater rather than less than 45° (Fig. 28). This relationship is to be expected if the foliation formed at a high angle to maximum compressive strain.

Other features also suggest flattening. The foliation in the matrix curves symmetrically at the edges of the porphyroclasts rather than asymmetrically as would occur if there had been appreciable shear along the foliation plane. Additionally, the porphyroclasts are bent at the edges as if by the flattening around of matrix foliation. This is similar to the modification of the edges of boudins as a result of flattening of the foliation around a boudin (See Ramsay, 1967, Fig. 3-43).

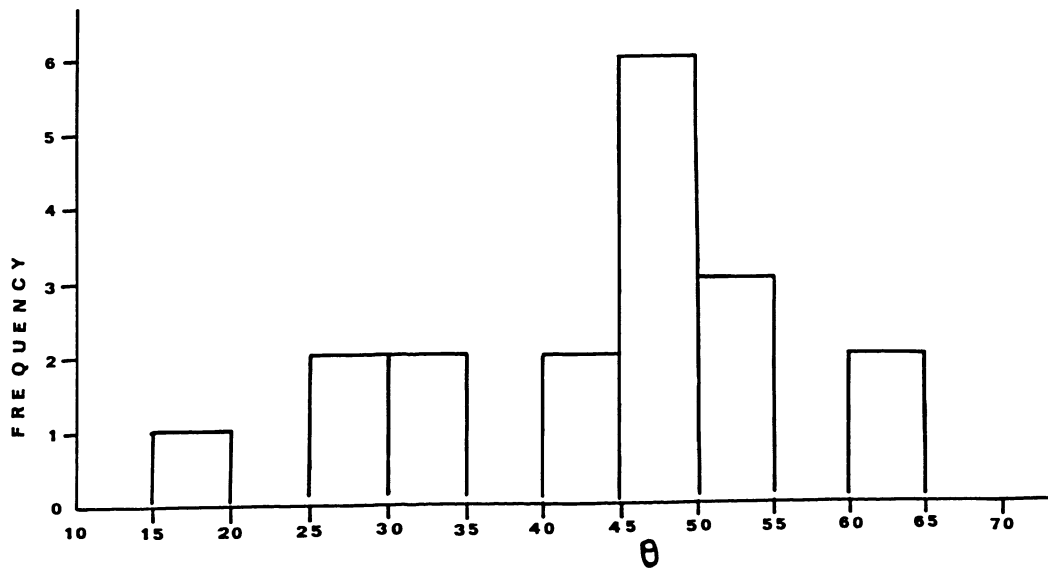


Fig.28 : Histogram of angle (θ) between shear fracture and S_{1m} mylonitic foliation. Data from feldspar porphyroclasts in two samples.

Mineral Growth

Certain features in the mylonites are not clearly understood and require further study. For example, the amphibolite facies grade of regional metamorphism, which has recrystallized almost completely the earlier fabrics in the southern metasediments and the northern gneisses, did not affect the syenitic mylonites to the same degree. In fact, that is why S_1 mylonitic foliation, even in isoclinal and disrupted F_2 folds is still preserved, whereas in the other rocks the S_1 fabric is obscured by later crystal growth during syn- D_2 metamorphism.

Thus, although most of the quartz and the matrix feldspar was recrystallized during syn- D_2 metamorphism, only a few of the larger porphyroclasts of feldspar are 'fresh' and the rest still exhibit original cataclastic textures. On the other hand, some nepheline porphyroclasts have been recrystallized. Also, while calcite has recrystallized completely, many biotite and muscovite crystals are bent and broken.

Probably, recrystallization was inhibited in the mylonites due to higher strain rates and drier conditions, especially in the syenite mylonites. Instead of recrystallization, the higher strain rate might have caused more 'granulation' of the porphyroclasts in S_1 during D_2 phase of deformation. Christie (1963) suggests that in the Moines in the Assynt region, Scotland, primary mylonitic rocks which developed during an early phase of deformation, were later crushed during a later phase of deformation which transformed them into secondary mylonitic rocks.

Such "continuous mylonitization" could explain the 'anomalous' textures in the mylonitic rocks in the present area.

Tectonic Significance of Mylonitic Zones

Mylonitic zones exist in metamorphic rocks of low grade, as in the Moines of Scotland (Barber, 1965), and of high grade, as near the Grenville Front (Dalziel and Bailey, 1968) and in the Peninsula Ranges of southern California (Theodore, 1970). Though they are closely associated with faults, their formation, as in the case of the study area, predates the faults (Moines, Grenville Front), and is very early in the deformational history of the area.

Notable examples of mylonites are found along the margins of orogenic belts. At such places, the basement is supposed to have behaved as a rigid block so that the maximum strain is confined to the discontinuity separating the basement and cover rocks. In the study area, the mylonite zones are also close to a boundary separating contrasting lithologies. But whether this boundary represents a basement - cover surface is not known.

Many mylonite zones occur in areas far away from lithological discontinuities. They could follow earlier fractures existing in underlying basement rocks. Ramsay and Graham (1970, p. 811) suggests that localization of deformation in 'shear zones' is due to initial strain softening of the rock, which later acquires a progressively decreasing yield strength or higher ductility as the deformation progresses. The 'strain softening' may be due to the introduction of mobile components

or to chemical change within the zones. Regardless of the mechanism the mylonite zones represent high strain zones in which deformation under high confining pressure occurred more rapidly than accompanying recrystallization.

The evidence for flattening across the foliation has been cited, but whether or not there was also displacement along the mylonite zones is unclear. Without the evidence of offsets, the problem may never be resolved, because in the case of extreme shear parallel to such zones the flattening plane closely parallels the shear zone (Ramsay and Graham, 1970, Fig. 8) .

DISCUSSION AND CONCLUSIONS

Tectonic Synthesis

From the evidence described a sequence of deformational, metamorphic, and intrusive events can be recognized as having affected the Grenville rocks of the study area. The metamorphic events are considered not to have occurred during distinct episodes separated by episodes without metamorphism, but rather during a single period of prograde regional metamorphism. The relative order and timing of the various events are outlined below.

During D_1 deformation, F_1 isoclinal folds were produced under low-grade metamorphic conditions, as indicated by fine-grained S_1 fabrics preserved in garnet porphyroblasts. F_1 were probably recumbent; this is inferred from the presence of open F_2 folds in S_1 that are upright, near symmetrical, and low plunging, so that before F_2 folding, S_1 appears to have been subhorizontal (p.28). No large-scale F_1 folds are recognized.

Before considering the D_2 structures it is convenient to define the relative timing of the intrusive events. The earliest intrusive recognized is the Mallard Lake metagabbro (cf Lumbers, 1967a,b), which, as shown by the S_2 foliation imprinted on a random fabric, was emplaced before the D_2 phase of deformation. The Bronson and Great Bend granitic plutons in the northern gneisses, which partly cross D_2 structures, and the Bessemer and Weslemkoon granitic plutons

in the southern metasediments, are foliated parallel to S_2 in the country rocks. Pegmatite veins associated with the Weslemkoon granitic pluton are folded gently and displaced only slightly along S_2 (Plate 31A,B), compared with adjacent metasedimentary layering in isoclinal F_2 folds, indicating that the pegmatites experienced only a part of D_2 strain. Hence, the granitic plutons, at least in part, were emplaced after the beginning, but before the end of D_2 deformation. Post D_2 intrusives include the unfoliated McArthurs Mills granite, and a few undeformed nepheline syenite pegmatites and feldspathic veins cutting across S_2 foliation.

The dominant and major structures are the penetrative D_2 structures, which continue without break from the southern metasediments into the northern gneisses. F_2 folds, upright to steeply inclined southeast, dominate the outcrop pattern, and the axial-surface foliation S_2 , a schistosity or gneissic layering, is present throughout the area.

The axial-surfaces of F_2 folds strike relatively constantly northeastward, but axial lineations L_2 vary in direction. Most notable are the variations toward the northern gneiss and granite complex and toward the Weslemkoon pluton, where L_2 changes from shallow or moderate plunges to plunge steeply down the dip of S_2 . Accompanying the steepening in plunge, the folds are increasingly flattened. It is suggested that the presence or emplacement of the relatively rigid plutons imposed an element of constrictional strain in the adjacent rock during the D_2 phase of deformation, and that L_2

rotated towards the down-dip direction of maximum extension. Certainly buckling and flattening were involved during F_2 folding, not merely laminar flow parallel to axial-surfaces as suggested for the main folds elsewhere in Grenville rocks (Wynne-Edwards, 1963).

The highest grade of regional metamorphism was essentially coeval with the D_2 phase of deformation, but it started prior to D_2 (when nepheline in the northern gneisses and sillimanite in the southern metasediments formed) and outlasted D_2 (post D_2 sillimanite formed in the southern metasediments). Because of the widespread growth and recrystallization of minerals during D_2 , earlier fabrics were destroyed, except those in the syenitic mylonites and those preserved in garnet porphyroblasts in the southern metasediments.

The rocks were evidently relatively cool when they were folded by open recumbent F_3 folds during D_3 . Minerals were bent or broken during this phase, and no new minerals grew, except perhaps retrograde chlorite.

Mylonite zones exhibit three phases of structures similar to those in adjacent rocks. ^{assuming the two sets of structures were coeval} Hence, the zones were initiated at an early stage in the deformational history of the area. The presence of potash feldspar and plagioclase that form augen structures in S_{1m} mylonitic foliation could indicate that within mylonite zones temperatures were high during D_1 . Perhaps the zones and the enclosing gneisses were lower in the tectonic pile than the metasediments. Contrary to earlier suggestions, the mylonites do not, at the present level of

erosion, form a structural boundary separating the northern gneiss complex from the southern metasediments, although they are near this boundary and could follow it at depth.

The K-Ar radiogenic dates obtained from aligned minerals indicate that the D₂ structural fabric in the feldspathic and nepheline gneisses formed during the Grenville orogeny. Though it has been speculated that some of the granitic plutons, and the feldspathic and nepheline gneisses near Bancroft represent a remobilized basement (Wynne-Edwards, 1964; also see Maps 1250A, 1251A, and 1256A of the Geological Survey of Canada), isotopic studies of strontium (Krogh and Hurley, 1968) and lead (Zartman, 1969) do not indicate a pre-Grenville sialic history. Thus isotopic and fabric information reveals no evidence for the presence of pre-Grenville rocks in the study area.

Regional Correlations

The major F₂ antiforms at Mayo Lake and Hermon (Fig. 5), continue eastward for 30 km, where the trends curve south-eastward parallel to the edge of the Weslemkoon batholith (Fig. 1). In this area culminations in plunge of major folds are considered by Evans (1964, p. 30) to be due to 'cross folding'. But he did not give any details.

Nepheline gneisses and metasediments 30 km northeast of the study area (Fig. 1, area C) were affected by three discontinuous phases of deformation and metamorphism (Appleyard, 1967, 1969), which are reflected by structures similar

in character and time relations to those described in the present study. Metamorphic reactions increased to those of the amphibolite facies during a static interval between D_1 and D_2 , when the metasediments were metasomatically replaced by nepheline bearing rocks, and acid intrusives were syntectonically emplaced during D_2 . Intrusion of syenitic rocks and migmatization of the rocks during a post- D_2 static interval indicate that the temperatures were still in the field of the amphibolite facies.

The D_2 structures continue beyond the western margin of the study area (Fig. 1, area B; Best, 1966). Although, Best (1966, p. 452) suggested that open folds with subhorizontal axial surfaces in calc-silicate rocks are F_1 structures, they appear to be F_3 folds affecting the S_2 foliation.

Forty kilometers south-southwest (Fig. 1, area A) the structures and inferred tectonic and metamorphic events (Carmichael, 1968, Table 1) are closely similar to those outlined for the Bancroft area: isoclinal F_1 folding under low-grade metamorphic conditions; granitic emplacement after F_1 and before F_2 , which involved buckling accompanied later by flattening, and the development of a penetrative axial-surface schistosity; and finally F_3 crenulations with subhorizontal axial surfaces. Although Carmichael states that the granitic emplacement was before F_2 , his evidence is compatible with intrusion during early F_2 . Unlike at Bancroft, where only small-scale F_1 folds are recognized, the major fold is an F_1 structure. A steep lineation axial to F_1 is

defined by elongated pebbles flattened in S_1 . The elongation is similar to that in the metaconglomerate band near the Weslemkoon pluton (Fig. 2). Carmichael (1968, p. 13) notes F_1 and F_2 folds are coaxial, a relationship that could apply in the Bancroft area.

Thirteen kilometers further to the south-southwest of area A (Fig. 1), Heidecker (1963) observed that towards a syntectonic granite, shallow lineations (stretched pebbles and aligned minerals) gradually steepen to become parallel to a subvertical direction of extension, in much the same way as L_2 steepens towards the granites in the study area. However, there is a basic difference in the kinematic interpretation of the structures, as he considered that the regional lineation is an 'A' lineation that parallels the slip direction, which is also considered as the direction of extension during shear or flow folding of the type visualized by Wynne-Edwards (1963).

Immediately to the south and southeast of the study area, detailed structural studies have not been carried out. However, F_2 folds are present and the metamorphic grade gradually decreases southeastward towards Madoc (Fig. 1) where Lumbers (1968, p. 32; 1969, Fig. 4) delineated andesine, oligoclase and zoned-albite isograds.

Further south (immediately east of area A, Fig. 1) in rocks of the greenschist facies, A. Fabbri (personal communications, 1968) has identified three generations of folds of similar style to those in the study area. Additionally, kinks of a later generation are present.

The lowest grade of metamorphic rocks in the Grenville Province occur near the northern edge of the Paleozoic cover near Madoc. No detailed structural work has yet been published, but in slates it is evident that a cleavage dips steeply to the southeast. The cleavage appears to be parallel to the axial trace of a synform, which plunges northeast and affects only the bedding (Hewitt, 1968, p. 9; Map 2154 of Ontario Department of Mines). On the basis of similar orientation, the slaty cleavage could be correlated with S_2 foliation, F_1 folds being absent. However, interference fold structures in marbles adjacent to the slates (P.H. Thompson, personal communications, 1971) indicate a more complex history, and the correlation of the slaty cleavage with the structures in the study area is not yet resolved.

Vankitasubramanyan (1969) delineates two main generations of folds in metasedimentary and metavolcanic rocks of greenschist to amphibolite grade 25 km east-northeast of Madoc. Small-and large-scale isoclinal recumbent early folds with northwest trending axes were refolded by major upright later folds trending northeast. The general fold style and structural sequence appears to be similar to that of D_1 and D_2 .

In the above areas, nowhere is there evidence of more than one period of regional metamorphism. However, Moore (1967), Sethuraman (1970), and Moore and Thompson (1971) suggest that immediately to the east and northeast of the Elzevir pluton (Fig. 1), a metavolcanic belt trending northwest was first folded under low-grade metamorphic conditions

by F_1 isoclinal folds and then uplifted and denuded. Subsequently, the volcanic rocks were submerged and overlain with angular unconformity by a basal conglomerate followed by calcareous and pelitic rocks, and as in the vicinity of Madoc (Venkitasubramanyan, 1969), during metamorphism to the amphibolite facies, deformation produced northeast trending major upright folds. The grade of metamorphism implies that there must have been burial to great depths after the sedimentation of the upper group of rocks.

Moore (1967, p. 66), and Moore and Thompson (1971, p. 45) state that all plutonic activity (including the emplacement of the Elzevir batholith) took place before deposition of the conglomerate. They consider that the metavolcanic rocks represent a projection northeastward along strike, of the low-grade zone centered north of Madoc.

Although the effects of the two main phases of deformation and associated metamorphism are similar to those of D_1 and D_2 in the study area, here no structural discordance has been observed between the basal metavolcanics (unit 1, Fig. 2) and the metasedimentary units. If the 'structural-metamorphic discordance' (Moore 1967) is present, but obscured by D_2 metamorphism and deformation, the tectonic history of the Grenville rocks in the Bancroft-Madoc region is more complex than now envisaged.

From the above accounts it is apparent that in areas from a wide region the tectonic history of the Grenville rocks has a basically similar pattern. Except, perhaps, for near Madoc, the metasedimentary and metavolcanic rocks were deformed

by two dominant phases of deformation, D_1 and D_2 , that at least in some areas are discontinuous in time. The D_1 phase occurred under low-grade regional metamorphic conditions and produced isoclinal folds, which are inferred to have been recumbent.

Regional metamorphism up to the amphibolite facies, though essentially syn- D_2 , started prior to D_2 and in some places outlasted it. The D_2 phase produced the most prominent folds, most of which are upright or steeply inclined. Syn- D_2 emplacement of numerous granitic plutons influenced local structural patterns.

Intrusion of mafic rocks occurred either during late D_1 or soon after, and there were felsic igneous intrusions before, during and after D_2 .

SUGGESTIONS FOR FUTURE WORK

A detailed investigation of the structures in the Wesslemkoon batholith lying largely to the southeast of the study area, and further consideration of their relationships to those in the metasediments which wrap around the pluton, should throw additional light on the tectonic history of this part of the Grenville Province.

Unlike in the study area, the gneisses immediately to the north and northeast (in Monteagle and Carlow Townships) contain numerous wide bands of schistose semi-pelitic metasediments. Investigation of the metamorphic mineral assemblages should reveal whether or not the metamorphic grade continues to increase northward as near Bancroft; and structural studies in the metasediments, gneisses, and plutons should help in understanding the age and evolution of the gneiss complex.

The intriguing fact that rocks thought to be the lowest part of the Grenville Group, that is the Tudor metavolcanics and associated slates, experienced the lowest grade of metamorphism (Lumbers, 1967b, p. 19), should be investigated by detailed structural and stratigraphic studies (especially of provenience and facies changes) immediately to the west and southwest of the Elzevir pluton. If the 'stratigraphic-structural discordance' indicated by Moore (1967) and others as occurring to the east of the pluton also exists to the west, it could suggest that the pluton and the surrounding volcanics might have formed a topographic high during sedimenta-

tion and acted as a source for the thick sequence of metasediments. Granitic pebbles in the conglomerate forming the base of the sediments supports this speculation. The stratigraphically higher sediments, being buried to a much deeper level away from the pluton than the volcanics and slates adjacent to it, underwent higher grade metamorphism.

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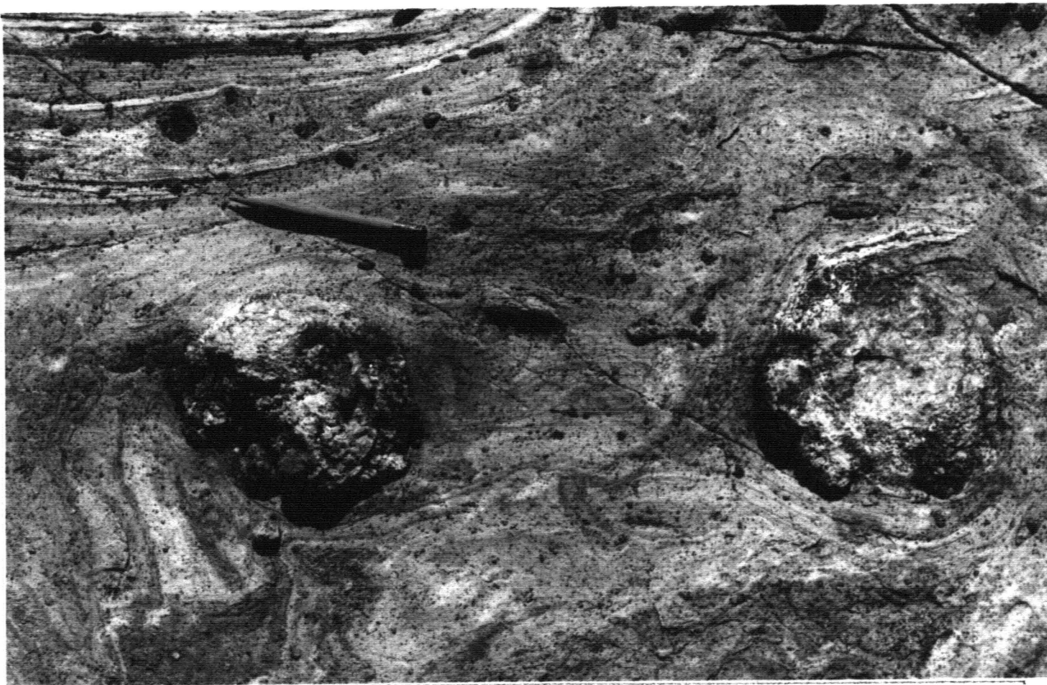
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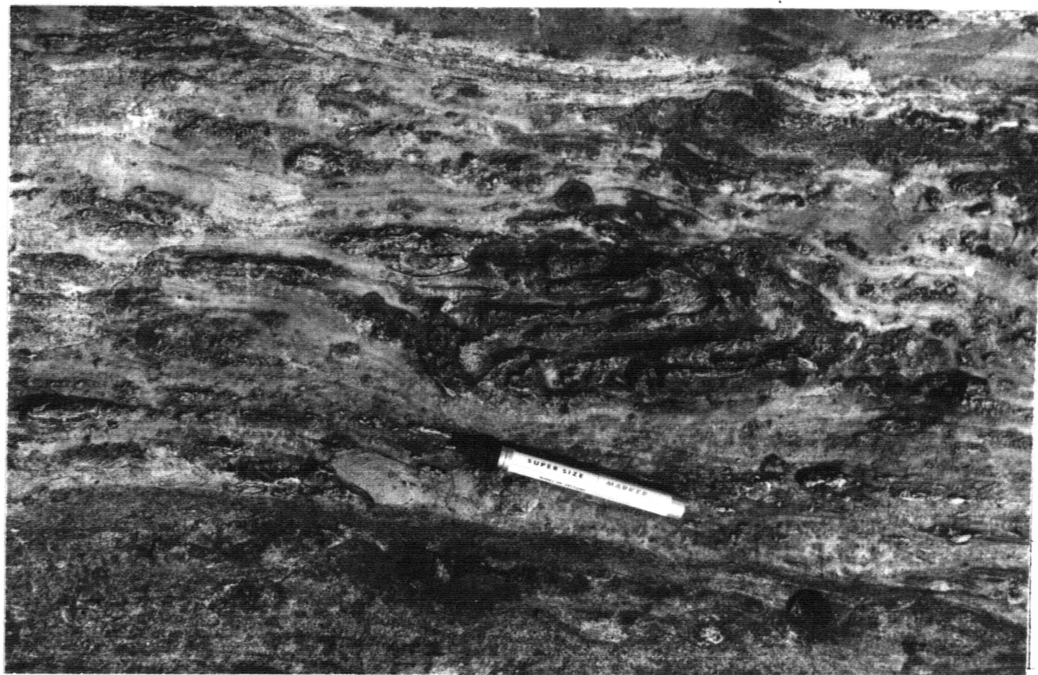


A

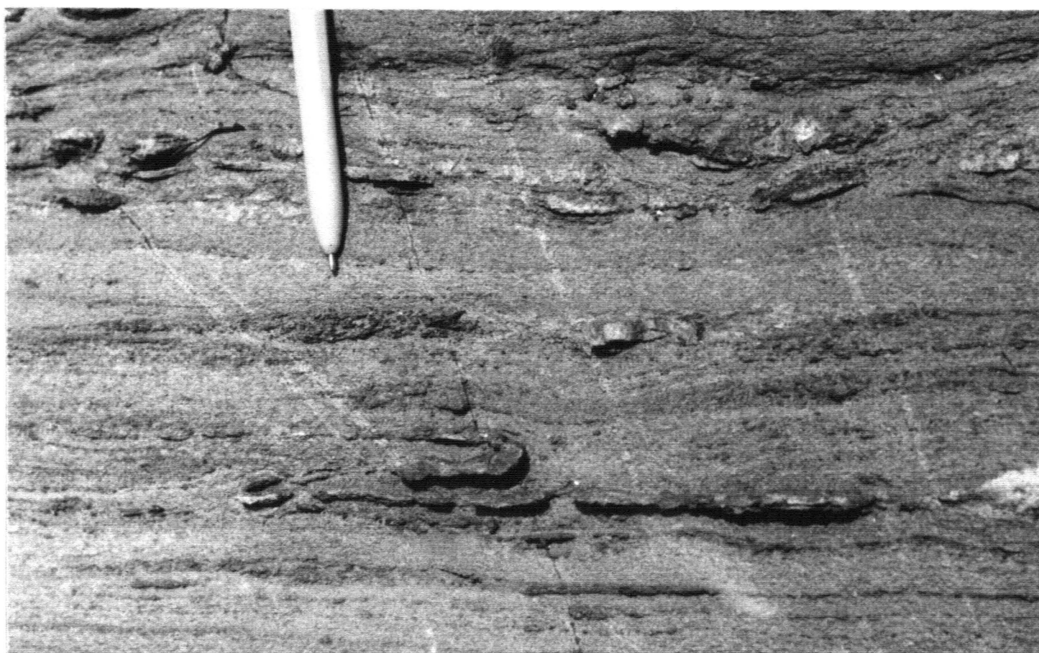


B

- A. Quartzitic, syenitic and granitic porphyroclasts in marble mylonite. Locality : 3km east of Bancroft on Highway 500.
- B. Syenitic and granitic rounded porphyroclasts in transposed mylonitic foliation S₂ in marble mylonite. Locality : 2.5 kilometers from Highway 28, on Monk road.

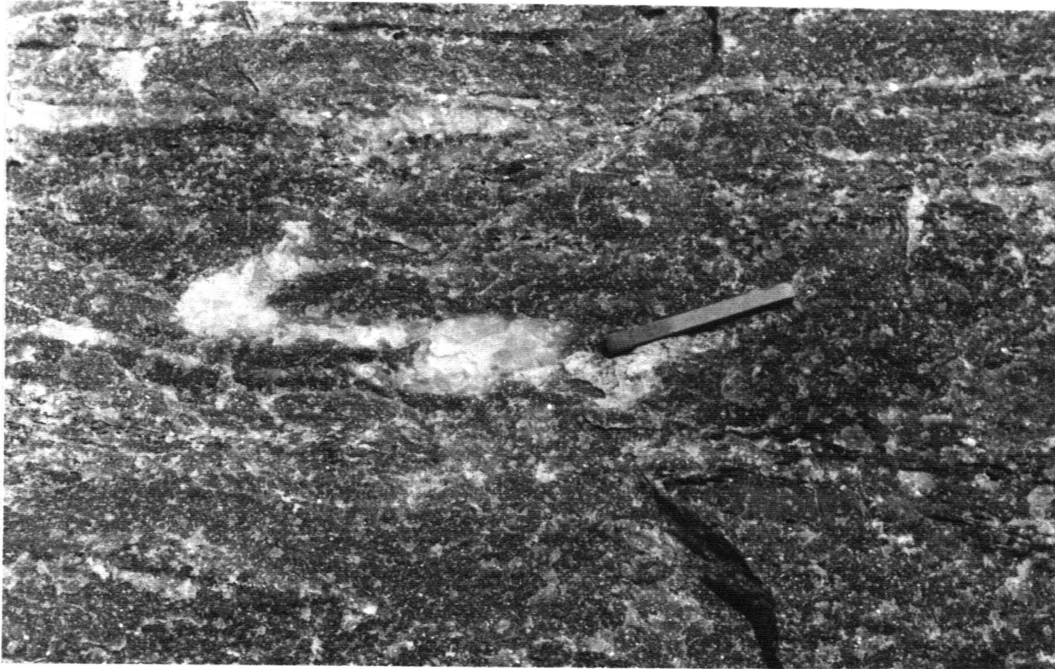


A

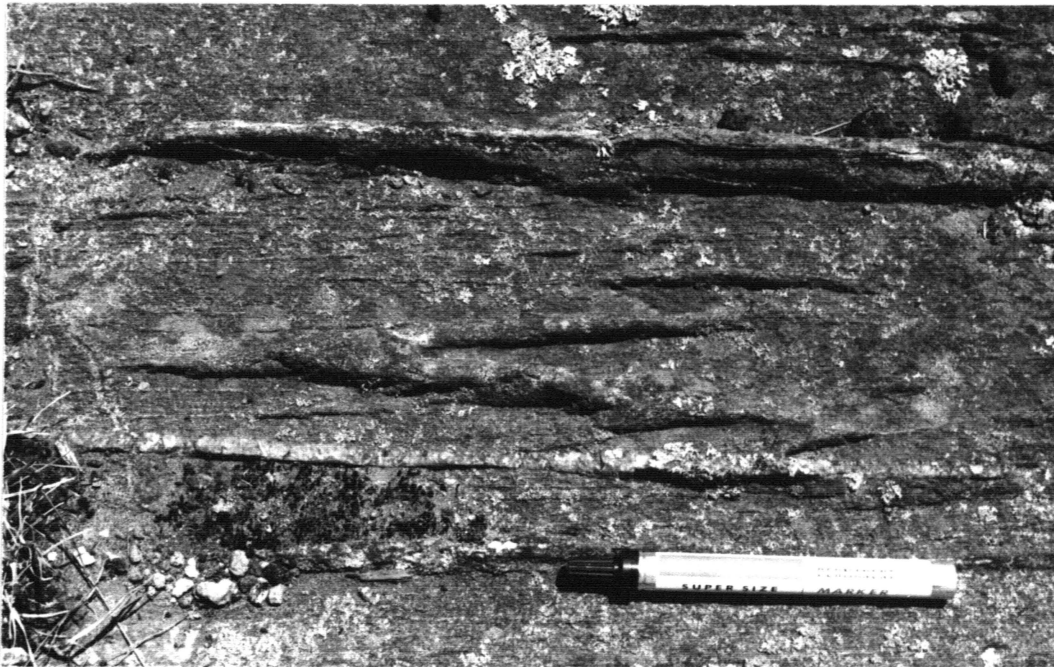


B

- A. F_1 fold in calc-silicate layers within marble; an axial-surface schistosity S_1 is present in marble. Locality : 2km southwest of Detlor.
- B. F_1 fold in calc-silicate layer within marble; an axial-surface schistosity S_1 is present in marble. No foliation within folded layer. Locality : 6km southwest of Mayo Lake.



A



B

- A. F_1 fold in calc-silicate layer within marble (rock Unit 3).
Locality : One kilometer south of Green Corners on
Highway 62.
- B. F_1 fold in quartzitic layer within quartz-feldspar-biotite
rock (rock Unit 5). Locality : 2km west southwest of
Mayo Lake.

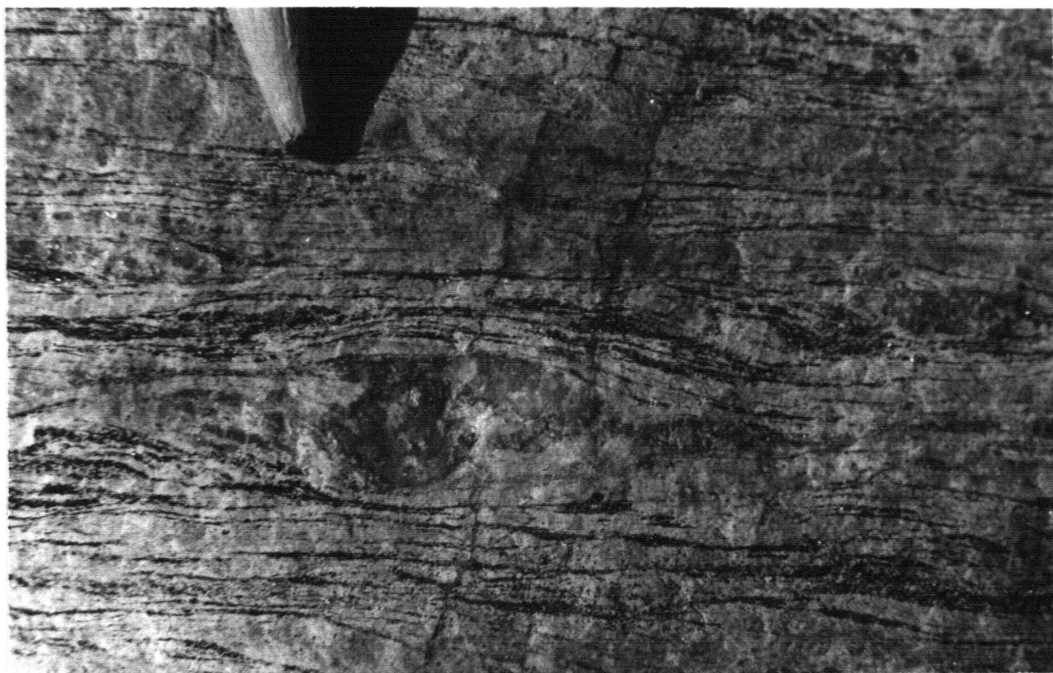


A.

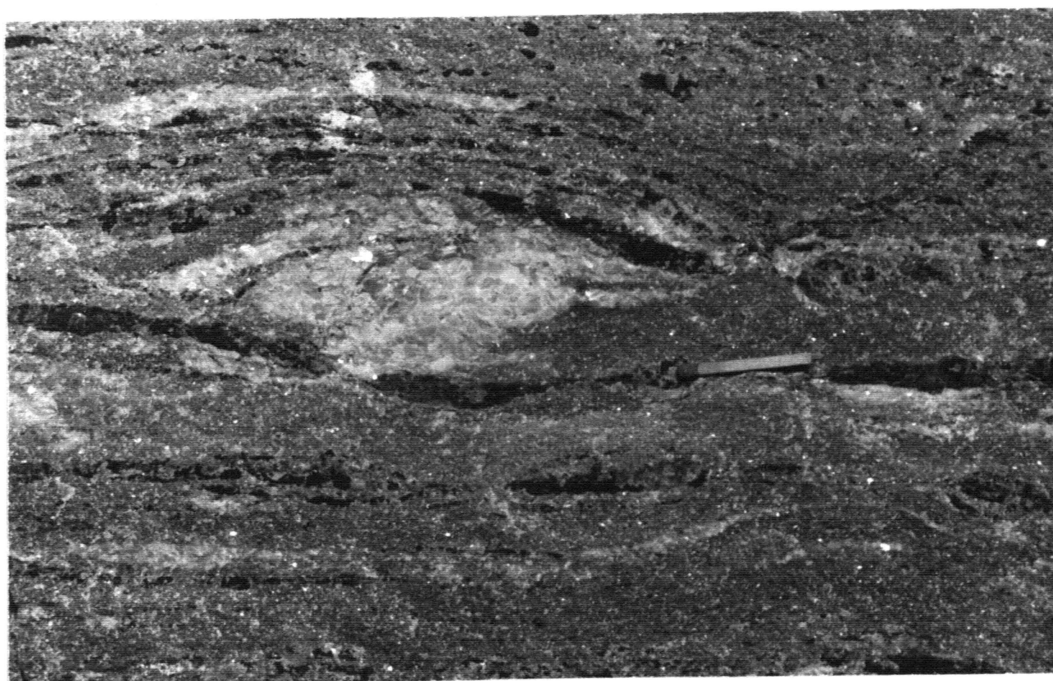


B

- A. F_1 fold in quartz layer in quartzitic rock (rock Unit 4); axial-surface schistosity (mica alignment) S_1 is present. Locality : 1.5 km southwest of Mayo Lake.
- B. Intersection of bedding (pebble free band) S_0 and S_1 schistosity in metaconglomerate. Locality : 0.5 km south of northeast corner of Mayo Lake.



A



B

- A. F_1 fold hinge in the form of 'bulge' in quartz layer in quartzitic rock (rock Unit 4). Locality : 1.5 km southwest of Mayo Lake.
- B. F_1 fold hinge in the form of 'bulge' in calc-silicate layer in marble (rock Unit 3). Locality : 6km southwest of Mayo Lake.

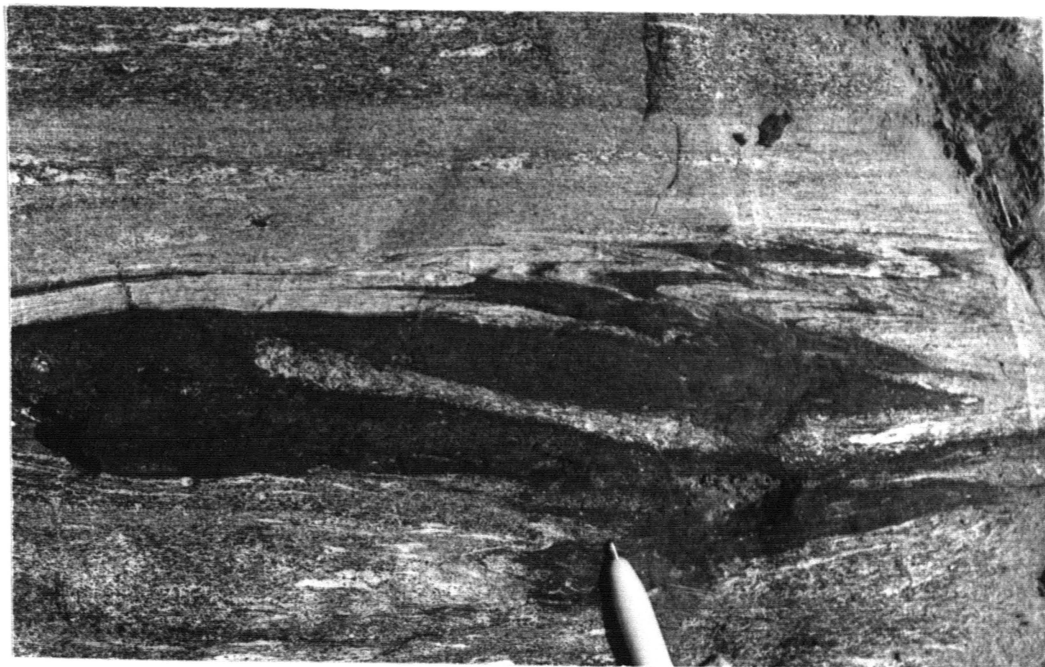


Long axis of pebble defines L_1 (parallel to hammer handle) on S_1 schistosity.
Locality : 0.5 km south of northeast corner of Mayo Lake.

- A. F_2 folds in quartz-mica layer within marble. S_1 foliation in the layer is bent. Locality : 2km south of Bessemer granite.
- B. F_2 fold in amphibolite layer within scapolite-calcite rock (rock Unit 7). Locality : 0.5 km west of Upper Hermon.

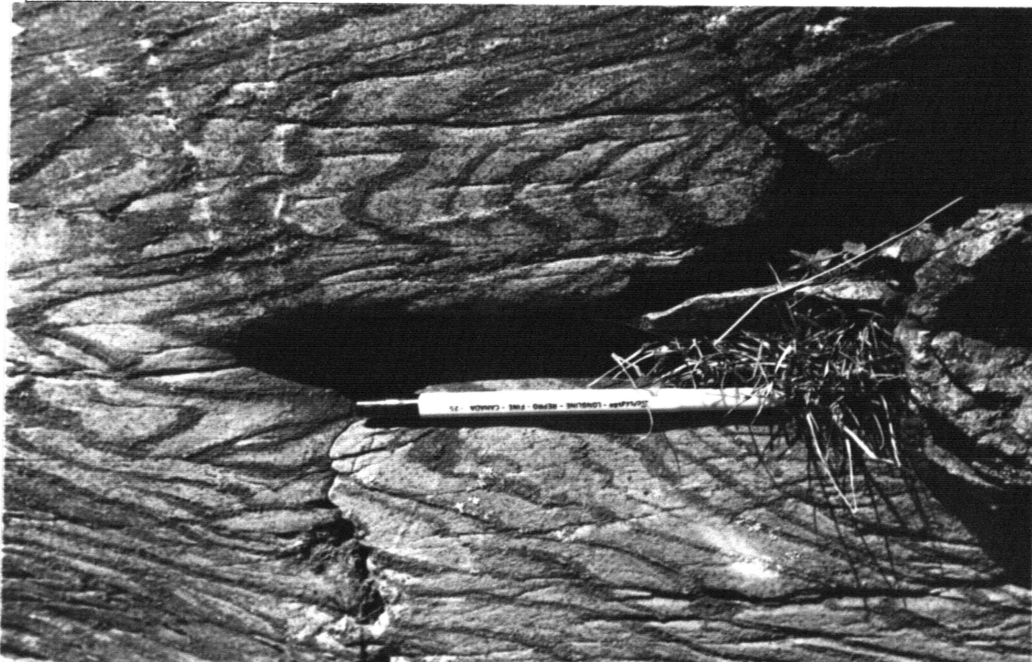


A

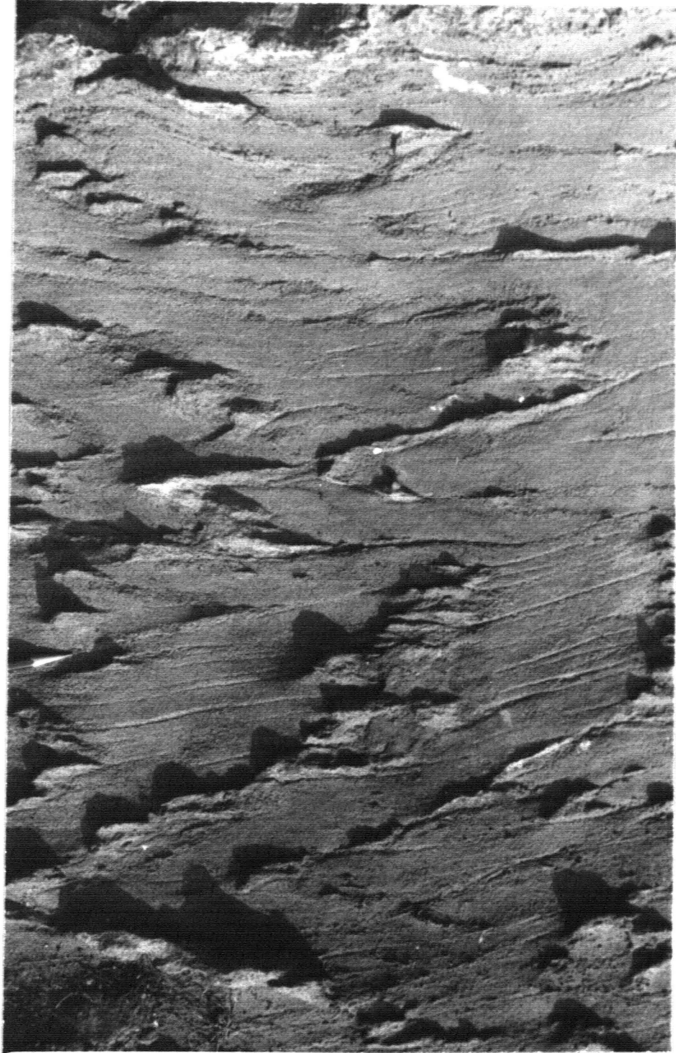


B

A



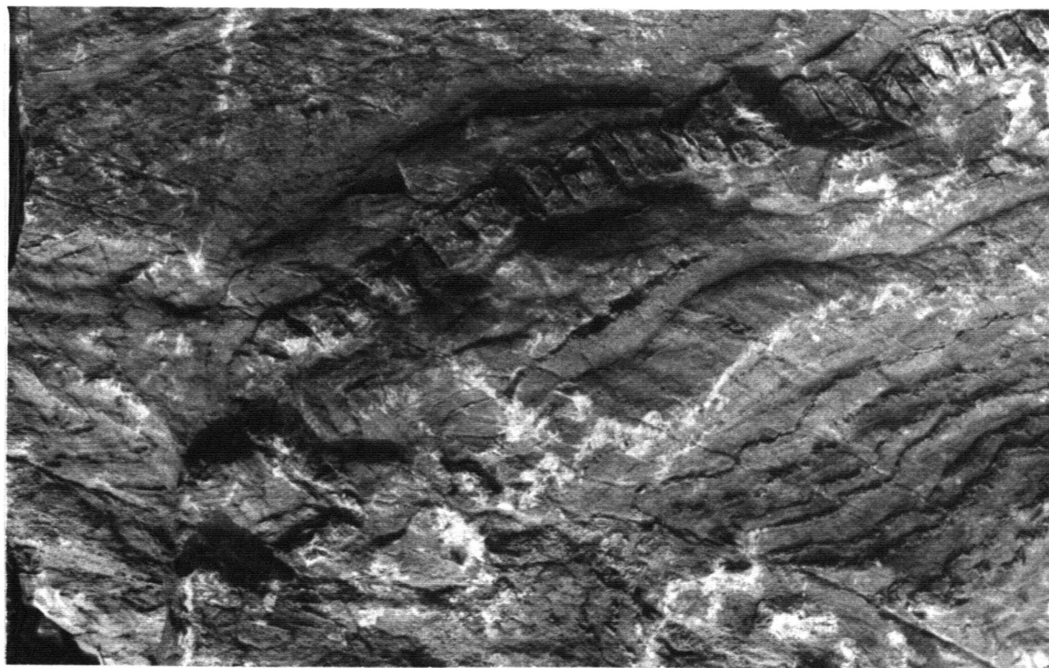
B



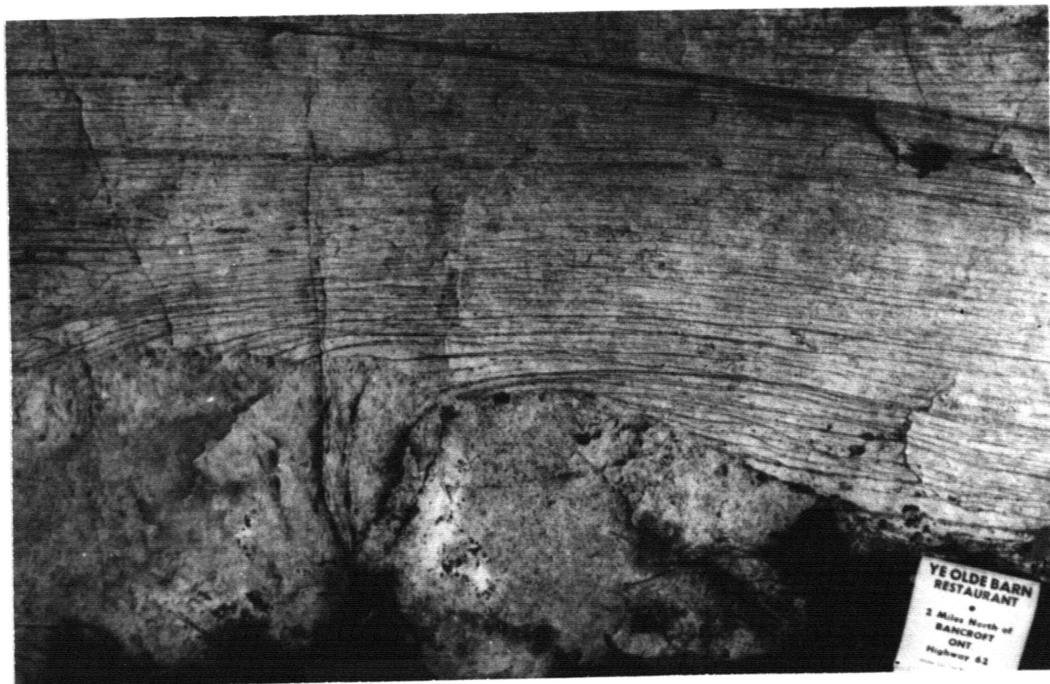
A. Axial-surface fracture cleavage S_2 in micaceous marble (rock Unit 3). Locality : 4 km southeast of Green Corners.

B. Axial-surface fracture cleavage S_2 along which calc-silicate layers are transposed in marble (rock Unit 3). Locality : One kilometer east northeast of Green Corners.

A



B



- A. Refraction of S₂ fracture cleavage across layers differing in competency in calc-silicate - marble rock. Locality : 3km south of Green Corners on Highway 62.
- B. Refraction of S₂ schistosity in feldspathic rock into thicker quartz-rich band (rock Unit 7). Locality : 1.5km southeast of Jamieson Lake.

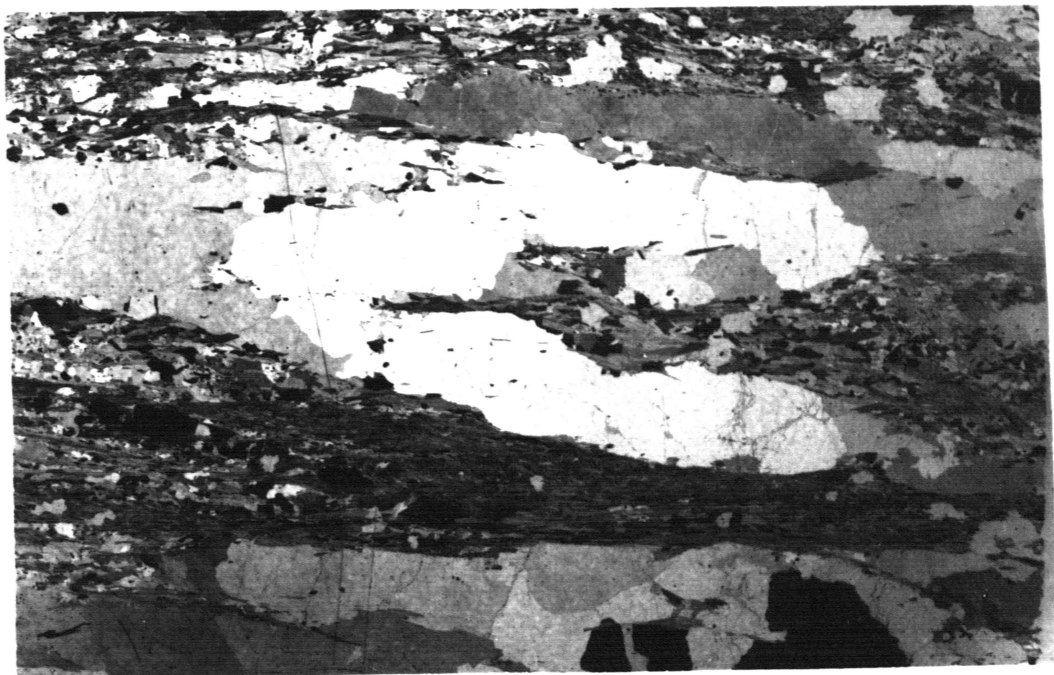
A



A. Recrystallization of calcite along S_2 foliation in open F_2 fold in calc-silicate rock (rock Unit 3). Locality : 1.5 km northwest of Mayo Lake.

B. Alignment of sillimanite along S_2 mica schistosity in semi-pelitic rock (rock Unit 7). Locality : 4.5 km north northeast of Green Corners.

B



1.5 cm



A. Pre-D₂ sillimanite needles bent by D₂; most sillimanite aligned in S₂. A few post-D₂ random crystals of sillimanite can also be seen. Locality : 1.8 km northwest of Jamieson Lake.

B. F₂ fold bending S₁ foliation in scapolite amphibole rock (Unit 7). Some hornblende crystals are aligned parallel to the axial-surface. Locality : 0.5 kilometer west of Upper Hermon.

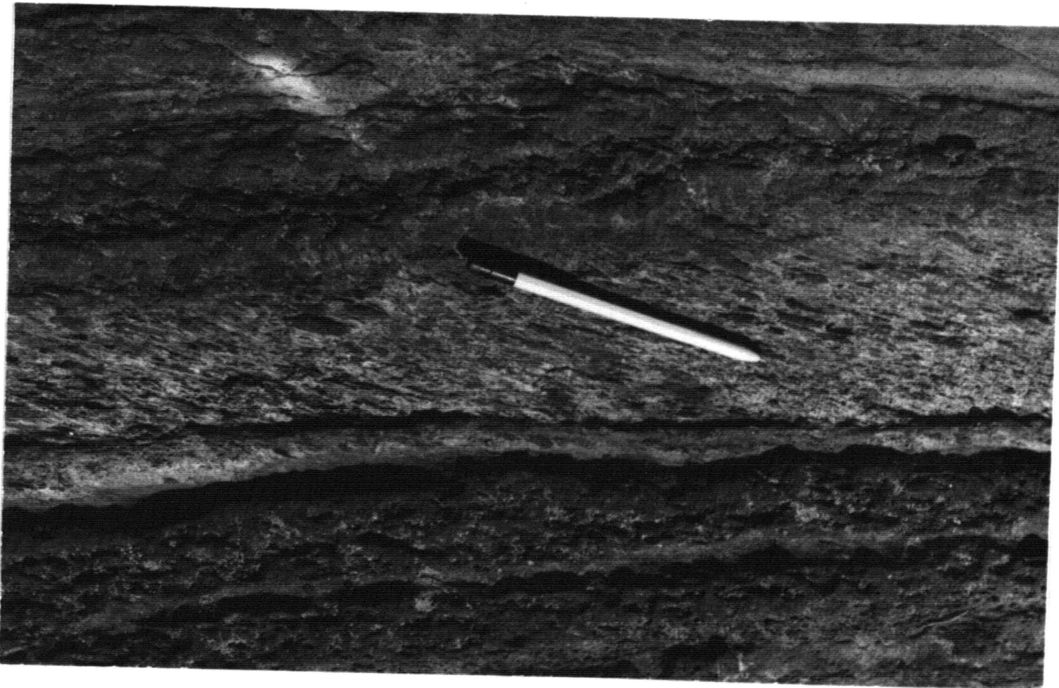


2 mm

B



A

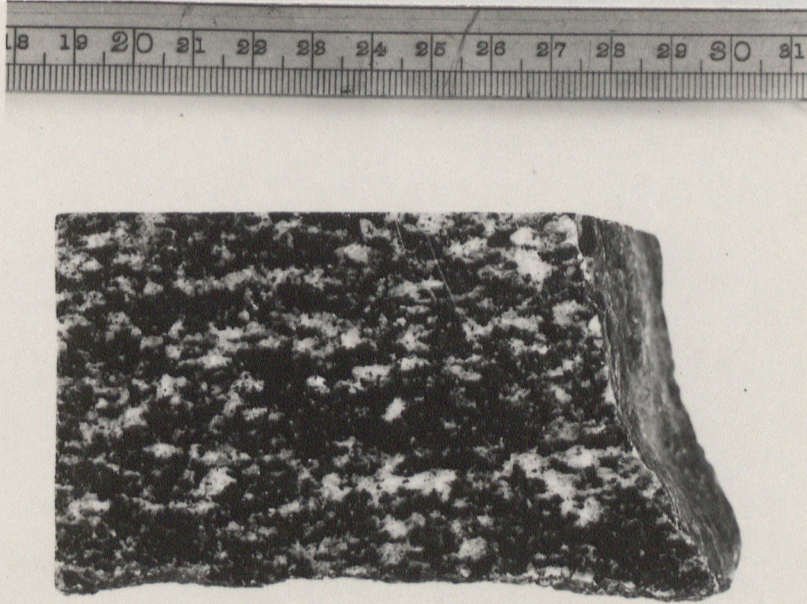


- A. S_2 (parallel to the pencil) oblique to S_1 stratiform foliation in scapolite hornblende rock (rock Unit 7). Locality : 0.5 km west of Upper Hermon.
- B. Detached F_2 hinges in calc-silicate - amphibole rock (rock Unit 7). Locality : 1.5 km west southwest of Hermon.



Fine S_1 quartz preserved in post D_1 garnet porphyroblast (lower left hand corner). S_2 , in which coarse sillimanite, biotite, muscovite and quartz are aligned, is warped around the garnet porphyroblast.

Locality : 2.5 km northwest of Detlor.



- A. Alternating dark (calc-silicate) and white (marble) ribbons defining L₂ in S₂ fracture cleavage surface (to the right of the pencil) in calc-silicate rock (rock Unit 2). Locality : 0.5 km north northeast of Detlor.
- B. Nepheline (white) and amphibole segregations defining L₂ on S₂ gneissosity in nepheline-amphibole gneiss (rock Unit 9). Locality : 1 km east southeast of Bancroft.

3 cm

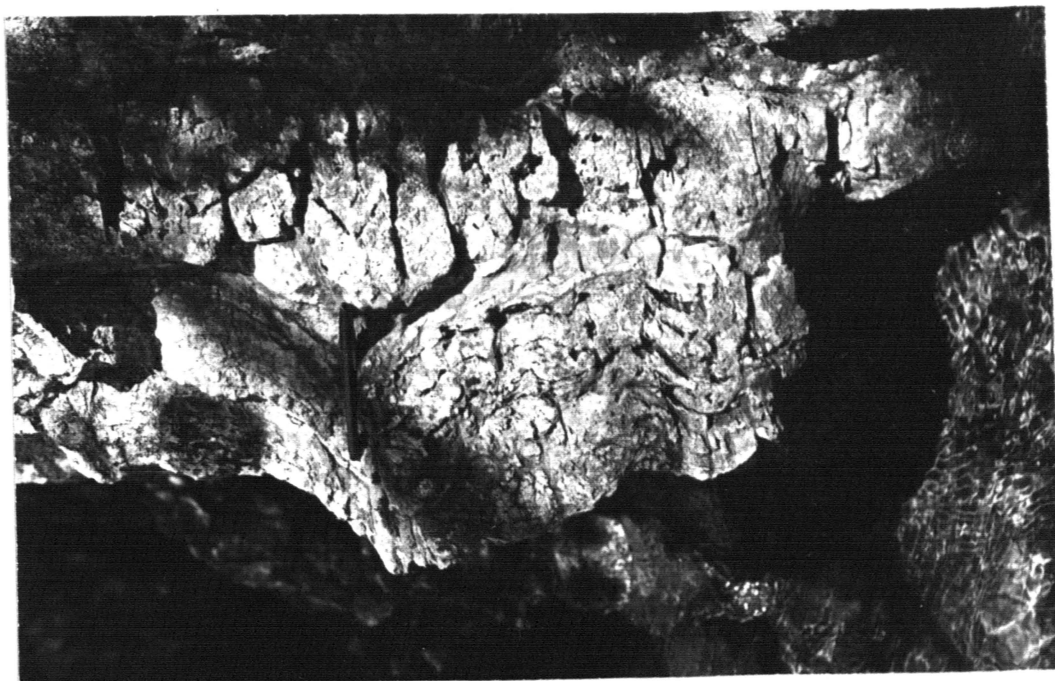


Amphibolite boudin aligned parallel to axial-surface of F_2 fold in calc-silicate - marble layers (rock Unit 3).
Locality : 2 km southeast of Bessemer granite.

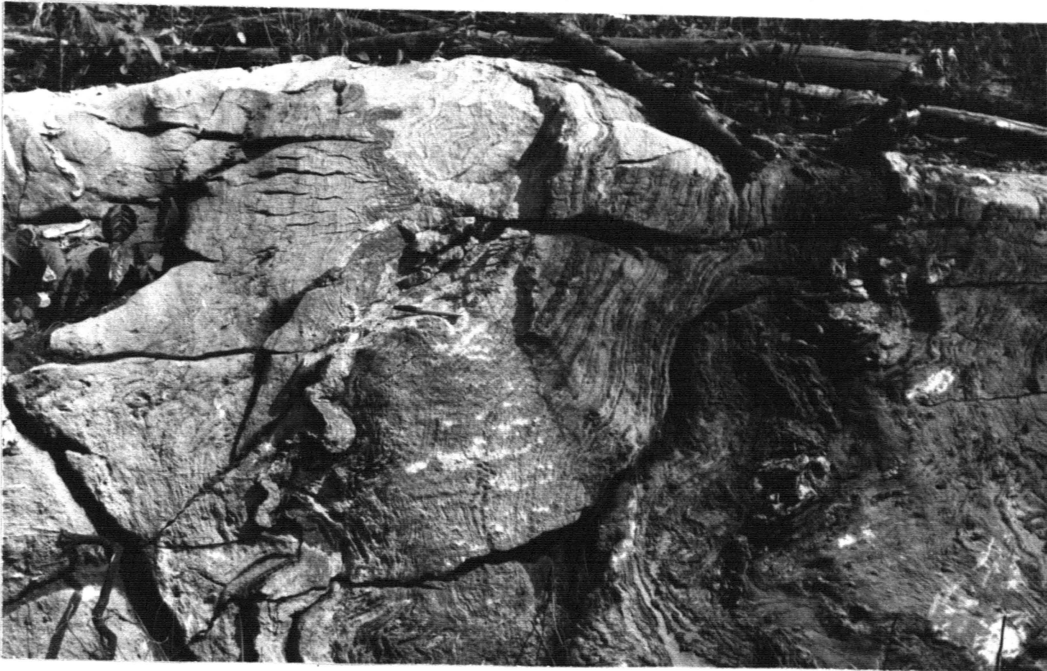
- A. F_3 folds in nepheline gneiss bending S_2 gneissosity. Area marked is enlarged in Plate 18A. to show S_3 axial-surface crenulation cleavage. Locality : Immediately to the south of Great Bend granite.
- B. F_3 folds in calc-silicate layers bending S_0 , S_1 and S_2 . Axial-surface fracture cleavage S_3 is present. Locality : Southern edge of L'Amable Lake.



A



B



A

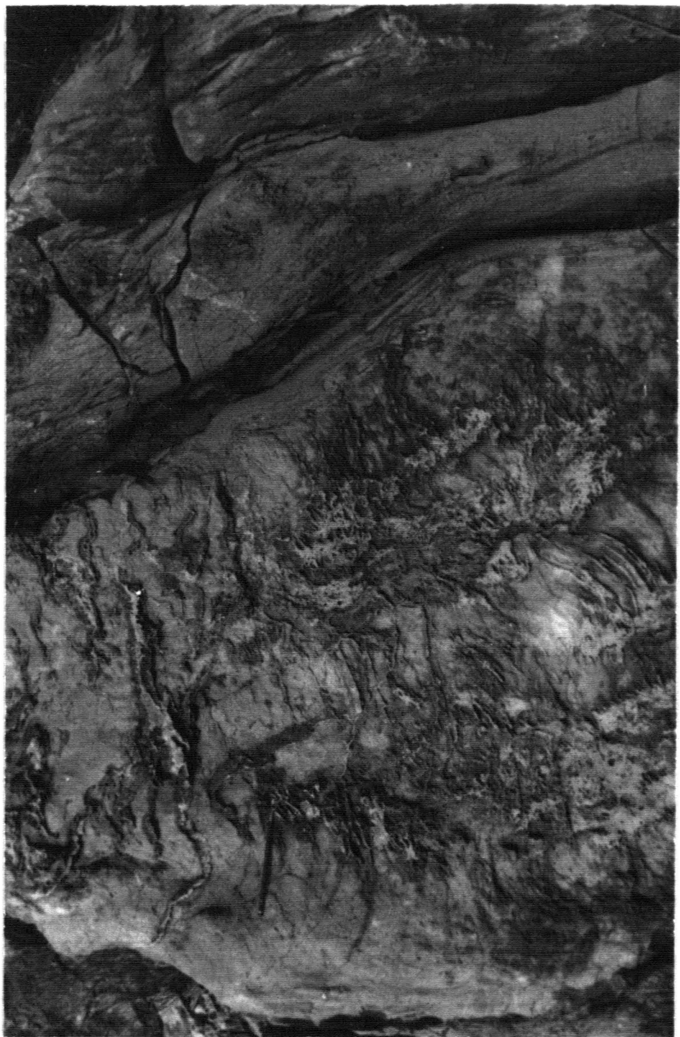


B

- A. Disharmonic F_3 folds in marble - calc-silicate rock. S_3 fracture cleavage is horizontal. Locality : 4 km south southwest of Mayo Lake.
- B. F_3 kinks in biotite (aligned in S_2 gneissosity) in nepheline amphibole gneiss (rock Unit 9). Locality : Immediately to the southeastern edge of Great Bend granite.

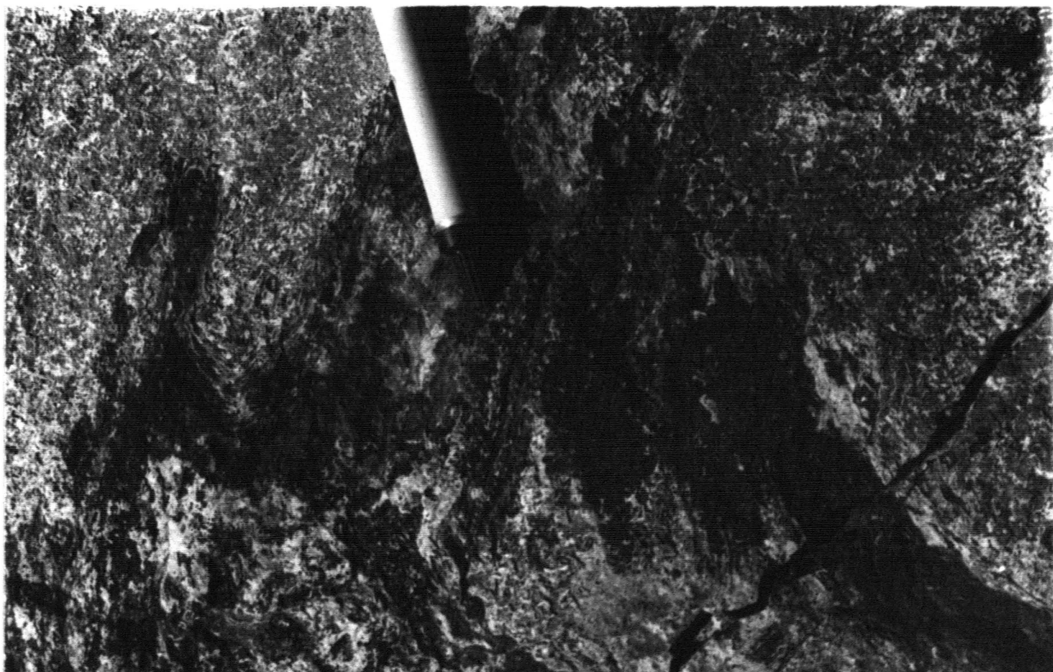
3 mm

B

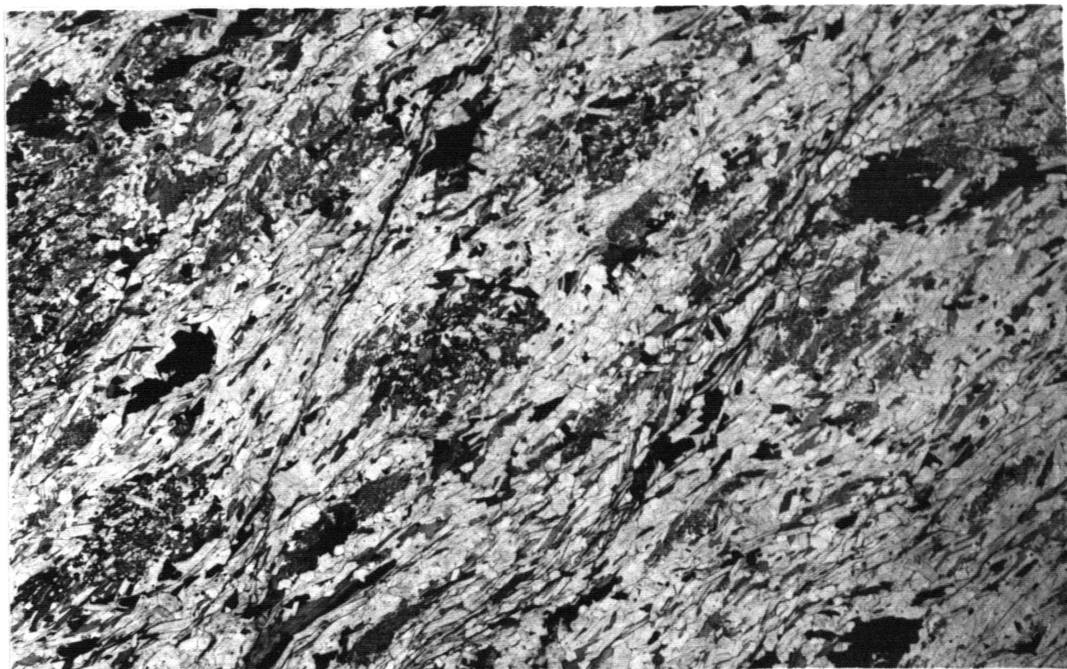


A. S_3 crenulation cleavage in amphibole-biotite rich layer in nepheline gneiss (rock Unit 9). Locality : Immediately to the south of Great Bend granite.

A



B. Recrystallized calcite and quartz along S_3 fracture cleavage in marble - calc-silicate rock. Locality : Near Green Corners on Highway 62.



A

1 cm



B

1.5 cm

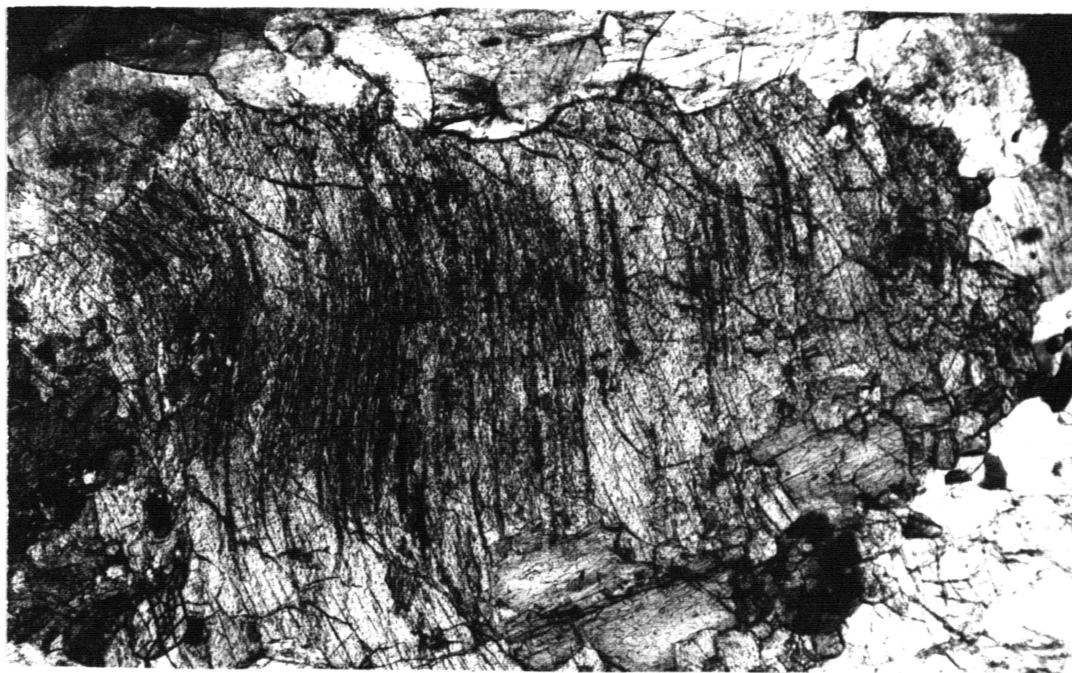
Pre-D₂ biotite porphyroblasts (black) in S₂ schistosity in semi-pelitic schist (rock Unit 7). Locality : 4 km north northeast of Green Corners.

A- Thin section $\perp$ to S₂ and L₂; porphyroblasts are equidimensional. B- Thin section $\perp$ to S₂ and $\parallel$ to L₂; porphyroblasts are elongated.



A

2 mm

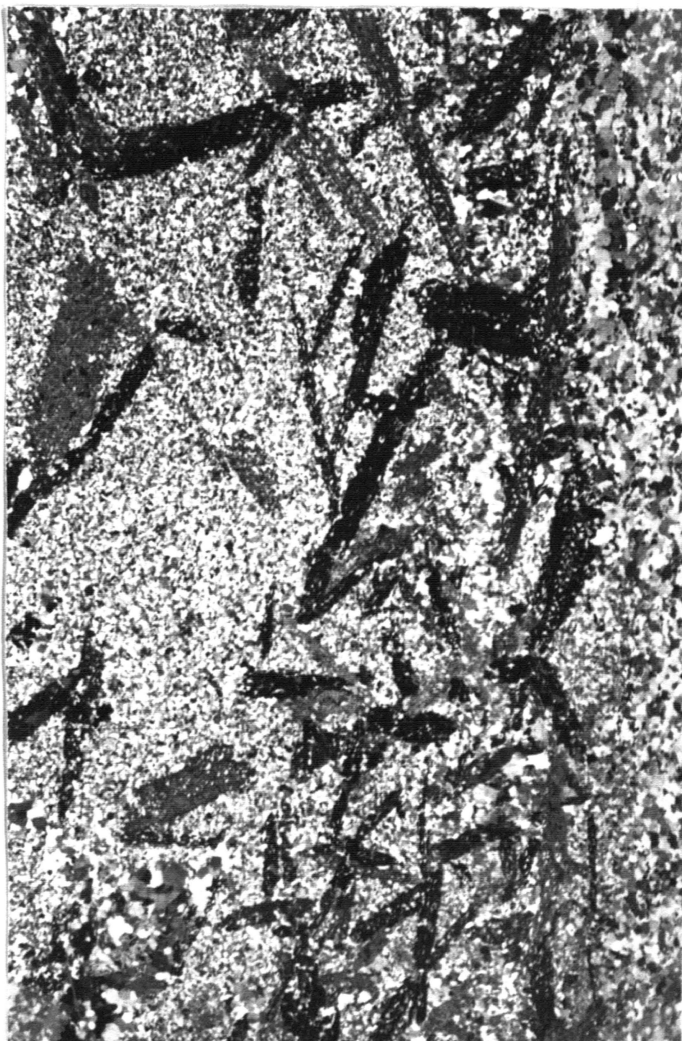


B

1 mm

- A. Post D_2 biotite porphyroblast enclosing S_2 schistosity in pyritic schist (rock Unit 5). Locality : At the central part of western edge of Mayo Lake.
- B. Pre D_2 or early D_2 hornblende crystal deformed by D_2 when the deformation lamellae were produced. Parts of the crystal without the lamellae have annealed during post D_2 times. Locality : 1 km east from the central part of the western margin of the Mallard Lake metagabbro.

- A. Post D_2 hornblende cutting across S_2 defined by alignment of flattened calcite crystals in 'feather amphibolite' (rock Unit 2). Locality: 2 km east of Green Corners.
- B. F_2 fold in quartz layer preserved in late- to post- D_2 garnet porphyroblast in semi-pelitic schist (rock Unit 7). S_2 foliation with sillimanite is warped around the porphyroblast. Note the coarse matrix outside the garnet. Locality: 1.5 km north northwest of Jamieson Lake.



A



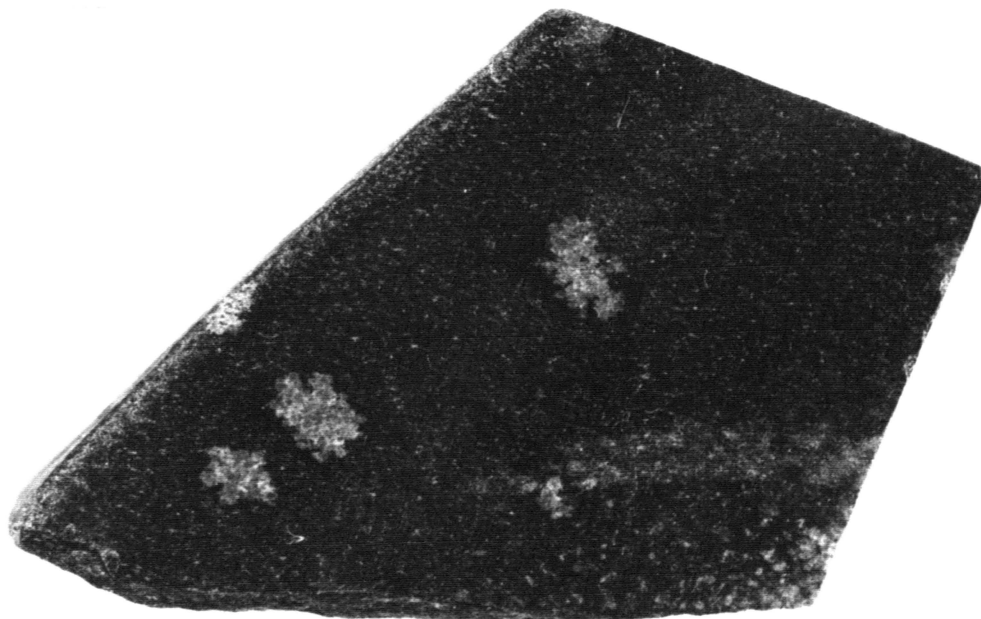
3 mm



B

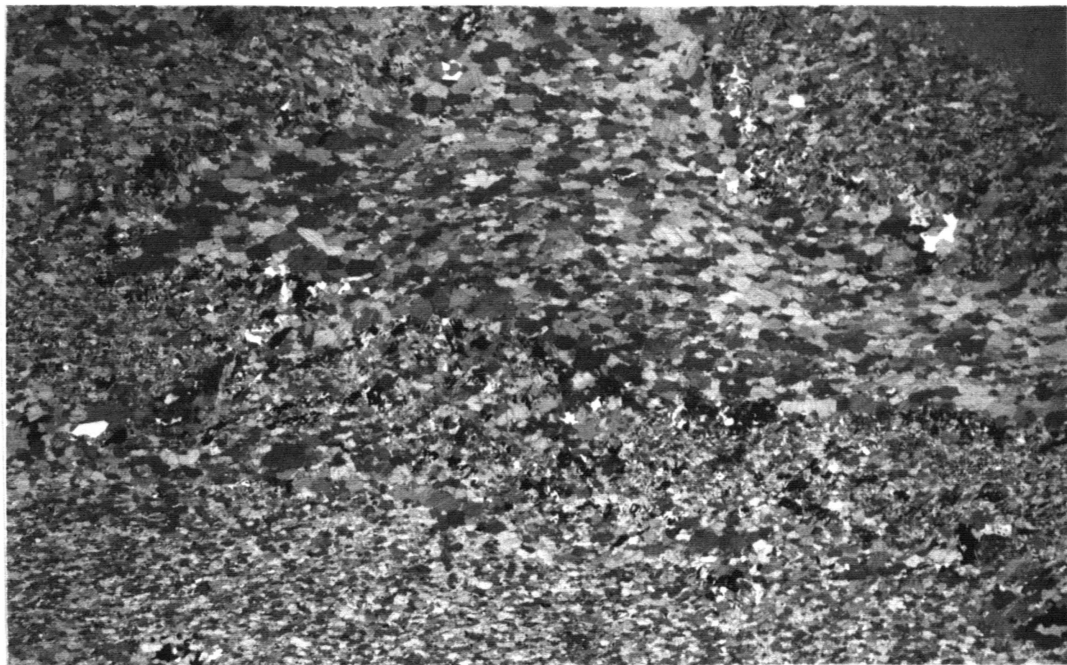


1 mm



A

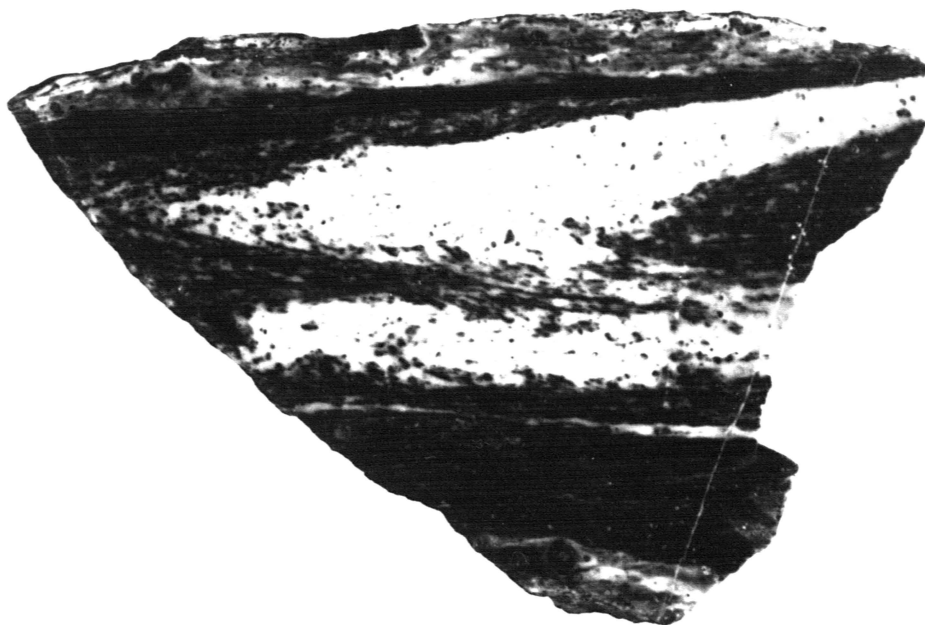
3 cm



B

1 cm

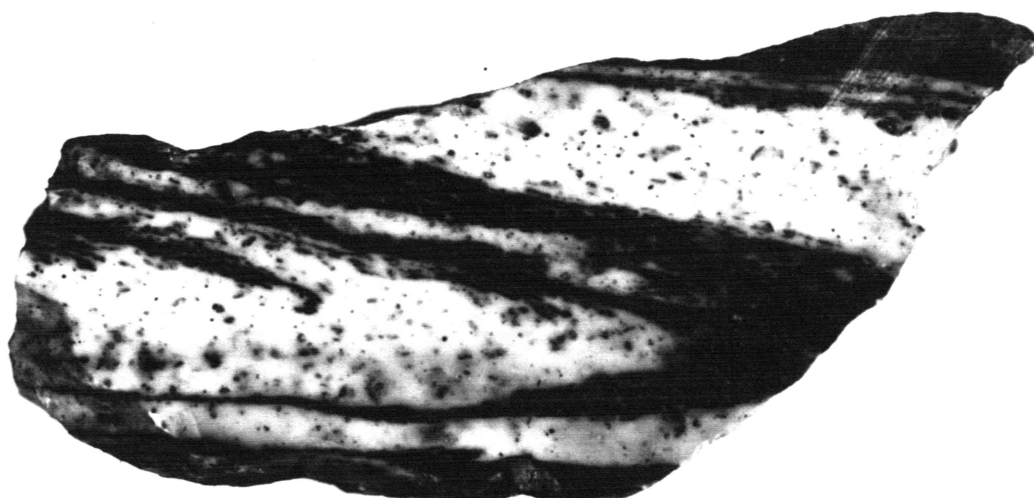
- A. Post D_2 "bird's foot" garnet enclosing S_2 foliation in amphibolitic band in marble (rock Unit 3). Locality : 3 km south southwest of Mayo Lake.
- B. Calcite crystals elongated parallel to the axial-surface of an F_2 fold in marble; the twin lamellae in the calcite crystals suggest flattening of S_2 during late D_2 . Locality : 4 km northeast of Jamieson Lake.



A



3 cm



B



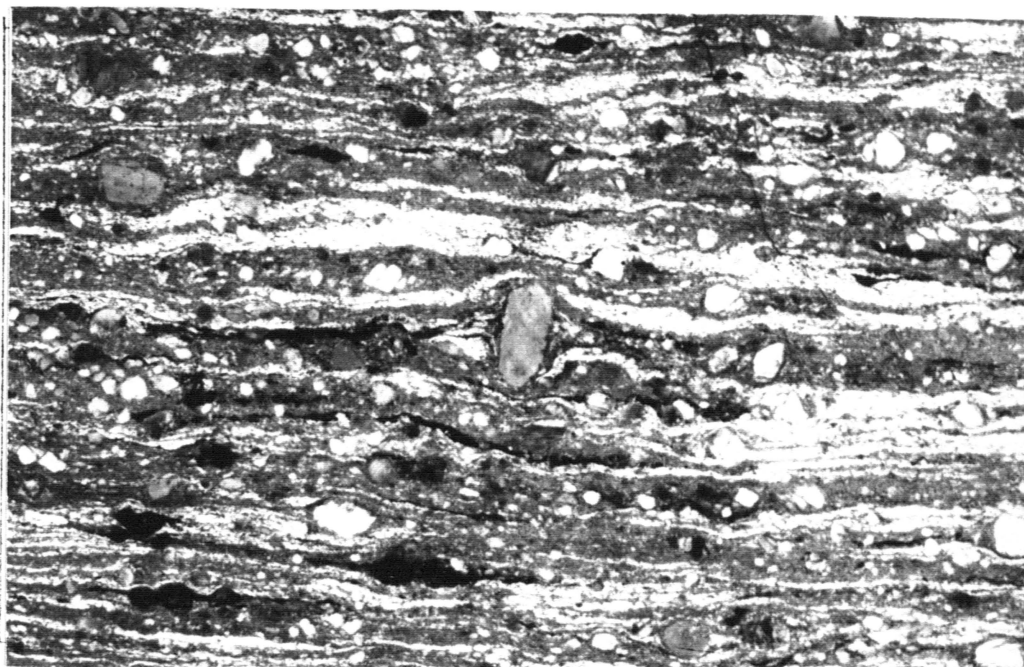
2 cm

F_{1m} fold in marble mylonite bending marble (white) and graphite-rich layers.

A. S_{1m} mylonitic foliation, defined by aligned lenses of calcite and streaks of graphite, lies parallel to the axial-surface of the fold. Locality : 2 km from Hwy. 28 on Monk road.

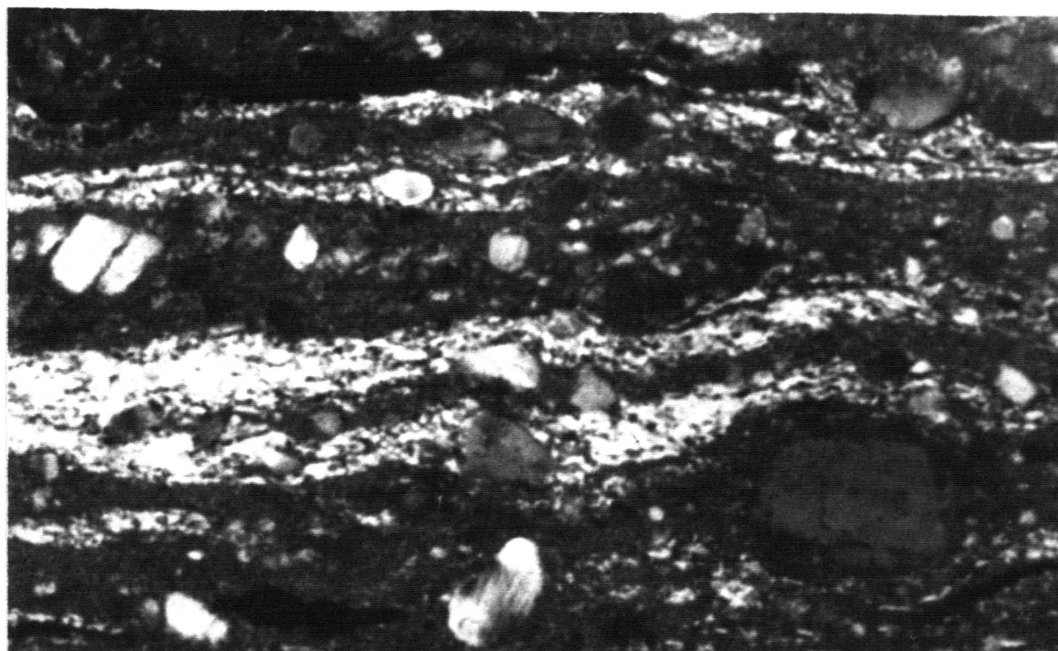
B. A parasitic fold can also be seen above the hinge.

Locality : Same as of A.



A

3 mm

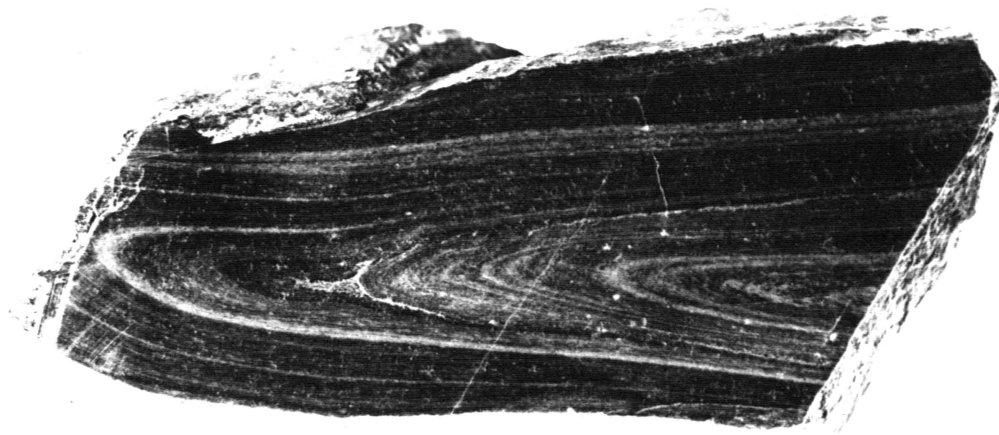
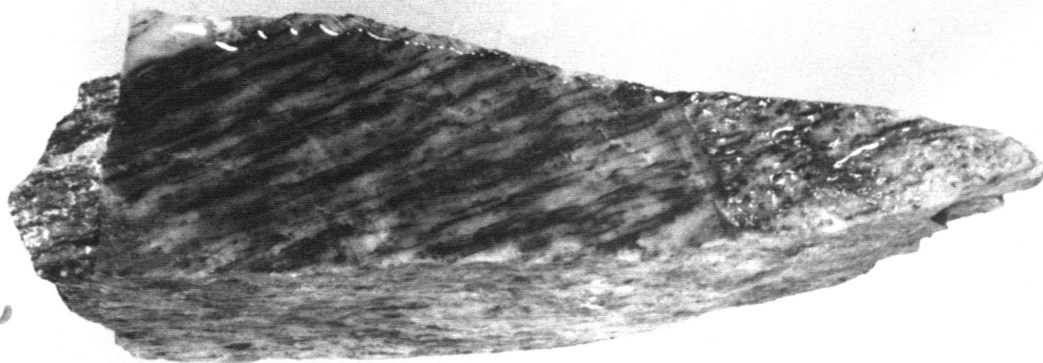


B

1 mm

- A. S_{1m} mylonitic foliation and parallel long dimensions of porphyroclasts of feldspars in syenitic mylonite (note the 'discordant' orientation of the porphyroclast at the center). Locality : 2.5 km from Highway 28 on Monk road.
- B. Enlarged portion of A, displaying the alignment of long axes of quartz grains in the matrix parallel to S_{1m} mylonitic foliation.

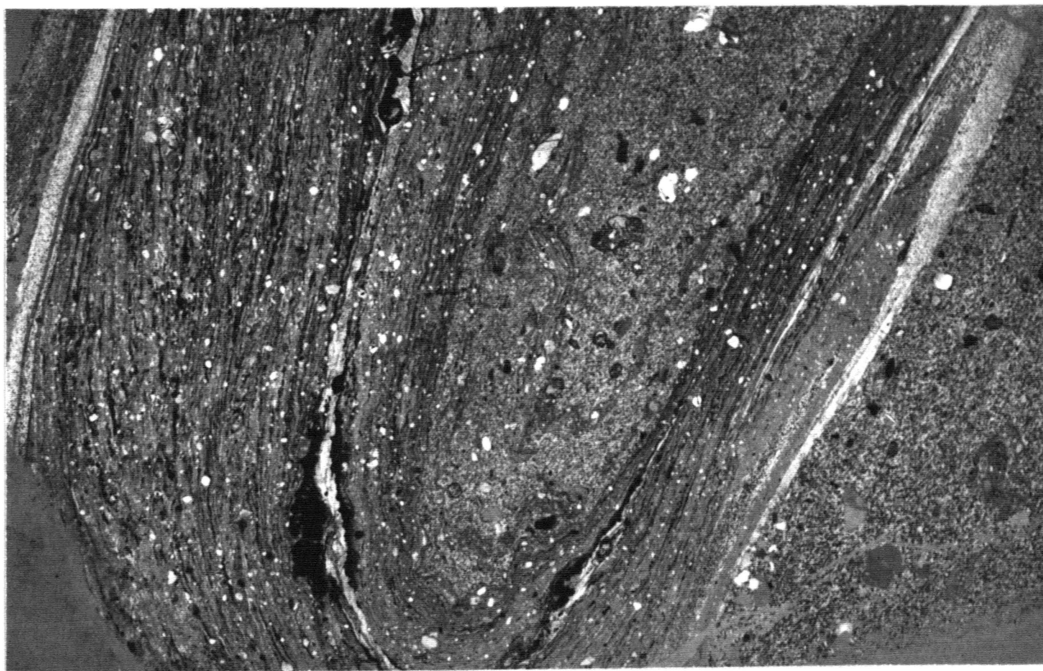
2 cm A 5 cm B



- A. L_{1m} axial lineation, defined by intersections of calcite-rich (white) and graphite-rich layers on S_{1m} mylonitic foliation in marble mylonite. Locality: 2 km from Highway 28 on Monk road.
- B. Near isoclinal F_{2m} fold bending S_{1m} mylonitic foliation in syenitic mylonite. Locality: 2.5 km from Highway 28 on Monk road.

B

8 mm



A

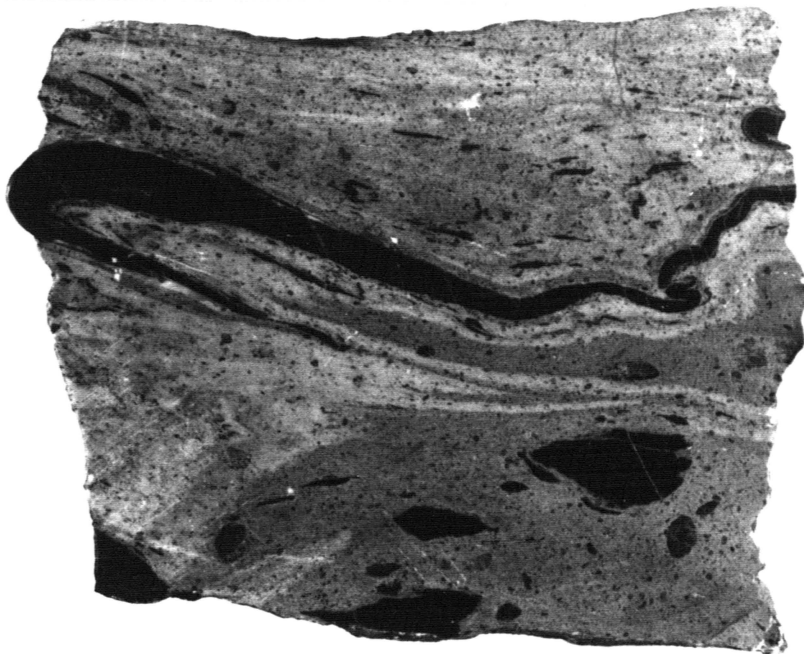
5 mm



A. Interfolial F_{2m} fold bending S_1 mylonitic foliation in syenite mylonite. The white layer is composed of quartz. Locality : 2.5 km from Highway 28 on Monk road.
 B. Near isoclinal F_{2m} fold bending S_1m mylonitic foliation in interlayered marble (in the core) and syenitic mylonite. Locality : Same as of A.

8 cm B

5 cm A



A and B - Detached F_{2m} fold hinges in syenitic mylonite layer enclosed in marble mylonite. S_{1m} mylonitic foliation is bent in the hinge area. S_{2m} foliation is parallel to the axial-surface of the fold. Locality : 2.5 km from Highway 28 on Monk road.

A. F_{2m} fold bending S_{1m} mylonitic foliation (virtually obliterated) in syenitic mylonite. Aligned long axes of quartz grains in quartz-rich layer define S_{2m} schistosity. (Photo taken from fold hinge in Plate 26A).

B. Isoclinal F_{2m} fold in S_{1m} mylonitic foliation, syenitic mylonite in marble. S_{2m} axial-surface schistosity is defined by the alignment of long axes of flattened calcite grains in the marble. Locality: 2.5 km from Highway 28 on Monk road.



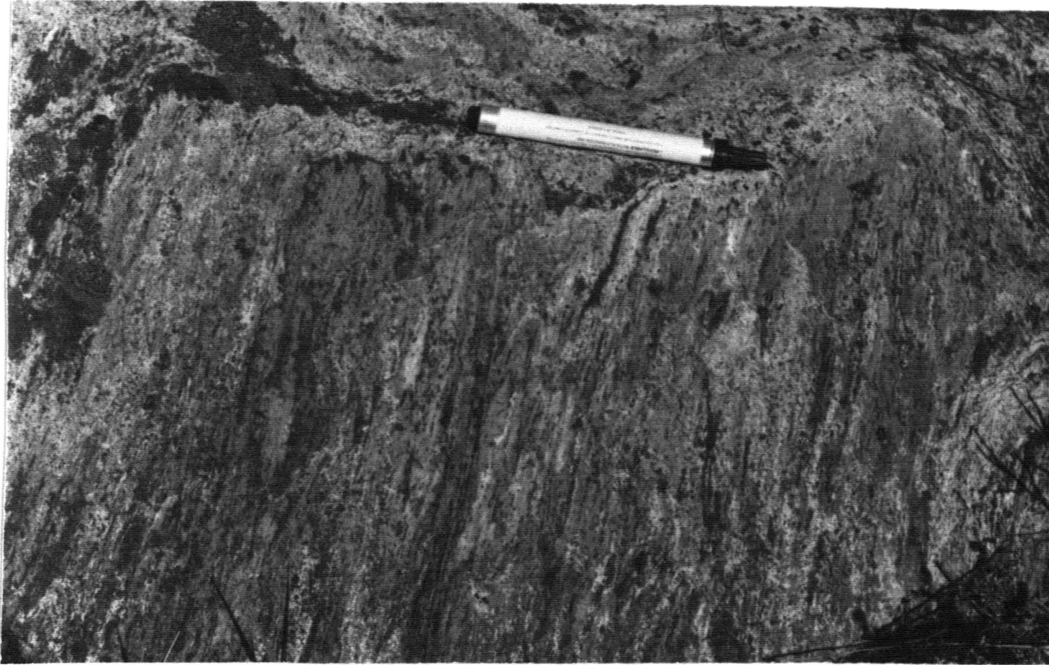
1 mm

A

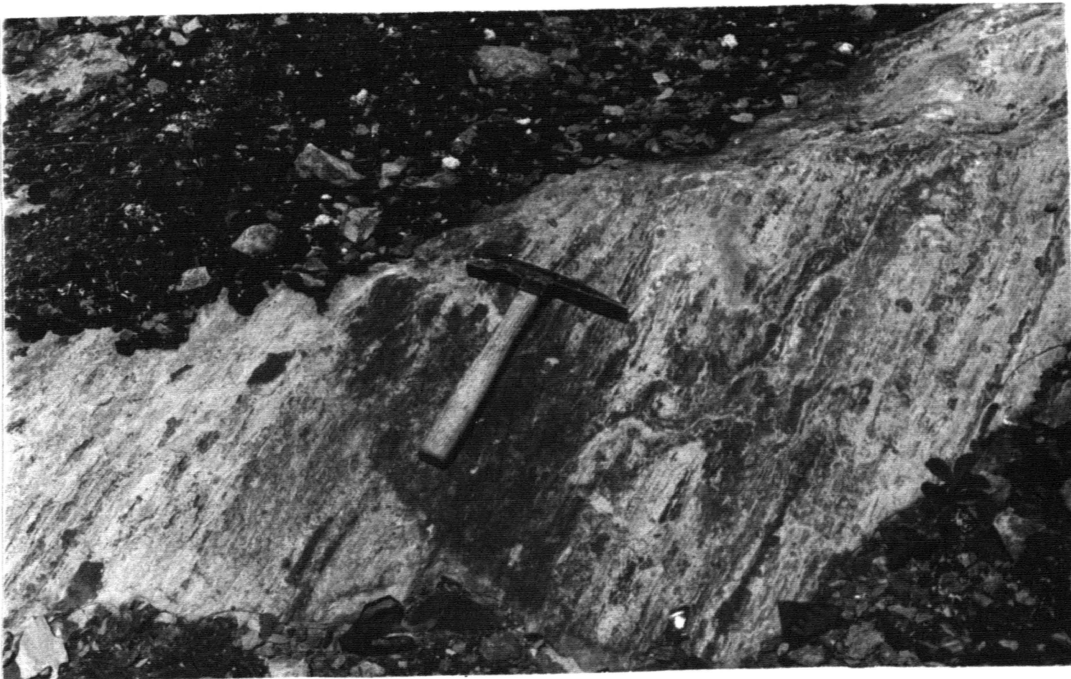


2 mm

B

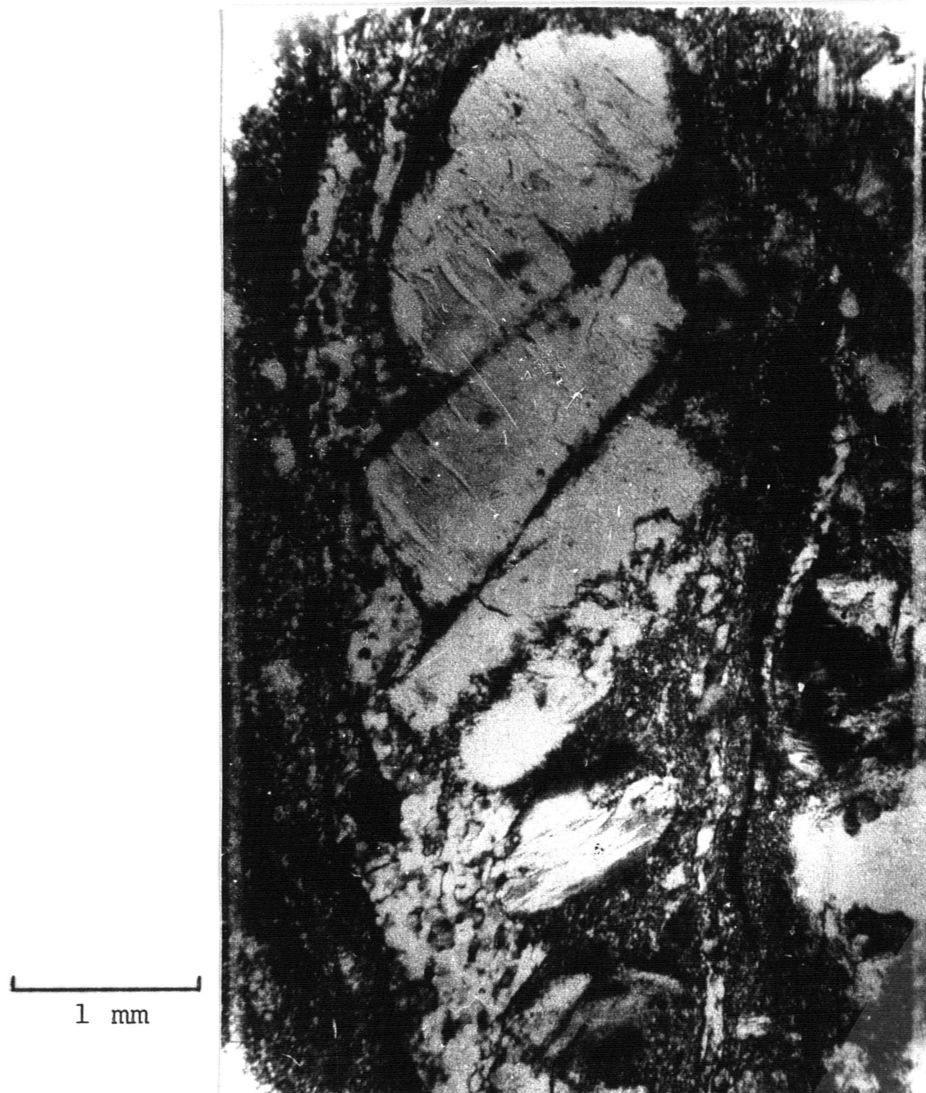


A



B

A and B - L_{2m} axial lineation defined by parallel streaks of carbonaceous and calcitic(white) material on S_{2m} schistosity in marble mylonite. Locality : 3 km east of Bancroft on Highway 500.



Displacement along parallel shear fractures in feldspar porphyroblast lying in S_{1m} mylonitic foliation in syenitic mylonite.

Locality : 2.5 km from Highway 28 on Monk road.

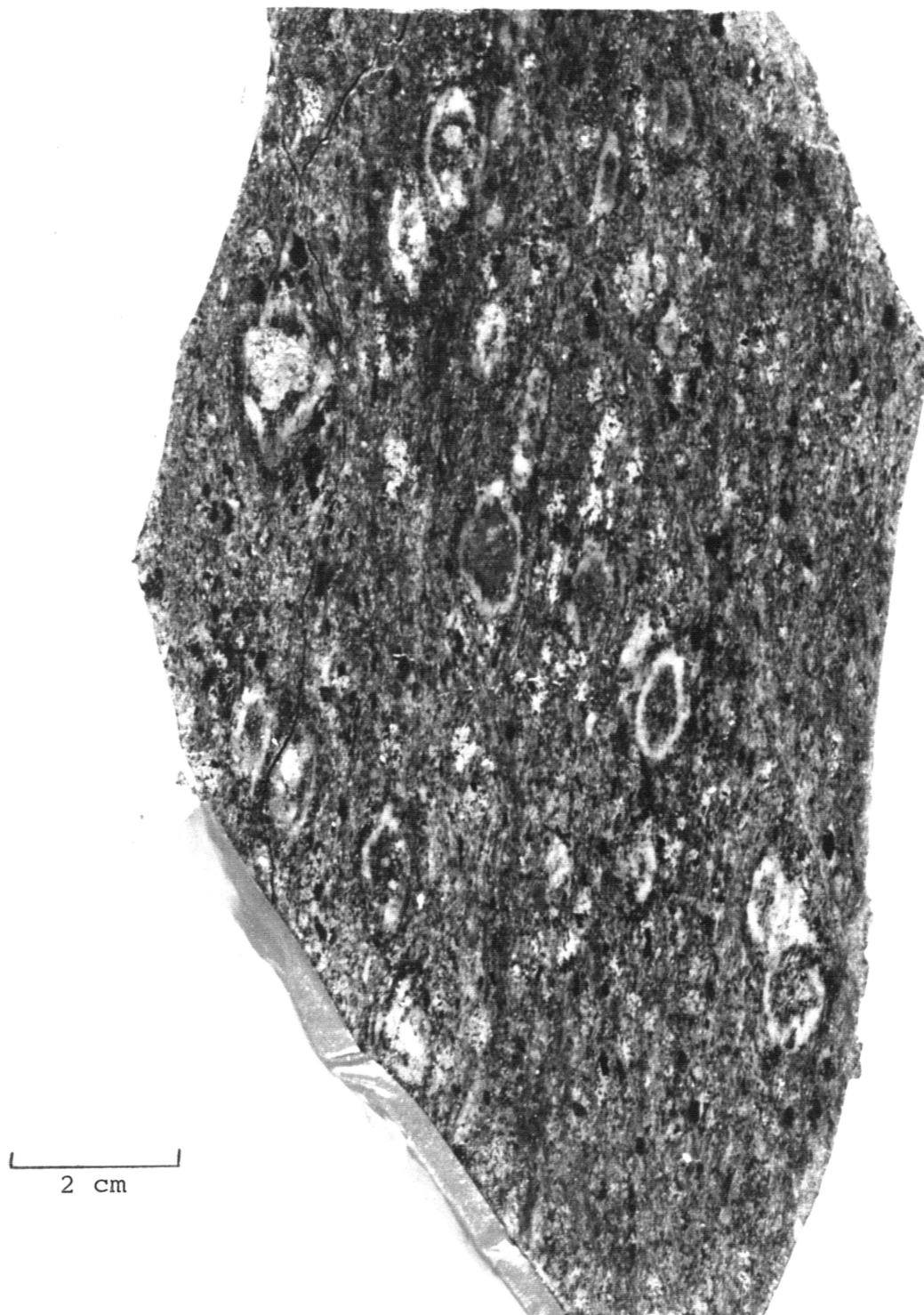


Granitic pegmatites folded with axial-surfaces parallel to S_2 schistosity in the semi-pelitic rock.

Locality : A. Close to the northeastern edge of the Weslemkoon pluton, 20 km east of the study area.

B. 2 km northeast of Mayo Lake.





Zoned porphyroblasts consisting of garnet, biotite and quartz, and dimensionally oriented in S_2 mica schistosity. A core consisting of garnet is always separated from an outer rim of sillimanite (white) by an internal zone consisting of biotite and quartz.
Locality : 2.5 km northwest of Detlor in semi-pelitic schist.

APPENDIX

Development of Sillimanite

Sillimanite almost always occurs in close association with quartz and biotite, irrespective of whether it is inferred to be a product of a staurolite-consuming reaction (Guidotti, 1966; Chakraborty and Sen, 1967), a kyanite-consuming reaction (Chinner, 1961), or a muscovite-consuming reaction (Evans and Guidotti, 1966). At and north of the sillimanite isograd, sillimanite associates with garnet, biotite and/or muscovite, plagioclase and quartz. Optical properties, determined with the aid of Universal Stage, of plagioclase indicate that its composition is near that of oligoclase-andesine.

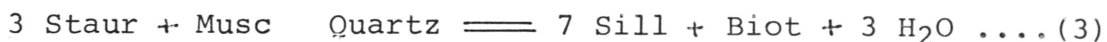
Sillimanite when associated with potash feldspar indicates upper amphibolite facies conditions (Turner, 1968, p. 311). Though sillimanite-potash feldspar association has been observed in Grenville metasediments elsewhere (Wynne-Edwards, 1967, Table 8; Rambaldi, 1971, p. 16), this association is not observed in the study area. Therefore, though textural evidences suggest that muscovite has been involved in the formation of sillimanite, absence of potash feldspar indicates that the reaction:

Musc + Quartz $\rightleftharpoons$ Sill + K-Feld(1)
did not take place.

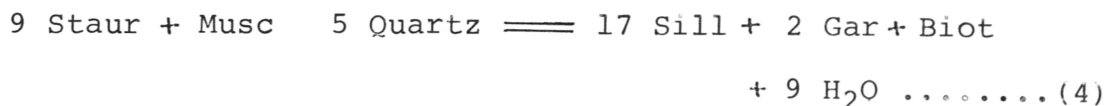
Kyanite is also absent in the present area. Therefore, unlike in the Barrovian sequence of the Dalradians in Scotland (Chinner, 1961) and area A (Fig. 1; Carmichael, 1969, p. 253)

either sillimanite did not develop 'from' kyanite or like in the former places it did develop from kyanite, but kyanite was completely consumed during the reaction. The latter suggestion seem improbable in view of other textural evidences observed in the rocks.

Textural relations between garnet and staurolite in sillimanite bearing schist about one mile north northwest of Jamieson Lake (Fig. 2), indicate a possible reaction by which sillimanite could have developed. In the sillimanite schist, 'remnants' of small grains of staurolite occur within and rarely outside the garnet grains. Biotite is the only mica present in the rock. As suggested for similar textural relations between these minerals by earlier workers, a single or a set of the following reactions could have taken place:



(Chakraborty and Sen, 1967, p. 219)



(Carmichael, 1969, p. 259)

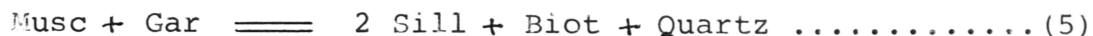
Therefore, in the sillimanite bearing pelitic schist north northwest of Jamieson Lake, either reaction (2) must have occurred, or absence of muscovite in the rock is due to the completion of reaction (3) or (4), or both (3) and (4). The latter possibility is more probable because at other places breakdown of muscovite is clearly apparent in the textures and some muscovite

is always present in these pelitic rocks with sillimanite.

Another interesting textural relationship between garnet, biotite, quartz and sillimanite indicates a different 'line' of development of sillimanite. Quartz and biotite are closely associated with garnet in two different textural relations. Firstly, they occur within garnet porphyroblasts as 'patches' and garnets enclosing them closely resemble the so-called 'attol garnets' (Rast, 1965, p. 94). Such attol garnets occurring in the Central Highlands of Scotland are explained by Rast (1965, p. 95) to have developed from pre-existing zoned garnets whose core was 'replaced' by quartz, feldspar and biotite. Whether such zoning existed in the garnets in the present area is not known. But textures suggest that the quartz and biotite within these garnet porphyroblasts have developed as a result of breakdown of the garnet.

Secondly, that the garnet had broken down to give rise to quartz and biotite, is also clear from the 'ellipsoidal reaction rims' well developed around the garnet porphyroblasts in the pelitic schist containing sillimanite and muscovite (Plate 32). Cores consisting of garnet and outer zone consisting of sillimanite are always separated by an inner zone consisting quartz and biotite. Quantitative proportions of the core and the inner zone appear to be inversely related. Some of these ellipsoids with these reaction rims do not contain any garnet at all, and the core is completely replaced by quartz and biotite.

Therefore, it is apparent that breakdown of garnet and muscovite appear to be related to the development of quartz, biotite and sillimanite. This reaction, as given by Kretz (1959, p. 375) can be written as:



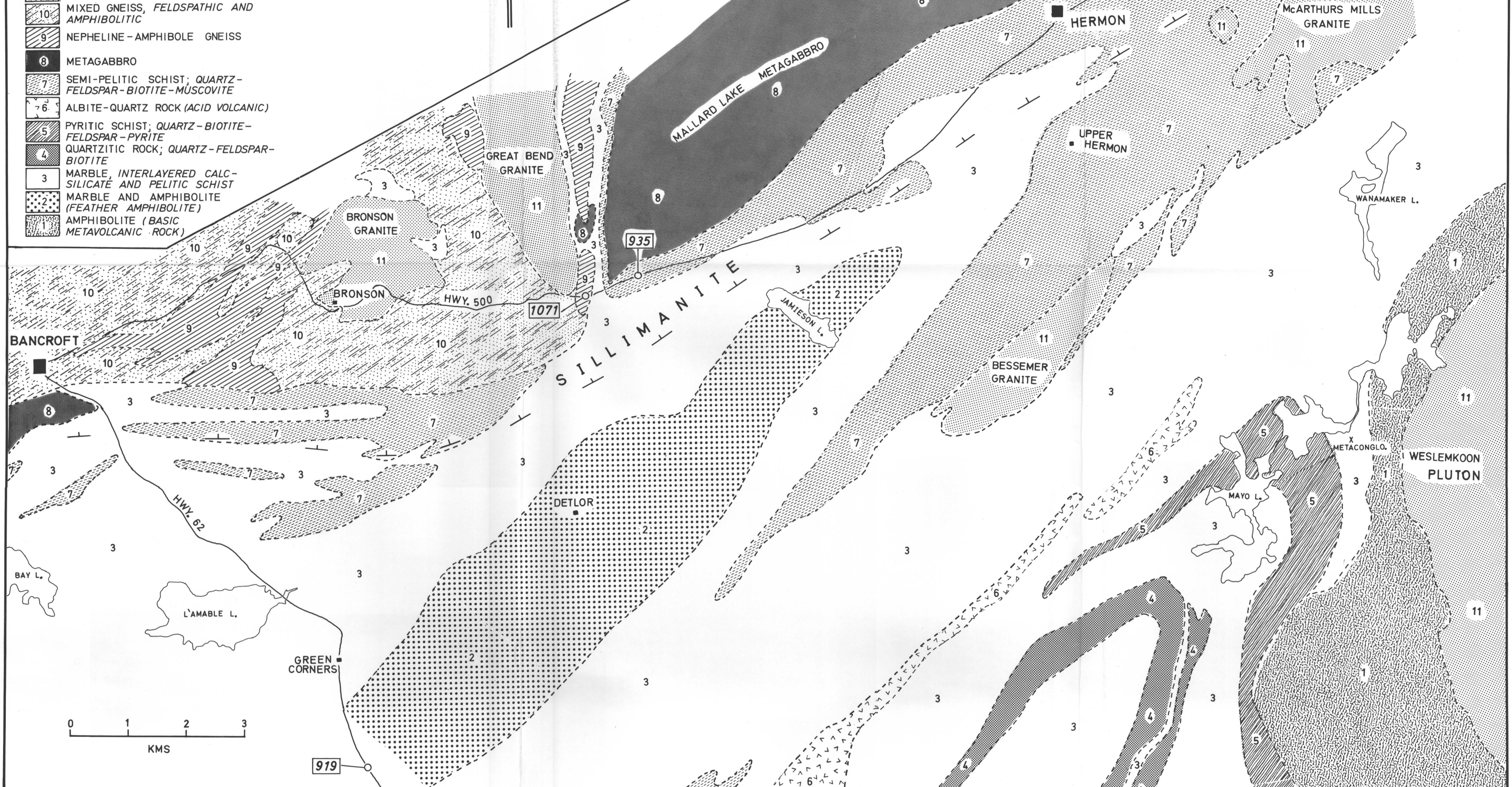
Obviously, this reaction was not complete and so all the five phases occur in the rock.

Therefore, sillimanite is explained here as to have developed along two 'lines' -- first one involving the destruction of muscovite and staurolite and production of garnet and sillimanite (reactions 2, 3, & 4), and second one involving the destruction of muscovite and garnet and production of sillimanite, biotite and quartz (reaction 5). As a result, it appears that at one instance garnet is produced and at the other garnet is consumed. This 'apparent contradiction' is due to different "times" and 'modes' of growth of sillimanite. Under static conditions feather-like and mats of sillimanite grew earlier by a reaction involving the destruction of muscovite and garnet. Later under dynamic conditions sillimanite in the form of prisms and needles grew by a reaction involving the destruction of muscovite and staurolite, and production of garnet.

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- 10 MIXED GNEISS, FELDSPATHIC AND AMPHIBOLITIC
- 9 NEPHELINE-AMPHIBOLE GNEISS
- 8 METAGABBRO
- 7 SEMI-PELITIC SCHIST; QUARTZ-FELDSPAR-BIOTITE-MUSCOVITE
- 6 ALBITE-QUARTZ ROCK (ACID VOLCANIC)
- 5 PYRITIC SCHIST; QUARTZ-BIOTITE-FELDSPAR-PYRITE
- 4 QUARTZITIC ROCK; QUARTZ-FELDSPAR-BIOTITE
- 3 MARBLE, INTERLAYERED CALC-SILICATE AND PELITIC SCHIST
- 2 MARBLE AND AMPHIBOLITE (FEATHER AMPHIBOLITE)
- 1 AMPHIBOLITE (BASIC METAVOLCANIC ROCK)



- $S_1 // S_0$ FOLIATION
- S_2 FOLIATION
- $S_2 // S_1$ FOLIATION
- S_3 FOLIATION
- F_1 FOLD AXIS
- L_1 LINEATION
- F_2 FOLD AXIS
- L_2 LINEATION
- F_3 FOLD AXIS
- L_3 LINEATION

- ANTIFORM
- SYNFORM

AXIAL TRACE OF MAJOR F_2 FOLDS

