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STRUCTURES AND TEXTURES OF METAMORPHIC ROCKS,  
OMPAH AREA, GRENVILLE PROVINCE, ONTARIO

by Toby Rivers

Thesis submitted to the School of Graduate Studies  
of the University of Ottawa as partial fulfillment  
of the requirements for the degree of Doctor of  
Philosophy in Geology

UNIVERSITY OF OTTAWA  
OTTAWA, CANADA, 1976

## ABSTRACT

Near Ompah, Ontario, metasedimentary and metavolcanic rocks of the Grenville Supergroup lie in a linear belt between three large gneissic granite plutons. During metamorphism and multiple deformation the rocks were folded about northeast trending axes, and a metamorphic gradient increasing from southwest to northeast was imposed.

Deformation was in three distinct phases, of which the first two formed major folds. Low plunging  $F_1$  folds are tight to isoclinal, and generally considerably flattened. The associated axial foliation ( $S_1$ ) and lineation ( $L_1$ ) are penetrative and widespread. Locally mylonite zones developed parallel to  $S_1$ . Strain during  $D_1$  was inhomogeneous, and as shown by deformed pebbles and crystal aggregates varied from flattening to constrictional types, the latter prominent in the hinge area of a major fold. Maximum extension was parallel to  $F_1$  fold axes throughout the area. The geometry of later folds suggests that  $F_1$  folds were originally gently inclined to recumbent. During  $D_1$  the grade of metamorphism varied from greenschist to amphibolite facies, and at any locality the end of  $D_1$  coincided with the metamorphic climax.

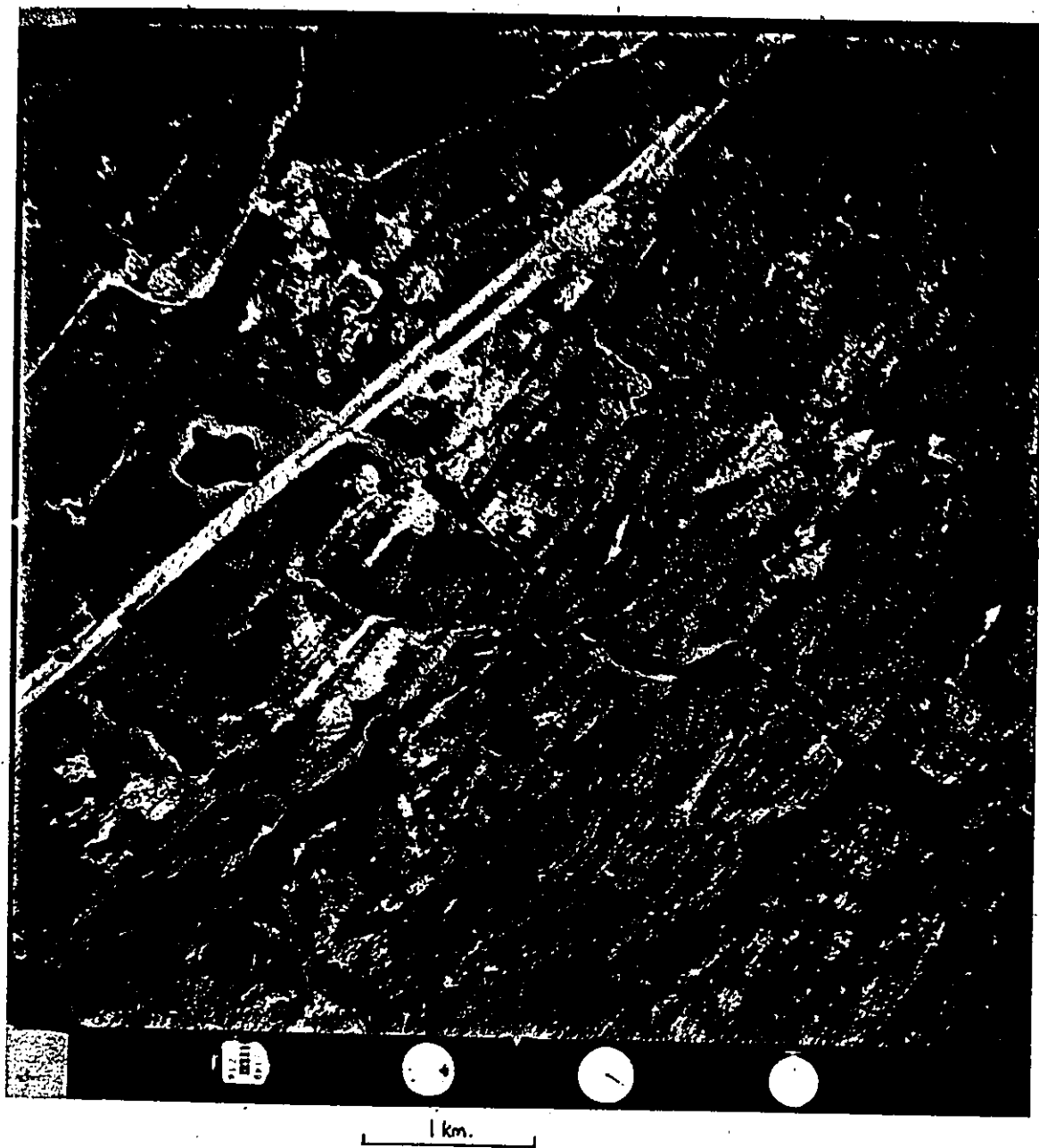
$F_2$  folds, coaxial with  $F_1$ , are low plunging, upright structures of variable scale, which formed at or soon after the metamorphic peak. They are generally isoclinal, but

less flattened than  $F_1$  folds. The associated  $S_2$  foliation and  $L_2$  lineation are not well developed over most of the area. Sporadically developed  $F_3$  folds are open, recumbent structures with little associated crystal alignment.

Textures of mineral assemblages in pelitic rocks have been used to interpret possible mechanisms for three prograde metamorphic reactions; staurolite breakdown in the presence of quartz, the polymorphic transformation of kyanite to sillimanite, and muscovite breakdown in the presence of quartz. The distribution of phases around the reactant species, together with evidence of breakdown of the reactants at lower grades than where the assumed products of thermal equilibrium reactions appear, suggests modification of thermal reactions by localized activities of elements such as  $K^+$ ,  $H^+$  and  $Na^+$ , and by the fugacity of water. These activities may have been in part determined by recrystallization of other mineral species not directly involved in the prograde reactions.

Muscovite crystals from a pelitic schist unit become on average larger and more equant with increasing metamorphic grade, the size and shape varying independently. The variation in shape may reflect a change in equilibrium shape with temperature. Accompanying the increase in size and change in shape, the preferred orientation of the muscovite crystals increases with grade of metamorphism.

Evidently crystals orientated at high angles to the foliation were progressively eliminated. It is considered that all the textural changes were achieved by recrystallization as textural equilibrium was more closely approach in rocks of high metamorphic grade.



Aerial photograph of the closure of the  $F_1$  Ompah synform.



Aerial photograph of the closure of the  $F_2$  Plevna synform.

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## INTRODUCTION

### Location and Rock Outcrop

The study area, near Ompah, Ontario (fig. 1) is covered by the Sharbot Lake, Mazinaw Lake and Clyde map sheets (Nos. 31C/15, 31C/14 and 31F/2 respectively) of the Canadian National Topographic 1:50,000 series. It includes all or parts of Clarendon, Palmerston, Miller, South Canonto and Lavant townships in the counties of Frontenac and Lanark, southeastern Ontario. Most of the area is covered by two regional geological compilations, map 2053 "Madoc Area" of the Ontario Department of Mines and paper 70-2 of the geology department of Carleton University, as well as by the detailed maps of Miller and Knight (1914) and Smith (1958). Access to the region is gained by highway 506 from the west, and 509 from the east and south; within the region a network of township and logging roads permits easy travel.

The area is about 1000 feet above sea level, but local relief is generally less than 200 feet. Outcrops are typically small and much of the bedrock is covered by glacial deposits and swamps. Road widening schemes during 1973 created some new exposures in a few localities in the southwest part of the area.

### Geological Setting and Previous Work

Although the rocks of this part of the Grenville Province (fig. 2) have been studied fairly continuously for over a

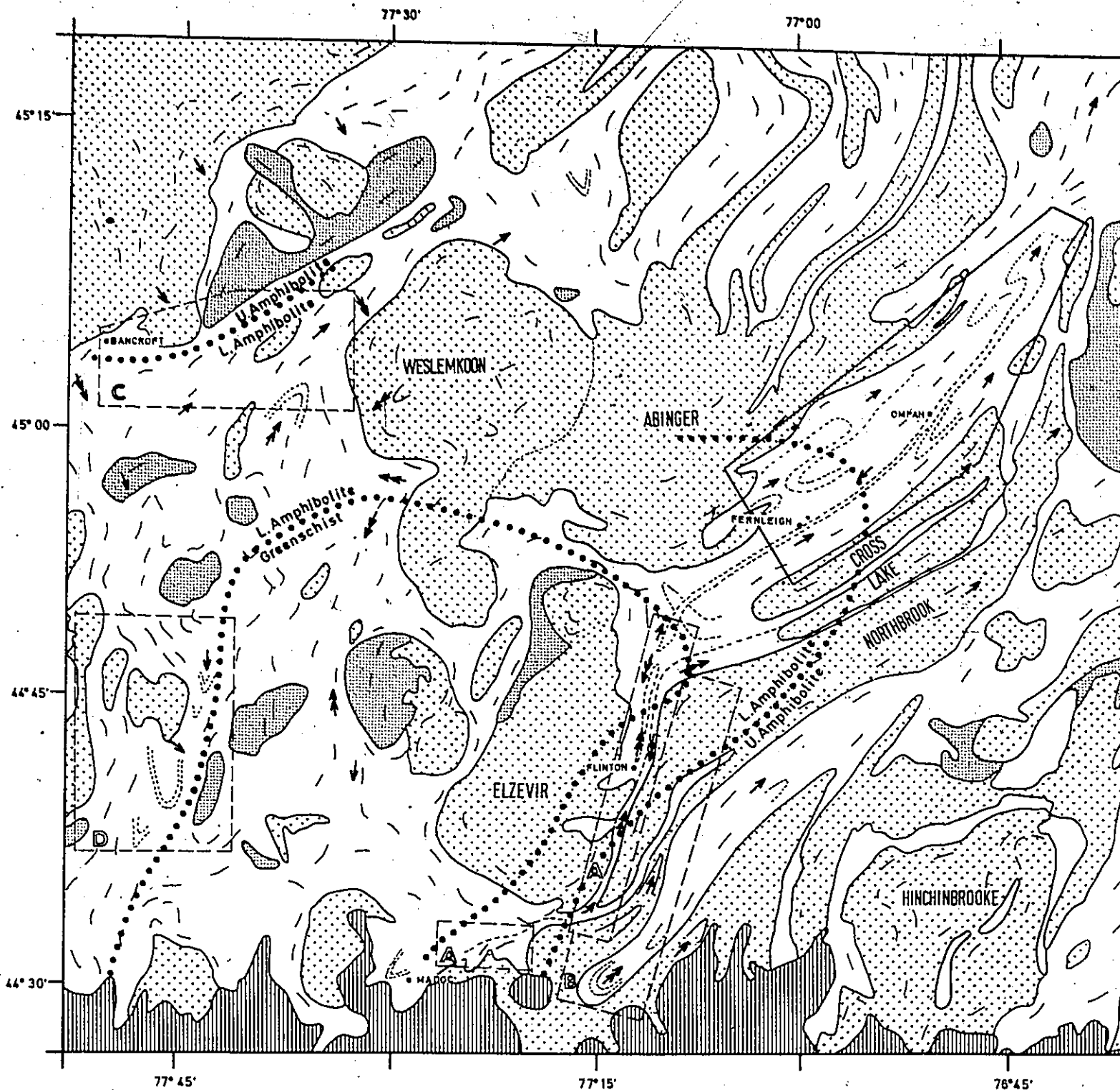
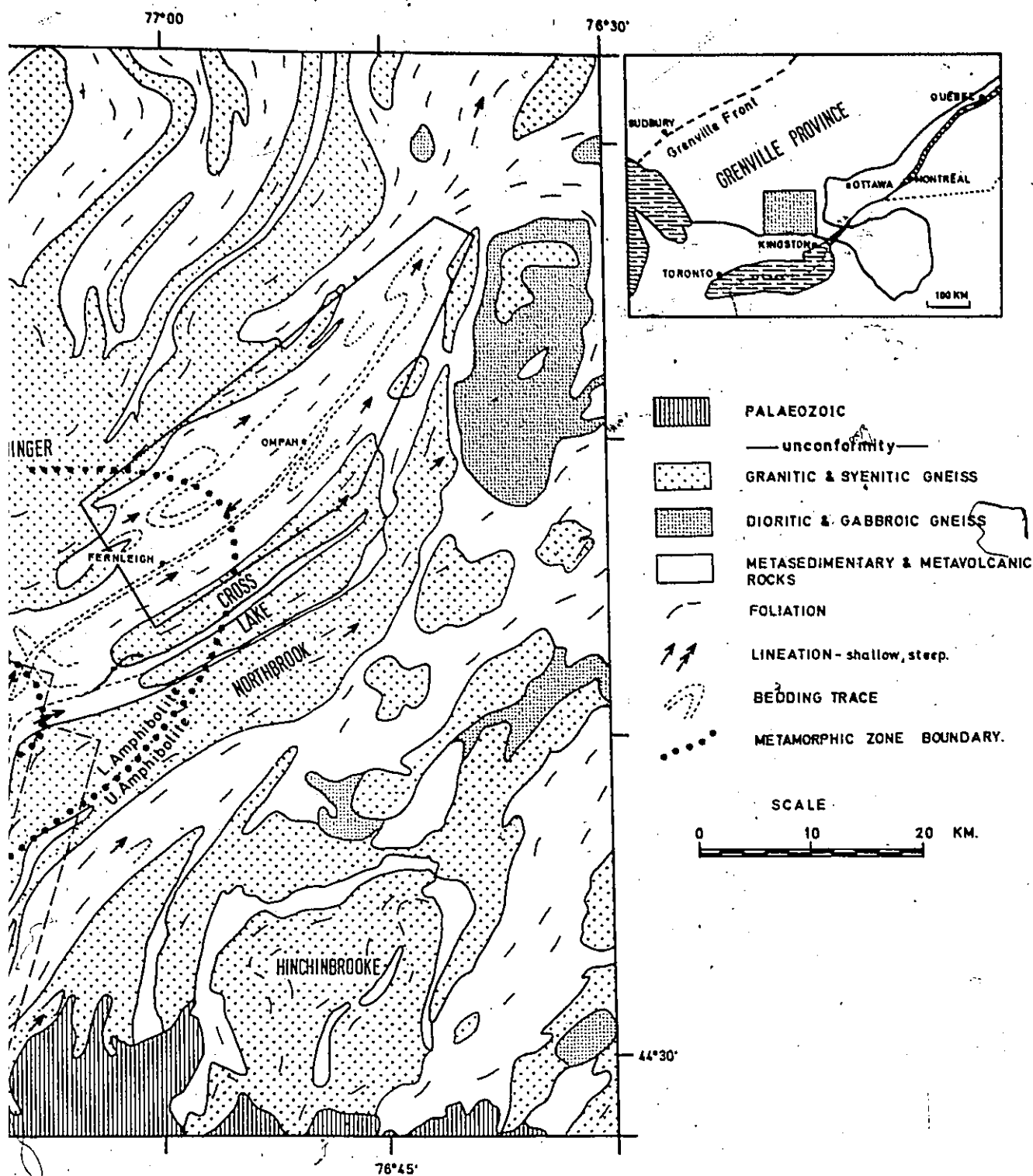


Fig. 2 Generalized geological-structural map of the Bancroft-Madoc region, mod from Lumbers (1964a, b), Carmichael (1968, area D), Venkitasubramanyan (1969, a Divi (1972, area C), Thompson (1972, area A) and Moore and Thompson (1972). It is outlined by solid line.

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p of the Bancroft-Madoc region, modified  
rea D), Venkitasubramanyan (1969, area B),  
and Moore and Thompson (1972). Study area

century, the overall stratigraphy of the region has only recently emerged in some detail, and still much remains tentative. Geologists as early as Logan (1863) recognized that there was an important boundary within the Grenville Province between the so-called "fundamental gneisses" (or "grey gneisses") of the basement and the younger metasediments and metavolcanics, now collectively known as the Grenville Supergroup, which stratigraphically overlie them. Some of the grey gneisses from near the northern boundary of the Grenville Province are reworked rocks of Archean and Aphebian ages (Emslie 1970); certain distinctive lithologies have been traced from the Superior and Churchill Provinces into the Grenville Province, where the earlier trends are altered and truncated by the penetrative northeast trending structural elements characteristic of the Grenville Province (Wynne-Edwards 1972). Isotopic age dating (Krogh and Davis 1969) has confirmed an Aphebian age (1725 myr.) for some of these rocks from well within the Grenville Province.

Over much of their outcrop the grey gneisses are gently dipping granitic and migmatitic rocks with a stratiform layering - they are considered to be basement to, and unconformably overlain by the Grenville Supergroup (Wynne-Edwards 1972). The contact between the gneisses and the cover rocks is marked by a linear zone of nepheline bearing syenites and associated rocks (eg. N.W. corner of fig. 2), but the

nature of the contact is not obvious in the field. A recent paper by Appleyard (1974) has shown that the contact zone is the site of several periods of intrusive activity, and has undergone a complex tectono-metamorphic history, which has obscured boundary relationships.

The overlying Grenville Supergroup consists of a series of metavolcanics from mafic to felsic in composition, a variety of clastic rocks and considerable carbonate. Apart from the intrusives, all the rocks in the Ompah area are assigned to this group. The rocks are Helikian in age, the oldest dated volcanics from the bottom of the sequence in the Bancroft region (where the contact with the basement is not exposed) being  $1310 \pm 15$  myr. (Silver and Lumbers 1966). Lumbers (1967a, 1967b) has proposed a comprehensive stratigraphy for the various formations of the Grenville Supergroup in the Bancroft-Madoc area. Although much of the region has undergone high grade metamorphism, a central portion around Madoc lies within the greenschist facies, and in this area it has been possible to interpret more fully the primary lithologies and textures. Lumbers has subdivided the Grenville Supergroup into the Herman Group, consisting mainly of metavolcanics, and the Mayo Group consisting of metasediments. Deposition of the two groups is thought to have been in part contemporaneous, and thus group boundaries do not represent isochronous lines. Briefly the suggested stratigraphy is as

follows: a thick sequence of pillowed, basic volcanics gives way upwards and laterally to felsic volcanics and sediments. Erosion of the volcanic rocks, especially the felsic ones, gave rise to the siliceous clastic sediments, whilst basic vulcanism continued sporadically. Much carbonate was precipitated at the same time.

Several conglomerate lenses occur within the Hermon and Mayo Groups, which in the past have been variably interpreted as representing either a major, structural unconformity, or merely a facies change and minor stratigraphic break. Lumbers (1967a) preferred the latter alternative, citing evidence that several of the lenses are at different stratigraphic levels. However recent work by Moore (1967), Sethuraman (1970), Moore and Thompson (1972), Thompson (1972) and Sethuraman and Moore (1973) has revised this interpretation for rocks in the east of the Bancroft-Madoc region, and a major unconformity has been defined. The younger rocks above the unconformity have been named the Flinton Group, after the village of Flinton, 40 kilometres southwest of Ompah (fig. 2). The unconformity surface is folded, and rocks of the Flinton Group outcrop in narrow, elongate, northeast trending synforms cross-cutting earlier lithologic contacts. Tentative correlation of these rocks over much of the eastern portion of the Bancroft-Madoc region has been documented (Moore and Thompson 1972). Near the village of Flinton (fig. 2),

1050  $\pm$  20 M.Y.

Intrusion of pegmatites. K-Ar dates  
"reset" at about this time.

1125  $\pm$  25 M.Y.

Culmination of regional metamorphism  
and deformation.

Deposition of Flinton Group.

#### UNCONFORMITY

Intrusion of potassic granite and  
quartz monzonite plutons\*

Intrusion of gabbroic, dioritic and  
syenitic plutons.

1250  $\pm$  25 M.Y.

Intrusion of trondhjemitic plutons.  
Low grade regional metamorphism and  
deformation?

1310  $\pm$  15 M.Y.

Deposition of Hermon and Mayo Groups.

#### UNCONFORMITY

Basement gneisses reworked rocks of  
Archean and Aphebian age.

---

Table 1      Inferred chronological relationships between  
deposition, plutonism and metamorphism/deformation in the  
Bancroft-Madoc area (Modified from Lumbers 1967b, Moore and  
Thompson 1972). Isotopic dates are from Krogh and Hurley  
1968 and Silver and Lumbers 1966.

\*      Emplacement of older potassic granite bodies is indica-  
ted. Other bodies are undeformed, indicating intrusion  
after regional metamorphism and deformation.

rocks of the Flinton Group rest unconformably on an erosion surface developed in intrusive rocks (the Elzevir pluton) as well as in rocks of the Herman and Mayo Groups. The lithologies of the Flinton Group have been described in detail by Moore and Thompson (1972) and Thompson (1972); they consist essentially of a series of metasediments of clastic and chemical origin, interpreted as shallow water marine and continental deposits. Volcanic rocks have not been recorded. In the Ompah area, Moore and Thompson (1972) consider that rocks of the Flinton Group outcrop in the Fernleigh belt, the Ompah synform, and in smaller proportions in the Shawenegog Lake antiform and the Plevna synform (fig. 3).

Igneous intrusion in the Bancroft-Madoc region is considered to have taken place in four distinct episodes (Lumbers 1967a, b). The earliest intrusions are those of trondhjemitic composition, and include the large Weslemkoon, Elzevir, Northbrook, Cross Lake and Hinchinbrooke plutons (fig. 2). They are considered to be epi- or meso-zonal features (Lumbers 1967b), and may have originated by mobilization and partial differentiation of grey gneisses which are basement to the Grenville Supergroup, as they show considerable chemical affinity to those rocks (Wallach 1974). The plutons are considered to be younger than the deposition of at least part of the Herman and Mayo Groups (xenoliths of volcanics from the Herman Group occur within the Elzevir batholith), but older

than the deposition of the Flinton Group, which in part rests unconformably upon them (Thompson 1972). Following intrusion of the trondhjemites, plutons of gabbro and related dioritic rocks were emplaced, with which are also associated undersaturated syenitic rocks of the region. Two small gabbro and diorite stocks are all that represent this group in the Ompah area. Subsequent to this, several large masses and many smaller bodies of potassic granite and quartz monzonite were intruded into the pile of plutonic and cover rocks. In the Ompah area this group is represented by the Abinger and Lavant plutons. Lumbers (1967b) suggests that these rocks were emplaced in the catazone before, during and after the culmination of regional deformation and metamorphism. Deposition of the Flinton Group post-dates the emplacement of at least some of these bodies, as cobbles of potassic granite are recorded in the basal conglomerate of that group (Thompson 1972). As regional metamorphism waned, pegmatite dykes were intruded in areas of high and intermediate metamorphic grade. The general age relationships between deposition, plutonism and regional metamorphism are summarized in table 1.

Northeast trending fold axes are common throughout much of the Bancroft-Madoc region, although in the northwest a southeast trending lineation is prominent. Details of the deformation and structural geometry are available from several

areas. Carmichael (1968), area D fig. 2, Venkitasubramanyan (1969), area B, Divi (1972), area C and Thompson (1972), area A, all record evidence of two major phases of deformation, and a third phase is noted by most of these authors; whilst Appleyard (1974), working near the contact between the grey gneisses and the cover rocks of the Grenville Supergroup about 50 km. northeast of Bancroft, gives evidence of at least five phases of deformation.  $F_1$  folds are typically isoclinal and may be of large scale (Carmichael, Venkitasubramanyan, Thompson), and were probably initially recumbent (Carmichael, Venkisubramanyan, Divi).  $F_2$  folds, which also vary from megascopic to mesoscopic in scale, are identified as northeast trending upright structures by most authors, and generally low plunging except in the vicinity of some plutons. (Divi 1972, Divi and Fyson 1973); however Appleyard considers  $F_2$  folds to be recumbent and southeast plunging in his area. Structures of the third deformation are sporadically developed, and may not be consistent in orientation throughout the region- however in areas C and D of fig. 2 recumbent  $F_3$  folds have been recorded. The dominant structural features in area C (fig. 2) and in the area described by Appleyard are  $F_2$  folds and associated structural elements, but Thompson considers that  $D_1$  structural elements predominate in area A. Most authors consider deformation and metamorphism were coeval, the peak of metamorphism coinciding with  $D_2$  deformation (Divi,

Thompson).

Two major metamorphic culminations are noted by Lumbers (1967b), and Moore and Thompson (1972) give evidence for both pre - and post-Flinton Group metamorphism. The earlier period ( $M_1$ ), which took place at approximately 1310 M.Y. ago was generally of lower grade (greenschist facies) than the later,  $M_2$ , (greenschist to upper amphibolite facies), which yields K-Ar dates of around 1125 M.Y. (Lumbers 1967b). However elsewhere in the Bancroft-Madoc region, evidence of only a single period of metamorphism, approximately contemporaneous with Lumbers'  $M_2$ , is recorded.

Studies by Carmichael (1968, 1969 and 1970) have given clear evidence of a metamorphic gradient from greenschist to amphibolite facies in area D, and Divi has discussed the position of the sillimanite isograd in area C. Isograds in the Fernleigh - Ompah area have been mapped by Moore (1967), and mineral equilibria in those rocks described by Hounslow and Moore (1966). Thompson (1972<sup>1973</sup>) continued similar studies to the southwest as far as Madoc, and these results have been compiled by Moore and Thompson (1972). The mineralogical and textural changes associated with metamorphism of the Cross Lake gneiss have been described by Chapman (1968) and Ehrlich et al. (1972), and Sethuraman (1970) has mapped isograds in metavolcanic rocks outcropping within and to the west of the Cross Lake pluton.

Most of the Bancroft-Madoc region is metamorphosed to the amphibolite facies. Kyanite and sillimanite are typically stable phases, indicating pressures above the aluminium silicate triple point. Sethuraman and Moore (1973) show that the facies series is between the classic Barrovian sequence in Scotland, and the Abukuma series in Japan; it is closely similar to the Idahoan series described by Heitanen (1967).

#### Scope of the Present Study

The objectives of this work are:

To investigate the structural geometry of the Ompah area and to compare the structural history with that in nearby areas in order to further define the regional tectonic history of this part of the Grenville Province.

To examine, in the Ompah area, the nature of the proposed unconformity at the base of the Flinton Group.

To determine the time relations between deformation, metamorphism and intrusion.

To determine, from an examination of crystal textures, possible mechanism of prograde metamorphic reactions.

To investigate some of the textural changes that accompany prograde metamorphism.

#### Acknowledgements

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patience and support during the project, though at times nearly exhausted, have been of indispensable aid in its completion.

## LITHOLOGY AND STRATIGRAPHY

### Introduction

The outcrop pattern of the lithologic units (fig. 1) is essentially similar to Smith (1958), though the boundaries and designation of some units have been modified. Broadly speaking the area comprises a northeast trending synform of metasedimentary and metavolcanic rocks lying between three large plutonic bodies, the Abinger, Lavant and Cross Lake gneisses.

The whole area has been metamorphosed to the amphibolite facies, and during the accompanying recrystallization the original textures were extensively modified. As a result the origin of some lithologies is not certain. Evidence of stratigraphic tops is rare, so the numbering of the units is not necessarily in stratigraphic order. Furthermore some units recur at more than one structural level, and these subdivisions have not been distinguished in fig. 1.

### Petrography

Unit 1 includes mafic to intermediate rocks with amphibole (hornblende) as the predominant mafic mineral. Outcrops of this unit define large folds around Plevna in the west and Folger in the east of the map area; many smaller bodies also occur. Unit 1a is amphibolite and plagioclase-hornblende gneiss, which occurs most commonly as small bodies

within marble. Unit 1b is restricted to a few outcrops immediately south of Plevna Lake; it is a metabasalt, which contains relict phenocrysts of plagioclase feldspar. Unit 1c contains significant biotite as well as amphibole. Unit 1d resembles unit 2, except that it does not contain inclusions. It has been grouped in unit 1 because of its association with rocks of this unit in the fold near Plevna, where the boundary relationships have not been mapped in detail. Unit 1e, chlorite-<sup>amphibole-</sup>quartz-feldspar schist, has limited outcrop in the area. An easily accessible outcrop occurs on the road between Ompah and Sunday Lake, in the centre of the Ompah synform\*. The origin of many of the rocks in unit 1 is uncertain. Whilst unit 1b, which contains relict phenocrysts, is indubitably a metavolcanic, the mineralogy of other rocks may have been significantly affected by neighbouring lithologies during metamorphism, eg. the common presence of amphibolite layers within marble may be due to the high partial pressure of  $\text{CO}_2$  during metamorphism rendering hornblende rather than biotite the stable phase. For an extensive discussion of the origins of mafic rocks see Jennings (1969).

Unit 2 is a quartzo-feldspathic hornblende, biotite garnet gneiss, frequently with a fine layering (< 1 cm) and

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\* for the location of this and other structural features, see figs. 3 and 5. The Ompah synform and the Fernleigh belt are composed of rocks ascribed to the Flinton Group (Moore and Thompson 1972) see p.5.

with inclusions of dioritic composition less than 1 m. in diameter. Rocks of this unit, are restricted to the Cross Lake antiform. Sethuraman and Moore (1973) consider that the lithology is that of a fragmental felsic rock. The layering, where present, is tightly folded and transposed, so that its origin (whether volcanic or metamorphic) is uncertain; probably it is a combination of both. Minor intercalations of pelitic and psammitic rocks, some alumino-silicate bearing, occur sporadically throughout the unit.

Unit 3. Conglomerate - this is subdivided into three units on the basis of the composition of the pebbles and the matrix. 3a, which is best exposed in the Ompah synform and Fernleigh belt, is a polymictic conglomerate with a variable matrix. Pebbles are of quartzite (generally fine grained, buff to dark grey in colour, the lighter coloured varieties sometimes possessing a layering,  $S_0$ ), granite, quartz (coarse grained - vein quartz?), marble, biotite schist, and amphibolite (the latter two compositions are often difficult to distinguish from the matrix in unpolished specimens). Granite pebbles, though variable in composition, typically contain 20 - 40% K. feldspar, and thus bear a closer resemblance to the Abinger and Lavant gneisses (unit 7a) than to the Cross Lake gneiss (unit 7b). In any one outcrop a certain variety of clast may be dominant; detailed mapping of the Ompah synform indicates that marble pebbles are more common in the

eastern part of the synform than the west, but no other consistent areal variations of pebble composition were noted.

The composition of the matrix in unit 3a is also variable. Where biotite is the predominant matrix mineral the rock is schistose, and garnets are not uncommon. Hornblende predominates over biotite in some parts, so that the matrix resembles unit 1c (an indication that all the rocks in unit 1 are not volcanic in origin). Less commonly the matrix is essentially an impure quartzite, with less than 20% mafic minerals. Bedding in this unit is generally absent on the scale of the outcrop; however pebble-free bands are occasionally seen and quartzite lenses are not uncommon in the centre of the Ompah synform. Thin lenses of marble are infrequent.

Pebble content of unit 3a varies from around 90% to zero (pebble free layers). Pebble size is another variable - the longest dimension of most pebbles is less than 30 cm., but in the Ompah synform there are a few boulders with the largest dimensions >1 m. Small pebbles are difficult to recognise in the field because deformation and recrystallization render borders indistinct. Smith (1958) noted that there is a general decrease in pebble size from west to east in the Ompah synform. Two other general observations were made of the conglomerate in the Ompah synform: (a) the composition of clasts of boulder size is invariably of granite, and the matrix is of biotite schist; and (b) that where clasts are

predominantly of calcite, the matrix is rich in hornblende.

Unit 3b is a boulder conglomerate, found in mappable units of limited extent only in the Plevna Lake area, where it is associated with unit 3a. It differs from unit 3a in that all the clasts in the rock are of boulder size, whereas in unit 3a where boulders occur, a variety of smaller clast sizes is also present. The boulders are of granite, and the matrix, which comprises less than 20% of the rock, is a pelitic schist, similar to unit 6b. According to Smith, Vogel and Spence (1969) the composition of the clasts is generally low in K. feldspar, and thus different from the underlying granite, though this was not verified in the present study.

Unit 3c, a carbonate conglomerate, is present in the Fernleigh belt as far east as the closure of the Ompah synform where it grades into unit 3a, and on the south limb of the Plevna synform. The matrix is carbonate (predominately calcite) with minor biotite, quartz, phlogopite and calc-silicate minerals. The clasts are of calcite, dolomite, quartz and calc-silicate minerals, the latter generally being dominant. Most pebbles are less than 30 cm. greatest dimension; bedding may be recognized in places as pebble-free layers (plate 1).

Unit 4a, buff to light grey quartzite, is confined as a mappable unit to the Ompah synform. Some of the quartzites

are cross-bedded (plate 2) and have been used for top determinations (see also Smith 1958, 1969); others however are too deformed to be useful for "way up" criteria. Typically the quartzite contains less than 10% feldspar, and minor biotite and muscovite. Outside the Ompah synform impure quartzites, containing more than 20% biotite, hornblende and calcite are common, and unit 4a is gradational with 6a, biotite-quartz schist. Unit 4b is calcareous quartzite. Common in small lenses throughout the area, it is mappable only in the centre of the Plevna synform and in the Clyde Forks region. Boundaries between this unit and marble (unit 5) are often difficult to locate in the field, but in several places they are defined as ridges on aerial photographs.

Marble, unit 5, is the most widespread rock type in the area. It includes both calcitic and dolomitic varieties, the former predominating. The rock varies in colour between white, buff and grey-blue, with pink and green shades being less common. Frequently a layering of a few centimetres width is present, caused by the segregation of silicate minerals such as tremolite, phlogopite, quartz, scapolite diopside and less commonly actinolite. Other minerals such as graphite and pyrite are locally common, and also generally segregated into discrete layers. Recrystallization during metamorphism and deformation has obliterated the original texture of the rock, so it is not known whether the segregation layering

is of metamorphic or sedimentary origin.

Unit 6 comprises schists and semischists considered to be of sedimentary or tuffaceous origin. 6a, which occurs as large elongate bodies north of the Fernleigh belt, is a biotite-quartz-plagioclase feldspar rock, which may contain hornblende, calcite, muscovite, K. feldspar and minor tourmaline. Unit 6b is a pelitic schist occurring at the northern boundary of the Fernleigh belt, and on the southern limb of the Plevna synform-its distinctive appearance is of considerable use in tracing the stratigraphy of the Fernleigh belt northeastwards in the Clyde Forks region, where the units become attenuated. It is frequently porphyroblastic. The principal matrix mineral is muscovite, though typically some biotite is present. Other matrix minerals are quartz, plagioclase and a variety of opaque oxides (magnetite, haematite and ilmenite being the most common); tourmaline is ubiquitous, though rarely exceeding 2% of the rock. A variety of porphyroblastic minerals may be present, some of which serve as indicators of metamorphic grade-these include plagioclase, biotite, staurolite, garnet, kyanite, sillimanite and intergrowths of magnetite and ilmenite. At high grades of metamorphism this unit is typically coarse grained, another feature which helps to distinguish it from adjacent units. Unit 6c is biotite-quartz-carbonate schist which outcrops in the Fernleigh belt. Amphibole may replace biotite in places,

and the rock possess a distinctive lamination in which biotite and hornblende are segregated from calcite and other minor felsic phases. Though the layering itself is probably of sedimentary origin, it has been repeated by small scale intra-folial folding, so that it is now a composite feature. Unit 6d also occurs in the Fernleigh belt. It is a biotite-quartz schist containing abundant pyrite, together with minor calcite and plagioclase feldspar. A distinctive unit in the southwest of the area, it appears to die out northeast of Ardoch.

Unit 7 consists of rocks of granitic (sensu lato) composition. 7a is a pink to white granite gneiss that occurs over much of the north and northeast parts of the map area. The mass to the north is named the Abinger granite gneiss after a township nearby, in which it is the dominant rock type. In the east it is referred to as the Lavant gneiss, following Smith (1958). The two bodies are essentially similar in composition, being potassic granites with a fairly consistent K. feldspar; plagioclase feldspar ratio of 2:1, and a quartz content of 20 - 35%. Biotite and muscovite together average about 5%, and opaques (usually magnetite) 2 - 3% (figures are from modal analyses by Smith, 1958). An indistinct layering caused by the segregation of biotite and opaques is commonly present, which defines a foliation. The grain size of the rock is variable, though generally in the

fine-medium range. Grain contacts show evidence of deformation and recrystallization; inter- and intragranular recrystallization of quartz and feldspar has taken place, and sutured boundaries between crystals are common. The margins of the Abinger body are frequently migmatitic (unit 7c), both metasedimentary and metavolcanic rocks being invaded by the granite. North of Palmerston and Canonto Lakes granite grades through an intermediate, banded rock containing about 40% biotite, amphibole, sphene and calcite to an amphibole-biotite gneiss. Feldspars are frequently corroded at the edges and distorted within, and some amphiboles have an alteration zoning.

Unit 7b is a grey to white massive rock of trondhjemitic composition which outcrops in the Cross Lake antiform. Typical mineralogy is plagioclase feldspar 50%, K. feldspar 5%, quartz 25 - 30%, biotite 10 - 15%, with minor hornblende, muscovite, calcite and epidote. Chapman (1968) has studied the metamorphism of this unit, and reports that muscovite, calcite and epidote present in the western part of the unit are replaced by hornblende in the east. Plagioclase composition, though variable ( $An_{18-36}$ ), is not an indicator of metamorphic grade, being dependent upon bulk composition. At the northeast end of the unit (high metamorphic grade), the rock is medium to coarse grained, and possesses a well developed lineation; towards the west, knots of mafic minerals occur

where Chapman reports relics of an original igneous texture.

7d. Coarse grained post-tectonic granitic rocks with an obvious igneous texture. Field examination shows that these rocks vary considerably in composition, but they were not studied in thin section. Most large outcrops are concentrated north of Palmerston Lake, although small outcrops are common over much of the area.

Unit 8 consists of gabbroic and dioritic rocks. These rocks, whilst present as only two small bodies in the Ompah area, outcrop as a large mass directly to the east of the map. 8a is metagabbro to meta-quartz gabbro, consisting of plagioclase and hornblende, with quartz, garnet, hypersthene and biotite present in some specimens. Plagioclase composition is generally between  $An_{30-40}$ , although  $An_{60}$  was recorded in some specimens by Smith (1958). Remnants of hypersthene crystals are typically mantled by hornblende, and a sub-ophitic texture is discernable in the least deformed specimens. In the central portion of the body situated in the axial region of the Plevna synform, partial alignment of hornblende and aggregates of feldspar produces a faint lineation in some specimens. Towards the margins of the body, near the contacts with amphibole-biotite gneiss (unit 1c), hornblende and biotite define a foliation in the gabbro, and it is in places difficult to distinguish the gabbro from unit 1c. Contacts with unit 1c are concordant on the map scale, but within a single outcrop are typically irregular, with blocks

of strongly foliated, micaceous rocks of unit 1c occurring well inside the gabbroic body; locally a brecciated border is apparent. Unit 8b, diorite to granodiorite, is intermixed with unit 8a in many outcrops; the gabbro of unit 8a is generally the dominant rock type, with a variable degree of veining of diorite and granodiorite. Peach (1958) has described a complete gradation between gabbro and granodiorite to the east of the study area (fig. 2).

### Stratigraphy

#### Correlation with the Flinton Group

The stratigraphic columns drawn by Moore and Thompson (1972) for rocks of the Flinton Group outcropping in the Flinton, Bishop Corners and Ompah synforms and the Fernleigh belt show considerable vertical variation in lithology as well as lateral facies changes. The basal unit in each case is, however, a metamorphosed quartz sandstone, conglomerate, or aluminous shale. Metaconglomerate and pelitic schist occur northwest of the Fernleigh belt in the Plevna area (units 6b and 3c in the Plevna synform and units 3a and 3b in the Shawenog Lake antiform). These rocks are assigned to the Flinton Group by Moore and Thompson (1972), but as can be seen from the cross sections (fig. 4), there is no basis for assuming that these units are at the same stratigraphic level as the similar lithologies in the Fernleigh and Ompah belts — the justification for any correlation is solely lithological. ○

The interpretation made here supports the conclusion of Lumbers (1967a) that the conglomerates are intraformational, and do not rest on an unconformity of regional significance.

Furthermore the Flinton Group as defined by Moore and Thompson contains only sedimentary rocks of fluvial, deltaic or shallow marine origin; no volcanics have been included. However rocks of definite (unit 1b) and possible (units 1a, 1c and 1d) volcanic origin occur in the Shawenegog, Plevna and Ompah fold closures structurally and possibly stratigraphically above metasediments of the Flinton Group. Logically, therefore, it would appear that these lithologies should also be assigned to the Flinton Group. If this is followed the Flinton Group may be very extensive, and the rocks difficult to distinguish from the underlying Hermon and Mayo Groups.

#### Possible Unconformity at the Base of the Flinton Group

Rocks assigned to the predominantly volcanic Hermon Group by Moore and Thompson (1972) comprise most of the rocks in unit 1 (fig. 1), whilst the large masses of marble (unit 5) and biotite-quartz-feldspar schist (unit 6a) may be part of the sedimentary Mayo Group. It is of interest to compare the basal contact of Flinton Group rocks in the Ompah area with the same contact from the type locality. Near the village of Flinton, granite pebbles from the basal conglomerate of the Flinton Group have been correlated with the underlying

rocks, and 16 km. northeast of Flinton, rocks of the Flinton Group cut across earlier lithologic contacts at a high angle (Sethuraman 1970, Thompson 1972). In the Ompah area the evidence is less convincing. Smith (1969) states that relatively simply deformed rocks within the Ompah synform overly with angular unconformity more complexly deformed rocks. However no evidence of either a contrasting deformational history or an angular unconformity were found in the present study. Contrasts in tectonic style are due to the difference in rheological properties of the rocks above and below the base of the Flinton Group (see Clifford and Rector, 1970, for a discussion of lithologic variation in strain response), and there is no evidence of earlier structures in the older rocks.

An apparent truncation of lithologic contacts at the base of the Flinton Group at the northeast end of Sunday Lake (Smith 1958, 1969) has been mapped in detail; it appears to be due to the attenuation of certain units (3a, 6a and 5) causing a convergence of contacts. The apparent cross-cutting contacts of the rocks of the Fernleigh belt with older rocks in the Clyde Forks area (Smith 1958) is also not substantiated (see fig. 1). Smith (1958) additionally reported unconformable contacts at the base of the Plevna synform, which again are not apparent from detailed mapping.

The geology of the Shawenog Lake antiform has been

discussed by Smith, Vogel and Spence (1969). At this locality granite boulder conglomerate and polymictic conglomerate overly a dome of granite gneiss (unit 7a). Although there is a superficial resemblance between boulders and underlying granite, modal analyses by the authors indicate significant mineralogical differences. Furthermore they point out that there is veining in the metasediments and a mineralogically distinct border facies in the granite, indicating that the granite was intruded into older sediments.

Thus there is no direct evidence for an unconformity. However it could be present, but obscured by the lack of outcrop and by parallelism of lithologic contacts imposed by several periods of coaxial deformation.

## STRUCTURAL GEOLOGY

### Introduction

#### Structural Summary

In the Ompah area two major deformational phases\* are recorded in the rocks, both of which produced a variety of structural elements such as foliations, lineations and folds. Evidence of a third, minor episode is also sporadically seen. These have been distinguished from each other on the basis of character, orientation and inter-relationship, no single criterion being adequate, table 2.

The first deformation, D1, gave rise to an inclined to sub-horizontal penetrative schistosity with accompanying inclined to recumbent NE-SW trending folds and lineations. A locally developed mylonitic layering may also have been formed at this time. These D1 structures were subsequently deformed during a second episode, D2, about vertical or steeply SE dipping axial surfaces, producing, folds with axes horizontal or gently plunging NE-SW. Third phase structures are sporadically developed, and do not occur in the majority of outcrops.

#### Difficulties in Interpretation

In unravelling the deformational history of metamorphic rocks, recognition of the nature of the folded surface is fundamental. In the area studied, in which penetrative recrystallization was synchronous with the first deformation,

\* see footnote p. 29

| OBSERVED<br>STRUCTURAL<br>ELEMENTS | DESCRIPTION OF FOLDS                                                                                                                                                                   |                          |                      | MINERAL<br>ALIGNMENT<br>PARALLEL TO<br>FOLIATION &<br>LINEATION   | FOLD GEOMETRY                                                        |                                                          | METAMORPHIC<br>GRADE                        |
|------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------------------|----------------------|-------------------------------------------------------------------|----------------------------------------------------------------------|----------------------------------------------------------|---------------------------------------------|
|                                    | SCALE*                                                                                                                                                                                 | INTERLIMB.<br>ANGLE**    | CLASS                |                                                                   | ORIENTATION<br>OF AXIAL<br>SURFACE                                   | ORIENTATION<br>OF FOLD<br>AXIS                           |                                             |
| D <sub>3</sub>                     | F <sub>3</sub> folds<br>S <sub>3</sub> Axial surface<br>S <sub>3</sub> Fracture cleavage<br>L <sub>3</sub> crenulation                                                                 | OPEN<br>TO<br>TIGHT      | 1B,<br><br>1C        | Absent                                                            | Strike:<br>040-050°<br>dip:<br>SE - NW<br>Gentle                     | Trend:<br>NE<br>Plunge:<br>sub-horizontal                | LOWER TO<br>MIDDLE<br>AMPHIBOLITE           |
|                                    | F <sub>2</sub> folds<br>S <sub>2</sub> foliation<br>S <sub>2</sub> axial surface<br>L <sub>2</sub> crenulation                                                                         | OPEN<br>TO<br>ISOCLINAL  | 1B,<br><br>1C,<br>1A | Absent, except<br>in siliceous<br>marble and<br>mylonites         | Strike:<br>040-060°<br>dip:<br>SE<br>Steep to<br>subvertical         | Trend:<br>NE<br>Plunge:<br>gentle<br>sub-horizontal      | MIDDLE TO<br>UPPER AM-<br>PHIBOLITE         |
| D <sub>1</sub>                     | F <sub>1</sub> folds<br>S <sub>1</sub> foliation<br>S <sub>1</sub> mylonitic layering<br>L <sub>1</sub> intersection of S <sub>0</sub> and S <sub>1</sub><br>L <sub>1</sub> elongation | TIGHT<br>TO<br>ISOCLINAL | 1C,<br>2,<br>3       | Present in all<br>rocks except<br>quartzite and<br>massive marble | Strike:<br>030-050°<br>Dip:<br>SE - NW<br>moderate to<br>subvertical | Trend:<br>NE<br>Plunge:<br>sub-horizontal<br>to moderate | GREENSCHIST<br>TO<br>UPPER AM-<br>PHIBOLITE |

Table 2. Summary of structural geology. For definition of terms see \* Turner and Weiss 1963;  
\*\* Fleuty 1964; \*\*\* Ramsay 1967.

Footnote to p. 28

\* A deformation phase is characterized by a distinct set of fabric elements. The words "phase" and "event" denote the time sequence of deformation, but do not necessarily imply that a significant time interval existed between D<sub>1</sub> and D<sub>2</sub> etc.

the original character of the folded surface is in places difficult to ascertain. Interpretations are also hampered by transposition of the folded surface into the axial surface foliation. Additionally linear structures are not readily assigned to the relevant phase of deformation because  $F_1$ ,  $F_2$  and  $F_3$  folds are all coaxial. Thus some individual structures remain enigmatic.

#### Analytical Technique

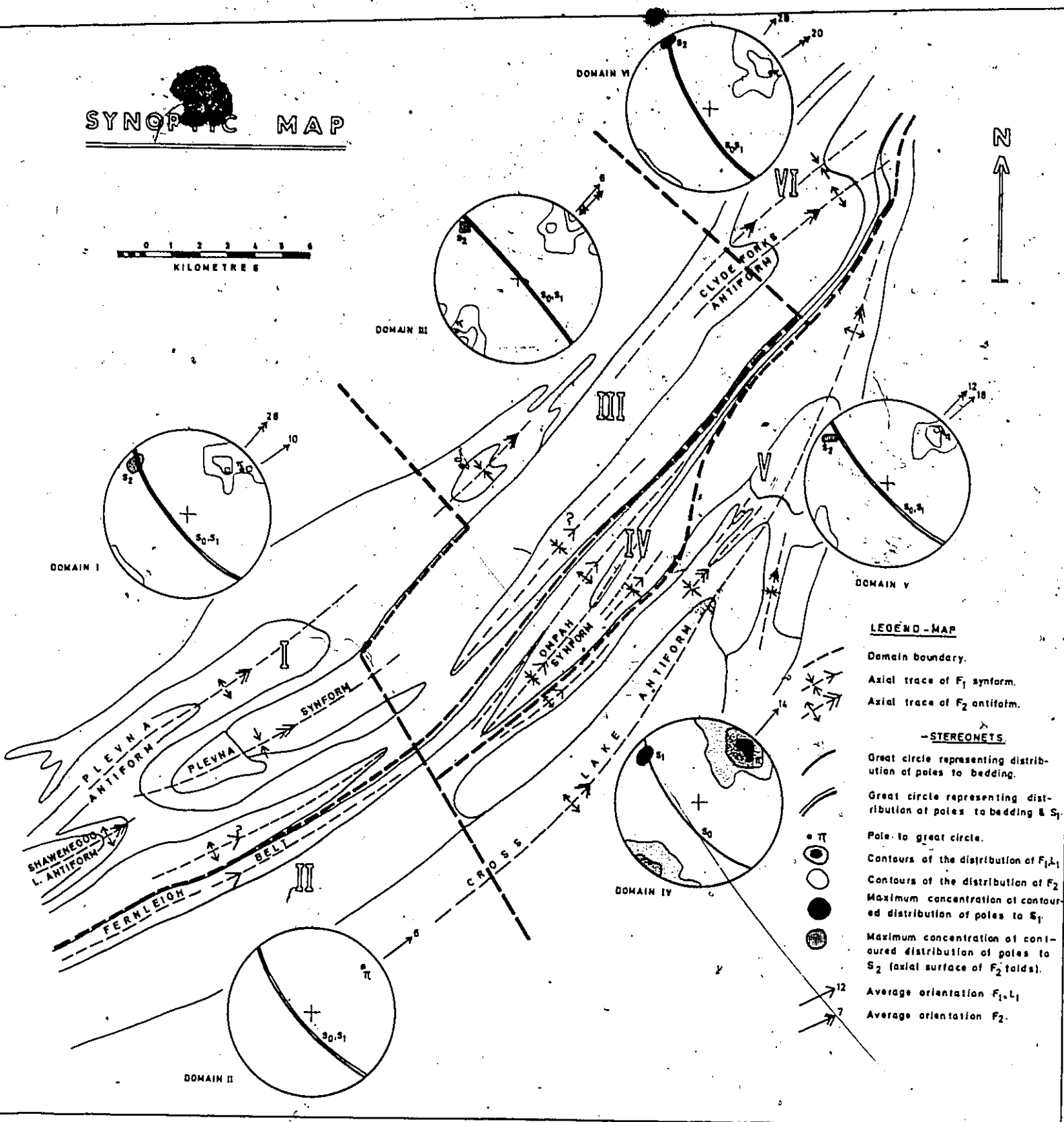
The structural map (fig. 3, in pocket) includes a representative selection of the measurements made in the field. The map has been divided into six structurally homogeneous domains. Originally nine such domains were delineated, all the rocks of the Flinton Group outcropping in the Fernleigh belt and the Ompah synform being separated from those that might lie unconformably beneath them. However stereograms of Flinton Group rocks include structures that appear identical in orientation to those in the adjacent rocks, giving no evidence for an angular unconformity; so the three domains containing Flinton Group rocks were amalgamated with adjacent domains. Several large-scale fold closures, which are apparent on aerial photographs, have been isolated into separate domains. The interpretation of the structural data is summarized in fig. 5.

#### D1 Structures

First phase structures are the dominant and most penetrative in the area; folds ( $F_1$ ), an axial surface foliation

# SYNOPTIC MAP

0 1 2 3 4 5  
KILOMETRES



## LEGEND - MAP

- Domain boundary.
- Axial trace of  $F_1$  synform.
- Axial trace of  $F_2$  antiform.

## -STEREONETS

- Great circle representing distribution of poles to bedding.
- Great circle representing distribution of poles to bedding &  $S_1$ .
- Pole to great circle.
- Contours of the distribution of  $F_1, L_1$ .
- Contours of the distribution of  $F_2$ .
- Maximum concentration of contoured distribution of poles to  $S_1$ .
- Maximum concentration of contoured distribution of poles to  $S_2$  (axial surface of  $F_2$  folds).
- Average orientation  $F_1, L_1$ .
- Average orientation  $F_2$ .

Fig. 5 Synoptic structural map of the Ompah area showing axial traces of major folds, and interpretation of stereonet data (see fig. 3).

( $S_1$ ), and an axial lineation ( $L_1$ ) are present, the foliation and lineation occurring in most outcrops.

#### Distribution and Orientation of $F_1$ Folds

The Ompah synform, which comprises the greater part of domain IV (fig. 3, in pocket) is defined by the outcrop pattern of conglomerate (unit 3a) and intercalated quartzite (unit 4a) layers. The only large-scale  $F_1$  fold that has been positively recognized, it has a wavelength of 5-6 km. and an amplitude of similar proportions (see cross-section, fig. 4, in pocket). Smith (1958) considered the fold to be a syncline from top determinations in cross-bedded quartzite layers eg. plate 2, but detailed mapping shows that the top determinations are ambiguous in some areas (see fig. 3), and elsewhere cross-beds are too deformed to be useful, so that the structure is better considered as a synform than a syncline. The detailed geometry of the Ompah synform is not completely understood, since exposure is sparse in critical areas. Overall the structure has a gentle plunge towards the northeast, and the axial surface is inclined steeply to the southeast. However it is apparent that there are several minor or parasitic folds (plate 3) superimposed on the main synformal structure; furthermore bedding-cleavage relationships indicate that portions of the southern limb are overturned to the northwest.

Immediately southeast of the Ompah synform and also part

of domain IV, there is another smaller synform in marble (unit 5) and ~~hornblende-biotite gneiss~~ (unit 10). Although exposure in this region is poor, it is concluded from the penetrative character of the axial surface foliation that this is also an  $F_1$  structure. Thus the intervening antiform must be very tightly appressed, as no evidence of a fold closure was found. Northwest of the Ompah synform is a narrow continuous linear feature defined by a distinctive series of lithologies consisting of polymictic conglomerate (unit 3a, grading southwest into carbonate conglomerate, unit 3c), calcareous biotite schist (unit 6c), pyritic schist (unit 6d, which dies out northeastwards), pelitic muscovite schist (unit 6b) and intercalated marble (unit 5). This feature, called the Fernleigh syncline by Moore and Thompson (1972), is referred to here as the Fernleigh belt. It has been identified as a  $D_1$  syncline about 30 km. southwest of Ompah, where the metamorphic grade is in the chloritoid-staurolite zone. At that locality its bilateral symmetry is evident, and the synclinal form is indicated by top determinations in cross-bedded quartzites. However to the northeast the bilateral symmetry of the structure is lost, and to account for the different metasedimentary sequences on either limb, Moore and Thompson suggest that there is a series of facies change across the unexposed hinge region. An alternative explanation, considered more appropriate in view of

the discontinuous nature of many of the small folds in the region, is that the syncline dies out northeastwards, in the Ompah area being replaced by a steeply southeast dipping, overturned sequence devoid of large-scale structures. The thinning of the outcrop of the Fernleigh belt to the northeast of Ompah may be in part related to this change of structure. Unfortunately neither hypothesis can be satisfactorily verified from the available field evidence.

Other large-scale  $F_1$  folds are more obscure. For instance the outcrop pattern of biotite schist (unit 6a) to the north of the Fernleigh belt suggests  $F_1$  folding, but bedding was not observed in the hinge region, and the plunge of  $L_1$  and small scale  $F_1$  folds, which might be used to assess the plunge of the major structure, is variable in this unit (fig. 3). In the quartzo-feldspathic ~~hornblende-biotite gneiss~~ (unit 2) on the north limb of the Cross Lake antiform, small  $F_1$  folds are present in many outcrops, but it is not clear whether the unit, which lacks mappable subunits, has been folded on a large scale or merely flattened. And between unit 2 and the conglomerate (unit 3a) of the Ompah synform a northeasterly closing fold is indicated on aerial photographs. Marble (unit 5) and calcareous quartzite (unit 4b), with minor amphibole gneiss (unit 1c) are exposed in this region. The fold was traced with difficulty in the field. It is defined by layers ( $S_0$ ) of marble and calcareous quartzite

near the Mississippi River. The shape of the fold to the northeast is not known.

Mesoscopic folds are common in the Ompah area, and some of these can be positively identified as  $F_1$  structures, particularly those situated in the hinge zones of large  $F_2$  folds, where the axial planes of the two sets of structures are divergent. Small intrafolial folds (wavelength less than 15 cm., amplitude less than 30 cm.) lying in  $S_1$  are present in thinly layered rocks, for example hornblende-biotite gneiss, although they are recognised only with difficulty in the field, due to their appressed form and to the discontinuous nature of the layering (fig. 11).

#### $F_1$ fold profiles

Small-scale  $F_1$  folds are generally tight to isoclinal, with angular hinges in schistose rocks. In ductile marbles the original layering has typically been completely transposed into  $S_1$ . Folds in competent rocks such as quartzite have larger hinge areas, and the hinges are rounded; nevertheless these folds too, are isoclinal. The true profile of the Ompah synform, constructed from the map pattern of the conglomerate (unit 3a) and quartzite (unit 4a), is depicted in fig. 6. The fold in conglomerate is of a class 3 type (though it closely approaches a class 2, or similar type), whilst that in quartzite is class 1c (Ramsay 1967). The percentage flattening of the quartzite (calculated by

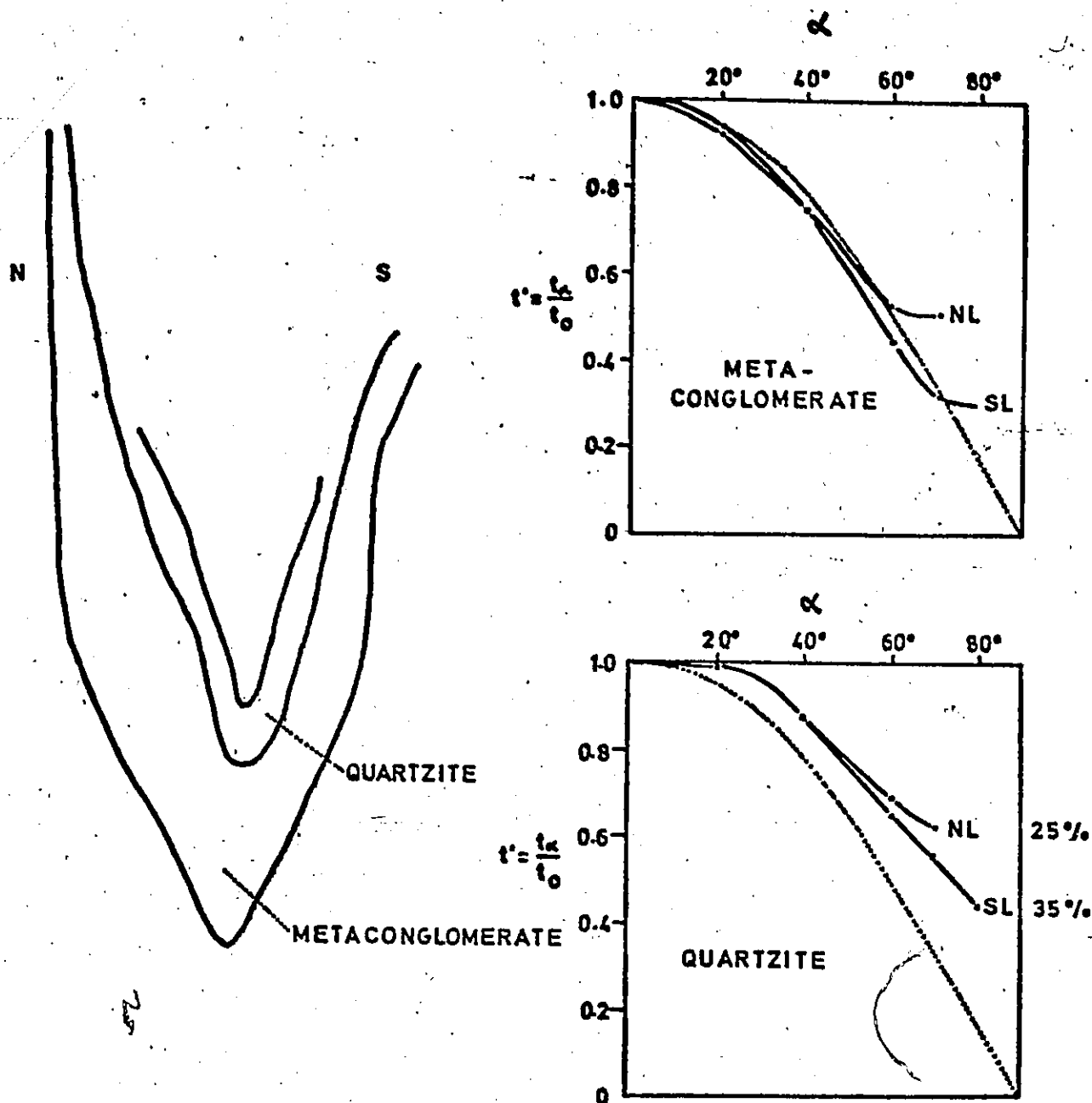


Fig. 6 True profile of the map pattern of metaconglomerate (unit 3a) and quartzite (unit 4a) in the Ompah synform, with graphs of  $t'$  against  $\alpha$  (Ramsay 1967) for the north and south limbs (NL and SL) of each unit. The percentage flattening of the quartzite unit is shown at the right of the graph.

the method of Ramsay 1962) is also shown in fig. 6. In both the conglomerate and quartzite units the south limb is the more deformed.

A road cut one and a half kilometres west of Ardoch in biotite-carbonate schist (unit 6c) of the Fernleigh belt reveals  $F_1$  folds which display a great variation in fold shape (fig. 7). An open antiform in which  $S_0$ , defined by both compositional and grain size variations, is the dominant foliation, is flanked on both sides by rocks in which  $S_0$  has been partly or completely transposed into  $S_1$ , and where folding is tight. In the central portion of the gently south-west plunging antiform (fig. 7B), bedding is cut by a fine cleavage  $S_1$ ; the bedding planes are not disrupted, but thin layers of calcite parallel to bedding are crenulated about the  $S_1$  surface. (In some layers  $S_1$  is marked by a series of calcite-filled tension cracks, but development of these must have been after folding and cleavage formation). On the northerly limb (fig. 7A),  $S_0$  is rotated into  $S_1$  in restricted zones between which it defines open, asymmetrical folds (class 1B, visual estimate). Cleavage is penetrative throughout. Relics of  $S_0$  are present in some zones where  $S_1$  is dominant, but not in others. On the south side of the fold (fig. 7C), the folds are of a smaller scale, isoclinal, class 1C and considerably flattened ( $> 30\%$  flattening-visual estimate). The variation in fold style is due in part to differences in thickness and rheology of the dominant member

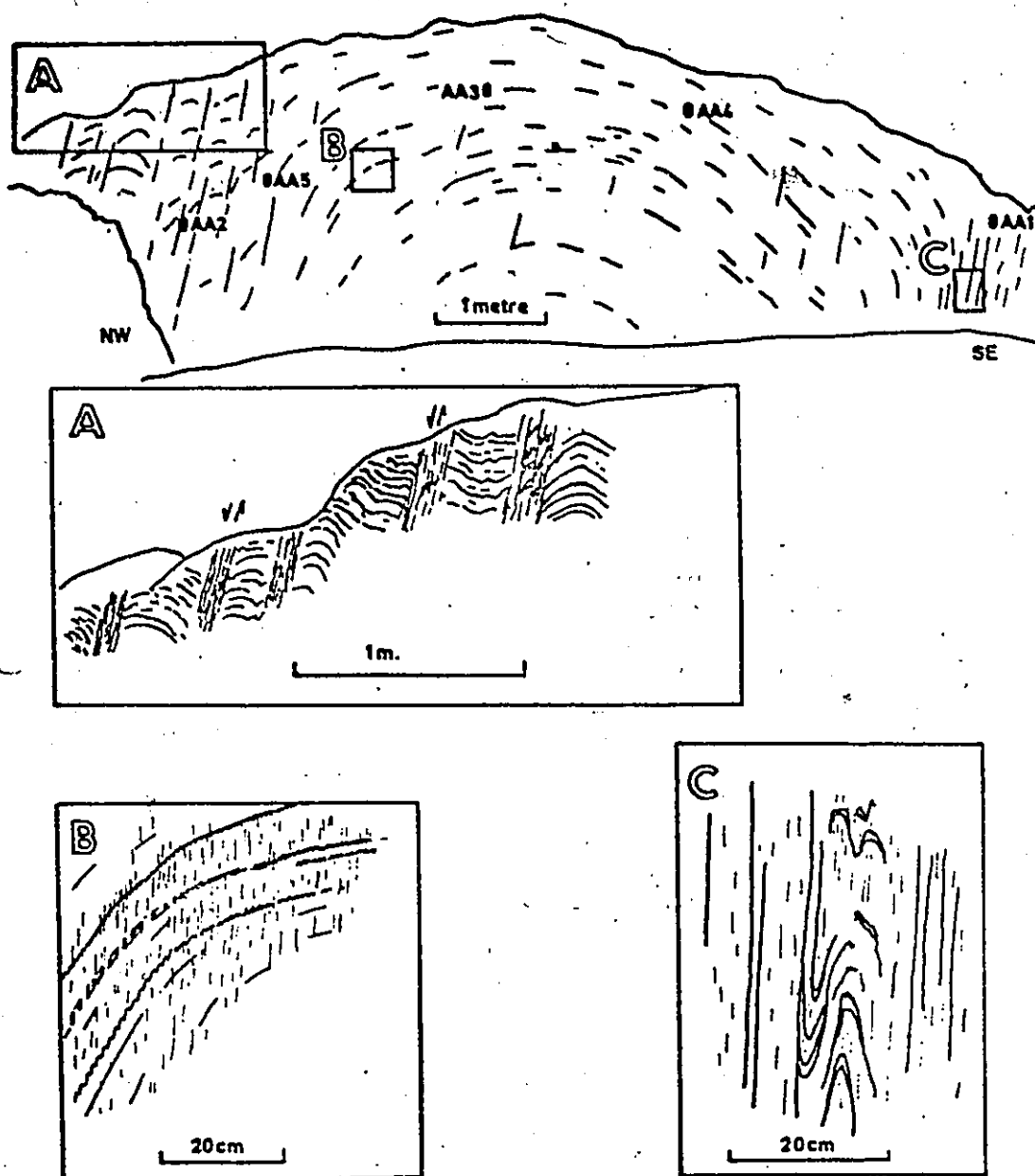


Fig. 7 Sketch of a mesoscopic  $F_1$  fold in layered biotite-quartz-carbonate schist (unit 6c) in a road outcrop near Ardoch (AA in fig. 3). The style of folding and degree of transposition of bedding into  $S_1$  foliation is variable about the fold. (Drawn from photographs). The numbers (AA1 et seq.) indicate the locations of specimens used for strain measurements based on the shape of tremolite-quartz aggregates.

during folding and to the different scale of folding. However it illustrates that some parts of a structure may appear considerably less deformed than others, and indicates that strain during  $D_1$  was inhomogeneous, even over small distances (see also estimates of strain from shapes of aggregates, p. 70).

### $S_1$ foliation

Associated with the  $F_1$  folds is an axial planar foliation  $S_1$ , generally penetrative, which due to the isoclinal form of most  $F_1$  folds is parallel to  $S_0$ , except in hinge areas. The combined  $S_0$ ,  $S_1$  foliation is defined by the crystallographic preferred orientation of mechanically anisotropic minerals (micas, amphiboles) and by the platy shape of other minerals with a less obvious mechanical anisotropy eg. quartz, feldspar, calcite, opaques.

The character of  $S_1$  in the fold hinges varies considerably and is in part dependent on lithology. For example in the Ompah synform, a strong axial planar foliation is developed in the schistose matrix of the conglomerate (unit 3a), whilst in quartzite (unit 4a) bedding ( $S_0$ ) is typically the only foliation present. In biotite-quartz-feldspar gneiss (unit 6a) to the north of the Ompah synform,  $S_1$  is reinforced by the transposition of  $S_0$  in some localities,  $S_0$  layers varying greatly in thickness due to disruption (fig. 8). In marble, (unit 5) development of  $S_1$  is greatly influenced by the



Fig. 8 Disruption and transposition of bedding into  $S_1$  foliation in biotite-quartz-feldspar-tourmaline gneiss (unit 6a). The dark layers are concentrations of tourmaline. (Drawn from a hand specimen).

presence of calc-silicate minerals; where these are present a penetrative  $S_1$  foliation is developed, in contrast to the massive nature of pure marble units. In the muscovite schist (unit 6b), the foliation is a well developed schistosity, which generally obliterates all traces of bedding. Measurements of the orientation of muscovite crystals within this unit show that the preferred orientation, and hence the development of the foliation, increases with grade of metamorphism (see Textural Variation with Metamorphic Grade).

An interesting example of  $S_1$  foliation, which casts light on its mode of formation, is seen in the phenocrystic metabasalt (unit 1b). The rock possesses a fine scale, though penetrative foliation  $S_1$ , which is enhanced by segregation of the mafic and felsic components of the groundmass (fig. 99). Apparently there has been considerable reorganization of material in the groundmass of the rock as a result of deformation. However lath-shaped phenocrysts of plagioclase are quite recognizable, and still exhibit a primary flow orientation oblique to  $S_1$ . In most places the metamorphic layering is undeflected by the phenocrysts, and these appear minimally affected by deformation, only a few exhibiting small fractures and offsets. Spence (1968) reports, however, that the phenocrysts of plagioclase have been converted to a state of equilibrium with the metamorphic conditions (ie. a reduction in an content) "without loss of habit or twinning". Thus

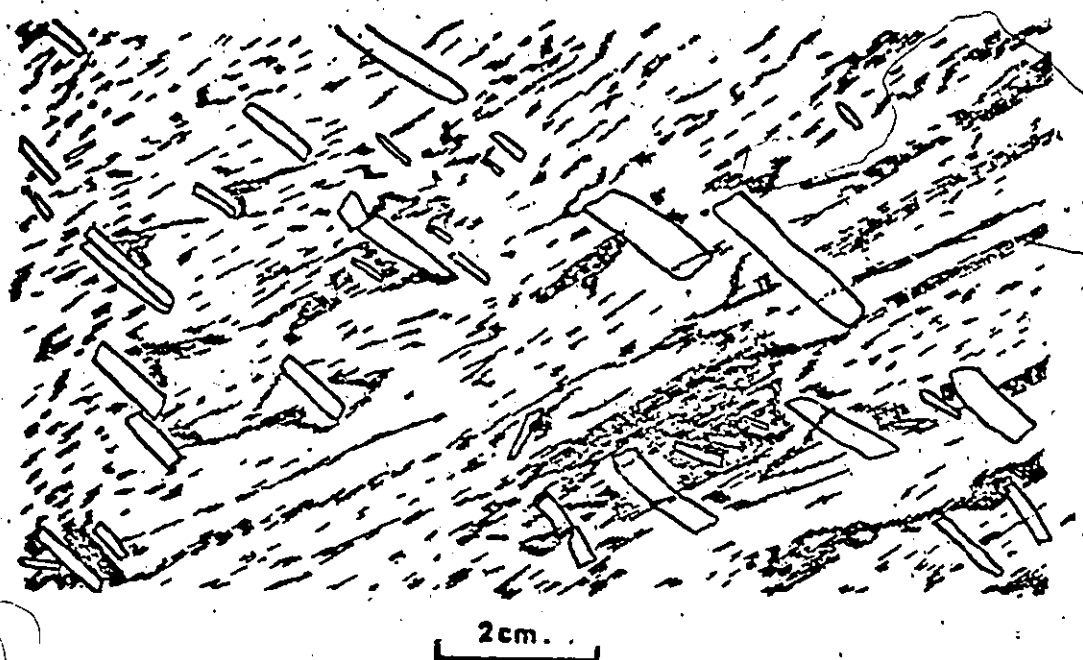


Fig. 9 Relict feldspar phenocrysts and  $S_1$  foliation in metabasalt. Phenocrystic metabasalt (unit 1b), in which a fine foliation ( $S_1$ ) is defined by alignment of biotite and amphibole and by the segregation of mafic and felsic components in the groundmass. The phenocrysts of feldspar exhibit a flow orientation oblique to  $S_1$ , which in a few places is slightly deflected around them.

despite the considerable rearrangement of material that has taken place, the formation of the foliation in this rock must have been a very delicate process, which did not cause obvious, small-scale distortion. Yet an estimate of the minimum amount of shortening that will cause a foliation visible to the naked eye is about 30% (Cloos, 1947), and Wood (1974) shows that the penetrative cleavage in slates is typically the result of 50 - 75% shortening perpendicular to the foliation.

#### $L_1$ lineation

A lineation  $L_1$ , parallel to the  $F_1$  fold axes is the dominant linear feature in the Ompah area. It is present in all the metasedimentary and metavolcanic units of the region, except the most massive and coarsely crystalline. In intrusive igneous rocks it is well developed in the Cross Lake gneiss (unit 9b), but only weakly developed in the Abinger granitic gneiss (unit 9a). Throughout the area, the orientation of  $L_1$  is very consistent-the plunge is gentle ( $< 30^\circ$ ), generally towards the northeast (fig. 3).  $L_1$  can be seen on  $S_0$  and  $S_1$  surfaces separately, and also on the combined  $S_0, S_1$  surface; because of this and the variety of rock types it has many morphological expressions. In amphibolite (unit 1a),  $L_1$  is defined by the alignment of amphibole crystals and felsic aggregates; individual amphibole crystals exhibit a preferred orientation, whilst feldspar and quartz crystals

in the aggregates are typically equant in shape. In banded quartz-biotite gneiss (unit 6a)  $L_1$  appears as a ridging or as a colour banding caused by the intersection of  $S_0$  on  $S_1$  surfaces. In some specimens  $L_1$  is also accentuated by a mica edge lineation on  $S_1$ ; this is due to the imperfect alignment of micas along  $S_1$ , resulting in a partial girdle of poles of (001) mica, the pole to the girdle coinciding with  $L_1$ . Another expression of  $L_1$  is visible in thin sections of fine-grained specimens of this lithology. Small porphyroblasts of biotite (1mm or less in diameter) are crystallographically randomly orientated, but morphologically are slightly flattened in the plane of  $S_1$ , and elongate parallel to  $L_1$  (plate 9). This feature will be discussed further in the section on the analysis of strain.

$L_1$  is also defined by porphyroblasts in some outcrops of hornblende-biotite gneiss (unit 6c) about 0.8 km. southwest of the fold closure of the conglomerate unit in the Ompah synform. These garnet crystals have typical dimensions of 10:1.5:1. and are parallel to a weakly developed  $S_0/S_1$  intersection lineation seen elsewhere in the outcrop. The foliation is not obvious in hand specimen, but in thin section it is defined by the preferred orientation of small biotite crystals; the garnets are poikiloblastic, overprinting the foliation.

In the pelitic muscovite schist (unit 6b), as in the more

micaceous<sup>an</sup> portions of the biotite-quartz-feldspar schist (unit 6a),  $L_1$  is defined by the alignment of mica edges and also, sporadically, by a compositional colour banding on  $S_1$ . These two linear elements are generally parallel, but in a few cases they were seen to diverge by as much as  $30^\circ$ . In these cases the  $S_0/S_1$  intersection lineation causing the colour banding is the steeper lineation. The reason for this divergence is not immediately clear. However a possible explanation may be that the colour banding lineation has been rotated into a steeper orientation during progressive flattening of small-scale mesoscopic folds (Ramsay 1962), whilst the mica edge lineation reflects the plunge of the major structure. Within unit 6d the mica edge lineation is less well developed at higher grades of metamorphism where the grain size is coarser. This observation can be reconciled with measurements of muscovite crystals from this lithology (see Textural Variation with Metamorphic Grade); with increasing grade of metamorphism the shape of muscovite (and biotite) crystals becomes progressively more equidimensional (fig. 43) which may detract from the mica edge lineation. Probably of greater significance though, is the measured increase in preferred orientation with grade of metamorphism (fig. 58). Since the mica edge lineation is due to the spread of (001) mica around the modal orientation of  $S_1$  (giving rise to a partial girdle of (001) mica on a stereonet),

a reduction in the spread of orientation (which tends to cluster the points on a stereonet) will weaken the lineation. However at high grades of metamorphism in this unit, a well developed lineation is frequently defined by the preferred orientation of sillimanite and occasionally by tourmaline crystals. Prismatic and fibrous sillimanite (plate 8) occur in the foliation aligned parallel to  $L_1$ ; and fibrous sillimanite also occurs in ellipsoidal pods similar to those illustrated by Divi (1972). These will be discussed further in a later section concerned with strain. Tourmaline, though ubiquitous in the muscovite schist unit at all grades of metamorphism, is visible in hand specimen only at high metamorphic grade. In thin sections of low grade rocks, small tourmaline crystals can be seen to lie in the plane of  $S_1$ , but they are not preferentially aligned parallel to  $L_1$ .

A striking  $L_1$  lineation defined by elongate pebbles is seen in the conglomerate of the Ompah and Plemma synforms and the Fernleigh belt (plate 4). Details of the shapes of the pebbles, and their implication for the type and amount of strain will be discussed later. In the schistose matrix of conglomerate (unit 3a) in the Ompah synform, a strong lineation is defined by biotite crystals, in places enhanced by the parallel alignment of amphiboles. These two lineations in the conglomerate, when seen together, are always coaxial, and are considered to parallel the plunge of the Ompah syn-

form, an  $F_1$  fold.

In the Cross Lake gneiss an L-S fabric is developed, in which the linear element, defined by the shape of aggregates and large, single crystals of quartz and feldspar, appears to be dominant. Where biotite is prominent, a foliation is also visible; but the small size of many of the mica crystals compared with those of quartz and feldspar, and their physical separation from one another often render this only a weakly developed feature.

The different morphological expressions of  $L_1$  therefore suggest that it is a combined intersection and extension feature ("stretch lineation").

#### Mylonites

About 0.5 km. north of the Fernleigh belt along the southern boundary of domain I is a zone of fine-grained laminated rocks outcropping for about 5 km., which is interpreted as a mylonite zone (fig. 3, in pocket). Although mostly grouped in unit 6a, the lithologies exhibiting the layering vary along strike, suggesting that it is not an original sedimentary feature. The best outcrops are in a road cut between Plevna and Fernleigh on highway 506, about 2 km. northwest of the village of Ardoch. There the lithology is a thinly layered (< 1 mm.) quartz-biotite-calcite-plagioclase-muscovite-haematite schist, with minor tourmaline; the layering is caused by the segregation of felsic

and mafic components, and in particular by the smearing of haematite and concentration of tourmaline crystals in the dark layers, plate 5. To the northeast of this outcrop, is a marble with a fine ( $< 2$  mm.) graphite and calc-silicate layering. Although segregation of graphite and calc-silicate minerals into layers in marble is common throughout the area, at this locality it is interpreted as being of mylonitic origin because (a) the banding is on a much finer scale than is generally observed in marble, and (b) the grain size of calcite varies as much as an order of magnitude from layer to layer (0.2 - 200mm.), being fine grained in the graphite and calc-silicate layers. Recrystallization and grain growth of calcite thus appear to have been impeded by the fine grain size of the graphite and calc-silicate phases. Further to the southwest along strike, a fine layering, is places almost paper-thin, is seen in a semi-pelite, as well as in other varieties of unit 6a. In all these examples, the rocks have recrystallized, and the mylonitic layering has been reformed during the second deformation into small-scale upright folds.

Other outcrops of fine grained rocks occur in the Panna synform and Shawenog Lake antiform, and these may also be mylonitic. They are finely layered ( $< 2$  mm.), and of several different lithologies eg. quartz-biotite-muscovite-tourmaline-calcite; quartz-feldspar-biotite-muscovite-opaque-tourmaline,

and frequently exhibit considerable variation in grain size from layer to layer (the layers without mafic phases being the coarser grained), as in marble.

All these rocks are blastomylonites, in which recrystallization has obliterated much of the original texture (Christie 1960). No porphyroblasts or relict grains were identified, suggesting either that the rock was fine grained when mylonitization occurred, or that mylonitization was very intense. The microstructure of the thicker layers in thin section is similar to the phyllitic texture illustrated by Bell and Etheridge (1973, p. 344), which they ascribe to the annealing of fine-grained mylonite, and they note that "there is still some segregation with elongate pods or layers alternately rich in biotite or muscovite", (p. 343), as is seen in the Ompah specimens. Characteristic of some of the Ompah specimens is the alignment of trains of tourmaline crystals in the mafic layers-these may represent the remnants of larger crystals broken down during mylonitization, whilst subsequent recrystallization and growth have allowed some crystals to develop euhedral outlines.

#### Attitude of $D_1$ Structures before $D_2$ Deformation

$F_1$  folds are considered to have been gently inclined to recumbent before  $D_2$  deformation. Small, scale recumbent  $F_1$  folds are common in thinly layered lithologies in the hinge zones of the Plevna antiform and synform, where their attitude

is considered not <sup>to</sup> have greatly changed during  $D_2$  and subsequent deformations.

Furthermore the present attitude of the rocks away from the hinge region of the megascopic  $F_2$  folds also indicates the  $S_1$  was gently inclined to sub-horizontal. The axial surfaces of  $F_2$  folds are steeply dipping, so on the assumption that the coaxial  $F_2$  folding is due to subhorizontally directed lateral compression, no folds would develop on  $S_1$  surfaces dipping more steeply than  $45^\circ$ . The combined symmetrical arrangement of the Plevna antiform and synform suggest that  $S_1$  was subhorizontal there before  $D_2$ , whilst the asymmetry of the  $F_2$  folding at Clyde Forks indicates a south-east dip for  $S_1$  before  $D_2$  (see cross-section, fig. 4).

#### $D_2$ Structures

##### Distribution and orientation of $F_2$ folds

The second deformation, though less penetrative on the local scale than the first, influenced much of the outcrop pattern on the geological map. Large-scale  $F_2$  folds are the dominant features in domains I, V and VI, whilst in domain III the distribution of poles to  $S_0$ ,  $S_1$  also suggests  $F_2$  folding (fig. 3, in pocket). Domain I includes the Plevna antiform and synform, both structures having a gentle ( $20 - 30^\circ$ ) northeast plunge, with bedding ( $S_0$ ) and cleavage ( $S_1$ ) being folded about vertical to steeply southeast dipping axial surfaces,  $S_2$ . From the structural and synoptic maps

(figs. 3 & 5) it can be seen that the poles to the  $S_0$  and  $S_1$  surfaces describe a common great circle, the pole ( $\pi$ ) of which defines the axis of the major folds.  $\pi$  falls within the spread of the contoured measurements of the axes of mesoscopic  $F_2$  folds. The average orientation of the plane of  $S_2$  (044/78SE) also passes through the spread of the contoured  $F_2$  measurements, and  $\pi$  lies close to this plane. Thus the mesoscopic  $F_2$  folds are coaxial and coplanar to the major structures.

Southwest along strike from the axis of the Plevna synform is the Shawenegog Lake antiform, and this, too, is an  $F_2$  structure. The folded surface is  $S_0$ ,  $S_1$ , and the plunge of the fold and the orientation of the axial surface are: trend 055°, plunge 15°, and strike 055°, dip 90° respectively, very similar to the Plevna antiform and synform to the northeast. The core of the Shawenegog Lake antiform is occupied by a granite body. The timing of the intrusion with respect to the formation of the fold is not clear; all contacts are intrusive and there is limited evidence of a contact aureole (Smith, Vogel and Spence 1969). The Shawenegog Lake antiform plunges into the Plevna synform less than a mile along strike to the northeast (see cross section A-A, fig. 4, in pocket). The disharmonic pattern of the folding suggest that the intervening lithology (predominantly marble) was very ductile during the second deformation. Transposition of biotite-

hornblende gneiss (unit 3a) parallel to the axial surface of the  $F_2$  folds may also have been facilitated by the ductility of the marble.

The Cross Lake antiform, which occurs in domain V, has a core of gneissic granite (the Cross Lake gneiss, unit 9b) over which the supercrystal rocks are deformed (fig. 3, cross section B-B). Chapman (1968) considers the gneiss to be a "phacolith-like sheet" intruded into metavolcanic and meta-sedimentary rocks. The whole sequence, including the region to the south beyond the boundaries of the map, is folded into an elongate dome, which plunges gently ( $10 - 15^\circ$ ) northeast in domain V. Bedding ( $S_0$ ) is rarely recognized, but it is apparent that the foliation ( $S_1$ ) in unit 1 is folded around the hinge of the fold (fig. 3). The outcrop pattern of the boundary of unit 2 with marble 5 km. to the southwest is evidence of the infolding of these two lithologies during  $D_2$  (fig. 1, in pocket). Mesoscopic  $F_2$  folds in this area typically plunge parallel to the major structure, but some have divergent (steeper) orientations. No small-scale re-folded folds were observed.

Domain VI chiefly comprises the Clyde Forks antiform and synform in the northeast part of the area. The folding is most clearly displayed by the outcrop pattern of the Benlough belt, which defines an asymmetrical S-shaped fold pair when viewed down plunge. The trend of this  $F_2$  structure is more

easterly than the other large scale  $F_2$  folds (trend  $057^\circ$ , plunge  $20^\circ$ ). Parasitic mesoscopic folds are quite common in certain lithologies, and some display the same sense of asymmetry as the major structure. This asymmetry is taken to indicate that the  $S_0$ ,  $S_1$  surface had a southeasterly component of dip before  $D_2$ .

In contrast to elsewhere, few mesoscopic and microscopic folds were recognized in domains II and III. In domain III the variable dips of  $S_1$  indicate that  $F_2$  or later folds are open and of a scale greater than the outcrop, but usually less than that which shows up on the map. By contrast, in domain II  $S_1$  generally dips steeply; small folds occur in calc-silicate layers in marble (unit 5), but the nature of the folded surface could not be determined with assurance. Fold vergence is typically not consistent over distances of more than a single outcrop. The lack of mesoscopic folding elsewhere may be explained by the massive nature of the coarsely crystalline marble which occupies much of these domains.

#### $F_2$ fold profiles

As the map pattern and the cross-sections show,  $F_2$  folds are generally more open than  $F_1$ . Using the map pattern (slightly modified see fig. 10) of the hornblende-biotite gneiss (unit 1c) of the  $F_2$  Plevna synform as an example, calculations of the amount of flattening of the folded unit yield values of the order of 10 - 15% at most (no estimate can be

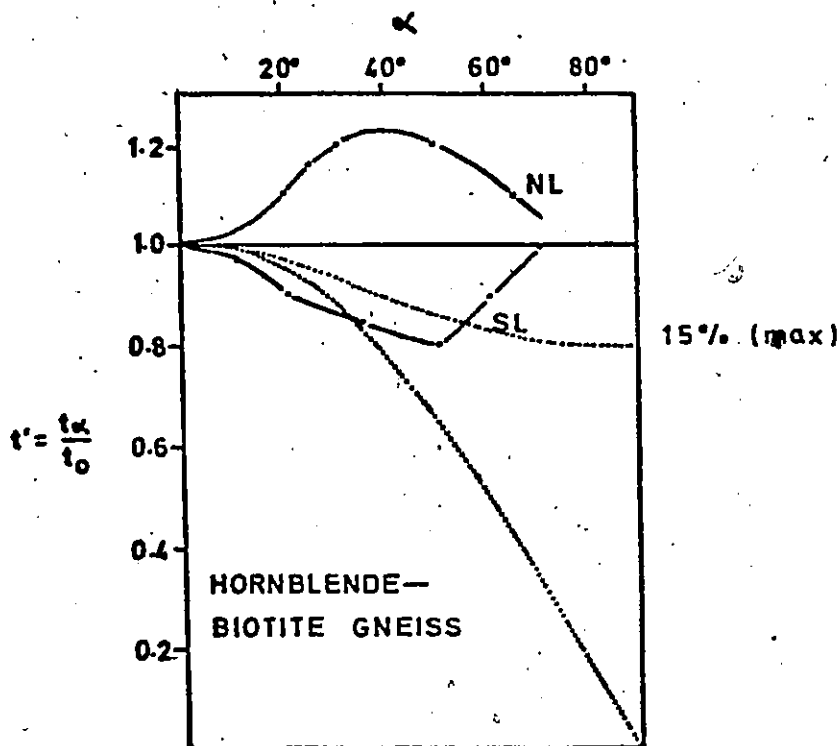
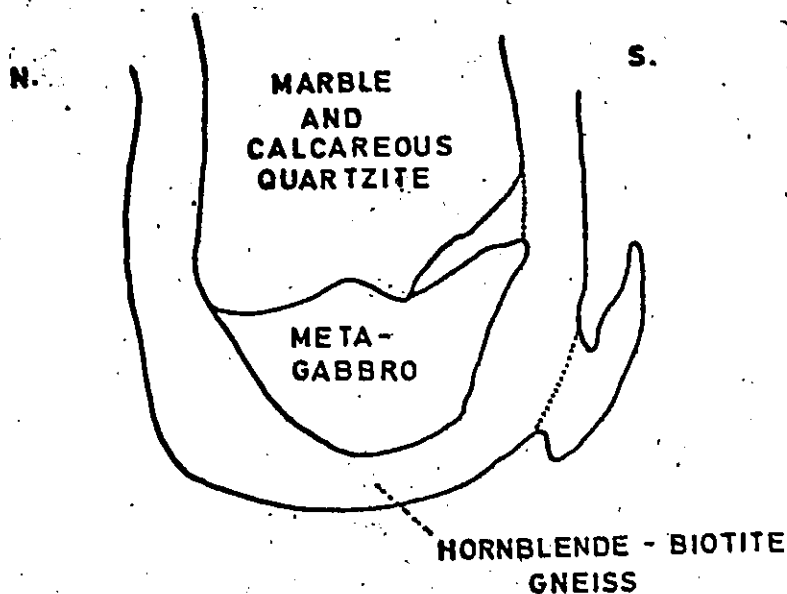


Fig. 10 Down plunge profile of the map pattern of hornblende-biotite gneiss (unit lc) in the F<sub>2</sub> Plevna synform, with graphs of  $t'$  against  $\alpha$  (Ramsay 1967) for north and south limbs (NL and SL). The percentage flattening is given at the right of the graph. For calculations of flattening the septum on the south limb of the fold was ignored (see dotted line).

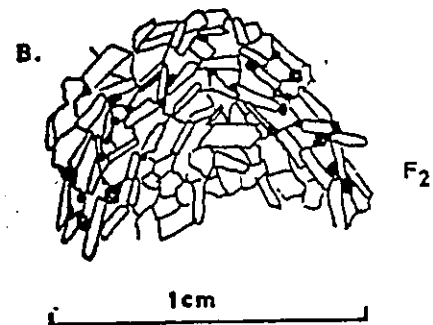
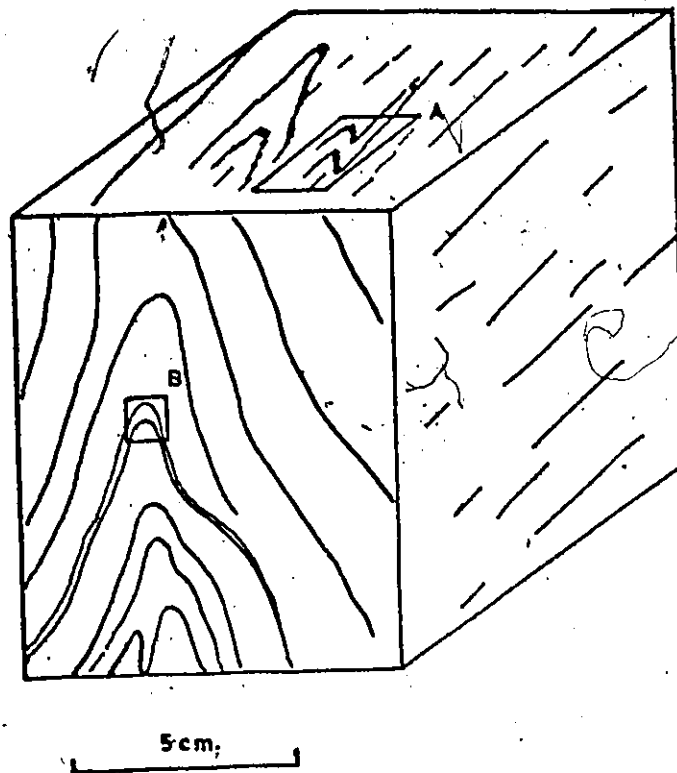


Fig. 11 Sketch of a specimen containing both  $F_1$  and  $F_2$  structures in biotite-quartz-feldspar-amphibole gneiss (unit 6a). In thin section the  $F_1$  fold (A) is seen to affect a compositional layering ( $S_0$ ), defined by a high concentration of amphiboles which are aligned parallel to the axial surface. In the  $F_2$  fold the  $S_1$  surface is folded. Locality 3.5 km. west of Ardoch.

made for the north limb, which increases in thickness away from the hinge ie. is class 1A). Folding is therefore nearly of the parallel (class 1B) type. A similar low percentage of flattening is indicated by the near uniform thickness of map units around other  $F_2$  folds eg. the Clyde Forks antiform, but the irregular boundaries of other large scale  $F_2$  folds do not permit even rough estimates of flattening.

Comparisons of the amount of flattening of different folds can be made if folding was initially parallel (class 1B) (Ramsay, 1967). Folding of quartzite is generally of this type, so it is of interest to compare the profiles of the quartzite unit of the  $F_1$  Ompah synform with the hornblende-biotite gneiss of the  $F_2$  Plevna synform. Whilst a direct quantitative comparison is not valid because of the different lithologies employed for the measurements, the estimate of 10 - 15% flattening of the  $F_2$  fold is in marked contrast to the 25 - 30% flattening of the  $F_1$  structure (fig. 6). In hand specimen  $F_2$  folds are also less flattened than  $F_1$  folds (fig. 11, plate 8).

#### $S_2$ foliation

$S_2$  is not commonly expressed as a well defined foliation in the Ompah area. Mesoscopic  $F_2$  folds are generally open warps and crenulations, the axial planar fabric being poorly if at all developed (many of the  $S_2$  measurements shown on the stereonet (fig. 3) are the orientations of the axial surfa-

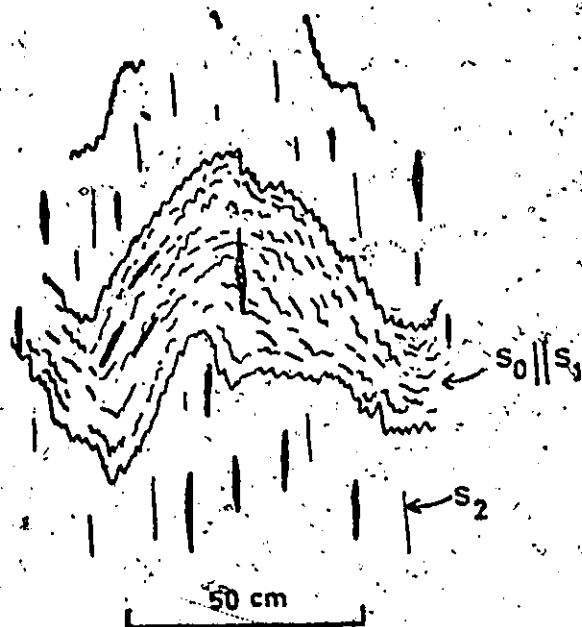


Fig. 12 Mesoscopic  $F_2$  fold in quartzofeldspathic gneiss (unit 1d), containing quartz segregations parallel to both  $S_1$  and  $S_2$ . Locality - hinge area of Plevna antiform, 0.75 km. south of Grindstone Lake. Drawn from a photograph.

ces of  $F_2$  folds). In the hinge of the Plemna ~~synform~~ a widely spaced (1 - 2 m.) open fracture cleavage occurs in unit 1c, but the folded surface,  $S_1$ , is the dominant foliation. In quartzo-feldspathic gneiss (unit 1d) in the hinge area of the  $F_2$  Plemna ~~antiform~~, mesoscopic  $F_2$  folds ~~define~~ the combined  $S_0, S_1$  surface.  $S_2$  is here more penetrative, and is enhanced by quartz segregation lenses (fig. 12).

Away from the hinge regions of  $F_2$  folds,  $S_2$  is subparallel to  $S_0$  and  $S_1$ , and thus difficult to identify. Several examples of fractures interpreted as  $S_2$  cut  $S_1$  at an acute angle. Some of these are filled with quartz or calcite, presumably a later effect. In the mylonite zones  $F_2$  folds are tighter than elsewhere,  $S_2$  is more penetrative and partly defined by aligned biotite crystals (plate 5). In the fold limbs there is also segregation of opaque minerals along the  $S_2$  foliation. The stronger development of  $S_2$  in these rocks, with more pronounced crystal orientation, may be due to a larger amount of strain energy having been available for recrystallization resulting from the intense strain during  $D_1$  mylonitization, since it is known that the amount of strain can greatly influence recrystallization rates, (Griggs, Paterson, Heard and Turner, 1960). In marble and calc-silicate rocks (unit 5)  $S_2$  is variably developed; within a single thin section some  $F_2$  microfolds have an associated axial planar fabric whilst others do not (fig. 13).

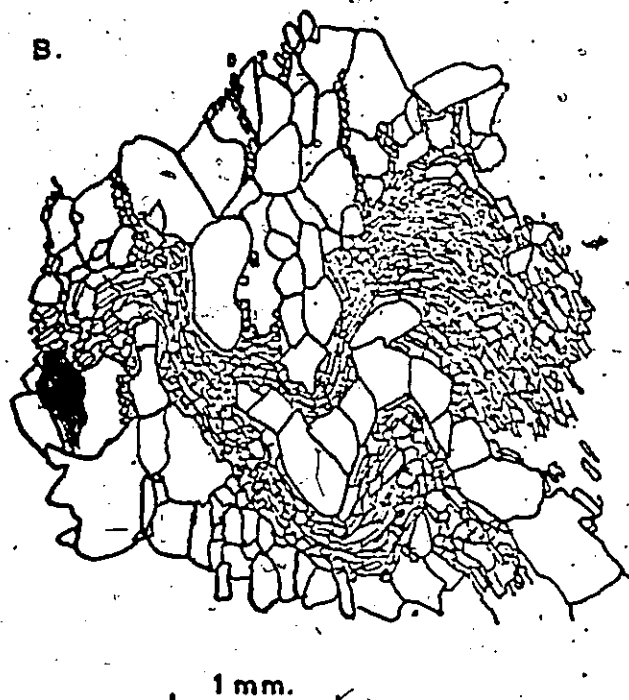
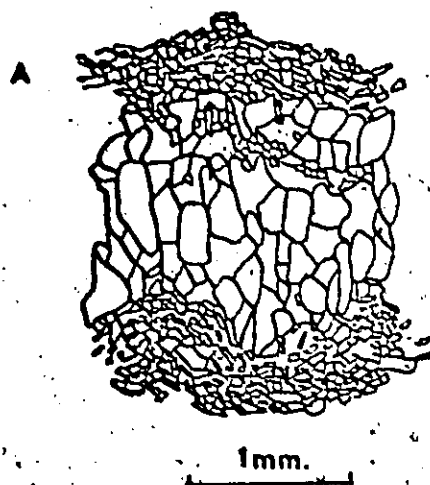
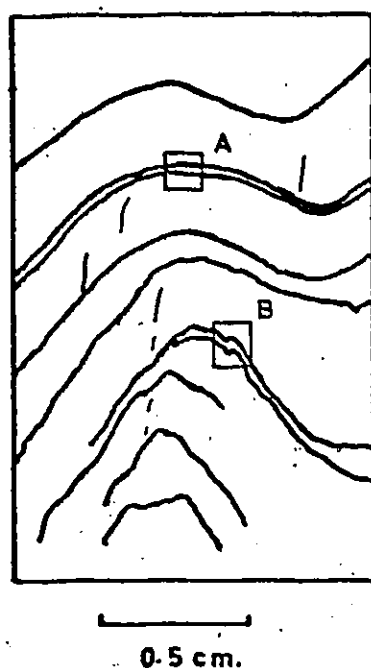


Fig. 13 Sketch showing variable development of  $S_2$  in a calcite-tremolite rock (unit 5). In A, small tremolite crystals are aligned parallel to  $S_1$ , the folded surface. Some calcite crystals (larger grains) are aligned in  $S_1$  and some in  $S_2$ , the axial surface foliation of the  $F_2$  fold. In B,  $S_2$  is defined by thin layers of tremolite as well as by inequant grains of calcite. Drawn from thin sections.

## $L_2$ lineation

As a consequence of the weak development of  $S_2$  in most rocks,  $L_2$  is not a common feature. It has little mineralogical expression, but rather is defined by the axes of  $F_2$  microfolds on the combined  $S_0$ ,  $S_1$  surface. In fig. 3  $L_2$  is grouped with  $F_2$ .

## Interference Patterns

Mesoscopic scale interference patterns between  $F_1$  and  $F_2$  folds are not common. In the hinge zones of the upright  $F_2$  Plevna antiform and synform, small  $F_1$  folds with subhorizontal axial surfaces are quite common in the more thinly layered units; but there are no associated small  $F_2$  folds. Folds of two generations in a single hand specimen are illustrated in fig. 11. The  $F_1$  fold was not apparent until after cutting the specimen. Transposition of the original layering ( $S_0$ ) during  $D_1$  has obscured  $F_1$  hinge areas; hence refolded structures may have been missed in the field. In plate 6 a mesoscopic refolded fold in marble (unit 5) outcropping between the Ompah synform and the Fernleigh belt is illustrated. The core of this structure is of marble, around which a thin layer of biotite schist forms a recumbent fold.  $F_2$  crinkles about steeply SE dipping axial surfaces are superimposed on this,  $F_1$  and  $F_2$  being subparallel.  $S_2$  is expressed in the marble as a fracture cleavage. Calc-silicate layers in the core of the structure are irregular, and do not define a

coherent interference pattern, probably reflecting the extreme mobility of calcite and calc-silicate phases during deformation.

### $D_3$ Structures

#### $F_3$ folds

A set of fold structures, sporadically developed, particularly around Plevna Lake and the villages of Ompah and Donaldson, is distinguished from  $D_2$  structures primarily on the basis of orientation. In contrast to  $S_2$  axial surfaces of similar NE strike, the axial surfaces dip less steeply than the folded surface,  $S_1$ .  $F_3$  folds typically die out rapidly along strike; near Plevna Lake conglomerate clasts flattened in  $S_1$  are sigmoidally folded about an  $S_3$  axial surface, but the fold persists only for about 0.5 m. (fig. 14). A larger scale open recumbent fold in the Fernleigh belt, which is apparent in cross-section C-C (fig. 4), may also be an  $F_3$  structure. Apart from this, all identified  $F_3$  folds are small (wavelength less than 1 m.). A crenulation sporadically developed in schistose lithologies (especially pelitic schist, unit 6c) throughout the area may also be a  $D_3$  structure. The typically steeply dipping foliation is folded about subhorizontal axial surfaces into open, non-penetrative microfolds.

Most  $F_3$  folds have rounded hinges, and are of the parallel (class 1B) type. However the folded clasts in conglomerate

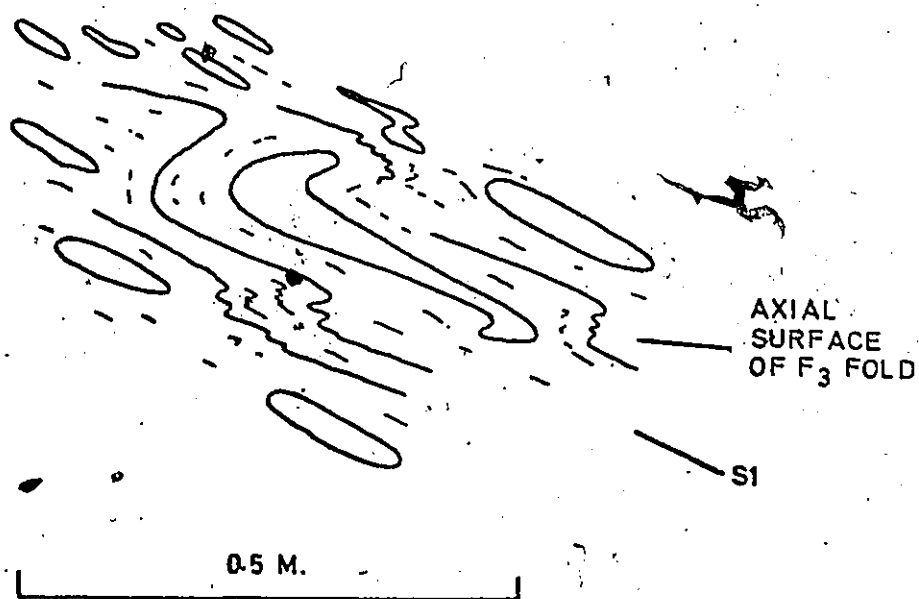


Fig. 14  $F_3$  folds in conglomerate near Plevna Lake. A granite pebble is folded, and considerably thickened in the hinge zone, but the folds die out within 1 m. Drawn from a field sketch.

at Plevna Lake show some thickening in the hinge region (fig. 14).

No  $F_3$  folds were found superimposed on  $F_2$  folds of a similar scale, so the age relationship between the folds is not definite. However sporadically developed, non-penetrative recumbent structures are a common late feature of many metamorphic terrains, including elsewhere in the region, eg. Divi (1972).

### $S_3$ fracture cleavage

Chlorite-filled fractures, eg. in biotite-quartz-feldspar gneiss (unit 6a) near Plevna Lake, are parallel to the axial surfaces of  $F_3$  folds. In competent rocks similar fractures appear to be the sole expression of  $D_3$  deformation. Growth of chlorite may be a retrograde effect, post-dating the formation of the cleavage.

### Strain Analysis of Pebbles, Aggregates and Porphyroblasts

In the study area several indicators of strain exist, and these have been used both quantitatively and qualitatively to obtain information about the local strain component of deformation.

### Pebbles

As already mentioned clasts in conglomerate (unit 3a) of the Ompah synform and the Fernleigh belt are highly deformed, and define a low plunging axial lineation. Most commonly

the intermediate and major axes of the pebbles lie in the  $S_1$  foliation plane, with the shortest axes perpendicular (the intermediate and minor axes may show some variability in this respect see plate 4).

Visual assessment of the shapes of clasts in the conglomerate unit of the Ompah synform at a large number of localities did not indicate any systematic change of pebble shape with location on the fold, or with distance along strike (ie. grade of metamorphism). Such a variety of axial ratios exists at a single locality that areal changes are effectively masked. The deformed shapes reflect the variable nature of the sediment (Mehnert 1939), and contributing factors include: (1) composition of the matrix, (2) composition of the clast and (3) proportion of clast to matrix. (1) and (2) control the relative ductilities of the matrix and clast, and where clasts vary in lithology, the strain calculated from the shapes also varies.

Because of the highly variable nature of the conglomerate unit, measurement of representative samples of the total population of clasts was not attempted. Instead, in order to determine the range and type of strain in the various conglomerate bodies of the Ompah area, measurements were confined to restricted types of pebbles in various matrix materials with different proportions of clasts. In the Ompah synform and the Shawenog Lake antiform measurements were made on

quartzite and granite pebbles (which possess similar axial ratios), whilst in carbonated conglomerate of the Plevna synform the pebbles measured were of calc-silicate composition.

The type of strain was obtained following the principles of Flinn (1956) and Wood (1974); and the amount of strain by calculating the natural octahedral shear (Nadai 1950, 1963, in Hossack 1968) - see Appendix I.

Type of strain. In the  $F_1$  Ompah synform, the type of strain ranges from constrictional, through approximately plane strain to flattening (fig. 15). In typical samples of constricted clasts eg.  $OS_1$ ,  $OS_2$ , the X axis is elongated by 100 - 300% of  $d$  ( $d$  is the diameter of a sphere of the same volume as the ellipsoid - see Appendix 1), and the Y and Z axes are reduced by 10 - 30% of  $d$  and 50 - 60% of  $d$  respectively<sup>1</sup>. The most pronounced constrictional strain is seen in the quartzite conglomerate with a contact fabric ( $OS_3$ ), where the X axis is elongated by about 500% of  $d$ , with Y and Z reduced by 40% and 75% of  $d$ . It is interesting to note that the highest constrictional strain ( $OS_1$ ,  $OS_3$ ) is from the hinge

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<sup>1</sup> These figures assume no volume change in the ellipsoid during deformation, and an initially spherical pebble shape. On the basis of density and other considerations, Ramsay and Wood (1973) suggest that volume loss during deformation does not exceed 10% in slates, and this may be qualitatively extended to the conglomerate of the present study; and Ramsay (1967) argues that for large amounts of strain the initial pebble shape is of little significance.

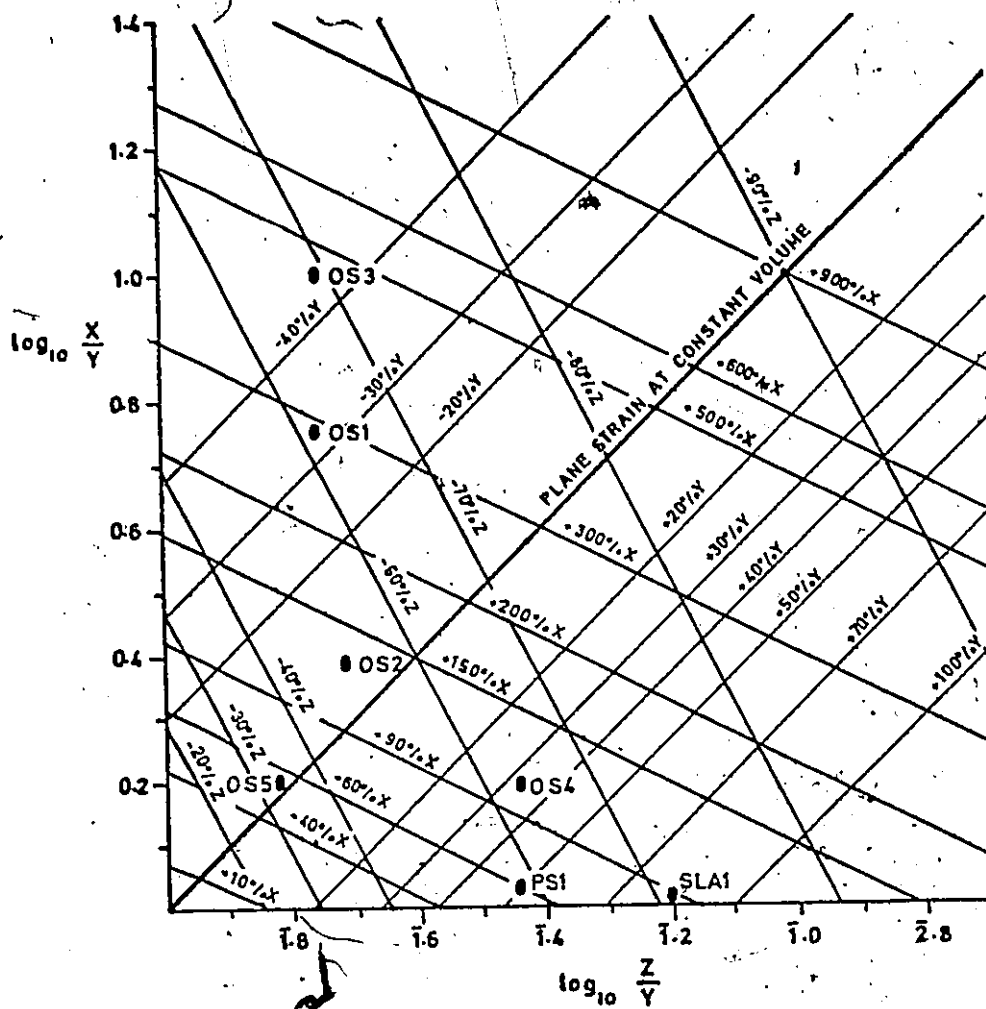


Fig. 15 Strain plot for deformed pebbles from metaconglomerate (after Flinn 1956 and Wood 1974). Sample numbers prefixed OS are from the  $F_1$  Ompah synform, PS is from the  $F_2$  Plevna synform and SLA from the Shawenégog Lake antiform - for exact locations see fig. 4.

area of the fold (fig. 3), and that the X axes of the clasts are parallel to the  $F_1$  fold axis.

Elsewhere in the Ompah synform, the pebbles tend towards an oblate shape. In  $OS_4$  the Z axis is reduced by about 60% of d, whilst the Y and X axes are increased by about 30% and 15% of d respectively. The shape of quartzite pebbles in conglomerate with an isolate fabric was not determined statistically, but several separate measurements of typical axial ratios made in the field indicate that distortion may approximate plane strain ( $OS_5$ ).

The samples from the  $F_2$  Shawenegog Lake antiform ( $SLA_1$ ) and  $F_2$  Plevna synform ( $PS_1$ ) are both flattened, almost to the exclusion of a linear element; no lineation was observed in the conglomerate at Shawenegog Lake. In  $PS_1$  and  $SLA_1$  the X and Y axes of the ellipsoid are increased between 50% and 100% of d, the assumed original diameter of the clast (fig. 15). The sample at Shawenegog Lake comes from the hinge zone of the  $F_2$  fold, and the clasts lie in the plane of the folded surface ( $S_1$ ) at a high angle to  $S_2$ . Evidently local strain associated with  $F_2$  folding was insignificant. This is in accord with the fold profiles which are of the parallel type (fig. 10), and in marked contrast to the constrictional strain in the hinge area of the  $F_1$  Ompah synform. The Plevna synform specimen ( $PS_1$ ) is from a limb of the fold, where  $S_1$  and the axial surface ( $S_2$ ) are subparallel. As to be expected

| Sample Number    | No. of measurements* | Clast composition    | Matrix composition/proportion          | Mean axial ratios of clasts | Total distortion (%) |
|------------------|----------------------|----------------------|----------------------------------------|-----------------------------|----------------------|
| OS <sub>1</sub>  | 30                   | quartzite            | impure quartzite (minor biotite) / 30% | 9.8 : 1.7 : 1               | 1.82                 |
| OS <sub>2</sub>  | 40                   | quartzite granite    | biotite schist / 40%                   | 4.7 : 1.9 : 1               | 1.27                 |
| OS <sub>3</sub>  | 20                   | quartzite            | impure quartzite / 5%                  | 24.7 : 2.4 : 1              | 2.50                 |
| OS <sub>4</sub>  | 40                   | <del>quartzite</del> | biotite schist / 15%                   | 6.6 : 3.6 : 1               | 1.54                 |
| OS <sub>5</sub>  | estimate             | quartz               | biotite-hornblende / 95% schist        | 3.0 : 1.5 : 1               | 0.90                 |
| PS <sub>1</sub>  | 30                   | calc-silicate        | carbonate / 40%                        | 3.8 : 3.5 : 1               | 1.23                 |
| SLA <sub>1</sub> | 40                   | quartzite granite    | biotite-hornblende / 35% schist        | 6.5 : 6.3 : 1               | 1.76                 |

\* Pairs of axial ratios

Table 3. Composition of measured clasts and matrix, and approximate clast: matrix proportions, together with mean axial ratios and total distortion factor (%). Sample numbers prefixed OS are from the Ompah synform, PS from the Plevna synform and SLA from the Shawenog Lake antiform; for locations see fig. 3.

if  $F_2$  folding has not affected the local strain, the flattening element on an  $F_2$  limb is similar to that of  $SLA_1$  from an  $F_2$  hinge zone.

In addition to the position on an  $F_1$  fold, the composition of the matrix may have an influence on the type of strain (table 3). The samples with the greatest constrictional element have a quartzose matrix, whilst these more flattened come from specimens with schistose matrices.

Amount of strain. As a general rule from visual estimates, the most deformed clasts are of carbonate composition, followed by schist, then quartzite and granite and finally crystalline (vein?) quartz. The results of the calculations of  $\gamma_0$ , the total distortion of the measured clasts, are summarized in table 3. Most values fall between  $\gamma_0 = 1.2 - 2.0$ , with the quartzite pebbles in a contact fabric ( $OS_3$ ) at  $\gamma_0 = 2.50$  and estimates of an isolate fabric ( $OS_5$ ) at  $\gamma_0 = 0.90$ , falling outside this range. The two specimens from the hinge zone of the Ompah synform ( $OS_1$  and  $OS_3$ ) which show the most pronounced constrictional element, are also the two which have suffered the greatest total distortion (table 3). However the flattened specimen from Shawenegog Lake antiform ( $SLA_1$ ) is only slightly less distorted than  $OS_1$ , so constrictional and flattening strains, though variable, appear to have been of the same order of magnitude throughout.

### Aggregates

In the  $F_1$  antiform at Ardoch (AA on fig. 3), small (2 - 7 mm. diameter) light coloured crystal aggregates which are present in certain layers, appear elliptical in shape on surfaces perpendicular to the fold axis. These aggregates are composed principally of tremolite and quartz, with minor calcite, whilst the darker matrix consists predominantly of biotite-calcite-quartz (plate 7). The  $S_1$  foliation, defined by the preferred orientation of biotite in the matrix, is deflected around the aggregates. The latter are considered to represent a recrystallized sedimentary feature, possibly sand pellets, and thus they might be expected to have approximated a spherical shape before deformation.

A total of 260 axial ratios of the aggregates were measured from five specimens, on several surfaces cut parallel and perpendicular to the fold axis (for location of specimens on fold see fig. 7). Most measurements were made in thin section, and whilst some aggregates have ragged edges which impede accurate measurement, the boundaries of the majority can be readily defined. Measurements were plotted as before.

Type of Strain. All five samples lie in the field of flattening strain, though none diverges greatly from plane strain (maximum increase of the Y axis is 12% of d) fig. 16. The Z axis is reduced by 15 - 40% of d, whilst the X axis is increased between 15 - 55% of d. The samples fall into two

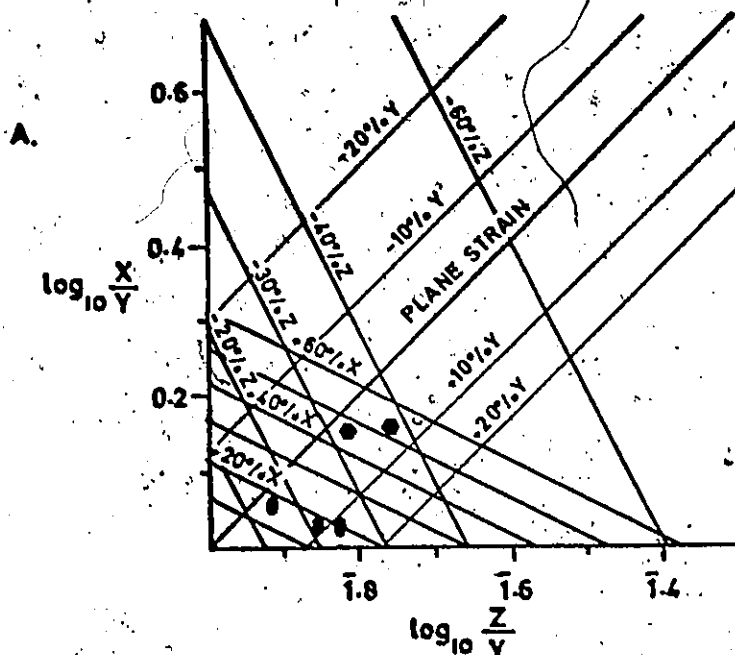


Fig. 16 Strain plot for deformed aggregates from Ardoch antiform. • - hinge region, • - limbs. For location of specimens on fold see fig. 7.

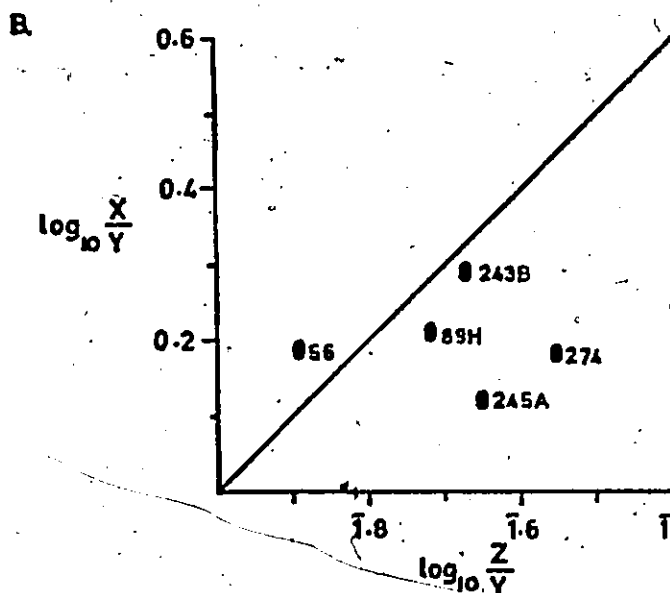


Fig. 17 Strain plot for deformed biotite and sillimanite porphyroblasts. Spec. 56 is from biotite-quartz gneiss (unit 6a) in the axial region of the Plevna antiform. Other specimens, from which both biotite and sillimanite porphyroblasts were measured, are from pelitic schist (unit 6b) of the Fernleigh belt. For exact locations see fig. 3.

| Sample | No. of<br>measurements<br>(axial ratios) | Mean Axial Ratios of Aggregates |      |   | Total Distortion ( $\chi_0$ ) |
|--------|------------------------------------------|---------------------------------|------|---|-------------------------------|
|        |                                          | X                               | Y    | Z |                               |
| AA1    | 33                                       | 2.54                            | 1.76 | 1 | 0.76                          |
| AA2    | 20                                       | 2.20                            | 1.54 | 1 | 0.65                          |
| AA3    | 82                                       | 1.57                            | 1.49 | 1 | 0.32                          |
| AA4    | 45                                       | 1.47                            | 1.39 | 1 | 0.34                          |
| AA5    | 80                                       | 1.38                            | 1.20 | 1 | 0.25                          |

Table 4. Mean axial ratios and total distortion factor ( $\chi_0$ ) for tremolite-quartz aggregates from the Ardoch Antiform. AA1 and AA2 are from the limbs, AA3-AA5 are from the hinge (for exact locations see fig. 7).

groups representing the limb and hinge zones of the fold. Divergence from plane strain is apparently not related to position on the fold, as the specimens which most closely approximate plane strain, AA<sub>2</sub> and AA<sub>5</sub>, are from the limb and hinge zone respectively.

Amount of strain.  $\gamma_0$  values for all 5 specimens are given in table 4; all are considerably lower than those for the conglomerate pebbles. Values from the limbs (AA<sub>1</sub> and AA<sub>2</sub>) are greater than those from the fold hinge (AA<sub>3</sub> - AA<sub>5</sub>), as can be qualitatively observed in fig. 16. This is in agreement with assessments of strain based on the variable development of S<sub>1</sub> cleavage and minor folds, which are less pronounced in the hinge area of the fold (fig. 7).

#### Porphyroblasts

Some porphyroblasts are useful in yielding qualitative information on the type of strain (see discussion of L<sub>1</sub>). They cannot be used for determination of the amount of strain because in many cases they were not present before deformation, and the factors which caused their ellipsoidal shape (deformation or growth under deviatoric stresses) cannot be distinguished. Randomly orientated porphyroblasts of biotite and pods of sillimanite (considered to have grown during D<sub>1</sub> see Time Relations) have been measured from sawn hand specimens and thin sections cut parallel and perpendicular to L<sub>1</sub>. Four specimens with biotite porphyroblasts and sillimanite pods

(plate 8) from pelitic schist (unit 6b) are from the Fernleigh belt, whilst a single specimen with biotite porphyroblasts (plate 9) is from biotite-quartz-feldspar schist (unit 6a) in the axial region of the  $F_2$  Plevna antiform (for location of specimens see fig. 3).

Type of Strain. The type of strain for all four specimens of pelitic schist (unit 6b), whether measured from biotite porphyroblasts or sillimanite pods, falls in the flattening field (fig. 17). Sillimanite pods in specimen 243B define the  $L_1$  lineation in hand specimen, demonstrating as do the conglomerate pebbles, that extension sufficient to give rise to a lineation visible to the naked eye may occur in the flattening field.

The biotite porphyroblasts from the specimen of unit 6a plot in the field of constrictional strain. As with the conglomerate pebbles, therefore, there is an element of extension parallel to fold axes, though in this case the major adjacent fold is an  $F_2$  structure. However strain measurements elsewhere eg. Shawenog Lake antiform, Ardoch antiform, Ompah synform, indicate that strain was predominantly imposed during  $D_1$ , so it is probable that the same was the case in this area too, with coaxial  $F_2$  folds being a later, super-imposed effect.

## Discussion

### Variation in Strain

Fig. 18A is a deformation plot in which the different tectonite classes are shown (after Flinn 1965), and it is apparent that almost the complete range of classes is represented by the deformed clasts, aggregates and porphyroblasts in the Ompah area. Most examples, however, lie in or near the fields of flattening or plane strain. Comparison of  $\gamma_0$  values for the clasts and aggregates (tables 3 and 4) also indicate a considerable range of total distortion. A similar range of strain type and magnitude is noted by Venkitasubramanyan (1969, area B, fig. 2). One explanation for this is that strain during  $D_1$  was inhomogeneous both in type and magnitude, and that the second deformation did not contribute greatly to the measured strain. This is supported by the correlation of strain with the intensity of development of  $D_1$  structures in the Ardoch antiform (fig. 7), and the flattening at high angle to  $S_2$  on the nose of the  $F_2$  Shawenegog Lake antiform, which, like other  $F_2$  folds is of the parallel type with little imposed flattening, p.53).

A second possibility is that during  $D_1$  deformation, strain was of the flattening type throughout the region; subsequently during  $D_2$  deformation, strain was also of a flattening type, but orientated at a high angle to  $S_1$ , causing extension parallel to  $L_2$ , the line of intersection of  $S_1$  and  $S_2$ .

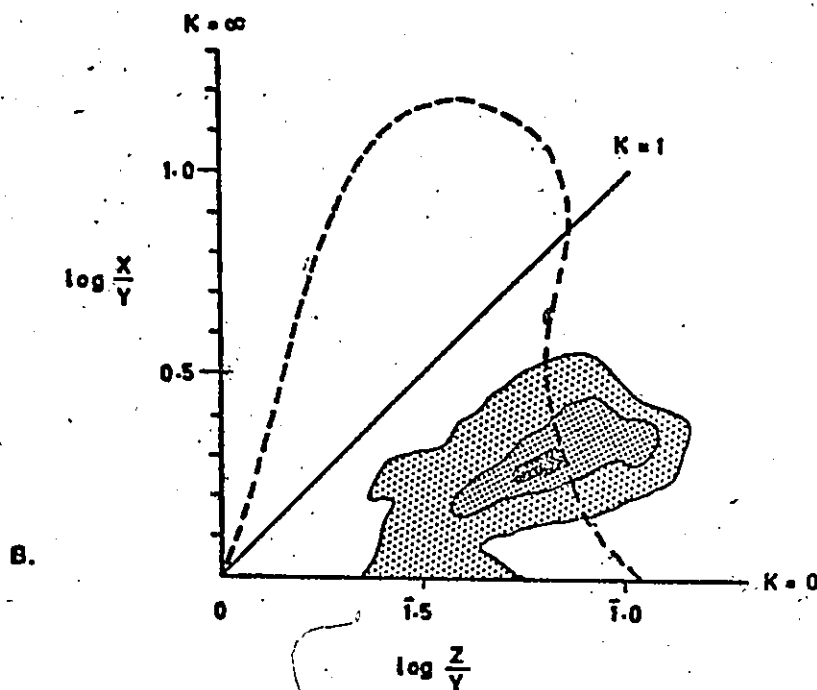
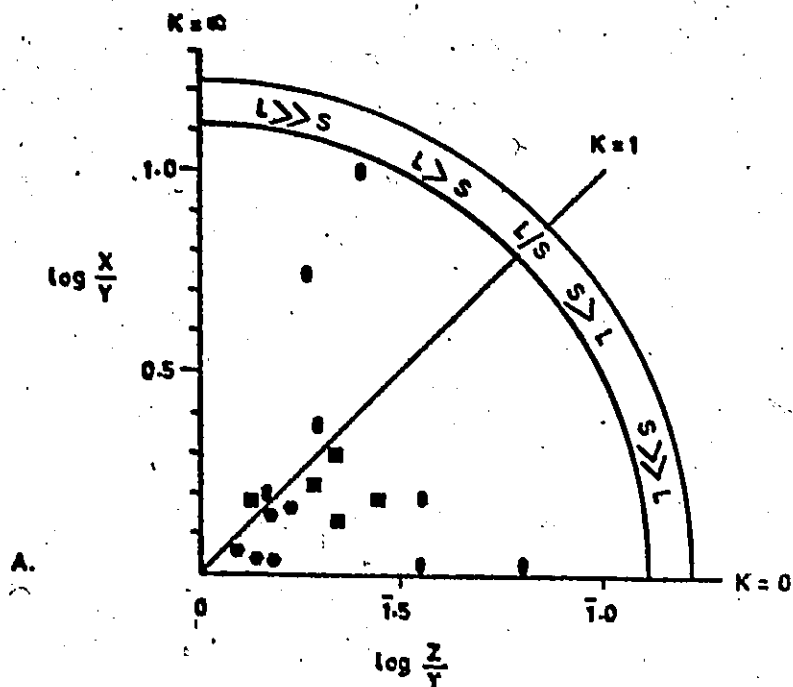


Fig. 18 A. Deformation plot showing tectonite classes for deformed pebbles ( $\circ$ ), aggregates ( $\blacksquare$ ), and porphyroblasts ( $\blacktriangle$ ) from the Ompah area.

B Comparison of the strain field characteristic of slate belts (contoured area - taken from Wood 1974) with that observed in the Bancroft-Madoc region (area enclosed by dashed line).

Recently documentation of this effect, in a conglomerate in the Alps, has been made by Thakur (1974). In his case the resultant type of strain recorded by the pebbles is constrictional, but presumably many other modifications of an initial strain type ( $S \gg L$  for example) are possible eg.  $S \gg L$  to  $S \rangle L$  or  $S \gg L$  to  $S/L$ , (see fig. 18A) so that the resultant strain does not necessarily fall in the constrictional field.

In the Ompah area, since  $F_1$  and  $F_2$  are coaxial, and the long axes of strain markers are parallel to fold axes, the superposition theory is feasible. The constrictional strain indicated by some of the clasts in the conglomerate at Ompah ( $OS_1$  and  $OS_3$ )<sup>10</sup> could be analogous to the results of Thakur (1974), with other  $D_1$  strain types being modified by an amount insufficient to move them from the flattening field. However several lines of evidence indicate that the first hypothesis is the more likely in the Ompah area. Calculations of the amount of flattening of quartzite (unit 4a) in the  $F_1$  Ompah synform (25 - 35%) (fig. 6) are less than the estimates from the Z axes of typical clasts in the adjacent conglomerate (shortened 45 - 70%) (fig. 15), and probably reflect the ductility contrast during deformation. Both values could represent the combined effects of  $D_1$  and  $D_2$  strain. However no evidence of  $S_2$  was found in the conglomerate (unit 3a), quartzite (unit 4a) or hornblende-biotite

gneiss (unit 1c) of the Ompah synform. The lack of  $S_2$  and the absence of folded pebbles suggest that strain was primarily accomplished during a single phase of deformation ( $D_1$ ).

In the Flinton belt, 30 km. southwest of the study area (fig. 2),  $F_2$  folds are at an oblique angle to  $F_1$  and the effects of the two deformations are more easily separated. Elongate pebbles in conglomerate lie in the  $S_1$  foliation surface, and plot within the field of constrictional strain (Thompson 1972). A comparison of Thompson's stereographic plots of the orientations of the  $L_1$  stretch lineations and the  $F_2$  fold hinges shows that the former have the steeper plunge; and so it seems unlikely that the constrictional strain and pebble elongation is due to the intersection of  $S_1$  with  $S_2$ . In fact Thompson remarks that in the area where there is most evidence of  $D_2$ , the pebbles have been flattened to "pancakes" and do not define a lineation at all, the reverse of what one would expect if the superposition of two flattening strains had combined to form a pronounced constrictional strain.

#### Comparison of $D_1$ and $D_2$ Deformations

The information presented indicates that  $D_1$  deformation consisted of folding and the formation of a penetrative low dipping foliation, accompanied by distortion which on a small scale, though variable both in type and magnitude, resulted in axial extension and a stretch lineation, especially in the

hinge region of  $F_1$  folds (fig. 15). In contrast, whilst the second deformation rotated the  $F_1$  folds and produced some large scale upright  $F_2$  folds in competent lithologies, it was predominantly a deformation by translation and rotation of large masses of rock as relatively rigid bodies-distortion was not penetrative on the crystalline (or pebble) scale. For example unit 1c, which is folded by the  $F_2$  Plevna synform, shows little evidence of internal  $D_2$  deformation (ie. it typically lacks a foliation  $S_2$ , and a lineation,  $L_2$  over much of its outcrop), its profile is only slightly flattened (fig. 10), yet the limbs are isoclinal (fig. 4). Folding such as this must have been accomplished at the expense of considerable, localized strain in the intervening rocks, especially marble, for which there is no evidence due to subsequent pervasive recrystallization.

#### Regional Correlations

The study by Thompson (1972) of Flinton Group rocks 40 km. southwest of the Ompah area (fig. 2) provides an obvious source for comparison with the present work. Broadly speaking the two structural sequences are compatible, though some minor differences emerge.

Deeply infolded keels of Flinton Group rocks trend NNE in the area around Flinton, compared to the NE trend at Ompah. In both areas the keels are  $F_1$  features which display remarkable continuity for their narrow width. In the Ompah

area this is easily understood, since  $F_1$  folds and  $L_1$  lineations (defined by  $S_0/S_1$  intersections and elongation) are gently plunging to sub-horizontal. In the Flinton area, however,  $L_1$  axial lineations (intersections of  $S_0$  and  $S_1$ ) vary from horizontal to almost vertical, and Thompson (1972, p. 67) records that "the elongation lineation is parallel to the hinge lines of the first phase folds in.....the Black River area (near Madoc-see fig. 2), but over the rest of the study area the elongation direction commonly plunges more steeply than the folds". Thompson interprets the spread of the orientation of  $F_1$  fold axes and  $S_0/S_1$  intersection lineations as due to inhomogeneous strain during  $D_1$ , causing a variable degree of rotation of  $F_1$  hinge lines (Ramsay 1962). Whilst this explanation is plausible, it is difficult to reconcile the steep plunges of the  $L_1$  lineations with the narrow, but continuous outcrop of rocks in the Flinton syncline, unless the small-scale folds are now not parallel to major folds. Following Thompson's assumption that the elongation lineation was parallel to the  $S_0/S_1$  intersection before additional strain, it is apparent that extension, initially parallel to  $F_1$  fold axes, is common to both regions.

Besides in the Flinton area where constrictional strain has been measured (Thompson 1972), linear fabrics in conglomerate have been reported by Carmichael (1968) and Divi (1972) in the Bancroft-Madoc region. These authors note

(in areas D and C respectively in fig. 2) that the longest axes of deformed pebbles are parallel to  $F_1$  fold axes, and they consider this to be the result of extension during  $D_1$  (whether strain was constrictional or flattening is not known, since neither author includes "Flinn plots" of their data). Venkitasubramanyan (1969) also notes that the long axes of deformed pebbles are parallel to fold axes in area B (fig. 2), but he does not indicate whether the folds are  $F_1$  or  $F_2$ . However the steep, reclined orientation of the long axes of the pebbles is parallel to that recorded by Thompson (1973) 15 km. further west, where extension is parallel to  $L_1$ , so a similar relationship appears likely. Extension parallel to  $F_1$  fold axes is therefore a characteristic of the  $D_1$  deformation in this region of the Grenville Province. Constrictional strain, however, may be restricted to the hinge regions of  $F_1$  folds.

Although axial extension is common, the scale and orientation of  $F_1$  folds are not constant from area to area. In the Actinolite-Kaladar area (B in fig. 2) large-scale low plunging  $F_1$  folds have been recorded by Venkitasubramanyan (1969), who considers them to have been originally northwest trending, and in the Whetstone Lake area (D in fig. 2).. Carmichael (1968) recognizes large-scale  $F_1$  folding with originally moderately plunging northeast trending axes. North and west of Bancroft a northeast trending structural

lineament. Separates Grenville Supergroup rocks from basement gneisses. Work in the Grenville Supergroup rocks by Divi (1972 area C, fig. 2), and in basement gneisses by Appleyard (1974 50 km. NE of Bancroft) and Francoeur (1975 40 km. west of Bancroft) has revealed mesoscopic, detached  $F_1$  fold hinges, which may have had a northeast original trend (Francoeur 1975), but no large scale  $F_1$  folds were identified.

Despite the varied orientation of  $F_1$  fold axes in the region, all the authors referred to except Thompson, consider that the  $S_1$  axial surface was originally gently dipping to horizontal<sup>1</sup>, and this is in accord with the present interpretation in the Ompah area. In the Flinton area, however, Thompson (1972) considers that  $S_1$  was steep before  $D_2$ , and that its attitude was not greatly altered during  $D_2$  deformation. It is not immediately clear what caused this divergent, steep orientation; possibly the presence of the large bodies of the Elzevir and Northbrook plutons and the extremely narrow outcrop of the Flinton belt are significant in this respect.

Folds of the second generation in the Ompah and Flinton areas are mutually parallel, being coaxial to  $F_1$  in the Ompah region and oblique to  $F_1$  around Flinton. In contrast to the Ompah area, in the region around Flinton  $F_2$  folds are

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<sup>1</sup> Recent mapping in the Clare River area (directly east of area B, fig. 2) has also indicated a subhorizontal attitude for  $S_1$ —J.F. Chappell (oral communication 1975).

restricted to the mesoscopic scale or smaller; the lack of megascopic folding may also be attributable to the presence of the Elzevir and Northbrooke plutons. Large scale north-east trending  $F_2$  folds have been recorded in area B (fig. 2) in both metasedimentary/metavolcanic sequences and in plutonic rocks intrusive into them (Venkitasubramanyan 1969).

Carmichael (1968) observes that  $F_2$  folds are coaxial with  $F_1$  folds in area D, fig. 2, and a similar conclusion is tentatively drawn by Divi (1972) near Bancroft. In both areas  $F_2$  folds are northeasterly trending, and the axial surfaces generally have a steep southeast dip, as in the Ompah area. It appears that  $S_2$  is a more penetrative feature around Bancroft (Divi 1972) than in the Ompah and Flinton areas.

Divi (1972) and Divi and Fyson (1973) note that  $F_2$  fold axes become progressively steeper as a syntectonic granite pluton is approached. Furthermore they show that there is a direct correlation between the plunge of the  $F_2$  fold axes and the profiles of the folds, with the steeper plunging folds near the pluton being more flattened than those with more moderate plunges away from the intrusion (flattening calculations by the method of Ramsay 1962). A similar increase in plunge of minor folds towards a pluton is also qualitatively recorded by Heidecker (1963) 50 km. SSW of Bancroft. Divi and Fyson (1973) suggest that the steep

plunges and flattened profiles near the margins of the pluton are best explained by assuming that extension parallel to  $F_2$  fold axes took place during rotation and flattening of the folds, in a manner discussed theoretically by Ramsay (1962).

$F_3$  folds are northeast trending, recumbent structures in areas C and D (fig. 2) as in the Ompah area, but around Flinton, Thompson (1972) recognizes northwest trending, vertical  $F_3$  folds. In basement gneisses northeast and west of Bancroft,  $F_3$  folds, which may be of very different age to those in the cover rocks, are upright, open structures with subhorizontal, generally northeast trending axes (Appleyard 1974, Francoeur 1975).

#### Speculations

Generalizing from the studies of deformation in cover (Grenville Supergroup) rocks in the Bancroft-Madoc region (fig. 2), it is apparent that the first phase folds were generally subhorizontal to recumbent, and were associated with a penetrative axial planar fabric, whilst in the second deformation folds were upright, and associated with a fabric which increased in intensity towards basement gneisses. Within the basement gneisses  $F_1$  folds are recumbent, and  $F_2$  folds may be either recumbent (Appleyard 1974) or with gently S.E. dipping axial surfaces (Francoeur 1975), with the earlier structures being largely obliterated.

The orientation of later folds in cover rocks is not consistent throughout the region, but recumbent northeast trending structures are common to several areas. In basement gneisses large scale open folds and warps with upright axial surfaces and low plunging axes are typical.

Marked differences in attitude of structures of different deformational episodes has been observed in other younger regional metamorphic terrains, and may be characteristic of different structural levels during deformation. Zwart (1964a, 1964b) and Fyson (1971) have shown that the orientation of folds may change radically in a continuous structural succession upwards from the "infrastructure" to the "superstructure". The infrastructure is characterized by recumbent folds, which gradually die out upwards and are replaced by upright folds in the superstructure. Fyson (1971) notes that folding at the two levels is generally not contemporaneous, and early recumbent structures in the infrastructure may be refolded by upright structures that are more prominent in the superstructure as the locus of folding migrates upwards. The upper levels of the superstructures are characteristically unaffected by the recumbent  $F_1$  folds, hence these rocks must have been eroded away in the Bancroft-Madoc region, exposing the level where the two sets of discordant structures are superimposed. In basement gneisses, where a deeper section is exposed, both  $F_1$  and  $F_2$  folds are recumbent, with  $F_3$  folds

being comparable in attitude to  $F_2$  in cover rocks.

The recumbent attitude of the early folds is considered to be a result of horizontal extension (folds are typically flattened), whilst during later upright folding, bulk extension has a large vertical component (Fyson 1971). In the Bancroft-Madoc area, the flattened profiles of recumbent  $F_1$  folds are a reflection of horizontal extension, which is especially prominent parallel to  $F_1$  fold axes; whilst the results of Divi and Fyson (1973) suggest that vertical extension is at least locally associated with upright  $F_2$  folding. The orientation of the recumbent  $F_3$  folds in cover rocks does not fit this pattern, and the reason for their attitude is not obvious. However it could be related to the upward push of nearby plutons.

The presence of an element of extension parallel to  $F_1$  fold axes may be characteristic of deep level tectonics. Recently Wood (1974) has assembled a large number of strain measurements made in Palaeozoic slate belts, where deformation style is characteristic of the superstructure. Plotting all the measurements on a strain graph (similar to fig. 15) and contouring the result, shows that most slates plot in the  $S \setminus L$  to  $S \gg L$  fields, and have a fairly restricted range of strain magnitude (fig. 18B). The relatively small range of both type and magnitude of strain may be due to vertical bulk extension being generally feasible in these terrains.

In contrast, the wider range of both type and magnitude of strain in the Bancroft-Madoc region (fig. 18A) suggests that stress relief was variable, at least during  $D_1$  deformation. This may be due in part to the different lithologies used for measurement, but the presence of large plutonic bodies during deformation is also probably significant. These may both locally reorientate stress directions and influence strain response. Thus in terrains where the infrastructure is exposed, metamorphic grade is high and plutonic activity is in part pre-tectonic, variable strain types and magnitudes may be typical, reflecting a more complex pattern of bulk extension directions during deep level deformation.

## METAMORPHISM

### Introduction

#### Grade of Metamorphism

A general increase in metamorphic grade northeastwards in the Ompah area was noted by Smith (1958), but details awaited the work of Moore (1967). On the basis of mineral assemblages in pelites (principally confined to the Fernleigh belt), Moore defined three metamorphic zones, all within the amphibolite facies, characterized by the index minerals (A) chloritoid - staurolite, (B) staurolite - kyanite and (C) sillimanite (fig. 19). Kyanite is also present in the lowest grade rocks (zone A). There is no separate kyanite zone, since the appearance of sillimanite coincides with the disappearance of staurolite. Garnet is not a useful mineral as a zone indicator, as it is present along the length of the belt; probably it has been derived in a variety of different metamorphic reactions. In the present study the position of the isograd between the staurolite - kyanite and sillimanite zones (Moore 1967) in pelitic schist (unit 6b) has been confirmed, and the position of another isograd representing the replacement of muscovite and quartz by the pair sillimanite - potash feldspar has been tentatively defined (fig. 19).

#### Composition of Pelites in the Fernleigh Belt

There is some variation in bulk composition of the pelitic

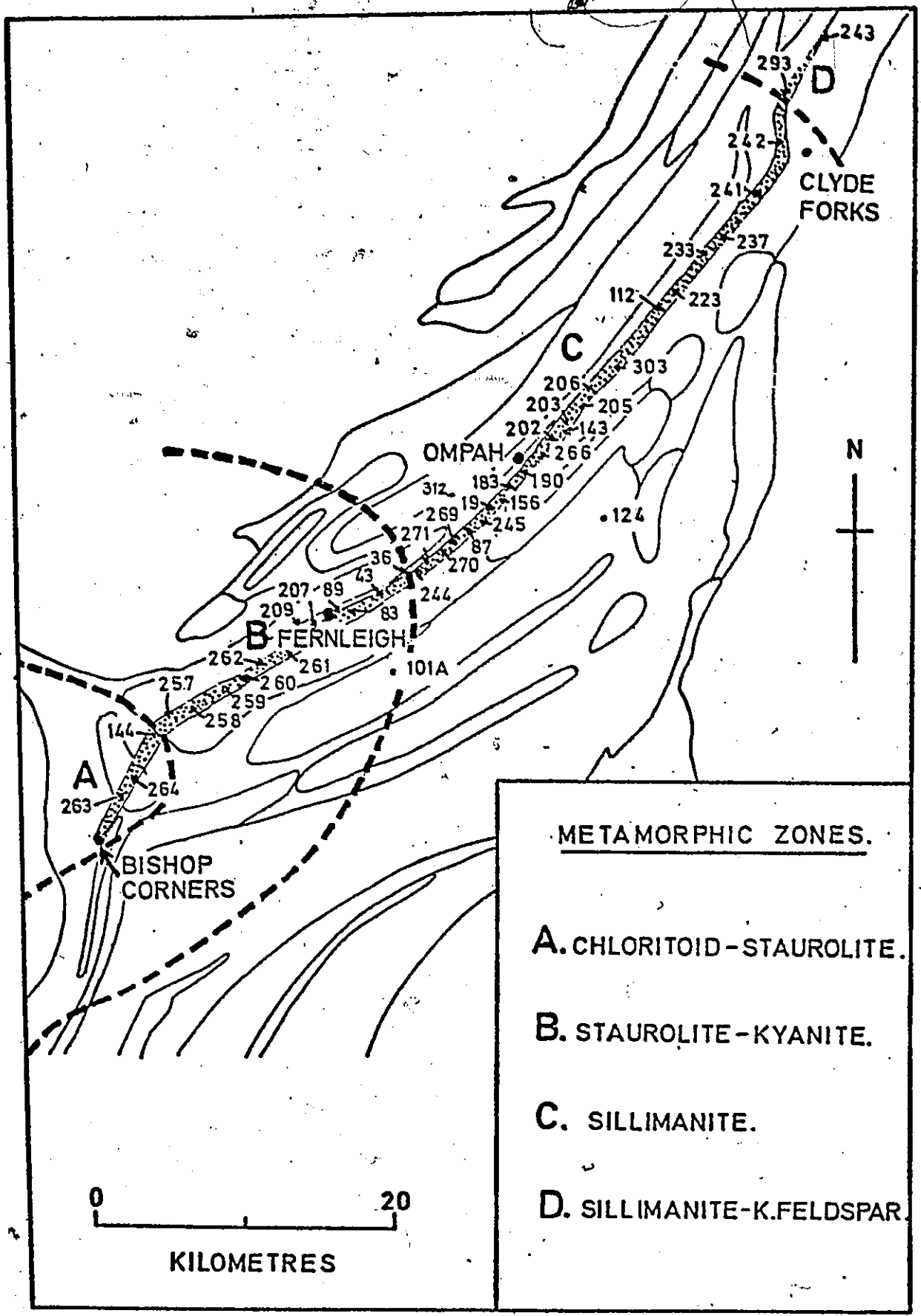


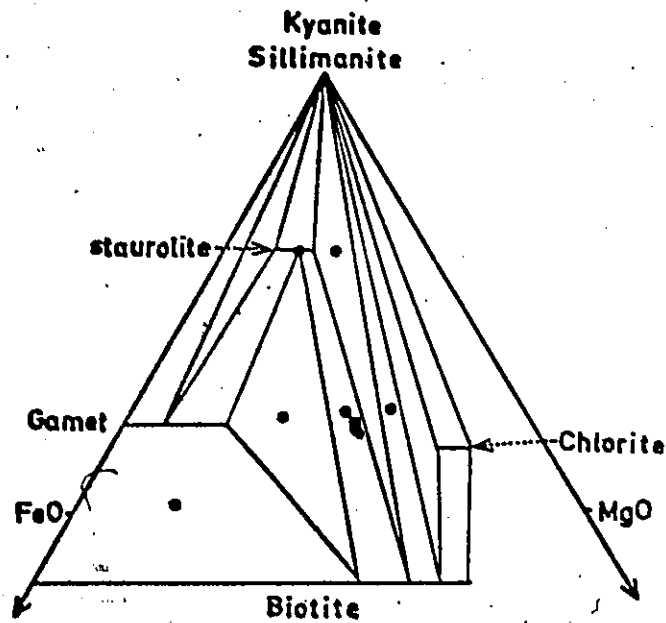
Fig. 19 Distribution of isograds in pelitic schist (unit 6b) of the Fernleigh belt (stippled), after Moore (1967) and this work. The locations of samples used for metamorphic and textural analysis are shown.

schist (unit 6b) in the Fernleigh belt, as is shown in fig. 20, which is taken from Hounslow and Moore (1967). In the AFM diagram it is apparent that the aluminium content, as well as the Fe : Mg ratio, shows considerable variation. Elsewhere in the same paper, Hounslow and Moore describe considerable differences in the rock oxidation state in sub-units of the pelitic schist (not shown in fig. 20). In the area under study the most magnesium rich assemblages (ie. chlorite-kyanite and chlorite-biotite) were not observed, any chlorite present being considered retrograde.

#### Reaction Mechanism

In order to understand the origin of isograd surfaces, it is necessary to have a knowledge of the mechanism of relevant metamorphic reactions. Many reactions have been conceived as isochemical changes, but in recent years the importance of a fluid phase carrying ionic as well as neutral species has been emphasized by some authors (Helgeson 1967, Eugster 1970). Eugster has distinguished between thermal and ionic equilibria, the former being isochemical except for the loss or gain of a neutral gas species such as  $H_2O$  or  $CO_2$ , whilst the latter are determined by local activities of mobile ionic components.

Carmichael (1969) has shown from a study of rock textures that certain common prograde reactions, which on a gross scale have generally been regarded as thermal equilibria, may in fact represent a series of interrelated ionic equilibria at



Assemblages include quartz,  
muscovite and oligoclase.

Fig. 20 (modified from Hounslow and Moore 1967). AFM diagram of pelitic assemblages in the Fernleigh belt considered in this work, represented on the AFM face of the AFMK tetrahedron by projection through muscovite. The magnesian assemblages noted by Hounslow and Moore (1967) were not encountered.

the crystalline scale.

In the sections which follow, the empirical method of Carmichael, together with the theoretical background of Eugster and Helgeson and others are used to draw up possible reaction mechanisms at isograd surfaces in the study area.

Reactions to be considered

Three thermal equilibrium reactions will be considered, and their applicability to the textures of the pelitic schist (unit 6b) in the Fernleigh belt are evaluated. The reactions are (a) staurolite breakdown in the presence of quartz, (b) the polymorphic transformation kyanite to sillimanite and (c) muscovite breakdown in the presence of quartz.

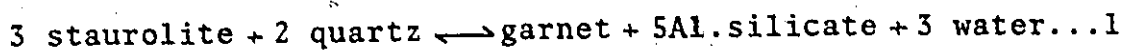
In the discussion which follows, the reactions are balanced for the following idealized mineral compositions.

|           |                                                       |            |                                                                  |
|-----------|-------------------------------------------------------|------------|------------------------------------------------------------------|
| Quartz    | $\text{SiO}_2$                                        | Biotite    | $\text{K}(\text{MgFe})_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$ |
| Albite    | $\text{NaAlSi}_3\text{O}_8$                           | Garnet     | $(\text{MgFe})_3\text{Al}_2\text{Si}_3\text{O}_{12}$             |
| Muscovite | $\text{KA}l_2\text{AlSi}_3\text{O}_{10}(\text{OH})_2$ | Staurolite | $(\text{MgFe})\text{Al}_4\text{Si}_2\text{O}_{10}(\text{OH})_2$  |

|             |   |                            |
|-------------|---|----------------------------|
| Kyanite     | } | $\text{Al}_2\text{SiO}_5$  |
| Sillimanite |   |                            |
| K. Feldspar |   | $\text{KAlSi}_3\text{O}_8$ |

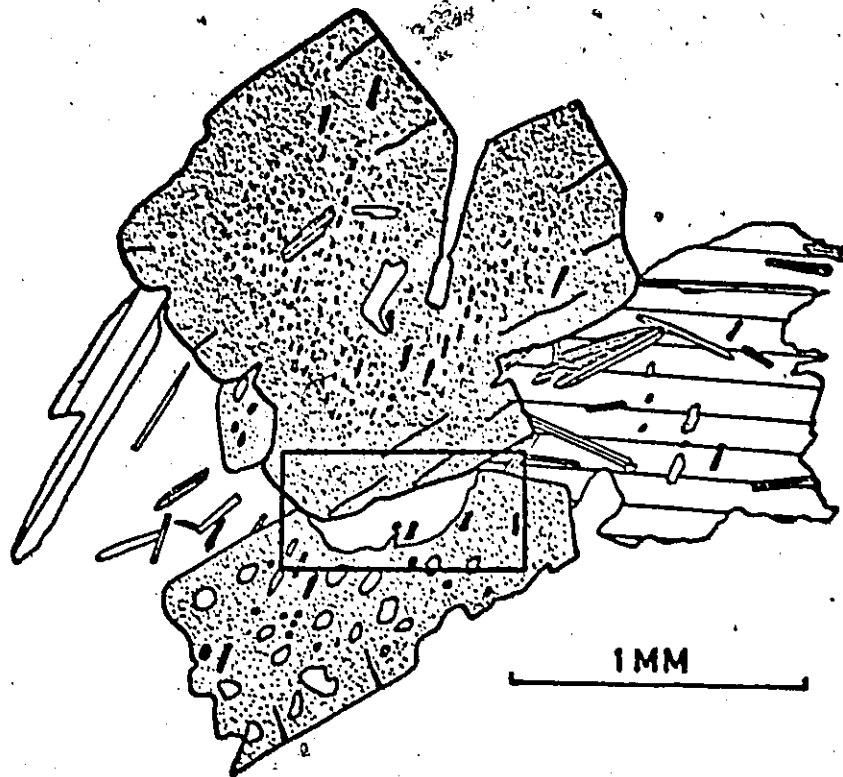
### Staurolite Breakdown Reaction

The reaction at this isograd has commonly been considered to involve the thermal equilibrium



However this reaction, as well as requiring  $\text{Fe}^{++}$  to enter and  $\text{Mg}^{++}$  to leave the system to balance the stoichiometric requirements of these elements in garnet and staurolite, also involves a substantial decrease in volume assuming all water is lost (17 - 23% depending upon Al. silicate), which may render it a somewhat improbable phenomenon. Possible evidence of this reaction, in which crystals of aluminium silicate and garnet occur adjacent to crystals of quartz and embayed staurolite, was found only once in the textures of the Fernleigh belt and apparently elsewhere it is not common, since Chinner (1965) writes "the frequency of its invocation seems to be due more to the elegance of its simplicity than to any indication of its actual occurrence". The sole example from the Fernleigh belt is illustrated in fig. 21. This too, is somewhat equivocal on two counts. Firstly the aluminium silicate product of the reaction, sillimanite, is extremely fine, such that its optical properties cannot be determined accurately. The second reason is that apart from this questionable occurrence of sillimanite, the specimen was considered to be from the kyanite - staurolite zone, albeit close to the isograd. However despite these drawbacks, the rela-

A.



B.

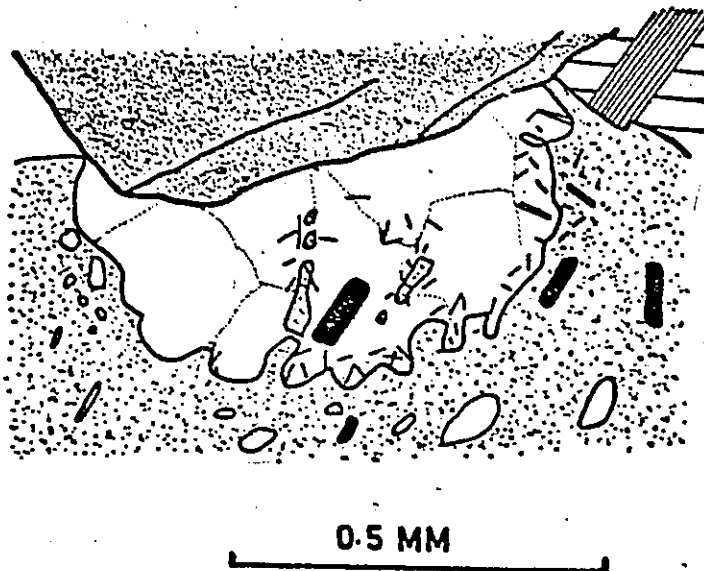
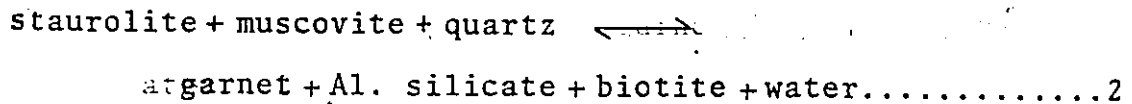


Fig. 21. Drawing of a texture illustrating evidence for staurolite breakdown to garnet and sillimanite. (A) A euhedral porphyroblast of garnet (heavy outline, close stipple) is contiguous with a porphyroblast of staurolite (light stipple) which is rounded and embayed in the area of the contact. Adjacent is a large crystal of biotite (widely spaced cleavage); which encloses small elongate crystals of muscovite (closely spaced cleavage). The contact between garnet and staurolite is enlarged in (B), where many small needles of sillimanite in quartz (blank) and staurolite are seen. Drawn from specimen 43B.

tive volumes of sillimanite and garnet in the texture, c. 1:100, clearly indicate that reaction 1, if it took place, could not have been isochemical on the scale of the texture, since the amount of sillimanite produced by the equilibrium reaction (1) would have been greatly in excess of that observed in the rock. Whilst the possibility exists that some of the garnet was present before the reaction began, the limited amount of sillimanite present in this rock suggests that in any case this was not an important staurolite breakdown reaction. It is to be noted, however, that intergrowths of garnet and sillimanite are common in the sillimanite zone and higher grades, though whether these are replacements after staurolite is not known.

Hounsflow and Moore (1967) and Carmichael (1969) instead proposed the reaction



Since solid solution in staurolite is known to be limited, and since the Mg/Fe ratio of staurolite is intermediate between coexisting biotite and garnet, this reaction might be expected to take place over quite a limited range of temperature and pressure. The occurrence of the six phase assemblage quartz-staurolite-muscovite-garnet-Al. silicate-biotite can therefore be used to demarcate the isograd. Carmichael reports that the occurrence of the six phase

assemblage in the Whetstone Lake area (area D on fig. 2) actually occurs over a distance of some 2 km., indicating that restricted element partition between the phases probably does take place. In the Fernleigh belt the six phase assemblage occurs over a distance of about 7.5 km. (between localities 207 and 269, see fig. 19), with either kyanite and/or sillimanite as the aluminium silicate present. However the lower grade occurrences of this assemblage (locality 207) may not be in equilibrium. Hounslow and Moore (1967) suggest (for their locality 7, which appears to be coincident with 207 in this study) that the garnet may be an "extra" phase stabilized by manganese. In the thin sections examined in the present study kyanite appears to be the unstable phase. The kyanite crystals are situated close to a small quartz vein, and show considerable evidence of resorption. If this locality is omitted, the other occurrences of the six phase assemblage (43, 271 and 269) span a distance of about 3.8 km (Oddly no specimens containing the six phase assemblage were found in either of localities 89 or 36, despite extensive sampling - this is presumably a bulk composition effect). The centre of the isograd zone so defined is coincident with the boundary between the staurolite-kyanite and the sillimanite zones, as drawn by Moore (1967) - fig. 19. Up grade from 269, the six phase assemblage is also found in localities 87, 156, 312 (not situated on the Fernleigh belt) and surprisingly at 243, the highest grade locality. However in

these, staurolite is present only as tiny rounded inclusions inside porphyroblasts of plagioclase feldspar - this texture, whilst clearly a relic of a lower grade of metamorphism, must be considered in disequilibrium, the staurolite being shielded from reaction by the plagioclase.

It is therefore evident that at close inspection in the field the isograd, when defined by a single thermal equilibrium reaction, is not a clearly marked line.

A possible mechanism for the reaction in rocks containing the six phase assemblage staurolite-muscovite-quartz-Al. silicate-garnet-biotite has been discussed in detail by Carmichael (1969). This will be briefly reviewed here, as it is probable that it is pertinent to at least some of the six phase assemblages in the Fernleigh belt. Carmichael notes the mutual separation of staurolite and garnet and staurolite and sillimanite. Instead staurolite is replaced by oligoclase and biotite (reaction C, fig. 22). Sillimanite commonly occurs as intergrowths with biotite and muscovite, and this assemblage is considered to grow at the expense of oligoclase (reaction D, fig. 22). Similarly the textures indicate that garnet grows from oligoclase and biotite (reaction A, fig. 22). Finally reaction B (fig. 22) has been written so that the system is closed on the scale of the thin section, except to water. Carmichael states (1969, p. 262 - 4) "The justification for linking the partial reactions to one another, and adding them together to get the net reaction, is the assump-

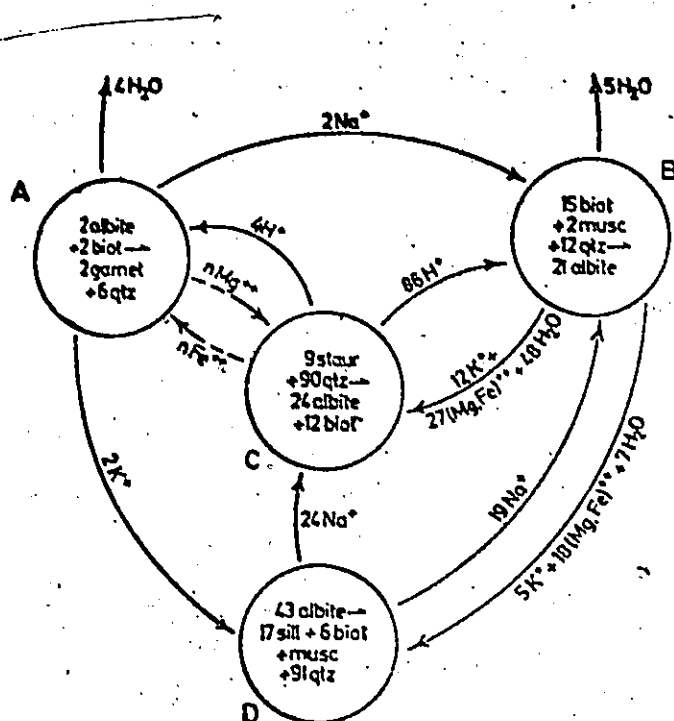
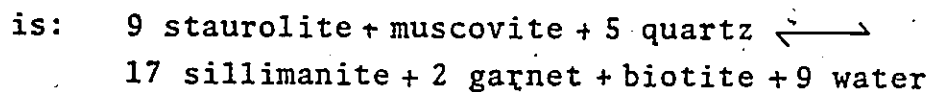


Fig. 22 (taken from Carmichael 1969).

Model of a possible reaction mechanism at the sillimanite-garnet-biotite isograd. If reactions at A, B, C and D go from left to right, the net change taking place in the system



tion that the system is closed, except to the escape of water. The partial reactions cannot proceed independently without appropriate changes in the bulk composition of the rock".

Carmichael's explanation would appear to be a plausible mechanism for the reactions which took place in the equilibrium (6 phase) assemblages. However in the Fernleigh belt it has been observed that in staurolite-muscovite-quartz bearing assemblages, in which staurolite is showing unmistakable signs of resorption and/or replacement, either andaluminium silicate or garnet occur with biotite much more commonly than the three phases occur together. It therefore appears that either staurolite breakdown was frequently not by a reaction such as 2, or that some of the products have diffused out of the local system, i.e. that the bulk composition of the rock must have changed. Some of these textures will now be discussed in detail, in an effort to discover something of the mechanism for these reactions.

At locality 207 (see fig. 19) in the staurolite-kyanite zone, specimen 207C consists of the assemblage muscovite-quartz-staurolite-plagioclase-kyanite-biotite-tourmaline-haematite-magnetite and ilmenite. Staurolite crystals are poikiloblastic, with inclusions of quartz and fine opaque ores, and in one example the porphyroblast is clearly being replaced by the pair kyanite-biotite (fig. 23). Garnet is not present in the thin section or hand specimen. The reaction

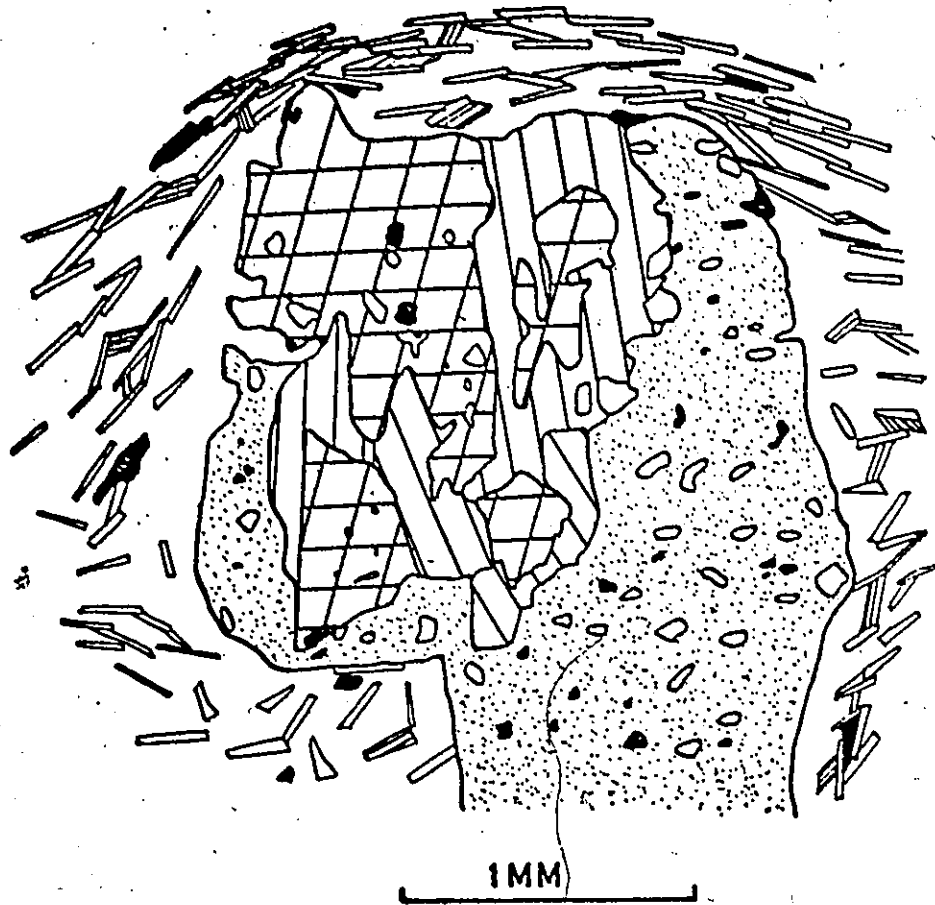
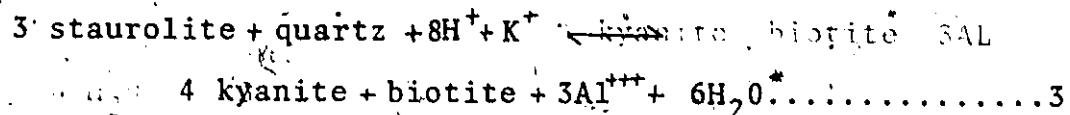


Fig. 23 Drawing of a texture in which a porphyroblast of staurolite (stippled) is partially replaced by an intergrowth of kyanite (slanted cross-hatch) and biotite (widely spaced cleavage). The matrix consists of quartz (blank), muscovite and opaques (black). Drawn from specimen 207A.

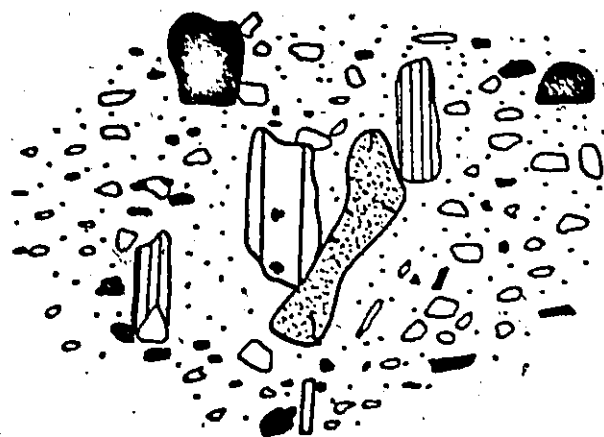
which may describe this texture is:



This equation satisfies the observation that replacement was essentially at constant volume, and that the products, kyanite and biotite, are formed in about equal proportions. The reaction, when proceeding from left to right, produces excess aluminium, which must leave the local reaction system. In nature the Fe: Mg ratio of biotite is lower than that of coexisting staurolite (Thompson 1957, Chinner 1965 and Hounsflow and Moore 1967). This could be achieved by an exchange reaction between newly formed biotite and unreacted staurolite whilst some of the latter still remains, but if the reaction goes to completion it will be necessary for some  $\text{Fe}^{++}$  to leave and /or  $\text{Mg}^{++}$  to enter the local reaction system. The excess aluminium on the right hand side of the equation may go to produce muscovite or additional kyanite outside the area of the texture. As to the origin of the  $\text{H}^+$  and  $\text{K}^+$  in this and other equations, Wintsch (1975a) has suggested that the recrystallization of muscovite can significantly alter the activities of these elements in the inter-

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\* Morey (1957) has produced experimental evidence of the solubility of silica in water, probably in the form of a complex  $\text{Si}(\text{OH})_4$ . However formation of hydrous aluminium complexes is not well known, so in this work aluminium will be considered to migrate as ions in the intergranular solution. An aluminium hydroxyl complex though, if formed, would not significantly alter the results.

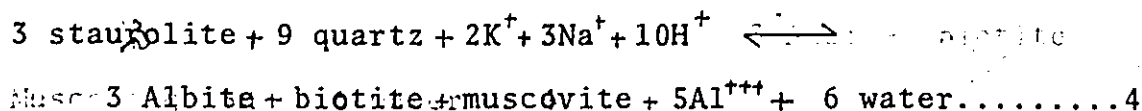


1 MM

Fig. 24 Drawing of a texture in which staurolite and quartz are replaced by oligoclase, biotite and muscovite. Rounded remnants of staurolite (close stipple) are surrounded by small crystals of biotite (widely spaced cleavage) and muscovite (closely spaced cleavage) - the whole texture is included in a 4 cm. long porphyroblast of plagioclase feldspar (oligoclase - open stipple), in which there are many inclusions of quartz (blank) and opaques (black). Drawn from specimen 89A.

granular solution. As new (001) surfaces of the mica are exposed by growth or deformation, the  $a_{K^+}/a_{H^+}$  ratio may fluctuate as instantaneous surface exchange reactions take place, in which  $H^+$  ephemerally replaces  $K^+$  in the muscovite lattice. As a result the  $K^+$  ions may be available for reactions such as 3. The  $H^+$  may also originate from the dissociation of water produced by dehydration reactions during prograde metamorphism. Following this line of reasoning an endogenous source of ions is available, so large scale metasomatism i.e. change in bulk composition, is not necessarily inferred when ionic reactions are postulated.

At locality 89, within the 3.8 km wide isograd zone, evidence of staurolite replacement by oligoclase, biotite and muscovite is found in a quartz-muscovite-oligoclase-biotite-staurolite-magnetite-haematite rock, specimen 89A. This rock contains large (up to 4 cm long) porphyroblasts of plagioclase feldspar, and within one of these the remains of a staurolite crystal, associated with biotite and muscovite is found, fig. 24. Staurolite occurs only as inclusions in the porphyroblasts, and not as a matrix mineral. The plagioclase, which elsewhere contains abundant inclusions of quartz, is free of inclusions in the region of the staurolite, attesting to the involvement of quartz in the reaction, so that



may explain this texture. The amount of  $Al^{+++}$  which leaves and  $Na^+$  which enters the reaction site may both be slightly reduced from the proportions indicated in equation 4, since oligoclase is more aluminous than the ideal albite composition used. As before, some magnesium may enter the system to accommodate the lower Fe: Mg ratio of biotite than staurolite, or excess  $Fe^{++}$  may crystallize as an opaque phase by altering slightly the magnetite: haematite ratio, or may leave the system. This reaction may have been quite a common one during prograde metamorphism in the Fernleigh belt, since inclusions of staurolite ( $\pm$  biotite and muscovite) are found locally in plagioclase porphyroblasts throughout the sillimanite and sillimanite - K.feldspar zones.

In another specimen from the same locality (89I), the mineralogy of which consists of quartz-garnet-staurolite-muscovite-biotite-oligoclase-ilmenite-magnetite-tourmaline, staurolite is replaced by garnet and quartz in the immediate region of the texture (fig. 25). Mineralogically this rock differs from those considered previously in that the amount of biotite is much greater than muscovite (ratio c. 9:1). Plagioclase is common as a matrix mineral, (plagioclase > quartz in some parts of the section), but it does not attain porphyroblast size. Both plagioclase and staurolite occur as rounded inclusions in garnet, suggesting that a reaction such as

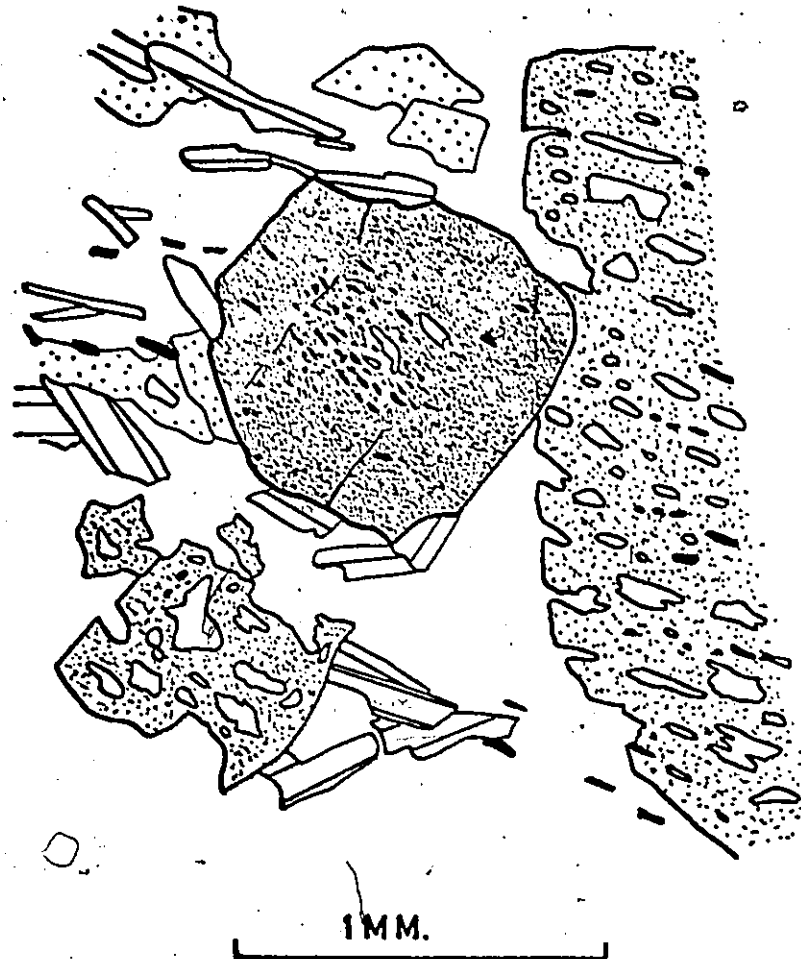
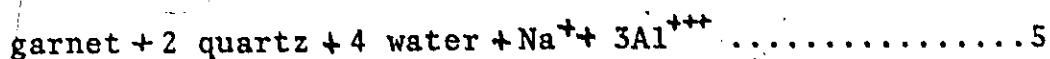
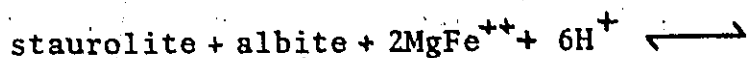


Fig. 25 Drawing of a texture illustrating the production of garnet and quartz at the expense of oligoclase and staurolite. A euhedral porphyroblast of garnet (heavy outline, close stipple) has partially replaced a porphyroblast of staurolite (intermediate stipple)- staurolite also occurs as smaller anhedral remnants. Biotite (widely spaced cleavage), oligoclase (light stipple) and quartz (blank) occur in the foliation. Drawn from spec. 89I.



may have taken place. In this reaction staurolite, which has approximately the same molar volume as garnet, may be replaced at constant volume by that mineral (see fig. 25). The reaction could have been written without involving as much transport of the ferromagnesian elements in the intergranular fluid. However this would produce a considerably greater excess of  $\text{Al}^{+++}$  on the right hand side of the equation, so considering the high proportion of biotite in the rock, reaction 5 is preferred. The excess aluminium and sodium on the right hand side of the reaction may combine to form muscovite outside the immediate region of the texture. In several cases muscovite crystals were observed orientated at high angles to the foliation, suggesting late growth possibly associated with the staurolite reaction.

The above examples represent only three of the more common staurolite breakdown textures in the Fernleigh belt. Other reactions, such as staurolite replacement by muscovite (Guidotti 1968) have been documented, and can be treated in a similar manner.

#### Polymorphic Transformation of Kyanite to Sillimanite

On a regional scale, the polymorphic aluminium silicate transformation

kyanite  $\longleftrightarrow$  sillimanite.....6  
describes this isograd. In the Fernleigh belt assemblages with both kyanite and sillimanite occur over a distance of 5.7 km. (localities 43, 36, 244, 271, 270, 269 and 87). If we omit the controversial occurrence of sillimanite at locality 43, described previously, this distance is reduced to 3.5 km.

If the polymorphic transformation (reaction 6) actually took place during prograde metamorphism, it would be expected that the textures would record the partially completed reaction of sillimanite growing on earlier kyanite crystals; and also that evidence of replacement of kyanite be limited to the 3.5 km. wide isograd zone where the two polymorphs occur together. In fact, in the Fernleigh belt neither is the case. On the outcrop scale Moore 1967, p. 332, notes (of locality 36), "some layers contain kyanite, some sillimanite", whilst in the thin sections examined in the present study in which the two polymorphs both occur, sillimanite has never been observed to be growing directly at the expense of kyanite. This observation has been made by others eg. Tozer (1955), Chinner (1961) and is probably of general significance. The textures of the specimens containing both the aluminium silicate polymorphs will be considered first.

Kyanite and sillimanite were found together in seven

thin sections (36I, 36L, 36Q, Q71C, 270C, 269 and 87A). All of these specimens also contain staurolite, but differ in other ways such as the muscovite:biotite ratio and the presence or absence of garnet. The texture of a typical example (36Q) is shown in fig. 26. The mineralogy of this specimen is: quartz-muscovite-biotite-sillimanite-kyanite-magnetite-haematite and tourmaline. The kyanite in the section is skeletal and embayed with quartz, but still retains some rational crystallographic forms. A knot of fibrous sillimanite\* is situated some 0.5 mm. away, and is separated from the kyanite by quartz and small crystals of an opaque phase. On the other side of the kyanite crystal are situated small, anhedral blebs of staurolite, which also show signs of replacement. The aluminium for the sillimanite may therefore be derived from both kyanite and staurolite. In all the seven examples, quartz is typically the dominant mineral replacing kyanite, with muscovite and biotite also being present in some of the textures. The Al:Si ratio of kyanite is higher than the ratios in the equivalent volume of quartz, muscovite and biotite or combinations of them, so it is neces-

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\* Elsewhere in the same thin section the sillimanite is prismatic, and intergrown with opaque phases. No regular pattern to the distribution of fibrolite (disordered structure) and prismatic sillimanite with metamorphic grade was discernable.

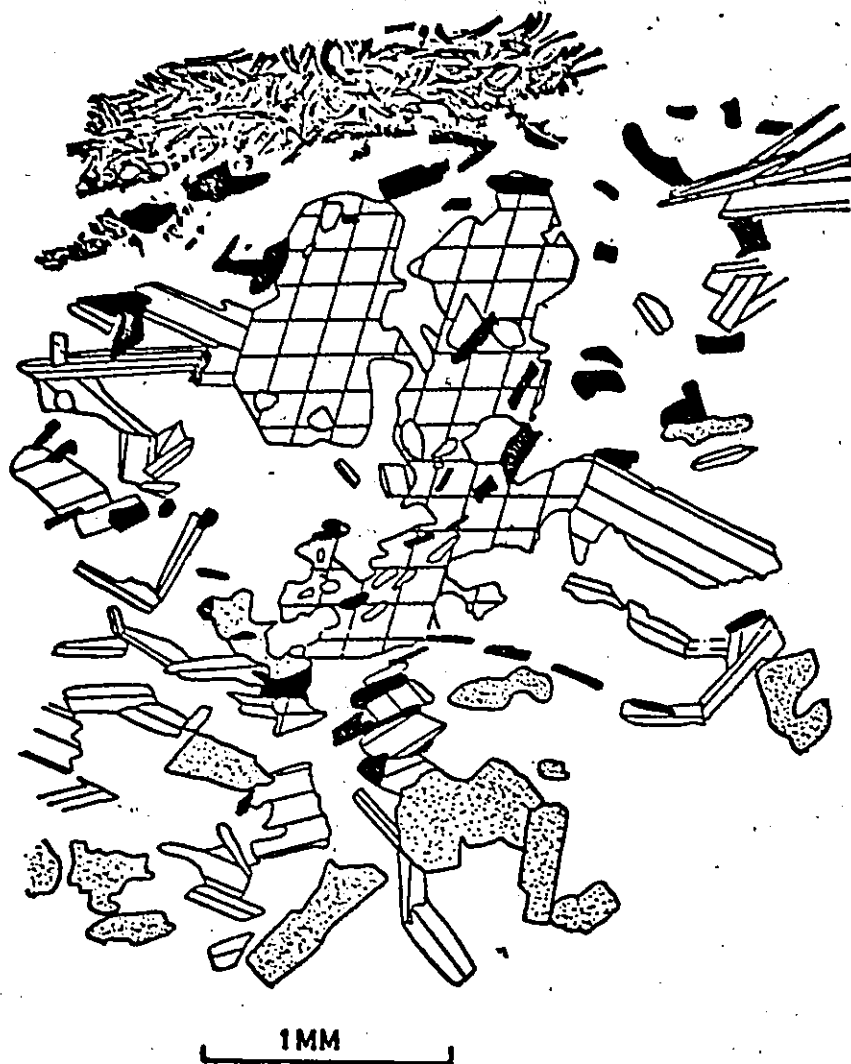
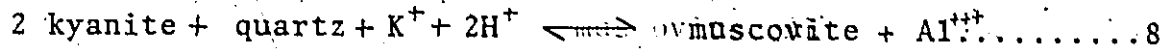
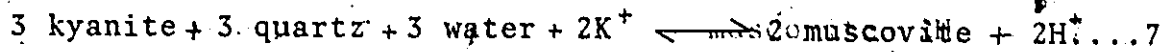


Fig. 26 Drawing of a specimen containing both kyanite and sillimanite. A skeletal kyanite crystal (slanted crosshatch) is separated by about 0.5 mm. from a knot of fibrous sillimanite (sillimanite is prismatic in top right of diagram). Kyanite replacement is mainly by quartz (blank). Anhedral blebs of staurolite (stippled) occur in the lower half of the diagram. Mica present is almost entirely biotite (widely spaced cleavage). Drawn from spec. 36Q.

sary either to involve silica (equations 7 and 8) or to remove the excess aluminium (equation 8) from the reaction site.



Equation 7 is after Wintsch 1975 (b)

For the texture illustrated in fig. 26, equation 8 is preferred since it provides aluminium for the growth of sillimanite, and it involves a smaller increase in volume on the product side than equation 7. In order that quartz be the dominant mineral replacing kyanite, as in fig. 26, it is necessary to introduce quartz from outside the region of the texture, probably in the form of an  $\text{Si}(\text{OH})_4$  complex.

In another of the specimens containing both kyanite and sillimanite (87A), droplets of kyanite are completely enclosed by porphyroblasts of muscovite. A similar texture has been illustrated by Chinner (1961), and equation 7 may represent this reaction.

However textures indicative of kyanite breakdown without the concomitant production of sillimanite are also quite common. Once again quartz-muscovite and quartz-biotite textures are seen, generally the phyllosilicate being the dominant phase by volume. In 89E, a quartz-muscovite-plagioclase-biotite-kyanite-magnetite-haematite rock, kyanite crystals are broken and show signs of replacement by biotite.

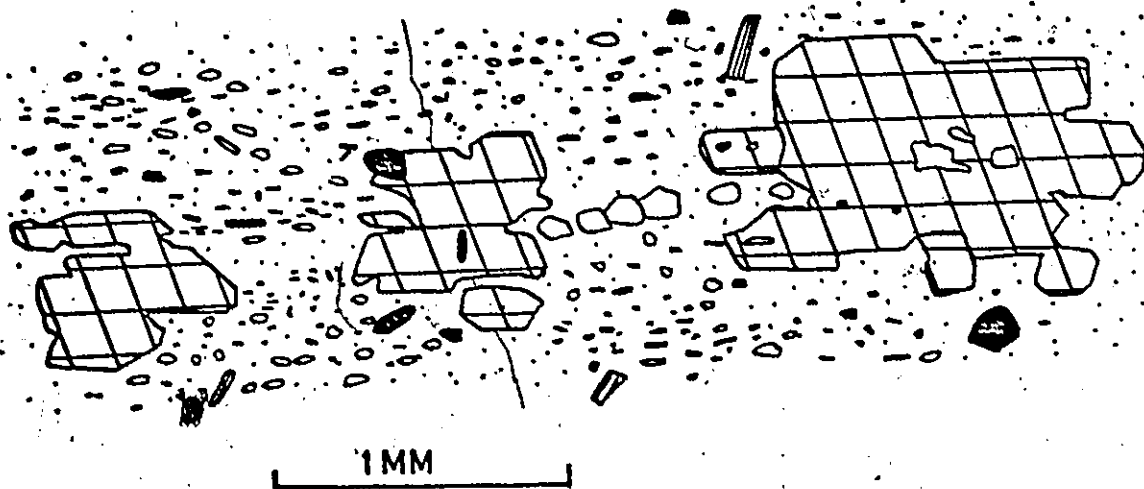
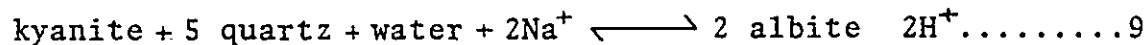


Fig. 27 Drawing of texture in which kyanite is replaced by oligoclase. Isolated remnants of kyanite in optical continuity are included in a porphyroblast of oligoclase feldspar (stippled background). Other inclusions are of quartz (blank) and opaques (black) and muscovite (closely spaced cleavage). Drawn from specimen 89H.

and quartz. An equation such as 7 or 8, but also involving the introduction of  $\text{FeMg}^{++}$  to the reaction site would provide a plausible mechanism for the formation of such a texture.

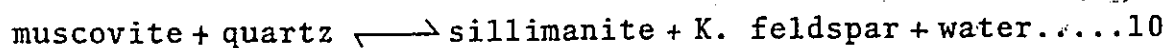
A different texture is seen in 89 H, in which kyanite crystals occur as porphyroblasts in a quartz-muscovite-opaque ores matrix, and also as inclusions in large porphyroblasts (2 cm long) of plagioclase feldspar. Evidence of replacement of kyanite by plagioclase is shown in fig. 27, in which several irregularly shaped kyanite crystals of the same optical orientation are separated by the feldspar. They are apparently remnants of a larger crystal, the long axis of which was lying in the plane of the foliation. Quartz is present abundantly as inclusions in the feldspar, so that a reaction such as



may describe this texture. The kyanite crystals in the matrix show signs of replacement by biotite, as described for specimen 89E. This would suggest that the porphyroblast and the matrix reacted as separate subsystems in this rock.

#### Muscovite Breakdown Reaction

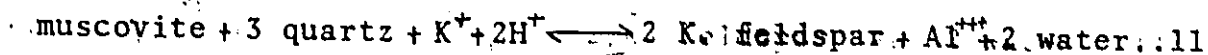
The thermal equilibrium



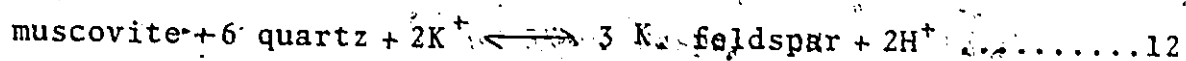
has been investigated experimentally by Evans (1965) and others, and the breakdown of muscovite in the presence of

quartz has been successfully duplicated in the laboratory. On this basis, therefore, the occurrence of the four phase assemblage muscovite-quartz-sillimanite-K feldspar can be used to delineate the isograd. In the Fernleigh belt only one occurrence of this assemblage was found (at locality 293). To the north-east of 293, muscovite, quartz and sillimanite are again seen to coexist in the absence of K feldspar. However well down-grade from 293, there is abundant evidence in this section of sillimanite growth at the expense of muscovite. Some of these textures will now be discussed in detail.

Textures of the four phase assemblage muscovite-quartz-sillimanite-K feldspar; other mineral phases present in specimen 293A are garnet-biotite-oligoclase-tourmaline. In this rock, whilst both K. feldspar and sillimanite individually are seen in juxtaposition with muscovite and quartz, the pair is not seen together, indicating that reaction 10, if it occurred, did not take place as a direct replacement of the reactants by the products. Potash feldspar crystals are anhedral, and some contain rounded inclusions of quartz together with muscovite and biotite, whilst others are essentially inclusion-free. Some muscovite crystals in contact with K. feldspar have ragged and sculpted boundaries, suggesting that muscovite has reacted to produce K. feldspar in situ. Accordingly a reaction such as

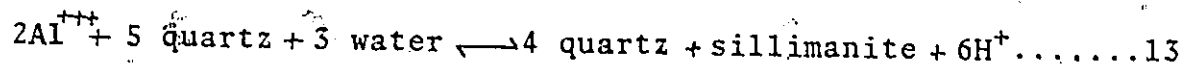


may represent this texture. Wintsch (1975b) has however suggested that the reaction



may also occur.

Sillimanite occurs as needles or irregularly shaped grains or grain aggregates within or protruding from quartz grains, indicating contemporaneous growth of these two minerals. Assuming that the aluminium comes from a reaction such as 11, the equation



may represent the mechanism by which a quartz-sillimanite intergrowth in the approximate volume ratio 1.5:1 could have formed. It is to be noted that quartz is present on both sides of the reaction, and that much of it therefore, is essentially recrystallizing.

Less commonly sillimanite occurs within crystals of muscovite, fig. 28. The sillimanite inclusions are not parallel to the cleavage in the muscovite, and the muscovite lattice is only slightly distorted by the sillimanite, suggesting that this too, is an intergrowth texture (c.f. Ashworth 1975). It is likely that the muscovite recrystallized from previously existing crystals (with the appropriate changes in minor element chemistry, see appendix 2), so that equation 14 may represent the reaction which gave rise to the

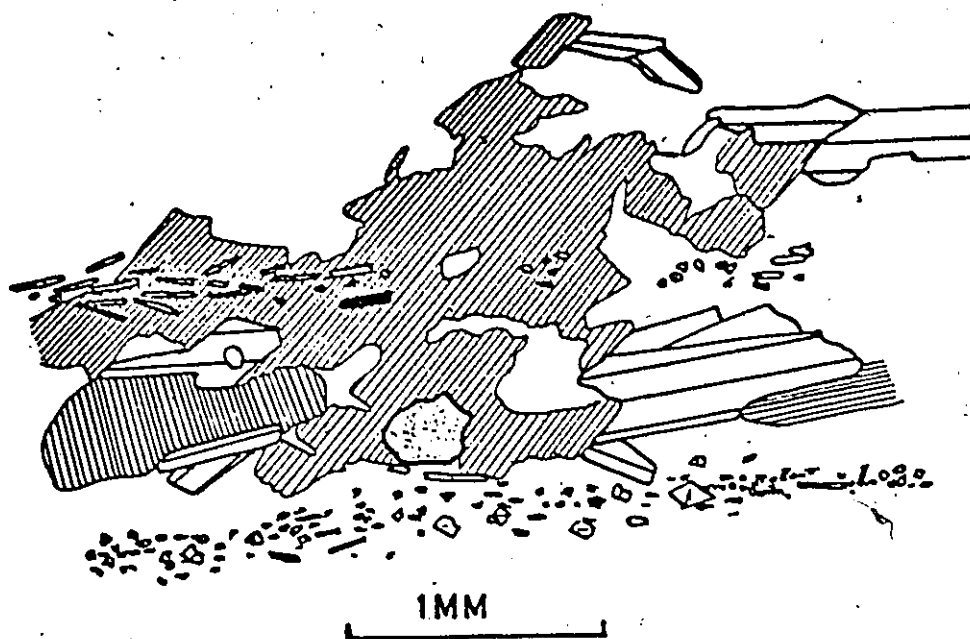
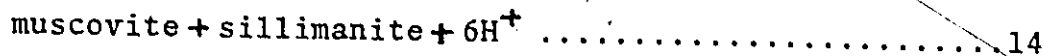
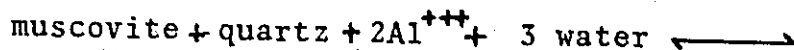


Fig. 28 Drawing of a texture in which a porphyroblast of muscovite (closely spaced cleavage) includes a train of sillimanite crystals parallel to  $S_1$ , a small garnet (heavy outline, stippled), and quartz (blank). Biotite crystals (widely spaced cleavage) lie parallel to  $S_1$ . Drawn from specimen 293A.

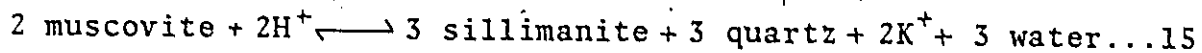
texture.



Since both 13 and 14 consume  $\text{Al}^{+++}$  when proceeding from left to right, they combine better with equation 11 than 12, and movement of aluminium can be limited to internal migration on the scale of the thin section. In both 13 and 14 the volume of the product assemblage is greater than that of the reactants. It is possible that the "excess" material can now be found in quartz and quartz-sillimanite veins which are common in the high grade rocks. By thus providing an endogenous origin for this material, the problems of long distance transport are circumvented (c.f. Yardley 1975).

Muscovite replacement by sillimanite without the concomitant production of potash feldspar is widespread in the sillimanite and sillimanite-K.feldspar zones. Sillimanite fibres and needles penetrate and distort the muscovite crystal lattice, the phyllosilicate layers being pried apart, and birefringence colours being lowered near the sillimanite. This texture is considered to be distinct from that in fig. 28 for the three reasons. In fig. 28, (a) the sillimanite inclusions are not parallel to the cleavage of the host mineral, (b) the muscovite lattice appears little distorted, and (c) the sillimanite is prismatic in form, all suggesting an intergrowth texture. In

this case, however, it appears that the sillimanite has been derived directly from the muscovite, so that a reaction such as that suggested by Wintsch (1975b) may represent the texture:



The products of this reaction occupy about 20% less volume than the reactants, so silica may be introduced to the site of the reaction as the products are formed, giving rise to the common quartz-sillimanite intergrowths in proportions typically seen in nature eg. quartz : sillimanite about 3:1.

#### Evidence of Contemporaneous

#### Recrystallization\*

Data suggesting that muscovite recrystallized during prograde metamorphism, based on crystal size distributions and crystal chemistry is given elsewhere (see Textural Variation with Metamorphic Grade and Appendix 2). It is likely that biotite followed a similar trend. On the basis of composition, recrystallization of plagioclase is known to take place with increasing grade of metamorphism (Turner 1968). There is also independent evidence in the Fernleigh belt of the recrystallization of staurolite and kyanite during prograde metamorphism. In the upper part of the staurolite-kyanite zone, crystals of staurolite and

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\* Recrystallization refers to the reconstitution of an existing phase, and may involve nucleation and growth processes, Spry 1969, p. 114.

kyanite attain considerable size (staurolite up to 2 cm. across, kyanite crystals occasionally being over 15 cm. in length, eg locality 89). Upgrade from this locality the two minerals persist locally for some 8 km., yet their crystal size progressively diminishes. In some cases the total volume of the mineral may not be less than at locality 89, but instead of one or two large porphyroblasts, many small crystals are present in the matrix. Whilst some of these may be physically separated remnants of a single large porphyroblast (in the case of kyanite in particular fragmentation of porphyroblasts appears to have been common), the optical orientation of others, as well as their wide distribution in the matrix of the rock, suggests that this is not always the case. For these, recrystallization from larger crystals appears likely, even though on the larger scale of the metamorphic belt these minerals were being destroyed and replaced by others. In minerals in which some solid solution is possible, recrystallization may have involved exchange reactions among coexisting phases eg. equations 3 and 4 etc. However for phases of essentially constant composition, recrystallization can be considered as isothermal.

Whether or not recrystallization of reactant phases is a general phenomenon during progressive metamorphism is not known. Atherton (1975), whilst noting that matrix minerals

recrystallize continuously throughout metamorphism, cites the chemical zonation common in porphyroblasts as evidence that the latter may remain as metastable remnants unless destroyed by penetrative deformation or high  $f \text{H}_2\text{O}$ . Parameters such as these might be expected to vary areally and temporally during metamorphism, so the degree of recrystallization of porphyroblastic phases may vary significantly within and between metamorphic terrains.

#### Discussion

All of the equations put forward to describe the reactions in rocks of the Fernleigh belt have been ionic equilibria, and almost all have involved water as a fluid or intergranular phase, indicating that  $f \text{H}_2\text{O}$  was an important variable during these reactions. The activities of elements in solution, particularly those of  $\text{K}^+$ ,  $\text{H}^+$ ,  $\text{Na}^+$  and  $\text{Al}^{+++}$ , as well as the transport of neutral complexes such as  $\text{Si}(\text{OH})_4$  are thus considered to have been crucial in determining in detail the mechanism of crystallization and recrystallization reactions. For the staurolite breakdown reactions, the effects of some of the postulated activities on staurolite stability are not well known; eg.  $f \text{H}_2\text{O}$ ,  $A_{\text{K}^+}$  and pH in equations 3, 4 and 5, and must await thermodynamic data. It is likely, though, that the effect of pH may be important. Eugster (1970) notes that the pH will drop when any hydrous phase crystallizes, unless there is an external source of  $\text{OH}^-$  ions.

Hydrous phases may therefore tend to be more stable in high pH environments. Kwak (1974) has compared staurolite breakdown textures from several metamorphic terrains, and he too concludes that reactions were ionic equilibria involving transport of some elements on at least the scale of the thin section. Whilst some of the textures described by Kwak are similar to those presented here, none are identical, suggesting that staurolite breakdown may follow a variety of different patterns. Indeed, the several possible reactions put forward in this work attest to the provinciality of subsystems during metamorphism. Staurolite stability is also governed to a limited extent by the oxidation ratio in the rock (Hounslow and Moore 1967); and as a consequence in rocks containing a variety of iron-bearing opaque oxides,  $a_{Fe^{++}}$  may be a critical factor. Furthermore Kwak (1974) notes that staurolite contains a significant proportion of zinc, so that  $a_{Zn^{++}}$  may also be a determinant factor. The stability of staurolite is therefore likely to be sensitive to the activities of a number of elements, and these activities may have exhibited considerable variation, both areally due to initial concentration differences, and temporally as a result of other ionic reactions taking place during progressive metamorphism. The result of these factors in combination is not known, but it is to be expected that their effects will not be simply additive. Simi-

larly the presence of other ions in the fluid phase, such as  $Ti^{+++}$ ,  $Fe^{+++}$ ,  $Mn^{++}$ ,  $Mg^{++}$ , or of any phase that is congruently dissolved cannot be easily predicted.

Ionic equilibria between muscovite, aluminium silicate and potash feldspar have been considered by Eugster (1970) and some of these are appropriate to the kyanite replacement textures in the Fernleigh belt. For instance Eugster shows that the reaction in which an aluminium silicate is replaced by muscovite, eg. equations 7 and 8, may be achieved by a drop in temperature, or an increase in  $f H_2O$ ,  $a_{K+}$  or pH, (fig. 29). Kwak (1971b) has calculated that the sequence of conversion of aluminium silicate to muscovite is kyanite, andalusite and finally sillimanite, and he considers  $a_{K+}$  to be crucial. The recrystallization of kyanite from large porphyroblasts to small matrix crystals in the Fernleigh belt may therefore be due to temporal variation in  $a_{K+}$ . The replacement of muscovite by sillimanite (equation 15) may also be treated in a similar manner, and can be achieved by a rise in temperature, or a decrease in  $f H_2O$ , pH or  $a_{K+}$  (fig. 29).

If, as suggested previously, variation of the  $a_{K+}/a_{H+}$  ratio is controlled in part by the deformation and recrystallization of mica (Wintsch 1975a), a form of chain reaction may be initiated with the original impetus derived from crystal deformation. During muscovite deformation,

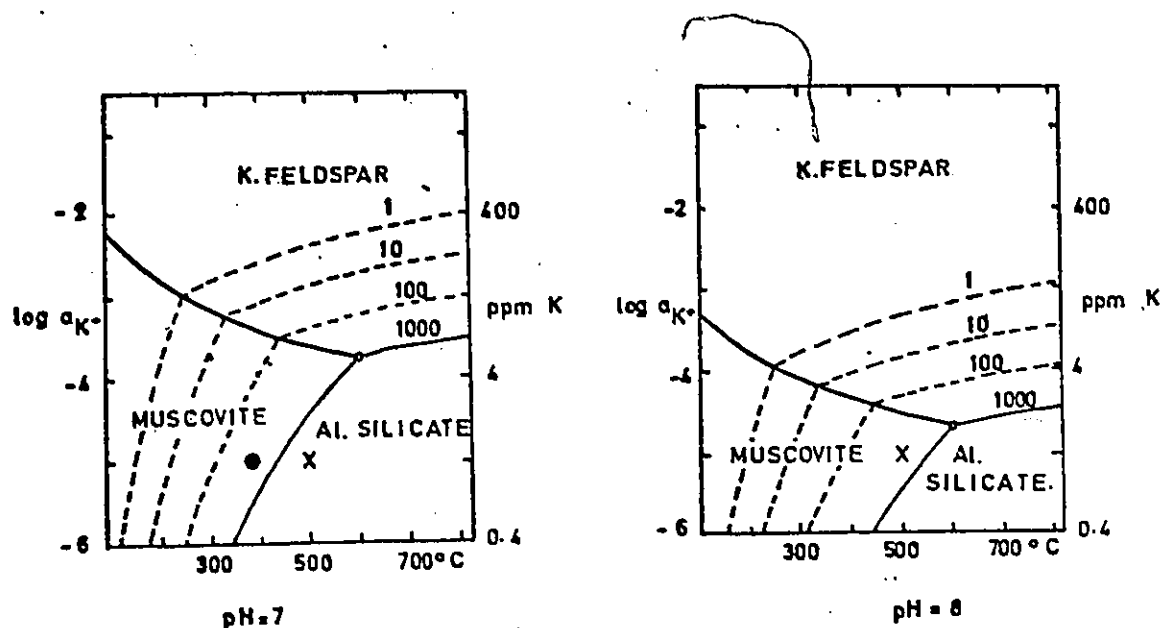


Fig. 29 (taken from Eugster 1970). Phase relations between muscovite, K. feldspar and aluminium silicate at pH of 7 and 8. Quartz and fluid are everywhere present. Solid lines are for  $f_{\text{H}_2\text{O}} = 1000$  bars;  $f_{\text{H}_2\text{O}} = 100, 10$  and  $1$  bar are represented by dashed lines. The potassium concentrations indicated on the right are approximate. The common replacement of muscovite by sillimanite (see text) may be achieved by a reduction in  $f_{\text{H}_2\text{O}}$  (• moves from the muscovite to the sillimanite field as  $f_{\text{H}_2\text{O}}$  drops from 1000 to 100 bars), or by a lowering of pH (X moves from muscovite to sillimanite fields as pH drops from 8 to 7). The reaction is also sensitive to temperature and  $a_{K^+}$ .

$a_{K^+}$  increases as new 001 ~~surfaces~~ of the crystal are exposed (an instantaneous surface exchange reaction in which  $K^+$  is replaced by  $H^+$ ), so that kyanite may become unstable and dissolve in the intergranular fluid, with the solid phase boundary moving into the muscovite field. However muscovite recrystallization, a rate controlled phenomenon, will gradually lower the  $a_{K^+}$ , so that kyanite may again become the equilibrium solid phase. By a mechanism such as this, recrystallization of both muscovite and kyanite is possible, without prior deformation of each kyanite crystal being a <sup>pre</sup> requisite. The location of the new nuclei will be controlled principally by local stress heterogeneities and the rates and flux of ionic transport. As previously mentioned, the recrystallization of muscovite, a hydrous phase, will cause a reduction in the pH of the intergranular fluid, as  $OH^-$  ions are bonded onto the crystal lattice. This change in composition of the fluid could equally be capable of altering mineral assemblages should the stability limits of solid phases have been reached, and may be pertinent to the replacement of muscovite by sillimanite below the potash feldspar isograd.

Likewise replacement of muscovite by potash feldspar can be achieved by both thermal and ionic reactions. Whether the occurrence of the four phase assemblage muscovite-quartz-sillimanite-K. feldspar represents a thermal equilibrium

over a localized "hot spot" is not known. The isolated nature of the occurrence, however, would suggest that a local ionic equilibrium is more likely, due to either increased  $a_{K^+}$  or pH or decreased  $f_{H_2O}$ .  $f_{H_2O}$  may have been high in the Fernleigh belt, permitting the stability of muscovite at the highest grades, whilst the comparative rarity of K. feldspar suggests that  $a_{K^+}$  was generally low.

### Conclusions

The fugacity of water and the activities of the various elements cited as potential variables in ionic reactions are not expected to vary systematically with grade of metamorphism, so such reactions should not strictly be considered "prograde" ie. due to increasing temperature and pressure. Intersecting isograds have been mapped utilizing similar principles (Carmichael 1970), indicating that these are rather iso-reactions lines than isograds. However the pattern of metamorphic zonation on the gross scale in the Fernleigh belt is such as to suggest that temperature and pressure were the dominant factors in their distribution. Therefore a combination of thermal and ionic equilibria appears realistic, in which the regional distribution of isograds is determined by the former, but localized mineral breakdown reactions are more usefully considered in terms of the latter. This conclusion is in accord with Carmichael (1969), Eugster (1970) and Kwak (1971a).

In the Fernleigh belt, the isograds defining the thermal stability of staurolite in the presence of quartz, and the aluminium silicate transformation kyanite  $\leftrightarrow$  sillimanite appear not to have been greatly affected by local activity gradients. However ionic breakdown of staurolite and kyanite may have resulted in the smearing out of the isograd which is apparent on detailed examination. The isolated occurrence of the pair K. feldspar and sillimanite may be due to a locally high activity of  $K^+$ , whilst the common replacement of muscovite by sillimanite, although not delineating an isograd, is considered to be due to a lowering of the  $pH$  or  $f. H_2O$  after the crystallization of muscovite.

## TIME RELATIONS BETWEEN DEFORMATION AND METAMORPHISM

### Introduction

Variations in the growth rates of matrix and porphyroblast phases, and in the relative rates of growth and strain may allow a variety of textures to develop simultaneously during syntectonic metamorphism. A re-examination of some of the usual assumptions in textural interpretation, especially those concerned with syntectonic mineral growth is therefore due.

### Assumptions in Textural Analysis

The textural criteria used for the separation of periods of mineral growth are well known, and have been reviewed at length by Spry (1969). The development of a foliation surface is considered to be syntectonic, though it may be enhanced by post-tectonic growth; and the timing of the growth of porphyroblasts is ascertained by reference to a foliation surface associated with a known deformation event. Wrapping of the foliation around a porphyroblast is considered evidence of pre-tectonic porphyroblast growth (fig. 30a), in contrast to growth post-tectonically, where no such deflection occurs (fig. 30b). The textures generally cited as indicative of syntectonic growth of porphyroblasts are relatively uncommon, despite the fact that many authors consider that deformation and mineral growth were essentially coeval in many metamorphic belts (eg. Spry 1969). The most frequently encountered

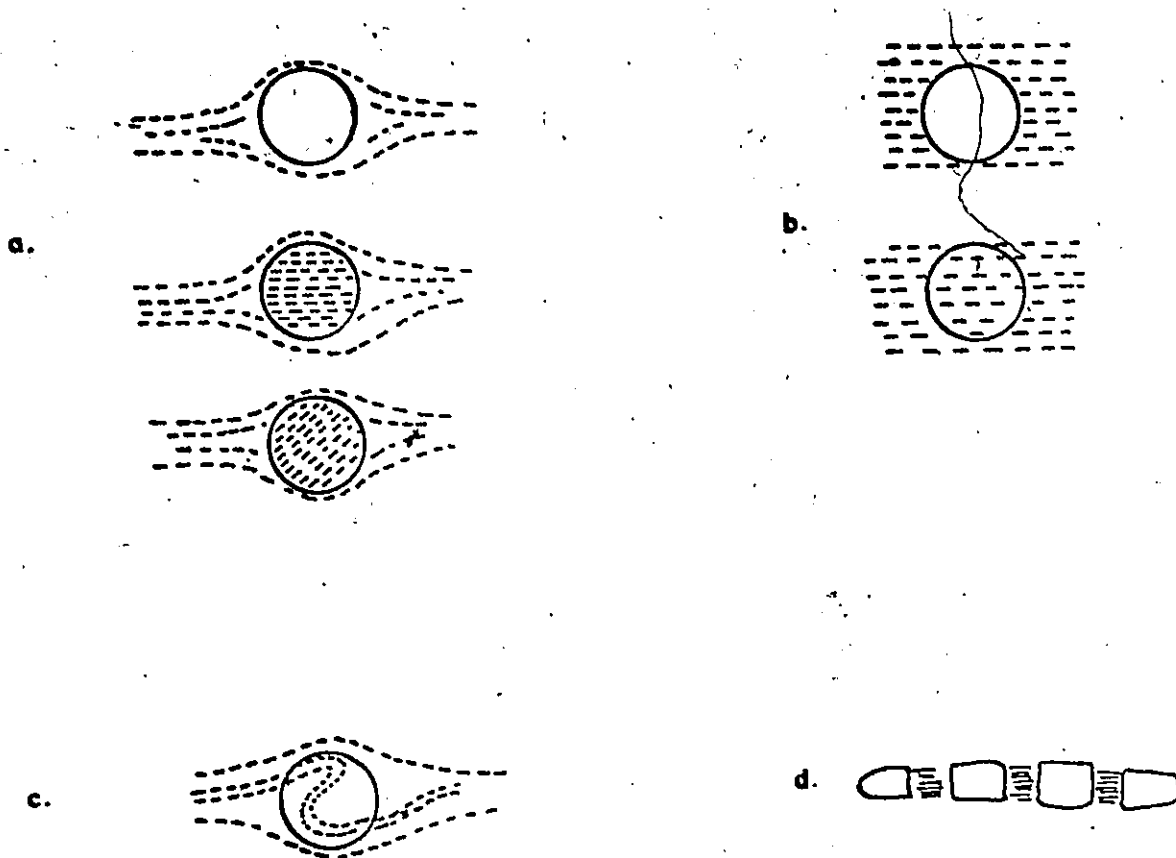


Fig. 30 Criteria for establishment of time relations between deformation and crystal growth. (a) Pre-tectonic growth of porphyroblasts, (b) post-tectonic growth of porphyroblasts, (c) syntectonic growth of porphyroblasts (snowball structure); and (d) syntectonic growth of porphyroblasts (paracrystalline microboudinage). In (d) a pre-tectonic crystal is pulled apart during deformation, and the spaces between the segments are filled by oriented syntectonic growth of fibres of another mineral. a - c after Spry (1969), d after Misch (1969).

syntectonic textures are the snowball (rotation or pinwheel) structure eg. Spry 1963, and paracrystalline microboudinage (Misch 1969) (figs. 30c and d).

#### Validity of Assumptions

The criteria for the timing of porphyroblast growth have recently been discussed by Misch (1971) and Ferguson and Harte (1975). Ferguson and Harte note that the absence of deformation effects in porphyroblasts is not definite evidence of post-tectonic growth. After detailed examination of a schist containing porphyroblasts of garnet and staurolite which possess an internal foliation ( $S_{int}$ ) and around which the matrix foliation ( $S_{ext}$ ) is typically truncated and/or slightly deflected, they conclude that truncation or deflection of  $S_{ext}$  by a porphyroblast, especially if it does not contain  $S_{int}$  (fig. 30b) is insufficient evidence for post or syntectonic growth. Furthermore they discuss cases in which two distinct deformation events produce foliation surfaces of similar orientation such that it is difficult to distinguish their effects on porphyroblast growth, and they note that post-tectonic annealing and grain growth can further hinder interpretation. Finally they observe that different rates of strain may also affect the texture; for instance different rates of rotation of porphyroblasts during growth, or of flattening of the foliation around porphyroblasts during growth may determine whether inclusion trails ( $S_{int}$ )

are straight or curved. Jones and Galwey (1962) note that variable rates of nucleation and growth of the porphyroblast phase may also be significant in this respect, and additionally that the nature of the porphyroblast-forming reaction may largely determine the presence, size and orientation and type of minerals included within porphyroblasts. It is apparent, therefore, that the current definition of syntectonic textures is too narrow; many crystals taken to be pre- or post-tectonic may in fact be syntectonic.

#### Textures and Deformation

In the Ompah area a variety of textures from different lithologies can be combined to indicate a relatively simple history of metamorphism and deformation, which two major deformation events,  $D_1$  and  $D_2$ , took place during a single extended period of metamorphism (fig. 32). Deformation was thus broadly syn-metamorphic, in common with most orogenic belts. In detail there is evidence to suggest that  $D_1$  started at a low grade of metamorphism, but continued till the metamorphic climax.  $D_2$  occurred at the metamorphic climax or soon afterwards.

During the study, in which over 200 thin sections were examined, it was apparent that some lithologies contained much more textural information than others, those with porphyroblastic phases being particularly useful. Thus the descriptions are limited to four rock types only; marble

(unit 5), pelitic schist (unit 4b), amphibole gneisses (unit 1c) and biotite schists (unit 6a), of which the latter two are considered together. Amongst them these lithologies show a variety of textures which are considered representative of the rocks in the area. The type of evidence and the assumed time relations between crystal growth, deformation phases and metamorphic grade are summarized in fig. 31.

#### Muscovite Schist

Although limited in areal extent, the muscovite schist (unit 6b) is particularly suitable for textural studies because many minerals within it are porphyroblastic.

Foliation Forming Minerals. Of the foliation forming minerals-muscovite is typically predominant. The crystals have a preferred orientation in the  $S_1$  foliation, and there is no evidence of their existence prior to  $D_1$ . Growth was thus syntectonic. Quite commonly porphyroblasts of plagioclase contain inclusions of muscovite ( $S_{int}$ ), which also exhibit a preferred orientation (plate 10). In some examples the inclusions are limited to the central portions of the porphyroblast, where they are of very fine grain size. The shape of the included crystals suggests that they grew at a low metamorphic grade, probably greenschist facies (fig. 43). The fine grain size may also be indicative of growth at low grade, whilst the much larger size of crystals in the external foliation ( $S_{ext}$ ) suggests that recrystallization (reduction

| DEFORMATION PHASE                   |                             | D1           | D2          | D3 |
|-------------------------------------|-----------------------------|--------------|-------------|----|
| METAMORPHIC GRADE                   |                             | GREENSCHIST  | AMPHIBOLITE | ?  |
| PELITIC SCHIST                      | MUSCOVITE                   | → ———— →<br> |             |    |
|                                     | QUARTZ, OPAQUES             | —————<br>    |             |    |
|                                     | BIOTITE                     | ? ———— →<br> |             |    |
|                                     | PLAGIOCLASE FELDSPAR        | ? ———— →<br> |             |    |
|                                     | STAUROLITE, KYANITE, GARNET |              |             |    |
|                                     | SILLIMANITE                 | —————<br>    |             |    |
| MARBLE                              | CALCITE                     | ? ———— →<br> |             |    |
|                                     | PHLOGOPITE                  | —————<br>    |             |    |
|                                     | TREMOLITE                   | —————<br>    |             |    |
| AMPHIBOLE GNEISS AND BIOTITE SCHIST | AMPHIBOLES                  | —————<br>    |             |    |
|                                     | BIOTITE                     | → ———— →<br> |             |    |
|                                     | GARNET                      |              |             |    |
|                                     | QUARTZ, OPAQUES             | ? ———— →<br> |             |    |

KEY: MATRIX

- fine grained || S1
- coarse grained || S1
- ———— → L1
- coarse grained || S1, folded & recrystallized.
- coarse grained S1, folded, crystals remain deformed.
- coarse grained || S2
- ———— equant
- ⊙ ———— equant, broken/bent or undulose extinction

PORPHYROBLAST

- no S<sub>int</sub>, S<sub>ext</sub> deflected around porphyroblast.
- S<sub>int</sub> | S<sub>ext</sub> S<sub>ext</sub> deflected around porphyroblast.
- S<sub>int</sub> | S<sub>ext</sub> S<sub>ext</sub> deflected around porphyroblast.
- broken, bent or with undulose extinction.
- overprinting of coarse grained S<sub>ext</sub> by porphyroblast.

Fig. 31 Schematic representation of time relations between crystal growth, metamorphic grade and deformation phases for dominant minerals in pelitic schist, marble and amphibole gneiss and biotite schist.

in numbers and increase in size of remaining crystals) has taken place there. Biotite crystals present in  $S_{ext}$  have similar textures to muscovite, although no fine grained crystals were seen included in porphyroblasts, so evidence of the crystallization of biotite at low grade is lacking. In the hinges of  $F_2$  folds, mica crystals generally follow the outline of the folded surface ( $S_1$ ) (fig. 11, plate 8) but in the majority of cases the micas are not themselves bent, indicating subsequent recrystallization. Micas are typically bent around small  $F_3$  crenulations however, with no subsequent annealing.

Other matrix minerals of the muscovite schist are quartz, feldspar and opaques. Some of these have a platy shape parallel to  $S_1$ , but equant grains, possibly indicative of recrystallization after deformation are more common. Fine grained, elongate crystals of quartz and opaques, as well as muscovite, define  $S_{int}$  in porphyroblasts of plagioclase feldspar, and so by analogy with muscovite, it is assumed that they started growth at a low metamorphic grade.

Porphyroblast Minerals. Some of the most conspicuous porphyroblasts which occur at all metamorphic grades are those of plagioclase feldspar. These may be up to 4 cm. in length, and they generally lie with their greatest and intermediate dimensions in the plane of  $S_1$ , the least dimension being perpendicular (that the porphyroblasts lie in  $S_1$  and not  $S_2$  has been determined where  $S_1$  and  $S_2$

are not subparallel eg. Clyde Forks region). Typically the porphyroblasts include a foliation ( $S_{int}$ ), whilst  $S_{ext}$  is variably deflected around them.  $S_{int}$  is generally parallel to  $S_{ext}$  away from the porphyroblast (plate 10), but in a minority of examples it is oblique, indicating that rotation between  $S_{int}$  and  $S_{ext}$  has occurred. In these latter cases the sense and amount of rotation is typically constant within a single specimen, though variable between specimens.  $S_{int}$  is thus considered to represent an early stage of  $S_1$ , prior to porphyroblast growth.

Not uncommonly the larger feldspar porphyroblasts have lattice deformities causing undulatory extinction, and in one section kinked crystals were observed. Elsewhere, however, evidence of late overgrowth of the coarse  $S_1$  foliation can be seen (plate 10). It is not easy to relate these phenomena directly to particular phases of deformation and mineral growth, and they are grouped here as post  $D_1$  features.

Porphyroblasts of kyanite, staurolite and biotite occur throughout the staurolite-kyanite zone, where they generally exhibit similar textural relations. Typically  $S_1$  is deflected around the porphyroblasts and, as in feldspars, some contain inclusion trails ( $S_{int}$ ). Because  $S_{int}$  is most commonly parallel to the  $S_1$  foliation outside the porphyroblast, it is taken to be  $S_1$  included near the end of  $D_1$  deformation. Porphyroblasts of kyanite and staurolite may be inclusion-

zoned i.e. they contain inclusions in the centre not at the periphery (plates 10 and 11), whilst in other specimens porphyroblasts of kyanite, staurolite and biotite are poikiloblastic, and overprint the  $S_1$  foliation. The larger kyanite porphyroblasts are orientated with their long axes in the foliation (though they are random in that plane), but as with staurolite porphyroblasts, smaller kyanite porphyroblasts may lie athwart the foliation. Porphyroblasts of biotite, though having a random crystallographic orientation, generally exhibit a morphological preferred orientation with their greatest dimension lying in the plane of  $S_1$ . As with feldspar, there is evidence of deformation of some porphyroblasts, especially those of biotite and kyanite; in plate 10 a porphyroblast of kyanite has been bent through about  $20^\circ$ , but subsequent annealing has restored optical continuity throughout the crystal.

These textures are taken to indicate growth during  $D_1$ , commencing after the early formation of  $S_1$  at low metamorphic grade. The end of growth is difficult to fix with regard to later deformation events, but the overprinting of a coarse foliation suggests that growth continued after  $D_1$ , and possibly during  $D_2$  in some cases (recrystallization of muscovite around fold hinges indicates that the temperature was still high at that time).

Garnet occurs throughout the muscovite schist (unit 6b).

where the bulk composition is appropriate, and suprisingly, considering the number of reactions in which it may be involved (see Metamorphism), the textural relations remain much the same throughout. Though euhedral outlines may be more common at low metamorphic<sup>grade</sup>, the textures are generally similar to the other porphyroblast phases described above, with  $S_{ext}$  deflected around the porphyroblast, and  $S_{int}$  generally parallel to  $S_{ext}$ . Where the two are not parallel (plate 12) visual estimates suggest that the angle between  $S_{int}$  and  $S_{ext}$  is greater in the low grade rocks. This may be due to the near spherical shape of euhedral porphyroblasts at low grades facilitating rotation (c.f. Ferguson and Harte 1975), whilst at higher grades porphyroblasts are typically inequant, lying in  $S_1$ . In some high grade specimens  $S_{int}$  is coarser at the edges of the porphyroblast than in the centre, which suggests overprinting of a coarse foliation. Thus garnet growth is considered to have started during  $D_1$ , and to have outlasted that event, but chronology relative to later deformation events is not firmly established.

Sillimanite occurs in flattened and elongate pods with or without quartz (plate 8, see also Divi 1972), around which the  $S_1$  foliation is deflected. Elsewhere it occurs as a matrix mineral in the foliation itself, and it is not uncommon to see sillimanite in the foliation deflected around a pod of the same mineral. Sillimanite intergrowths with garnet,

muscovite, tourmaline and opaques are also seen, and indicate contemporaneous crystallization of these phases. Sillimanite growth is therefore considered to have started during  $D_1$ , but in common with other minerals, to have outlasted that event.

Porphyroblasts of muscovite occur in the sillimanite and sillimanite-potash feldspar zones. These may include the  $S_1$  foliation, in places defined by trains of sillimanite needles (fig. 28). As with biotite, the porphyroblasts exhibit a morphological preferred orientation parallel to  $S_1$ . These textures suggest growth during  $D_1$  deformation.

#### Marble

In some specimens of marble, particularly those containing minerals such as phlogopite, tremolite and quartz, the effects of  $D_1$  are well developed. The  $S_1$  foliation is defined by the morphological preferred orientation of calcite crystals and by the crystallographic preferred orientation of phlogopite and tremolite, and  $L_1$  by the long axes of tremolite crystals. It is therefore assumed that these minerals grew during  $D_1$  deformation. The imprint of  $D_2$ , which is generally less evident, may also be seen in some thinly layered lithologies where it gives rise to grain elongation of calcite and preferred orientation of tremolite parallel to  $S_2$  (fig. 13). In other examples tremolite crystals lying in  $S_1$  were bent during  $D_2$ , and were not later recrystallized. Other hand-

specimens of small-scale  $F_2$  folds display complex textures in thin section, in which neither  $S_1$  nor  $S_2$  are evident. Calcite crystals have a range of grain size (generally  $< 2$  mm.), and the variously shaped crystals have serrated and indented boundaries, with multiple sets of twin lamellae frequently being developed in a single crystal. Calc-silicate phases are generally fine grained, and exhibit local preferred orientation patterns which are neither parallel to  $S_1$  nor the axial surface of  $F_2$  folds as seen in hand specimen. These are evidently disequilibrium textures, and are considered to be the result of  $D_2$  or later strain superimposed on an  $S_1$  fabric. In outcrops of marble which are massive or possess a widely spaced layering, calcite crystals are typically equant, and calc-silicate phases are dispersed throughout the rock and randomly orientated. In these more homogeneous lithologies, evidence of structural elements are hard to find, and so the timing of crystal growth cannot be related to deformation.

Evidence to support growth of calcite and calc-silicate phases during  $D_1$ , and in places during  $D_2$  is therefore present. Timing of mineral growth with respect to  $D_3$  has not been established.

#### Amphibole Gneisses and Biotite Schists

Foliation Forming Minerals. In amphibole gneisses (unit 1c) and biotite schists (unit 6a) the foliation,  $S_1$ , and the lineation,  $L_1$ , are defined by the preferred orientation of

biotite and amphibole crystals, indicating that growth of these minerals was essentially during  $D_1$ . Feldspar, quartz and calcite also crystallized at this time, as deduced from their inequant shape parallel to  $S_1$  in some sections. These minerals were deformed <sup>during</sup>  $D_2$ , as in a few examples bent and fractured crystals lying parallel to  $S_1$  were seen. However, in the majority of cases undeformed crystals lie parallel to  $S_1$ , suggesting recrystallization during or after  $D_2$ . In fine grained mylonitic rocks an axial planar foliation ( $S_2$ ) defined by biotite crystals was developed (plate 5).

Porphyroblast Minerals. Some amphiboles and biotite occur as porphyroblasts in amphibole gneisses and biotite schists. Hornblende porphyroblasts are typically almost equant, randomly orientated and poikiloblastic, with very little deflection of the foliation around the crystal; whilst those of actinolite, present in mafic pods and lenses of unit 1c within marble, occur as radiating, apparently undeformed, aggregates. Growth of porphyroblasts of hornblende is therefore considered to have taken place during and after  $D_1$ , whilst those of actinolite may be all post  $D_1$ .

Biotite porphyroblasts generally exhibit a morphological preferred orientation (plate 9), in the same manner as in pelitic schist (unit 6b). They are typically slightly poikiloblastic, with the external foliation ( $S_1$ ) deflected around the porphyroblast, indicating, as in unit 6b, growth during

and after  $D_1$ . Porphyroblasts of diopside are common in some of the more calcareous lithologies. Equant

poikiloblastic crystals, around which the foliation is slightly deflected, probably grew during and after  $D_1$ , in common with those of hornblende and biotite.

Porphyroblasts of garnet are common in these lithologies, and they display a variety of textures. Some are zoned with respect to the presence or absence of inclusions; generally the central portions contain inclusions, whilst the peripheries are inclusion-free. In other examples, garnet porphyroblasts include the  $S_1$  foliation throughout the crystal, the outer inclusions having a larger grain size, suggesting late overprinting of a coarse foliation. The shape of the porphyroblasts is frequently inequant, and in extreme examples the garnets define the  $L_1$  lineation (see p.44 ). The degree of deflection of the foliation around the porphyroblasts is highly variable between specimens, though approximately constant within a single specimen. This variation does not appear to be related to grade of metamorphism, but may be dependent upon lithology, as a trend was observed for the deflection to be more pronounced in schistose lithologies containing a high proportion of biotite and quartz. The included  $S_1$  fabric is taken to indicate that garnet porphyroblasts grew during  $D_1$ , and the overprinting of a coarse foliation suggests late  $D_1$  or post  $D_1$  growth, but the timing with respect to  $D_2$  and later

deformations has not been established with certainty. The inclusion zoning in some porphyroblasts may indicate that the garnet-forming reaction changed during prograde metamorphism.

#### Deformation and Metamorphic Grade

From the minerals and textures described and summarized in fig. 31, it is possible to infer time relations between deformation and metamorphic grade (fig. 32). During  $D_1$  deformation,  $S_1$  foliation started forming at low metamorphic grade in pelitic schist (unit 6b), and though evidence is lacking, it is assumed that a similar fabric developed in the other rocks. As the metamorphic grade increased variably across the area, the  $S_1$  foliation coarsened by recrystallization of matrix minerals, and frequently became included in porphyroblasts of variety of mineral species. Thus the highest grade "index" mineral in any specimen overprints the foliation. Because straight segments of the  $S_1$  foliation are included by most porphyroblasts, either  $S_1$  was not rotated during growth of the porphyroblasts or growth was fast relative to rotation. Later during  $D_1$ , flattening of the foliation around the porphyroblasts occurred; at the same time some porphyroblasts suffered rotation relative to the foliation, and others were deformed into ellipsoidal shapes.

$F_2$  folding apparently occurred at or slightly after the metamorphic climax, but because superposed  $F_2$  folds lack a penetrative  $S_2$  foliation, the metamorphic grade at this time is not precisely known. Minerals defining the  $S_1$  foliation

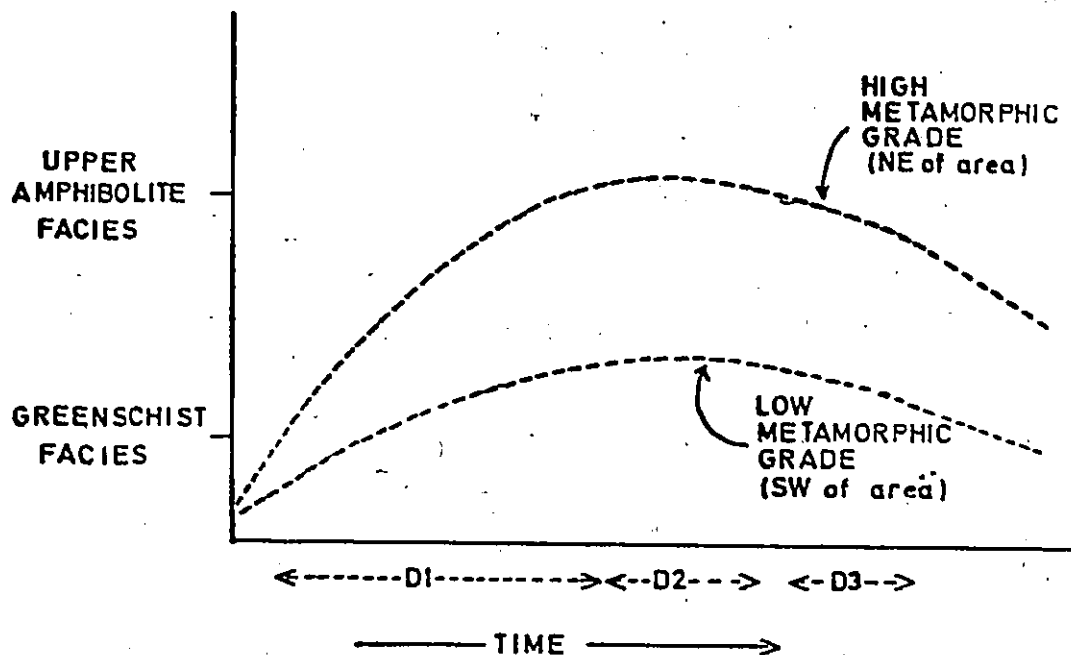


Fig. 32' Schematic diagram relating the timing of deformation to grade of metamorphism.

were bent and broken, with the subsequent recrystallization of some, but not all crystals during and after  $F_2$  folding. Syntectonic overprinting of the  $S_1$  foliation by some porphyroblasts eg. plagioclase feldspar, staurolite, garnet, sillimanite etc. in pelitic schist, garnet, biotite and amphibole in amphibole gneisses and biotite schists, may have occurred but in the absence of a penetrative  $S_2$  foliation, possible fabrics are indistinguishable from those considered to indicate syn-or-post  $D_1$  growth. Marble was ductile during  $D_2$  (see Structural Geology), so the typical lack of a foliation defined by inequant minerals may be taken as evidence of recrystallization after  $D_2$  strain. The experimental evidence of Griggs, Paterson, Heard and Turner (1960) indicates that annealing recrystallization of calcite is strongly dependent upon the amount of strain induced in the rock before annealing; for instance annealing recrystallization of marble strained 20% at room temperature begins at  $500^\circ\text{C}$ , whilst 40% strain at room temperature results in major recrystallization at the same  $500^\circ\text{C}$  temperature. In the Ompah area temperatures of at least  $500^\circ\text{C}$  are indicated by the mineral assemblages, and local areas of high strain exceeding 40% have been measured eg. fig. 15; thus extensive annealing of marble is likely to have taken place.

The grade of metamorphism during  $D_3$  is not known precisely. Muscovite crystals in pelitic schist bent by  $F_3$  folds are not

annealed, suggesting that temperatures were lower than during  $D_2$ . The thickening of the hinge area of some  $F_3$  folds eg. fig. 14, indicates that quartz was mobile during  $D_3$ , but this is possible at low temperatures in a hydrous environment (Morey 1957).

The time relations between deformation and metamorphism are summarized in fig. 32. From the textural evidence it is not possible to tell whether there were significant time gaps between  $D_1$ ,  $D_2$  and  $D_3$ , or whether the three deformations were essentially part of a single process, as have been suggested for deformation sequences in other areas, eg. Fyson 1970.

#### The Timing of Plutonism with Respect to Deformation

In the northern part of the area the boundary between the Abinger granite gneiss and the metasedimentary and meta-volcanic rocks is migmatitic. The migmatites possess a penetrative foliation similar to  $S_1$  elsewhere in the area, so it is probable that intrusion predated the end of the first deformation. Within the main body of the Abinger granite gneiss, though, a foliation is only weakly developed, hence relationship are not completely clear.

The Cross Lake gneiss possesses a strong linear fabric correlated with  $L_1$  elsewhere in the area, thus, as for the Abinger body, intrusion predated at least the end of  $D_1$  deformation.

Two gabbroic stocks, possibly connected at depth to the

large mass 5 kms. east of Clyde Forks (fig. 2) occupy axial region of the  $F_2$  Plevna synform and another  $F_2$  synform to the northeast (fig. 3). Contacts with the surrounding rocks are brecciated in places, and a poorly developed foliation and lineation are present in the central portions of the bodies. Whether or not these features correlate with  $D_1$  or  $D_2$  structures in the surrounding rocks is not immediately evident. However the presence of a lineation may be indicative of  $D_1$  deformation, as  $L_1$  is well developed throughout the area. If so, emplacement would be before the end of  $D_1$  deformation, and in any case the fabrics certainly indicate emplacement before the end of  $D_2$  deformation. The presence of the stocks may have influenced fold development during  $D_2$ , hence their locations in the hinge areas of  $F_2$  folds.

The larger pegmatite intrusions in the Ompah area are all post-tectonic (ie. post  $D_2$ -their relationship to  $D_3$  has not been established), although many smaller, locally derived granitic veins were probably formed during deformation and high grade metamorphism.

#### Comparison with Other Areas

##### Deformation and Metamorphic Grade

Within the Grenville Province of eastern Ontario, information concerning time relations between deformation and metamorphism is available in several areas. It is to be noted, however, that when deformational events in different areas are

compared, no inference is made that these were synchronous. As remarked in the introduction, Lumbers (1967b) and Moore and Thompson (1972) consider that there were two metamorphic culminations in the Bancroft-Madoc region. Textures in the Ompah area provide evidence of a single prograde metamorphic event, which is correlated to the later ( $M_2$ ) metamorphism of Lumbers. However if the early metamorphism ( $M_1$ ) attained only greenschist facies, as suggested by Lumbers, its imprint may well have been lost during the subsequent, higher grade event. The detail of the time relations between deformation and metamorphism in the following paragraphs are all concerned with metamorphism correlated with the later ( $M_2$ ) episode.

In the Flinton area, Thompson (1972) considers that metamorphic grade during  $D_1$  was variable, and that  $D_2$  took place at the metamorphic culmination; the time relations are thus the same as those inferred for the Ompah area.  $D_3$  around Flinton, although of different orientation and attitude to  $D_3$  in the Ompah area, is also considered to be a retrograde (greenschist) feature. Carmichael (1968) and Divi (1972), referring to areas D and C (fig. 2), both consider that  $D_1$  was entirely a low-grade phenomenon, with the metamorphic grade rising between  $D_1$  and  $D_2$ , with  $D_2$  being synchronous with the metamorphic culmination.  $D_3$  is considered to be a retrograde feature by both these authors. Thus the time relations between deformation and metamorphism, although margi-

nally different to those in the Ompah and Flinton areas to the east, depict a similar overall pattern, which may be characteristic of this region of the Grenville Province.

Outside the Grenville Province, the pattern of low to medium grade metamorphism during recumbent folding, followed by upright folding at the metamorphic culmination may be quite common. Johnson (1963) reports such a pattern from the Caledonides in Scotland, and Zwart (1960) from the Hercynian of western Europe.

In the section on structural geology it was suggested that early recumbent folds may be characteristic of the infrastructure of a tectonic pile, with later upright folding predominating at higher levels. Assuming this model is correct, and considering that  $D_1$  was initiated at low metamorphic grade, it would appear that there is no direct correlation between metamorphic grade and structural level, as the metamorphic front must have been imposed on the rocks after they had descended to the infrastructure. In fact, Fyson (1971) notes that several other tectono-metamorphic sequences have also been recorded, so it is unlikely that there is a direct link between style of folding and grade of metamorphism.

#### Deformation and Plutonism

Lumbers (1967b) considers that most of the plutonic rocks in the Bancroft-Madoc region were emplaced before the culmi-

nation of regional metamorphism and deformation, although he observes that some potassic granites may be syn- and post-deformational features (table 1). In area C (fig. 2) (Divi 1972), the earliest intrusive recognized is of gabbroic composition, emplacement being pre- $D_2$ , which is in accord with the possible time relations of gabbro bodies in the Ompah area. Granite plutons were intruded syn- and post  $D_2$  deformation. Granite emplacement in area D (fig. 2) was after  $F_1$  and before  $F_2$  folding (Carmichael 1968). More complex histories have been recorded in the basement gneisses 50 km. northeast and 40 km. west of Bancroft (Appleyard 1974, Francoeur 1975). Francoeur, for instance, recognizes five distinct granitic phases whose age decreases southeastwards as the cover rocks are approached.

The sequence of plutonism and regional metamorphism in the Ompah area is thus in accord with the general scheme of Lumbers (1967b), (table 1), but details are at variance with those from other areas. This is to be expected: in a region where basement cover relations may have been an important factor during deformation, and plutonism took place in several stages over an extended period of time.

## TEXTURAL VARIATION WITH METAMORPHIC GRADE

### Introduction

Apart from the use of textures for determining time relations between deformation and metamorphism (see previous section), it has long been recognized that textural studies can yield much information about the deformation and metamorphic processes themselves. Recently there has been a trend towards basing descriptions on statistical analyses of some of the more easily measured textural parameters; this approach, as well as reducing subjective bias and facilitating comparison with other results similarly obtained, renders a wider range of data available, and may allow discrimination between several possible theoretical models.

The texture of a crystalline rock can be characterized by the inter-relationship of four variables- (a) size, (b) shape, (c) orientation and (d) spatial distribution of the constituent mineral grains. Variation of any or all of these parameters with metamorphic grade may provide information on the direction of textural changes during prograde metamorphism. Such changes, whilst taking place at lower energy thresholds than chemical reactions in which a new mineral species is crystallized, may nevertheless be significant in determining the course of metamorphic reactions because of their influence on reaction kinetics.

Quantitative studies of crystal textures by DeVore

(1956, 1959), Galwey and Jones (1963, 1966), Jones and Galwey (1964, 1966) and Kretz (1966a, 1966b) have yielded much information on metamorphic processes. However only recently has an attempt been made to relate such studies to a metamorphic gradient. Ehrlich et al. (1972) examined the shape of feldspar crystals from a plutonic body which exhibits a metamorphic gradient across it (the Cross Lake gneiss, fig. 2); and Jones and Galwey (1972) measured the shape of biotite crystals from four Barrovian zones in Glen Clova, Scotland.

In this work, variation in the size and shape of muscovite crystals with metamorphic grade is analysed, and the results are examined with regard to the equilibrium shape hypothesis. The data are also used to test certain nucleation and growth models suggested by Kretz (1974). Analyses of measurements of crystal orientation were also performed. Orientation measurements, when linked to the size and shape of individual crystals yield new information about the process of cleavage or foliation formation in medium grade rocks.

In order that rocks from a wide range of metamorphic grade could be sampled, the study area was extended to the southwest to the village of Bishop Corners (fig. 19); thus from low to high the metamorphic grade varies from the chloritoid-staurolite zone, through the staurolite-kyanite zone and the sillimanite zone to the sillimanite-potash feldspar zone. The isograds are at high angles to the litho-

logic boundaries (Moore 1967), and thus it is possible to follow the same unit through all the metamorphic zones.

#### Procedure

Sampling. Oriented samples were collected along the length of the Fernleigh belt from the pelitic schist unit (6b). The sampling interval varied between 1 and 4 km. (see fig. 19). Biotite sample localities for two specimens from the Cross Lake gneiss (unit 7b) are also indicated in fig. 19. A third specimen is from south-west Quebec, near Otter Lake, some 110 km. to the north-east, where the metamorphic grade has reached the sillimanite-K. feldspar zone (R. Kretz, personal communication).

Measurement. The maximum length (L) and thickness (T) of the crystals were measured in thin section; for the muscovite crystals the following procedure was set up in an attempt to standardise the measurements.

(a) Over the greater part of the Fernleigh belt, the  $L_1$  mica edge lineation on the  $S_1$  foliation surface is horizontal.

Vertical sections were cut perpendicular to this lineation (fig. 33). When information from the third dimension was

required, measurements were made on horizontal sections cut parallel to  $L_1$ .

(b) All measurements were made with a universal stage, allowing rotation of the (001) face of the crystal parallel to the microscope tube. The thickness recorded is therefore the

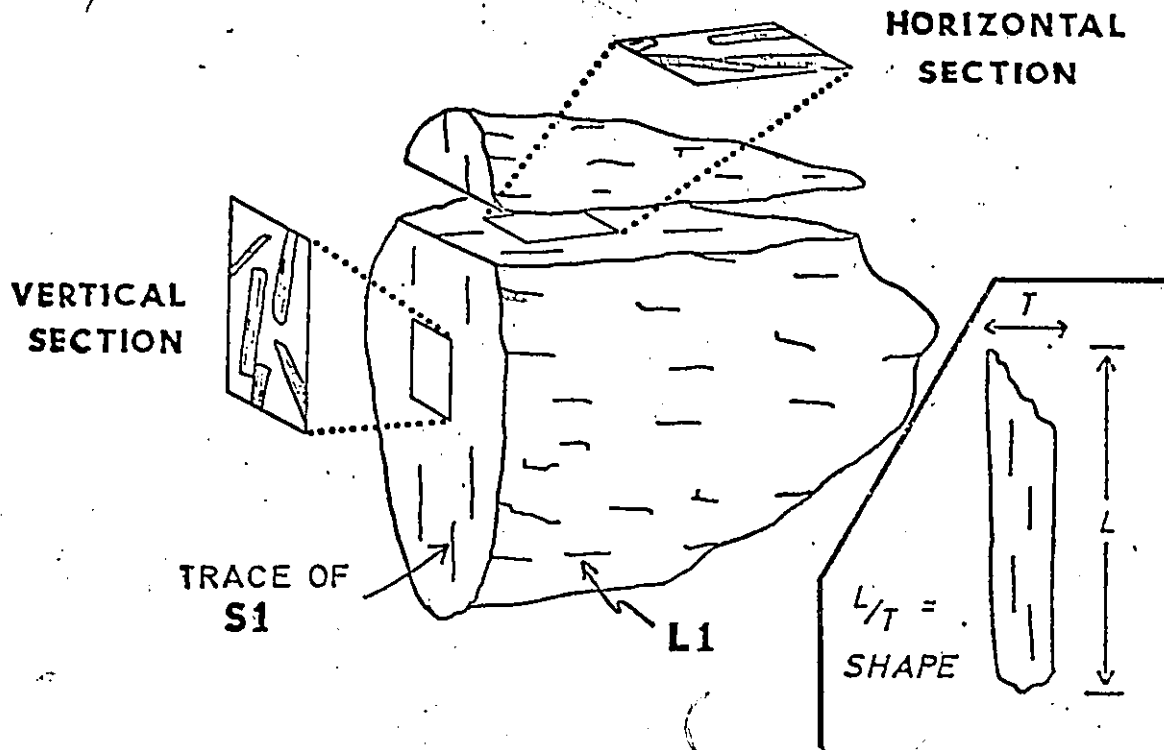


Fig. 33 Sectioning and measurement procedure in textural analysis. Vertical and horizontal sections were cut perpendicular and parallel to the  $L_1$  lineation. The shape ratio was calculated from the maximum measured dimensions of the muscovite crystals.

true c-axis dimension of the crystal. The length, however, is not a rational crystallographic plane, but an arbitrary (hk0) section through the crystal; it may be less than the true length by a variable amount, since the crystal may not have been sectioned through its greatest diameter. However since this "cut effect" is considered to affect all populations of measurements equally, a quantitative allowance was deemed unnecessary.

(c) Straight-line traverses were made across thin sections. Only crystals with distinct boundaries were measured. The U-stage greatly aids resolution, but frequently where two or more mica crystals, with their cleavages parallel, are in contact, it was difficult to separate them, and no measurements were made. With this proviso, which in effect restricts measurements to the quartz-mica domains, all muscovite crystals encountered on traverses were measured. The spacing between traverses was dependent upon crystal size, being arranged so that no crystal was measured more than once. In both the traverse method and in the sectioning of the rock, another form of bias is introduced, namely that the larger crystals are more likely to be traversed or cut than the smaller ones. However, the effect of this is thought to be small, especially where there is only a small range of grain size in the specimen, so no allowance was made.

(d) Within the pelitic schist unit there is some variation

in composition, mainly expressed by the proportions of the minerals present. This variable was minimized as a factor affecting the results by using samples of similar mineralogy. Approximate mineral proportions are given below:- muscovite 40-60%; quartz 40-60%; magnetite/haematite/ilmenite 2-8%; plagioclase 3-10%; biotite 5-20%; tourmaline less than 1%; kyanite, sillimanite, staurolite, garnet and chloritoid 0-10%.

(e) Minimum sample size was obtained by sample splitting of a population of 200 measurements. The results suggest that 100 measurements is sufficient to roughly characterise the population (table 5).

Sample variation at a single locality must be known before comparisons between localities can be made. Table 5 shows the measured and calculated parameters of four specimens from the same locality (sillimanite grade). Variation between specimens renders a single sample of 100 measurements inadequate to properly characterize the locality, but it is felt that combined results of the 500 measurements provide a good estimate.

A method of incorporating local variation without at the same time involving 500 measurements at each locality was therefore sought. A suitable method is a moving average technique, in which the results of measurements from adjacent localities are compiled into a single estimate. The measurements from each locality are used more than once, as the

| Specimen no.                         | NN  | Mean T<br>(mm) | Mean L<br>(mm) | Mean L/T | Mean LxT<br>(mm <sup>2</sup> ) |
|--------------------------------------|-----|----------------|----------------|----------|--------------------------------|
| 143A1                                | 100 | 0.15           | 0.73           | 5.89     | 0.126                          |
| 143A2                                | 100 | 0.26           | 0.93           | 5.09     | 0.347                          |
| 143A1, 143A2                         | 200 | 0.20           | 0.83           | 5.49     | 0.237                          |
| 143C                                 | 100 | 0.17           | 0.81           | 5.79     | 0.162                          |
| 143F                                 | 100 | 0.20           | 1.22           | 7.58     | 0.303                          |
| 143E                                 | 100 | 0.18           | 0.78           | 5.09     | 0.180                          |
| 143A1, 143A2,<br>143C, 143F,<br>143E | 500 | 0.19           | 0.90           | 5.89     | 0.224                          |

Table 5. Determination of sample size by sample splitting, and the assessment of sample variability at a single locality, (location 143, sillimanite grade). The first two rows show the results of splitting a sample of 200 measurements; in the third these have been combined. The remaining three rows are a comparison of muscovite statistics from four different rocks at the same specimen locality. The combined means for all 500 measurements are given in the last row.

central point of the estimate moves along the belt, fig. 34. In this work 47 small samples of 100 measurements each were compiled into 13 moving average estimates of 500 measurements.

(f) Horizontal sections. Six sets of 100 measurements each were made on horizontal sections. After this, since a consistent relationship with measurements from the vertical sections was obtained (dimensions in horizontal sections being about 10% greater than those in vertical sections), a further two samples were calculated on a proportional basis using mean data from the corresponding vertical sections.

Due to the low sample density, measurements from the horizontal sections cannot be incorporated into moving average populations. However, their similarity to measurements from the vertical sections suggest that they represent an acceptable approximation of the population at that locality.

(g) Biotite. Two samples of 100 measurements each were made on biotite from the Cross Lake gneiss. The measurements are intended primarily for comparative use, and the sampling and sample size procedure were not rigorously applied. The greater homogeneity of the gneiss, however, suggests, that the effects of compositional variation would be less than in the pelitic schist. These results will be considered separately, after those for muscovite.

(h) Crystal orientation. Six samples of similar composition, but varying metamorphic grade, were selected for analysis.

# MOVING AVERAGE TECHNIQUE

A,B,C- adjacent specimen localities progressing up grade.

1,2. - sample numbers.

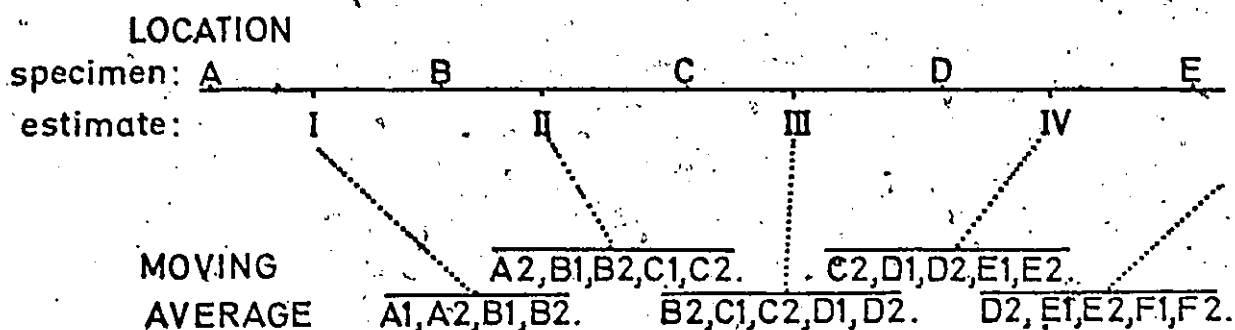


Fig. 34 Moving average method of sample combination. Each moving average estimate of 500 measurements (400 at each end of the belt) incorporates samples of 100 measurements from several adjacent localities. In cases where only one of the two samples was required from a single locality an arbitrary selection was made.

The specimens were sectioned as previously described (fig. 33). Vertical sections only were used. The orientation of (001) muscovite was measured, utilising two rotation axes of the universal stage. 200 measurements were made on each sample, and traverses were regularly spaced across the thin sections, all crystals traversed being measured. Readings were plotted and contoured by computer.

For evaluation of the effect of crystal size and shape on preferred orientation, two sets of 200 measurements were made on specimens from different metamorphic grades (one from the staurolite-kyanite zone, the other from the sillimanite zone). Measurements were again confined to the quartz-mica domains. The readings were subdivided into three groups on the basis of size (for both the T and L parameters), and for shape (L/T ratio); each group of readings was then plotted on a separate stereonet.

Analysis. A standard statistical approach, much of which is already in use in palaeontology (Imbrie 1956), was employed for analysis of the data. In both the samples and the populations, all measurements were processed by computer in the following manner. After conversion of T and L into millimetres, the shape ratio (L/T) and the area of the cut face of the crystal (LxT) were calculated. The mean, standard error of the mean, standard deviation and range of T, L, L/T, and LxT were then compiled. Following this, a corre-

lation table was calculated and each correlation tested for a significant difference from zero. If a correlation was established, the following statistical parameters were computed:-correlation coefficient, significance of correlation coefficient, gradient of the reduced major axis, constant of the reduced major axis, error of slope of the reduced major axis and the relative and absolute dispersion about the reduced major axis. Subsequently graphs were plotted of each variable against each other (six graphs in all), and histograms compiled for all four variables. This information is sufficient both to describe a single population and to demonstrate differences between populations.

Estimates of crystal volume were made by combining measurements from horizontal sections with those from vertical sections. Calculations were based on rectangular shapes, and mean sample values were used. The rectangular shapes, whilst not a realistic analogue, produces no bias towards any particular size or shape group and is easily calculated from the measured parameters. The use of mean values is considered justified if the measured parameters have unimodal distributions. Surface area calculations were also based on similar assumptions.

For evaluation of the degree of preferred orientation, contouring was done by the usual Schmidt equal area method, and also by a statistical method suggested by Kamb (1959). In this latter technique in which the area of the counter is

not constant, but varies according to the number of readings, the distribution is compared statistically to that of a similar number of randomly distributed points; the contours in standard deviations therefore estimate the degree of deviation from a random distribution. The contoured results on an equal area net projection are more useful than the contours of the Schmidt method in that the area enclosed by a contour, as well as its concentration, is taken into account. Thus when sample size is the same, a direct numerical comparison of the maxima, in terms of standard deviations, gives a valid comparison of the distributions. The data may then be plotted graphically to facilitate comparison between samples.

Where sample sizes are not the same (eg. in the evaluation of the effect of crystal size and shape on preferred orientation), a qualitative estimate of the degree of preferred orientation is made by measuring the spread of points about the modal value (contoured by the Schmidt method) for each group.

#### Crystal Size and Shape

##### Individual Samples-Muscovite.

The results obtained from the large and small samples show essentially the same trends. The correlation coefficients established between variables for the moving average results are given in table 6. Despite the spread of L/T

|                                    | Population<br>Number | Thickness<br>against<br>Length | Thickness<br>against<br>Shape<br>(L/T) | Length<br>against<br>Shape | Thickness<br>against<br>Area<br>(LxT) | Length<br>against<br>Area | Shape<br>against<br>Area |
|------------------------------------|----------------------|--------------------------------|----------------------------------------|----------------------------|---------------------------------------|---------------------------|--------------------------|
| Sillimanite<br>K. feldspar<br>zone | 13                   | 0.63                           | -0.37                                  | 0.25                       | 0.86                                  | 0.81                      | N.S.                     |
|                                    | 12                   | 0.67                           | -0.27                                  | 0.36                       | 0.89                                  | 0.83                      | N.S.                     |
|                                    | 11                   | 0.68                           | -0.27                                  | 0.34                       | 0.89                                  | 0.83                      | N.S.                     |
|                                    | 10                   | 0.64                           | -0.29                                  | 0.38                       | 0.87                                  | 0.84                      | N.S.                     |
| Sillimanite<br>zone                | 9                    | 0.65                           | -0.37                                  | 0.26                       | 0.89                                  | 0.80                      | N.S.                     |
|                                    | 8                    | 0.53                           | -0.40                                  | 0.25                       | 0.85                                  | 0.73                      | -0.19                    |
|                                    | 7                    | 0.66                           | -0.40                                  | N.S.                       | 0.86                                  | 0.78                      | -0.23                    |
|                                    | 6                    | 0.68                           | -0.30                                  | 0.40                       | 0.89                                  | 0.84                      | N.S.                     |
|                                    | 5                    | 0.54                           | -0.45                                  | 0.36                       | 0.83                                  | 0.81                      | N.S.                     |
|                                    | 4                    | 0.63                           | -0.45                                  | N.S.                       | 0.85                                  | 0.84                      | N.S.                     |
| Staurolite<br>kyanite<br>zone      | 3                    | 0.57                           | -0.39                                  | 0.21                       | 0.89                                  | 0.65                      | N.S.                     |
|                                    | 2                    | 0.54                           | -0.46                                  | 0.33                       | 0.89                                  | 0.77                      | N.S.                     |
| Chloritoid<br>Staurolite<br>zone   | 1                    | 0.56                           | -0.40                                  | 0.32                       | 0.92                                  | 0.64                      | N.S.                     |

Table 6. Correlation table for moving average data for muscovite. Populations are numbered from low grade (1) to high grade (13). All correlations given are significant at the 99.5% significance level. N.S. = no significant linear correlation.

ratios in large samples (figs. 35 and 36) correlation between T and L is consistent within a narrow range; correlation coefficients 0.53-0.68 (mean 0.61); a closely similar mean correlation coefficient, but a wider range is obtained from the samples of 100 taken individually. Absolute dispersion (the spread of the data about the reduced major axis) is lower in the low grade samples, reflecting the smaller size of the crystals. Relative dispersion however, which is independent of size, is approximately the same in all samples.

Correlation between T and L/T (ie. size versus shape) is lower than between T and L (table 6), the range of correlation coefficients in large samples being 0.27-0.46 (mean 0.37). However 40% of the small samples of 100 measurements from the high grade rocks (sillimanite grade and above) showed no significant correlation between T and L/T. An examination of this plot graphically (fig. 37) suggests that the scatter of points defines an area rather than a curve, which would explain the low and insignificant correlation coefficients.

Consistent high correlations are recorded between area LxT against both T and L. Area LxT versus shape L/T shows no significant linear correlation in small samples. A similar lack of correlation exists in all but two of the large sample distributions (table 6).

Histograms of T, L, L/T and LxT are typically unimodal,

CHLORITOID-STAUROLITE ZONE

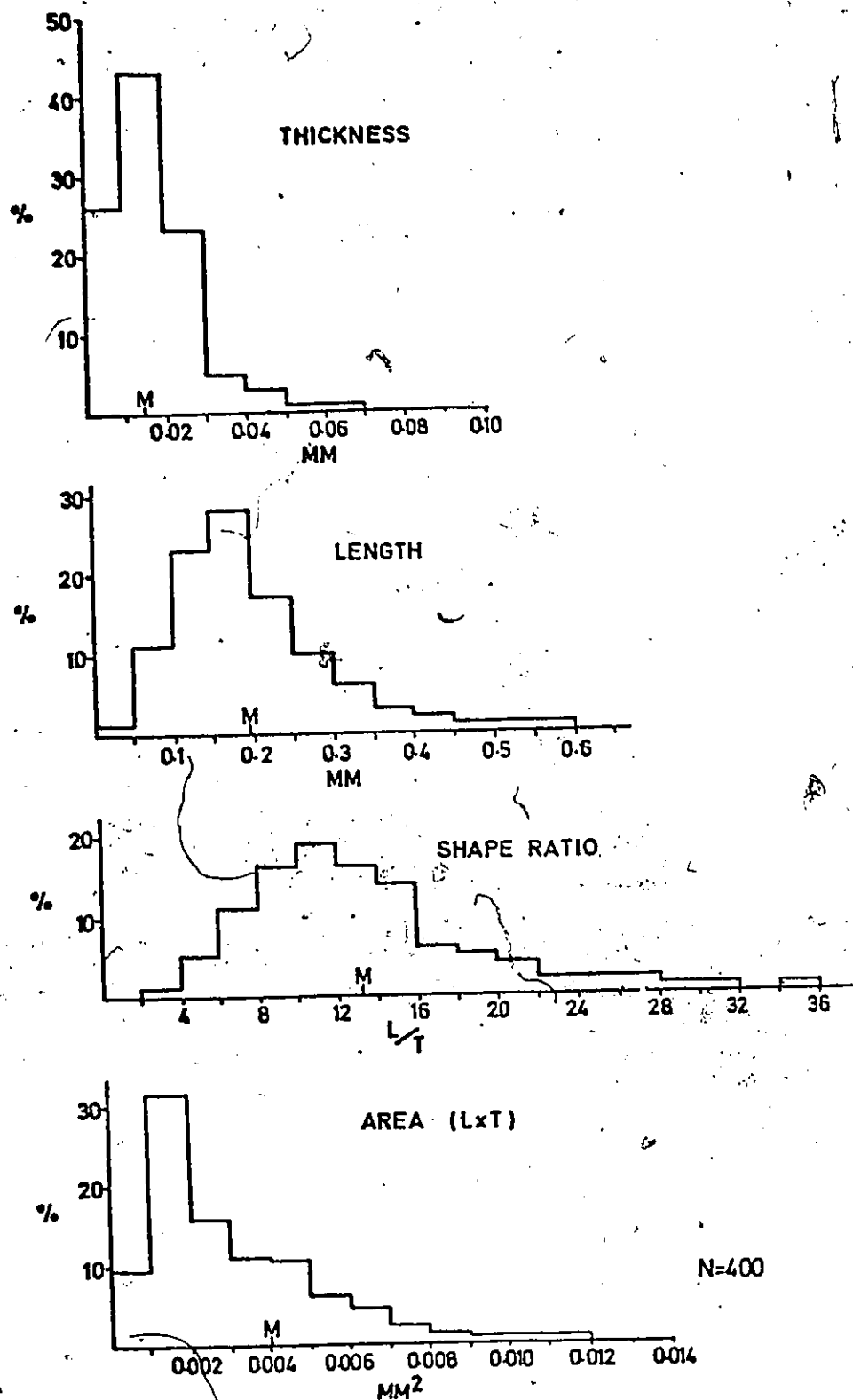


Fig. 35 Histograms of T, L, L/T and LxT for the lowest grade moving average sample (no. 1). M = mean value.

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SILLIMANITE ZONE

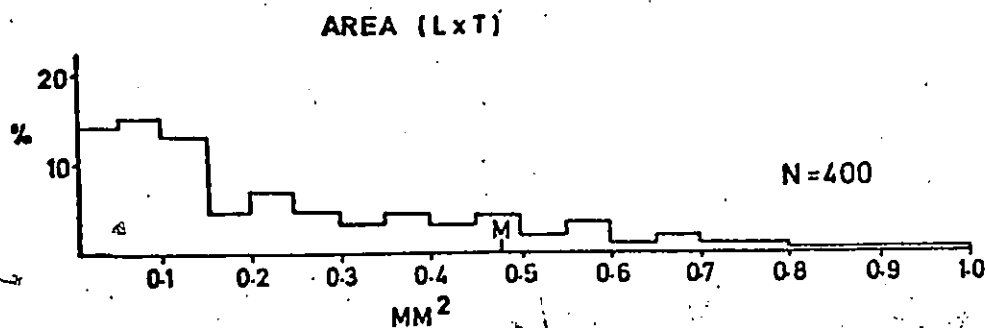
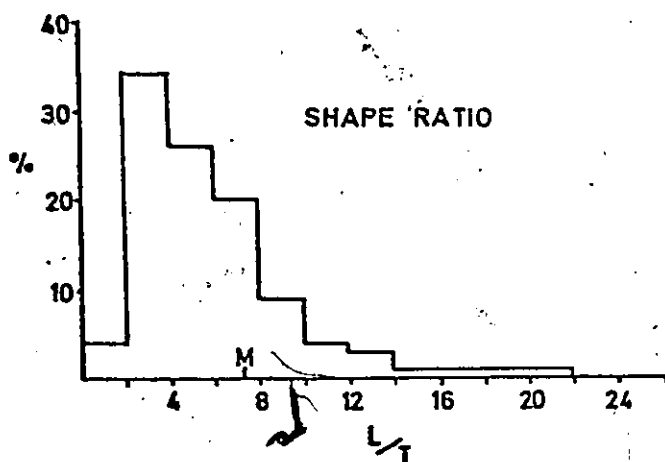
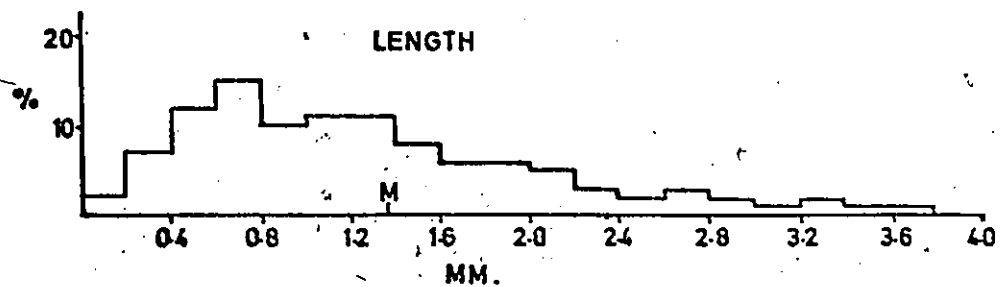
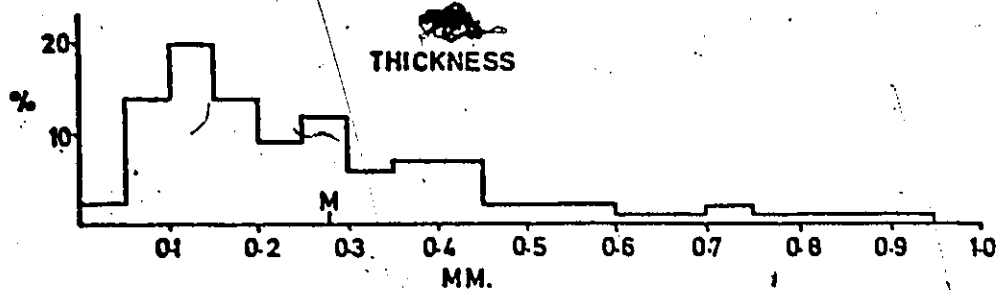


Fig. 36 Histograms of T, L, L/T and LxT for the highest grade moving average sample (no. 13). M=mean value. Note that all the scales are not the same as in fig. 35.

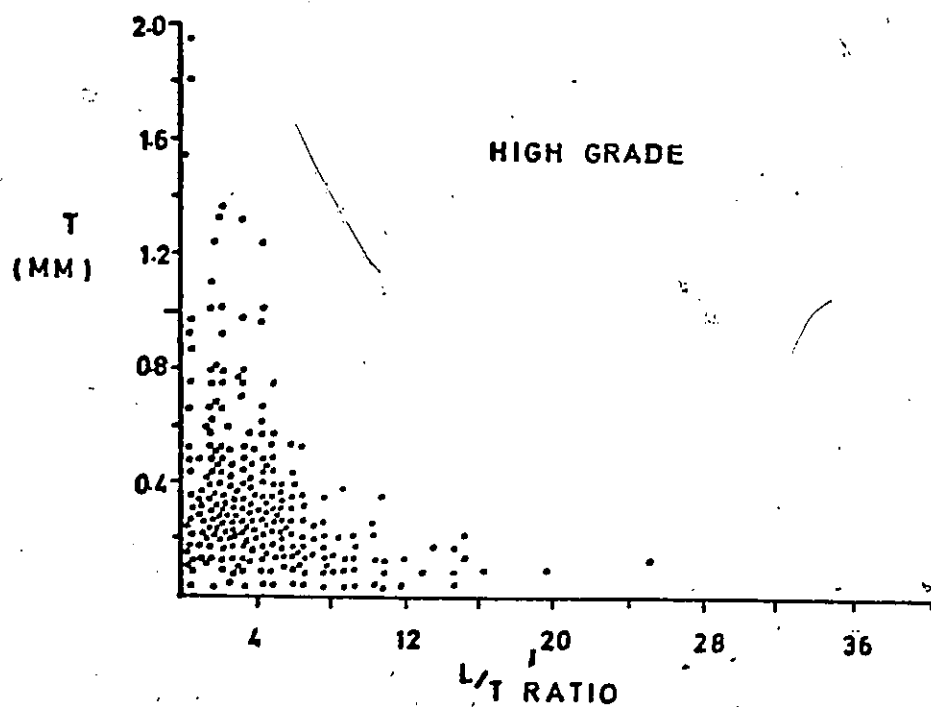
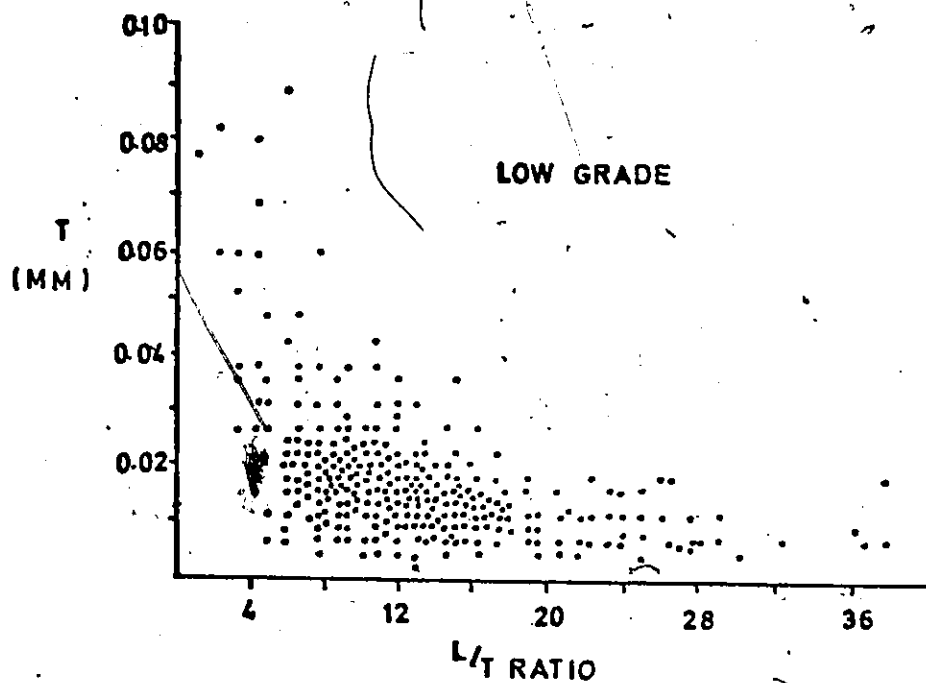


Fig. 37. Graphs of muscovite size (T) against shape (L/T) for the lowest and highest grade moving average samples (1 and 13 respectively). 400 points per graph. Note that the scale for T is not the same in each graph.

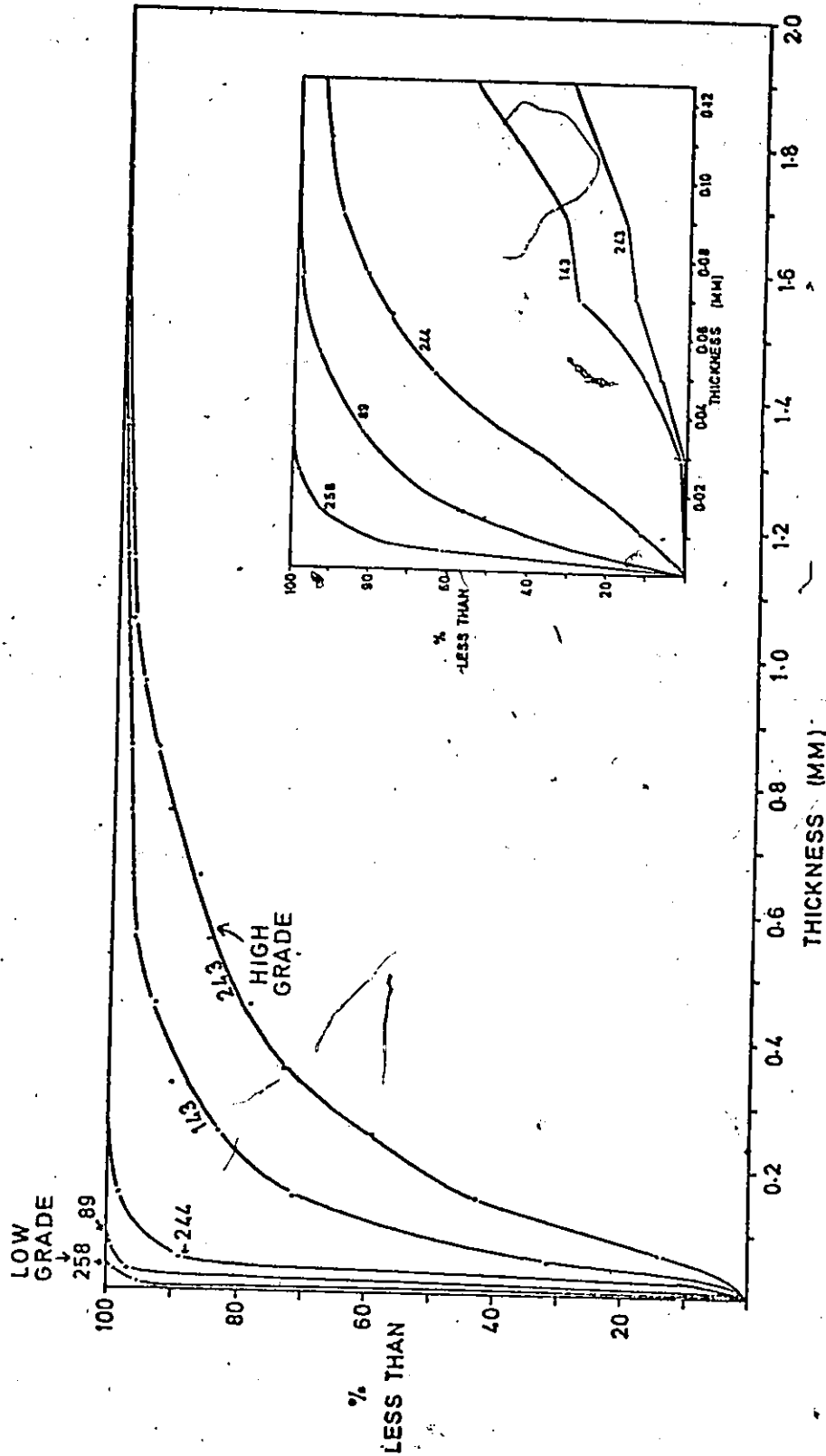


Fig. 38 Cumulative histograms of muscovite crystal size (T) for 5 samples from different metamorphic grades (258 - low grade, 243 - high grade; for exact localities see fig. 19). 200 measurements per sample. In the inset diagram the T scale has been enlarged to show the shape of the low grade distributions. (% less than indicates the percentage of crystals in a sample with a thickness less than a given T value).

though minor secondary modes are quite common, especially in small samples (figs. 35 and 36). Most histograms are positively skewed, the mean being slightly greater than the mode. Cumulative distributions of both large and small samples have a similar form. In fig. 38 cumulative histograms for five samples, in which all measurements were made from small volumes of rock, have been compiled into a single diagram. The different slopes of the curves are considered to be due essentially to the different size ranges of the crystals involved. An enlargement of the left-hand portion of the diagram has been made (fig. 38 inset); the irregularities in the curves of the high grade specimens (143 and 243) are probably due to low representation in the sample, and are not thought to have any particular significance. These data will be used later for the discussion of nucleation models.

When plotted on probability paper, cumulative distributions are approximately log-normal, fig. 39 and 40. Cumulative size distributions, given by thickness  $T$ , are less regular than those for shape  $L/T$ , the latter consistently defining two intersecting lines of different gradient. The reason for this is not known.

#### Variation Between Samples - Muscovite.

There is a large increase in crystal size ( $T$ ) with grade of metamorphism (fig. 41); mean  $T$  from the highest grade

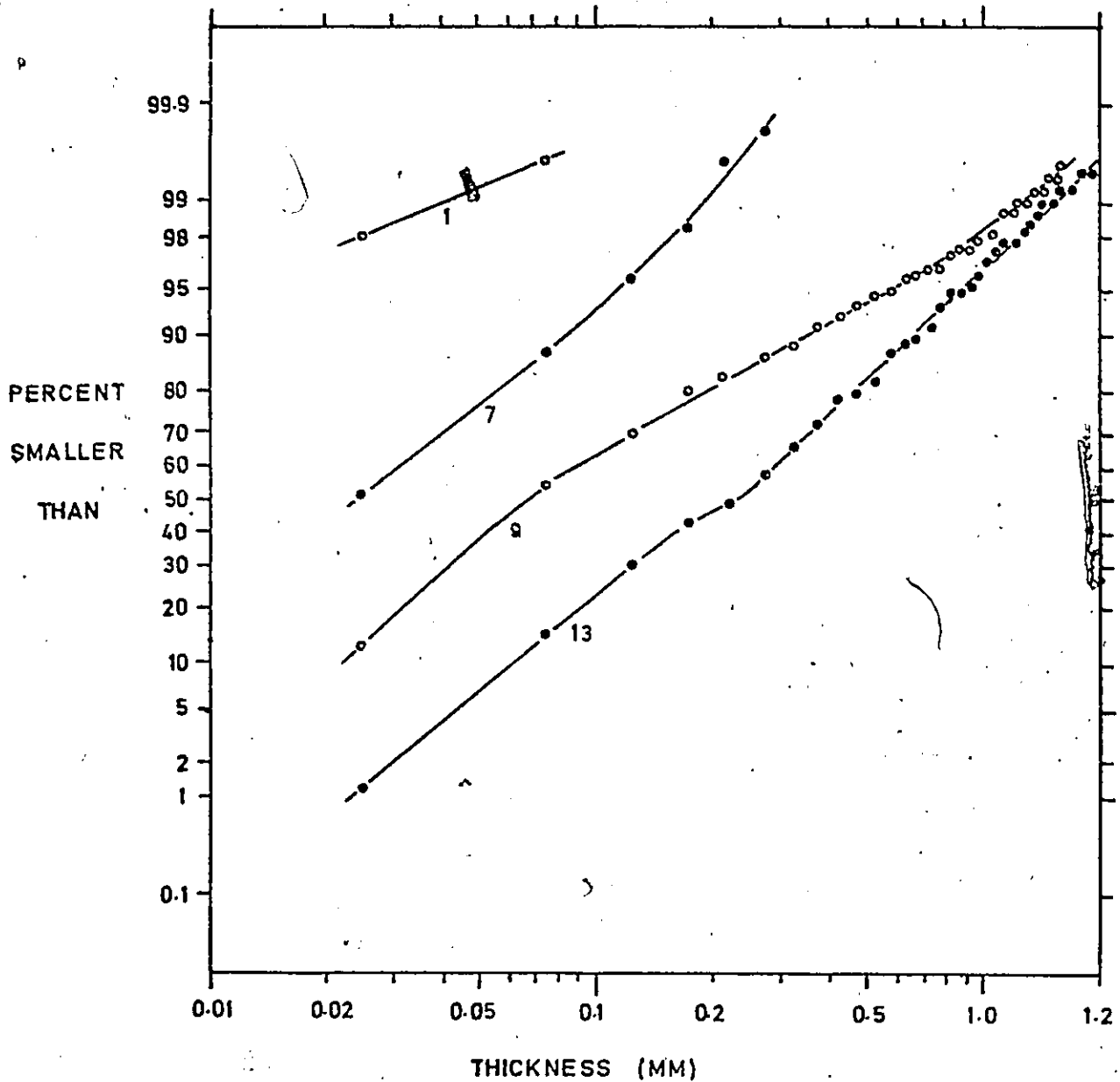


Fig. 39 Cumulative size distribution plotted on log. probability paper, for four moving average samples (1 - low grade, 13 - high grade).

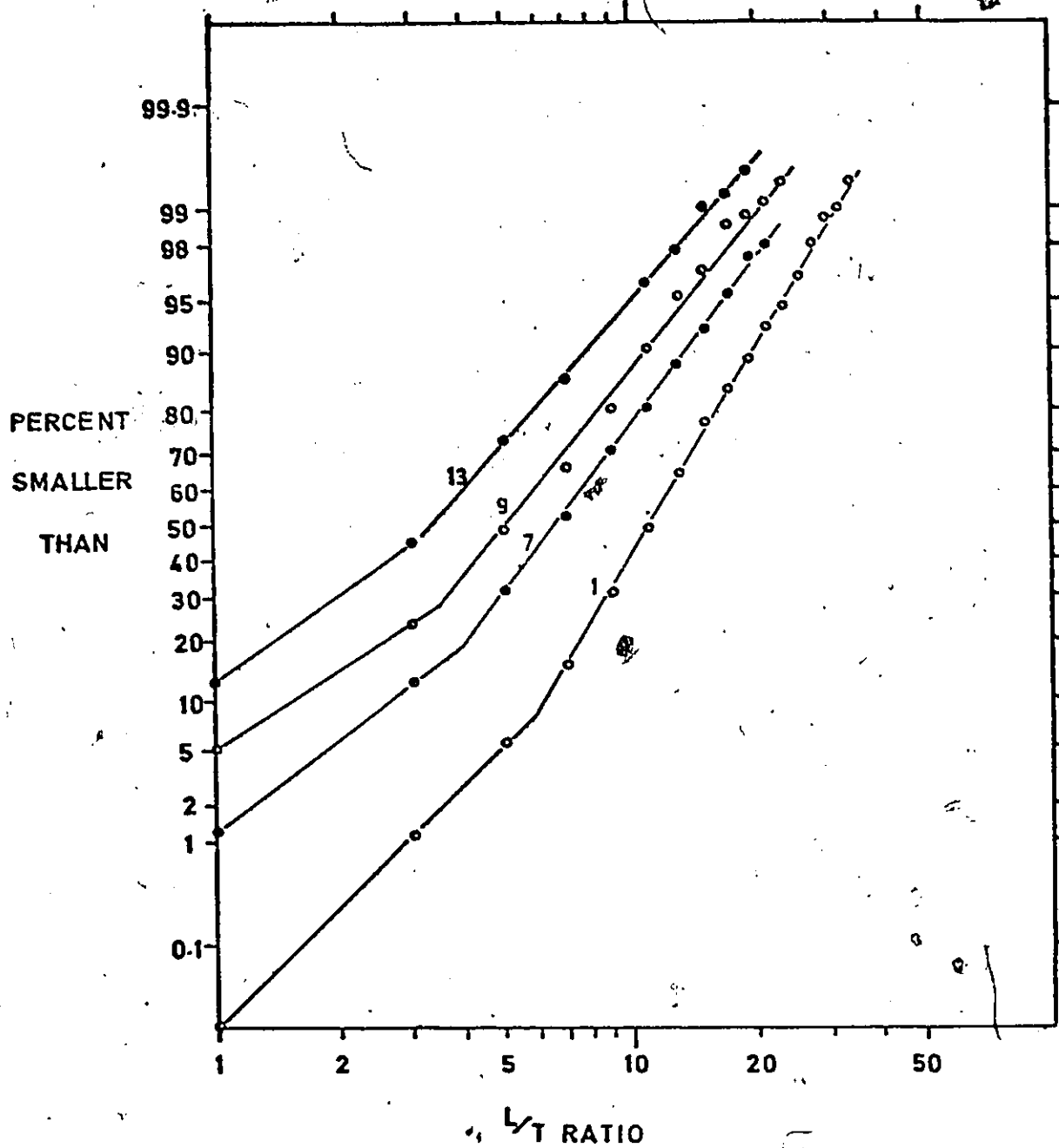


Fig. 40 Cumulative shape distributions plotted on log. probability paper, for four moving average samples (1 - low grade, 13 - high grade).

samples being approximately nineteen times that from low grade (0.326 mm. and 0.017 mm. respectively). Corresponding increase of L and LxT also occur.

It is of interest to know if there is an associated change in the kurtosis of the histograms with grade of metamorphism. However this cannot be measured directly because of the great range in size along the metamorphic belt necessitates changing the histogram interval. Thus a statistic was sought which would characterize the spread of the population. It was found that the standard deviation/mean enables a useful comparison to be drawn between samples. Quantitatively for any given mean, the greater the value of  $SD/M$  the more platykurtic the distributions. Thus if  $SD/M$  remains approximately constant as the mean increases, the spread of the distributions must have increased proportionately (kurtosis decreased). Similarly greater  $SD/M$  values associated with increasing means indicates a greater spread as the mean increases, whilst decreasing  $SD/M$  values as the mean increases may indicate either that the spread is the same or is decreasing.

$SD/M$  for the T parameter shows no systematic trend with metamorphic grade (increasing size, therefore increasing M), so it is concluded that the kurtosis increases proportionately with the mean - table 7. However for L there is a trend towards more leptokurtic distributions at low metamorphic grades.

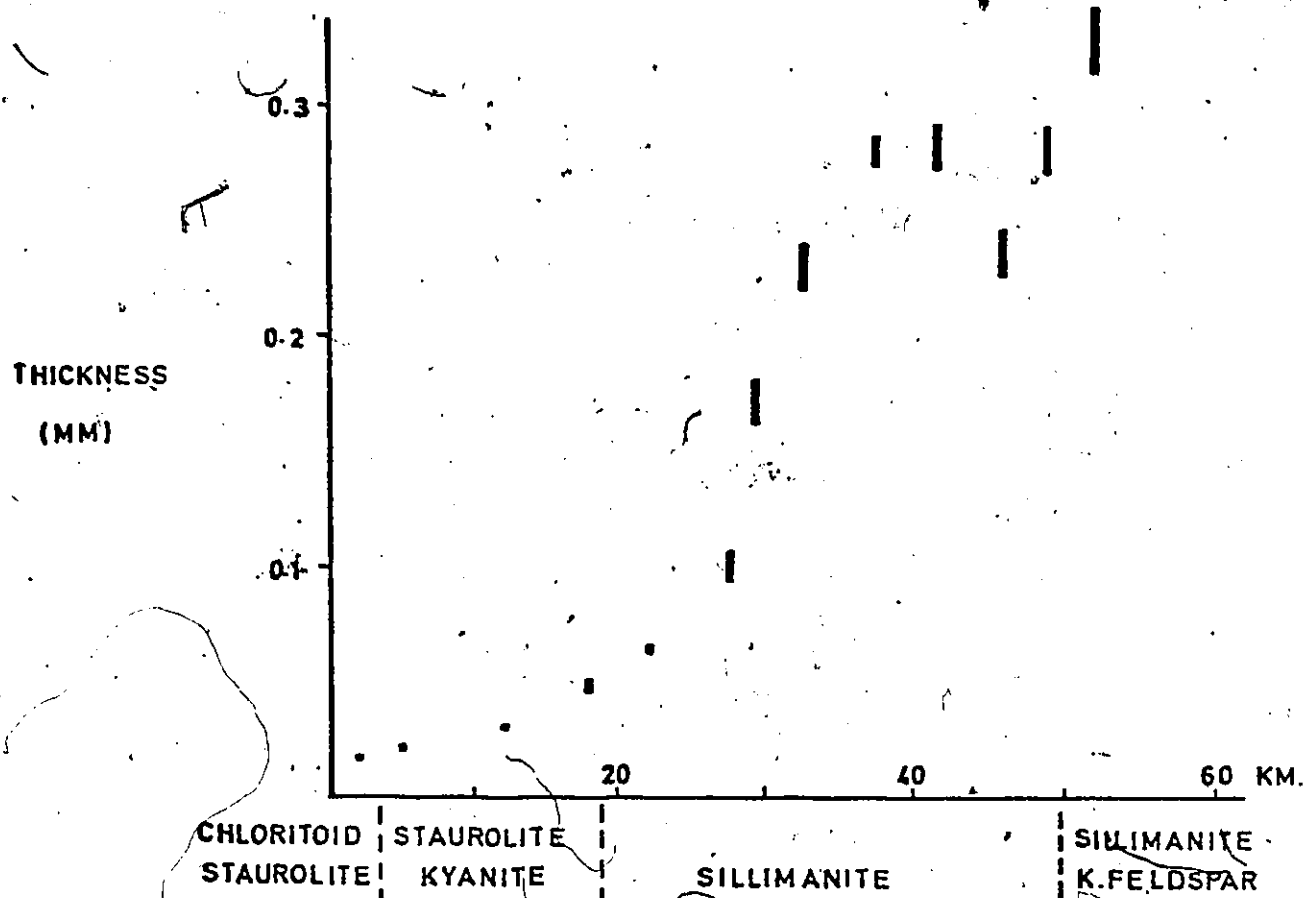


Fig. 41 Mean thickness of muscovite crystals (plotted as the standard error of the mean) against grade of metamorphism - moving average data.

| Population No. | Thickness (T) | Length (L) | Shape (L/T) |                                                             |
|----------------|---------------|------------|-------------|-------------------------------------------------------------|
| 13             | 0.86          | 0.72       | 0.62        | HIGH GRADE:<br>large crystals<br>with low shape<br>ratios.  |
| 12             | 0.75          | 0.74       | 0.60        |                                                             |
| 11             | 0.81          | 0.76       | 0.55        |                                                             |
| 10             | 0.77          | 0.73       | 0.52        |                                                             |
| 9              | 0.81          | 0.64       | 0.57        |                                                             |
| 8              | 1.07          | 0.68       | 0.72        | LOW GRADE:<br>small crystals<br>with large<br>shape ratios. |
| 7              | 1.29          | 0.70       | 0.64        |                                                             |
| 6              | 0.64          | 0.47       | 0.48        |                                                             |
| 5              | 0.65          | 0.55       | 0.52        |                                                             |
| 4              | 0.78          | 0.67       | 0.60        |                                                             |
| 3              | 0.96          | 0.60       | 0.55        |                                                             |
| 2              | 0.73          | 0.44       | 0.43        |                                                             |
| 1              | 0.76          | 0.47       | 0.43        |                                                             |

Table 7. Standard deviation/Mean values for moving average samples.  $SD/M$  used as a measure of the kurtosis of the distributions; for fuller explanation see text.

As expected from the increase in T and L, the average volume of muscovite increases with ascending grade of metamorphism. The rate of increase, fig. 42, is rapid up to the top of the sillimanite zone, at which grade it declines quite abruptly.

Average crystal surface area also increases rapidly with metamorphic grade, Fig. 42. However the rate of increase with metamorphic grade is less, and the declining rate starts lower in the sillimanite zone than for volume.

A systematic change in muscovite shape ratio L/T occurs along the Fernleigh belt, figs. 43A and B. At the lowest grade of metamorphism the modal shape ratio is 11:1 (mean 13:1) changing to 3:1 (mean 5:1) at the high grade end (moving average data). It can be seen that the change in shape is a gradual one. At the high grade end of the metamorphic belt it appears that a plateau has been reached, with the five end samples all having the same modal shape ratio. In fig. 44, the histograms for shape ratio (moving average results) have been compiled into a single diagram. It is evident that the histograms become increasingly leptokurtic at the higher grades of metamorphism, reflecting a narrower spread of shape ratios in the high grade rocks.  $SD/M$  values for crystal shape (table 7) are approximately the same at all grades, indicating that the spread of the distributions increases as the mean L/T increases (ie. as metamorphic grade

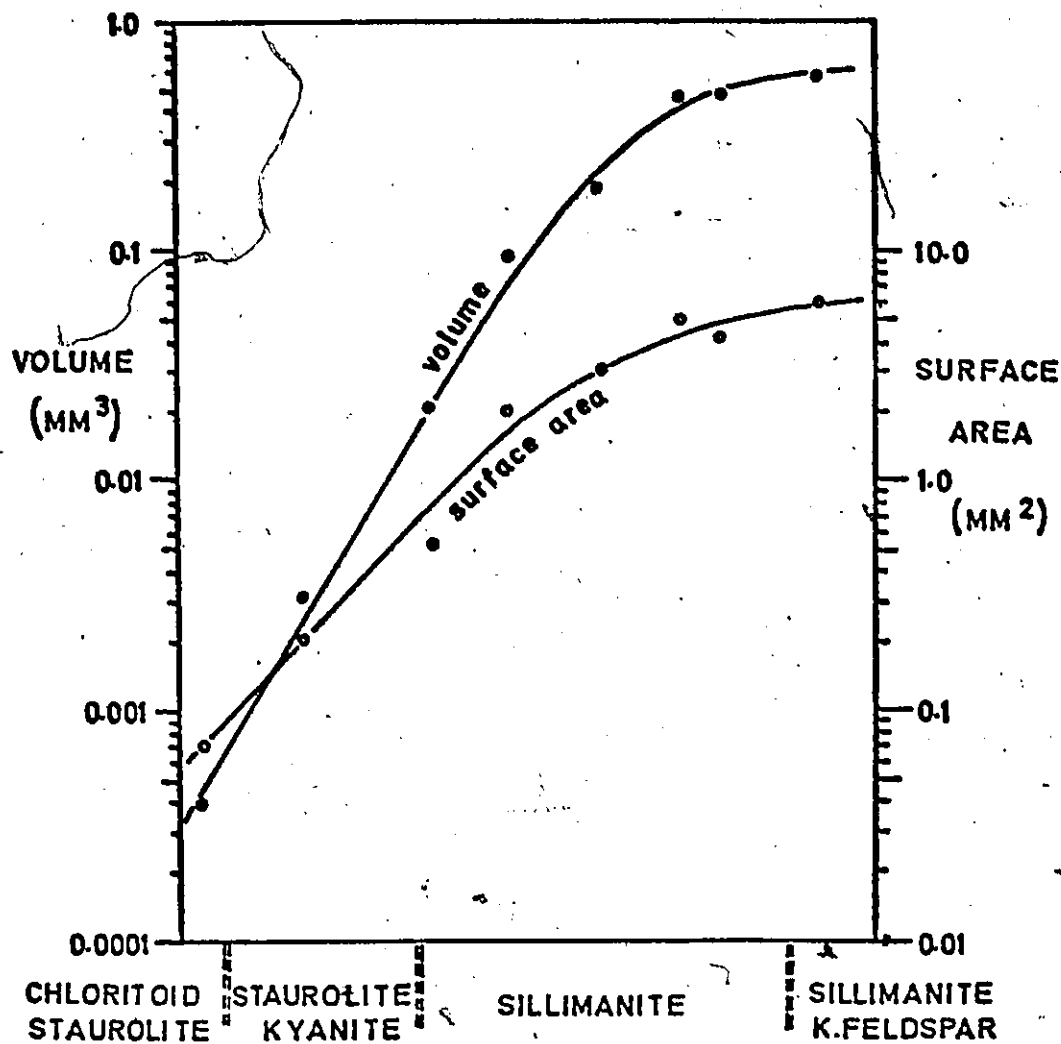


Fig. 42 Mean volume and mean surface area of muscovite crystals plotted against grade of metamorphism.

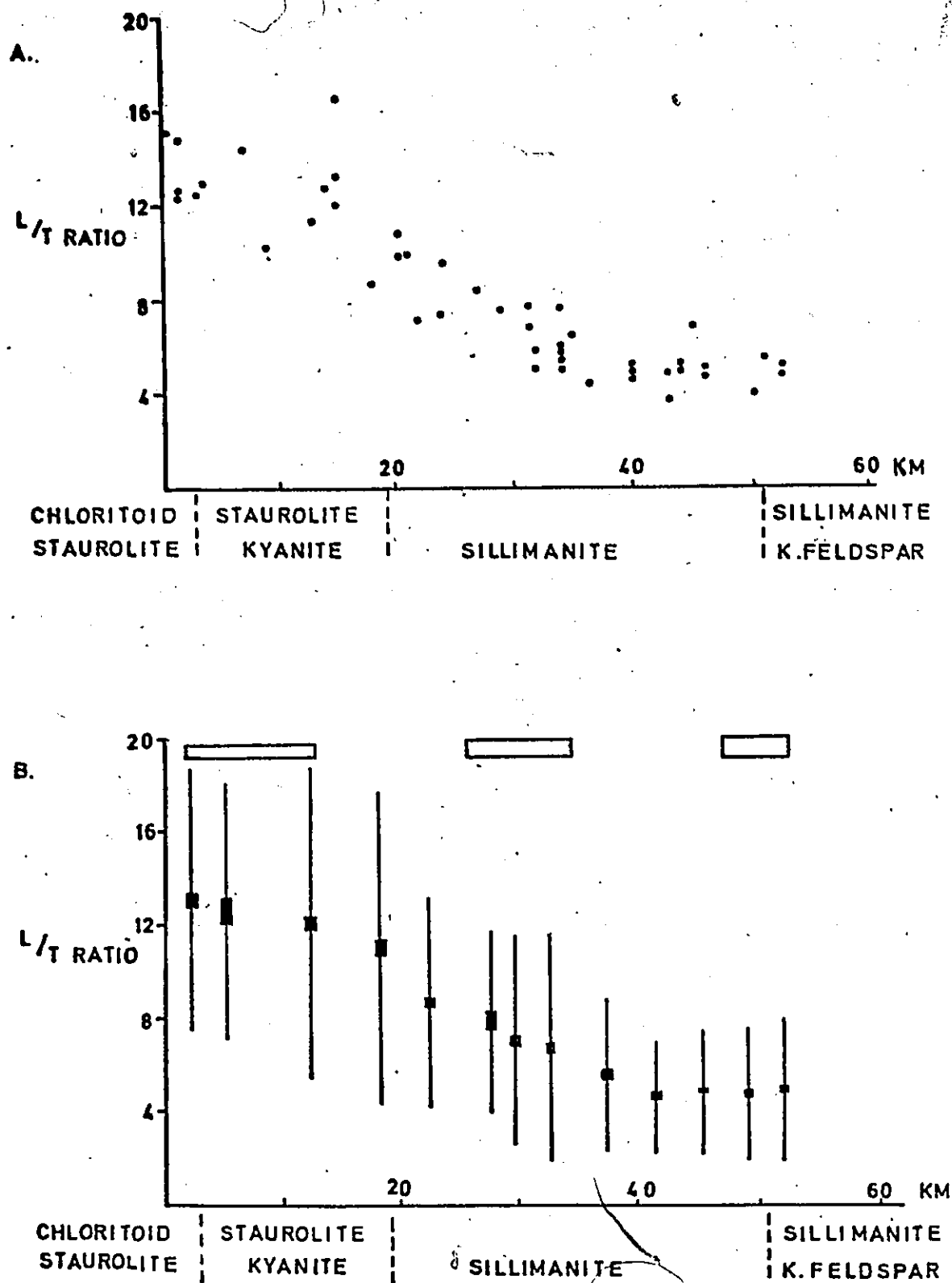


Fig. 43 Muscovite shape plotted against grade of metamorphism.  
 A. Mean shape ratio (L/T) for 47 small samples ( $n = 100$ )  
 B. Shape ratio (expressed as the standard error of the mean  $\pm 1$  standard deviation) for 13 moving average samples.  
 Typical crystal shapes, as seen in section, are represented above the graph.

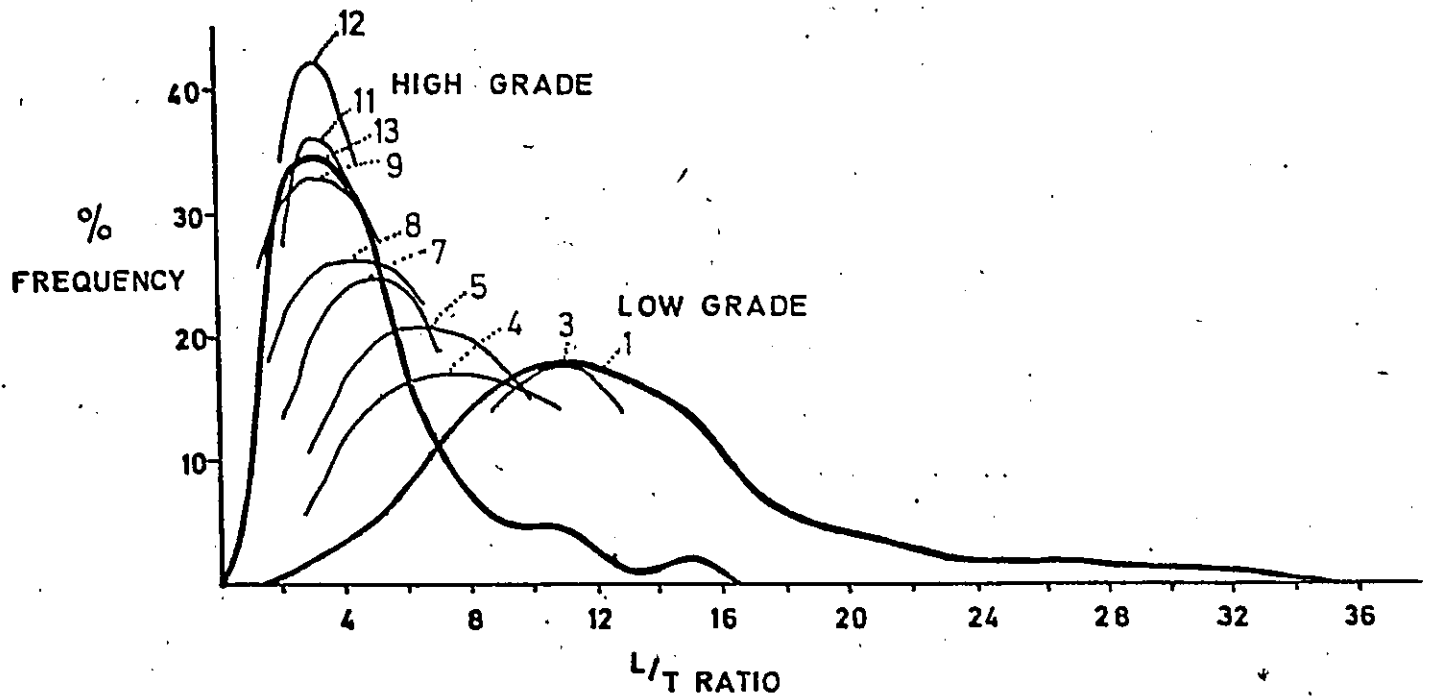


Fig. 44 Compilation of histograms of shape ratios, moving average data (1 - low grade, 13 - high grade).

decreases). This is apparent visually in fig. 44.

The cumulative curves for crystal shape (fig. 40), which consistently define two intersecting log-normal lines, show that the skewness of the histograms is consistent throughout the metamorphic gradient examined. The point of intersection of the two lines occurs at progressively lower L/T ratios as the metamorphic grade increases, but does not coincide with the modes of the histograms.

Fig. 45, in which the shape ratio of muscovite crystals (used as a measure of metamorphic grade) is plotted against the mean surface area/volume ratio, shows that the  $SA/VOL$  ratio is much reduced for crystals at high grade (low grade is 17 times high grade).

Some interesting results were obtained from the comparison of measurements of muscovite crystals enclosed in porphyroblasts of plagioclase feldspar with those from crystals outside the porphyroblasts. Whilst the results do not have the statistical significance of those reported earlier, the trends are apparent. In the low grade rocks (chloritoid-staurolite and staurolite-kyanite zones) the inclusions are present only in the central portions of the porphyroblasts. These enclosed crystals possess modal L/T ratios which correspond approximately to the chloritoid-staurolite zone (figs. 46A and B). The other specimens measured (C and D of fig. 46), in which the included crystals are dispersed

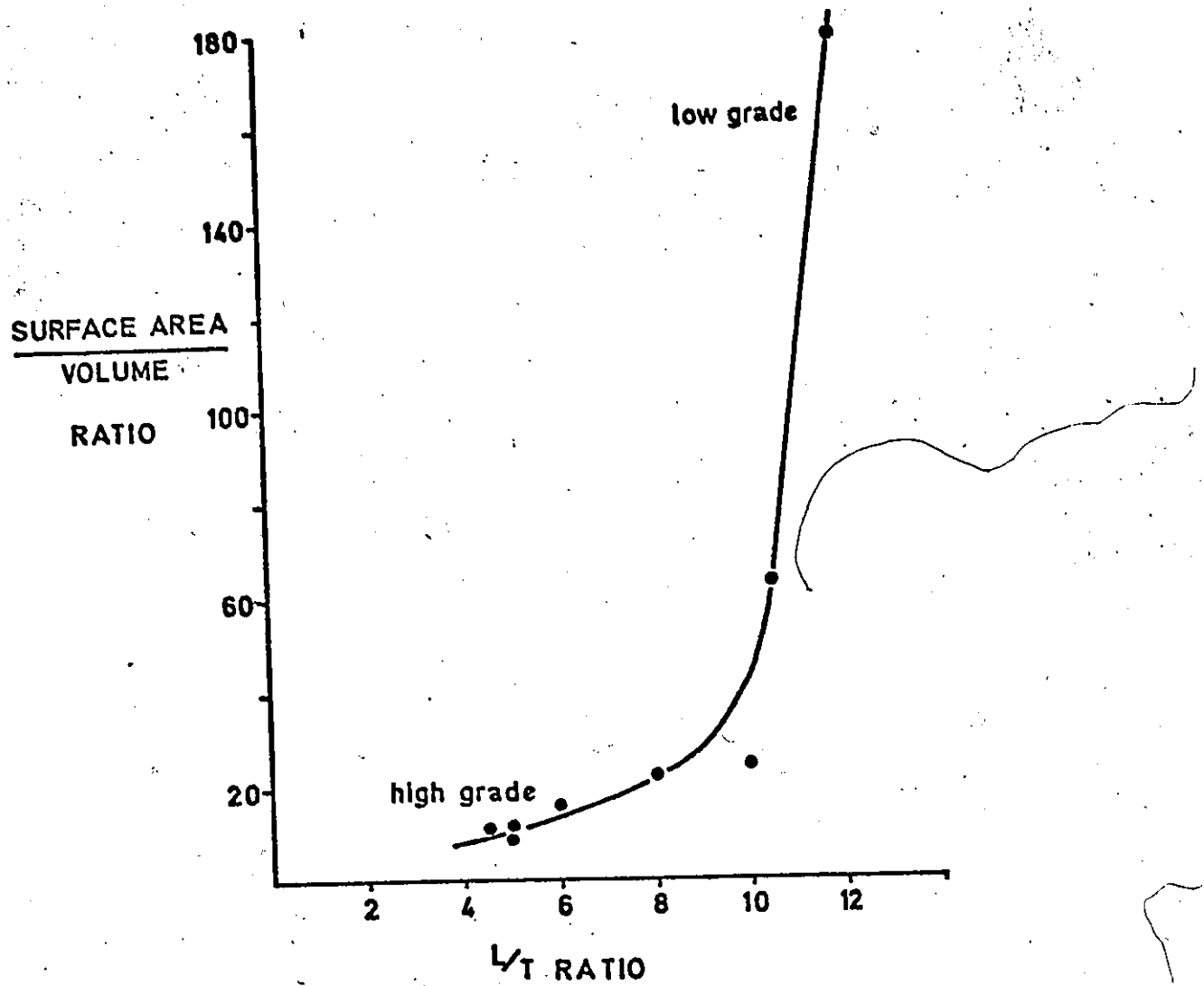


Fig. 45 Mean shape ratio of muscovite crystals of various metamorphic grades plotted against their surface area/volume ratios.

throughout the porphyroblasts, come from the sillimanite zone. In D the shapes of the included crystals cannot be distinguished from those in the foliation outside the porphyroblast; whilst in C a range of shapes is seen, possibly representing an intermediate step between B and D.

It is evident from fig. 47 that the mean size of the crystals enclosed in the porphyroblasts is less than the mean for those in the foliation. This is particularly marked in the highest grade sample (D of fig. 46), where the mean T (foliation mica) is five times mean T (included mica), yet their shape ratios are indistinguishable. It is concluded, therefore, that crystal shape and size are independent variables.

In fig. 48 information from the plots of T against L for the lowest and highest grades of metamorphism (moving average data) have been compiled into a single graph. The large difference in the ranges of the two distributions is apparent. The gradient of the reduced major axes are also significantly different, reflecting the different modal shape ratios of the muscovite crystals at each end of the metamorphic belt. The constants of the reduced major axes (ie. the intercepts with the y-axis) are consistently slightly negative (average -0.1).

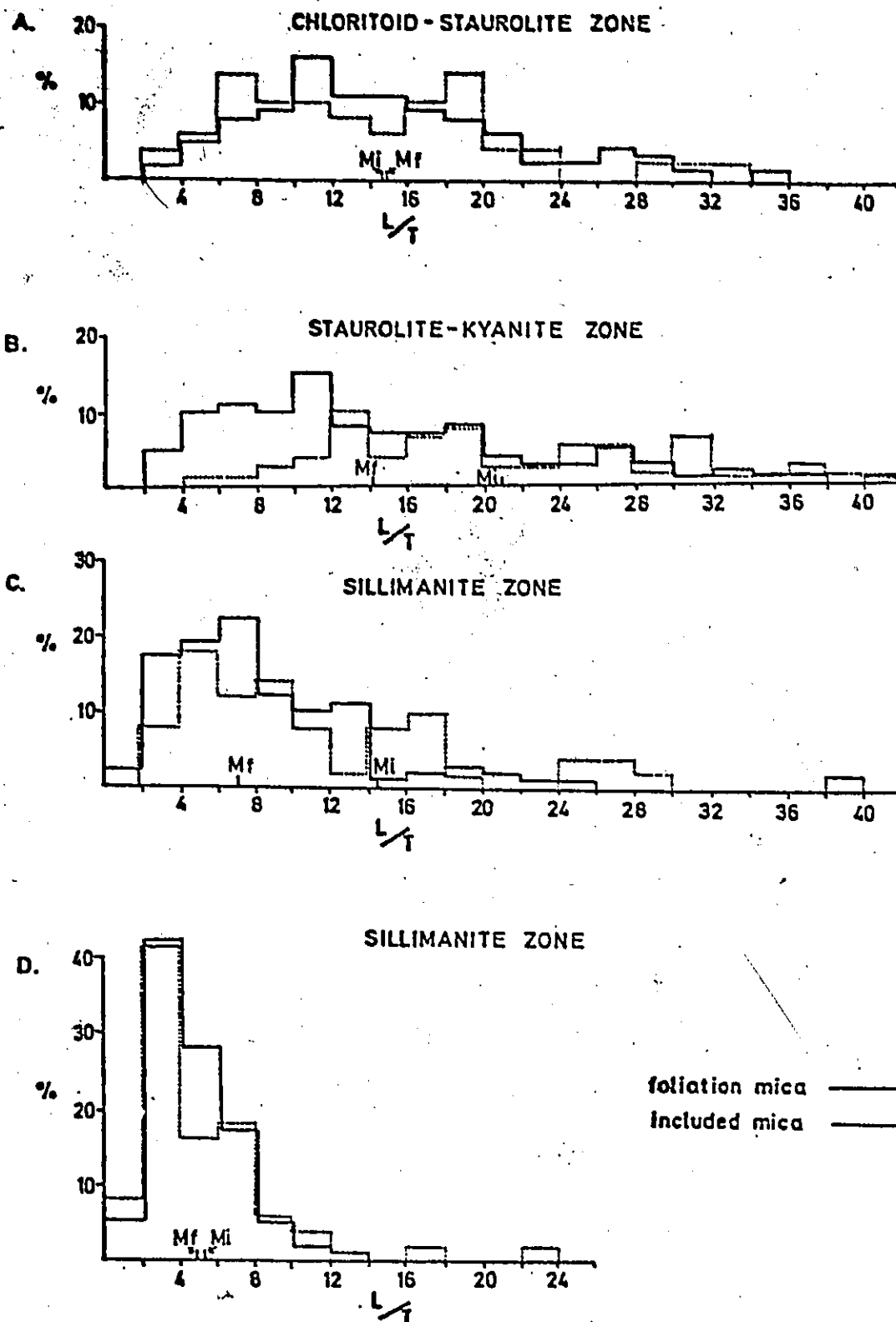


Fig. 46 Comparison of histograms for shape ratio of muscovite crystals occurring in the foliation with those included in porphyroblasts. Metamorphic grade increases from A to D. N (foliation mica) = 100; N(included mica) = 50 in each case. Mi = mean (included mica); Mf = mean (foliation mica).

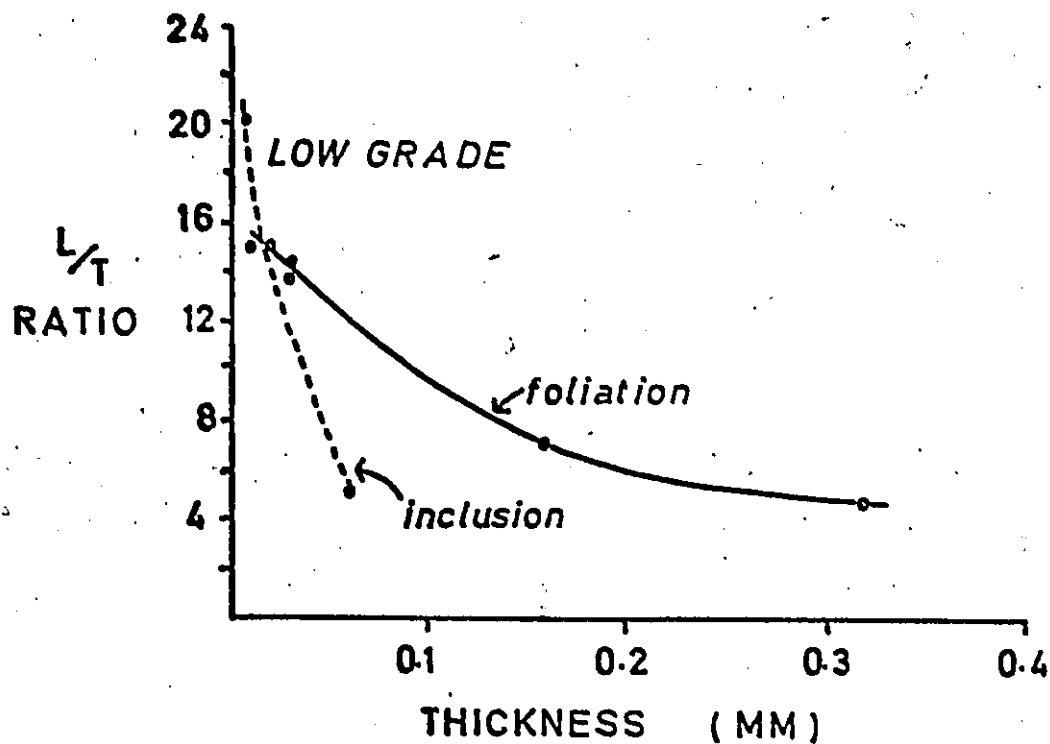


Fig. 47 Graph of size (thickness) against shape ratio for muscovite crystals occurring in the foliation and in porphyroblasts of plagioclase feldspar.

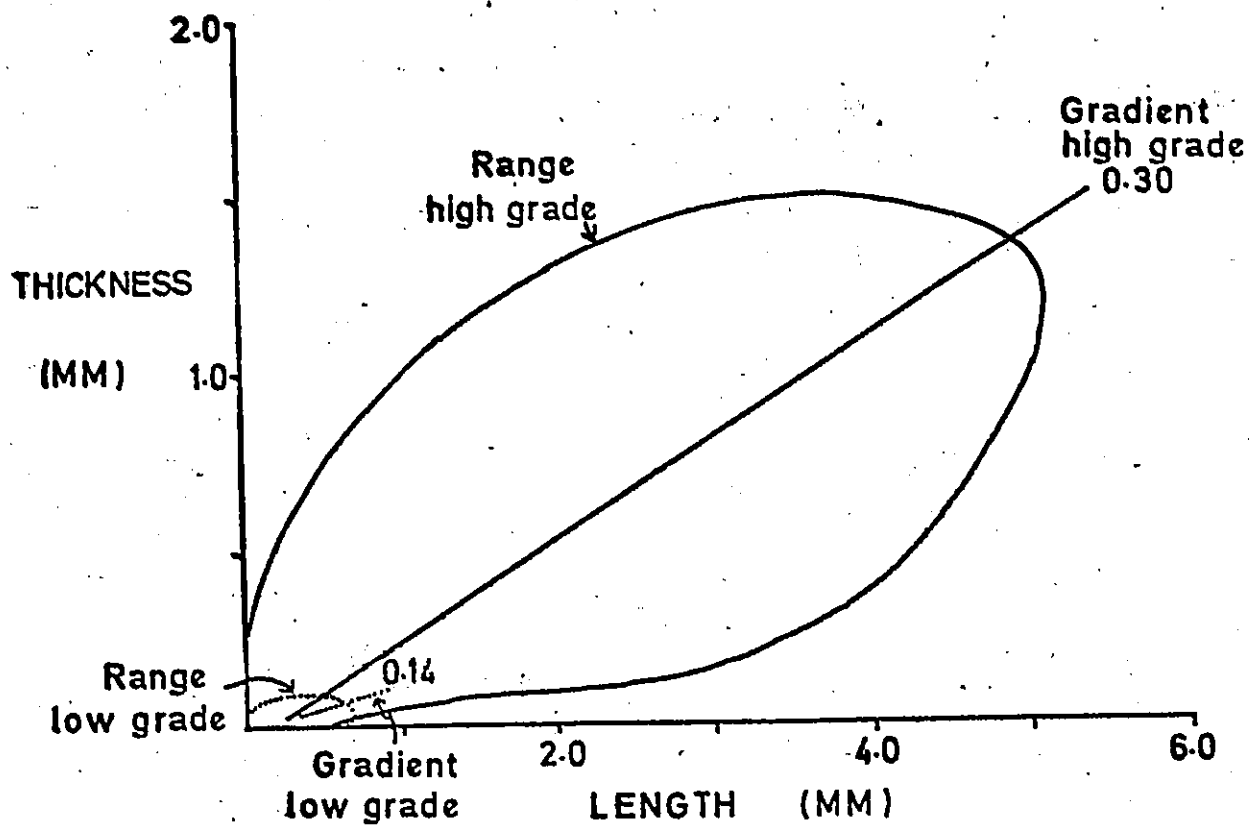


Fig. 48 Comparison of the ranges of distributions and the gradients of reduced major axes for the graphs of T against L for the highest and lowest grade moving average samples.

| Thickness<br>vs<br>Length | Thickness<br>vs<br>Shape | Length<br>vs<br>Shape | Thickness<br>vs<br>Area | Length<br>vs<br>Area | Shape<br>vs<br>Area |
|---------------------------|--------------------------|-----------------------|-------------------------|----------------------|---------------------|
| 0.92                      | -0.86                    | -0.73                 | 0.96                    | 0.93                 | -0.73               |

Table 8. Correlation table based on mean T and mean L statistics from 47 samples of 100 measurements each, taken from along the length of the metamorphic belt. All correlations are significant at the 99.5% significance level.

Table 8 is a correlation table based on the mean L and T values for the 47 small samples from along the length of the Fernleigh belt. When compared with table 6 several differences are apparent. Correlation between T and L is higher than in the individual samples and populations. However more significantly a strong correlation exists between the measures of crystal size ie. T, L, LxT, and crystal shape L/T, in marked contrast to the samples and populations from a single metamorphic grade. In fig. 49 this data for the moving average results has been plotted using thickness as the indicator of size. Similar distributions of points occur in the graphs of L and LxT against L/T. The line joining the points from low grade to high grade defines the "prograde growth trend" for muscovite. Low grade crystals are characterized by their small size and tabular shape, whilst at high metamorphic grades crystals are on average larger and more equidimensional in form.

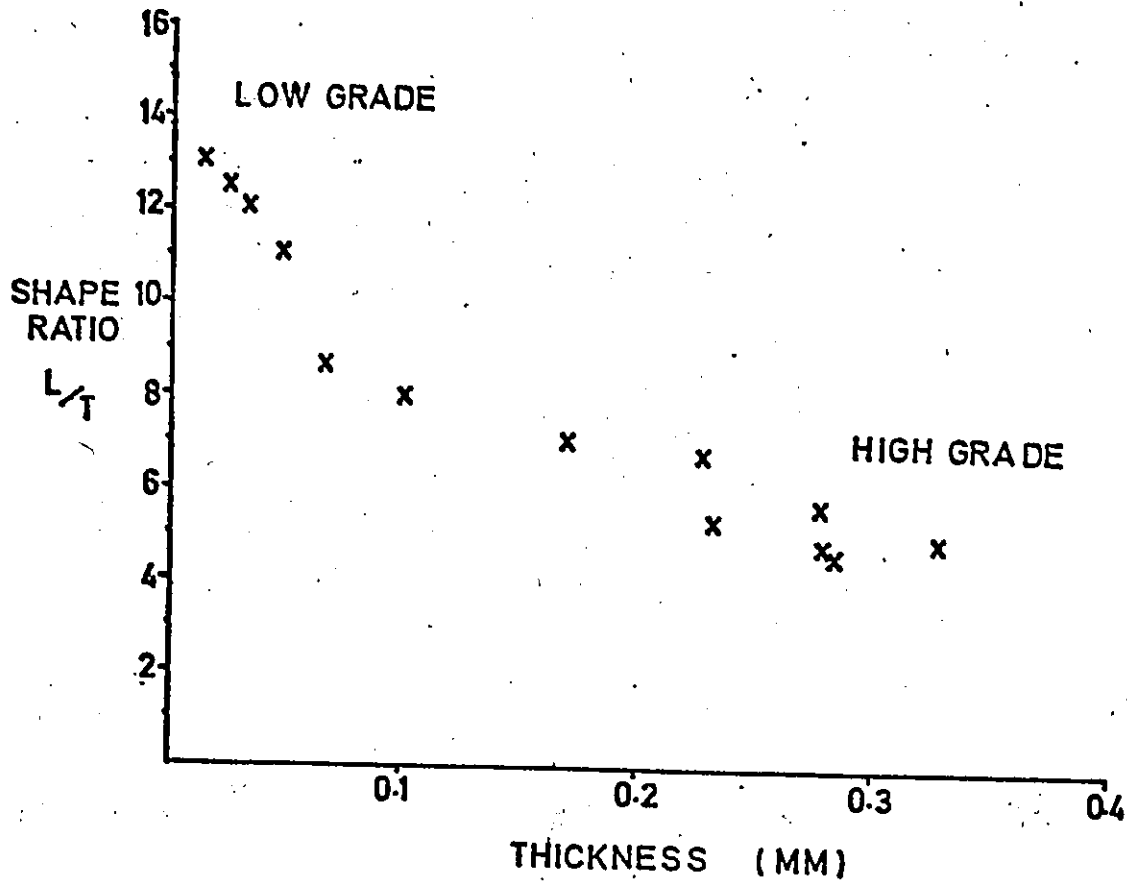


Fig. 49 Mean size (thickness) of muscovite plotted against mean shape ratio (moving average data). The distribution of points from low grade to high grade defines the, "prograde growth trend".

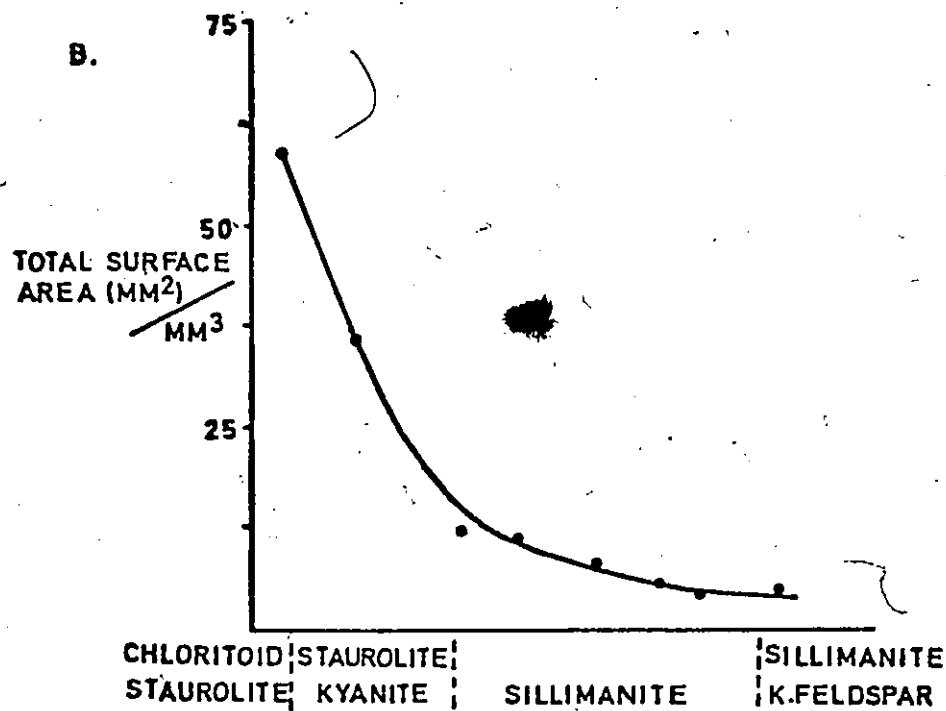
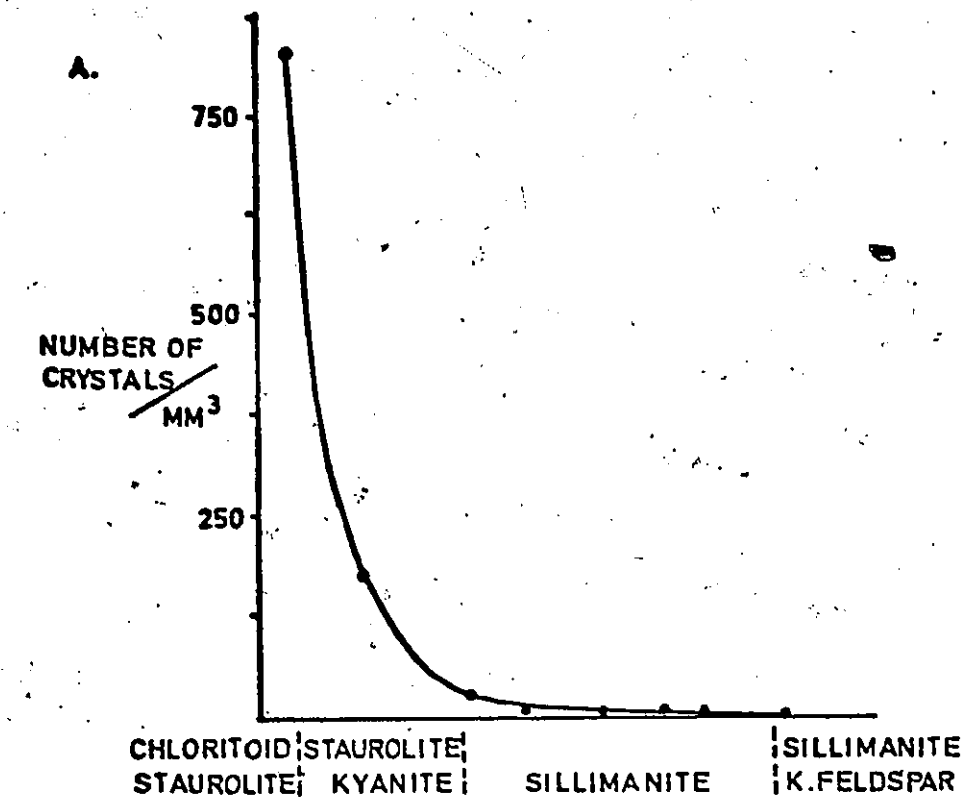


Fig. 50

A Estimated number of crystals per unit volume of rock (assumptions stated in text) plotted against grade of metamorphism.

B Estimated total surface area of muscovite crystals per unit volume of rock plotted against grade of metamorphism.

Utilising the estimates of average volume per crystal at different grades of metamorphism (fig. 42), the number of crystals per unit volume of rock may be calculated (fig. 50A, this calculation is based on the assumption that the rock contains 50% muscovite by volume). If this value is then combined with that for the mean surface area per crystal, we arrive at an estimate of the total surface area of muscovite per unit volume of rock. This has been plotted in fig. 50B, and it is apparent that there is a substantial reduction in total surface area/unit volume of rock with increasing metamorphic grade. Thus the rock as a whole, with regard to muscovite, shows a similar, but enhanced trend to that of the individual crystal.

#### Crystal Size Distribution.

Combining the estimates of the total numbers of muscovite crystals per unit volume of rock (fig. 50A) with the data from the cumulative histograms for crystal size (fig. 38), an estimate of the actual number of crystals in each size range in a given volume of rock can be made. This data is presented graphically in figure 51. Considering the solid portions of the curves, the rate of increase in numbers of crystals in each successively larger size range appears to lessen slightly with decreasing metamorphic grade (although the actual numbers of crystals involved in the low grade specimens is greater by several orders or magnitude). All

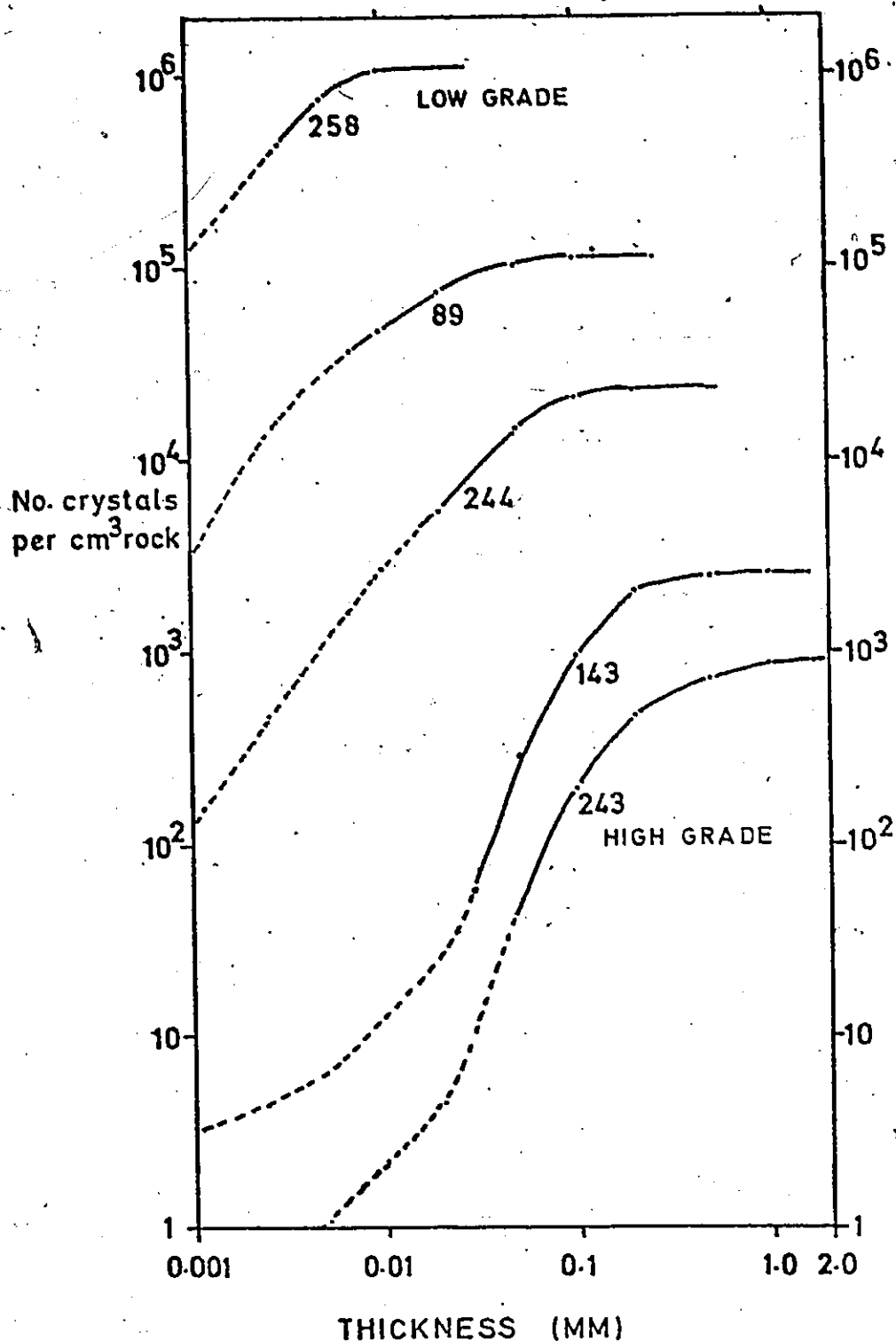


Fig. 51 Estimated number of muscovite crystals per cubic centimetre of rock at different metamorphic grades. Estimates are derived from the data on crystal volume and from the crystal size distributions. The dotted portions of the curves are interpreted from the cumulative histograms.

curves flatten out towards the right-hand side, reflecting the relatively small numbers of crystals in the larger size ranges of each sample. The dotted portions of the curves are interpreted from the cumulative histogram curves (fig. 38); curves 243 and 143 suggest that the rate of increase of numbers of crystals with increasing size is greatly reduced at the small size ranges of the populations.

However despite the variation in size ranges of the crystals from different metamorphic grades, the shape of the distributions is essentially the same. The moving average results, fig. 39, when plotted on log-probability paper, do however show some variability. Whilst this may be partly due to the method of plotting the data, it appears that this is not the complete explanation since no systematic trend could be found.

#### Biotite.

The results for the size and shape of biotite from the gneisses are summarized in table form (table 9). Correlation values (table 9A) are similar to those for muscovite (table 6), although no correlation was established between thickness and shape for any of the biotite samples. Histograms (not included) are unimodal, and slightly positively skewed as for muscovite. The mean values for crystal shape (L/T) (table 9B) show a similar, though less marked decrease with increasing metamorphic grade to those of muscovite. Biotite size, howe-

| Specimen<br>Number | Thickness<br>vs<br>Length | Thickness<br>vs<br>Shape | Length<br>vs<br>Shape | Thickness<br>vs<br>Area | Length<br>vs<br>Area | Shape<br>vs<br>Area |
|--------------------|---------------------------|--------------------------|-----------------------|-------------------------|----------------------|---------------------|
| 865-72             | 0.67                      | N.S.                     | 0.38                  | 0.88                    | 0.89                 | N.S.                |
| 124(V)             | 0.68                      | N.S.                     | N.S.                  | 0.86                    | 0.89                 | N.S.                |
| 101A(V)            | 0.65                      | N.S.                     | N.S.                  | 0.82                    | 0.91                 | N.S.                |

Table 9A. Correlation table for biotite measurements. All correlations given are significant at the 99.5% significance level. N.S. - no significant linear correlation.

| Specimen N <sup>o</sup> . | Mean T | Mean L | Mean L/T | Mean LxT |
|---------------------------|--------|--------|----------|----------|
| 865-72                    | 0.11   | 0.34   | 3.45     | 0.04     |
| 124(V)                    | 0.09   | 0.28   | 3.58     | 0.03     |
| 101A(V)                   | 0.28   | 1.02   | 4.02     | 0.35     |

Table 9B. Mean values for biotite measurements. 100 measurements per sample. Results are arranged in order of descending metamorphic grade.

ver, shows no consistent trend with grade of metamorphism; the lowest grade specimen has the largest mean size, a result in marked contrast to those for muscovite.

#### Interpretation.

Crystal size. A general correlation between crystal size and metamorphic grade has been known for a long time eg. Leith 1905. Turner and Verhoogen (1960), who discuss the problem in some detail state: "coarse grain is favoured by high temperature of metamorphism and probably also by protraction of the period over which the metamorphic temperatures are maintained, even when the latter are relatively low" (p. 597). Although in the Fernleigh belt muscovite size varies regularly with grade of metamorphism, crystal size is thought not to be a universally suitable indicator of metamorphic grade due to the possibility of protracted growth at low temperatures, as indicated by Turner and Verhoogen, and also due to the possibility of reduction of crystal size during subsequent deformation events. The results for biotite, in which the largest crystals come from the lowest metamorphic grade (table 9B) support this interpretation.

The rate of increase of crystal volume with metamorphic grade, rapid at low grades, slows down in the sillimanite zone, suggesting that an optimum size range has been reached, (fig. 42). Although theoretically crystal size

could increase without limit during metamorphism, this is not generally observed to be the case. There appears to be an upper limit above which single large crystals are replaced by aggregates of smaller crystals of equivalent volume. For muscovite in the study area, the optimum thickness appears to be between 0.2-0.4 mm. (fig. 41).

Crystal shape. It has long been known theoretically that the shape of a crystal may be influenced by the extensive variables of its physical environment, and this has led to the concept of an "equilibrium shape" of a crystal. The subject has been considered by Herring (1953) and others, and the geological application discussed by Kretz (1966a). The equilibrium shape of a crystal is defined as that shape which possesses a minimum sum total of the interfacial free energies\* for a given volume (Curie 1885). Wulff (1901 - in Buckley 1951, p. 107) added a rider to Curie's hypothesis. He showed that the distances from the centre of the crystal to the crystal faces are proportional to the relative rates of growth of the

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\* The interfacial free energy is the difference between the energy of the atoms at the surface and those in the interior of the crystal. It is the work required to form a unit area of surface by a reversible isothermal process, and is a function of the viscosity of the system. Interfacial free energy values are about 1/3 those of surface energy, Spry 1969, p. 25-36.

crystal faces. Then, on the assumption that the rates of growth of each face are proportional to their interfacial energy/surface area ratios, it follows that the distance of a crystal face to the centre of the crystal is proportional to its interfacial free energy. Thus if the equilibrium shape of the crystal is known, it is possible to calculate the relative free energies of the crystal faces (Wulff Theorem).

In the light of the equilibrium shape hypothesis, several interpretations for the data from the Ompah area appear possible. In all the histograms for shape ratio, a spread of shapes is recorded (eg. figs. 35, 36). Kretz (1966a), who obtained a similar distribution from the measurements of 615 phlogopite grains extracted from a marble considered that either (1) only the smallest crystals in the population achieved a near equilibrium shape or (2) that none of the crystals possessed an equilibrium shape, or if any ~~did~~ it was fortuitous.

However another interpretation, which places more importance on the modal peaks of the histograms is also possible. The assumption is made that these represent the equilibrium shape of the crystals, deviations from the equilibrium shape being due to sluggish reaction rates (such relics would have shape ratios greater than the mode) or to other physical factors which will be considered later.

The results for muscovite from the Ompah area suggest

that the third possibility is the more likely in view of the evident significance of the modes of the histograms which vary systematically, both in position and kurtosis, with metamorphic grade (fig. 44). Thus, if the modes of the histograms represent the equilibrium shapes, it is apparent that a greater proportion of crystals achieved equilibrium shapes at high grade than at low; or in other words, that a wider range of shapes is tolerated at lower grades of metamorphism.

Deviations from equilibrium shape are caused by factors which influence the relative interfacial energies of the bounding faces of the crystal. Such factors include :- (a) species and orientation of neighbouring crystals, (b) crystal composition, (c) pressure, (d) adsorption effects and (e) temperature. The influence of these factors in controlling the shape of muscovite crystals in the Fernleigh belt will now be considered.

(a) The effect of the orientation and species of the neighbouring crystal has been discussed by Herring (1953), Kretz (1966a) and Vernon (1968). Kretz noted that the ability to form a crystal face varies significantly both between crystal species (the Becke crystallographic series) and also for different crystallographic orientations of the same species. However in the case of muscovite from the Fernleigh belt, neighbouring crystal species and orientation is not considered to be a significant factor, due to the common ob-

servation that muscovite, regardless of its crystallographic orientation, imposes its form on both quartz and feldspar against which it was generally measured. This situation is therefore equivalent to growth in an isotropic medium (Kretz 1966a).

(b) Composition as a factor affecting crystal shape has been discussed by Swalin (1962) and Kretz (1966a). Kretz notes that the common forms of different species of garnet may be correlated to their composition, and a compositional effect on shape ratio is known in the pyroxene group (Helmstaedt 1976, personal communication). In order to determine the influence of this variable on crystal shape, the compositions of several muscovite crystals of varying shape ratio from each of seven specimens from different metamorphic grades were analysed by means of an electron microprobe. The analyses are included in appendix 2. The results indicate that whilst there are regular changes in muscovite composition with metamorphic grade, there are no systematic changes within a single sample which can be related to crystal shape. It is concluded, therefore, that muscovite composition does not affect the shape of the crystal.

(c) Theoretical considerations indicate that the interfacial energy and hence the equilibrium shape, is pressure dependent - Swalin (1962, p. 180 et seq.). However the results of Jones and Galwey (1972) suggest that mica shape is not pressure sensitive since they report a change in shape

across both regional and thermal metamorphic gradients, which might be expected to have had very different pressure régimes. In the Fernleigh belt no evidence was found to indicate a systematic pressure gradient along the length of the belt, and so it is concluded that pressure differences, if operative, were not dominant in determining muscovite shape.

(d) Adsorption of a different chemical species onto the surface of the crystal may have a significant effect on the interfacial energy, and hence the equilibrium shape. Experimental results show that the surface energy of (001) muscovite in high vacuum is much greater than in air, which is greater again than in water (Herring 1953). Water is generally assumed to be present on grain boundaries in metamorphic rocks as an adsorbed phase, possibly also penetrating the outer layers of the crystal lattice (Laffite 1957, p. 204). With increasing grade of metamorphism an overall decrease in the amount of water in the rock is anticipated. However the amount of water present per unit area of grain boundary may not decrease as rapidly as the total water in the rock, as there is a general trend towards reduction of grain boundary area with increasing metamorphic grade. eg. fig. 42. Unfortunately the role of adsorption in metamorphic processes, whilst understood in principle, is not easy to investigate in the rocks themselves; and in the Fernleigh belt in particular its influence on crystal shape must remain

speculative.

(e) Temperature dependence of surface tension was predicted by Curie (1885), and by the Wulff theorem. It has been discussed more recently by Herring (1953) and Kretz (1966a). Herring constructed a plot of crystal orientation against surface tension, known as the  $\chi$ -plot, fig. 52. In this plot the downward pointing cusp represents a stable low energy face surrounded by two relatively higher energy faces. With increasing temperature the long range order of the atoms on the high energy surface is disrupted (it assumes some of the properties of a liquid), and the surface free energy is reduced. However disruption of the stable low energy faces may fail to take place below the melting point. Thus the relative difference in surface free energies declines, and the equilibrium shape of the crystal will change. Kretz (1966a) notes that the results of experimental work show that for some solids the surface free energy is sensitive to temperature, whilst for others it is practically insensitive. The Wulff theorem (in Buckley 1951) predicts that crystal shape should be constant at a given temperature. In the Fernleigh belt the unimodal histograms for muscovite shape have been interpreted as indicating that an approach to constancy of shape was attained at any given metamorphic grade (a closer approach at high grade than at low). Thus it follows that the shifting position of the mode is due to

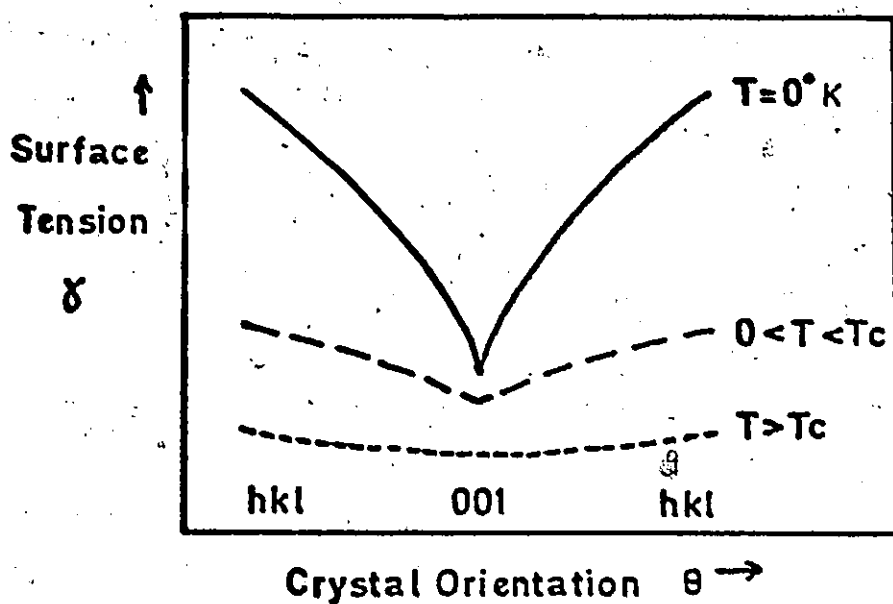


Fig. 52 The  $\gamma$ -plot, after Herring (1953). The surface tension of a crystal face is plotted against crystallographic orientation. The downward pointing cusp represents a relatively low energy crystal face, eg. (001) in micas, surrounded by two higher energy faces, eg. (hkl) in micas.  
 $T_c$  = critical temperature.

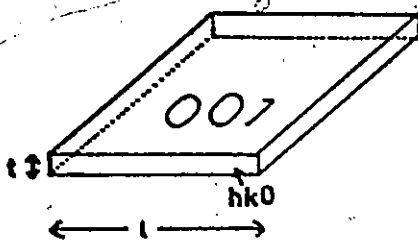
systematic temperature variation along the belt. The spread of the histograms about the modal values may be due to the influence of some or all of the other variables affecting equilibrium shape mentioned earlier.

Utilizing the principles of the Wulff theorem, and following the procedure outlined by Kretz (1966a, p. 84), the total surface free energy ( $F_s$ ) at equilibrium will be a minimum when

$$L/T = f_{hk0}/f_{001}$$

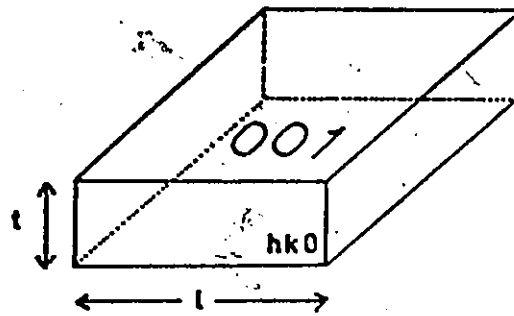
( $f$  is the specific interfacial free energy associated with the  $hk0$  and  $001$  faces of the crystal). There is, therefore, a relative change of interfacial free energies between  $hk0$  and  $001$  from 11:1 at low grade to 3:1 at high grade respectively (c.f. fig. 43). Associated with this there is a change in the surface area proportions of  $hk0:001$ , due to the change in crystal shape. Using rectangular analogues for the shape of muscovite crystals, calculations show that the proportion of the area of the  $hk0$  faces to total surface area increases from about 15% at low grade to 36% at high grade, fig. 53.  $001$  is the stable, low energy crystallographic orientation for micas. Proportional increase in the surface area of the relatively high energy  $hk0$  faces must therefore become progressively more favourable at higher metamorphic grades. Thus the observed changes in muscovite shape with grade of metamorphism fit the theoretical model of Herring

LOW GRADE



$L/T$  RATIO = 11

HIGH GRADE



$L/T$  RATIO = 3

PERCENT SURFACE AREA OF  $hk0$  FACES:

15.4 %

36 %

Fig. 53 Relative dimensions of rectangular analogues of muscovite crystals from low and high grades of metamorphism. The relative percentages of the total surface area contributed by the  $(hk0)$  faces have been calculated.

(1953), as illustrated by the  $\chi$ -plot.

It is therefore concluded that the systematic variation in mica shape along the Fernleigh belt is primarily due to change in temperature. The shape change is explained by postulating that the 001 and hk0 faces of muscovite have different temperature coefficients; and the  $\chi$ -plot suggests that the hk0 faces will be the more temperature sensitive and therefore more responsive to a metamorphic gradient.

The declining surface area to volume ratio with increasing metamorphic grade (fig. 45) is of interest in this context. In isometric crystals, a reduction in the SA/VOL ratio will reduce the total surface free energy for the same crystal volume. However the mica group is not isometric, and as indicated by Herring (1953) interfacial energy differences between crystal faces may be considerable. It is therefore of some surprise to see a similar trend to that predicted for isometric crystals. It appears though, that reduction in the surface area/volume ratio is a significant driving force during textural rearrangement, as can be ascertained from a comparison with clays, the presumed sedimentary parent. An aggregate of clay particles  $10^{-4}$  cm across has a grain boundary area of  $10^5$ - $10^6$  cm<sup>2</sup>/cm<sup>3</sup> of sediment (Spry 1969, p. 23); this is greater by 2-3 orders of magnitude than rocks from the chloritoid-staurolite zone in the study area.

A corollary of the equilibrium shape model is that

crystal shape is not necessarily related to crystal size; ie. that at a given temperature and pressure, a crystal will have a certain shape regardless of its size. This can be tested with the data from the Fernleigh belt. Although along the length of the belt the increase in muscovite size is accompanied by a reduction in shape ratio, the data for biotite (table 9) and for the muscovite inclusions in feldspar porphyroblasts (fig. 47) show that the two trends are not directly interdependent. Detailed analyses of single populations leads to similar conclusions. Correlations between shape  $L/T$  and area  $LxT$  (a measure of size) are generally not significant (table 6, col. 6), whilst those of thickness ( $T$ ) (another measure of size) and shape ( $L/T$ ) are low (table 6, col. 2). As already mentioned, the distributions of points on the graphs of thickness against shape in fig. 37 define areas rather than curves and similar distributions are obtained from samples of 100 measurements from a single rock. Whilst part of the scatter of the distributions is undoubtedly due to the cut effect (see p.152), it is felt that this does not account for all of the observed spread (consideration of the graphs shows that for a plot to define a curve rather than an area on the graph, the thinner crystals would have to have been selectively shortened by the cut effect by a greater amount than the thicker ones). The distributions therefore do reflect real characteristics of the

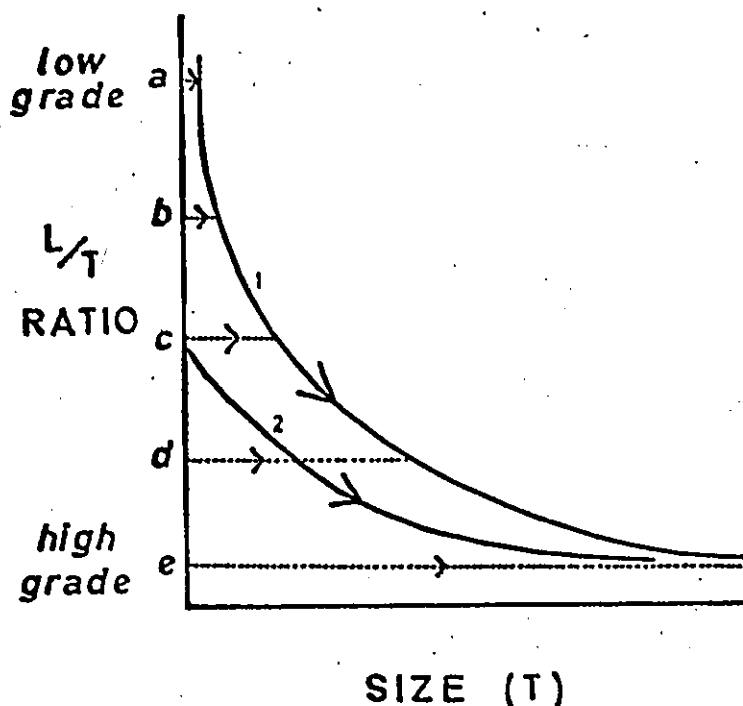


Fig. 54 Schematic diagram of possible growth paths of muscovite crystals at various metamorphic grades. Dotted lines a-e represent growth at constant metamorphic grade. Curves 1 and 2 represent cumulative growth paths after prograde metamorphism (1 - nucleation at low grade, 2 - nucleation at intermediate grade).

population; and indicate that, in the high grade case for example, many small crystals have as low L/T ratios as the large crystals. Thus whilst there is a tendency for the shape ratio to be smaller as crystal size increases, the converse (ie. that smaller crystals have larger shape ratios) is not necessarily true.

It appears therefore that the trends of increasing size and reduction in shape ratio - the trends of prograde metamorphism - are not both followed by individual populations growing at a specific metamorphic grade. Rather, in accord with the equilibrium shape model and the Wulff theorem, a particular shape is attained regardless of size. Possible growth paths at different metamorphic grades are shown in fig. 54. Deviations from modal shapes are ascribed either to the slow rate of metamorphic transformations, or to factors such as adsorption of foreign mineral species onto the crystal surface.

#### Discussion and Comparison with Other Results.

Jones and Galwey (1972) show a similar trend for the variation in shape of biotite with metamorphic grade to those presented here for muscovite and biotite. Possibly a calibrated scale might be set up whereby mica shape ratios could be used to give an assessment of metamorphic grade. However more work is necessary to establish whether shape ratios from different metamorphic terrains can be correlated. It is

apparent from a comparison of the trends in fig. 55 that the range in shape variation for biotite is less than that for muscovite, suggesting that any such attempt must restrict comparison to a single mineral species.

At the present time the main application of the method is that, in the absence of index minerals, it depicts the trend of prograde metamorphism. Since minerals of the mica group are widespread in metamorphic rocks, and the index minerals generally have a more restricted compositional range, there may be terrains where this is the only readily available indicator of metamorphic grade.

Another advantage which is apparent in fig. 43 is that the zones which appear mineralogically homogeneous when mapped as isograds (for example the staurolite-kyanite and the sillimanite zones of the Fernleigh belt) are seen to be regions in which there is continuous change in the shape of muscovite. The muscovite is therefore a more satisfactory indicator of continuous change than the isograds. Also frequently the spacing between the isograds is irregular, and the metamorphic zones vary greatly in areal extent. In these cases trends may be apparent from muscovite shapes which are not evident from isograd maps.

Drawbacks to the use of mica shape are also apparent from fig. 43, in which it can be seen that the mean shape ratio does not change between the upper part of the sillimanite zone and the sillimanite - K. feldspar zone. The 3:1

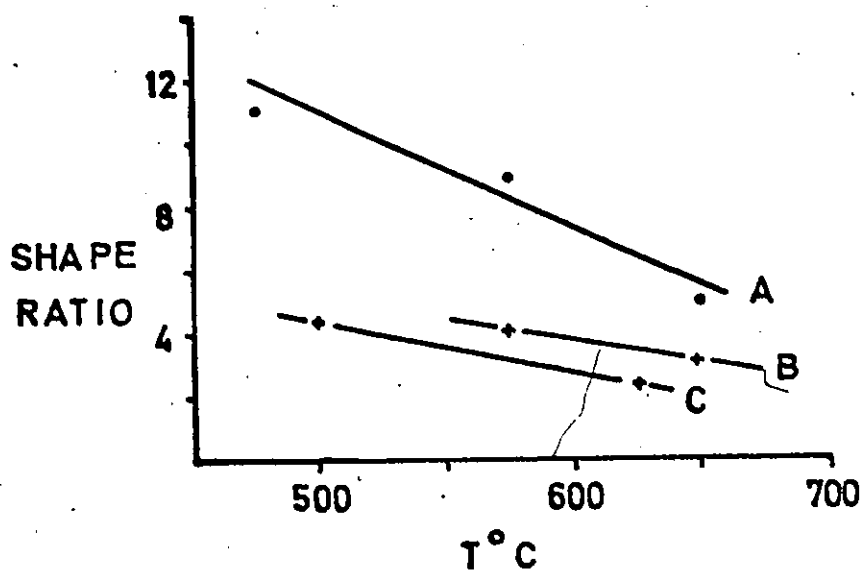


Fig. 55 Tentative calibration between shape ratio of mica and grade of metamorphism. A - muscovite (this study); B - biotite (this study); C - biotite (Jones and Galwey 1972, Barrovian zones of Glen Clova, Scotland). Temperatures are estimated from the metamorphic reactions, and are necessarily approximate; data from Turner 1968.

shape ratio appears to be stable over a range of temperature. A similar problem exists at the low grade end of the belt, where the 11:1 modal shape ratio also spans an isograd. Thus the change in shape is only apparent over a range corresponding approximately to the amphibolite facies.

Jones and Galwey (1972) offered a different interpretation concerning the growth paths of crystals to that suggested here. From their data they concluded that the growth path followed by crystals nucleating at high grade was the same as that for those which nucleated at low grade - ie. all crystals followed what has been termed the prograde growth path, from small tabulate crystals to larger, relatively equidimensional crystals. This led them to consider kinetic controls on crystal growth; more specifically that the rate of advance of different crystal faces was not the same, so that the shape change could occur at a variety of temperatures. This differs from the interpretation put forward in this work, in which the rates of advance of different crystal faces, whilst not the same, remain in constant proportions at a given grade of metamorphism. The change in crystal shape is thus related to metamorphic grade, and can be expressed as the temperature coefficient of growth of the crystal face.

#### Nucleation and Growth

The kinetics of nucleation and growth of new mineral

species in a rock has been discussed by Galwey and Jones (1963), Jones and Galwey (1964, 1966) and Kretz (1966b, 1973 and 1974). In these papers several possible models are suggested, and these can be tested with the muscovite data. Kretz (1974) put forward three models for crystal growth and four for nucleation. The growth models are: (a) constant rate of linear growth, (b) constant rate of increment of area and (c) constant rate of increment of volume per crystal. A method for distinguishing between these models is described for which it is necessary to gather information on the chemical profiles of compositionally zoned crystals of different sizes. This has not been done in the present study, so one of the growth models has been arbitrarily chosen. A constant rate of increment of area of muscovite crystals along the metamorphic belt is not considered probable in the present study since the surface area/volume ratio (fig. 45) and the  $hk0/001$  ratio (fig. 53) both change with grade of metamorphism. Kretz's (1974) data show that the growth rate for garnet in a small specimen of schist from Yellowknife was somewhere between rates (a) and (b) above. For this study a constant rate of linear growth (with respect to the T dimension) will be assumed; and this will be used to derive information about the nucleation rate of muscovite in the pelitic schist.

Kretz (1966b) points out that if an assumption of constant linear growth is made, the time taken for growth is

proportional to crystal size. Thus with a time scale based on a dimension (in this case garnet diameter) of the largest crystal in his population of measurements, the rate of production of new crystals, the nucleation rate, can be expressed graphically in terms of crystal size, and if necessary the equations to fit the curve may be derived. (It is of interest to note that the curve obtained is actually an upside-down version of the cumulative histograms of the crystal dimension). The curves obtained by Kretz (1966b and 1973) and by Jones and Galwey (1966) indicate that nucleation began slowly, increased exponentially and then declined in a similar manner. It is the aim of this section of the study to compare the characteristics of populations of measurements from different grades of metamorphism in an attempt to determine the effect of a metamorphic gradient on the nucleation rate.

In discussions of nucleation and growth of crystals, models are constrained to small volumes of rock - thus the moving average technique is not applicable. The five sets of measurements used in this study were all obtained from thin sections of small volumes of rock from different metamorphic grades (fig. 38). Furthermore previous work on this subject (cited above) was concerned with crystals physically isolated from each other in the rock eg. garnet crystals in a schist or gneiss, phlogopite crystals in marble etc. As a

result the theories invoked to explain the results have assumed that nucleation was random (that is that the position of a nucleus was not affected by its neighbours) and that there was no interaction between crystals during growth. These constraints may not be applicable to the present study, where interaction between crystals during growth is to be expected.

#### Nucleation Models

Two nucleation models will be considered assuming a constant rate of linear growth of individuals.

Model A. Using the data in fig. 51, and following the reasoning of Kretz (1966a), a time scale has been set up in terms of growth of the largest crystal (which in this case occurs in the sample from the highest metamorphic grade). Assuming that the linear rate of growth is constant at all grades, the rate of production of other, smaller crystals (the nucleation rate) is proportional to crystal size (fig. 56). It is apparent from this model, in which size (thickness) reflects the time elapsed, that the duration of metamorphism at low grade was a fraction of that at high grade.

Model B. Utilizing the same data, the duration of metamorphism is considered to have been the same for all metamorphic grades, but the growth rates differed. The rate of crystal growth at a single metamorphic grade remained constant, but varied with grade of metamorphism. In this model the

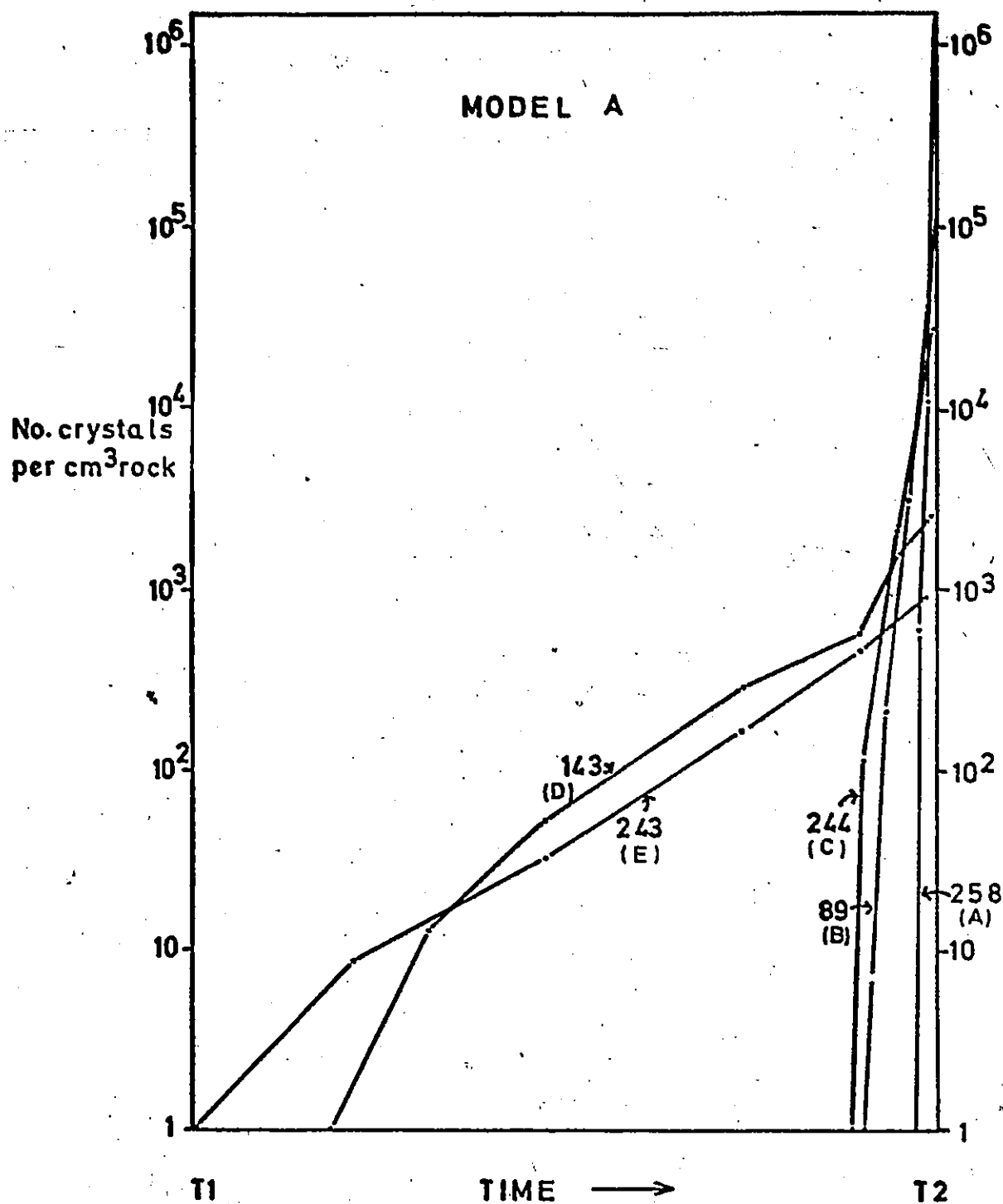


Fig. 56. Nucleation model assuming a constant rate of linear growth at all metamorphic grades. The time scale is set up with respect to the largest crystal present, and the duration of crystal growth is proportional to size. Largest crystal starts growing at T1, smallest crystal stops growing at T2. A - low grade, E - high grade. Numbers refer to sample locality, see fig. 19.

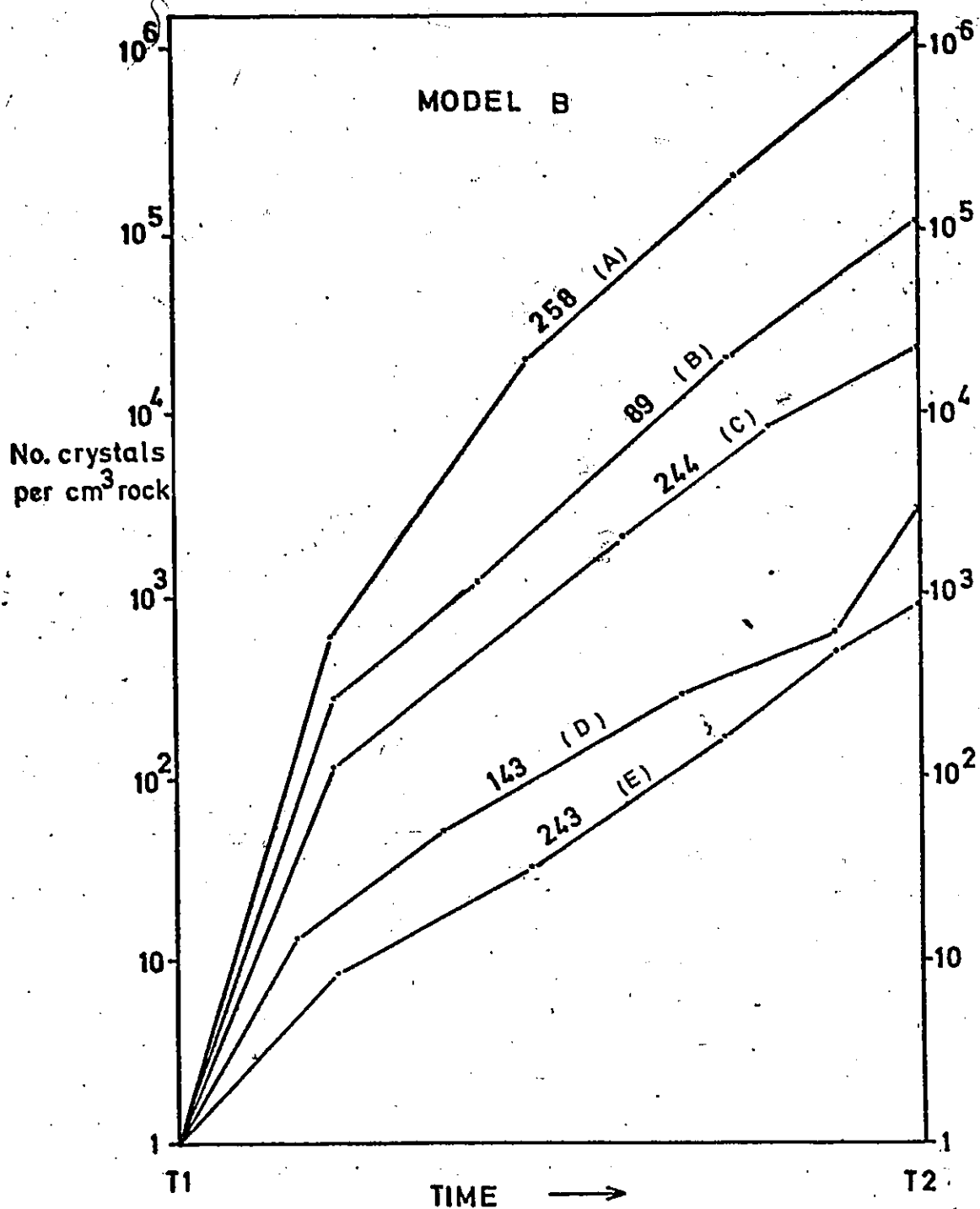


Fig. 57 Nucleation model assuming that the duration of metamorphism is the same at all metamorphic grades. The rate of linear growth within a single sample is constant, but the rates between samples are different. A - low grade, E - high grade.

T1 - largest crystal in each sample starts to grow.  
T2 - smallest crystal in each sample stops growing.

disparity in the rates of nucleation between the high and low grade specimens is reduced.

#### Discussion.

Models A and B represent only two end members of a spectrum of possibilities, and there are reasons to believe that neither represents the actual history of the rocks. Indirect evidence suggests that the linear growth rates (increase of the T dimension) will be different at different metamorphic grades. If an assumption is made that the rate of recrystallization (per volume of muscovite recrystallised) is the same at different grades of metamorphism, it becomes apparent that the average crystal at high grade, with a low L/T ratio, will increase its T dimension relatively faster than the average crystal at low grade (high L/T ratio), due to the different crystal shapes, and irrespective of size. Thus it is probable that model A is not directly appropriate. Geological considerations suggest that model B also is not a probable analogue since the duration of metamorphism at high grades was probably longer than that at low grades. Thus an intermediate model involving different durations of metamorphism and varying rates of crystal growth at different grades appears likely.

However neither model explains how the large number of small crystals at low grade is transformed into the much smaller number of larger crystals at high grade, assuming,

as the textural evidence suggests, that metamorphism was progressive, and that such a transformation has actually occurred. The original size distributions of muscovite were probably determined early in the metamorphic history of the rock, perhaps during zeolite facies conditions (Winkler 1967). However during subsequent prograde metamorphism, it is likely that muscovite was involved in a variety of reactions (see Metamorphism). The original size distributions determined by the relative rates of nucleation and growth may have been considerably altered during these reactions, particularly if small crystals were preferentially selected for reactions in which muscovite was involved (because of their higher surface energy/volume ratio). Nucleation of the muscovite later in the reactions may have been facilitated on existing muscovite crystals, thus gradually reducing the numbers and increasing the size of the remaining crystals. Small changes in muscovite composition occurred during this process (Appendix 2).

Another process which almost certainly affected the crystal size distribution is deformation. Metamorphism in the Fernleigh belt is considered to be syntectonic (see Time Relations), so deformation of the crystal lattices would have occurred during metamorphism. It is possible that the strained crystals may also have been preferentially selected for reaction, because their higher strain energy would lower reaction thresholds. Deformation, however, is generally con-

sidered to reduce crystal size, so it must be coupled with a mechanism by which growth is concentrated at progressively fewer sites. Annealing or annealing recrystallization would provide such a mechanism. Spry (1969, p. 71) states "annealing and similar recrystallization is driven first by the strain energies in the original strained aggregate, and then by the interfacial surface energies..... it involves the movement of grain boundaries such that some crystals increase in size at the expense of others". The processes of annealing in metals have been given extensive study eg. Burke and Turnbull 1952, and the behaviour of carbonates (Griggs et alia, 1960) and silicates (Vernon 1968) are thought to be similar. The annealing temperature is about half the melting point in degrees Kelvin (Spry 1969, p. 116), conditions certainly fulfilled during amphibolite grade metamorphism of a pelitic rock.

Recently mica deformation and subsequent recrystallization have been examined experimentally in some detail (Etheridge, Hobbs and Paterson 1973, Etheridge and Hobbs 1973). In the latter paper the results of electron microprobe analyses of mica crystals before and after recrystallization were reported. These showed small but consistent changes in minor element concentrations. This led the authors to state; "strain energy contributes to the process (recrystallization), but another energy source is necessary. This energy source can be found in the changes in the chemical composition which accompany recrystallization".

accompany recrystallization" (Etheridge and Hobbs 1973, p. 121). Since these changes involve transfer of elements into or out of the system, and since there is independent evidence that such transfer may have taken place during prograde metamorphic reactions (see Metamorphism), such a hypothesis seems likely. Thus it appears probable that the original size distributions of muscovite have been considerably modified by recrystallization, the energy for which was derived primarily from strain energy, surface energy and the free energy of reaction.

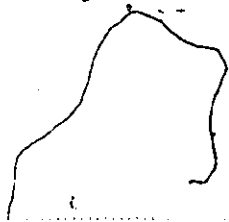
#### Crystal Orientation

The development of a preferred orientation of phyllosilicate minerals in slates and schists has been the subject of debate for over a century, see Siddans 1972 and Wood 1974 for reviews of the literature. The debate has centred around two principal issues: (a) the relation of the principal axes of stress and strain to the plane of schistosity of cleavage, and (b) the mechanism by which the phyllosilicate minerals become preferentially orientated in that plane. In recent years (a) appears to have been satisfactorily answered, and it is generally agreed that the plane of cleavage or schistosity develops normal to the direction of the maximum principal compressive stress ( $\sigma_1$ ), and that it is normal to the minimum principal strain axis ( $\epsilon_3$ ) - and that it is a plane of flattening. The second question, however, is still contro-

versial. Evidence for the rotation of pre-existing grains during "dewatering" of the sediment has been given by Maxwell (1962), Clarke (1970a, 1970b), Powell (1972) and Alterman (1973); rotation of mica grains may also take place at low metamorphic grades, facilitated by pressure solution of quartz, Williams (1972). The field evidence of Groschong (1972) and Carrara (1972), and the experimental evidence of Means and Williams (1974) support this as a feasible mechanism.

Recrystallization as the mechanism giving rise to a preferred orientation (by which nucleation and growth is restricted to favourably orientated crystals, those unfavourably orientated being eliminated) has been considered by Kamb (1959), Brace (1960) and recently by Wood (1974), Etheridge et al. (1974), Etheridge and Hobbs (1973) and Etheridge and Lee (1975). It appears probable that the final answer will be a combination of several processes. For instance Helm and Siddans (1971 p. 258) state: "a more realistic model of slaty cleavage must allow for solution, diffusion, nucleation and growth of minerals in a primary anisotropic fabric subject to successive non-coaxial strain increments, together with the rotation of existing grains and the effects of mimetic recrystallization". Moench (1966), for example, favours a combination of rotation during dewatering with subsequent mimetic recrystallization.

It is the object of this section of the study to try to



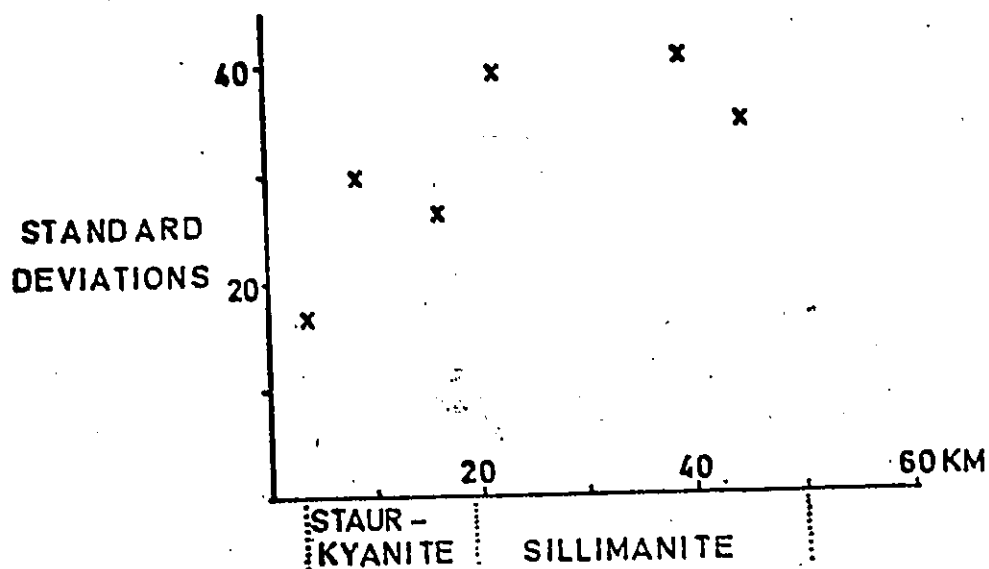


Fig. 58 Variation in the degree of preferred orientation of muscovite (001) with metamorphic grade. The degree of preferred orientation for six specimens (for each  $N=200$ ) is estimated by the Kamb method (1959), and is expressed as the maximum difference, in standard deviations, of the population distributions from the distribution produced by random sampling of a population with no preferred orientation.

distinguish between the models presented briefly above, and to decide which of them is applicable to the schistosity as developed in the pelitic schist unit of the Fernleigh belt. The orientation of muscovite will be considered in relation to metamorphic grade, and also to the shape and size of the crystals themselves.

Results:

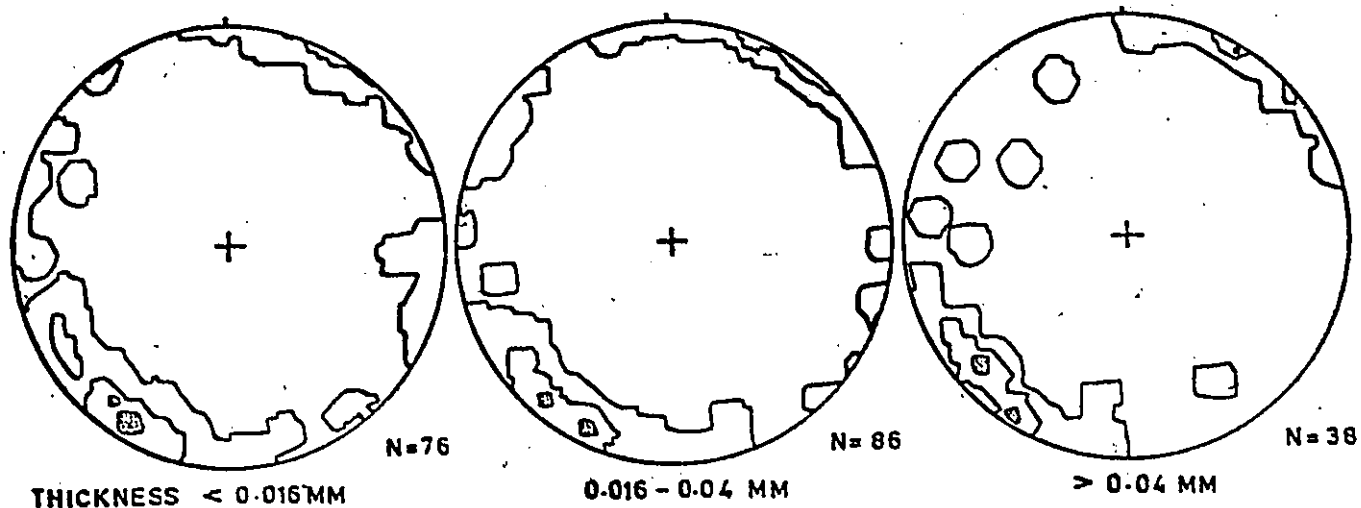
Orientation and metamorphic grade. In the six samples measured there is a clear trend of increasing preferred orientation from the staurolite-kyanite to the sillimanite zones (fig. 58). Within the sillimanite zone the degree of preferred orientation may remain approximately constant. It is interesting to note that the maxima of the Schmidt contouring method do not show a similar trend, but rather a scatter (data not included).

Orientation and crystal size. No correlation between thickness (T) and orientation was found in either of the samples (fig. 59), or between length and orientation in the specimen from the sillimanite zone (fig. 60B). However in the specimen from the staurolite-kyanite zone the longer crystals are less scattered in orientation than the shorter ones (fig. 60A).

Orientation and crystal shape. A weak correlation between muscovite orientation and shape ratio is seen (fig. 61), in that the crystals with the high L/T ratios tend to have a

A. SPEC. N°: 2628

STAUROLITE-KYANITE ZONE



B. SPEC. 112L

SILLIMANITE ZONE

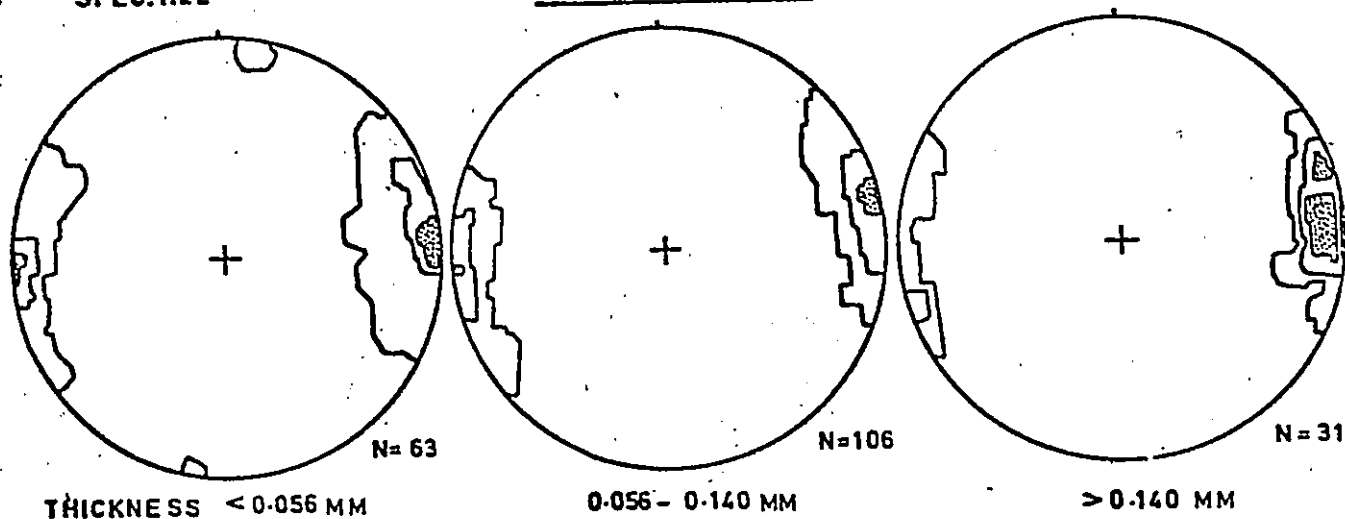
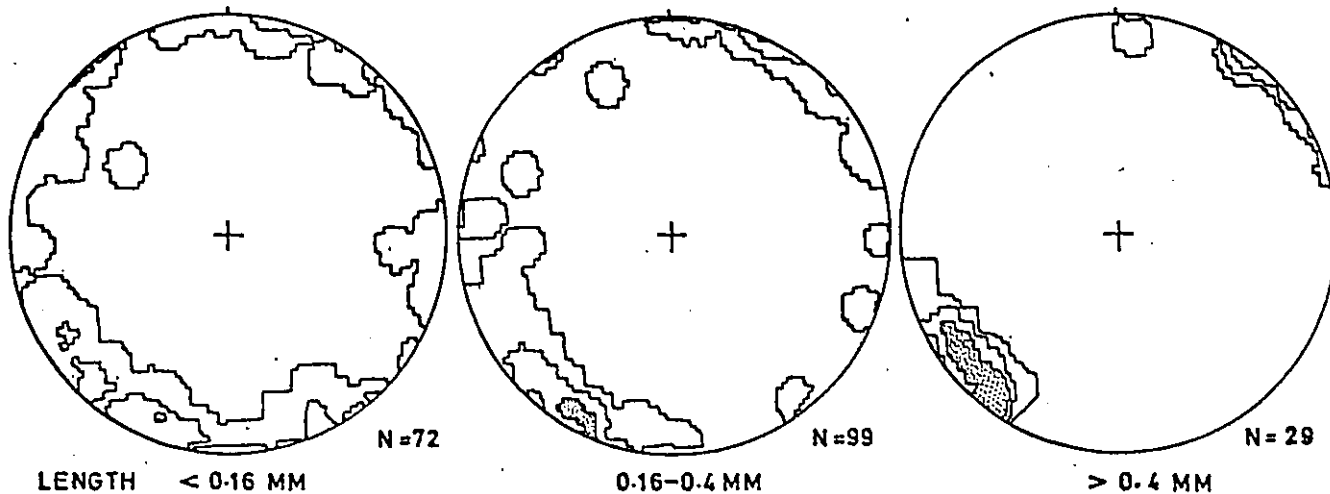


Fig. 59 The orientation of muscovite crystals in two samples (for each N = 200) from different metamorphic grades, subdivided on the basis of thickness (T) into three groups each. (Lower hemisphere equal area projection, contours by the Schmidt method; contour interval 1%, 10% and 20% per 1% area).

A. SPEC. NO: 262B

STAUROLITE-KYANITE ZONE



B. SPEC. NO: 112L

SILLIMANITE ZONE

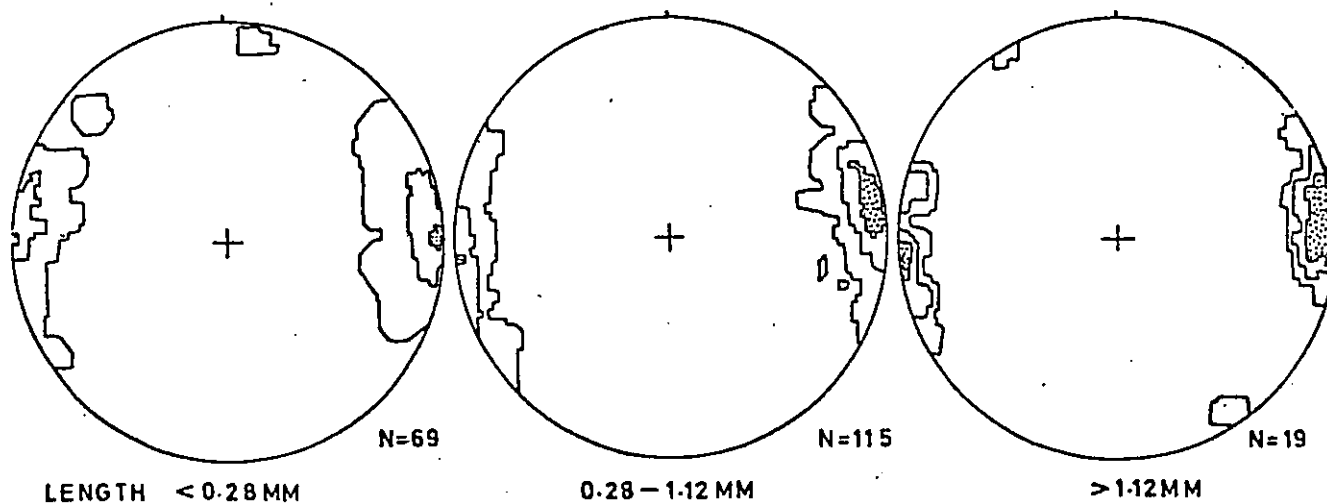
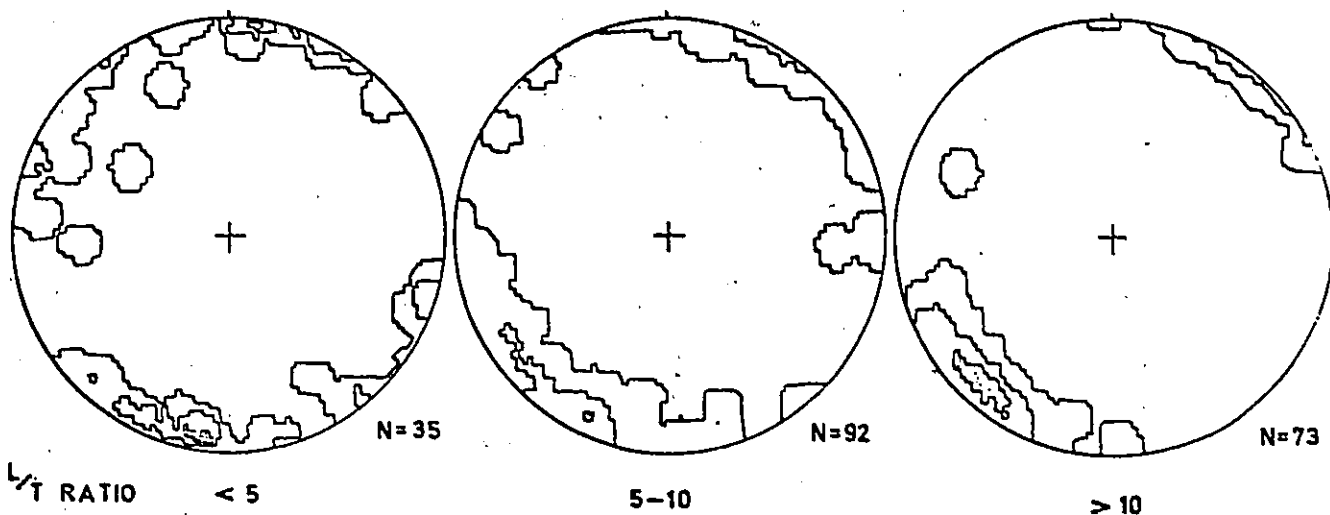


Fig. 60 The orientation of muscovite crystals in two samples (for each N=200) from different metamorphic grades, subdivided on the basis of length (L) into 3 groups each. (Lower hemisphere equal area projection, contours by the Schmidt method; contour interval 1%, 10% and 20% per 1% area).

A. SPEC. NO: 262 B

STAUROLITE KYANITE ZONE



B. SPEC. NO: 112 L

SILLIMANITE ZONE

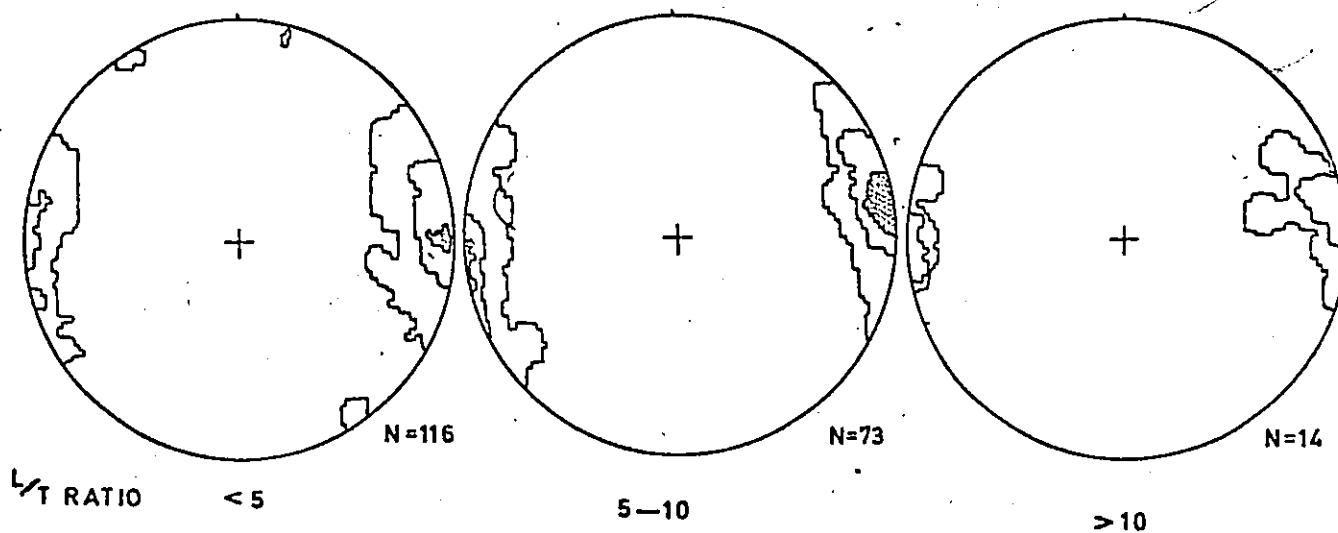


Fig. 61 The orientation of muscovite crystals in two samples (for each  $N = 200$ ) from different metamorphic grades, subdivided on the basis of shape (L/T) into 3 groups each. (Lower hemisphere equal area projection, contours by the Schmidt method, contour interval 1%, 10% and 20% per 1% area).

less scattered orientation than those with low ratios. The effect is more marked in the specimen from the staurolite-kyanite zone. However the small number of measurements with an L/T ratio of greater than 10 from the sillimanite zone specimen renders the results inconclusive in this sample.

#### Discussion.

The higher degree of preferred orientation with increasing grade of metamorphism (fig. 58) was not anticipated. Qualitative assessments in the past have tended to suggest that the opposite is the case; eg. "It is a matter of common observation that there is some tendency for the degree of preferred orientation to lessen with increasing grade", (Jones and Galwey 1973, p. 237). Whether the trend of fig. 58 is generally valid must await measurements from other areas. The previous common observations have been qualitative assessments of average orientations that are inherently inaccurate for comparative purposes unless in each case the field of view contains approximately the same number of crystals. In this case, where the mean area  $L \times T$  per crystal at the high grade end of the belt is 150x that at the low grade end, the field of view should vary accordingly. It is therefore a problem of scale, and previous assessments may not be meaningful.

This result, when considered in combination with the related trends of increase in size and reduction in numbers

of crystals (figs. 41 and 50A respectively) is considered to support a recrystallization hypothesis for the development of schistosity in the Fernleigh belt. It is apparent that at higher grades, as the total number of crystals decreases, the number orientated at high angles to the schistosity decreases by a proportionately greater amount. The most stable (minimum energy) orientation for muscovite is with (001) perpendicular to  $\sigma_1$ , the greatest principal stress direction, normal to the foliation. The stable position has therefore been approached to a greater degree in the high grade rocks. Evidence that recrystallization was a continuous process during metamorphism has been given from arguments based both on chemical grounds, and on the results of the crystal size distribution. It therefore seems most probable that the elimination of unfavourably orientated crystals was by a process of recrystallization. The conclusion of this process is expected to be more nearly attained at high metamorphic grades than at low because of the higher temperature and probably longer time available for recrystallization.

The implications of this result suggest that the conclusions of Tullis and Wood (1975) may be somewhat precipitate. In several specimens of slate from the Cambrian slate belt of Wales, they calculated the strain from the shape of deformed reduction bodies and correlated this to the degree of preferred orientation of the constituent muscovite crystals

(measured by x-ray goniometry). They found that "there is a one-to-one relationship between the shape of the strain ellipsoid and the preferred orientation predicted by the March model" (March 1932 has shown mathematically how the degree of preferred orientation by rotation of pre-existing grains can be correlated with strain). Thus Tullis and Wood considered that the degree of preferred orientation could be used as a measure of strain. However, as apparent from the structural configuration (fig. 3), <sup>and strain measurements (fig 7)</sup> there is no evidence to suggest that strain varied systematically along the length of the Fernleigh belt, so variation in strain is not considered a probable cause of the variation in preferred orientation. Rather the degree of preferred orientation is in part determined by the grade of metamorphism. Hence use of the degree of preferred orientation to predict strain may only be applicable to rocks of the same metamorphic grade.

The correlations between length and orientation and shape and orientation at intermediate metamorphic grade (staurolite-kyanite zone) (figs. 60A and 61A) show that crystals orientated at high angles to the foliation ~~are~~ generally those with a shorter more equidimensional form. Oertel (1970) has shown how during post-tectonic growth mica crystals orientated parallel to a plane of schistosity may grow longer than those at high angles to that plane, as impingement of growth at crystal boundaries stops more crystals of the latter than the former orientation (fig. 62). If this principle is

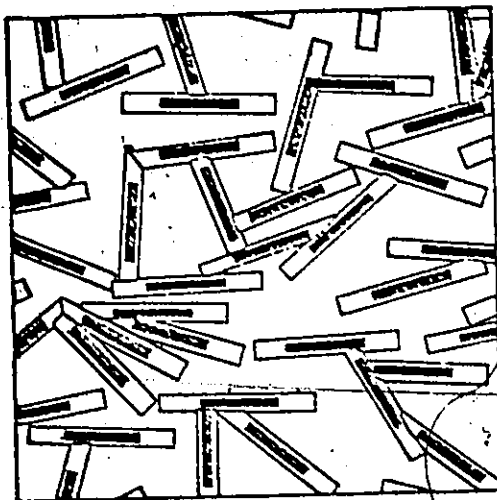


Fig. 62 (Taken from Oertel 1970). Enhancement of a preferred orientation by post-tectonic growth of mica crystals. Central black portion of crystals, which have a preferred orientation, have an L/T ratio of 8. During growth to twice the original size (outlined), impingement stops more crystals that deviate strongly from the plane of preferred orientation than crystals that are nearly parallel to it, thus the degree of preferred orientation is enhanced.

extended to syntectonic growth, the effect may be more marked as crystals orientated at high angles to the foliation will be preferentially strained. The accumulation of stress and strain energy in these crystals is an energy source for recrystallization, during which process strained crystals will either assume a more favourable orientation parallel to the schistosity or be eliminated. This process must have occurred in the high grade specimen, where there are fewer crystals at high angles to the schistosity (figs. 60B and 61B).

Etheridge (1972) has pointed out that many schistose rocks are inhomogeneous, consisting of quartz-rich and mica-rich domains. From linked measurements of mica (biotite) shape and orientation from each of the two domains, he showed that crystals in the mica-rich domains tend to have a stronger preferred orientation and a higher shape ratio (they are more elongate) than those from the quartz-rich domains. Etheridge therefore concluded that lithology and orientation are significant factors determining mica shape. Higher shape ratios in the mica-rich domains were also observed in the present study; however in this case measurements were confined to a single lithology (ie. quartz-rich domains), and although the more elongate crystals have a stronger preferred orientation within the staurolite-kyanite zone, this relationship has not been demonstrated within the sillimanite zone (fig. 60). This observation, together with the association of the low

L/T ratio crystals and better developed preferred orientation at high grade indicate that mica shape was not primarily determined by orientation in the Fernleigh belt.

The development of the schistosity in the Fernleigh belt is therefore considered to be primarily a result of recrystallization. However in this study all rocks examined possessed a well developed metamorphic fabric, and whether or not the initial cleavage formation at low grade was due to rotation, eg. Moench (1966) is not clear. Recent work of Oertel and Phakey (1972) and Etheridge and Lee (1975) suggests that (re)crystallization can be a significant mechanism even at low grades of metamorphism. The recrystallization mechanisms of Etheridge et al. (1974) and Etheridge and Lee (1975) are considered compatible with the present results if combined with a mechanism by which unfavourably orientated crystals are eliminated by recrystallization, as outlined above.

#### Summary

Muscovite crystals exhibit the following textural changes along the Fernleigh belt, which are considered to be a result of the metamorphic gradient:

- (1) The crystal size increases, and the surface area/volume ratio decreases with increasing metamorphic grade.
- (2) The crystal shape ratio (L/T) decreases, and the form of the histogram distribution becomes more

leptokurtic (less dispersed) with increasing metamorphic grade. This is interpreted as indicating that proportionately more crystals achieved the equilibrium shape at high grade than at low.

- (3) The crystal growth paths at different metamorphic grades are not the same; this is a function of the changing equilibrium shape with grade of metamorphism (essentially temperature).
- (4) The distributions of crystal sizes retain a similar form along the length of the metamorphic belt. However the number of crystals per unit volume of rock is reduced by about 3 orders of magnitude at high grade.
- (5) The preferred orientation of muscovite crystals increases with increasing grade of metamorphism.
- (6) In medium grade rocks crystals that are longer and with higher shape ratios tend to have a more preferred orientation. At high grades fewer crystals at high angles to the foliation are present.

Taken individually it has been argued that these changes have taken place in order to reduce the free energy of the crystal towards a minimum for a given metamorphic environment. Taken together, and also considering the small, but systematic changes in minor element composition of muscovite (Appendix 2), the total free energy of the rock is reduced,

and a minimum free energy condition of textural equilibrium is approached. Combining the results listed above indicates that a closer approach to textural equilibrium has been attained in the high grade rocks, where processes of annealing and recrystallization were more effective than at low grade.

## CONCLUSIONS

In the Ompah area plutonic bodies surround a northeast trending synform of metasedimentary and metavolcanic rocks. These rocks have undergone three phases of deformation which may be of regional significance. First phase folds, which are of variable scale and are associated with a penetrative axial foliation and lineation, are northeast trending, tight to isoclinal structures. Profiles of  $F_1$  folds are flattened, and there is typically maximum extension parallel to  $F_1$  fold axes. The attitude of later  $F_2$  folds suggests that  $F_1$  and  $S_1$  were originally recumbent. Similar structures have been reported by Divi (1972), (area C, fig. 2) and Carmichael (1968), (area D, fig. 2) west and southwest of the Ompah area, where they are also considered to have been originally recumbent.

Later  $F_2$  folds, also variable in scale, are low plunging, upright structures with northeast trending axes, and include the large, map-scale Plevna antiform and synform. They are less flattened than  $F_1$  folds, and the associated axial lineations and foliations are not well developed in the Ompah area.  $F_2$  folds of similar orientation and style, but limited to outcrop scale, are described by Thompson (1972) (area A, fig. 2) 40 km. west of Ompah, but further west  $D_2$  structures, though of similar attitude, are more penetrative and obliterate most signs of  $D_1$  deformation (Divi 1972):

Sporadically developed  $F_3$  folds in the Ompah area are open crenulations and warps about northeast trending, low-dipping axial planes; axial lineations and foliations have no mineralogical expression. Similar structures are reported by Divi and Carmichael west and southwest of Ompah.

The shape and size of inclusions of muscovite and other minerals in porphyroblasts of plagioclase feldspar suggests that  $D_1$  deformation began at low grade metamorphism in the Ompah area, and continued till the metamorphic climax.  $D_2$  deformation occurred at or soon after the metamorphic peak, and  $D_3$  is considered to have been a later, retrograde phenomenon. Similar relative sequences are reported by Carmichael (1968), Divi (1972), and Thompson (1972) from west of the study area.

Plutonism in the Ompah area predated regional metamorphism and deformation, a pattern considered typical of the region by Lumbers (1967b).

Metamorphism reached amphibolite facies grade throughout the area. The grade rises from southwest, and the area has been subdivided into three metamorphic zones characterised, in pelitic assemblages, by the index minerals staurolite and kyanite, sillimanite and sillimanite - K. feldspar.

Textures of mineral assemblages in pelitic rocks suggest that the mechanism of prograde metamorphic reactions may frequently involve phases present on both sides of the isograd

defining the metamorphic reaction. Ionic transfer of elements and neutral solutions between reacting phases may permit the breakdown of muscovite in the presence of quartz at a lower temperature than the thermal stability limit marked by the first appearance of the pair sillimanite - K. feldspar. Staurolite stability and the kyanite to sillimanite transformation may also be ionically controlled. Furthermore, the discovery of the pair sillimanite - K. feldspar at a single locality only suggests that the appearance of K. feldspar may be influenced by the ionic concentration of  $K^+$  in the fluid phase. The concentration of elements in the fluid phase may be controlled by in situ recrystallization of matrix phases such as muscovite, the rate of which is determined by deformation. Support for these as feasible mechanisms for prograde reactions is found in the work of Eugster (1970), Carmichael (1968, 1970), Wintsch (1975b) and others.

Textures of muscovite crystals in a pelitic schist show several systematic trends with grade of metamorphism. Crystals become on average larger and more equant with increasing metamorphic grade. However crystal size and shape are not dependent variables, as crystal shape is primarily determined by the temperature of metamorphism, which controls the equilibrium shape. Preferred orientation of muscovite crystals is enhanced with increasing grade of metamorphism, as crystals

orientated at high angles to the schistosity are progressively eliminated. Thus the development of a schistosity may be influenced by grade of metamorphism. Recrystallization processes have allowed a closer approach to the minimum energy condition of textural equilibrium in high grade than in low grade rocks.

#### Suggestions for Future Work

Detailed investigation of the area east of Clyde Forks, including the gabbroic pluton in Dalhousie and Lanark townships, would shed additional light on the time of emplacement of the pluton and the tectonic history of this part of the Grenville Province. The Robertson Lake shear zone (Smith 1958) is located at the western margin of this body and may be related either to pluton emplacement or to later deformation.

Mapping northeast of Clyde Forks will establish whether the Fernleigh belt extends further in that direction, and also whether the metamorphic grade continues to rise north-eastwards. The location of assemblages containing the mineral pair sillimanite - K. feldspar will be of particular interest in this regard.

Strain measurements from several areas within the Bancroft-Madoc region indicate that flattening strain with maximum extension parallel to  $F_1$  fold axes is characteristic of this region of the Grenville Province. With the recent

publication (Wood 1974) of possible associations between tectonic environments and the type and amount of strain, it would be of interest to establish statistically a general strain pattern for the region.

Finally, and on a more general level, further work evaluating the variation of crystal textures with metamorphic grade will be useful. Such techniques may yield valuable information on the mechanism of recrystallization and grain growth in metamorphic rocks.

LIST OF REFERENCES

(Includes those articles referred to in Appendices I and II).

- Alterman, I.B., 1973, Rotation and dewatering during slaty cleavage formation: some new evidence and interpretations: *Geology*, v. 1, p. 33 - 36.
- Appleyard, E.C., 1974, Basement/Cover relationships within the Grenville Province of eastern Ontario: *Can. Jour. Earth. Sci.*, v. 11, p. 369 - 379.
- Ashworth, J.R., 1975, The sillimanite zones of the Huntly-Portsoy area in the northeast Dalradians, Scotland: *Geol. Mag.*, v. 112, p. 113 - 136.
- Atherton, M.P., 1975, Crystal growth models in metamorphic tectonites: *Progr. of Abstracts, Roy. Soc. Meeting, "Natural Strain and Geological Structure"*, p. 17.
- Bell, T.H. and Etheridge, M.A., 1973, Microstructure of mylonites and their descriptive terminology: *Lithos*, v. 6, p. 337 - 48.
- Bence, A.E. and Albee, A.L., 1968, Empirical correction factors for the electron microanalysis of silicates and oxides: *Jour. Geol.*, v. 76, p. 382 - 403.
- Brace, W.F., 1960, Orientation of anisotropic minerals in a stress field: discussion: *Geol. Soc. Amer. Mem.*, v. 79, p. 9 - 20.
- Buckley, H.E., 1951, *Crystal Growth*, Wiley and Sons, New York.
- Burke, J.E. and Turnbull, D., 1952, Recrystallization and grain growth: *Progr. Metal. Phys.*, v. 3, p. 220 - 242.
- Butler, B.C.M., 1967, *Chemical Study of minerals from the Moine*

schists of the Ardnamurchan area, Argyllshire, Scotland:

Jour. of Pet., v. 8, p. 233 - 67.

Carmichael, D.M., 1968, Structure and progressive metamorphism in the Whetstone Lake area, Ontario, with emphasis on the mechanisms of prograde metamorphic reactions: Ph.D. thesis, Univ. of California, Berkeley.

\_\_\_\_\_, 1969, On the mechanism of prograde metamorphic reactions in quartz-bearing pelitic rocks: Contrib. to Min. and Petrol., v. 20, p. 244 - 67.

\_\_\_\_\_, 1970, Intersecting isograds in the Whetstone Lake area, Ontario: Jour. Petrol., v. 11, p. 147 - 181.

Carrara, A., 1972, Structural geology of Lower Paleozoic rocks, Mt. Albert area, Gaspé peninsula, Québec: Ph.D. thesis, Univ. of Ottawa.

Chapman, D., 1968, Petrology, structure and metamorphism of a concordant granodiorite gneiss in the Grenville Province of southeastern Ontario: Ph.D. thesis, Rutgers Univ. New Jersey.

Chinner, G.A., 1961, The Origin of sillimanite in Glen Clova, Angus: Jour. Petrol., v. 2, p. 312 - 323.

\_\_\_\_\_, 1965, The kyanite isograd in Glen Clova, Angus, Scotland: Mineral. Mag., v. 34, p. 132 - 143.

Christie, J.M., 1960, Mylonitic rocks of the Moine Thrust zone in the Assynt district, northwest Scotland: Trans. Geol. Soc. Edin., v. 18, p. 79 - 117.

Clark, B.R., 1970 (a), Origin of Slaty cleavage in the Coeur d'Alène district, Idaho: Geol. Soc. Amer. Bull., v. 81, p. 3061 - 72.

- Clark, B.R., 1970 (b), Mechanical formation of preferred orientation in clays: Amer. Jour. Sci., v. 269, p. 250 - 66.
- Clifford, P.M. and Rector, R.J., 1970, Fabric and compositional effects on strain in rocks: Progr. of abstracts, south-central 4<sup>th</sup> Ann. Meetg., Geol. Soc. Amer., v. 2, p. 277.
- Cloos, E., 1947, Oolite deformation in South Mountain fold, Maryland: Geol. Soc. Amer. Bull., v. 58, p. 843 - 917.
- Curie, M.P., 1885, Sur la formation des cristaux et sur les constants capillaires de leurs différentes faces: Bull. Soc. franç. Minéral., v. 8, p. 145 - 50.
- Deer, W.A., Howie, R.A. and Zussman, J., 1969, An Introduction to the Rock-forming Minerals: Longmans.
- DeVore, G.W., 1956, Surface chemistry as a chemical control of mineral association: Jour. Geol., v. 64, p. 31 - 55.
- , 1959, Role of minimum interfacial free energy in determining the macroscopic features of mineral assemblages I - the model: Jour. Geol., v. 67, p. 211 - 227.
- Divi, S.R.R., 1972, Structural analysis of Grenville rocks near Bancroft, Ontario, Canada: Ph.D. thesis, Univ. of Ottawa.
- and Fyson, W.K., <sup>1973</sup><sub>A</sub> Folds and strain in Grenville metamorphic rocks, Bancroft, Ontario, Canada: Geol. Soc. Amer. Bull., v. 84, p. 1607 - 18.
- Ehrlich, R., Vogel, T.A., Weinberg, B., Kamilli, D.C., Byerly, G. and Richter, H., 1972, Textural variation in petrogenetic analysis: Geol. Soc. Amer. Bull., v. 83, p. 665 - 76.

- Emslie, R.F., 1972, Grenville Province: from Geology of Canadian Shield, in Geology and Economic Minerals of Canada, Dept. of Energy, Mines and Resources, Canada.
- Etheridge, M.A., 1973, Variation of the length/thickness ratio of biotite crystals in metamorphic rocks: a discussion: Jour. of Geol., v. 81, p. 234 - 36.
- , Hobbs, B.E. and Paterson, M.S., 1973, Experimental deformation of single crystals of biotite: Contrib. to Min. and Petrol., v. 38, p. 21 - 36.
- and ———, 1974, Chemical and deformational controls on recrystallization of mica: Contrib. to Min. and Petrol., v. 43, p. 111 - 24.
- , Paterson, M.S. and Hobbs, B.E., 1974, Experimentally produced preferred orientation in synthetic mica aggregates: Contrib. to Min. and Petrol., v. 44, p. 275 - 94.
- and Lee, M.F., 1975, Microstructure of slate from Lady Loretta, Queensland, Australia: Geol. Soc. Amer. Bull., v. 86, p. 13 - 22.
- Eugster, H.-P., 1970, Thermal and ionic equilibria among muscovite, K. feldspar and alumino-silicate assemblages: Fortschr. Miner., v. 47, p. 106 - 123.
- Evans, B.W., 1965, Application of a reaction-rate method to the breakdown equilibria of muscovite and muscovite plus quartz: Amer. Jour. Sci., v. 263, p. 647 - 67.
- Ferguson, C.C. and Harte, B., 1975, Textural patterns at porphyroblast margins and their use in determining the time relations

of deformation and crystallization: Geol. Mag., v. 112,  
p. 467 - 81.

Fleuty, M.J., 1964, The description of folds: Proc. Geol. Assoc.,  
v. 75, p. 461 - 92.

Flinn, D., 1956, On the deformation of the Funzie conglomerate,  
Fetlar, Shetland: Jour. Geol., v. 64, p. 480 - 505.

———, 1965, On the symmetry principle and the deformation  
ellipsoid: Geol. Mag., v. 102, p. 36 - 45.

Francoeur, D., 1975, Petrographic and Structural evolution of  
granites and gneisses of the Grenville Province, Glamorgan  
and Monmouth townships, Ontario: M.Sc. thesis, Univ. of  
Ottawa.

Fyson, W.K., 1970, Structural relations in metamorphic rocks,  
Shuswap Lake area, British Columbia: Geol. Assoc. of Canada,  
spec. paper no. 6, p. 107 - 22.

———, 1971, Fold attitudes in metamorphic rocks: Amer.  
Jour. Sci., v. 270, p. 373 - 82.

Galwey, A.K. and Jones, K.A., 1962, Inclusions in garnets:  
Nature, v. 193, p. 471 - 2.

——— and ———, 1963, An attempt to determine  
the mechanism of a natural mineral forming reaction from an  
examination of the products: Jour. Chem. Soc., p. 5681 - 6.

——— and ———, 1966, Crystal size frequency dis-  
tribution in garnets in some analysed metamorphic rocks from  
Mallaig, Inverness, Scotland: Geol. Mag., v. 103, p. 143 -  
52.

- Griggs, D.T., Paterson, M.S., Heard, H.C. and Turner, F.C., 1960, Annealing recrystallization in calcite crystals and aggregates: Geol. Soc. Amer. Mem. 79, p. 1 - 38.
- Groschong, R.H., 1972, Deformation by pressure solution: Progr. of Abstracts, northeast section 7<sup>th</sup> Ann. Meetg. Geol. Soc. Amer., p. 17.
- Guidotti, C.V., 1968, Prograde muscovite pseudomorphs after staurolite in the Rangeley-Oquossoc areas, Maine: Amer. Mineral., v. 53, p. 1368 - 76.
- , 1970, The mineralogy and petrology of the transition from the lower to upper sillimanite zone in the Oquossoc area, Maine: Jour. of Petrol., v. 11, p. 277 - 336.
- , 1973, Compositional variation of muscovite as a function of metamorphic grade and assemblage in metapelites from N.W. Maine: Contrib. to Min. and Petrol., v. 42, p. 33 - 42.
- Heidecker, E., 1963, The tectonic significance of structures in some Grenville rocks: Ph.D. thesis, Queen's Univ., Kingston.
- Heitanen, A., 1967, On the facies series in various types of metamorphism: Jour. of Geol., v. 75, p. 187 - 214.
- Helgeson, H.C., 1967, Solution chemistry and metamorphism: in Researches in Geochemistry, vol. II, P.H. Abelson editor, Wiley and Sons.
- Helm, D.G. and Siddans, A.W.B., 1971, Deformation of a slaty, lapillar tuff in the English Lake District: discussion: Geol. Soc. Amer. Bull., v. 82, p. 523 - 532.

Herring, C., 1953, The use of classical macroscopic concepts in surface energy problems: in Structure and Properties of Solid Surfaces, R. Gomer and C.S. Smith editors, Chicago Univ. Press, Chicago.

Hossack, J.R., 1968, Pebble deformation and thrusting in the Bygdin area, (southern Norway): Tectonophys., v. 5, p. 315 - 39.

Hounsflow, A.W. and Moore, J.M., 1967, Chemical petrology of Grenville schists near Fernleigh, Ontario: Jour. of Petrol., v. 8, p. 1 - 28.

Imbrie, J., 1956, Biometrical methods in the study of invertebrate fossils: Amer. Mus. Nat. Hist. Bull., v. 102, p. 217 - 52.

Jennings, D.S., 1969, Origin and metamorphism of part of the Herman Group near Bancroft, Ontario: Ph.D. thesis, McMaster Univ., Hamilton, Ontario.

Johnson, M.R.W., 1963, Some time relations of movement and metamorphism in the Scottish highlands: Geol. en Mijnb., v. 42, p. 121 - 42.

Jones, K.A. and Galwey, A.K., 1964, A study of possible factors concerning garnet formation in rocks from Ardara, Co. Donegal, Ireland: Geol. Mag., v. 101, p. 76 - 93.

\_\_\_\_\_ and \_\_\_\_\_, 1966, Size distribution, composition and growth kinetics of garnet crystals in some metamorphic rocks from the west of Ireland: Quart. Jour. Geol. Soc. (London), v. 122, p. 29 - 44.

Jones, K.A. and Galwey, A.K., 1972, Variation of the length/thickness ratio of biotite crystals in metamorphic rocks: Jour. of Geol., v. 80, p. 112 - 120.

\_\_\_\_\_ and \_\_\_\_\_, 1973, Variation of the length/thickness ratio of biotite crystals in metamorphic rocks: a reply: Jour. of Geol., v. 81, p. 236 - 38.

Kamb, W.B., 1959, Ice petrofabric observations from Blue Glacier, Washington, in relation to theory and experiment: Jour. Geophys. Res., v. 64, p. 1891 - 1909.

Kretz, R., 1966 (a), Interpretation of the shape of mineral grains in metamorphic rocks: Jour. of Petrol., v. 7, p. 68 - 94.

\_\_\_\_\_, 1966 (b), Grain-size distributions for certain metamorphic minerals in relation to nucleation and growth: Jour. of Geol., v. 74, p. 147 - 73.

\_\_\_\_\_, 1973, Kinetics of the crystallization of garnet at two localities near Yellowknife: Can. Mineral., v. 12, p. 1 - 20.

\_\_\_\_\_, 1974, Some models for the rate of crystallization of garnet in metamorphic rocks: Lithos, v. 7, p. 123 - 31.

Krogh, T.E. and Davis, G.L., 1969, Old isotopic ages in the northwestern Grenville Province, Ontario: in Age Relations in High Grade Metamorphic Terrains, H.R. Wynne-Edwards editor, Geol. Assoc. Can. Spec. Paper No. 5, p. 189 - 93.

\_\_\_\_\_ and Hurley, P.M., 1968, Strontium isotope variation and whole rock isochron studies, Grenville Province of

Ontario: Jour. Geophys. Res., v. 73, p. 7107 - 25.

Kwak, T.A.P., 1971 (a), Justification for both ionic and thermal reactions in Grenville Province pelitic rocks near Sudbury, Ontario, Canada: Can. Jour. Earth Sci., v. 8, p. 1333 - 54.

\_\_\_\_\_, 1971 (b), The selective replacement of aluminium silicates by white micas: Contrib. to Min. and Petrol., v. 32, p. 193 - 210.

\_\_\_\_\_, 1974, Natural staurolite breakdown reactions: Contr. to Min. and Petrol., v. 44, p. 57 - 80.

Laffitte, P., 1957, Introduction à l'étude des roches métamorphiques et des gîtes métallifères: Masson et Cie., Paris.

Leith, C.K., 1905, Rock cleavage: U.S. Geol. Surv. Bull., No. 239, p. 1 - 216.

Logan, W.E., 1863, Geological Survey of Canada: Report of progress from its commencement to 1863:

Lumbers, S.B., 1967 (a), Stratigraphy, plutonic activity and metamorphism of the Ottawa River remnant in Bancroft-Madoc area of the Grenville Province of southeastern Ontario: Ph.D. thesis, Princetown Univ., U.S.A.

\_\_\_\_\_, 1967 (b), Geology and mineral deposits of Bancroft-Madoc area: in Geology of parts of eastern Ontario and western Quebec, S.E. Jenness editor, Geol. Assoc. Canada guidebook., p. 13 - 29.

March, A., 1932, Mathematische Theorie der Regelung nach der Korngestalt bei affiner Deformation: Zeitschr. Kristallograph., v. 81, p. 285 - 98.

- Maxwell, J.C., 1962, On the origin of slaty and fracture cleavage in the Delaware Water Gap area, New Jersey and Pennsylvania: Geol. Soc. Amer. Mem., Buddington volume, p. 281 - 311.
- Means, W.D. and Williams, P.F., 1972, Compositional differentiation in an experimentally deformed salt-mica specimen: Geology, v. 2, p. 15 - 16.
- Mehnert, K.R., 1939, Die Meta-Konglomerate des Weisenthaler Greiszuges in Sächsischen Erzgebirge: Mineral. Petrogr. Mitt., v. 50, p. 194 - 272.
- \* Miller, W.G. and Knight, C.W., 1913,
- Misch, P., 1969, Paracrystalline microboudinage of zoned grains and other criteria for synkinematic growth of metamorphic minerals: Amer. Jour. Sci., v. 267, p. 43 - 63.
- \_\_\_\_\_, 1971, Porphyroblasts and crystallization force: some textural criteria: Geol. Soc. Amer. Bull., v. 82, p. 245 - 252.
- Moench, R.H., 1966, Relation of  $S_2$  schistosity to metamorphosed clastic dykes, Rangeley-Phillips area, Maine: Geol. Soc. Amer. Bull. v. 77, p. 1449 - 62.
- Moore, J.M., 1967, Grenville structure, stratigraphy and metamorphism, southeastern Ontario, Northbook-Ompah area: in Geology of parts of eastern Ontario and western Quebec, S.E. Jenness editor, Geol. Assoc. Canada guidebook, p. 59 - 71.
- \* \_\_\_\_\_ and Thompson, P.H., 1972, The Elinton Group, Grenville Province, eastern Ontario, Canada: Proc. XXIV Internat. Geol. Congress, Montreal, section 1, p. 221 - 29.

- Morey, G.W., 1957, The solubility of solids in gasses: Econ. Geol., v. 52, p. 225 - 51.
- Oertel, G., 1970, Deformation of a slaty, lapillar tuff in the Lake District, England: Geol. Soc. Amer. Bull., v. 81, p. 1173 - 88.
- \_\_\_\_\_ and Phakey, P.P., 1972, The texture of a slate from Nantlle, Caernavon, north Wales: Texture, v. 1, p. 1 - 8.
- Peach, P.A., 1958, The geology of Darling township and part of Lavant township: Ann. Rept. Ontario Dept. Mines, v. 65, p. 47 - 60.
- Powell, C. McA., 1972, Tectonically dewatered slates in the Ludlovian of the Lake District, England: Geol. Jour., v. 8, p. 95 - 111.
- Rambaldi, E., 1970, Distribution of elements among feldspars, micas and epidote in some metamorphic rocks near Bancroft, Ontario: Ph.D. thesis, Univ. of Ottawa.
- Ramsay, J.G., 1962, Geometry of similar type folds: Jour. Geol., v. 70, p. 309 - 27.
- \_\_\_\_\_, 1967, Folding and Fracturing of Rocks: McGraw - Hill, New York.
- \_\_\_\_\_ and Wood, D.S., 1973, The geometric effects of volume change during deformation: Tectonophys., v. 16, p. 263 - 77.
- Sethuraman, K., 1970, Petrology of Grenville metavolcanics in the Bishop Corners - Donaldson area: Ph.D. thesis, Carleton Univ. Ottawa.
- \_\_\_\_\_ and Moore, J.M., 1973, Petrology of metavolcanic

- rocks in the Bishop Corners - Donaldson area, Grenville Province, Ontario: Can. Jour. Earth. Sci., v. 10, p. 589 - 614.
- Siddans, A.W.B., 1972, Slaty cleavage - a review of research since 1815: Earth Sci. Reviews, p. 205 - 232.
- Silver, L.T. and Lumbers, S.B., 1965, Geochronological studies in the Bancroft - Madoc area of the Grenville Province, Ontario, Canada: Abstract, Geol. Soc. Amer. Spec. Publ. No. 87, p. 156.
- Smith, B.L., 1958, The geology of the Clarendon-Dalhousie area: Ann. Rept. Ontario Dept. Mines, v. 65, p. 1 - 46.
- , 1969, The Ompah syncline in relation to the Grenville stratigraphy of eastern Ontario: in Age Relations in High Grade Metamorphic Terrains, H.R. Wynne-Edwards editor, Geol. Assoc. Canada Spec. Paper No. 5, p. 225 - 28.
- , Vogel, T.A. and Spence, W.H., 1969, Precambrian metaconglomerate mantling a granite dome at Plevna Lake in the Grenville Province of southeastern Ontario, Canada: Geol. Soc. Amer. Bull., v. 80, p. 297 - 302.
- Spence, W.H., 1968, Relict plagioclase phenocrysts from meta-volcanic rocks mantling a granite dome, and the petrology of the associated rocks in the Grenville Province of southeastern Ontario: Ph.D. thesis, Rutgers Univ., U.S.A.
- Spry, A., 1963, The origin and significance of snowball structure in garnet: Jour. Petrol., v. 4, p. 211 - 22.
- , 1969, Metamorphic Textures: Pergaman Press, Oxford.

Swalin, R.A., 1962, Thermodynamics of Solids: Wiley and Sons, New York.

Thakur, V.C., 1974, Analysis of type of tectonic strain in polyphase deformed Molare area, Tessin, Switzerland, using deformed pebbles: Geol. Rundschau, Bd. 63, p. 326 - 34.

Thompson, J.B., 1957, The graphical analysis of mineral assemblages in pelitic schists: Amer. Mineral., v. 42, p. 842 - 58.

Thompson, P.H., 1972, Stratigraphy, structure and metamorphism of the Flinton Group in the Bishops Corners area, Grenville Province, eastern Ontario: Ph.D. thesis, Carleton Univ., Ottawa.

\* Thompson, P.H., 1973

Tozer, C.F., 1955, The mode of occurrence of sillimanite in the Glen district, Co. Donegal: Geol. Mag., v. 92, p. 310 - 20.

Tullis, T.E., and Wood, D.S., 1975, Correlation of finite strain from both reduction bodies and preferred orientation of mica in slate from Wales: Geol. Soc. Amer. Bull., v. 86, p. 632 - 38.

Turner, F.J., 1968, Metamorphic Petrology: McGraw-Hill, New York.

——— and Weiss, L.E., 1963, Structural Analysis of Metamorphic Tectonites: McGraw-Hill, New York.

Venkitasubramanian, C.S., 1969, Large scale superimposed folds in the Precambrian of the Actinolite-Kaladar area, southeastern Ontario: Ph.D. thesis, Queen's Univ., Kingston.

Vernon, R.H., 1968, Microstructures of high-grade metamorphic rocks at Broken Hill, Australia: Jour. of Petrol., v. 9, p. 1 - 22.

- Wallach, J.L., 1974, Origin of the Hinchinbrooke gneiss, and its relationship to the Grenville Group rocks of southeastern Ontario: Progr. of Abstracts, Geol. Assoc. Canada Ann. Meetg. St-John's Nfld, p. 96.
- Williams, P.F., 1972, Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia: Amer. Jour. Sci., v. 272, p. 1 - 47.
- Winkler, H.G.F., 1967, Petrogenesis of Metamorphic Rocks: Springer-Verlag, New York.
- Wintsch, R.P., 1975 (a), Feldspathization as a result of deformation: Geol. Soc. Amer. Bull., v. 86, p. 35 - 38.
- , 1975 (b), Solid-fluid equilibria in the system  $KAlSi_3O_8 - NaAlSi_3O_8 - SiO_2 - H_2O - HCl$ : Jour. Petrology, v. 16, p. 57 - 79.
- Wood, D.S., 1973, Patterns and magnitudes of natural strain in rocks: Phil. Trans. Roy. Soc. London, v. 274, p. 373 - 82.
- , 1974, Current views on the development of slaty cleavage: Ann. Review of Earth and Plan. Sci., v. 2 p. 369 - 401.
- Wynne-Edwards, H.R., 1972, The Grenville Province, in Variations in Tectonic Styles in Canada, R.A. Price and J.W. Douglas editors, Geol. Assoc. Canada Spec. Paper No. 11, p. 264 - 334.
- Yardley, B.W.D., 1975, On some quartz-plagioclase veins in the Connemara schists, Ireland: Geol. Map., v. 112, p. 183 - 190.
- Zwart, H.J., 1960, Relations between folding and metamorphism:

in the central Pyrenees, and their chronological succession:

Geol. en. Mijnb., v. 39, p. 163 - 180.

Zwart, H.J., 1964 (a), The structural evolution of the Palaeozoic of the Pyrenees: Geol. Rundschau, Bd 53, p. 170 - 205.

\_\_\_\_\_, 1964 (b), The development of successive structures in the Devonian and Carboniferous of Devon and Cornwall:

Geol. en. Mijnb., v. 12; p. 516 - 26.

---

Miller, W.G. and Knight, C.W., 1913, The precambrian geology of southeastern Ontario: Ont. Bur. Mines, V.XXII (2).

Thompson, P.H.; 1973, Mineral. zones and isograds in "impure" calcareous rocks, an alternative method of evaluating metamorphic grade: Contrib. to Min. and Petrol., v. 42, p. 63-180.

Plate 1. Bedding,  $S_0$ , in carbonate conglomerate (unit 3c) defined by variations in matrix composition and pebble content.  $S_1$  foliation is oblique to  $S_0$ .

Locality: 4 km. southeast of Plevna on highway 506.

Plate 2. Cross-bedding in quartzite (unit 4a). The sedimentary <sup>is</sup> is to the left of the photograph, and the unit is overturned.

Locality: 3.2 km. south, southeast of Ompah, in the Ompah synform.

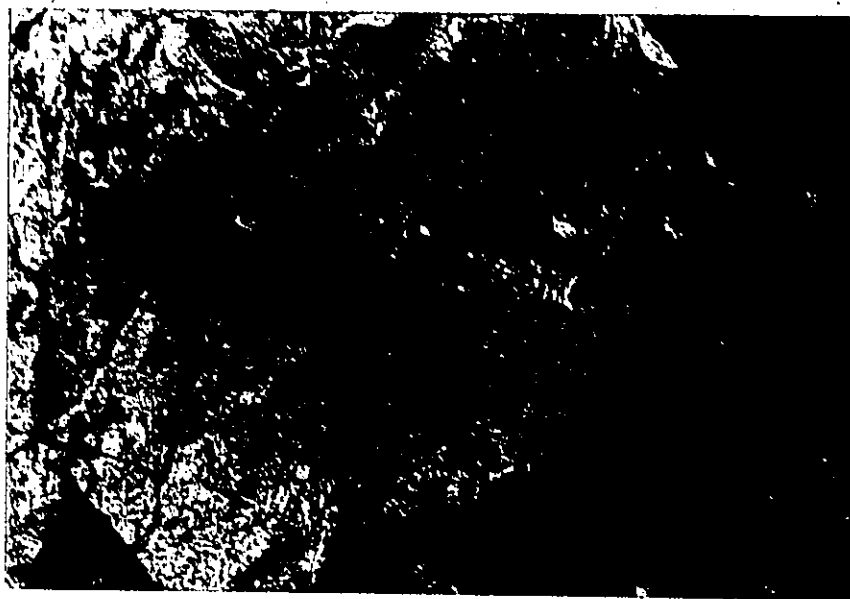


Plate 3A. Recumbent mesoscopic <sup>F<sub>1</sub></sup> fold in quartzite  
(unit 4a)..

Plate 3B. Close up of the hinge region showing S<sub>1</sub>  
foliation in more micaceous layers. In  
the top left of the photograph some  
layers are cross-bedded.

Locality: 3.4 km. south of Ompah, in  
the Ompah synform.



3A

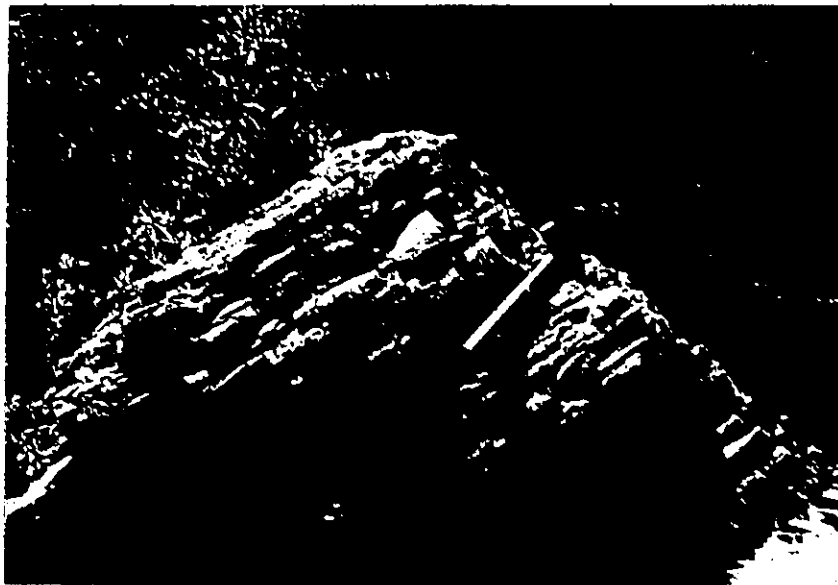


3B

Plate 4A. Deformed pebbles in a loose block of conglomerate, viewed perpendicular to the foliation  $S_1$ , and the lineation  $L_1$ .

Plate 4B. viewed parallel to  $L_1$ .

Locality: Mississippi River, 3.8 km. south of Ompah, in the Ompah synform.



4A



4B

Mylonitic Rocks

Plate 5A. Mylonitic layering in quartz-biotite-calcite-opaque-rock (unit 6a), showing possible attenuated  $F_1$  fold hinges. Scale bar represents 5 mm.

Plate 5B. Mylonitic layering as in A, folded by  $F_2$  folds. Micas are aligned in  $S_2$ . Scale bar represents 5 mm.

Plate 5C. Mesoscopic  $F_2$  folds in marble mylonite (unit 5).

Plate 5D. Same as C; close-up showing nature of folded mylonitic layering.

Locality: 2.2-2.5 km. northeast of  
Ardoch on highway 506.

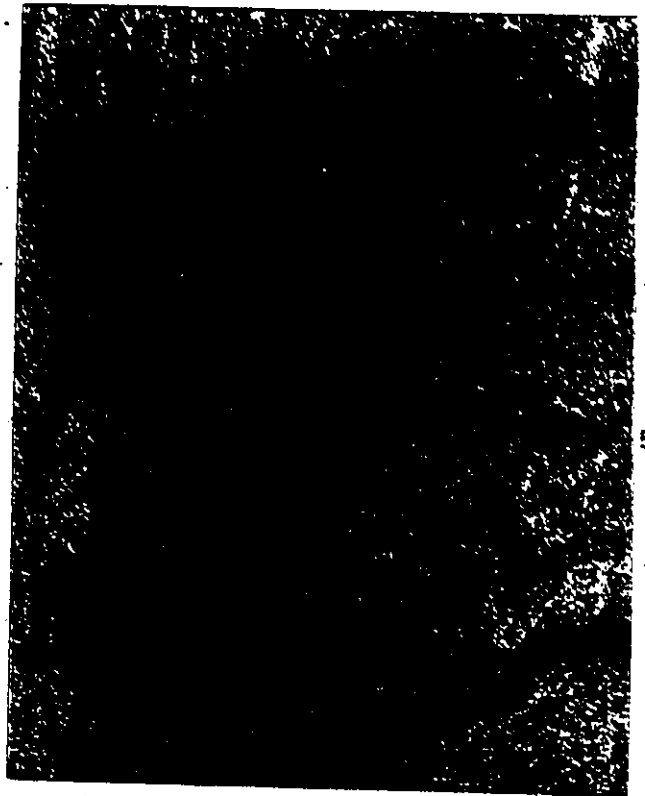
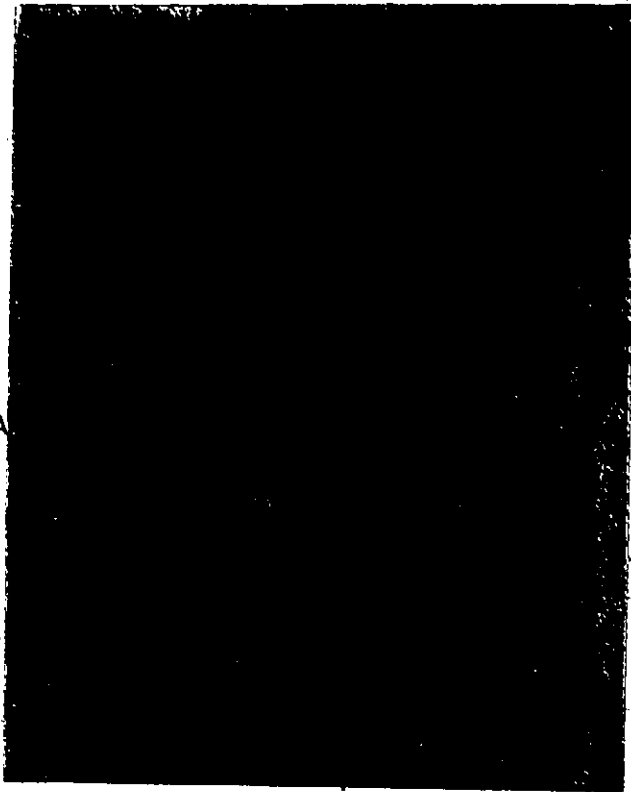


Plate 6. Recumbent mesoscopic  $F_1$  fold with superimposed upright  $F_2$  folds and  $S_2$  fracture cleavage. The core of the  $F_1$  fold is of marble, around which a thin layer of biotite schist is folded.

Locality: 2.9 km. south of Ompah, in marble (unit 5) between Ompah synform and Fernleigh belt.



6

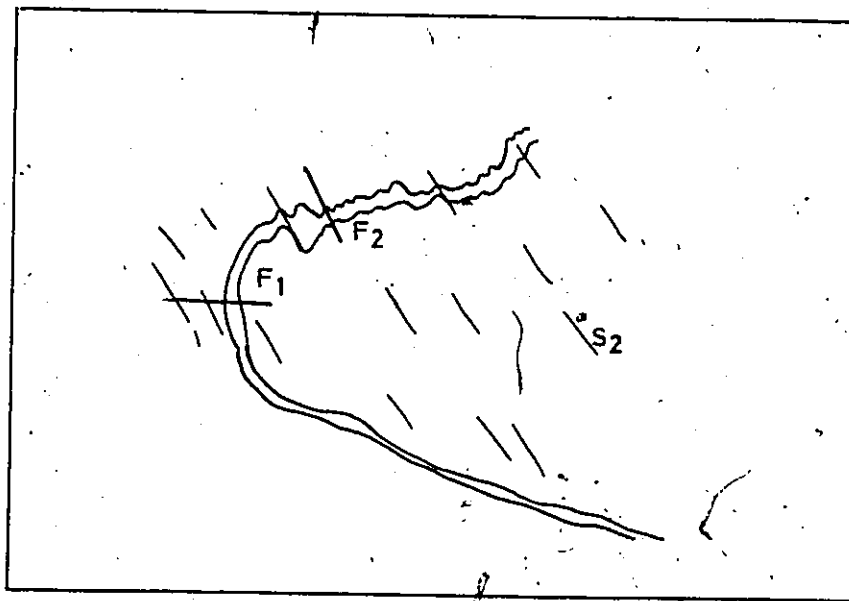
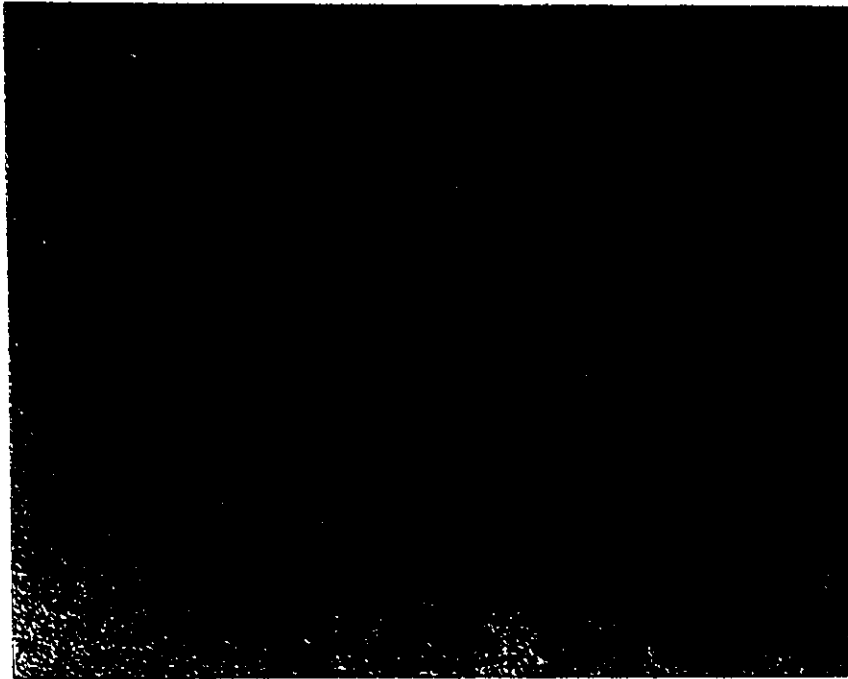


Plate 7A. Deformed tremolite-quartz aggregates in biotite-quartz-carbonate schist (unit 6c). Section cut parallel to  $L_1$ , perpendicular to  $S_1$ . Scale bar represents 5 mm.

Plate 7B. As above, section cut perpendicular to  $L_1$  and  $S_1$ . Scale bar represents 5 mm.

Locality: 1.5 km. west of Ardoch.



7 A



7 B



Plate 8. Sillimanite pod (dark grey) aligned in coarse grained  $S_1$  foliation, which is deformed by  $F_2$  folds. Most micas have annealed after  $D_2$  deformation. Section perpendicular to  $L_1$  and  $S_1$ . Scale bar represents 5 mm. Specimen is from pelitic schist (unit 6b) of the Fernleigh belt, 1.7 km. west of Folger.

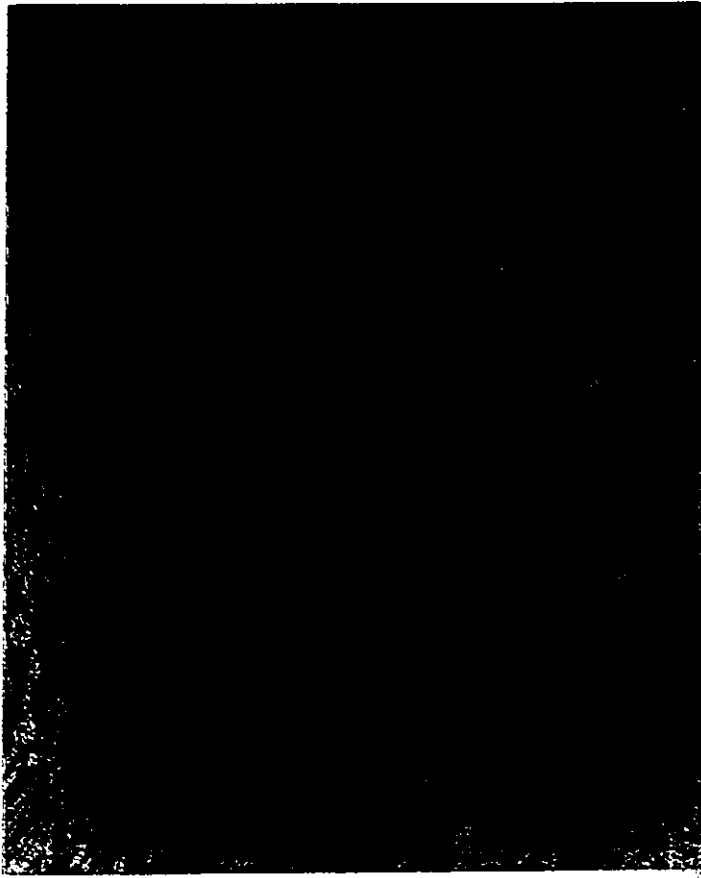


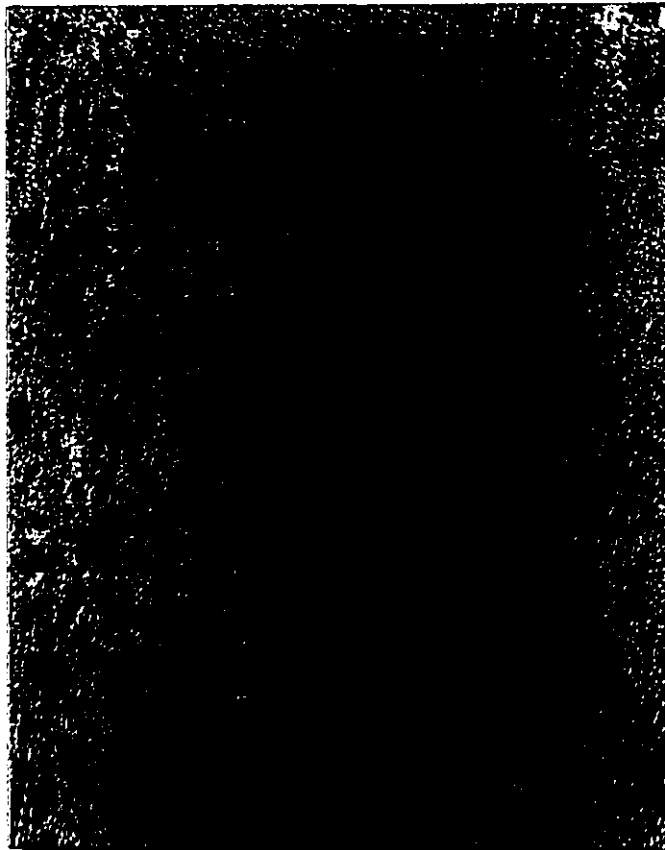
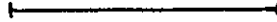
Plate 9A. Deformed biotite porphyroblasts in biotite-quartz schist (unit 6a).  
Section perpendicular to  $L_1$  and  $S_1$ .  
Scale bar represents 5 mm.

Plate 9B. As above, section cut parallel to  $L_1$ , perpendicular to  $S_1$ . Scale bar represents 5 mm.

Locality: 1.3 km. west of Plevna.



9 A



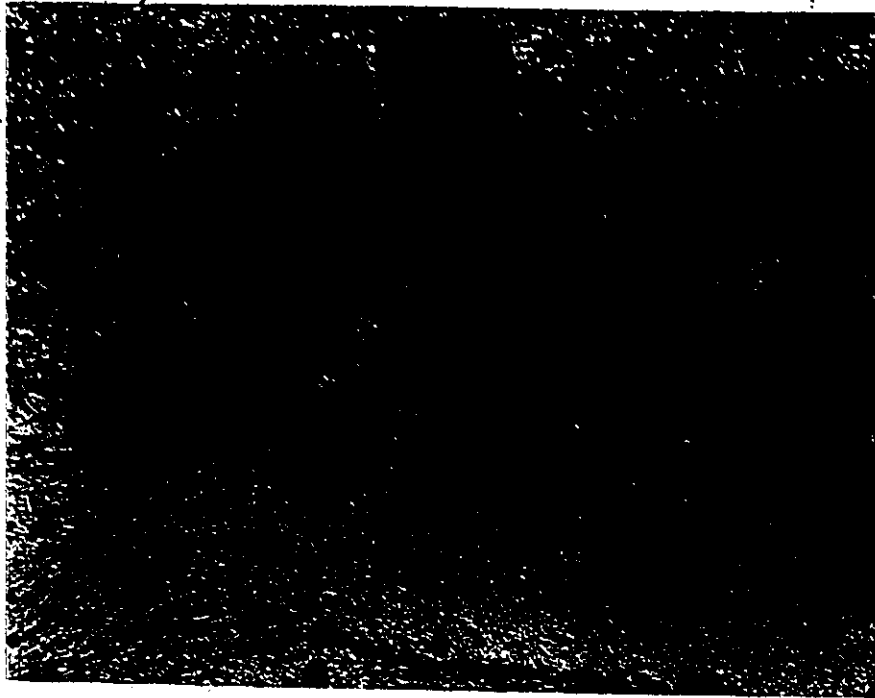
9 B



Plate 10A. Textural relations around a porphyroblast of plagioclase feldspar in pelitic schist (unit 6b). Part of a large porphyroblast of plagioclase feldspar, which contains fine grained inclusions ( $S_{int}$ ) of quartz, opaques and some muscovite, as well as larger crystals of opaques and kyanite. Other porphyroblast phases are kyanite and biotite, the latter being morphological aligned in  $S_1$ .  $S_1$  foliation ( $S_{ext}$ ), principally defined by muscovite, is deflected around all porphyroblasts, and  $S_{int}$  is parallel to  $S_{ext}$  away from the porphyroblasts. Section cut perpendicular to  $L_1$  and  $S_1$ . Scale bar represents 5 mm.

Plate 10B. Enlargement of part 10A showing the oriented inclusions in plagioclase, which are of larger size at the periphery of the porphyroblast. The porphyroblast of kyanite, which is inclusion zoned, has been bent and subsequently annealed. Scale bar represents 2 mm.

Locality: 0.6 km. northeast of Fernleigh, under hydro lines.



10A



10B

Plate 11. Textural relations around a porphyroblast of staurolite in pelitic schist (unit 6b). The porphyroblast includes smaller porphyroblasts of plagioclase and opaques morphologically aligned in  $S_1$ , as well as orientated fine grained inclusions ( $S_{int}$ ) of quartz and opaques.  $S_{int}$  is also present in the plagioclase, where it is parallel to, but of finer grain size than that in staurolite. Staurolite is inclusion zoned, and  $S_{int}$ , which is coarser at the edge of the porphyroblast, is continuous with the external foliation ( $S_1$ ) defined by muscovite crystals. Section perpendicular to  $L_1$  and  $S_1$ . Scale bar represents 5 mm.

Locality: 0.6 km. northeast of Fernleigh, under hydro lines.

Plate 12. Textural relations of garnet porphyroblasts in pelitic schist (unit 6b). A coarse grained foliation  $S_1$ , defined by biotite, sillimanite, quartz and minor muscovite, is deflected around porphyroblasts of garnet. The garnets contain fine grained planar inclusion trails ( $S_{int}$ ) which are not parallel to  $S_1$  outside the porphyroblast. Section perpendicular to  $L_1$  and  $S_1$ . Scale bar represents 5 mm.

Locality: 2.2 km. north of Clyde Forks.



11



12



## APPENDIX I

### Measurement of Strain

The measurement of strain can be subdivided into two parts, one concerned with the type<sup>1</sup> of strain, the other with the magnitude of strain.

#### Type of strain

The deformation ellipsoid is defined as having three orthogonal axes  $X \gg Y \gg Z$ , the type of strain being a function of the relative lengths of these axes. A method of depicting the three-dimensional type of strain graphically in two dimensions was described by Flinn (1956). The three axes of the deformation ellipsoid define the three principal finite strains  $(1 + e_1)$ ,  $(1 + e_2)$  and  $(1 + e_3)$ , where  $e_1 \gg e_2 \gg e_3$  are the three principal extensions;  $e$  is related to the axial ratios of the pebble  $e_1 = \frac{X - d}{d}$ , where  $d$  is the diameter of a sphere of the same volume as the ellipsoid. The ratio of  $\frac{(1 + e_1)}{(1 + e_2)} = a$  is plotted as the ordinate ( $\frac{X}{Y}$  in terms of the axial ratios of the pebble); and  $\frac{(1 + e_2)}{(1 + e_3)} = b$  is plotted as the abscissa ( $b = \frac{Y}{Z}$ ). Flinn then defined  $K$  as the ratio

---

<sup>1</sup> Type of strain refers to the shape of the deformation ellipsoid, and has been referred to as the symmetry of strain by Flinn (1956), Hossack (1968). However the present nomenclature is preferred, as it avoids confusion with the symmetry of the tectonite fabric, as described by Turner and Weiss (1963).

$$K = \frac{a - 1}{b - 1} = \frac{(1 + e_3)(e_1 - e_2)}{(1 + e_2)(e_2 - e_3)};$$

so that the line  $K = 1$  bisects the graph with a gradient of

1.5. Any point plotting on the line  $K = 1$  has undergone plane strain, in which there is no change in the Y dimension of the ellipsoid compared with the diameter of a sphere of the same volume during deformation. Points plotting above and to the left of the line  $K = 1$  are constricted or prolate ellipsoids, whilst points falling beneath and to the right of the line have undergone extension of the Y dimension during deformation, and are flattened or oblate ellipsoids.

Recently Wood (1973, 1974) has suggested that a logarithmic version of this plot is advantageous, since all lines representing equal changes in length are straight. Furthermore the data show a greater spread if  $\log \frac{Z}{Y}$  (rather than  $\log \frac{Y}{Z}$ ) is plotted as the abscissa; accordingly these adaptations are followed in this work.

#### Amount of Strain

The Flinn plot allows an assessment of the amount of strain that has occurred in a clast parallel to the three principal axes of strain ellipsoid, but it does not enable the total amount of strain in the clast to be calculated (except in a qualitative way in that the greater the strain, the further away the points plot from the origin). Nadai

(1950, 1963, in Hossack 1968) has shown how a quantitative estimate of the amount of strain can be made, irrespective of the type of strain. Thus the method is suitable for a comparison of the strain in oblate and prolate ellipsoids.

The strain unit used is that of logarithmic or natural strain ( $\epsilon$ ), where  $\epsilon_1 \gg \epsilon_2 \gg \epsilon_3$ .  $\epsilon$  is related to  $e$  by the relation  $\epsilon = \log_e (1 + e)$ . The values for the natural strain ratios of the ellipsoid are then combined in the equation

$$\gamma_0 = \left( \frac{2}{3} \right) \left[ (\epsilon_1 - \epsilon_2)^2 + (\epsilon_2 - \epsilon_3)^2 + (\epsilon_3 - \epsilon_1)^2 \right]^{1/2}$$

where  $\gamma_0$  is the natural octahedral shear.  $\gamma_0$  is an absolute measure of the total distortional or strain component of deformation, proportional to the amount of work necessary to produce the distortion, and hence is useful for comparison of the magnitude of strain in oblate and prolate ellipsoids.

## APPENDIX II

### Muscovite Analyses

In the chapter entitled "Textural Variation with Metamorphic Grade" it was shown that the shape of muscovite crystals varies systematically with grade of metamorphism, and some of the factors which may have contributed to this variation were briefly discussed. Amongst these was the chemical composition of the muscovite itself, and it was with this in mind that the following measurements were attempted. Several crystals of different shape ratio were analysed in each of seven thin sections taken from across the metamorphic zonation examined in the study ie. chloritoid-staurolite, staurolite-kyanite, sillimanite and sillimanite- K. feldspar zones, so that the compositional variation within and between samples could be assessed. Specimen locations are shown in fig.A-1.

A comprehensive study of the variation of muscovite composition as a function of assemblage (ie. bulk composition) and grade of metamorphism was made by Guidotti (1973) in samples from N.W. Maine, U.S.A., and this provides a very suitable source for comparison of the data. Guidotti noted that the role of the assemblage was important in controlling muscovite composition (c.f. Butler 1967, Rambaldi 1970) and useful comparisons between specimens from different rocks

could only be made in "limiting assemblages"\*. In the context

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\* A limiting assemblage is one in which the number of phases present equals the number of components required to describe those phases. Hence phase compositions will be a function of temperature and pressure (Guidotti 1973).

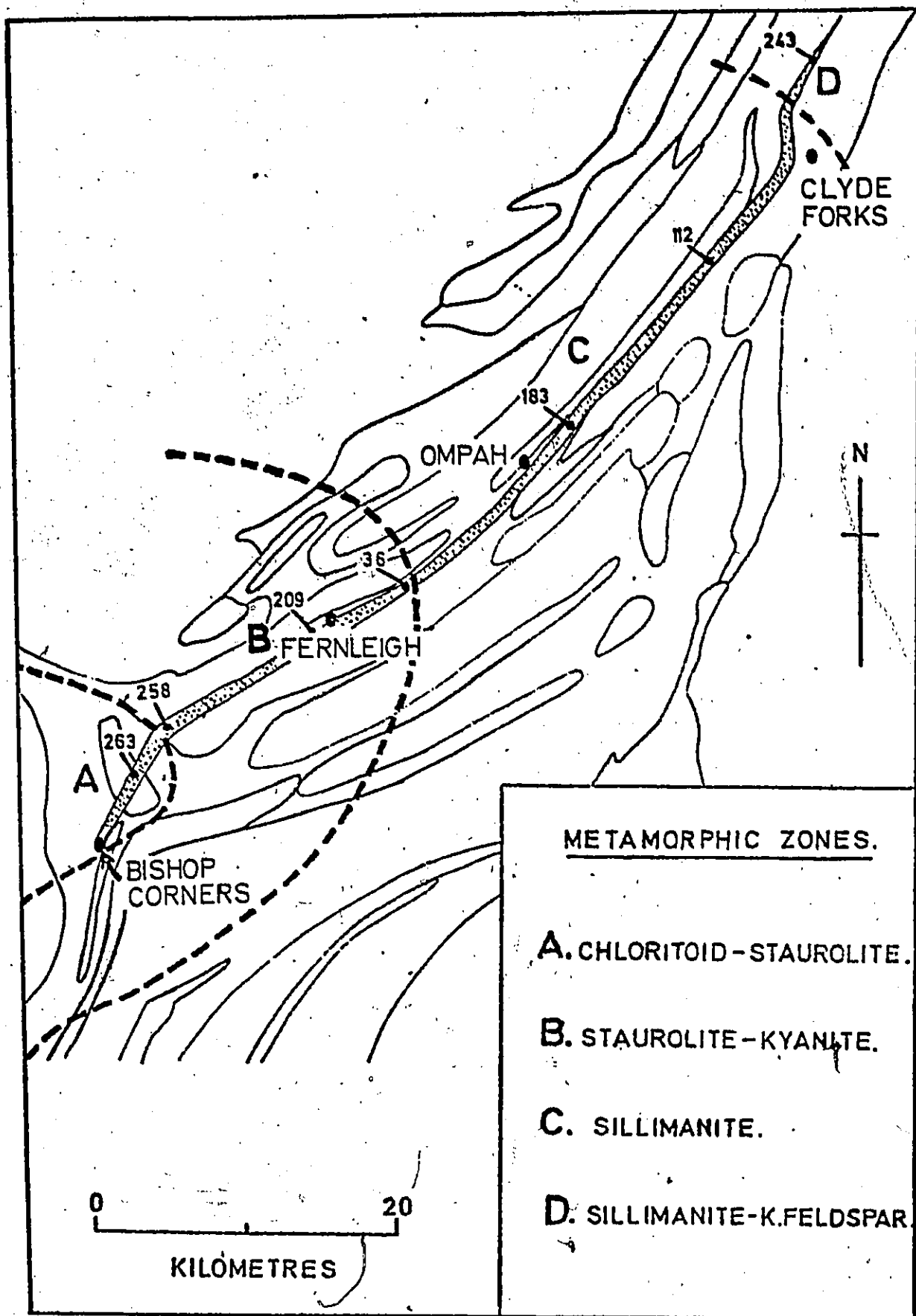


Fig. A-1 Metamorphic zonation in the Ompah area. The Fernleigh belt is stippled, and numbers refer to the location of samples in pelitic schist (unit 6b) used for muscovite analysis.

of his study, such assemblages were termed "high aluminium assemblages", since they contain an aluminous phase such as kyanite, sillimanite or staurolite. Accordingly in this study all analyses were performed on high aluminium assemblages, so that any consistent variation observed between samples may be considered a result of the metamorphic environment.

The bulk composition of the rocks considered in this study is represented on the triangular A-K-Na diagram, fig.A-2. It should be pointed out here that the highest grade specimen (243), although northeast of the sillimanite - K. feldspar isograd, does not itself contain K. feldspar (see Metamorphism) and that the aluminous phase present in the lowest grade specimen (263) is staurolite, in contrast to an aluminium silicate in the others. No paragonite was encountered in any of the specimens, and no true phengites were found. (phengites have an Si:Al ratio of greater than 3:1, Deer, Howie and Zussman 1969). All assemblages can therefore be represented in the same segment of the triangular diagram, so the effects of bulk composition on crystal chemistry are minimized.

#### Analytical Methods

Muscovite analyses were performed with an electron micro-

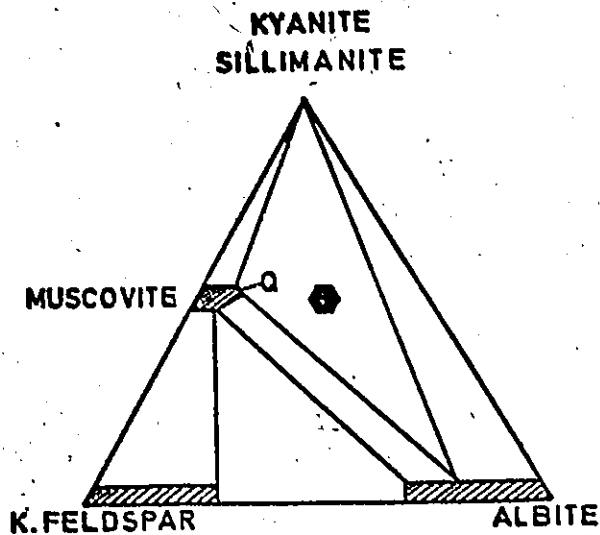


Fig.A-2 (after Guidotti 1973).  
A-K-Na diagram showing approximate composition of assemblages analysed (●). Lowest grade specimen (263) coexists with staurolite (neither aluminium silicate is present). Movement of tie lines about point Q allows muscovite composition to change - see text.

probe\*, the results being processed using the Bence and Albee (1968) correction factors. Unfortunately, due to inexperience with the machine, standards were not analysed sufficiently often, and some drift occurred between measurements, so that analysis totals did not sum to 100%. However the error is proportional, and can be largely offset if it is assumed that all analyses contain the same amount of water. Accordingly a value of 4.65% H<sub>2</sub>O was adopted, this value being the mean of that reported in several muscovite analyses culled at random from the literature. The results are thus comparable within themselves, and the trends may be regarded as real.

Results:

Results of the analyses are presented in table 1 on the following pages

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\* An A.R.L.-AMX energy dispersive unit at Queen's University Kingston was used for the analyses; samples were run for 10 elements. Operating conditions were 100  $\mu$  amps specimen current, 15 kv. accelerating potential and a 1-3  $\mu$  beam size, with a 120 second counting time. Two analyses per crystal were made in the coarser grained rocks (ie. all specimens except 263, 258 and 209), four crystals per specimen being analysed. In the fine grained rocks measurements were made on eight different crystals. From the duplicated results on single crystals it is estimated that the results are reproducible within at most 5% (3% for the majority of analyses) for Si, Al and K. For elements present in low concentrations eg. Na, Mg, the limits of precision of the machine are almost reached, and accuracy may be somewhat reduced. Nevertheless for these elements too, the results show a high degree of consistency between crystals of the same specimen, and are generally reproducible 2/10% for Na, 3/10% for Mg in crystals on which more than one analysis was performed.

Table A-1

## Specimen 263G

## Assemblage

Quartz-muscovite-biotite-staurolite-opaque-plagioclase-chlorite-(retrograde).

| Muscovite Analyses             |       |       |       |       |       |       |       |       |             |             |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------------|-------------|
| Analysis No.                   | 1     | 2     | 3     | 4     | 5     | 6     | 7     | 8     | 1-5<br>MEAN | 6-8<br>MEAN |
| L ratio<br>T                   | 4.5   | 10    | 10    | 4     | 10    | 8     | 5     | 7     |             |             |
| SiO <sub>2</sub>               | 47.43 | 47.32 | 47.06 | 47.20 | 48.28 | 50.99 | 50.54 | 55.25 | 47.45       | 52.26       |
| TiO <sub>2</sub>               | 0.35  | 0.31  | 0.40  | 0.59  | 0.34  | 0.00  | 0.27  | 0.31  | 0.40        | 0.19        |
| Al <sub>2</sub> O <sub>3</sub> | 33.96 | 35.18 | 34.37 | 34.47 | 34.45 | 32.45 | 32.75 | 29.24 | 34.49       | 31.48       |
| Fe(total)                      | 2.42  | 1.78  | 2.03  | 2.14  | 1.64  | 1.71  | 1.74  | 1.49  | 2.00        | 1.65        |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00        | 0.00        |
| MgO                            | 0.55  | 0.41  | 0.46  | 0.43  | 0.47  | 0.56  | 0.31  | 0.36  | 0.46        | 0.41        |
| CaO                            | 0.00  | 0.00  | 0.63  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00        | 0.00        |
| Na <sub>2</sub> O              | 1.05  | 0.91  | 0.66  | 0.95  | 0.77  | 0.82  | 0.79  | 0.73  | 0.87        | 0.78        |
| K <sub>2</sub> O               | 9.56  | 9.41  | 9.71  | 9.53  | 9.35  | 8.79  | 8.93  | 7.92  | 9.51        | 8.55        |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00        | 0.00        |
| Anhydrous total                | 95.32 | 95.32 | 95.32 | 95.31 | 95.30 | 95.32 | 95.33 | 95.30 |             |             |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.30 | 6.24 | 6.26 | 6.27 | 6.36 | 6.67 | 6.63 | 7.14 | 6.29 | 6.81 |
| Al <sup>iv</sup> | 1.70 | 1.76 | 1.74 | 1.73 | 1.64 | 1.33 | 1.37 | 0.86 | 1.71 | 1.19 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.63 | 3.71 | 3.65 | 3.67 | 3.72 | 3.68 | 3.69 | 3.60 | 3.68 | 3.66 |
| Ti               | 0.04 | 0.03 | 0.04 | 0.06 | 0.04 | 0.00 | 0.03 | 0.03 | 0.04 | 0.02 |
| **Fe             | 0.27 | 0.20 | 0.23 | 0.24 | 0.18 | 0.19 | 0.19 | 0.16 | 0.22 | 0.18 |
| Mg               | 0.11 | 0.17 | 0.09 | 0.08 | 0.09 | 0.11 | 0.06 | 0.07 | 0.11 | 0.08 |
| Σ                | 4.05 | 4.11 | 4.01 | 4.05 | 4.03 | 3.98 | 3.97 | 3.86 | 4.05 | 3.94 |
| Na               | 0.27 | 0.23 | 0.17 | 0.24 | 0.20 | 0.21 | 0.20 | 0.18 | 0.22 | 0.20 |
| K                | 1.63 | 1.59 | 1.64 | 1.61 | 1.57 | 1.50 | 1.49 | 1.30 | 1.61 | 1.43 |
| Σ                | 1.90 | 1.82 | 1.81 | 1.85 | 1.77 | 1.71 | 1.69 | 1.48 | 1.83 | 1.63 |
| Σ Al             | 5.33 | 5.47 | 5.39 | 5.40 | 5.36 | 5.01 | 5.06 | 4.46 | 5.39 | 4.84 |
| Σ Mg + Fe        | 0.38 | 0.37 | 0.32 | 0.32 | 0.27 | 0.30 | 0.25 | 0.23 | 0.33 | 0.26 |
| Na/(Na+K)×100    | 14.2 | 12.6 | 9.4  | 13.0 | 11.3 | 12.3 | 11.2 | 12.2 | 12.1 | 11.9 |
| Mg/(Mg+Fe)×100   | 28.9 | 45.9 | 28.1 | 25.0 | 33.3 | 36.7 | 24.0 | 30.4 | 32.2 | 30.9 |

\* The means for analyses 6-8 are given separately, because of their higher silica content and generally somewhat different characteristics (see text)

\*\* Total Fe as FeO in this and other analyses

Table A-1. Analyses of muscovite crystals of various shape ratios from different metamorphic grades.

Table A-1 (con'd)

## Specimen 258A

Assemblage  
quartz-muscovite-biotite-kyanite-plagioclase  
Muscovite Analyses

| Analysis No                    | 1     | 2     | 3     | 4     | 5     | 6     | 7     | 8     |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio<br>T                   | 8     | 5     | >15   | 4     | 15    | 10    | 9     | 6     | MEAN  |
| SiO <sub>2</sub>               | 47.42 | 48.11 | 47.34 | 46.42 | 47.40 | 46.60 | 47.98 | 48.91 | 47.52 |
| TiO <sub>2</sub>               | 0.40  | 0.00  | 0.34  | 0.86  | 0.41  | 0.00  | 0.26  | 0.29  | 0.32  |
| Al <sub>2</sub> O <sub>3</sub> | 34.83 | 34.87 | 35.10 | 34.41 | 35.06 | 36.84 | 34.62 | 34.08 | 34.97 |
| Fe(total)                      | 2.19  | 1.99  | 2.12  | 3.04  | 2.04  | 1.97  | 2.07  | 1.86  | 2.16  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.46  | 0.37  | 0.65  | 0.61  | 0.28  | 0.41  | 0.52  | 0.25  | 0.44  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.53  | 1.45  | 1.52  | 1.48  | 1.69  | 1.29  | 1.67  | 1.65  | 1.53  |
| K <sub>2</sub> O               | 8.49  | 8.53  | 8.24  | 8.49  | 8.44  | 8.21  | 8.20  | 8.28  | 8.36  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.32 | 95.32 | 95.31 | 95.31 | 95.32 | 95.32 | 95.32 | 95.32 |       |

Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.27 | 6.34 | 6.24 | 6.18 | 6.26 | 6.14 | 6.32 | 6.43 | 6.27 |
| Al <sup>iv</sup> | 1.73 | 1.66 | 1.76 | 1.82 | 1.74 | 1.86 | 1.68 | 1.57 | 1.73 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.69 | 3.76 | 3.70 | 3.58 | 3.72 | 3.86 | 3.70 | 3.71 | 3.72 |
| Ti               | 0.04 | 0.00 | 0.03 | 0.09 | 0.04 | 0.00 | 0.03 | 0.03 | 0.03 |
| **Fe             | 0.24 | 0.22 | 0.23 | 0.34 | 0.23 | 0.22 | 0.23 | 0.20 | 0.24 |
| Mg               | 0.09 | 0.08 | 0.13 | 0.12 | 0.05 | 0.08 | 0.11 | 0.05 | 0.09 |
| Σ                | 4.06 | 4.06 | 4.09 | 4.13 | 4.04 | 4.16 | 4.07 | 3.99 | 4.08 |
| Na               | 0.39 | 0.38 | 0.39 | 0.38 | 0.44 | 0.33 | 0.42 | 0.42 | 0.39 |
| K                | 1.44 | 1.43 | 1.39 | 1.45 | 1.42 | 1.37 | 1.37 | 1.39 | 1.41 |
| Σ                | 1.83 | 1.81 | 1.78 | 1.83 | 1.86 | 1.70 | 1.79 | 1.81 | 1.80 |
| Σ Al             | 5.42 | 5.42 | 5.46 | 5.40 | 5.46 | 5.72 | 5.38 | 5.28 | 5.44 |
| Σ Mg + Fe        | 0.33 | 0.30 | 0.36 | 0.46 | 0.28 | 0.30 | 0.34 | 0.25 | 0.33 |
| Na/(Na+K)×100    | 21.3 | 21.0 | 21.9 | 20.8 | 23.7 | 19.3 | 23.5 | 23.2 | 21.8 |
| Mg/(Mg+Fe)×100   | 27.3 | 26.6 | 36.1 | 26.1 | 17.9 | 26.6 | 32.4 | 20.0 | 26.6 |

Table A-1 (con'd)

Specimen 209C

## Assemblage

Quartz-muscovite-biotite-kyanite-staurolite-plagioclase-opaque

## Muscovite Analyses

| Analysis No                    | 1     | 2     | 3     | 4     | 5     | 6     | 7     | 8     |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio                        | > 15  | 5     | 7     | 2     | > 10  | 10    | 10    | 4     | MEAN  |
| SiO <sub>2</sub>               | 47.40 | 47.64 | 47.54 | 47.38 | 47.57 | 47.34 | 46.81 | 47.72 | 47.42 |
| TiO <sub>2</sub>               | 0.23  | 0.37  | 0.44  | 0.32  | 0.21  | 0.43  | 0.39  | 0.37  | 0.34  |
| Al <sub>2</sub> O <sub>3</sub> | 35.30 | 35.34 | 35.03 | 35.53 | 35.22 | 35.12 | 34.81 | 35.25 | 35.20 |
| Fe(total)                      | 2.12  | 2.17  | 2.07  | 1.79  | 2.11  | 2.17  | 2.61  | 1.87  | 2.11  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.49  | 0.33  | 0.34  | 0.51  | 0.44  | 0.56  | 0.54  | 0.44  | 0.45  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.89  | 1.79  | 2.04  | 2.06  | 2.01  | 1.98  | 2.27  | 1.91  | 1.99  |
| K <sub>2</sub> O               | 7.90  | 7.68  | 7.85  | 7.73  | 7.75  | 7.71  | 7.87  | 7.76  | 7.78  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.33 | 95.32 | 95.31 | 95.32 | 95.31 | 95.31 | 95.30 | 95.32 |       |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.25 | 6.26 | 6.27 | 6.25 | 6.26 | 6.25 | 6.21 | 6.28 | 6.25 |
| Al <sup>iv</sup> | 1.75 | 1.74 | 1.73 | 1.77 | 1.74 | 1.75 | 1.79 | 1.72 | 1.75 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.73 | 3.74 | 3.71 | 3.74 | 3.73 | 3.72 | 3.65 | 3.75 | 3.72 |
| Ti               | 0.02 | 0.04 | 0.05 | 0.03 | 0.02 | 0.02 | 0.04 | 0.04 | 0.03 |
| **Fe             | 0.23 | 0.24 | 0.23 | 0.20 | 0.23 | 0.24 | 0.29 | 0.20 | 0.23 |
| Mg               | 0.10 | 0.07 | 0.07 | 0.10 | 0.09 | 0.11 | 0.11 | 0.08 | 0.09 |
| Σ                | 4.08 | 4.09 | 4.06 | 4.07 | 4.07 | 4.09 | 4.09 | 4.07 | 4.08 |
| Na               | 0.48 | 0.45 | 0.53 | 0.53 | 0.51 | 0.51 | 0.58 | 0.48 | 0.51 |
| K                | 1.33 | 1.30 | 1.31 | 1.30 | 1.30 | 1.30 | 1.34 | 1.30 | 1.31 |
| Σ                | 1.81 | 1.75 | 1.84 | 1.83 | 1.81 | 1.81 | 1.92 | 1.78 | 1.82 |
| Σ Al             | 5.48 | 5.48 | 5.44 | 5.51 | 5.47 | 5.47 | 5.44 | 5.47 | 5.47 |
| Σ Mg + Fe        | 0.33 | 0.31 | 0.30 | 0.30 | 0.32 | 0.35 | 0.40 | 0.28 | 0.32 |
| Na/(Na+K) x 100  | 26.5 | 25.7 | 28.8 | 29.0 | 28.3 | 28.2 | 30.2 | 27.0 | 28.0 |
| Mg/(Mg+Fe) x 100 | 30.3 | 22.6 | 23.3 | 33.3 | 28.1 | 31.4 | 27.5 | 28.6 | 28.1 |

Table A-1. (con'd)

## Specimen 36N

## Assemblage

Quartz-muscovite-biotite-kyanite-plagioclase-tourmaline-opaque

## Muscovite Analyses

| Analysis No                    | 1A    | 1B    | 2A    | 2B    | 3A    | 3B    | 4A    | 4B    |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio                        | > 15  | > 15  | 4     | 4     | 6     | 6     | > 15  | > 15  | MEAN* |
| T                              |       |       |       |       |       |       |       |       |       |
| SiO <sub>2</sub>               | 47.22 | 53.05 | 47.07 | 47.28 | 47.79 | 47.64 | 47.21 | 47.80 | 47.43 |
| TiO <sub>2</sub>               | 0.28  | 0.37  | 0.40  | 0.50  | 0.37  | 0.42  | 0.31  | 0.00  | 0.33  |
| Al <sub>2</sub> O <sub>3</sub> | 34.36 | 30.56 | 34.28 | 34.56 | 35.02 | 34.88 | 34.12 | 34.27 | 34.50 |
| Fe (total)                     | 2.72  | 1.96  | 2.60  | 2.42  | 2.01  | 2.12  | 2.61  | 2.30  | 2.40  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.69  | 0.72  | 0.79  | 0.65  | 0.41  | 0.43  | 0.84  | 0.57  | 0.63  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.75  | 1.42  | 1.78  | 1.66  | 1.48  | 1.76  | 1.76  | 1.81  | 1.71  |
| K <sub>2</sub> O               | 8.30  | 7.24  | 8.40  | 8.24  | 8.23  | 8.08  | 8.46  | 8.58  | 8.33  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.32 | 95.32 | 95.32 | 95.31 | 95.31 | 95.33 | 95.31 | 95.33 |       |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.26 | 7.04 | 6.25 | 6.26 | 6.29 | 6.28 | 6.27 | 6.33 | 6.28 |
| Al <sup>iv</sup> | 1.74 | 0.96 | 1.75 | 1.74 | 1.71 | 1.72 | 1.73 | 1.67 | 1.72 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.63 | 3.91 | 3.61 | 3.65 | 3.73 | 3.70 | 3.61 | 3.68 | 3.65 |
| Ti               | 0.03 | 0.04 | 0.04 | 0.05 | 0.04 | 0.04 | 0.03 | 0.00 | 0.03 |
| **Fe             | 0.30 | 0.22 | 0.29 | 0.27 | 0.22 | 0.23 | 0.29 | 0.26 | 0.27 |
| Mg               | 0.14 | 0.14 | 0.16 | 0.13 | 0.08 | 0.08 | 0.17 | 0.11 | 0.12 |
| Σ                | 0.44 | 0.36 | 0.45 | 0.40 | 0.30 | 0.31 | 0.56 | 0.37 | 0.40 |
| Na               | 0.46 | 0.37 | 0.46 | 0.42 | 0.38 | 0.45 | 0.46 | 0.47 | 0.44 |
| K                | 1.40 | 1.23 | 1.43 | 1.39 | 1.39 | 1.36 | 1.43 | 1.44 | 1.41 |
| Σ                | 1.86 | 1.60 | 1.89 | 1.81 | 1.77 | 1.81 | 1.89 | 1.91 | 1.85 |
| Σ Al             | 5.37 | 4.87 | 5.36 | 5.39 | 5.44 | 5.42 | 5.34 | 5.35 | 5.38 |
| Σ Mg + Fe        | 0.44 | 0.36 | 0.45 | 0.40 | 0.30 | 0.31 | 0.56 | 0.37 | 0.40 |
| Na/(Na+K) x 100  | 24.7 | 23.1 | 24.3 | 23.2 | 21.5 | 24.9 | 24.3 | 24.6 | 23.9 |
| Mg/(Mg+Fe) x 100 | 31.8 | 38.9 | 35.6 | 32.5 | 26.7 | 25.8 | 30.4 | 29.7 | 31.4 |

\* excluding analysis 1B.

Table A-1 (con'd)

183C

## Assemblage

Quartz-muscovite-plagioclase-sillimanite-tourmaline

## Muscovite Analyses

| Analysis No                    | 1A    | 1B    | 2A    | 2B    | 3A    | 3B    | 4A    | 4B    |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio                        | 4     | 4     | 1.75  | 1.75  | 9     | 9     | 1     | 1     | MEAN* |
| SiO <sub>2</sub>               | 47.20 | 45.26 | 47.11 | 46.93 | 47.04 | 47.20 | 47.45 | 47.29 | 47.17 |
| TiO <sub>2</sub>               | 0.53  | 0.35  | 0.48  | 0.39  | 0.25  | 0.31  | 0.43  | 0.28  | 0.38  |
| Al <sub>2</sub> O <sub>3</sub> | 53.58 | 53.08 | 53.37 | 53.48 | 53.91 | 53.59 | 53.43 | 53.55 | 53.54 |
| Fe(total)                      | 2.85  | 5.86  | 2.93  | 3.21  | 2.96  | 3.01  | 2.65  | 3.01  | 2.95  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.69  | 1.07  | 0.89  | 0.94  | 0.77  | 0.91  | 0.82  | 0.57  | 0.81  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.10  | 1.14  | 1.17  | 1.08  | 0.93  | 0.99  | 1.12  | 1.23  | 1.09  |
| K <sub>2</sub> O               | 9.36  | 8.56  | 9.36  | 9.29  | 9.45  | 9.31  | 9.40  | 9.38  | 9.36  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.31 | 95.32 | 95.31 | 95.32 | 95.31 | 95.32 | 95.30 | 95.31 |       |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.29 | 6.12 | 6.29 | 6.27 | 6.27 | 6.29 | 6.32 | 6.31 | 6.29 |
| Al <sup>iv</sup> | 1.71 | 1.88 | 1.71 | 1.73 | 1.73 | 1.71 | 1.68 | 1.69 | 1.71 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.57 | 3.40 | 3.54 | 3.54 | 3.60 | 3.57 | 3.57 | 3.59 | 3.57 |
| Ti               | 0.05 | 0.04 | 0.05 | 0.04 | 0.03 | 0.03 | 0.04 | 0.03 | 0.04 |
| **Fe             | 0.32 | 0.64 | 0.33 | 0.36 | 0.33 | 0.34 | 0.30 | 0.34 | 0.33 |
| Mg               | 0.14 | 0.22 | 0.18 | 0.19 | 0.15 | 0.18 | 0.16 | 0.12 | 0.16 |
| Σ                | 4.08 | 4.33 | 4.10 | 4.13 | 4.11 | 4.12 | 4.07 | 4.08 | 4.10 |
| Na               | 0.29 | 0.29 | 0.31 | 0.28 | 0.24 | 0.26 | 0.29 | 0.32 | 0.28 |
| K                | 1.59 | 1.47 | 1.59 | 1.59 | 1.60 | 1.59 | 1.60 | 1.60 | 1.59 |
| Σ                | 1.88 | 1.76 | 1.90 | 1.87 | 1.84 | 1.85 | 1.89 | 1.92 | 1.88 |
| Σ Al             | 5.28 | 5.28 | 5.25 | 5.27 | 5.33 | 5.28 | 5.25 | 5.28 | 5.28 |
| Σ Mg + Fe        | 0.46 | 0.89 | 0.51 | 0.55 | 0.48 | 0.52 | 0.46 | 0.46 | 0.49 |
| Na/(Na+K) x 100  | 15.4 | 16.5 | 16.3 | 15.0 | 13.0 | 14.1 | 15.3 | 16.7 | 15.1 |
| Mg/(Mg+Fe) x 100 | 30.4 | 24.7 | 35.3 | 34.6 | 31.3 | 34.6 | 34.8 | 26.1 | 32.4 |

\* excluding analysis 1B.

Table A-1-(con'd)

112A

Assemblage

Quartz-muscovite-sillimanite-plagioclase-opaque-tourmaline

## Muscovite Analyses

| Analysis No                    | 1A    | 1B    | 2A    | 2B    | 3A    | 3B    | 4A    | 4B    |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio                        | 3     | 3     | 1.5   | 1.5   | 5     | 5     | 3.5   | 3.5   | MEAN  |
| SiO <sub>2</sub>               | 48.35 | 47.29 | 47.17 | 47.50 | 47.21 | 47.58 | 47.33 | 47.29 | 47.47 |
| TiO <sub>2</sub>               | 0.34  | 0.00  | 0.63  | 0.39  | 0.61  | 0.41  | 0.64  | 0.47  | 0.44  |
| Al <sub>2</sub> O <sub>3</sub> | 33.23 | 33.88 | 34.04 | 33.76 | 33.19 | 33.40 | 33.25 | 33.47 | 33.53 |
| Fe (total)                     | 2.55  | 2.77  | 2.49  | 2.69  | 2.92  | 2.79  | 2.90  | 2.93  | 2.76  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.63  | 0.90  | 0.84  | 0.85  | 0.92  | 0.84  | 0.83  | 0.89  | 0.84  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.04  | 1.04  | 0.97  | 0.76  | 0.92  | 0.91  | 0.90  | 0.89  | 0.93  |
| K <sub>2</sub> O               | 9.19  | 9.45  | 9.18  | 9.38  | 9.54  | 9.39  | 9.46  | 9.38  | 9.37  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.33 | 95.33 | 95.32 | 95.33 | 95.31 | 95.32 | 95.31 | 95.32 |       |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.41 | 6.30 | 6.27 | 6.31 | 6.30 | 6.33 | 6.31 | 6.30 | 6.32 |
| Al <sup>iv</sup> | 1.59 | 1.70 | 1.73 | 1.69 | 1.70 | 1.67 | 1.69 | 1.70 | 1.68 |
| $\Sigma$         | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.61 | 3.62 | 3.60 | 3.60 | 3.52 | 3.57 | 3.54 | 3.56 | 3.58 |
| Ti               | 0.03 | 0.00 | 0.07 | 0.04 | 0.06 | 0.04 | 0.06 | 0.05 | 0.04 |
| **Fe             | 0.28 | 0.31 | 0.27 | 0.30 | 0.33 | 0.31 | 0.32 | 0.33 | 0.31 |
| Mg               | 0.12 | 0.18 | 0.17 | 0.17 | 0.18 | 0.17 | 0.17 | 0.18 | 0.17 |
| $\Sigma$         | 4.04 | 4.11 | 4.11 | 4.11 | 4.09 | 4.09 | 4.09 | 4.12 | 4.10 |
| Na               | 0.27 | 0.28 | 0.24 | 0.20 | 0.24 | 0.23 | 0.23 | 0.23 | 0.24 |
| K                | 1.55 | 1.60 | 1.55 | 1.60 | 1.62 | 1.60 | 1.60 | 1.60 | 1.59 |
| $\Sigma$         | 1.82 | 1.88 | 1.79 | 1.80 | 1.86 | 1.83 | 1.83 | 1.83 | 1.83 |
| $\Sigma$ Al      | 5.20 | 5.32 | 5.33 | 5.29 | 5.22 | 5.24 | 5.23 | 5.26 | 5.26 |
| $\Sigma$ Mg + Fe | 0.40 | 0.49 | 0.44 | 0.47 | 0.51 | 0.48 | 0.49 | 0.51 | 0.47 |
| Na/(Na+K) x 100  | 14.8 | 14.9 | 13.4 | 11.1 | 12.9 | 12.6 | 12.6 | 12.6 | 13.1 |
| Mg/(Mg+Fe) x 100 | 30.0 | 36.7 | 38.6 | 36.2 | 35.3 | 35.4 | 34.7 | 35.3 | 35.3 |

Table A-1

243C

Assemblage

Quartz-muscovite-biotite-sillimanite-plagioclase-opaque

## Muscovite Analyses

| Analysis No                    | 1A    | 1B    | 2A    | 2B    | 3A    | 3B    | 4A    | 4B    |       |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| L ratio                        | 10    | 10    | 1     | 1     | 6     | 6     | 4     | 4     | MEAN  |
| SiO <sub>2</sub>               | 47.39 | 46.91 | 47.46 | 47.35 | 47.53 | 47.35 | 47.93 | 46.88 | 47.35 |
| TiO <sub>2</sub>               | 1.01  | 0.67  | 0.60  | 0.49  | 0.77  | 0.73  | 0.89  | 0.77  | 0.74  |
| Al <sub>2</sub> O <sub>3</sub> | 33.12 | 33.74 | 34.11 | 34.39 | 33.66 | 34.27 | 33.51 | 33.79 | 33.82 |
| Fe (total)                     | 2.51  | 2.72  | 2.56  | 2.08  | 2.40  | 2.37  | 2.38  | 2.71  | 2.46  |
| MnO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| MgO                            | 0.81  | 0.67  | 0.59  | 0.79  | 0.81  | 0.57  | 0.48  | 0.96  | 0.71  |
| CaO                            | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Na <sub>2</sub> O              | 1.34  | 1.21  | 1.16  | 1.18  | 1.00  | 0.95  | 0.97  | 1.27  | 1.13  |
| K <sub>2</sub> O               | 9.16  | 9.41  | 8.83  | 9.04  | 9.14  | 9.07  | 9.16  | 8.94  | 9.09  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  |
| Anhydrous total                | 95.34 | 95.33 | 95.31 | 95.32 | 95.31 | 95.31 | 95.32 | 95.32 |       |

## Formula based on 22 oxygens

|                  |      |      |      |      |      |      |      |      |      |
|------------------|------|------|------|------|------|------|------|------|------|
| Si <sup>iv</sup> | 6.31 | 6.25 | 6.30 | 6.27 | 6.31 | 6.28 | 6.35 | 6.24 | 6.29 |
| Al <sup>iv</sup> | 1.69 | 1.75 | 1.70 | 1.73 | 1.69 | 1.72 | 1.65 | 1.76 | 1.71 |
| Σ                | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 | 8.00 |
| Al <sup>vi</sup> | 3.51 | 3.56 | 3.64 | 3.64 | 3.58 | 3.64 | 3.59 | 3.54 | 3.59 |
| Ti               | 0.10 | 0.07 | 0.06 | 0.05 | 0.08 | 0.07 | 0.09 | 0.08 | 0.08 |
| *Fe              | 0.28 | 0.31 | 0.28 | 0.23 | 0.27 | 0.27 | 0.27 | 0.31 | 0.28 |
| Mg               | 0.16 | 0.13 | 0.11 | 0.16 | 0.16 | 0.11 | 0.10 | 0.19 | 0.14 |
| Σ                | 4.05 | 4.07 | 4.09 | 4.08 | 4.09 | 4.09 | 4.05 | 4.12 | 4.08 |
| Na               | 0.35 | 0.32 | 0.30 | 0.30 | 0.26 | 0.24 | 0.24 | 0.34 | 0.29 |
| K                | 1.56 | 1.60 | 1.49 | 1.53 | 1.55 | 1.53 | 1.55 | 1.52 | 1.54 |
| Σ                | 1.91 | 1.92 | 1.79 | 1.83 | 1.81 | 1.77 | 1.79 | 1.86 | 1.84 |
| Σ Al             | 5.20 | 5.31 | 5.34 | 5.37 | 5.27 | 5.36 | 5.24 | 5.30 | 5.30 |
| Σ Mg + Fe        | 0.44 | 0.44 | 0.39 | 0.39 | 0.43 | 0.38 | 0.37 | 0.50 | 0.42 |
| Na/(Na+K)×100    | 18.3 | 16.7 | 16.8 | 16.4 | 14.4 | 13.6 | 13.4 | 18.3 | 16.0 |
| Mg/(Mg+Fe)×100   | 36.4 | 29.6 | 28.2 | 41.0 | 37.2 | 29.0 | 27.0 | 38.0 | 33.3 |

### Discussion of the data

With the exception of three analyses (6, 7 and 8) in 263G and one each in 36N (analysis 1B) and 183C (analysis 1B), the data appear to be comparable within themselves, and suitable for a study of compositional changes associated with crystal shape and metamorphic gradient. The anomalous analyses from the low grade specimen (263G) may be the result of an increase in phengite content, by which some Al is replaced by Si, though the Mg and Fe contents, which are generally higher in phengites do not reflect this. It is probable that the analyses 1B from 36N and 183C are faulty measurements, since in both cases they bear little relation to 1A; the higher Fe content of one of them may be the result of the beam being focused on a minute inclusion of an opaque iron oxide, of which there are many in these rocks. These anomalous data have therefore been omitted from the calculations of the means, since for comparative purposes they are not considered representative.

No consistent relationships were obtained between the composition and the shape of the crystals in the measurements from a single locality, indicating that, in the case of muscovite, chemistry is not a significant factor in determining crystal morphology. However between specimens some trends do emerge. Some of these are depicted graphically in figs. A-3 and A-4. Of the major elements the  $Si^{IV}$  and  $Al^{IV}$  contents

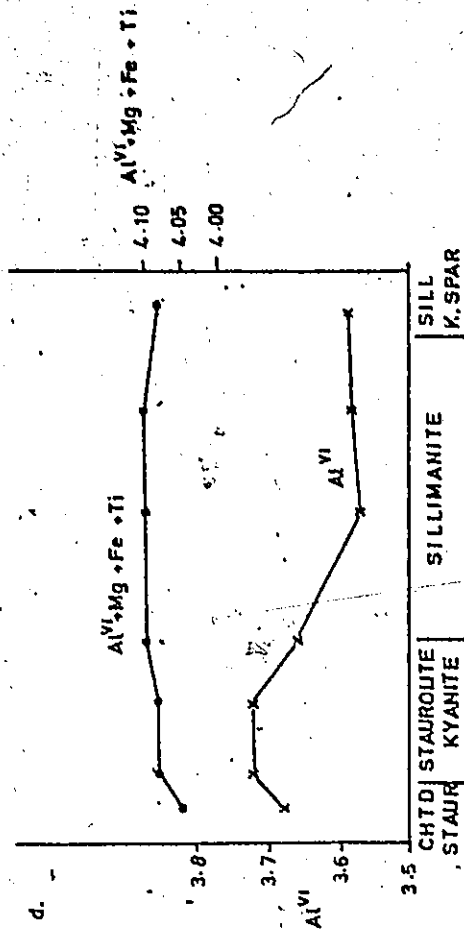
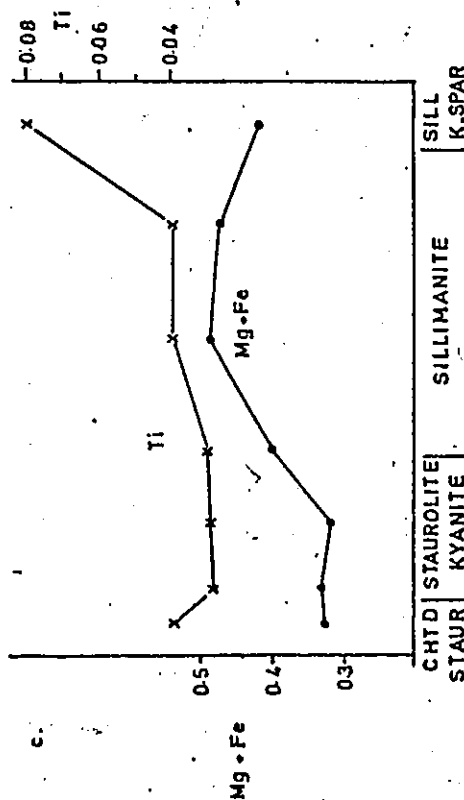
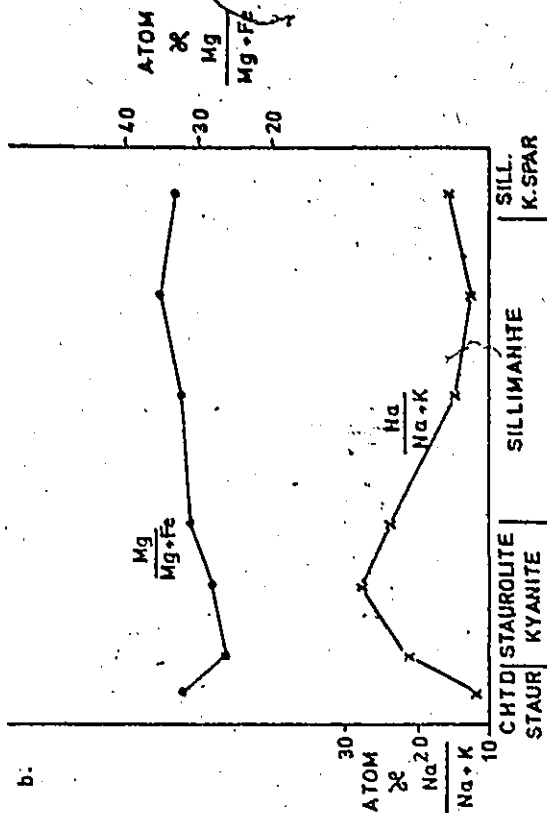
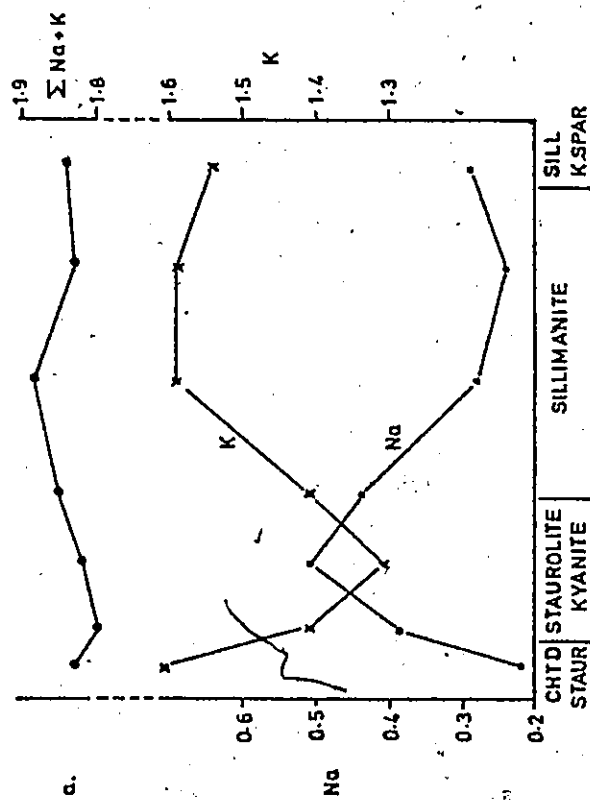


Fig. A.3 Analysis of the compositional variation of muscovite with grade of metamorphism. Ionic proportions of muscovite are calculated on the basis of 22 oxygens.

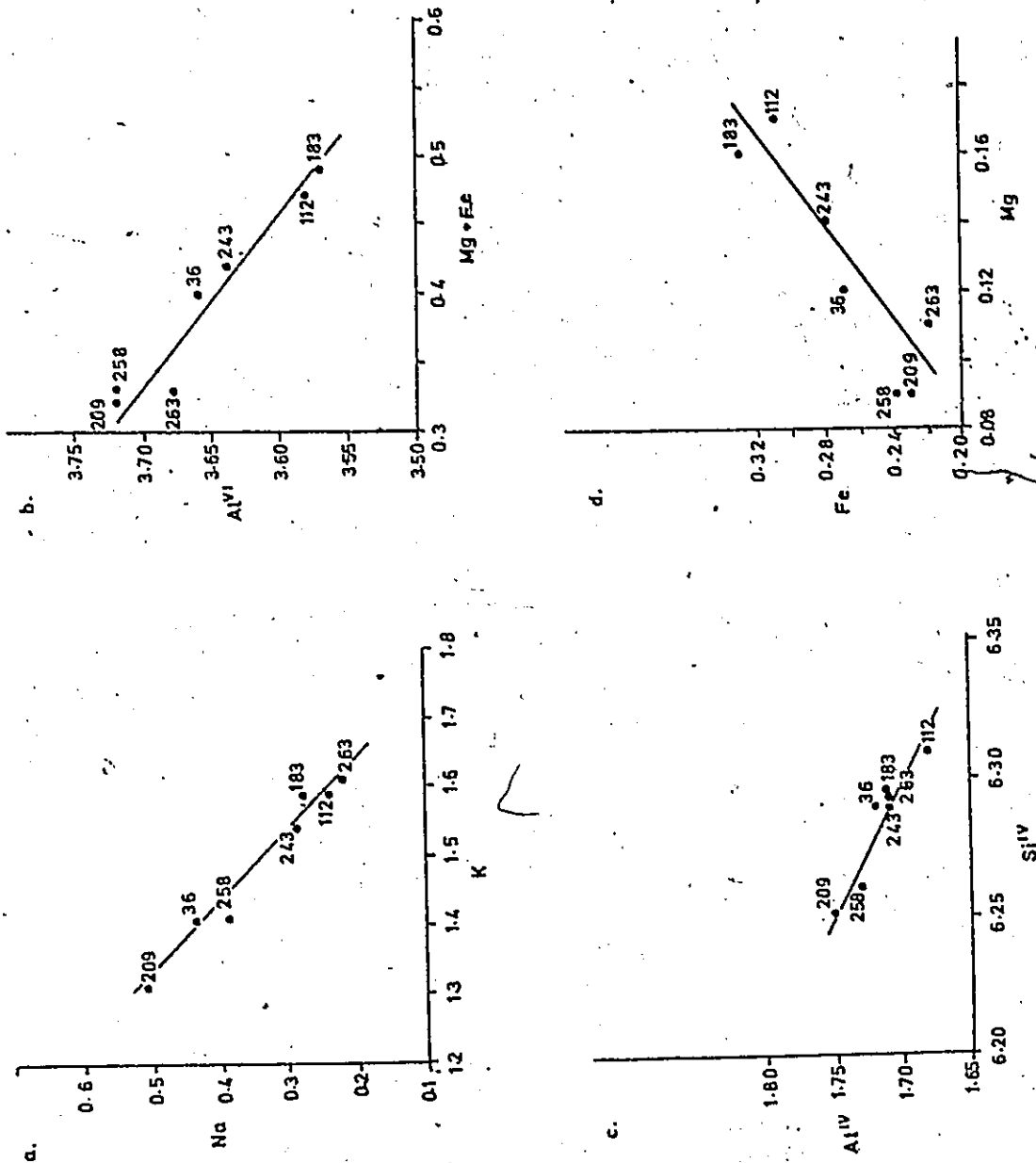


Fig. A.4 Analysis of the compositional variation of muscovite with grade of metamorphism. For location of specimens with reference to grade of metamorphism see fig. A-1. Lines between the points are drawn by eye.

of muscovite show only minor variation across the metamorphic gradient examined; when plotted against each other the points fall on a line, although not in order of metamorphic grade (fig. A-4c).  $Al^{VI}$  decreases with increasing metamorphic grade, at least between the staurolite-kyanite and sillimanite zones (fig. A-3d). Total alkalis,  $Na+K$ , remain approximately constant (fig. A-3a), and graphs for Na and K are inversely proportional reflecting the substitution of one element for the other. The ratio  $Na/(Na+K) \%$  increases from the chloritoid-staurolite zone to the staurolite-kyanite zone, and thereafter declines gradually (fig. A-3d). In fig. A-4a Na is plotted against K, and all specimens fall on a line, with the staurolite-kyanite zone specimens grouping at the top left and specimens from higher and lower zones clustering at the bottom right. Fe+Mg and Ti both increase with grade of metamorphism (fig. A-3c) in a trend which appears to compensate for the reduction in  $Al^{VI}$  in the higher grade zones (figs. A-3d and A-4d). The ratio  $Mg/(Mg+Fe) \%$  is lowest in the staurolite-kyanite zone, but increases only slightly to the sillimanite-K. feldspar zone. Fig. A-4d, in which Mg is plotted against Fe, shows that both elements increase with grade of metamorphism, as the high grade samples are grouped in the top right of the graph. This suggests that substitution of one element for the other may be limited, so some of the Fe present may be  $Fe^{+3}$ .

### Comparison with Other Data

The data of Guidotti (1973) on the compositional variation of muscovite with grade of metamorphism provide an excellent source for comparison. However it should be born in mind that (a) the pressure in the Maine examples is inferred to be somewhat lower than in the Ompah area, ie. below the aluminium silicate triple point (Guidotti 1970); and (b) the range of metamorphic grade sampled from the staurolite to upper sillimanite zones is slightly less than in this study; in particular, the lowest grade in the Ompah area, the chloritoid-staurolite zone, or equivalent, is not represented.

In general the analyses from the two areas are very similar, though the Maine specimens seem to be slightly lower in Si and higher in Al than the Ompah ones. This may be a function of the lower pressure environment, since Guidotti (1975) also includes for comparison some high pressure equivalent of his Maine specimens from Vermont and St. Paul Island Nova Scotia, which both contain a higher phengite content. Furthermore, sillimanite zone specimens from the northeast Dalradians of Scotland, inferred to have coexisted at lower temperature with andalusite, possess silica and alumina contents comparable to those in the Maine examples (Ashworth 1975).

Over the metamorphic grade studied, and in the high aluminium assemblages, Guidotti (1973) reports the following trends in muscovite composition with grade of metamorphism: "(a) the sum of Fe + Mg increases, as does Ti; (b) Na decreases and K increases, so that the ratio  $\frac{Na}{Na+K}$  decreases, and (c) there is a possible, but rather uncertain suggestion that Al, especially in the octahedral sites, decreases. There is also a similar suggestion for Si".

If the staurolite zone of Guidotti (1973) is taken as the equivalent of the staurolite-kyanite zone in the Ompah area, certain similarities appear between the two sets of analyses. Most conspicuously the declining Na/Na+K ratio with increasing metamorphic grade is common to both groups of samples in the medium and higher metamorphic zones. Interestingly, though, this trend reverses in the lowest grade specimens from the Ompah which have no metamorphic equivalent from Maine for comparison. Other trends common to the two sets of analyses are the increase in Mg + Fe and Ti, and the decrease in  $Al^{VI}$  with increasing grade of metamorphism. However no systematic trend in reduction of  $Al^{IV}$  and  $Si^{IV}$ , as tentatively reported by Guidotti (1973) was apparent in the Ompah examples.

Variation of the Na/Na+K ratio in limiting assemblages is partly controlled by the composition of coexisting plagioclase. As can be seen from fig. A-2, any change in the

composition of plagioclase will be reflected by reciprocal small changes in the composition of muscovite by movement of the tie line geometry around point Q (Guidotti 1973). Thus muscovite variation may be due, at least in part, to changes in plagioclase composition during progressive metamorphism. However the reversal in trend of the Na/Na+K ratio evident in the Ompah analyses (fig. A-3b) is more difficult to explain, since the Na content of plagioclase is not known to peak in the staurolite zone.

In the muscovite specimens from the Ompah area the decrease in  $Al^{IV}$  is approximately balanced by the increase in Mg, Fe and Ti (fig. A-3d): A similar situation was observed in analyses of muscovites from Maine by Guidotti (1973), who suggests that the following substitution may take place: "the increase in  $\Sigma Mg + Fe$  is not fully balanced by increase of Si and decrease in Al — that is an increase of the phengite content.....

... Instead some of the decrease of  $Al^{IV}$  that should result from increase of  $\Sigma (Mg+Fe)$  in the octahedral sites seems to be offset by  $(Ti^{+4})$ , requiring substitution of  $(Al^{+3})^{IV}$  for  $(Si^{+4})^{IV}$ ". Whether a similar network of substitutions occurred in the Ompah specimens is not known— however as already mentioned, the  $Si^{IV}$  content does not vary systematically with metamorphic grade, so it appears unlikely that it is greatly affected by the  $\Sigma Mg + Fe$ , as suggested by Guidotti.

## Conclusions

In limiting assemblages (1) muscovite crystals from a single specimen show only a small variation in composition; and composition is not a factor that determines crystal shape in muscovite.

(2) With increasing grade of metamorphism Mg, Fe and Ti in muscovite increase, and  $Al^{VI}$  may tend to decrease.

(3) The ratio  $\frac{Na}{Na+K}$  increases to the staurolite-kyanite zone, then decreases, and so may be a useful indicator of metamorphic grade.

(4) Phengite content shows slight, but unsystematic variation across the metamorphic gradient examined. All specimens are higher in Si and Fe and lower in Al than those of Guidotti (1973) and Ashworth (1975), however, whose measurements are from rocks inferred to have come from lower pressure regimes. Phengite content therefore appears to be more sensitive to the pressure than to the temperature of the metamorphic environment.

The observed variations in muscovite composition with grade of metamorphism are in general agreement with those of Guidotti (1973); however the wider range of metamorphic grade examined in this study has enabled the conclusions to be extended to lower grades where the trends are somewhat different.

FI

GE

IN

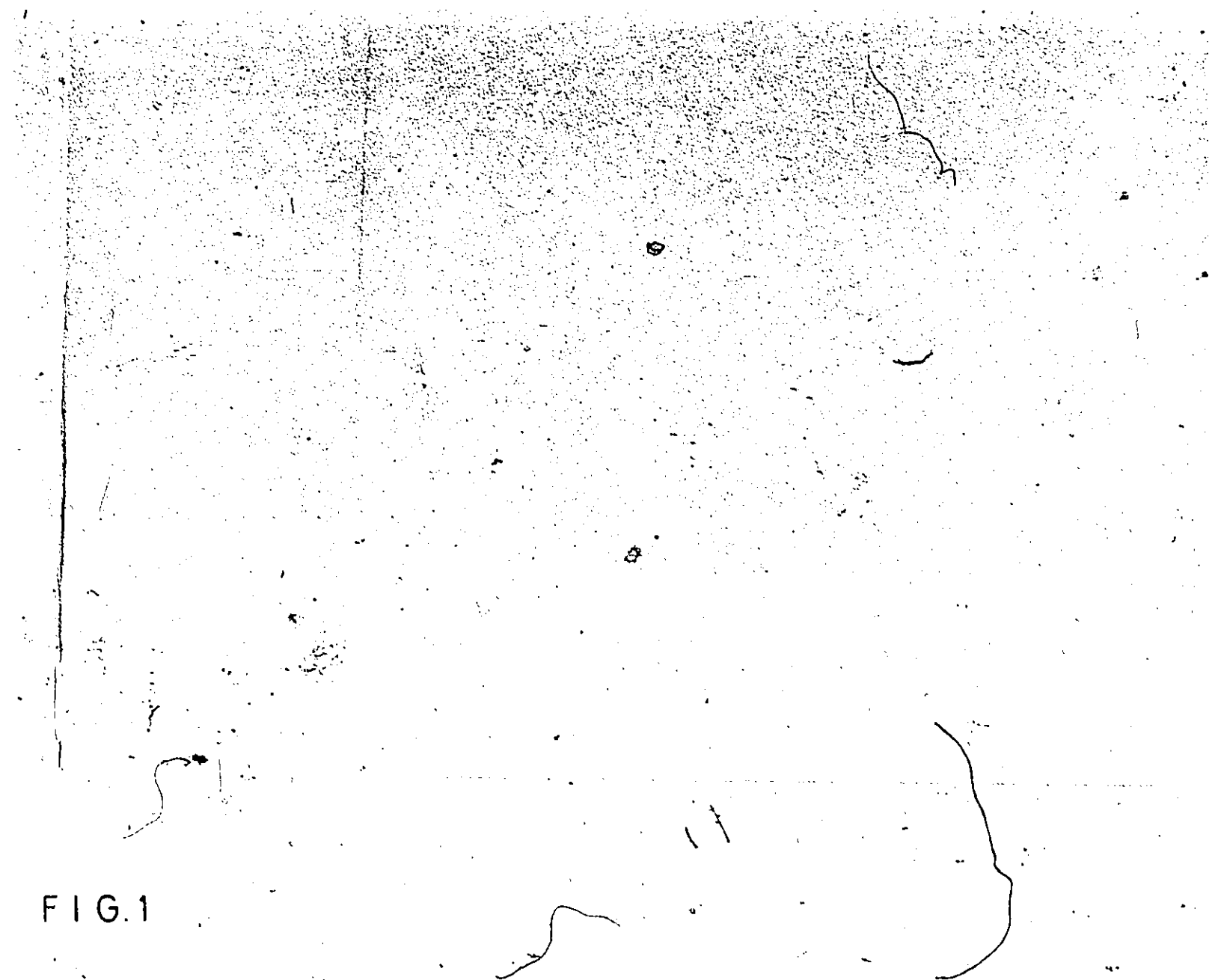


FIG.1

**GEOLOGICAL MAP OF THE PRE  
IN THE OMPAH AREA, S.E. ONT.**

GEOLOGY MODIFIED FROM SMITH (1958) BY T. RIVERS

BASE MAP FROM NATIONAL TOPOGRAPHIC 1:50,000 SE

**CAMBRIAN ROCKS**

**ARIO.**

1972.3.4.

RIES.

(30)

179

C

LAVAN  
LONG

REDHORSE  
L



59



G N E I S

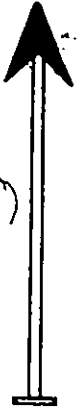
J

25

A

6.0

N



A B I N G E R

7a

G R A S T O N

7a.d

1781

A diagram showing a line segment with points B, E, and I. Point B is at the top left, E is in the middle, and I is at the bottom right. The letters G, N, S, and S are scattered around the diagram.

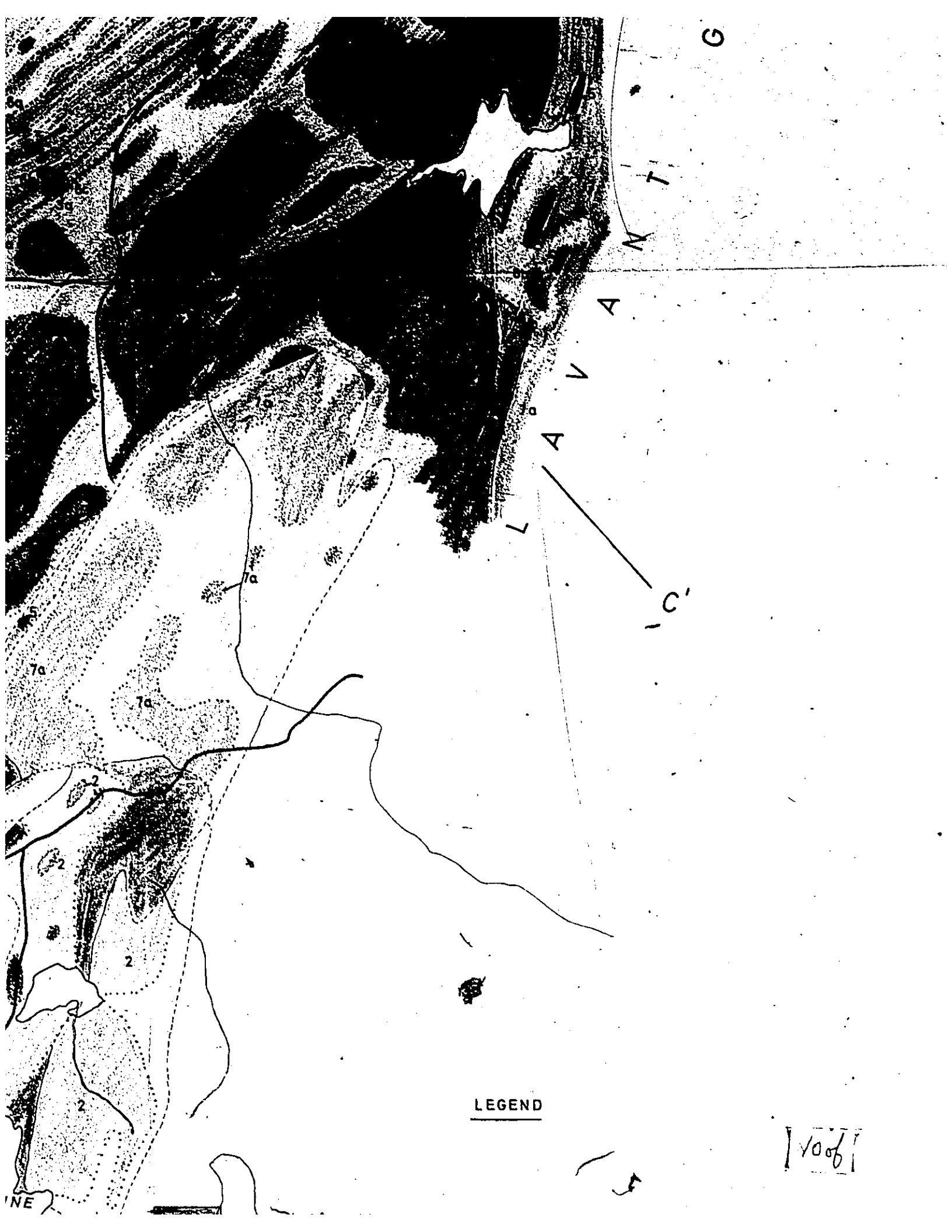
7c.7d

70

6 d

7d'



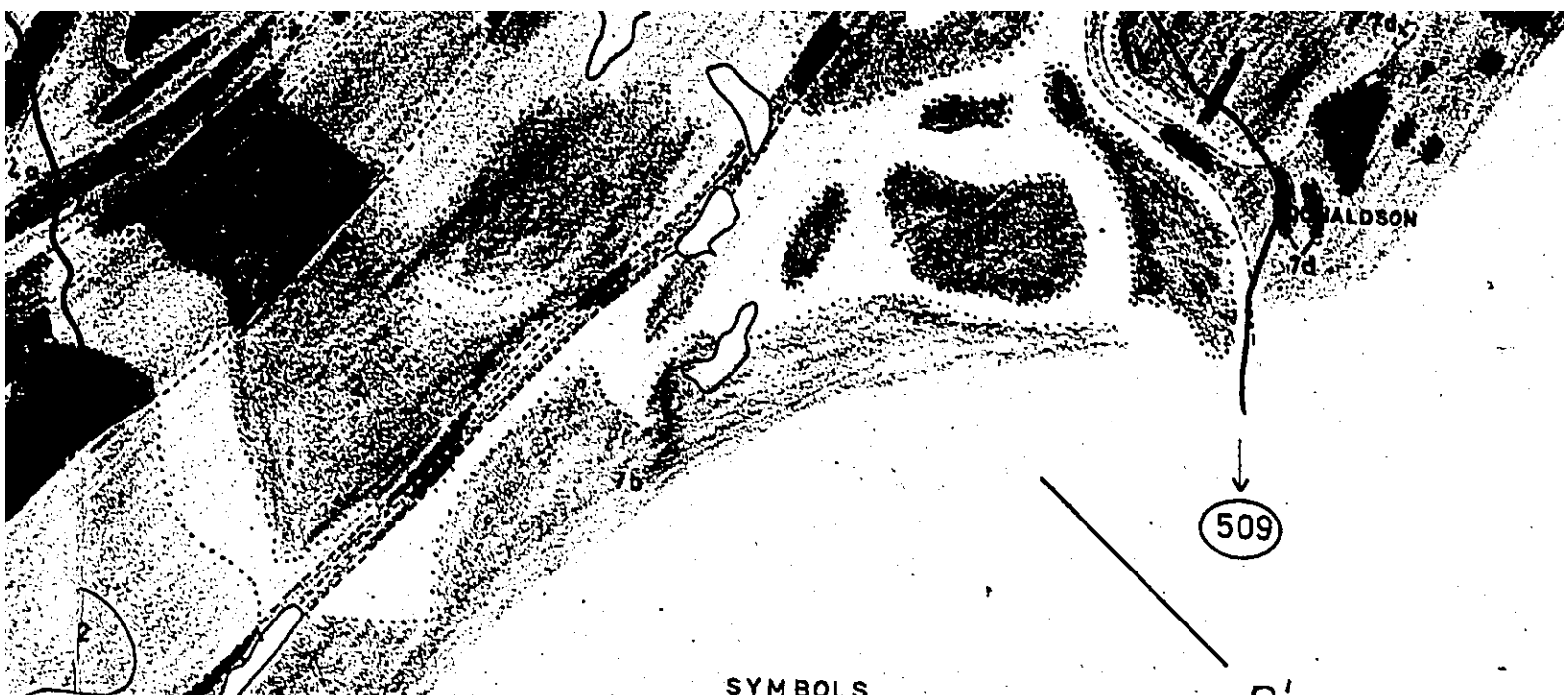




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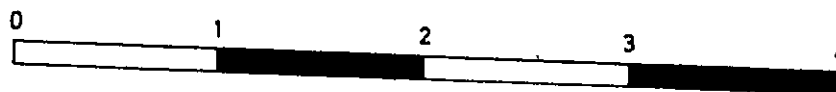


SYMBOLS

- Road.
- ^ ^ Hydro line.
- ⬢ Outcrop, including areas of semi-continuous outcrop.
- Geological boundary.
- - - Geological boundary (inferred).
- A A' Line of section.

SCALE

KILOMETRES.



[1408]



7a Pink, potassic granite and granite gneiss, esp. Abinger and Lavant plutons.

7b Grey granodioritic to trondhjemitic gneiss esp. Cross Lake body.

7c Migmatite

7d Pegmatite.



6a Biotite-quartz-plagioclase feldspar, hornblende, calcite, K-feldspar schist/gneiss.



6b Pelitic (muscovite-quartz-opaques-plagioclase feldspar) schist.

6c Biotite-quartz-carbonate schist.

6d Biotite-quartz-pyrite schist.



5 Calcitic and dolomitic marble, graphite, pyrite, quartz, diopside, tremolite, phlogopite, scapolite; includes some interlayered amphibolite.



4a Quartzite

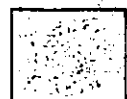
4b Calcareous quartzite.



3a Conglomerate - pebbles of quartz, quartzite, granite, marble and biotite schist in a schistose matrix of biotite, quartz, amphibole, plagioclase feldspar and calcite.

3b Conglomerate - boulders of granite in a pelitic schist matrix.

3c Conglomerate - pebbles of calc-silicate in a carbonate matrix.



2 Quartzofeldspathic hornblende-biotite gneiss, frequently layered and containing inclusions of diorite; includes thin interbeds of pelitic and psammitic schist.



1a Amphibolite and plagioclase feldspar, amphibole gneiss.

1b Metabasalt with relict plagioclase feldspar phenocrysts.

1c Amphibole-biotite gneiss and biotite-amphibole schist.

1d Quartzofeldspathic biotite gneiss, typically with a coarse grained quartz phase; associated with 1c.

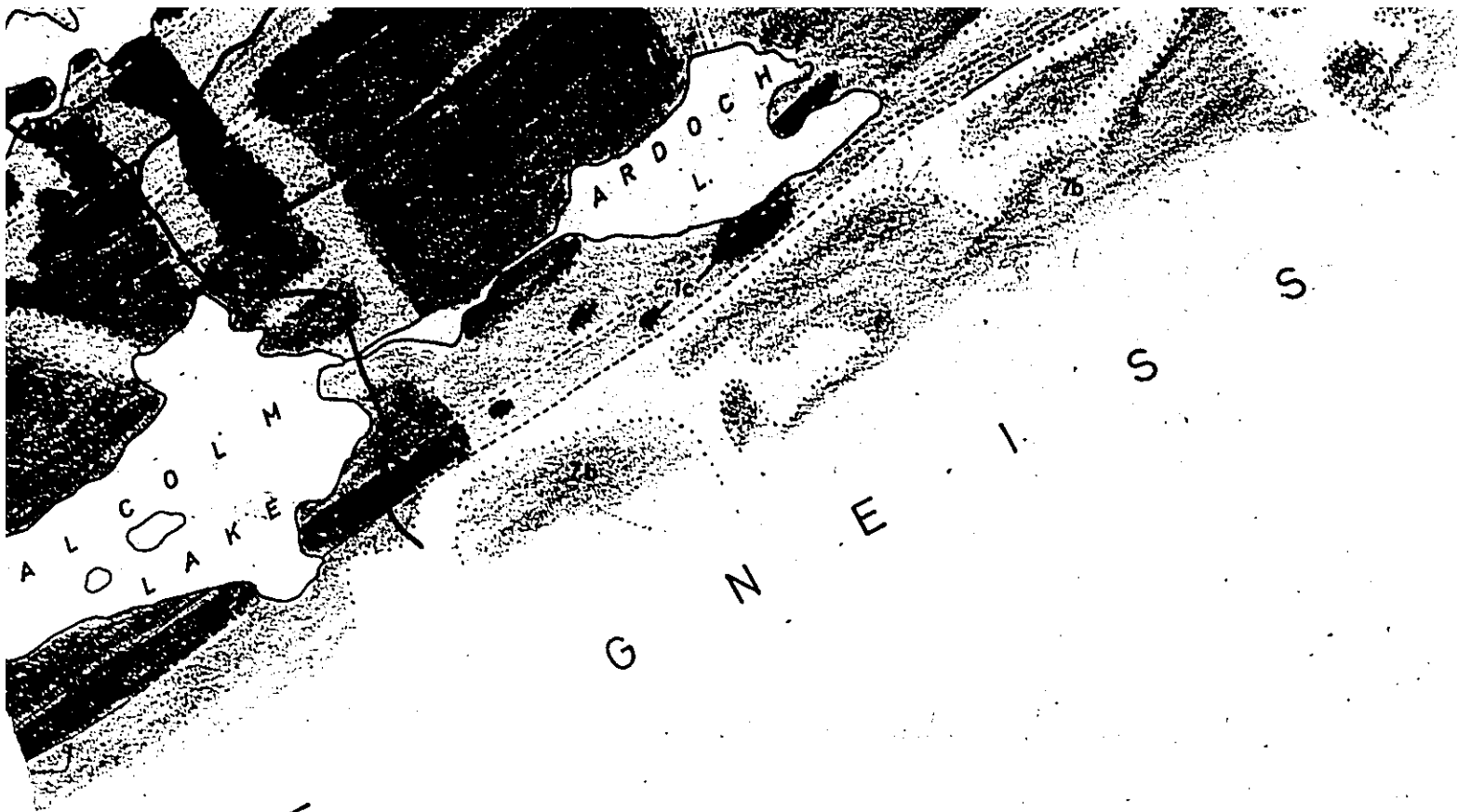
1e Chlorite schist.

15





17.F



180F

- 
- 3a Conglomerate - pebbles of quartz, quartzite, granite, marble and biotite schist in a schistose matrix of biotite, quartz, amphibole, plagioclase feldspar + calcite.
- 3b Conglomerate - boulders of granite in a pelitic schist matrix.
- 3c Conglomerate - pebbles of calc-silicate in a carbonate matrix.

- 
- 2 Quartzofeldspathic hornblende-biotite gneiss, frequently layered and containing inclusions of diorite; includes thin interbeds of pelitic and psammitic schist.

- 
- 1a Amphibolite and plagioclase feldspar, amphibole gneiss.
- 1b Metabasalt with relict plagioclase feldspar phenocrysts.
- 1c Amphibole-biotite gneiss and biotite-amphibole schist.
- 1d Quartzofeldspathic biotite gneiss, typically with a coarse grained quartz phase; associated with 1c.
- 1e Chlorite schist.

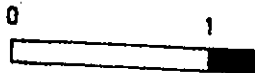
NOTE: Units not numbered in stratigraphic order.

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FIGURE 3.

# STRUCTURAL MAP (

SCALE:-



# OF THE OMPAH AREA

KILOMETRES

2

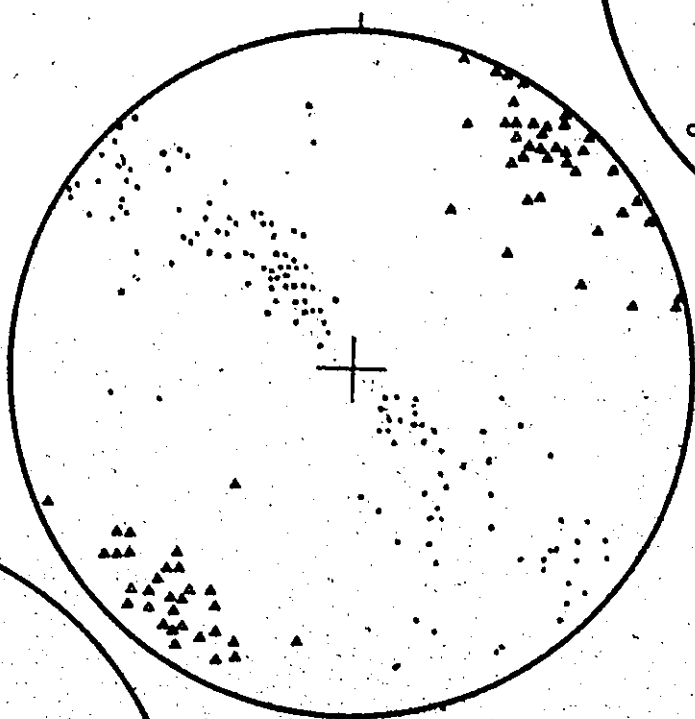
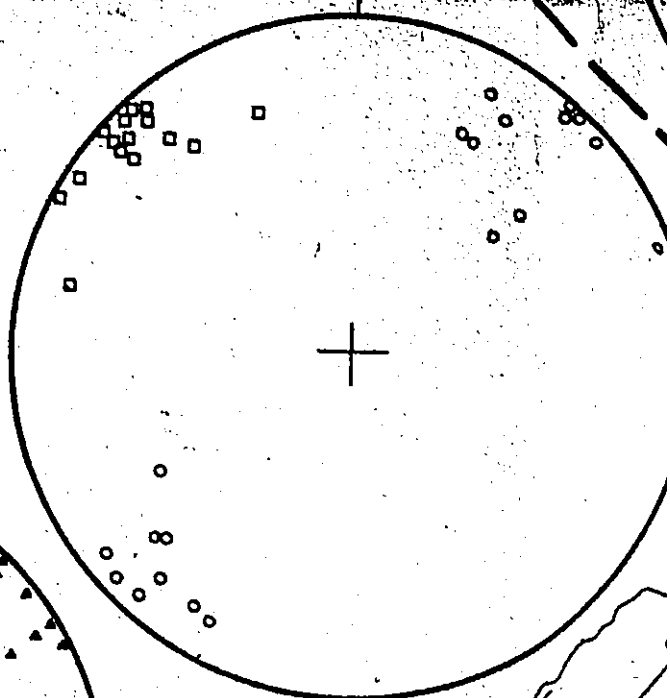
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4

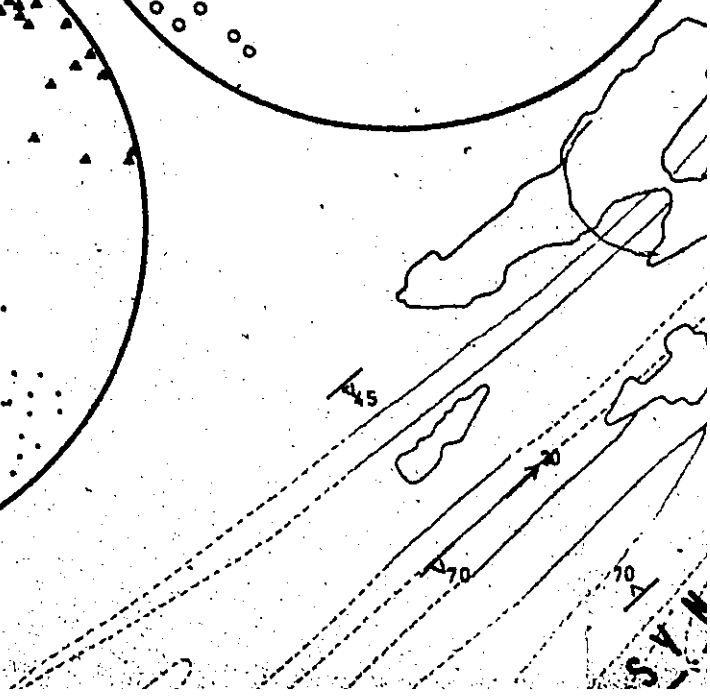
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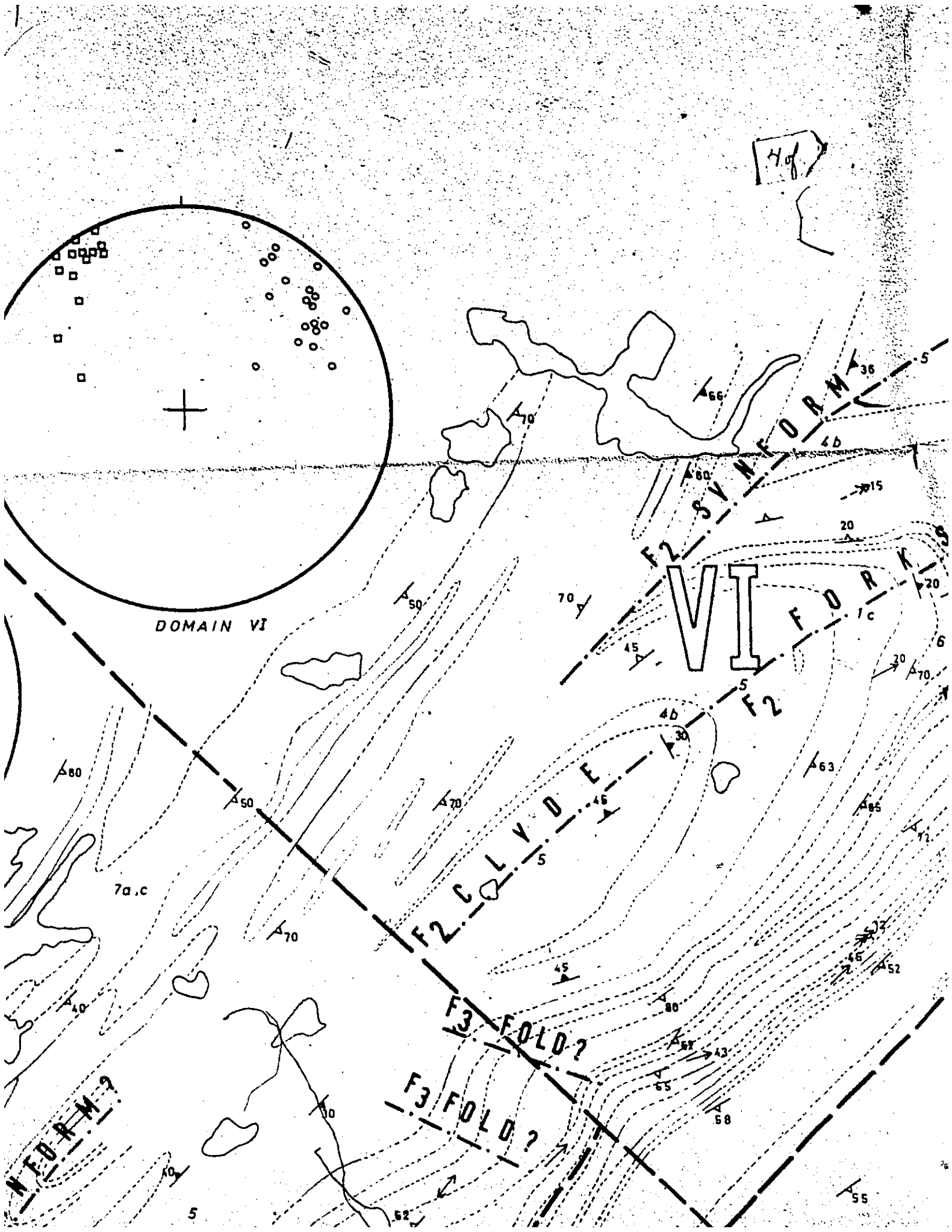
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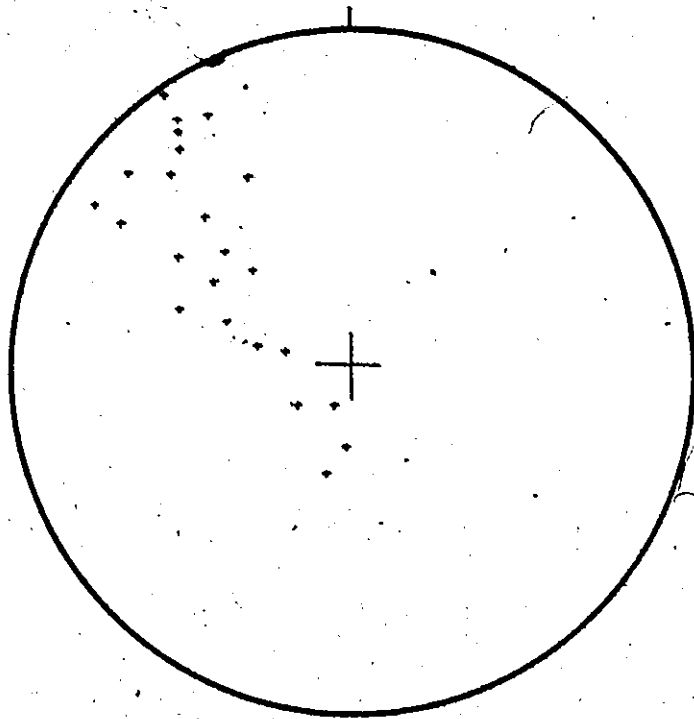
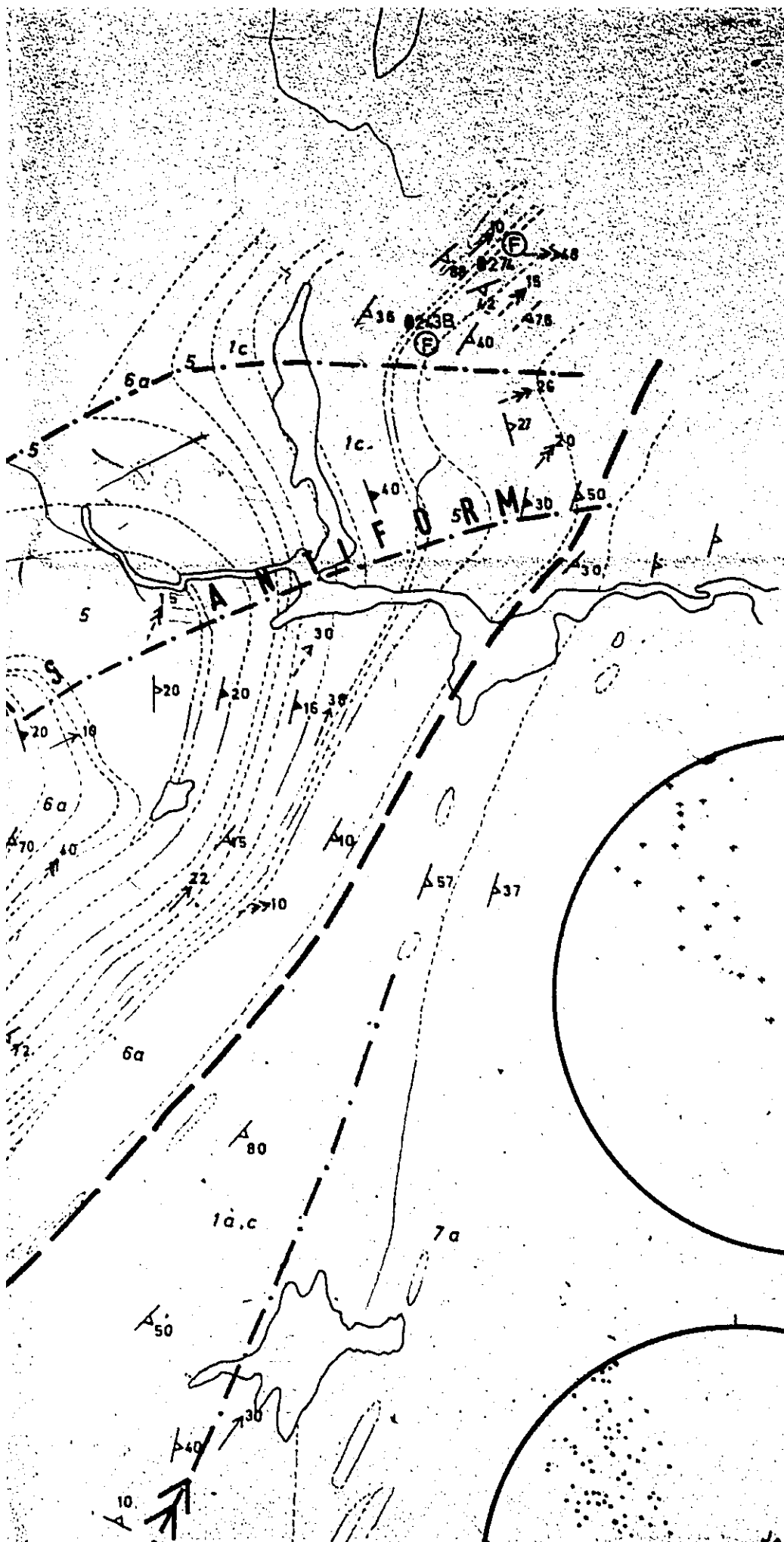
304



DOMAIN III







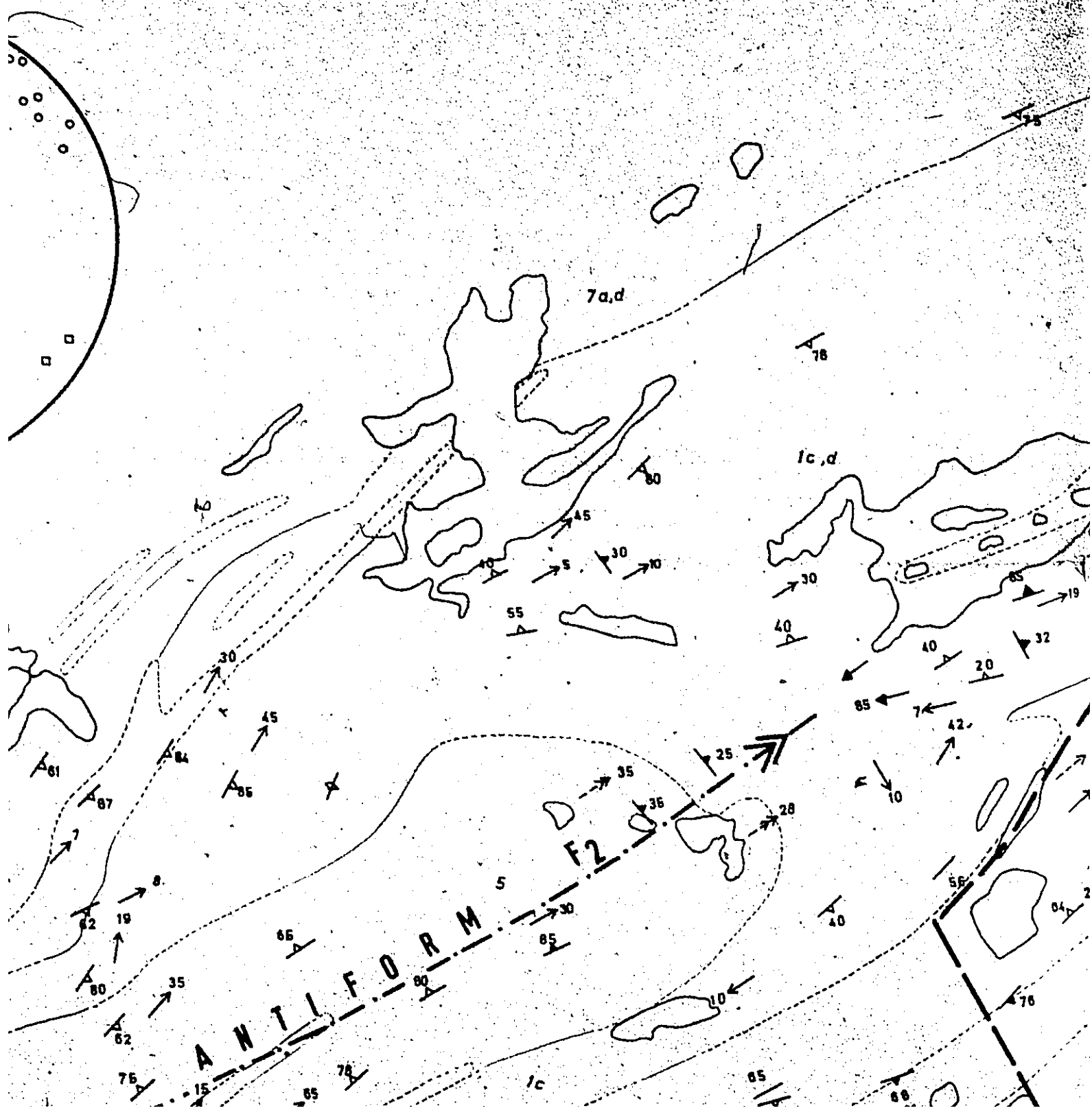
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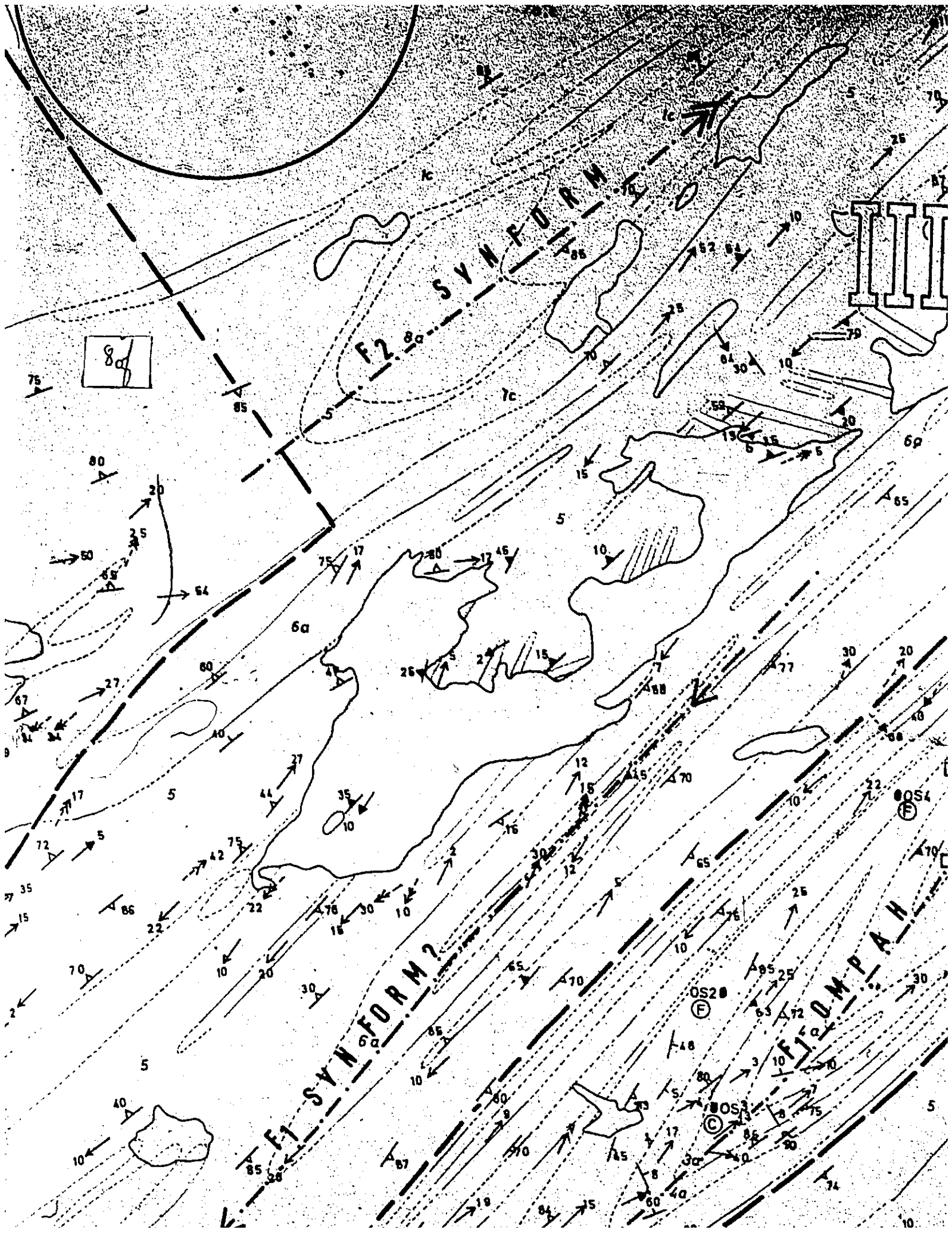
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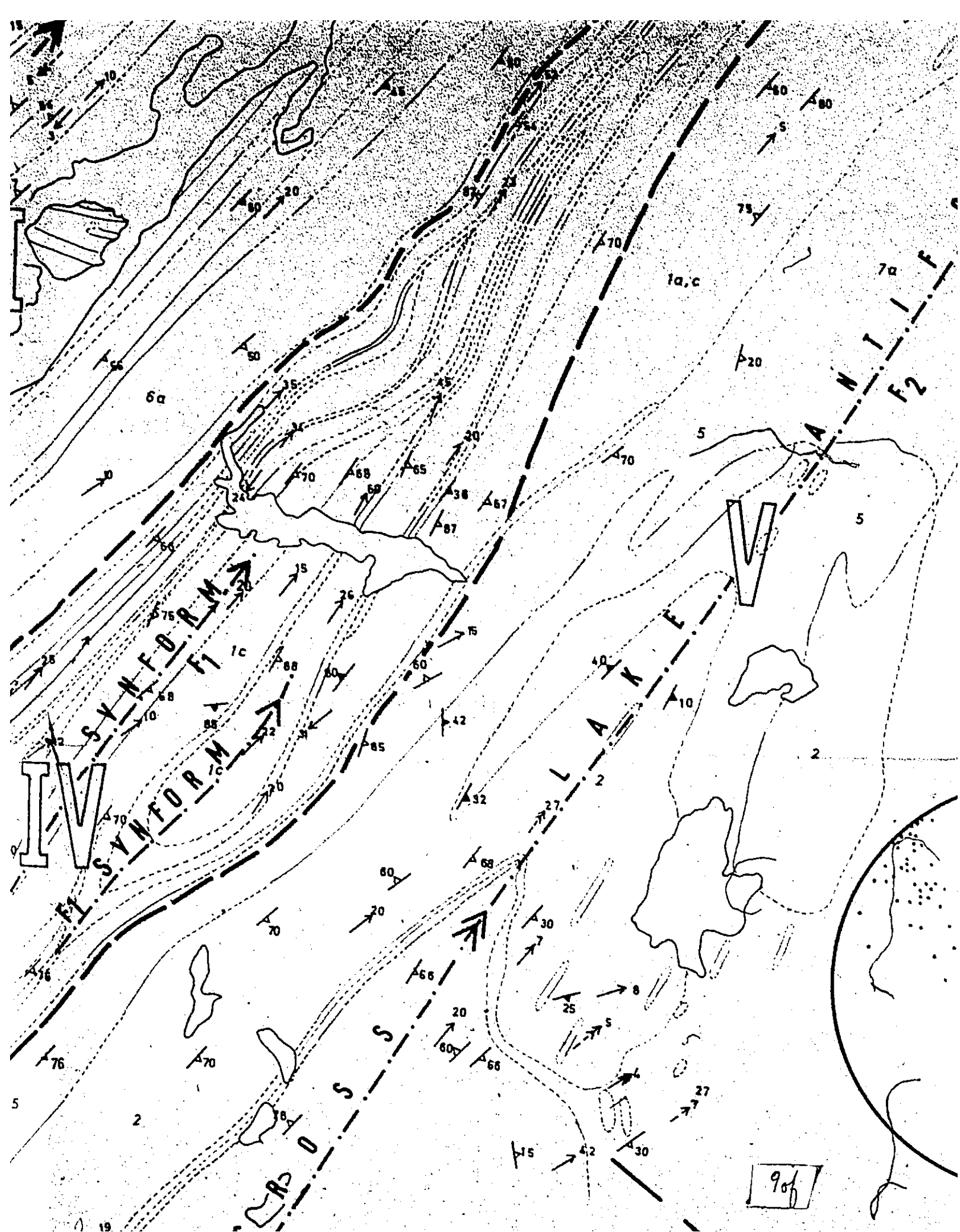


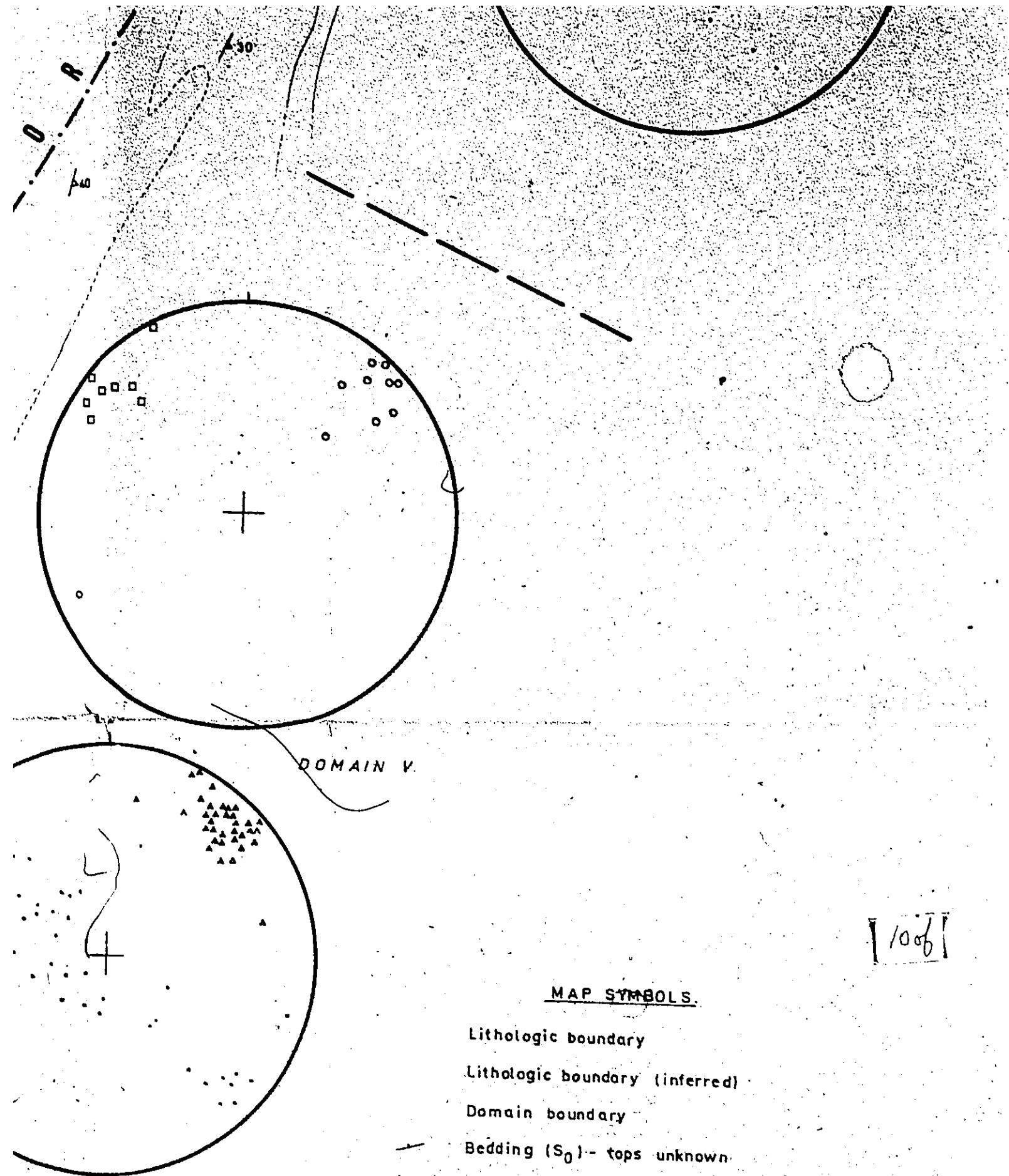
DOMAIN 1

70









# MAP SYMBOLS.

- Lithologic boundary
- - - Lithologic boundary (inferred)
- - - Domain boundary
- Bedding ( $S_0$ ) - tops unknown
- Bedding ( $S_0$ ) - facing in direction of dip.
- Bedding ( $S_0$ ) - facing in direction of loop, overturned.
- Foliation  $S_1$

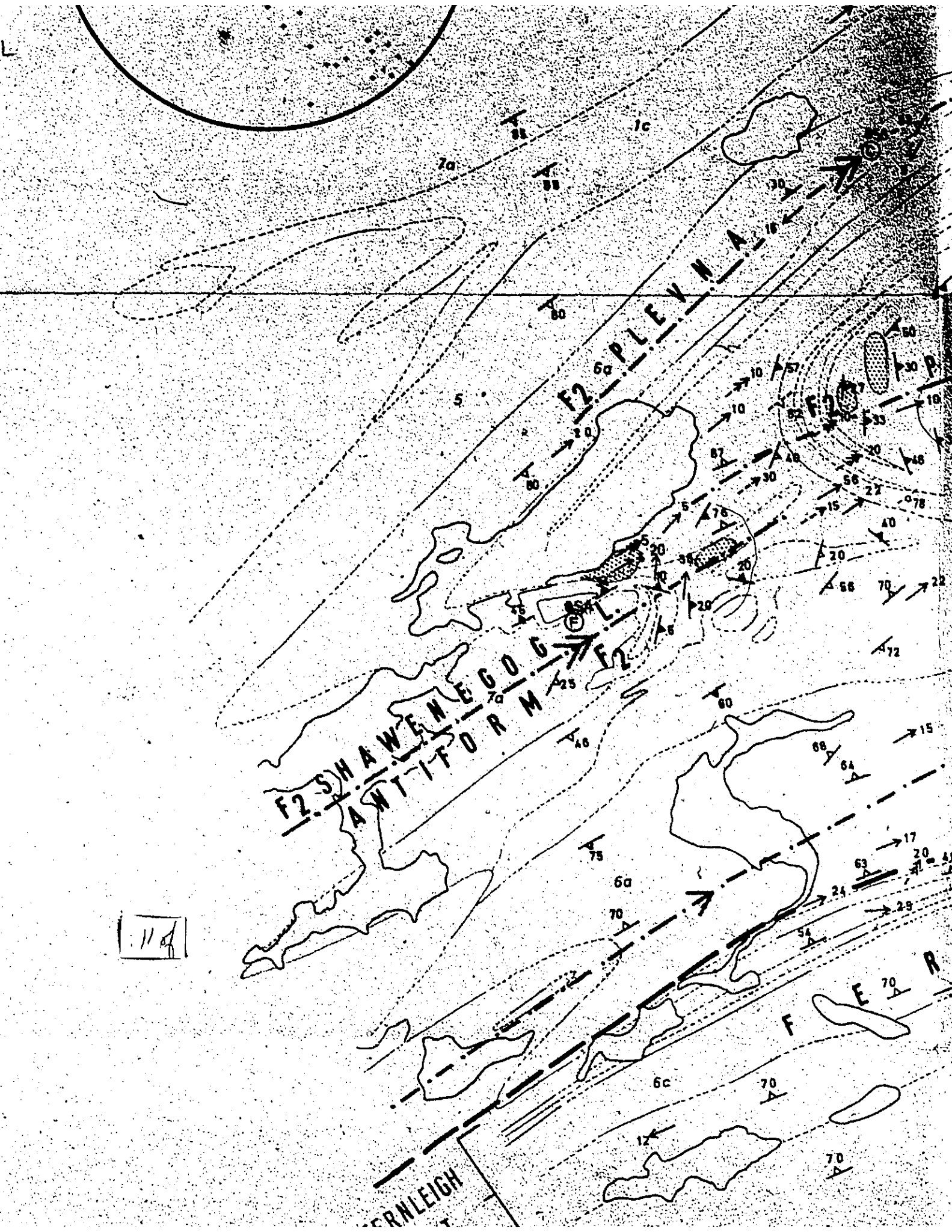
114

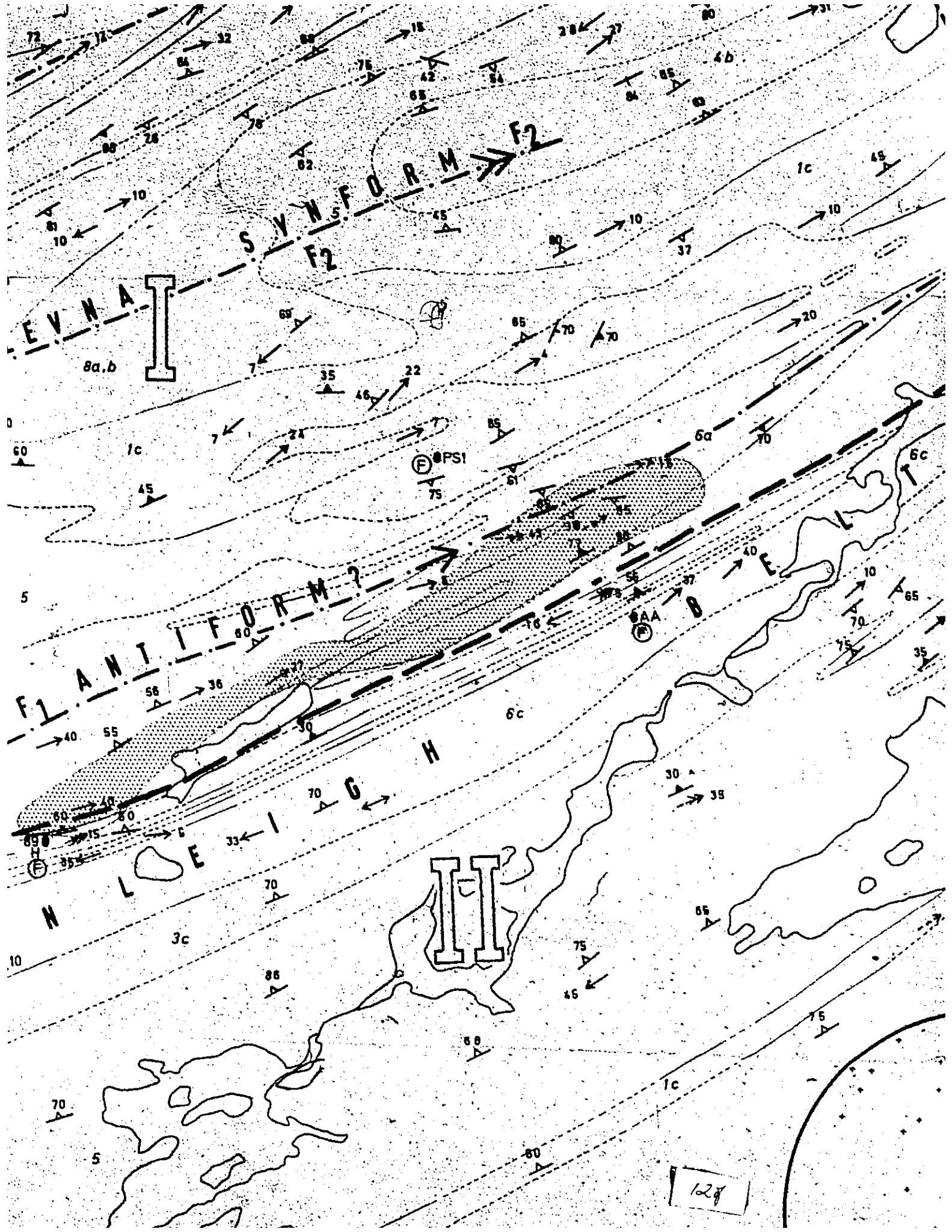
F2 SHAWNEGO  
ANTIFORM

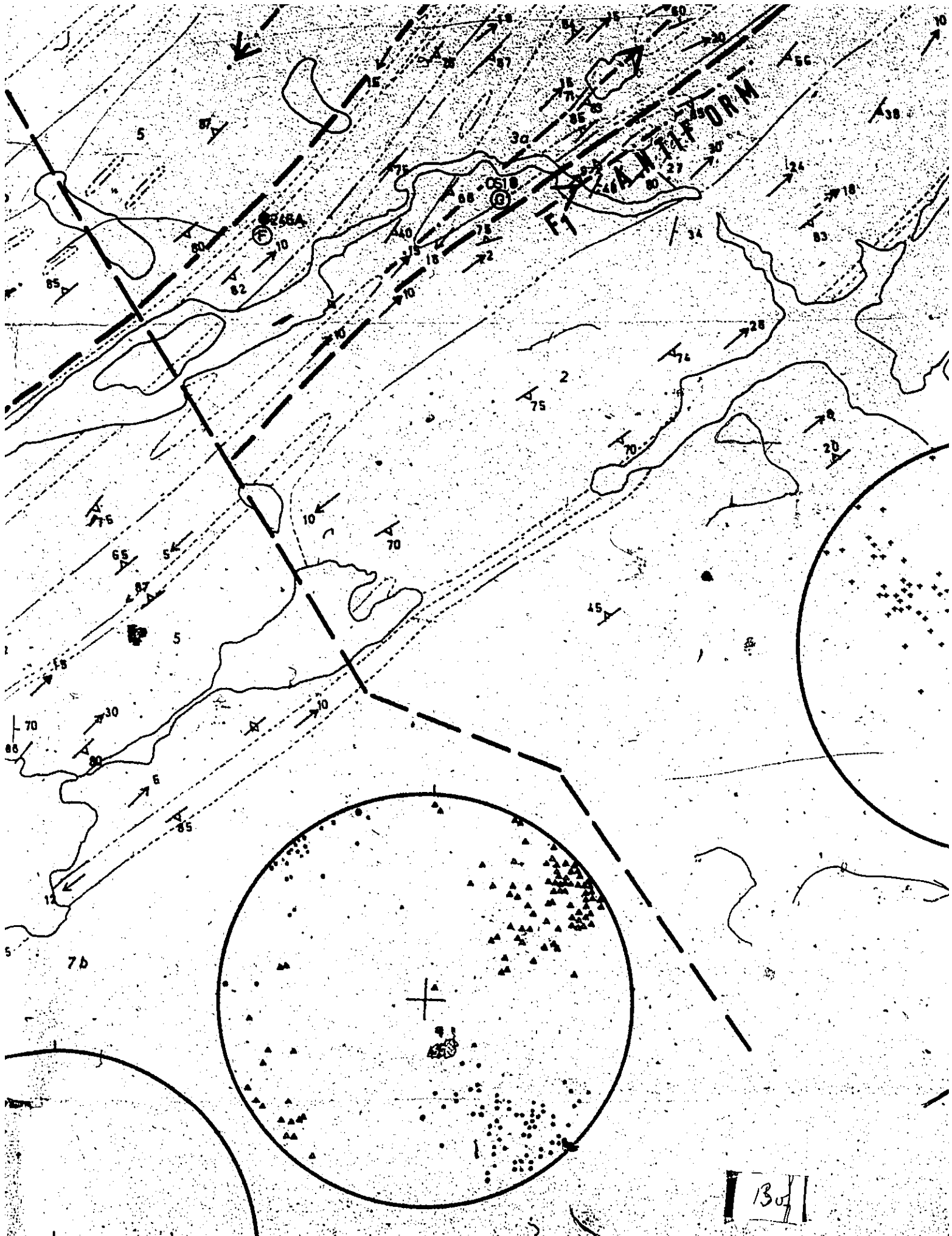
F2 PLYMOUTH

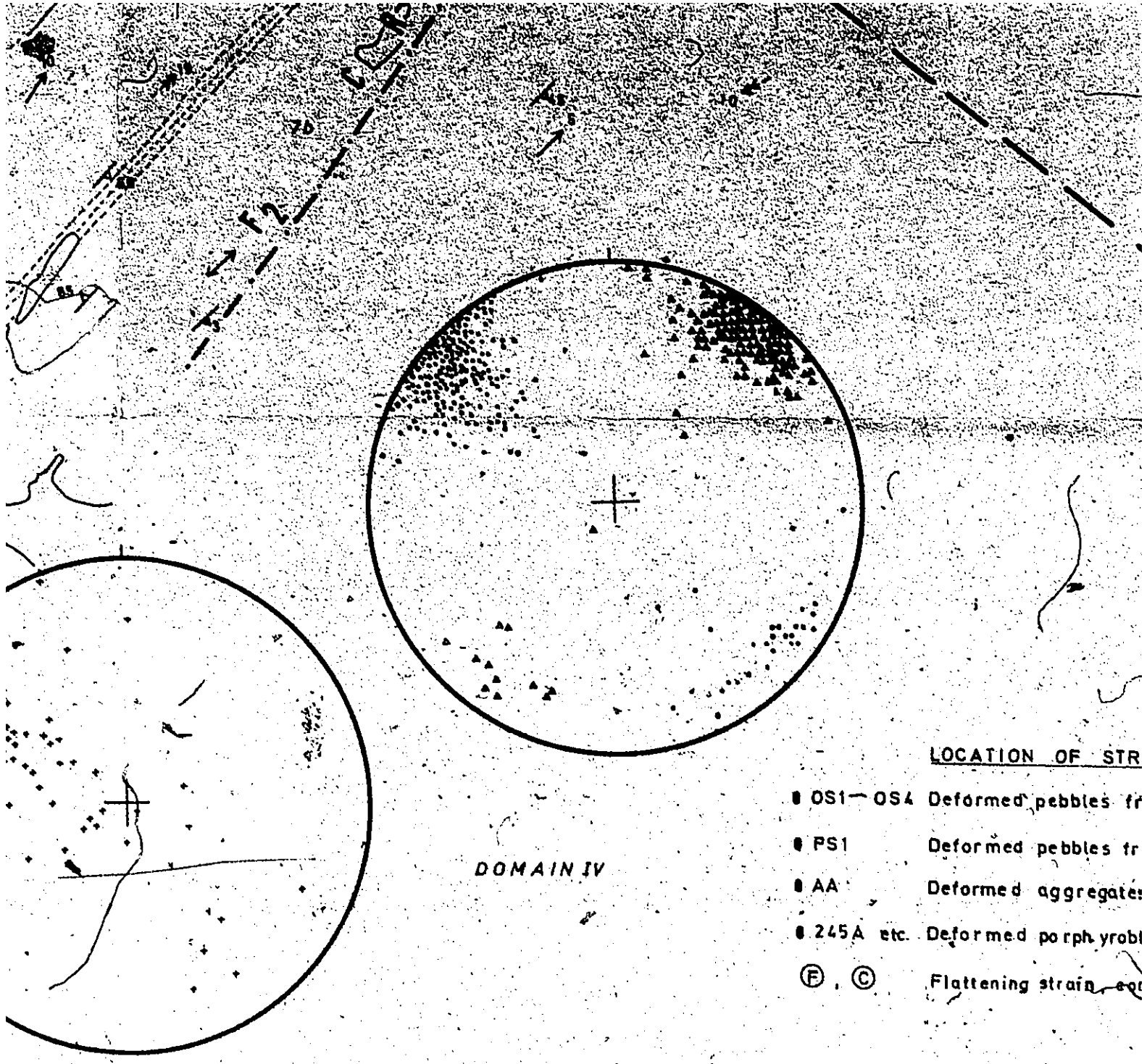
ERNLEIGH

FERR









DOMAIN IV

LOCATION OF STR

- OS1-OS4 Deformed pebbles fr
- PS1 Deformed pebbles fr
- AA Deformed aggregates
- 245A etc. Deformed porphyrobl
- ⓕ, Ⓢ Flattening strain eq

- Bedding ( $S_0$ ) - facing in direction of dip.
- ↘ Bedding ( $S_0$ ) - facing in direction of dip, overturned.
- Foliation  $S_1$
- Foliation  $S_1$  vertical
- Foliation  $S_1$  parallel to bedding  $S_0$
- Lineation  $L_1$
- ↔ Lineation  $L_1$  horizontal
- Fold axis  $F_1$
- Foliation  $S_2$
- Lineation & fold axis  $L_2$  &  $F_2$
- Lineation & fold axis  $L_3$  &  $F_3$
-  Mylonites

Lithologies numbered as on geological map - major units only are indicated.

MEASUREMENTS

npah synform  
 evna synform  
 antiform near Ardach  
 ional strain

STEREONET SYMBOLS

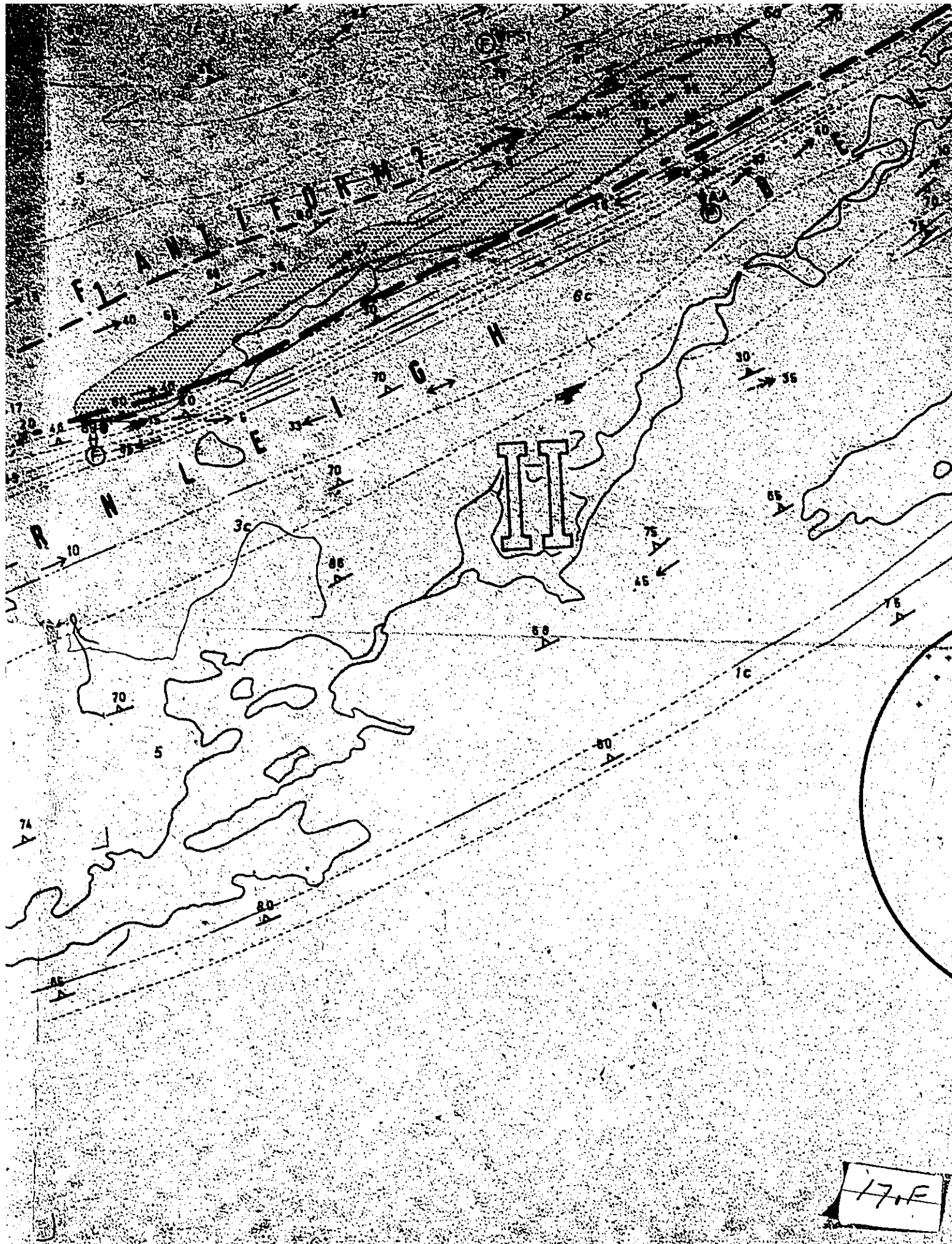
- + Pole to  $S_0$
- Pole to  $S_1$
- ▲  $L_1$  and  $F_1$
- Pole to  $S_2$
- $L_2$  and  $F_2$

150 F

F2.5 HAWKWOOD  
ATTORNEY

FERNLEIGH  
BELT

16 of



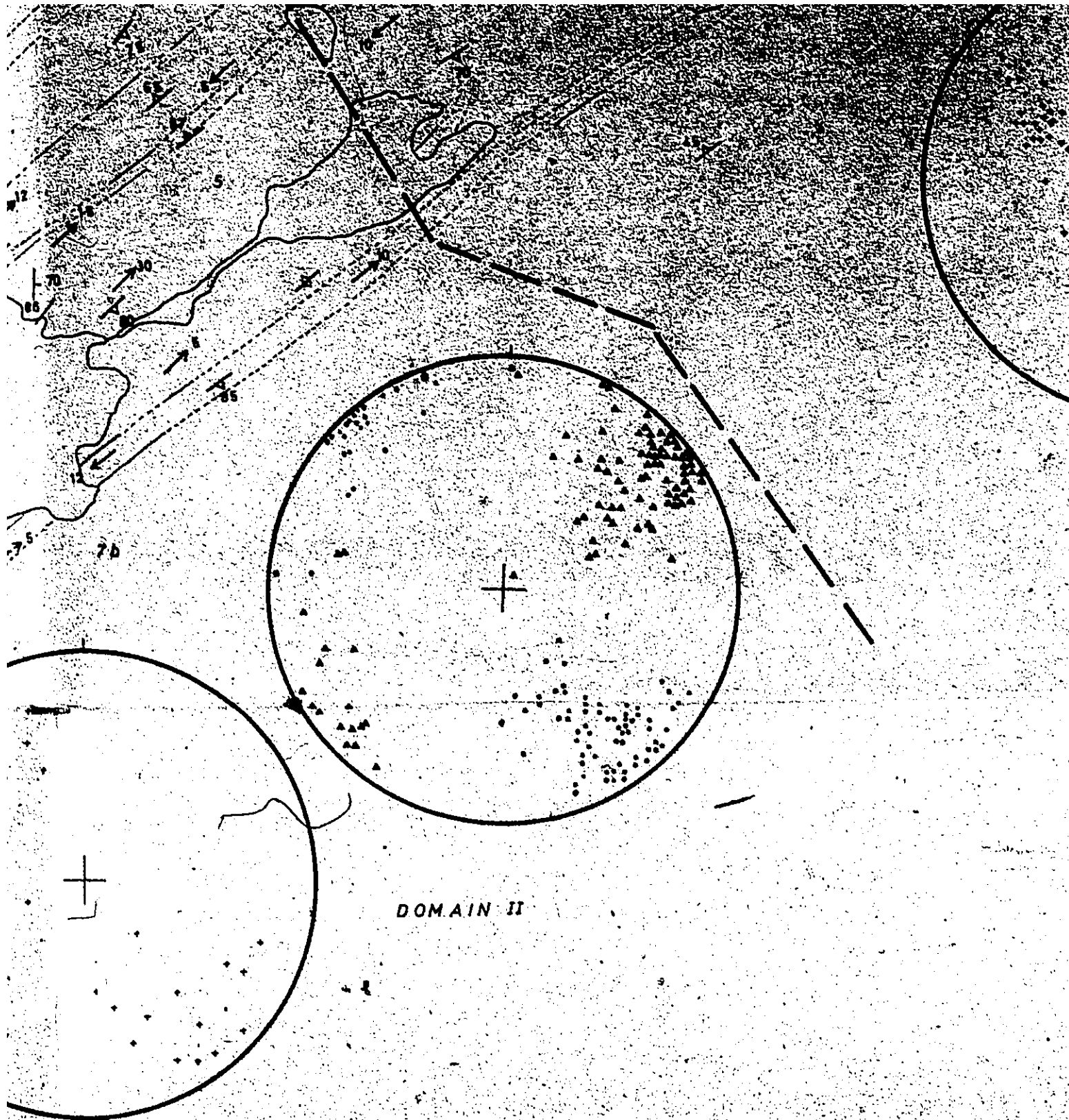


FIG 2

CROSS SECTION

TRUE SECTIONS PERPENDICULAR TO PLUNGE OF  
FOR LOCATION OF SECTION LINES SEE GEOLOGICAL

C

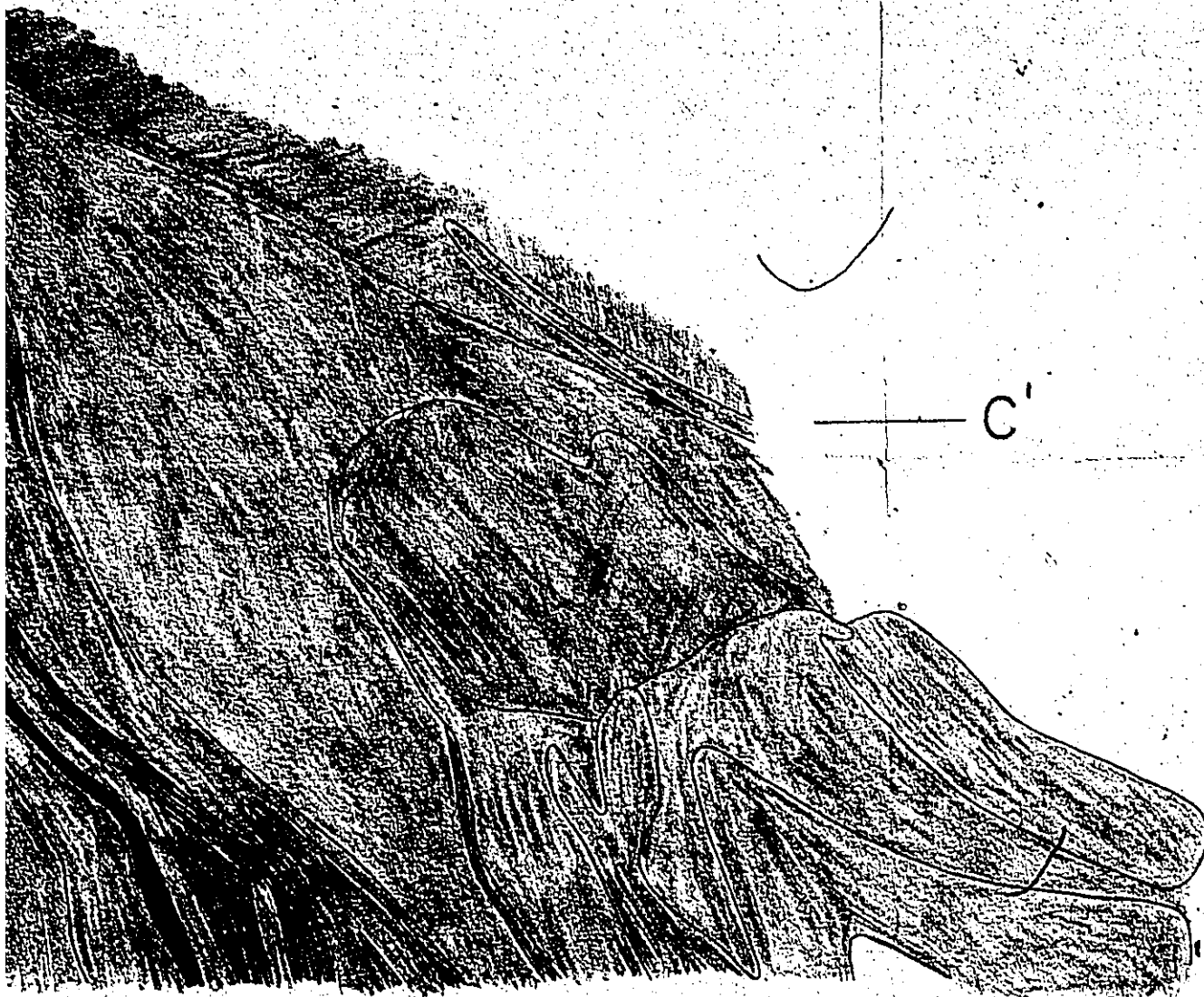


101

IN S.

MAJOR FOLDS

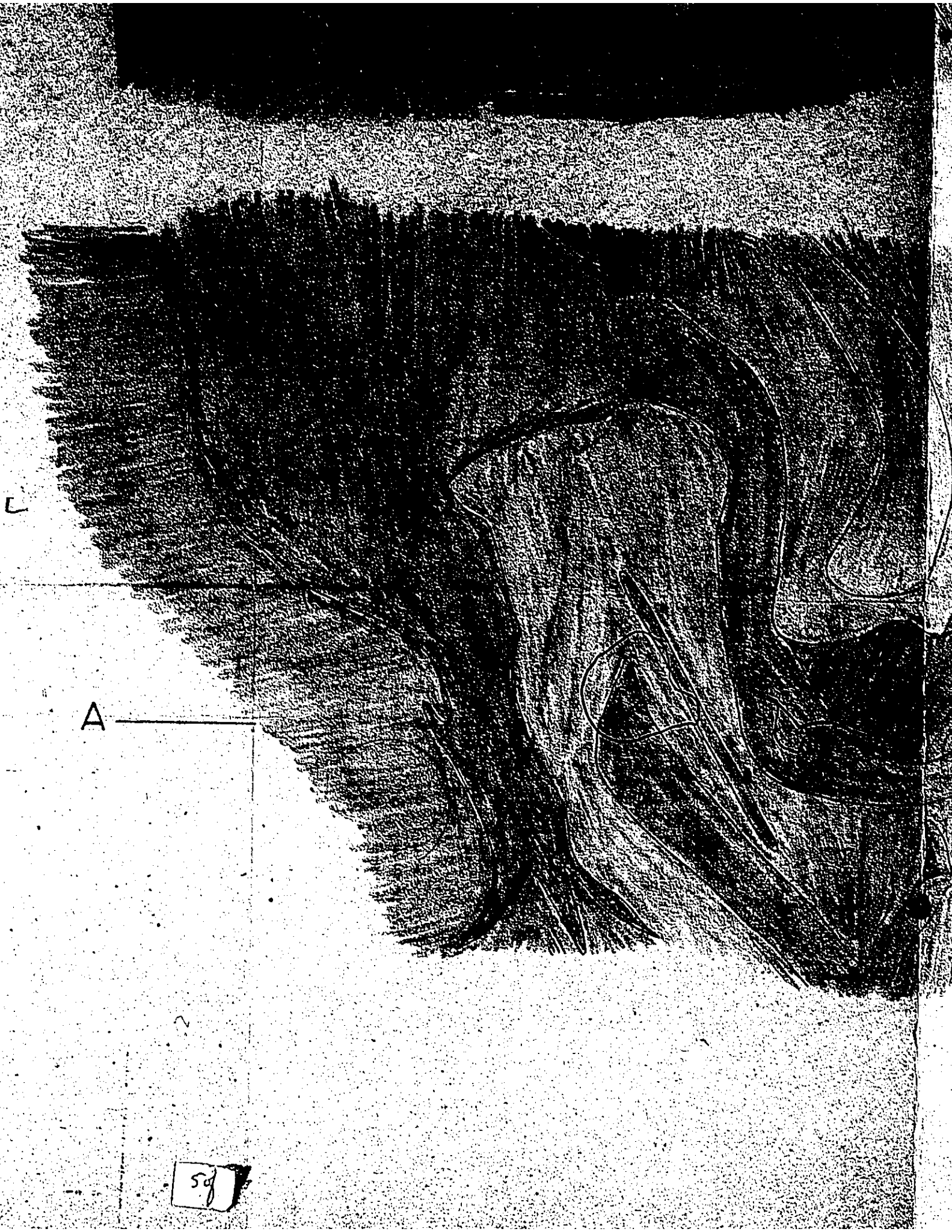
MAP



B

307

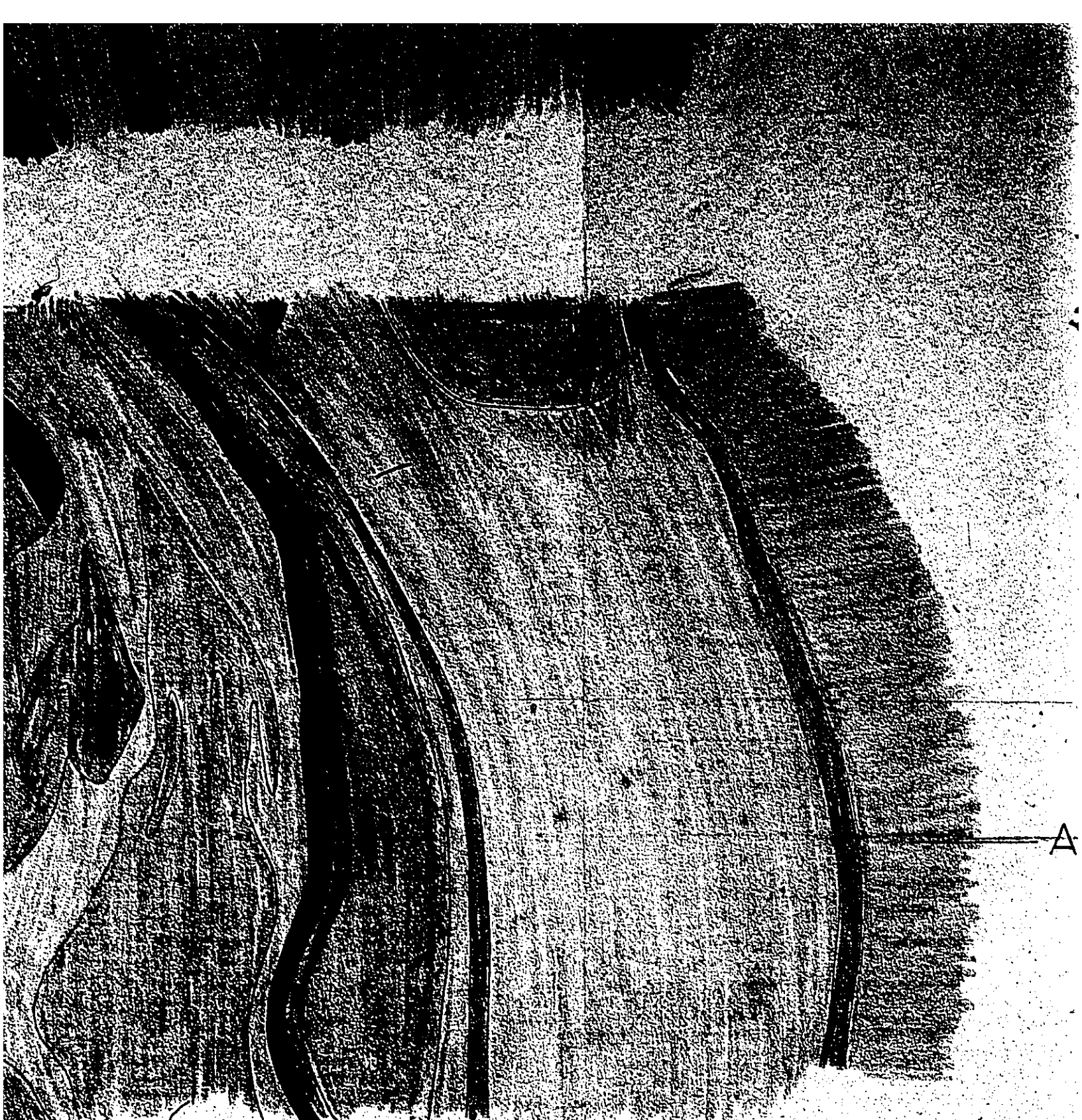




L

A

59



—after Moore & Thompson  
1972.

606