METAMORPHIC PETROLOGY AND GEOCHEMISTRY ACROSS THE GRENVILLE PROVINCE - SOUTHERN PROVINCE BOUNDARY

DILL TOWNSHIP, SUDBURY, ONTARIO

By

Teunis Adrianus Pieter KWAK, B.Sc., M.Sc.

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"Let us keep the discoveries and indisputable measurements of physics. But let us not become bound and fettered to the perspective of final equilibrium that they seem to suggest."

Pierre Teihard de Chardin:
"The Phenomenon of Man"

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TITLE: Metamorphic Petrology and Geochemistry across the Grenville Province -- Southern province boundary, Dill township, Sudbury.

AUTHOR: Teunis Adrianus Pieter Kwak, B.Sc. (University of British Columbia)
M.Sc. (University of British Columbia)

SUPERVISOR: Dr. P. M. Clifford

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SCOPE AND CONTENTS: The geology and geochemistry of an area in Dill township, Sudbury, Ontario are studied in detail. Emphasis is placed on the nature of the boundary between the two provinces and the geochemistry of a sequence of pelitic schists in the Grenville province.

The boundary appears to be marked by a mylonite zone separating pelitic schists containing andalusite-staurolite assemblages from kyanite-sillimanite-staurolite ones.

The pelitic schists in the Grenville province show a steep increase in temperature and pressure extending from the lower almandine amphibolite to the lower granulite facies. Electron microprobe, spectrographic and wet chemical analyses were performed on kyanite, sillimanite, biotite, muscovite and garnet and indicated many systematic changes of elemental distribution coefficients with changing metamorphic grade.

ABSTRACT

A detailed metamorphic petrological and geochemical study of a part of Dill township was undertaken with emphasis on the nature of the Grenville province - Southern province boundary and the geochemistry of the pelitic schists in the Grenville province.

The boundary between the provinces is marked by a mylonite zone (approximately 100 yards wide) along which gabbro and a coarse grained granite were emplaced. There is a significant break in metamorphic grade across the boundary for andalusite-staurolite assemblages occur on the Southern province side and kyanite-sillimanite-staurolite assemblages occur on the Grenville province side. A pervasive mineral lineation is very noticeable near the boundary and probably marks the direction of movement on the mylonite zone (fault).

Aluminum-silicate and rutile bearing pelitic schist samples were collected across the mapped, successive isograds of sillimanite appearance, staurolite disappearance, K-feldspar appearance, muscovite disappearance and kyanite disappearance. Fourteen samples along a 10 kilometer traverse were selected for detailed geochemical study using the electron microprobe, emission spectrograph and "rapid" wet chemical methods. It was found that there were many systematic changes of elemental distribution coefficients between phases and elemental abundances within phases across the traverse.

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CHAPTER I

GENERAL GEOLOGIC FEATURES

A. INTRODUCTION

Many publications on metamorphic rocks exist but none, to date, deal with the interaction of all the factors involved. This is understandable in view of the fact that many variables occur, development of instrumentation and related sciences have not been sufficient in the past and the system of publication is not conducive to lengthy studies.

To properly evaluate a metamorphic assemblage one must know all the variables and constants involved. This has been attempted in the present study. In many studies, assumptions are often stated (or made but unstated) for which little evidence exists. This has been avoided as much as possible and assumptions are stated wherever necessary.

The study deals mainly with a group of samples of pelitic schists which were collected in the summers of 1965 and 1966 from an area near Sudbury, Ontario, Canada (see figure 1).

A 65 square-kilometer area lying athwart the boundary between the Southern and Grenville provinces was mapped. The samples were collected from a sequence of amphibolite facies rocks belonging to the Grenville geological province. Much attention was also paid to the boundary between the provinces.

In the main part of the study, the following features of the pelitic schists were determined:

(1) Well defined positions in the field of metamorphic isograds.

(2) The degree to which phases were chemically homogeneous within individual grains.

(3) The degree to which grains of the same phase were chemically identical within the same section.

(4) The trace and major element composition of most of the phases present.

(5) The interrelation of textural and compositional factors.

B. REGIONAL GEOLOGY

Precambrian rocks of different compositions, metamorphic grades and structures which formed at various ages, underlie most of the North American continent. The best exposures of these rocks are in the Canadian Shield of Canada. This shield has been subdivided into geological provinces on the basis of large scale structural features which, within narrow limits, have uniform potassium-argon dates. These dates are thought to represent the last period of orogenesis affecting the province. It must be pointed out that potassium-argon dating only measures the time at which the rocks had cooled sufficiently to retain argon gas within the mineral lattices and need not represent a metamorphic or orogenic event.

The highly metamorphosed Grenville province has a fairly uniform 950 million year potassium-argon age date while the less metamorphosed Southern province shows ages widely varying with an average near 1700 million years. The boundary between the two provinces, the "Grenville Front", has been traced from Georgian Bay in southern Ontario to Labrador in Quebec. This feature was studied in a fair amount of detail and appeared mainly to be a mylonite zone in the area.

The upper almandine amphibolite facies Grenville Province rocks, in the general area, are composed of gneisses, schists, amphibolites, and rare marbles, all of which have been intruded by minor amounts of pegmatitic, basic and ultrabasic bodies. Very little is known of the structure and lithology of the Grenville province due mainly to its economically "barren" nature.

The upper greenschist to lower almandine amphibolite facies Southern province rocks, located to the northwest of the Grenville Front, contain the economically important Sudbury "nickel irruptive" as well as numerous uranium-bearing conglomeratic horizons. These features have encouraged much academic and commercial study of the rocks.

The Southern province rocks are composed of fine grained quartzites, schists, phyllite and rare marble, all of which have been intruded by gabbro, granite and finally, diabase dykes.

Previous Work in the Area

This study is mainly of a geochemical nature and will not dwell at too great a length on the intricacies of local rock unit nomenclature which has been ably summarized by Henderson (1967).

Southern Province

The rocks now ascribed to the Southern province were studied by Quirke and Collins in their classic 1930 publication "The Disappearance of the Huronian". These authors, as well as many recent students (Frarey, 1966; Young and Church, 1966; Henderson, 1967), believe that identifiable rock units can be traced for many kilometers. Thomson (1957), and Phemister (1961) restudied the area around Sudbury and, among other conclusions, felt that too much inhomogeneity of rock types occurred to make Quirke and Collins' conclusions tenable. Card (1964), working to the northwest of the area studied here, believes there is too much structural complication for detailed stratigraphic work.

This latter author showed that metamorphism in the Southern province consists of nodes of high grade metamorphism very similar to ones described by James (1955) from northern Michigan.

Grenville Province

The Grenville "Front" discontinuity was first defined by A. Murray (1856). Bell (1891) entered the Grenville rocks into Logan's "Laurentian" Archean system. Coleman (1914) believed the rocks were correlative with the "Grenville" series common near Bancroft, Ontario, on the basis of rarely occurring marble layers. Quirke and Collins (1930) stated that most, if not all, the rocks were upgraded and metasomatized Huronian equivalents. This view was also upheld by Phemister (1960), who believed that a large granite body located between the two provinces was the termination of a metasomatizing "front". Recently Henderson (1967) has shown that the latter granite was largely formed from a magma. He also believes the Grenville gneisses in the area to be reformed Southern province rocks.

C. GENERAL METAMORPHIC AND LITHOLOGIC SETTING

i. Introduction

It is often the numerically minor and erratic deviations from the average of a collection of data which warrant the most attention in science and, as a consequence, as much "reasonable" deviation as possible should be recorded and studied. One maps rock units primarily to determine features needed for the particular study in hand. This selective study of features may be criticized from the point of view of being pragmatic, but this is in the best scientific tradition and has yielded utilitarian fruit in many studies.

On both counts small, distinctive outcropping layers of quartzite, pelitic schist or marble are important, as they give much insight on the form of protolith, environment of formation of protolith and, most significant for this study, the subsequent metamorphic history. In this study, small outcrops were slightly enlarged on the map so they could be represented and not "lumped" into large uninterpretable units. The pelitic schists were studied because of their sensitivity to varying metamorphic grades.

ii. Rock Nomenclature

There are many problems associated with the naming of rocks, particularly metamorphic ones. Geologists suffer greatly from two of Francis Bacon's chief sources of ignorance, namely "respect for custom" and "the vain, proud unteachableness of our dispositions". Definitions based on the origin of protoliths, present texture, present mineral composition, etc. are, in the literature, intermixed in illogical profusion, so that definitions or references to definitions are essential to any report. The rock types mentioned here, conform to the definitions referred to by Shaw (1957).

iii. Southern Province Rocks

(a) Areal Distribution and Stratigraphy

The Southern province rocks, including the ones in the area studied (fig.1), are believed to consist of traceable rock units, according to some writers, as mentioned in the last section. Undoubtedly all the writers have some valid points but unfortunately, the answer one accepts is built partially on one's previously acquired prejudices and, in this case, partially on insufficiently studied geology. It should be pointed out that many beds, the "metagreywackes" particularly, are discontinuous and no amount of "unfolding" and "unfaulting" will cause them to be continuous. As a consequence, for correlation, a detailed petrologic study of them is necessary. It is also true that a lithologic group consisting primarily of quartzite and "metagreywacke" extends from well into the northern United States to the northern extent of the Southern province in Canada.

The Southern province rocks underlie the northwestern part of the area studied and fall into Collins' (1938) Mississagi formation. If Henderson's (1967) units, which occur a few miles to the southwest, can be extrapolated to this area, the rocks would fall into his lower quartzites, closest to the Grenville rocks, and lower "greywacke" near Richard and McFarlane Lakes.

(b) Lithology

Both metamorphic and igneous rocks underlie the Southern province part of the area studied. The metamorphic rocks consist of phyllites, schists, quartzites and very rarely occurring amphibolites. Metamorphosed gabbros as well as unaltered diabase dykes occur. A large granite body occurs at the contact of the Southern with the Grenville province. This pluton has been put in the "Between the provinces" classification as it does not seem typical of either province. Pelites, particularly aluminum silicate bearing ones, are most sensitive to metamorphic change at these grades and have received the most attention. The rocks will be described first by their name and secondly, by their various assemblages. These are as follows:

Schists and Phyllites

Quartz-plagioclase-muscovite-K-feldspar-biotite
" " -chlorite- " " -muscovite
" " -biotite-staurolite-andalusite

Amphibolites

Quartz-plagioclase-hornblende-epidote

Quartzites

Quartz-plagioclase-K-feldspar-biotite
" " "
" " -muscovite-K feldspar-biotite
" " -K-feldspar-muscovite

The distribution of the rock types may be seen in figure 1.

Schists and Phyllites

(1) Quartz-plagioclase-muscovite-K feldspar-biotite assemblage

Rocks containing this assemblage occur scattered within quartzite units especially near McFarlane and Richard lakes. In handspecimen the rock cleaves readily along its laminar bedding. The grey to brown weathering rock shows a poor foliation which may occur at a small angle to the bedding. Near the Grenville province boundary a strong lineation formed mainly by quartz rods is developed. This feature is absent or poorly developed one mile northwest of the boundary. The rock is interbedded with the K-feldspar quartzite.

In thin section, fine grained (5 mm or less) anhedral K-feldspar (microcline), plagioclase (average composition of An_5) and quartz occur surrounded by flakes of subhedral biotite (medium to dark chocolate brown pleochroism) and muscovite. Although the micas display a general parallelism of their long dimensions, they generally "sweep" around the more competent salic minerals. The quartz commonly occurs as aggregates which display non-similar extinguishing members. Also present are minor amounts of pyrite partially altered to a limonitic secondary product, and minute euhedral zircon crystals. K-feldspar is often sericitized. The micaceous minerals do not exceed the salic ones in modal abundance.

(2) Quartz-plagioclase-chlorite-K feldspar-muscovite-biotite assemblage

This assemblage is identical to the last described except for the occurrence of pale greenish chlorite. The mineral (clinochlore ?) occurs as minute grains commonly associated with biotite but it is not possible to ascertain whether it is primary or secondary. A number of green coloured phyllites were observed which had much chlorite of apparently primary origin so that the assemblage may well be stable.

(3) Quartz-plagioclase-biotite-staurolite-andalusite assemblage

Rocks of this assemblage are found mainly along the south shore of both McFarlane and Richard lakes. This dull brownish green weathering rock occurs in massive or well bedded outcrops. The bedding is shown by interbedded quartzose layers or layers of varying grain sizes. The most striking feature of the rock is the weathered, large, apparently altered porphyroblasts of andalusite and staurolite (figure 2) which occur in a very fine grained matrix. These large crystals do not display any preferred alignment of their long axis suggesting they did not grow under a directed stress.

There has been some debate about the identification of these and similar crystals in other areas along the Grenville-Southern boundary. In the area studied, staurolite and andalusite are definitely identified as coexisting, and are inferred to be in equilibrium. The identifications were based upon:

i) euhedral crystal form; the staurolite displays "cross" twins and the andalusite orthorhombic prisms.

ii) the optical properties - particularly the yellow pleochroism of staurolite and the very low birefringence of andalusite.

iii) major element analysis by use of the electron microprobe analyzer.

Figure 2.

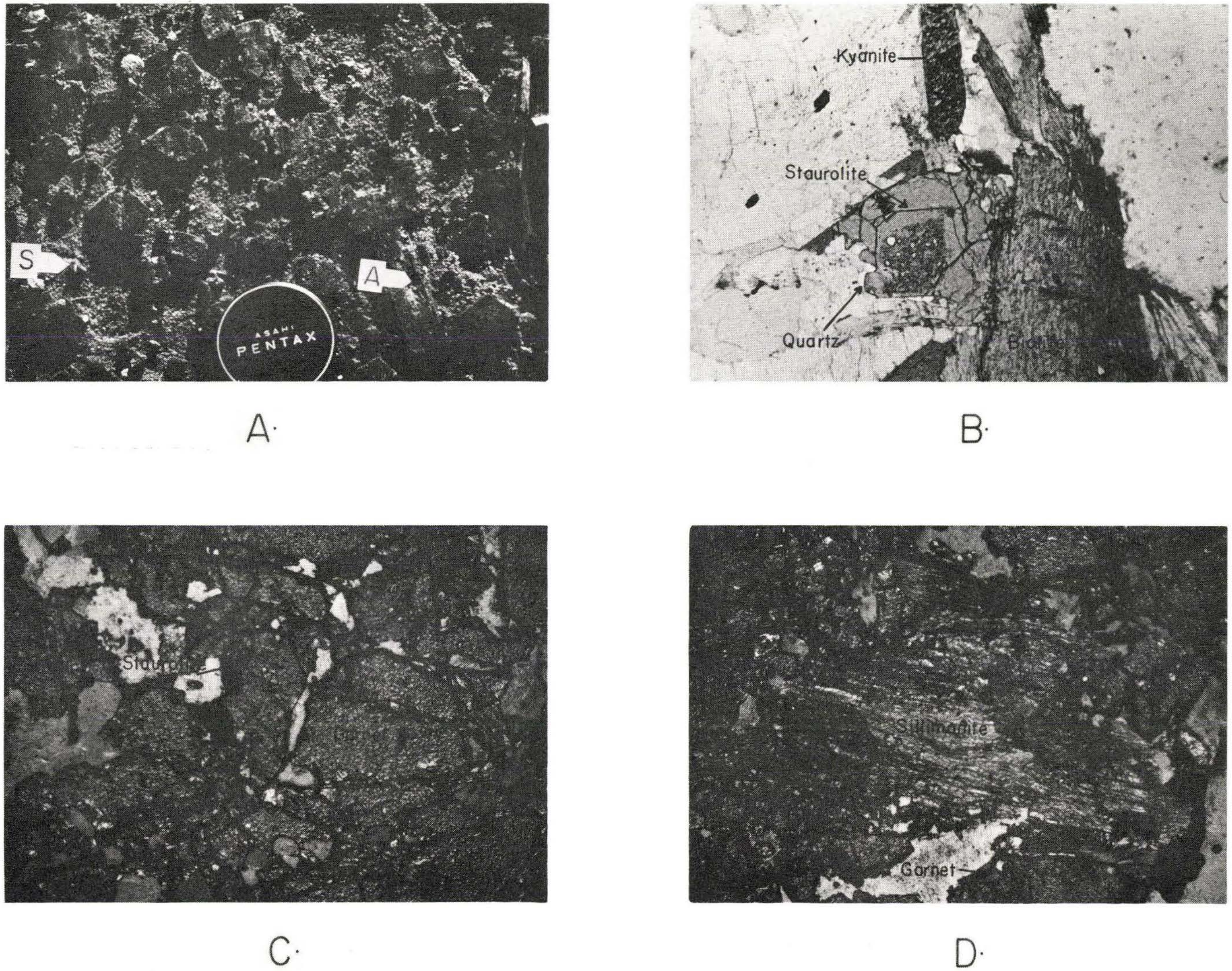
A. Photograph of large orthorhombic andalusite (=A) and cross twinned staurolite (=S) porphyroblasts in a fine grained micaceous matrix. The sample came from an outcrop occurring on the southeast shore of McFarlane lake.

B. Photomicrograph (approx. 40X) showing staurolite having a poikilitic core and clear outer rim as well as being mantled by clear quartz. The sample comes from an outcrop located on the northwest part of the Grenville province.

C. Photomicrograph (approx. 40X) showing anhedral staurolite inclusion within garnet. The sample was collected near the intersection of the Dill lake road and Highway 69.

D. Photomicrograph of sillimanite-biotite "clot" (approx. 40X) within garnet. The "clot" has much the same occurrence and shape as the staurolite inclusion in 2C but the former occurs only at higher grades than staurolite.

Figure 2


S
A
ASAHI
PENTAX
A·
Kyanite
Staurolite
Quartz
Biotite
B·
Staurolite
C·
Sillimanite
Garnet
D·

In thin section the porphyroblasts, as well as large subhedral biotite crystals, are extremely poikilitic. The andalusite and staurolite contain up to 50 percent inclusions and the biotite, 20 percent. This feature makes identification in the outcrop extremely difficult, particularly if the minerals are also partly altered.

In thin sections the porphyroblasts contain numerous quartz and plagioclase inclusions which form a relict bedding, continuous with that in the surrounding matrix. This suggests the crystals have not been rotated.

Andalusite forms orthorhombic crystals (approx. 2.5cm.x.3cm.x.3cm.). The staurolites are smaller in size and twin according to the cross and skew twin laws. Both of these minerals contain large biotite grains near their edges as well as similar quartz and plagioclase grains throughout. This fact suggests that the growth of staurolite and andalusite was initiated before the biotite. The two minerals could not expel or assimilate all the salic inclusions during any time of their growth. Some of the staurolite and most of the andalusite crystals exhibit much fine grained secondary micaceous alteration near their edges.

Medium to deep brown pleochroic biotite porphyroblasts (2mm. wide flakes) occur scattered throughout. The mineral commonly contains many kink zones suggesting mechanical deformation.

The fine grained matrix between the porphyroblasts is composed of interlocking, small, anhedral grains of plagioclase (An_8 av.), quartz, muscovite, cubes of pyrite and zircon crystals. The textures are generally similar to the other fine grained schists described. Locally, near the large porphyroblasts, coarse grained aggregates of clear quartz have been produced, probably by recrystallization.

Schistose quartzite

The rock type, in the field, commonly occurs spatially between quartzite and schist layers. In form it is transitional between the two units as it consists of foliated quartzite commonly interbedded with schist. (Volumetrically schist is 1/3 to 2/3 of the rock.)

K-feldspar quartzite

This rock type makes up most of the metasedimentary rocks of the Southern province. In outcrop, it weathers to a white to tan colour, is well bedded, and commonly displays good crossbedding.

Occasionally a faint trace of a foliation can be observed which may or may not be parallel to the bedding. A strong lineation is developed especially near the Grenville-Southern boundary.

The rocks consist of up to 1/3 K feldspar, much plagioclase and essential quartz. Accessory amounts of subhedral chocolate brown biotite, subhedral muscovite, euhedral zircon and cubes of pyrite were observed. Generally the textures are the same as those in the schists.

Amphibolite

A few small layers of amphibolite occur in the southern part of the area of Southern province studied. Their surface extent is very small and they are not marked on figure 1. In form the bodies are dyke-like and contain no interlayers of other materials. The bodies display a good hornblende lineation, no foliation, and no bedding.

The rock in thin section is fine grained and displays an even grain size distribution. It consists of interlocking anhedral crystals of quartz, hornblende (pleochroic in blue-greens), plagioclase (An_{23}), epidote (pistacite ?) and pyrite.

The rock is considered to have had a protolith which was a concordant basic dyke. This is inferred from the following evidence:

(1) homogeneity of mineralogy, modal abundance of minerals, and grain size.

(2) sharp contacts with surrounding quartzites.

Gabbro

This rock type occurs in large,homogeneous,prominently weathering outcrops throughout the Southern province part of the area studied (see figure 1). It is a common rock type in the Southern province and has been described from areas as far south as Georgian Bay. The outcrop patterns of the rock type are fold forms which will be described later. The light to dark green weathering rock displaces no small scale structures except for chloritized fractures.

The mineralogy varies greatly. Some sections contain primary igneous minerals while others show large amounts of alteration to minerals of the greenschist facies. Primary pyroxene, and plagioclase (av. An_{40}) display a mutual subophitic texture. It is suspected that an opaque mineral such as magnetite was once present but now only limonitic masses occur. Nickel and copper sulphides have been observed in a number of places.

In its more altered forms, the rock consists mainly of sericitized plagioclase and chloritized (pyroxene ?) metacrysts. Occasionally unaltered pyroxene "eyes" (optically continuous in some small areas) occur surrounded by chloritic alteration, rarely occurring amphibole (actinolite ?) and epidote.

Diabase

There are a number of diabase dykes in the Southern province area. These vary in width from a few centimeters to thirty meters and may be continuous at least up to two kilometers. The dykes extend for many kilometers into the Sudbury basin (A. Larochelle, 1966). The rock type forms bold outcrops which weather to a dark brownish green.

The dykes have fine grained margins which contain angular to rounded xenoliths of quartzite.

In thin section, the coarse grained center part of the dykes displays a typical "diabasic" texture. The minerals present are plagioclase (av. An_{62}), titanaugite (pinkish pleochroic) and rectangular, opaque grains (ilmenite ?). The minor alteration products include chlorite, sericite, and limonite.

(c)Metamorphism

The area of Southern province rocks studied is relatively small (approximately 12.5 square kilometers) so for a more thorough understanding of the regional metamorphic picture, pertinent literature has been reviewed. Grades of metamorphism from lower greenschist to lower almandine amphibolite facies have been described from Southern province rocks near Sudbury. The occurrence of the distinctive metamorphic minerals staurolite, chloritoid, andalusite and garnet is ubiquitous in rocks of the Southern province from Georgian Bay to Lake Wanapitei (see Quirke and Collins, 1930; Thompson, 1953; Phemister, 1961; Grant, 1964; Young and Church, 1966; Frarey, 1966; and Henderson, 1967).

The metamorphism of the Sudbury area is generally believed to be of the "regional" type, as defined by Barrow (1893), although the possibility that it is of the "contact" type of Eskola (1939) and due to the effects of granite or the ubiquitous "Sudbury" gabbros cannot be ruled out entirely. This latter possibility is supported by the fact that the minerals are randomly oriented and consist of low pressure, low to moderate temperature assemblages both reminiscent of hornfelsic environments. However, the idea is considered unlikely on the grounds, that many rocks containing the highest grade assemblages occur too far from observed igneous contacts; the distinctive metamorphic minerals are too widespread in the Southern province; the mineral staurolite is not common in well studied contact metamorphic zones; and most intrusions, including the Sudbury gabbros, appear to be metamorphosed to the same grade as the surrounding metasediments. Many magmatically derived bodies exist in the Southern province but no clear relation to a staurolite bearing aureole has been attributed to them, although not many detailed metamorphic studies have been done in the area.

Card (1964), working near Agnew Lake in the Sudbury district, has shown metamorphism there to consist of "high temperature nodes superimposed on a lower or intermediate grade background" (p. 1014). The highs are of the lower almandine amphibolite facies (staurolite exists) whereas the "background" is low to middle greenschist facies. Card believes the highs to be produced by "pipes of heat conduction by upward flowing hot fluids" (p. 1029).

James (1955) described nodes of high metamorphic grade in similar rocks in northern Michigan. The nodes terminated above the sillimanite zone from a chlorite zone background. The heat source is believed to be unobserved magma.

Both these studies deal with nodal highs although nodal lows possibly occur in the area. Writers are less prone to describe lows, especially in greenschist facies rocks, because minerals which are easily identified and show small metamorphic changes, occur only at the higher grades. It seems significant that in the studies mentioned above and most others dealing with Barrovian, Abukuma, Saxonian and Idahoan metamorphic sequences (Hietanen, 1967), no direct evidence of a magmatic source for the causes of the metamorphism is observed. Most metamorphic sequences also terminate in nodal highs.

One is thus left with the following possibilities:

(1) upward flowing hot fluids

(2) the occurrence of "hot spots" in the crust

(3) differential uplift.

The hypothesis of hot fluids may be true but no direct evidence is possible unless some component of the fluid has remained,which has not been demonstrated. The hypothesis of hot spots has some points in its favour. It has been demonstrated that some areas at the surface today emit more heat than others. However, nothing on the scale of that necessary for the different grades has been observed. No evidence of magmatic sources exists but this model is appealing.

The last hypothesis, which suggests the differential uplift of rocks of depth zoned, increasing metamorphic grade, seems the most likely. There is certainly much precedent for inferring significant vertical movements in the earth's crust. The nearby Grenville province contains high pressure granulite facies rocks which now occur at the earth's surface. The widths of the observed nodes in the Southern province are large enough to make such a theory seem practical although no deflection of the stratigraphy has been shown. A major problem with this hypothesis is the apparent lack of domal structures such as accompany similar gneissic domes. Perhaps the mapping was done from mainly a metamorphic point of view and not a structural one. The hypothesis might well explain why the narrower node (30 kilometers diameter) described by Card does not extend to as high a grade over the background as the wider node (45 kilometers diameter) of James.

In summary, it seems the Southern province rocks display an irregular metamorphic topography possibly due to differential (magmatic ?) hot spots or vertical movements. The mineralogical evidence seems compatible with metamorphism of low pressure, moderate temperature Abukuma type (Miyashiro, 1961) or the slightly higher pressure Idahoan type (Hietanen, 1967).

The entire area of Southern province rocks in this study lies metamorphically above the biotite isograd and is believed to be of the lowest almandine amphibolite facies. At least part of the area definitely can be shown to be of lower almandine amphibolite facies because of:

(a) the occurrence of staurolite in pelites (a necessary criterion as defined by Francis (1956));

and (b) the occurrence of oligoclase in a small amphibolite interbed (a necessary criterion as defined by Eskola (1939)).

These occurrences are not spread evenly over the area (none north of highway 69 in the area studied) but it is assumed the host rock composition was not appropriate in these parts of the area.

The lower almandine amphibolite facies, in a regional sense, seems to be typical for the area immediately adjacent to the contact with the Grenville province and corresponds to the most intense part of Card's node. Away from these higher grade rocks, a greenschist background exists both near Agnew Lake and probably immediately northwest of the area mapped in this study (inferred from observations of chlorite porphyroblasts along highway 69, especially in the Ramsay Lake conglomerate near Ramsay Lake in Sudbury).

iv. Grenville Province

(a) Areal distribution and Stratigraphy

The rocks of the Grenville province, which occur in the southeastern part of the map area, have been correlated with the rocks of the "Grenville group" of the "Hastings basin" by Coleman (1914). This classification was based mainly on the rare occurrences of thin marble and calc-silicate granulite layers. The present study has indicated more of these layers but certainly none quantitatively comparable to the Grenville Group proper. The ground between this area and the Hastings basin is essentially unmapped.

Very little is known about the Grenville province near the area studied. Phemister and assistants (Phemister et al. 1961) mapped approximately 38 square kilometers along the Grenville-Southern boundary mainly in Dill township. Pearson (1961) described in detail a kyanite prospect a few kilometers northeast, near Coniston.

In this study, three major units have been defined in the Grenville province. These units trend parallel to the boundary between the provinces which is parallel to the strike of the beds in both the provinces. The major units are fairly uniform in width. They are defined as follows:

Unit 1 - a 2.7 kilometer thick unit of felsic gneiss containing much amphibolite and occurring nearest the Grenville-Southern boundary, and

Unit 2 - a 1 km. thick unit of mainly garnet and diopside amphibolite with many aluminum silicate bearing pelitic schists, quartz-feldspar gneiss and a few calc-silicate granulite horizons. This unit occurs near the center of the map area.

Unit 3 - A 3 kilometer thick unit occurring farthest from the boundary. This unit consists of mainly K-feldspar gneiss with a small amount of amphibolite, calc-silicate and aluminum silicate pelitic horizons.

The central unit (2) is very distinctive by virtue of its high aluminous mineral content. The boundaries between these major units are somewhat arbitrary. It is inferred that the units correspond to major sedimentological variations and may be treated as metamorphosed "formations".

It is difficult to determine the sequence prior to metamorphic recrystallization. On the northeast shore of a small beaver pond located in the northeastern tip of the Grenville area, a quartzose gneiss layer contains apparent crossbedding which suggests a way-up direction to the northwest. No folding could be observed to have altered this orientation and from it one could suggest that unit 3 is the first, and 1, the last deposited.

(b) Lithology

It must be apparent to any reader who has worked on high grade metamorphic rocks that the division of rock units, and the subsequent positions of boundaries on a map, are very susceptible to personal bias. Nevertheless, the division is one of the most essential features of any realistic metamorphic study. The degree to which subdivisions of main rock types (schist etc.) should be introduced, is problematical. If too much "lumping" is done (i.e. the units are too vague so that ones of different origins, etc. occur as a single unit) then the use of the map is seriously hampered. If too much "splitting" occurs (i.e. very much subdivision into varietal types) the accuracy of the work would probably be affected, especially when working quickly over large areas. Of the two, the writer prefers the latter, as to use more generalized units can always be done if units are thought to be too similar. The converse is impossible without further mapping.

In this study, the divisions used were gneisses, schists and amphibolites. The subdivision of these units has been on the basis of mineralogy, in the case of schists and amphibolites, and degree of uniformity of composition of the layering and variation of modal abundances, in the case of the gneisses. The groupings are somewhat arbitrary and all gradations occur. This is especially true of the gneisses and some of the amphibolites.

The only metamorphic rocks encountered which do not have a layered structure are pyroxene and anthophyllite amphibolites. All others, including the marbles, are compositionally and texturally layered and strictly should be termed gneisses. This was not done in many of the units as it was felt that nearly all the rock units would be called gneiss, making the use of the word unnecessary.

The following subdivisions have been defined and the descriptions of the rocks will be presented in the order given here.

Gneisses -

- K-feldspar gneiss
- K-feldspar-Hornblende-Biotite gneiss
- Quartz-Feldspar gneiss

Schists -

- Muscovite-K-feldspar schist
- Muscovite-Kyanite schist
- K-feldspar-Sillimanite schist

Amphibolites -

- **Common amphibolite**
- **Garnet amphibolite**
- **Diopside amphibolite**
- **Anthophyllite amphibolite**

Calc-silicate marble (granulite)

Intrusions

- Basic
- Ultrabasic
- Pegmatite

Gneisses

The gneisses are areally the major rock type in the Grenville province (figure 1). They are found in all the major units but are especially abundant in the last. All the gneisses except the quartz-feldspar gneiss contain pegmatitic veinlets and the term "migmatite" could properly be used for them. The minerals in the pegmatitic veinlets are of granitic proportions and the veinlets occur only within the gneissic units. This suggests the veinlets have been produced from the bodies they are found in.

K-feldspar gneiss

This unit is defined on the following mutually complementary factors in the field.

(i) A high proportion of K-feldspar. A crude approximation of 30% or more apparent K-feldspar was used in the field, however, in thin section the true content was less than 20%

Figure 3.

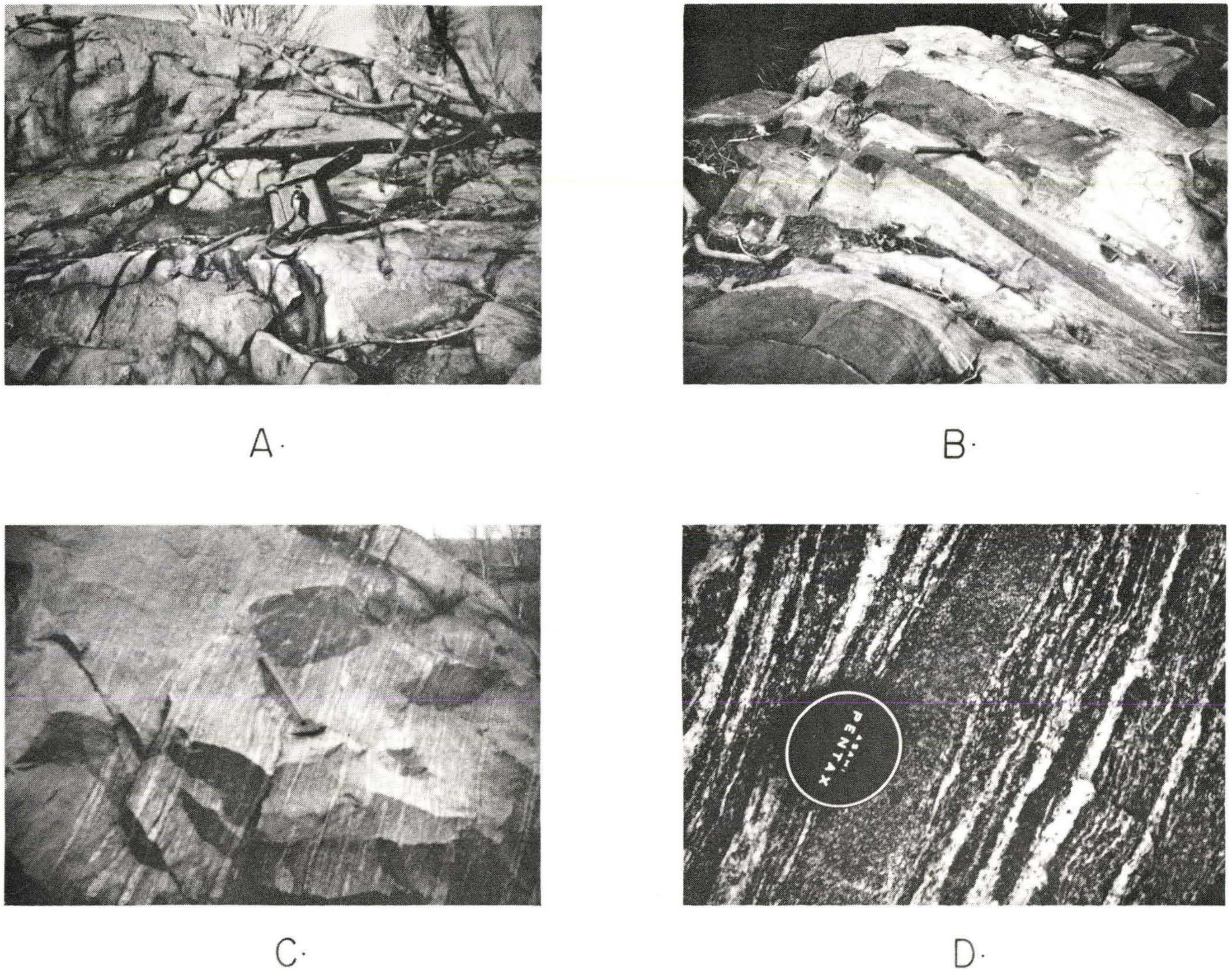
A. A homogeneous, massive (orthogneiss ?) gneiss outcrop located in the northwestern part of the Grenville province rocks studied.

B. Alternating mafic and felsic layers in a gneiss outcrop located near the boundary with the Southern province.

C. Lit-par-lit type gneiss located on highway 69 near the Dill lake road.

D. K-feldspar-Biotite-Hornblende gneiss from the southeastern part of the Grenville province area studied.

Figure 3



mainly due to the influence of perthitic intergrowths.

(ii) A low proportion of more mafic interbeds (less than 20 visual percent). The latter factor was the main consideration in the rock's definition.

In outcrop, the following gneissic types occur:

(1) fairly homogeneous rock approaching a granitic texture (figure 3A);

(2) alternating salic and mafic layers to the scale of centimeters to meters wide (figures 3B and 3C);

(3) alternating layers containing numerous oval feldspar eyes (augen gneiss).

The K-feldspar gneiss grades into the K-feldspar-hornblende-biotite gneiss unit. In the field the former unit is very seldom in contact with large bodies of mafic rich rock such as amphibolite.

The white to pink weathering rock has a well developed layering, good foliation and excellent mineral lineation. The foliation is usually parallel to the layering except in the hinge zones in certain folds where it traverses the layering. The mineral lineation is commonly down the dip,except in the hinge zones in some folds,where it has been apparently rotated around the fold surface. The lineation becomes less noticeable a few miles from the Grenville-Southern boundary, a feature common to all the metamorphic rocks in both provinces. The layering in nearly all cases is interpreted as relict bedding. In thin section the following assemblages have been observed:

Quartz-plagioclase-K-feldspar-biotite
" - " - " - " -hornblende
" - " - " - " -garnet
" - " - " - " - " -hornblende
" - " - " -garnet-hornblende

The major minerals, K-feldspar (microcline), plagioclase (An_{10} average) and quartz form an intertongued fine to medium grain sized mosaic. The grains in this mosaic have an elongate fabric. Many of the salic mineral grains showed a slight oscillatory extinction. Common accessories are subhedral biotite (of light to dark chocolate brown pleochroism), hornblende (of medium to dark green pleochroism), euhedral inclusion filled garnet, euhedral sphene and zircon. Some limonitic cubes (after pyrite ?) were observed.

The feldspars have been sericitized to varying degrees. Hornblende and,to a lesser extent,biotite have been altered to chlorite (penninite ?).

No systematic mineralogical changes were observed in this rock throughout the area. The rock is considered to have had a quartz sandstone protolith because of its high quartz content (50% av.) and extremely variable mineral composition.

K-feldspar-Biotite-Hornblende gneiss

This rock type occurs throughout the area. It is defined as having:

(1) less than 30% apparent K-feldspar;

(2) greater than 20% interbedded layers of more mafic rock types.

The unit is transitional between K-feldspar gneiss and various amphibolitic rock units (both spatially in the field and in definition, e.g. fig.3D). The petrology is the same as that of the K-feldspar-gneiss. The protolith of this rock type was probably an impure quartz sandstone because of its high quartz content and extremely variable composition, both along and across strike.

Quartz-feldspar gneiss

The unit is distinguished from the other gneissic units by the extremely high proportion of quartz. This mineral makes up to 90% of some sections and is never less than 60%.

This rock occurs mainly within the major unit (2). Common interlayers are aluminum silicate bearing schist, up to 12 centimeters thick, layers of calc silicate bearing marbles, biotite schist and garnet amphibolite.

In hand specimen, the white weathered rock shows a medium grain size and is conspicuous by the large amount of sugary clear quartz. A well developed mineral lineation, moderate foliation and observable layering are common features.

The dominant minerals in thin section are quartz, K-feldspar (microcline), plagioclase (An_{14} av.) with accessory amounts of biotite (light to dark brown pleochroic), garnet, muscovite and zircon. The interrelations of these minerals are typical of the gneissic rocks so far mentioned.

Common assemblages are:

Quartz-plagioclase-K-feldspar-biotite
" " " " -muscovite
" " " " -garnet
" " " " -muscovite

An interesting feature is the occurrence of K-feldspar-rich patches which seem spatially related to fractures in the unit located in outcrops near the intersection of Highway 69 and major unit (2). The feature is suggestive of metasomatism. The rock undoubtedly had a nearly pure quartz sandstone protolith.

Schists

Schists occur distributed throughout the area although a high concentration occurs in major unit (2). The schists commonly form interbeds which are from a few centimeters to tens of meters wide and many extend for hundreds of meters along strike. The schists have received the most attention in this study because their mineralogical changes are most sensitive to metamorphic change and will be more fully described in section IIB.

Muscovite-K-feldspar schist

This rock type is defined on the basis of the association of muscovite and K-feldspar. It lacks aluminum silicates. The incompatability of aluminum silicates with K-feldspar at the almandine amphibolite facies of metamorphism has been noted by many authors (e.g. Fyfe et al, 1962, p. 229). Kyanite, sillimanite and K-feldspar do occur in apparent equilibrium in a few samples beyond the muscovite breakdown isograd.

In outcrop the rock type weathers to a tan colour and displays augen-shaped pink bodies of feldspar. These augen occur in a coarse grained matrix of large, distinctive muscovite flakes. Lineation is rarely present probably due to the lack of appropriate mineralogy to express it. The layering is generally parallel to the foliation.

The only assemblage is

Quartz-plagioclase-K-feldspar-muscovite-biotite-garnet

In thin section the rock is medium to coarse grained and has excellent parallelism of micas. Occasionally microfolds are displayed as well as swirls of micas around garnets and K-feldspar rich aggregates. The micas consist mainly of muscovite with very fine grained interleaved biotite (light to dark greenish brown). Some areas of the slides contain extremely fine grained aggregates of muscovite which are suggestive of mechanical deformation.

The garnets are oval shaped in cross-section and display poikilitic cores. The borders and fractures within the garnets are mantled by secondary clear chlorite (penninite ?). Quartz, plagioclase (An_8 av.) and sericitized K-feldspar, which form intermica aggregate bodies, display prominent undulatory extinction. Many homogeneous, oscillatory-extinguishing quartz grains are in microfolds which suggests they were "bent" this way. Accessories include euhedral apatite and zircon, pyrite cubes, graphite, and brown rutile.

Muscovite-Kyanite schist

This rock unit is in form, distribution and structure much the same as the last one mentioned. The following assemblages have been observed:

Quartz-plagioclase-muscovite-kyanite-garnet-biotite
" - " - " - " - " - " -staurolite-sillimanite
" - " - " - " - " - " -sillimanite
" " " " " " - " - K-feldspar

In thin section the rock type resembles the K-feldspar-muscovite schist with the general exclusion of K-feldspar augen and the occurrence of staurolite, kyanite and sillimanite.

The stauralite has mutual boundaries with all the phases of the assemblage. The staurolite isograd (figure 6) was determined where the mineral no longer had mutual boundaries although it still persists as armoured relicts in garnet (figure 2C). The loss of the mineral marks the boundary of the Staurolite-Quartz subfacies. The staurolite forms prismatic crystals which often display a poikilitic core surrounded by a clear yellow outer zone (figure 2B). There is often an outer rim of quartz occurring around and as embayments within the grains.

Kyanite occurs as grey to blue subhedral crystals scattered throughout the sections. In one section a distinctive blue core was observed. The mineral commonly is mantled by a coarse grained aggregate of muscovite especially in samples from near the breakdown isograd. A very fine grained, micaceous, secondary product has been observed as a mantle around the kyanite in some sections. The latter mantling is not to be confused with the former. A spectral scan by the electron microprobe analyzer indicated peaks for silicon, potassium and aluminum, suggestive of muscovite. The alteration does not seem to have any relation to the coarse grained muscovite. In the lowest grade sample (T-79-1), kyanite contains an inner poikilitic core surrounded by a clear rim.

In general, the kyanite is intergrown with plagioclase, quartz, biotite, and garnet, and apparently is stable with respect to these minerals.

Sillimanite occurs as fibrous grains commonly associated with biotite. It occurs as a sheaf of crystals (fibrolite) and has mutual boundaries with all the other minerals.

Two types of related pegmatitic bodies occur within outcrops of the Muscovite-Kyanite schist. One type is a kyanite-bearing one which contains large (to 8 cm. long) aligned, blue kyanite crystals in a quartzose matrix (figure 4A). The other consists of fibrolite bodies up to 5 cm. wide (figure 4B).

The kyanite-bearing bodies are up to 12 centimeters wide and extend for meters generally parallel to, although occasionally crossing the layering. The rock is composed mainly of quartz (75%) and kyanite (15%) with minor amounts of plagioclase, biotite, and muscovite. The rock type only occurs within the kyanite-bearing muscovite schists and mainly below the staurolite breakdown isograd. The rock forms knife edge contacts with the surrounding schist (see figure 4A).

Sillimanite pods consist of fibrolite which is intimately intergrown with the normal schist mineralogy, particularly quartz. The contacts of this rock are somewhat gradational into the surrounding schists. Layers having widths near 5 centimeters have been measured, which extend for many meters parallel to the foliation.

The sillimanite bodies contain no kyanite and the kyanite bodies, no sillimanite. Both types occur in a number of outcrops but no intersections of the two types have been observed. In both bodies the minerals are aligned parallel to those in the schist.

The sillimanite pods seem to occur first at the sillimanite first appearance isograd although less than 1% sillimanite was observed in the host rock. This feature has been described for Scottish rocks by Chinner (1960, 1961).

K-feldspar-Sillimanite Schist

This rock type is defined on the basis of the K-feldspar content and general lack of kyanite and muscovite. Areally it occurs in major unit 3 and above the muscovite breakdown. The rock forms interbeds of various thicknesses some of which are only a few centimeters thick.

Some kyanite and/or muscovite has been observed in a few samples located near the kyanite breakdown isograd. These minerals show textures interpreted as disequilibrium textures and will be discussed in section IIB. The following stable assemblage occurs:

Quartz-plagioclase-K-feldspar-sillimanite-biotite-garnet

In thin section the K-feldspar, which occurs as microcline, forms small anhedral grains intergrown with the general groundmass, and in perthitic intergrowth in large augen. No direct relations of this mineral to muscovite have been seen. Biotite is extremely fine grained and commonly has a greenish cast to the chocolate brown colour. Sillimanite occurs as stubby crystals scattered throughout the rock and has no apparent relation to biotite as it does in the other aluminum silicate bearing schists. Fibrolite lenses (macroscopic) and sheafs (microscopic) were not observed. Fibrous sillimanite has been observed included within garnet but the sillimanite outside the garnet is stubby. This situation may be interpreted as sillimanite of "low grade form" occurring as a relict inside garnet. The pinkish subhedral garnets commonly display a well defined poikilitic core with a wide outer clear zone. Near their edges, the garnets have large irregular quartz masses.

The plagioclase and quartz occur as a fine grained, granular intergrowth with microcline, biotite and sillimanite. Accessory minerals seen in various sections are pyrite, chalcopyrite, pyrrhotite, magnetite, rutile, apatite and zircon. Magnetite occasionally has an outer rim of rutile suggestive of a reaction. Mild chloritization (penninite ?) occurs within the garnet. Some of the feldspars are partly sericitized.

Biotite Schist

A small number of units were mapped as biotite schist on the basis of a high biotite content (> 30%) and schistose appearance. The rock type grades spatially (and conceptually) into both amphibolite and K-feldspar-hornblende-biotite gneiss.

In outcrop, the rock type usually displays a strong foliation, weak lineation, and a moderate amount of layering. It weathers to a dull brown. It is present as thin, unmapped layers in other rock types.

Common assemblages are:

Quartz-plagioclase-biotite-hornblende-garnet
" - " - " -garnet-biotite
" - " -K-feldspar-biotite-garnet

In thin section the brown or greenish brown biotites form an intertongued intergrowth with hornblende, plagioclase, quartz and, in some areas, K-feldspar or perthite. Common accessories are apatite, sphene, opaques and zircon. All the schists probably had protoliths which were shales of various compositions.

Figure 4.

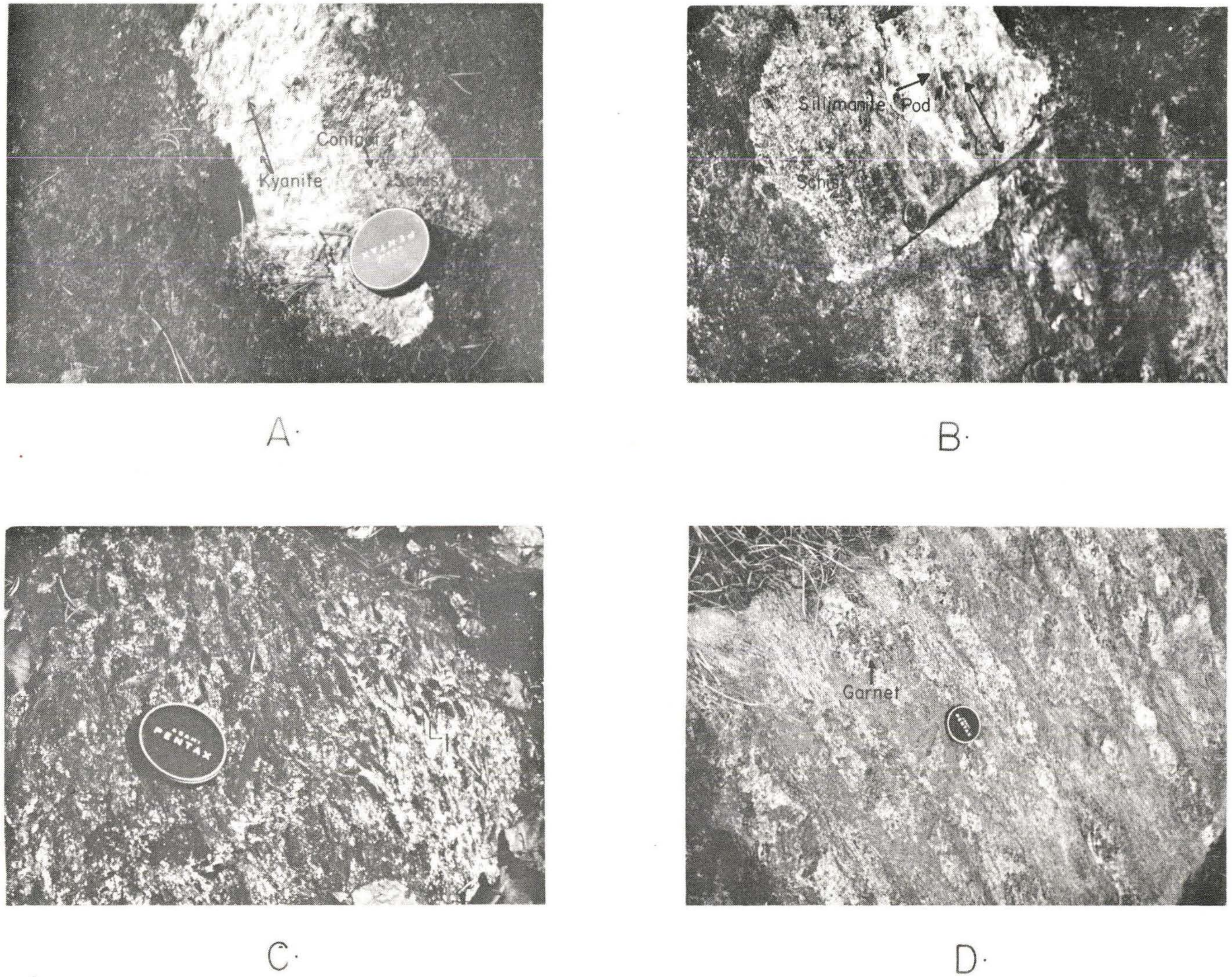
A. Kyanite-quartz body within kyanite-muscovite schist (Local T-78-1). Notice the knife edge contact near the lens cap.

B. Sillimanite pod within kyanite-muscovite schist. This sample is from an outcrop located near the boundary with the Southern province (Local T-78-1).

C. Aligned kyanite grains indicating L_i direction (Local T-78-1).

D. Large garnet crystals within sillimanite-K-feldspar schist located 1 kilometer east of intersection of highway 69 and Dill lake road.

Figure 4


Kyanite
Contact
Schist
A.
Sillimanite
Pod
Schist
B.
L₁
C.
Garnet
D.

Amphibolite

Approximately a third of the Grenville province rocks may be classified as amphibolite of various types. All the amphibolites except the diopside and anthophyllite bearing ones, contain numerous small felsic interlayers. As a consequence, most amphibolites are in a strict sense amphibolite gneisses. However it is felt their amphibole content warrants the amphibolite classification. The amphibolites are subdivided on the basis of distinctive (or lack of distinctive) mineralogy.

Common Amphibolite

"Common" amphibolite on figure 1 refers to such a rock which contains only the necessary minerals to be called amphibolite, i.e. hornblende and plagioclase. Garnet may be present to 3 percent and biotite to 20 percent. The rock type grades into biotite schist, garnet amphibolite and K-feldspar-hornblende-biotite gneiss. Garnet and common amphibolite grade into each other both across and along strike and again the extremely arbitrary nature of the contacts must be stressed. The inconsistent composition is cited as evidence that this unit and the garnet amphibolite are sedimentary in origin.

In thin section, the dominant minerals are hornblende, plagioclase, garnet, K-feldspar, and quartz. Accessory minerals include biotite, zircon, apatite, rutile, sphene, opaque (ilmenite ?) and secondary chlorite.

The hornblende forms an intergrown, fine to medium grained mosaic with the other major constitutents. The stubby crystals are commonly pleochroic (light to dark brownish green in colour). Some large elongate hornblende crystals occurring in amphibolite a few hundred yards south of the summit of Candy Mountain display a very poikiloblastic texture. This hornblende is associated with large sphene and plagioclase (oligoclase) crystals.

The garnets are generally large (0.5 cm. wide on the average) and are brownish red in colour. They often are bounded by felsic minerals which give the garnet an "eye" like appearance. The garnets display a good euhedral form and the cores of some garnets have been observed in a graphic intergrowth with optically continuous quartz. Many of the inclusions within the garnets are opaque minerals (ilmenite ?) and sphene. Some peripheral chlorite (penninite ?) has been observed. The salic minerals form a mosaic intergrowth with the garnet and hornblende. Biotite, when present, is usually medium to dark reddish brown. K-feldspar, commonly occurring as perthite, occurs in many sections. In some places (e.g. in the central major unit near highway 69) dykes and pods of K-feldspar bearing amphibolite occur. The K-feldspar rich areas are closely related to fractures in the rocks and may be metasomatic in origin.

Garnet Amphibolite

This amphibolite unit is defined as containing more than 3% garnet and no diopside or anthophyllite. It is found throughout the area as minor beds and, especially in the central unit (2), as large outcrops. The unit is extremely variable in modal percentage of garnet and is gradational to biotite schists and "common" amphibolite. Many interbeds of K-feldspar-hornblende-biotite gneiss and other amphibolites occur within the unit.

In outcrop the rock usually weathers to a dull green brown and is conspicuous by the weathered out garnets. A good hornblende lineation is common although the foliation is usually poor. This rock and the non-varietal amphibolite often contain concordant pegmatitic veins rich in large albitic plagioclase, quartz, hornblende and sphene.

Common assemblages are:

Quartz-plagioclase-biotite-hornblende-garnet
" - " -hornblende-garnet
" - " -K-feldspar-biotite-hornblende

In thin section the garnet amphibolite is identical to the non-varietal type except for greater garnet content.

Diopside Amphibolite

This unit is defined by its pyroxene content and near total lack of garnet (and anthophyllite). The rock type occurs scattered throughout the area as large massive outcrops. The contacts with the surrounding rock are sharp and usually very few or no interlayers of other units occur (figure 5B).

No foliation or layering is developed. A good lineation occurs, usually parallel to that in the surrounding rocks. A large body near the center of the area has lineations differing in attitude from those in the enclosing rock.

The only assemblage observed is

Quartz-plagioclase-hornblende-pyroxene.

In thin section, the rock displays an even, granular texture of interlocking hornblende, plagioclase (andesine), and minor amounts of quartz. In one outcrop, large, dull green pyroxene clots up to one centimeter wide occurred. These may represent a primary igneous feature.

A number of samples show undeniable primary igneous textures, such as subophitic relations between large sericitized plagioclase grains and hornblende mantled augite and/or hypersthene. All gradations of mantling of (igneous ?) pyroxenes occurred. In many sections optically continuous "eyes" of pyroxene occur, scattered within a hornblende host.

The diopside amphibolite is inferred to have had a gabbroic protolith because of:

1. uniformity of composition
2. lack of interlayers of non-amphibolite compositions
3. sharp contacts with surrounding country rock
4. retention of some igneous textures.

Anthophyllite Amphibolite

This unit is defined by its fibrous (anthophyllite) amphibole content. The fibrous amphibolite totalled less than 30% of the rock and perhaps the term granofels should have been used. Only two small outcrops have been observed, both of which occur in the central major unit (2). The contact between the anthophyllite amphibolite and the surrounding amphibolites are sharp. No interlayers of other compositions occur within the unit.

The rock displays a slight to moderate lineation, moderate foliation and no layering. The lineation and foliation are consistent in attitude with those in the enclosing rocks.

In hand specimen the dark brown weathering rock is conspicuous by the occurrence of large pink garnets (to 2.5 cm. in diameter) and the fibrous amphibole. This amphibole is not observed in any other rocks type.

The only assemblage observed is

Quartz-plagioclase-anthophyllite-garnet-biotite.

In thin section, the coarse grained rock contains garnet, biotite, plagioclase (oligoclase), quartz and anthophyllite as the dominant mineralogy. Accessory minerals include rutile, opaque (ilmenite ?) and secondary chlorite.

The clove brown, parallel extinguishing anthophyllite occurs intimately associated with reddish brown biotite. The latter mineral commonly shows scalloped edges when in contact with the amphibole. The quartz and plagioclase form a granular mosaic. Large clear garnet crystals, containing numerous biotite flakes but no anthophyllite, are common in the rock.

The unit is considered to have been derived from a plutonic protolith along lines of evidence similar to those for the diopside amphibolite except for the lack of primary igneous textures.

Calc silicate Marble

Rocks were identified as belonging to this unit if they contained primary calcite. These rocks would fall under Shaw's (1957) definitions of marble (for pure calcite marbles with some garnet and/or diopside) to calc-silicate granulite (for diopside granulitic rocks). The former weather to a white to tan colour and the latter a medium brownish green. The calcite weathers first, so that the surface is extremely uneven depending on the original modal abundance and distribution of the minerals. This rough weathered surface is a good guide for identification in the field.

The rocks are distributed fairly randomly and commonly occur in very thin layers. Some of these layers are very continuous and can be traced for a 0.5 kilometer (central part of unit 2). The rock is interlayered with more salic bands and commonly a centimeter wide hornblende selvage occurs around the marbles. The following assemblages are seen:

Garnet-plagioclase-calcite-amphibole
" -diopside-epidote-calcite-quartz
" - " - " - " - " -plagioclase
diopside-plagioclase-calcite-quartz-epidote
" -epidote-quartz-calcite
garnet-amphibole-plagioclase-epidote-calcite-quartz-biotite
diopside-calcite-plagioclase-amphibole-quartz
diopside-calcite
diopside-quartz-calcite
diopside-quartz-calcite-amphibole-epidote

In thin section the composition of the marbles is extremely variable. The essential calcite normally occurs as an interstitial anhedral mineral. Near equant grains of greenish tinged diopside are common and volumentrically constitute a large part of most slides. In one sample collected near the east central part of the area, diopside occurs in a myrmekitic intergrowth with an unidentified mineral (quartz ?). The exact form of amphibole is problematical although the external form and colour is typical of amphiboles in the tremolite-actinolite series. The mineral varies somewhat in properties in different slides, although a blue green to medium green pleochroism and anhedral crystal form are usual properties. Garnet occurs as anhedral crystals in units very rich in calcite. Unlike any other garnets mentioned, it is reddish brown and most likely andradite rich. Epidote forms elongate, anhedral crystals. It is identified as pistacite mainly by its strong birefringence and parallel extinction. Plagioclase (An_{41} av.) and quartz appear as an intergrown mosaic between the diopside crystals. Biotite (medium to dark greenish brown) is a rare constituent.

The minor minerals include euhedral apatite and pyrite, some ilmenite and numerous radioactive opaques (identified by shattered aureole around grains and glassy appearance). In a trenched pegmatitic vein within a large marble layer near the Wanapitei river, two radioactive minerals were identified. One was identified as allanite by its brown appearance and crystal form. The other was black and formed elongate crystals suggestive of fergusonite. The close association of radioactive minerals and marble is very common in the Grenville province, especially near Bancroft.

There appears to be no systematic change of marble mineralogy with increasing metamorphic grade.

Basic Intrusions

A few small, oval and dyke-like bodies of intrusive rock were observed. These appear randomly distributed in the area and variable in composition. The outcrops are a dull greenish or brownish colour and are devoid of any structure except cleavage. Each body appears to be of a constant composition and has sharp contacts with the surrounding metamorphic rocks.

Figure 5.

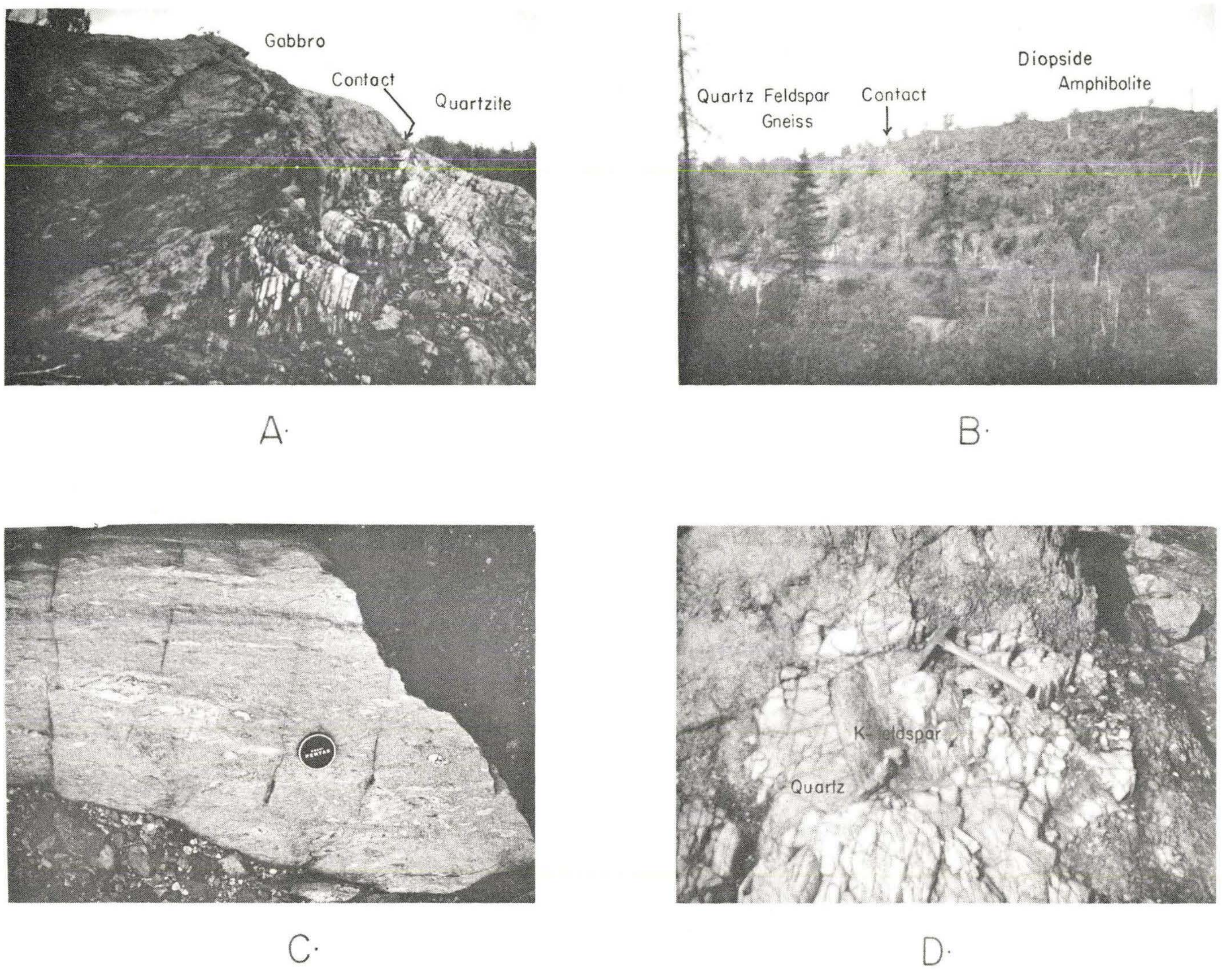
A. Photograph of a contact between gabbro and quartzite in Southern province rocks near highway 69.

B. Photograph of a contact between diopside amphibolite ("metagabbro"?) quartz-feldspar gneiss (meta-quartz sandstone?) in Grenville province rocks.

C. Photograph of biotite-rich amphibolite containing polymictic angular fragments from a local on highway 69 near the center of the area (Grenville province)

D. Photograph of coarse grained pegmatite containing a large perthite crystal (Grenville province).

Figure 5


Gabbro
Contact
Quartzite
A·
Quartz Feldspar Gneiss
Contact
Diopside Amphibolite
B·
C·
K-feldspar
Quartz
D·

The following types have been observed:

norite

quartz norite

gabbro

The norite occurs in the southeastern part of the area. Some of the small bodies have undergone so much uralization that they should perhaps be called amphibolites although their texture and at least half of the constituent minerals are still igneous. The contacts are sharp and no metamorphic structures are observed.

In thin section, the elongate, subhedral crystals of hypersthene contain exsolution lamellae. Plagioclase has commonly a high anorthite content (An_{70}) and is poorly twinned. In one section a small amount of quartz has been identified thus yielding a quartz norite. Brown or green hornblende mantles the pyroxenes to varying degrees. This suggests a hydrous reaction at fairly elevated temperatures. Rare opaque grains are generally of very fine grain size.

Near the southeastern part of the map area, a number of small dykes occur. These are very fine grained and highly uralitized. The plagioclase is sericitized and marginally altered to a fine grained mosaic of salic minerals. The pyroxenes are almost completely altered to blue-green hornblende. Many primary features still persist, particularly the subophitic texture. Minor amounts of equant opaque minerals are present. This rock and a gabbroic body near the Grenville-Southern boundary appear to have undergone "moderate" metamorphism.

Ultrabasic intrusions

These bodies are identical to the basic ones in form, weathering characteristics, and lack of structure. Many of them display large olivine porphyroblasts on weathered surfaces.

The following types have been observed

harzburgite

amphibole peridotite

The harzburgite forms a small body within major unit 3 (near highway 69). The equant anhedral olivine in the rock has been serpentinized especially in fractures and at the margins of grains. Many opaque grains occur in the serpentine. The hypersthene has been less altered. This alteration is probably due to normal weathering processes.

The amphibole periodotite occurs in the northern part of major unit (1) near the Grenville-Southern boundary. It is typified by highly serpentinized equant olivine grains, less altered orthopyroxene and essentially unaltered amphibole. The amphibole is clear and uncoloured, forming subhedral elongate prisms, and has an extinction angle of less than 10 degrees. It is randomly oriented, which suggests it did not grow in a directed stress field and is thought to be a primary phase. Many minute opaque grains (magnetite ?) were observed.

Pegmatitic intrusions

Most of the Grenville rocks contain very coarse grained dykes and sills which commonly have a composition related to the host rock. The kyanite bodies mentioned are found only in kyanite-bearing rock. Amphibolites commonly contain small bodies of coarse grained hornblende-plagioclase-quartz composition. Near the Wanapitei river, a large marble outcrop contains a coarse grained pegmatite containing diopside, brown garnet (andradite ?), plagioclase, quartz and large allanite/fergusonite crystals.

The various gneiss units contain the largest amount of coarse grained pegmatites consisting of K-feldspar (perthite), plagioclase, quartz, muscovite and biotite (see figure 5D). Many of these pegmatites have been mined for muscovite. These"gneissic"pegmatites appear to be formed by partial melting or a segregation effect, for they can be traced into unaltered gneiss.

The fact that the host rock commonly determines the composition of the pegmatitic bodies strongly suggests a local origin and is good evidence for the isochemical behaviour of rock units.

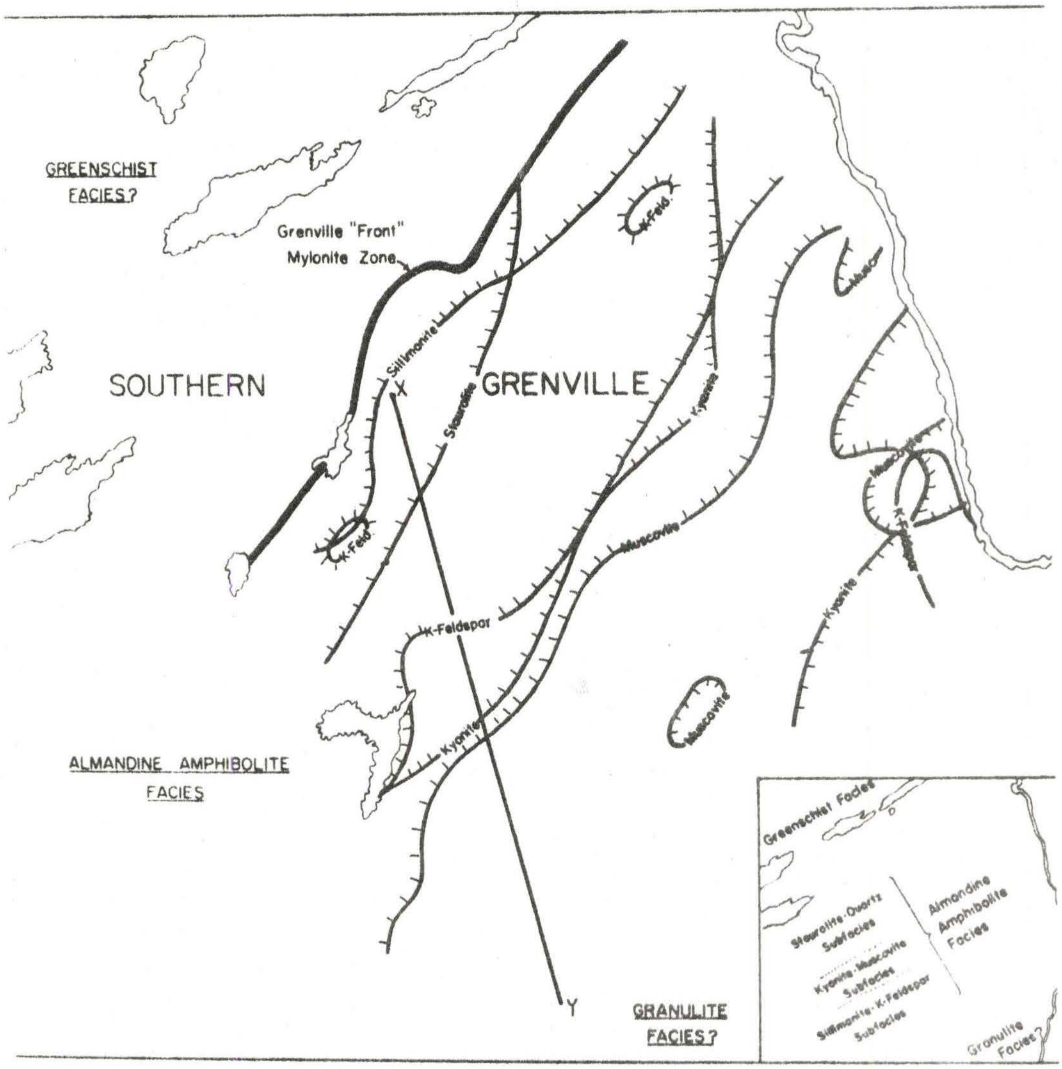
The size of some of the pegmatites is enormous, making them mappable. "Candy Mountain" is a large pegmatite near the center of the area, and is approximately 3/4 kilometers long and hundreds of meters wide. Often these pegmatites contain a faint foliation which is parallel to that in the enclosing rock.

(c) Metamorphism

The Grenville province contains rocks showing many grades of metamorphism and lithologies. An upper almandine amphibolite to granulite metamorphism probably exists in the Grenville except near Madoc, Ontario, where it is lower. The best review of the metamorphism of the Grenville province is in the Geological Association of Canada guidebook, 1967. The nodal highs, common to the Southern province, have not been identified in the Grenville.

The rocks underlying the Grenville province part of the area mapped are mainly within the sillimanite zone, as described for Scottish rocks by Barrow (1893) or the almandine amphibolite facies (Eskola, 1939). There seems to be a linear increase of metamorphism from the Grenville-Southern boundary to the extreme southeastern part of the area studied,

Figure 6



MAP OF ISOGRADS & METAMORPHIC GRADES

Map of the area studied showing the various isoglads as deduced from the mineralogy of the pelitic schists. The line x - y marks the position of a traverse from the lowest grade rocks at x to the highest grade ones at y and will be explained in section II,C. The hachurs occur on the low grade side of the isograds.

which will be discussed in Chapter II.

The near ubiquitous occurrence of sillimanite makes the exact classification of the metamorphic grade difficult. The lowest grade of the area is just below the breakdown of staurolite and is tentatively assigned to the staurolite-quartz subfacies of the almandine amphibolite facies (Francis, 1956). The highest grade is near the almandine amphibolite-granulite boundary. This is inferred from:

1. the passing of the muscovite breakdown isograd (figure 6);
2. the occurrence of the assemblage kyanite-orthoclase.

No metamorphic hypersthene has been found in the area although it can be seen in a road cut 25 kilometers south.

The metamorphic variation within the Grenville rocks is inferred from "isograds". Unfortunately this term has been interpreted and defined in different ways and the following two definitions were accepted for two different isograd relations:

> "(A) Isograds based on the first entrance of a particular mineral, minerals or polymorph. This group includes the classic isograds of Barrow.
>
> (B) Isograds based on the disappearance of a phase previously present."
>
> Atherton, 1965, p. 172.

As one traverses from the Grenville-Southern boundary to the extreme southeast, the following isograds are encountered in aluminum silicate bearing pelitic schists (figure 6):

(1) <u>Sillimanite isograd</u> - first appearance of

of sillimanite

(2) Staurolite isograd - disappearance of staurolite;

(3) K-feldspar isograd - first appearance of K-feldspar;

(4) Kyanite isograd - disappearance of kyanite

(5) Muscovite isograd - disappearance of muscovite

These isograds are well-defined by both hand contouring and a statistical trend surface analysis which will be fully described in section, II,C.

It can be seen on figure 6 that the isograds are somewhat irregular and may even cross (e.g. staurolite and sillimanite isograds). This may be due to the values of one (or more) of the intensive parameters (e.g. P_{H_2O}) which makes up the general metamorphic grade changing in an independent manner to the other intensive parameters, or differences in bulk compositions of the rocks. The latter possibility will be explained in section II,A. The irregularity of the muscovite isograd may be due to a decrease in general metamorphic grade and/or a local high P_{H_2O}. The irregular kyanite isograd, especially that occurring on the right hand side of the map (figure 6) probably indicates irregular fluctuations of metamorphic grade in that area. The general metamorphic grade along the line X-Y seems to increase regularly and can be explained by a fairly simplified (linear ?) gradient. North of this line, the gradient is more irregular.

The Grenville province metamorphism may be subdivided into the following subfacies as defined by Francis (1956):

(a) Staurolite - quartz subfacies, to the staurolite breakdown isograd;

(b) Kyanite - muscovite subfacies, to the muscovite isograd;

(c) Sillimanite - K-feldspar subfacies, from the muscovite subfacies to the extreme southeast

The metamorphism of the Grenville province rocks contrasts with that of the Southern province by the lack of andalusite in the Grenville province and kyanite (± sillimanite) in the Southern province.

V. Rock occurring at the boundary between the provinces

Rocks have been placed in this subdivision because they are compositionally or texturally distinct and are not observed in rocks away from the boundary in either province. The following units are included:

1. A gabbro body which contains K-feldspar and quartz crystals of indubitable post-intrusive origin.
2. Granitic bodies emplaced after the gabbroic body.
3. Cataclastized and mylonitized Superior and Grenville rocks.

(a) Gabbro

This rock is distinguished in the field from the Sudbury gabbros by virtue of its weathering colour. The Sudbury gabbros weather to a medium green while this body weathers to a brownish one. The gabbro often exhibits small weathered out impressions reminiscent of olivine crystals. The two different gabbros have not been seen in contact but their difference becomes apparent when one visits two outcrops of either gabbro located in the large swamp which marks part of the Grenville-Southern boundary south of highway 69. The distinctive gabbro outcrops fairly continuously within the boundary area between the two provinces.

Figure 7.

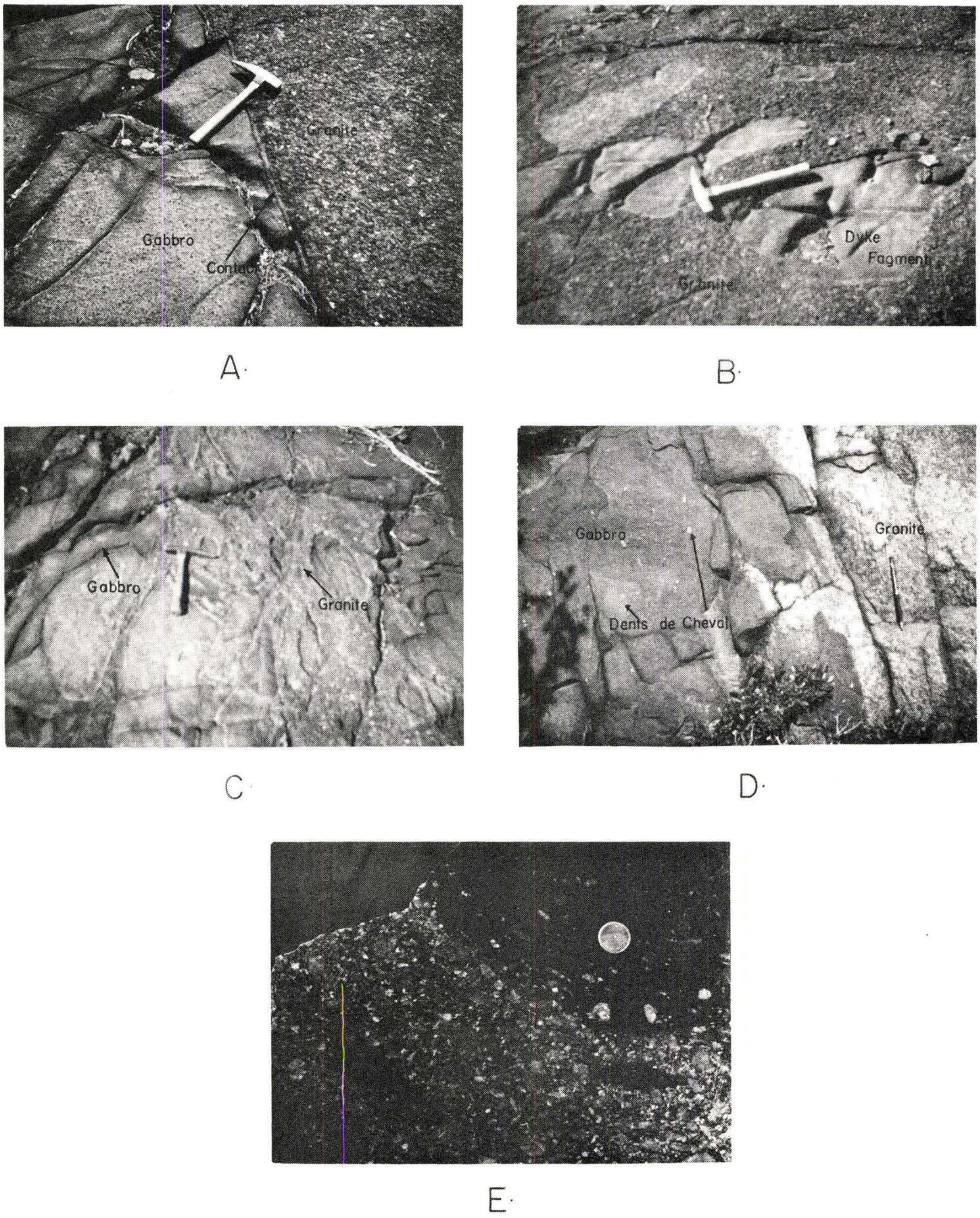
A. Photograph of sharp contact between gabbro and coarse grained granite located a kilometer southwest of the intersection of highway 69 and the boundary.

B. Photograph of fine grained dyke fragments within coarse grained granite (location is the same as 7A).

C. Photograph of interpenetration of granite within gabbro (located near 7A).

D. Photograph of "Dents de Cheval" within gabbro at granite-gabbro contact (located near 7A).

E. Photograph of cut and polished specimen showing the gradational boundary between granite and gabbro.



A.

B.

C.

D.

E.

Figure 7

In thin section, the gabbro consists of altered greenish pyroxene, unaltered brown biotite, plagioclase (An_{47}), equant opaque mineral(s) and much apatite. Serpentine clots, suggestive of altered olivine, exist in some areas. The biotite does not appear to have formed at the expense of any other phase and is probably primary. The composition of the gabbro is comparable to that in the diabase dykes occurring in the Southern province.

The unaltered gabbro grades into gabbro containing K-feldspar and quartz "metacrysts" as granitic bodies are approached. Some of the rock, especially near the farm house 200 meters northeast of the intersection of highway 69 and the boundary, is so rich in the two felsic minerals as well as having been partially mylonitized and altered, that the correlation of the rock as being gabbro is extremely difficult to maintain.

(b) Granitic rock

This unit may well be continuous to the southwest with the Killarnean granite located near Georgian Bay. The large body best marks the boundary between the two provinces and is inferred to have been intruded after the major period of metamorphism in the Grenville.

The origin of the body has been variously interpreted with the general area of study. Phemister *et al.*, (1961) believed it to be entirely formed by metasomatism of pre-existing Southern province and gabbroic intrusive rocks. Henderson (1967) studying much the same ground as Phemister, does not adhere to the latter writer's theory and believes the granite was emplaced in magmatic form at its present level. In the area studied the granite has:

1. fairly sharp contacts with surrounding rocks in some places (figure 7A);

2. compositions that fall in the minimum of the system Alb-Qtz-K-felds. (Bowen and Tuttle, 1950)
3. cross-cutting aplitic dykes of similar composition to the granite. These dykes were not observed in either the Southern or Grenville provinces proper.

In addition it may be noted:

4. large unaltered K-feldspar and quartz crystals occur within normal gabbroic rock. The crystals may or may not be related to the proximity of granite outcrops (figure 7D and E);
5. the extreme interpenetration of the granite in the gabbro (figure 7C);
6. the apparent occurrence of mafic rich granite when enclosed in mafic rock, and mafic poor granite, when enclosed in mafic poor rock (e.g. quartzite). This feature has not been quantitatively verified and needs much more study;
7. irregular gabbroic fragments and gabbroic pyroxene "metacrysts" making up a substantial part of some granite outcrops;
8. a linear array of fine grained gabbroic dyke fragments which occur within coarse grained granite. This suggests the fragments have not been rotated (figure 7B).

The features numbered 1 to 3 are generally indicative of a granite of magmatic origin and the latter ones (4 to 8) are more indicative of metasomatism. It is apparent that none of the features are unequivocal.

It appears that at least part of the K-feldspar bearing rock is metasomatically formed and there are strong inferences part is intrusive.

This leaves one with the problem "is it mainly metasomatic with small amounts of local mobilization or is it mainly intrusive with marginal metasomatism?". In the author's opinion, the granite in the area studied contained much marginal metasomatism but it is generally believed from a magmatic source.

The granitic rock varies in composition between granite and quartz monzonite although granite is more common in the area studied. Thin sections of rock, when observed from locals well away from contacts, contain the following minerals: perthitic microcline, green hornblende, green biotite, plagioclase (An_{15} av.), epidote and pyrite. The perthite consists of patchy plagioclase masses within large microcline host crystals. The quartz occurs as aggregates of elongate, irregularly shaped individuals which appear to "flow" around the feldspars and mafic minerals. Plagioclase forms small subhedral crystals scattered throughout the groundmass. The mafic minerals are generally anhedral and occur in concentrations between the larger feldspar grains. The epidote (pistacite ?) occurs throughout the slide but particularly in small, cross cutting fractures. This suggests it may be secondary. All of the granite observed was mechanically deformed.

(c) Cataclasites and Mylonites

The Grenville-Southern boundary in the area is marked by a zone of up to 200 meters wide of cataclastic and mylonitic rock. On approaching the contact between the provinces, one notices an increase of cataclastic textures until true flinty mylonites are observed. Two generalized sections are shown in figure 8. As can be seen, both mechanically deformed Southern and Grenville rocks occur separated by a zone containing

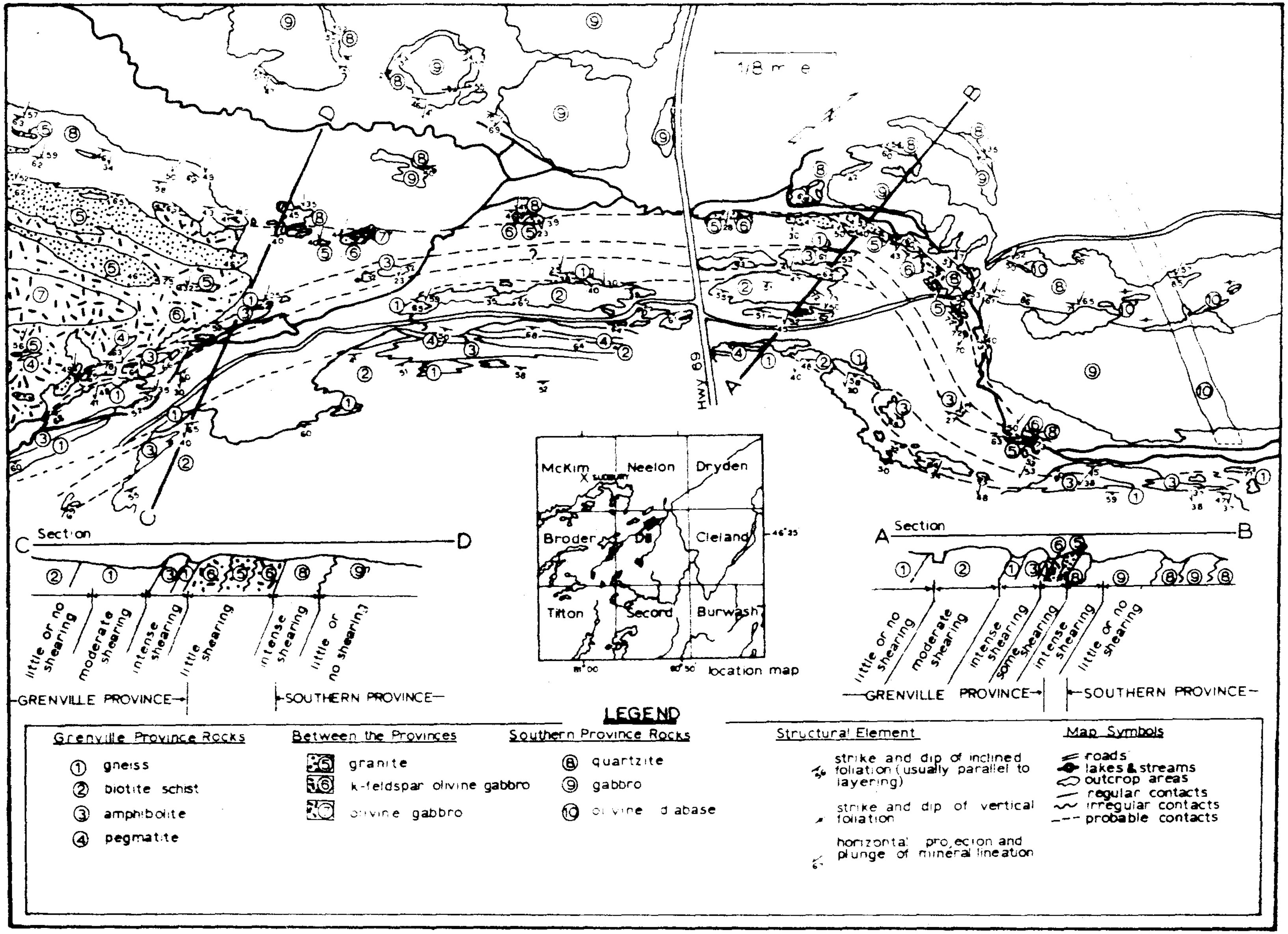


Figure 8

Detailed map showing the relations along the Grenville - Southern province boundary near the centre of Dill township.

partly deformed granite and gabbro. This zone may be only 40 meters wide but is so persistent that the mylonites derived from the rocks of the two provinces were not seen in contact.

The mylonites derived from the two provinces can be distinguished on the following basis: 1) Grenville province mylonites and cataclasites are inconstant in composition as one proceeds across strike. This feature is typical for the Grenville stratigraphy. 2) The Southern province mylonites are all quartzose and fairly homogeneous which is typical for metasedimentary rocks of that province near the boundary.

The largest part of the granite-gabbro complex occurs near the southwestern part of the mylonitic zone while most of the cataclastic and mylonitic rock occurs on the southern side of the body. Large blocks of mylonite occur within only slightly deformed granite in the extreme southwestern part of the granite-gabbro body. The layering of the mylonite blocks is not parallel to the foliation in the granite or to the trend and layering of the mylonite outside the granite. This feature suggests that at least some of the mylonitization occurred before the formation of the granite. Mylonitized Grenville province rocks occur practically in contact with very slightly deformed gabbro.

The mylonites and cataclasites are similar to those of the Moine thrust in northwestern Scotland as described by Christie (1963). The first observed indications of mechanical deformation in the field appears to be the fracturing and "smearing" out of large feldspar crystals within pegmatites. Some of the phyllosilicates in rocks near the boundary also show deformation in the form of kink zones within the crystals. This is seen within an eighth of a mile from the granite body. The garnets in many different rock types appear ellipsoidal although garnets of this form are found at significant distances (kilometers) from the boundary. Quartz exhibits undulatory extinction first on approaching the boundary and feldspar soon follows this behaviour.

The shearing effects increase until in the area nearest the other province, flinty mylonites occur. The salic minerals in rocks seem to be ground down more finely than ones rich in mafic ones; thus one finds salic mylonites within identifiable biotite schists. Schists display many cataclastic features but still contain complete, although highly altered and undulatory extinguishing, garnets and plagioclase. The amphibolites display no textures seen anywhere else in the Grenville. They are generally very fine grained and appear to have recrystallized. The amphiboles are bluish-green while the normal amphiboles in the Grenville are medium (not bluish) green. The plagioclase is low oligoclase while that in the Grenville is much higher.

Thus, the following appeared under the prevailing conditions:

1. amphibolites recrystallized;
2. most rocks of granitic composition became true flinty mylonites with no recrystallization;
3. the schists were least affected possibly due to pre-existing foliation capable of transmitting movements more easily than in other rocks.

The flinty mylonites display all the features described by Christie. Small augen of feldspar and hornblende occur sporadically distributed in most sections (figure 9E). Some layers have been so milled down as to be aphanitic in appearance. The rock type contains numerous small cross-cutting fractures which have been filled with anhedral quartz crystals. The colour of the mylonites varies from pinkish to black. In a few rocks found in the zone of most intense deformation, some large crystals have persisted (figure 9B, C) but even these have undergone much marginal brecciation. The laminated mylonites have been isoclinally folded (figure 7D) and display good lineation which is parallel to that in the Grenville and Southern provinces.

The granite and gabbro in the very center of the most mylonitized area are only partly affected. The occurrence of K-feldspar porphyroblasts and rotated mylonitic blocks indicate that most of the mylonitization occurred prior to the emplacement of the intrusions.

The gabbros and granites have been mechanically deformed enough to display undulatory extinction, marginal brecciation of feldspar grains and flowage of quartz but not enough to mill down the large (to 2.5 cm. long) K-feldspar crystals.

D. STRUCTURAL SETTING

i) Large Scale features

The term "megascopic" applies to features larger than outcrop size. The structural features are shown in figure 10.

(a) Southern province.

In the Southern province two large folds exist, readily identifiable by the outcrops of Sudbury gabbro (figure 2). These fold forms may be:

1. part of the same fold;
2. two folds with a very tightly folded intervening synform.

Figure 9.

A. Photograph showing the extreme proximity of the Grenville and Southern provinces.

B. Photograph of large feldspar augen within mylonitic ground mass.

C. Photomicrograph of B (approximately 40x) showing large augen.

D. Photograph of folded, finely laminated mylonite.

E. Photomicrograph of D (approximately 40x) showing small augen and layers of aphinitic "ultra" mylonite.

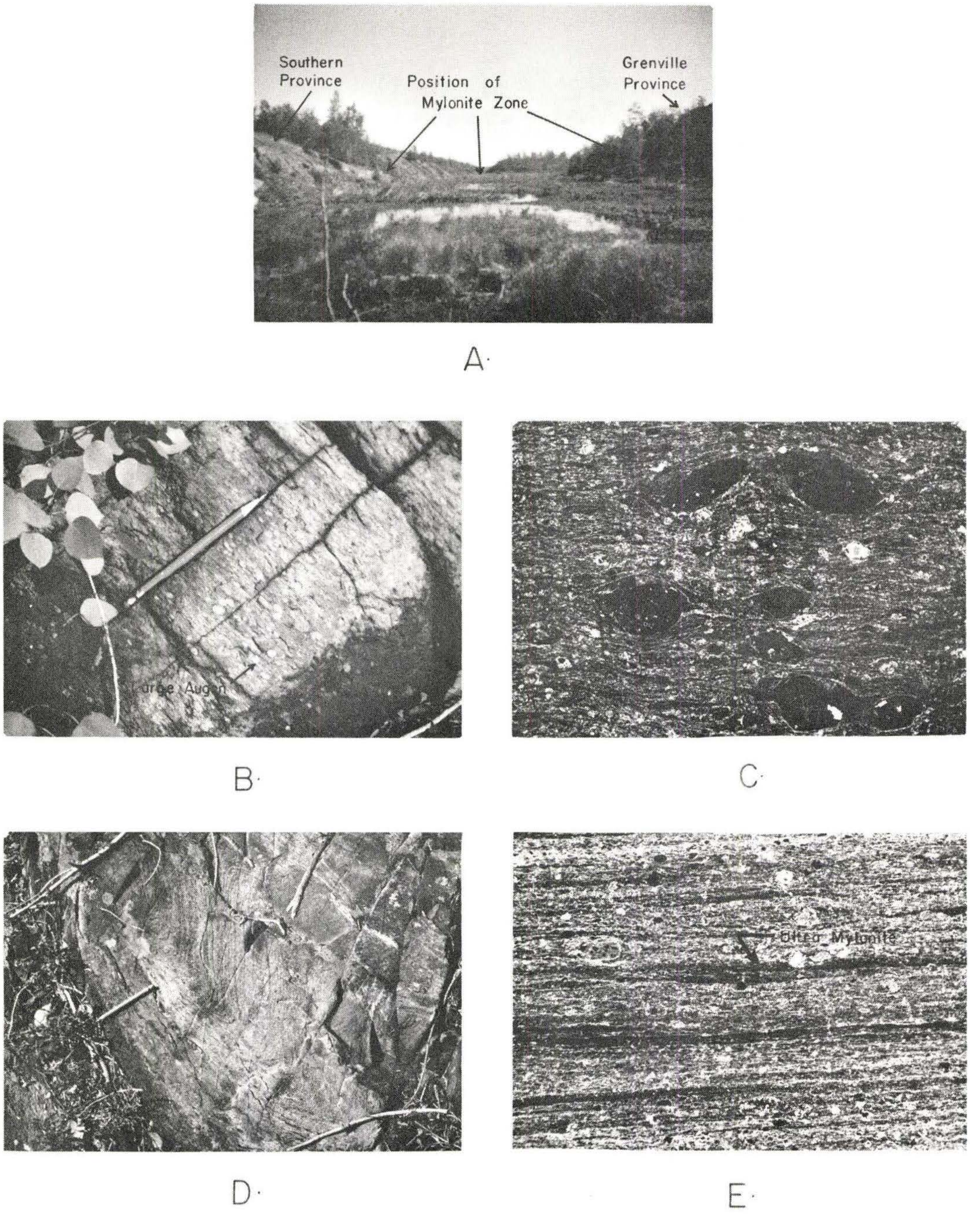

Southern Province
Position of Mylonite Zone
Grenville Province
A.
Large Augen
B.
C.
D.
Ultra Mylonite
E.

Figure 9

The latter explanation is preferred although either is possible. The tightening of the possible synform may be related to the warp in the foliation and layering present near highway 69 and the boundary between the provinces.

The gabbro bodies have been metamorphosed to a certain extent but no textures or structures related to the major folding were observed. Whether the gabbroic bodies intruded previous fold forms or were folded, is not well known. Young (personal communication) believes that in a similar situation near Georgian Bay, the gabbros intruded previous fold forms. The warping of the fold limbs probably occurred after the main period of folding.

In the Southern province, the strike of the major sedimentary units is parallel to the axial planes of the major or fold forms. These both have strikes which are parallel to the boundary between the provinces.

(b) Grenville province.

The Grenville province in the area studied displays a fairly uniform distribution of structural features. There are, however, a number of major "warps" in foliation and layering which are difficult to explain. The warps seem to "swirl" around major gneissic bodies. Three possibilities exist:

1. They represent isoclinal folds.
2. They represent a "gneissic dome" like structure.
3. They are produced by compression around resistant gneissic bodies.

The isoclinal possibility may be valid but usable marker horizons do exist and do not deviate as would be expected in an isoclinally folded terrain. The gneissic dome possibility is difficult to demonstrate as the foliation dips very steeply over the whole area. The last possibility seems most appealing as it best explains why the competent gneisses are at the centers of the worls.

ii) Small scale features

The small scale structural features are those smaller than an outcrop.

a) Southern province

The Southern province rocks studies contains few minor folds. Those seen are isoclinal and have foliation parallel to their axial plane. In these there is a well defined mineral lineation parallel to the fold axis.

The foliation in the Southern province rocks is poorly developed, undoubtedly due to the low grade of metamorphism and the general lack of appropriate compositions to produce phyllosilicates. Where measurable, the strike of the foliation is generally parallel to the Southern-Grenville boundary. The dips are steeply southeast and generally parallel to those in the Grenville province. Usually the foliation is parallel to the bedding except in a few areas which are interpreted as being hinge zones of folds. The foliation is generally parallel to the limbs of the few isoclinal folds observed. The poles to foliation appear to lie on a great circle.

There is a pervasive mineral lineation in the area which is generally parallel to the fold axes. This mineral lineation is nearly completely lost 3 kilometers away from the boundary between the provinces. The lineation appears to be parallel to the elongation of stretched pebbles in a conglomerate layer occurring on the northwest shore of Richard Lake, from which it may be inferred the lineation for at least part of its history coincided with a transport direction. This latter feature has also been suggested by Henderson (1967).

Fractures are a common feature in the area. These occur at various orientations but the dominant one has a strike near 130° and a 80° dip southwest. The fractures (or joints) of this set are spaced from inches to feet apart and have been regarded as fracture cleavages in other areas (Spaven, 1966). The contoured plots of the poles to the fractures are shown in Figure 10. The average of the poles represents a plane which contains the mineral lineations and which is approximately normal to the average foliation plane. This suggests the fractures are related to the folding.

The diabase dykes in the area studied and in the general Sudbury area have a strike which is parallel to the dominant fracture set. This suggests the fractures were the loci for diabase intrusion and predate the intrusions.

A few kink zones were noted in the Southern province, but these were only measured near the boundary at highway 69.

In summary, the Southern province can be shown to have undergone a period of intense deformation which has yielded a limited number of isoclinal folds and a period of near brittle deformation. Other periods cannot be demonstrated, which may be due to the rock type's competent nature.

b) Grenville province

The Grenville province rocks in the area studied have been subdivided into four domains. Two of the domains border the boundary between the provinces and are separated at the major inflection in the boundary. The other two parallel these first two and occur farther into the Grenville province. The purpose of this separation is to see if there are any systematic structural changes on approaching the boundary.

The first observed period of folding was identified in the field as having mineral lineations parallel to the fold axis (figure 10). The fold types also were more isoclinal than those of other periods. The mineral lineation lies very close to the great circle which is the average of the poles to foliation. The axes of the first folds are nearly coincident with the mineral lineation and the average of the poles to axial planes is within the plot of the poles to foliation.

The second observed period of folding was identified on the basis of:

(a) folding of previous first period folds (see figure 11A).

or (b) folding of mineral lineations (see figure 11B).

These second period folds are more open than the first type. The axial planes of these folds are nearly parallel to those of the first period

Figure 11.

A. Photograph of folded first period fold (left hand side of photograph).

B. Second period fold with folded mineral lineation (to the right of hammu handle.

C. Isoclinal first period folds.

D. Extreme quartz-feldspar rodding near mylonite zone.

E. Photomicrograph (approximately 40 x) of folded sillimanite (fibrolite) by second period fold.

F. Photomicrograph (approximately 40 x) of "rolled" garnet (sample T-78-1).

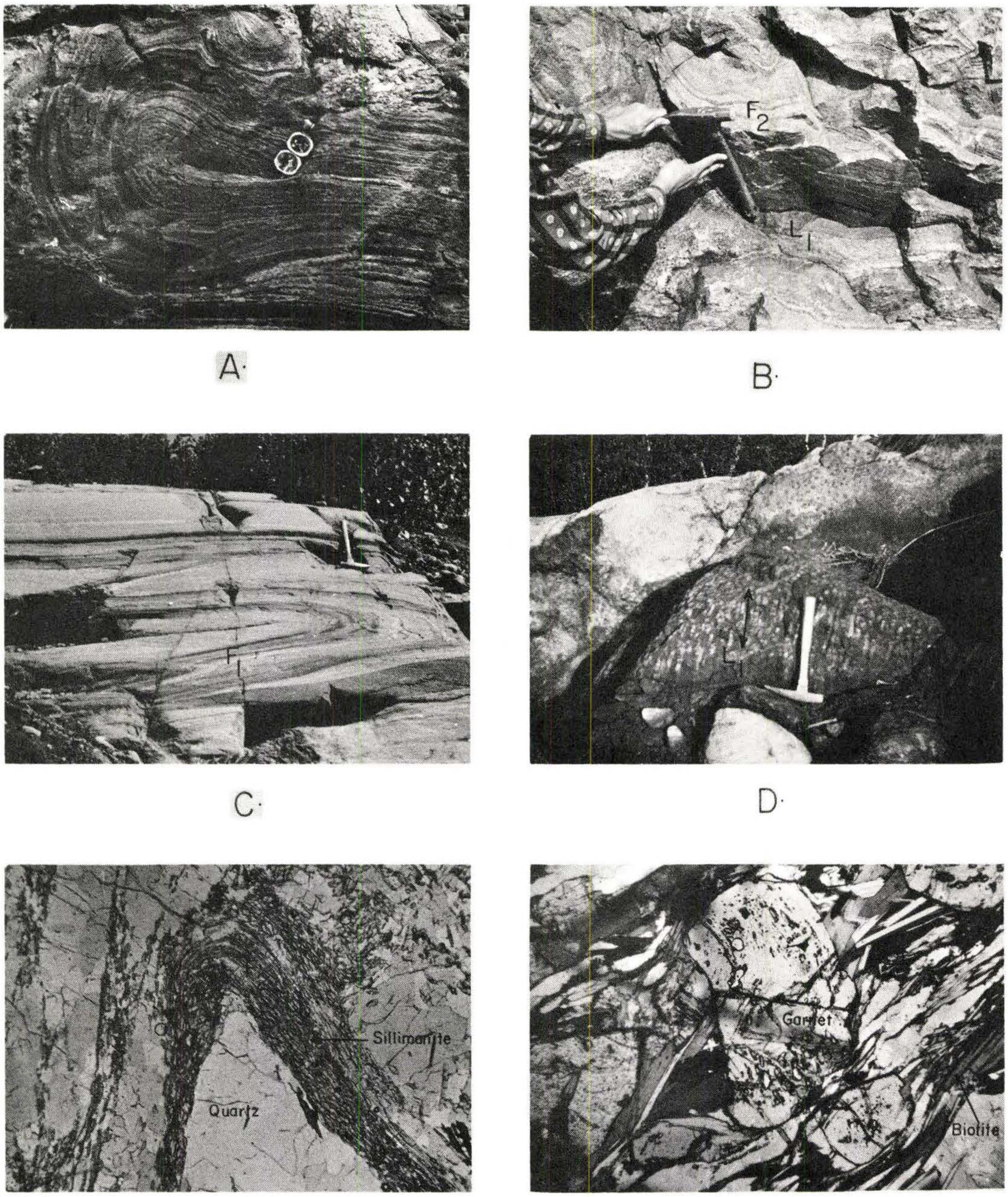


Figure 11

although the axes are more northerly trending. The second period of folding is considered responsible for the folding of the mylonites. This latter conclusion is based on the fact that mylonites which have normal Grenville stratigraphy contain crushed minerals which in the undeformed Grenville province constitute the mineral lineation. This mineral lineation is parallel to the first fold axes. The style of the second period folding and that in the mylonites are also similar.

The interpretation of two periods of folding is preferred over a continuing single period as suggested by Wynne Edwards (1963) because of the mylonitization between the two. The concentric form of the second period type also suggests that conditions were different. The time break between the two periods is unknown although a fairly small one is suggested.

The third period of deformation produced conjugate kind bands (figure 10) mainly within the thinly laminated mylonites. Features such as this have been described by Ramsay (1962), Donath (1964) and Paterson and Weiss (1966).

Many Grenville layers show a late gentle warping. The precise attitude of the broad warps could not be measured because of their large size. The largest such warp is demonstrated by the deflection of the mylonite near highway 69. The kink bands are particularly well developed in this area which suggests there is a relation between the two features.

The foliation in the Grenville province strikes northeast-southwest, parallels the boundary and dips steeply southeast. As in the Southern province rocks, there is a distinct down dip lineation: however, unlike the Southern province, there are numerous minor folds. It can be demonstrated by the stereonet plots (figure 10) that the foliation and the first period of folding are closely related.

There are a number of "gash veins" within the Grenville province which are in form, elongate elliptical gash-like features, they seem to form at an angle of approximately 30 degrees sinistral to a dominant fracture set. The fracture set is the same as the one described in the Southern province and is inferred to belong to the first period of folding.

The Grenville-Southern Province Boundary

In the study area, the following features mark the boundary between the Southern and Grenville provinces:

(a) a distinct break in metamorphic grade;

(b) a 100 to 300 meter wide zone of cataclasites and mylonites;

(c) a coarse grained granite which lies between mylonites derived from Grenville and Southern province rocks.

Regionally, there is a steep metamorphic gradient in the Grenville province which decreases from near the lower granulite facies, 10 kilometers southeast of the boundary to the staurolite-quartz subfacies of the amphibolite facies at the boundary. A decrease of metamorphic grade from the boundary to the northwest is also inferred in the Southern province.

There is also a decrease in the amount of folding going from Grenville into Southern province rocks. A distinct penetrative mineral lineation occurs in both provinces within a few kilometers of the mylonite zone. In the Southern province it is lost a few kilometers northwest while in the Grenville province it marks the geometric "b" axis of first period isoclinal folds.

The break in metamorphic grade is not obvious in terms of metamorphic facies because the staurolite-quartz subfacies occurs on both sides of the boundary. However, staurolite coexists with andalusite in the Southern province while in the Grenville province staurolite coexists with garnet, kyanite and sillimanite. The rocks in the Southern province have retained most of their primary sedimentary features while such features rarely are observable in the Grenville province rocks. The inferred temperature break is approximately 100°C and pressure, 2 kilobars (described in section II.G)

The difference in the amount of folding may simply be caused by differing lithologies. The Southern province consists mainly of very homogeneous quartzite while the Grenville province lithology is heterogeneous, containing many schist and amphibolite layers.

The pervasive mineral lineation was undoubtedly related to the movement on the mylonite zone (as suggested by Henderson, 1967) and probably is a reinforcement of previously existing axial lineations of first period folds.

The Southern province rocks are believed to have been deposited between 2500 and 2100 million years ago (Van Schmus, 1965). They were metamorphosed during the "Hudsonian" orogeny (Stockwell, 1964) which gave rise to the "Penokean" fold belt about 1700 million years ago.

The Grenville province rocks were mainly deposited prior to 1700 million years ago as indicated by Rb-Sr dates on gneisses (Krogh in Davis, et al., 1967; Grant, 1964). The same rocks share the general K-A date of 900 to 1100 million years for the Grenville province as a whole (Stockwell, 1964). Some Rb-Sr ages of near 1000 million years on granite have been reported near the Grenville-Southern boundary (Grant, 1964).

The Grenville-Southern boundary in the general area of this study has been interpreted by various authors. Two different hypotheses have evolved:

(a) The abrupt discontinuity is caused by the termination of "movement, recrystallization, remelting and feldspathization each acting as a trigger to the other". The boundary is the "front" of the above activity (Phemister, 1960).

(b) The abrupt discontinuity is caused by "a dynamo-thermal metamorphic gradient, east of which Southern province rocks were transformed during the Grenville orogeny into gneiss and amphibolite with little or no addition of exogenous material, but with considerable chemical and mechanical reworking" (Henderson, 1967).

Phemister stresses the feldspathization which he believes transformed some of the Southern Province quartzites and gabbros into granitic rocks. Henderson believes the same granitic rocks are intrusive in their present setting and essentially obscure the isograds between the sillimanite zone in the Grenville province and the biotite zone in the Southern province.

To the present author, the following possible interpretations exist:

1. That the Grenville-Southern boundary marks the geographic termination of the effects of the Grenville orogeny.
2. That movement on the mylonite zone has brought high grade rocks which were metamorphosed during the Grenville orogeny into juxtapoisiton with Southern province rocks, not affected by the Grenville orogeny.
3. That movement on the mylonite zone has brough high grade "Penokean" rocks into juxtaposition with their lower grade equivalents.

The first possibility has been suggested by both Henderson and Phemister and seems a "reasonable interpretation for places where wide areas of "Killarnean" granite exist. However, this is not the case in this area.

The geographic termination of orogenic effects in as sharp a manner as is suggested by features on both sides of the boundary in the area studied seems unlikely. For the "Front" to be the termination of a pervasive metasomatism as suggested by Phemister also seems unlikely.

The second possibility is that the termination of the Grenville orogeny was more gradational but that movement on the shear zone was obscured the gradation. This would necessitate having a termination of the Grenville orogeny within kilometers of the present Southern province rocks or movement on the shear zone of many kilometers.

The final possibility involves the movement on the mylonite zone bringing high and lower grade "Penokean" or earlier metamorphic rocks into juxtaposition. This model is preferred. If the curve representing metamorphic grade, non-stepped, be adjusted to give a uninterrupted curve, it implies a relatively small movement on the mylonite zone. The granite body needs simply to have been emplaced in the central part of the zone of movement marking the boundary. The extrapolated (not measured) variation in age dates between the two provinces could be accounted for by Grenville province rocks being subjected to temperatures above those at which argon can be retained in a certain crystal structure, for longer times than the lower grade equivalent Southern province rocks. Movement on the mylonite zone could have brought the rocks up, above the necessary geotherm at which argon can be retained and a discrepancy of K-A dates for the provinces was produced.

CHAPTER II

THE GEOCHEMISTRY OF THE PELITIC SCHISTS

A. INTRODUCTION

The purpose of the study is to explain the changes in the composition, shape and amount of minerals in the pelitic schists with increasing grade. The schists are homochemical to the first approximation (to be discussed in Section II, A, ii) and the changes arose from various solid state reactions. Conclusions about the chemical species participating in the reactions will be found from the following types of evidence:

(a) textural evidence;

(b) modal abundance variations of minerals;

(c) chemical variations of minerals.

Interpretation of any one of these features is equivocal but together they constitute strong evidence for various reactions.

The features in the study are closely interrelated so that certain features are mentioned in early sections which will only be explained later.

i) Sampling

The samples collected in this study occurred exclusively within the Grenville province. The aluminum silicate bearing schists are distributed sporadically throughout the area (Figure 1) in layers from a few centimeters to tens of meters wide and from a few meters to tens of meters long. The rock type was usually found within sequences of garnet amphibolite and garnet-bearing quartz feldspar gneiss.

When all the study material cannot be removed, one must take samples and is consequently faced with the query "Are the samples representative of the material as a whole?" Many sampling schemes exist and one of the most systematic involves an equilateral grid having only the bias of orientation on the geographical map and spacing of the grid points. Such a scheme was not applicable in this study as only a very small part of the area is underlain by pelitic schists and many large areas exist where no pelitic schist occurs.

The hand specimens were collected from all observed aluminum silicate-bearing rocks. Each sample was approximately 10 cm. square. The samples were selected with the bias that where two aluminum silicates occurred in the outcrop, a hand specimen containing the two was collected. Fourteen samples, collected along a traverse across the center of the area, were used for chemical determinations.

ii) Compositions of the Whole Rock Samples

In studies involving the chemical variations of minerals with changing metamorphic grade, it is usually assumed that the rock bulk compositions are constant. An estimate from the mineralogy is not entirely satisfactory as certain elements are included in phases in minor amounts (e.g., MnO and CaO) and can not be determined. Whole rock analyses are thus necessary although even then the loss of such mobile components such as H_2O from rocks with increasing grade would cause variations from ideal compositions.

To decide if two rocks are different because of bulk compositional or metamorphic differences is extremely difficult with the present state of knowledge. Usually, a graphical analysis is used where the **minerals** of two rocks are plotted in terms of a simplified composition diagram and apparent inconsistencies are pointed out. Greenwood (1966) approached the problem by mathematically extending the graphical approach to a larger number of dimensions. The variation in composition, to a first approximation, should not be large enough to cause the exclusion of a phase from an assemblage (e.g., Staurolite probably needs a fairly small range of composition to be stable). The apparent homogeneity will be judged subjectively in this study and is open to reinterpretation when a meaningful method to measure the effects of inhomogeneity is proposed.

All the rocks were crushed to less than 100 mesh and analyzed for their major element contents (see Appendix II for the analyses).

The variations of the oxide compositions are shown graphically in Figure 12 and are listed in Appendix II .

$\underline{SiO_2}$

The variation of silica in samples across the traverse is somewhat irregular. This irregular variation will have little effect on the subsequent mineralogies as SiO_2 is in excess and forms its own phase, quartz. The rocks near the end of the traverse (T-293-3 and T-296-1A in Figure 10) were low in silica but high in quartz because of the quantitatively abundant garnet.

$\underline{Al_2O_3}$

The alumina variation is within a few weight percent of 20 percent. The samples near the end of the traverse are slightly lower than those near the beginning.

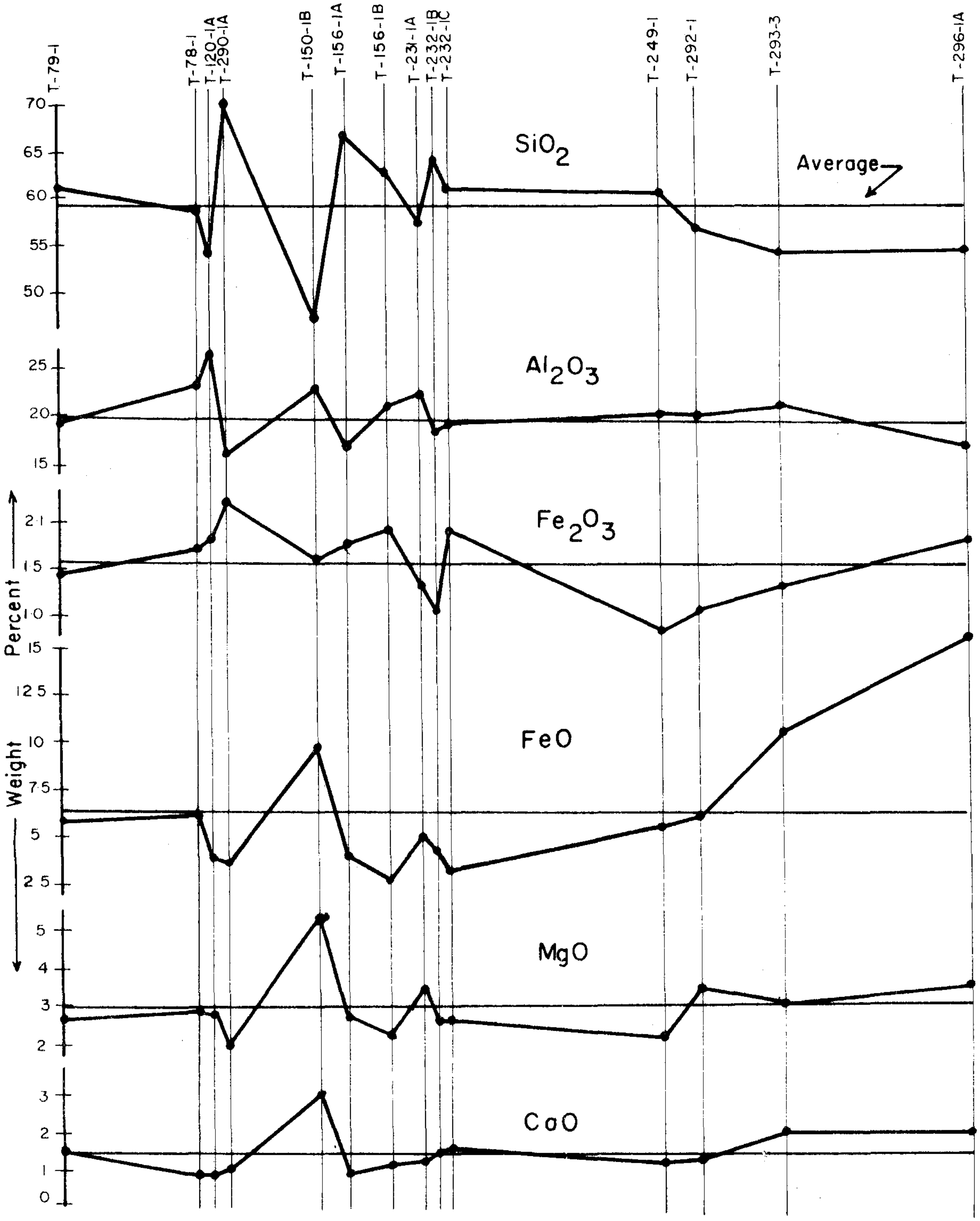
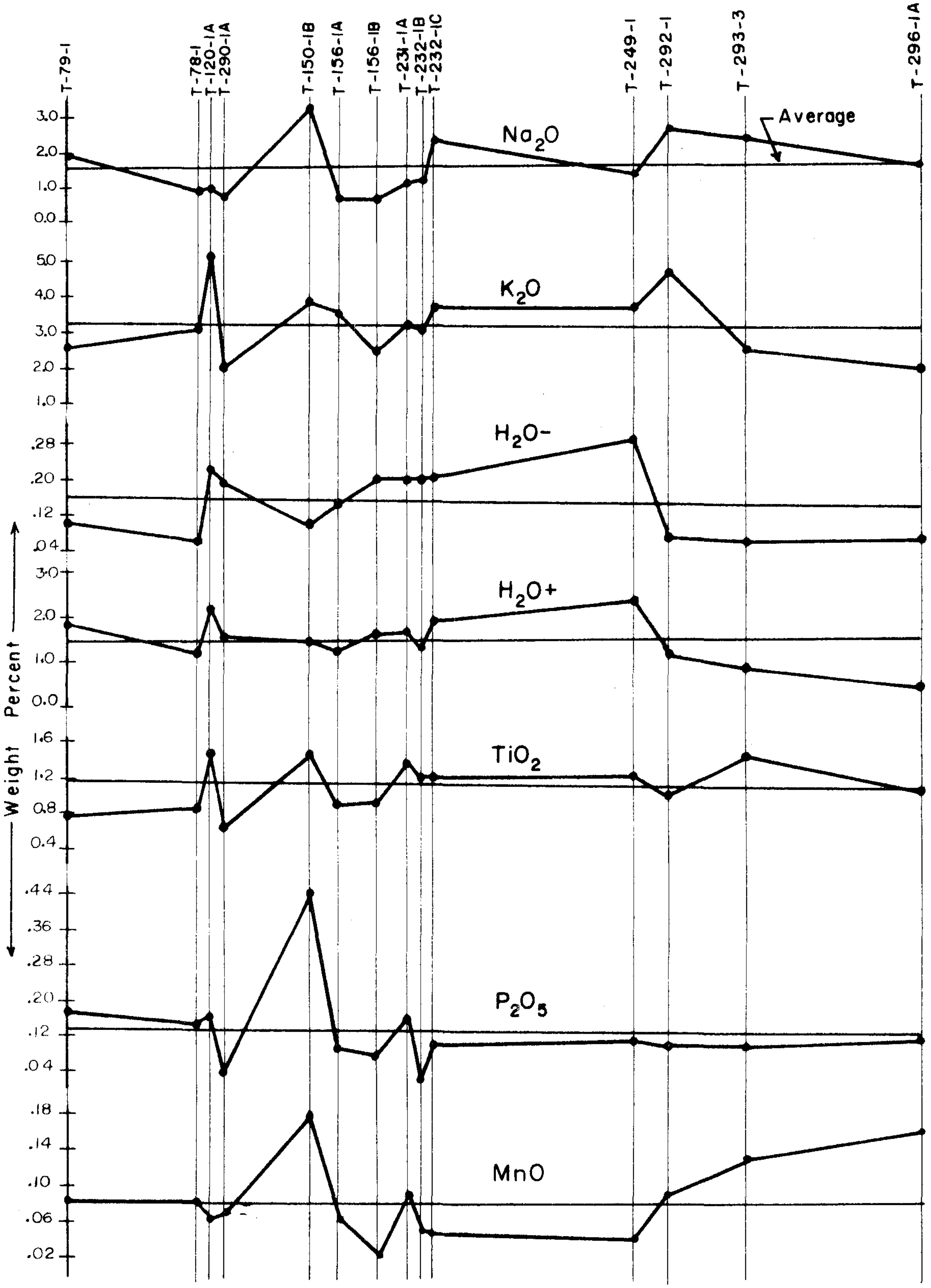


Fig. 12 Variation of Oxide Weight Percent in Whole Rock Samples with Metamorphic Grade


T-79-1
T-78-1
T-120-1A
T-290-1A
T-150-1B
T-156-1A
T-156-1B
T-231-1A
T-232-1B
T-232-1C
T-249-1
T-292-1
T-293-3
T-296-1A
Average
Na$_2$O
3.0
2.0
1.0
0.0
K$_2$O
5.0
4.0
3.0
2.0
1.0
H$_2$O-
.28
.20
.12
.04
H$_2$O+
3.0
2.0
1.0
0.0
TiO$_2$
1.6
1.2
0.8
0.4
P$_2$O$_5$
.44
.36
.28
.20
.12
.04
MnO
.18
.14
.10
.06
.02
Weight Percent

Fe_2O_3

This oxide varies greatly, although a general trend towards decreasing values near the end of the traverse seems to hold (except for Sample T-296-1A). Some of the positive variation from the average may be due to limonitic contamination (e.g., Samples T-120-1A and T-232-1C).

FeO

This oxide shows a general increase in content near the end of the traverse. Sample T-150-1B is anomalous to the samples found near it in being enriched in FeO. The increase in FeO near the end of the traverse is much more than can be accounted for by the reduction of Fe_2O_3 and appears to be a major chemical inhomogeneity.

MgO

The distribution of this oxide is fairly constant in all the samples except for T-150-1B which is anomalously high.

CaO

This oxide is similar to MgO, being of fairly constant distribution except in Sample T-150-1B where CaO is anomalously high.

MnO

MnO shows a variable content in the samples and is especially enriched in Sample T-150-1B and the last three samples in the section.

Summary

Samples T-150-1B, T-293-3 and T-296-1A are anomalously enriched in mafic elements (Fe, Mn and Mg). The latter two samples are **impoverished** in Si, Na and K. Samples T-120-1A, T-156-1A and T-290-1A are enriched in salic constituents (Na, K, Si) over mafic ones. Samples T-150-1B and

T-296-1A are the most anomalous of the samples but a general homochemical model for the important oxides MgO, CaO, MnO, and Al_2O_3 in the samples is inferred to a first approximation.

iii) Significance of Deviation of Rock Samples from being Homochemical

In studying the geochemical behaviour of a sequence of rocks of variable metamorphic grade, one is faced with the problem "are the observed mineralogical differences between two or more samples from different areas due to bulk compositional (extensive variables) or metamorphic (a composite of different interacting intensive variables) differences?"

It is probably true that no two rocks occur which are of identical composition although cogenetic rocks produced by the same igneous event (i.e., a basalt flow) and subsequently metamorphosed,may have differences within the error of measurement and could be considered identical. To test equilibrium partitions of elements between co-existing minerals, it is desirable to have a sufficiently different bulk rock composition for different samples so as to produce a linear and not an equant distributions plot at one grade of metamorphism, as an equant plot is non-determinate. Conversely, we presumably need a limited variation in bulk composition so that the rocks maintain the same phases except for those produced as a result of metamorphic changes.

The problem of whether the compositions of the samples have remained constant through geological time is always to be considered. There is much evidence for the mobility of some constituents in rocks (e.g., pyrometasomatic ore deposits) and also a whole theoretical basis for such mobility (Korzhinskii, 1959). Metasomatic additions and subtractions of components

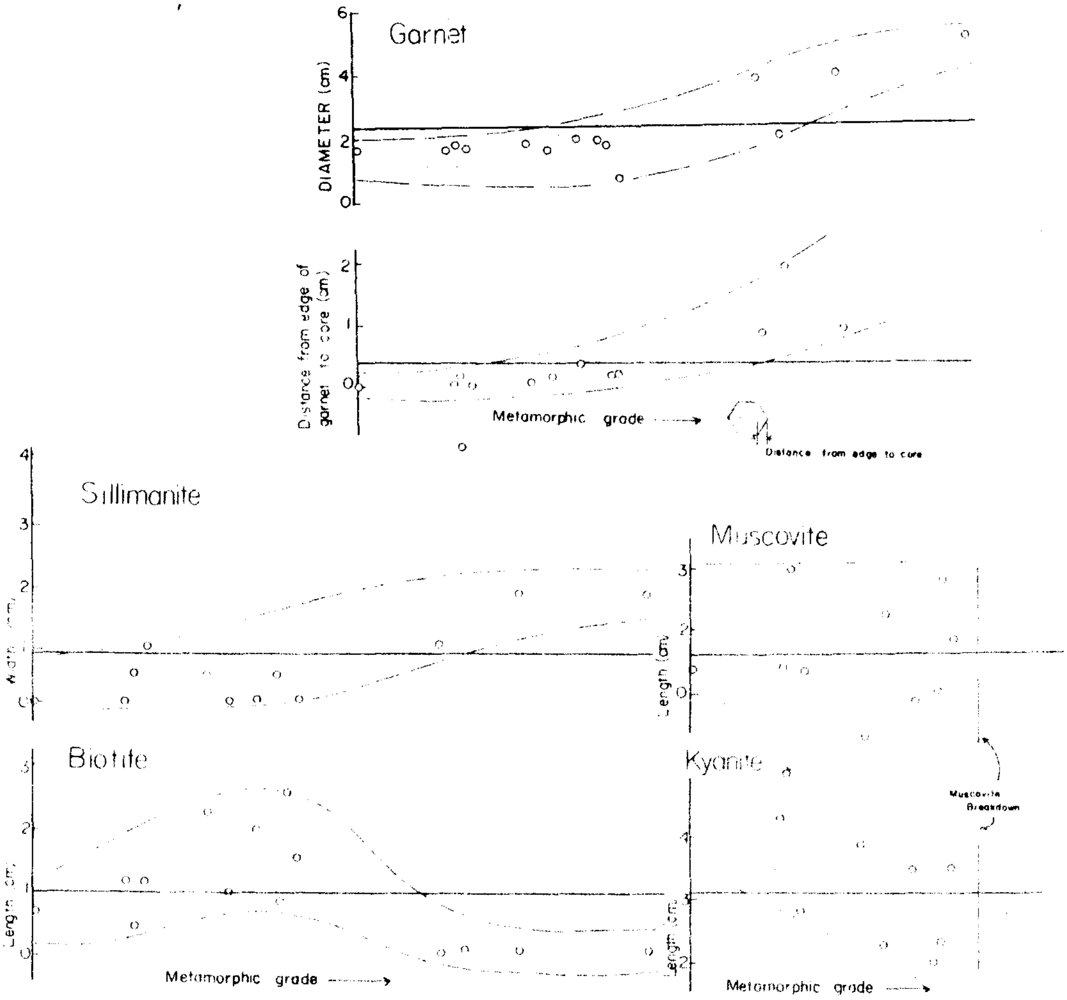


Fig. 13 Variation of grain sizes of **biotite** sillimanite, muscovite, kyanite and garnet across the metamorphic section. The increase in width of the outer clear zone of garnet is also shown.

have been amply described with respect to the granite problem. High grade metamorphic terrains are considered by some to be subject to much metasomatic change (e.g., Engel and Engel, 1958) and not so much by others. A guide as to which is true in the area studied was the occurrence of numerous pegmatitic bodies which had compositions closely related to the host rock and not to surrounding rocks. The pegmatitic bodies in the pelitic rocks were kyanite-bearing quartz sills and fibrolite lenses, as already mentioned. These two features occur up to the breakdown isograd of muscovite after which they were no longer observed. This appears to be unequivocal proof that at least some silica and alumina moved from the host rock, to form independent rock types and either went back into the same host rock or was lost to surrounding rocks when the muscovite isograd was reached. The "fairly" uniform alumina content of the whole rocks would suggest that the subtraction or addition of the kyanite and sillimanite bodies have not interfered with the bulk composition too much. To be exact, one would have to consider these bodies, but their sampling would cause many problems. Any addition or subtraction of other constituents is unknown. The assumption is made that little metasomatic change has occurred with the possible exceptions of water, oxygen, hydrogen, carbon and sulphur.

The pelites had an argillaceous protolith which formed in clastic sedimentary environments and was very susceptible to varying bulk chemical inconsistencies. Ideal sampling would be done on a continuous bed extending from low to high grades. Few beds of constant shale composition and extending for many miles exist in the observed sedimentary sections and even fewer are known in metamorphic ones. The beds studied are not interconnected and undoubtedly vary in composition. It is believed that a large enough number of samples was collected (approximately 50) to make meaningful inferences about isograds and changing modal amounts of a phase.

B. DESCRIPTION OF MINERALS AND TEXTURES

The minerals comprising the pelitic schists will be described in this section. When physical measurements are shown they are of the average of ten widely spaced grains in the polished thin sections. The "length" refers to the longest dimension of the grains and the "width", to a measurement perpendicular to the length.

i) Staurolite

Staurolite was identified optically and from electron microprobe studies. The mineral occurs in contact with all the minerals in the assemblage only near the Grenville-Southern boundary. A half kilometer southeast of the boundary it only occurs as inclusions within garnet (Figure 2C) and can not be considered in equilibrium with the other phases. When the volume percentage is less than 0.8, staurolite only occurs as inclusions in garnet and this percentage defines the staurolite isograd (see Figure 6).

Below the staurolite isograd, rectangular sections of this mineral display a poikiloblastic core, surrounded by a clear outer rim (Figure 2B) and commonly have a conspicuous corona of interlocking quartz grains. This latter texture cannot be explained.

The reaction involving the breakdown of the mineral is difficult to assess but the loss of staurolite in the rocks seems to coincide with a rise in content of sillimanite (discussed quantitatively in Section II, C). This will be discussed later in Section II, F.

The occurrence of staurolite in rocks will depend on the FeO and, to a lesser degree, the MgO content as well as the necessary Al_2O_3 and SiO_2.

Samples T-79-1 and T-78-1 contain staurolite which has mutual boundaries with all the other phases and is considered a stable mineral in the assemblage. The two samples are not anomalously different in whole rock composition **to the other samples which suggests that the occurrence of the mineral is due to metamorphic and not bulk compositional conditions.**

Near the center of the traverse, staurolite does not occur within garnet, and conspicuous masses of sillimanite occur which are, in shape, similar to the inclusions of staurolite observed within garnet (Figure 2D). This implies a reaction by which staurolite partly breaks down to sillimanite.

Kyanite

This mineral forms subhedral to euhedral translucent crystals which vary greatly in colour **(Table 1).** The colour variations could not be correlated with metamorphic grade or composition.

Only near the Grenville-Southern boundary do kyanite crystals have a poikiloblastic core which is well demonstrated by the first sample on the traverse (T-79-1).

The grain size of kyanite decreases with increasing metamorphic grade (Figure 13) which is probably in response to metamorphic grade.

Kyanite, especially when found near the center of the traverse (x-y in Figure 6), is intimately intergrown and enclosed within an aggregate of muscovite crystals (Figure 15B). Kyanite may also exist as optically continuous, isolated rounded bodies within muscovite suggesting a near complete reaction of kyanite to form muscovite (Figure 15A). The view that kyanite has formed from muscovite is unlikely because of:

(a) the rounded form and indentations of the kyanite bodies suggesting alteration;

TABLE I

The colour of kyanite, garnet and biotite in the samples analyzed

	Kyanite	Garnet	Biotite
T-79-1	pale blue	dull medium red	light chocolate brown
T-78-1	pale to medium blue	dull medium red	light chocolate brown
T-120-1A	pale blue to white	dull medium red	light chocolate brown
T-290-1A	very pale blue-green to white	pinkish red	light chocolate brown
T-150-1B	sky blue to pale blue	medium dark red	medium chocolate brown
T-156-1A	pale blue-green	medium pinkish red	light chocolate brown
T-156-1B	pale blue-green	medium red	light chocolate brown
T-231-1A	pale blue	dull medium red	chocolate brown
T-232-1B	medium blue	medium red	light chocolate brown
T-232-1C	medium blue	dull red	light chocolate brown
T-249-1		pinkish red	deep chocolate brown
T-292-1		pinkish medium red	dark brown
T-293-3		pinkish ruby red	chocolate brown
T-296-1A		pinkish ruby red	greenish deep chocolate brown

(b) the significant separation, in the thin section, between different optically continuous **kyanite bodies.**

(c) the lack of muscovite aggregates such as shown in Figure 15B, at lower metamorphic grade.

The conclusion, however, leaves us with the question - where does the potash come from? The only other potash-bearing phase is biotite and it is inferred this mineral is one of the reactants. A model of potash being derived metasomatically is unlikely as the potash content of the rocks where the texture is most common is not greater than anywhere else in the section. There is no direct evidence of biotite being consumed, expect perhaps the common occurrence of sillimanite needles in biotite, which may form from the alumina liberated by an adjacent or part of the same reacting biotite grain.

If biotite is a reactant, it should not, in a classical sense, be unstable with respect to muscovite and the minerals which do not take part in the proposed reaction. However, away from the kyanite-muscovite aggregates, the muscovite and biotite form textures which at lower grades, (where the aggregates do not exist) have been inferred to indicate equilibrium. It is apparent that either biotite has differentially lost potash with perhaps an increase of soda or some of the biotite has been consumed by reaction. If the latter is true, Fe and Mg in the biotite probably have formed garnet for this is the only other major Fe-Mg bearing mineral. One could postulate that a potash phase such as K-feldspar' was previously present and has been totally consumed but no K-feldspar occurs in the pelitic schists at lower grades where the kyanite-muscovite aggregates do not exist. The chemical data pertinent to the problem will be reviewed in II, E.

Kyanite persists in the rocks beyond the point at which muscovite is lost. Textural evidence beyond this muscovite breakdown isograd on the map strongly suggests the transition kyanite→sillimanite (see Figure 14D).

Sillimanite

Both kyanite and sillimanite were identified by optical and X-ray means. Sillimanite, which is clear and uncoloured in all thin sections, has a purplish cast in the field which was later shown to be due to microscopic crystals of biotite.

Near the beginning of the traverse (near the Grenville-Southern boundary) sillimanite occurs most commonly as fibrolite masses. Beyond the breakdown of muscovite the sillimanite shows a sudden decrease in length and the fibrolite masses are replaced by scattered stubby crystals (Figure 13). One sample collected near the end of the traverse contained isolated stubby sillimanite crystals in the ground mass and fibrous crystals occurring as inclusions in garnet (Figure 15C). It is inferred that the fibrous sillimanite formed while the rocks were still of the metamorphic grade at which the fibrous variety is commonly found (below the kyanite breakdown isograd in this study).

The intergrowth of sillimanite with biotite has been interpreted by authors in other studies as due to (a) processes of biotite breakdown releasing alumina,and (b) nucleation of sillimanite onto biotite (Chinner, 1961). The source of the alumina in the latter process is believed by Chinner to be due to the breakdown of kyanite. The first process is preferred because:

(a) the orientation of the sillimanite fibers in the biotite

Figure 14

A. Photomicrograph (approximately 40x) showing optically continuous rounded masses of kyanite within a coarse-grained muscovite host.

B. Photomicrograph (approximately 40x) showing an altered kyanite crystal mantled by muscovite crystals.

C. Photomicrograph (approximately 40x) showing kyanite in contact with sillimanite. Neither mineral appears to be altered to the other.

D. Photomicrograph (approximately 40x) showing kyanite rimmed and partially altered to sillimanite. Kyanite is lost in the metamorphic sequence at higher grades than where this section was collected.

Figure 15

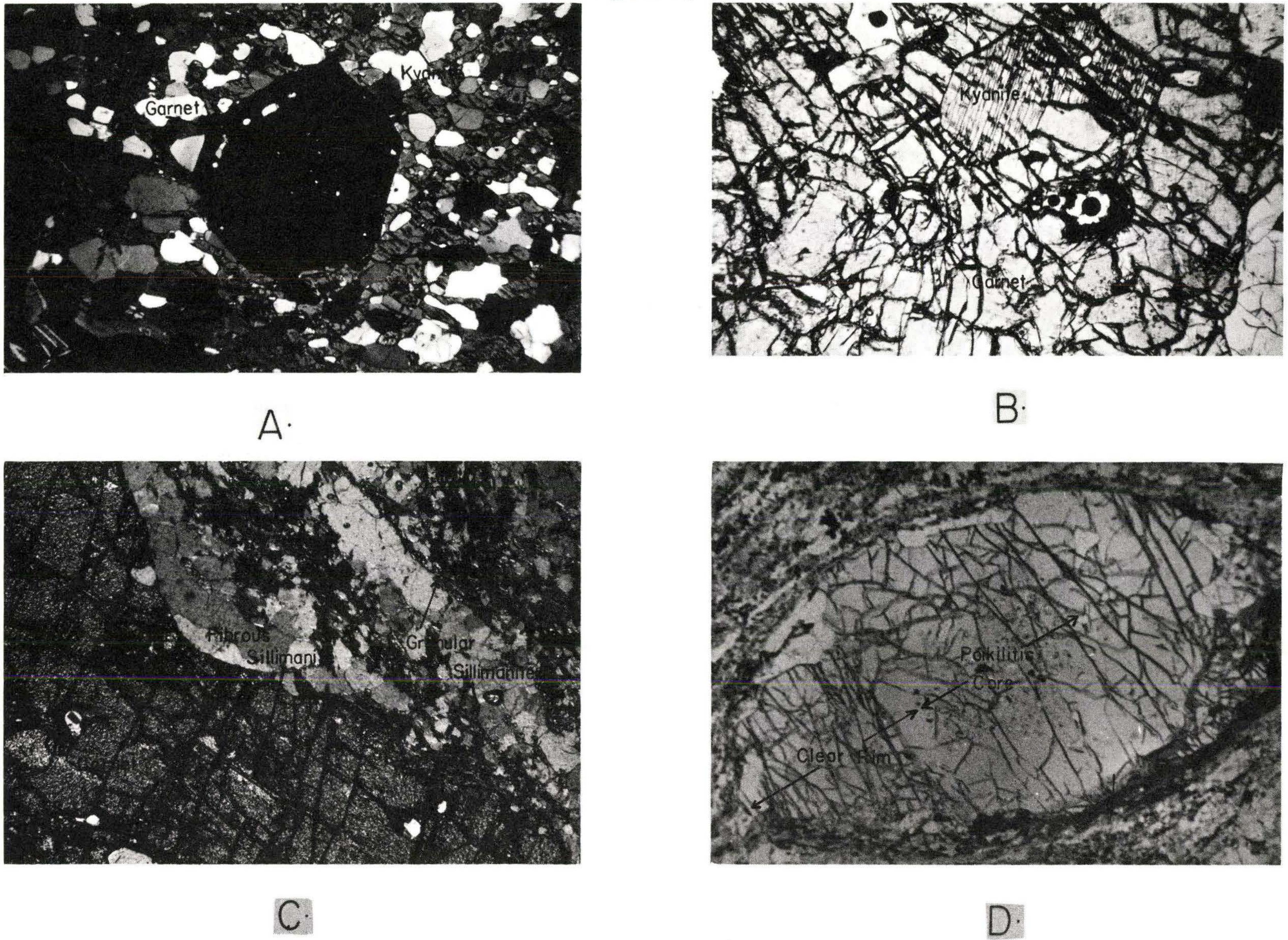
A. Photomicrograph (approximately 40x) showing a euhedral garnet included within a quartz inclusion-filled kyanite crystal. The sample was collected near the boundary between the two provinces.

B. Photomicrograph (approximately 40x) showing a euhedral kyanite crystal within garnet. The sample was collected near the center of the area.

C. Photomicrograph (approximately 40x) showing fibrous sillimanite included within garnet while stubby sillimanite occurs outside of garnet.

D. Photomicrograph (approximately 40x) showing garnet having poikilitic core and clear outer rim. The garnet is from beyond the kyanite isograd.

Figure 15


Garnet
Kyan
A·
Kyanite
Garnet
B·
Fibrous
Sillimani
Granular
Sillimanite
C·
Poikilitic
Core
Clear Rim
D·

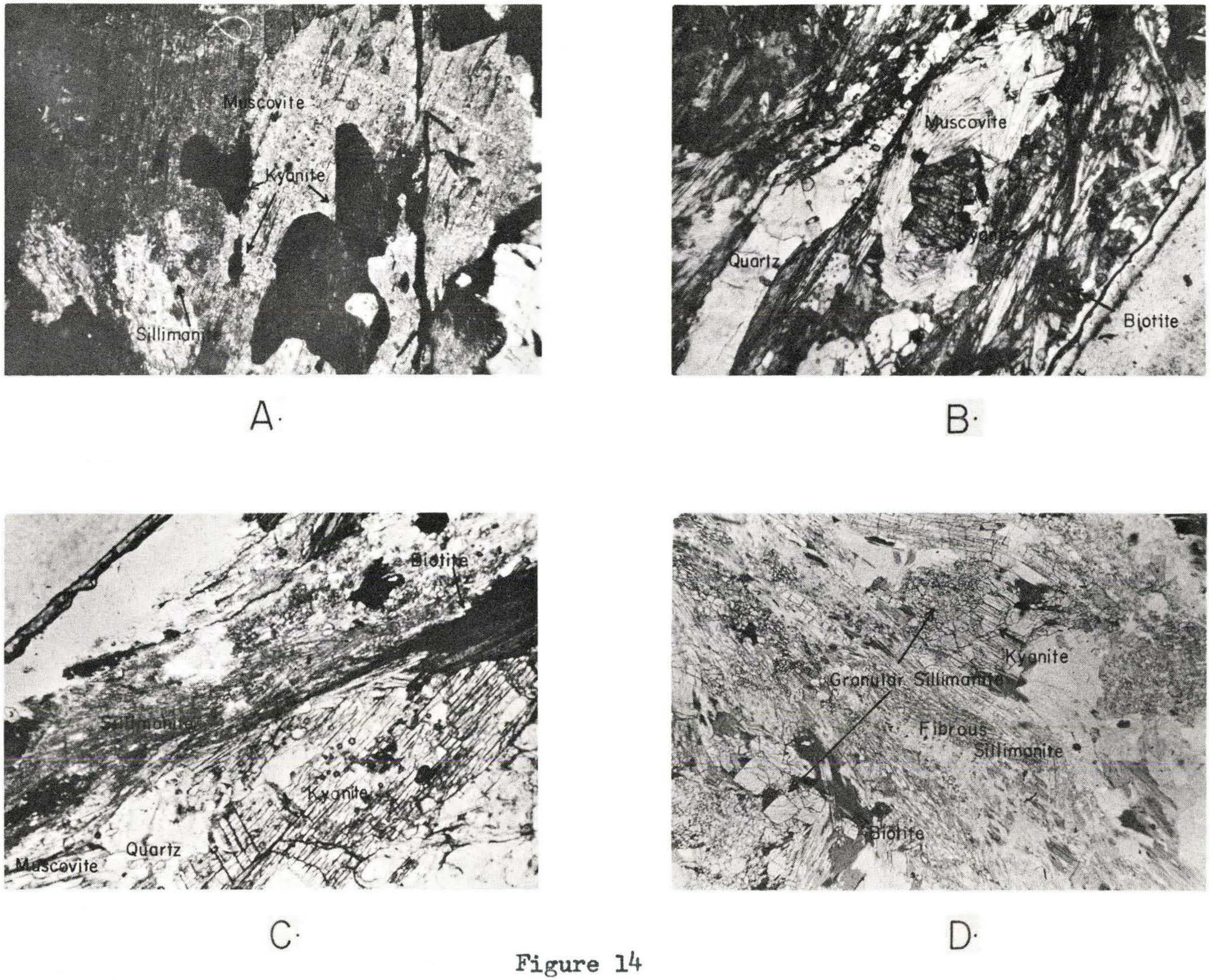

Muscovite
Kyanite
Sillimanite
A.
Muscovite
Kyanite
Quartz
Biotite
B.
Biotite
Sillimanite
Kyanite
Muscovite
Quartz
C.
Granular Sillimanite
Kyanite
Fibrous Sillimanite
Biotite
D.

Figure 14

is not related to any constant lattice direction of the biotite.

(b) the intergrowth is not observable above the muscovite breakdown isograd.

Sillimanite does not occur as inclusions within staurolite, kyanite, garnet or feldspars below the staurolite isograd suggesting its growth was initiated after the crystallization of these minerals. Although it may be crystallographically and chemically stable with respect to the rest of the assemblage, sillimanite does occur within some garnets near the end of the traverse which suggests that garnet has grown since the original **growth of sillimanite in the rocks.** This feature was only observed after the kyanite and muscovite breakdown isograds were passed. Sillimanite needles are usually deflected around euhedral kyanite crystals which suggests the former mineral grew after the latter.

Garnet

This mineral occurs in almost all the rocks observed. The first occurrence of garnet in the field is coincident with the boundary between the provinces which corresponds to Henderson's (1967) garnet isograd. It has been shown (Section I, E) that the boundary in this map area is a mylonite zone and it is inferred that movement on this zone has brought garnet bearing almandine amphibolite facies rocks in juxtaposition with lower grade rocks of the Southern province. The first occurrence of garnet is not an isograd in the classical sense in that it does not imply a reaction but that the first occurrence is due to faulting processes. These processes have essentially removed from view lower grade rocks where garnet very likely is present. This model is preferred over one of garnet being produced by a reaction at the mylonite zone because it is likely that garnet would form long before the almandine amphibolite facies was reached, and unlikely that the garnet isograd would form exactly at the mylonite zone.

In thin section the mineral forms euhedral grains in phyllosilicate rich layers and anhedral grains in quartz-feldspathic layers (Figure 2). Near the end of the traverse, garnets have large (approximately 4 mm. wide on the average) quartz crystals near their peripheries many times larger than the quartz in the surrounding groundmass.

Garnet commonly contains alternate inclusion and clear zones (see Figure 15D) demonstrating periods of growth during which it could or could not expel or assimilate impurities.

Figure 13 shows that garnet increases in grain size from the beginning to the end of the traverse. The mean distance from the edge of a garnet crystal to the edge of the first inclusion zone in each section was measured and plotted in Figure 13. The increase in width of the clear outer rim with increasing distance up the metamorphic grade shows the garnets went through the same inclusion filled stages but that the higher grade ones continued growing while the lower grade ones did not. It may be argued that this increase is due to an increase of FeO near the end of the traverse. However, it must be noted that the lowest grade sample (T-79-1) has almost an identical whole rock composition to one of the highest grade samples (T-292-1) although the average diameter of garnets in the latter sample is nevertheless much greater.

The interrelation of kyanite and garnet is of interest in that a sample near the beginning of the traverse (T-78-1) shows euhedral garnet crystals occurring as inclusions in a kyanite host (Figure 15A) while samples at the center of the traverse show euhedral kyanite crystals in a garnet host (Figure 15B). This may mean that garnet was present in the rocks prior to kyanite, kyanite nucleated and grew, enclosing garnet crystals, and finally garnet continued growing and incorporated some kyanite crystals as inclusions. At the higher grades, garnet shows a pinkish cast to its normal red colour which was not observed prior to the muscovite breakdown (Table 1). This is thought to be due to an enrichment of MgO which will be demonstrated in Section III, E.

All the minerals were found included in garnet except muscovite up to the latter's breakdown.

Biotite

Biotite occurs evenly distributed throughout the samples and in contact with all the minerals, but is most commonly intergrown with muscovite. The colours of the mineral (Table 1) are very similar in the study although a greenish tinge was observed at the highest grades.

The grain size of biotite varies greatly with metamorphic grade (Figure 13) showing a sharp general decrease beyond the muscovite breakdown isograd.

Muscovite

The mineral is not found included within garnet and staurolite but is found in contact and included within all the other minerals of the assemblage. The relations with kyanite and biotite have already been mentioned earlier in this section. The grain size of the mineral is very changeable (Figure 13) along the section and may be related to such factors as K_2O availability and P_{H_2O}.

Near the center of the area, muscovite is lost and K-feldspar appears which implies a reaction involving the two. No mutual textures were observed, however.

Plagioclase

This mineral, which occurs in every thin section, forms anhedral irregularly interlocking boundaries with the other minerals. The change of the albite twin law orientation was used to determine the compositions of the mineral (Figure 16). It is believed that the anorthite content increases slightly although it must be born in mind that the geological empirical rule that anorthite content increases with increasing grade holds true for rocks of identical composition. Samples such as T-150-1B contain more CaO than the average and consequently also contain plagioclase of a higher anorthite. The mineral shows a marked decrease in grain size at grades higher than the muscovite breakdown. Near the center of the traverse the mineral contained irregular masses of K-feldspar which were especially distinct on staining of the sections.

Quartz

This mineral occurs in all the sections as anhedral, interlocking crystals. A significant decrease in grain size was observed along the traverse above the muscovite breakdown isograd. Occasionally, quartz grains occur which contain sillimanite fibers.

K-feldspar

This mineral occurs from the middle to the high grade end of the traverse. It is in two forms which are:

(1) as small anhedral microcline crystals within the groundmass;

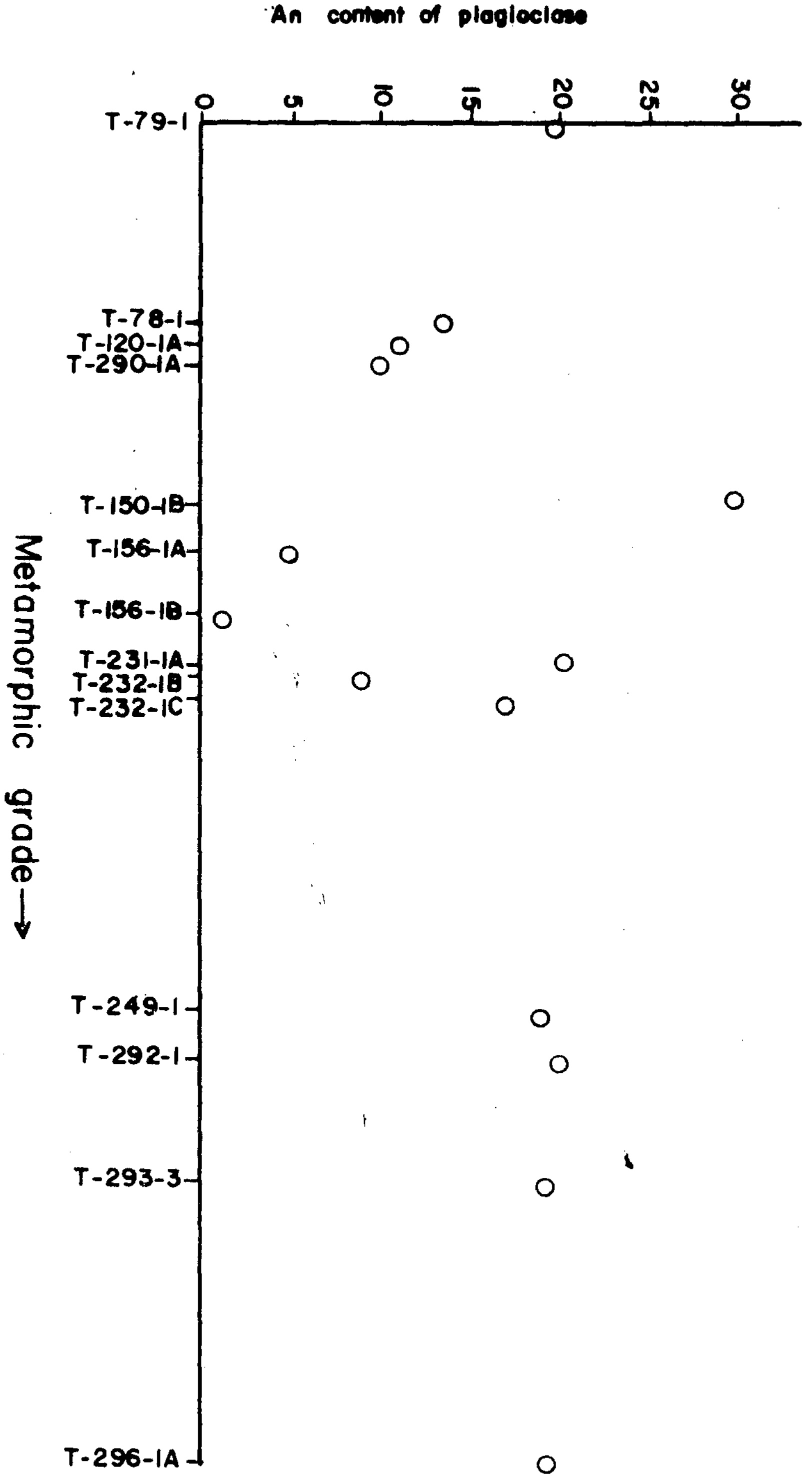


Figure 16 Plot of sample collection local along the metamorphic gradient versus the anorthite content of plagioclase.

and

(2) as spindle-shaped microcline exsolution lamellae in plagioclase.

The second type is restricted to larger plagioclase grains.

Accessories

Apatite forms clear colourless hexagonal prisms associated with biotite and occurring in small amounts in all the rocks.

Euhedral prisms of zircon occur mainly in biotite grains and characteristically are surrounded by a dark brown biotite halo.

Rutile occurs commonly as small stubby brown crystals (.1 mm. wide). Rutile has not been observed as inclusions within the phyllosilicates but is common in kyanite and garnet.

Graphite, when present, occurs intimately intergrown with the phyllosilicates. It can be seen within all the other minerals, especially kyanite.

Pyrite occurs as yellow metallic cubes in a few sections.

Pyrrhotite occurs only as partial rims around pyrite grains. These occur principally as inclusions within garnet. The rimming may suggest it formed at the expense of pyrite.

A few anhedral grains of greenish yellow chalcopyrite occur in a few samples.

Magnetite occurs as euhedral cubic crystals within all the rocks and was especially common at the higher grades. The mineral was not observed in some polished sections but upon putting a magnet through a crushed whole rock sample, a few grains were removed.

A few minute metallic grains **of an unknown minerals were observed but these were of too small a size to be identified.**

C. MINERAL MODAL ABUNDANCE VARIATIONS

i) Introduction

In the last section (II,B) it was inferred that mantling of kyanite by muscovite represents a reaction which was arrested before completion. With increasing metamorphic grade, it is probable that in an arrested reaction more of the product (e.g., muscovite in the example cited) and less of the reactant (e.g., kyanite) would be present because reaction rates would increase with increasing metamorphic grade. This section shows the systematic variations of modal abundances with varying metamorphic grade and with other data will be used to suggest certain reactions.

All rock samples containing aluminum silicates and occurring in the Grenville province were thin sectioned and the modal abundances of the constituent mineral phases were determined by use of a point counting stage, taking 1000 regularly spaced points. Modal abundance graphs versus number of points counted levelled off at approximately 300 points for the particular grain sizes of these rocks so that the 1000 points used were amply representative. (See Appendix II for modal abundances.)

The modal percentages were plotted on a map and contoured by hand. Since hand contouring is susceptible to personal bias, a mathematical type of contouring was also used and the two methods were compared (Figure 17).

Trend surface analysis involves fitting successive multidimensional surfaces represented by algebraic equations of a geographic map of the localities (coordinates) at which observed magnitudes of a variable have been obtained. The general method has been reviewed by Krumbein and Graybill (1965) and Krumbein (1959). It does not appear to have been used in metamorphic areas.

Although the method has been criticized (e.g., Chayes and Suzuki, 1963) it was found useful in this study because of the following reasons:

(1) contours of the surfaces fitted are objective;

(2) the observations are smoothed so that local inconsistencies are largely discounted.

The problem as to which successive surface to adopt is a difficult one. How well each surface fits the data is determined by finding the percentage of sums of squares accounted for by each surface. An analysis of variance (Krumbein and Graybill, 1965, p. 336-337) was used. The first surface to attain a 99% confidence level or the highest surface computable using the program of Chi (1965) was adopted as being representative. The highest surface obtainable is one represented by a cubic equation.

It was assumed that any successive surface after the one to attain a 99% confidence level would not add a great deal to the interpretation.

The algebraic formulae, coordinates and magnitudes of variables were determined as in Chi (1965). The trend surface contours, hand-drawn contours and positions of sample locals are shown in Figure 17. The modal abundances are given in Appendix II and the equations of the trends etc. in Appendix III.

ii) Analysis of Results - Modal Abundance Variations

Kyanite and sillimanite abundances are best described by linear trend surfaces as shown in Figure 17. These surfaces are inverse to each other and their strikes are, within a small acute angle, parallel. As can be seen, the hand-contoured trends are both more northerly trending for both minerals and show a more abrupt decrease of kyanite modal abundance than the trend surface ones.

The inverse relationships of the trends may indicate that kyanite inverted to sillimanite with increasing grade although the textural data mentioned in the last section would indicate the relationship is more complex.

Machine and hand trends for staurolite are similar and nearly linear. The small modal abundances render any interpretation questionable from this data but the decrease of the mineral is coincident with an increase of sillimanite. The sillimanite increase can, however, not be attributed to staurolite breakdown alone if the few samples containing staurolite as a stable part of the assemblage are representative and have such low abundances of staurolite. Muscovite and K-feldspar show an inverse relationship and their respective trends are suggestive of a reaction involving the production of K-feldspar at the expense of muscovite. The fact that the feldspar maximum is only a third of the height of the muscovite maximum can be accounted for by the fact that K-feldspar contains approximately three times as much K_2O as muscovite. The FeO and MgO from the muscovite would probably form part of the clear outer rims of garnet which are common at this position on the traverse. Na_2O would undoubtedly be accounted for in the K-feldspar. The maximum of muscovite near the center of its trend may be partly caused by the chemical inhomogeneity in the whole rock **sample** of K_2O as suggested by sample T-120-1A (see Section IIA for exact values of K_2O). The positive inflection near the center of the trend surface map near the muscovite isograd, may represent the production of muscovite by the inferred reaction with kyanite.

The maximum at the center of theK-feldspar trend is also interpreted as being caused by chemical inhomogeneity which is **reinfored by the** whole rock compositions of the higher grade samples (T-292-1 and T-249-1).

The trends of biotite and garnet should show inverse relationships since garnet abundance increases across the area and biotite is the only other abundant mafic phase. A reaction involving the two minerals cannot be clearly suggested from the trends as there appears to be a distinct increase of FeO in the highest grade samples (see Section IIA for numerical data).

Quartz and plagioclase are both very erratic in distribution which may be partly attributed to variations of the SiO_2 and Na_2O contents of the whole rock samples. The decrease of plagioclase and the increase of quartz near the right hand side of their trends may result from errors in point counting in that many of the grains were too small for certain identification.

iii) Determination of the Position of the Traverse

It was found expedient to select a traverse across the various isograds and look at representative samples along the traverse. To do this two problems were encountered:

(1) where should the section be chosen?

(2) as no line can be drawn on the map on which enough samples will plot, how do we project the locales of various samples on the map onto a proposed line?

The first problem was solved by drawing a line from the lowest grade point to the highest grade one. It was assumed that where the maximum modal abundance of the staurolite trend occurred, the minimal metamorphic grade existed. The maximum modal abundance of sillimanite on its trend (Section III, C, i) was considered the maximum metamorphic grade. The model is reinforced by the following evidence:

(1) the trend of kyanite;

and (2) internal zoning of garnet (to be described in Section II, D) which is most extreme below the staurolite isograd, consistent with a model of increasing diffusion rates and homogenization with increasing metamorphic grade;

The resulting section is at a high angle to the various isograds. The line on the map was drawn through the lowest grade sample (T-79-1) and one of the highest (T-293-3) as shown on Figure 18.

The second problem was solved in the following way. As the kyanite, staurolite and sillimanite occurrences were mainly used to define the metamorphic gradient, a translation on the map of sample positions parallel to the modal abundance contours of one of these minerals and onto the proposed line was done (Figure 18). The modal abundance trend of sillimanite was chosen as this mineral was the only one of the three which occurs over almost the entire area. The modal abundance trends are nearly parallel for kyanite, sillimanite and staurolite, so that a rotation along any one of their trends or a composite of some or all would not change the section significantly. The reason the muscovite and K-feldspar were not used was that the occurrence of these minerals is dependent upon P_{H_2O} and also, their constituent oxide bulk compositions (K_2O, Na_2O) are more varied than the Al_2O_3 and SiO_2 necessary for kyanite and sillimanite. Figure 19 shows the variation of modal abundances for the various mineral studied across the traverse.

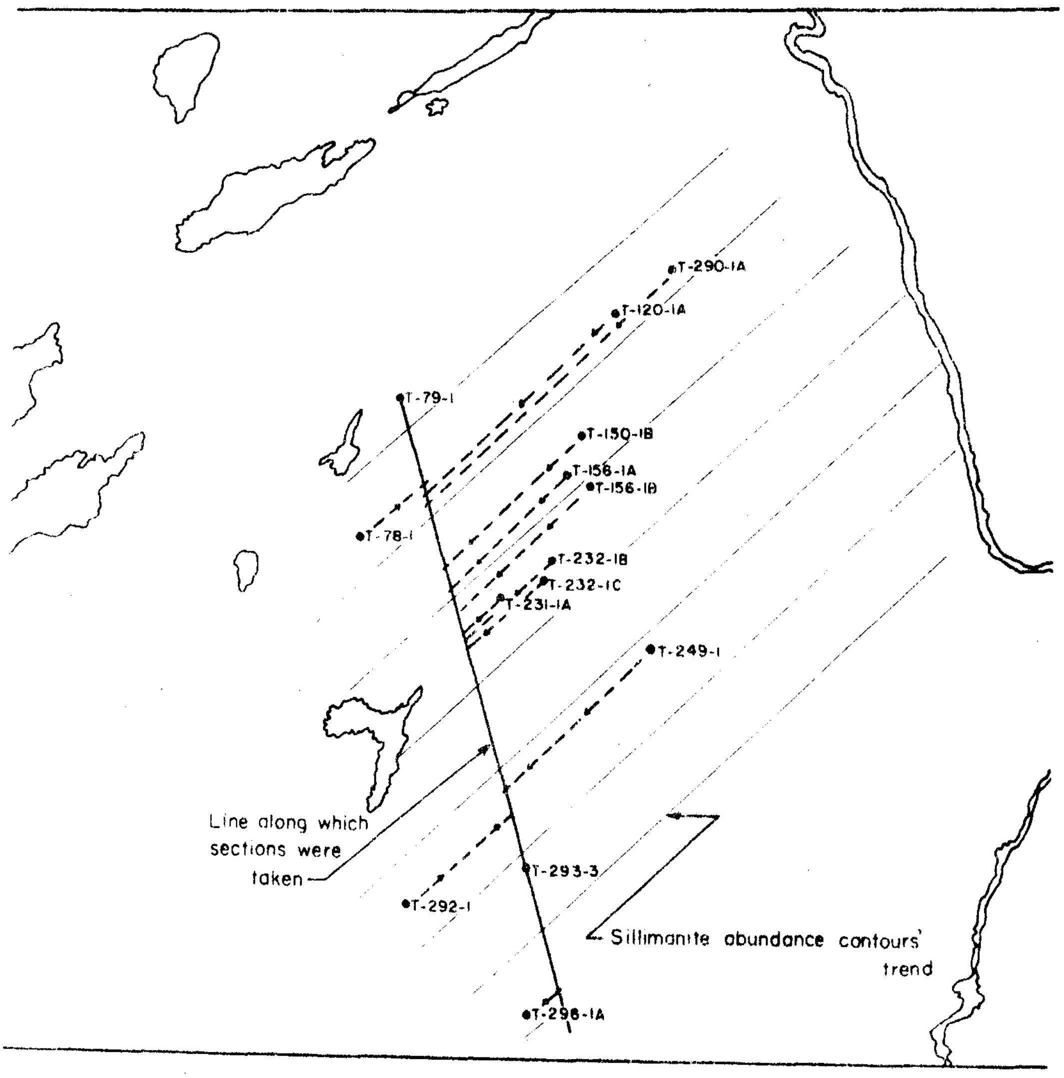


Figure 18 Map of the area studied showing the translation of samples onto the traverse. The line is shown as x-y on Figure 6 and shows the position of the isograds with respect to it.

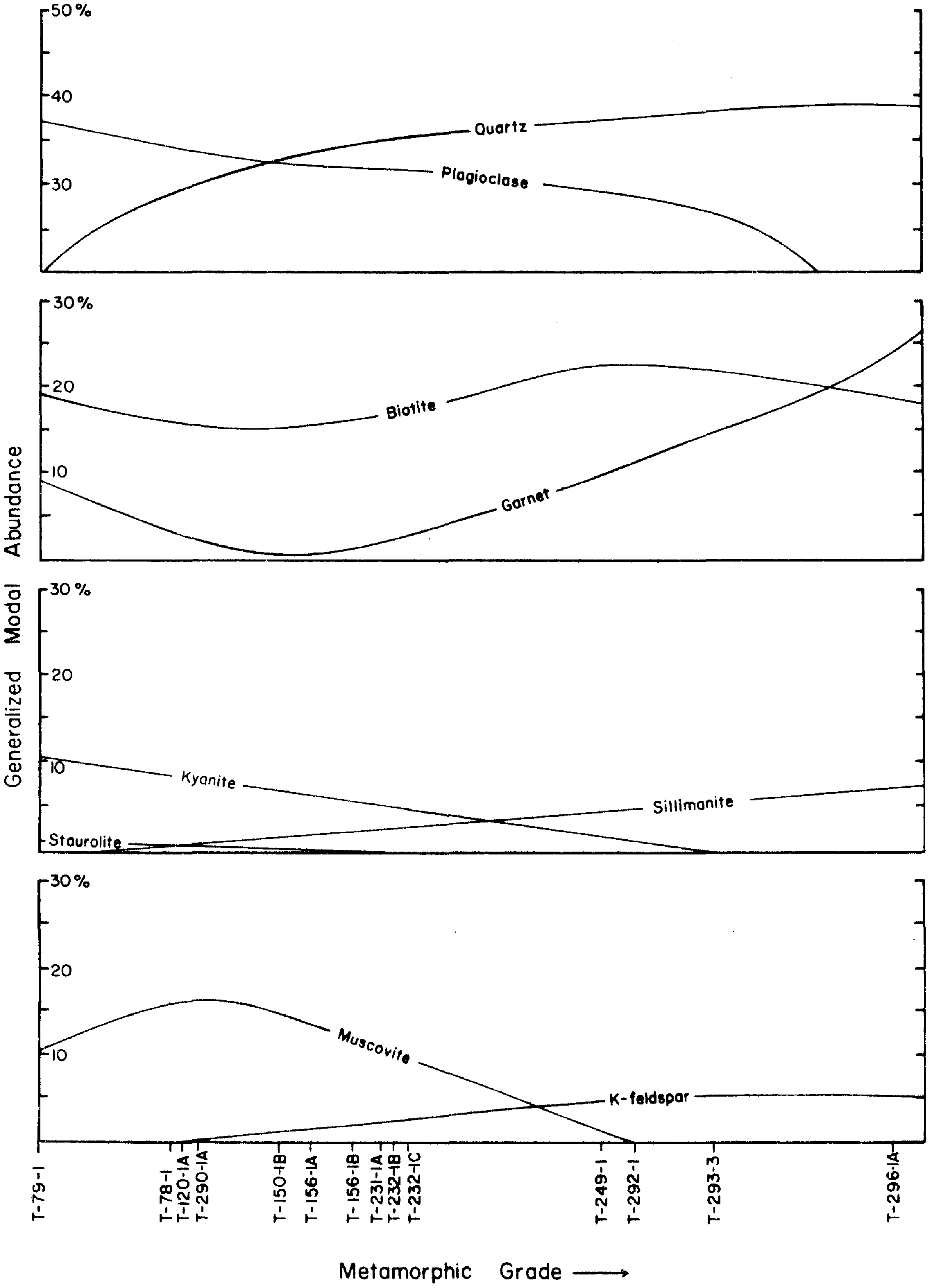


Figure 19 Sections of trend surfaces of modal abundance of minerals in pelitic schists across traverse.

One could suggest that the production of kyanite and sillimanite is dependent upon P_{H_2O} as these minerals are formed from hydrous phases but no evidence of this occurs.

D. MINERAL GRAIN HOMOGENEITY

i) Introduction

Students of element partition normally assume of a hand specimen that:

(a) all mineral grains are internally homogeneous in composition;

and (b) all mineral grains of the same mineral are of the same composition.

There is ample evidence of internal chemical zoning in ore, skarn and igneous minerals but generally this has not been observed to exist in metamorphic rocks and it is assumed that the time interval, during crystallization, was great enough for diffusional processes to go to completion. Implicit in such a view is that there has been no partial re-equilibration of elements between phases at temperatures and pressures lower than those at which the assemblage originally crystallized. This is partially substantiated by the fact that high grade mineralogy persists for long times at the low grade conditions of the earth's surface.

The grain size of metamorphic rocks is usually small (commonly <1 cm) so that analyses of parts of such grains was difficult with the instrumentation available prior to 1955. The electron microprobe analyzer (EMP) is now available and permits analysis of very small volumes. The advantages of the machine are:

(1) areas as small as 5 microns in diameter can be conveniently analyzed;

(2) the method is non-destructive;

and (3) because polished thin sections of rocks can be used, minerals involved in reaction textures can be analyzed.

In this study, a polished thin section was prepared from each of fourteen hand specimens. The minerals were analyzed by use of an Acton Electron Microprobe Analyzer (Acton Laboratories, Acton, Mass., U.S.A.).

In order to measure point to point composition changes across a mineral grain, there are two ways of recording radiation intensities which are:

(1) by counting the intensities at a particular point on the grain and manually moving the stage to a new point,

and (2) motor-driven scanning.

Method (1) is very accurage because a careful focus can be ensured at each spot; Method (2) is more affected by dissimilar qualities of polish, possible areas of surface impurities, electronic variation, and focus effects due to bevelled edges, but it superior to the first by being rapid. Figure 20 shows the results of using both methods to measure the Mn intensity in a garnet crystal. The average of the points by method (1) is very close to the average of the data for method (2). The latter method was mostly used for the following observations but was checked periodically during measurement for electronic drift etc. by referring to a standard.

Most of the results are left in relative units. The Mn, Ca, Fe and Mg contents of one garnet (Figure 23) were analyzed semiquantitatively in the following way. The garnets from each of the samples from which the polished thin sections were made, were analyzed chemically (Appendix II). Those garnet samples which contained no chemical inhomogeneities within the

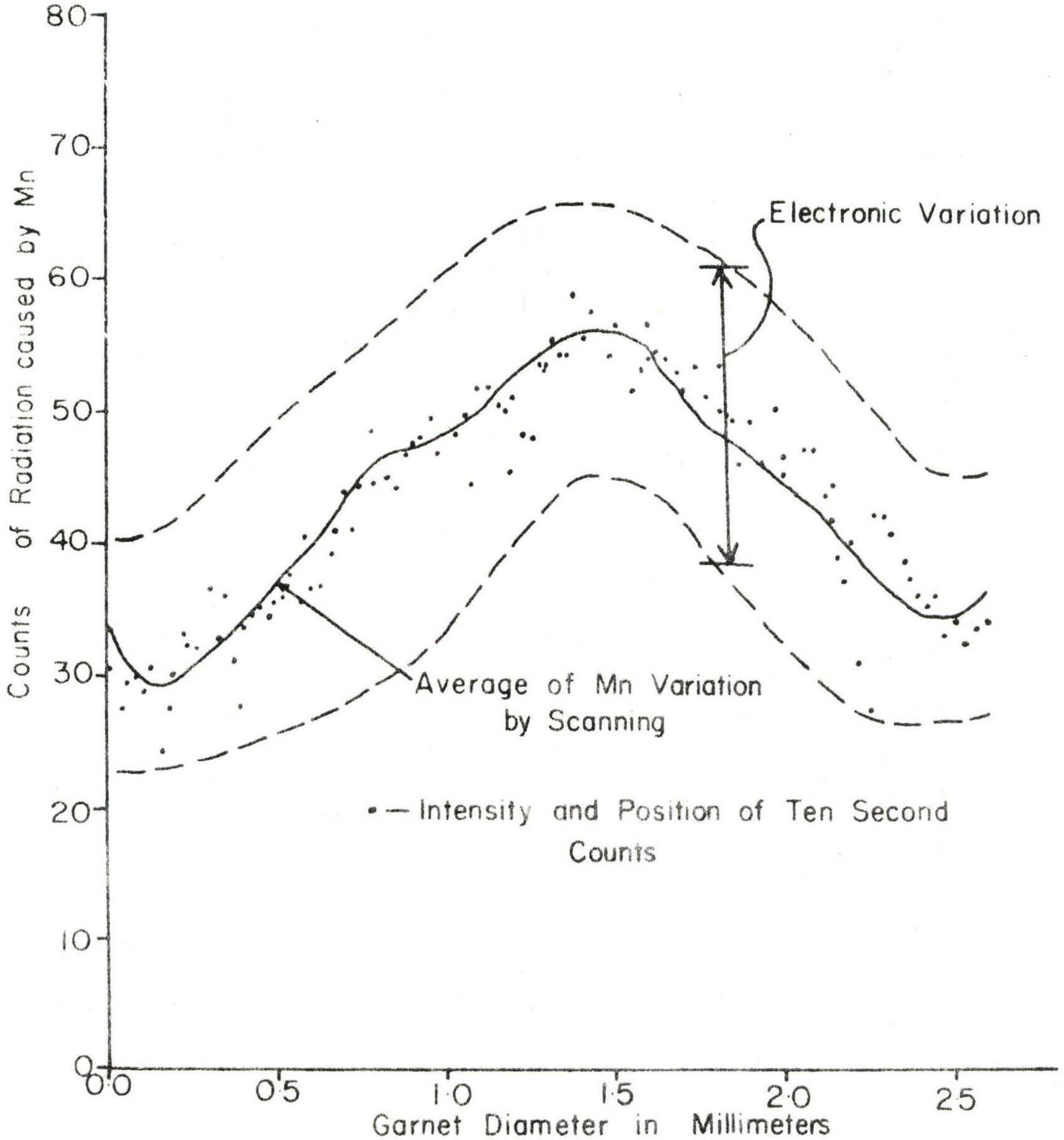


Figure 20 Graph showing counts of radiation of Mn across a grain as measured by counting intensities at specific points and scanning.

same section were used as standards (T-296-1A and T-293-3). The two samples used met these requirements as well as being of sufficiently differing compositions to allow linear working curves for Fe, Mg, Mn and Ca to be produced. The linearity of radiation versus elemental content working curve was assumed.

In each section, where possible, five grains each of staurolite, garnet, biotite and muscovite were analyzed for various elements. In a few sections, kyanite and sillimanite were also analyzed for Ti and Fe. Care was taken to measure grains as widely separated from each other as possible. The polished thin sections were only 2.5 centimeters in diameter but this appears to be an adequate representation for meaningful interpretations.

The reproductivity of the data is difficult to show as electronic drift occurs in the machine from day to day. The measurements for each polished is accurate because a standard was referred to before and after each garnet was analyzed. Figure 21 shows the typical scatter across a scanned trend. The scatter (marked "electronic variation" on Figure 21) is mainly caused by electronic fluctuations in the machine and source. The Figure (21) shows the trends of Mn and Mg which are inverse at the peripheries of the grain. This will be referred to later as Type 2 chemical zoning. The exact inverse relationship in form and intensity has been verified a number of times on the same traverse on different days.

ii) Homogeneity of Individual Grains

Biotite

The individual grains of this mineral were analyzed, and appear to be essentially homogeneous, for Fe and Mg (Figure 22B). Sample T-290-1A contains a large grain in which some chemical difference exists.

Muscovite

The mineral grains were homogeneous with respect to Fe and K. The constant Fe content suggests the element is an integral part of the muscovite structure, and is not present in inclusions. (See Figure 22A)

Staurolite

The mineral was analyzed for Fe and Mg in three types of occurrences:

(1) extremely poikilitic large crystals associated with andalusite and occurring in the Southern province (analyzed by counting fixed points and not scanning);

(2) clear staurolite crystals associated with kyanite and sillimanite and occurring in the Grenville province;

(3) "armoured" relict inclusions in garnets of the Grenville province.

In all these occurrences the grains were internally homogeneous even when crystals one centimeter wide, were analyzed. No difference was detected in composition between staurolite of the poikilitic core and the large outer rim in type (2) staurolite. These results contrast with observations by Hollister (1967) who reported "sector" zonation in staurolite from British Columbia, Canada.

Kyanite

The mineral was analyzed for Fe and Ti content in the lowest grade sample (T-79-1). If the lowest grade sample is free of chemical inhomogeneities, then presumably any higher grade ones would also be, since diffusion rates would be greater at higher grades. The kyanite grains

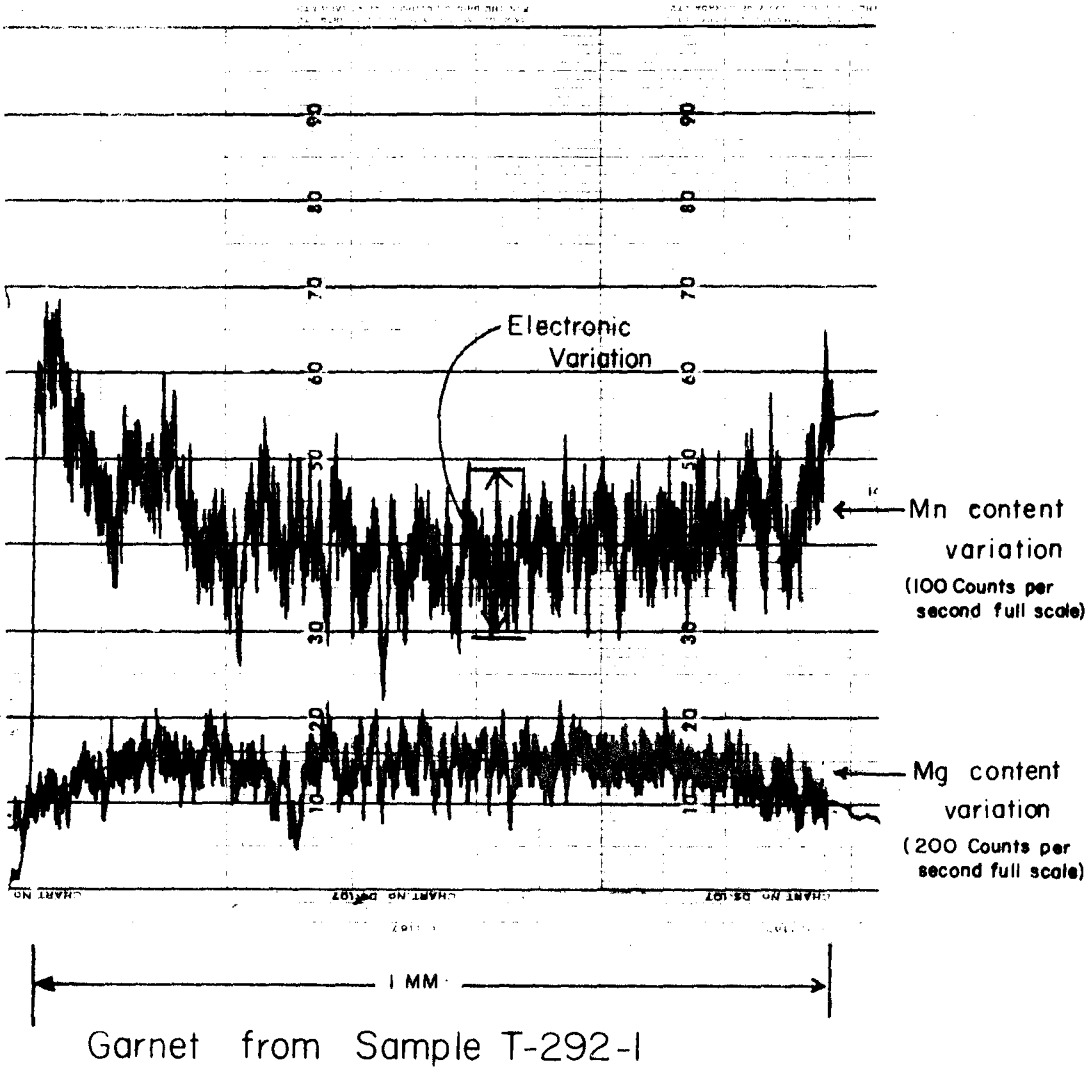


Figure 21 **Typical** scanned plot for Mn and Mg across a garnet. (Note the inverse **relation** of the elements at the edge **of the** grain.)

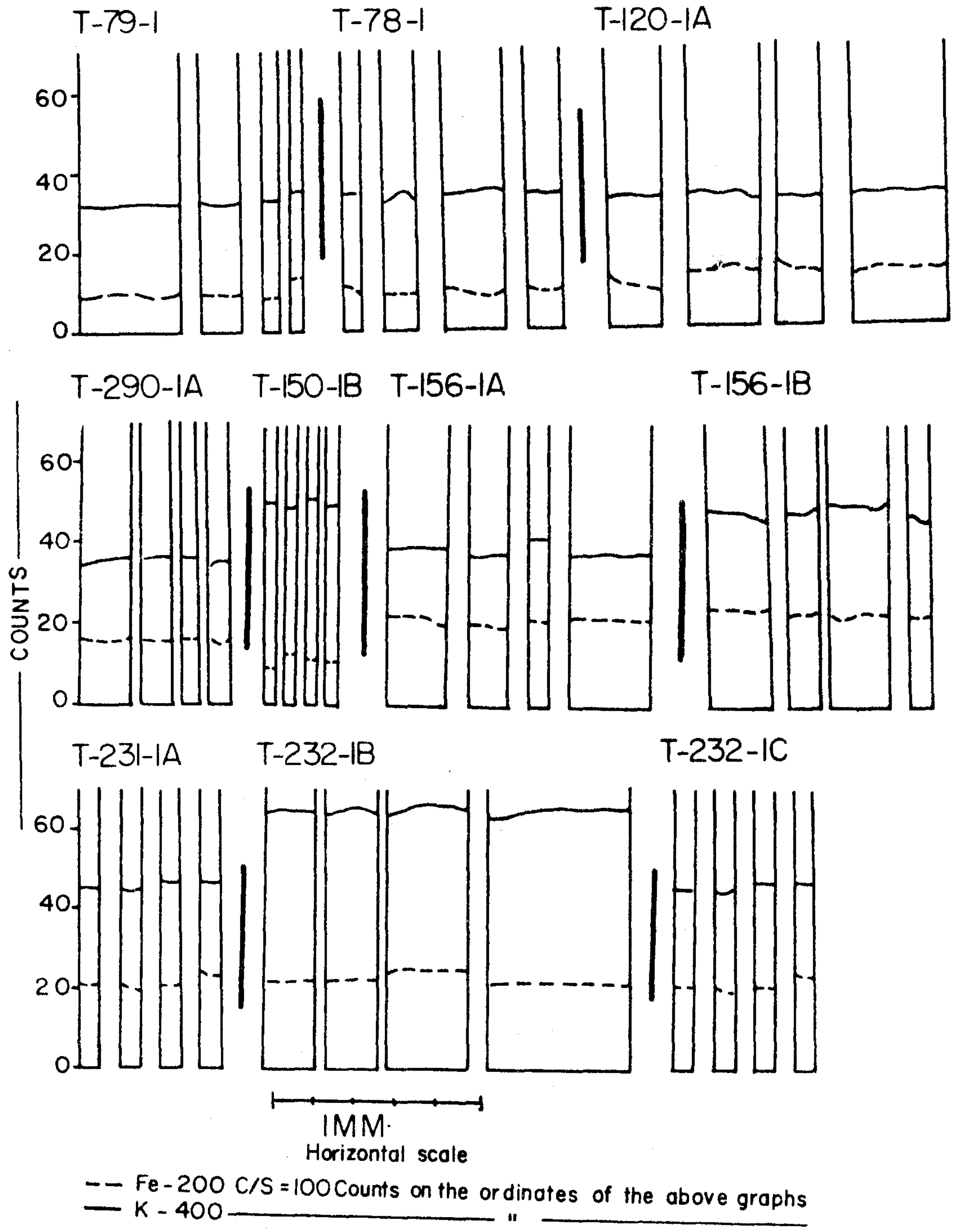


Figure 22A Variation of Fe (---) and K (——) across muscovite grains.

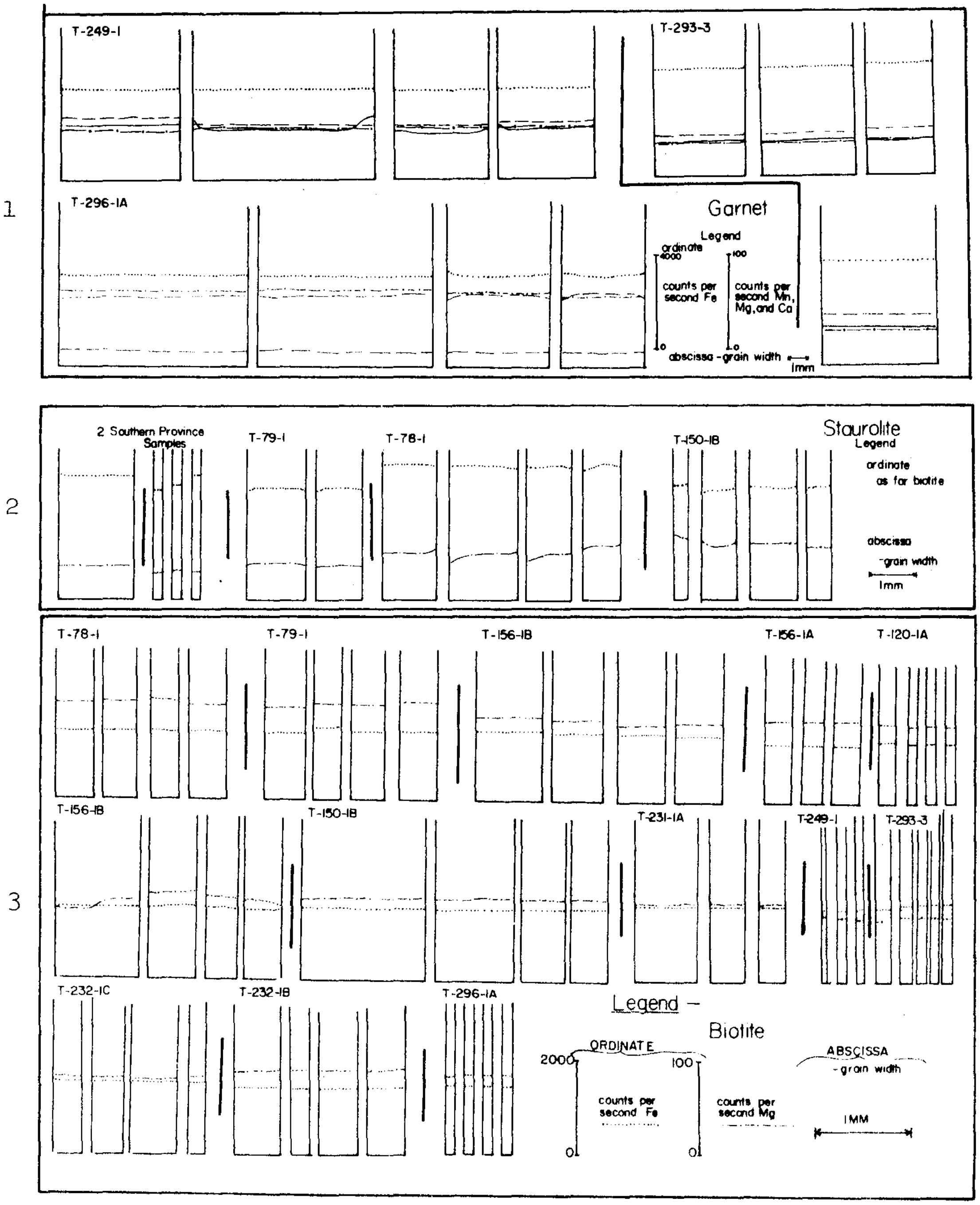


Figure 22B (1) Variation of Fe (...), Ca (——), Mn (--), and Mg (-·-·) across garnet grains. Garnets in Sample T-292-1 is shown in Figure 25.

(2) Variation of Mg (-·-·) and Fe(...) across staurolite grains. The staurolite in Sample T-150-1B occurs as inclusions in garnet.

(3) Variation of Fe and Mg in biotites.

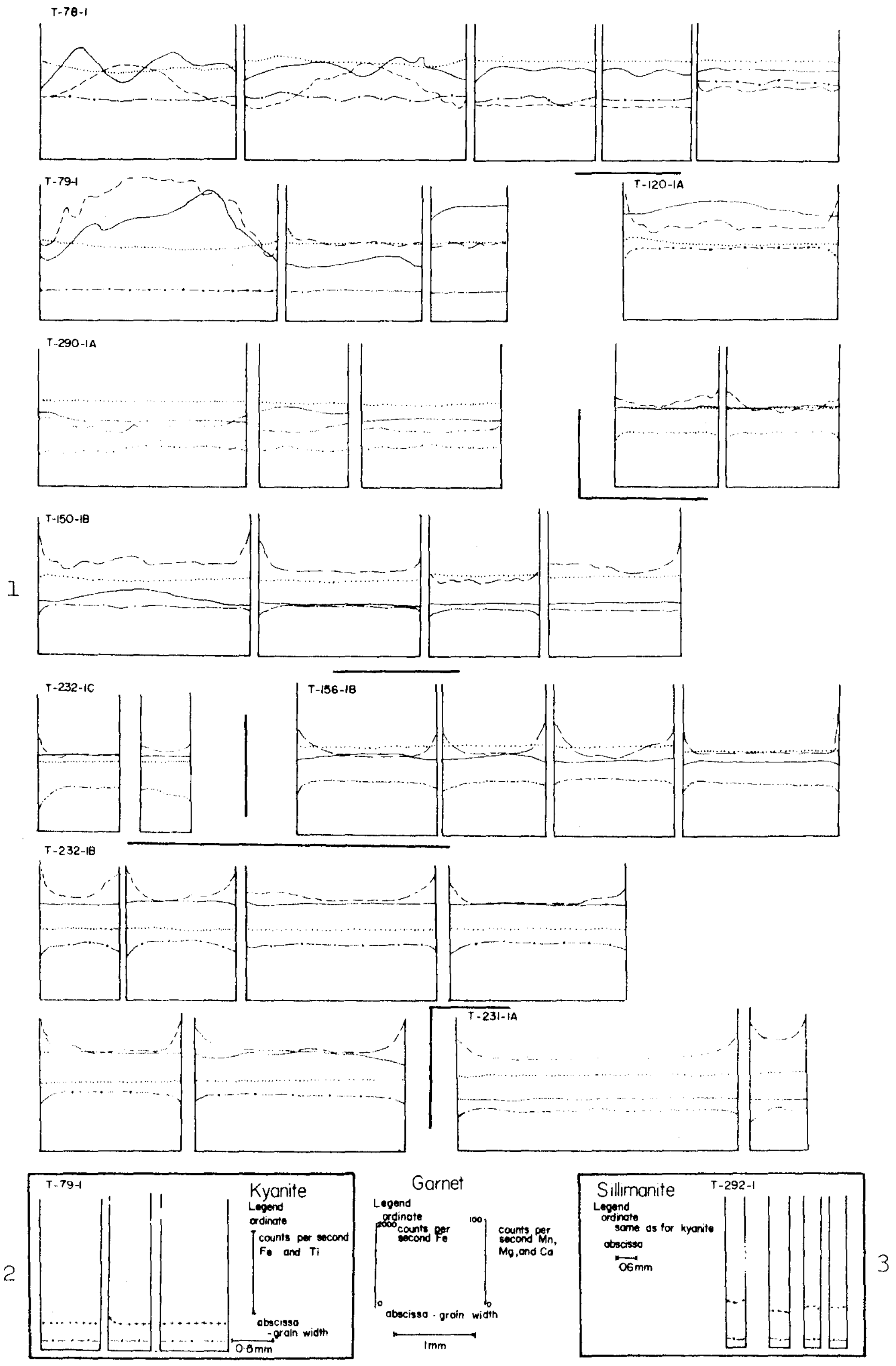


Figure 22C (1) Variation of Fe (...), Ca (——), Mg (-·-·) and Mn (---) across garnet grains collected from rocks occurring below the muscovite isograd.

(2) Variation of Ti (-x-) and Fe (+++) across kyanite grains in the lowest grade sample (T-79-1).

(3) Variation of Ti (-x-) and Fe (+++) across sillimanite grains from sample T-292-1.

contained poikilitic cores and the traverse circumvented each of the cores. The distributions of Fe and Ti are constant in each grain, which suggests the elements are an integral part of the structure of the minerals. (Figure 22C)

Sillimanite

It was not possible to analyze sillimanite from samples of lower metamorphic grade than the muscovite breakdown isograd, because the grain sizes of such samples are too small. Sillimanite crystals were analyzed for Ti and Fe in Sample T-292-1 and were found to show a minor amount of chemical inhomogeneity (see Figure 22C).

Garnet

Most of the garnet crystals in nearly all the sections are relatively inhomogeneous in their distribution of Mg, Mn and Ca and have small variations in the values of the intensities of Fe.

There is a distinct difference in the zoning of garnets occurring in samples metamorphically lower and higher than the staurolite isograd. The samples from below the isograd (T-79-1 and T-78-1) contain Mn variation which decrease steeply from the center of the crystal to the edges. Ca in the same crystals is of low concentration at the center, goes through an intermediate high zone and is low again at the edge of the crystal. The Mn, Ca and various inclusion zones within a garnet crystal in Sample T-78-1 are shown in Figure 23. Sample T-78-1 has garnets which show a crude correlation between inclusion free zones and higher values of Mn in the garnet. It is possible that the clear zones in the garnet crystals represent times in the crystal's growth during which growth was slow enough for inclusions to be expelled or assimilated. The supply of Mn to the garnet was rapid or the growth of the garnet was slow during the times the inclusion filled areas grew.

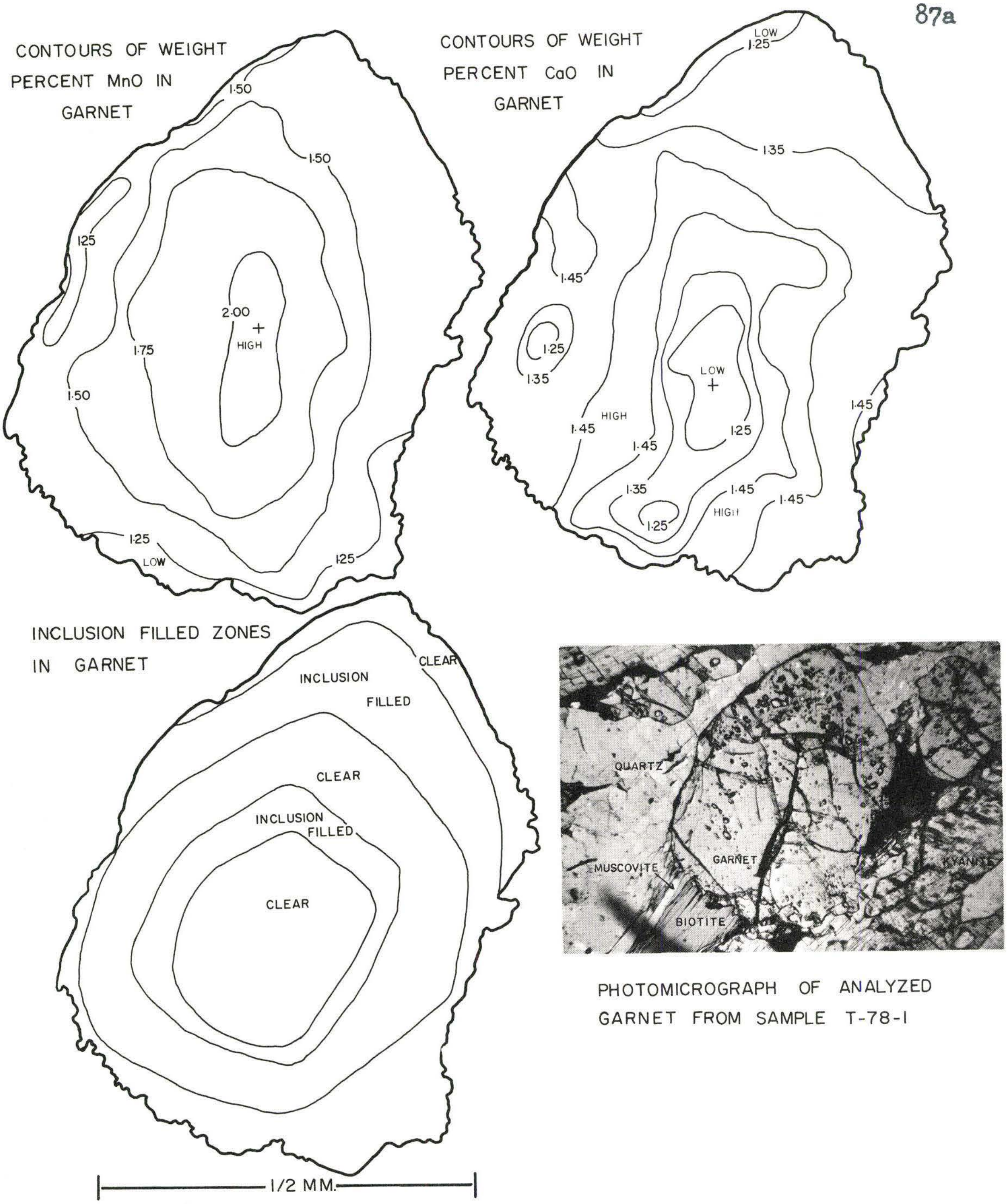
The reason for the discrepancy between the values of the chemical analysis of Mn by rapid methods (0.89 weight percent for T-78-1 in Appendix II) and the data derived by trend surface analysis is unknown. It is unlikely that matrix effects upon the analysis by the electron microprobe are significant enough to cause the differences. It may be possible that some garnets in the slide are a different composition to the one studied here but this has not been tested for.

Mn contents increase slightly near the edges of the crystals in samples below the staurolite isograd (garnets 1 and 2, T-79-1; and garnets 1 and 3, T-78-1 in Figure 22C). The orientation of the garnet slice available in a random thin section must be considered here. A cut near the center would show the maximum variation of elemental intensities while one near the edges would show very little variation.

Garnets in 12 polished thin sections from above the staurolite isograd were chemically homogeneous with respect to Mn, Ca, Mg and Fe in their centers although many showed a steep increase of Mn and a decrease of Mg at their borders. The zoning of Mn and Ca in the interiors of the garnet grains which is described in Figure 23 will be referred to as Type (1) while that at the grain boundaries described in Figure 21 will be referred to as Type (2).

The high Mn content near the center of garnet has been described by a number of authors (Crawford, 1966; Hollister, 1966; Atherton and Edmunds, 1966; and Evans and Guidotti, 1966). Atherton believes the zonation is produced by changing equilibrium conditions during the growth of the garnets. Hollister (1966) shows that the Rayleigh Fractionation model is applicable. In the garnets of the present study, the Mn distribution is a "bell" curve with small perturbations from a smooth surface. As the perturbations appear to correlate with inclusion zones, they may reflect times when there were variations of metamorphic grades. The total Mn variation may also be related to supply variations as Hollister (1966) suggests or a combination of supply and response to the prevailing metamorphic grade. The intensities of Mn on opposite edges of the same grain are often different which indicates the two edges are not in chemical equilibrium. This is attributed to differing amounts of supply of Mn between the two edges.

The distribution of calcium in Figure 23 may be due to the sudden increase in supply of calcium near that part of the garnet's growth and metamorphic history. The sudden change in supply may be related to the breakdown of a mineral such as epidote. An inverse relationship at the peripheries of garnet grains, called Type (2) zoning in this study, has been described by Crawford (1966) who has found that an increase of Mn is matched by a decrease of Ca; plagioclase immediately in contact with the garnet shows an enrichment of calcium. Crawford believes the relationship is not due to a retrograde metamorphic effect but "more complex reactions between the various metamorphic phases" (p. 283). Atherton and Edmunds (1966) explain similar "irregular" zoning as being due to "regional followed by thermal metamorphic conditions" (p. 192). Hollister (1966) explains the steep increase of Mn at the garnet peripheries as being caused by "effects on the fractionation factor as the rate of garnet growth approaches zero". Evans and Guidotti (1966, p. 47) ascribe the reason for the zoning of garnet in rocks from western Maine, to "a reaction between muscovite and garnet margins to produce biotite, quartz and sillimanite". They also believe



PHOTOMICROGRAPH OF ANALYZED GARNET FROM SAMPLE T-78-1

Figure 23 The variation of MnO and CaO in a garnet crystal from Sample T-78-1. The contours are on the basis of 11 traverses across the garnet. The center of the inclusion-filled zones matches the center of the MnO and CaO zonation.

that the growth of biotite would consume more pyrope than almandine and considerably less spessartite, resulting in marginal changes identical to those observed. The absence of crystal faces on garnet was cited as evidence supporting the last reaction quoted above and this latter of the two explanations was preferred.

Figure 24 shows the variation of the volumetric ratio biotite/garnet and the average height of the Mn at the edges of the garnet crystals. This feature strongly supports a model of element exchange between the two mineral phases because when the ratio is near unity, there is little or no zoning of type (2) while, when the ratio is large, a great amount of the zoning occurs.

In this study, the inverse relationship of Mn and Mg (and/or Ca) is believed related to a re-equilibration between at least biotite and garnet after the main period of metamorphism, because:

(1) there is a good correlation between the volume percentage of biotite present in the rock and the average peak height of Mn and Mg of Type 2 zoning (see Figure 24);

(2) biotite compositions change as garnets are approached in some sections (described in the next section);

(3) there is a steep (near exponential) increase of Mn and decrease of Mg, suggestive of diffusional mechanisms;

(4) the increase of Mn in garnet is commonly indicative of low metamorphic grades while an increase of Mg is indicative of a high one.

The exchange of Mg and Mn between garnet and some source outside the hand specimen was considered improbable because of the constant MgO values in

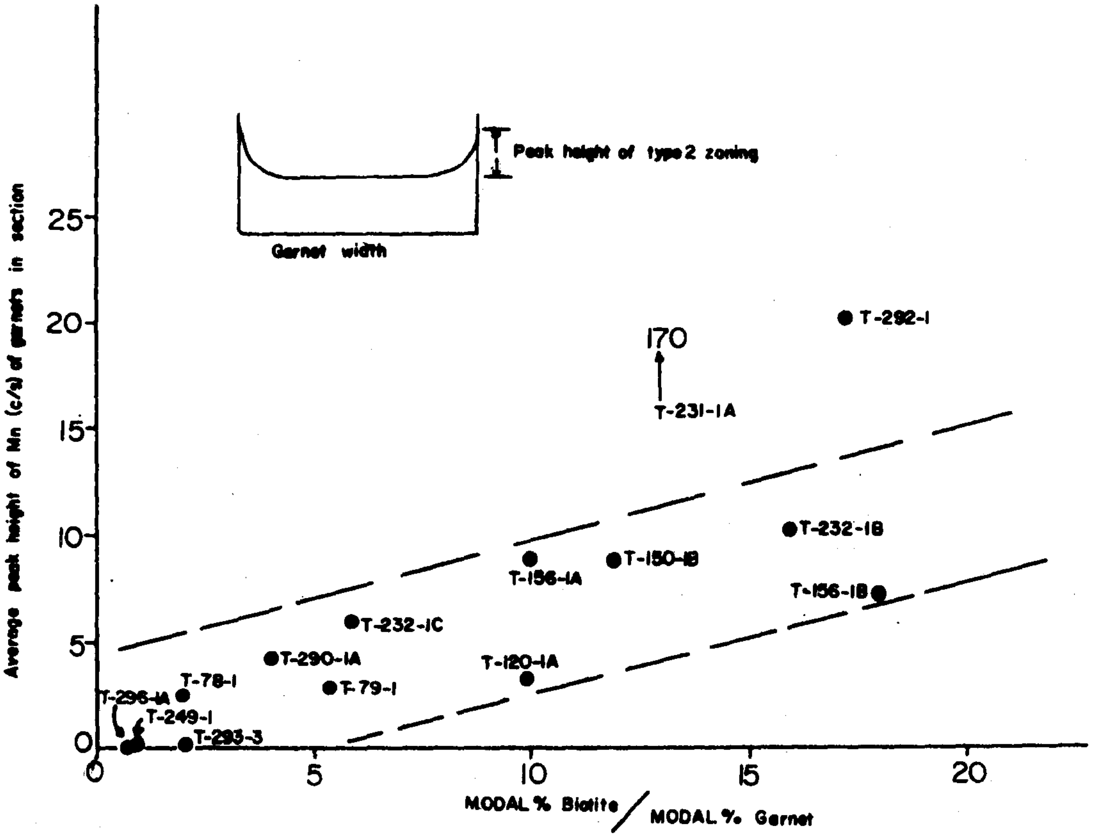


Figure 24 Graph showing the ratio of the modal abundances of biotite/garnet versus the average peak height in counts per second of Mn in Type 2 zoning. The dashed line bounds the inferred trend. The anomalous behaviour of T-231-1A and T-292-1 is believed due to non random distribution of the two minerals in the hand specimen.

the rocks (Section II, A, ii) and the evidence stated in (2) above.

Summary

Grains of biotite, muscovite, kyanite and sillimanite were found to be relatively free of inhomogeneities while garnet was not. Two types of zoning are observed in garnet:

(a) zoning in which the center of the crystals are inhomogeneous with respect to Ca and Mn;

and (b) zoning in which the crystal edges show variations mainly of Mn and Mg.

The first type is believed caused by a mixture of variation of Mn supply during the growth of the garnet and the metamorphic grade present during the time when each successive layer was being added to the garnet.

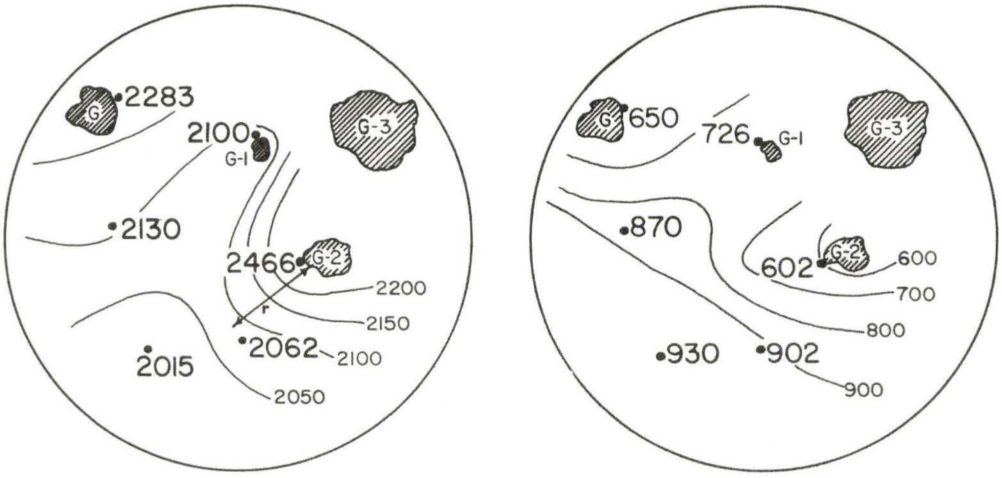
The second type of zoning is believed to be caused by re-equilibration of Mn and Mg between biotite and garnet at grades different to those when the minerals crystallized (crystallization was presumably at the intensive parameters indicated by the inferences from the presence of kyanite and sillimanite). This could have been brought about by an increase in P_{O_2} which would cause the biotite to become more Mg rich (Wones and Eugster, 1965).

iii) Grain to Grain Composition Differences of the Same Mineral within a Thin Section

The results for the measurements in this section of the study are shown in Figures 22A, B and C.

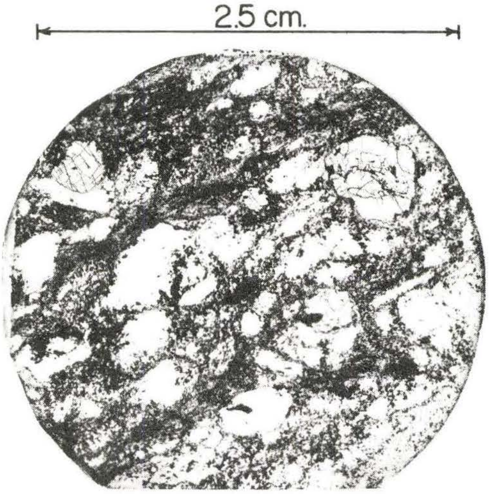
Biotite

The use of the electron microprobe in geology is fairly recent and only in the work of Evans and Guidotti (1966) have chemical differences



Mg content of Biotites

Mn content of Biotites



Photomicrograph of polished thin section

LEGEND

▨ -garnet crystal

2100• -counts per hundred seconds at this point

—2100 -contour of counts per second of element

←r→ -radius of equilibrium system

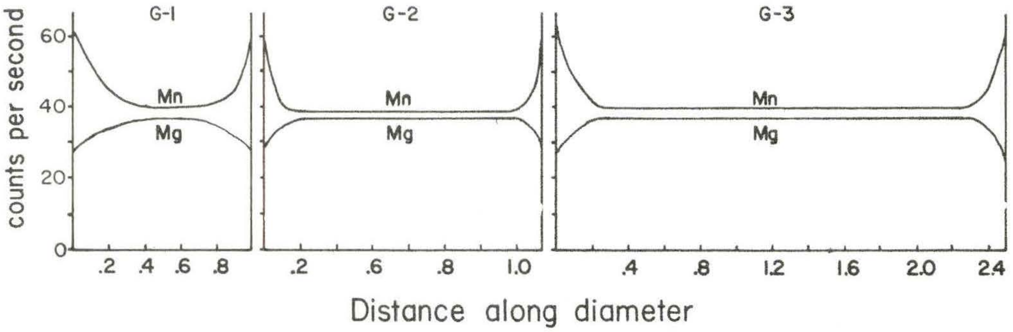


Figure 25 Figure showing the variation of Mn and Mg content of biotite in relation to proximity to garnet. The bottom of the figure shows the inverse relationship of Mg and Mn zoning in garnet.

(in MgO, MnO, K_2O and TiO_2) between biotites occurring within and outside a garnet crystal, been recorded.

In this study, different biotite grains in the same section display some chemical variations for Mg and Mn. Sample T-292-1 showed the greatest variation of composition of different grains of biotite in the same section and a detailed study of this phenomenon is shown in Figure 25. As can be seen, the Mg content of biotite increases and the Mn content decreases, as a garnet crystal is approached, over distances of 0.5 centimeters. The trends of Mg and Mn compositions of the biotites are inverse to each other which further implies the exchange reaction between garnet and biotite mentioned in the last section (II, D, ii).

Muscovite

Different muscovite grains in the same section show few differences in the content of K and Fe.

Staurolite

Staurolite contains few chemical inhomogeneities of Fe and Mg both with respect to different grains within one section and ones of different sections.

Kyanite

In the one section analyzed (T-79-1), different kyanite crystals contained the same amount of Ti and Fe per unit volume. Sample T-79-1 is the lowest grade sample and as element diffusion is expected to increase with increasing grade, it is inferred the other samples from higher grades were also homogeneous for different crystals in sections.

Sillimanite

This mineral showed grains of slightly differing compositions of

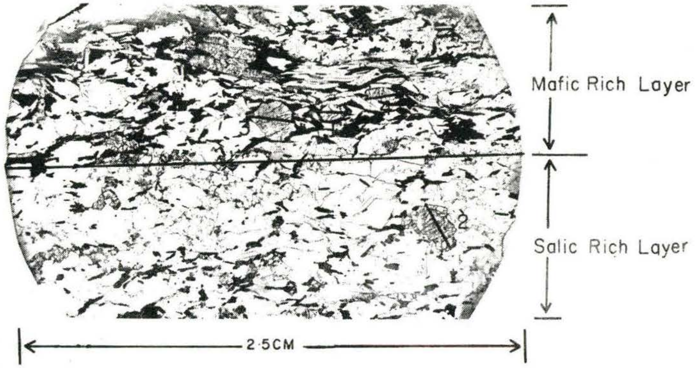
Ti and Fe in the one section analyzed (in Sample T-292-1).

Garnet

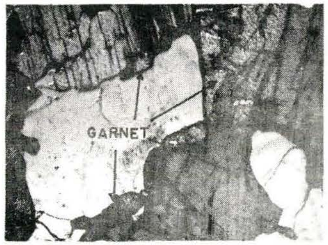
In general, above the staurolite isograd, centers of garnets are of the same abundance of Ca, Fe, Mn and Mg, even when different crystals up to 2.5 centimeters apart are analyzed. This homogeneity is of interest when Sample T-156-1A is considered for it displays two different forms of garnet occurring in two compositionally different layers (Figure 26). The garnet form occurring in the mafic rich layer is euhedral and displays zones of inclusions while the garnet form in the felsic layer is extremely anhedral, since it has grown between crystal **boundaries** and does not contain small inclusions. The contents of Mg, Mn, Ca and Fe were the same in both types. This shows that during the main period of metamorphism and garnet growth, diffusion was extensive enough so that bulk chemical dissimiliarities between different layers in a 2.5 cm. diameter area were not important enough to lead to different garnet compositions.

Summary

Differences in abundances of Fe, Mg, Mn and Ca in biotite and, to a lesser degree garnet grains, occur in the same section. Different kyanite, sillimanite and muscovite grains are, to a first approximation, of identical composition with respect to Ti and Fe. The inhomogeneity of the biotite is believed due to re-equilibration of elements between biotite and garnet at lower grade conditions to when the minerals originally crystallized. Many of the compositions of the biotites are constant within the same section which is possibly due to their being measured too close to garnet boundaries or that overlapping equilibrium "cells" occur. The garnets are nearly always



BELOW— Photomicrograph showing garnet occurring along grain boundaries in salic rich layer



APPOX·30X

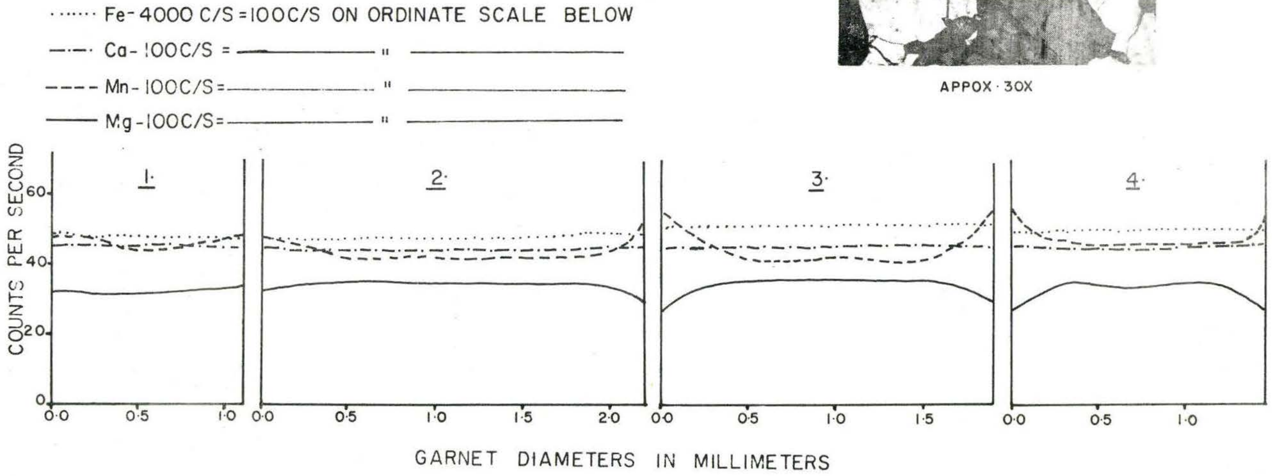


Figure 26 Diagram showing the uniformity of composition of Fe (-•-•), Mg (___), Mn (---) and Ca(•••) at garnet centers. Garnet 2 is from a salic rich layer while garnets 1, 3 and 4 are from a mafic rich one.

inhomogeneous except for those in Samples T-296-1A and T-293-3.

It is possible that some bias due to differing compositions of the same phase in the same hand specimen may have been introduced during the separation necessary for the wet chemical analyses. The minerals muscovite, biotite, garnet and sillimanite were separated on the basis of their magnetic and density properties and these would be affected by chemical variations as described in this section. If crystal edges and small grains were lost during the crushing of the samples due to their intergrowth with other phases, compositional bias will develop in the case of garnet because the edges of grains of this mineral are very different in composition from the central part.

iv) Discussion of Biotite - Garnet Relations

In the preceding section it has been shown that the composition of biotite grains changes in the thin section as garnet grains are approached. This is thought to be caused by the exchange of Mg and Mn between the two minerals after the period of metamorphism during which they originally crystallized. The degree of exchange will have an effect on the subsequent distribution coefficients of Mg and Mn between the two minerals, and the samples which should show the most anomalous values will be indicated.

The degree of exchange which has occurred in a hand specimen between the two minerals will depend upon:

1) the size of the garnets;
2) the number of garnets;
3) the relative positions of individual garnet and biotite grains in the hand specimen;

4) the ease of diffusion of Mg and Mn through the rock matrix. It is assumed that diffusion times are similar and that all compositional extremes from pure Mn to pure Mg end members in both phases are possible at the particular intensive parameters during which the exchange occurred.

The size of the garnets is important because from geometric considerations it can be shown that the area for exchange of ions will increase almost exponentially with a linear increase of the diameters of the garnets. The average diameters of the garnets in the fourteen polished thin sections used are shown in Figure 13. As can be seen, Samples T-249-1, T-293-3 and particularly, T-296-1A contain the largest garnets while T-232-1C and T-231-1A contain the smallest.

The number of garnets will affect the degree to which biotite has exchanged ions because it determines the number of centers the exchanging ions can proceed to. Figure 27 shows the number of garnets present in each of the fourteen polished thin sections 2.5 cm. in diameter. Samples T-78-1, T-293-3 and T-150-1B contain many garnets while T-232-1C, T-231-1A and T-79-1 contain few.

The position of the various mineral grains in the hand specimen will partly determine the degree of exchange of ions because this will determine the length of the mean free path of diffusion of ions. The fourteen polished thin section are shown in Figure 28. Most of the samples show a fairly even distribution of garnets and biotites except T-156-1A, where the biotite appears to be concentrated mainly in a mafic layer; T-292-1, where few garnets exist in half the section; and T-232-1C and T-231-1A, where the few garnets that do exist only do so in a small part of the section.

The ease of diffusion of ions between the two reacting phases is

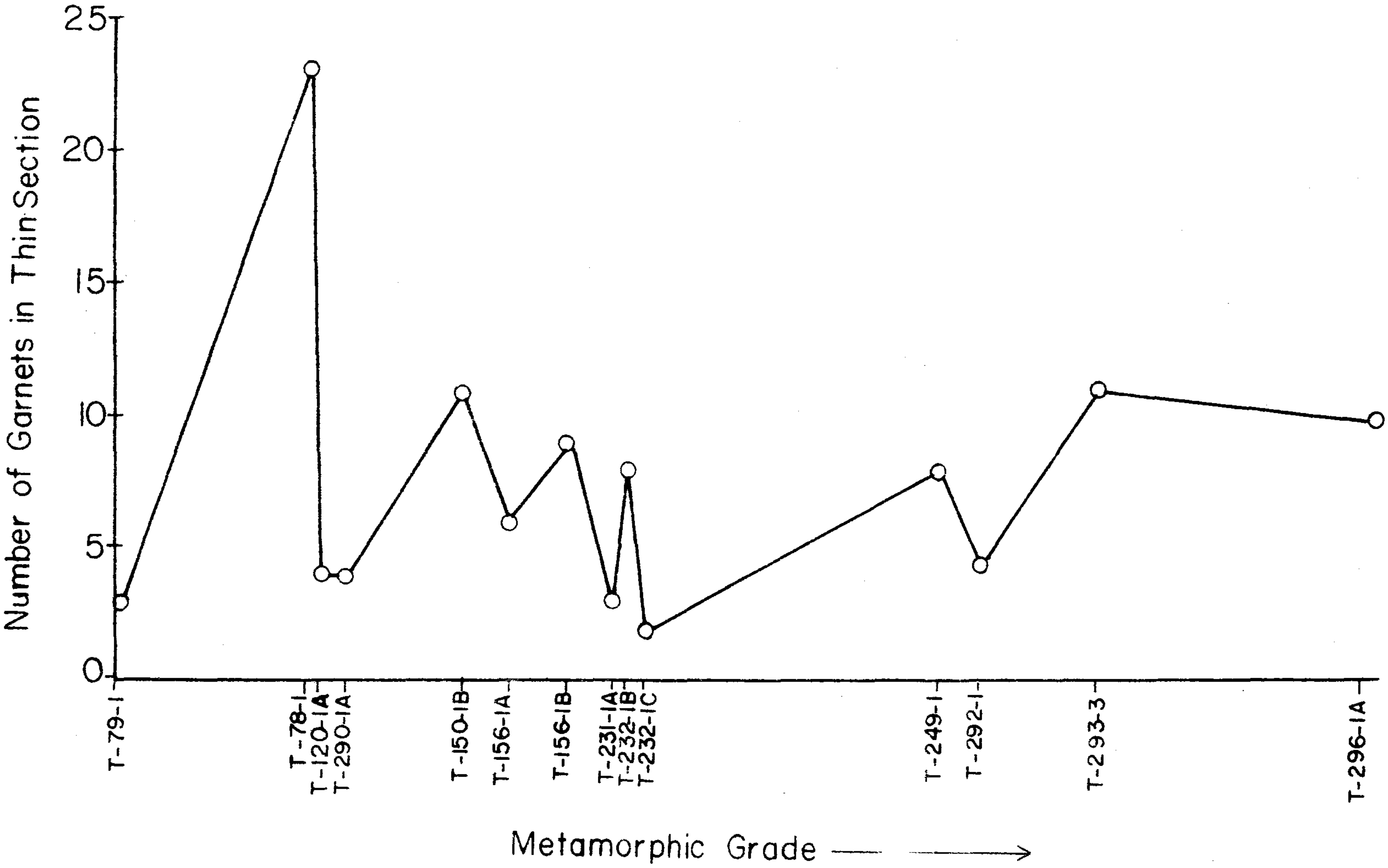


Figure 27. Variation of the number of garnets in the sections versus Metamorphic Grade in a 2.5 cm. diameter circular section.

difficult to assess as the precise mode of diffusions in unknown. The following possibilities exist:

(1) diffusion along intergrain channel ways

and (2) diffusion by lattice site exchange.

No proof exists for either of these. In both of these modes, grain sizes will be a major factor and the second mechanism will be influenced by the stability of ions occurring at various lattice sites.

It appears from relations of the chemical inhomogeneities mentioned in the preceding two sections (II, D, ii and iii) that Mg and Mn can diffuse with greater ease through approximately 0.5 centimeters of rock than through approximately 100 microns of garnet at the prevailing metamorphic grade(s).

Summary

The relative importance of each of the four factors mentioned at the beginning of this section (II, D, ii) is difficult to assess but certain samples are anomalous with respect to more than one of them. Garnets and biotites in Samples T-231-1A and T-232-1C should show the least amount of Mn and Mg exchange because these samples are low in amount of garnet, number of garnets and the existing garnets are close together. Samples T-296-1A and T-293-3 should show the greatest amount of ionic exchange to new metamorphic conditions because these samples are rich in garnets, the garnets are large and well-distributed. The distribution diagrams to be explained in Section III, E, iv,confirm the conclusions of this section fairly well.

Figure 28. Photographs (2x) showing the 14 samples used and distribution of garnet and biotite.

1. T-79-1
2. T-78-1
3. T-120-1A
4. T-290-1A
5. T-150-1B
6. T-156-1B
7. T-156-1A
8. T-231-1A
9. T-232-1B
10. T-232-1C
11. T-249-1
12. T-292-1
13. T-293-3
14. T-296-1A


1


2


3


4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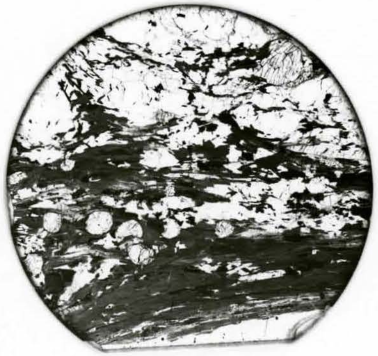
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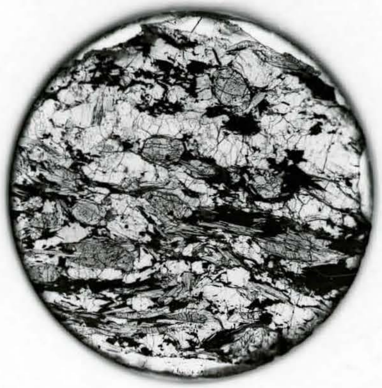

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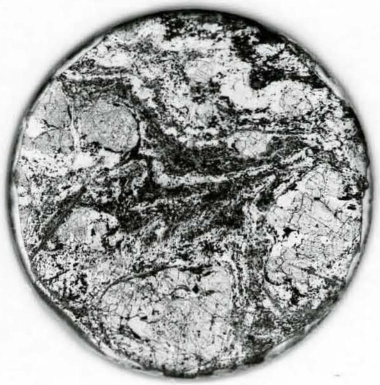

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E. GEOCHEMISTRY

i) Introduction and Theoretical Considerations

Atherton (1965) has cited isograds involving three distinct groups of chemical reactions which are: (a) polymorphic transformations, (b) single mineral variations, and (c) mineral pair variations. These are subdivided into such divisions as "single mineral - variable composition, varies systematically with grade, e.g. garnet". Such subdivisions may be useful but they do not involve all the relations present, specifically the interaction with other phases by which garnet lose manganese and possibly gains magnesium. Atherton's main divisions are thus accepted but his subdivisions will not be used.

The polymorphic transformation of kyanite to form sillimanite has been well documented in this study (section II,B). The single mineral isograds, namely those of the breakdown of staurolite and muscovite and the first appearance of K-feldspar are also well documented. This part of the study deals mainly with the isograds of type (c) which according to Atherton (1965, p. 183) are not dependent upon "variation of host-rock composition and --- volumetric amounts of these minerals". It has been suggested by Albee (1965b) and Kretz (1960) that the type of assemblage has an effect.

Theoretical Considerations

The thermodynamic laws are often used in presenting arguments about the concept of equilibrium in chemical systems. The state of equilibrium is attained when the system "shows no further tendency to change its properties with time" (Moore, 1964, p. 6).

Equilibrium may refer to the structural state or the chemistry of mineral phases. Presumably structural equilibrium is more difficult to attain as significant energy barriers must be overcome for the complete re-arrangement of the silica tetrahedra and alumina octahedra in geological systems. The only way to test structural equilibrium is to synthesize phases in the laboratory and determine their stability fields. This was beyond the scope of this study and only the chemical equilibrium between phases was determined.

Chemical equilibrium is difficult to attain in metamorphic rocks because diffusion is very slow in the ionically bound silicates. From the definition of equilibrium and considerations of the second law of thermodynamics, one can intuitively see that a certain measurable quantity must be at a constant value in the system at equilibrium. Gibbs (1906) stated that at equilibrium we can write a series of equations of the type

$$\left.\begin{array}{l} t' = t'' = t''' = \text{etc.} \\ p' = p'' = p''' = \text{etc.} \\ \mu_1' = \mu_1'' = \mu_1''' = \text{etc.} \\ \text{-------------------------} \\ \mu_n' = \mu_n'' = \mu_n''' = \text{etc.} \end{array}\right) \qquad (1)$$

(omitting electrical and gravitational effects) where $t' = t'' = t'''$, etc.; $p' = p'' = p'''$, etc.; and $\mu_n' = \mu_n'' = \mu_n'''$ are the potentials for different intensive parameters. Chemical potential is defined as

$$\mu_i = \left.\frac{dG}{dn_i}\right)_{T,p,n_x,n_y,\text{ etc.}} \qquad (2)$$

(from Prigogine
(and Defay, 1954

where G = Gibbs free energy

n_i = molar quantity of component i

The equations have been summarized by Gibbs into the following "free energy function":

$$dG = TdS - PdV + \mu_i\ dn_i \text{ --- } \mu_n\ d\mu_n \qquad (3)$$

(at constant T,P)

where T = temperature (t' etc. above)

S = entropy

P = pressure (p' etc. above)

V = volume

In geological rock systems the thermal and mechanical parts of equation (3) are usually zero so that the equation reduces to:

$$dG = \mu_i\ dn_i' - \mu_n\ dn_n'$$

If one considers two phases, α and β containing c components; with a change of intensive parameters an exchange of a component i will occur so that the affinity for reaction

$$A = \mu_i^{\alpha} - \mu_i^{\beta} = 0 \qquad (4)$$

(Prigogine & Defay, 1954)

at equilibrium.

Albee (1965) states however that

> "equilibrium does not require that the chemical potential of each component be the same in both phases; it merely requires that the free energy at constant P and T be minimum; or dGβ" (p. 275)

Gibbs (1906) clearly states, however, that for a system at equilibrium

> "the potential for each component substance must be constant throughout the whole mass" (p. 65).

Gibbs' definition will be adhered to here as it is by other authors (e.g. Phinney, 1963) For non-ideal situations the effect of the non-ideality is expressed in the following way:

$$\mu_i = \mu_i^{\circ}\ (T,P\ \text{etc.}) + RT \ln x_i \gamma_i \qquad (5)$$

(Prigogine & Defay, 1954)

where

μ_i = chemical potential at new condition

μ_i° = " " at standard state

R = gas constant

x_i = mole fraction of component i

and γ_i = fugacity coefficient,

and for two phases in which component i is present

$$\left.\begin{aligned} \mu_i^{\alpha} &= \mu_i^{\circ\alpha}\ (T,P) + RT \ln x_i^{\alpha}\gamma_i^{\alpha} \\ \text{and } \mu_i^{\beta} &= \mu_i^{\circ\beta}\ (T,P) + RT \ln x_i^{\beta}\gamma_i^{\beta} \end{aligned}\right) \qquad (6)$$

From equation (4)

$$A = \mu_i^{\circ\alpha} - \mu_i^{\circ\beta} + RT \ln \frac{x_i^{\alpha}\gamma_i^{\alpha}}{x_i^{\beta}\gamma_i^{\beta}} \qquad (7)$$

and from Gibbs free energy considerations

$$\mu_i^{\circ\alpha} - \mu_i^{\circ\beta} = RT \ln K_i\, T_p \qquad (8)$$

which, when substituted into (7), yields

$$A = RT \ln K_{i\,(T,P)} + RT \ln \frac{x_i^{\alpha}\,\gamma_i^{\alpha}}{x_i^{\beta}\,\gamma_i^{\beta}}$$

and at equilibrium A = 0 so that

$$\frac{x_i^{\alpha}\,\gamma_i^{\alpha}}{x_i^{\beta}\,\gamma_i^{\beta}} = K_{i\,(T,P)} \qquad (9)$$

This equation shows that the activity ($= x_i\,\gamma_i$) ratio of component i soluble in phases α and β is a constant at equilibrium.

The value of γ_i is nearly impossible to predict and must be measured. However, Henry's law predicts that at infinite dilution γ_i also equals unity and we may therefore write

$$\frac{x_i^{\alpha}}{x_i^{\beta}} = K_{i\,(T,P)} = D_{i\,(T,P)} \qquad (10)$$

This is the Nernst distribution law and D_i is the distribution coefficient at a specific T and P.

At trace element concentrations, the Nernst law should be valid and a linear plot of points should be produced for a particular element partitioned between two phases at equilibrium. This is assuming that the trace element occurs in a compound stable in both phases. Kretz (1961) states that the bulk composition may have the effect of producing a non-linear relationship. This view is not accepted in this study. A non-linear plot is possible, however, if the concentration of the element is too great to follow the Nernst distribution law distribution in the particular mineral structures involved.

The distribution coefficient (in equations 9 and 10) will vary with changing intensive parameters so that if we look at distribution coefficients of elements between the same mineral pairs from different metamorphic conditions, a systematic variation should occur. This result is also implicit if equilibrium has been attained between the two phases at each P,T interval. It must be pointed out again, however, that the term "metamorphic grade" covers only the effect of the major intensive parameter(s) on the mineralogy and may, for the distribution of a particular element, not be as significant as another intensive paramenter such as P_{O_2}.

Thermodynamic laws are, in essence, gross generalizations of a host of interrelated physiochemical factors.

Goldschmidt formulated the following rules:

(a) For two elements to enter into a crystal structure, the size of the ions must be comparable.

(b) In ionic crystals, if sizes are similar and charges identical, then the smaller ion will preferentially displace a larger ion.

(c) If sizes of ions are similar, a more highly charged ion will displace an ion with smaller charge.

These rules have not been challenged seriously (Burns and Fyfe, 1966).

The main criticism has been with regqrd to the deviation from ionic behaviour of many ions. Ahrens (1964) and Ringwood (1955) have modified Goldschmidt's rules, particularly with respect to non-ionic behaviours.

Fyfe (1951) suggested a model of "stereochemical control" or considerations of electronegativity in cases where covalent bonding is substantial.

Fersman (see Doerffel, 1953) studied lattice energy values and found elements of similar lattice energy contributions hehaved in a similar geochemical fashion.

Burns and Fyfe (1965) maintain that as "we introduce any medium more complex than the gas phase, such as an aqueous fluid or silicate melt, any analysis of value must involve an estimate of thermodynamic functions for the steps of a cycle". They state that a cycle of the following type must be considered

$$\begin{tikzcd}
X^{+}_{\text{fluid}} + YZ_{\text{solid}} \arrow[d, "\Delta H_{\text{solvation}}"] & & XZ_{\text{solid}} + Y^{+}_{\text{fluid}} \\
X^{+}_{\text{gas}} + YZ_{\text{solid}} \arrow[dr, "U"] & & XZ_{\text{solid}} + Y^{+}_{\text{gas}} \arrow[u, "\Delta H_{\text{solvation}}"'] \\
& X^{+}_{\text{gas}} + Z^{-}_{\text{gas}} + Y^{+}_{\text{gas}} \arrow[ur, "U"'] &
\end{tikzcd}$$

The extent of this reaction will depend upon:

(a) the relative lattice energies U, of the solids;

(b) the relative solvation energies $\Delta H_{\text{solvation}}$ of the ions;

and (c) the overall change in free energy (ΔG).

The authors, from work by Latimer (1952), show that for many reactions involving alkali and alkali earth salts, a good balance of potentials can be indicated.

Undoubtedly fluid-gas-solid reactions occur in metamorphic rocks as can be readily inferred from dehyrdation reactions such as the breakdown of muscovite. Burns and Fyfe state that such factors as the knolwedge of the relative lattice energies are absolutely essential and "empirical short cuts of the past cannot be satisfying until such data are compiled".

Burns et al. (1964) stated that recent work to explain the structures, stabilities and general properties of the transition elements lies in the following theories:

(1) the electrostatic theory, including recent crystal field modifications;

(2) the valence bond theory;

and (3) the molecular orbital theory.

Of these three the first is easiest to understand and the most amenable to accurate quantitative calculations. This theory is, however, only for point charge distribution and is not applicable to situations involving covalent bonding.

Certain generalized trends of substitution coefficients have been deduced for igneous situations (Curtis, 1964; Burns and Fyfe, 1964; Williams, 1959) and metamorphic ones (Burns et al., 1964; Schwarcz, 1966).

One major problem with using this type of analysis is the inability to know what form an ion had prior to its introduction into the crystal structure. There are severe limitations to these theories which are, in practical terms, not very useful.

From the previous discussion it is obvious that the partition of elements between phases is potentially useful to determine:

(a) the degree to which equilibrium has been attained;

and (b) the values of intensive paramenters during crystallization.

The temperatures and pressures were undoubtedly uniform in a hand specimen size areas but the chemical potentials may or may not have been (see section II, D). For the chemical potential to be constant, the molar content of a component must be constant in all areas of a mineral grain. This condition was not met in this study.

The diffusion of an element follows the Arrhenius diffusional equation, which is

$$D = C_o \ E^{-Q/RT}$$

where C = diffusion coefficient

C_o = a constant

and Q = activation energy which varies with the site and the particular ions involved.

Metamorphic rocks usually have been subjected to at least two significant periods of different T, P, etc. including

(a) the one at which they formed their present mineralogy

and (b) the one at which they now occur.

If an element is very stable in lattice positions in two phases, it may show the partition coefficient indicative of the intensive parameters during period (a) of their crystallization

or

$$\frac{X^{\text{component a}}_{\text{phase 1}}}{X^{\text{component a}}_{\text{phase 2}}} = D_{T,P}\text{, etc. at crystallization.}$$

Another element may be more unstable in lattice positions at the type (b) conditions and re-equilibrate between the phases so that

$$\frac{X^{\text{component b}}_{\text{phase 1}}}{X^{\text{component b}}_{\text{phase 2}}} = D_{T,P}\text{ at the new conditions.}$$

It is also possible that an element may equilibrate between two phases at type (b) conditions but not all the phases or

$$\frac{X^{\text{component c}}_{\text{phase 2}}}{X^{\text{component c}}_{\text{phase 3}}} = D_{T,P}\text{ at the new conditions,}$$

and the partition of component c between phases 1 and 2 would give a mixed value of the distribution coefficient, or

$$\mu^{\text{component c}}_{\text{phase 1}} \neq \mu^{\text{component c}}_{\text{phase 2}}$$

In reality, as suggested from the information in section III, D an element may only partially re-equilibrate at new conditions and a gradient will be set up in a sample so that

$$\mu^{\text{component b}}_{\text{grain 1 of phase 1}} \neq \mu^{\text{component b}}_{\text{grain 2 of phase 1}}$$

and only adjacent rims of grains of two phases are in equilibrium. The subsequent partition plots will not be those of a simple curve but a two dimensional group which may or may not have an axis of slope near 45° for trace elements in a log-log diagram. This may introduce some scatter in data. Also, even if equilibrium partition can be shown, it may still not be that of the conditions during the original crystallization.

Mineral Structures and Site Occupancy

In classical thermodynamics, phases are treated as homogeneous masses in which no systematic distributions of various ions exists. This is probably due to the fact that the theory was determined from ideal gas interactions where no bonding exists. In solid crystalline structures, however, there are distinct lattice positions for various ions and a knowledge of the crystal structures is necessary.

Garnet

This mineral consists of silicon-oxygen tetrahedra existing as independent groups linked to octahedra of trivalent ions (Al^{+3}, Fe, Cr and Ti). Divalent ions (Mg, Fe^{+2}, Mn, and Ca) are situated in the interstices within the Si-Al network so that each ion is surrounded by eight oxygen ions (Strunz, H., 1957). Some replacement of Si by P or Ti in the tetrahedral position has been inferred (Deer et al., 1966).

Kyanite

The mineral consists of chains of Al-O octahedra which are linked together by the remaining Al, Si and O ions. Oxygen atoms are arranged in a slightly distorted close-packed cubic array. Si is coordinated by 4 oxygen ions Al by 6. The Si lies between 4 oxygens and gives a structure with independent SiO_4 tetrahedra (Naray-Szabo et al, 1929). Pearson and Shaw believe the elements Mg, Ti, Ni, Co, Mo, Sn, Cu, Zr, Li and V occur at aluminum sites,Be might replace Si and the positions of B and Ba are uncertain.

Sillimanite

Sillimanite consists of Al-O octahedra parallel to the Z axis. Within the chains, octahedra of 6 oxygen ions around Al share edges with neighbouring octahedra on either side. Al is half in octahedral and half in tetrahedral coordination (Hey et al., 1931). Pearson and Shaw 1960 suggest the elements Mg, Ti, Ni, Co, Mo, Sn, Cu, Zr, Li and V occupy 6 coordinated positions. Fe as well as Mg, Mn and Ti substitute for Al.

Muscovite and Biotite

These minerals consist of composite sheets in which layers of octahedrally coordinated cations are sandwiched between two identical layers of linked (Si, Al)O_4 tetrahedra. Interstitial ions include K, Na, Ca, Ba, Rb and Cs, octahedrally bound ions include Al, Mg, Fe, Mn, Cr, V, Ti and Li and tetrahedrally bound ions include Al, Si and perhaps some Fe^{+3} and Ti (Deer et al, 1966).

(ii) Ti in muscovite and Biotite

The chemical analysis of the minerals is given in appendix II.

The content of Ti in biotite has been claimed to be directly dependent on the prevailing intensive parameters during crystallization (Phillips, 1930; Ramberg, 1952; DeVore, 1955; Engel and Engel, 1960). It may be useful as a metamorphic indicator or geothermometer.

Figure 29.

The variation of the modal proportions of Ti, Fe^{+2} + Mg + $Al_{oct.}$, Fe^{+2} + Mg + Ti + $Al_{oct.}$ and Si in biotite versus metamorphic grades. The molar proportions were calculated from wet chemical analyses as shown in appendix II, 2).

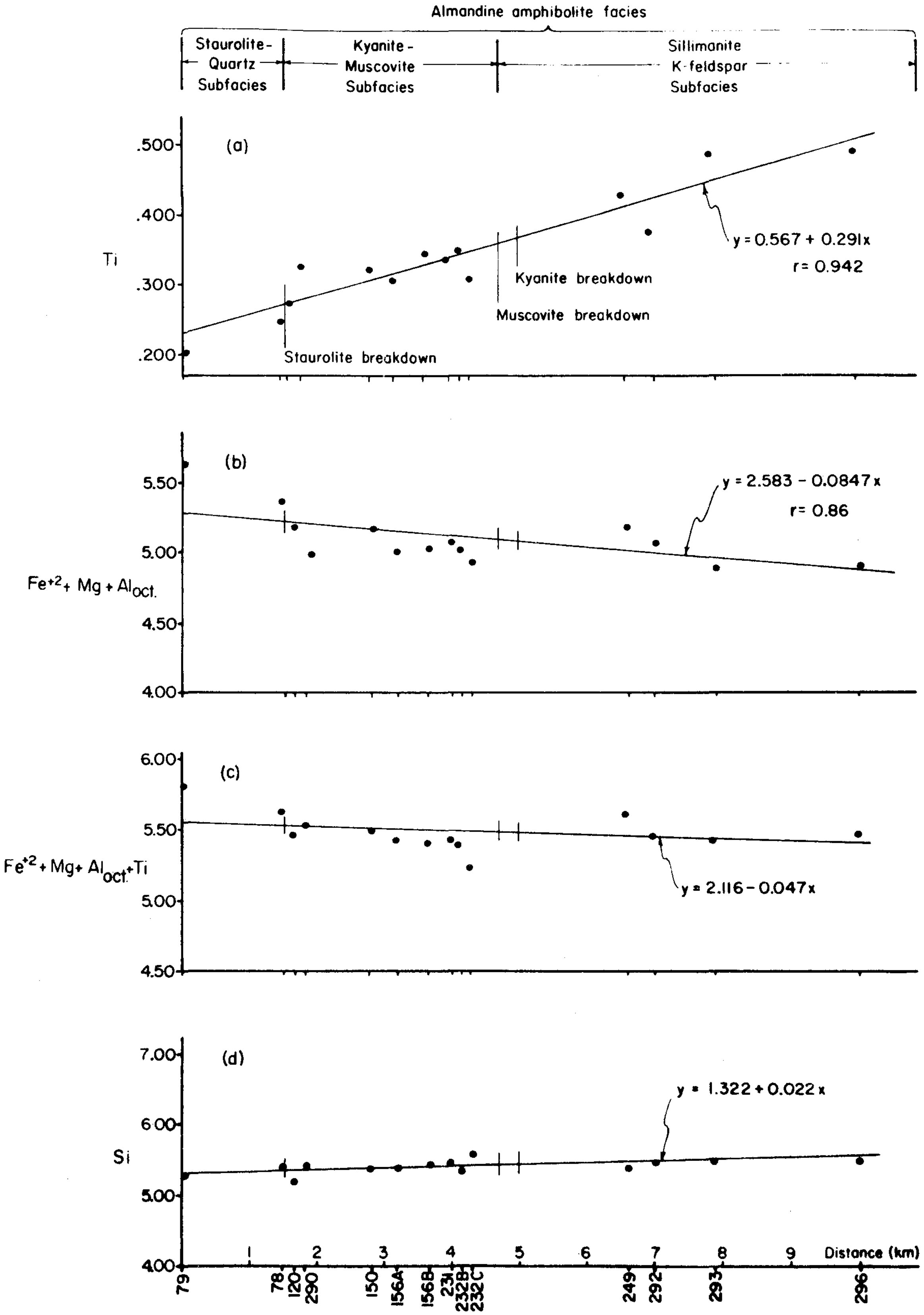

Almandine amphibolite facies
Staurolite-Quartz Subfacies
Kyanite-Muscovite Subfacies
Sillimanite K-feldspar Subfacies
(a)
Ti
.500
.400
.300
.200
y = 0.567 + 0.291x
r = 0.942
Kyanite breakdown
Muscovite breakdown
Staurolite breakdown
(b)
Fe^{+2} + Mg + $Al_{oct.}$
5.50
5.00
4.50
4.00
y = 2.583 − 0.0847x
r = 0.86
(c)
Fe^{+2} + Mg + $Al_{oct.}$ + Ti
6.00
5.50
5.00
4.50
y = 2.116 − 0.047x
(d)
Si
7.00
6.00
5.00
4.00
y = 1.322 + 0.022x
1
2
3
4
5
6
7
8
9
Distance (km)
79
78
120
290
150
156A
156B
231
232B
232C
249
292
293
296
Metamorphic Grade

Figure 29.

The position and ionic species of titanium in the biotite structure were dicussed by Kunitz (1936) who suggested Ti^{+4} replaces Si^{+4} in the tetrahedral position although large ionic size differences exist (Si^{+4} = 0.41Å: Ti^{+4} = 0.68Å). Engel and Engel (1960, p. 31) have suggested that Ti^{+4} does not replace Si^{+4} but that Ti^{+3} (0.76Å) replaces octahedrally bound Al^{+3} (0.51Å) and Fe^{+2} (0.76Å). Saxena (1960) believes Ti occurs in octahedral as well as tetrahedral positions in biotite. He suggests that at high temperatures Ti^{+4} can enter the tetrahedral site in significant amounts except when excess Al is available. This implies that the Al content (and other elements) of the rocks up to a certain concentration, will determine the Ti content of biotite (and muscovite). Ramberg (1952, p. 45) feels that relatively large amounts of TiO_2 are necessary in the biotite structure for it to be stable at high temperatures. This conclusion is contrary to the experimental data of Wones and Eugster (1965, p. 1243) who have grown Ti-free biotite of compositions on the annite-phlogopite join at temperatures of up to 900°C. Kretz (1958, p. 390) believes that Ti enters the same sites as Fe, Mg and Mn in biotite. Butler (1967, p. 258) implies that Ti occurs in similar sites as Fe^{+3} but not Mg^{+2} or Fe^{+3} in bitotie or muscovite.

There is some evidence that Ti^{+3} as well as Ti^{+4} is present in the biotite structure (Schwarcz, 1966, p. 507) but this has not been confirmed.

Biotite

Figure 29 shows the variation of the molar proportion of Ti in biotite versus metamorphic grade. A computer program was used to determine the best fitting curve for the data; a linear rather than quadratic or cubic solution was preferred. The correlation coefficients were not included for figures 1c and 1d because y is almost independent of x as can be seen.

Figure 1b shows the decrease of the other octahedrally-bound cations (Al^{+3} + Fe^{+2} + Mg). The negative linear gradient is inferred to be in response to the increasing Ti content and may suggest that all the Ti occurs in octahedral positions. This may also be suggested from the near inverse symmetry of the points on the two graphs. Fe^{+3} was not included with the other octahedrally bound ions partly because of its low abundance in uncontaminated specimens (e.g. 78) and partially because of contamination by limonitic surface stain on some of the concentrates. Mn is of too low an abundance to be important.

It should be noted that the difference over the metamorphic interval of Al^{+3}_{oct} + Fe^{+2} + Mg is greater (-0.43 mole) than Ti (-0.27 mole) which may be explained by excess charge supplied by tri--or quadravalent Ti replacing positions previously held by divalent ions and/or an increase of octahedral vacancies. The total octahedral occupancy (Al^{+3} + Fe^{+2} + Mg + Ti in figure 1c) also decreases.

The proportion of Si shows a slight increase (approximately +.15 mole see figure 1d) which means that whether 1) the net tetrahedral valency increases, and octahedral and/or interlayer cationic valency decreases by an increase in vacancies and/or divalent substitution for trivalent ions; 2) or the net tetrahedral valency remains constant with an increase in the number of vacancies previously held by Al^{+3}. The in- of Si would strongly suggest that Ti does not replace it although Ti could possibly exist in the tetrahedral layer substituting for Al^{+3}. Generally an increase in Si^{+4} in biotites is associated with an increase of octahedral vacancies (H. Eugster, personal communication, 1968) and although this may also hold for Ti^{+4} in the tetrahedral position, it is considered unlikely.

Muscovite

The Ti content of muscovite versus increasing metamorphic grade along the traverse is shown in figure 30. The two points well off the linear curve are anomalous due to contamination by rutile (as determined by inspection of the separates). A corresponding change of other ions in muscovite cannot be demonstrated because of the small Ti variation along the traverse and the relatively large degree of contamination.

In biotites, the size difference and the fact that Si increases in content across the traverse strongly suggests Ti^{+4} does not replace Si^{+4} as Kunitz (1936) suggested. It is believed that Ti^{+4} (or Ti^{+3}) occupies the same sites as all the other main octahedrally bound cations and not only some of them as various authors have suggested.

Engel and Engel, 1960 show a linear increase of from .36 to .60 moles Ti in biotite over an inferred temperature interval of from 500 to 600°C. These values are significantly higher than the values in this study although the muscovite breakdown isograd was also near the center of the section. The differences between the two studies may be caused by incorrect geothermometry or the fact that the Ti excess phase in Engel and Engels' study is ilmenite while in this study it is rutile. The quantity of Ti in the biotite structure should depend on the chemical potential (μ) of Ti in the system (among other factors) which will be governed by the chemical potential of Ti in ilmenite and rutile so that:

$\mu^{Ti}_{Biotite}$ in an ilmenite buffered system $>$ $\mu^{Ti}_{Biotite}$ in a rutile buffered system.

From this it is suggested that suitable phases must be present to use the data and also points out the extreme importance of knowing the accessory minerals, including the opaque minerals, in a metamorphic study. There was no direct relation between bulk composition (including TiO_2 content) and Ti substitution in this study as Saxena (1966) infers for similar rocks. (See Kwak, 1968, in press).

Al^{IV} is generally thought to increase in biotite with metamorphic grade (e.g. Butler, 1967, p. 26) but the opposite was found in this study.

The linearity of Ti in biotite and muscovite is puzzling as usually solubilities follow an exponential trend. The reason for the linearity of trend is not known but seems to follow directly the inferred linear increase of P and T.

The partition coefficient $D^{Ti}_{Biotite/Muscovite}$ decreases linearly 37 per cent over an interval of 100°C and does not appear to be sensitive to the change of assemblage (loss of staurolite).

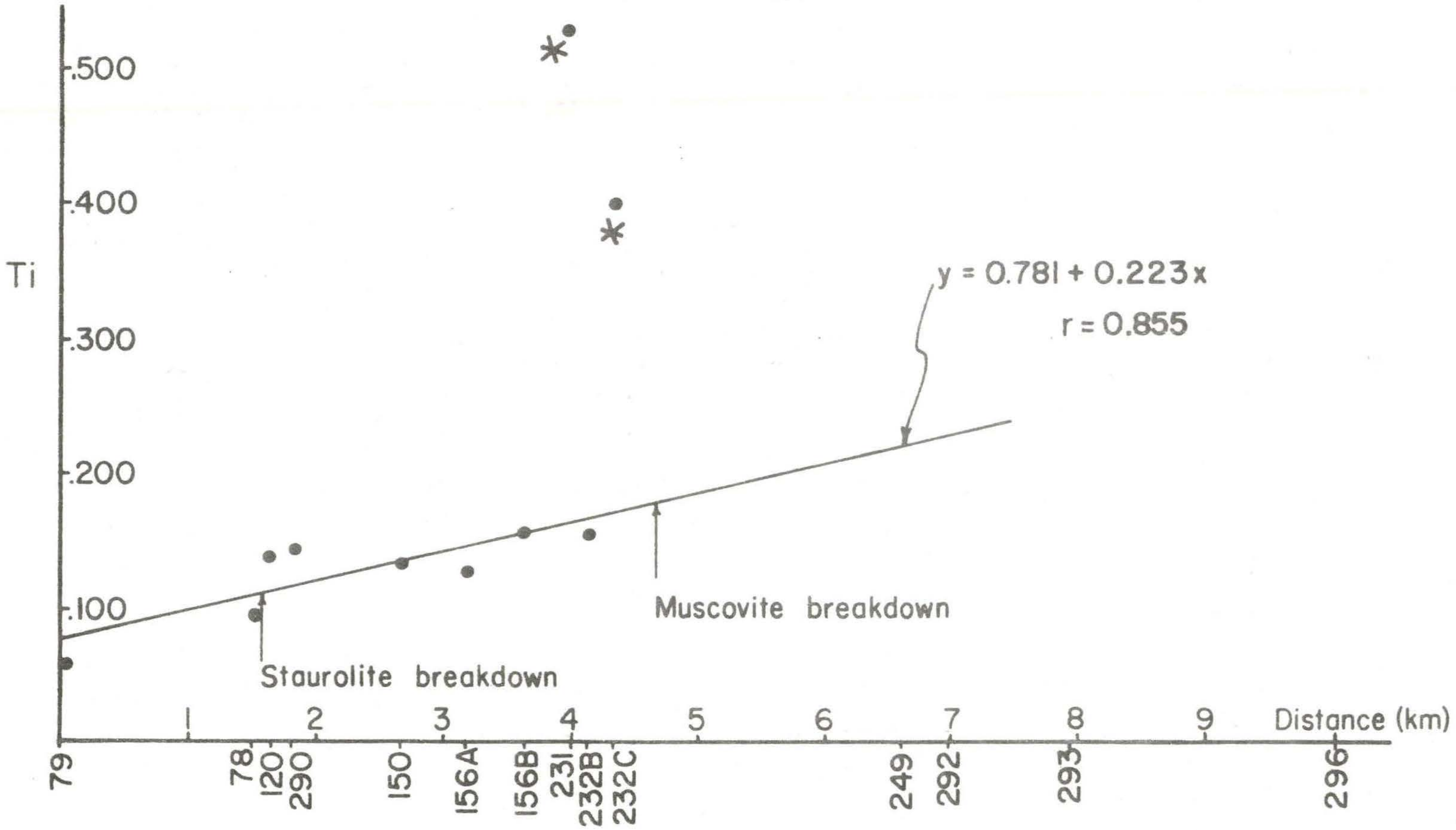


Figure 30.

The variation of Ti (molar proportion) in muscovite versus metamorphic grade. The samples marked * are believed anomalous due to rutile contamination (see Appendix I).

The linear increase of Ti in biotite and muscovite suggests its substitution is independent of the kind of other ions present at the site.

iii) Garnet compositions and metamorphic grade

(a) Mn in Garnet

There are many phenomineralogical rules present in the metamorphic rock literature involving garnet. Commonly it is believed that the spessartite molecule in garnet decreases with increasing grade (Miyashiro, 1961). Such "laws" may be useful but do not necessarily state all the conditions present for their validity.

With an increased concern with chemical processes occurring during metamorphism, it becomes apparent that distribution coefficients (see section II,i) are more significant than element abundances and the results of the garnet composition studies should be explained in the framework of general thermodynamic rules.

Presumably, if garnet is the only mafic phase in a rock it would show little or no change in Mn content with changing metamorphic grade.

If, however, it is present with another mafic phase, one could expect a change of composition or amount or both, with changing grade. In the present study, a reaction which produces garnet at the expense of biotite is envisioned. If the amount of garnet increases and the amount of Mn in the rock remains constant, the Mn content of garnet will decrease with increasing grade. Figure 31 shows the variation of Mn in garnet with increasing metamorphic grade. As can be seen, there is a maximum content near the center of the section and not at the lowest grade part. Clearly

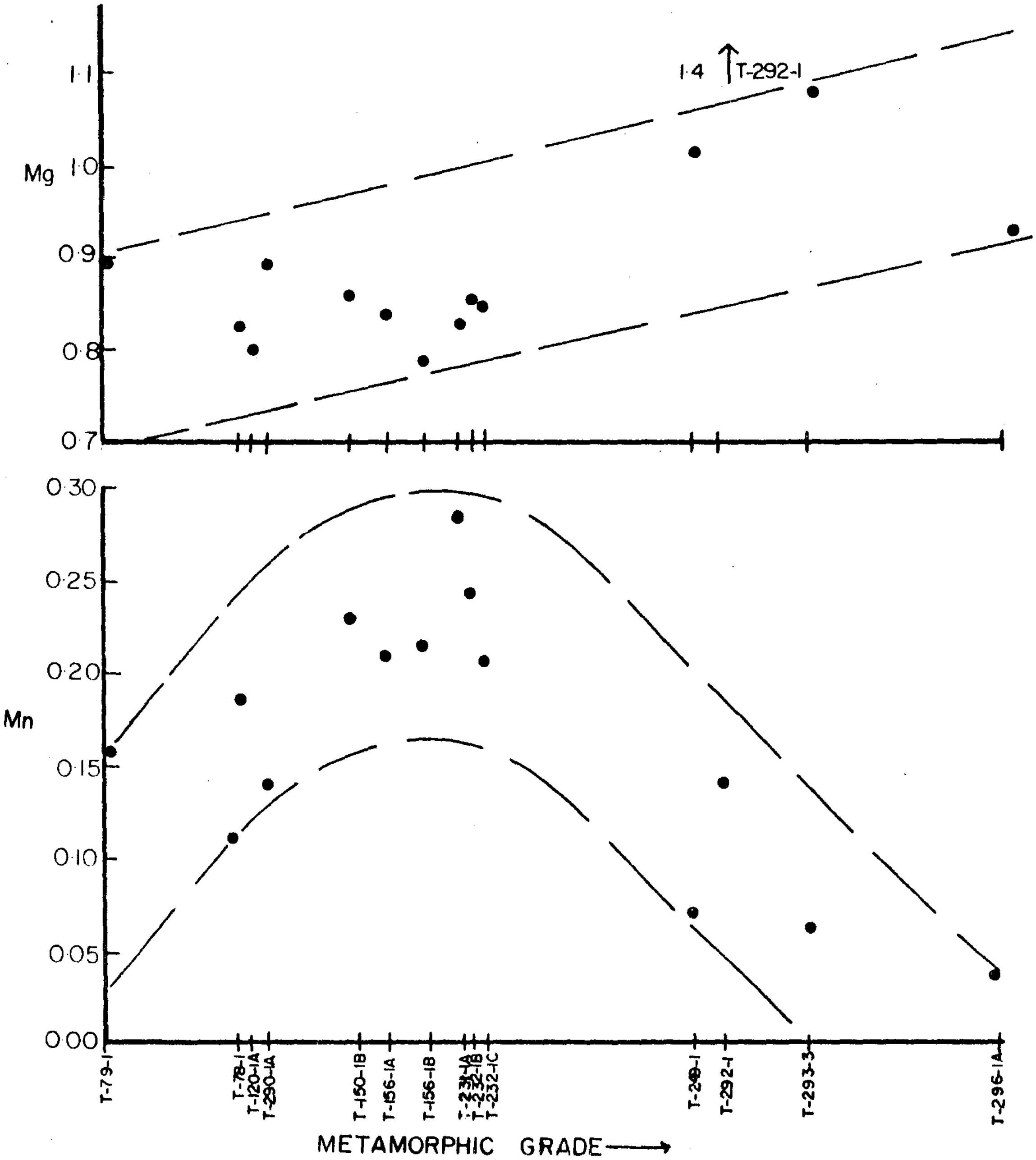


Figure 31.

The variation of Mg and Mn molar proportion in garnet versus metamorphic grade. The analyses are shown in Appendix II, 2.

Miyashiro's empirical law concerning spessartite garnet does not always hold. The effect of pressure on Mn in garnet is not well known but this should not be too important as spessartite-rich garnets have been found at high metamorphic grades (e.g. Evans and Gurdotti, 1966).

The high grade end of the section is, however, the absolute minimum. The distribution cannot be explained by inhomogeneity of MnO in the whole rock, for the content of MnO volumetric unit in the rock decreases near the center of the section. This clearly shows that the important factor is the amount of garnet; which will be affected by such features as MgO and K_2O contents to produce biotite, P_{O_2} values to produce FeO and biotite and perhaps even P_{H_2O} values which influence such minerals as muscovite and in turn, biotite.

(b) Mg in Garnet

Evans (1965) believed that "the use of the pyrope content of garnet as **a piezometer** should be restricted to an estimate of the minimum pressure of formation" (p. 1297). He also stated that growth of garnet could be due to temperature or pressure changes. These conclusions were based mainly from the evidence of Boyd and England's 1962 study in which they were able to grow pyrope only at pressures greater than 10 kb. The other garnet end members could be grown at room pressures by various experimenters. Yoder and Chinner (1960, p. 81-84) believed the pyrope content of garnets is a potentially useful barometer. Figure 31 shows the magnesium content of garnets in this study to be an increasing function of metamorphic grade.

The effect of P_{O_2} or biotite compositions and the subsequent increase of FeO in the mineral with decreasing P_{O_2} (Wones and Eugster, 1965) would affect the garnet composition by making them more pyrope rich (Mg is lost from biotite).

It is believed that total pressure and temperature are not the only factors which determine the content of Mg and Mn, respectively, in garnet.

IV Distribution of Elements Between Phases

Kretz (1959, 1960, 1961) stated that the abundance of a given element in a phase will depend on the content in that phase of elements in sites which the given element can enter. He mentioned a specific case (1958, p. 352), where for garnet, one could consider either

$$(1)\quad \frac{\text{number of atoms of element in 8 coordinated position}}{\text{total number of 8 coordinated positions}}$$

or

$$(2)\quad \frac{\text{number of atoms of element in 8 coordinated position}}{\text{total number of 8 coordinated positions minus the number of positions occupied by non-exchangeable ions}}$$

Kretz gives the example of the ratio $Mg/Fe^{+2}+Mg+Mn$ in garnet and omits Ca because it is considered "unexchangeable". The justification for such a view was that an acceptable plot was produced, which essentially makes the argument circular. In this study it was assumed that all the ions at a particular site were exchangeable.

The major and trace element partition is thought to indicate equilibrium if the points fall on a simple curve (having a slope near 45°) or show a systematic change of partition coefficient with varying metamorphic grade.

For trace elements, the partition of elements between coexisting minerals has been plotted for samples occurring metamorphically above and below the muscovite isograd. It must be remembered that there is a significant metamorphic change across the area represented by each plot so that deviations oi points from a linear plot will occur in response to the intensive parameter change, as well as to analytical error. It must be pointed out that the partition of elements between minerals may be in response to various conditions during the history of the assemblage as mentioned in Section II, E. The partition may be in part related to conditions of the highest grade of metamorphism and in part, to lower grades as is implied by the biotite-garnet relations (section II,D,IV).

The partition of trace elements between metamorphic minerals have been discussed by many authors and some feel they can be used to support a model of equilibrium (Albee, 1965), others believe the use of trace elements is very restricted (Turekian and Phinney 1962). The verification of the use of trace element to show equilibrium still needs experimental data to confirm results from field studies and extrapolations from theoretical considerations. The author realises the many possible limitations of inferences from trace element data but they still seem, along with supporting textural evidence etc., a useful medium to use.

The following section deals with the partition of major and trace elements between kyanite, sillimanite, biotite, muscovite and garnet. There are no known data on elemental partition involving kyanite and sillimanite in the literature. All distributions involving kyanite and muscovite are for samples collected below the muscovite isograd.

(a) Kyanite-Sillimanite

Up to the muscovite breakdown isograd the two minerals show mutual textures which are indicative of equilibrium (figure 14,C) although kyanite does not seem to be in equilibrium with muscovite, while sillimanite does, and sillimanite does not seem to be in equilibrium with biotite, while kyanite does. A few outcrops exist where kyanite and sillimanite show textures suggesting a polymophic inversion (explained in section II,B, and shown in figure 14d).

Figure 32 shows the trace element distribution between the minerals in hand specimens collected below the muscovite isograd. The concentrations of elements are almost equal in the two phases, with a slight preference for the sillimanite lattice. The point groups are not linear although for the elements Cr, Be, Fa, Zr, V, Ti, Cu, Rb, Mn, and Sr, the groups have a maximum dimension which has a slope near 45°. The plots of the elements Li,Y,and Fe are considered indeterminate. Many of the trends of the trace elements show a very steep slope (near vertical on figure 32) for which no explanation is known.

The precision of the trace element analyses are given in appendix II and as be seen, some of the scatter of points may be due to absolute precision not having been attained although the deviations are not great enough to significantly alter the generalized trends.

To a first approximation, chemical equilibrium has been attained between kyanite and sillimanite up to the muscovite isograd. From this one can suggest that P-T conditions of metamorphism were on or fluctuating near those at the kyanite-sillimanite univariant in the Al_2SiO_5 system.

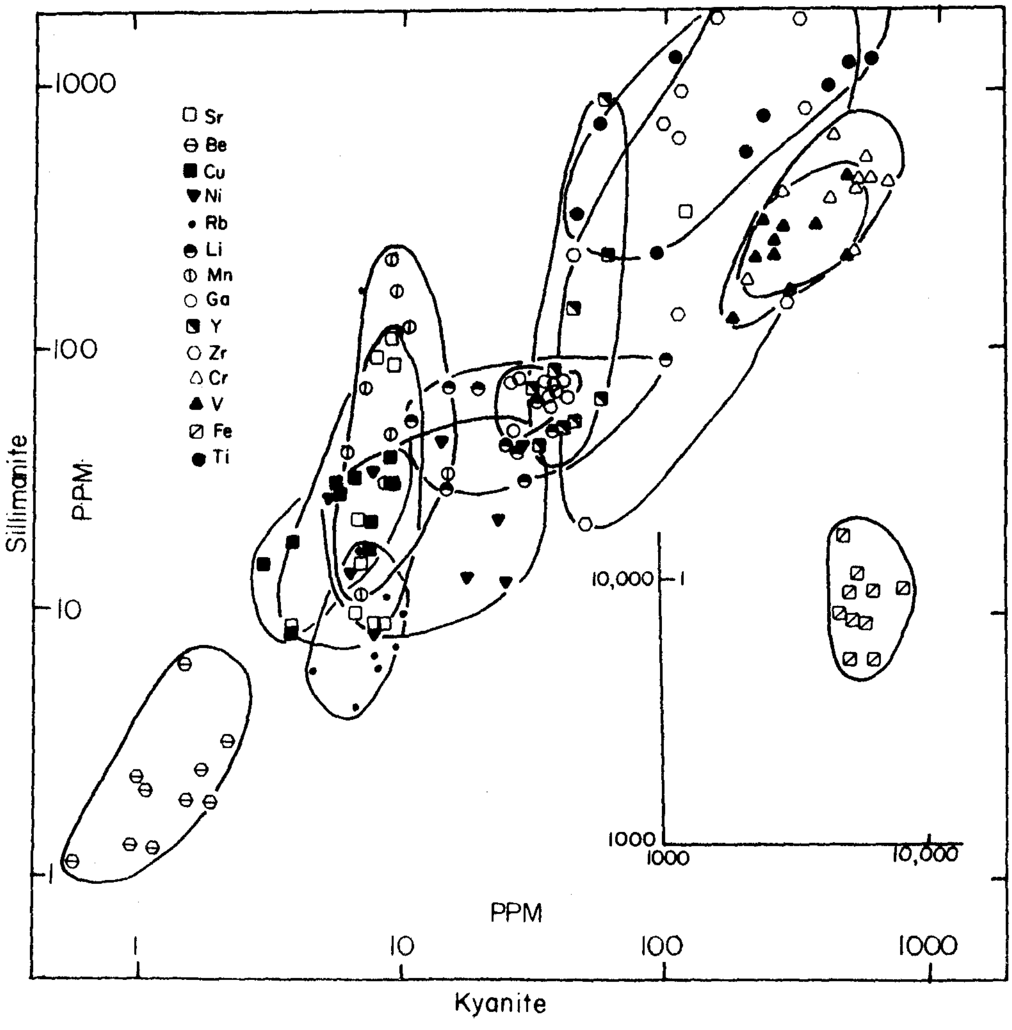


Figure 32.

The distribution of trace and minor elements between kyanite and sillimanite in samples up to the muscovite isograd.

Kyanite-Biotite

In thin section the two minerals have mutual textures from which one can infer they are stable with respect to each other. The partition of trace elements between kyanite and biotite is shown in figure 33. As can be seen, the trace elements are favoured in the biotite lattice. The trace element distributions of Y, Cr, V, and, Li are suggestive of equilibrium while Be, Ni, Rb, Ga, Zr, and Sr are indeterminate and Cu is rejected as indicating equilibrium partition. Part of the reason that the partition is fairly scattered is because biotite grains in the polished thin section contains variations of major element contents which probably are also reflected in the trace element contents, (as is described in section II,D,IV).

Kyanite-Muscovite

The relations between the two minerals have been described (section III,B) and kyanite appears to be altered partly to muscovite. The trace elements are favoured in the muscovite structure which is probably a result of a more "open" structure in muscovite (figure 34). The partitions of Cr and Cu are accepted as indicating equilibrium, Y, Ga, Ni, and Rb are indeterminate and V, Be, Li, Si and Rb, have point groups suggestive of non-equilibrium.

Kyanite-Garnet

The two minerals show mutual textures which are suggestive of their being stable with respect to each other. The trace elements are generally favoured in the garnet structure except for V, Li, Be, Ga, Rb and Co (figure 35).

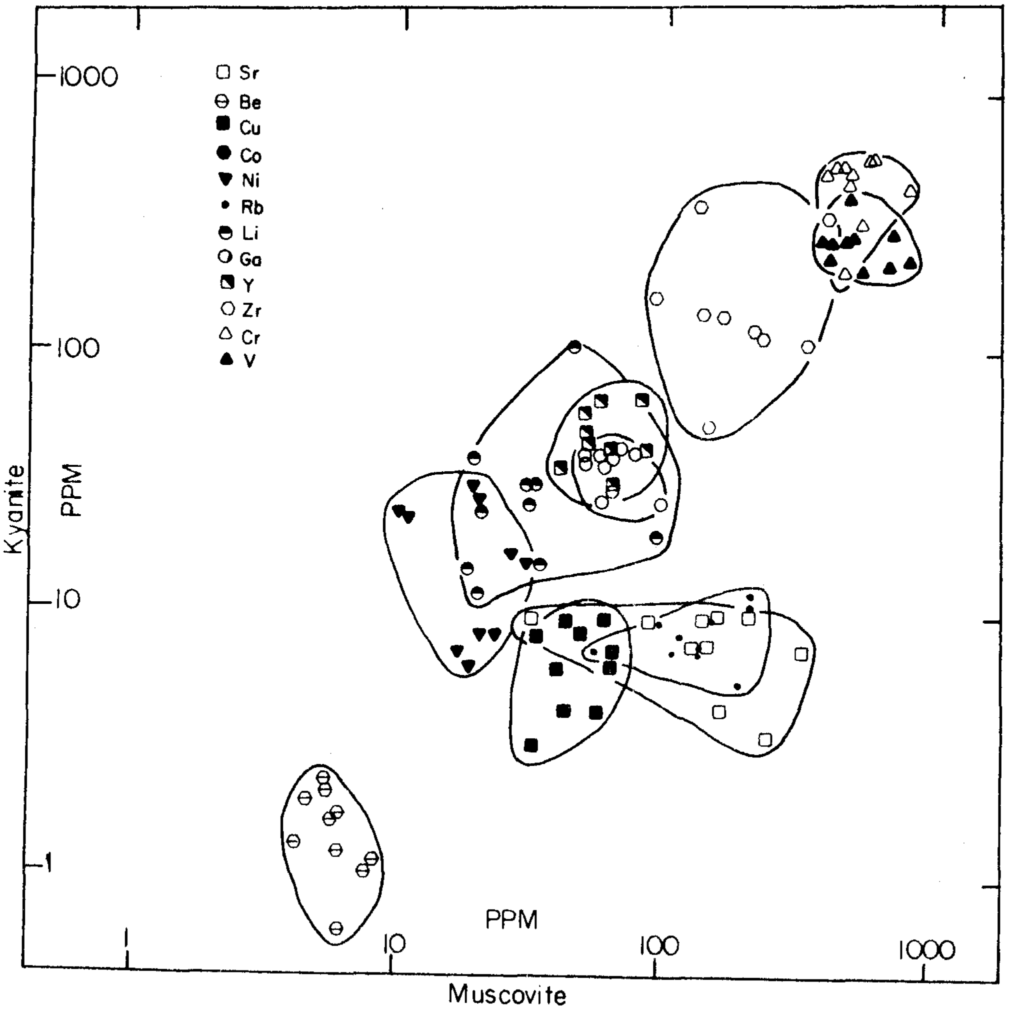


Figure 34.

Partition of trace elements between kyanite and muscovite.

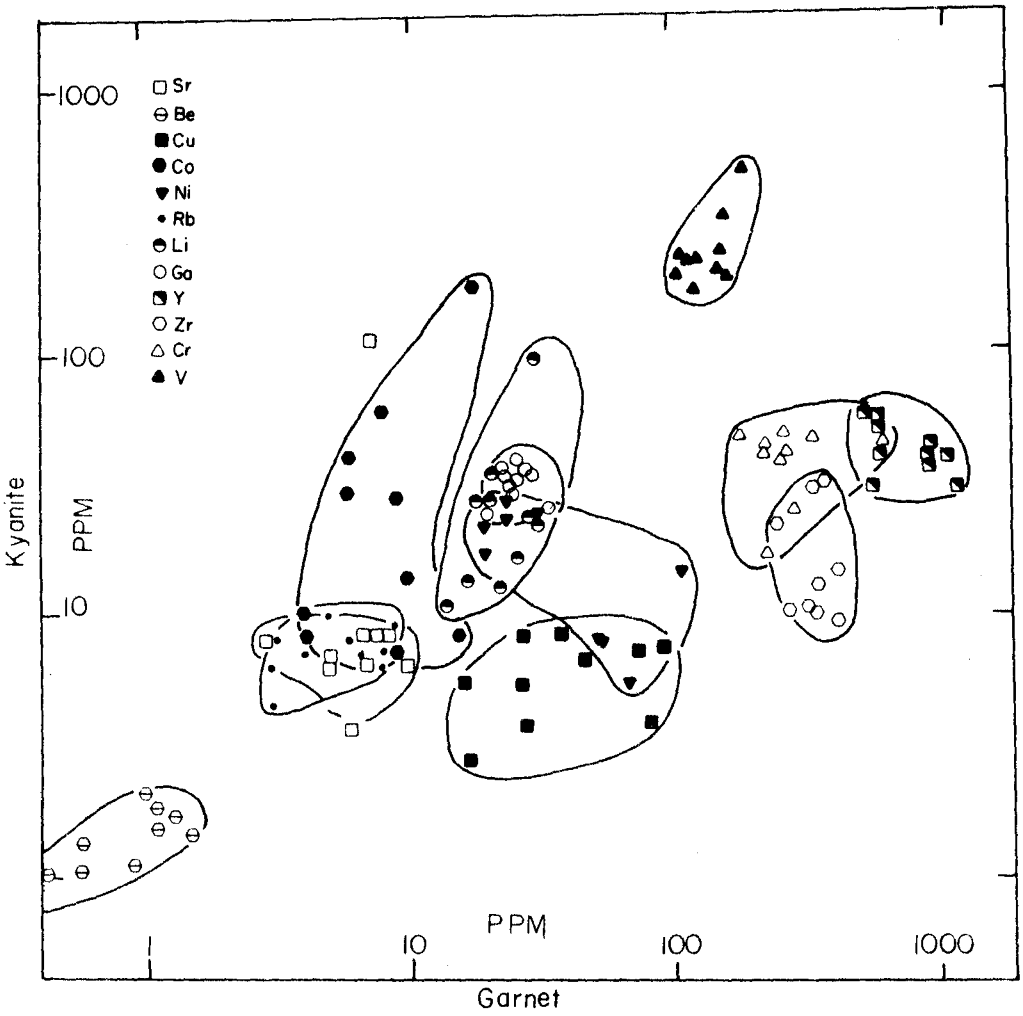


Figure 35.

The partition of trace elements between kyanite and garnet.

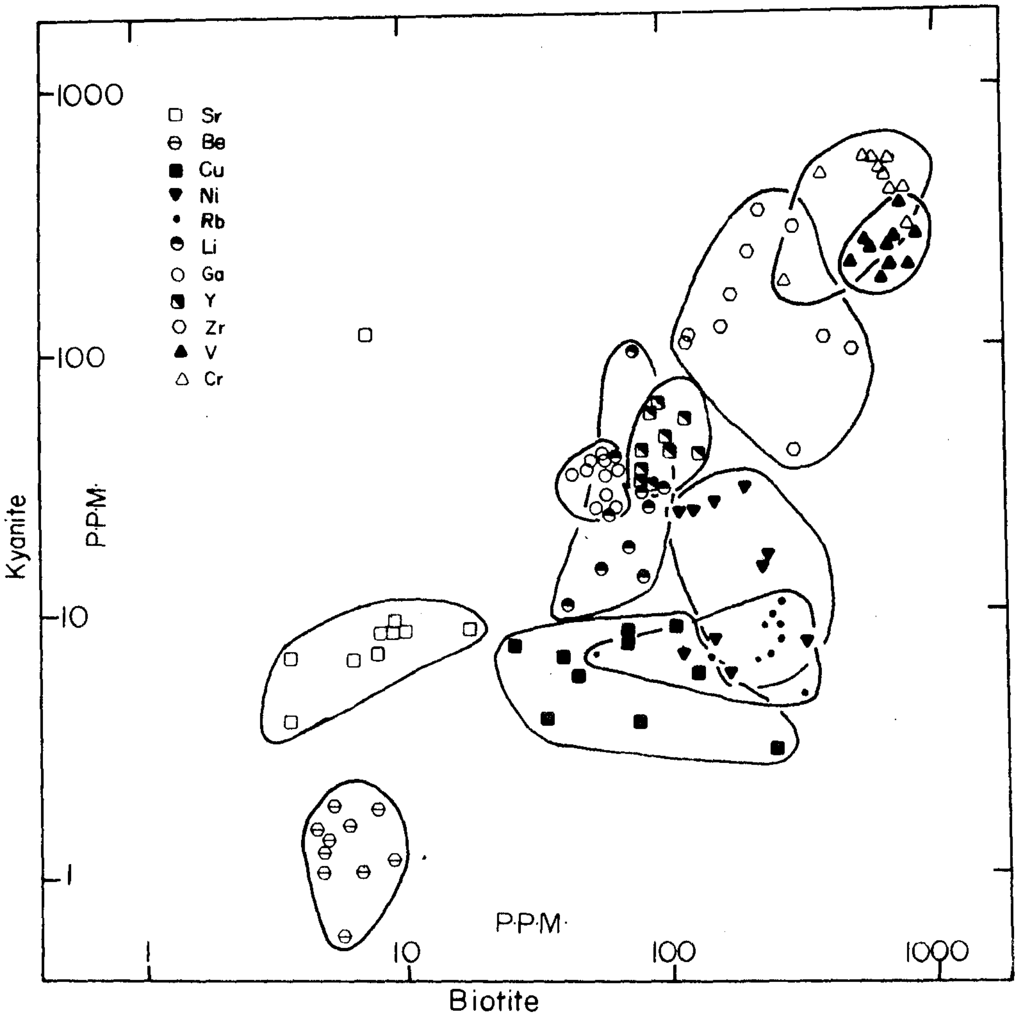


Figure 33.

The partition of trace elements between kyanite and biotite in samples collected up to the muscovite isograd.

The distributions of the elements V, Cr, Be, Co and Li are indicative of equilibrium Ga and Sr, are indeterminate and Cu, Zr, and Y are rejected as showing equilibrium.

Muscovite-Garnet

The mutual textures between the two minerals are inferred to be indicative of equilibrium. The elements Y, Sc, Co, Ni, and Cu are favoured in garnet while Rb, Be, Sr, Ga, Li, Cr, and V are favoured in muscovite (figure 36). The trace element plots of Cr, V, Sc, Ni, Ga, and Co are indicative of equilibrium partition; Sr, Y, Li, Rb, and Cu are indeterminate; and Be is inferred to show a non equilibrium partition.

The partition of Mg between the two minerals is shown in figure 37 and as can be seen by the inferred dashed trend of $D^{Mg}_{Ga/Musc.}$, there is a tendency for muscovite to become enriched in the element with respect to garnet, with increasing grade.

Mn shows a fairly scattered distribution although there is a vague trend towards concentation of Mn in garnet with increasing grade. Mn and Mg show inverse relationships and are inferred to have exchanged these elements with increasing metamorphic grade.

The scatter of points of Mg and especially Mn from a simple curve is partly due to analytical error, for the Mn content of muscovite is extremely small (0.0X weight percent) and accuracy is difficult to obtain at this level. Sample T-231-1A shows a significant deviation from a general increasing trend in figure 38 which is thought to be caused by a more complete re-equilibration of Mn to lower grade conditions as mentioned in section II, D, IV.

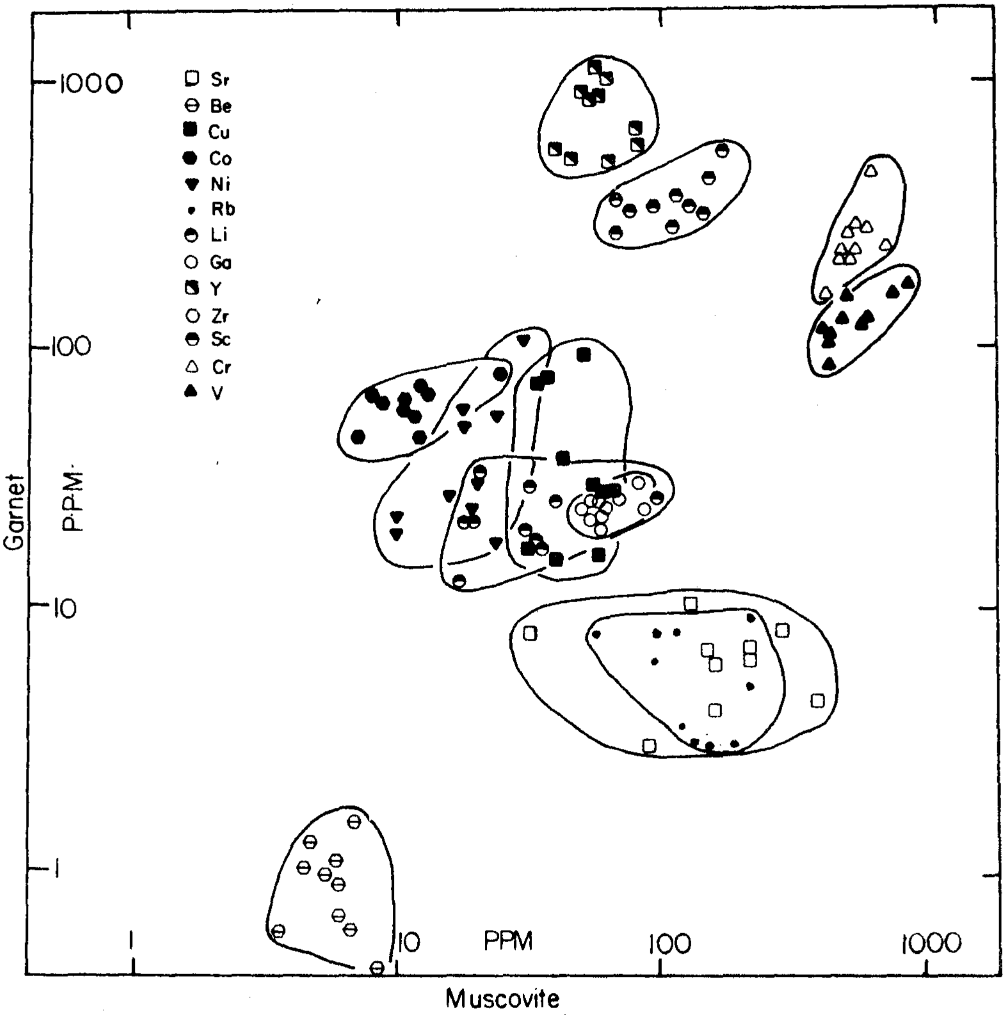


Figure 36.

The partition of trace elements between garnet and muscovite.

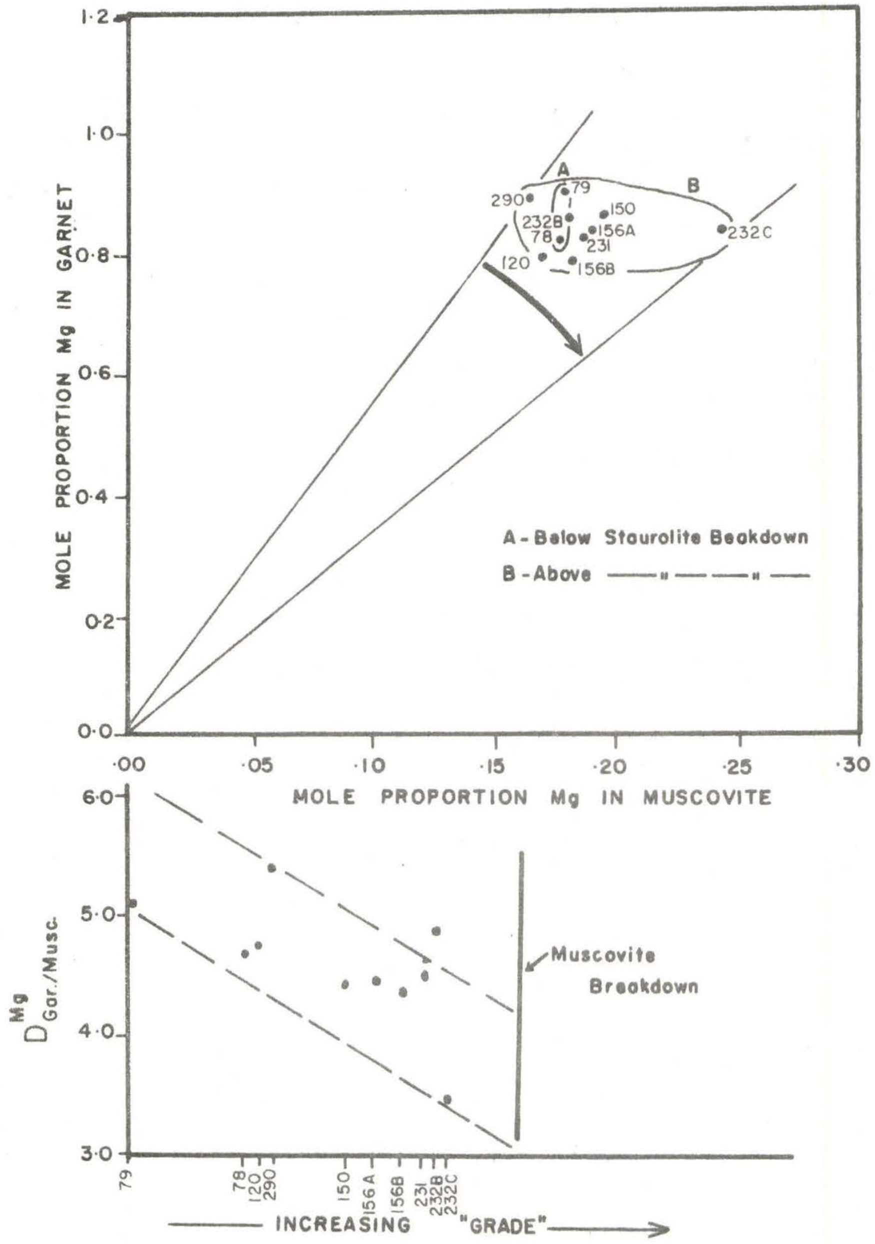


Figure 37.

The partition of Mg between garnet and muscovite. The heavy arrow in the upper part of the diagram shows the change of the distribution with increasing grades. The bottom part of the diagram shows the inferred trend of changing $D^{Mg}_{Gar./musc.}$ by dashed lines.

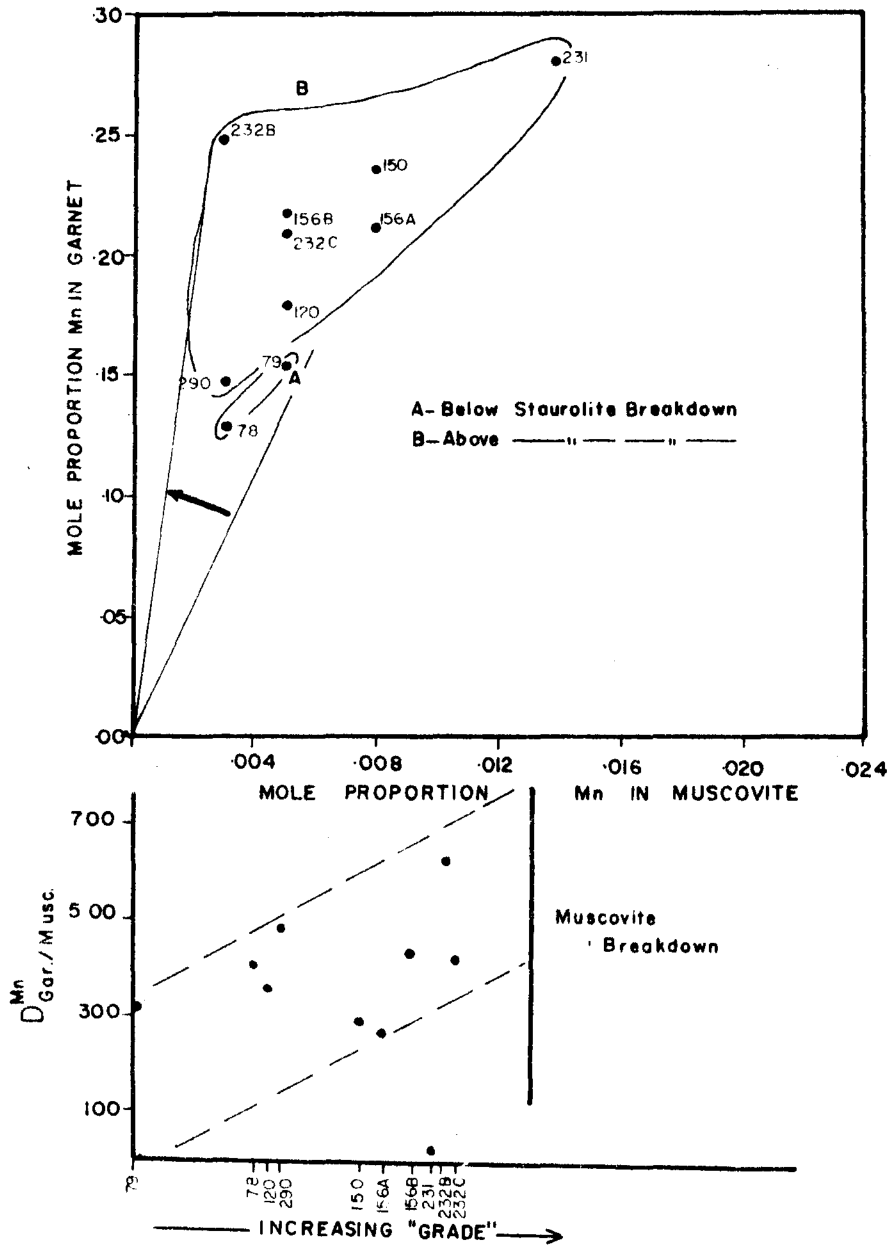


Figure 38.

This figure shows the distribution of Mn between garnet and muscovite. The dashed lines indicate the inferred trend.

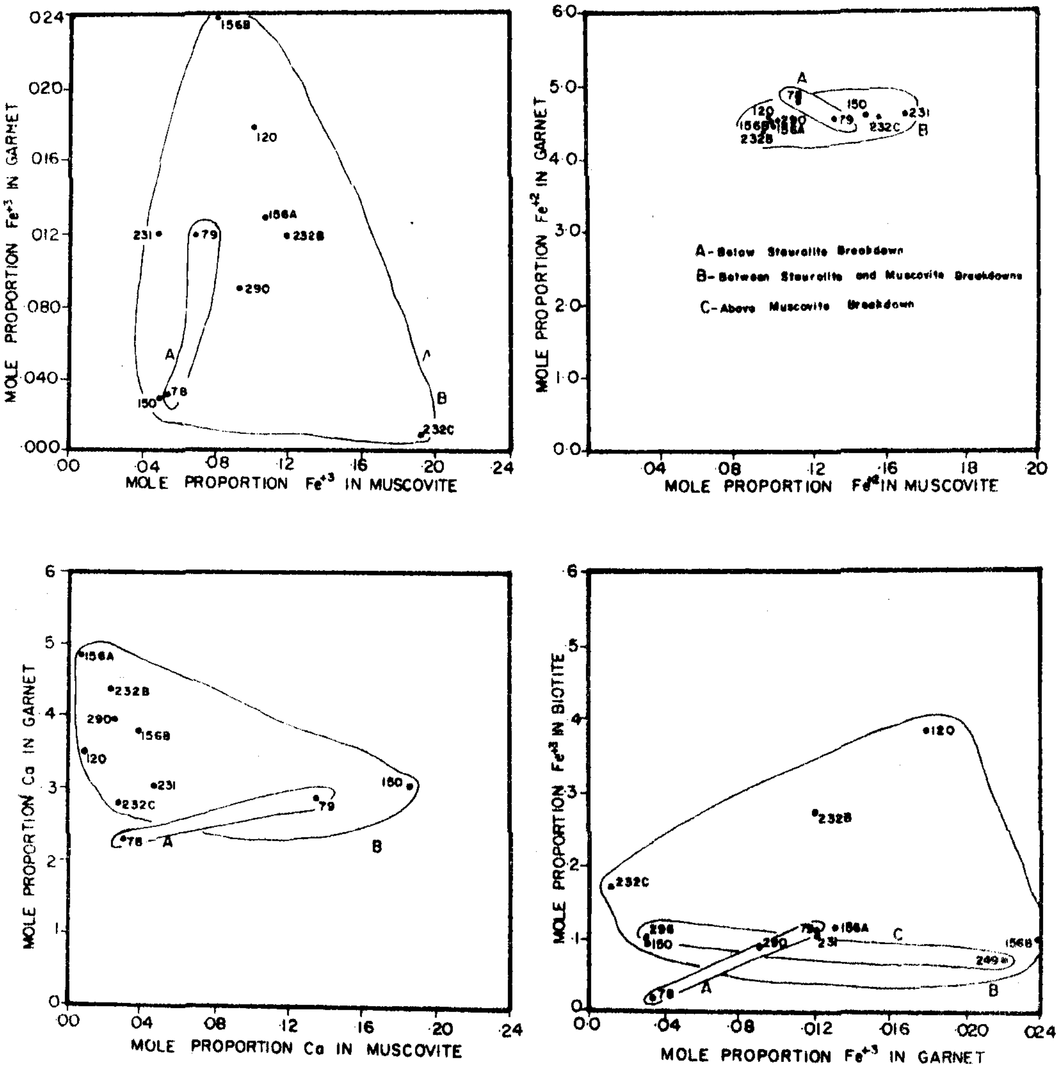


Figure 39.

The partition of Fe^{+3}, Fe^{+2} and Ca between muscovite and garnet and the partition of Fe^{+3} between biotite and garnet.

Fe^{+2} plots in a fairly small area and favours garnet over muscovite (figure 39). No relation of distribution coefficient with metamorphic grade is inferred.

Fe^{+3} is distributed fairly randomly and widely between muscovite and garnet (figure 39). The ion is generally favoured in the garnet phase. No correlation of distribution coefficient with metamorphic grade was observed which may be due to analytical error.

Muscovite seems to be the most impure with respect to limonite stain (Fe^{+3} rich) however, the fact that the ion is favoured in unstained garnet, suggests it to have the ion present at a lattice site.

Ca shows a fairly scattered plot of values between garnet and muscovite and is favoured in garnet. There is some suggestion of the distribution coefficient of garnet/muscovite increasing with metamorphic grade.

Sillimanite-Biotite

The two minerals occur intimately associated up to the muscovite isograd and have been inferred to indicate disequilibrium (in section III, B). Above the isograd, the minerals appear to be in equilibrium from a textural point of view.

As can be seen in figure 40, Zr, Ga and Sr, are favoured in sillimanite while Rb, Be, Cu, Li, Cr, V, Co and Y are favoured in garnet.

Below the muscovite isograd Cr, Cu, Sr, and Ni show equilbrium partition; Li, Y, V, Zr, and Co are indeterminate and Be and Rb show disequilibrium point groups. Above the muscovite isograd, the elements Ga, Y, Cu, Co, Be, V, and Sr show equilibrium partition, Rb and Cr are indeterminate and Ni and Li indicate disequilibrium partition.

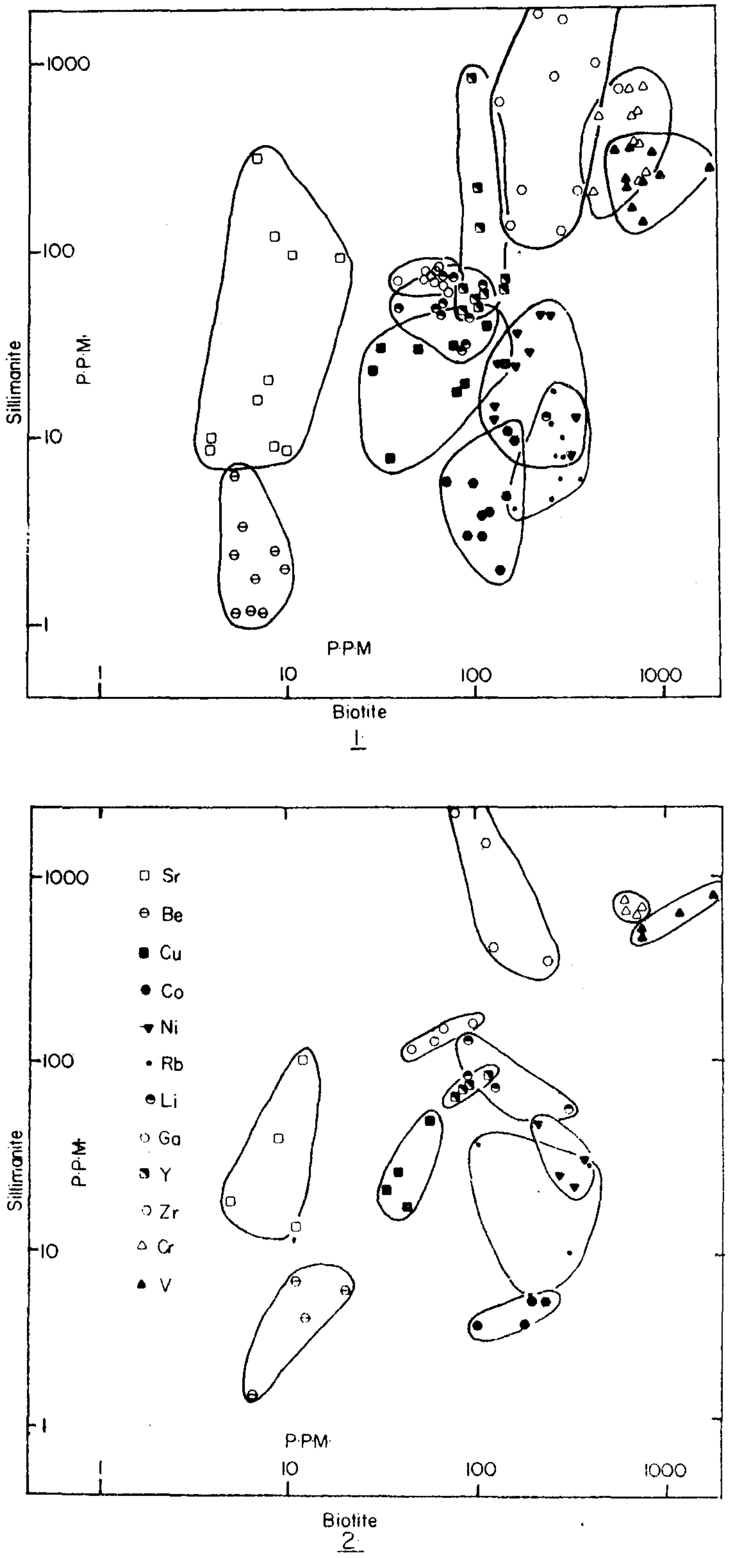


Figure 40.

The partition of trace elements between biotite and sillimanite below (1) and above (2) the muscovite isograd.

Sillimanite-Muscovite

The two minerals show mutual textures from which one can infer that the minerals are in equilibrium. In figure 41, the trace elements Zr, Cr, Ni, Li, Ga, and Y are favoured in sillimanite while Rb, Be, Sr, Cu, V, and Co are favoured in muscovite.

The partition of Ni, Cu, Li, Be, Sr, Rb and Cr show equilibrium partition and Ga, Y, Zr, V, and Co are indeterminate.

Sillimanite-Garnet

Garnet and sillimanite appear to be texturally in equilibrium in all the thin sections studied. In figure 42, Co and Y are favoured in garnet while Ba, Zr, Cr, Ni, V, Li, Cu, Ga, Sr, Be and Rb are favoured in sillimanite. Below the muscovite isograd, the partition of Be, Cr, Ga and Ba are inferred to show equilibrium, Ni, Co, V, Li and Zr are indeterminate and Cu, Y and Rb are inferred to show disequilibrium.

Above the muscovite isograd, Li, V, Co, Cu, Ni, and Ba are inferred to show equilibrium; Ga, Y, Cr, Rb and Sr are indeterminate and Be is inferred to show disequilibrium.

Muscovite-Biotite

Some of the textures between the two minerals are suggestive of equilibrium (section III, B) while others are not.

Figure 43 shows that Sr, Zr, Cr, Ga, Be and Sc are favoured in muscovite while Li, Co, Ni, Rb, V, and Cu favoured in biotite. The partition of Co, Ni, Sc, Rb, V, Cr, and Zr to show equilibrium, Be, Sr, Ga, Li and Y are indeterminate while the partition of Cu shows disequilibrium.

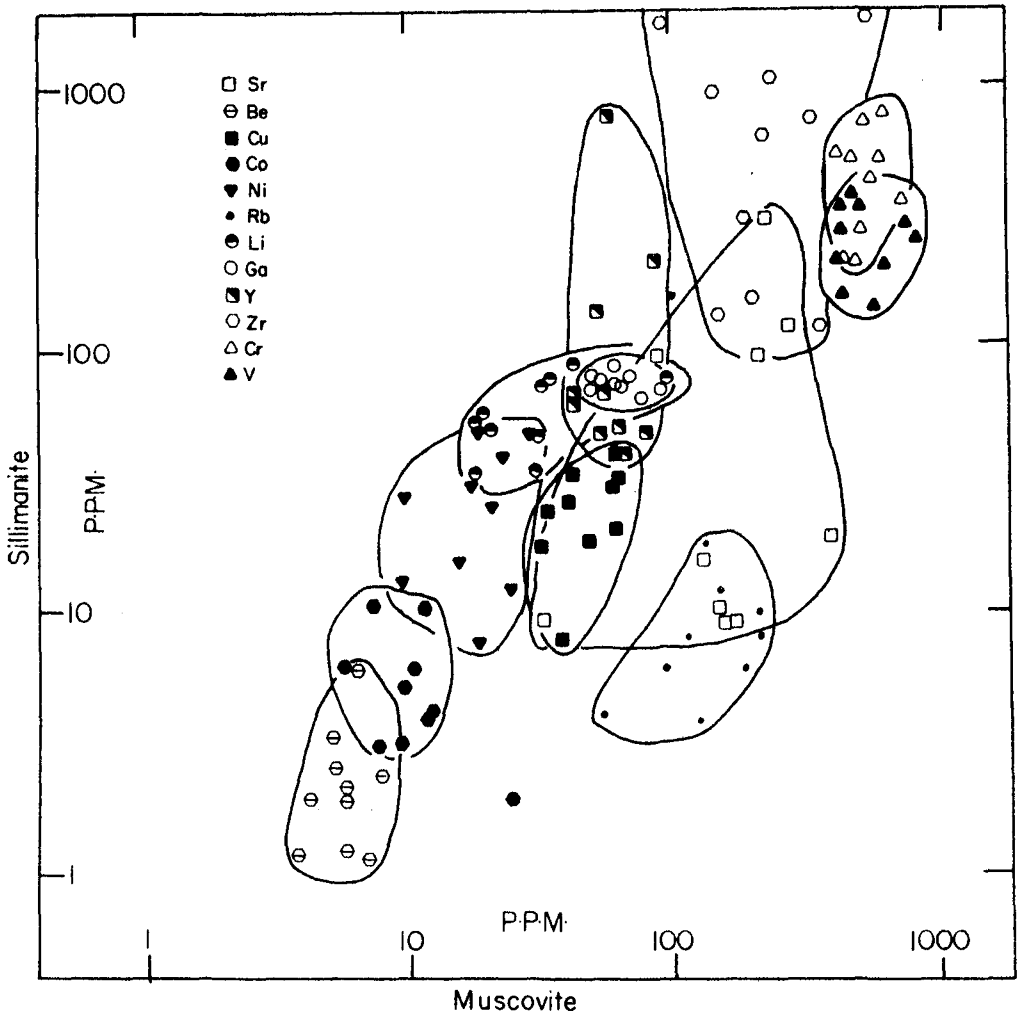


Figure 41.

Partition of trace elements between sillimanite and muscovite.

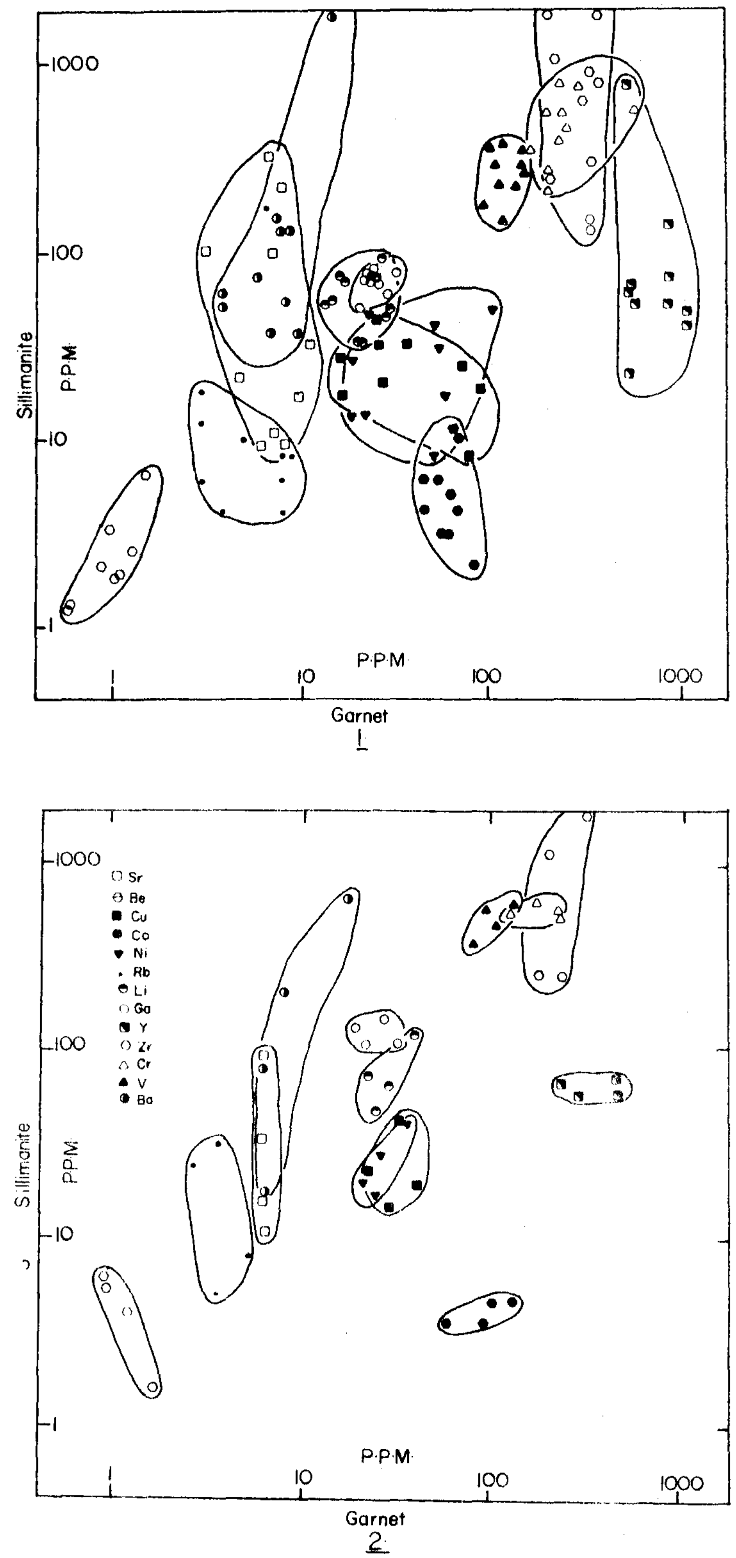


Figure 42.

Partition of trace elements between garnet and sillimanite from samples below (1) and above (2) the muscovite isograd.

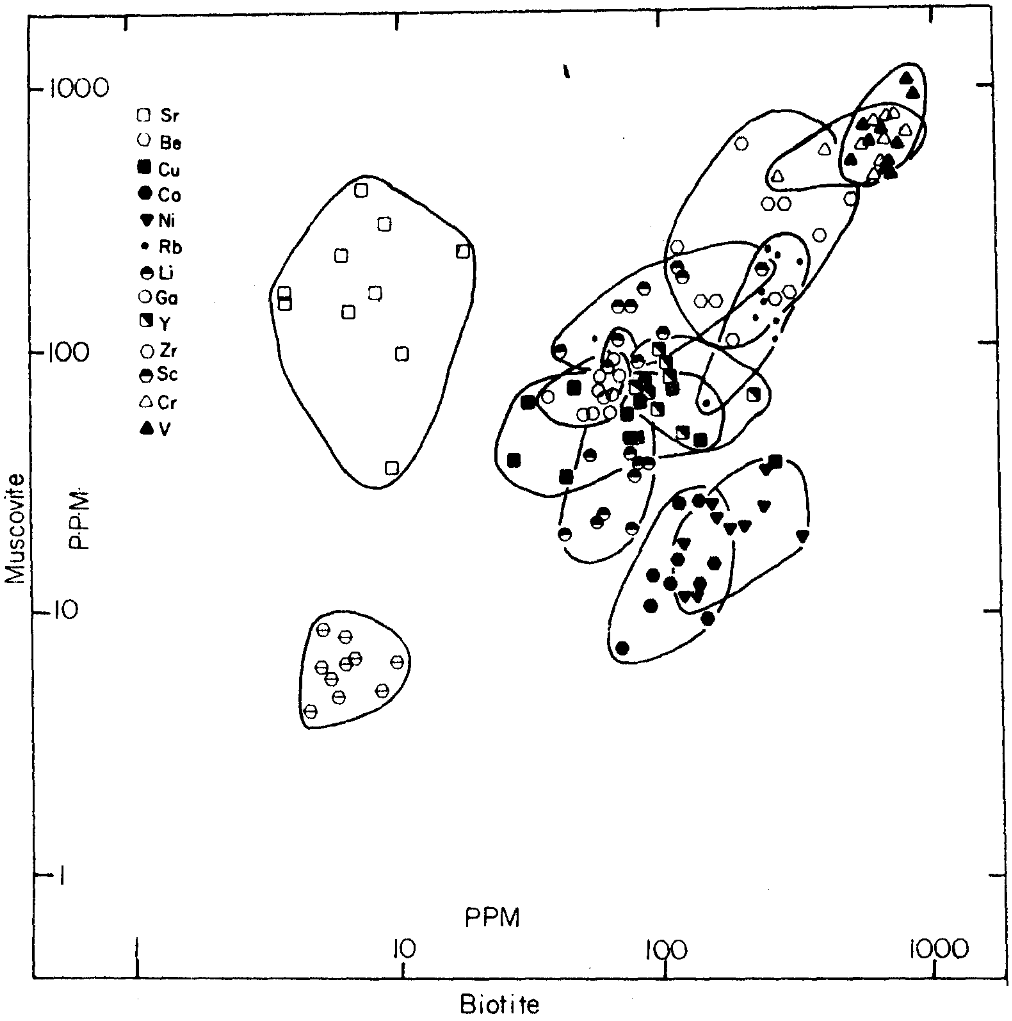


Figure 43.

Partition of trace elements between muscovite and biotite.

The distribution coefficient $D^{Mg}_{Bio./Musc.}$ decreases across the metamorphic interval as can be seen in figure 44. Sample T-232-1C is anomalously low which may be caused by a factor such as mentioned in section III, D, IV.

The distribution coefficient $D^{Mn}_{Bio./Musc.}$ probably increases with increasing metamorphic grade (figure 45). The Mn determinations were probably very susceptible to analytical error as the wet chemical methods are not very accurate at the .0X% level.

The partition of the Fe^{+2} is favoured in biotite. No correlation between metamorphic grade and distribution coefficient was observed (figure 46).

The distribution of Fe^{+3} is very erratic and does not indicate a preference in either phase (figure 46). It must be remembered that muscovite is stained by limonite in some sections which may account for the anomalous Fe^{+3} values (see appendix II, 2).

The partition of Na shows a fair scatter of points although an inference of an increase of the coefficient $D^{Na}_{Bio./Musc.}$ (figure 47) may be made.

Figure 46 shows the distribution of Ti between the two minerals. Samples T-232-1C and T-231-1A were contaminated by rutile (see appendix II) The figure suggests that the distribution of Ti is not very dependent upon metamorphic conditions. The element is strongly favoured in biotite.

The partition of K favours neither mineral and no regular change of the value of $D^{K}_{Bio./Musc.}$ was found (figure 47).

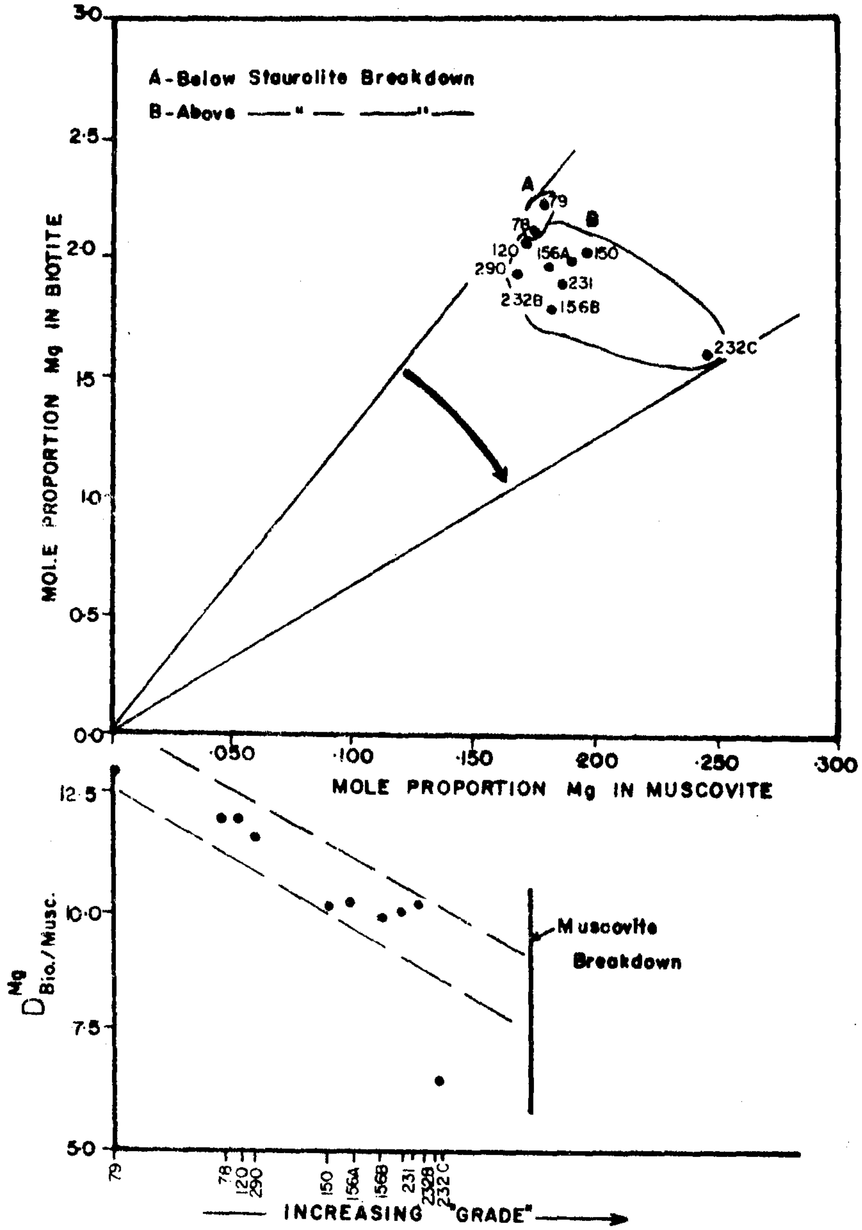


Figure 44.

The partition of Mg between muscovite and biotite. The dashed lines indicate the inferred trend.

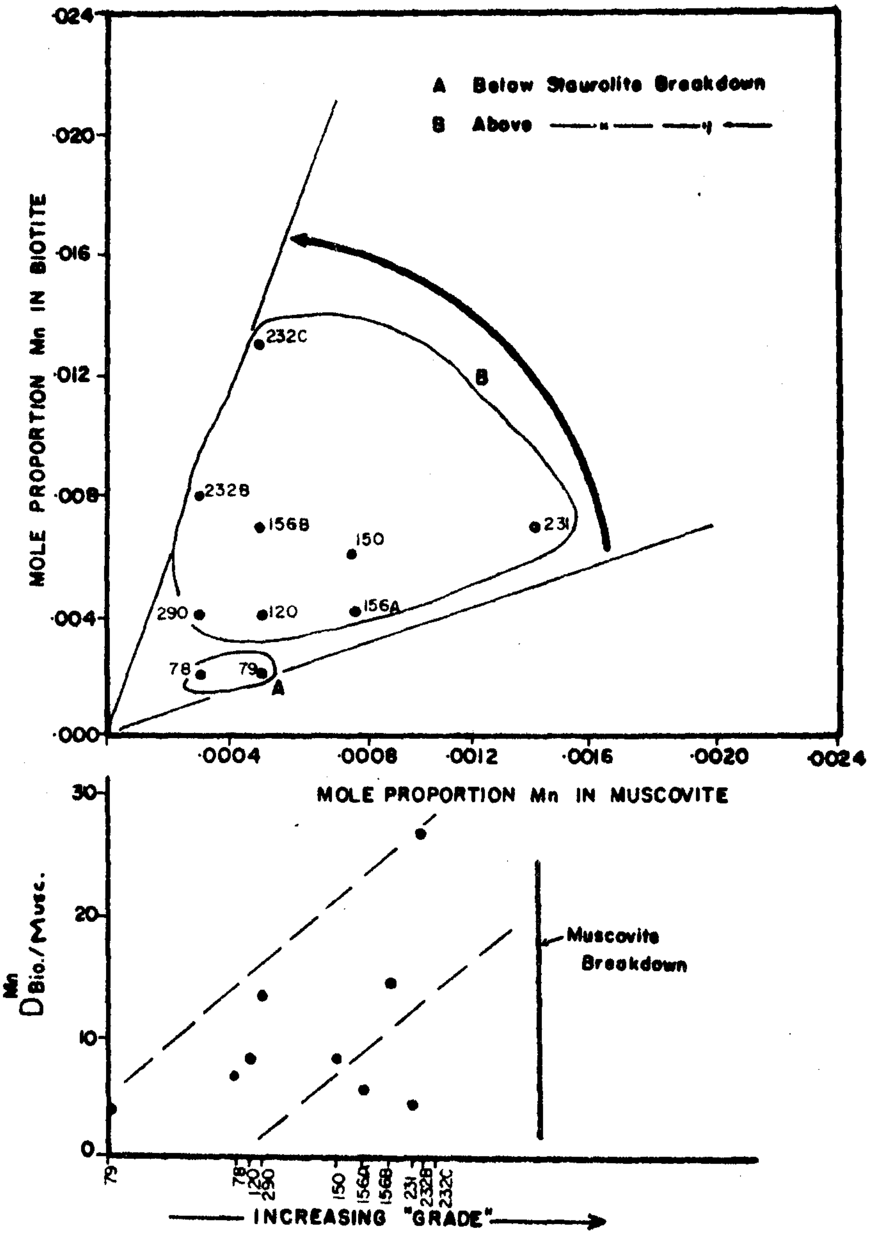


Figure 45.

Partition of Mn between muscovite and biotite. The dashed lines indicate the inferred trend.

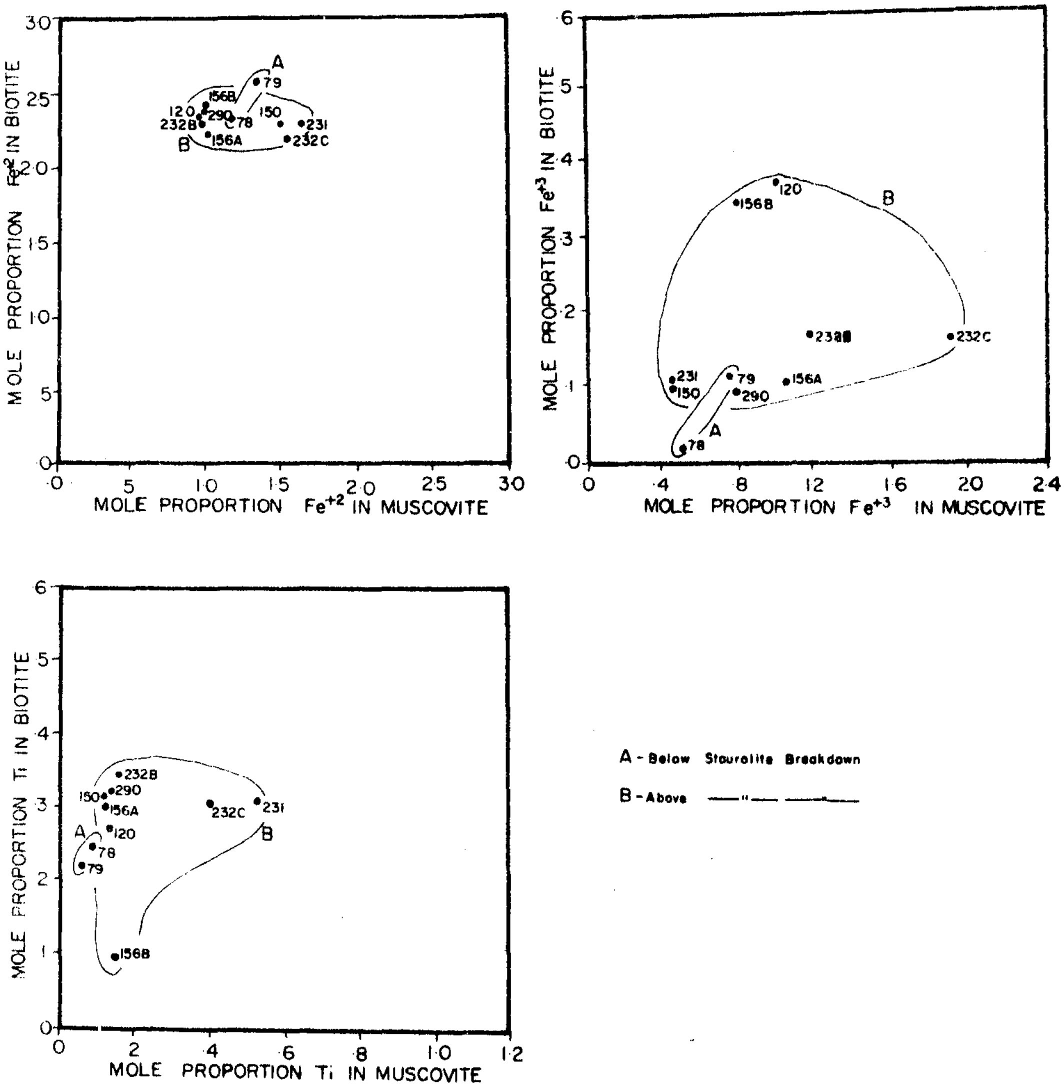


Figure 46.

Partition of Fe^{+2}, Fe^{+3} and Ti between biotite and muscovite.

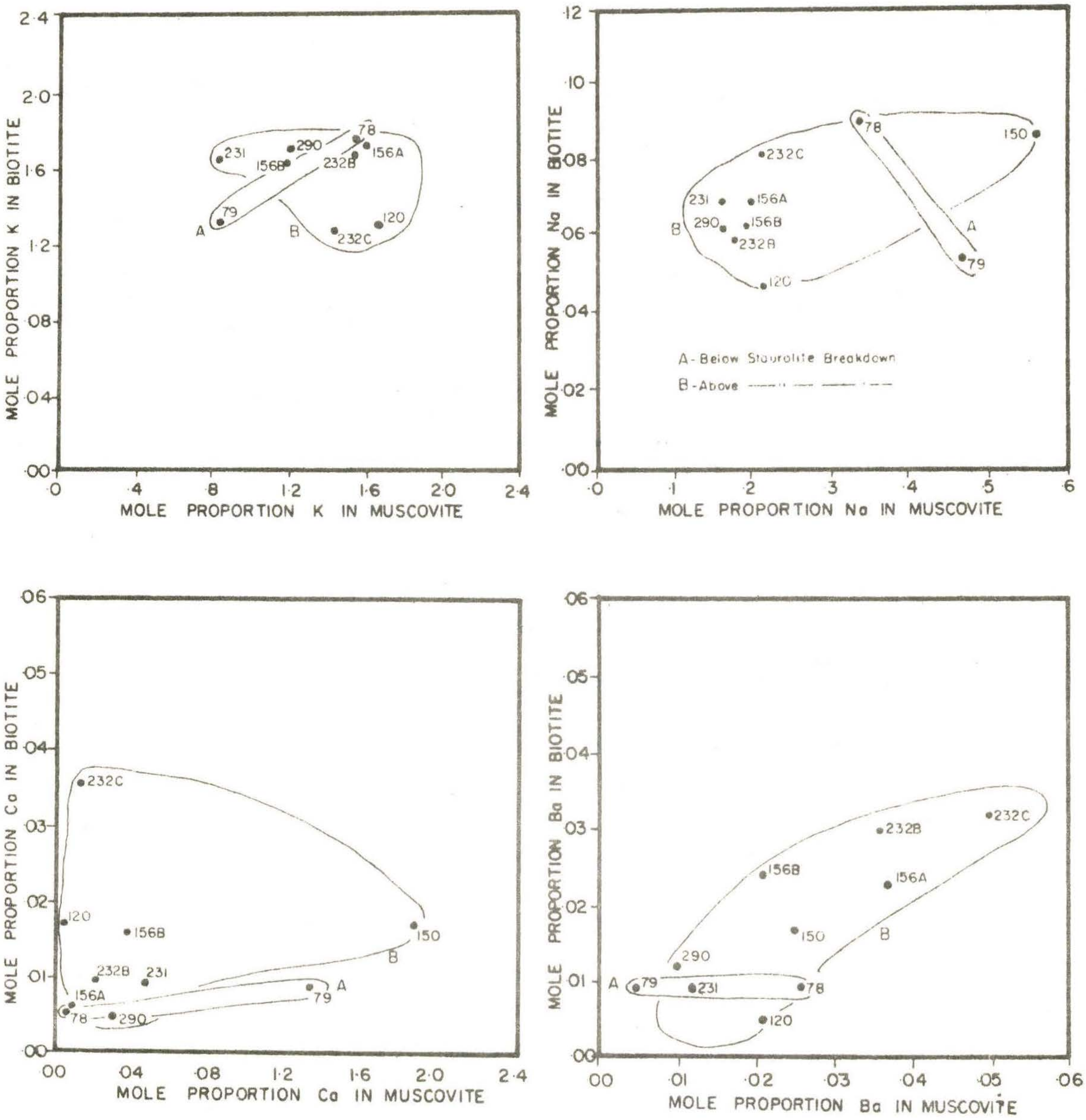


Figure 47.

Partition of K, Na, Ca and Ba between biotite and muscovite.

Ba is preferred in muscovite over biotite and the distribution trend of samples from above the staurolite isograd is suggestive of equilibrium. The D decrease of $D^{Ba}_{Bio./Musc.}$ with grade is inferred

Ca shows a great scatter of points. There is some suggestion of $D^{Ca}_{Bio./Musc.}$ increasing with metamorphic grade. Ca is favoured in biotite (figure 47).

Biotite-Garnet

The trace element partition between co-existing garnet and biotite at various grades have been determined by many authors (DeVore, 1955a and b; Lambert, 1959; Engel and Engel, 1960; Turekian and Phinney, 1962; and Albee, 1965a).

DeVore (1955a, b) believes that Ni, Co, Cu, Zr, Pb, V, Sc, La, V, Mn and Ti do not occur at lattice sites but on growth surfaces, in imperfections, dislocations, and various interfaces. His data are shown in figure 48.

Engel and Engel (1960) have plotted data for samples collected from the upper almandine amphibolite facies (figure 48).

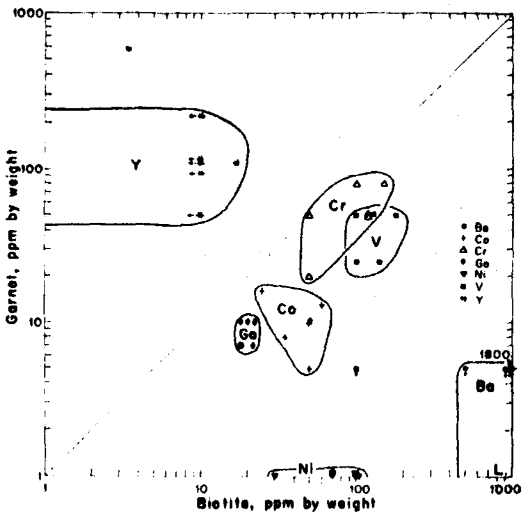
Lambert (1956) gives data from the almandine amphibolite facies (figure 48).

Kretz (1959) also gives data from the almandine amphibolite facies. Turekian and Phinney (1962) believe that samples of hand specimen size rarely show equilibrium partitions of trace elements in metamorphic sequences.

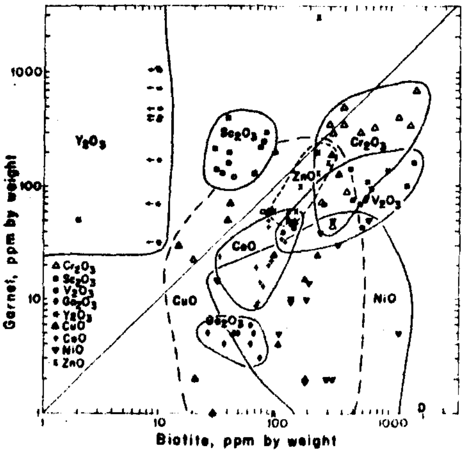
Albee (1965) believes that most point clusters for the various elements have a near 45 degree slope and are suggestive of equilibrium partition.

Figure 48.

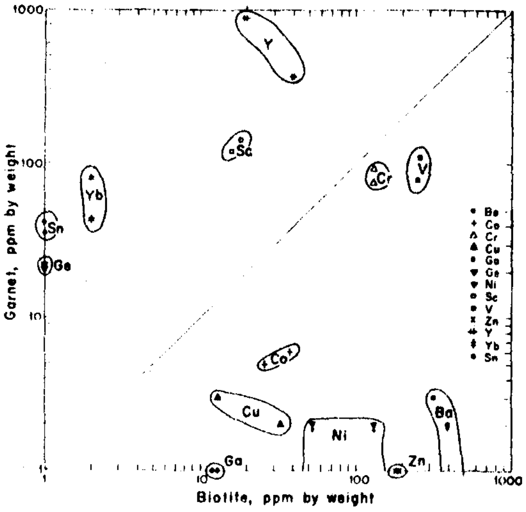
The partition of trace elements between biotite and garnet as determined by various authors.



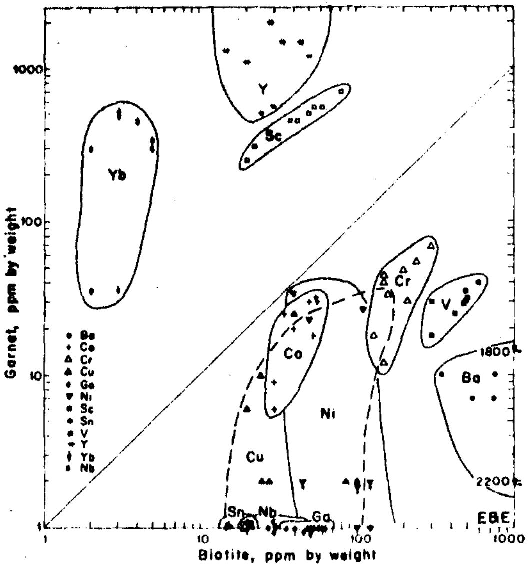
Distribution of minor elements between garnet and biotite (Lambert,



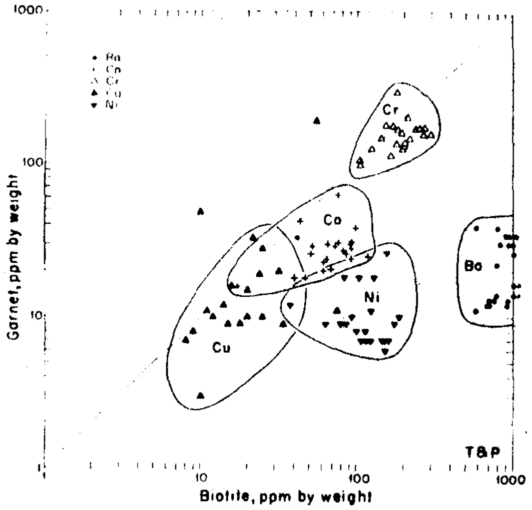
Distribution of minor elements between garnet and biotite (DeVore, 1955 a, b).



Distribution of minor elements between garnet and biotite.



Distribution of minor elements between garnet and biotite (Engel & Engel, 1960).



Distribution of minor elements between garnet and biotite (Turekian & Phinney, 1962).

Figure 48.

In the present study, below the muscovite isograd, it was inferred that Co, Cr, Ni, V, and Ga indicated equilibrium; Be and Li were indeterminate and Cu, Sc, Rb and Sr were rejected as showing equilibrium. Above the muscovite isograd Co, Sc, Y, Ni, Cr, V and Be were accepted as having equilibrium partitions; Sr is indeterminate and Rb, Ga, Li and Cu were rejected as showing equilibrium partition. Y, Sc, and Zr are concentrated in garnet while Rb, Be, Sr, Ga, Cu, Li, Ni, Cr, U and Co, are concentrated in biotite (figure 49).

The relations of major element partition between biotite and garnet have been inferred to be useful as geothermometers (Ramberg, 1952; Miyashiro, 1953, 1957; Kretz, 1959 and Frost, 1962). Albee (1965) has summarized the results of various workers and, on the basis of seventy garnet-biotite pairs from samples of known assemblages and metamorphic grade, reached certain conclusions. These are that the distribution coefficient of Mn, Fe and Mg depend upon:

1. the metamorphic grade;
2. the Mn content of garnet and
3. the particular assemblage present

The effect due to the prevailing intensive parameters (metamorphism) is predictable from thermodynamics (see section II, E, i). The reason the other two factors are important cannot be understood within the framework of thermodynamics.

The manganese content in garnets may be viewed as an extensive parameter and the value of the distribution coefficient should only be related to the intensive parameters. This argument can be circumvented if the Mn content

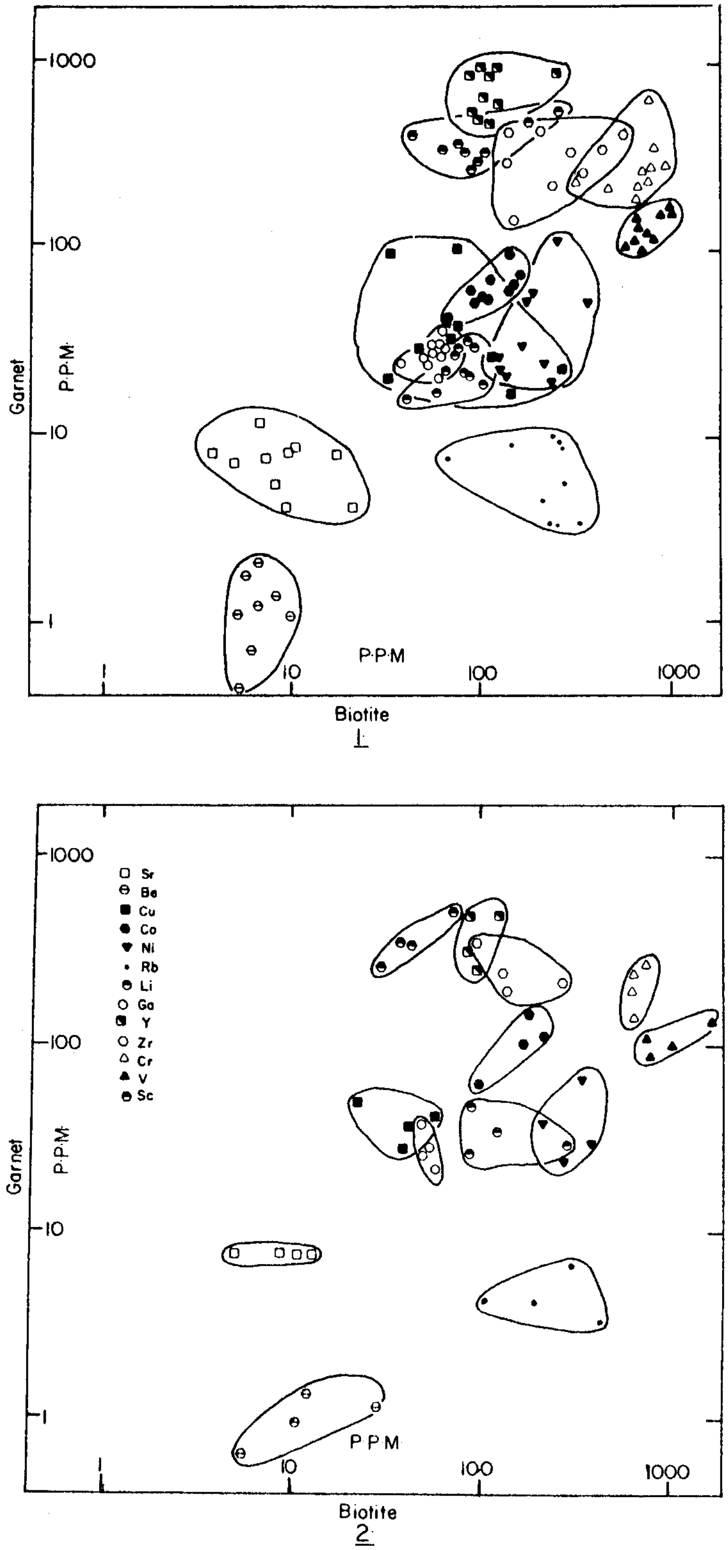


Figure 49.

Partition of trace elements between biotite and garnet from samples below (1) and above (2) the muscovite isograd.

of garnet is directly related to an intensive parameter such as pressure (as inferred by Yoder and Chinner, 1960, p. 81-84): however, this notion has been rejected in section II, E, iii, in the present study. Albee discusses the ratio Mn/Mn+Fe+Mg, assuming that these ions only interchange with each other and not with others at the same sites, an assumption that may well be incorrect. The effect of differing assemblages has been documented by Albee (1962, 1965) but no such effect was observed in this study.

The distribution of Mg between garnet and biotite is shown in figure 50. The three groups of points from each of the three subfacies present in the field, do not overlap and are elongate close to the 45° position. This feature is inferred to result from an equilibrium partition of Mg. In many studies (e.g. Hounslow and Moore, 1967), the points tend to fall on curved line but in the present study, a straight line is the best approximation.

The bottom of figure 50 shows the decrease of $D^{Mg}_{Bio./Gar.}$ with increasing grade. The points near the top of the zone marked by dashed lines, are interpreted to have partially exchanged Mg to a large extent while those points along the bottom of the zone are interpreted as having exchanged Mg to a lesser extent.

Mg is favoured in biotite at the conditions of the re-equilibrations (mentioned in section II, D, iv) while Mn is favoured in garnet.

The distribution coefficient of Mn ($D^{Mn}_{Bio./Gar.}$) between the two phases increases with increasing metamorphic grade (figure 51). Samples

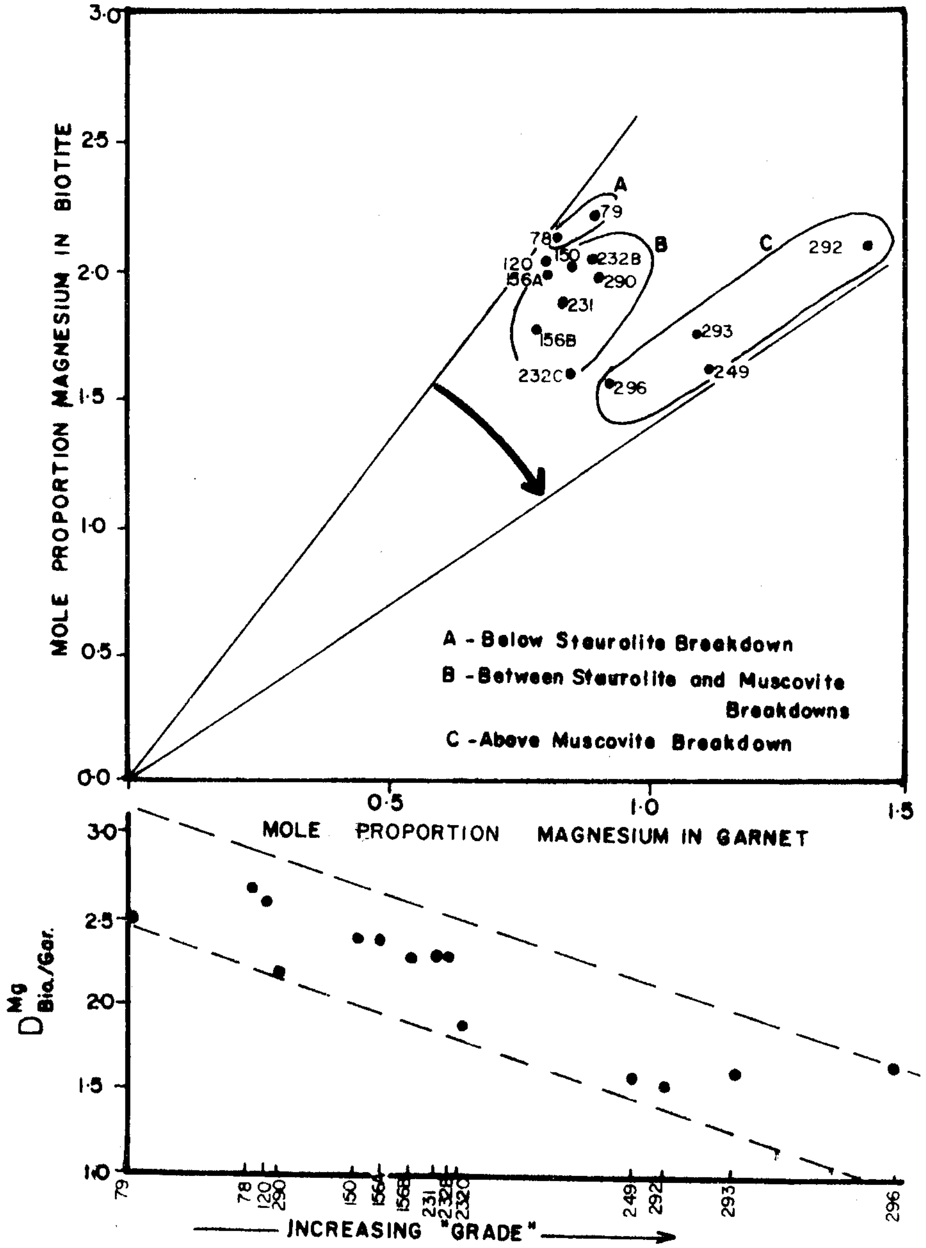


Figure 50.

The distribution of Mg between garnet and biotite. The dashed lines indicate the inferred trend.

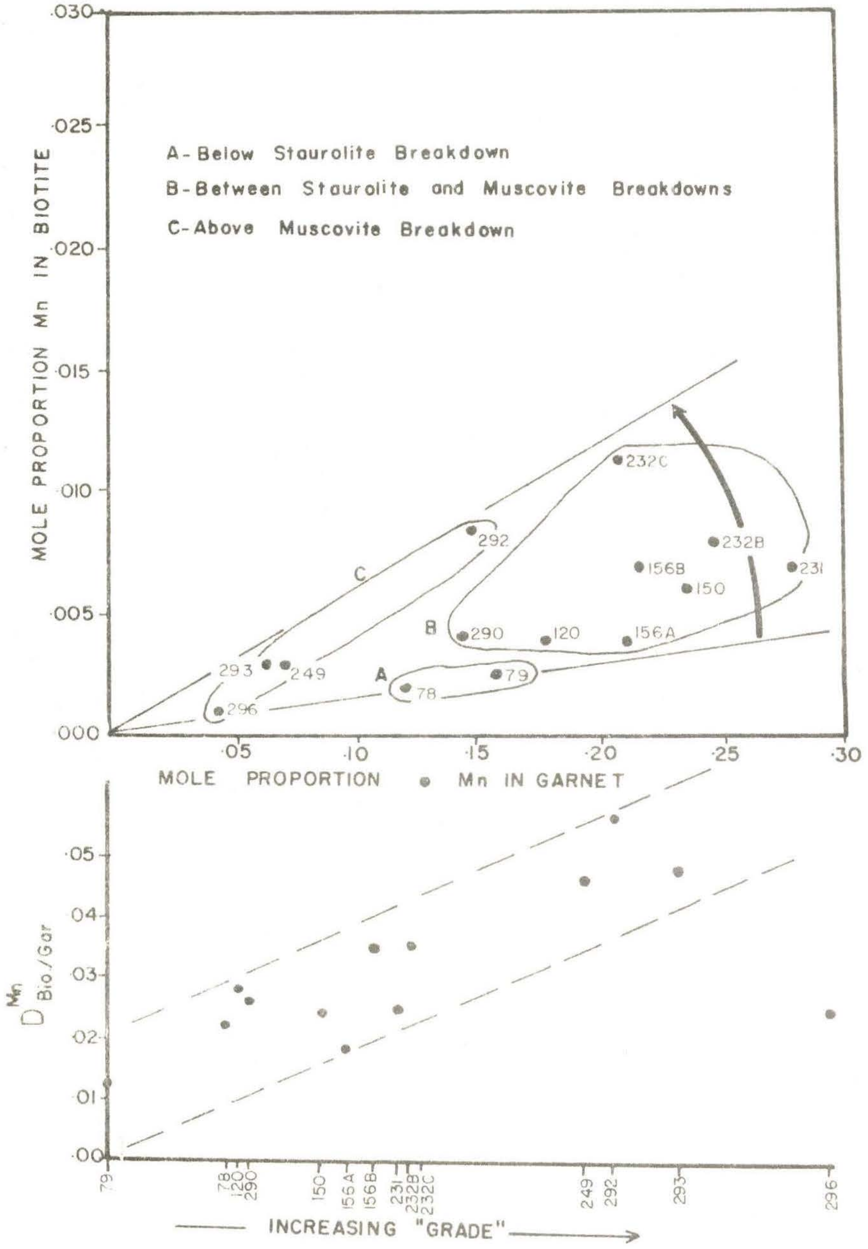


Figure 51.

The distribution of Mn between garnet and biotite. The dashed lines indicate the inferred trend.

T-296-1A and T-232-1B fall off the generalized trend due to the factors mentioned (p.) in section II, D, iv). These two samples are somewhat anomalous in most of the major element data.

Figure 52 shows the variation of Fe^{+2} for samples from the three subfacies present in the area studied. The distribution coefficient $D^{Fe^{+2}}_{Bio./Gar.}$ appears to decrease with metamorphic grade, which is similar to behaviour of the coefficient for Mg. Sample T-296-1A again plots off the generalized trend.

The distribution of Ca is shown in figure 53 and as can be seen, a fair amount of scatter exists. The variation of $D^{Ca}_{Bio./Gar.}$ shows an increase with metamorphic grade. Samples T-296-1A and T-232-1C again are anomalous on opposite sides of the generalized trend.

The fact that the change of the distribution coefficient is similar to those of Fe^{+2}, Mg and Mn would mean that Ca should also be included in an exchange reaction. This is not done by most authors (e.g. Kretz, 1960; Albee, 1965).

Summary

(i) Trace elements

Transition metal trace elements which include Mn, V, Ni, Cr, Co, Sc and Y, probably substitute for Fe^{+2} and Mg, and their partition has been inferred to indicate equilibrium especially between the minerals: muscovite, biotite and garnet. The alkali and alkali earth trace elements which include: K, Na, Rb, Be, Li, Ba and Sr do not generally show good equilibrium partition. Zr and Cu most commonly show highly scattered plots which are not indicative of equilibrium partition. This may in both cases

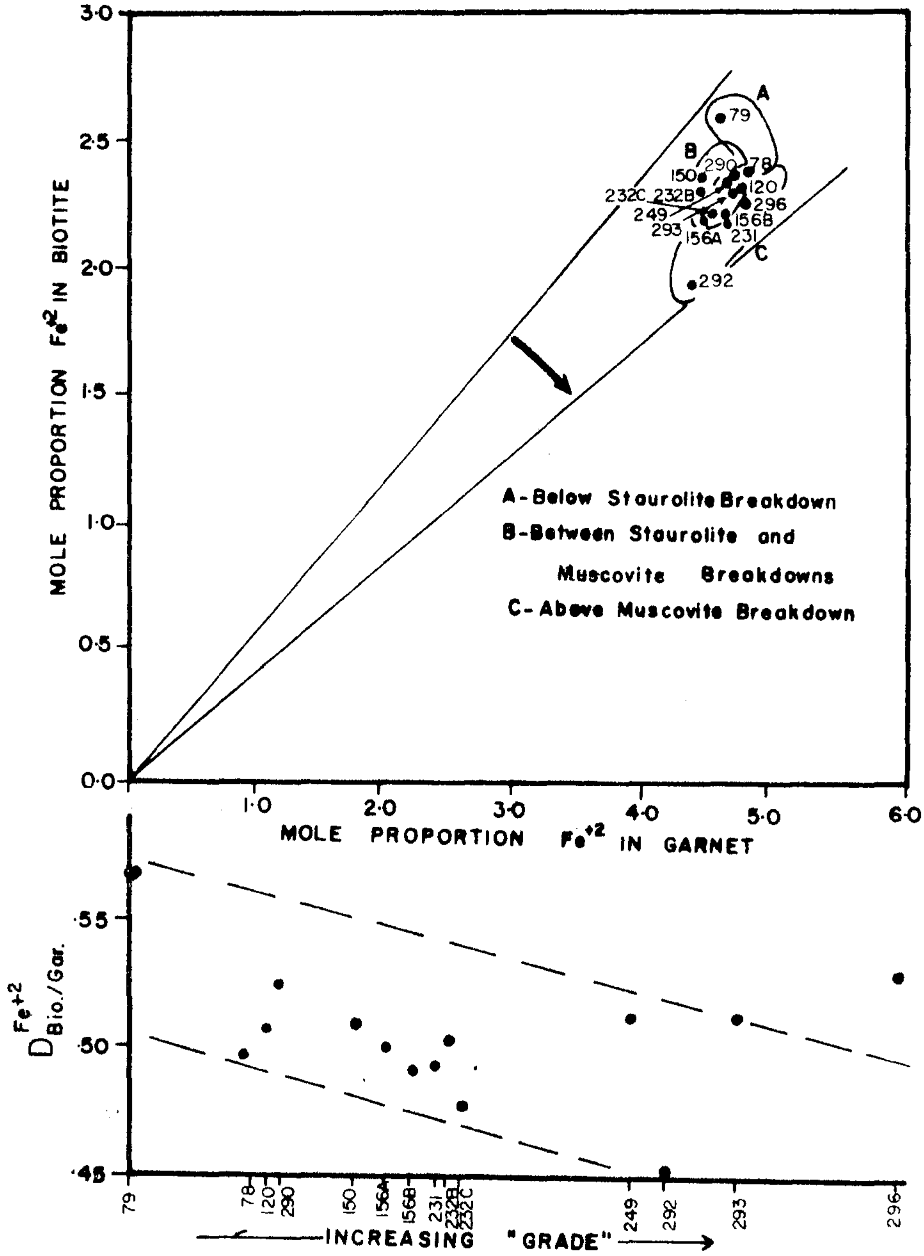


Figure 52.

The partition of Fe^{+2} between garnet and biotite. The dashed lines indicate the inferred trend.

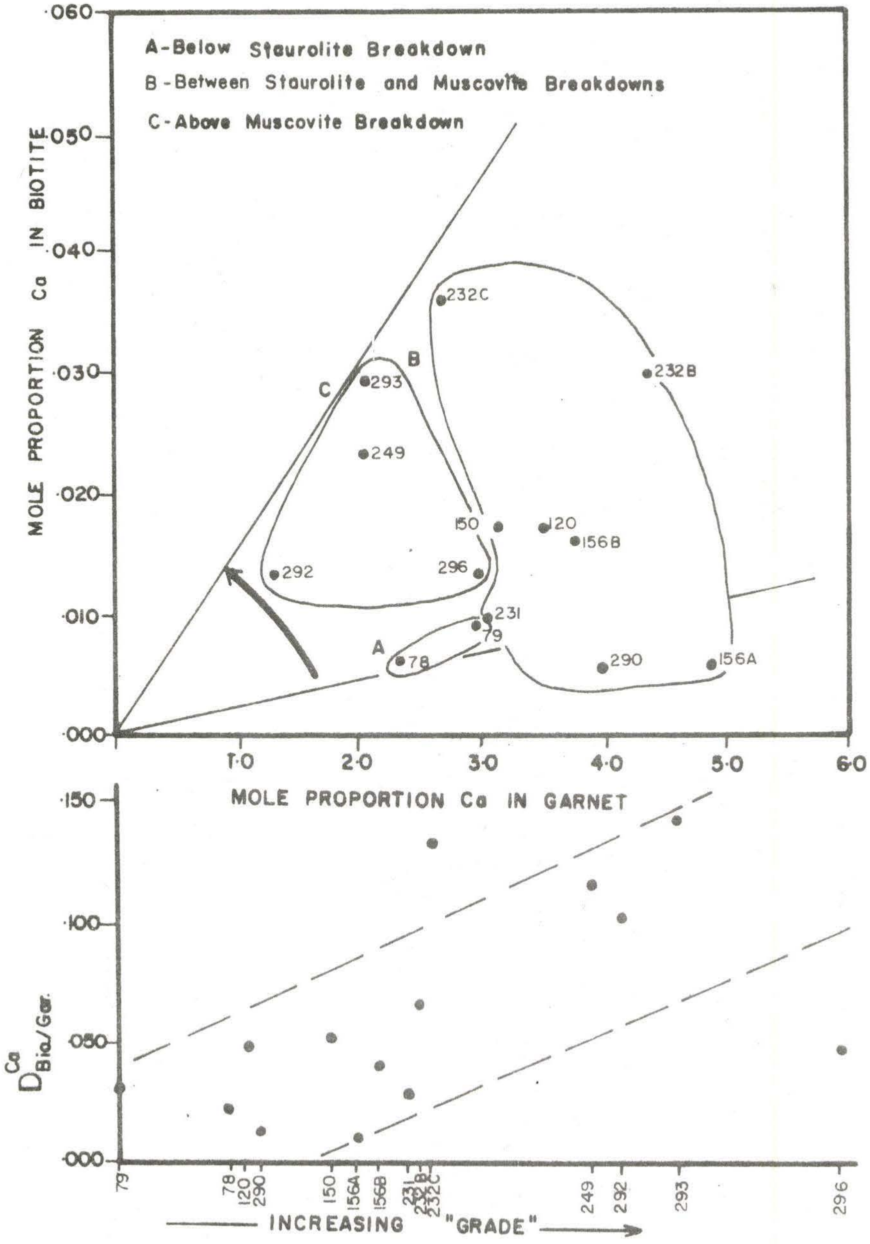


Figure 53.

The partition of Ca between biotite and garnet. The dashed lines indicate the inferred trend.

be caused by impurities of zircon and sulphide mineral (chalcopyrite ?) respectively.

The poor partition of the alkali and alkali earth metals may show that the latter group of elements are more exchangeable between the phases or the equilibrium systems may be very erratically distributed. The implications for such techniques as Rb-Sr dating are obvious. The transition elements may show good partition distributions because of their very stable 6 or 8 coordinated positions.

(ii) Major elements

The distributions of the elements Mg and Mn suggest systematic inverse relationships between garnet, biotite and muscovite which strongly suggests an exchange reaction between the phases at increasing grade. D^{Ca} and D^{Fe+2} also showed inverse relationships between biotite and garnet but it could not be shown between these minerals and muscovite.

It is inferred that if the partition coefficients change systematically with metamorphic grade, equilibrium has been maintained during that metamorphism and has been quenched in since that time. Partial re-equilibrium of ions to lower grade conditions are inferred to cause the major deviations from linear trends (explained in section III, D, iv).

As can be seen on most of the trends of distribution coefficients versus grade, no inflections occurred at any of the isograds across the section. Near the muscovite breakdown isograd the assemblage changes by the loss of muscovite and kyanite and the appearance of K-feldspar but no relation of this to the trends was shown.

The other major elements did not show any systematic changes of partition coefficients with metamorphic grade.

F. IMPLIED REACTIONS

When dealing with mineral assemblage-compositional changes with changing metamorphic grade, one should study all the phases involved. Possible metasomatic effects, together with the slight chemical differences from rock to rock, make it difficult to assign coefficients to formulae in metamorphic reactions. Two distinct types of partial reactions may be postulated which are:

(1) a reaction by which crystallographically new phases are produced from ones previously existing. This may or may not involve bulk chemical changes.

and

(2) chemical changes of phases previously present with small crystallographic changes.

Type (1) reactions

The following minerals were involved in this type of reaction:

(a) Staurolite

Two reactions have been proposed by various authors and compiled by Winkler (1967):

(1) 3 staurolite + 2 quartz → 1 almandine + 5 kyanite + 3 H_2O (p. 101).

and

(2) 1 staurolite + 1 muscovite + biotite + quartz → Al_2SiO_5 + Biotite$_2$ (richer in Fe^{+2}) + H_2O. (p. 101).

It is difficult to decide which seems the most reasonable but

Chinner (1965), prefers the latter for he states

> "Staurolite disappears by reaction with other phases long after it becomes incompatible with quartz."

Such a view may be true of some assemblages but in the present study staurolite persists as inclusions in garnet when it is no longer stable with respect to the rest of the assemblage.

Equation (1) is preferred because there is textural, modal abundance and chemical data that muscovite was stable and increasing in amount along the traverse, near its breakdown isograd.

Kyanite

Two types of reactions governed by the breakdown of this mineral. The first is the polymorphic inversion to sillimanite as

(3) kyanite → sillimanite

The only evidence of this inversion is near the kyanite isograd where sillimanite occurs as rims around kyanite. Below the muscovite isograd kyanite and sillimanite are, to a first approximation, chemically and structurally in equilibrium.

The second reaction, involving muscovite, is inferred from the excellent textural relations and general lack of equilibrium partition of trace elements between kyanite and muscovite.

The following form of reaction seems acceptable

(4) kyanite + biotite + SiO_2 (quartz) → muscovite + pyrope rich almandine (+ sillimanite ?).

There may be an excess of alumina on the right-hand side of the reaction which, with silica, forms sillimanite. Sillimanite needles have been found associated with altered biotite. The growth of muscovite in the rocks is puzzling in this part of the metamorphic sequence as the P_{H_2O} decreases (to be explained in section G) over the same interval at which the mineral increases in modal abundance. There is a possibility that the growth of muscovite is in response to an increasing partial pressure of a halogen bearing gas (F?). The muscovite breakdown is abrupt and the reaction involving it with kyanite is inferred to have been operative up to the muscovite breakdown.

Muscovite

The breakdown of this mineral described by the following reactions:

(5) 1 muscovite + 1 quartz → 1 orthoclase + 1 sillimanite + H_2O
(Winkler, 1967, p. 111)

(6) muscovite + quartz + Na rich plagioclase → sillimanite + Na bearing alkali feldspar + a more basic plagioclase + H_2O
(Guidotti, 1963)

and

(7) 1 muscovite + 1 biotite → 1 almandine + 2 K-feldspar + $2H_2O$
(Winkler, 1967, p. 111).

In this study the muscovite breakdown isograd is marked by the appearance of K-feldspar (5 to 10 modal percent) and a large increase in almandine modal content. Plagioclase increases in anorthitic content and a reaction of the following type seems reasonable.

(8) muscovite + quartz + biotite_1 + Na rich plagioclase + Ca bearing almandine → sillimanite + Na bearing alkali feldspar + a more basic plagioclase + almandine + biotite_2 + XH_2O.

This reaction is preferred although the other reactions are considered possible.

Type (2) reactions

Two examples of this type of reaction are possible in this study. These are:

(9) $\text{biotite}_1 + \text{garnet}_1 + \text{muscovite}_1 \xrightarrow{\text{inc. P.T.}} \text{biotite}_2 + \text{garnet}_2 + \text{muscovite}_2$

Numerical values can be entered by using the analytical data. Beyond the muscovite breakdown the following reaction occurs:

(10) $\text{biotite}_1 + \text{garnet}_1 \xrightarrow{\text{inc. P.T.}} \text{biotite}_2 + \text{garnet}_2$

It must be remembered that all other phases equilibrate with these. The continuous equations involve an exchange of Mn, Ca, Fe^{+2} and Mg.

The fact that different grain sizes of a mineral such as sillimanite occur in a section has been attributed by various authors (e.g. Rutland, 1965) to different periods of growth of the mineral due to different reactions. Such a view would suggest that during each reaction a nucleation of the new phases is effected and not simply a growth onto previously formed grains of the same phase. This interpretation may be texturally believable but kinetically is questionable as the energy needed to form the new nuclei may be much greater than to simply form on previously formed lattices. This should hold unless diffusional distances are too great. Two types do seem acceptable if one is isolated by, for example, inclusion in garnet.

Figure 54

Graph showing the extent of distance "implied" reactions were active over:

1. 3 staurolite + 2 quartz $\rightarrow$
   1 almandine + 5 kyanite (or sillimanite)
   + 3 H_2O.
3. kyanite $\rightarrow$ sillimanite.
4. kyanite + biotite + SiO_2 (quartz) $\rightleftharpoons$
   muscovite + almandine (+sillimanite ?).
8. muscovite + quartz + biotite + Na rich
   plagioclase + Ca bearing almandine $\rightarrow$ sillimanite
   + Na bearing alkali feldspar + a more basic
   plagioclase + almandine + $biotite_2$ + XH_2O.
9. biotite + $garnet_1$ + $muscovite_1$ $\rightarrow$ $biotite_2$ +
   $garnet_2$ + $muscovite_2$.
10. $biotite_1$ + $garnet_1$ $\rightarrow$ $biotite_2$ + $garnet_2$.

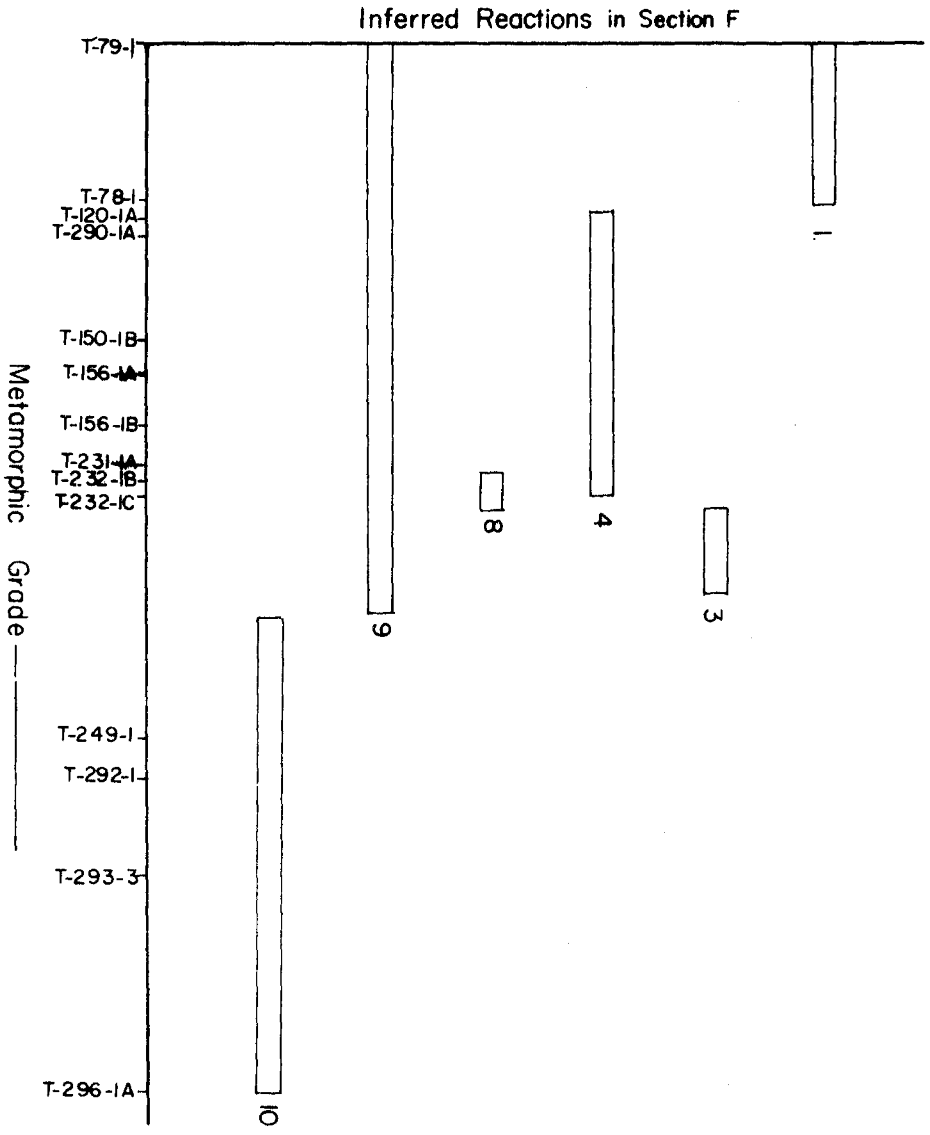

Inferred Reactions in Section F
T-79-1
T-78-1
T-120-1A
T-290-1A
T-150-1B
T-156-1A
T-156-1B
T-231-1A
T-232-1B
T-232-1C
T-249-1
T-292-1
T-293-3
T-296-1A
Metamorphic Grade
10
9
8
4
3
1

Figure 54.

Both garnet and sillimanite have formed in crystals of varying sizes although two or more periods of nucleation are not envisioned; rather a growth of a phase to large size in response to a higher supply in that region is preferred. This is inferred from the increase of the clear outer rim of the garnets with increasing metamorphic grade. The systematic changes of some of the distribution coefficient values and the trace element data suggest that most of the minerals are in equilibrium even though some have grown after others (e.g. sillimanite after kyanite, some garnet after kyanite). This can be explained by equilibration of elements after or during the growth period and points out that conclusions from chemical and relative growth period data are not necessarily compatible. The reactions which have been suggested are shown in figure 54. The textural evidence concerning kyanite and muscovite and the inferred identification of biotite as one of the reactants and garnet as one of the products poses a great problem. From the chemical data (section II, E, iv) one can infer that equilibrium partition of trace and some major elements has occurred between minerals which texturally appear to be reactants and products. This inconsistency could be explained by a period of production of crystallographically new phases at the expense of the reactants following by a period of re-equilibration of elements. The partition of the major elements, however, shows a fairly systematic change along the traverse which means the re-equilibration occurred in a varying P-T gradient which was most likely the same one expressed by the phases present. If the re-equilibration of elements occurred after the original crystallization, the P-T gradient must have been steep and near parallel to the gradient present during the original crystallization which is unlikely but possible. The re-equilibration may have occurred during the Grenville "orogeny" while the original crystallization occurred during the Hudsonian orogeny. The "re-equilibration" mentioned above must not be confused with the re-equilibration inferred from the grain inhomogeneity data.

It is unlikely that the kyanite-muscovite texture is a retrogressive feature as the values of $D^{Mg}_{Gar./Musc.}$ and $D^{Mg}_{Bio./Musc.}$ change systematically with metamorphic grade. If the feature is retrogressive, it must have formed before the period of re-equilibration of elements mentioned above. Chinner (1966, p. 168) considers of a similar texture involving andalusite instead of kyanite that muscovite is a "late" phase and implies it is not retrogressive.

G. DISCUSSION OF KYANITE AND SILLIMANITE RELATIONS

Two aspects of co-existing phases are important when considering equilibrium which are:

(1) crystallographic factors

(2) chemical factors

Presumably a change of the chemistry of a phase will also have crystallographic changes but these are usually very small.

To test whether two phases are crystallographically at equilibrium one either produces them together in the laboratory or finds them commonly associated without mutual textures suggesting one is reacting to form the other.

The co-existing kyanite and sillimanite in this study had mutual textures suggesting stability up to the muscovite isograd after which kyanite altered to sillimanite. From this it is inferred that up to the muscovite isograd the two minerals are crystallographically at equilibrium.

To test whether two phases are chemically in equilibrium involves inferences from thermodynamic postulates like the Neunst distribution "Law". The distribution of trace elements in kyanite and sillimanite was measured and, although many possible limitations can be thought of, the distributions suggest chemical equilibrium has been attained for most of the elements measured.

Thus, applying the only tests usable, a conclusion that kyanite and sillimanite are in equilibrium is inevitable. If one uses precidence as a judgement, it becomes apparent that co-existing kyanite and sillimanite are fairly commonly described in the literature. The frequency of the association being described is undoubtedly out of proportion to the areal proportion of rocks which are of high enough metamorphic grade to contain the two polymorphs. However, the occurrence of the two minerals is probably too frequent to simply be accounted for by fortuitous occurrences of conditions at the kyanite-sillimanite univariant being attained in the earth's crust. We are thus left with kinetic arguments as to why the two polymorphs are so common; arguments involving other components having an effect on the stability fields; that conditions of the kyanite-sillimanite univariant are common thermobaric regimes in the earth crust; or, some combination of these factors. From experimental work (e.g. Bell, 1963) it becomes apparent that the formation of kyanite and sillimanite in experimental studies are sluggish so that this factor may be significant. There is no known relation of other components such as a TiO_2 or Fe_2O_3 on the stability of the polymorphs. Undoubtedly there will be some effect but the low abundances such as those described in this study (see appendix II) would probably rule this out to a first approximation.

Chinner, 1966, believes it to be a "difficult" if not impossible question to answer (p. 122) whether the occurrence of sillimanite indicates the attainment of pressure/temperature conditions corresponding to the divariant sillimanite field. He sites the evidence of where coarse grained muscovite occur replacing kyanite and not sillimanite (p. 120, figure 4,B in Chinner, 1966) that sillimanite is stable relative to the other polymorph and conditions were those of the sillimanite stability field.

Most authors believe that when the different Al_2SiO_5 polymorphs occur in the same rock, the temperature and pressure were fluctuating near the invariant point or divariant line. Hietanen (1956) believes that the occurrence of kyanite, sillimanite and andalusite in specimens from Boehls Butte quadrantle, Idaho indicated fluctuating conditions near the invariant point. Woodland (1963) believed of a similar situation in northeastern Vermont that "a kyanite-sillimanite-andalusite-bearing assemblage suggests that these polymorphs were probably formed under closely similar conditions during thermal metamorphism but not necessarily simultaneous".

The view taken in the present study is that the conditions were also near the kyanite-sillimanite univariant.

H INTENSIVE PARAMETERS

Linearity of Metamorphic Grade

A knowledge of the profile of the change of metamorphic grade is necessary to understand the mineralogical changes observed across the area. The change to successively higher grade subfacies as one approaches the southeast of the area has been explained in Chapter I and one can infer that the major factors which control the mineralogy increase in that direction. Pressure is inferred to have increased because assemblages stable only at high pressures were observed in the southeast of the area (kyanite-orthoclase). From the degree of element partition change and the breakdown isograds of various minerals one can infer the temperature also increased.

The changes in P and T are inferred to have been linear from the following evidence:

(a) the equilibrium relations between kyanite and sillimanite implies the metamorphic gradient is close to or at the kyanite-sillimanite univariant curve. This univariant curve is nearly linear as is discussed in the next section.

Various other features are linear also although no theoretical basis for the linearity is known.

(b) The linearity of the kyanite and sillimanite and the near linearity of the staurolite trend surface modal abundance trends (section II, C).

(c) The linear change of Ti in biotite and muscovite across the area (section II, E, iii).

(d) The linear change of partition coefficients of some elements in some phases across the area (section II, E, v).

The inference from the kyanite-sillimanite equilibrium relationsip is the best evidence for the linear profile model.

Intensive Parameters

The importance of knowing the past intensive parameters under which a particular assemblage was stable has been mentioned in section III, E, i and will now be discussed.

(a) Temperature and Pressure

Past temperatures and pressures are difficult to determine in metamorphic rocks as this demands a complete knowledge of the formation of the indicators used - the metamorphic minerals.

Many geothermometers have been suggested and used but few are calibrated sufficiently to give accurate results. Three types are as follows:

(a) isotopic fractionation (e.g. oxygen isotopes fractionated between quartz and magnetite);

(b) the substitution of an excess component into a phase (e.g. the calcite-dolomite geothermometer);

(c) the breakdown and first occurrence of mineral phases in the field (e.g. biotite isograd).

Type (a) was not used in the present study. Type (b) was tried with respect to the possible muscovite-paragonite and calcite-dolomite geothermometers. Ten muscovites were analyzed by X-ray diffraction but no indication of the presence of paragonite was detected. No dolomite could be found associated with the calcite in hand specimens of marble. Type (c) was used. Most of the isograds in the area studied are of the breakdown type which according to Atherton (1965, p. 172) should be most representative because "isograds" (of this type) "do not involve nucleation which according to some metamorphic geologists is the most important single factor in mineral development". This statement is acceptable if no new phases have been nucleated, but usually the loss of one phase is concurrent with the formation of a new phase in metamorphic rocks.

The determination of temperature by use of type (c) geothermometers needs a knowledge of the conditions of synthesis of the minerals in question. These conditions are determined in the laboratory and inferences from them are subject to some of the following restrictions:

(a) ideal pure compositions have been used, to date, although these seldom, if ever, occur in nature;

(b) the assumption that total pressure equals fluid pressure may be true in the laboratory but may not be usable in the fields;

(c) the time for reaction in the laboratory is often too small to test equilibrium convincingly;

(d) because of the mineral grain sizes often involved, chemical analyses and identification of phases are very prone to analytical error.

Regardless of these possible limitations, implications from laboratory studies are still the <u>only</u> means of suggesting past temperature and presure values.

The intersections of two univariant lines on a P-T diagram determine the values of these intensive parameters at that intersection and can be related to field studies if the univariant lines can be shown to relate to field isograds.

In this study, the intersections of the staurolite and muscovite breakdown curves with the kyanite-sillimanite univariant line were the means of determining the temperatures and total pressures. From the equilibrium partition of trace elements (section II, E, iv) and textural implications between the two minerals (section II, B) one can infer that up to the muscovite breakdown isograd, kyanite and sillimanite coexist at equilibrium. The extrapolation beyond the kyanite breakdown isograd is on the basis of the continuing linear increase of Ti in the phyllosilicates and distribution coefficients.

The univariant lines have been determined by various workers and will now be discussed.

The aluminum silicate system has received much attention mainly because of the apparent property of being sensitive only to temperature and total pressure. There are a large number of laboratory and field determinations of the position of the phase areas on a P-T plot (Bell, 1963; Aramaki and Roy, 1954, 1962; Miyashiro, 1949; Kennedy, 1956; Clark et al, 1957; Clark, 1961; Khitarov et al, 1963; Walbaum, 1965;

Newton, 1966; and Holm and Kleppa, 1966). To determine which values to accept is extremely difficult. Some of the determinations are suspect because the stability of a phase was determined by the growth of chemical onto nucleii of that phase.

The minimum temperature and pressure must have been greater or equal to the P-T conditons at the Al_2SiO_5 invariant point as can be inferred from the co-existence of kyanite and sillimanite and the total lack of andalusite. The apparent equilibrium co-existence of kyanite and sillimanite and the increasing metamorphic grade would suggest the thermobaric gradient must be fluctuating near or be on the linear kyanite-sillimanite univariant curve. An exhaustive discussion of the numerous experimental determinations of phase areas in the Al_2SiO_5 system is not in the scope of the present study and the compellations of Fyfe (1967), Hietanen (1967) and Winkler (1967) will be used. Fyfe discussed the experimental results of most of the existing work and concluded a triple point near 475°C and 2.4 kilobars. Hietanen and Winkler assumed on the basis of a few experimental studies and geological implications, a triple point of approximately 475°C and 5 kilobars and 575°C and 6.5 kilobars respectively. The $dT_{/dP}$ data for the kyanite-sillimanite univariant curve is near $50°C/_{kb}$ for all three authors.

Richardson (1966) has experimentally produced ferroan staurolite at greater than 750°C and 2 kilobars P_{H_2O}. Hoschek (1967) has produced staurolite at 545°C and between 4 and 7 kilobars P_{H_2O} and finds the mineral still stable at 650°C. Hietanen (1967) has suggested a staurolite breakdown at near 500°C and has inferred it to be essentially pressure independent in the range of 4 to 7 kilobars total pressure.

Velde (1966) has shown that the muscovite breakdown curve is essentially linear from 685°C at 2 kilobars to 725°C at 8 kilobars P_{H_2O} which closely corresponds to the experimental data of a number of other writers. Hietanen (1967), on a compellation from various authors, shows a breakdown temperature of near 600°C and a near constant pressure variation. If Richardson's and Hoschek's data are used, the staurolite breakdown indicates a temperature which is probably greater than the muscovite breakdown temperature. The muscovite breakdown isograd, however, occurs at a substantially higher metamorphic grade (see figure 1). This anomalous behaviour may be due to the fact that $P_{Total} \neq P_{H_2O}$ or the effect of impurities present in either of the two minerals although the explanation preferred is that the experimental determination of staurolite stability are too high.

Using Hietanen's and Fyfe's recommendations regarding the Al_2SiO_5 system to be used and Hietanen's staurolite and muscovite breakdown data, a temperature gradient of approximately 475 to 725°C may be adopted. The pressure difference would be 4.8 to 7.5 kilobars for Hietanen's data and approximately 3 to 5.5 kilobars for Fyfe's data. If Velde's muscovite breakdown is used, the temperatures would be 475 to approximately 900°C which is geologically unreasonable. The pressures would be 4.8 to 11

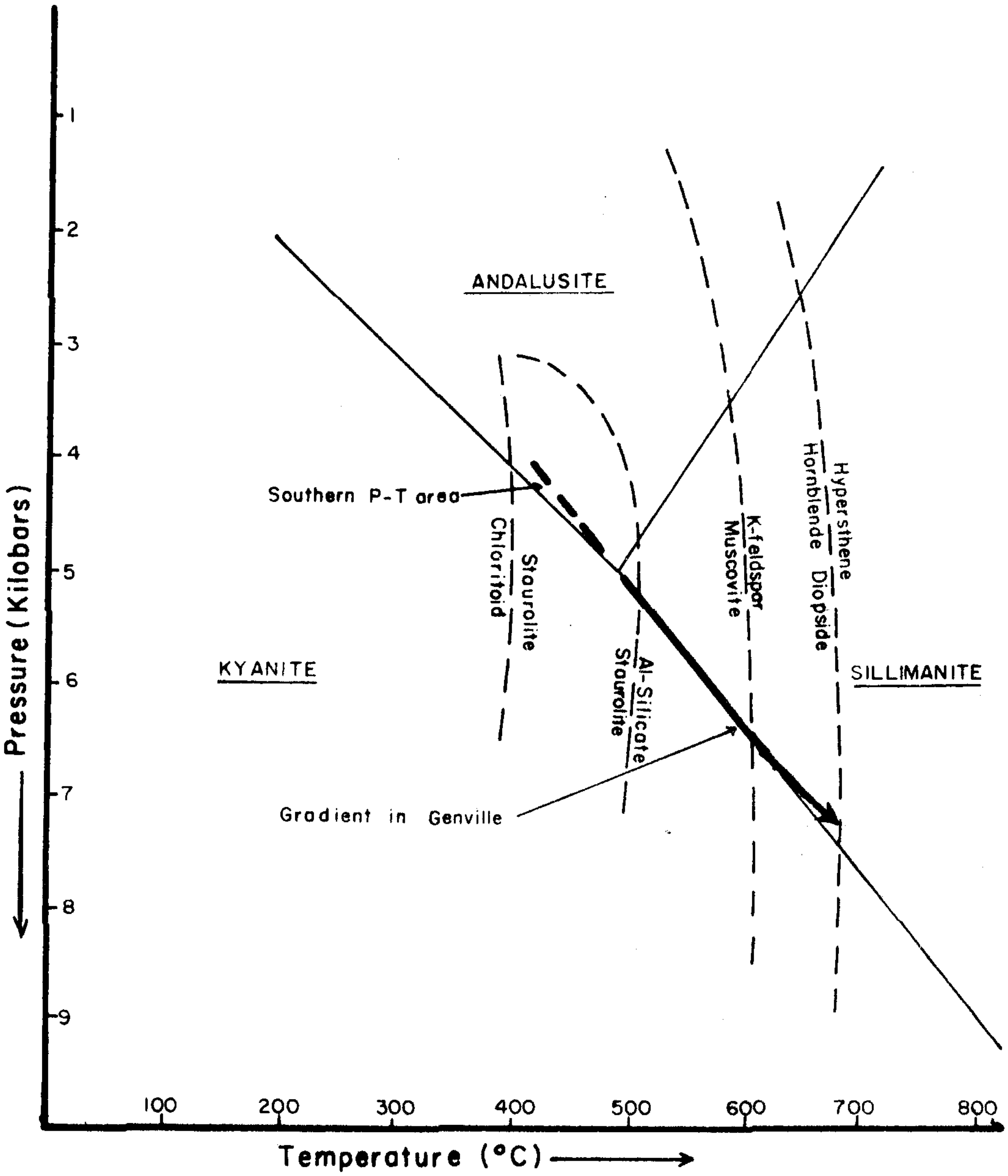


Figure 55.

Parts of petrogenetic grid of Hietanen (1967). The heavy dark arrow is the inferred P-T gradient in the Grenville and Southern provinces in this study.

kilobars using Hietanen's data and 3 to approximately 7.5 kilobars using Fyfe's data. Winkler's recommendations concerning the Al_2SiO_5 system can only be used with Hoschek's staurolite data and Velde's muscovite data. The temperature in this case would be at least 650°C to 750°C near the beginning of the traverse. The pressure differential across the area would probably be small.

It can be shown that the height of a column of rock equal to the length of the traverse would exert a pressure almost exactly equal to the pressure differential using Fyfe's and Hietanen's data but not Velde's and probably not Winkler's.

Inferences from Hietanen's compellation are used fór the Po_2, f_{H_2} and P_{H_2O} data.

It is of interest that if the gradient, which is inferred to be very close or on the kyanite-sillimanite univariant, is projected to lower P-T conditions, the field of andalusite-staurolite is intersected. This assemblage is present in the Southern province pelitic rocks near the boundary between the two provinces. Only a very small pressure differential would be needed (0.5 kilobars ?) which may suggest a displacement of a few kilometers on the mylonite zone, (see figure 55).

The inferred temperatures and pressures in the Grenville province are shown in figure 56.

Figure 56.

Variation of temperature, pressure and water pressure across the area studied. The water pressure determinations will be explained later in this section.

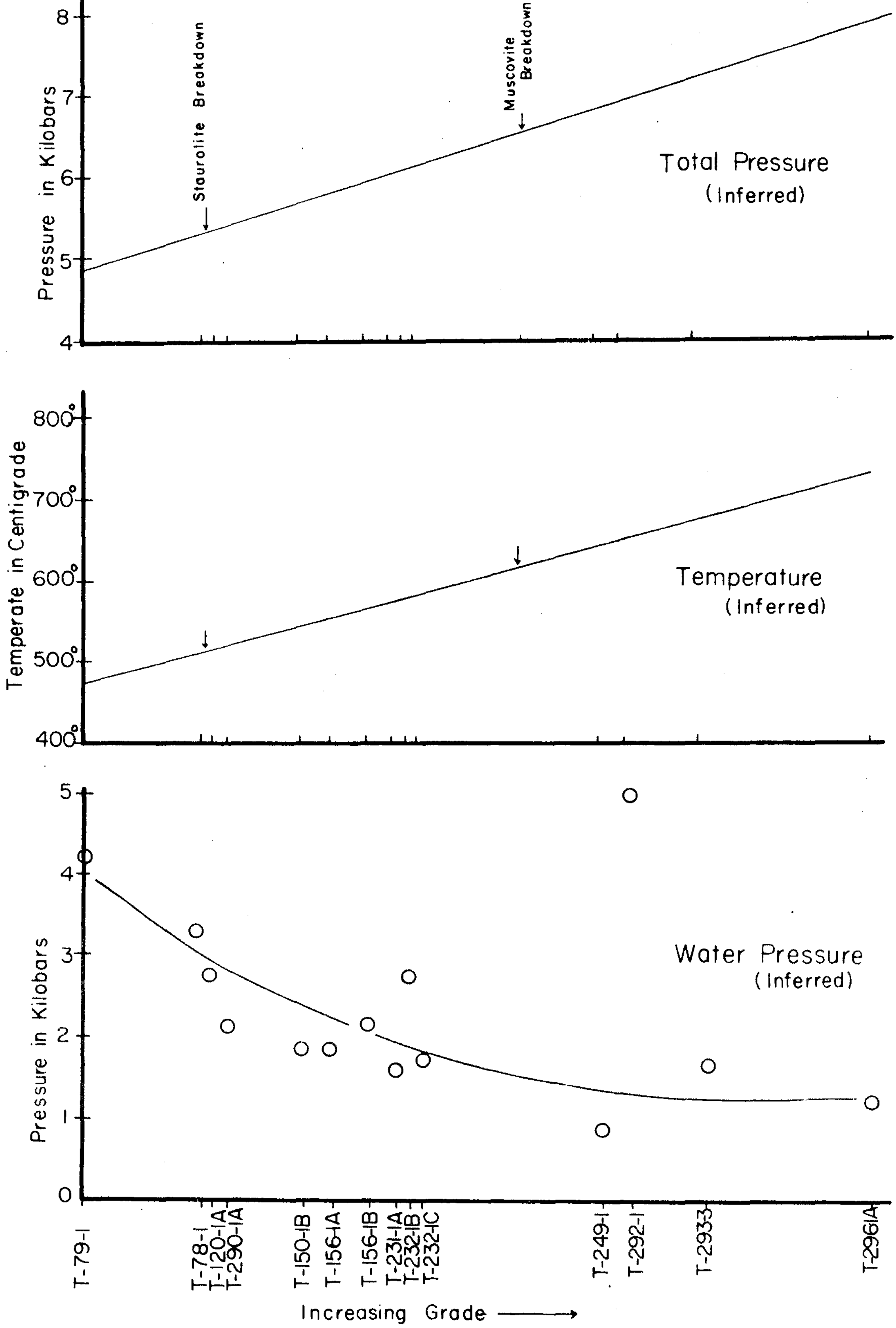

Pressure in Kilobars
8
7
6
5
4
Staurolite Breakdown
Muscovite Breakdown
Total Pressure
(Inferred)
Temperate in Centigrade
800°
700°
600°
500°
400°
Temperature
(Inferred)
Pressure in Kilobars
5
4
3
2
1
0
Water Pressure
(Inferred)
T-79-1
T-78-1
T-120-1A
T-290-1A
T-150-1B
T-156-1A
T-156-1B
T-231-1A
T-232-1B
T-232-1C
T-249-1
T-292-1
T-293-3
T-2961A
Increasing Grade

Figure 56.

(b) Partial vapour pressures and fugacities

It is not known whether the total vapour pressure present in rocks during mineral crystallization is equal to the total pressure but this has been assumed by some writers (e.g. Buddington and Lindsley, 1964, p.). This assumption is not used in the present study.

Some of the partial vapour pressures may have been small (e.g. P_{O_2}) but their effects on the subsequent mineralogy were undoubtedly significant and their determination is essential to a study of this kind. As these are past partial pressures, one can only determine their values by the effect they had on the prevailing mineralogy from laboratory studies. Few of these studies exist to date.

Another problem is that the exact species of gases present in metamorphic rocks during metamorphism are not known. In this study, gases or fluids containing S, P, halogens, O, H, H_2O are inferred.

Sulphur

Pyrite, pyrrhotite and chalcopyrite were observed in some of the polished thin sections and probably formed under a partial pressure of S_2 gas. Toulmin and Barton (1964) have shown that if the assemblage pyrite-pyrrhotite exists and the temperature is known, one can estimate a partial fugacity of S_2 (figure 57). Sample T-78-1 and T-150-1B contained both pyrite and pyrrhotite as well as chalcopyrite. These two samples allow one to infer a fugacity of 10^{-4} and 10^{-3} atmospheres using temperature estimates of approximately 500°C and 550°C respectively. Samples T-292-1 and T-156-1A contain pyrite which means

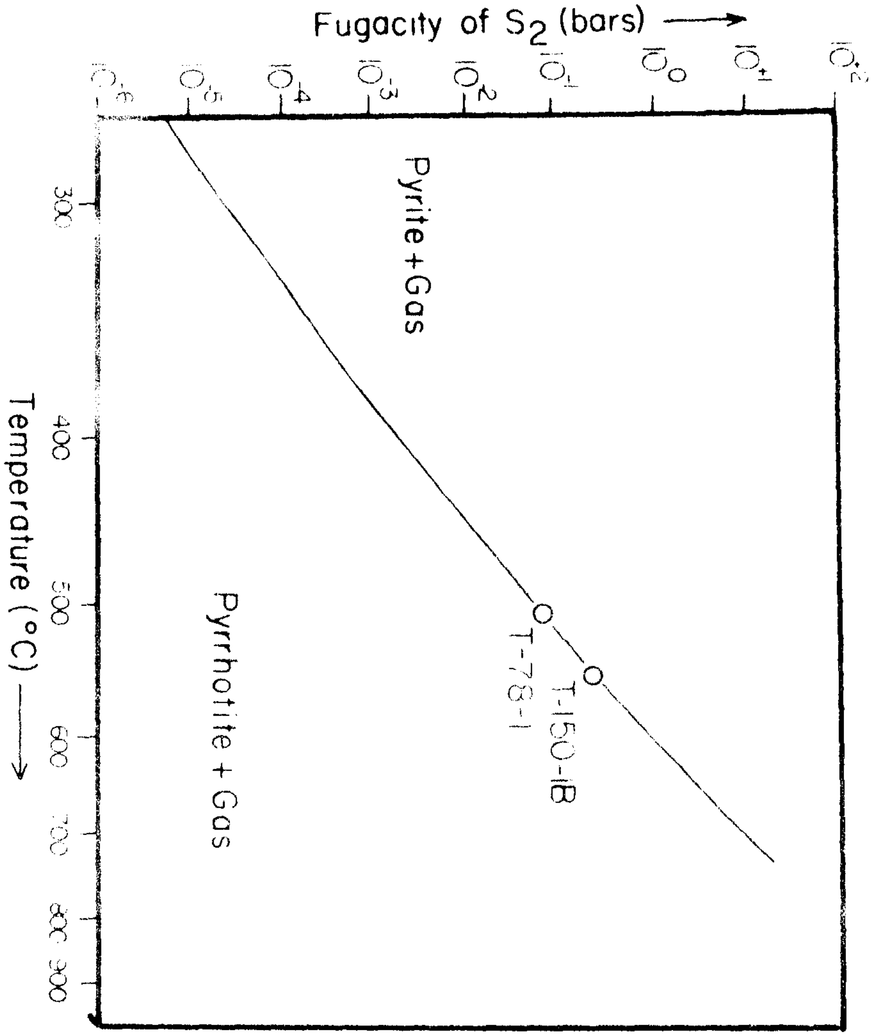


Figure 57.

Stability fields of pyrite and pyrrhotile in F_{S_2} - temperature space after Toulmin and Barton (1964).

their S_2 fugacities have a minimum value of $10^{+.3}$ and $10^{-2.5}$ using appropriate temperatures. If these four samples are indicative of the other samples with respect to f_{S_2}, then the increase of f_{S_2} across the area is as shown in figure 58.

(b) Phosphorous bearing species

Little is known about the effect of a phosphorous bearing species on the products of a metamorphic episode. The P_2O_5 content of the rocks in this study is fairly uniform which may mean the P_P species was also fairly uniform.

(c) Halogen bearing species

There were probably partial pressures of halogen bearing species but no evidence for them is known except the inferred halogen contents of the phyllosilicates.

(d) Fugacity of oxygen

Two ways of determining partial oxygen fugacities have been proposed:

1. by use of Fe-Ti oxides (Buddington and Lindsley, 1964)

and 2. by use of biotite (Wones and Eugster, 1965).

To use the former the assemblage must indicate a phase on the magnetite-ulvöspinel join and another on the hematite-ilmenite join in the Fe-Ti-O ternary system. These requirements were not met in the present study.

To use biotite as an oxygen barometer, the following requirements are necessary and are met in this study:

Figure 58

Variation of f_{S_2}, f_{H_2} and f_{O_2} across the metamorphic interval.

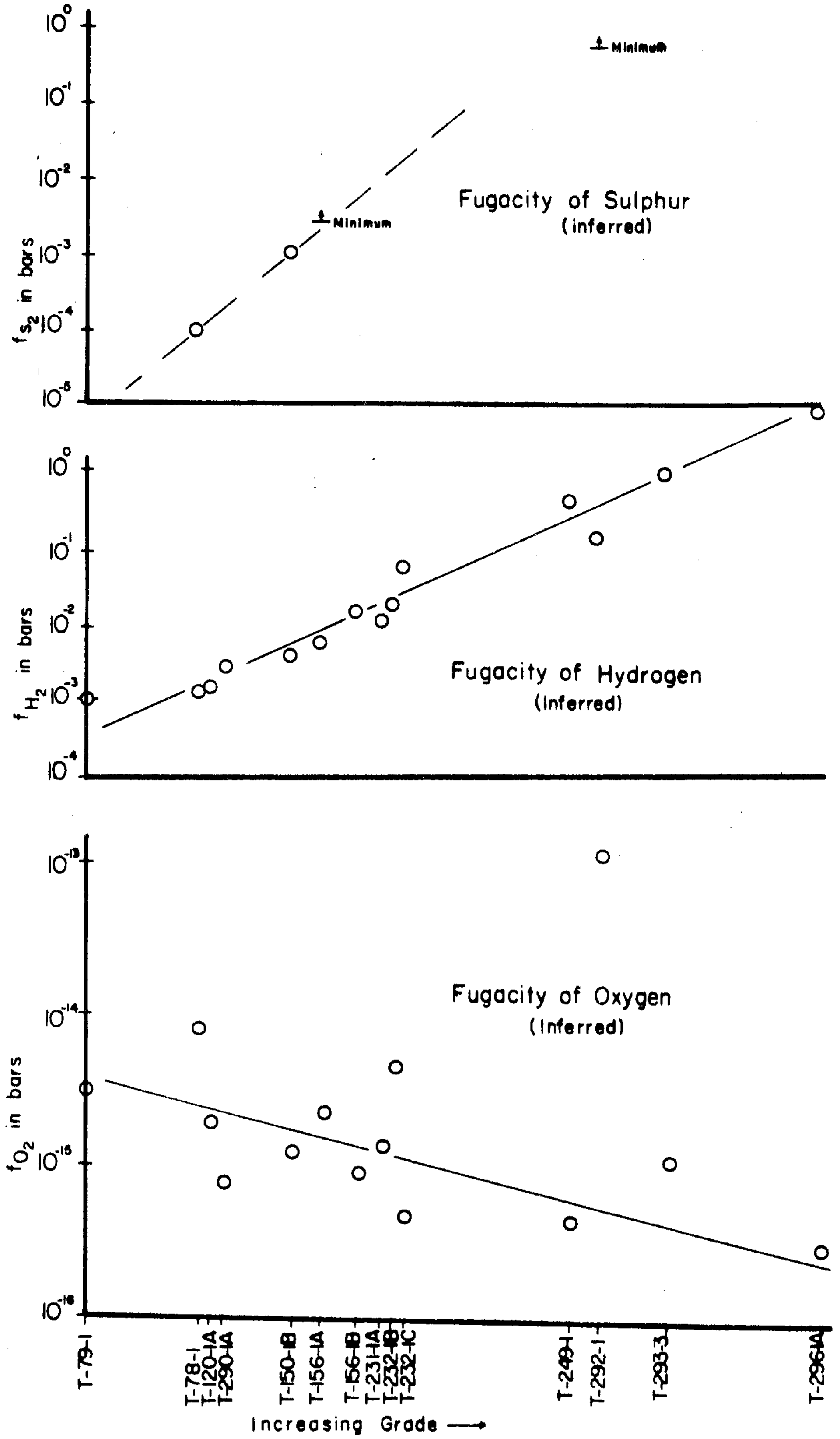

10^{0}
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
f_{S_2} in bars
Minimum
Minimum
Fugacity of Sulphur
(inferred)
10^{0}
10^{-1}
10^{-2}
10^{-3}
10^{-4}
f_{H_2} in bars
Fugacity of Hydrogen
(inferred)
10^{-13}
10^{-14}
10^{-15}
10^{-16}
f_{O_2} in bars
Fugacity of Oxygen
(inferred)
T-79-1
T-78-1
T-120-1A
T-290-1A
T-150-1B
T-156-1A
T-156-1B
T-231-1A
T-232-1B
T-232-1C
T-249-1
T-292-1
T-293-3
T-296-1A
Increasing Grade ⟶

Figure 58.

(1) biotite compositions lie on the annite-phlogopite join;

(2) the assemblage includes: biotite-K-feldspar-magnetite; biotite-muscovite-corundum (or aluminum silicate-quartz)-magnetite.

(3) an independent determination of temperature;

(4) the Fe^{+2}/Fe^{+2} + Mg ratio of the biotites.

Figure 59 shows the relations of temperature, oxygen fugacities and Fe^{+2}/Fe^{+2} + Mg ratios of synthetic biotites. The Fe^{+2}/Fe^{+2} + Mg ratios are given in table 2. Samples occurring metamorphically below the breakdown of muscovite lie outside the stability field of magnetite and in that of hematite although only magnetite occurs in the hand specimens. This anomalous situation could be due to:

(1) error in the experimental determinations of the magnetite stability field;

(2) the effect of impurities in minerals;

(3) analytical error in determining the biotite compositions in this study.

The effect of impurities is strongly emphasized by Wones and Eugster (1965).

M. J. Rutherford (1967, personal communication) is working on annites of various alumina contents and he believes the phase boundary of the magnetite-biotite-sanidine-gas field may be advanced to higher oxygen fugacities and decreased temperatures by as much as 60°C if the soda con-

Figure 59.

Variations of log oxygen fugacity versus temperature after Wones and Eugster (1965). The samples are those in this study and the $Fe^{+2}/Fe^{+2} + Mg$ ratios are shown in table 2. The samples enclosed within the rectangles contained graphite and the other ones did not. The graphite breakdown curve is from French and Eugster (1965). The heavy dark arrow indicates increasing metamorphic grade.

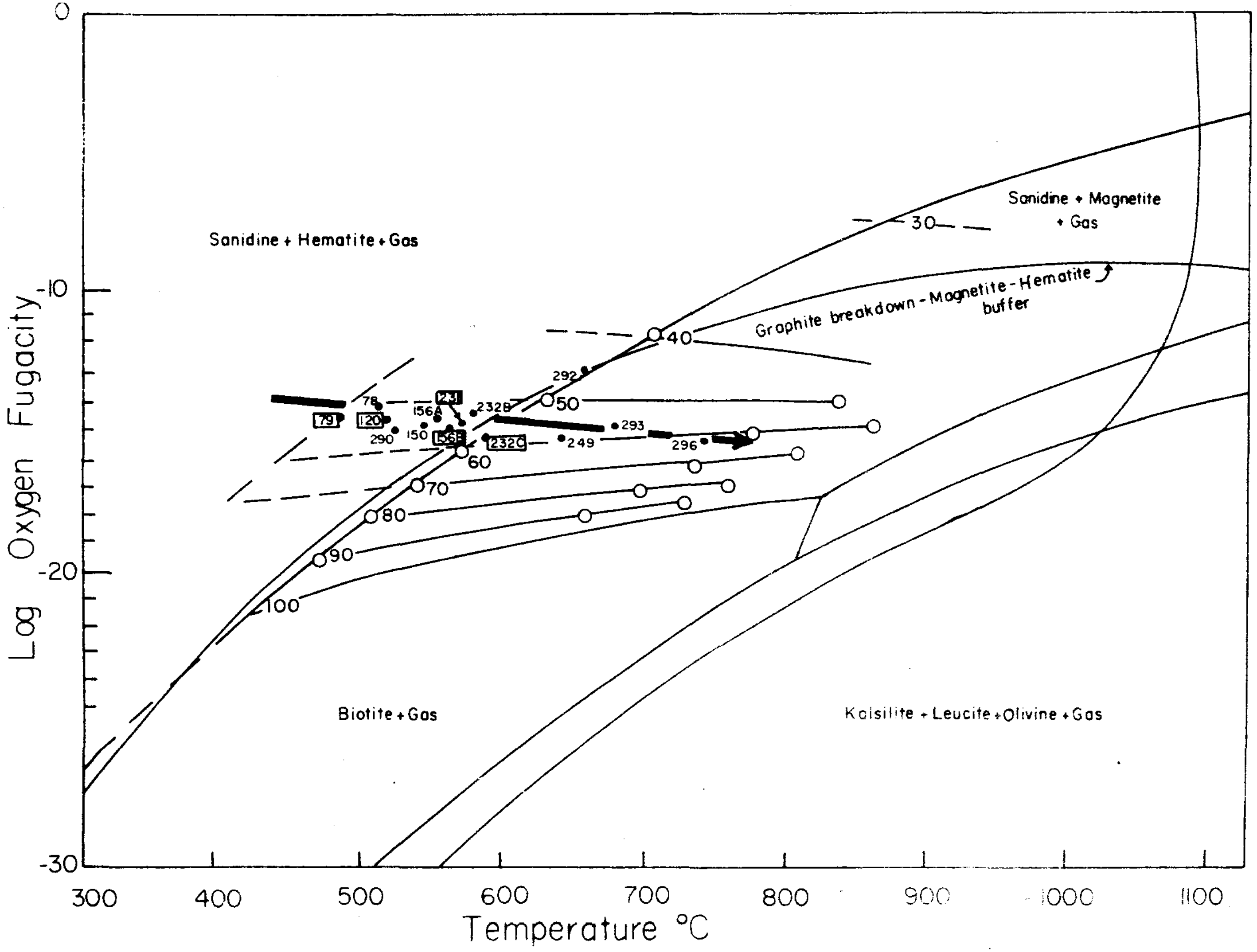

0
-10
-20
-30
Log Oxygen Fugacity
300
400
500
600
700
800
900
1000
1100
Temperature °C
Sanidine + Hematite + Gas
Sanidine + Magnetite + Gas
Graphite breakdown - Magnetite - Hematite buffer
Biotite + Gas
Kalsilite + Leucite + Olivine + Gas
30
40
50
60
70
80
90
100
292
78
231
156A
232B
79
120
290
150
156B
232C
249
293
296

Figure 59.

tent in biotite is high. In the present study, the biotites at the lower grades (below the muscovite breakdown isograd) average .068 mole proportion soda while at higher grades 0.58 is the average. This difference is small indeed. The soda content of biotites are given in table 2. The difference seems small but is consistent with Rutherford's inference that the magnetite field is enlarged by soda being included in the biotite structure.

The anomalous occurrence of the samples in figure 59 may also be caused by the Ti content of the biotites. The element increases in abundance from the low to the high grades in this study as mentioned in section II, E, iii. The effects of Ba, Ca, Mn and Fe^{+3} may be important individually or in a combination of their effects (analyses given in appendix II). Sample T-292-1 is the most anomalous as it occurs off the generalized trend in figure 59. The sample is anomalously high in Ba (0.45 molar proportion while the average for all the samples is less than half that), high in Na for the samples found near it (see table 2) and low in Ti with respect to the samples found near it. The sample may be anomalous due to the effects of impurities of may simply have formed at a lower P_{O_2} than the samples near it.

It must be pointed out that the exchange of Mg between biotite and garnet after the crystallization of the minerals may have a significant effect on the P_{O_2} determinations.

Table 2.

Content of the modal proportion of Fe^{+2}/Fe^{+2} + Mg, Na_2O, and BaO in biotite

Sample	Fe^{+2}/Fe^{+2} + Mg	Na_2O molar proportion	BaO molar proportion
T-79-1	.54	.055	.009
T-78-1	.53	.090	.009
T-120-1A	.53	.048	.005
T-290-1A	.55	.062	.012
T-150-1B	.54	.088	.017
T-156-1A	.53	.069	.023
T-156-1B	.56	.062	.024
T-231-1A	.54	.069	.009
T-232-1B	.53	.059	.030
T-232-1C	.58	.082	.032
T-249-1	.59	.055	.014
T-292-1	.46	.045	.045
T-293-3	.57	.069	.018
T-296-1A	.62	.055	.012

(e) Fugacity of hydrogen

The requirements to determine f_{H_2} from biotite data are the same as for the oxygen determination. The plots of the values are shown on figure 60. The large black arrow indicates a very rapid increase from lower grades (10^{-3} bars) to high grades (10^{+1} bars) and the samples plot nearly in the same sequence as they do on the metamorphic section. The exception to this is sample T-292-1 which has a lower f_{H_2} value than the samples found near it.

(f) Partial pressure of H_2O

Wones and Eugster (1965) have presented a formula by which P_{H_2O} can be determined:

$$\log f_{H_2O} = \frac{3428 - 4212\,(1-x)^2}{T} + \log x + 0.5 \log f_{O_2}$$

$$+ 8.23 - \log a_{KAlSi_3O_8} - \log a_{Fe_3O_4} \quad (\pm 0.20)$$

where $x = Fe^{+2}/Fe^{+2} + Mg$ in biotite

T = temperature

a = activity

This equation presupposes all the requirements necessary for a f_{O_2} determination of biotite. For the present study, the compositions of the K-feldspar (or muscovite plus aluminum silicate) and magnetite are unknown. If their activities are unity, the log f_{H_2O} would be a minimum. The values of the activity of K-feldspar used were those cited by Wones and Eugster (1965). These values (.81 at 500°C and .79 at 600°C) were determined from Orville's (1963) data and, in the present study, were

Figure 60.

Variation of log f_{H_2} across area versus temperature; the heavy dark arrow indicates increasing grade. The near parallel lines indicate the ratio of $Fe^{+2}/Fe^{+2} + Mg$ in percent.

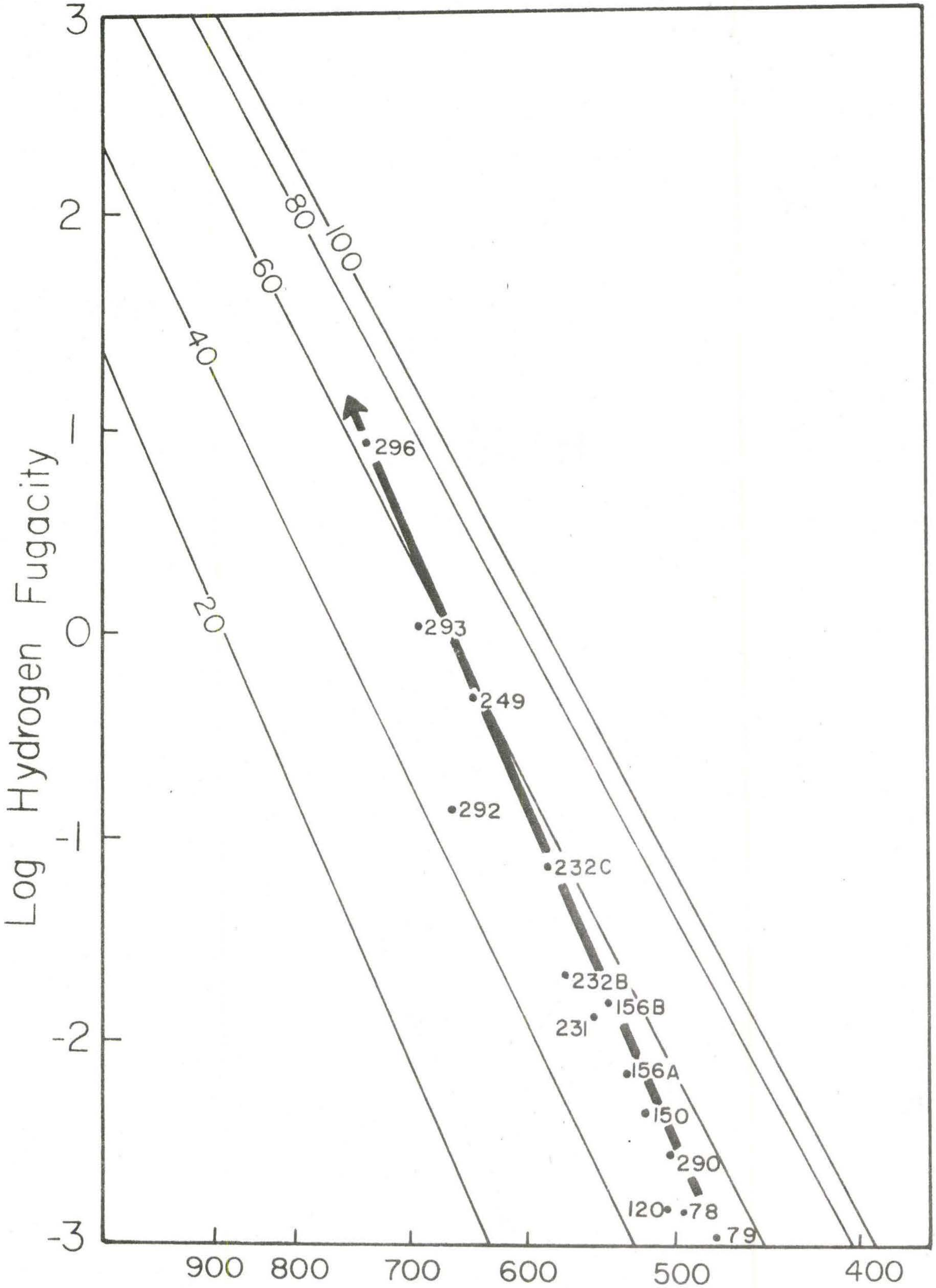

3
2
1
0
-1
-2
-3
Log Hydrogen Fugacity
900
800
700
600
500
400
Temperature °C
20
40
60
80
100
296
293
249
292
232C
232B
156B
231
156A
150
290
120
78
79

Figure 60.

extrapolated linearly to .77 at 700°C. The magnetite at this temperature and f_{O_2} will be nearly pure (interpretation from Buddington and Lindsley, 1964) and the near unity activity will cause the term to be negligible.

If the system behaves ideally, the fugacities determined from the biotite data will equal the pressure and the fugacity coefficients will equal unity. The relation for non-ideality is as follows:

$$\text{Pressure (P)} = \frac{\text{fugacity (f)}}{\text{fugacity coefficient } (\gamma)} \quad \text{(from Moore, 1962, p.193)}$$

The fugacity coefficients used in the present study were those of Holser (1954) and are for pure H_2O. Holser's values apply up to 800°C but only 2000 bars pressure. When plotted on a graph of water fugacity versus pressure for different temperatures, it becomes apparent that the graph levels off at 1500 bars and an extrapolation to the pressures inferred in this study are reasonable (see figure 61). Near the muscovite breakdown, the P_{H_2O} takes a sudden drop in values. This is consistent with theoretical considerations of the breakdown of the mineral and reinforces the data on the partial pressure. P_{H_2O}, f_{O_2} and f_{H_2} are shown in figures 56 and 58. The first two decrease almost exponentially from the lowest to the highest grades while f_{H_2} increases exponentially.

(g) Carbon bearing species

Five of the fourteen whole rock samples selected contain carbon in the form of graphite. The use of this mineral in metamorphic rocks has been explained by Miyashiro (1964) and experimental breakdown curves, using various oxygen buffers, have been determined by French and Eugster (1965).

Figure 61.

Values and extrapolations (dashed lines) of fugacity coefficients after Holser (1954).

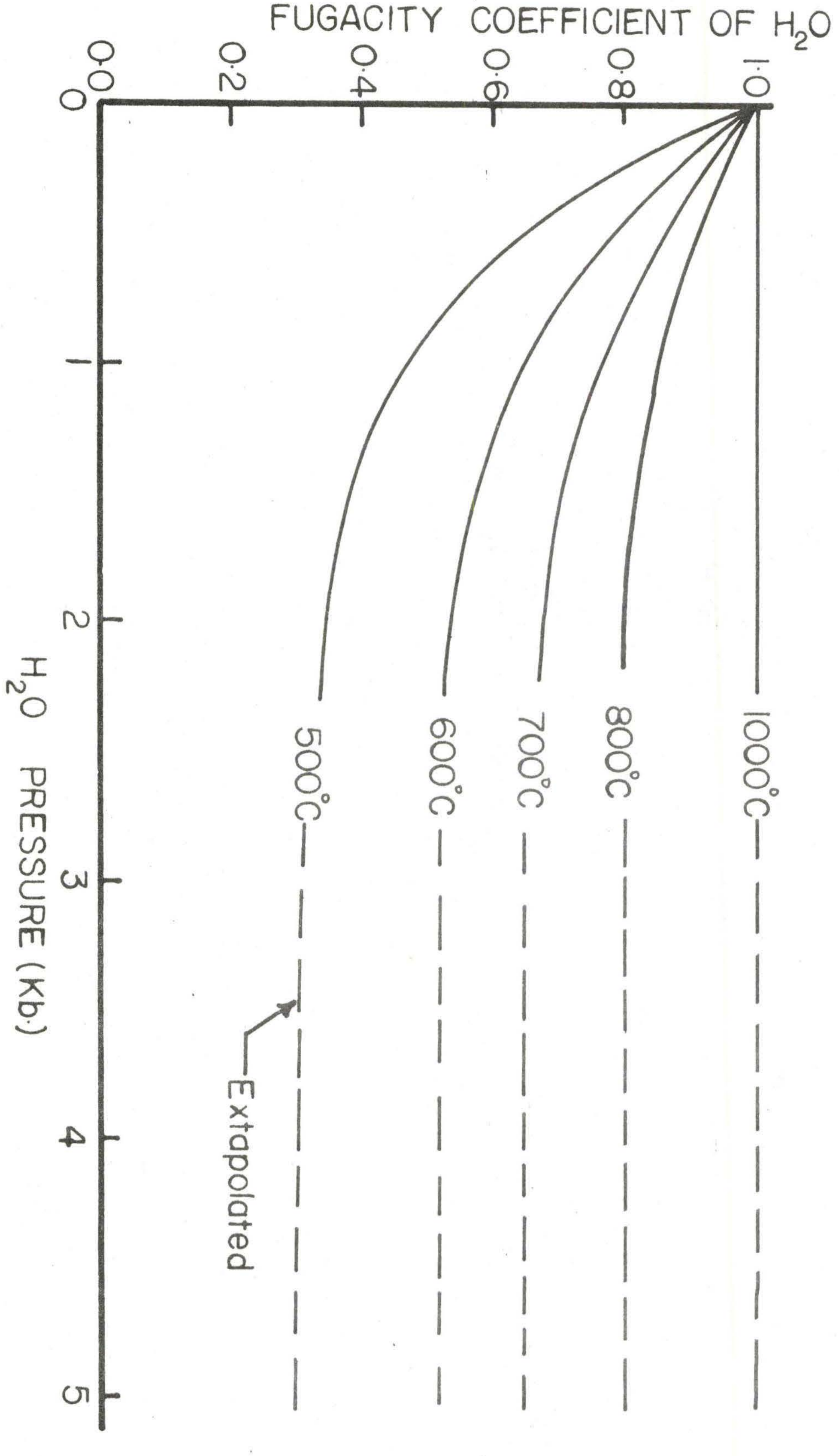

FUGACITY COEFFICIENT OF H_2O
0·0
0·2
0·4
0·6
0·8
1·0
0
1
2
3
4
5
H_2O PRESSURE (Kb.)
500°C
600°C
700°C
800°C
1000°C
Extapolated

Figure 61.

The breakdown curve occurring at the highest partial pressure of oxygen, the hematite-magnetite buffered one, has been reproduced in figure 59. From this curve we can delimit the maximum values of P_{CH_4}, P_{CO} and P_{CO_2} and using the author's data it can easily be shown that the sum of the carbon species partial pressures did not exceed one bar.

On figure 59 the samples containing graphite are enclosed in rectangles but, on the basis of the biotite data, plot beyond the hematite-magnetite buffer curve. Possible reasons for this are:

(a) impurities in the graphite

and (b) errors in the biotite data mentioned in the last section.

The samples containing graphite do not seem to be anomalous to the samples not containing the mineral so that either:

1. the samples without graphite formed under a partial pressure of carbon bearing gas which was identical to the samples containing graphite;

or 2. the effect of the partial pressure of carbon bearing gas was negligible.

(d) Oxidation Ratio

Chinner (1960) found that a ratio of $\frac{2Fe_2O_3}{2Fe_2O_3+FeO}$ x 100% was a good index of the degree of oxidation of the rocks he studied. He found that the ratio varied greatly for individual layers in close proximity to each other within an almandine amphibolite facies terrain and concluded (using inferences from Fe-Ti oxides) that the layers had been closed with respect to oxygen pressure during metamorphism.

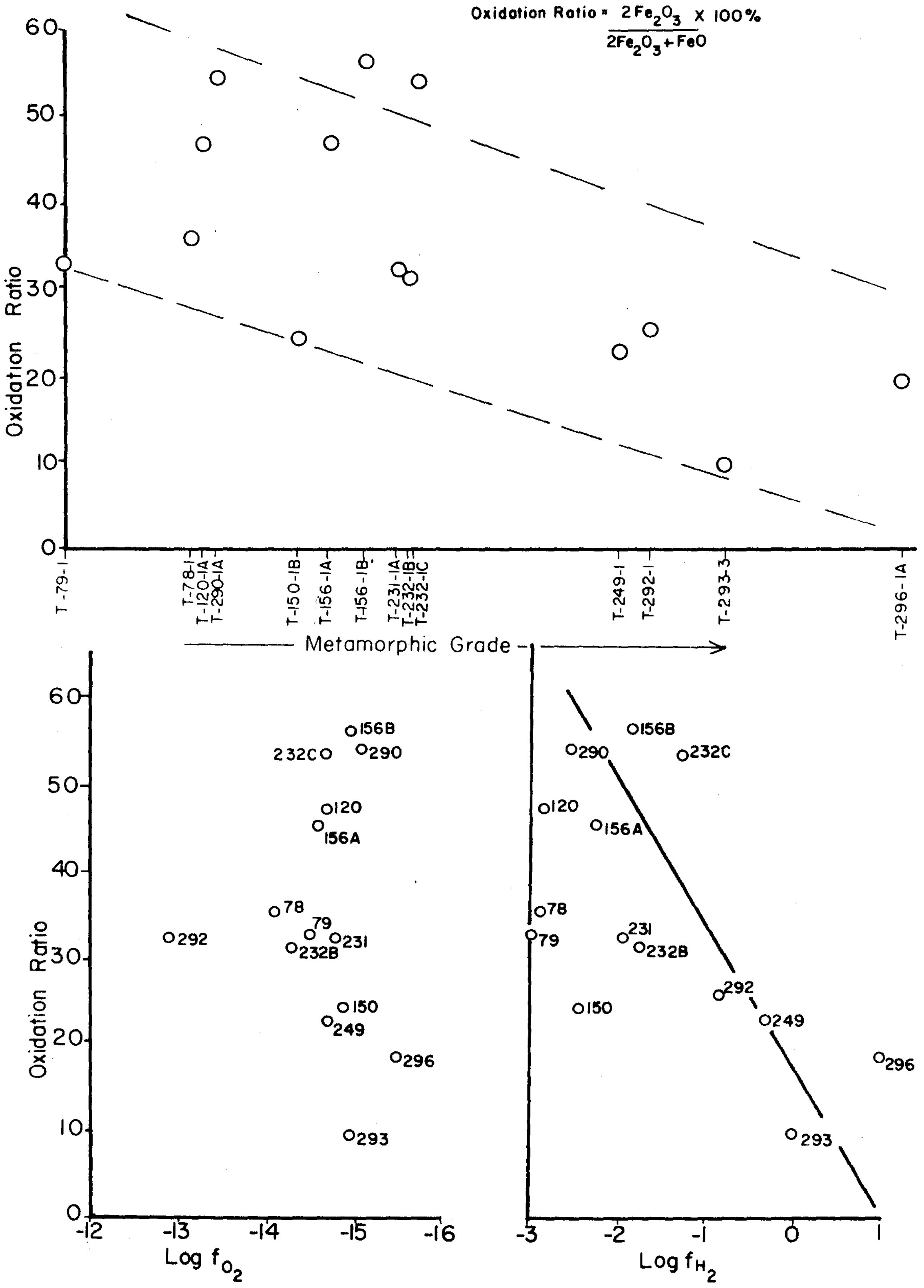


Figure 62

Variation of oxidation ratio versus metamorphic grade, log f_{O_2} and log f_{H_2}.

In the present study, the oxidation ratios for the 14 whole rock samples are plotted on figure 62 and a crude decrease of values with increasing metamorphic grade can be observed, as has been observed in numerous other areas. The decrease of the oxidation ratio is paralleled by an increase in f_{H_2} although no clear relation to f_{O_2} can be shown.

Summary of Intensive Parameters

The inferred temperatures and pressures show an increase from 475 to 725°C and 4.8 to 7.7 kilobars, respectively. The change of these parameters is assumed to be linear.

The partial fugacity of S_2 is inferred to change from approximately 10^{-5} at the lowest grades to 10^{-1} bars near the center of the section. The partial vapour pressure of phosphorous and halogen bearing species is unknown but may have been substantial. The partial fugacity of oxygen decreases from $10^{-14.5}$ to $10^{-15.5}$ bars towards the high grade end of the map area. F_{H_2} increases exponentially from $10^{-3.5}$ to $10^{'}$ at the high grade end. This is matched by steep decrease of P_{H_2O} from 4 kilobars to near 1 kilobar at the high grade end. The partial pressure of carbon bearing species is unknown but must have been less than 1 bar.

The oxidation ratio as defined by Chinner (1960) decreases across the area to the high grade end of the section. There appears to be a correlation between oxidation ratio and inferred f_{H_2} in that the ratio decreases as f_{H_2} increases.

The total fluid pressure is much less than the total pressure which suggests that the assumption $P_{fluid} = P_{total}$ may be erroneous in other studies.

CHAPTER III

CONCLUSIONS

A. GENERAL GEOLOGY

i) Metamorphism

The Southern province consists of rocks of the staurolite-quartz subfacies of the almandine amphibolite facies near the boundary with the Grenville province and the upper greenschist facies a few kilometers northwest.

The Grenville province contains, progressively away from the boundary, the staurolite-quartz subfacies, kyanite-muscovite subfacies, and sillimanite-K-feldspar subfacies of the almandine amphibolite facies and, 10 kilometers southeast of the boundary, the lower granulite facies. This is inferred to represent a linear temperature increase of 475 to 725°C, linear pressure increase from 4.8 to 7.7 bars, an exponential P_{H_2O} decrease from 4 kilobars to 1 kilobar, an exponential f_{H_2} increase from 10^{-3} to 10^{1} bars and a f_{O_2} exponential decrease from $10^{-14.5}$ to $10^{-15.5}$.

ii) Protoliths

The Southern rocks contain many features indicative of miogeosynclinal sequences. The protoliths were mainly pure quartz sands with some feldspathic ingredients and a few interbedded impure pelitic (shales ?) sediments.

The Grenville province lithology is indicative of more immature sediments. The gneisses were probably impure quartzites and greywackes as they contain usually about 50 to 60% quartz. A few layers still contained breccia fragments and lithology changes were numerous. Most of the rocks were probably formed under marine conditions for siliceous limestones (now calc silicate marbles) occur as small scattered layers throughout the section. There are a few layers of very pure sandstone but these occur mainly within the central part of the map area and may represent a time in the sedimentation when better sorting and mechanical weathering could occur. Pelitic (shale ?) layers are found throughout but were especially closely associated with the pure quartzites. It is not inferred that volcanic sediments are numerous, if present, although some rocks are inferred to have been intrusive gabbros and gabbroic dykes.

Grenville-Southern Boundary

Of the various models to explain the boundary described, the one preferred attributes the boundary to a break, marked by a mylonite zone, which brought Penokean? high grade metamorphic rocks into juxtaposition with lower grade equivalents. This is explained in section I,E.

B. GEOCHEMISTRY OF GRENVILLE PELITIC ROCKS

i) Modal Abundance changes

Modal abundances of minerals occurring in aluminum silicate bearing pelites show systematic changes across the metamorphic interval. Trend surface analysis maps for each mineral are fairly consistent with hand contoured maps and allow a more objective appraisal of the changes. The modal abundances of sillimanite and kyanite change linearly and are inverse to each other in trends. Staurolite shows a near-linear decreasing trend with metamorphic grade. Muscovite and K-feldspar show inverse trends to each other. Garnet increases sharply with increasing grade. Biotite, quartz and plagioclase trends are somewhat erratic.

ii) Grain homogeneity

Garnet is very inhomogeneous in composition within individual garnets and two types of zoning have been observed.

(a) zoning of Ca and Mn from the center of the garnet to the edge;

and (b) zoning of Mg and Mn at the outer rim of the garnets.

The first zoning is believed to be in response to garnet growth from the rock matrix. The second is in response to re-exchange of ions with biotite.

Beyond the staurolite isograd, type a (above) zoning is not observed and the interior of all the garnets in the same section are at a constant value of oxide content, indicating a previous homogenization.

The degree of Mn (and Mg?) zoning of type (b) can be related to the ratio biotite/garnet. Biotite inhomogeneities of composition of different grains in the same section occur. Compositions change on approaching garnet crystals and values are inverse to those of zoning type (b) in garnet. Individual grains of kyanite, sillimanite, muscovite an biotite are homogeneous.

iii) Geochemistry

There is no clear linear change of garnet compositions with increasing metamorphic grade (section F, ii).

Ti in phyllosilicates shows a linear increase in response to increasing metamorphic conditions. In biotite this is paralleled with a linear decrease of the value of the function $Fe^{+2} + Mg + Al$.

Trace and major element partition indicate that most coexisting phases were chemically approximately in equilibrium with respect to most elements. Some, such as kyanite-muscovite, which have obvious disequilibrium textures, also have chemical disequilibrium partitions. The change of distribution coefficients ($D_{P,T\ etc.}$) with changing intensive parameters are probably linear for:

(a) Mn and Mg between muscovite and biotite.

(b) Mn and Mg between muscovite and garnet.

(c) Mn, Mg, Ca and Fe^{+2} between biotite and garnet.

Between biotite and garnet, Mn and Ca partially exchange for Fe^{+2} and Mg with increasing grade.

It is inferred from the partial vapour pressures that the rocks behave essentially as closed systems with respect to such parameters as f_{O_2}, f_{H_2} and P_{H_2O}. The values of these parameters could conceivably be partially controlled by buffer systems in the surrounding gneisses and amphibolites.

APPENDIX I

CHEMICAL CONSIDERATIONS

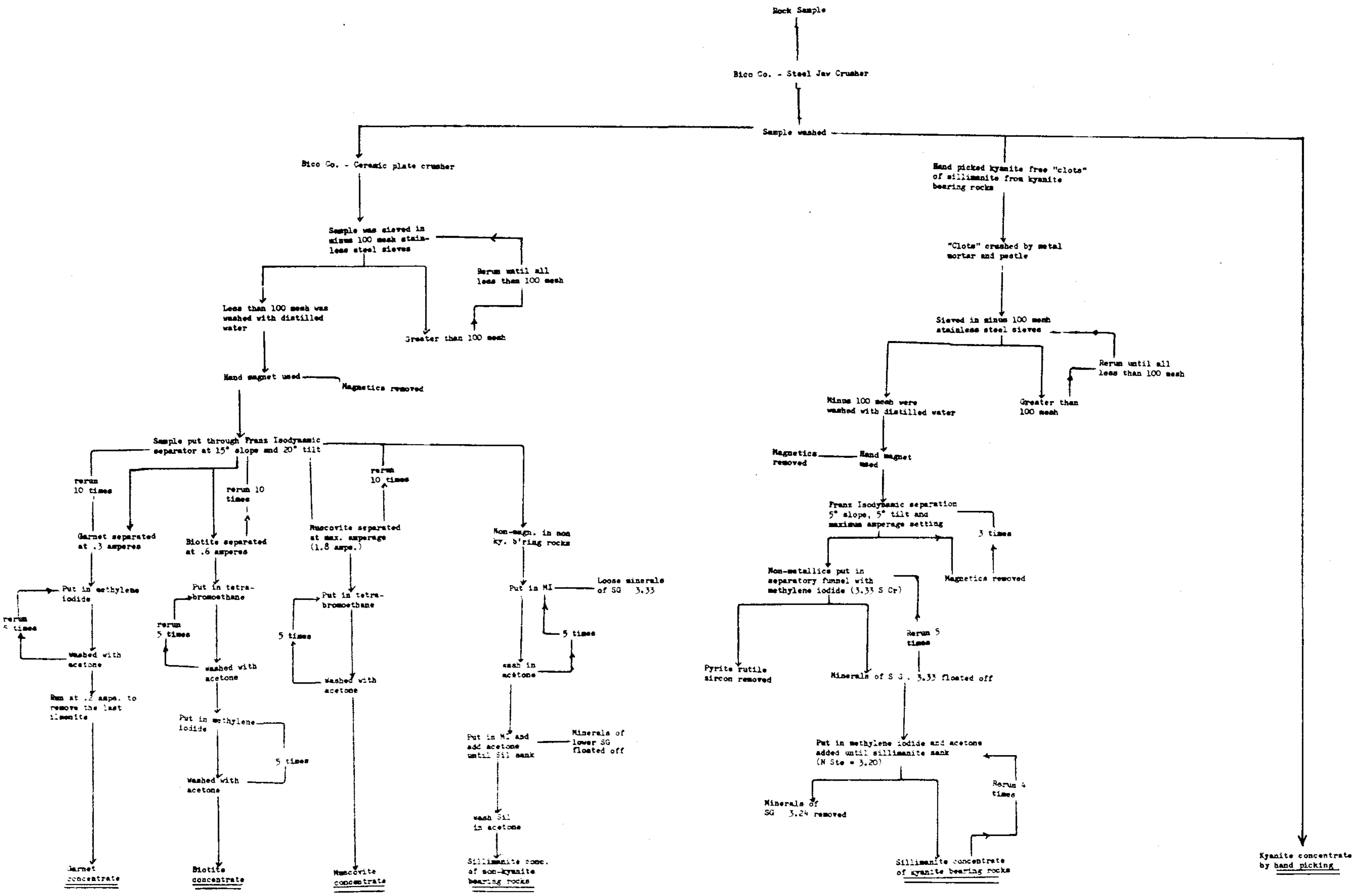
1. Separation techniques

Each of the fourteen selected rock samples was cut into four portions; one for mineral separation, one for bulk chemical analysis, one for thin sectioning and one for polished sectioning. The portion allotted to mineral separation was treated by hand picking, magnetic and heavy liquid techniques to remove concentrates of sillimanite, kyanite, biotite, muscovite and garnet. Each sample was broken into one inch fragments before crushing and was carefully inspected so that weathered material could be removed. The exact separation techniques are shown in the accompanying flow sheet (figure 64).

Many difficult problems were encountered during the separations of which the main one concerned removal of inclusions from the 100 to 200 mesh concentrates. An obvious solution would seem to be to crush the grains to smaller sizes but it was found that separations became difficult to perform on grain sizes much less than 200 mesh. The main inseparable contaminants that fell into this category were: minute rutile crystals within many of the concentrates, biotite flakes interlayered within muscovite, minute fibers of sillimanite in muscovite and biotite, and microscopic inclusions of quartz and biotate in garnet.

Another problem was the difficulty of separating two minerals of similar properties. Coarse-grained, prismatic, blue kyanite was separated from fine-grained fibrous sillimanite by hand picking clean kyanites and

Figure 64: Flow Chart of Mineral Separations



kyanite free "clots" of sillimanite. The separation of ilmenite from garnet was facilitated by repeated passes through the Franz Isodynamic Separator. The settings (near .2 amps at 15° slope, 20° tilt) had to be varied slightly as it appeared the separator was indicing magnetism in the minerals. Rutile, zircon, ilmenite and pyrite were separated with great difficulty from sillimanite by repeated settlings in methylene iodide and finally, hand picking.

Some samples still contained alteration products in spite of the measures taken to avoid this. Garnet sample T-120-1A contained inseparable secondary chlorite. Most of the muscovites and some of the biotites and garnets had surface films of limonitic material.

An acid wash, as used by some (Albee et al., 1965), was not used as it was felt that this might alter particularly the phyllosilicate's interlayered composition. A method involving a Blackstone laboratories ultrasonic probe was tried, to remove the films, but after lengthy exposures (hours) no apparent effect was observed. It was finally decided, as the film was quantitatively a very small portion of the grains, it be retained.

2. Homogeneity of concentrates

The desired purity of a mineral concentrate will depend on the purpose of the study. The magnitude of any error depends on the concentration of the desired element as well as the volume of the contaminant so that fine grained, common contaminants such as rutile, ilmenite and zircon will have a profound affect on the composition of the host concentrate or Fe, Ti and Zr.

The homogeneity of mineral concentrates is a vaguely mentioned subject by most authors possibly because of the difficulty of measuring it. The parameter one wishes to measure is the percentage volume of the constituents but, at best, a modal abundance based on a grain's two dimensional area is effected.

The measurement of the volumes is difficult due to the irregular shape and non-uniform volumes of all the grains even though they occur between two Sieve Sizes (e.g. 100 to 200 mesh).

Impurities can be listed under two categories; those which occur as idstinct foreign grains and those as inclusions within grains of the desired mineral. The counting of the number of distinct foreign grains in the concentrate, as done by some authors (e.g. Chiang, 1965) is, in my opinion, too crude a method as it assumes absolute uniformity of grain parameters. A point counting method which takes into account at least two dimensions of the grains was adopted in this study. It still has the unfortunate limitation of not representing the third dimension and as a consequence, for example, a large flat-lying thin biotite flake may be represented as having twice the volume of a smaller equant garnet fragment. Probably the most accurate way to determine volumes is to encase "randomly" oriented grains within clear plastic and cut a thin section. This was attempted but abandoned because of the mechanical difficulty of making sections and the relatively greater loss of amount of concentrate to that in the method employed. A common polished mount would not be satisfactory as one would have difficulty to identify non-opaque minerals.

Table 3.

IMPURITIES IN MINERAL CONCENTRATES

No.	Bio.	Graph.	Musc.	Gar.	Chlor.	Opaq.	Zirc.	Sil.	Rut.	Qtz.	Staur.	Plag.	Ky.	Total For.	Conc. & inc.	Remarks
Biotite																
78	98.8	.1	.5	.5	.1	-	-	-	-	-	-	-	-	.7	99.3	No stain (limonite)
79	87.6	.4	-	.9	11.1	-	-	-	-	-	-	-	-	5.9	94.1	No stain
290	99.4	-	-	.1	-	-	-	.5	-	-	-	-	-	.2	99.8	5% of grains slightly stained
120	91.9	.8	.7	-	6.3	.1	-	.2	-	-	-	-	-	2.2	97.8	No stain
232-1C	97.7	.3	.5	1.1	-	-	-	-	.1	.3	-	-	-	2.1	97.9	No stain
150-	98.7	.1	.7	-	-	-	.2	.3	-	-	-	-	-	.6	99.4	No stain
156-1A	98.1	.2	1.2	.2	-	-	.2	-	-	-	.1	-	-	.9	99.1	3% of grains slightly stained
156-1B	96.0	.5	.2	.4	-	.1	.1	2.7	-	-	-	-	-	.5	99.5	5% " " " "
231	98.5	.4	.2	-	-	-	.1	.8	-	-	-	-	-	.2	99.8	No stain
232-1A	98.1	.4	.3	.6	-	-	.1	.4	-	-	.1	-	-	1.0	99.0	10% moderately stained
296	98.2	-	-	.2	-	-	-	1.6	-	-	-	-	-	.4	99.6	1% slightly stained
293	97.2	-	-	1.0	-	.4	.1	1.2	-	-	-	.1	-	1.1	98.9	No stain
292	97.2	-	-	.7	-	-	-	2.1	-	-	-	-	-	.9	99.1	No stain
249	97.6	-	-	.3	2.0	-	-	-	-	.1	-	-	-	.6	99.4	No stain
Garnet																
78	.1	-	-	99.8	-	.1	-	-	-	-	-	-	-	.2	99.8	No stain, many minute inc.
79	-	-	-	99.9	.1	-	-	-	-	-	-	-	-	-	100.0	No stain, many minute inc.,12% chlori-
290	.2	-	-	99.7	-	-	-	.1	-	-	-	-	-	.3	99.7	2% of grains slightly stained /tized
120	-	-	-	99.3	-	.7	-	-	-	-	-	-	-	.7	99.3	No stain
232-1C	.8	-	-	98.2	-	1.0	-	-	-	-	-	-	-	1.0	99.0	No stain
150	.1	-	-	99.5	-	.4	-	-	-	-	-	-	-	.5	99.5	No stain
156-1A	.1	-	-	99.7	-	.2	-	-	-	-	-	-	-	.3	99.7	No stain, 5% grains inc.,inc=5% of gr.
156-1B	-	-	-	98.8	-	1.2	-	-	-	-	-	-	-	1.2	98.8	5% slightly stained,some inc.(minute)
231	-	-	-	99.6	-	.4	-	-	-	-	-	-	-	.4	99.6	No stain
232-1B	.3	-	-	99.9	-	.6	-	.2	-	-	-	-	-	1.1	98.9	7% of grains slightly stained
296	.2	-	-	98.9	-	.9	-	-	-	-	-	-	-	1.1	98.9	2% of grains " "
293	.2	-	-	99.5	-	.3	-	-	-	-	-	-	-	.5	99.5	No stain
292	.1	-	-	99.8	-	.1	-	-	-	-	-	-	-	.2	99.8	No stain
249	.1	-	-	99.1	.1	.6	-	-	-	.1	-	-	-	.6	99.4	10% of grains stained
Muscovite																
78	.4	.4	98.4	-	-	-	-	.2	-	-	-	-	.6	1.0	99.0	30% slightly stained
79	.6	.4	95.8	-	-	-	-	2.2	-	-	-	-	1.0	2.7	97.3	15% moderate stain
290	-	-	98.0	-	-	-	-	1.3	-	-	-	-	.7	1.1	98.9	30% " "
120	.4	.4	98.2	-	-	-	-	.6	-	-	-	-	.4	.7	99.3	20% " "
232-1C	.6	1.2	96.6	-	-	-	-	1.2	.2	-	-	-	.2	.7	99.3	50% moderate to intensely stained
150	1.6	.6	96.4	-	-	-	-	1.1	.1	-	-	-	.2	1.0	99.0	5% slightly stained
156-1A	.5	-	98.8	-	-	-	-	.6	-	-	-	-	.1	.5	99.5	5% slightly stained
156-1B	1.5	1.0	93.8	-	-	-	-	3.3	-	-	-	-	.4	.7	99.3	10% " "
231	1.0	2.2	85.0	-	-	-	-	4.6	.1	-	-	-	7.1	9.8	90.2	15% " "
232-1B	-	.2	96.0	-	-	-	-	3.2	-	-	-	-	.6	.8	99.2	50%moderate stain

No.	Bio.	Graph.	Musc.	Gar.	Chlor.	Opaq.	Zirc.	Sil.	Rut.	Qtz.	Staur.	Plag.	Ky.	Total For.	Conc. & inc.	Remarks
Sillimanite																
78	.1	-	-	-	-	-	.1	99.8	-	-	-	-	-	.2	99.8	No stain
79	-	-	-	-	-	-	-	99.5	-	.5	-	-	-	.5	99.5	alt. of f.g. musc. on all,20% stained
120	-	.5	.5	-	-	-	.1	96.3	.1	.5	-	-	1.0	2.7	97.3	3% stained slightly
290	-	-	-	-	-	-	-	99.8	-	.2	-	-	-	.2	99.8	5% " "
232-1C	.2	-	.3	-	-	-	1.0	98.0	.5	-	-	-	-	2.0	98.0	30% slightly to mod. stained
150	1.3	-	-	-	-	1.0	-	84.2	-	.5	-	-	3	15.8	84.2	10% slightly stained
156-1A	1.0	.5	-	-	-	-	.2	93.1	.2	3.0	-	-	3.0	7.9	9.21	10% " "
156-1B	-	-	.5	-	-	-	-	98.6	.1	.3	-	-	.5	1.4	98.6	10% " "
231	.1	-	1.0	-	-	-	.5	95.4	-	-	-	-	3.0	4.6	95.4	10% " "
232-1B	.3	.2	-	-	-	-	-	99.6	.1	-	-	-	-	.4	99.6	5% " "
296	.1	-	-	-	-	-	-	99.0	-	.9	-	-	-	.9	99.1	1% " "
293	-	-	-	-	-	-	.3	99.6	.1	-	-	-	-	.4	99.6	No stain
292	.5	-	-	-	-	-	.3	99.1	-	.1	-	-	-	.4	99.6	No stain
249	-	-	-	-	-	-	-	99.7	.1	.2	-	-	-	.3	99.7	No stain
Kyanite																
78	no impurities											-	100.0	-	-	No stain
79	-	1.0	-	-	-	-	-	-	-	-	-	-	99.0	-	100.0	Some musc. alt. on 50% of grains
290	-	.1	-	-	-	-	-	-	-	-	-	-	99.9	-	100.0	No stain
120	-	.2	.5	-	-	-	-	-	1.0	-	-	-	98.3	-	100.0	Slightly stained (20%)
232-1C	-	-	-	-	-	-	-	-	.5	-	-	-	99.5	-	100.0	No stain
150	no impurities											-	100.0	-	100.0	No stain
156-1A	"	"										-	100.0	-	100.0	10% slightly stained
156-1B	"	"										-	100.0	-	100.0	No stain
231	-	.1	-	-	-	-	-	-	.2	-	-	-	100.0	-	100.0	No stain
232-1B	-	-	-	-	-	-	-	-	-	-	-	-	99.7	-	100.0	No stain

Bio. = Biotite
Graph. = Graphite
Musc. = Muscovite

Total For. = Total Foreign Grains

Gar. = Garnet
Chlor. = Chlorite
Opaq. = Opaque

Zirc. = Zircon
Sil. = Sillimanite
Rut. = Rutile

Conc. + inc. = Concentrate plus inclusions

Qtz. = Quartz
Staur. = Staurolite
Plag. = Plagioclase
Ky. = Kyanite

In this study, mechanically split samples of mineral concentrates were mounted in Canada balsam between a glass slide and a cover glass and 1000 equally spaced points were counted using a mechanical stage. 1000 points were accepted as representative as this was well beyond the levelling off of the point counts versus mineral concentration graphs.

Determining the quantity of impurities included within the host mineral is extremely difficult. Allowances were made, where possible, for included grains.

The problems associated with limonitic films have already been mentioned. The films were quantitatively described on the basis of percent of grains with film and "relative" intensity of film thickness.

The homogeneity of the mineral concentrates is shown in the accompanying table 3.

3. Spectrographic method

The mineral concentrations were mechanically split into portions for major element chemistry, trace element chemistry and grain mounts for impurity determinations. Approximately 50 milligrams allotted to trace element determination were accurately weighed by use of a Vereenidge Draddfabireken Berman blanace and ground to a fine powder, with acetone, in an agate mortar and pestle. The grains in the mortar were inspected before and after the grinding and any impurities were removed by use of a wetted needle. When all the grains were sufficiently fine, an equal weight of buffer containing 0.025% $PdCl_2$, 49.975% graphite and 50.000%

Cs_2CO_3 was added. The buffer and sample were mixed for 3 to 5 minutes in acetone and on drying of the mixture, six cratered anode electrodes were filled and packed.

The rapid deliquescence of caesium carbonate is a great problem as in weighing, the added water decreases the actual amount of carbonate, and in burning, erratic and explosive burns are produced.

The problem was lessened by heating the carbonate to 200°F for an hour before weighing and heating the buffer-sample mixture a final 2 minutes just before arcing of the electrodes (see Payne, 1966).

The spectrographic parameters are as follows:

Spectrograph - Jaco 21 foot grating spectrograph, Wadsworth mount, first order dispersion 5.2 Å/mm.

Condensing optics (distances from slit)

0.0 cm. Cylindrical lens (horizontal axes) focal length 25 cm.

16.1 cm. Cylindrical lens (vertical axis) focal length 6.7 cm.

27.5 cm. Disphragm with 5 mm. aperture.

58.1 cm. Spherical lens, focal length 10 cm.

72.5 cm. Arc location

Arc gap - 4 mm.

Slit width - 30 microns.

Slit height - 8 mm.

Intensity Reduction - 7-step rotating sector (log intensity ratio = 0.2) at the slit. Two mesh screens used for the range 2400 - 4800 Å or yellow filter for the range 5700 - 8300 Å at the diaphragm.

Voltage - 225 volts D.C., open circuit.

Current - 9.5 amps.

Exposure - Samples burned to completion (50-70 seconds burning time).

Electrodes - Anode - Ultra Carbon Corp. preformed 1/8" diameter rod; plain crater 1/16" x 3/8".

- Cathode - National Carbon Co., special graphite; 1/8" diameter rod.

Electrode Stand - ARL arc-spark stand, water cooled jaws.

Gas jet - Stallwood jet used with a mixture of 79% Argon and 21% oxygen for the range 2400 - 4800 Å, or air stream for the range 5700 - 8300 Å at 18 SCHF.

Cooling - Water at 15 GPH.

Photographic plates - Eastman Kodak type SA1, range (2400-3600 Å)
" " " IIIF, " (3600-4800 Å)
" " " I N, " (5700-8300 Å)

Processing - Kodak D-19 Developer, one half minute water wash, 3 minutes at 20°C. Fixer, 5 minutes at 20°C, one half hour water wash at 10 degrees centigrade.

Photometry - Jaco microphotometer, three steps measured, background set to 100 on first step read correction read on third step.

4. Major element techniques

Many of the conclusions could not have been satisfactorily shown without the determination of major elements. Mr. J.R. Muysson very ably determined the compositions of biotite, garnet, and muscovite by the following methods:

Al_2O_3 - colorimetrically with Alizarin Red-S on Solution prepared by acidifcation of a NaOH fusion of the sample (as in L. Shapiro and W.S. Brannock - USGS, Bulletin 1036-C).

SiO_2 - colorimetrically as the blue complex of silico-molybdate after fusion as for Al_2SiO_3 (Muysson's modification, unpublished, of Shapiro and Brannock's method).

TiO_2 - colorimetrically with "Tiron" (Shapiro and Brannock's method).

Total Fe - colorimetrically with O-phenonthroline (Shapiro and Brannock) but using differential spectro-photometric measurements when necessary.

FeO - slightly modified procedure of Reichen and Faher (USGS, Bulletin 1144-B).

MnO, Na_2O, K_2O, CaO, MgO, BaO - by atomic absorption spectrometry.

H_2O - by method of Cruft, Ingamells and Muysson (Geochim. et Cosmochim. Vol. 29, No. 5).

The methods used to determine the whole rock analysis are as follows. They were performed at the University of Manitoba, Geology Department, by K. Ramlal.

SUMMARY OF METHODS USED IN THE ANALYTICAL LABORATORY

UNIVERSITY OF MANITOBA - GEOLOGY DEPARTMENT

Element	Method
Si Al Fe (Total) Mg Ca K Ti Mn Zr	X-ray fluorescence Spectrometry. Weighted sample plus $Li_2B_4O_7$ plus La_2O_3. Heated in graphite crucible at about 1050°C for ½ hour. Resulting glass bead with H_3BO_3 (Total weight 2.1000 grams) ground to -200 mesh and then compressed to 50,000 p.s.i. Elements then simultaneously analysed on multi channel ARL X-ray Spectrometer.
Na_2O, Cu, Ni, Co Cr, Pb, Zn & Trace Metals	Atomic Absorption Spectrophotometry. Rock dissolved with HF, H_2SO_4, HNO_3 in platinum crucibles - Perkin Elmer 303. A.A.S.
P_2O_5	Colorimetry - solution as for Na_2O above. The absorption at 430 mμ of molybdivanadophosphoric acid complex - Coleman Spectrophotometer.
FeO	Rock decomposed with HF and 1:4 H_2SO_4 solution titrated with $K_2Cr_2O_7$ using Sodium Diphenylamine sulfonate as indicator.
H_2O^-	Determined by heating the sample to constant weight at 110°C.
H_2O Total	Determined by a modification of the Penfield method using PbO as a flux for minerals which are only decomposed at higher temperatures.
H_2O^+	H_2O Total $-H_2O^-$
S	Determined by heating samples in an induction furnace with oxygen flowing through combustion chamber. SO_2 evolved is then titrated. Leco Induction furnace and automatic Titrator.
CO_2	Sample decomposed by HCl and heat. CO_2 evolved passed through drying train and collected on Ascarite.

APPENDIX II

PART 1.

MODAL ABUNDANCES OF MINERALS IN PELITIC SCHISTS

TABLE 4

MODAL PROPORTION OF MINERALS IN THIN SECTIONS IN PELITIC SCHISTS (GRENVILLE PROVINCE)

	musc.	staur.	garnet	sil.	plag.	bio.	kfds.	ky	qtz.	rut.	op.	ap.	px.
T-78-1	16.1	.9	8.3	.3	9.0	19.3	.4	14.8	30.8	.1	.1	.1	-
T-79-1	9.5	2.0	3.8	.1	27.3	20.4	0.0	11.8	24.4	.6	.2	.1	-
T-120-1A	23.0	-.1	3.1	.1	34.7	9.7	-	11.6	17.4	.1	.1	-	-
T-150-1B	2.3	-	4.8	5.1	40.5	43.1	-	.1	3.6	.2	.3	.1	-
T-150-2	2.6	.4	12.4	1.7	-	20.3	4.0	21.2	32.6	.7	-	-	-
T-156-1A	19.7	-	2.5	1.8	9.2	22.4	.5	5.2	39.2	.1	-	-	-
T-156-1B	2.4	-	3.4	11.2	16.4	24.7	-	4.5	36.2	.1	1.2	-	-
T-158-3	65.0	-	4.0	-	.6	16.4	-	11.8	-	.3	1.5	.3	-
T-163-1A	17.7	-	3.4	1.3	1.1	27.7	-	42.7	-	.4	-	-	-
T-167-2	1.1	-	3.1	1.1	12.5	15.6	-	11.6	54.8	.1	-	-	-
T-192-1A	7.0	-	.2	.2	25.9	32.5	-	20.4	14.0	.1	-	-	-
T-195-1A	30.1	-	12.5	.6	23.4	23.6	-	2.4	6.3	.8	.2	.1	-
T-195-1A	10.7	-	-	.4	32.1	24.1	-	15.0	17.8	.1	.1	.1	-
T-195-1C	2.3	.1	25.2	.3	23.2	10.5	.4	7.8	30.8	.3	.1	-	-
T-210-2	5.7	-	8.7	-	41.7	25.5	.1	13.8	3.9	.7	.9	-	-
T-217-1	.9	-	.7	10.9	21.0	8.3	.2	-	57.9	.1	.1	-	-
T-227-3	-	-	2.8	3.5	15.6	37.5	8.3	.5	31.7	.1	-	-	-
T-230-1	.1	.1	6.5	.2	27.6	32.8	-	5.6	27.5	.2	.1	.1	-
T-231-1A	5.4	-	.2	5.7	16.0	34.0	-	11.5	26.8	.5	.2	-	-
T-232-1A	15.6	-	2.6	3.8	18.8	26.7	-	3.8	28.4	.2	.1	-	-
T-232-1C	15.8	-.1	2.6	2.5	29.3	15.5	-	3.6	29.8	.1	.4	.2	-
T-235-1A	25.0	.5	.2	1.0	25.6	.7	4.8	2.2	40.2	.1	-	.1	-

	musc.	staur.	garnet	sil.	plag.	bio.	kfds.	ky	qtz.	rut.	op.	ap.	px.
T-253-2	4.6	-	1.3	8.1	11.3	18.9	1.0	.2	54.6	-	-	-	-
T-249-1	1.1	-	21.8	7.0	10.3	21.8	2.7	-	35.8	.1	.1	-	-
T-254-3	10.4	.1	3.4	.2	30.0	5.0	1.6	5.3	43.3	.2	.5	-	-
T-256-1	-	-	8.4	.3	23.9	27.0	15.0	-	25.4	.1	-	-	-
T-258-9	1.3	-	-	8.9	1.0	14.2	.8	-	73.8	-	-	-	-
T-270-3	-	-	16.7	1.0	29.0	34.4	-	1.9	14.8	-	-	-	-
T-275-1	-	-	7.1	10.8	4.9	47.7	.3	12.5	16.8	-	.1	-	-
T-276-1	.9	-	2.2	-	19.0	25.9	.5	12.1	24.5	.1	-	-	-
T-278-1	14.3	-	-	5.4	14.8	10.0	2.9	-	52.6	-	.2	-	-
T-280-2	.8	-	16.0	2.9	28.5	33.6	-	-	17.5	.1	.4	-	.1
T-281-3	-	-	5.8	1.7	30.7	26.9	-	-	34.0	-	.7	-	-
T-285-2	1.4	.6	.8	6.2	50.9	22.4	-	8.3	8.5	.2	.6	-	-
T-285-6	.2	-	1.0	1.4	42.9	8.1	-	-	45.3	-	-	.1	-
T-286-1	7.5	-	.5	.4	5.0	20.1	10.0	-	56.5	-	.1	-	-
T-286-4	38.7	-	.5	.1	16.6	19.5	.3	-	24.2	-	-	-	-
T-290-1A	2.2	.1	4.2	2.4	11.9	18.6	-	7.0	53.8	.1	.1	-	-
T-291-2	23.3	-	8.5	2.0	10.1	49.7	2.8	-	3.6	-	-	-	-
T-292-1	-	-	2.1	5.0	15.9	36.9	31.8	-	9.0	.1	.1	-	-
T-293-3	-	-	13.7	15.6	9.1	27.1	1.3	-	32.1	.1	.8	-	-
T-296-1a	-	-	26.8	8.5	9.4	20.1	1.6	-	33.0	.1	-	-	-

Opaques-minerals in the 14 pelitic schists studied:

No.	Graphite	Magnetite	Pyrite	Phyrrhotite	Chalcopyrite
Staurolite quartz subfacies					
T-78-1	-	-	x	x	x
T-78-1	x	-	-	-	x
Kyanite-muscovite subfacies					
T-120-1A	x	-	-	-	-
T-150-1B	-	-	x	x	x
T-156-1A	-	-	x	-	-
T-156-1B	x	-	-	-	-
T-232-1B	-	-	-	-	-
T-232-1C	x	-	-	-	-
T-231-1A	x	-	-	-	-
T-290-1A	-	-	-	-	x
Sillimanite-K-feldspar subfacies					
T-293-3	-	-	-	-	-
T-249-1	-	-	-	-	-
T-292-1	-	-	x	-	-
T-296-1A	-	x	-	-	-

APPENDIX II

PART 2.

COMPOSITIONS OF THE MINERALS

Standard and Analysis Lines - Precision of Analyses Lines

Standard	Analysis Line	Precision Coefficient of Variation in Percent
Pd 3242.7	Ga 2943.6	9.8
	Be 3130.4	13.8
	U 3185.4	9.9
	Fe 3244.2	8.3
	Cu 3274.0	13.8
	Y 2327.9	10.3
	Zr 3391.9	17.5
Pd 3221.2	Co 3453.5	11.0
	Ni 3414.7	10.6
Pd 3609.5	Mn 4034.4	13.1
	Si 3911.8	12.5
	Cr 4254.3	11.6
	Ti 4305.9	14.0
Cs 4555.4	Ba 4554.0	12.6
	Sr 4607.3	14.3
Cs 6723.2	Li 6103.6) 6707.8)	8.1
Cs 7609.0	Rb 7800.2	11.0

TABLE 5

GARNET COMPOSITIONS IN WEIGHT PERCENT AND PPM

	T-79+1	T-78-1	T-120-1A	T-290+1A	T-150-1B	T-156-1A	T-156-1B
SiO_2	37.52	37.62	37.07	37.45	37.43	37.55	37.09
TiO_2	.08	.15	.19	.03	.21	.11	.31
Al_2O_3	21.18	20.99	21.17	21.14	21.00	21.15	21.12
Fe_2O_3	.11	.02	.16	.07	.02	.11	.21
FeO	33.79	35.27	34.14	33.90	33.95	32.94	33.92
MnO	1.12	.89	1.30	1.08	1.75	1.55	1.61
MgO	3.75	3.43	3.32	3.73	3.58	3.52	3.27
CaO	1.72	1.36	2.01	2.31	1.83	2.83	2.19
K_2O	.19	.03	.06	.02	.02	.01	.01
Total	99.46	99.76	99.42	99.73	99.79	99.77	99.73
PPM (averages of triplicate analyses)							
Be	.6	.6	.4	1.3	.9	.7	1.1
Ga	24	34	23	24	30	21	26
Cr	178	265	222	229	217	287	252
V	96	145	110	161	153	121	118
Li	28	15	21	17	18	20	29
Ni	19	54	20	23	24	58	55
Co	48	87	55	62	67	64	70
Cu	18	83	27	28	75	38	98
Y	598	610	993	577	520	923	1135
Sc	307	380	360	322	437	385	270
Zr	330	353	411	244	365	276	330
Sr	7	5	7	7	8	6	10
Ba	17	9	8	7	6	6	10
Rb	7	4	3	8	8	3	3

	T-2311-1A	T-232B-1B	T-232-1C	T-249-1	T-292+1	T-293-3	T-296-1A
SiO_2	37.24	37.26	37.15	37.61	37.85	37.46	37.21
TiO_2	.08	.26	.20	.12	.01	.17	.22
Al_2O_3	21.23	21.12	21.06	21.23	21.60	21.20	20.94
Fe_2O_3	.08	.10	.01	.19	.00	.00	.03
FeO	33.79	33.03	34.39	34.49	32.52	34.71	35.20
MnO	2.06	1.80	1.53	.53	1.10	.48	.32
MgO	3.46	3.57	3.52	4.28	6.02	4.60	3.86
CaO	1.78	2.54	1.58	1.18	.79	1.20	1.72
K_2O	.02	.02	.01	.08	.02	.03	.04
Total	99.74	99.70	99.45	99.71	99.91	99.85	99.54
Be	1.5	1.0	1.1	1.3	1.8	.9	1.0
Ga	29	24	26	37	24	21	28
Cr	645	265	343	198	260	253	143
V	155	108	127	118	91	103	132
Li	26	31	22	46	25	33	29
Ni	109	30	27	64	40	60	23
Co	72	58	46	142	62	113	105
Cu	27	17	16	48	26	35	39
Y	592	1128	933	503	513	338	270
Sc	567	342	317	369	500	350	258
Zr	224	403	345	197	345	222	216
Sr	3	4	8	7	7	7	7
Ba	4	15	9	20	9	8	8
Rb	8	9	5	4	6	3	4

TABLE 6

GARNET - MOLECULAR FORMULAE

as done in D.H. & Z. p.515; on the basis of 24 (O) (48 anionic charges)

		79		78		120		290	
Tetrahedral	Si	5.922	6.000	6.070	6.070	5.985	6.000	6.006	6.006
	Al	.078		-		.015		-	
Octahedral	Al	3.940	3.964	3.990	4.017	4.007	4.057	3.993	4.007
	Ti	.012		.024		.032		.005	
	Fe^{+3}	.012		.003		.018		.009	
Interstitial	Fe^{+2}	4.543	5.893	4.755	5.935	4.600	5.927	4.541	5.973
	Mg	.899		.824		.799		.891	
	Mn	.156		.121		.179		.144	
	Ca	.295		.235		.349		.397	

		150		156A		156B		231	
Tetrahedral	Si	6.014	6.014	6.013	6.013	5.971	6.000	5.989	6.000
	Al	-		-		.029		.011	
Octahedral	Al	3.974	4.011	3.990	4.020	3.973	4.047	4.011	4.033
	Ti	.034		.017		.050		.009	
	Fe^{+3}	.003		.013		.024		.012	
Interstitial	Fe^{+2}	4.548	5.955	4.406	6.040	4.562	6.039	4.538	5.952
	Mg	.858		.836		.783		.830	
	Mn	.236		.212		.217		.280	
	Ca	.313		.486		.377		.304	

	232B		232C		249		292	
Tetrahedral								
Si	5.980)	6.000	5.992)	6.000	6.017)	6.017	5.985)	6.000
Al	.020)		.008)		-)		.015)	
Octahedral								
Al	3.974)	4.028	3.996)	4.029	4.002)	4.043	4.022)	4.023
Ti	.042)		.032)		.019)		.001)	
Fe^{+3}	.012)		.001)		.022)		.000)	
Interstitial								
Fe^{+2}	4.429)	5.987	4.634)	5.960	4.610)	5.902	4.294)	5.993
Mg	.853)		.847)		1.018)		1.419)	
Mn	.246)		.208)		.072)		.147)	
Ca	.439)		.271)		.202)		.133)	

	293		296	
Tetrahedral				
Si	5.903)	6.000	5.994)	6.000
Al	.097)		.006)	
Octahedral				
Al	3.839)	3.869	3.972)	4.010
Ti	.030)		.035)	
Fe^{+3}	.000)		.003)	
Interstitial				
Fe^{+2}	4.541)	5.888	4.734)	5.996
Mg	1.078)		.923)	
Mn	.066)		.044)	
Ca	.203		.295	

TABLE 7

MUSCOVITE COMPOSITIONS IN WEIGHT PERCENT AND PPM

	79	78	120	290	150	156A	156B
SiO_2	56.00	45.98	45.82	52.14	50.34	46.40	51.34
TiO_2	.63	.92	1.27	1.36	1.29	1.23	1.56
Al_2O_3	27.30	34.26	33.50	31.10	30.41	33.08	30.49
Fe_2O_3	.48	.53	1.04	.92	.49	1.04	.80
FeO	1.27	1.02	.86	.93	1.36	.90	.95
MnO	.005	.003	.005	.003	.007	.007	.005
MgO	.94	.89	.85	.87	.99	.95	.92
CaO	.97	.04	.06	.18	1.22	.05	.28
Na_2O	1.90	1.26	.81	.63	2.19	.73	.75
K_2O	5.29	9.00	9.51	6.79	7.16	9.33	6.79
H_2O+	3.48	4.07	4.39	3.64	2.65	4.24	4.08
H_2O-	.12	.10	.12	.28	.15	.08	.19
BaO	.1	.5	.4	.2	.3	.7	.4
Total	98.49	98.57	98.64	99.04	98.56	98.74	98.56
PPM							
Be	4.2	7.8	8.8	5.0	6.5	6.5	4.9
Ga	61	95	65	68	89	66	57
Cr	460	773	517	517	595	610	578
V	466	643	493	918	840	640	463
Li	44	20	22	39	36	34	35
Ni	26	20	11	11	21	20	26
Co	13	27	13	10	9	12	15
Cu	64	42	67	63	38	47	55
Y	49	87	58	46	65	69	65
Sc	116	100	78	88	174	130	78
Zr	375	154	109	156	154	201	245
Sr	249	413	233	162	305	172	143
Rb	109	135	145	122	104	161	204

	231	232B	232C
SiO_2	48.16	46.92	44.04
TiO_2	5.37	1.53	3.88
Al_2O_3	31.54	32.88	31.09
Fe_2O_3	.49	1.18	1.89
FeO	1.42	.85	1.36
MnO	.013	.003	.005
MgO	1.01	.91	1.20
CaO	.34	.15	.10
Na_2O	.65	.66	.82
K_2O	4.94	9.11	8.13
H_2O+	3.83	4.56	4.61
H_2O-	.34	.06	.23
BaO	.2	.7	.9
Total	98.30	99.51	98.26
PPM			
Be	6.2	5.7	6.5
Ga	57	68	78
Cr	486	682	700
V	567	487	535
Li	105	23	20
Ni	34	23	18
Co	14	12	7
Cu	70	35	45
Y	95	71	61
Sc	194	142	168
Zr	602	384	267
Sr	97	171	35
Rb	60	237	232

TABLE 8

MUSCOVITE-MOLECULAR FORMULAE on the basis of 44 anionic charges Foster (1960)

		79		78		120		290	
Tet.	Si	7.098	8.000	6.221	8.000	6.187	8.000	6.764	8.000
	Al	.902		1.779		1.813		1.236	
Oct.	Al	3.197	3.6375	3.665	4.1043	3.514	4.0205	3.514	4.0143
	Ti	.060		.095		.139		.141	
	Fe^{+3}	.068		.053		.100		.091	
	Fe^{+2}	.133		.115		.097		.101	
	Mg	.179		.176		.170		.167	
	Mn	.0005		.0003		.0005		.0003	
Interlayer									
	Na	.464	1.498	.003	1.916	.211	1.898	.156	1.212
	K	.852		1.552		1.636		1.121	
	Ca	.133		.005		.009		.025	
	Ba	.0049		.026		.021		.010	

		150		156A		156B		231	
Tet.	Si	6.652	8.000	6.257	8.000	6.768	8.000	6.323	8.000
	Al	1.348		1.743		1.232		1.677	
Oct.	Al	3.392	3.9148	3.512	4.0358	3.487	4.0035	3.203	4.1264
	Ti	.130		.127		.154		.530	
	Fe^{+3}	.047		.105		.079		.047	
	Fe^{+2}	.150		.101		.102		.160	
	Mg	.195		.190		.181		.185	
	Mn	.0008		.0008		.0005		.014	
Interlayer									
	Na	.560	1.970	.191	1.839	.189	1.406	.165	1.051
	K	1.210		1.604		1.189		.827	
	Ca	.185		.007		.039		.047	
	Ba	.015		.037		.020		.012	

MUSCOVITE-cont.

		232B		232C	
Tet.	Si	6.277	8.000	6.043	8.000
	Al	1.723		1.956	
Oct.	Al	3.458	4.0033	3.067	4.0625
	Ti	.153		.400	
	Fe^{+3}	.117		.190	
	Fe^{+2}	.095		.155	
	Mg	.180		.245	
	Mn	.0003		.0005	
Interlayer					
	Na	.171	1.779	.210	1.705
	K	1.550		1.430	
	Ca	.022		.015	
	Ba	.036		.050	

TABLE 9

BIOTITE COMPOSITIONS - WEIGHT PERCENT AND PPM

	T-78-1	T-79+1	T-120-1A	T-290-1A	T-150+1B	T-156-1A	T-156-1A
SiO_2	35.80	34.67	33.64	35.98	35.83	36.37	36.47
TiO_2	2.22	1.75	2.32	2.84	2.81	2.68	3.01
Al_2O_3	19.39	19.59	19.55	19.36	19.53	19.48	19.88
Fe_2O_3	.72	.95	3.17	.68	.83	1.00	.80
FeO	18.77	20.17	18.00	18.86	18.41	17.73	17.78
MnO	.019	.061	.034	.032	.050	.033	.055
MgO	9.46	9.74	8.87	8.64	8.95	8.93	7.89
CaO	.04	.07	.10	.03	.11	.04	.10
Na_2O	.32	.19	.16	.21	.30	.24	.21
K_2O	9.16	6.74	6.53	8.79	8.96	9.12	8.42
H_2O+	3.63	5.71	5.68	3.46	3.41	3.45	3.55
H_2O-	.13	.12	.50	.22	.17	.11	.23
BaO	.2	.2	.1	.2	.3	.4	.4
Total	99.86	99.96	98.65	99.30	99.66	99.58	98.80
PPM							
Be	7.3	5.3	5.3	8.6	9.9	6.3	5.9
Ga	63	59	65	65	73	63	55
Cr	735	653	468	301	770	920	678
V	653	693	640	913	1000	725	777
Li	42	73	67	62	107	90	93
Ni	315	236	138	124	214	193	165
Co	137	110	99	92	150	145	119
Cu	35	31	117	88	29	79	78
Y	101	115	107	88	96	87	96
Sc	41	98	63	89	125	73	88
Zr	156	285	200	331	261	131	135
Sr	8	7	19	4	10	4	7
Rb	220	55	262	291	288	254	355

	T-231-1A	**T-232-1B**	**T-232-1C**	**T-249-1**	**T-292-1**	**T-293-3**	**T-296-1A**
SiO_2	36.42	35.78	37.03	35.68	37.19	36.16	36.11
TiO_2	2.94	3.06	2.70	3.64	3.32	4.21	3.73
Al_2O_3	19.71	19.48	19.62	20.91	19.54	19.12	18.88
Fe_2O_3	.93	1.51	1.49	.61	.42	.43	.84
FeO	17.75	17.73	17.54	18.59	15.57	18.60	19.81
MnO	.054	.062	.118	.030	.059	.021	.007
MgO	8.43	8.61	7.01	7.11	9.95	7.89	6.93
CaO	.06	.09	.23	.14	.09	.17	.08
Na_2O	.24	.20	.28	.19	.16	.24	.19
K_2O	8.58	8.67	6.51	5.19	9.12	8.81	9.16
H_2O+	3.53	3.56	4.45	6.38	3.19	3.07	2.86
H_2O-	.29	.34	.68	.24	.23	.22	.19
BaO	.2	.5	.5	.2	.1	.3	.2
Total	99.13	99.59	98.16	98.91	98.94	99.24	98.99
PPM							
Be	5.2	5.6	6.8	12	6.1	10.9	19.3
Ga	57	39	64	47	50	69	96
Cr	740	700	800	665	790	775	680
V	873	557	637	750	770	1303	1800
Li	78	65	87	91	90	128	312
Ni	251	163	123	343	212	376	294
Co	161	107	73	189	98	222	174
Cu	50	276	149	23	39	42	57
Y	102	119	242	125	89	88	97
Sc	253	80	90	37	73	41	29
Zr	235	573	442	133	89	126	265
Sr	11	9	10	12	9	11	5
Rb	157	270	298	110	300	408	188

TABLE 10

BIOTITE MOLECULAR FORMULAE - on the basis of 44 anionic charges after Foster, 1960

	79		78		120		290	
Tetrahedral								
Si	5.298	8.000	5.396	8.000	5.226	8.000	5.429	8.000
Al	2.702		2.604		2.774		2.571	
Octahedral								
Al	.824	5.943	.895	5.648	.805	5.824	.869	5.597
Ti	.202		.249		.271		.322	
Fe^{+3}	.110		.018		.370		.079	
Fe^{+2}	2.575		2.362		2.327		2.379	
Mg	2.215		2.122		2.047		1.944	
Mn	.002		.002		.004		.004	
Interlayer								
Na	.055	1.384	.090	1.865	.048	1.363	.062	1.769
K	1.311		1.760		1.293		1.690	
Ca	.009		.006		.017		.005	
Ba	.009		.009		.005		.012	

	150		156A		156B		231	
Tetrahedral								
Si	5.388	8.000	5.408	8.000	5.499	8.000	5.475	8.000
Al	2.612		2.592		2.501		2.525	
Octahedral								
Al	.846	5.580	.830	5.436	1.028	5.476	.964	5.534
Ti	.318		.300		.341		.332	
Fe^{+3}	.094		.112		.091		.105	
Fe^{+2}	2.312		2.202		2.238		2.229	
Mg	2.004		1.978		1.771		1.887	
Mn	.006		.004		.007		.007	
Interlayer								
Na	.088	1.839	.069	1.837	.062	1.720	.069	1.744
K	1.717		1.729		1.618		1.644	
Ca	.017		.006		.016		.009	
Ba	.017		.023		.024		.009	

BIOTITE-cont.

		T-232-1B		T-232-C		T-249-1		T-291-1	
Tetrahedral	Si	5.393)	8.000	5.612)	8.000	5.437)	8.000	5.513)	8.000
	Al	2.607)		2.388)		2.563)		2.487)	
Octahedral	Al	.851)	5.546	1.115	5.408	1.191	5.660	.926	5.490
	Ti	.346)		.307)		.416)		.369)	
	Fe	.171)		.170)		.070)		.060)	
	Fe	2.230)		2.220)		2.365)		1.928)	
	Mg	1.940)		1.583)		1.615)		2.199)	
	Mn	.008)		.013)		.003)		.008)	
Interlayer									
	Na	.059)	1.782	.082)	1.406	.055)	1.098	.045)	1.831
	K	1.665)		1.256)		1.006)		1.728)	
	Ca	.029)		.036)		.023)		.013)	
	Ba	.030)		.032)		.014)		.045)	

		T-293-3		T-296-1A	
Tetrahedral	Si	5.436)	8.000	5.473)	8.000
	Al	2.564)		2.527)	
Octahedral	Al	.820)	5.440	.842)	5.438
	Ti	.476)		.426)	
	Fe	.050)		.097)	
	Fe	2.327)		2.507)	
	Mg	1.764)		1.565)	
	Mn	.003)		.001)	
Interlayer					
	Na	.069)	1.803	.055)	1.845
	K	1.688)		1.765)	
	Ca	.028)		.013)	
	Ba	.018)		.012)	

TABLE 11

SILLIMANITE COMPOSITIONS IN PARTS PER MILLION
(averages of triplicate analyses)

	T-79-1	T-78-1	T-120-1A	T-290-1A	T-150-1B	T-156-1A	T-156-1B
Be	1.2	1.2	2.4	2.5	2.1	1.2	1.9
Ga	75	76	80	82	63	51	77
Ti	1090	227	1353	360	753	860	550
Cr	225	392	518	211	262	443	768
V	174	221	287	260	281	142	229
Li	93	52	55	75	70	33	46
Ni	13	8	26	13	48	30	39
Co	4	2	6	3	11	5	4
Cu	31	8	42	20	24	33	18
Y	63	52	138	65	890	52	48
Mn	110	30	34	52	232	42	12
Zr	135	143	1800	222	900	305	653
Sr	315	21	99	10	122	9	16
Ba	3750	54	155	38	121	75	38
Rb	172	4	18	8	6	12	6
Fe	1.5%	.52%	.98%	.76%	1.2%	.76%	.53%

	T-231-1A	T-232-1B	T-232-1C	T-249-1	T-292-1	T-293-3	T-296-1A
Be	6.5	3.3	1.8	3.4	1.3	5.4	4.7
Ga	79	74	72	91	98	117	129
Ti	5000	1483	1.1%	1390	283	1700	675
Cr	560	519	760	522	436	488	472
V	345	352	383	398	329	495	533
Li	76	49	32	102	64	57	42
Ni	47	25	15	16	36	24	18
Co	10	3	6	4	3	4	3
Cu	31	17	26	17	20	13	39
Y	225	40	73	62	53	54	59
Mn	183	32	75	16	13	15	18
Zr	2000	777	1077	228	1850	1133	268
Sr	98	9	9	81	30	10	14
Ba	63	52	138	555	184	70	101
Rb	4	8	10	27	7	21	4
Fe	.97%	.75%	.97%	.76%	.52%	.74%	.97%

TABLE 12

KYANITE COMPOSITIONS IN PARTS PER MILLION
(averages of triplicate analyses)

	T-79-1	T-78-1	T-120-1A	T-290-1A	T-150-1B	T-156-A
Be	1.3	1.0	1.1	1.9	1.2	.6
Ga	29	26	39	36	37	26
Ti	375	90	465	43	67	229
Cr	510	415	485	184	445	280
V	275	217	255	211	264	184
Li	99	11	38	15	30	30
Ni	17	8	24	25	30	6
Co	-	-	-	7	-	-
Cu	7	4	9	4	8	9
Y	57	43	47	34	61	43
Mn	12	9	15	9	9	6
Zr	312	135	165	49	349	111
Sr	116	7	9	7	9	4
Ba	181	7	64	11	31	45
Rb	7	7	7	8	9	9
Fe	.65%	.49%	.73%	.41%	.49%	.47%

	T-156-1B	T-231-1A	T-232-1b	T-232-1C
Be	2.0	1.6	2.3	1.7
Ga	36	39	36	42
Ti	218	509	93	102
Cr	403	475	522	510
V	246	361	217	240
Li	26	18	24	14
Ni	8	15	27	7
Co	-	-	-	-
Cu	8	6	3	6
Y	41	62	32	40
Mn	7	10	9	7
Zr	116	232	100	110
Sr	7	9	9	9
Ba	15	11	9	30
Rb	5	7	10	11
Fe	.57%	.48%	.48%	.59%

APPENDIX II

PART 3.

WHOLE ROCKS ANALYSIS

TABLE 13

WHOLE ROCK - WEIGHT PERCENT

	79	78	120	290	150	156A	156B
SiO_2	60.95	58.80	53.53	69.65	47.45	66.35	63.70
Al_2O_3	19.87	23.36	26.80	16.25	23.20	17.20	21.60
Fe_2O_3	1.45	1.70	1.78	2.20	1.56	1.74	1.92
FeO	5.92	6.24	3.98	3.72	9.76	4.02	2.84
MgO	2.70	2.89	2.80	1.96	4.40	2.80	2.27
CaO	1.49	.76	.81	.94	2.95	.83	1.09
Na_2O	1.78	.83	.88	.65	3.15	.53	1.03
K_2O	2.69	3.12	5.17	2.13	3.89	3.59	2.50
H_2O-	.09	.06	.22	.19	.10	.15	.21
H_2O+	1.81	1.23	2.15	1.51	1.40	1.31	1.59
CO_2	-	-	-	-	.11	-	-
TiO_2	.78	.84	1.46	.61	1.45	.90	.93
P_2O_5	.17	.14	.16	.04	.44	.09	.07
MnO	.08	.08	.06	.07	.18	.06	.02
Total	99.77	100.05	99.62	99.92	100.04	99.57	99.77

	231	232B	232C	249	292	293	296
SiO_2	57.50	63.95	60.95	61.30	57.20	53.60	53.90
Al_2O	22.84	18.81	19.90	20.80	20.80	21.50	17.60
Fe_2O_3	1.33	1.06	1.88	.84	1.09	1.32	1.80
FeO	5.52	4.64	3.24	5.64	6.28	10.60	15.80
Mg O	3.44	2.67	2.60	2.16	3.48	2.99	3.49
CaO	1.10	1.36	1.49	1.04	1.28	1.94	1.76
Na_2O	1.00	1.07	2.27	1.35	2.56	2.29	1.49
K_2O	3.31	3.18	3.74	2.72	4.75	2.77	2.05
H_2O-	.20	.21	.24	.30	.08	.07	.08
H_2O+	1.67	1.39	1.94	2.26	1.16	.85	.39
CO_2	-	-	-	-	-	.02	-
TiO_2	1.38	1.22	1.23	1.25	1.04	1.46	1.04
P_2O_5	.16	.02	.10	.11	.10	.10	.11
Mn O	.09	.05	.05	.04	.09	.11	.16
Total	99.54	99.63	99.63	99.81	99.91	99.62	99.67

TABLE 14

PRECISION AND ACCURACY OF MAJOR ELEMENTS

Analysed at the University of Manitoba

Constituent	Concentration %	Instrument Precision		Accuracy of Replicates
		σ	STD. DEV.	σ
SiO_2	60.0	.12		.20
Al_2O_3	9.0	.05		.13
Fe_2O_3 (Total)	10.0	.017		.03
MgO	4.0	.04		.10
CaO	10.0	.02		.07
K_2O	2.6	.01		.01
MnO	.41	.01		.01
TiO_2	.48	.02		.02
Na_2O	4.20	.01		.05
H_2O (Total)	1.60	.03		.06
CO_2	1.15	.05		.12
P_2O_5	0.20	.01		.01

Appendix III

Trend Surface Data

The equations of the trend surfaces are given in U-V coordinates where U is a axis of NE-SW attitude in figure 17 and U is perpendicular NW-SE. The origin is the upper corner of the rectangle in figure 17. The values of U vary from 0 at the origin to 9.9 at the south western corner. U varies from 0 at the origin to 9.9 at the north eastern corner. The values of the modal abundances are given in appendix II and the positions of the locals from which the samples were collected are given on figure 17.

TABLE 15

Equations of trend surfaces used

Kyanite

linear

$$x = 14.81 - 0.82U - 1.04\ V$$

K-feldspar

linear + quadratic

$$x = -0.84 - 1.45U + 1.58V + 0.21U^2$$
$$-0.02UV - 0.13\ V^2$$

Staurolite

linear + quadratic

$$x = 0.40 + 0.21U - 0.33V - 0.004\ U^2$$
$$-0.02UV + .03V^2$$

Muscovite

linear + quadratic + cubic

$$x = -3.96 + 12.92U + 4.62V - 3.09U^2$$
$$+ 0.14UV - 1.23\ VU + 0.22\ U^3$$
$$-0.05U^2V + 0.001\ V^2U$$
$$+0.07\ U^3$$

Biotite

linear + quadratic + cubic

$$x = 25.08 + 8.34U - 8.03U - 2.86U^2 + 1.56UV$$
$$+ 1.06U^2 + 0.19\ U^3 - 0.12U^2V - 0.12V^2U$$
$$- 0.03\ U^3$$

Quartz

linear + quadratic + cubic

$$x = 21.29 - 20.52U + 12.25U + 5.88U^2 -$$

linear + quadratic cubic

$$x = 1.73UV - 0.93\,U^2 - 0.33U^3 - 0.15\,U^2V$$
$$+ 0.28\,V^2U - 0.02\,V^3$$

Garnet

linear + guadratic

$$x = 12.27 + 0.62U - 4.14U - 0.33U^2$$
$$+ 0.56UV + 0.28\,U^2$$

Sillimanite

linear

$$x = 0.39 + 0.12 + 0.67U$$

Plagioclase

linear + quadratic + cubic

$$x = 41.09 - 7.99U - 9.15V + 1.68\,U^2 + 0.54\,UV$$
$$+ 1.55U^2 - 0.15\,U^3 + 0.13U^2V - 0.19V^2U$$
$$- 0.05V^3$$

TABLE 16

ANALYSIS OF VARIANCE OF MODAL ABUNDANCES KYANITE

	Sum of Squares	Df	Mean Square	F	Confidence level %
Due to Linear	704.436	2	352.2	6.13	99.5%
Deviation from Linear	2239.689	39	57.4		
Due to Quadratic	727.495	3	242.747	4.16	97.5%
Deviations from Quadratic	2216.63	38	58.2		
Due to Cubic	1133.854	4	283.4	5.78	99.8%
Deviations from Cubic	1810.288	37	49.0		

K-FELDSPAR

Due to Linear	196.642	2	98.321	3.6	95%
Deviation from Linear	1063.988	39	27.3		
Due to Quadratic	306.406	3	102.102	4.35	99%
Deviation from Quadratic	954.224	38	25.1		
Due to Cubic	484.501	4	121.12	5.77	99.9%
Deviations from Cubic	776.138	37	21.0		

QUARTZ

1st trend - Due to Linear

2nd trend - Due to Quadratic

Deviations not given in computor program

Cubic - Due to	3035.294	4	758.8	2.83	95%
Deviations from Cubic	9921.789	37	268.0		

SILLIMANITE

Due to Linear	135.569	2	67.784	5.35	99%
Dev. from Linear	493.501	39	12.65		
Due to Quadratic	170.030	3	56.610	4.70	99%
Dev. from Quadratic	459.040	38	12.05		
Due to Cubic	240.734	4	60.18	5.89	99.9%
Dev. from Cubic	388.335	37	10.5		

PLAGIOCLASE

Due to Linear	212.593	2	106.29	.673	25%
Deviation from Linear	6168.952	39	158.0		
Due to Quadratic	478.413	3	159.4	1.025	40%
Dev. from Quadratic	5903.133	38	155.5		
Due to Cubic	834.859	4	208.46	1.41	75%
Dev. from Cubic	5546.650	37	148.		

BIOTITE

Due to Linear	266.44	2	133.22	1.058	50%
Dev. from Linear	4919.803	39	126.1		
Due to Quadratic	306.7	3	102.2	.799	25%
Dev. from Quadratic	4879.543	38	128.		
Due to Cubic	927.884	4	231.67	1.80	75%
Dev. from Cubic	4258.378	37	115.1		

MUSCOVITE

Due to Linear	596.93	2	298.26	1.792	75%
Deviations from Linear	6482.29	39	166.3		
Due to Quadratic	918.84	3	309.26	1.97	75%
Dev. from Quadratic	6160.37	38	162.		
Due to Cubic	1639.62	4	409.9	2.73	95%
Deviations from Cubic	5439.60	37	147.0		

GARNET

Due to Linear	84.53	2	42.26	.8	25%
Dev. from Linear	2103.475	39	54.0		
Due to Quadratic	722.016	3	240.972	6.23	99.5%
Dev. from Quadratic	1465.990	38	38.6		
Due to Cubic	764.256	4	191.064	4.96	99.5%
Dev. from Cubic	1423.748	37	38.5		

STAUROLITE

Due to Linear	.977	2	.488	3.392	95
Dev. from Linear	4.068	39	.144		
Due to Quadratic	3.293	3	1.096	23.75	>99.99
Dev. from Quadratic	1.752	38	.0462		
Due to Cubic	4.158	4	1.		100%
Dev. from Cubic	0.887	37			

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LEGEND

Mappable major geological units

- ▨ gabbro-Southern province
- ▨ quartzite and schist-Southern province
- ▨ granite and olivine gabbro-between the provinces
- ▨ mylonite and mylonite gneiss-between the provinces
- ▨ mainly gneiss-Grenville province
- ▨ mainly garnet amphibolite with interlayered amphibolite, quartzite and pelitic schist-Grenville province

Structural symbols

●-------first fold axes

◉-------poles to first fold axial planes

○-------second fold axes

◎-------poles to second fold axial planes

□--(or F if more than 50 poles)---poles to foliation

■--(or L if more than 50 readings)--lineations

▼--(contoured if more than 50 poles)--poles to joints

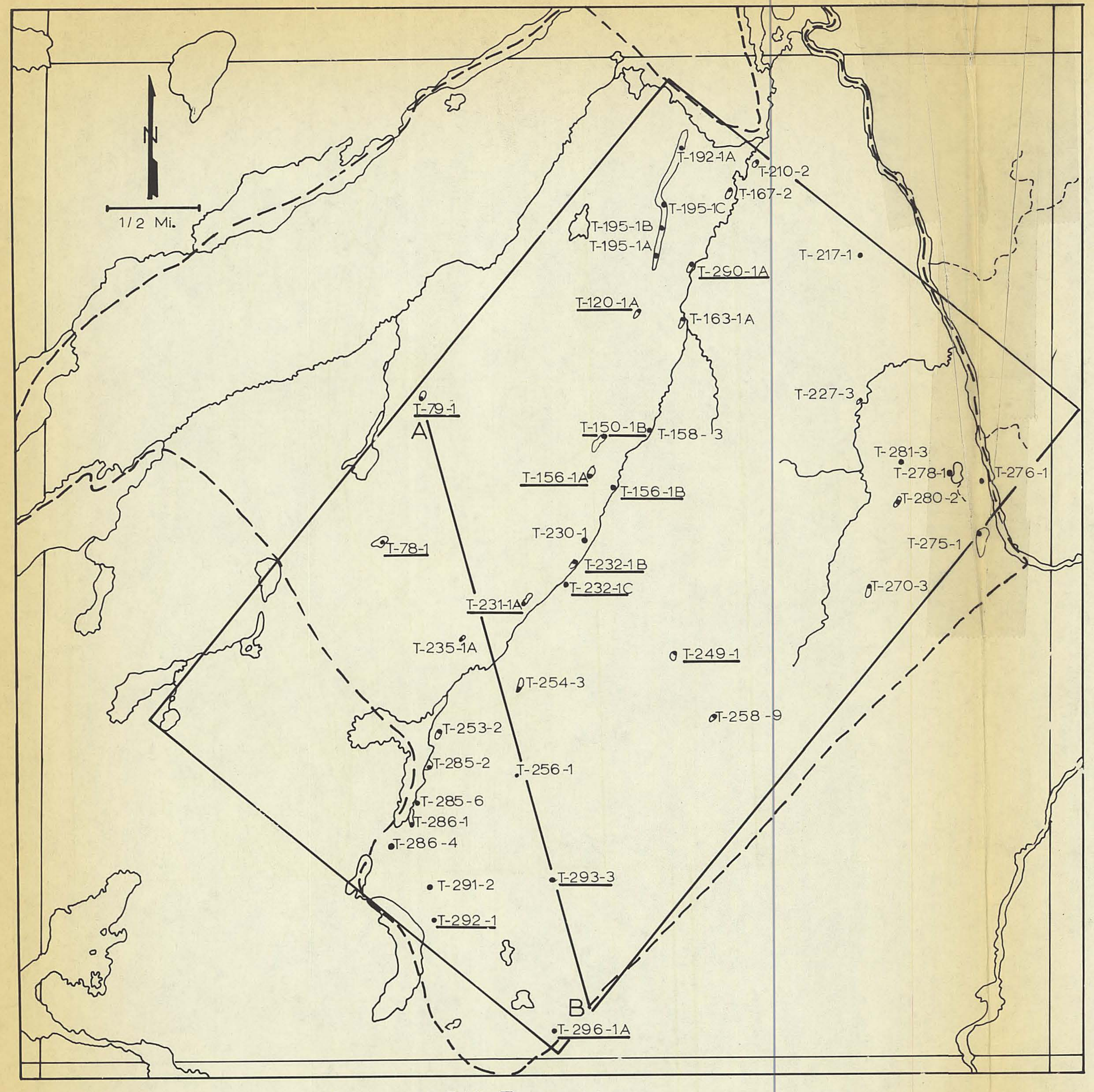
X-------poles to gash veins

▲-------poles to kink zoes-all in mylonite, nort corner of area C

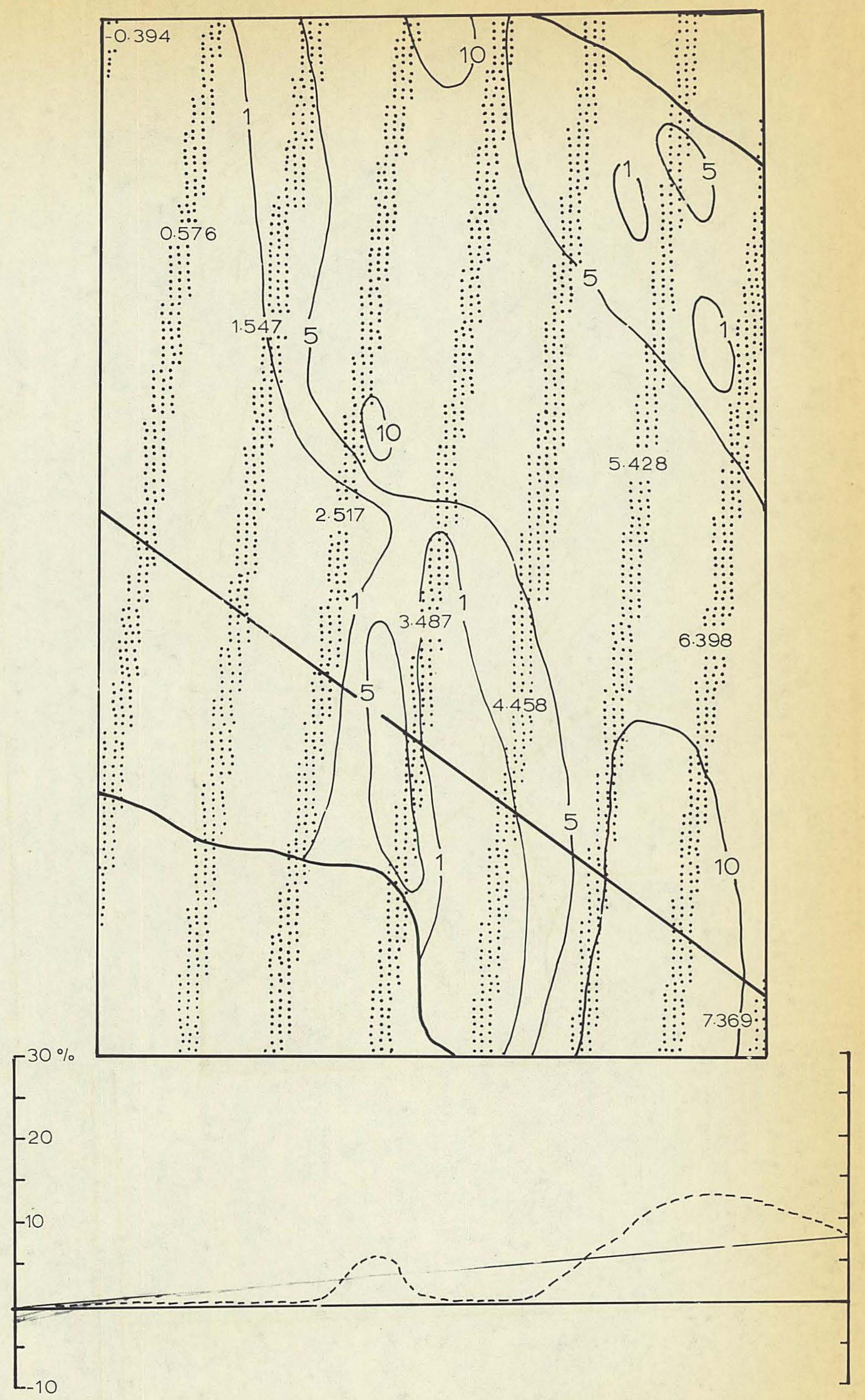
Map symbols

streams and laes

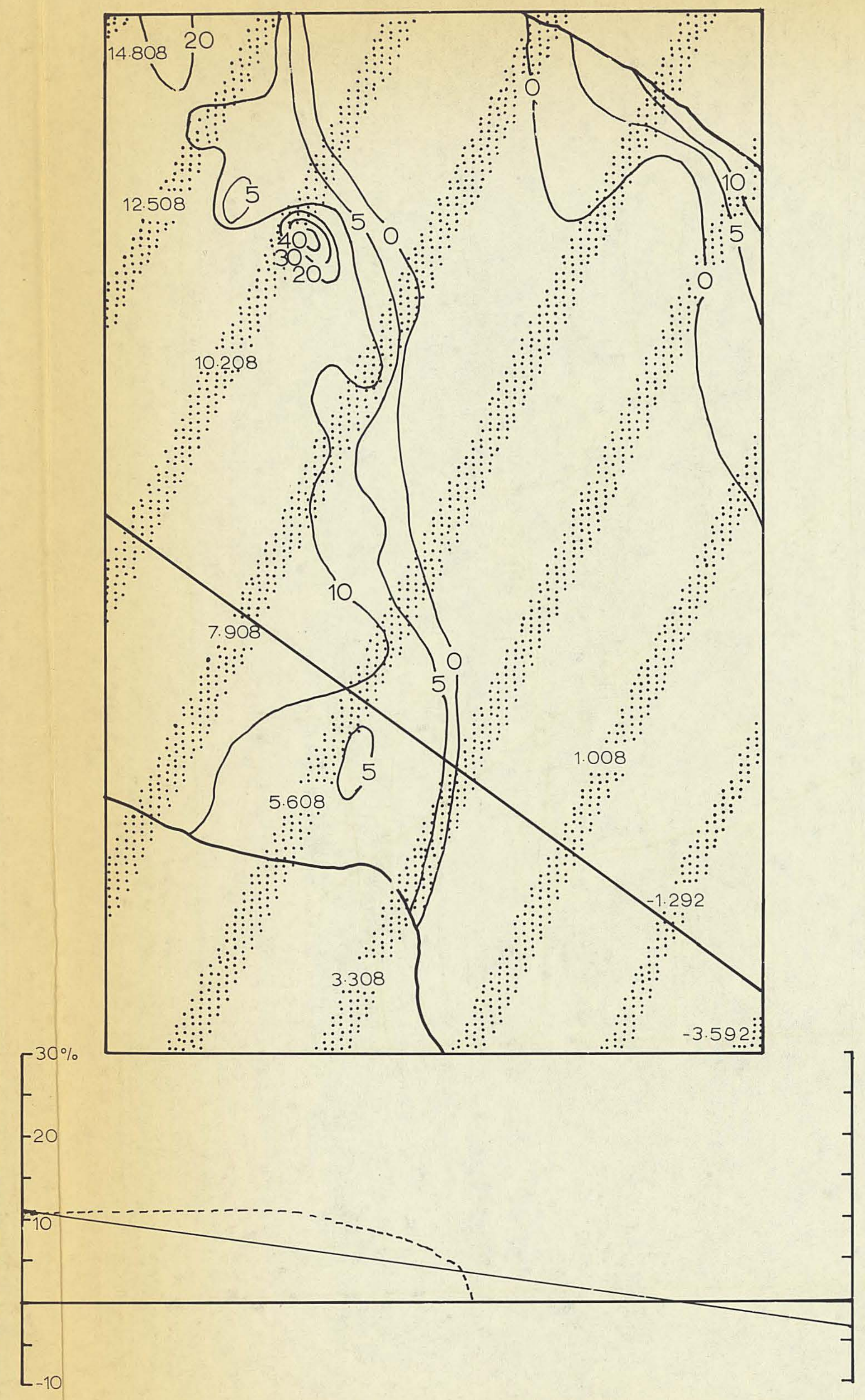
north pole of streographic projection



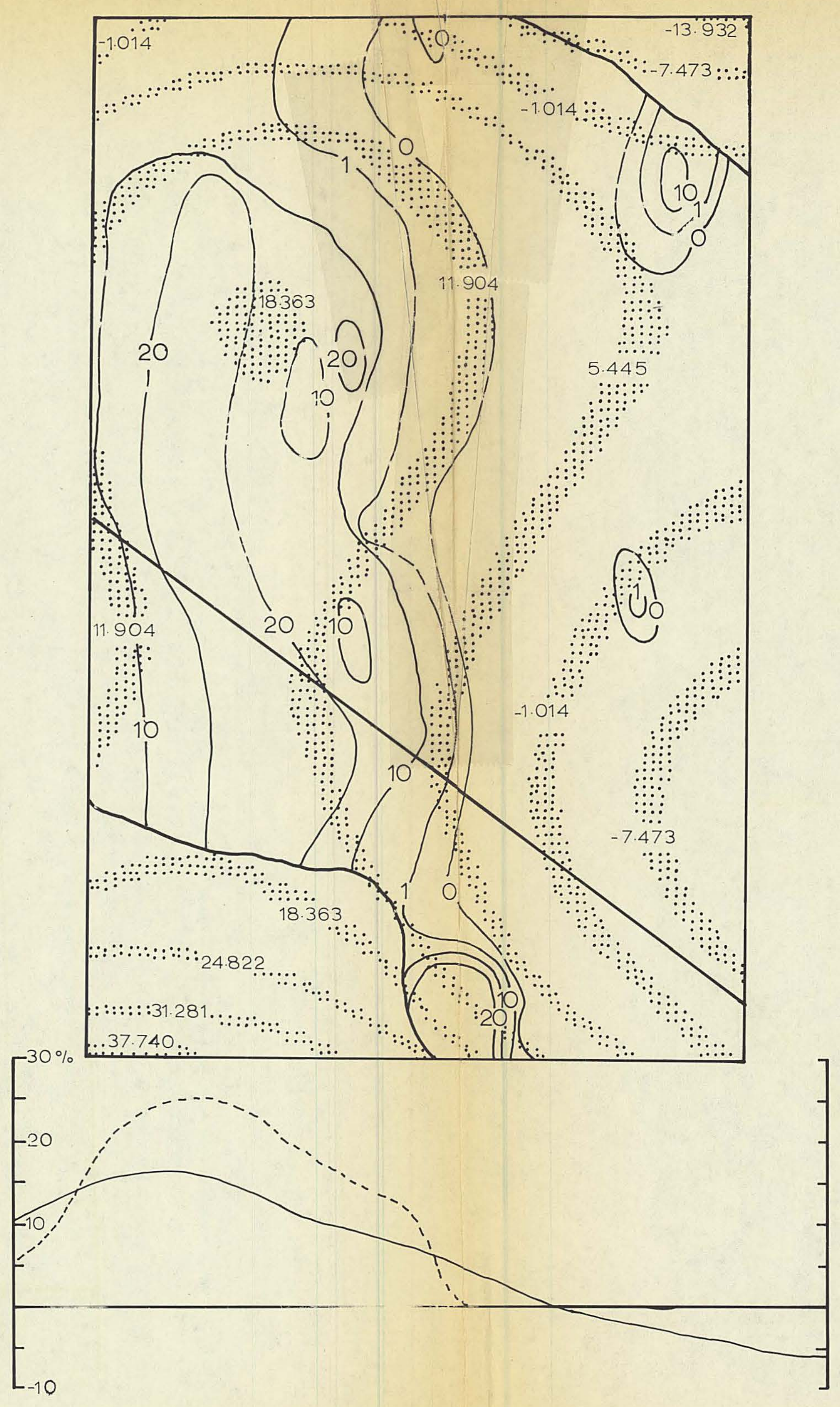
Map of Dill Township



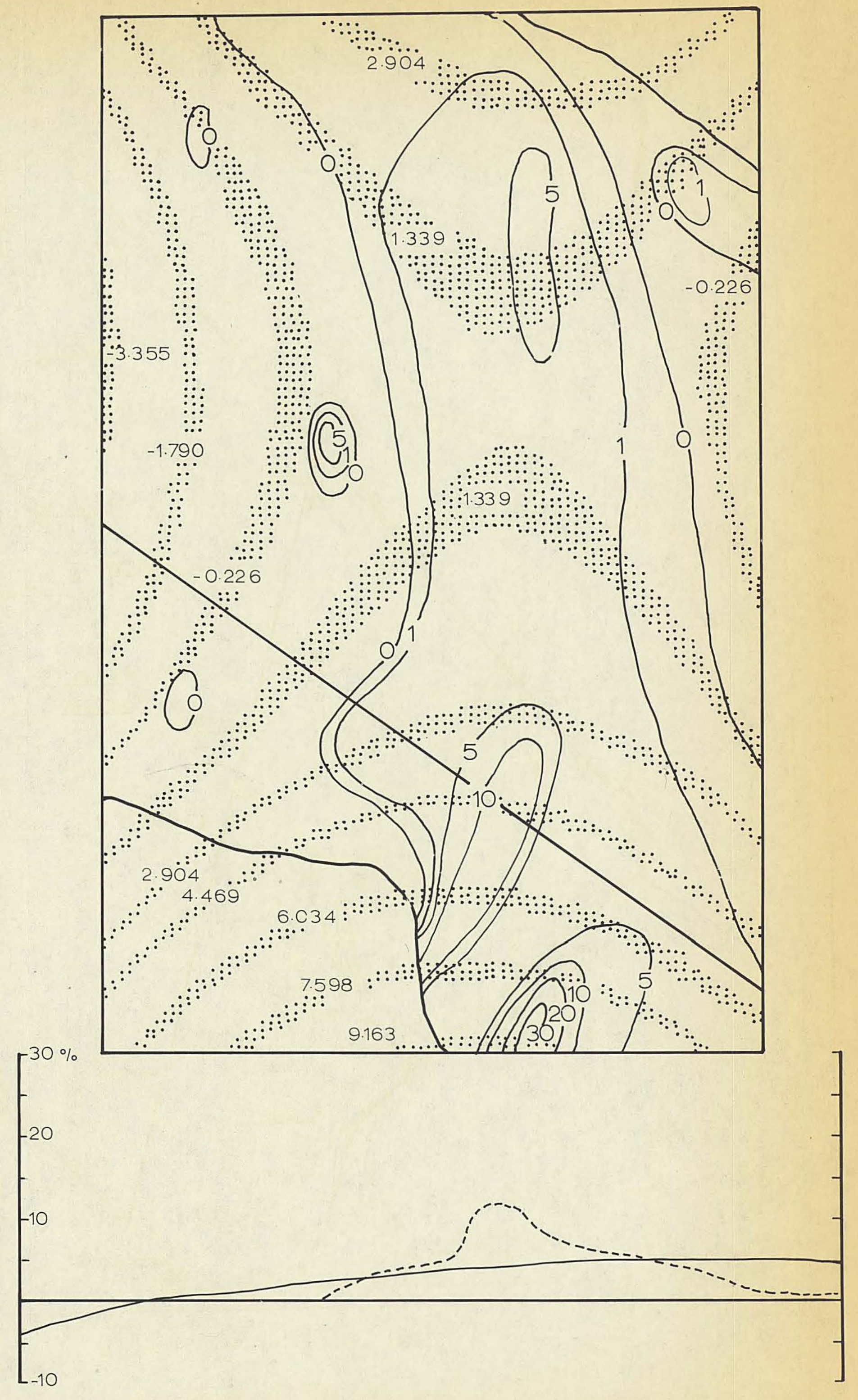
Sillimanite



Kyanite



Muscovite



Potash Feldspar

HAND AND TREND SURFACE CONTOURS FOR THE MODAL DISTRIBUTION OF MINERALS IN ALUMINUM SILICATE BEARING PELITIC SCHISTS IN PART OF DILL TOWNSHIP

Legend

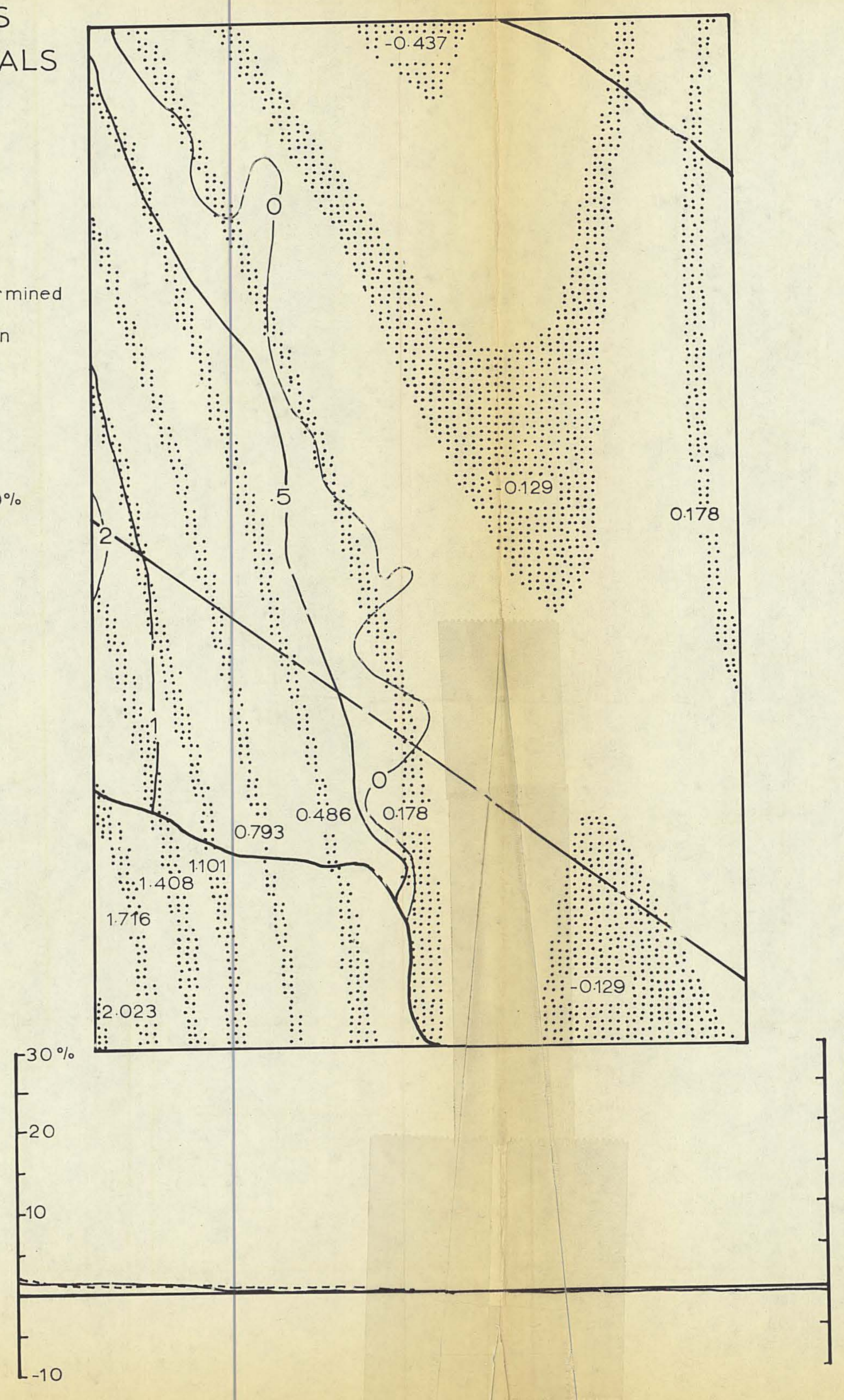
A. Map of Dill Township

— Rectangular area from which trend surfaces were determined
– – Area mapped
A—B Line from which sections below contoured maps were taken
T-78-1, etc. Number and location of samples used
T-78-1, etc. Samples used for chemical analysis

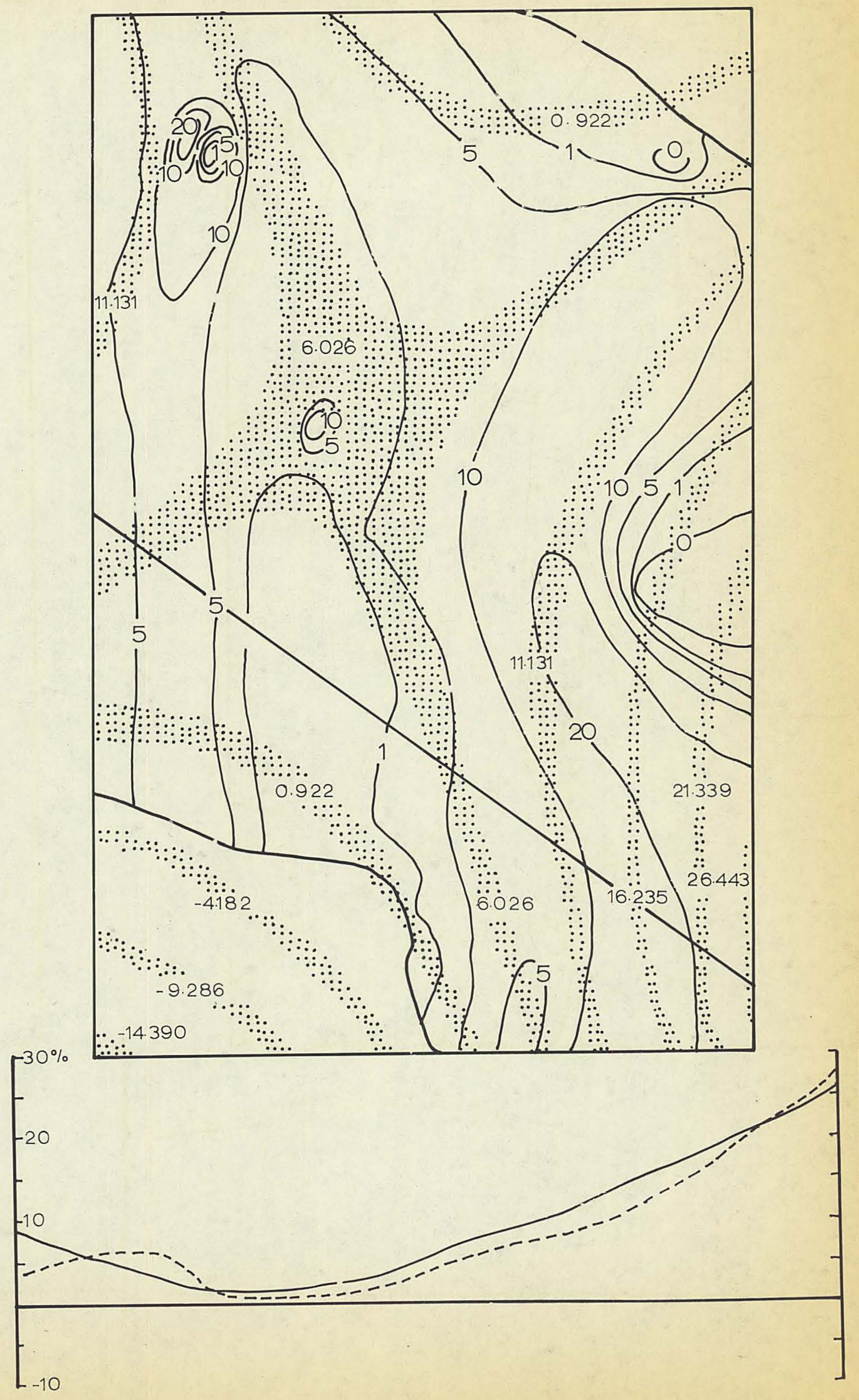
B. Mineral Distribution Maps

1. MAPS
- Hand contoured
- Contours on machine computed trend surfaces
  - the surface was chosen which first displayed a 99% confidence level or was the maximum surface the program could compute
  - for equations, confidence levels, etc. see appendix

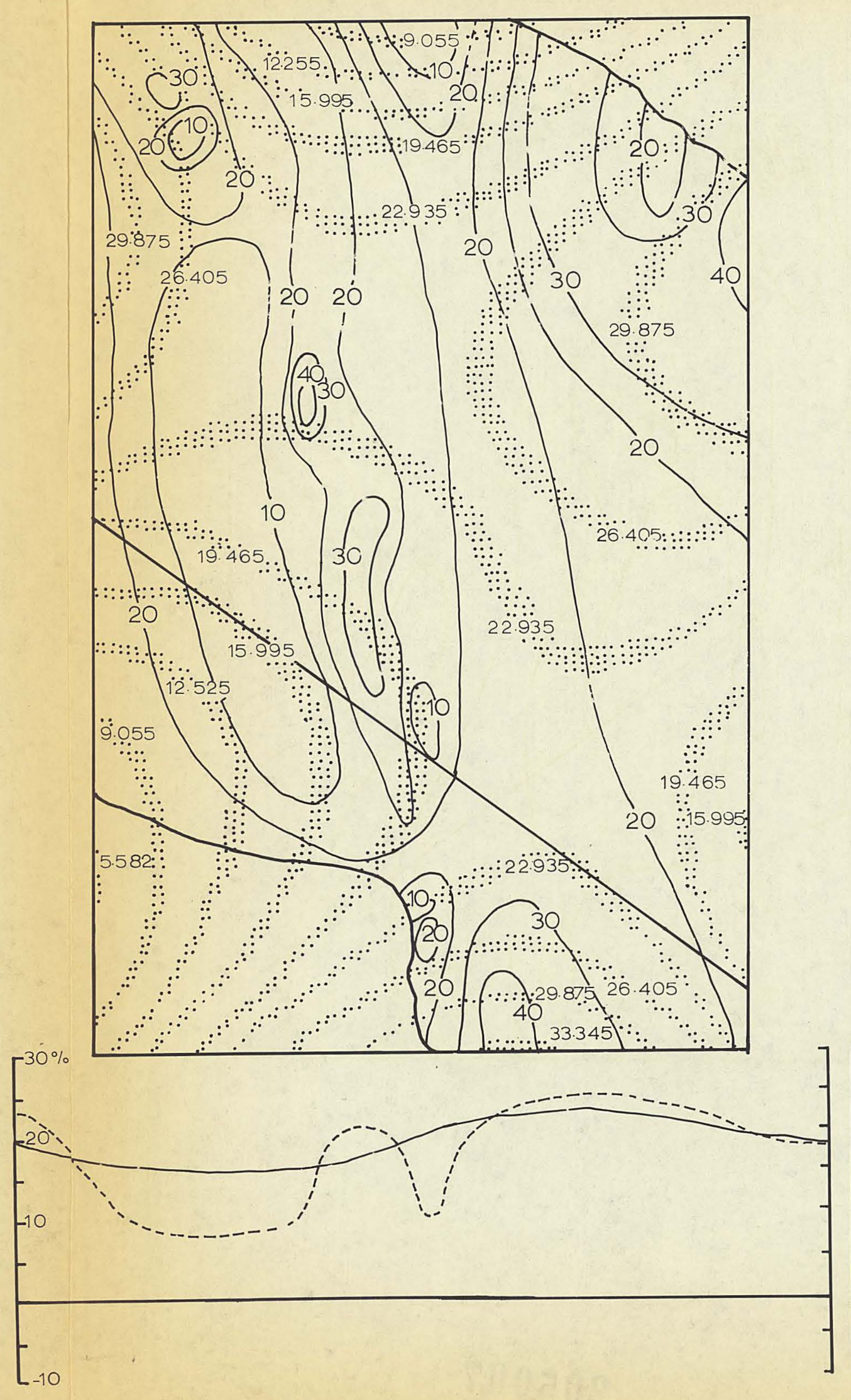
2. SECTIONS
- Hand contoured
- Contours on trend surface



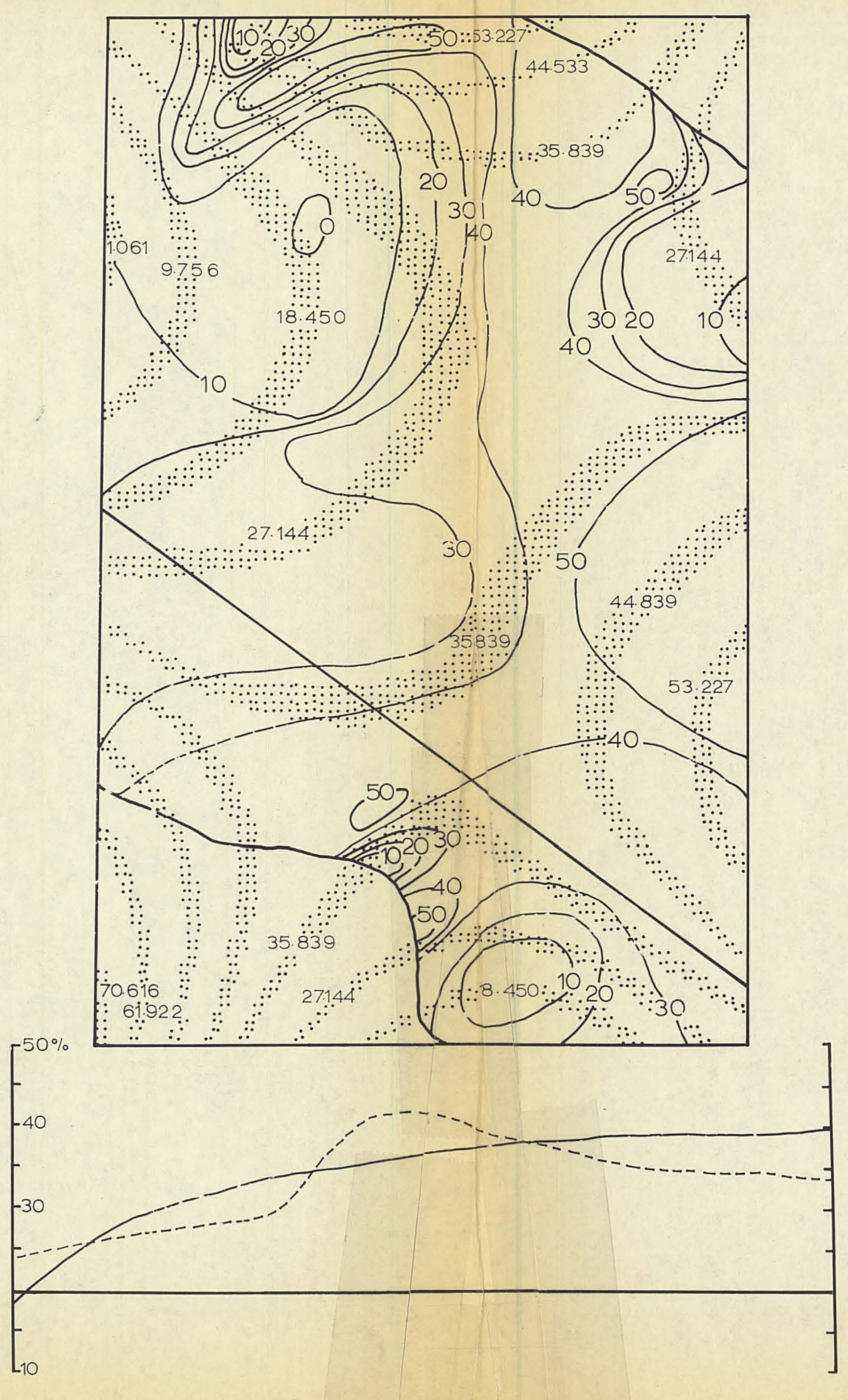
Staurolite



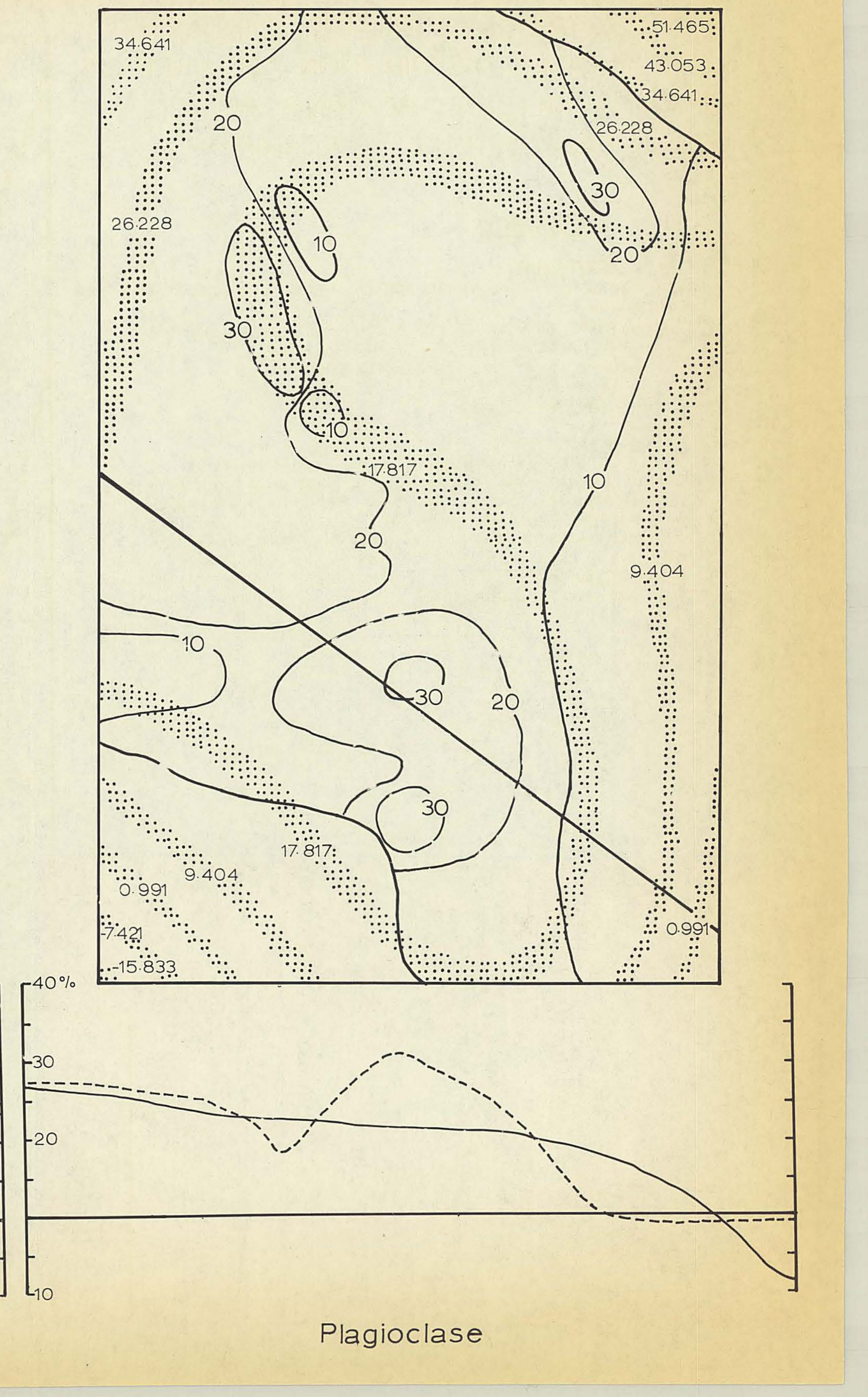
Garnet



Biotite



Quartz



Plagioclase

LITHOLOGIC MAP
OF PART OF DILL TOWNSHIP

Legend

Grenville Province

1	Ultrabasic Intrusions
2	Basic Intrusions
3	Pegmatites
4	K-feldspar gneiss
5	K-feldspar-hornblende-biotite gneiss
6	Quartz-feldspar gneiss
7	Muscovite - K-feldspar Schist
8	Muscovite-kyanite schist
9	Sillimanite-K-feldspar schist
10	Biotite schist
11	Anthophyllite amphibolite
12	Diopside amphibolite
13	Garnet amphibolite
14	Common amphibolite
15	Calc-silicate marble

Rock Units Between Provinces

16	Mylonite
17	Granite and quartz monzonite
18	K-feldspar - quartz (metasomatized ?) gabbro
19	(Olivine ?) gabbro

Southern Province

20	Diabase dikes
21	Gabbro
22	Schist and/or phyllite
23	Schistose quartzite
24	Quartzite

Symbols

Contact - good control
Contact - poor control
Outcrop area greater than 50%
Lakes
Railway

Scale: miles
0 1/4 1/2 3/4 1