The petrology and structure of the north eastern Neria area, south-west Greenland

Preston, R. M. F.

How to cite:

Preston, R. M. F. (1969) *The petrology and structure of the north eastern Neria area, south-west Greenland*, Durham theses, Durham University. Available at Durham E-Theses Online: http://etheses.dur.ac.uk/9275/

The Petrology and Structure of the North-Eastern Neria Area, South-West Greenland

by

R.M.F. Preston, B.Sc. Hatfield College

A thesis submitted for the degree of Doctor of
Philosophy in the University of Durham, May 1969

Abstract

The rocks of two areas of the pre-Cambrian basement of S.W. Greenland are described. The suite of plutonic rocks comprises gneisses, amphibolites, ultra-basic rocks, schists, and granites. Particular attention is devoted to the petrology and geochemistry of the amphibolites and it is shown that mineralogical differences between the amphibolites of the two areas cannot be attributed to differences in bulk rock composition. The origin of the amphibolites is discussed in the light of X-ray fluorescence rock analyses and it is concluded that they are probably igneous in origin.

The amphibolites of one area contain stable garnet and clinopyroxene, but garnet and stable clinopyroxene are not found in the amphibolites of the other (none of the amphibolites contains stable orthopyroxene). This difference in mineralogy is interpreted as being the result of "wet" metamorphism near a large autochthonous granite, and "dry" away from it. The grade of metamorphism is revealed by the presence of sillimanite in the schists as being of the upper amphibolite facies. The presence of cordierite in the same rock shows that pressure was only moderate.

The country rocks were migmatised throughout what was probably a long period of plutonism. The plutonism ended with the formation of a large autochthonous granite. Small allochthonous granites surround the autochthonous mass.

The large and small scale structures of the areas are described, and the deformational history elucidated as far as is possible. Three periods of plutonic deformation are recognised, with the main migmatisation having occurred during the second.

Numerous dykes, some of which are fresh olivine dolerites and others metadolerites, intrude the plutonic rocks and these also are briefly described.

Acknowledgements

My thanks are due to my supervisor, Dr. C.H. Emeleus, for obtaining permission for me to work in Greenland and for subsequent valuable assistance and discussion of the manuscript, and to the then Head of Durham University Geology Department, Dr. K.C. Dunham, for permission to work in the Department and for assistance in obtaining a grant. I am also very much indebted to Dr. K. Ellitsgaard-Rasmussen, Director of the Greenland Geological Survey, for permission to work in Greenland and for the facilities provided for field work there. I also benefited from discussion with Survey geologists in Copenhagen. I was supported during the three years of this research by a Science Research Council N.A.T.O. Studentship which enabled me to spend one year at Aarhus University, Denmark. I thank Professor A. Berthelsen for permission to work in Aarhus, and Dr. F. Kalsbeek for much assistance while I was working there and for hospitality at the time of a later visit. Finally I wish to thank the technical staff of Durham University Geology Department for the preparation of thin sections and photographs.

CONTENTS

Page

Petrography of Unlaminated Amphibolites

Petrography of the Ultra basic rocks

Page

Dyke Petrography

Chapter Ten - Structure

Planar structures

Faults and mylonites

List of Figures

List of Tables

INTRODUCTION

Although most of Greenland is covered by ice-cap - generally called Inlandice by the Danes - there is a stretch of ground around the margin free of permanent ice, except for small local ice-caps and glaciers leading down to the fjords. The ice-free margin may be up to 200 km. wide, but 50 km. is a more usual width. The present study has been carried out on two parts of this margin in the Neria Area S.W. Greenland; Figures 1 and 2 show the regional and local geography respectively. The smaller area mapped occupies approximately 70 sq. km. at the end of a peninsula between two fjords some 35 km. south-south-east of Frederikshåb. The larger is an inland area of about 200 sq.km. and is situated 40 km. east of the coastal area.

The investigation was carried out as part of the GGU* programme of systematic mapping in S. Greenland. The Survey has, since 1956, mapped many thousands of square kilometres between Kap Farvel and Frederikshåb on a scale of 1:20,000. The work is now being extended to the north of Frederikshåb.

The field work occupied the summers of 1964, '65 and '66 with three months being spent in the field

*Grønlands Geologiske Undersøgelse (Geological Survey of Greenland)


81°
80°
64°
52°
40°
24°
12°
0°
81°
ELLESMERE ISLAND
Peary Land
Nyeboe Land
Washington Land
Independence Fjord
Danmarks Fjord
Nord
KRONPRINS CHRISTIANS LAND
GRØNLANDSHAVET
78°
78°
INGLEFIELD LAND
THULE
Jøkelbugt
GERMANIA LAND
Danmarks Havn
75°
75°
MELVILLE BUGT
Shannon
Clavering Ø
Andrées Land
Kejser Franz Josephs Fjord
72°
72°
UPERNAVIK
Traill Ø
Mesters Vig
Kong Oscars Fjord
SVARTENHUK
Jameson Land
Renland
SCORESBYSUND
UMANAK
NUGSSUAQ
QUTDLIGSSAT
DISKO
Gaaseland
69°
69°
GODHAVN
DISKO BUGT
JAKOBSHAVN
EGEDESMINDE
CHRISTIANSHÅB
Agto
Kangerdlugssuaq
DAVISSTRÆDET
HOLSTEINSBORG
DANMARKSSTRÆDET
66°
66°
Søndre Strømfjord
SUKKERTOPPEN
ISLAND
ANGMAGSSALIK
GODTHÅB
63°
63°
Fiskenæsset
Ravns Storø
TINGMIARMIUT
GRØNLAND
FREDERIKSHÅB
Neria
KILOMETRES
0
100
200
300
400
500
IVIGTUT
NARSSAQ
60°
60°
JULIANEHÅB
NANORTALIK
Kap Farvel
52°
48°
40°
36°
32°
28°

Fig.1

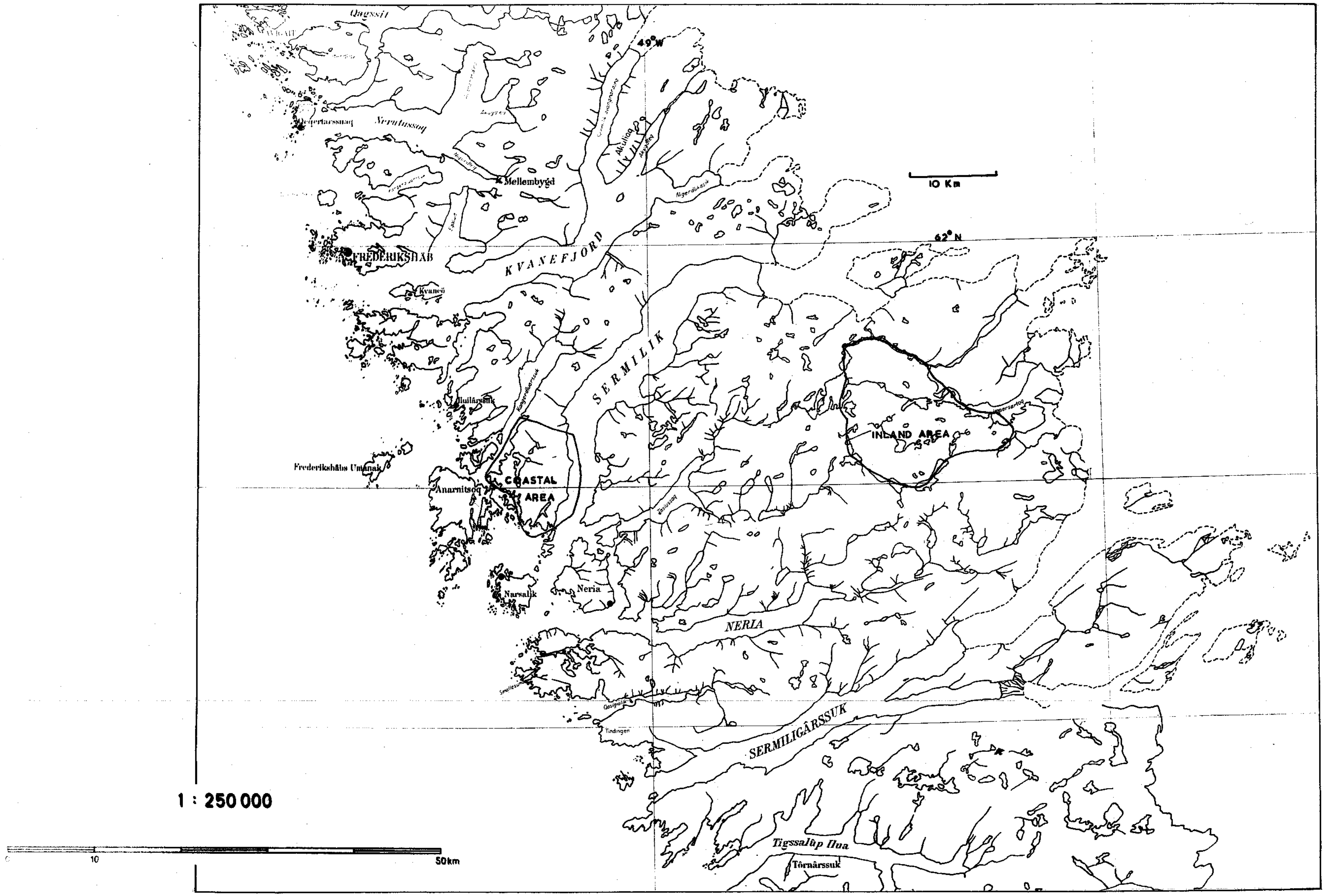


Figure 2: The Frederikshab district

each season. Most of the time was spent on the inland area. The coastal area extends up to 600 m. altitude, and the inland area ranges from 600 m. to 1400 m. Below approximately 600 m. heather, willow scrub, and moss grow thickly in the valleys and the rock faces are usually encrusted with lichens. The lichens can seriously impede observation of the details of the rocks, and at low altitudes most of the very detailed observations must be made on a clear zone which occurs along the shore between the land and sea floras. Above 600 m., however, the vegetation is very sparse and where there are no boulder fields or local ice-caps the exposure is magnificent.

Previous work on the areas is negligible and nothing has been published. A small-scale reconnaissance map produced by A. Berthelsen showed the main structures of the Coastal Area, but not of the Inland Area. This map was made from aerial reconnaissance and aerial photographs and was not published. Geologically, therefore, the areas were very much terrae incognitae.

Regional Setting of the Geology

Until the publication in 1938 of C.E. Wegmann's work on the structural divisions of S.W. Greenland very little was known about the large scale structural units of the territory. The work of N.V. Ussing in 1912 and earlier pioneers was much more local and petrological in aspect than Wegmann's.

Wegmann divided the Pre-Cambrian rocks of S.W. Greenland into two main groups and called them Ketilidian and Gardar. The Gardar rocks consist of almost unmetamorphosed sediments, volcanics and intrusives which overlie and intrude the much older Ketilidian rocks. The place of the Gardar rocks in the chronology of the region is not much questioned, but the chronology of the Ketilidian and pre-Ketilidian rocks is not completely understood.

The Ketilidian was divided into two groups by Wegmann: a basement migmatite infrastructure, and an overlying, slightly metamorphosed suprastructure. Originally Wegmann considered the infrastructure to be the gneissified equivalent of the supracrustals. However, when mapping started at the head of Sermiligarssuk Fjord (fig.2) in 1964 the supracrustals were found to overlie the gneisses with a strong unconformity. Thus it became clear that the gneisses are pre-Ketilidian in age. The details of the chronology of the Sermiligarssuk-Ivigtut area have been given by Higgins and Bondesen (1966) and Windley, Henriksen, Higgins, Bondesen, and Jensen (1966). The first paper traces the development of ideas about the chronology since Wegmann's account. Apart from dykes and granites the rocks of the Neria Area, i.e.

N. of Sarmiligarssuk Fjord, are all pre-Ketilidian in age. The terms supracrustal and infracrustal are frequently used by Scandinavian geologists to describe rocks formed at different levels in the Earth's crust. Supracrustal rocks are those that were formed at the surface of the Earth. They include sediments and volcanics. Supracrustals may be slightly metamorphosed, but they will not have been gneissified. Infracrustal rocks are those formed under plutonic conditions and occur as basement rocks of an area. Infrastructure and suprastructure are the names given to the structural units formed by the infracrustal and supracrustal rocks.

South of Sermiligarssuk the gneisses and supracrustals have been affected by metamorphic events of Ketilidian and Sanerutian ages. Sanerutian is the name given to a period of plutonism which occurred later than the Ketilidian, but its exact status is not yet decided. The effects of these two events increase to the south and as a result the differences between the infrastructure and the suprastructure have become blurred. It is for this reason that there was early confusion about the relationship between them. In the infrastructure the increase in metamorphism is shown by changes in the basic dykes which intrude the gneisses. Dykes in the Neria Area are almost completely undeformed, but may be slightly metamorphosed. Southwards, however, they become progressively more deformed and metamorphosed, until south of Ivigtut the only remnants of the dykes are sheared out amphibolite lenses concordant to the gneiss structures.

Although there are no visible effects of the Ketilidian orogeny in the Neria Area (with the possible exception of the weak metamorphism of the pre-Gardar dykes), K/Ar age determinations give ages of around 1800 m. y. (Larsen, 1966). This date is only slightly older than that of the Ketilidian orogeny, which has been dated at around 1600 m.y (ibid). The significance of the 1800 m.y. date is not yet understood, but it may correspond to a separate thermal event that affected the rocks of the Godthåbsfjord area (McGregor, 1966), or it may be an intermediate date formed by slight heating of the rocks by the Ketilidian metamorphism and a partial expulsion of the argon. Further north, from around Fiskenaesset and Godthåb (fig.1), K/Ar dates of 2300 m.y. and 2600 m.y. (Larsen, 1966) have been obtained from rocks belonging to the same group as the pre-Ketilidian rocks of the Neria Area. Rb/Sr whole rock ages should become available in the future and from these it should be possible to obtain the age of formation of the Neria gneisses.

The metadolerite dykes of the Neria Area are thought to be pre-Ketilidian in age as only two dykes are known to cut the unconformity between the infracrustals and supracrustals, even though there are dozens of dykes in the infracrustals to the north and west. In the Midternaes area at the head of Sermiligarssuk Fjord where the unconformity is least deformed there are unfortunately very few dykes and none can be traced up to the unconformity. However, if the dykes be Ketilidian or younger a number should be found in the Ketilidian rocks. Further south, in Kobberminebugt, dykes have been found in the Ketilidian rocks and a complicated history of metamorphism, deformation and granitisation has been elucidated by Watterson (1965), but these dykes are usually only 1 - 2 m. thick whereas those in the infrastructure are up to 50 m. thick. Therefore correlation is difficult, but it would appear that Watterson's first period dykes are equivalent to the two dykes seen cutting the border between the infrastructure and suprastructure and are probably younger than those below the unconformity.

CHAPTER ONE

STRATIGRAPHY

Four main rock types make up the country rock of the areas studied. They are, in decreasing order of abundance, acid gneiss, amphibolite, garnet-mica-schist, granite and pegmatites. There are also ultra-basic lenses and calc-silicate rocks, but these are present in much smaller amounts. Hornblende and biotite gneisses are by far the most abundant rocks, comprising approximately 80-90% of the total rock mass. On account of this, detailed stratigraphy could not be worked out because the gneiss is very uniform in composition over the area as a whole, while being very heterogeneous on the hand specimen scale. However, a broad division into major stratigraphical units can be made.

The rocks of the Inland Area can be divided into several large stratigraphical units which are also found in the surrounding areas mapped by other geologists. In the Coastal Area two main divisions can be erected. Correlation between the two areas cannot yet be carried out with very much certainty.

Inland Area

Six major stratigraphical units are recognised in the Inland Area:

Hornblende Gneiss	highest
Boundary Layer Amphibolite	
Upper Biotite Gneiss	
Laminated Amphibolite	
Brown Schist	
Lower Biotite Gneiss	lowest

The terms "upper" and "lower" refer only to structural position and must not be considered as indicating the relative ages. There is no evidence available at present that indicates the direction of younging.

Hornblende Gneiss

This occurs in the north-western corner of the Inland Area and on the large scale forms a synform that closes to the north-east. The synform is at least 50 km. long and strikes approximately N.E. - S.W.

Boundary Layer Amphibolite

This amphibolite is one of the many concordant amphibolites occurring in the gneiss succession, but it is given a name because it is particularly important, coming as it does between the Hornblende Gneiss and Upper Biotite Gneiss (See Map I). It was named by Dr. Zdenek Misar, who mapped the area to the west where it is particularly well developed. In the Inland Area the Boundary Layer Amphibolite is only a few metres thick and is often boudinaged, but elsewhere it becomes much thicker and may have ultra-basic rocks associated with it (personal communications from other GGU geologists).

Upper Biotite Gneiss

This forms a very large unit between the Boundary Layer Amphibolite and the Laminated Amphibolite. It is also seen in grid squares D4 and G5 (Map I) where it outcrops in the synform between the Laminated Amphibolite and the Brown Schist horizon, but here it is much thinner.

In the main development of the Upper Biotite Gneiss a progressive change across the strike can be observed. In the upper part near the Boundary Layer Amphibolite several concordant amphibolites occur, and two or three of them have associated ultra-basic lenses. Lower in the succession the amphibolite horizons become thinner and less abundant and do not have any ultra-basic rocks.

It is probable that there is a major break in the Upper Biotite Gneiss, but at the present state of knowledge it is best to consider it as one unit. The reasons for postulating a break and its nature and position will be discussed later.

Laminated Amphibolite and Brown Schist

These usually occur together and provide the best marker horizon in the whole succession. As can be seen from Map I they define a south-westerly plunging antiform and an adjacent synform to the east. It is probable that there is only one brown schist horizon, but this cannot be proved beyond doubt. To the north-east M. Henri Masson, who has mapped that area, has concluded on structural grounds that there is only one brown schist horizon. In the Inland Area the evidence points to the same conclusion. Brown schist is not found in the north-west development of the Upper Biotite Gneiss whereas if there were two schist horizons the eastern one should be repeated to the north-west. Furthermore, although laminated amphibolite is not found all the way along the strike of the eastern brown schist horizon it is present in places (such as in grid square E1, Map I). Thus it is most likely that there is only one brown schist horizon. It is on this assumption that the interpretation of the structure is based.

Lower Biotite Gneiss

This outcrops in the core of the major antiform and in the east of the Inland Area. In the core of the antiform no amphibolite horizons occur, but in the eastern development of the Lower Gneiss amphibolites are found. There are some layers of hornblende gneiss in this succession, but they are very much subordinate to the biotite gneiss.

Coastal Area

Only two stratigraphical units are recognised in the Coastal Area:

Hornblende Gneiss
Biotite Gneiss

Most of the Coastal Area lies within the Hornblende Gneiss Succession. Biotite-gneiss is found only in the northern part. An amphibolite horizon with ultra-basic lenses separates the two. This horizon is probably the continuation of the Boundary Layer Amphibolites along the northern side of the Hornblende Gneiss synform.

Correlation between the two areas

Correlation is very tenuous at the present state of knowledge, but it is possible that the amphibolite between the two gneiss units of the Coastal Area may be the same as the Boundary Layer Amphibolite of the Inland Area, and that the Hornblende Gneiss successions in both areas are the same, and that the Biotite Gneiss of the Coastal Area correlates with the upper part of the Upper Biotite Gneiss. The justification for this proposal is the similarity between the coast succession and the upper part of the inland one coupled with the knowledge that the Hornblende Gneiss occupies a large synform which runs between the two areas. However, only when the geological map of the whole Neria Area becomes available will it be possible to put this very tentative correlation on a firm basis.

CHAPTER TWO

GNEISSES

Types of Gneiss

A dual classification of the gneisses is employed: a broad petrographical division into hornblende - or biotite-bearing gneiss, and a structural classification that can be applied to both types. The structural classification is one that has been evolved over several years by GGU geologists and is used because petrographical differences between gneisses are often so small as to be unusable in the field. The details were given by Berthelsen (1961), but the main types are described below.

Banded gneiss is by far the most common type and is made up of alternating, well defined layers of mafic and felsic rock. The bands can vary in thickness from millimetres to several tens of centimetres in thickness. The mafic layers have an appreciable proportion of dark mineral whereas the felsic layers may be almost devoid of mafic minerals. Fig. 3 shows typical banded gneiss.

Veined gneiss occurs commonly and is characterised by many irregular sub-parallel felsic veins composed of quartzo-feldspathic material. The veins are usually concordant, but may be cross-cutting to the gneiss structures.

Fig. 3 Hornblende gneiss showing banding on several scales. Hornblende Gneiss Succession, grid square 1G, Map II.

Both these varieties of gneiss may be deformed on the mesoscopic scale (Turner and Weiss, 1963,p.15).

Agmatitic gneiss is gneiss that has a breccia like appearance with a felsic groundmass enclosing irregular, more basic fragments. Either or both of the parts of this rock type has a preferred orientation, usually a foliation.

Nebulitic gneiss is gneiss in which the lithological differences have become vague and ill-defined.

Homogeneous gneiss has no banding but has a preferred orientation of the platy or prismatic minerals. As the name suggests there is no lithological variation. This can be an original feature or can be caused by a completion of the processes which gave rise to nebulitic gneiss.

Biotite Gneiss

Varieties other than the normal banded gneiss generally do not form large areas, but occur only locally. In grid squares A6 and B6 (Map I) the gneiss is more finely banded than usual and in places has almost completely lost its banding and is only foliated. This is the largest area of homogeneous gneiss, but it is only 3-4 sq.km. in extent. An area of this size composed of homogeneous gneiss probably represents part of the original succession that was particularly uniform in composition and which has been preserved. Other occurrences of homogeneous gneiss are much smaller. They generally occupy a few hundred square metres or less, and can be seen to have been formed by homogenisation or granitisation of an originally banded or more basic rock. These will be described with the granite of the area.

The other types of gneiss apart from the banded variety also have irregular distributions and their occurrence may be related to local features. Thus agmatitic gneiss occurs along the ridge between grid squares E4 and F4 (Map I). This is caused by veining related to the leucocratic granite which occupies the southern end of the ridge. Similarly, veined gneiss occurs in the deep valley along the south-west side of the Inland Area, and this is associated with the Neria Granite to the south and east.

The mafic and quartzo-feldspathic layers in the banded gneiss are not necessarily of the same thickness at the same locality. Usually the felsic layers are thicker than the mafic. When the banding is on the large scale, i.e. the individual layers are up to 20-30 cm. thick, the contacts between them are usually knife sharp. On the small scale, however, with layers 1-2 mm. or less in thickness there is not such a sharp boundary between them, and the layers may then be composed of relic clots of biotite intergrown with the quartz and feldspar grains. Within the large scale banding there will be a smaller scale banding and this decreasein the scale of the banding can go right down to the scale of individual mineral grains. The mafic layers are generally not banded, but if they are the banding is much more regular and finer than that in the felsic layers. Individual layers cannot be traced for any great distances. They die out, disappear because of folding, or become boudinaged. The biotite-rich and hornblende-rich bands in particular are very susceptible to becoming boudinaged, whereas the felsic bands rarely show this feature. The foliation of the rock is usually parallel to the banding, and this is also found in boudins where the internal foliation follows the curved edge of the boudin. The gneiss foliation outside the boudins also is curved round the boudins and it is clear that the mafic bands were much more competent than the felsic during one period of deformation and as a result they were boudinaged and the felsic rock was then able to flow around them. Complicated internal deformation of the boudins has occurred in a number of localities. The internal foliation and layering is isoclinally folded, usually about axial plains parallel to the regional trend of the foliation. The presence of folds inside a boudin indicates that there was an early period of deformation during which the rock was less competent than it was in the later deformation when it was boudinaged. (Berthelsen, 1957, 1960). The banding is folded on all scales from that of the hand specimen up to that measured in kilometres or even tens of kilometres. Where the bands are thick the folds developed may be metres in amplitude but when they are thin the folds are much tighter and irregular, (fig.4.)

Thus the overall picture that emerges of the gneisses is of a thick succession of quartz-feldspar rock with interlayered mafic bands. The succession is very uniform in composition on the large scale, but on the hand specimen or outcrop scale is heterogeneous. The gneisses are believed to be derived from sedimentary rocks and the banding represents original differences in composition that have been accentuated and separated by metamorphic differentiation and shearing stress. Thus the bands do not correspond to an original banding but they are parallel to the original bedding.

Hornblende Gneiss

The amount of hornblende gneiss in the biotite gneiss succession increases slightly as the Boundary Layer Amphibolite in the north-west is approached, but it is still very much less abundant than biotite gneiss. Hornblende gneiss occurring in the biotite gneiss succession is distinctly different from that of the main hornblende gneiss succession. Where it occurs in the biotite gneiss

Fig. 4 Tight smallfolds in finely banded biotite gneiss. Grid square B6, Map I.

Fig. 5 Very melanocratic part of Hornblende Gneiss succession. Grid square S6, Map II

hornblende gneiss is essentially the same as the biotite gneiss except that it contains hornblende as mafic mineral rather than biotite. The proportion of mafic to felsic material is roughly the same, with the felsic minerals forming the main part of the rock and the hornblende occurring in only minor amounts. In the hornblende gneiss succession proper, however, the mafic minerals are very much more important and the mafic bands often predominate over the felsic. Fig. 5 shows part of the hornblende gneiss that is particularly mafic and irregularly banded. Homogeneous gneiss is absent in the hornblende gneiss succession, but veining and other granitisation phenomena occur.

Gneiss Petrography

There is great uniformity about the petrography of the acid gneisses as quartz and plagioclase usually comprise 70-90% of the rock (all modal percentages are volume percent). The very mafic rich bands have a higher proportion of biotite, but again there is very little variation within the mafic bands. The texture of the gneiss is usually interlobate-hemigranoblastic (Berthelsen, 1960, p.24), but porphyroblasts of feldspar are not uncommon, and where there has been nearby mylonitisation quartz is recrystallised into many small grains. The foliation visible in hand specimen usually cannot be seen

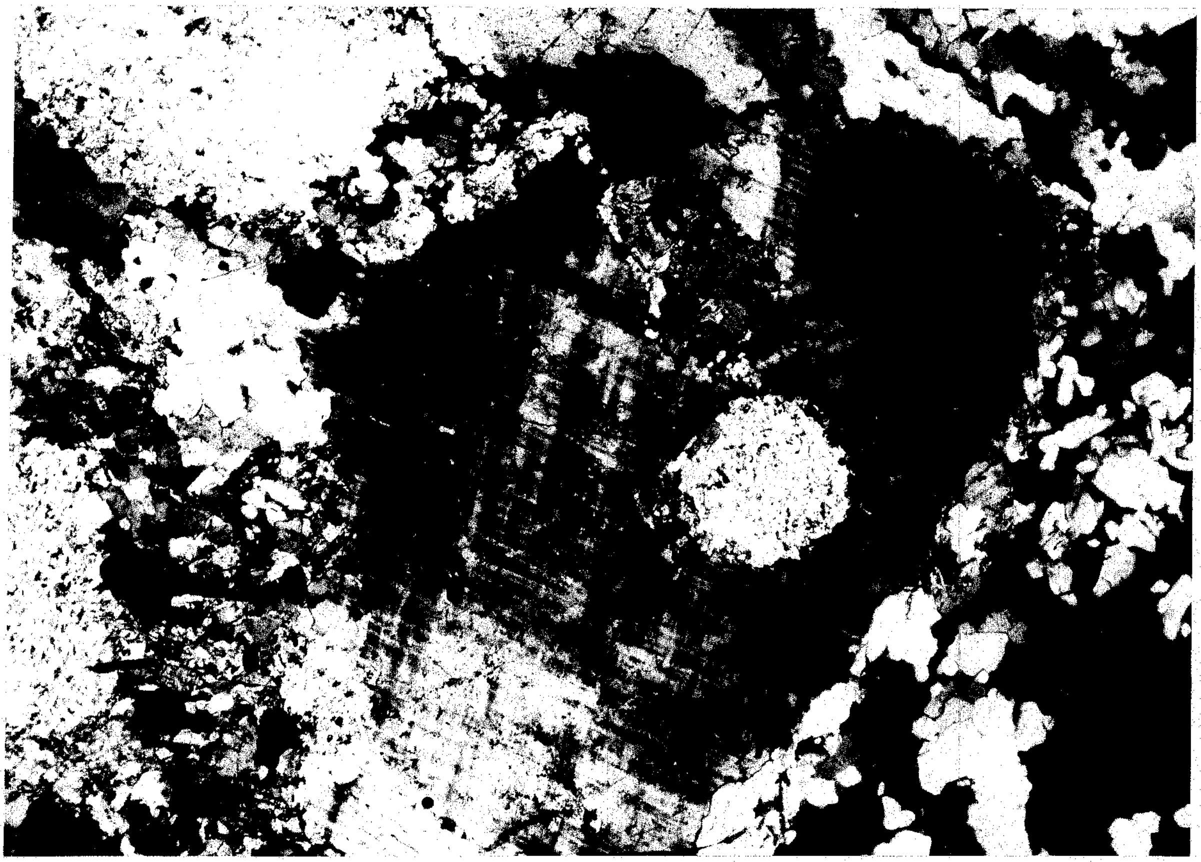


Fig. 6 Microcline porphyroblast enclosing small quartz grains and a sericitised plagioclase grain in biotite gneiss. Specimen 71041, XN, X 10.

in thin sections of felsic gneiss because there is so little biotite, but in the biotite-rich bands there is a stronger planar structure.

Plagioclase

The composition of the plagioclase is between An_{15} and An_{25} in biotite gneiss, but in hornblende gneiss compositions of An_{30} have been recorded. Several initial determinations of plagioclase were done using the universal stage and the curves of Slemmons (1962) and all the grains examined were found to have low temperature optics. Thereafter low temperature optics were assumed and all plagioclase determinations were made by measurement of the extinction angle in sections cut perpendicular to 'a'.

Plagioclase often has the appearance of being porphyroblastic even though it does not form grains significantly larger than the other minerals, usually 1-2 mm. The crystals are rounded and equidimensional, unlike the quartz grains which have very irregular shapes. Several generations of plagioclase can sometimes be distinguished, but there are no systematic compositional differences and no evidence to suggest that they were formed during different plutonic events. The apparent different generations are probably the result of continuous formation over a period of time.

The plagioclase has been altered in many rocks. Saussurite is the usual alteration product, but in a few rocks sericite is the alteration product. The intensity of alteration is variable and has no regularity. It is probably related to the proximity of mylonite and fault zones, but this cannot always be demonstrated.

Albite twinning is present in all the plagioclases of all gneisses, but in some it is not obvious and where alteration has been strong the twinning may be obscured. Carlsbad twinning never occurs.

Alkali Feldspar

Perthitic microcline is present in many of the gneisses, but its presence is apparently haphazard and the distribution of microcline throughout the rock mass as a whole cannot be related to any controlling factor on grounds of spatial distribution. Table I gives the modes of several gneisses and it can be seen that the proportion of microcline ranges from 0-30%.

The microcline crystals range in size from very small irregular interstital patches to large porphyroblasts several centimetres in diameter (fig.6). Rounded relics of plagioclase are often enclosed in large microclines, but they are never in optical continuity with each other or the microcline host. All

Specimen	Plagioclase	Quartz	Microcline	Biotite	Amphibole
71041	34.5	37.7	22.0	6.0	
71050	50.6	33.4	8.5	7.5	
71061	64.6	27.6		7.7	
71069	48.6	34.8			16.6
71079	35.1	34.1	27.4	1.8	
71081	50.4	28.4	9.9	11.3	
71101	46.3	33.6	6.3	13.8	

	SiO_2	Al_2O_3	FeO	CaO	MgO	Na_2O	K_2O
71041	74.2	13.1		1.8		3.6	2.9
71050	72.7	15.2	1.4	2.6	0.6	4.7	1.8
71061	70.2	17.0		3.3		5.7	0.9
71069	74.1	12.0		4.2	2.7	4.5	0.1
71079	74.0	14.0		1.9		3.8	3.5
71081	69.9	16.2	2.1	2.6	0.9	4.7	2.4
71011	71.4	15.6	3.6	2.4	1.4	6.8	2.4

Table I

Modes and calculated oxide analyses of gneisses

Mineral compositions obtained from Deer, Howie and Zussman, 1962.

Plagioclase:	Analysis 12, p. 113, vol.4
Microcline:	Analysis 7, p. 38, vol.4
Biolite:	Analysis 11, p. 64, vol.3
Amphibole:	Analysis 10, p. 252, vol.2

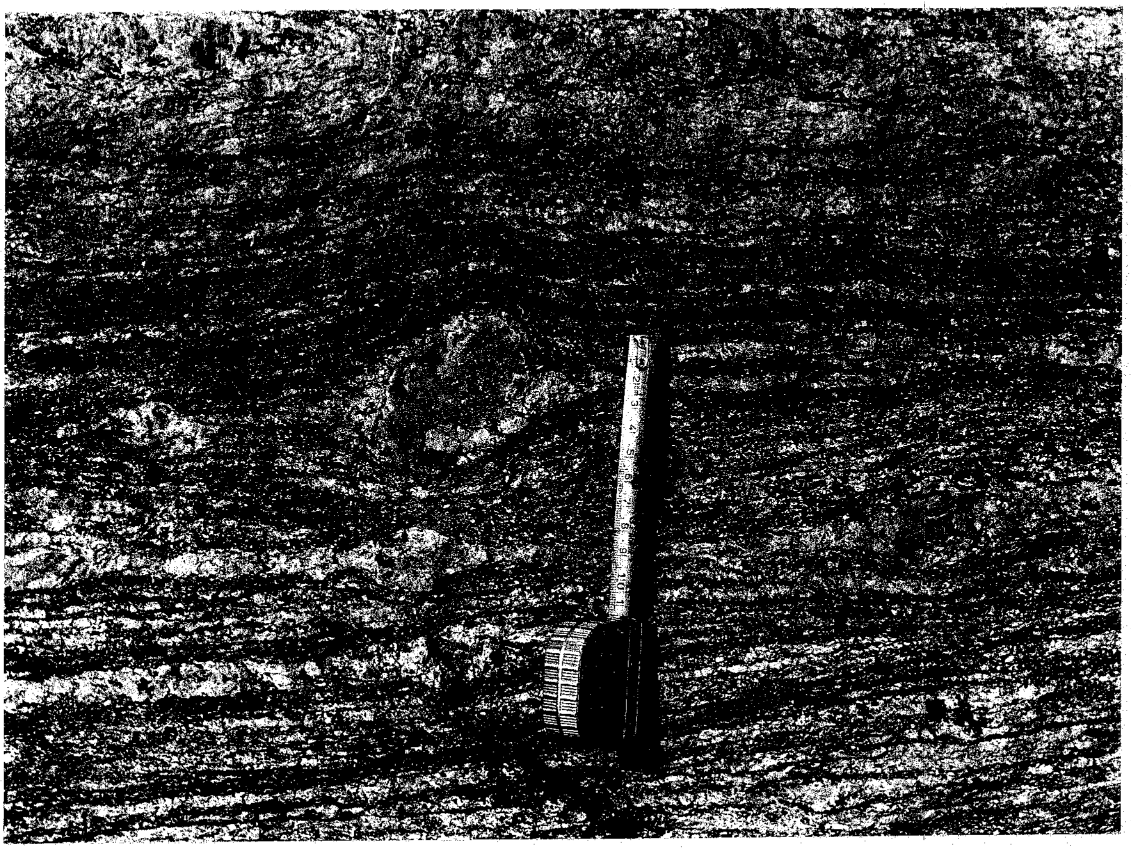


Fig. 7 Large microcline porphyroblast developed across the foliation in biotite gneiss.
U. Biotite Gneiss, grid square E6, Map I.

these features show that the microcline developed later than the plagioclase and gradually enclosed plagioclase grains. The microcline did not develop by replacement of pre-existing plagioclase. In hand specimen the late development is shown by the way large porphyroblasts cut the foliation (fig.7). The foliation is slightly deformed round the porphyroblast but this is probably the result of a very minor shearing after the formation of the porphyroblast. If there had been a major stress imposed after the formation of the porphyroblast it is probable that a pressure shadow would have been formed in which quartz would have crystallised.

Quartz

Quartz forms 30-40% of most of the felsic gneisses and is very variable in grain size. Where rock has been mylonitised quartz has recrystallised into a mass of small granules each less than 1 mm. across. Individual granules are not strained. Most commonly, however, quartz occurs in large, very irregular lobate patches that are highly strained. Plagioclase crystals are usually embayed by the quartz and occasionally trains of small recrystallised quartz grains cut across plagioclases.

Smoothly rounded grains of 'drop' quartz are enclosed in some of the feldspars - usually in the microcline porphyroblasts. These quartz grains are up to 0.5 mm in diameter. It is impossible to say for certain whether these quartz grains crystallised before, during or after the crystallisation of the microcline, but in view of the ease with which quartz becomes strained or recrystallised it is more likely that the drop quartz crystallised contemporaneously with, or shortly after the microcline. If it had been enclosed during the growth of the porphyroblast then strained quartz should also have been enclosed; this is not observed. If the drop quartz had crystallised long after the formation of the microcline, and after all deformation had ceased, then it should also be found outside the porphyroblasts; but it is not. Therefore, it is probable that it was formed during or after the formation of the microcline porphyro-blasts, but before the final deformation occurred.

Biotite

Biotite is the most common mafic mineral but it rarely exceeds 10% of the rock outside the mafic-rich bands. It is present in small shreds and has largely been chloritised. The shreds run in trains between the feldspars and quartz grains and have been curved round the resistant feldspars by subsequent deformation. The shreds of biotite are aligned generally parallel to one another, but they are so small and few that they do not impart a foliation to the rock.

Amphibole

Amphibole-bearing gneisses usually contain hornblende, but occasionally the amphibole is actinolite. In amphibole-bearing gneiss of the biotite gneiss successions the amphibole is usually very poorly preserved. The hornblende or actinolite is often partly replaced by biotite and has had the crystal outlines largely destroyed. In some of these rocks the proportion of mafic mineral is very little greater than it is in the normal biotite gneiss.

In the hornblende gneiss succession, however, the amphibole is quite different. A well preserved dark green hornblende is usually present and the proportion of mafic minerals may be as much as 30-40%. Biotite is also present in some of these rocks and it cuts the hornblende where the two are in contact. There is nothing to suggest, however, that the biotite was formed during a much later period than the hornblende. They are probably contemporaneous and the biotite cuts the hornblende because it has stronger growth capabilities in certain directions than does the hornblende.

Accessories

Epidote is frequently present, either as very small grains formed from alteration of the plagioclase or disseminated throughout the rock as a whole. The latter type occurs as individual grains or as glomeroblasts composed of radiating growths. Vein epidote is abundant and in some of the larger veins crystals up to 1 cm. long have formed.

Apatite and zircon occur in most of the gneisses, and calcite is occasionally present. The zircons are typical of those found in sedimentary rocks (Poldevaart, 1955). They have rounded and sometimes broken ends and are dark brown in colour with scratched surfaces. Since zircon crystals are generally considered to be resistant to the effects of metamorphism their appearance after metamorphism can be used as an indicator of the origin of the original rock. Poldevaart found that in arenaceous sediments the zircons are comparatively large - 0.15- 0.25mm. - and in the finer grained sediments correspondingly smaller. Measurement of 200 zircons from four of the Neria gneisses gave average lengths of less than 0.1 mm. (See Table II).

Specimen	$\bar{l}$	$\bar{b}$	$\bar{l}/\bar{b}$
71050	80.1	35.9	2.23
71056 (g)	76.9	39.4	1.95
71061	98.9	52.2	1.89
71086	95.1	42.7	2.23
71092	90.0	51.3	1.75
71097 (g)	88.7	37.0	2.40
71109	104.5	61.2	1.71

Table II

Mean lengths and breadths (in microns) of zircons from gneisses and discordant granites (g). 200 separated zircons measured from each specimen

This is within the range given by Poldevaart for zircons from siltstones (0.05-0.15mm.). However, it is by no means certain that zircons do pass unscathed through all metamorphism. A study by Kalsbeek and Zwart (1967) has indicated that metamorphosed and migmatised gneisses from the Pyrenees contain smaller zircons than anatectic granites and unmigmatised gneisses from the same area. Therefore considerable caution is necessary when using zircons as indicators of the origin of metamorphic rocks. However, it is most likely that the small brown zircons of the Neria gneisses went through a sedimentary phase before being metamorphosed, but it is not possible to go a step further and say from which sediments they originate.

CHAPTER THREE

AMPHIBOLITES

The rocks included in this chapter are rocks of basic composition which have been metamorphosed and are generally concordant to the acid gneisses. The most common mineral assemblage is that of simple amphibolite, viz. hornblende and plagioclase, with accessory quartz. Relic clinopyroxene occurs in the amphibolites of the Inland Area, but in the amphibolites of the Coastal Area clinopyroxene is in equilibrium and forms an important part of the rocks. Garnet also is abundant in the coastal amphibolites but is never present in the inland ones.

In the field two main groups of amphibolites were recognised - those with garnets and those without. The ungarnetiferous amphibolites were further subdivided on the basis of their appearance into fine grained laminated amphibolites and medium to coarse grained massive amphibolites. The garnetiferous amphibolites are always of the massive coarse-grained type and are confined to the Coastal Area and usually have associated ultra-basics. The ungarnetiferous amphibolites occur in the Inland Area and only those near the Boundary Layer Amphibolite have associated ultrabasics. The laminated amphibolites never occur with ultrabasics, but do have a mica schist horizon along one side of them.

Laminated Amphibolites

The fine grained laminated amphibolites are dull black in colour and contain small scale alternating bands of mafic minerals and felsic minerals as do the gneisses, but in these amphibolites the bands are never more than 1-2 cms. in thickness and are usually only one mineral grain thick (Fig.8).

Crystals or aggregates of crystals of dull pale green pyroxene can often be seen easily in hand specimen in this type of amphibolite. The pyroxene is often concentrated into lenses or nodules in the rock, but in thin section individual grains may occur scattered irregularly throughout the rock. The concentrations of pyroxene crystals are frequently associated with bands of white feldspar (fig.9 shows a pyroxene-feldspar segregation that has been folded). The mode of occurrence in hand specimen suggests that the pyroxene is relic and microscopic examination shows that this is in fact so (see below).

The laminated amphibolite and the adjacent brown schist horizon act as an excellent marker horizon and define the major structure of the Inland Area as a plunging antiform with the closure in the south-west corner of the area and two parallel limbs striking N.E. - S.W. (Map I). Along the western limbs of the fold the amphibolite outcrop

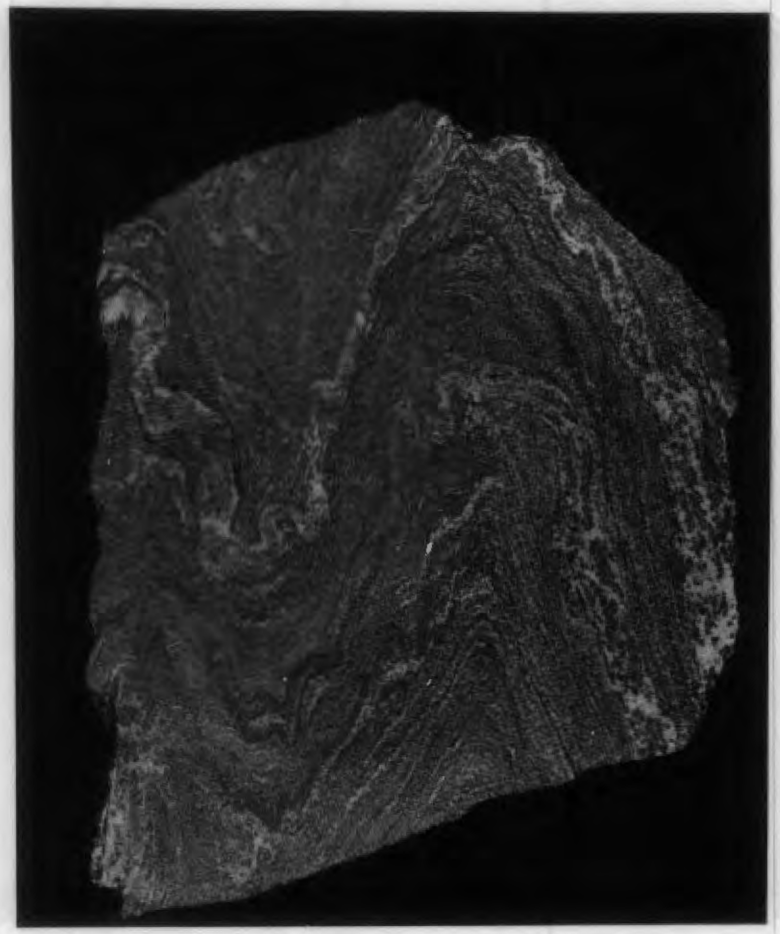


Fig. 8 Laminated amphibolite showing the typical small scale mineralogical banding. The wider felsic bands are composed of relic pyroxene and plagioclase. Specimen is 10 cm. across.

Fig. 9 Folded relic pyroxene-plagioclase inclusion in Laminated Amphibolite. Grid Square E6, Map I.

Fig. 10 Xenolith of laminated amphibolite enclosed in pegmatitic material. Grid square D6 Map I.

is 110 - 120 m. wide. But approximately 100 m. of that is taken up by pegmatite and only some 20-30 m. of amphibolite immediately adjacent to the brown schist is free from large masses of pegmatite. The pegmatite is almost one continuous body with isolated xenoliths of amphibolite enclosed in it (fig.10). The strike and thickness of the amphibolite is constant along the limbs, but it thickens in the hinge zone to about 3-400 metres and here a smaller proportion is pegmatite (fig.11). As the amphibolite horizon comes out of the hinge region along the eastern limb the proportion of pegmatite again increases (fig.12) until it takes up an important part of the amphibolite.

Where the eastern limb approaches the major East-West fault a change occurs in the amphibolite. The proportion of pegmatite decreases although there are still many large pegmatite masses present, and the amphibolite itself alters until to the north of the fault it is almost unrecognisable as an amphibolite. The colour changes from black to light grey as the hornblende is replaced by a colourless mineral, and only small relics of the normal amphibolite are found. These take the form of small indistinct patches, with dimensions measurable in centimetres, of black hornblende-bearing rock set in the grey rock. With the change in colour the laminations disappear and the rock becomes homogeneous.

Fig. 11 Pegmatite bodies in laminated amphibolite in hinge of major antiform. Note highly cleaved nature of the amphibolite. Grid square C3, Map I.

Fig. 12 General view of Laminated Amphibolite horizon along eastern limb of major antiform showing the considerable amount of pegmatitic material in the Amphibolite.

In the Neria Granite in the southern part of the area there are rafts of amphibolite, some of which have been appreciably altered during the formation of the granite. Most of the rafts belong to a horizon of unlaminated amphibolite, but others belong to what was probably a laminated horizon that has been granitised (fig.13). The darkest rock in fig. 13 is unaltered amphibolite, the grey rock a hornblende-bearing rock that is partly granitised amphibolite, and the palest rock is granite. This amphibolite is thought to belong to the Laminated Amphibolites, on account of its proximity to the brown schist horizon. However, this correlation cannot be made with certainty because the amphibolite in the granite is not continuous along the brown schist.

In grid squares G3, H3 and 4 (Map I) a laminated amphibolite horizon occurs that does not have brown schists along one margin. The rock has the typical appearance of the laminated amphibolite and probably originated in the same way. The rock to the east of the amphibolite is a very homogeneous granitised rock cut by regularly shaped small pegmatites. It has the appearance of having been formed by granitisation, in which case the original rock may have been brown schist. In the Neria Granite laminated amphibolite has been granitised more completely than brown schist, but from work to the north

Fig. 13 Amphibolite (Laminated Amphibolite?) in Neria Granite showing various stages of granitisation. Grid square E1, Map I

it is known that in some instances brown schist is granitised more easily than amphibolite (personal communication by H. Masson). The amphibolite is folded about a flat-lying axial plane and a south-east plunging axis and gapes to the south.

Massive Unlaminated Amphibolites

The unlaminated amphibolites are more variable in grain size than the laminated amphibolites: some are fine-grained with crystals only a few millimetres in size, while others may have crystals 1-2 cms. across. Garnet shows the greatest range in size - it usually has similar grain size to the rest of the minerals but occasionally porphyroblasts several centimetres in diameter may occur. There is also a greater range in composition: from normal amphibolite composed of plagioclase and hornblende to hornblendite, but the hornblendites usually contain some pyroxene in addition to hornblende.

The absence of lamination makes these amphibolites immediately distinguishable from the laminated variety. Instead of being segregated into distinct layers the mafic and felsic minerals are usually uniformly distributed throughout the rock. In some garnetiferous bands on the coast, however, a banding produced by concentration of the felsics may be apparent, but this is on a larger scale than that seen in the true laminated amphibolite. The individual bands are thicker than one mineral grain and the mafic bands are composed of hornblende and plagioclase, rather than being monomineralic.

Coast Area

Most, if not all, of the amphibolite horizons on the coast are garnet-bearing, but the garnets are not uniformly distributed through any one band. A hand specimen may be homogeneous, but this is unusual and it is most common to find that the garnets are irregularly distributed, even throughout a hand specimen. It sometimes appears that the garnets are concentrated in the felsic bands in the amphibolites. This is the result of their being more obvious in a groundmass of almost pure feldspar, but it is certain that they have grown to large size in the felsic bands. Alternating bands of garnet-free and garnet-bearing amphibolite, parallel to the strike, occur within individual amphibolite horizons but they are not sharply defined and are not continuous for any distance along the strike.

These amphibolite horizons are generally thinner and much more variable in thickness than the laminated amphibolite. They are strongly boudinaged and migmatised by granitic material, both on the large and small scales.

There has been some rotation of the blocks of amphibolite by the intrusion of the neosome at some localities, but at others the amphibolitic palaeosome has not been disoriented during the migmatisation. This can be demonstrated by measuring the foliation planes in the amphibolite and plotting them on a stereogram. Where there has been no disorientation the poles to the foliation planes will still form a regular pattern, either a concentration around one point or, if there has been local folding, a great circle giving the local fold axis. Where there has been a large influx of neosome, as in the laminated amphibolite horizon, rotation appears to have occurred, but because the amphibolite blocks are so widely scattered it is difficult to prove it. Where, however, the amount of introduced material is much less than the palaesome then rotation does not appear to have occurred. (Kalsbeek, 1967).

The amphibolite horizons on the coast are less strongly migmatised and boudinaged than those inland. The horizon separating the area of dark amphibolite gneiss from that of normal hornblende gneiss, for example, is particularly well-preserved. It can be followed west from the coast of Sermilik Fjord around the closure and becomes indefinite only as it approaches the fjord again (Map II). For most of its length it maintains a uniform thickness and is not boudinaged or migmatised. This horizon is composed of hornblendite with a strong schistosity. Other coastal amphibolites are boudinaged on a large scale. They have been necked and separated into individual boudins which may be up to 2 km. in length and 200 m. wide. The individual boudins often show pinch and swell structure similar to that seen in boudinaged pegmatites (Map II).

A small peninsula in the southern part of the Coast Area (4D, Map II) is composed predominantly of a strongly schistose amphibolite that in its regularity of banding is similar to the laminated amphibolite (fig.14). However, this amphibolite also is garnetiferous. Within the amphibolite and concordant with it is a band of acid gneiss approximately 10 m. thick. The lower (western) contact between amphibolite and gneiss is sheared and the upper gradational. The shearing suggests the presence of a strike fault causing a repetition of strata and the gradational contact indicates that the original basic rock was deposited as a supracrustal. This would give rise to the very regular banding, as in the case of the laminated amphibolite.

Fig. 14 Finely banded amphibolite. The folded discordant veins visible above the hammer appear concordant when viewed in a horizontal plane. Grid square 3D, Map II.

Inland Area

The unlaminated amphibolites of the Inland Area have small outcrops compared with those of the coast or the laminated amphibolite. Only in the north-western part of the area is a large expanse of this type of rock encountered, and that is present only by a freak of the folding. The horizon is only a few metres thick where it first appears on the western margin of the area, but it soon enters a region of intricate folding and is then spread out to several hundred metres thickness. Within this region ultrabasic lenses and large pegmatites are concentrated. The ultrabasic lenses occur both inside and around the margin of amphibolite and since it is usual for the ultrabasics to be concentrated along one margin of an amphibolite horizon they were probably brought inside it by being enclosed during the folding. North-east of this region the amphibolite once more becomes insignificant in thickness and very shortly disappears under large boulder fields and moraine.

Close below this large amphibolite, to the south-east geographically, are other amphibolites which show some interesting features. These are uniformly dipping sheets and all have associated ultra-basic lenses and may, in fact, belong to the same isoclinally folded horizon, but they cannot be seen to join. Followed along the strike to the north-east, the amphibolite becomes boudinaged and the amount of felsic material increases. Fig. 15 shows a typical boudin cut and surrounded by younger felsic material. The small veins inside the boudin are both pink and white and they do not occur outside the boudin and therefore were formed earlier than the event which caused the amphibolite to become boudinaged. Further north-east still, the amphibolite becomes granitised and changes its colour from dark-green to grey (fig.16). Relics of dark-green amphibolite are numerous and often define small folds in the rock (fig.16). The contacts between the grey amphibolite and normal amphibolite remains are usually not gradational but extremely sharp (fig.17). The paler rock has a much higher proportion of felsic minerals than the normal amphibolite and the mafics include biotite in addition to hornblende. These changes all take place over a distance of 200m. The amphibolite cannot be traced further north-east because of poor exposure but it may connect with an amphibolite seen some distance away to the north-east. If this is the case then it shows that the granitisation is of restricted extent because the amphibolite to the north-east shows no granititsation features.

The folded amphibolite in grid square B6 (Map I) may be the continuation of the granitised one just described, but there are differences between the two.

Fig. 15 Boudin of massive amphibolite containing veins formed prior to the formation of the boudin. Grid Square C7, Map I

Fig. 16 Granitised amphibolite. Palimpsest structures are preserved by less granitised more mafic bands. Same amphibolite horizon as seen in fig. 15

Fig. 17 Ungranitised (dark) and partly granitised (pale grey) amphibolite. Grid Square D7, Map I.

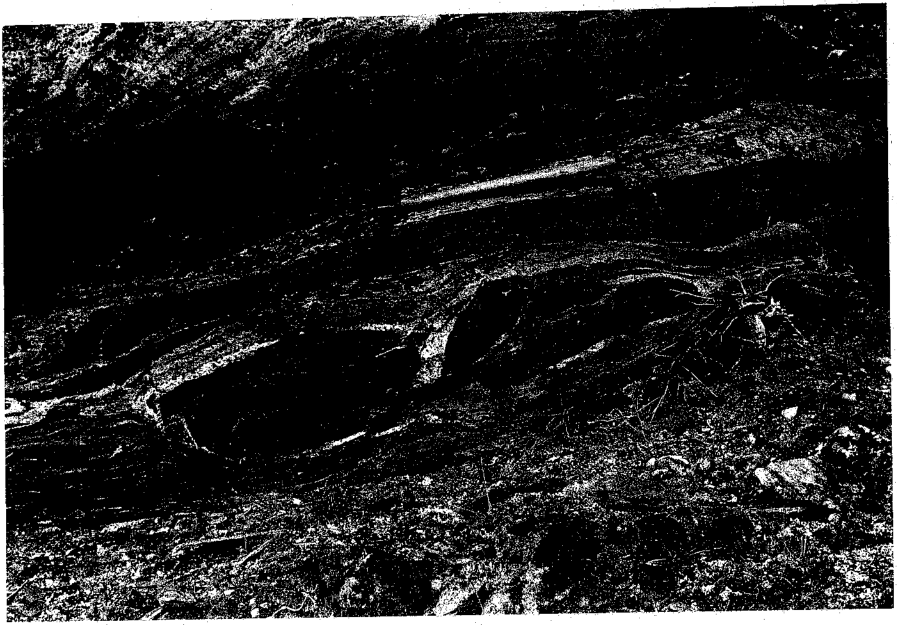


Fig. 18 Coarse grained green amphibolite boudins inhost of fine grained blue amphibolite. Grid square B6, Map I.

The folded amphibolite contains a high proportion of bright green amphibolite. The bright green colour is produced by larger grain size and smaller proportion of felsics than the more usual blue amphibolite. At one locality coarse-grained green amphibolite occurs in finer grained, more feldspar-rich blue amphibolite. Similar features have also been seen on the coast. The blue and green amphibolites have different rheological properties, as can be seen in fig.18 where two boudins of coarse-grained green amphibolite occur in a host of blue fine-grained amphibolite. Quartz has recrystallised in the sheltered space between the boudins. The two types of amphibolite are probably the result of very minor difference in composition of the original rock.

All other amphibolites in the Inland Area, i.e. those to the south-east of the ones described above, do not have ultra-basic lenses associated with them. Two parallel amphibolites to the east of the laminated amphibolite and brown schist horizon (running through grid squares F4, G4 and H4) are frequently well-exposed and show good migmatite relationships. These two amphibolites are probably part of the same isoclinally folded horizon. Fig. 19 shows a general view of the horizon where it outcrops on a hillside (the width of the outcrop is ca. 30 m. and it dips almost vertically).

2-3 kms. along the strike to the west there is a higher proportion of granitic material (fig.20). Between these two localities, however, invasion and disruption of the amphibolite by granitic material has been even stronger, and only isolated xenoliths of amphibolite are seen enclosed in foliated granite.

The rock comprising this amphibolite is a black amphibolite frequently containing biotite in addition to hornblende, plagioclase and salite. The amphibolite xenoliths are usually foliated and have a small scale banding of the felsic and mafic minerals. There is little veining within the xenoliths and no gradations between granite and amphibolite have been seen. Small concentrations of dull green pyroxene and altered feldspar occur within the amphibolite in a similar manner to those seen in the laminated amphibolite (Fig.9).

In the main granite body in the south of the Inland Area there are xenoliths of amphibolite that are relics of a once continuous amphibolite horizon. The xenoliths occur along a line striking approximately north-south, but are completely separated from each other by granite. The amphibolite is really an ultra-mafic rock composed predominantly of hornblende and biotite. Around the xenoliths there is often a thin skin of biotite separating the amphibolite from the granite. Any foliation

Fig. 19 Migmatised concordant amphibolite. Grid Square G.4, Map I.

Fig. 20 Same amphibolite as is shown in fig. 19, but much more strongly migmatised.

the xenoliths once had has been destroyed except where there is biotite present and they appear to have been rotated since their long axes may be at any angle to the strike of the band of xenoliths as a whole. Partly granitised or palimpsest xenoliths have not been seen, only the extremes occur - very mafic amphibolite and granite.

Petrography of Laminated Amphibolites

Dark green hornblende and plagioclase comprise most of the rock, in some sections as much as 80-90%. Quartz is usually present and also relic salite. Apatite, sphene and an ore mineral are common accessories; zircon is found very rarely. Epidote occurs frequently but as a vein mineral or in finely disseminated grains scattered throughout the rock. Alteration of the amphibolites is restricted in extent and is associated with proximity of epidote veining so that any specimen, or even thin section, may contain completely fresh and strongly altered parts.

Ideally the texture is eugranoblastic saccharoidal (Berthelsen 1960, p.24) but where there has been considerable alteration the crystal margins become diffuse and irregular. There are no remains of original igneous texture. Lobate grains are not uncommon but amoeboid texture has not been observed in these rocks. Occasionally

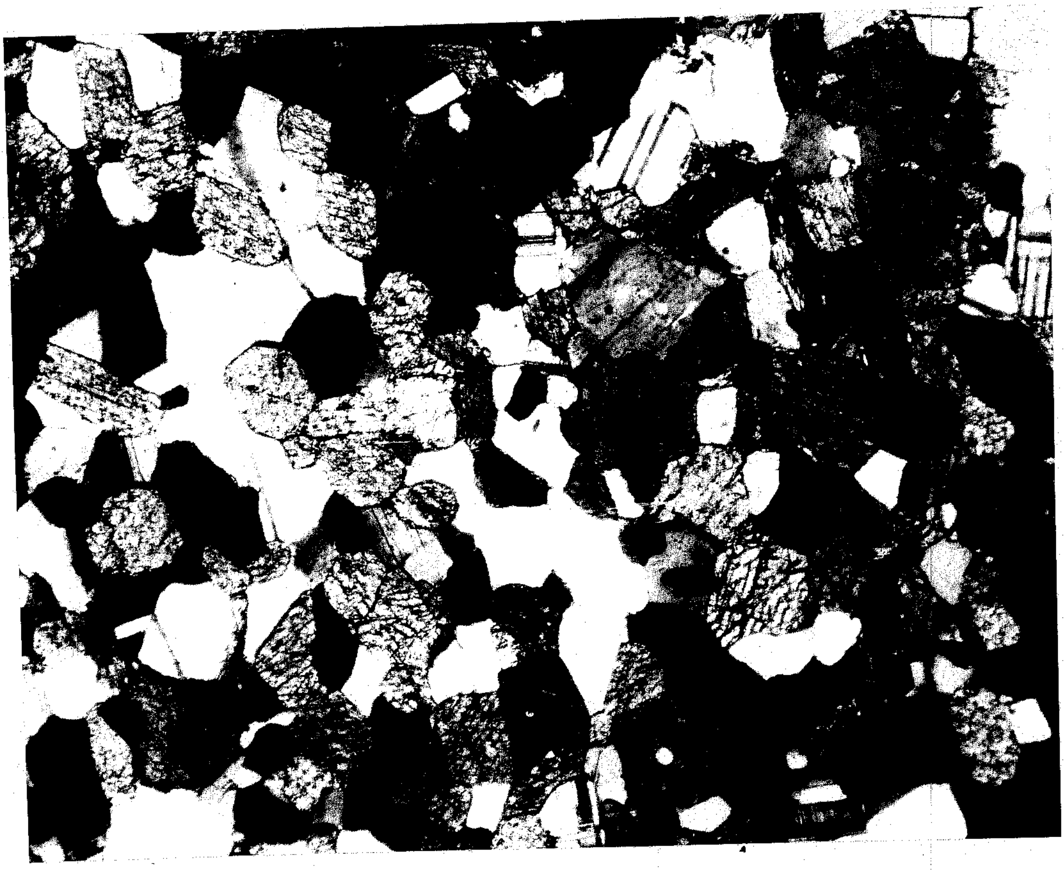


Fig. 21 Photomicrograph of laminated amphibolite showing well developed granular texture and uniform grain size. Specimen 71042, XN, X 12.

feldspar porphyroblasts may occur but these are not large and cannot be seen in hand specimens. The hornblendes are generally aligned parallel to each other within the plane of the lamination and this sometimes imparts a weak lineation or foliation to the rock, but this is extremely weak compared to the foliation produced in the biotite-bearing rocks. The grain size is usually 0.5-1.0 mm.(Fig.21).

Amphibole

By far the most common amphibole is a dark green pleochroic hornblende which usually comprises 50-75% of the rock. Table III contains the main optical properties of the mafic minerals of the amphibolites and the chemistry of the hornblendes is described in Chapter Four.

The hornblende usually has complete, uneroded crystal edges and cross-sections in particular may show extremely well-developed hexagonal outlines. Inclusions of quartz and apatite are contained in the hornblendes but they are not common. The hornblende is clearly in equilibrium in the laminated amphibolites.

The other amphibole is a colourless actinolite-tremolite. This mineral occurs in large irregularly shaped porphyroblasts with diffuse and ragged crystal boundaries, and also as rims to older pyroxene relics. Where the hornblende and plagioclase are fresh and in apparent perfect equilibrium actinolite is usually not found. It occurs only where the

	Colour	$2V_x$	$C \wedge Z$	N
Hornblende	pleochroic scheme:- Z deep olive green Y dark brown X pale brown $Z > Y > X$	85°±	13°±	
Actinolite-tremolite	colourless	70°-80°±	13°±	
Salite	green	120°-125°	40°	1.702

<u>TABLE III</u> <u>Optical properties of main mafic minerals in the laminated amphibolites</u>

rock shows signs of alteration. Around the pyroxene it forms wide diffuse rims without any definite boundaries. The actinolite is clearly older than the pyroxene but its age relative to the hornblende is not clear. It could either be an older mineral phase out of equilibrium with its surroundings, as is the pyroxene, or it could be younger than the hornblende formed by retrograde metamorphism that had only just commenced.

Pyroxene

The pyroxene is a salite and from the available optical data (Table III) contains Ca, Fe, Mg in the proportions 45%, 25%, 30%. It occurs in isolated patches throughout the rock as rounded, irregularly shaped grains that may be up to 3 mms. long. These grains are all out of equilibrium and contain incipient alteration products which are mainly ore minerals and hornblende. In a number of grains, small patches of green hornblende, scattered throughout the grain but in optical continuity, can be seen to have formed.

Feldspar

Feldspar comprises generally 30-40% of the rock but may be less than 20%. It is always plagioclase, usually of composition around An_{45}. Generally the plagioclase occurs as pellucid equidimensional grains with straight sides and often ideal triple junctions (Rast, 1964). The grains are usually smaller than the other minerals and are rarely more than 1 mm. across.

Alteration of the plagioclase to sericite or saussurite (usually the latter) occurs incompletely and is usually confined to a small part of a specimen.

The plagioclase twinning is secondary (Vance, 1961), but it rarely shows evidence of later deformation, such as bending or kinking of the twin planes, such as is very common in the gneisses.

Porphyroblasts of plagioclase occasionally occur, but these are not common and are usually strongly altered, even when the groundmass plagioclase is perfectly fresh, so that their composition cannot be determined. They often occur with the pyroxene and appear to be relics from the same period in which that mineral was formed.

Quartz

Quartz is present in nearly all sections of laminated amphibolite but rarely exceeds 1% of the total rock. It occurs as irregular interstital grains and usually has somewhat undulose extinction. Compared to that found in the gneisses, however, the amphibolite quartz is comparatively little strained and strongly lobate or sutured grains indicative of mylonitisation are never found.

Accessories

These include sphene, apatite, ore minerals, and, very rarely, zircon. Apatite is by far the most abundant and occurs usually as rounded grains which may be up to 1 mm. in diameter. Ore minerals are present in most amphibolites as small rounded grains but are never abundant, and are particularly uncommon in the laminated amphibolites. Magnetite is the usual ore mineral but ilmenite also occurs. Sphene commonly rims ore grains. Zircon is very rarely encountered and when it is, it is in the form of small rounded and scratched grains. These are typical of a sedimentary environment as zircons from basic igneous rocks are almost non-existent and occur as very small irregular shards (Poldevaart, 1956).

Epidote is present in most specimens of amphibolite either as alteration of plagioclase or as discrete grains scattered throughout the rock. It may occur as small grains thickly distributed or as larger grains thinly spread. The large grains rarely, if ever, show good crystal form, but instead are aggregates which sometimes are composed of fibrous epidote (Misch, 1965).

Petrography of unlaminated amphibolites

Coast amphibolites

The presence of garnet and, frequently, abundant fresh salite are the most obvious differences between the coast amphibolites and the inland ones, but there are also more subtle differences in the mineralogy.

Hornblende is less olive green than in the inland amphibolites and in many rocks the blue-green variety predominates, although olive-green hornblende is never totally absent. It occurs in corroded straggling grains and is often intergrown with broken up garnets. Mobilised quartz has intruded some hornblende crystals along the cleavages and pushed the crystal apart, but without rotating the pieces so that they remain in optical continuity. Colourless actinolite occurs very rarely and then only as small rims to hornblende crystals and not as discrete crystals. Amphibole comprises 40-70% of the rock. Fig.22 shows the main textural features.

Salite is not invariably present in the coast amphibolites but few lack it totally. It occurs as very pale green rounded grains and usually contains no alteration products. In some grains, however, incipient alteration to hornblende can be seen, but this has not developed to the scale it has in the inland amphibolites. On the coast salite appears to be in equilibrium with its

Fig. 22 Salite (light grey) intergrown with hornblende (dark) in a coast amphibolite. The felsic mineral is predominantly plagioclase. Specimen 70930, pp L, x 10.

surroundings. Like hornblende it is not uncommonly found intergrown with the remains of garnets and in some rocks appears to have replaced the garnet, e.g. 70930. Salite may form up to 25% of an amphibolite. The R.I. of the clinopyroxene is slightly greater than that of the inland salite. It is therefore slightly more iron rich.

Garnet occurs in crystals that have usually been broken around the edges, but have often retained a whole centre. There are rarely inclusions, (drop quartz around the edges is the most common) and there are never enough to warrant describing the garnets as skeletal. There is no evidence of alteration of garnet. It is either present as a completely fresh mineral or absent. There are no intermediate stages in which another mineral can be seen to be developing from it. The proportion of garnet in these rocks is usually not in excess of 10%, but there may be 2 or 3 times that amount in some garnet-rich bands.

Plagioclase shows less variation in composition than in the inland amphibolites. It ranges from An_{30} to An_{45} but most are more calcic than An_{40}. Some amphibolites have zoned plagioclases, a feature that has not been seen in any inland area rocks. One crystal in 70938 varies in composition from An_{45} in the centre to An_{32} at the rim. The zoning is restricted to the extreme edge of the crystal and is gradual. There is little alteration of the plagioclase and no amphibolite contains completely saussuritised plagioclase, as do some inland amphibolites. The amount of plagioclase present is very uniform, usually 25-30% of the rock.

Quartz occurs much as it does in the inland amphibolites; in small interstitial grains, with the occasional larger one, slight strain extinction, and evidence of late mobilisation. It forms 2 - 3% of the rock. Drop quartz is occasionally trapped in hornblende crystals.

Sphene is usually present and often rims magnetite grains. It may form as much as 1% of the rock in some specimens, e.g. 70807. In this rock sphene occurs as rounded grains, often in clusters, with distinct pleochroism from pale pink-brown to pale green-brown.

Inland unlaminated amphibolites

In thin section these usually appear essentially similar to the laminated amphibolites, with hornblende and plagioclase forming the bulk of the rock. The texture is granoblastic, occasionally weakly nematoblastic. In most specimens this type of amphibolite is coarser grained than the laminated amphibolite and usually has grains 1-2 mms. across.

There is, in fact, much greater variation in the mineralogical composition of these amphibolites than there is in the laminated type. In some of the unlaminated amphibolites felsic minerals are almost totally absent and in others biotite may be abundant. Pyroxene is nearly always better preserved and more abundant than it is in the laminated amphibolites. This is also true of tremolite-actinolite. On account of this variety it will be more convenient to describe these amphibolites by rock types, rather than mineral by mineral.

The simplest are composed of hornblende, plagioclase and quartz with the usual accessories, and may differ little from the laminated amphibolites when seen in thin section. 71009 from the western part of the area is typical of this type of amphibolite. It has a nematoblastic texture with well-shaped hornblende grains and interstitial feldspar and quartz. The hornblende has similar optics to those in the laminated amphibolites but is less olive green in colour. No colourless amphibole, pyroxene, or biotite occurs in this rock.

The plagioclase is around An_{33} in composition and has slightly distorted twin lamellae from post-crystallisation deformation. Quartz is not abundant and occurs as small irregular interstitial grains. The quartz appears to have been mobile later than the other minerals as it has very little undulose extinction and some has formed "U" shapes around other minerals.

Sphene and epidote are the accessory minerals but they are not at all abundant. Tiny (0.1mm.) rounded grains of zircon are present in the rock but there are only a few grains in the whole slide. They are brown and scratched and appear to be from a sedimentary environment.

Most of the other unlaminated amphibolites contain another mafic mineral in addition to hornblende. The amphibolites in the northern part of the Inland Area with associated ultrabasic lenses almost invariably contain salite. The clinopyroxene is older than the hornblende and usually contains alteration products, but is sometimes clear. The alteration products are hornblende or an opaque mineral, probably magnetite.

The formation of hornblende starts along planes in the salite crystals oblique to the cleavages and is first seen as small parallel slivers of high birefringence. At the earliest stage they cannot be seen in ordinary light. The slivers gradually increase in size and coalesce into patches of green hornblende that are all in optical continuity. The complete replacement of salite by hornblende has not been seen and that stage may not have been reached. The other form of alteration consists of the development of small grains of an ore mineral in the clinopyroxene crystals. There is apparently no other change associated with the formation of the ore mineral and either or both types of alteration can occur in the same crystal,(fig.23.)

Some of the inland amphibolites, e.g. 71125, are almost completely devoid of felsic minerals. There may be only a few small grains of quartz and plagioclase. In such rocks the quartz grains are concentrated along the boundaries of the hornblendes, with a few grains developed inside the hornblende crystals. Most of the amphibolites, however, have the usual development of quartz and plagioclase which is generally around An_{32} in composition.

Also in these rocks are sphene, epidote, biotite, apatite, and occasionally calcite. In some sphene is relatively abundant occurring in clots of rounded pleochroic brown grains of high relief which may each be up to 0.5mm. in diameter. Apatite comprises 1-2% as in most amphibolites, and epidote and biotite are sometimes present in small amounts. In one rock, 71125, calcite and quartz occur in small veins. The veins act as centres for the alteration of salite; where they cross salite grains, the mineral is strongly altered and offshoots of the calcite spread out into it. The calcite veins rarely cut hornblende crystals and may therefore be older than the hornblende, but a more probable reason for their absence is that the hornblende was stable in the conditions prevailing during the

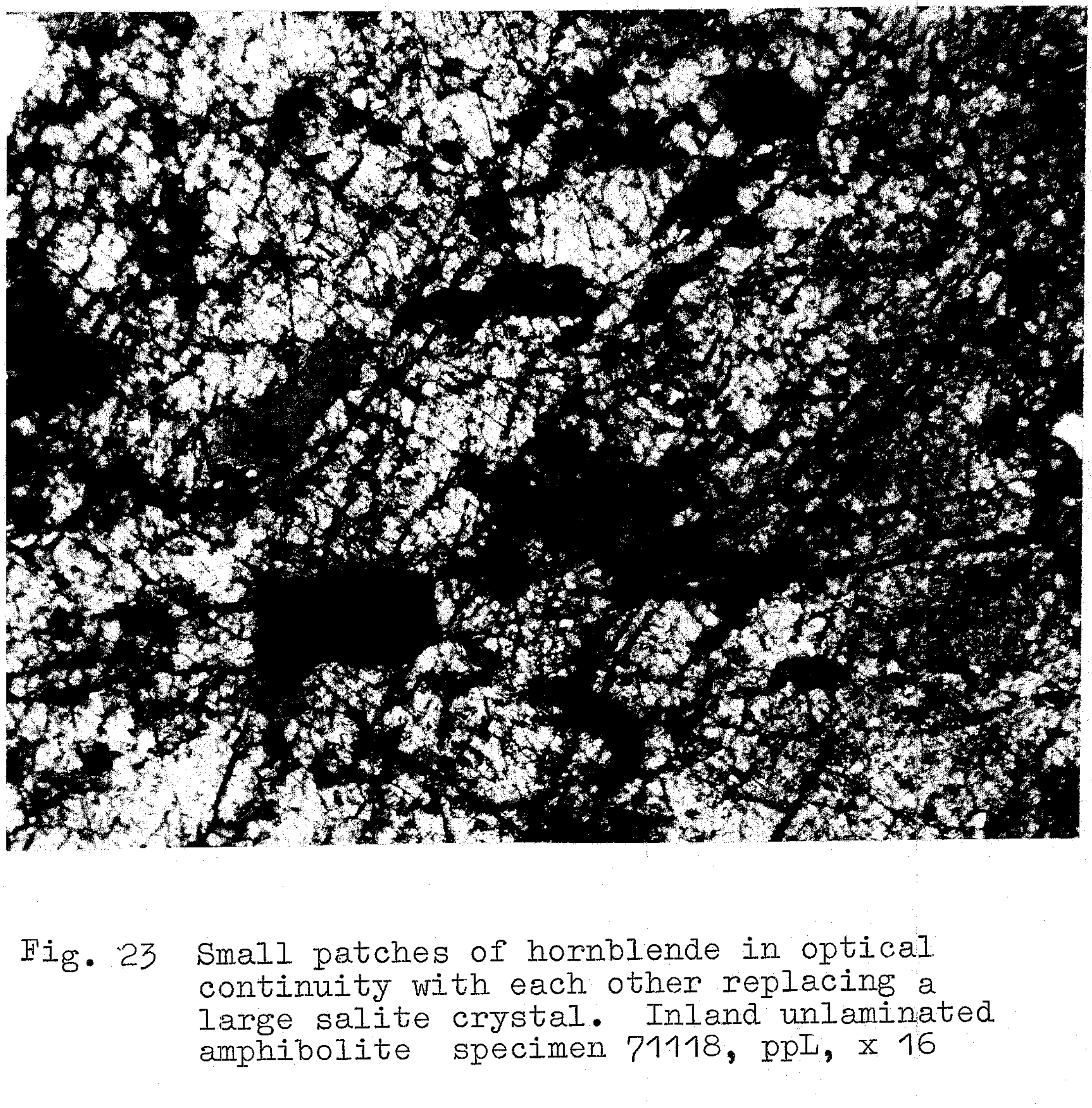


Fig. 23 Small patches of hornblende in optical continuity with each other replacing a large salite crystal. Inland unlaminated amphibolite specimen 71118, ppL, x 16

formation of the calcite and therefore would not break down to permit the growth of calcite. Or again, it could be a physical control - the pyroxene, being older and more cracked than the hornblende, would be more easily veined.

In grid square C7(Map I) one of these amphibolite horizons has been granitised, as described above (P.48) In thin section biotite can be seen to have replaced hornblende as the dominant mafic mineral. Biotite occurs in small flakes, rarely more than 0.5 mm. long, that define a weak foliation. The laths are generally ragged and usually concave towards plagioclase. They have a slightly mottled appearance between crossed polars and often have pleochroic haloes. Dark brown is the usual colour in the maximum absorption position but some flakes have stripes parallel to the cleavage a much paler golden brown in colour. These stripes do not show up at all between crossed polars.

The hornblende is blue-green in colour and occurs as small irregularly shaped patches. In a relic piece of unaltered amphibolite the hornblende is much better preserved, but even here good crystal outlines are lacking and the grains are rarely more than 1 mm. in diameter.

The plagioclase is of similar composition to that in the unaltered amphibolites and is around An_{32}. Curved twin lamellae show the plagioclase has suffered slight distortion since crystallisation, but it is generally very fresh with only a small amount of incipient saussuritisation.

Quartz occurs as in most amphibolites and the usual accessories are present.

The hornblende in the migmatised amphibolite of grid squares F4, G4 and H4, Map I, is the usual dark green variety and comprises 50% of the total rock volume. It occurs in well preserved crystals that usually have uncorroded sides and occasionally proper crystal form. Blue-green hornblende rims some of the dark green variety and there are a very few blue-green hornblende crystals, which are much less well preserved than the dark green ones. A colourless amphibole, tremolite-actinolite, is also present but this is very rare.

Biotite occurs as irregularly shaped patches and may form up to 20% of the rock. In some slides, esp. 51034, it is largely altered to a colourless mineral of moderate relief, prismatic form, and blue interference colours. This mineral has developed parallel to the cleavage of the biotite and is probably chlorite. Small magnetite grains are frequently abundant in the biotite and also are found rimming it.

Plagioclase forms up to 20% of the rock and is often slightly sericitised. The composition is more calcic than is usual; in one specimen it is An_{65}. Quartz may also occupy up to 20% of the rock, which is much more than is usually found in amphibolites. It occurs as large irregularly shaped grains usually with undulose extinction. "Drop" quartz is found in several of the hornblende grains. Small fingers of quartz intrude into many of the hornblendes and it is clear that quartz has again been mobile at a late stage.

Apatite is relatively abundant in this amphibolite - up to 1% of the total volume - and occurs in grains 0.5 mms. across. The rock is also unusually rich in ore mineral. This is magnetite and there is as much as 4-5% in the rock. Zircon grains are not uncommon, particularly in specimen 51034, fig.24, whereas they are almost completely absent in the other amphibolites. The zircons are scratched and broken and are typical of those that have had a sedimentary period in their history. The high proportion of zircon in this amphibolite may be related to the high quartz content if this originated partly by the admixture of a sand fraction from a sediment to the original basic rock.

Another unusual amphibolite occurs a short distance south-east of the amphibolite described. This has a very localised distribution and has not been followed very far along the strike. It is a black massive amphibolite with the feldspars concentrated into large porphyroblasts. These porphyroblasts develop over a

Fig. 24 Large, worn zircon crystal in migmatised biotite-rich amphibolite. Specimen 51034, ppL, x 64.

distance of a few metres and when first seen are numerous and diffuse, but rapidly become larger, better defined and fewer in number.

In thin section the rock is quite unlike any other amphibolite. There are apparently three generations of amphibole:- two of nearly colourless tremolite-actinolite and a dark green hornblende. The relationships between these are complex. The hornblende is intermediate in age and crystals of it occur embedded in very irregular porphyroblasts of the older actinolite. It is well preserved but it does not occur directly in contact with the older actinolite - there is always a rim of younger colourless amphibole.

The original outlines of the older actinolite are no longer visible and the mineral occurs as large optically continuous patches between younger minerals. It has extremely weak green pleochroism and numerous small alteration products. The younger actinolite has formed by alteration of the hornblende and also by recrystallisation of the older. It forms a continuous rim around the hornblendes, but where it extends into the older actinolite it develops a spiky habit. The direction in which it grows is related to the orientation of the nucleus hornblende and not to the orientation of the older actinolite.

Clinopyroxene occurs as large rounded crystals with alteration products inside them. It appears to be between the older actinolite and the hornblende in age, (fig.25)

Epidote is abundant in the rock. It occurs as clusters of small grains with a granular appearance and not as an alteration product of the other minerals. Vermicular intergrowths of epidote in epidote occur in many clusters. Sphene and magnetite occur in clusters with sphene usually rimming a magnetite core.

Plagioclase is mainly concentrated into large porphyroblasts, but not all the plagioclase material was incorporated and smaller grains of plagioclase are clustered around the porphyroblasts. The large porphyroblasts are often entirely altered to epidote and their composition cannot be determined, but the composition of the smaller grains is An_{45}. No quartz has been seen in this rock.

Amphibolite inclusions in the Neria granite

In thin section comparatively little alteration has occurred of those amphibolites caught up in the large granite body in the southern part of the Inland Area. The amphibolite next to the brown schist horizon in grid square E1 is probably the remains of one of the laminated amphibolites on the grounds of both position and appearance.

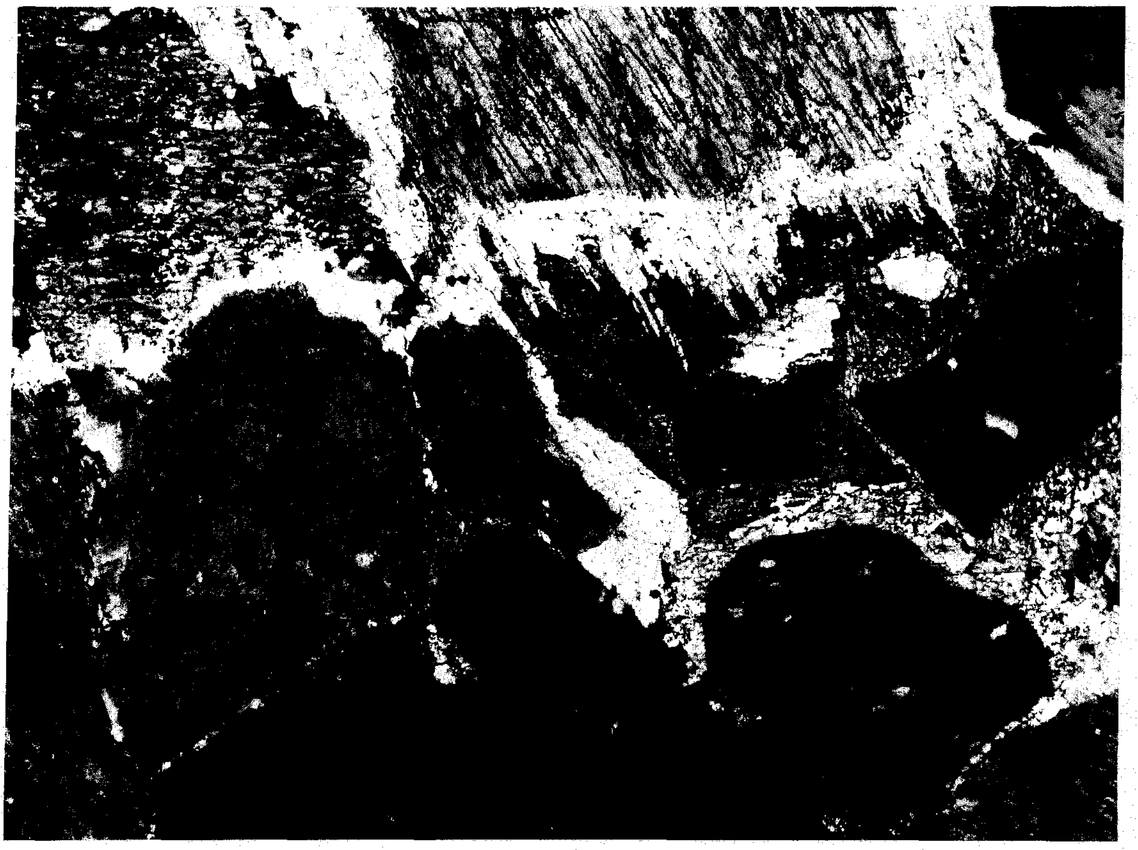


Fig. 25 Spiky younger actinolite growing out from a hornblende crystal into matrix of older actinolite. The block area is a salite area older than the adjacent hornblende. Specimen 51022, x N, x 10.

In thin section it is very similar to laminated amphibolites found outside the granite except that it is slightly finer grained. The hornblende occurs in slightly ragged crystals but is still olive-green in colour.

Plagioclase occurs generally as small crystals less than 1 mm. across, but there are a few small porphyroblasts up to 1.5 mm. across which enclose quartz and small hornblende grains. The composition is in the range An_{30-35}. Throughout the plagioclase epidote has developed, as discrete prismatic crystals up to 0.2 mm. long and not as a saussuritisation product. Small olive green flakes of biotite have started to develop in the amphibolite and this is probably the result of the effect of the formation of the granite on the amphibolite.

The other amphibolite that occurs as xenoliths in the granite is almost ultrabasic in composition. Some of the xenoliths are composed solely of hornblende and biotite, while in others there is a small amount of plagioclase that has been completely sericitised. The hornblende in these rocks is very much paler in colour than that in ordinary amphibolites. The pleochroic scheme is Y green, Z pale green, X neutral, $Y>Z>X$. There is a faint tint of blue in the green of this hornblende and this indicates that the hornblende is approaching pargasite in composition. Hornblende forms

60-70% of the rock. Many of the hornblendes in this rock have very fine faint lines, with the appearance of lamellae, parallel to what should be a cleavage direction. In some cross-sections it is possible to see one normal coarse cleavage and the unusual fine one in the same crystal.

Biotite occurs in segregated bands and with a weak foliation. It cuts across the hornblendes and is probably younger than it. The pleochroism is dark to pale brown. There are opaque inclusions along the cleavage planes of many grains and small weak pleochroic haloes. The plagioclase, where it occurs, is so completely altered to colourless mica that only the shapes of the micaceous areas indicate that plagioclase was originally present.

CHAPTER FOUR

CHEMISTRY OF THE AMPHIBOLITES

Fifty-five amphibolites were analysed by X-ray fluorescence for nine major elements, viz. Si, Al, Fe, Mg, Ca, Na, Ti, Mn (Total iron was calculated as Fe_2O_3). Specimens were collected for analysis from both areas and from all types of amphibolites. Figs. 26 & 27 show the localities of the analysed amphibolites.

The hand specimens for analysis were prepared in the usual way. They were first split to a suitable size by hydraulic splitter and the selected sample broken down in a jaw crusher. The sample thus obtained was halved and one half crushed in a disc mill for 5 - 10 minutes until it was reduced to flour. Pellets were prepared by mixing 2-3gm with a few drops of mowiol and compressing them with a boric acid backing in a hydraulic ram.

Twenty-five rock standards of widely spread compositions were used and the actual percentage compositions of the amphibolites were calculated by computer, using a programme developed by J.G. Holland (Holland and Brindle, 1966). The error for Na is about $\pm$ 10% and for other elements $\pm$2-3%.

In order to construct ACF and AFM diagrams FeO must be known and this was determined indirectly from the compositions of the constituent hornblendes of four of the amphibolites. The FeO content of the hornblendes was found

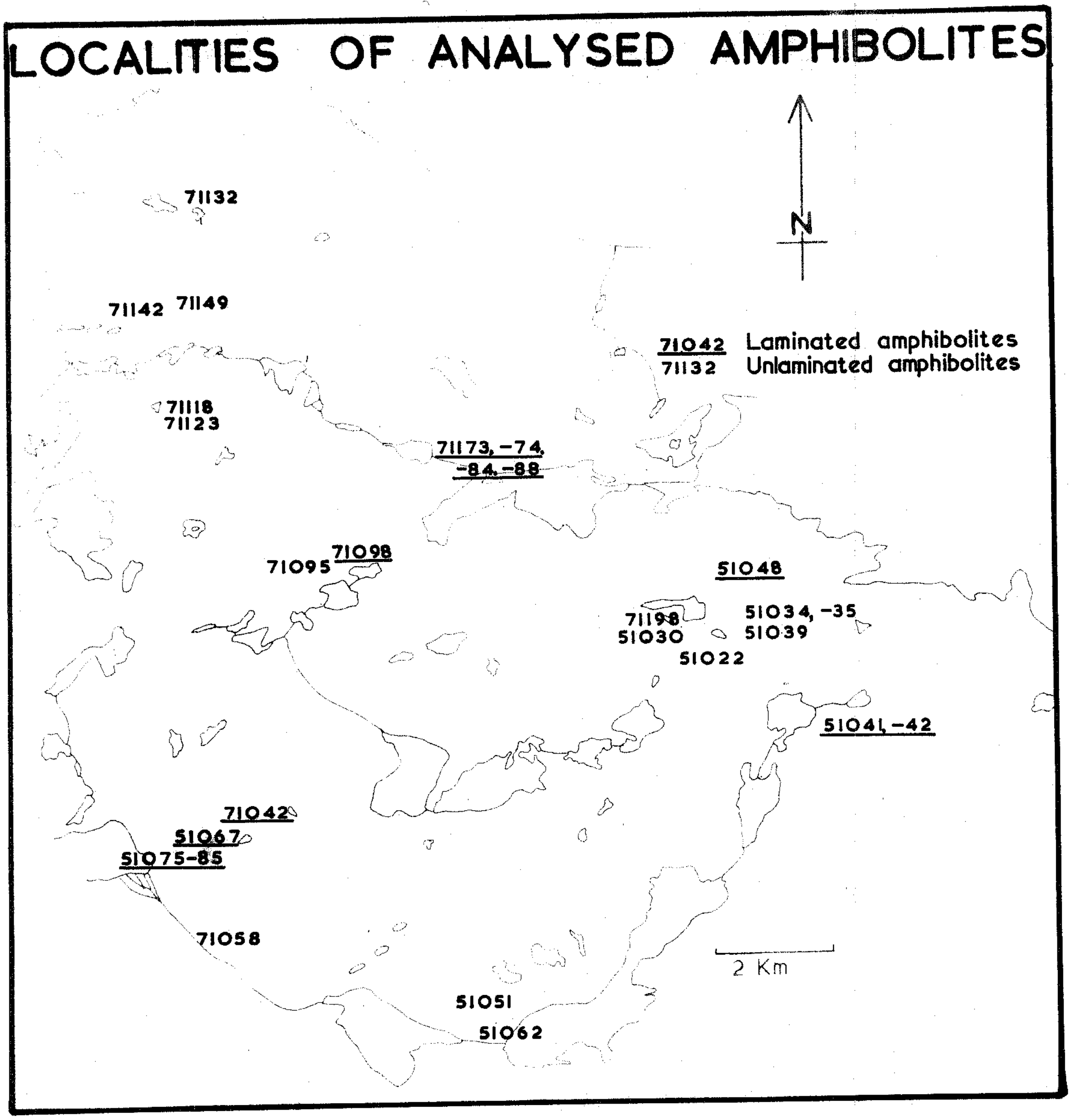

LOCALITIES OF ANALYSED AMPHIBOLITES
N
71042 Laminated amphibolites
71132 Unlaminated amphibolites
71132
71142
71149
71118
71123
71173, -74, -84, -88
71095
71098
51048
71198
51030
51034, -35
51039
51022
51041, -42
71042
51067
51075-85
71058
51051
51062
2 Km

Fig.26

LOCALITIES OF ANALYSED AMPHIBOLITES

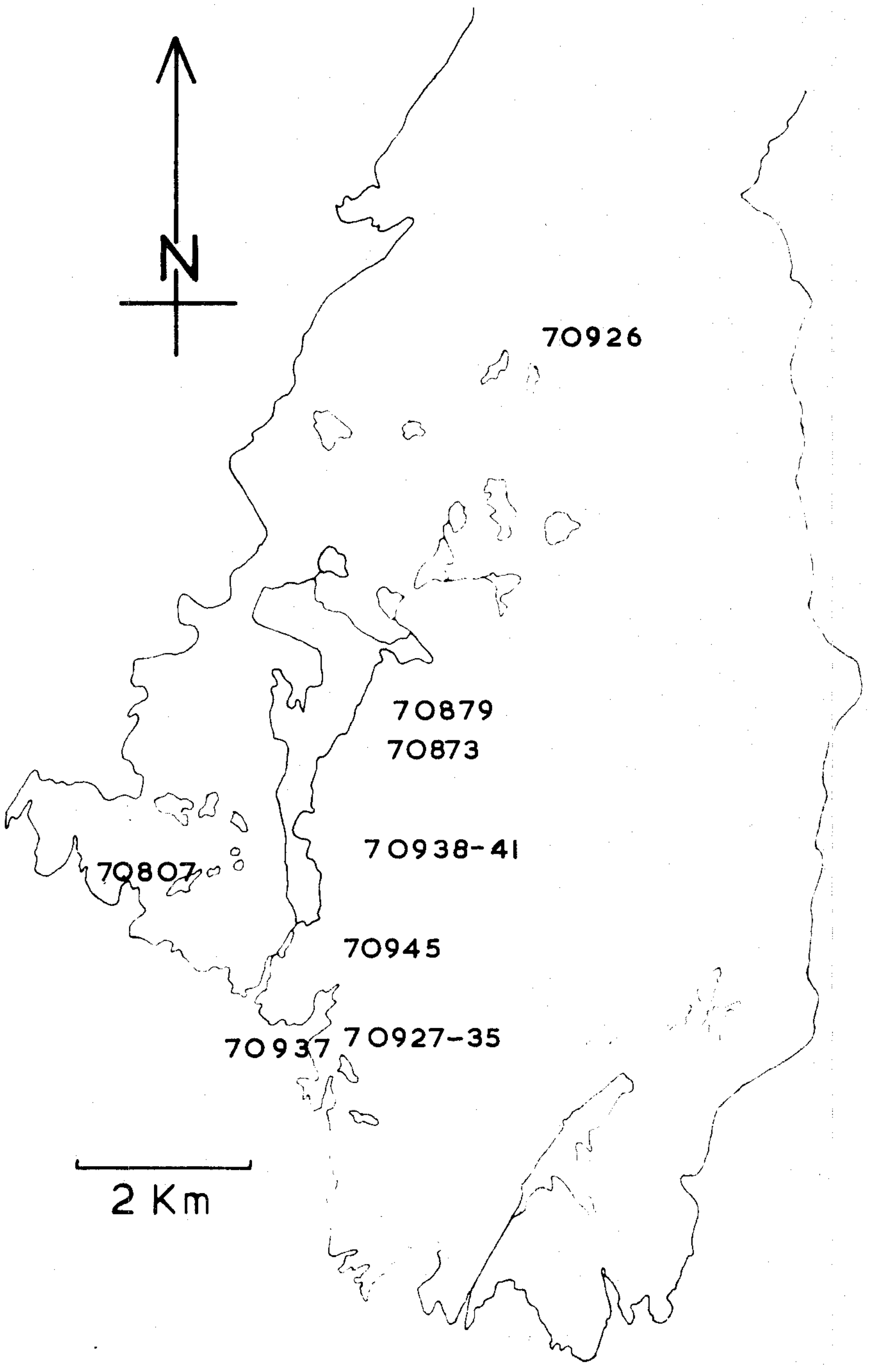


Fig.27

by titration against potassium permanganate and the FeO/Fe_2O_3 ratio calculated. The two hornblendes from the inland area had FeO/Fe_2O_3 ratios 3.85 and 4.82, with a mean of approximately 4.3. The two hornblendes from the coastal area had ratios of 2.25 and 2.35, with a mean of 2.3. These values were then applied to the appropriate groups of rocks. FeO could not be determined directly from the rock powder because in the crushing in the disc mill it was heated and became slightly oxidised. Several rock powders were dissolved and titrated, but very erratic results were obtained.

The analyses were divided into three groups corresponding to those used in the field and the means and standard deviations for each element calculated. It was found that the coastal amphibolites had the smallest variations from the mean, and the inland unlaminated generally the largest. This result was to be expected since the laminated and coastal amphibolites are the most uniform on geological evidence. The laminated amphibolites are all of similar appearance and are all from the same horizon, and the same is largely true of the coastal amphibolites with 14 out of 17 being from what is probably the same horizon. The variation in the inland unlaminated group is so great that it is necessary to sub-divide it. The sub-division based on composition corresponds well with sub-groups that are distinguishable on petrographical and structural grounds. Table IV gives the amphibolite analyses and table V the means and standard deviations of the three main groups.

Between the inland laminated and coastal amphibolites the differences are very small. A t-test calculation for each element shows that for only Na and Mn is there a statistically significant difference between the two groups. Table VI gives the t values for the amphibolites and the values t must be for 95% and 99% significance levels (obtained from Davies, 1954, table C). Na is higher in the coastal amphibolites - 2.56% as opposed to 2.03%; and Mn lower 0.21% against 0.25%. These differences are only statistically significant and since the amounts concerned are so small they are not of geological importance. The fact that there are no important compositional differences between these two amphibolite groups shows that the different mineralogies are the result of the rocks having attained equilibrium under different physical conditions. There are no differences in composition between coastal amphibolites with garnets and those without. Fig.28 shows the coast amphibolites plotted on an ACF diagram, and it is clear that the garnet-bearing rocks do not fall on opposite sides of the An-Hbl tie line. Thus this plot also supports the hypothesis that bulk rock composition was not the deciding factor in determining whether or not

TABLE IV Amphibolite analyses (All recalculated to 100%)

A. Inland laminated amphibolites

Specimen	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO
51041	54.24	14.32	3.37	8.42	3.40	10.71	2.93	0.63	1.68	0.29
51042	46.61	15.19	4.72	11.78	6.76	9.15	1.72	1.84	1.96	0.26
51067	48.69	15.22	3.85	9.62	7.19	10.74	2.35	0.54	1.54	0.28
51075t	49.94	13.02	3.64	9.09	7.60	13.15	1.81	0.48	1.04	0.23
51075b	50.29	13.47	3.64	9.07	7.18	12.73	1.91	0.47	1.02	0.22
51076	48.74	14.78	3.57	8.91	6.91	13.05	2.28	0.47	1.05	0.23
51077	48.23	14.06	3.73	9.32	7.57	13.31	2.13	0.33	1.09	0.24
51078	48.32	17.71	2.96	7.39	3.83	16.88	1.56	0.14	0.87	0.34
51079	55.26	17.56	2.21	5.52	4.09	9.77	3.42	0.53	1.44	0.20
51080	47.72	16.17	3.36	8.38	6.27	14.83	1.82	0.31	0.90	0.25
51081	47.65	15.69	3.51	8.75	11.95	9.60	1.71	0.19	0.76	0.19
51081	47.67	15.73	3.52	8.76	11.87	9.58	1.70	0.20	0.78	0.19
51082	42.07	18.65	3.95	9.86	10.77	11.16	1.37	0.83	1.10	0.24
51083	42.63	17.72	3.96	9.88	11.07	11.35	1.29	0.84	1.01	0.24
51084	48.07	15.51	3.45	8.62	10.55	10.61	1.80	0.32	0.87	0.19
51085	47.92	15.63	3.45	8.61	10.75	10.42	1.88	0.27	0.87	0.20
71042	50.89	15.41	3.38	8.43	6.25	11.32	2.54	0.48	1.07	0.22
71098	48.66	12.44	3.85	9.60	7.72	13.65	2.01	0.39	1.36	0.31
71173	50.46	13.69	4.00	9.99	7.01	10.68	2.32	0.24	1.38	0.22
71174	44.14	15.63	4.81	12.02	6.90	12.45	1.48	0.59	1.67	0.32
71184	48.92	13.64	3.42	8.53	7.94	14.06	1.56	0.86	0.97	0.26
71188	51.18	13.82	9.54	5.74	11.48	2.68	0.61	0.61	0.87	0.26

TABLE IV (continued)

B. Coast amphibolites

Specimen	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO
70807	48.43	14.68	3.72	9.28	5.45	13.11	3.02	0.77	1.30	0.24
70873	51.55	13.37	3.86	9.65	6.80	10.32	2.47	0.69	1.09	0.20
70879	46.19	11.44	3.72	9.29	19.22	7.30	1.67	0.31	0.70	0.17
70926	48.37	15.09	3.60	8.98	8.32	11.67	2.46	0.45	0.87	0.20
70927	50.67	13.83	3.41	8.51	8.57	10.80	2.69	0.66	0.66	0.20
70928	50.54	13.77	3.49	8.73	8.58	10.88	2.58	0.52	0.70	0.20
70929	51.13	13.97	3.37	8.43	7.85	11.40	2.58	0.37	0.69	0.19
70930	50.50	13.52	3.85	9.61	7.09	11.52	2.39	0.22	1.09	0.22
70931	50.21	13.25	3.91	9.77	7.01	11.30	2.42	0.22	1.69	0.22
70933	51.23	13.86	3.54	9.35	6.99	10.09	2.36	0.72	1.45	0.21
70934	55.60	15.53	2.45	6.10	6.00	9.46	3.32	0.69	0.69	0.16
70937	51.91	13.03	3.67	9.15	6.13	11.25	2.54	0.72	1.37	0.22
70938	49.59	15.86	3.34	8.33	9.07	9.73	2.68	0.45	0.78	0.18
70939	52.51	14.91	3.20	8.00	7.38	10.24	2.52	0.31	0.77	0.17
70940	49.60	14.12	3.48	8.70	8.88	11.88	2.04	0.42	0.67	0.21
70941	50.51	13.56	3.98	9.93	6.50	11.11	2.54	0.55	1.12	0.20
70945	47.17	14.65	3.90	9.73	9.04	11.01	2.42	0.72	1.13	0.22
71180	50.36	14.79	3.38	8.43	7.85	11.34	2.51	0.34	0.76	0.25
70935	47.60	2.85	3.27	8.17	35.05	2.33	0.16	0.10	0.23	0.24

TABLE IV (continued)

C. Inland unlaminated amphibolites

Specimen	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO
51022	46.16	14.35	4.61	11.51	7.71	11.58	1.66	1.36	1.83	0.19
51030	46.07	14.35	4.29	10.72	8.70	8.90	1.69	3.01	2.07	0.21
51034	50.56	11.82	4.54	11.33	6.09	10.91	1.45	0.69	2.47	0.14
51035	45.22	15.45	4.21	10.52	11.26	6.71	1.74	2.40	2.31	0.18
51039	49.04	14.61	4.19	10.45	6.99	9.10	2.30	1.10	2.03	0.19
51048	49.76	14.93	3.45	8.63	9.84	10.07	1.84	0.38	0.91	0.21
51051	48.77	7.71	3.60	8.97	18.04	9.10	0.77	2.18	0.61	0.24
51062	46.00	16.07	3.92	9.80	9.00	11.20	1.42	1.41	0.93	0.24
71058	49.23	14.10	3.90	9.72	8.02	10.51	2.28	0.82	1.19	0.24
71095	54.25	12.61	3.37	8.41	7.90	8.19	2.68	1.54	0.87	0.17
71118	51.52	4.95	4.10	10.22	12.88	13.53	0.74	0.56	1.22	0.29
71123	53.63	6.83	5.05	12.59	10.80	6.87	1.26	0.87	1.80	0.30
71132	50.49	14.14	3.98	9.93	7.06	9.93	2.65	0.27	1.34	0.22
71142	51.15	13.18	4.88	12.19	4.50	8.88	2.55	0.61	1.81	0.25
71149	51.95	13.10	3.92	9.77	6.34	10.19	2.57	0.58	1.36	0.23
71198	47.68	13.77	4.36	10.87	7.79	9.69	2.31	1.07	2.24	0.21

TABLE V

A. Inland laminated amphibolites

	Mean	Standard Deviation
SiO_2	48.60	3.09
Al_2O_3	15.20	1.64
Fe_2O_3	12.76	1.85
MgO	7.50	2.32
CaO	11.96	1.87
Na_2O	2.03	0.53
K_2O	0.54	0.35
TiO_2	1.17	0.32
MnO	0.25	0.04

B. Coast amphibolites

SiO_2	50.58	1.81
Al_2O_3	14.22	0.80
Fe_2O_3	12.41	1.25
MgO	7.50	1.10
CaO	11.01	0.85
Na_2O	2.56	0.27
K_2O	0.52	0.18
TiO_2	0.99	0.31
MnO	0.21	0.02

C. Inland unlaminated amphibolites

SiO_2	49.47	2.65
Al_2O_3	12.56	3.13
Fe_2O_3	14.50	1.61
MgO	8.93	3.11
CaO	9.71	1.66
Na_2O	1.87	0.62
K_2O	1.56	0.56
MnO	0.22	0.04

TABLE VI

t-test values for amphibolite groups A,B & C.

	t_{AB}	t_{AC}	t_{BC}
SiO_2	2.27	0.87	1.37
Al_2O_3	2.18	3.22	2.05
Fe_2O_3	0.64	2.92	4.05
MgO	0.01	1.56	1.73
CaO	1.89	3.69	2.76
Na_2O	3.71	0.82	4.09
K_2O	0.23	3.32	3.40
TiO_2	1.68	2.61	3.53
MnO	3.68	1.20	1.21

	99%	95%
t_{AB}	2.73	2.03
t_{AC}	2.73	2.03
t_{BC}	2.75	2.04

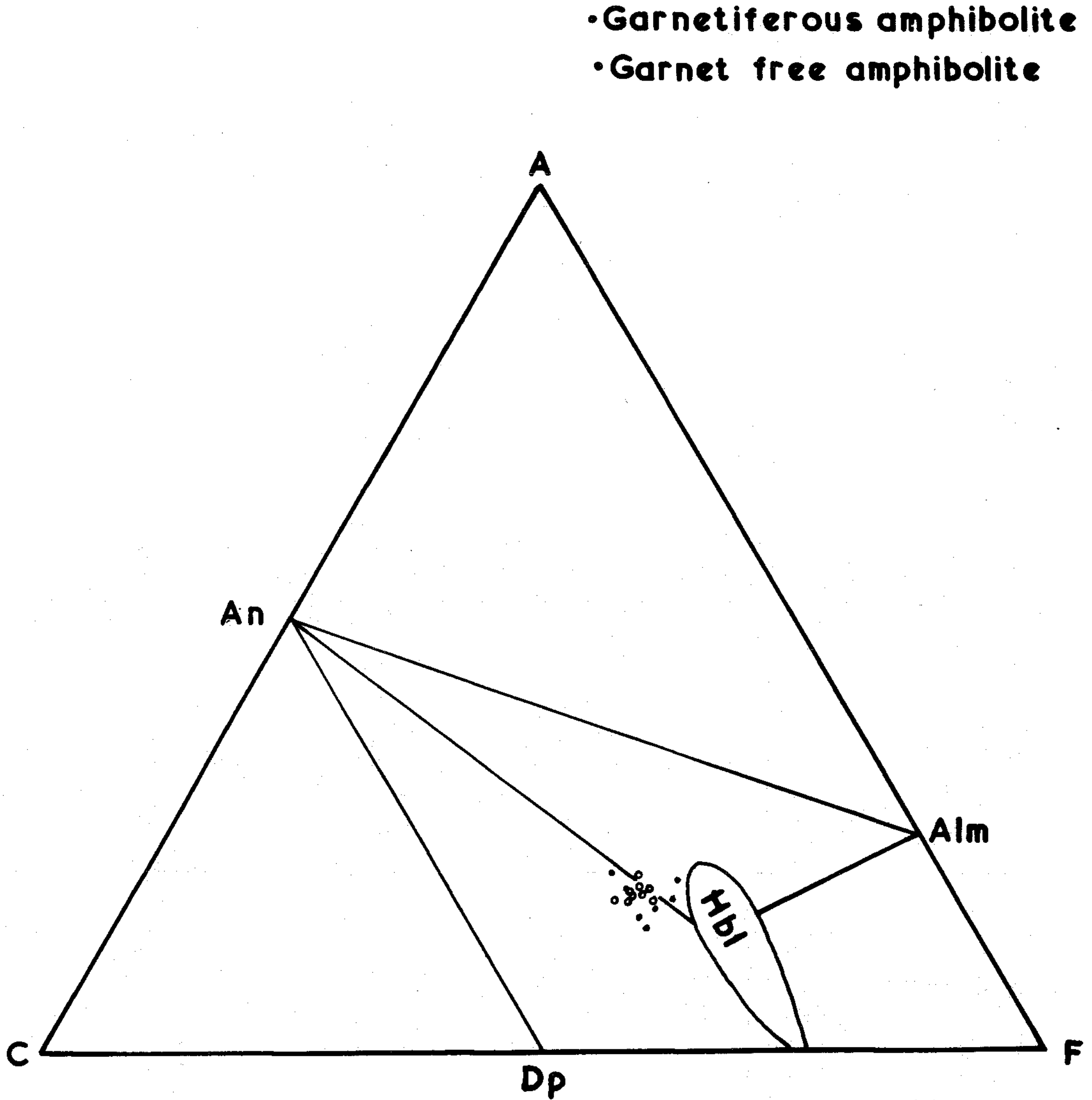


Fig. 28 ACF diagram for coast amphibolites

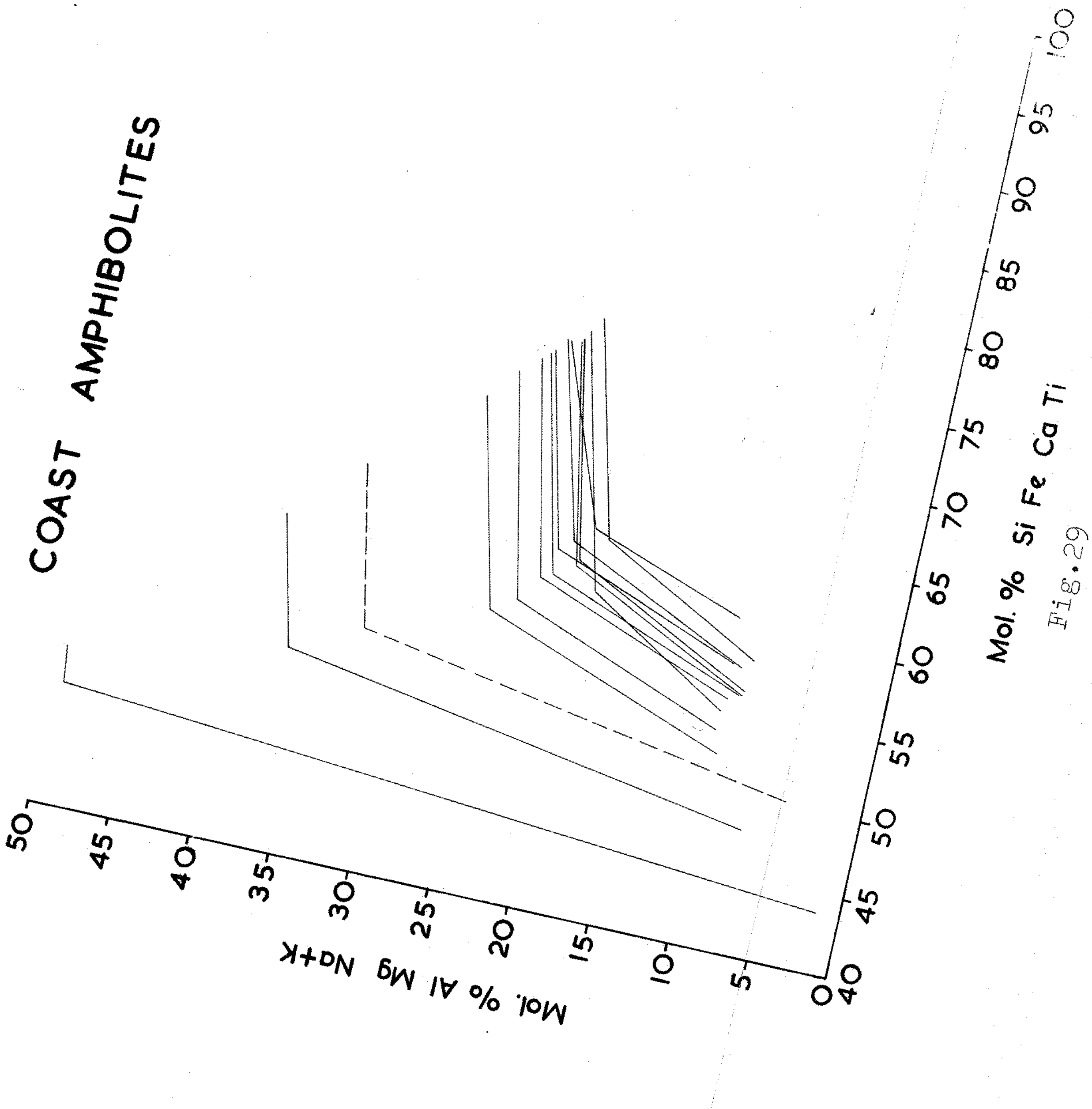
garnet would form. This is the opposite conclusion to that reached by de Waard (1965) in his study of Adirondack metabasites.

Compositionally the amphibolites are very similar to basic igneous rocks and it is probably from such rocks that they originated. There can be no doubt that those amphibolites with closely associated ultra-basic rocks originated as igneous rocks and as the differences between them and the amphibolites without ultra-basics are usually negligible, it is most likely that the latter also originated as basic igneous rocks. The nearest quoted basalt compositions to these amphibolites are 3 tholeiite basalts from Japan (Kuno, 1960).

A very convenient way of expressing rock compositions graphically for comparative purposes is to use a vectorial plot devised by Korzhinskii, (1959). By means of this plot all the elements can be used instead of only two or three. The weight percentages of the elements to be used are recalculated as molecular percentages. These percentages are plotted on squared paper alternately parallel to the abscissa and ordinate. Thus in this case Si, Fe, Ca, and Ti were plotted parallel to the abscissa and Al, Mg and Na+K parallel to the ordinate. Two lines inclined to each other are obtained for each analysis. The point at which the first line starts gives the molecular percentages of Si and Al, the slope of the line gives the ratio Mg/Fe, and its length an indication of the amounts of Fe and Mg present. The slope and length of the second line give the same information for alkalies and Ca. The amount of Ti is given by the distance parallel to the abscissa between the end of the second line and the line joining 100% on the abscissa to 100% on the ordinate.

The slope of a single line, the resultant of the two already drawn, gives the ratio (Mg + alkalies)/(Fe+Ca). But since Na+K is never more than 4% the slope is effectively Mg/(Fe+Ca). This single line plot shows very clearly the variation from basic to ultrabasic since the latter are particularly distinguished by the very high proportion of Mg they contain.

The plot with both lines drawn is more informative, however, and that one is used here. Fig.29 shows this plot for the coastal amphibolites. The dashed lines belong to a rock from the inland unlaminated group that is included in this diagram because it is intermediate in composition between basic and ultra-basic and helps bring out the gradation between the two. The diagram shows that at first as the rocks become more ultrabasic (moving from right to left) both Mg and Fe increase as Si decreases. The increase in Mg and Fe is shown by the line relating to these elements maintaining the same



COAST AMPHIBOLITES

Fig.29

slope but increasing in length and moving towards the ordinate. If only Mg increased then the slope would increase also. Ca and alkalies remain constant as is shown by their line maintaining the same slope and length. When Mg increases to almost 50 mol.%, however, other elements start to decrease in addition to Si, and when the extreme ultra-basic composition (e.g. 70935) is reached Al, Ca and alkalies are all very much reduced.

Fig.30 shows a vectorial plot for the inland laminated amphibolites and it is clearly very similar to that of the coastal amphibolites. The few rocks that have unusual compositions are immediately obvious. Rock 51078 for example can be seen at a glance to be rich in Ca and correspondingly low in Fe and Mg. This is because the specimen used in the analysis contained part of a pegmatite vein and was therefore much richer in plagioclase than the normal amphibolite. Although there is more variation in the laminated group, the analyses as a whole fall very much in the same area on the diagram as do the coastal amphibolites. But more important than this, both plots show the same trend from basic to ultra-basic composition. Evans and Leake (1960) have shown that trends are more important than individual compositions in determining the origins of amphibolites. Thus the similarity between the trends of the two groups

INLAND LAMINATED AMPHIBOLITES

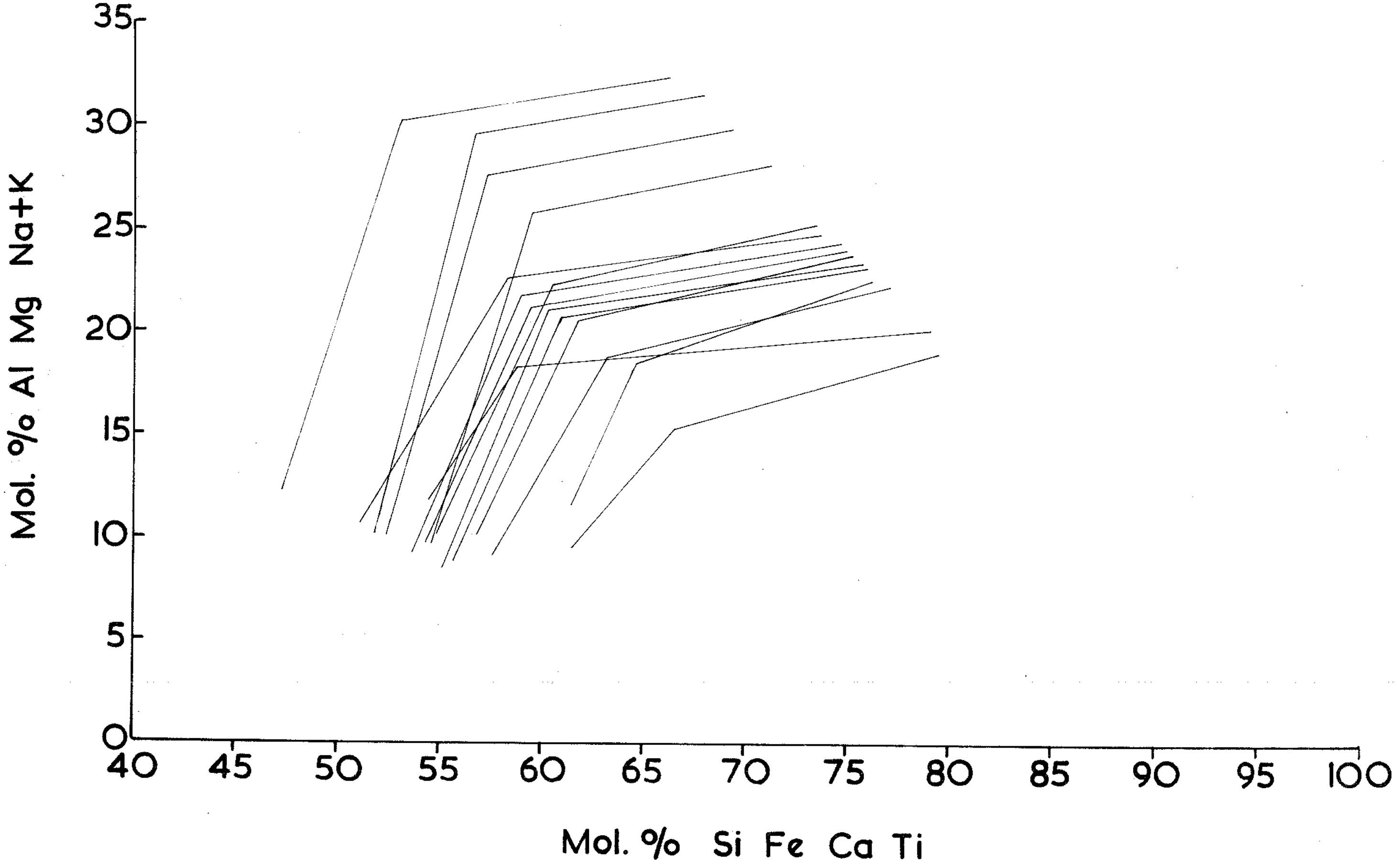


Fig.30

is taken as very strong evidence that they both originated as basic igneous rocks. Fig. 31 shows the inland unlaminated amphibolite plot which is much more irregular.

The trends of certain Niggli values are also useful in indicating the origins of metamorphosed rocks. Leake (1964) in a study of the Connemara striped amphibolites (which appear, incidentally, to be identical with the inland laminated amphibolites) found that plots of k and ti against mg were the most useful in differentiating between sedimentary and igneous trends. Figs. 32 and 33 show k and ti respectively against mg for all the analysed amphibolites. The fields or trends of Karroo dolerites and certain sediments (from Leake, 1964) are also shown, and although the amphibolite analyses do not completely follow the igneous trends they are much further from the sedimentary ones.

The most distinctive members of the inland unlaminated groups are specimens 51030, -34, -35, -39, and 71198. These are all derived from the black biotite bearing amphibolite described above (p. 53) and can thus already be recognised as being different from the rest of the amphibolites. Compositionally their most striking feature is a very high concentration of Ti. They all have more than 2% TiO_2, wherease all the others have less, and usually appreciably less, than this. The

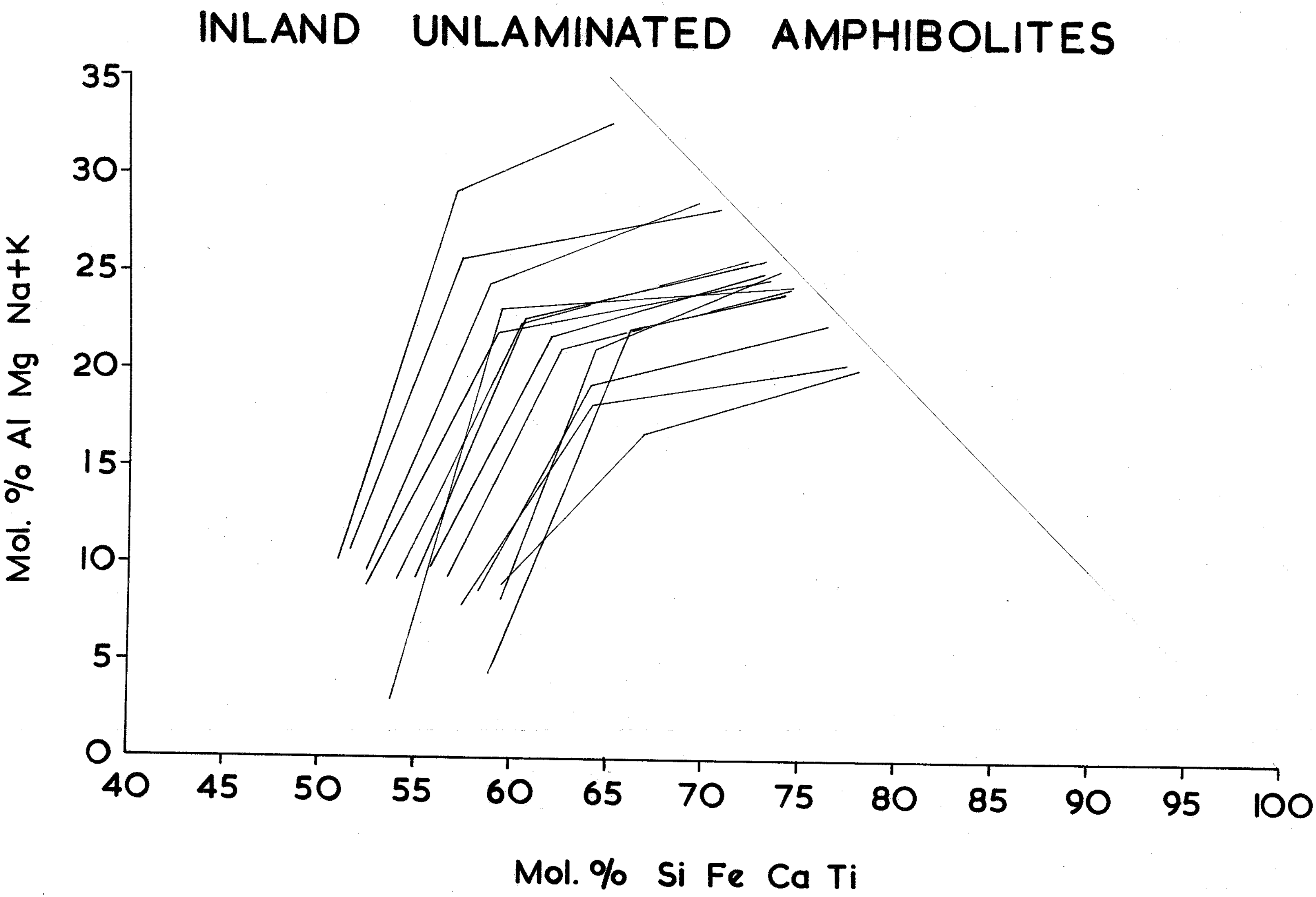

INLAND UNLAMINATED AMPHIBOLITES
Mol. % Al Mg Na+K
35
30
25
20
15
10
5
0
40
45
50
55
60
65
70
75
80
85
90
95
100
Mol. % Si Fe Ca Ti

Fig.31

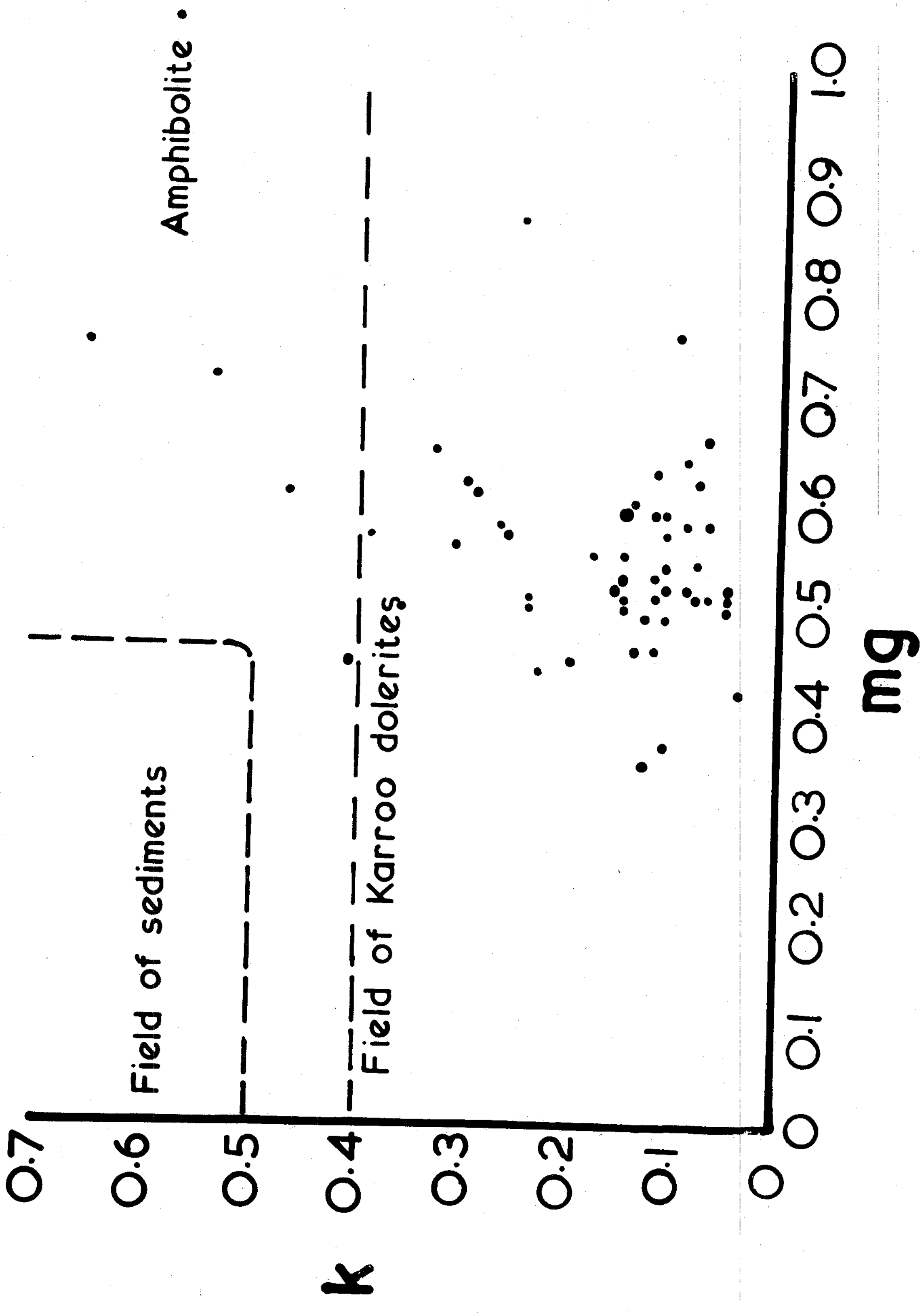

Amphibolite
Field of sediments
Field of Karroo dolerites
k
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1.0
mg

Fig.32

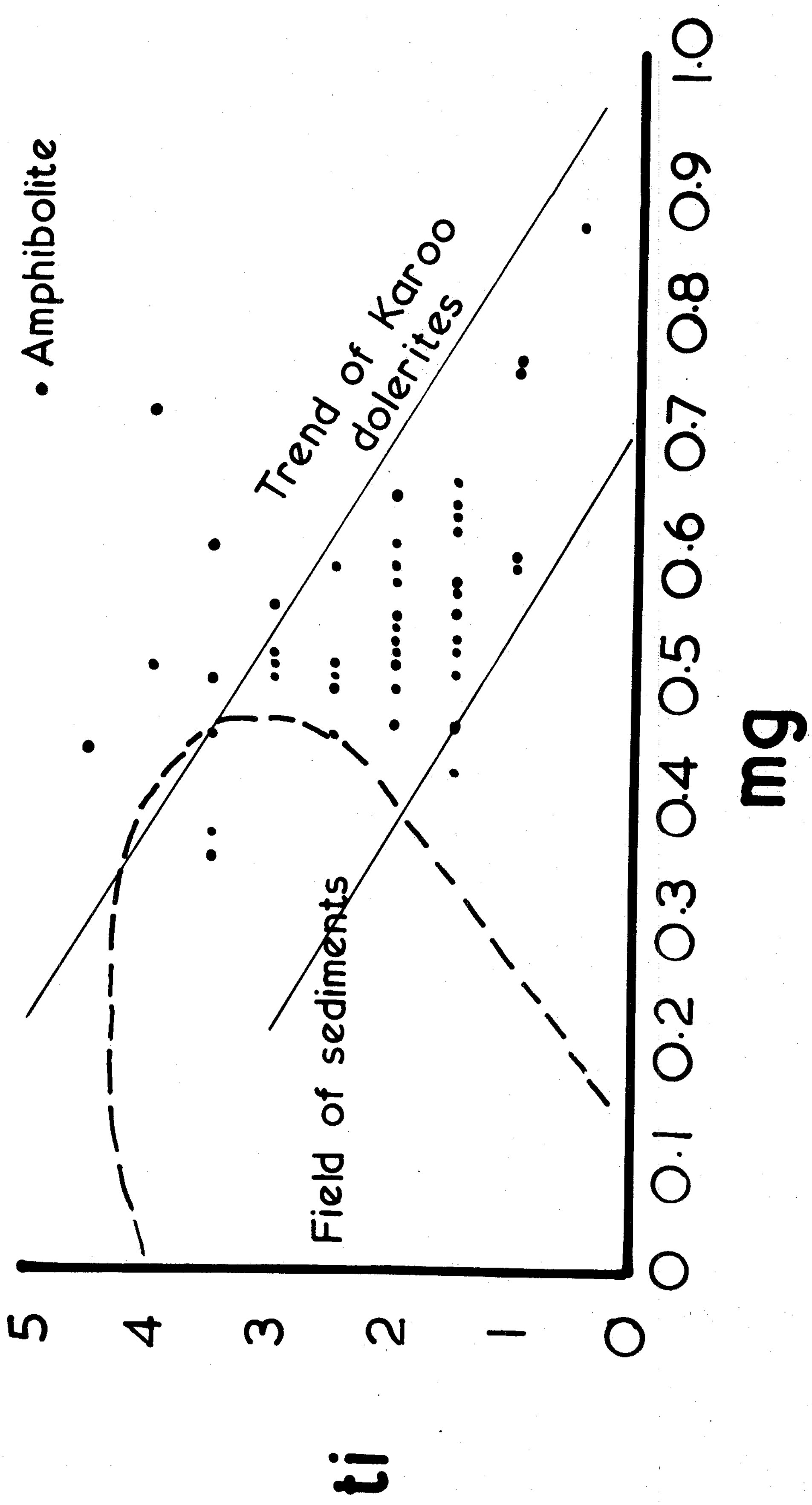

Amphibolite
Trend of Karoo dolerites
Field of sediments
ti
0
1
2
3
4
5
mg
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1.0

Fig.33

only amphibolites to approach these in Ti content are those that occur in the same general region, i.e. the eastern part of the Inland Area. Specimens 51022, -41, -42, all have over 1.7% TiO_2, but only one other amphibolite from the two areas contains this much Ti. Specimens 51041 and 51042 have been included in the inland laminated group because they look like the laminated amphibolites. But they cannot be seen to be part of the same horizon on stratigraphical grounds, nor is there an horizon of brown schist associated with the amphibolite from which these two rocks come. The very high Ti content of these rocks also suggests that they do not belong to the laminated group because Chayes (1965) has shown that Ti is the only reliable indicator of different origins of basalts. He found that out of a large number of analysed basalts the content of TiO_2 could be used to differentiate between oceanic and cicum-oceanic basalts with 93% efficiency. Other oxides, or combinations of oxides excluding TiO_2, gave little better efficiency than would be achieved by chance. The division between the two types came at 1.75% TiO_2. Oceanic basalts would have more, and circum-oceanic basalts less, than this. Chayes concluded that this difference in TiO_2 content indicated two separate sources for the basalt types, but his purely statistical work could not prove it. Be that as it may, it would appear that large differences in the TiO_2 content of basalts indicate at least formation in different environments. In the particular case of the Neria area the presence of high Ti values from the amphibolites of one part of the area would suggest that these were not originally part of the same group as all the other amphibolites.

The five amphibolites with very high TiO_2 values (over 2%) have other compositional features which distinguish them from the rest, and the mineralogical peculiarities of these five can be related to these differences. Table VII shows the mean composition, standard deviations, and t-test values of the very high Ti amphibolites compared with the laminated group. The laminated group is used for comparison rather than the coastal group because in addition to being a fairly homogeneous group it was metamorphosed under the same conditions as the high Ti group. Any differences in present compositions will therefore reflect original differences and not differences caused by differentiation during non-isochemical metamorphism.

There are significant differences in the amounts of Fe, Ca, K, Ti, and Mn between the two groups and these account for the presence of biotite in the high Ti group. There is approximately 30% less CaO in the high Ti group

TABLE VII

High Ti amphibolites (Nos.51030,-34 -35, -39, 71198)

	Mean	Standard Deviation	t
SiO_2	47.71	2.17	0.59
Al_2O_3	14.00	1.40	1.45
Fe_2O_3	15.14	0.49	2.87
MgO	8.17	2.17	0.56
CaO	9.06	1.53	3.10
Na_2O	1.90	0.39	0.50
K_2O	1.65	1.00	3.97
TiO_2	2.26	0.58	5.48
MnO	0.19	0.03	2.97
	Inland laminated group		
SiO_2	48.60	3.09	
Al_2O_3	15.20	1.64	
Fe_2O_3	12.76	1.85	
MgO	7.50	2.32	
CaO	11.96	1.87	
Na_2O	2.03	0.53	
K_2O	0.54	0.35	
TiO_2	1.17	0.32	
MnO	0.25	0.04	

and since hornblende requires a lot of Ca a dearth of that element would inhibit its development. On the other hand biotite requires a higher proportion of K and Ti than hornblende and the higher proportion of these elements in the high Ti group would therefore encourage the growth of biotite.

Engel and Engel (1962a) describe changes in amphibolite compositions in rocks from the Adirondack Mountains in New York State and conclude that they were caused by the change in metamorphic grade from almandine-amphibolite facies to granulite facies. No systematic changes in composition can be found in the Neria amphibolites, but this is to be expected because it does not appear that the grade of metamorphism on the coast was different from that inland.

If only the mean values for each oxide are considered and the standard deviations ignored then there appear to be bigger differences between the different groups of Neria amphibolites than there are between the Emeryville and Colton amphibolites described by Engel and Engel. But in the Adirondacks the changes can be seen to have occurred steadily over some distance, and it would appear that this is a case of geological evidence outweighing statistical insignificance.

Hornblende chemistry

There is no more variation in the hornblende composition than there is in the analysed amphibolites, and what variations there are can often be related directly or indirectly to the rock composition. Table VIII gives the details of the chemistry of the hornblendes. The value for H_2O^+ was estimated to be about 2%, so this was added to the XRF analyses and all the oxides, including water, were recalculated to 100%. The analyses were recalculated by computer on a basis of 24 oxygen atoms, using a program written by R. Phillips. Most are called tschermakite and a few pargasite, but the difference in name is not usually significant because they all fall almost midway between these two end members with the exception of 70879 and 71118.

It can be seen immediately from table VIII that there are few differences between the coast hornblendes and those from the inland laminated amphibolites, and none of the differences reaches the 99% significance level. The 95% significance level is exceeded by the differences between the values for Fe, Mg, Na, Ti and Mn. The coast hornblendes contain 7-8% less Fe_2O_3, 15% more MgO, 20% more MgO, 20% more Na_2O, 20% less TiO_2, and 30% less MnO than the inland laminated hornblendes.

The inland unlaminated hornblendes are more

TABLE VIII Hornblende Analyses

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO
Coast										
H.70879	43.85	14.31	3.49	7.10	16.75	10.88	2.05	0.38	1.07	0.13
H.70926	44.70	10.61	5.89	11.95	11.72	11.51	1.35	0.63	1.37	0.25
H.70930	44.98	10.04	6.10	12.38	11.32	11.90	1.16	0.48	1.53	0.11
H.70934	44.83	10.33	5.61	11.39	12.34	11.59	1.22	1.01	1.38	0.31
H.70938	45.76	10.47	5.39	10.93	13.21	11.36	1.17	0.43	1.00	0.27
Inland Unlaminated										
H.71009	43.90	9.78	6.71	13.61	10.40	11.64	0.94	1.19	1.46	0.36
H.71095	45.65	9.13	5.87	11.93	12.26	11.78	1.01	0.98	1.09	0.30
H.71118	46.95	6.99	5.69	11.56	13.29	12.29	0.96	0.79	1.18	0.30
H.71132	45.46	10.14	6.32	12.83	11.06	11.19	1.33	0.27	1.11	0.29
Inland Laminated										
H.51067	45.95	9.41	6.14	12.47	10.83	11.62	0.96	0.64	1.59	0.39
H.51080	44.34	10.74	6.33	12.84	9.61	12.36	1.16	0.65	1.63	0.34
H.71098	44.48	10.12	6.35	12.89	10.10	12.11	1.02	0.73	1.78	0.42
H.71184	45.90	9.02	6.05	12.28	11.00	12.17	0.96	0.63	1.64	0.34

TABLE VIII(cont'd) Basic formulae of analysed hornblendes

H.70879	$*_{0.23}\mathrm{Na}_{0.77}\mathrm{Ca}_{2.00}\mathrm{Mg}_{3.95}\mathrm{Al}_{1.05}\mathrm{Si}_{6.19}\mathrm{Al}_{1.81}\mathrm{O}_{22}(\mathrm{OH})_2$
H.70926	$*_{0.59}\mathrm{Na}_{0.41}\mathrm{Ca}_{1.91}\mathrm{Na}_{0.09}\mathrm{Mg}_{3.81}\mathrm{Al}_{1.19}\mathrm{Si}_{6.49}\mathrm{Al}_{1.51}\mathrm{O}_{22}(\mathrm{OH})_2$
H.70930	$*_{0.73}\mathrm{Na}_{0.27}\mathrm{Ca}_{1.85}\mathrm{Na}_{0.15}\mathrm{Mg}_{3.87}\mathrm{Al}_{1.20}\mathrm{Si}_{6.54}\mathrm{Al}_{1.46}\mathrm{O}_{22}(\mathrm{OH})_2$
H.70934	$*_{0.55}\mathrm{Na}_{0.45}\mathrm{Ca}_{1.92}\mathrm{Na}_{0.08}\mathrm{Mg}_{3.88}\mathrm{Al}_{1.12}\mathrm{Si}_{6.50}\mathrm{Al}_{1.50}\mathrm{O}_{22}(\mathrm{OH})_2$
H.70938	$*_{0.62}\mathrm{Na}_{0.38}\mathrm{Ca}_{1.97}\mathrm{Na}_{0.03}\mathrm{Mg}_{3.93}\mathrm{Al}_{1.07}\mathrm{Si}_{6.58}\mathrm{Al}_{1.42}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71009	$*_{0.46}\mathrm{Na}_{0.54}\mathrm{Ca}_{2.00}\mathrm{Mg}_{4.04}\mathrm{Al}_{0.96}\mathrm{Si}_{6.50}\mathrm{Al}_{1.50}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71095	$*_{0.47}\mathrm{Na}_{0.53}\mathrm{Ca}_{2.00}\mathrm{Mg}_{4.19}\mathrm{Al}_{0.81}\mathrm{Si}_{6.66}\mathrm{Al}_{1.34}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71118	$*_{0.45}\mathrm{Na}_{0.55}\mathrm{Ca}_{2.00}\mathrm{Mg}_{4.38}\mathrm{Al}_{0.62}\mathrm{Si}_{6.84}\mathrm{Al}_{1.16}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71132	$*_{0.60}\mathrm{Na}_{0.40}\mathrm{Ca}_{1.97}\mathrm{Na}_{0.03}\mathrm{Mg}_{4.00}\mathrm{Al}_{1.00}\mathrm{Si}_{6.63}\mathrm{Al}_{1.37}\mathrm{O}_{22}(\mathrm{OH})_2$
H.51067	$*_{0.65}\mathrm{Na}_{).35}\mathrm{Ca}_{1.96}\mathrm{Na}_{0.04}\mathrm{Mg}_{4.02}\mathrm{Al}_{0.98}\mathrm{Si}_{6.71}\mathrm{Al}_{1.29}\mathrm{O}_{22}(\mathrm{OH})_2$
H.51080	$*_{0.60}\mathrm{Na}_{0.40}\mathrm{Ca}_{1.95}\mathrm{Na}_{0.05}\mathrm{Mg}_{3.89}\mathrm{Al}_{1.12}\mathrm{Si}_{6.52}\mathrm{Al}_{1.48}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71098	$*_{0.66}\mathrm{Na}_{0.34}\mathrm{Ca}_{1.91}\mathrm{Na}_{0.09}\mathrm{Mg}_{4.00}\mathrm{Al}_{1.07}\mathrm{Si}_{6.54}\mathrm{Al}_{1.46}\mathrm{O}_{22}(\mathrm{OH})_2$
H.71184	$*_{0.71}\mathrm{Na}_{0.29}\mathrm{Ca}_{1.90}\mathrm{Na}_{0.10}\mathrm{Mg}_{4.10}\mathrm{Al}_{0.97}\mathrm{Si}_{6.70}\mathrm{Al}_{1.30}\mathrm{O}_{22}(\mathrm{OH})_2$

heterogeneous than the other two groups as are the corresponding amphibolites. The larger standard deviations (table IX) show this greater variety in composition, and table VIII shows that hornblende 71118 has the most extreme composition by being very low in Al_2O_3. The dearth of Al means that there can be very little substitution of Al for Mg in the Y site. This, together with Na occupying the A site, places the mineral in the edenite series.

Hornblende 70879 is situated well within the pargasite field, distinctly separate from all the other coast hornblendes. 70879 is approximately 40% richer in Al_2O_3, 40% richer in MgO, and 50% richer in Na_2O. Most of the excess is accounted for by 70879 being some 35% poorer in Fe_2O_3 than the other coast hornblendes.

These are the main points of the hornblende compositions and they can be explained by reference to the chemical and mineralogical compositions of the host amphibolites. Differences in the conditions of metamorphism in the two areas do not appear to have influenced the hornblende compositions directly, but only indirectly by controlling the mineral phases that could form.

F. Coast hornblendes

	Mean	Standard Deviation	t(FG)	t(FH)
SiO_2	45.068	0.412	0.632	0.201
Al_2O_3	10.362	0.211	1.890	1.349
Fe_2O_3	17.410	0.819	1.458	2.735
MgO	12.148	0.713	0.519	3.365
CaO	11.590	0.197	0.533	2.443
Na_2O	1.225	0.076	1.632	3.111
K_2O	0.637	0.227	0.718	0.188
TiO_2	1.320	0.195	0.777	2.830
MnO	0.235	0.075	1.672	2.879

G. Inland unlaminated hornblendes

SiO_2	45.490	1.082	0.422
Al_2O_3	9.010	1.221	1.014
Fe_2O_3	18.630	1.196	0.286
MgO	11.752	1.110	1.904
CaO	11.725	0.392	1.232
Na_2O	1.060	0.158	0.341
K_2O	0.807	0.341	0.731
TiO_2	1.210	0.148	4.735
MnO	0.312	0.028	2.361

H. Inland laminated hornblendes

SiO_2	45.168	0.759
Al_2O_3	9.822	0.660
Fe_2O_3	18.838	0.381
MgO	10.385	0.561
CaO	12.065	0.273
Na_2O	1.025	0.082
K_2O	0.663	0.040
TiO_2	1.660	0.072
MnO	0.372	0.034

TABLE IX Means, standard deviations and t-test values for analysed hornblendes

The difference between the inland and coastal hornblendes is the result of clinopyroxene and garnet being present in the coast amphibolites and absent in the inland amphibolites. In a rock containing almandine a large proportion of the iron in the system will be incorporated into the garnet, thus leaving less available for other minerals. This will account for the slightly lower amount of iron in the coast amphiboles than is found in the inland ones. There being good diadochy between ferrous iron and magnesium, the hornblende will readily take up more magnesium to make up for the deficiency in iron. Hence it is to be expected that the coast amphibolites will be slightly richer in MgO than the inland ones. In the case of Na_2O the hornblendes vary sympathetically with the rocks since neither, garnet or salite take significant amounts of sodium. The lower amounts of TiO_2 and MnO in the coast hornblendes are the result of the rocks containing sphene and garnet respectively. Sphene is common in the coast amphibolites, but not in the inland ones and since it can contain up to 40% TiO_2 even 1% sphene will make an important difference to the amount of titanium available for other minerals. Similarly manganese preferentially enters almandine, leaving less for the hornblende.

The two unusual hornblendes, 71118 and 70879, are remarkable because they occur in rocks of unusual compositions. Amphibolite 71118 is low in Na_2O and Al_2O_3 and the hornblende is correspondingly poor in these oxides. But because there is no plagioclase in the rock the hornblende in fact contains more of these elements than might be expected because it does not have to compete for them.

In the case of 70879 the rock composition more indirectly affected the hornblende composition. The rock contains appreciably more MgO than the other analysed amphibolites and so does the hornblende, but this simple relationship does not hold for the other elements. Because hypersthene is the only other mineral present in the rock there was less iron, but more sodium, calcium, and aluminium available for the hornblende than if plagioclase had been able to form. It is clear that orthopyroxene formed because of the rock composition and not because granulite facies metamorphism took place.

The main conclusion reached about the hornblendes in the Neria Area therefore is the same as that reached by Engel and Engel (1962b) in their examination of Adirondack amphibolites and hornblendes and is that bulk rock composition, directly or indirectly, influences hornblende composition much more than changes in temperature and pressure of metamorphism.

CHAPTER FIVE

ULTRA-BASIC ROCKS

Ultra-basic rocks are widespread over most of the Neria Area, but they are not uniformly distributed and they are not found at all in some parts. This is the case in the two Areas described here. Ultra-basic rocks are found in most parts of the Coastal Area, but in the Inland Area they occur only in the upper part of the Upper Biotite Gneiss Succession - in the north-west of the Area geographically. The main occurrences of ultra-basic rocks are indicated on Maps I and II. Ultra-basics are known to occur above the Upper Biotite Gneiss outside the Inland Area, but they do not occur below that horizon.

The Ultra-basic rocks take the form of unconnected pods or lenses which range in length from metres to hundreds of metres and have elliptical or circular horizontal sections. The lenses are invariably associated with amphibolite and, except where extreme deformation has occurred, are usually along one margin of the amphibolite. Not all the amphibolite horizons have associated ultra-basics, but most of those in the Coastal Area do. The distribution of the lenses is not uniform along the amphibolite sheets and while they may be common in one part they may disappear along the strike. Fig. 34 shows a typical small lens on the margin of one of the inland amphibolites.

In appearance the ultra-basics are massive fine grained rocks rusty brown on weathered surfaces and dark grey or blue-grey on fresh. Some of the rocks have nodular weathered surfaces caused by the weathering out of harder finer grained portions. Individual mineral grains usually cannot be seen by the unaided eye, but in some specimens mica flakes can be seen. Around the margins of some lenses, and lining fissures, tremolite has developed (fig. 35). The tremolite occurs as pale green fibrous crystals grown perpendicularly from the margin of the lens or fissure. It appears that tremolite is a late alteration product of earlier formed minerals, but in thin section even later minerals can be seen.

In addition to these ultra-basic rocks there are also a few ultra-mafics to be found. In hand specimen the ultra-mafics look like hornblendites, but in addition to hornblende there may be biotite or pyroxene present. Such rocks are not abundant. The best example occurs on the Coastal Area and it occurs as a thin horizon (ca. 15 m. thick) in a very hornblende rich gneiss. It has been seen at several localities along approximately the same strike, but the exposure between is very poor and it

Fig. 34 Typical small u.b. lens (under hammer in centre of photograph) between amphibolite (foreground) and biotite gneiss (background). Grid square B6, Map I.

Fig. 35 Fibrous tremolite developed radially from margin of an u.b. lens. Grid square B6, Map I.

has not been possible to link up the isolated outcrops. The strike of the outcrops is very close to that of the gneisses and it is difficult to determine whether it is part of the gneiss succession or a later intrusive. The amphibolite lenses found in the Neria Granite in the Inland Area are similar in composition and mineralogy to this rock (see above p.77).

Petrography of the ultra-basics

There is a variety of mineral assemblages to be found in the ultra-basic rocks at present and it is probable that both original mineralogy and the effects of metamorphism have produced this variety. For the purpose of description they can be broadly grouped on mineralogical grounds, but these groupings do not necessarily correspond to genetic groups. The two main groups used are a) ultra-basics composed of olivine, pyroxene, and/or amphibole, and b) ultra-basics that have been serpentinised or steatitised.

Non-serpentinised or-steatitised ultra-basics

Olivine is generally the most abundant mineral, but the amount present is variable and may be between 50% and 90% of the rock. The olivine composition has not been determined exactly, but it is magnesium rich, as is to be expected, having a 2V of nearly 90° and a negative sign. It is probably in the region of Fa_{15}

The olivine crystals have rounded subhedral outlines and do not appear to have been altered, although veins of serpentine, talc and in a few cases brucite cut some of them. The grains are usually 1-2 mm in diameter. There is no reaction between olivines and surrounding minerals and therefore they were probably stable under the conditions of metamorphism. The olivines may in fact be relics of the original igneous rock, but triple junctions of 120° where three olivine crystals meet suggest they may have been recrystallised. The absence of strain lamellae points to the same conclusion.

Both clino and ortho-pyroxene occur in these ultra-basics. The clinopyroxene is diopside and the ortho-pyroxene hypersthene. Diopside occurs as prismatic crystals of approximately similar size to the olivines. Identification is not always certain because much of the diopside has been converted to a colourless pale green hornblende. Thus it is impossible to make an estimate of the amount of diopside present. In one section (71146) large irregular patches of diopside occur partly enclosing olivines. The diopside patches show slight strain extinction.

The hypersthene has its usual distinctive pleochroism - X pink, Z very pale green and Y colourless. Usually only one cleavage is well developed so that many grains have no apparent cleavage and this, combined with their high relief, makes them look like olivine. However, the pleochroism and small negative 2V of the hypersthene usually enable the two to be distinguished. The shapes of the grains of the two minerals are usually identical.

An amphibole is often present and the habit of many of the crystals suggests that they may have been formed by alteration of diopside. The amphibole has the same prismatic form as the diopside, but it is often pleochroic and in the occasional cross-section the two amphibole cleavages can be seen. The optics of the amphibole are similar to those of the amphibolite hornblendes except that it has a much weaker green in the position of maximum absorption. This is a reflection of the different composition of the hornblende, which in turn reflects the different bulk rock composition.

Phyllosilicates are present in some of these ultra-basics, but they are not abundant, nor do they occur as large crystals. Where found they form colourless plates with dull grey interference colours. Phyllosilicates are better developed in some of the serpentinised ultra-basics and will therefore be described in detail with those rocks.

Accessory minerals are brucite and opaques. Iron oxides and chromite form the opaque minerals. Chromite is usually slightly more abundant than the iron oxides and occurs as rounded embayed grains. The iron oxides present are haematite and ilmenite, frequently intergrown with each other. Ilmenite occasionally occurs as lamellae in chromite and has the appearance of being younger than the chromite.

Brucite occurs as plates that have filled interstices between silicate grains, and also as a vein mineral. It resembles carbonate, but its low birefringence and uniaxial positive figure indicate that it is in fact brucite.

Serpentinised and steatitised ultra-basics

The soft steatitised ultra-basics are the rocks used by the Greenlanders as a sculpting medium. Formerly soapstone, as the rock is known locally, was used to make lamps and other practical objects as it was easily hollowed out or otherwise worked, but with the introduction of metal to Greenland this use rapidly declined and it is now used only for the production of ornaments.

The degree of hydrous alteration of the ultra-basics varies from incipient to almost complete. Both talc and serpentine may occur in the same rock, but the distribution of the two types of alteration is not haphazard. In the Coastal Area talc predominates over serpentine, and in the Inland Area serpentine is the main alteration product.

Relic olivine, pyroxene and amphibole occur in these altered ultra-basics as small scattered corroded grains.

Serpentine occurs as aggregates of sub-microscopic fibres. The fibres generally have grown out from a short spine and now form ovoids of serpentine. In other cases amphibole crystals have acted as nuclei for the serpentine and the fibres radiate out from the crystal. Talc occurs as circular aggregates of small, highly birefringent flakes. It has developed more haphazardly throughout the rocks than serpentine. Where the two occur together talc has overgrown the serpentine and is clearly younger.

In some of the strongly serpentinised ultra-basics of the Inland Area a phyllosilicate is moderately abundant. It is weakly pleochroic from colourless to extremely pale brown. The position of maximum absorption is with the cleavage trace perpendicular to the vibration direction of the polariser, i.e. the opposite to that usually observed in micas. In basal, or nearly basal, sections the mica can be seen to contain inclusions of rutile needles arranged in a triangular pattern. This suggests the mineral is, or was, phlogopite. The interference clours are anomalous deep blue and brown which make it impossible to obtain an interference figure. These anomalous colours are probably the result of the phlogopite having been chloritised. The phlogopite is clearly a late metamorphic mineral and may post-date the formation of the serpentine. The phlogopite is the only mineral in the ultra-basics to show post-crystallisation deformation. The crystals are often crossed by kink bands which have not been observed in any of the other minerals.

The opaque minerals are generally similar to those of the unserpentinised ultra-basics, but in some magnetite is obviously present in appreciable quantities because hand specimens will deflect a compass needle. The chromite-bearing rocks do not do this.

CHAPTER SIX

BROWN SCHIST

The rock described under this heading is not, in fact, a true schist and is more accurately described as biotite-sillimanite gneiss, but "brown schist" was a convenient field term to separate it from the acid gneiss, and it is retained here. The brown schist is a compact, but usually strongly foliated, red-brown rock, very rich in dark brown biotite and sometimes extremely garnetiferous. It is found only in the Inland Area and is a particularly striking rock which can be seen very easily from great distances on account of the contrast between its deep colour and the white of the gneiss. It forms an excellent marker horizon and is, in fact, better than the amphibolites as there is only one brown schist horizon and it is much less tectonically deformed than most amphibolites, although its thickness varies from place to place. It is much more resistent to granitisation than amphibolite and it can be traced easily through the Neria Granite in the south of the Area, although it has been intensely veined by intrusion of granitic material.

The brown schist can vary from red-brown to grey in colour depending on the colour and amount of biotite present. North of the eastern end of the major E-W fault, where two brown schist horizons run parallel to each other, there is a gradation between brown schist and amphibolite. As described above, (p.38) the amphibolite in this part of the Area is seen only as small blocks included in a much paler grey rock. This grey rock is brown schist that contains less biotite than usual and more felsic minerals. The amphibolite has been mixed with this and a gradational contact developed.

The amount of garnet and the freshness of that mineral are extremely variable. In the hinge zones of the major SW plunging folds the brown schist is extremely rich in fresh red garnets. The garnets have not developed their proper crystal form but are rounded and are rarely more than 1 cm. in diameter. But even here there are bands in the schist completely devoid of garnets. At localities along the limbs of the folds there may be no garnets present over large areas, but these are unusual. In certain bands within the schist individual garnets may grow to almost 2 cms diameter while adjacent bands will be full of many small garnets only millimetres in diameter. Sillimanite can be seen in many specimens of brown schist as felted mats of fine fibres. It is very unevenly distributed throughout the rocks and a hand specimen apparently rich in sillimanite may give a slice containing no sillimanite at all and the reverse may also occur.

Petrography of brown schists

The brown schists are composed of biotite, quartz, plagioclase, garnet, cordierite, and sillimanite, with apatite, zircon and opaque minerals as accessories.

Biotite forms 20-30% of most brown schists and is a deep red-brown, quite different from the green-brown biotites of the gneisses. It is pleochroic from colourless to brown. It occurs as ragged flakes with a very strong preferred orientation. Inclusions and pleochroic haloes are common. In a few specimens the biotite has been bleached. This is found particularly in specimens from near the major E-W fault and it is often associated with chloritisation of the garnet and was probably caused by alteration associated with the formation of the fault. Biotite appears to be unstable in all brown schists and therefore has a very ragged appearance.

Plagioclase is present in most specimens of brown schist but rarely exceeds 10% of the total rock volume. The plagioclase composition is extremely variable. In most sections it is in the range An_{30-45} but it may be as sodic as An_{25} or as calcic as An_{70}. The more calcic values were obtained from rocks from the south-west part of the area. In the plagioclases with high anorthite content, i.e. around An_{70}, very fine parallel lines inclined approximately 20° to (010) have been seen. It is not known what these are, whether a cleavage or a form of growth in the crystal. They have been observed in other high anorthite plagioclases from rocks such as gabbro anorthosite (Kalsbeek, personal communication) but no explanation of them has yet been advanced.

As far as can be determined optically there is little variation in the plagioclase composition within each specimen. Plagioclase usually occurs as interstitial grains, but in some sections plagioclase porphyroblasts are beginning to develop. Plagioclase almost invariably has convex margins to biotite and appears to have been growing as biotite was breaking down. Incipient sericitisation occurs in some plagioclases, but usually they are quite fresh. No alkali feldspar or muscovite occurs in brown schist, and biotite is the only potassium-bearing mineral. Some specimens of brown schist were stained with sodium cobaltinitrite in case alkali feldspar had been mistaken for cordierite or plagioclase, but no alkali feldspar was found.

Cordierite is present in most sections but it is sometimes difficult to distinguish it from plagioclase. Yellow pleochroic haloes around zircon grains and hexagonal twinning occur occasionally and enable the mineral to be positively identified. Cordierite usually occurs as small prophyroblasts but in places larger irregularly-shaped porphyroblasts have developed which enclose other minerals such as biotite and plagioclase, (fig. 36). Micrometric analysis of a number of brown schist sections indicates that cordierite may occupy up to 30% of the rock, but results obtained for cordierite (and therefore also for plagioclase) may not be very accurate.

Garnet forms up to 20% of the rock and is usually fresh and free from inclusions. Drop quartz, plagioclase, and small biotite flakes may all be found inside garnets, but usually they have effectively cleared themselves. Included grains do not have any preferred orientation, or spiral or sigmoidal arrangement, therefore no conclusions can be drawn about changes in the orientation of the foliation during the growth of the garnets. The garnets grew post-tectonically as they cut across the biotite and sillimanite crystals rather than deflect them. There has, however, been a slight deformation subsequent to the growth of the garnets because biotite flakes near the top and bottom of the garnet grains have been bent round them (fig. 37). Garnet is the most susceptible of all the brown schist minerals to alteration. Near the E-W fault it is frequently completely altered to chlorite or sericite and quartz, and in many other localities chloritisation has begun along cracks in the crystals. In the Neria Granite to the south, garnet is again the first mineral to break down and usually all that remains

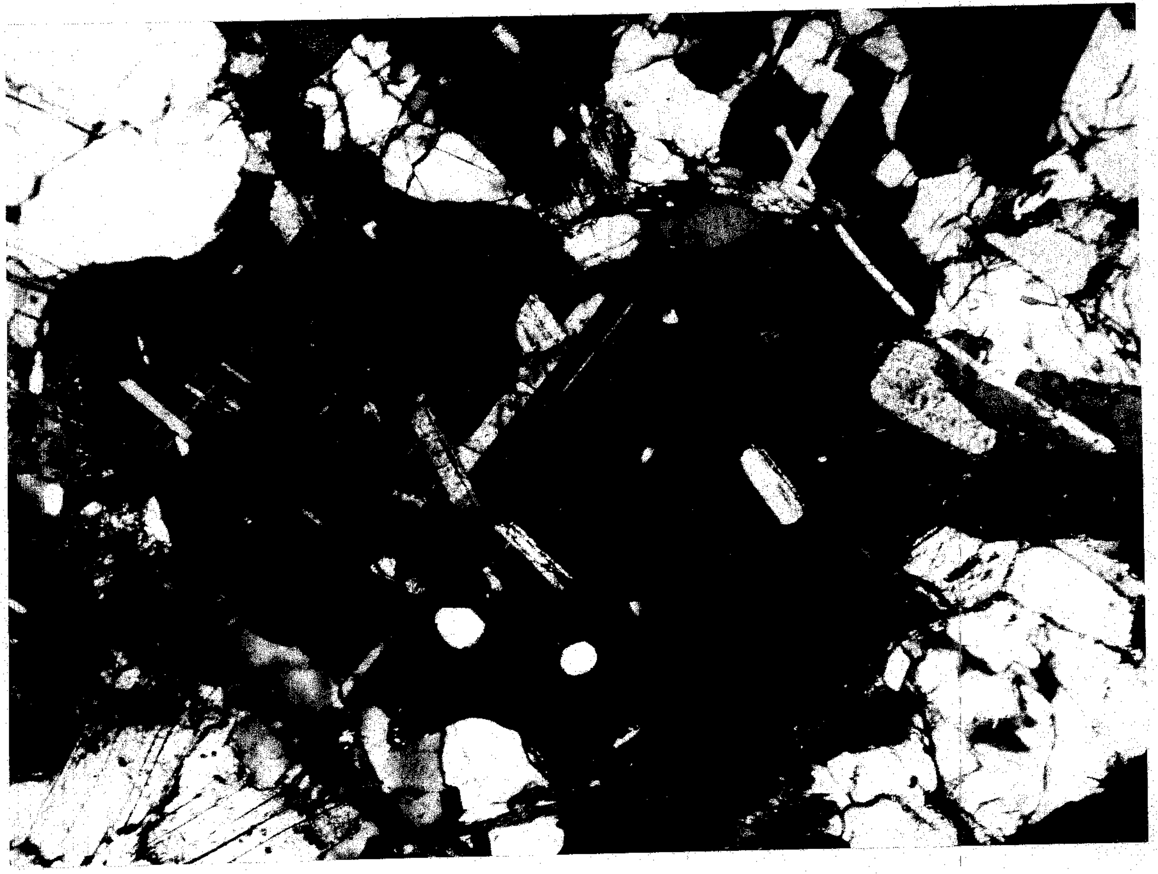


Fig. 36 Cordierite porphyroblast enclosing quartz and biotite grains in brown schist. Specimen 71033, x N, x 12.

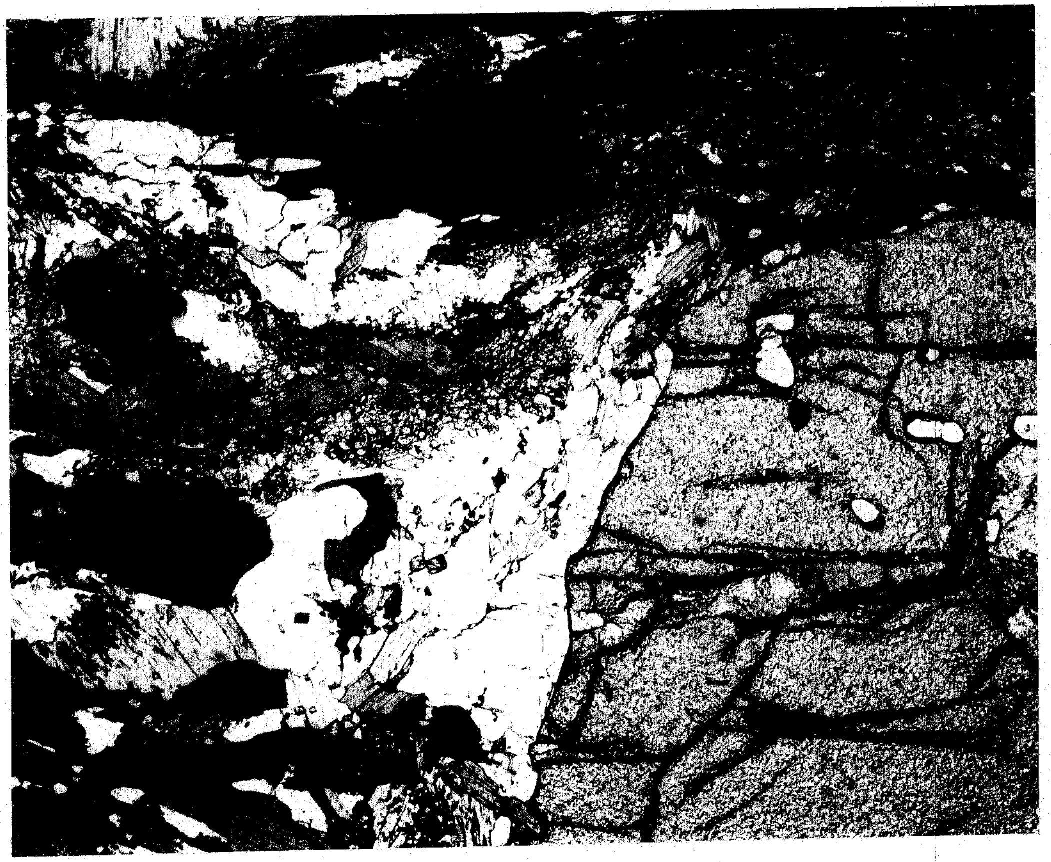


Fig. 37 Biotite crystals and trains of sillimanite deflected around a garnet crystal in brown schist. Specimen 71045, ppL x 10

are pseudomorphs of finely divided mica. At D2 7670 (Map I), which is just inside the granite, there is virtually no alteration of any of the minerals. But at El 8382, approximately 2 km. inside the granite, there garnet has decayed to chlorite.

Sillimanite occurs both as felted fibrous masses and also as discrete crystals which may be up to 0.5 mm. across in cross-section. The greatest volume of sillimanite observed in any sections was 15%. It is usually closely associated with biotite and has clearly formed later than it, but it is impossible to say whether it is forming from the biotite or merely nucleating there. The latter is the more likely (Chinner, 1961). Sillimanite is older than the garnet and the relationship between them is similar to that between garnet and biotite (fig.38). No kyanite has been observed in any brown schists.

Quartz occurs as irregularly-shaped grains with considerable range in size. Some recrystallised quartz occurs as masses of very small crystals, while older quartz occurs as strained straggling grains that may enclose other minerals. Highly lobate margins are not found and there is no evidence of any mylonitisation having occurred. The amount of quartz present is variable and has a range of approximately 20-50% of the rock. This probably reflects different proportions of sand in the original sediment.

Fig. 38 Large euhedra of sillimanite grown in intimate association with biotite. Brown schist 71045, ppL, x 64

Small rounded zircon grains occur rarely in the schists. They are typical of zircons that have been in a sedimentary environment, but are smaller and much less abundant than the zircons extracted from the gneisses.

Opaque minerals are ilmenite, which occurs as skeletal grains often intergrown with biotite; a sulphide, probably pyrites, which occurs as massive irregular grains; and another skeletal mineral which may be graphite.

Composition of the brown schists

Micrometric analysis of four sections of brown schist gave the compositions shown in Table X. Chemical analysis of a combined sample from the same four specimens gave a result very similar to the mean of the four micrometric analyses. Discrepancies between the chemical and micrometric analyses are probably the result of errors in the point counting. Opaque minerals were not counted and since ilmenite is present this would cause inaccuracies in the Ti percentage obtained. The much higher value obtained for Ca by chemical analysis is probably attributable to misidentification of plagioclase. If plagioclase were identified as cordierite in the counting, or if it be generally more calcic than it appeared to be from optical examination, then the micrometric values for Mg and Ca would be too high and low respectively. In general the brown schists appear to be similar to the haematite-free sillimanite gneisses from Glen Cova described by Chinner (1960), except for the dearth of alkalies in the brown schists, but even some of the Glen Cova rocks are deficient in these elements.

The oxidation state of the brown schists is also similar to that of the haematite-free Glen Cova gneisses. The oxidation ratio (mol. percent. $(2Fe_2O_3x100)/(2Fe_2O_3 + FeO)$) of the brown schists is 15.6. This falls comfortably within the range of the oxidation ratios of the Glen Cova gneisses that are devoid of haematite and have deep red-brown biotite.

Tourmaline associated with brown schist

Along the brown schist-gneiss contact tourmaline is often found. It occurs in two forms: either as large crystals intergrown with quartz in lens shaped pods up to 5 m long, or as smaller crystals replacing the usual minerals of the schist. In the latter case a hard, black, shiny rock is produced which sometimes forms a distinct horizon some 20 cm thick near the apparent base of the schist. The quartz-tourmaline pods occur right along the contact and the tourmalinised rock within 1-2m of the contact, but never in the gneiss. Pure quartz veins are often intimately associated with the tourmalinised rock, but the quartz and tourmaline are not intergrown as they are in the pods. The veins have margins of black brown schist in which tourmaline has formed (fig. 39). The tourmaline is genetically associated with the veins because the thickness of the tourmalinised margin is proportional to the thickness of the quartz vein; this can be clearly seen in fig.39.

In thin section the tourmaline has formed as rounded equant grains and it replaces all the other minerals, including garnet, which was the last formed metamorphic mineral. Therefore the formation of the tourmaline was a very late stage event. Fig. 40 is a photomicrograph showing the tourmaline replacing the other minerals of the schist.

It is most likely that the boron required to form the tourmaline was introduced with the quartz and not derived from the schist itself; however the latter possibility cannot be completely dismissed. If sufficient boron were available in the schist then it would probably have become mobile very early in the metamorphism and tourmaline would have formed much earlier. Zwart (1958) has described a similar occurrence of tourmaline-bearing schist from the Central Pyrenees and he concluded that the boron was introduced with the quartz.

The distribution of the tourmaline appears to be structurally controlled. It is found along the margin

Fig. 39 Quartz vein bounded by black tourmalinized brown schist. The thickness of the tourmalinized margin is related to the thickness of the vein. Grid square D5, Map I.

Fig. 40 Photomicrograph of a garnet crystal in brown schist (R.H.s of photograph) being replaced by tourmaline (L.H.s. of photograph). Specimen 71099 PPL, x 10.

of the schist that faces into the core of the antiform. It has been found at many places along the contact but not in a continuous band. None has been found along either of the contacts of the eastern schist horizon. It would appear most likely that the boron rich solutions percolated up through the gneisses until they reached the brown schist where a favourable environment for the crystallisation of tourmaline must have existed.

Specimen	Quartz	Cordierite	Biotite	Sillimanite	Garnet	Plag.
71033	48.5	28.9	21.9			0.6
71047	29.9	8.9	31.4	7.8	16.9	5.8
71044	48.5	10.3	24.3		7.2	9.7
71045	23.2	20.8	25.6	15.2	6.9	8.1

Specimen	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	H_2O
71033	70.9	14.2	0.3	5.6	4.9	0.1	0.3	1.9	0.6		1.2
71047	57.4	19.7	0.4	12.1	4.1	1.0	0.5	2.7	0.6	0.2	1.2
71044	69.5	13.0	0.4	7.7	3.3	1.4	0.5	2.1	0.7	0.1	1.1
71045	55.3	27.9	0.6	8.2	4.6	0.8	0.9	2.2	0.8	0.1	1.3
Mean	63.3	18.7	0.4	8.4	4.2	0.8	0.5	2.2	0.8	0.1	1.2
chem. mean	61.0	17.5	1.5	7.3	3.3	2.5	0.7	2.0	1.2	0.1	1.1

TABLE X

Modal and calculated oxide analyses of brown schists with one chemical analysis

Mineral analyses obtained from Deer, Howie and Zussman(1962)

Cordierite analysis 6, p.276, vol.1
Sillimanite analysis 3, p.123, vol.1
Garnet analysis 4, p. 86, vol.1
Plagioclase
71033 analysis 1, p.114, vol.4
71045 " 11, p.115, vol.4
71047 " 14, p.117, vol.4
71044 " 1, p.118, vol.4

Biotite analysis 12, p. 64, vol.3

Chemical analysis by Ib Sørensen

CHAPTER SEVEN

GRANITES AND PEGMATITES

In this chapter the rocks of approximately granitic composition that cut the country rocks are described. A simple classification of these younger rocks is used in the field. They are described as pegmatites and aplites or granites depending on the form of the body and the mineralogy of the rock. Essentially the terms as used here are structural terms - pegmatite or aplite is used to describe those bodies that have been deformed with the country rock and granite those that are younger than the main period of deformation. The structural classification usually corresponds to a mineralogical one in that pegmatites and aplites contain little or nothing apart from quartz and feldspar, while granites are finer grained and contain essential biotite or hornblende. It is not hard and fast, however, as undeformed pegmatites and deformed granites occur.

The granitic rocks of the Neria Area range in compositions from granite *sensu stricto* to granodiorite, but the calcium rich members of the suite are the most abundant. They are all called granite in the field for convenience and because they are all genetically related. Petrographically homogeneous gneiss and granite (strictly granodiorite or adamellite) are identical and they can be distinguished only when the field relations are known. Most of the granites were formed as the deformation of the rocks finished.

Granites

Small granites

The granite bodies occurring in the two areas are of varying size and irregular distribution. Some are small veins centimetres wide and metres long, while the largest is hundreds of square kilometres in extent. On the coast granites are found at very few localities, and then only as small veins or irregularly shaped bodies, but inland granite bodies up to several hundred square metres in area are found frequently. Both discordant and concordant contacts with the country rocks occur in the same body. The granite in grid square C7 (Map I) shows such features. It is approximately pear-shaped in plan and has discordant contacts except along its southern margin. In side elevation the granite has an almost vertical root that is sub-parallel to the gneiss banding, and then spreads out horizontally across the gneiss structures. The granite has in places a very weak foliation that is generally parallel to the gneiss foliation, and there are also large feldspar porphyroblasts scattered irregularly throughout it.

Similar granites, with features identical to the one described above, although differing in shape, occur at many localities in the north-west part of the Inland Area, but one in grid square E6 (Map I) shows many of the features particularly well, (Fig.41). The inclusions in this body are in various states of preservations; some are partly assimilated, while others are completely unaltered. In fig.42 two amphibolitic inclusions can be seen, and the presence of these shows that movement occurred during the formation of the granite because there is no amphibolitic gneiss adjacent to the granite in its present situation. Feldspar porphyroblasts occur in this granite (fig.43), and there is also a part of it that is much coarser grained than the rest (fig.44).

Usually the only structure these granites have is an extremely weak foliation. But a small granite in grid square F5 has a strong foliation and a weak lineation. The granite is discordant and cuts across small folds in the gneiss. The foliation is parallel to that in the surrounding gneiss and the lineation is parallel to the axes of the major folds of the area (plunging south-west) which were formed during the last phase of deformation.

The granites described above are clearly intrusive in origin, but there are a number of small granite bodies that appear to have formed by replacement. These

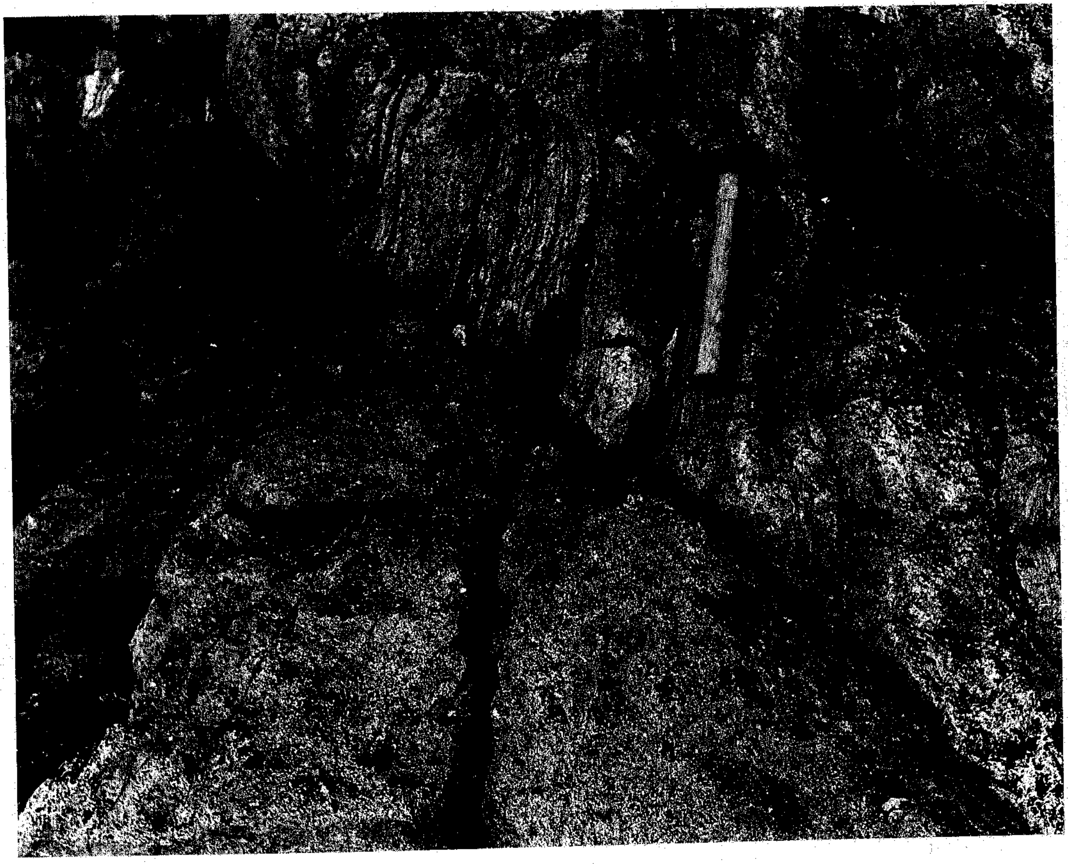


Fig. 41 Discordant contact of granite against banded biotite gneiss. Grid square E6, Map I

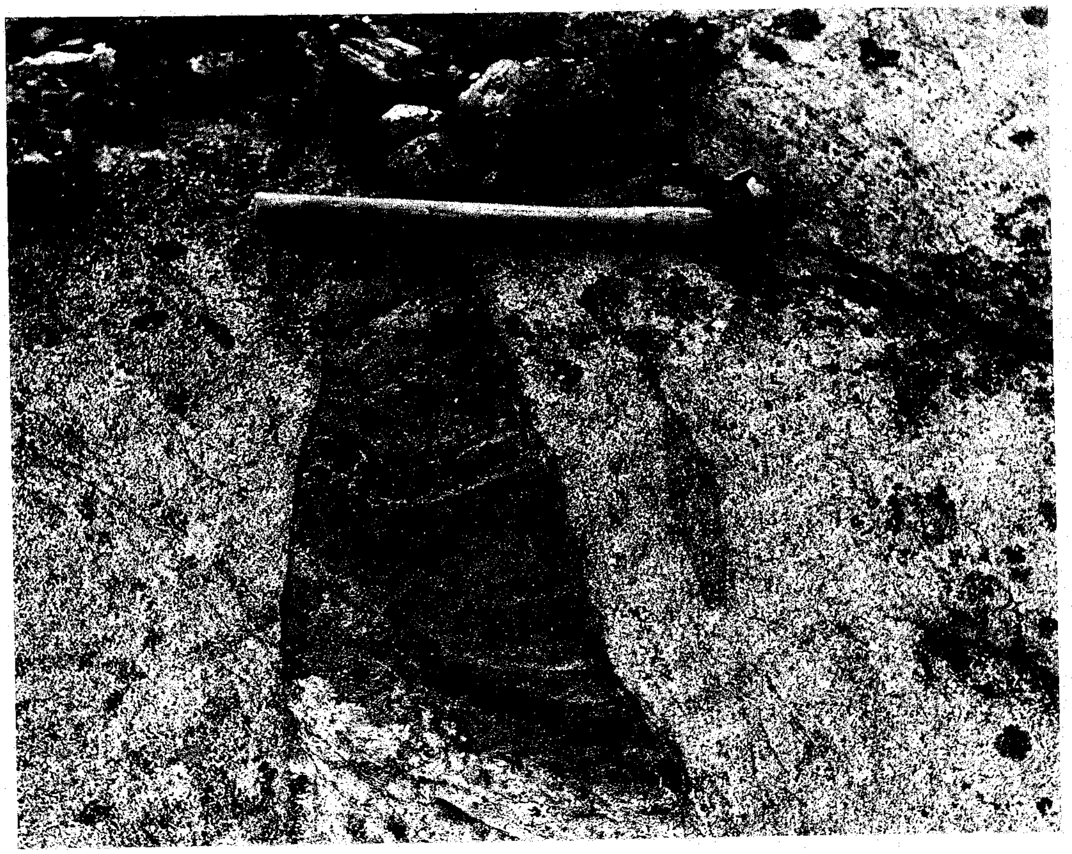


Fig. 42 Amphibolite inclusions in granite. The small inclusion on the right has been partly transformed, whereas the large one shows no evidence of alteration. Grid square E6, Map I.

Fig.43 Microcline porphyroblasts in granite.
Grid square E6, Map I.

Fig.44 White coarse grained granite in finer grained dark granite. Grid square E6, Map I.

are generally in the form of dykes and may be up to 5m. thick. Fig. 45 shows part of one of these dyke-like bodies that has been faulted and veined by epidote. The replacive nature of this body is shown by the presence of palimpsest isoclinal folds of the pre-existing banded gneiss. Replacement granite bodies are distinguished from intrusive ones by the complete absence of any shearing along their margins. Although they are approximately parallel to the gneiss banding, in detail their margins are often discordant to the banding. Their contacts are sharp and transition zones do not occur.

In the same area as this replacement dyke (grid square E4 Map I) a black biotite schist is partly transformed to granite, (fig.46). The contact between granite and schist is sharp and the only intermediate stage seen is the incipient development of feldspar porphyroblasts above the granite. The porphyroblasts rapidly decrease in abundance upwards. Nuclei and bands of granite have developed in the schist ahead of the main mass of granite. In the granite itself there are relics of schist that have not been completely digested (fig.47). In thin section the relics are seen to be composed largely of chloritised garnets.

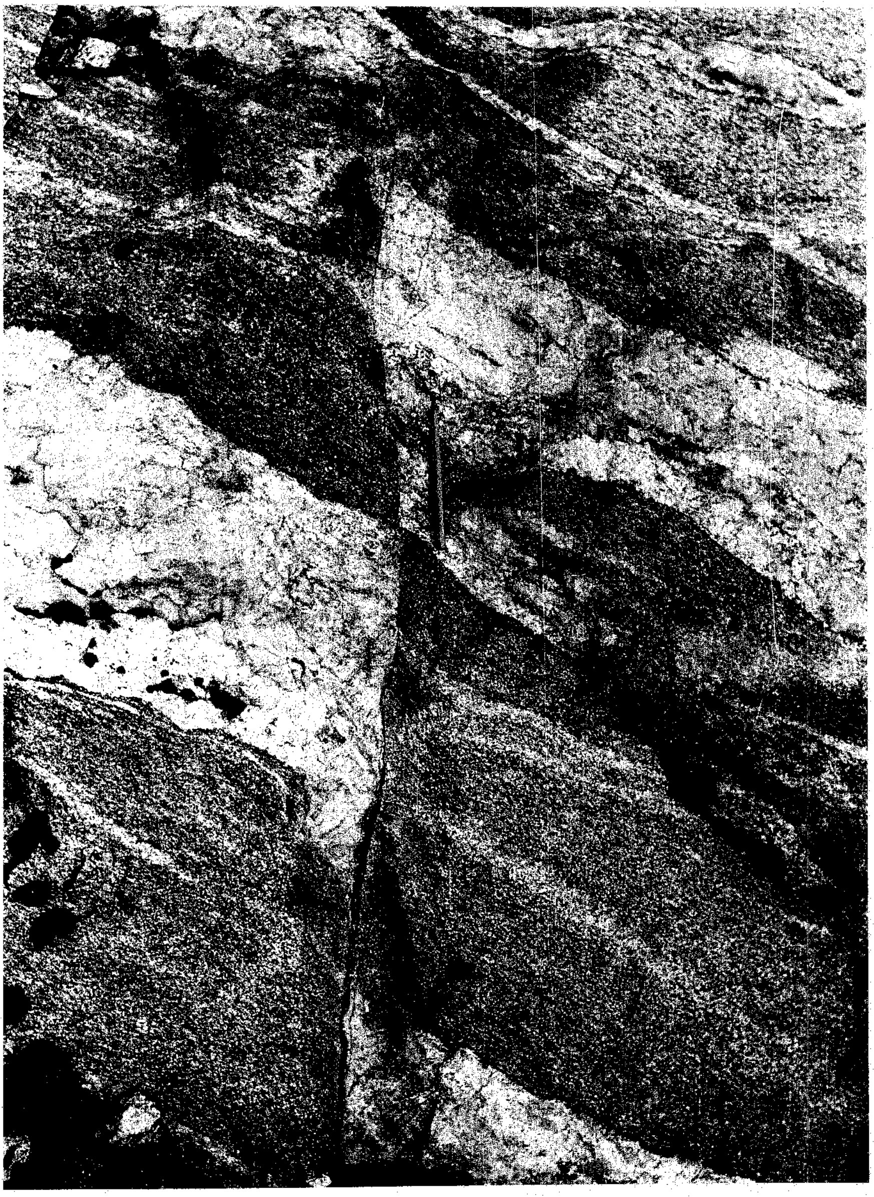


Fig. 45 Part of a replacement granite body. Palimpsest folds of the original compositional banding can still be seen. The fault parallel to the pencil is filled with epidote. Grid square E4, Map I

Fig.46 Brown schist partly granitised from below. The centres of nucleation above the main body of granite can clearly be seen above the hammer handle. Grid square G4, Map I

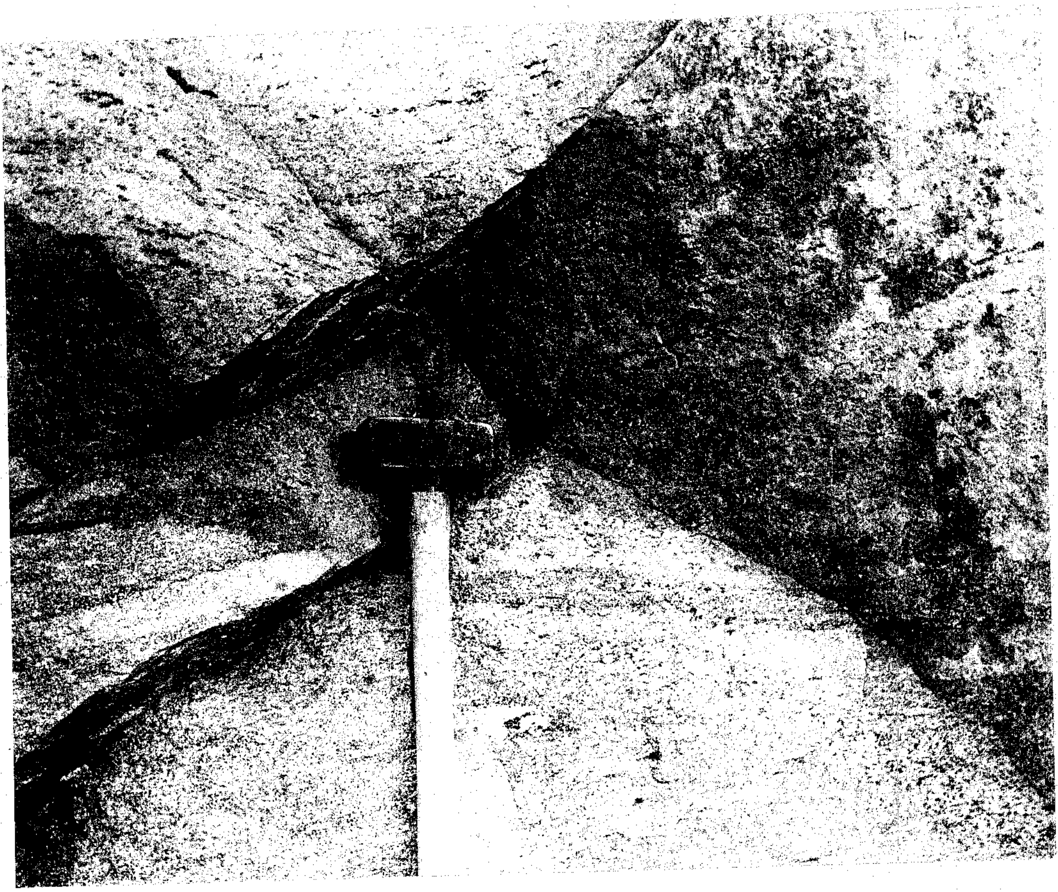


Fig. 47 Relic schlieren of brown schist below the granite front seen in fig. 46

Neria Granite

The large granite body occurring in the southern part of the Inland Area, called the Neria Granite, is largely replacive in origin, but structurally it is quite different from the small discordant granites described above. It does not have a sharp contact with the surrounding gneisses, but instead there is a gradual transition from granite to normal banded gneiss over a distance of 1-2 kms. The contact relations are best seen in the large U-shaped valley forming the southern boundary of the Area. An increase in the intensity of veining of the gneiss occurs up the valley, and simultaneously the structures of the gneiss become less sharply defined. By the time the 720 lake is reached these two effects have changed the rock completely and no structures, except for a weak foliation, are left. It is not known whether the foliation at present seen in the granite is a relic from the gneiss or a new foliation formed by recrystallisation, under stress. There is comparatively little veining in the granite itself and large areas may be completely homogeneous. But at a few localities white, sometimes orthite-bearing, coarse-grained pegmatites intrude the granite.

More common than pegmatites, however, is a younger granite intruding the main, older granite. The younger granite is seen nearly everywhere but it is particularly abundant in the side valley branching off to the north from the main one. The younger granite is much lighter in colour than the older and usually occurs in irregular masses, but in the side valley there is a large dyke of younger granite cutting vertically through the older (fig.48). The contact relations between the two granites show clearly that the younger intruded the older as very mobile material. It must be stressed that "granite" is not being used in the strict petrographical sense as the older granite is granodiorite or even quartz diorite in composition. The younger is rarely more potassic than adamellite and is usually a granodiorite.

The older granite does not intrude the brown schist horizon to any appreciable extent, and from the fact that the schist horizon can be traced intact for a great distance it is clear that very little movement took place during the formation of the older granite. It is probable that slight movement occurred because the style of the folding inside the granite is more irregular than that outside. At one point the schist horizon is only some 10-20 m. thick, and this thinning may be the result of stretching during the formation of the granite. Any movement that did occur was only local and it is not

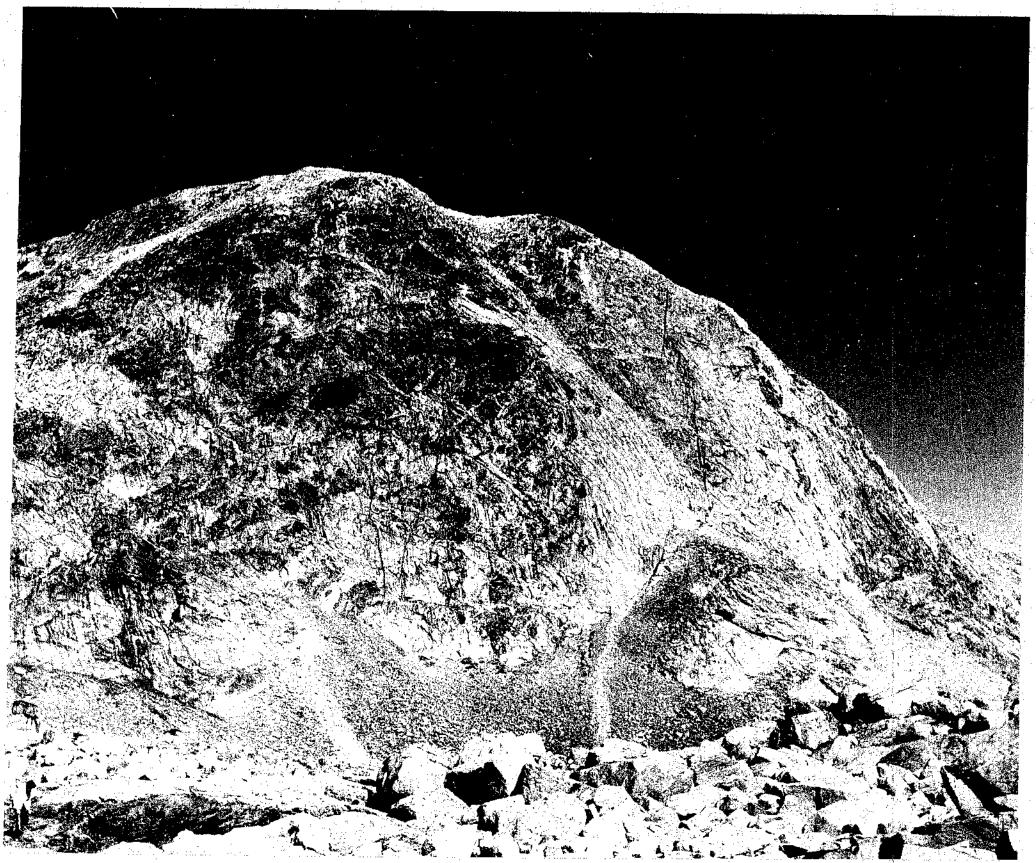


Fig. 48 A wide "dyke" of white younger granite cutting older Neria Granite. Grid Square D2, Map I. The mountain is 500 m. high

suggested that the granite was emplaced by intrusion of a solid mass in the manner of a salt dome.

To the north of the Neria Granite there are two areas of coarse grained leucocratic granite (Map I). Some of the contacts of these bodies are sharp and can be easily mapped, whereas others are gradational. The form the gradation takes is a gradual increase in the volume of granitic material present in the country rock, until isolated blocks of gneiss are found in a large mass of homogeneous pink granite. The blocks of country rock appear to have been rotated since their internal foliation is often widely divergent from the regional trend of the gneiss foliation. There is no gradation between the granite and the gneiss inclusions in it: the contacts are completely sharp. The sharp contacts of the granite mass as a whole are quite discordant to the gneiss structures. No roots have been observed to these masses and they have the form of flat-lying bodies capping the ends of two ridges.

These two granites are clearly late bodies, but their relationships to the others are not clear as they do not intersect any other granites. It is unlikely that they are much younger than the Neria Granite and locally abundant orthite-bearing pegmatites in the two northern granites suggests that there is a connection between them and the Neria Granite. The connection may be that they are a late intrusive stage of the main granite as the field relations show clearly that they are intrusive and not transformative in origin, their coarse grain and leucocratic character would also fit in with such an origin.

Petrography of the Granites

Small granites

True granite, i.e. with alkali feldspar forming at least two thirds of the feldspar, occurs at only two localities, both in the northern part of the Inland Area. They are both late discordant granite bodies. The body described above from grid square C7, Map I, is composed predominantly of quartz, alkali feldspar, and plagioclase. The amount of alkali feldspar (microcline) in the rock varies throughout the body and is related to the grain size of the rock. The later, coarse-grained parts of the granite are much richer in microcline than the earlier, finer-grained parts. It is clear that enrichment in potassium occurred later in the development of the granite because microcline is present in all specimens, but in the finer-grained specimens microcline is present as small interstitial patches and porphyroblasts are rare or absent. In the coarse-grained granite microcline and quartz comprise nearly the whole of the rock. Plagioclase is very subordinate and mafic minerals are almost completely absent. Table XI shows the modal compositions of the coarse- and fine-grained granites from this body. The microcline porphyroblasts may exceed 5 mm. in diameter in this rock. In the finer-grained, more granodioritic, rocks the grain size is usually 2-4 mm. Vein perthite is common in the microcline.

Plagioclase is present usually as rounded porphyroblasts, but there are also more elongate grains that have vaguely lath-like forms. The composition is usually in the range An_{25-30}. Many of the non-porphyroblastic plagioclases have Carlsbad or combined Carlsbad-Albite twinning. This is taken as evidence of the presence of a liquid melt at one stage. It has been found (Tobi, 1961) that Carlsbad twins do not occur in metamorphic plagioclase except in albite porphyroblasts in rocks metamorphosed under greenschist facies conditions. Therefore, part at least of these small granites must have crystallised from a silicate melt. The plagioclase porphyroblasts, however, probably developed in what was essentially a solid rock immediately after crystallisation. Slight sericitisation of the plagioclase has occurred and only the smaller porphyroblasts are completely altered.

Quartz is present in large quantities in all the granites and always shows signs of strain. Two generations

TABLE XI

Modal analyses of granite

	Quartz	Microcline	Plagioclase	Biotite	Epidote	Sphene	Orthite
71166	36	42	21	0.3			
71165	27	5	48	18		1	0.3
51018	30	30	29	9	0.2		1

are present - an older and more abundant highly strained generation, and a younger strain-free generation of drop-quartz. The latter is most often seen enclosed in feldspar porphyroblasts but it also occurs in the enclosing groundmass. The strained quartz occasionally shows evidence of very weak mylonitisation, such as very small recrystallised grains and highly lobate or even amoeboid larger grains. Vermicular quartz occurs in a small number of feldspar grains, but true myrmekite along the interfaces between alkali feldspar and plagioclase is almost completely absent.

The mafic mineral is biotite in this granite and it has formed under stress because in thin section the flakes are all aligned roughly parallel, although no foliation may be visible in the outcrop. Much of the biotite is fresh and dark brown, but some of it has been chloritised and is now green. The latter has unusual purple interference colours in basal sections, which are probably caused by interaction between the normal isotropic appearance of basal sections and the interference colours of incomplete chloritisation. Except in the coarse-grained parts of the granite, biotite is present in larger and better preserved flakes than the biotite in the felsic parts of the gneisses.

The granite has a well-developed suite of accessory minerals comprising sphene, orthite, zircon, apatite, epidote, and ore minerals. Sphene is particularly striking. It occurs as abundant, well-formed, diamond-shaped crystals and also as irregular and rounded ones. The largest crystals may be up to 1mm. long and be twinned. The pleochroism is usually very strong pale-brown (Y) to colourless (X). One large, well-shaped grain has a strongly pleochroic centre and an almost colourless non-pleochroic rim.

Orthite is less abundant than sphene but also occurs in unusually large and well-formed crystals. A few of the larger grains have good crystal faces and angles developed. Orthite crystals also may be up to 1 mm. long. Many of the smaller grains are completely metamict, but the large ones frequently have a non-metamict core and a metamict rim. Where orthite abuts against biotite there is a black rim in the biotite where radiation from the orthite has destroyed the biotite crystal structure. The metamict orthite is golden-brown in colour and isotropic, while the non-metamict parts are pleochroic olive green (Y), and pale brown (Z).

Zircon crystals are more abundant in some specimens of the granite than in the surrounding gneisses, but they are not the large euhedral crystals that are found in some high level igneous granites. They occur in this granite as clear crystals that are exceptionally up to 0.5 mm long. The corners are usually rounded but the angles are still visible. Some grains, however, have completely rounded, and others broken, ends. The number of zircons with well-preserved prism faces is much greater than in the gneisses. There is little scratching of the surfaces of the zircons and they do not have the brown appearance of those found in the gneisses.

Ore minerals comprise up to 2% of the rock in the finer-grained varieties and consist of haematite and pyrites, with possibly also magnetite and ilmenite. Haematite is not usually found as an original constituent in the rocks of the area as a whole, and does not appear in the norms of the analysed amphibolites. But it is common in many of the fault and mylonite zones as a vein mineral or finely disseminated particles. It is probable that the haematite in this granite has been introduced after the formation of the granite, and may be related to the haematite found in the near-by E-W fault.

The small granite in grid square F5 (Map I) is very similar in thin section to the rocks just described, but has less abundant accessory minerals. The modal composition is given in table XI and it can be seen that the rock is an adamellite with a very low proportion of biotite. The value for biotite is probably low because the section was cut perpendicular to the lineation and foliation. Thus only thin, very irregular ends of biotite crystals appeared in the section. The orthite appeared to have had its crystallisation controlled by the stress field acting on the granite as most of the crystals in the section appeared in cross-section and are clearly elongated parallel to the lineation.

The presence of 30% microcline in this rock suggests that it was formed as one of the later granites because microcline is always associated with the later granites, but the structural evidence suggests that in fact it was formed earlier than the other discordant granites. The explanation of this contradiction may lie in the situation of the granite. It lies in the centre of the major antiform where the shearing stress may have lasted longer than on the limbs, which is where most of the other granites occur.

The descriptions given above is generally applicable to all the discordant granite bodies, with only minor variations. The microcline will be more or less well developed. The accessories may not be so varied or abundant. Some may contain pegmatitic segregations. They are all of different and irregular shapes and sizes. Apart from these differences, however, they are so similar that it is clear they are all related and have a common origin.

Neria granite

The mineralogy of the older granite in the Neria Granite is related to the composition of the original rock from which it was derived. The most common paragenesis is quartz, plagioclase and biotite, which is the same compositionally, but not texturally, as the acid gneiss. Where a former amphibolite occurred, hornblende is found in the granite. Amphibolite relics occur in grid squares D1 and the older granite surrounding them contains approximately 10% green hornblende in addition to the biotite. The hornblende is clearly out of equilibrium and is a relic from the original amphibolite. Biotite occurs as well-formed laths, usually dark brown in colour, and imparts a weak foliation to the rock. Pleochroic haloes are common in the biotites. In some specimens of the older granite it forms as much as 25-30% of the rock. Skeletal ilmenite is often found intergrown with it. Ilmenite is the only common ore mineral in this granite.

The almost complete exclusion of alkali feldspar from the older granite shows that it is a granodiorite. Plagioclase forms 30-40% of the rock usually and occurs as interlocking grains giving the rock a granoblastic texture. The usual grain size is 1-2 mm. but some plagioclase crystals may have grown to twice the average to form small porphyroblasts. Combined Carlsbad-albite twinning is common and from grains with such twinning a composition of An_{35-37} is usually obtained. The presence of such twins supports the field evidence indicating melting and localised magmatic intrusion. Saussuritisation has occurred in much of the plagioclase, but it is by no means complete.

Quartz occurs in the manner usual to granites of this type. The crystals are a little smaller than the plagioclase and they are strained. Sutured margins do not occur on any of the quartz grains. Small clusters of unstrained recrystallised quartz occur scattered throughout the rock, but they comprise a very small proportion of the total quartz. Vermicular quartz is present in some small microcline grains, but it is not frequently found.

There are fewer accessories than in some of the discordant granites. Apatite, epidote, and zircon are still found, but in smaller quantities, except for epidote which is an alteration product anyway. Sphene has not been seen and orthite is very rare and forms only very small crystals when it is present.

The younger granite is typified by a smaller proportion of mafic minerals than the older, a weaker foliation, and the presence of microcline. The last is the most significant difference, but not all specimens of the younger granite contain microcline uniformly distributed throughout. However, they usually do and it may form 30-40% of the rock. Its mode of occurrence does not differ appreciably from that of the microcline in the discordant granites. Etching and staining specimens containing the older and younger granites in contact with one another shows that the microcline is indeed confined to the younger granite except for very small microclines concentrated along the margin of the older granite. This is explained by permeation of pottassium from the younger into the older granite. Microcline appears from microscopic examination to have formed after all the other minerals had crystallised and this is in accord with it being the low temperature form of potassium feldspar.

Plagioclase usually forms less than 10% of the rock, but one specimen contained only a few small interstitial grains of microcline and had 20-30% of plagioclase. Plagioclase in the younger granite is between An_{25} and An_{30} in composition. This rock could be seen to be younger granite because it intruded a darker granite. At the locality where it occurred, however, the original country rock was apparently amphibolite, and it is possible that the younger granite produced from this, while being more felsic than the older granite, was more basic than younger granite produced from acid gneiss. There is usually stronger saussuritisation in the plagioclases of the younger granite than in those of the older.

Biotite is less abundant than it is in the older granite and often does not impart any foliation to the younger granite. Otherwise it is very similar in its mode of occurrence to that in the older granite. Accessory minerals are the same as those found in the older granite and are again poorly represented. Opaque minerals may be completely absent in some specimens of the younger granite.

Pegmatites

Masses of quartz-feldspar material cut all the country rocks of the Neria Area and locally form a very high proportion of the total rock volume. The forms of these masses are very variable and range from parallel-sided dyke -like bodies of uniform strike and dip, through ptygmatically folded and necked sheets to large irregular bodies of indefinite shape. Both concordant and discordant pegmatites occur, but the former are frequently the result of deformation after emplacement. This can clearly be seen when apparently concordant pegmatites can be viewed parallel to the strike. Fig. 14 shows such bodies: a horizontal section shows a series of apparently concordant thin pegmatite veins, but a vertical section perpendicular to the strike shows discordant veins that have been folded and sheared into approximately conformity to the layering of the host rock. Clearly, for this to happen the strike of the veins cannot deviate widely from that of the plane of shearing (Watterson 1967). Most pegmatite veins show some deformation but a few are apparently undeformed (fig.49), but this does not mean that they were emplaced after all deformation was finished because a flat plane does not necessarily become curved during shearing deformation.

Approximately planar bodies are most abundant in the gneisses, but they are also found cutting all the other rock types whereas the much larger and more irregularly shaped masses are almost entirely confined to amphibolite and brown schist host rocks. Fig.50 shows some very large bodies in the brown schists.

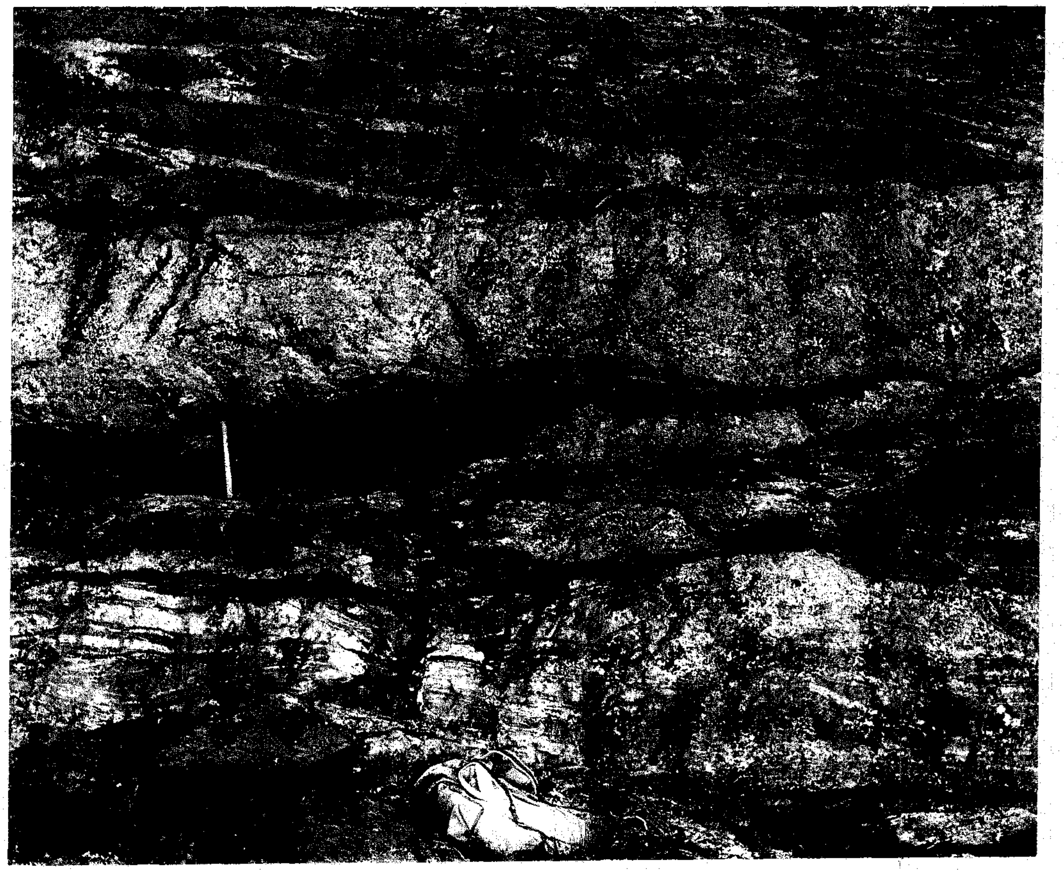


Fig. 49 An apparently undeformed discordant pegmatite cutting banded gneiss in the Upper Biotite Gneiss Succession. Grid square B5, Map I

Fig. 50 Large lenticular masses of pegmatitic material enclosed in brown schist. Grid square D4, Map I

With the large vertical exposure (some 350 m) it can be seen that the pegmatite bodies do not continue in depth and therefore cannot have been emplaced by injection of magma from below.

The largest development of pegmatitic material is in the Laminated Amphibolite, as already described (page 38) it occupies a major part of the horizon. There is no evidence to suggest that it was formed by transformation of the amphibolite. All contacts between the two rocks are sharp. The pegmatite is foliated and the foliation is curved round the included blocks of amphibolite (fig. 51) which not infrequently appear to have been rotated as the strike and dip of the foliation of the blocks bears no resemblance to that of the local dip and strike of the amphibolite horizon.

At only one locality has a reaction rim been observed between a pegmatite and its host rock (apart from the quartz veins bordered by tourmalinised brown schist) and that is in the Neria Granite. In this example a pink quartz-feldspar pegmatite 20 cm thick has a margin of biotite rich rock one to two biotite flakes wide.

Pegmatites can occasionally be related to folds. Fig. 52 shows a number of pink quartz-feldspar

Fig. 51 Sharp contact between pegmatitic material and enclosed block of laminated amphibolite Grid square F6, Map I

Fig. 52 Pink axial plane pegmatites cutting small-folded amphibolite gneiss. Grid square 5F, Map II

pegmatites that have developed along the axial planes of small folds in the rock.

Pegmatite mineralogy

The mineralogy of the great majority of pegmatites in the areas is very simple and is comprised of quartz and plagioclase. Alkali feldspar is present in a few pegmatites, which are usually pink or red instead of white, but it has never been seen to be more abundant than plagioclase. Muscovite is the most common other mineral and it usually occurs as thick books of flakes a few centimetres across. In the large masses in the brown schists where the grain size is smaller biotite occurs and is uniformly distributed throughout the rock as small flakes. At one locality on the coast of Sermilik Fjord a number of small actinolite-bearing pegmatites were found, but apart from these the only other unusual mineral found in a pegmatite is the orthite of the eastern part of the Inland Area. The orthite occurs as well formed crystals up to 1-2 cm long. It is largely metamict, but some of the larger grains retain a shiny black unaltered core. The feldspar crystals around the metamict orthite are frequently split by radiating cracks running out from the orthite. These are the result of expansion of the orthite during its self-destruction.

Age of the granites and pegmatites

Pegmatites

Pegmatites were formed throughout a long period of time as is shown by the relationships between different pegmatite veins, and between veins and folds. Folds of all ages deform and are cut by leucocratic veins so it is clear that such veins were formed throughout the whole of the plutonic events to which the rocks were subjected, but production was not necessarily continuous. The latest pegmatites were formed very late indeed as they cut the Neria Granite, which is thought to have been formed as the final product of the pre-Ketilidian plutonism.

Granites

The period during which granites were formed was only short. They are certainly younger than the main period of folding and migmatisation because they can in numerous places be seen to cut second period folds and pegmatites. No granite has been seen to cut the latest folds, but neither is there any evidence to suggest that the granites were older than this latest folding. Since the Neria Granite is a very large body and since the last folding gave rise to extremely large folds it is probable that large scale deformation of the Neria Granite would have resulted had the folding succeeded the formation of the granite. Therefore, it is concluded that the granites were formed at the end of the plutonic events to which the Neria rocks were subjected.

The other limit to the age of the granite is imposed by the metadolerite dykes which are abundant in the region. In the Neria Area the dykes are all younger than the granite, but further south the same dykes have been demonstrated to be older than Ketilidian granites (Watterson, 1965). Therefore, the Neria Area granites are pre-Ketilidian.

Formation of the granite

For the purposes of description the granites have been separated into small discordant bodies and the Neria Granite, but because they were formed at the same period of time it is probable that they are all genetically related and had a common origin.

The features of the Neria Granite described above point to the Granite being a catazonal granite (Buddington, 1959) which formed as an end product of the regional metamorphism. The field and textural evidence all indicate that the granite was formed by recrystallization of the acid gneisses with only small scale movement of elements. Most of the granite is biotite-bearing, but locally it is hornblendic where it was formed from amphibolite or hornblende gneiss. The homogenisation of the gneiss banding took place in both the solid and liquid states. Textural evidence such as porphyroblastic plagioclase suggests a liquid phase was not involved, but the presence of Carlsbad twins indicates otherwise.

For melting to have occurred volatiles (primarily water) must have become available because in the dry at 5Kbars pressure (see below p 181) granite will not melt until a temperature much in excess of 700°C has been attained. In wet conditions, on the other hand, granite starts to melt at between 650°C and 700°C at 5Kbars (Tuttle & Bowen 1958, p.122). The water content of the Neria Gneisses is not more than 2% as biotite is the only water-bearing mineral present. At 5 Kbars and 700°C very little granite magma will form with only 2% water, and it is for this reason that the whole region was not transformed to granite. Thus an influx of water is the most likely reason for the localisation of the Granite. But there is no clue as to its source.

The experimental work done by Winkler and von Platen predicts that "granite" melts should be produced in high metamorphic terrains provided there is ample water (Winkler, 1965). Winkler (op. cit., p.195) describes the melting of a gneiss (gneiss B) which is very similar in composition to the Neria gneisses (50% quartz, 40% plagioclase, 10% biotite). The final melt obtained from this rock is almost trondjemitic in composition, as is the older part of the Neria Granite. When the younger part of the Neria Granite is taken into account, however, the experimental data does not fit the field evidence. According to Winkler the earliest melts are richer in K than the later, whereas in the field the younger granite is richer in K. This is a general rule, and from many shield areas, e.g. Central Africa, Scandinavia, Scotland, the older granites are deficient in K (granodiorites, trondjemites, etc) and the younger are less deficient (adamellites, granites).

The presence of potash feldspare in younger granites in general may be more easily explained by reference to the quartz-albite-orthoclase system. As water vapour pressure decreases the ternary minimum for the system moves away from the albite apex (Tuttle and Bowen, 1958, fig.38) and this may result in later formed granites containing potash feldspar. This may be the reason for the presence of microcline in the Neria Granite, but many of the microcline grains are porphyroblasts and their occurrence in the younger granite is very similar to their occurrence in the country gneisses. There is not, however, any spatial relationship between the distribution of granite and the presence of microcline. The porphyroblastic microcline is best explained by postulating a late, widespread and haphazard K metasomatism. The K must have been introduced from some outside sources as otherwise it would have been expected to become mobile sooner. There is at present no explanation for why the K preferentially avoided the older granite. The problem of the microcline is similar to that posed by the orthoclase porphyroblasts of the Shap Granite, but at least at Shap the distribution of the porphyroblasts is clearly related to the position of the granite boss.

It is of interest to note that Winkler's work provides an explanation for the resistance of the brown schists to granitisation. The mineral assemblage of the brown schists is very similar to the restate phase which he suggests may be left after the partial melting of Al-rich sediments. Therefore it is to be expected that the brown schists will not have melted very much. The main change that has occurred in the brown schist in the Granite is the chloritisation of the garnet, probably a result of excessive amounts of water entering the rock. Also sillimanite has not been observed in brown schist obtained from inside the Granite, nor has cordierite been positively identified.

CHAPTER EIGHT

ORIGINS AND METAMORPHISM OF THE NERIA COUNTRY ROCKS

Gneisses

It has been shown that the gneisses of this region are extremely heterogeneous on the small scale, so that a hand specimen can never be representative of the succession as a whole. But on the other hand, when each gneiss succession is considered as a whole it is remarkably homogeneous. The Lower Biotite Gneiss is indistinguishable from the Upper on petrographical grounds, and neither can different parts of the same succession be distinguished. The Hornblende Gneiss can be differentiated from the Biotite Gneisses, but one part is indistinguishable from another.

These characteristics of small scale heterogeneity and large scale homogeneity are typical of sedimentary sequences. The composition and mineralogy of the gneisses suggests that an impure sandstone was the most likely origin.

The absence of any original textures hinders identification of the original sediment, but on the grounds of composition and rock association the source was probably a greywacke. Arkose is also a possibility, but greywacke as defined by Pettijohn (1957) fits the available evidence better. Pettijohn distinguished between arkose and greywacke largely on textural and mineralogical grounds, but he also describes compositional differences. The two rock types are similar in the broad features of their chemistry, but there is an important difference in the Na and K contents. The oxidation state of Fe is also significant, but that cannot be used in rocks that have been subjected to high grade regional metamorphism. Greywackes almost invariably contain more Na than K, whereas it is the other way round in arkoses. Table I gives modal composition and derived chemical compositions of six of the Neria Gneisses. All, including the microcline-bearing gneisses, are richer in Na than K. The present bulk composition of the gneisses including pegmatites is probably the same as that of the original sediment, but to obtain this composition an allowance must be made for the pegmatites. Although an accurate estimate of the volume of pegmatite material cannot be made it is clear that because the pegmatites contain very little alkali feldspar the amount of K contained in them will not significantly change the Na/K ratio of the gneisses as a whole. For this reason the Na/K ratio obtained from the gneiss analyses is probably meaningful.

The suite of metamorphosed rocks occurring in the Neria Area is one that would be expected in association with greywackes. The thick pile of sandstones with numerous concordant basic igneous rocks, either intrusive or extrusive, the occasional shaly horizon, and very few thin impure calcareous bands fits well into a eugeosynclinal environment where greywackes are much more likely to be found than arkoses.

Brown Schists

The same restrictions apply to the brown schists as to the gneisses; only a few of the usual criteria are available for identifying the original rock. There is no reason to believe the schist is a metamorphosed igneous rock; the small scale heterogeneity and large scale homogeneity seen in the gneisses are also found in the schist. The greater abundance of mica indicates that the rock is richer in Al and Fe than the gneisses and the general impression in that the rock was originally an argillaceous sediment.

It is impossible to give an exact name to the original sediment because to do that it is necessary to know the grain size of it. However, it is clear from the presence of zircons that the sediment was not a pure argillite, but had sandy layers in it. Zircons in sediments are associated with the sand fraction

(Pettijohn, 1957) and pure shales contain virtually no zircons at all (Poldevaart, 1955). The micrometric analyses of four schist specimens, table IX, reveal how the grain size varies since the silica/alumina ratio is related to the grain size (Pettijohn, 1957, p.343). The ratio varies from 5:1 to 2:1 reflecting the interlayered coarse grained sandy layers and fine grained shaly layers.

The very low Na content (see table IX) of the brown schists is an unusual feature. Less than 1% Na_2O is far lower than the percentages quoted in Pettijohn for what are thought to be similar rocks, see tables 61 and 62. The alumina/soda ratio is used as an index of the maturity of sediments and in the brown schists values of up to 50 are found. This figure is so large that it is probably meaningless, but it is probable that the sediment was very mature. This is in marked contrast to the gneisses, where it is between three and four, thus indicating a very low degree of maturity.

The change from immature sandstone (gneiss) to mature shale (schist) could be a result of a change in the rate of erosion of the source rocks, the argillite being deposited during a period of slow erosion which followed the wearing down of the surrounding land during the deposition of the sandstones. The change back to coarse grained sedimentation, i.e. the Upper Biotite Gneiss, would be caused by uplift of the land and subsequent rapid erosion. The amphibolite over-lying the schist may be the result of limited vulcanity associated with the uplift.

Amphibolites and ultra-basics

The origin of the amphibolites has already been briefly discussed in Chapter Four and there it was concluded on compositional grounds that they were ultimately of igneous origin. The very constant field relationships between the amphibolites and ultra-basics indicate that they are genetically related and that the ultra-basics also were originally igneous rocks, and not, as Sørensen (1953 & 1954) has suggested, metasomatised sediments. The association of eugeosynclinal sediments is one in which basic and ultra-basic igneous rocks would be expected to occur, e.g. the Franciscan of California (Fyfe, 1967).

Some of the amphibolites may have passed through a sedimentary stage in their development. The laminated amphibolite of the Inland Area and the coast amphibolite described above (pp.35 & 45) are thought to have been sorted by the sea. If the lamination were the result of tectonic activity then it should have been developed generally in the amphibolites, and the fact that it is not suggests that these laminated amphibolites had a different history from the unlaminated in the pre-plutonic stage. A source for such rocks could be ash directly from a volcano, or a sediment from erosion of an igneous mass on land (Kalsbeek, 1965). Walton (1966) has described relic pillows from an amphibolite in the Frederikshåb area, but no such structures have been seen in the amphibolites described here.

The comparatively simply mineralogy of the Neria ultra-basic rocks is quite different from that of the Scourian ultra-basics described by O'Hara (1961). The explanation proposed by Watson (Sutton & Watson, 1951) is more likely to fit the Neria rocks than O'Hara's. The fact that ultra-basic lenses are usually concentrated along one of the margins of amphibolites in the Neria Area suggests that they were formed by differentiation from a basic magma, rather than the amphibolites were formed by reaction between ultra-basic magma and hot country rock. The present lenticular form of the ultra-basic masses is the result of later tectonism. Basic intrusions as thick as many of the present amphibolite horizons could not give rise to ultra-basic differentiates but the present thickness of the amphibolites is no guide to their original thickness. Watterson (1965) has shown how extreme the flattening may be in some cases.

It is thought unlikely that the ultra-basic rocks of the Neria Area are of the Alpine type and were introduced separately from the basic igneous rocks (the present amphibolites). Typical Alpine peridotites are indeed found in environments similar to the pre-Ketilidian basement of S.W. Greenland (e.g. Turner and Verhoogen, 1960). In addition to the invariably close relationship between basic and ultra-basic rocks in the Neria Area there is also stratigraphical evidence to be considered. Ultra-basic rocks are not found below the Upper Biotite Gneiss, and are best developed in the Hornblende Gneiss succession. If the ultra-basic rocks were derived from depth by tectonic activity they should be found at all levels of the area.

Metamorphism

It is clear from the petrographic descriptions that the rocks occurring in the two areas attained their present form under different external conditions. This is shown by the fact that rocks of the same chemical compositions have different mineralogical compositions. It is probable that there is a gradation between the two areas, but as they are not adjacent this has not been seen. The intervening country has been mapped by other geologists and it is expected that their work will reveal a gradual change, both from east to west and from north to south.

It is not possible to determine the differences that exist between the two areas with complete certainty because the rocks that are necessary to enable this to be done are not found in both areas. Acid gneiss (biotite- or hornblende-bearing) is the most abundant rock and occurs everywhere, but it gives very little information about the conditions under which it attained equilibrium. All that this rock reveals is that it crystallised under temperatures greater than those encountered in the greenschist facies. This is determined from the composition of the plagioclase, which is too calcic to have formed under greenschist facies conditions (Turner & Verhoogan, 1960, p.533). The ultimate temperature and pressure attained cannot be determined; nor have variations of temperature and pressure within the wide limits of the amphibolite facies left any trace.

Ideally both basic and pelitic (or semi-pelitic) rocks should be present in the same place, but in the Neria Area semi-pelitic rocks are confined to the north-eastern parts. They are found in the Inland Area and to the north of it, but not to the west and to the east the Inlandice conceals the rocks. The horizon of semi-pelites can be traced through the Neria Granite to the south for some distance, but it is eventually lost under the ice.

Rocks of basic composition i.e. amphibolites, are found everywhere in the Neria Area and thus comparisons between different parts of the Area must be based on information obtainable from them. Although basic rocks are considerably more sensitive to variations in temperature than are acid gneisses, they are less sensitive than pelites, particularly in the upper part of the amphibolite facies. Thus when only amphibolites are available it is impossible to say to which subfacies of the amphibolite facies they belong.

The data available for determining the conditions of metamorphism are basically:-

i) The presence of sillimanite in the brown schist,
ii) The presence of cordierite in the brown schist,
iii) The presence of garnet in the brown schist,
iv) The absence of orthopyroxene in basic rocks,
v) The distribution of garnet in basic rocks.

From i) the temperature of metamorphism must have been at least 500°C (Newton, 1966). From iv) the temperature must have been less than 700°C (Winkler, 1967). From ii) and iii) but especially from ii), the pressure must have been between 3.5 and 5.5 Kbars if the temperature limits are 500°C to 700°C (Schreyer and Yoder, 1964). Because the brown schists are deficient in K neither muscovite nor orthoclase were able to develop, and thus it is impossible to determine whether the rocks belong to the muscovite or orthoclase subfacies of the amphibolite facies. Figs. 53 and 54 taken from Winkler (1967) and Chinner (1962) respectively show the stability relationships thought to apply to the brown schists. Fig. 53 also has the inland amphibolites plotted on it, and from this it appears that theoretically a large proportion of them should contain diopside. The fact that salite is rarely present is further evidence that something other than bulk rock composition, pressure and temperature had a large influence on the mineralogy of these rocks.

In the Coast Area all that can be determined with certainty is that the temperature was not more than 700°C because again the basic rocks do not contain more than one pyroxene. Those rocks that contain orthopyroxene either alone or with clinopyroxene, are not of basic composition, but are richer in Mg or are ultra-basic. According to Shido (1958) amphibolites which contain garnet were formed under higher pressure than those that do not, but she does not give a pressure limit below which garnet will not form. Hietanen (1956) records garnetiferous amphibolites and corierite-bearing schists from the same district of Idaho. Therefore, the presence of garnet in amphibolite does not mean the

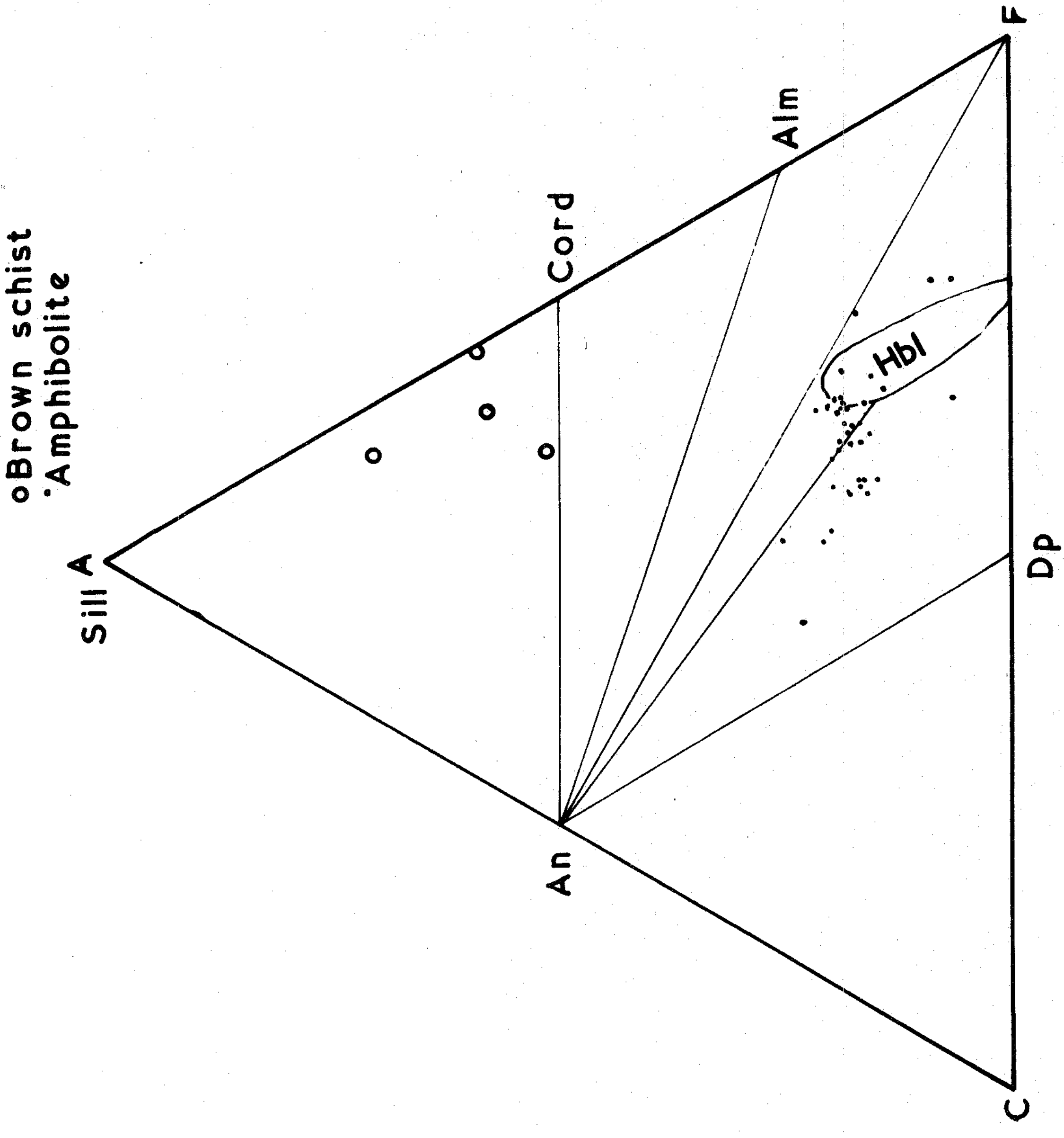


Fig. 53

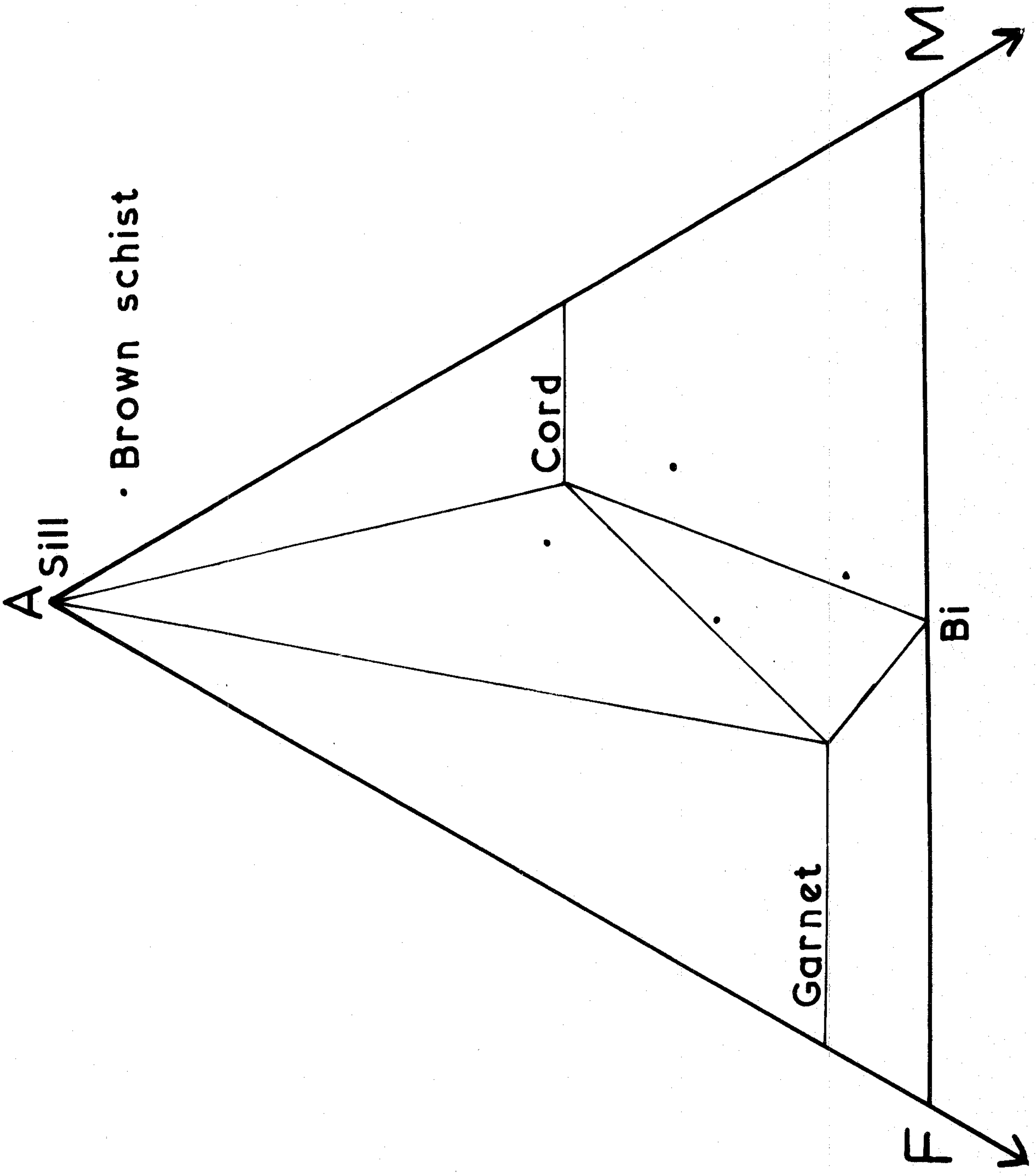


Fig. 54

upper limit of the stability field of cordierite has been exceeded. Extrapolating this to the Neria Area it indicates that the confining pressure in the Coast Area was not necessarily greater than that in the Inland Area. Although it cannot be proved that the confining pressure in the two areas was the same, it is more likely that it was than was not because there is a relatively short distance separating the two areas. In Japan high and low pressure metamorphic belts lie side by side and less than 50 km. separate areas of high and low pressure metamorphism (Miyashiro, 1961), but it is unlikely that this is the case in the Neria Area because that Area is part of the Canadian Shield and is not situated on the margin of a continent.

From information provided by other geologists working in the Neria Area it appears that the amphibolites become garnetiferous to the north of the Neria Granite as well as to the west. To the south the geology is made more complex by the appearance of younger supracrustals and the distribution of garnets there is not known. But at least to the north and west garnetiferous amphibolites appear to be concentrically arranged around the Granite. It is suggested, therefore, that water became available in the area where the Granite is now found and that as this spread out through the gneisses and amphibolites the anhydrous hornblende - salite-almandine-plagioclase assemblage was replaced by the hydrous hornblende-plagioclase assemblage. Comparison of mineral compositions shows that combinations of almandine and salite plus water are roughly equivalent to tschermakite, with probably changes in the quartz and plagioclase contents of the rocks.

Metamorphic facies series

The concept of metamorphic facies series introduced by Miyashiro in 1961 has been gradually refined until at present his original five facies series have been increased to eight (Hietanen, 1967). All the different facies series pass through the same range of temperature but they form under different pressure conditions. It is the pressure, and thus fundamentally the geothermal gradient, that defines each series. Each of the eight series recognised at present is but an example of the parageneses that are to be expected, and there is probably a continuous gradation from the very low pressure contact metamorphism to the extremely high pressure Alpine metamorphism.

To determine the type of metamorphism operative in any area it is necessary to be able to see a gradation from the low grade to the high. If this can be done it is possible to trace the mineralogical changes that occurred in the different rock types as the metamorphism progressed. But the pre-Ketilidian basement rocks of S.W. Greenland have all been metamorphosed to the same grade and thus the progressive development of the metamorphism cannot be traced. The situation is complicated by the fact that the rocks must have been held at high grade for some time as nearly all the relic minerals from the lower grades were destroyed. The only exception to this is the presence of some pre-hornblende actinolite in certain amphibolites, but this only confirms that the rocks were not metamorphosed under high pressure conditions. The brown schists do not contain any relics such as staurolite, kyanite, or andalusite which would have been much more helpful in revealing the path by which the rocks reached their present mineralogical compositions. By comparison with other metamorphosed terrains it is possible to come to some conclusions about the type of metamorphism that was operative in the Neria Area. Reference to fig. 1 of Hietanen (1967) shows immediately that any one of four types of metamorphism - Japan, Buchan, Pyreneean, or Idahoan - could have led to the distinctive assemblage of the Neria rocks, viz. the biotite-quartz-plagioclase-sillimanite-cordierite-almandine assemblage of the brown schists.

Amphibolites from the Central Abukuma Plateau (Japan type of metamorphism of Hietanen) do not contain garnet (Miyashiro, 1958; Shido, 1958). Shido and Miyashiro decided that the absence of garnet is the result of low pressure metamorphism and not "wet" metamorphism. Therefore it is probable that metamorphism in the Neria Area took place at higher pressure than that which prevailed in the Abukuma Plateau. Basic igneous rocks were not involved in the metamorphism of the Bosost district, Central Pyrenees (Zwart, 1962) and therefore direct comparison with Bosost cannot be made. But from the Merrimac Area, California, Hietanen (1951) has described rocks which she states belong to the Pyreneean type of metamorphism (Hietanen, 1967), and in these rocks garnet is not found in the amphibolites. Nor is garnet present in the metabasites of the type area of Buchan type metamorphism (Read, 1952). Therefore there is no evidence that pressure was lower in the Neria Area than in the Ythan Valley or Bosost. The brown schists of the Neria Area have a higher proportion of garnet than do the schists of the Bosost region (Zwart, 1962). This may be an indication that pressure was higher in the Neria Area, but garnet is abundant in the Abukuma Plateau schists where pressure is believed to have been lower than at Bosost. Therefore the abundance of garnet in schist cannot be taken as a guide to the pressure.

It is most probable that the Idahoan type of metamorphism prevailed in the Neria Area since it is clear that Barrovian type metamorphism took place at higher pressure than occurred in the Neria Area. This is proved by the absence of cordierite from the schist of the South-East Highlands of Scotland (Barrow, 1893). Read (1927) has described rocks of Barrovian type from Deeside and he, unlike Barrow, described the petrography of the metabasites. It appears that garnet is almost entirely absent; one specimen is recorded as having skeletal garnet euhedra. This part of the S.E. Highlands is intensely granitised (Barrow, 1893; Read, 1927), as is the eastern part of the Neria Area, and it is thus probable that the metamorphism was "wet" and that even under the high pressure conditions that prevailed there anhydrous garnet was unstable in rocks of basic composition. It appears, therefore, that the absence or presence of garnet in a metabasite does not prove that the rock was formed under low or high pressure, but is more probably an indication that the metamorphism was "wet" or "dry".

Parageneses of the Metamorphic rocks

Gneisses

There is very little to be said about these because they are so insensitive to changes in temperature and pressure. The only clear feature of these rocks is that quartz recrystallised at a very late stage during localised mylonitisation and that plagioclase and K-feldspar grew porphyroblastically on more than one occasion.

Amphibolites

The amphibolites also have a simple mineralogy and there is only little evidence to suggest that there were several periods of metamorphism in which minerals were produced. In the Inland Area the absence of garnet and the presence of relic clinopyroxene has already been noted and ascribed to the influence of water from the Neria Granite. The formation of the granite was a late event and therefore the destruction of garnet and salite must also be late, i.e. after the three recognisable periods of folding were finished. This change in mineralogy was probably not accompanied by a general recrystallisation of the amphibolites. If it had been, then the fine scale lamination in the laminated amphibolite would probably have been destroyed. The tremolite-actinolite crystals found occasionally in the amphibolites of the Inland Area are usually relics from a lower grade of metamorphism through which the rocks passed in order to reach the amphibolite facies, but in one particular amphibolite tremolite-actinolite was formed retrogressively after the hornblende (fig.25). The retrograde actinolite occurs only where hornblende and older actinolite crystals are in close contact and this accounts for the absence of retrograde actinolite elsewhere because actinolite crystals found in other amphibolites are isolated and surrounded by rims of alteration products.

Brown Schists

These rocks are more complex mineralogically than amphibolites or gneisses and more conclusions can be drawn concerning their development than can be drawn for any other rocks. Biotite is the oldest metamorphic mineral present (the few small zircon crystals are much older, but they are relics from the sedimentary stage and have not been recrystallised in the metamorphism) and was developed during the first period of folding when the foliation was imposed on the rocks. Biotite did not develop to any great extent in the second period of folding. Fig.55 shows a second phase fold in the brown schist with a strong axial plane foliation. In thin section the foliation is seen to be composed of

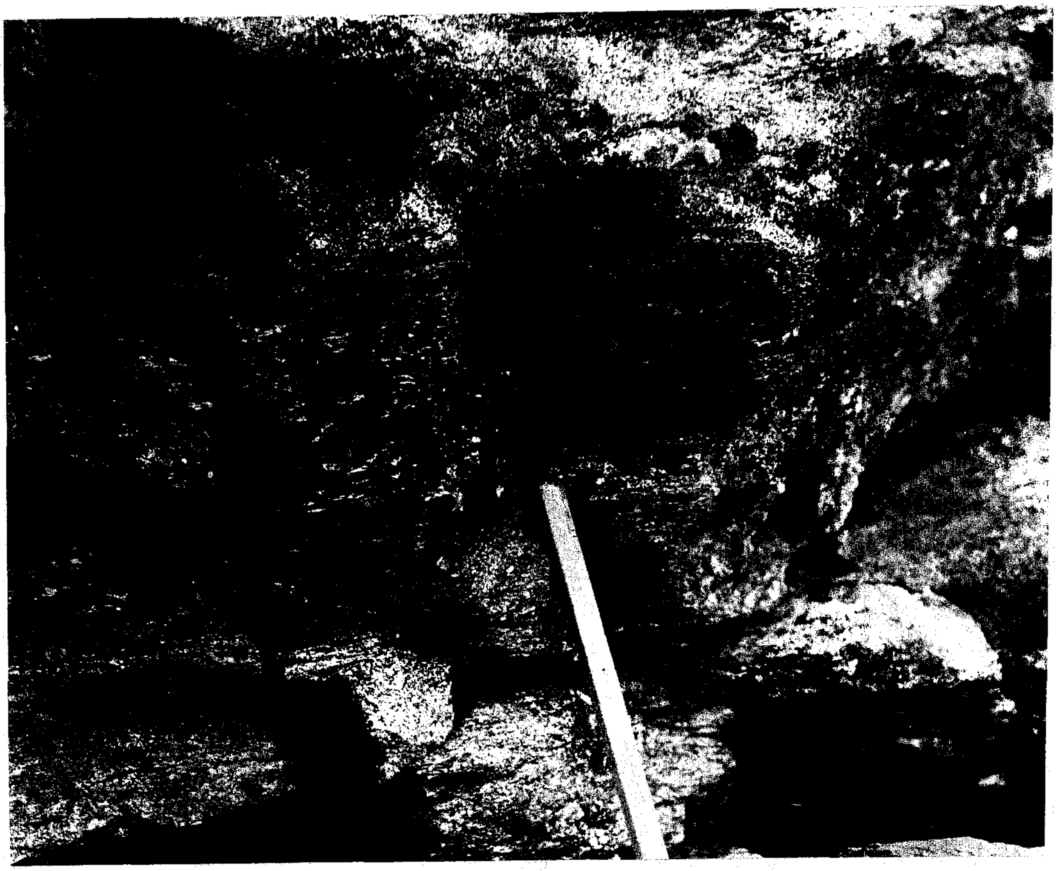


Fig. 55 Garnet crystals grown across the axial plane foliation of a second period small fold. Grid square E6, Map I.

small crystals of quartz and plagioclase. Scattered throughout the rock are a few plates of biotite whose orientation can be at any angle to the new foliation. There is no sillimanite or cordierite in this rock so no conclusions can be drawn concerning the relationship between them and the folding. Garnet was formed later than the second phase of folding because well preserved garnet crystals cut the axial plane foliation of the fold described above (fig.55). The presence of garnet in the neosome of the migmatised brown schist also indicates that the formation of garnet occurred after the second folding. The garnet could not have been formed before the migmatisation because in view of the ease with which it can be chloritised it would not survive the transformation of brown schist into a quartz-feldspar rock.

The time of formation of cordierite and sillimanite cannot be determined precisely, but it probably was between the first and second periods of folding. These two minerals are younger than biotite and older than garnet and they are also older than the migmatisation as they are never found in the neosome of the migmatitic brown schist. The formation of cordierite is favoured by conditions free of stress (Harker, 1939) and therefore it is probable that it was formed after the deformation that produced the first phase folds was finished. Sillimanite is a high temperature mineral and therefore it would have formed late in the metamorphic process, when the highest temperatures were reached, shortly after cordierite.

Table XII (p.243) summarises the mineral parageneses and relates them to the main structural and plutonic events in the Neria Area.

CHAPTER NINE

DYKES

There are many dykes cutting the gneisses and they range in thickness from tens of centimetres to over 100m. They can be divided into three main groups: the pre-Gardar (Kuanitic), Gardar, and post-Gardar.

Field Relations

Kuanitic dykes

The Kuanitic dykes were formed over a longer period than the others and while most of them are probably pre-Ketilidian, some of those to the south of Sermiligarssuk Fjord are almost certainly post-Ketilidian. The age relations of these dykes to the gneisses on the regional scale are discussed in Bridgwater (1965,pp20-22).

The Kuanitic dykes are dark green or black in colour, are frequently weathered to a hard smooth surface and are occasionally sheared. The shearing is often restricted to the margins, but some are strongly sheared throughout. There is no uniformity to the shearing, the dyke may be sheared for the whole of its length or only locally sheared. Shearing is often particularly intense where two dykes cut, and in such circumstances the age relationships cannot be determined. Many of the dykes are porphyritic. The phenocrysts are plagioclase and vary in size from little more than the groundmass size to large nuggets several centimetres across. The most common arrangement is for the phenocrysts to be concentrated in a band along the centre of the dyke and the margins to be non-porphyritic (fig.56). The distribution of phenocrysts is not always uniform along the dykes. The dyke seen in fig. 56 for example, is non-porphyritic some two kilometres along the strike to the west.

The different generations of Kuanitic dykes found in the Inland Area are:-

Youngest:

NE	Two 10 m. thick dykes
SE-ESE	Several dykes, up to 25-30m thick.
NE	Main generation of dykes, up to 50m. thick, and often make en echelon breaks.

Oldest:

N(approx.)	Two or three pale green dykes less than 10 m. thick, and strongly epidotised.

On the Coastal Area the generations are different:-

Youngest:

E	One thick dyke more than 50 m. thick.

Two generations

NE	One generation strikes 060 and the other 040 The 040 dykes are younger than the 060.

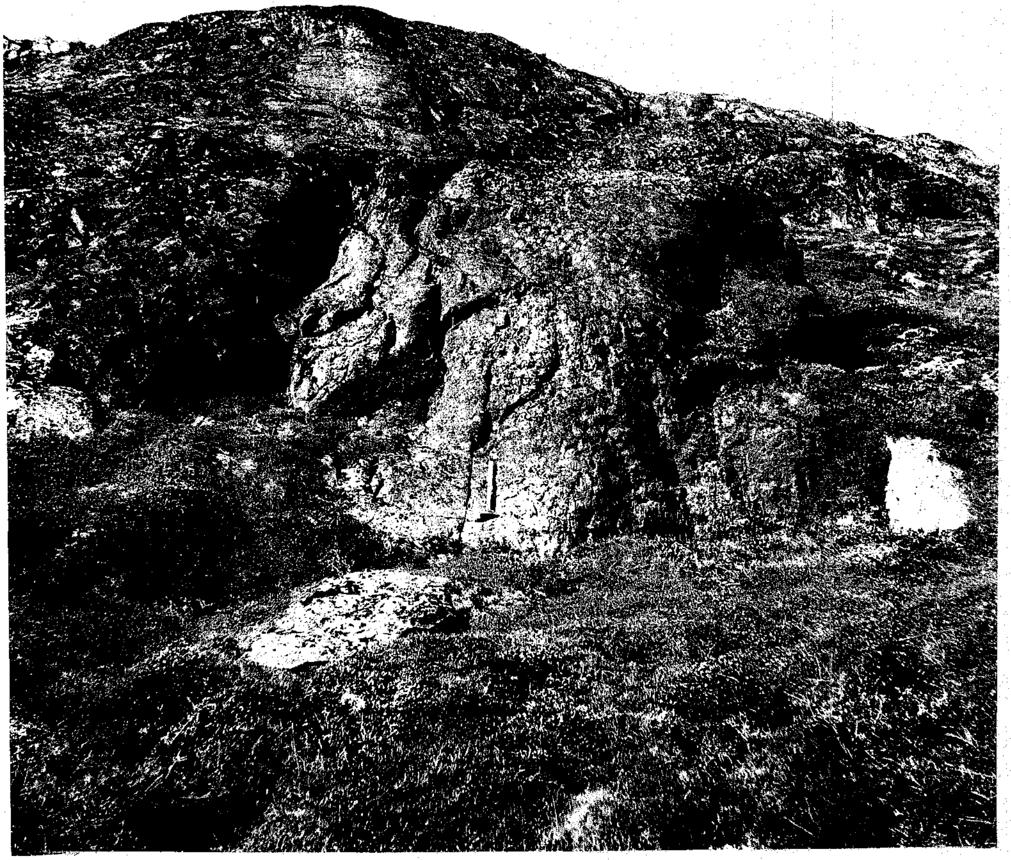


Fig.56 Porphyritic pre-Gardar dyke. The main concentration of phenocrysts can be seen along the centre of the dyke to the right of the hammer. Grid square 5E, Map II.

Oldest:

E Several dykes approximately 5 m. thick. These are parallel to the youngest dyke.

Gardar dykes

These are much less abundant than the Kuanitic dykes and counted in ones and twos rather than in scores. They have a distinctive red brown colour and usually have dolerite sandy weathering. This, combined with the fact that they are unmetamorphosed dolerites makes them easily recognisable in the Inland Area. On the coast, however, it is more difficult because most of the dykes there are unaltered and a number of the Kuanitic dykes have the sandy weathering. The behaviour of the dykes at faults taken in conjunction with their strike willusually show whether the dyke is Gardar or not. Gardar dykes usually strike NE or SE, and a dyke with one of these directions that crosses faults without being displaced is probably Gardar. There is only one Gardar dyke on the Coastal Area and that strikes north-east.

Inland there are four large (60-70m. thick) Gardar dykes striking NE and several smaller (ca.10m. thick) ones parallel to this direction and at right angles to it. Where intersections can be seen the NE group is younger than the SE. Porphyritic and non-porphyritic Gardar dykes are found.

In the Inland Area a number of small dykes (up to 5 m thick) striking between 120° and 160° are found. They are a distinctive grey-brown colour and always have porphyritic centres and non-porphyritic margins (fig.57). They cut Kuanitic dykes and are cut themselves by the NE-striking Gardar dykes, but it is not known whether they are early Gardar dykes or late Kuanitic dykes. In view of the fact that these grey-brown dykes are mainly fresh dolerites it is probably more likely that they are early Gardar dykes.

Post-Gardar Dykes

T.D. (Trap Diabase) is the name given to a group of unmetamorphosed dolerite dykes striking almost N-S along the coast. They are not found inland at all, but on the coast may extend for many tens of kilometres. Four of them cut across the Coast Area, but they are not continuous. They can be 30-40 m. thick, but there are often breaks of several kilometres. Sometimes they disappear abruptly and sometimes they gradually thin to nothing over some distance. They are very similar to Gardar Dykes in appearance, but are slightly redder in colour and there is often a distinct vegetation change over them, (The grass is literally greener on the sandy soil produced by the weathering of these dykes).

Fig. 57 Grey brown porphyritic dyke. Grid square C6 Map I. (Figs. 56 and 57 show clearly the contrast in amount of exposure between the Coast and Inland Areas).

These dykes are younger than all the others and are also younger than the faulting. There are T.Ds. displaced by faults but these occurrences are extremely rare. Where they cut faults, however, there is often a slight lateral displacement of the dyke, but this is caused during the intrusion of the dyke and is not faulting.

The age of these dykes is not known with certainty. They were ascribed to the Tertiary at first, but after dating of one from the Frederikshab area they are now provisionally put in the Mesozoic (O.Larsen, 1966).

Dyke Petrography

Kuanitic dykes

Petrographically these dykes are predominantly olivine dolerites. Usually the plagioclases occur as well preserved laths and ophitic texture is common. On the coast some dykes have been subjected to a weak autometamorphism or local mylonitisation. This was not a regional metamorphism because younger dykes may be altered while older dykes are still fresh dolerites. The form the alteration takes is a uralitisation of the pyroxenes, but the igneous texture is often preserved. Inland, however, a weak regional metamorphism has affected the Kuanitic dykes. This is shown by all the pyroxenes being altered to an unidentifiable fibrous amphibole. These dykes are best described as metadolerites as the igneous textures are usually preserved. A few dykes have been more strongly altered and are now true amphibolites with no relic igneous textures.

Gardar dykes

These dykes also are olivine dolerites and show no evidence of having been metamorphosed. The augite frequently has a brownish tint and is therefore slightly titaniferous. Accessory minerals include apatite, iron oxides, and occasionally biotite. In some dykes the apatite and iron ore may be very abundant.

The grey-brown porphyritic dykes show signs of having been slightly metamorphosed. The pyroxenes show incipient uralitisation, but they are much less altered than the pyroxenes in the Kuanitic dykes from the Inland Area. Therefore they are later than the slight regional metamorphism which affected the Kuanitic dykes.

Post-Gardar dykes

The youngest dykes are again olivine dolerites and show only a very weak and locally developed autometamorphism. Petrographically they are indistinguishable from the Gardar dykes.

Discordant Amphibolites

All the discordant amphibolites found in the Neria Area are thin sheets generally not more than 1 m. thick. They may have any dip from horizontal to vertical and cannot be traced for more than a few metres along their strike. There are undoubtedly many more than have so far been recognised because any which happen to be concordant with the gneiss structures cannot be distinguished from amphibolite bands which were part of the original succession. The Kuanitic dykes which have been amphibolitised are not classed as discordant amphibolites because they have not been subjected to plutonic conditions in which complete recrystallisation occurred, whereas the true discordant amphibolites have been completely recrystallised and are no longer recognisable as igneous rocks on petrographic grounds.

Few discordant amphibolites have been found in the Neria Area as a whole and in the areas described here they are most common along the southern and eastern shores of the Coast Area. Elsewhere in the Coast Area none has been found and this is almost certainly because of the poor exposure. The two that have been found in the Inland Area indicate that they are much less common there than they are on the coast; exposure is so good inland that they would have been seen if present. The distribution of discordant amphibolites is related to the stratigraphy because the only good example seen in the Inland Area occurs in the Hornblende Gneiss Succession in the north-west part of the Area, and of course all those found on the coast also occur in the Hornblende Gneiss Succession.

Fig. 58 shows the Inland Area amphibolite viewed along the strike in a vertical plane. This amphibolite has been subjected to very strong shearing which has produced the folded shape and the sheared out upper limb. It is only because the hinge has been preserved that the body is now recognisable as a discordant amphibolite. If only the upper limb or the outcrop on a horizontal surface could be seen then the body could not be recognised for what it is. From just one occurrence of a discordant amphibolite it is impossible to say when it was formed and it may have been folded during any of the periods of deformation that affected these rocks.

The discordant amphibolites found on the coast are all less deformed than the body described above, and some are still straight, parallel-sided dykes that have been recrystallised without being deformed. No discordant amphibolite has been seen to be cut by pegmatites, but at several localities an amphibolite has been seen to cut

Fig.58 Folded discordant amphibolite cutting hornblende gneiss. Grid square B 8 Map I.

pegmatites (fig. 59). No intersections have been observed between discordant amphibolites and there is no evidence of an event such as metamorphism, deformation, or migmatisation affecting some and not others. The inland folded amphibolite is possibly older than the coastal ones, but the evidence is not strong enough for such a conclusion to be made definite.

Hornblende and plagioclase are the essential minerals of the discordant amphibolites, while quartz may be found in small amounts. Biotite is sometimes present in amounts almost equal to the hornblende. Foliations or lineations are rarely seen because platy minerals are not abundant and the hornblende is usually chunky, rather than prismatic, in habit. In those discordant amphibolites in which there is a foliation it is developed parallel to that in the surrounding gneiss. The gneiss foliation and other structures do not curve towards the strike of the discordant amphibolites at the point of contact, but are cut off abruptly. This is evidence that the amphibolites were not syntectonic because when dykes are intruded into rocks under shear it appears that the gneiss foliation next to the dyke and the internal foliation of the dyke become sigmoidal (Watterson, 1965).

Fig. 59 Discordant amphibolite cutting second period folds with axial plane pegmatites. Grid square 5F, Map II

CHAPTER TEN

STRUCTURE

Introduction

In the elucidation of the structure of complexly deformed rocks such as the pre-Ketilides of S.W. Greenland, the techniques of structural analysis described by Turner and Weiss (1963) are used extensively. The form of the overall large scale structure can be largely found by normal lithological mapping. But at best this will show only the geometrical form of the rocks at present. It does not give much information about the structural development of the rocks. In order to determine the sequence of events which led up to the rocks having their present form it is necessary to use small scale features and the relations between them and the large structures. Structures visible on the microscopic, mesoscopic and macroscopic scales, as defined by Turner and Weiss (1963, pp.15 and 16) are all used. The microscopic work has been confined to descriptions of textures and no fabric measurements have been made.

The structural elements used in the analysis are penetrative planes and lineations. Planar structures are banding, foliation, and axial planes. Linear structures are fold axes and lineations formed by elongation of mineral grains and intersecting planes. Rodding is comparatively rarely found and mullions not at all. Boudins produced by necking of competent layers enclosed in a less brittle rock are abundant but they are usually visible in only two dimensions and appear equi-dimensional. They do not provide another linear feature of the rocks.

Planar Structures

Banding

The dominating planar structure is the regional banding which has been described in the section on gneisses (p. 15) and is defined by differences in the composition of adjacent bands in the rocks. In certain situations it can be seen that there are two generations of banding. Usually, however, it is impossible to tell whether the banding at any one place is the older or the younger, as the two are quite indistinguishable unless they can be seen together with an angular discordance between them.

Foliation

Planar structures younger than the banding are foliations caused by parallel growth of platy minerals. This is best developed in the brown schist where there is a very high proportion of biotite. In the gneisses the mafic-rich bands are strongly foliated for the same reason while the quartzo-feldspathic bands containing only small shredded biotites have very weak foliation. In amphibolites also the foliation is very weak on account of the almost universal absence of platy minerals in that rock. Usually the foliation is parallel to the banding, but in the hinge zones of small folds the axial plane foliation (where one is developed) cuts across the banding. At one locality, the axial plane foliation of one fold cuts across the limbs of an earlier fold (fig.60). In migmatites two non-parallel foliations are occasionally found. In these instances, it is clear that the foliation in the granitoid neosome is younger than that in the more basic palaeosome. Fig. 61 shows a clear example of the discordance between the two foliations.

Linear Structures

Lineations can be seen in many of the rocks but it is not present so generally as foliation. When measuring a lineation it is necessary to be careful

Fig. 60 Axial plane foliation (parallel to pencil) of a younger fold (above pencil) cutting across the hinge of an older fold. Grid square F4, Map I.

Fig. 61 Neosome foliation cutting across palaeosome foliation (parallel to pencil). Grid square F4, Map I.

that it is measured on a foliation surface. If what appears to be a lineation is seen on a surface that is not a foliation surface (or parallel to a foliation) then either it is not a lineation, but merely an apparent lineation formed by the intersection of the surface and the foliation, or, if it is a lineation, then a false plunge and strike will be obtained.

Three types of lineation occur:- a mineral lineation formed by the growth of minerals with their long axes parallel to one particular direction; a structural lineation formed by the intersection of two penetrative S planes; a crenulation or microfolding of a foliation surface. Foliations and banding are penetrative S planes and thus the intersections of two foliations, or a foliation and the banding can produce a lineation. This is particularly seen in small folds where the intersection of the axial plane foliation and the banding around the hinge results in the formation of a lineation parallel to the axis of the fold.

Mineral lineations are not as commonly seen in the amphibolites as would be expected because the hornblende grains are usually equi-dimensional. Thus, although microscopically the hornblendes are clearly aligned parallel to one direction, in hand specimen or outcrop no lineation may be visible. Lineations are in fact more frequently found in the gneisses. They are not uniformly distributed throughout the areas. In some parts lineations may be seen almost everywhere, but elsewhere they may be most uncommon. These lineations in unfolded gneisses are mineral lineations formed by the growth of feldspars and quartz with their long axes parallel to the P min. of the strain ellipsoid in the rocks.

The third type of lineation, crenulations on a foliation surface, form the strongest lineation (see fig.62). These are not as commonly encountered as the other types of lineation, but when they do occur they plunge approximately parallel to the local smallfold axes and are therefore related genetically to them.

Smallfolds

By smallfolds are meant those folds which can be seen in their entirety on an outcrop or in a large hand specimen, and which range in amplitude and wavelength from centimetresto metres. They are revealed by folding of the layers of different composition within the rock, and when the rock is homogeneous no smallfolds can be produced although it may be deformed in exactly the same manner as a heterogeneous rock strongly smallfolded.

Fig.62 Small scale crenulations in hornblende gneiss plunging parallel to the pencil. Grid square E5, Map I.

The most usual style of smallfold is isoclinal, with parallel or subparallel limbs and appreciable thinning of the limbs and thickening of the hinges. The folding occurs within the banding and only those bands actually folded are affected. Not infrequently the limbs are sheared and truncated by a younger banding (fig. 63). These have been called intrafolial folds by Windley (1966). They are probably the earliest folds preserved in this region. Other folds have pegmatites intruded along their axial planes. An example of this occurs on the coast of Sermilik Fjord in square 5E (Map II), (Fig.52).

More open, less isoclinal folds occur in some rocks, (fig. 64) and this style may be related to the mode of formation of the fold or to the mechanical properties of the rocks. The folds seen in fig.64 are open and almost concentric because of viscosity differences within the amphibolites and because of the relative thicknesses of the more and less viscous quartz of the amphibolite (Ramsay 1967, Ch.7). But the concentric open fold seen in fig. 65 has been produced by folding in which there was no, or very little, shearing. This is demonstrated very clearly by the presence of a smaller isoclinal fold, with an axial plane foliation, that has been folded around the later concentric fold.

Fig.63 Intrafolial fold with a new banding partly developed along the lower limb. Grid square B8, Map I.

Fig.64 Open concentric folds developed in massive amphibolite. The less viscous darker material in the lower part of the photograph has developed "pinched" antiforms between the lobate synforms. Grid square B7, Map I.

The large fold in fig. 65 is one of the latest generations of smallfolds as it folds the foliation as well as the banding, instead of having an axial plane foliation as have the older folds.

The axial planes of the smallfolds are usually parallel to the banding at any one place because the folding has been isoclinal on all scales. In those small areas where the folding has been peculiarly active and has resulted in intricate outcrop patterns with many smallfolds the axial planes and axes of the smallfolds often have very irregular distribution. Statistical analysis by means of stereographic plots sometimes aids in elucidation of their orientation, but not in every case.

In some instances the rocks have been very fluid during the deformation and the style of the resulting smallfolds reflects this. Fig. 66 shows such smallfolds and it can be seen that the axial plane are curved thus making it impossible to measure their orientation. However, such folds are exceptional. The different styles of smallfolds cannot be related to individual phases of folding because folds of the same style were formed in the same rocks at different periods. Only the last folds have a distinctive style and that is because there was only weak shearing, if any at all, associated with their formation.

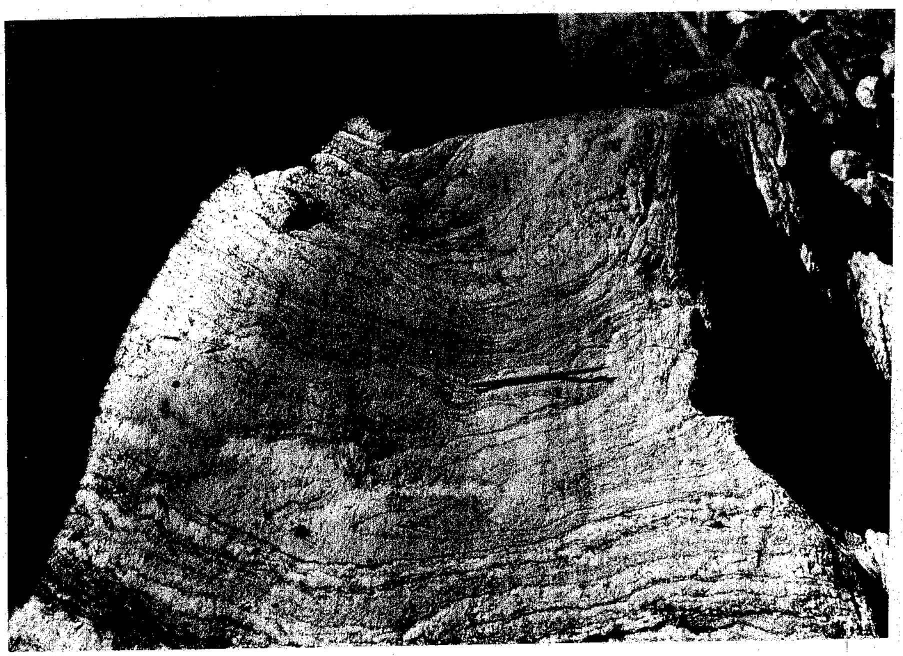


Fig. 65 Open concentric fold in biotite gneiss refolding a tigher older fold. Grid square F4, Map I.

Fig.66 Small fold with curved axial plane. Grid square D4, Map I.

Inland Area

The overall structure of the area is a plunging antiform with subsidiary folds on its flank and nose. The antiform closes in the south-west corner of the area and plunges south-west. To the N.E. the axis flattens and eventually begins to plunge N.E. The flattening can be detected on the Inland Area by means of constructed fold axes (fig.67), but the N.E. plunge appears only on the area to the north. This antiform is only part of much larger and more complicated structures that can be seen only when a larger area than the present 200 sq.kms. is considered. The antiform is revealed by the shape of the outcrop of the brown schist and laminated amphibolite horizon. None of the other amphibolites can be traced round this fold and neither can amphibolites to the north-west be correlated with amphibolites to the south-east of the limbs.

The north-west limb has a uniform north-westerly dip except where it has a slight kink and where it is disrupted by the large east-west fault. The south-east limb has a similar strike to the other but has a more variable dip, although it is never far from vertical. There is no axial plane foliation to this fold and the foliation within the brown schist around the hinges is concordant with the folded rocks. Around the hinges a

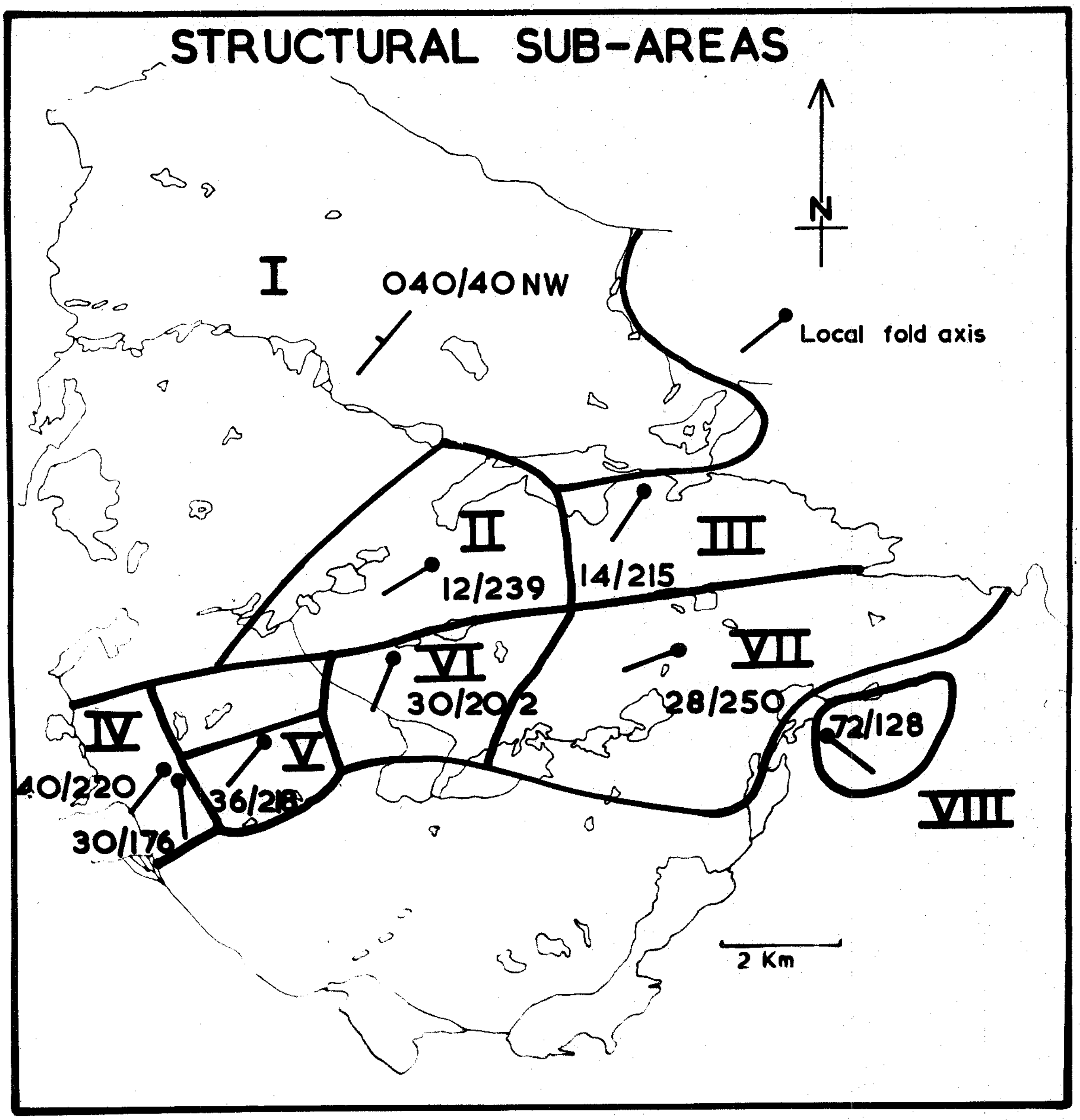
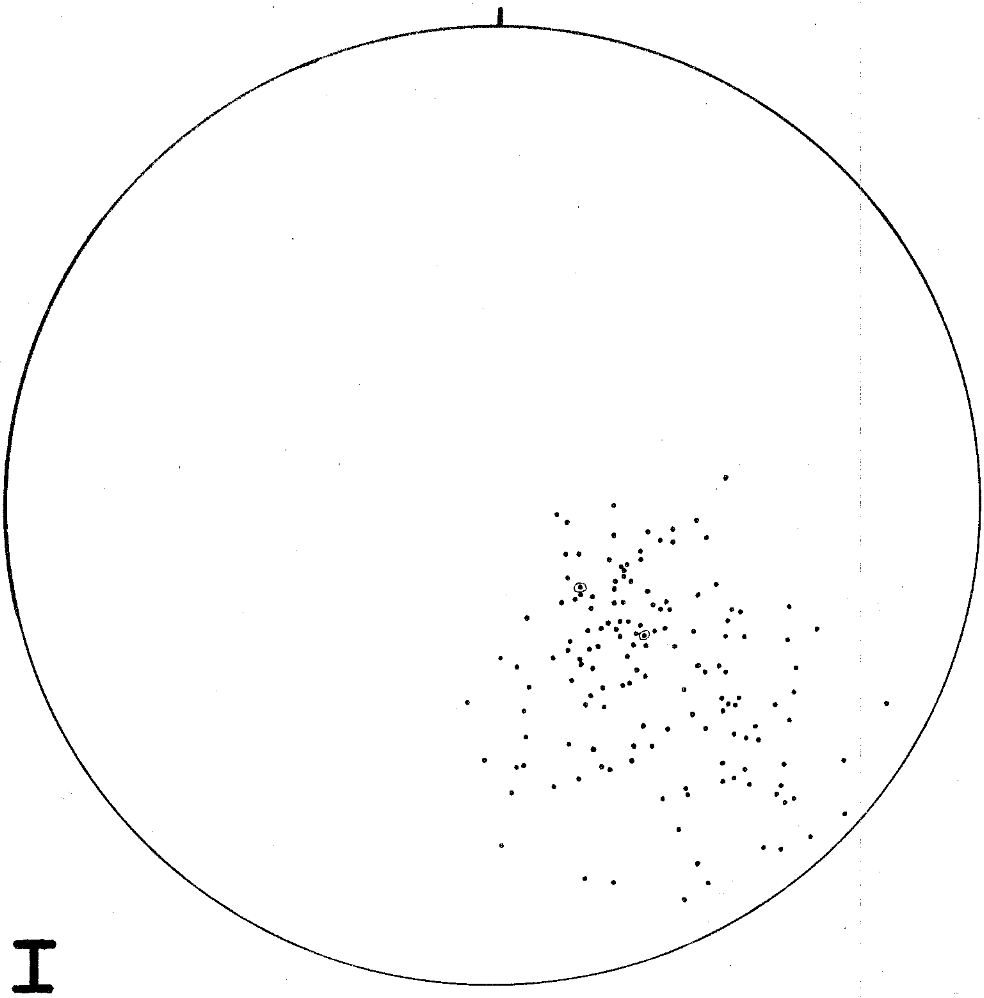


Fig.67 Inland Area

number of smallfolds occur within the schist, all with their axial planes parallel to the foliation. Thus the major fold was formed later than these minor folds.

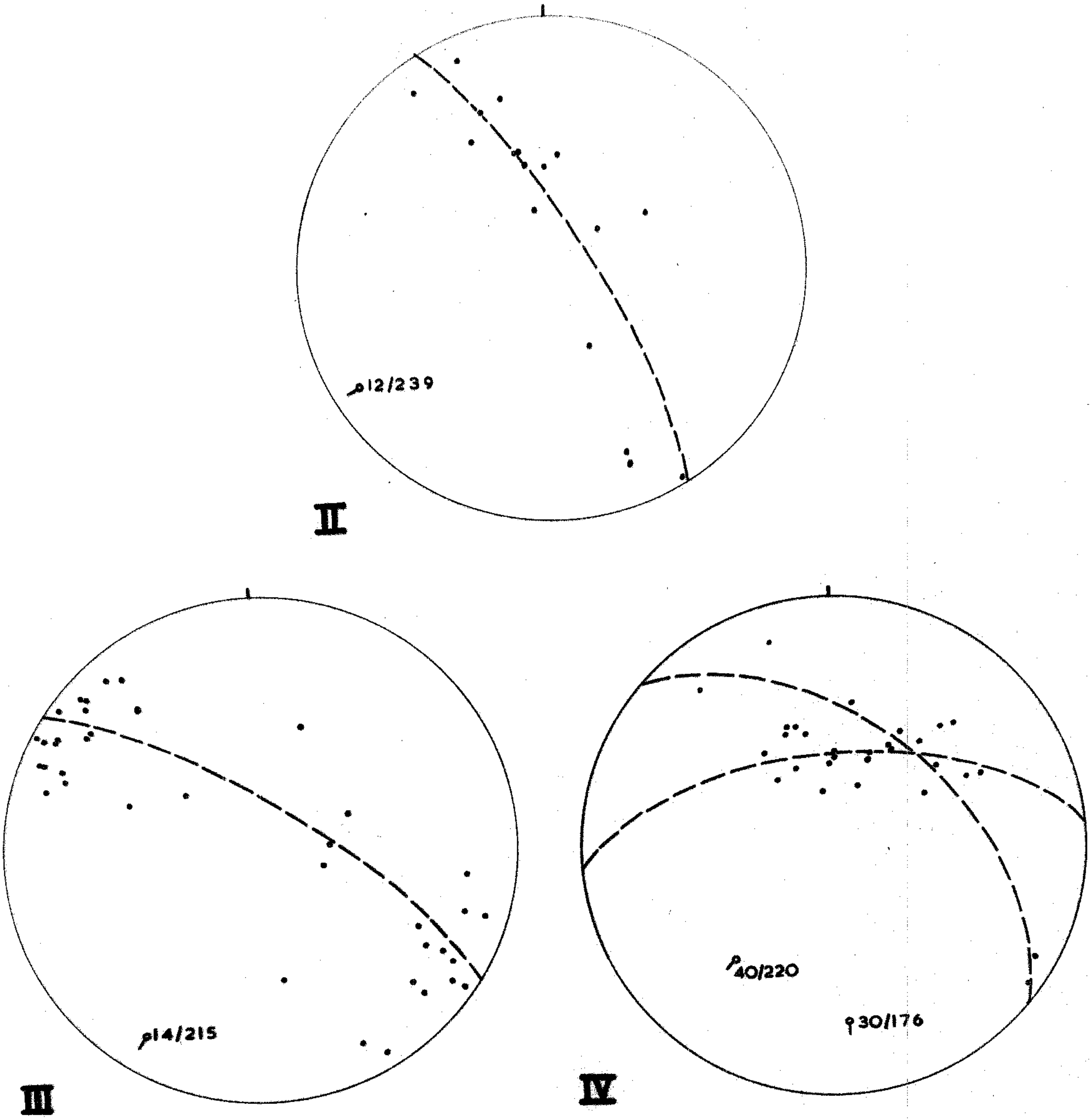
The isolated folded amphibolite in the eastern part of the Area is folded about an axial plane that dips south-east and an axis that plunges steeply to the south-east (sub- area VIII, fig.70). The connection between this amphibolite and the main antiform is not known, but there are reasons for believing that this south-east plunging reclined fold belongs to an older generation of folds than the main antiform.

Following the methods of Turner & Weiss, the Area has been divided into eight sub-areas, each of which is homogeneous with respect to the regional banding. The intricately folded amphibolite horizons occurring in grid squares B6 and 7 have been excluded because they are too complex to be included in an analysis on a large scale. The fact that only eight sub-areas are required indicates that the overall structure is fairly simple. The large area in the south not subdivided is occupied by the Neria Granite. The limits of the sub-areas are shown in fig.67, and the π diagrams for each one in figs. 68, 69 and 70. In sub area I the very uniform N.W. dip gives rise to a strong concentration of points



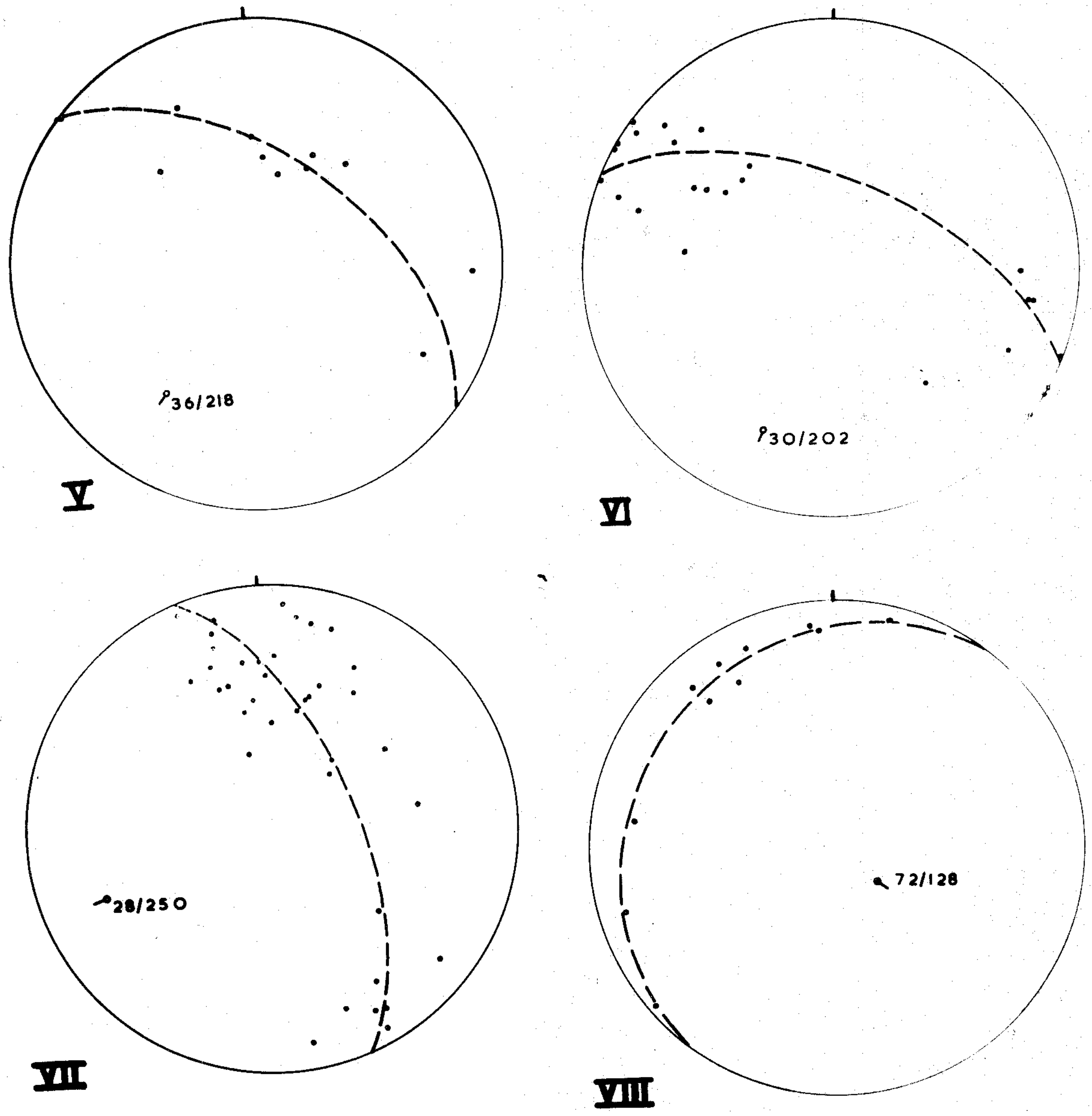
π S DIAGRAM

Fig. 68 Inland sub-area



π S DIAGRAMS

Fig.69 Inland sub-area



π S DIAGRAMS

Fig. 70 Inland sub-areas

and a great circle cannot be constructed. For all the other sub-areas, however, the poles to the foliation planes are dispersed and a great circle can usually be constructed. In sub-area IV two great circles can be drawn, showing that there folding was more complicated. The two points in the south-east part of the stereogram are important because they are derived from the unfolded north-west limb of the antiform in sub-area IV. They define that part of the fold and therefore any great circle giving the axis of the fold must include them. The spread of points from the centre and eastern parts of the fold lie on a great circle giving a south-easterly plunging axis. South-westerly plunging axes dominate throughout the area as a whole and only sub-area VIII has a different main axis. The plunge of the axes in the other areas become progressively less to the north-east.

The foliation planes used in sub-area VIII were derived solely from the amphibolite horizon that occurs there and it is thought that this axis is older than the south-west plunging axes of the other areas. From the western end of the migmatitic amphibolite in sub-area VII, measurements made of the internal foliation of the amphibolite blocks and the foliation of the enclosing gneiss neosome indicates two different fold axes. Fig.71 shows the great circles obtained. The amphibolite is

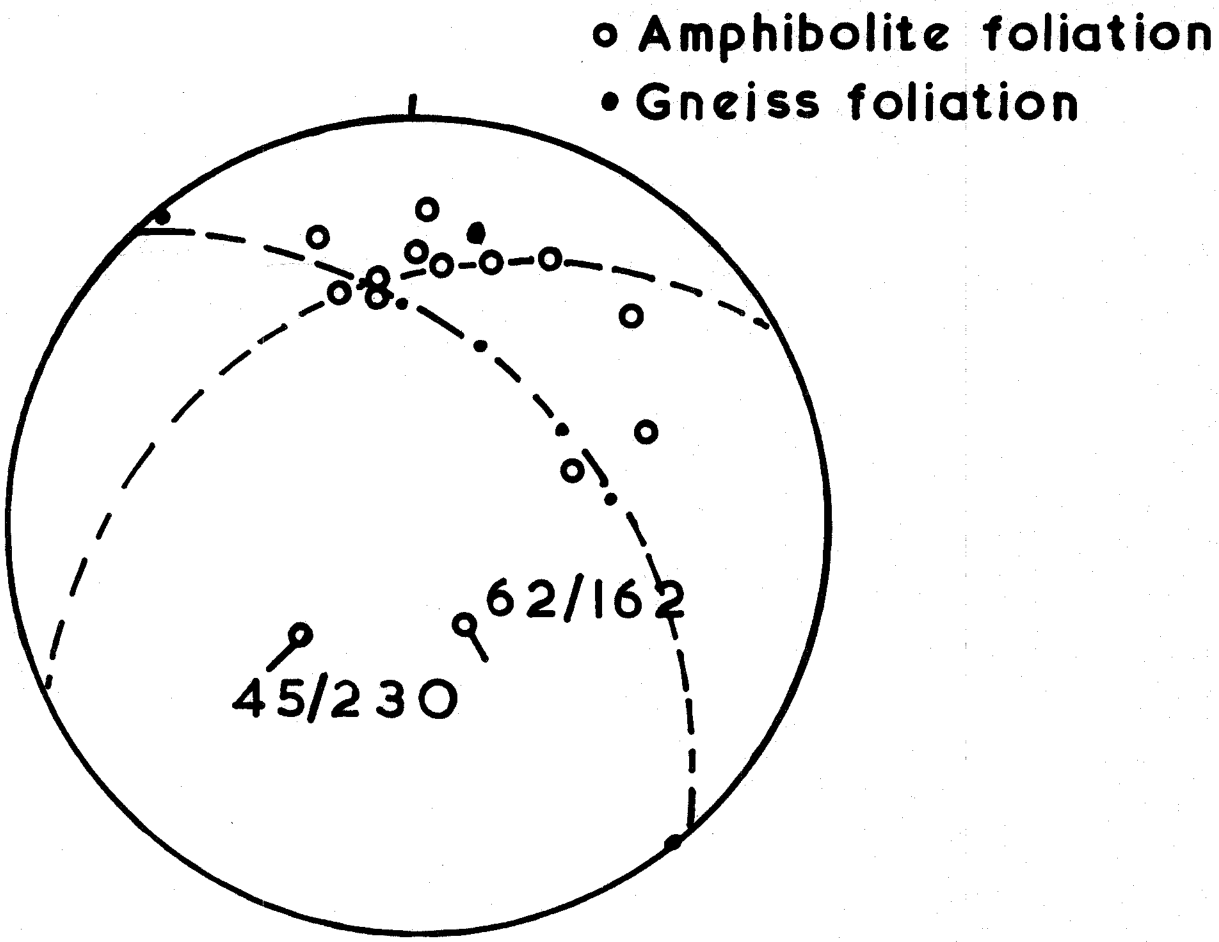


Fig. 71 πs diagram from migmatised amphibolite

apparently folded about a south-easterly plunging axis and the neosome about an axis plunging south-west. This is interpreted as the result of two different periods of folding. The earlier one was pre-migmatitic and folded the already existing amphibolite about the south-east axis. Introduction of the gneiss neosome was associated with or succeeded by folding about a south-west plunging axis which had only slight effect on the amphibolite. An alternative interpretation is that the unfolded, but foliated, amphibolite was intruded by the gneiss whose foliation was developed oblique to that of the amphibolite. On folding by one period of deformation two axes would result. However, if this were the case the two axes so obtained should lie in the axial plane of the folding. The plane that contains the two axes dips 66° almost due south (080/66S). This is different from the constructed axial plane for the latest folding and although there are appreciable inaccuracies in all these constructions it is probable that the south-east plunging axis does not belong to the same phase of folding as the south-west plunging axes.

The small fold axes and lineations have a very regular distribution. Those to the north-west of the major antiform plung north-west, those along the centre of the antiform plunge generally south-west and those to the south-east plunge south-east. Fig. 72 shows the general distribution of the small scale linear structures. The linear structures have not been plotted on stereograms because their orientations are easily seen from the map - a classic example of H.H. Read's dictum "Map it, and if it's there it will come out." The concordance between these small scale fold axes and lineations and the local fold axes constructed from foliation measurements is very close.

Two axial directions almost perpendicular to each other is a very common arrangement and has been described on numerous occasions (e.g. Ayrton, 1963; Windley, 1966, etc). It is usually interpreted as being the result of two periods of deformation, but the geometrical distribution of the axial directions alone does not prove this, (Ramsay, 1967). In the Neria Area, however, the evidence provided by the migmatisation shows that there was a time interval between the SE-NW folding and the later NE-SW folding.

Coast Area

The structure of the Coast Area is less well understood than that of the Inland Area. There are two reasons for this: firstly the Coast Area is small in relation to the large scale structures, and secondly, the small scale

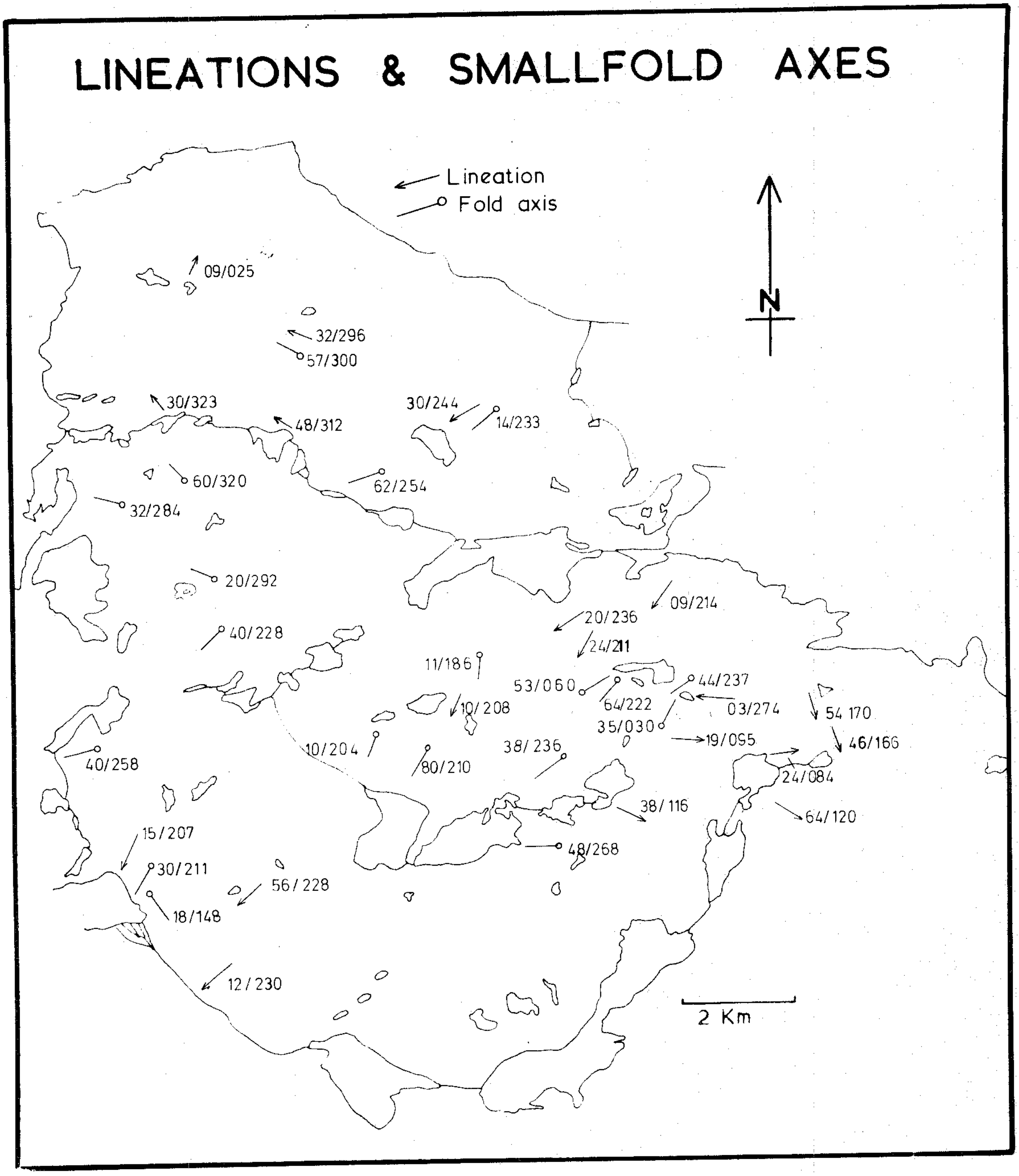


Fig. 72 Inland Area

structures, and secondly, the small scale structures are less helpful than they might be because away from the shore they cannot be seen. The main features of the structures of the Area can be seen on Map II and fig.73 which shows the attitudes of the local fold axes.

The large eastward plunging synforms and antiform are part of an even larger structure that continues to the east of Sermilik Fjord (personal communication by D.Friderichsen). The very basic parts of the succession in the cores of the folds are seen to join on the eastern side of the fjord. It is also possible that the vertically plunging fold at the southern end of the peninsula is part of this overall structure, but this cannot be demonstrated.

The western and northern parts of the Area are relatively straightforward. In sub-area II the strike of the rocks is uniformly north-south along the long lake, but north of this lake the strike swings round to become east-west (Map II). Fig. 74 shows the stereograms from which the local fold axes were constructed.

The separated lenses of amphibolite running up the centre of the Area are the result of large scale boudinisation. The rounded ends of amphibolite are not

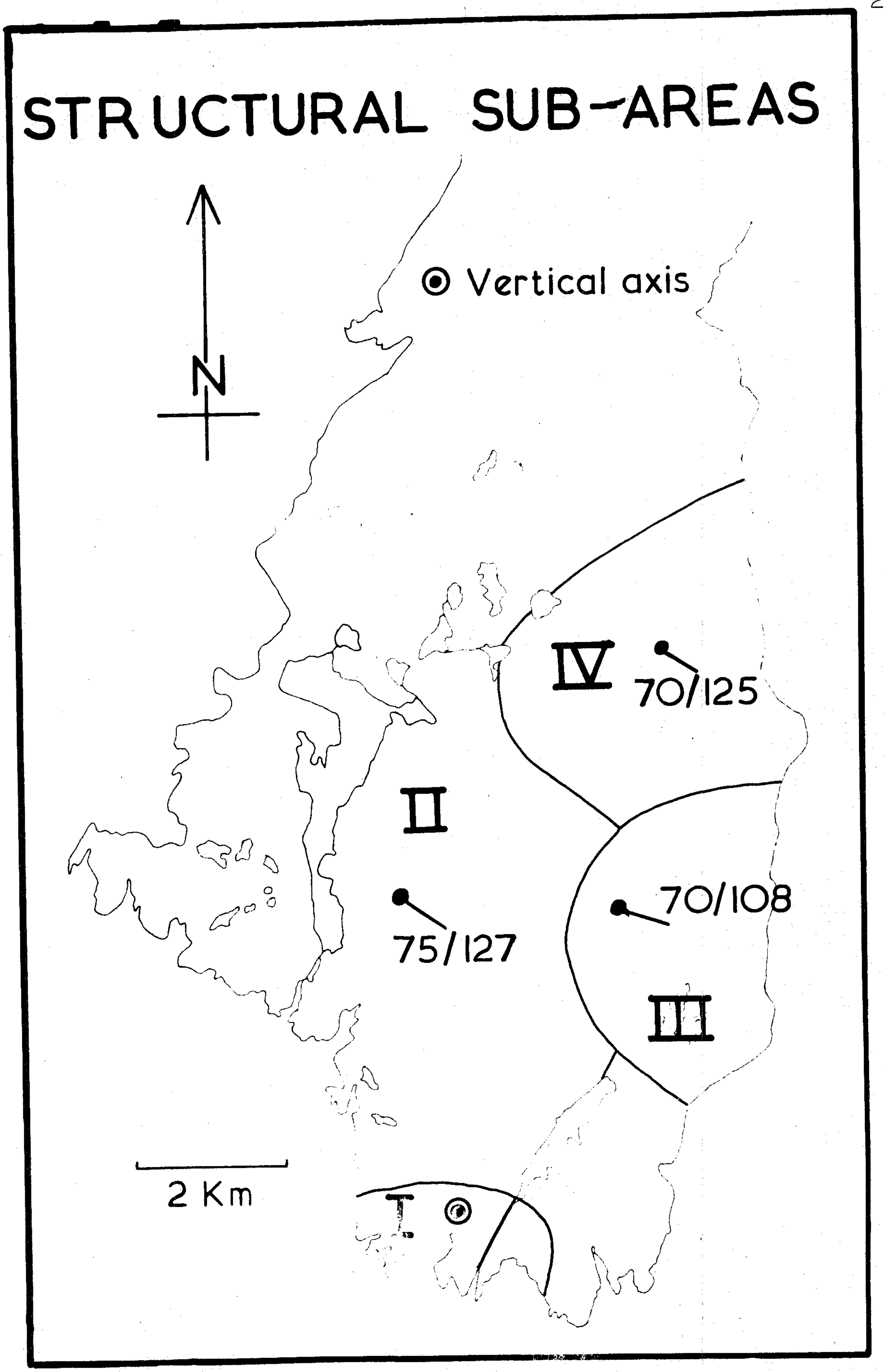


Fig. 73 Coast Area

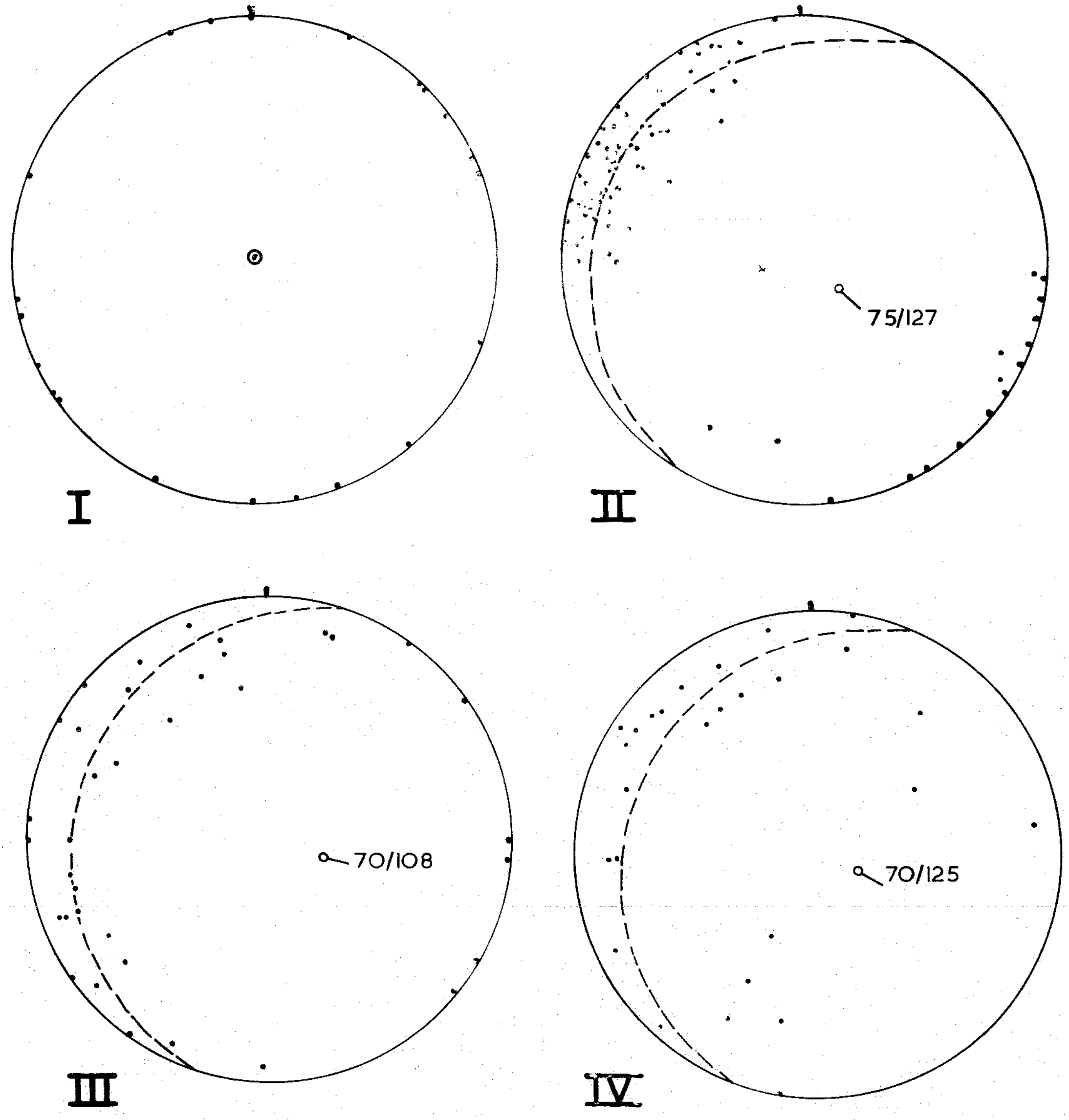


Fig.74 πs diagram for structural sub-areas of Coast Area

tight fold closures, but the separated parts of once continuous horizons. The gneiss foliation between the amphibolite lenses does not run across between them, but pinches in and out as it does on the small scale along pegmatites which show pinch and swell structures. Some of the lenses show incipient pinching which has not yet produced complete separation.

Faults and mylonites

Introduction

These two features are frequently found together, but they are distinguished by consideration of the amounts of shearing, recrystallisation, and displacement. Generally, mylonites are considered to be older than the dyking and are revealed by a strong recrystallisation of the rocks with sometimes little, or no, displacement; whereas most of the faults are younger than the Kuanitic dykes and are not necessarily accompanied by strong recrystallisation of the rocks.

Movement, with or without extensive shearing, took place along the same lines during widely separated periods. This is most clearly seen by the influence of faults on dykes: where Gardar dykes are displaced by the same fault that displaces Kuanitic dykes the former are almost invariably less displaced than the latter.

No faults within the Area have been seen to displace the youngest dykes (the T.D.s)

Inland Area

Faults

The main faults are shown on Map I. Faults with displacements of less than 100m are not shown. The fault with by far the greatest lateral displacement strikes ENE across the middle of the Area. Correlation across this fault is possible only because of its effect on the brown schist and laminated amphibolite horizon. The displacement is so great that only one of the large Kuanitic dykes can be seen on both sides of the fault, the two others cut by the fault are displaced out of the Area. The country rocks and Kuanitic dykes show a dextral displacement of 3-4 km on the western side of the Area and a slightly larger displacement of 4-5 km on the eastern side. The dyke which can be seen on both sides of the fault intersects it in grid squares B4 and D4. The close agreement between the displacement of the country rocks and the dykes shows that the fault was probably not formed until after the Kuanitic dykes were emplaced. Movement also took place after the Gardar dykes were emplaced. At this time the fault moved in the same sense as previously by approximately

1 km. The two NE striking Gardar dykes running up from the SW corner of the Area provide good correlation.

This fault forms a strong linear feature through the topography except where it cuts the high ground in the centre of the Area. East and west of the centre it forms a deep valley, but through the knot of hills it forms only a small notch. There is a certain amount of recrystallisation of the rocks along the fault and specularite mineralisation is general along its length.

The only other fault with a large displacement runs across the northern half of the Area. It is one of five ESE striking faults that all have dextral displacements. The total displacement is approximately 2km, of which half is taken up by the central fault (see Map I). The Kuanitic dykes and country rocks are displaced by equal amounts (1km), but one small red dolerite dyke is displaced sinistrally. This dyke, being fresh dolerite is probably Gardar in age and therefore it may indicate that there were two sets of movement along this fault - an early dextral one of little over 1 km and a later sinistral one of about 200 m. However, it is known that some dykes run along fault planes and thus appear to be older than the fault when in fact they are younger. This may have happened in this case, but is impossible to tell because the fault plane is covered by boulders and a small permanent snow patch.

A NE striking fault (grid square E7 Map I) with a sinistral displacement of some 400m is younger than the ESE set of faults. The age relations are shown by the relationships of different generations of Kuanitic dykes to the faults. Two NE striking dykes are displaced dextrally by an ESE fault and along the fault runs a younger Kuanitic dyke which is in turn displaced sinistrally by the NE striking fault.

Mylonites

Mylonite zones in the Inland Area are not more than 10-20 m wide and are recognised by the usual porcelanous texture of the rocks. A green colour derived from finely disseminated chlorite is commonly present as is a red staining from haematite. Most of the mylonites have produced very little or no lateral displacement of the rocks. The mylonite running through grid squares F2 and E3 (Map I) has produced no visible displacement, but considerable shearing did take place along it because even the Gardar dykes are sheared by it (but not displaced laterally). This mylonite probably is the cause of the deep valley WNW of the 870m lakes, but because the floor of that valley is choked with boulders and there is no displacement of the brown schist the mylonite cannot be seen.

Coast Area

Faults

There are few faults of any size in the Coast Area. The largest occurs in the SE of the Area and has a lateral displacement of 2-300m (see Map II). Two other dykes striking normal to the first have sinistral displacements of 1-200m each. Both these sets of faults are younger than the youngest Kuanitic dykes, but their ages relative to each other or to the Gardar dyke are not known

Mylonites

There are several small mylonite zones such as were found in the Inland Area, but there is also a broad zone of mylonitisation which occupies the westernmost peninsula of the Area (see Map II). Here a zone almost 1km across has been strongly recrystallised. The recrystallisation has not been to a uniform degree throughout the whole width of the zone and in places the former character of the rocks can still be seen, but in most of the zone the rocks are pale green in colour and porcelanous in texture. Small folds very commonly occur in the mylonite. These are clearly related to the formation of the mylonite as their axial planes are parallel to the strike of the mylonite as a whole.

In thin section the rocks can be seen to have been almost completely crushed. Those parts of the rocks that are predominantly composed of quartz, such as old pegmatites, are made up of a mosaic of very small, largely strain-free quartz crystals. In more feldspar rich rocks rounded remains of feldspar grains are surrounded by a groundmass of very much finer grained quartz and feldspar. Only rocks that were originally rich in micas or amphiboles have retained any recognisable mafic minerals, and even these have been reduced in size and bleached.

Movement recurred along the mylonite at intervals as Kuanitic dykes have been mylonitised. The main mylonitisation took place before the dykes were intruded because the dykes are much less altered than the country rocks. Only one side of the mylonite zone occurs in the Area, therefore it is impossible to determine how much, if any, lateral displacement there has been.

CHAPTER ELEVEN

CHRONOLOGY

The methods of normal stratigraphy are of limited assistance in determining the plutonic development of basement complexes, but it is still important to establish a stratigraphical succession. If a recognisable succession is not set up then the determination of the structure is made much more difficult because of the numerous ways in which parts of the succession can be changed: by tectonic activity, transformation or metamorphism. Once the stratigraphy is known and marker horizons found it becomes possible to extrapolate from them and thus identify featureless gneisses. In the Inland Area studied here, for example, the brown schist and laminated amphibolite are an excellent marker; and once these two are identified in an unexplored part of the Area the Upper and Lower Biotite Gneiss Successions are immediately known and the problem of linking the structures of the new ground with the known structures of the old is made much simpler.

Having established the stratigraphy, however, it becomes necessary to use completely different criteria to determine the sequence of events. The criteria used are the relationships between the structural elements, periods of metamorphism, migmatisation and granitisation, if possible basic dyking, and radiometric dating. It can never be known for certain in infracrustal rocks that all the events have been recognised; there will always be the possibility that features imposed by the earliest events have been completely obliterated by the later.

The oldest recognisable plutonic features of the Neria Area rocks are the banding and pre-migmatisation folds. The banding seen at the present may or may not be the first imposed on the rocks, and may in fact be of more than one age. Fig.63 shows that banding has indeed been imposed on more than one occasion, but except around the hinges of folds the old and new bandings are parallel, or so nearly so that they are inseparable in the field. At the same time as the banding was produced the rocks were metamorphosed and the platy minerals gave rise to the foliation.

Following these first plutonic events the migmatisation and second phase folds were formed. Both major and minor folds belonging to this period are found with axial directions striking NW-SE. Individual small-folds of this period often have an axial plane foliation if there is no axial plane pegmatite developed, but a new foliation was not imposed on a regional scale as a pre-existing foliation was, in certain circumstances, preserved during this period (see above p. 225 & fig. 71).

The SW-NE folds followed the migmatisation. This is shown by the way in which the huge pegmatites bodies in the Laminated Amphibolite are folded along with the Amphibolite around SW plunging axes. After this period of deformation the granites were emplaced to be followed by the formation of mylonite zones, cooling, dolerite dyke intrusion, and faulting. There was a further slight metamorphism of the Inland Area between the intrusion of the Kuanitic and Gardar dykes, but this had no effect on the mineralogy of the country rocks (see above p.201).

The discordant amphibolites cannot be placed with any certainty in this chronology, but because they are unmigmatised they were emplaced after the migmatisation and the fact that they are wholly metamorphic in texture shows that they were present before strong metamorphism ceased. Because they are so few in number it is impossible to get enough information to decide whether they were emplaced in cold rocks which were subsequently reactivated, or whether they were emplaced in hot rocks and crystallised directly as metamorphic rocks.

The significance of discordant amphibolites in basement complexes is debatable, but it is clear that it cannot be taken for granted that they indicate a break in plutonic conditions (Watterson, 1965). It is not suggested here, therefore, that the few found in the Coast Area prove that the country rocks were subjected to two different plutonic periods separated by cooling and uplift. The less complicated history of a single, but long, plutonic period culminating in the formation of granite is thought to be more probable. Table XII summarises the chronology of the Area.

TABLE XII Summary of Chronology

Structural Event	Plutonic Event	Mineral Paragenesis	
		Amphibolite	Brown Schist
	Intrusion of Gardar dykes Faulting Weak regional metamorphism in Inland Area Intrusion of Kuanitic dykes Mylonitisation along particular zones		
Final phase of folding (NE-SW)	Formation of granite Introduction of discordant amphibolites	Destruction of garnet & salite in Inland Area	Local chloritisation of garnet
Second phase of folding (NW-SE) Folds with axial phase foliation	Continuous metamorphism and migmatisation with main period of migmatisation contemporaneous with second phase of folding	Hornblende-salite-garnet-plagioclase	Growth of garnet Growth of sillimanite Growth of cordierite
Early folds, often with sheared limbs. Formation of new banding Earliest recognisable banding and foliation		Growth of tremolite-actinolite	Growth of biotite

References

Ayrton, S.N., 1963: A contribution to the geological investigations in the region of Irigtut, S.W. Greenland. Medd. om Grønland, Bd. 167, Nr.3.

Barrow, G., 1893: On an intrusion of muscovite-biotite gneiss in the South-eastern Highland of Scotland, and its accompanying metamorphism. Quart. J. geol. Soc. Lond. v.49, p.330.

Berthelsen, A., 1957: The structural evolution of an ultra- and polymetamorphis gneiss-complex, West Greenland. Geol. Rdsch., Bd. 46, p.173.

————., 1960: Structural studies in the Pre-Cambrian of Western Greenland; Pt. II: Geology of Touqussoq nuna. Medd. om Grønland, Bd. 123, Nr.1

————., 1961: Structural classification of gneisses - as used in team work in S.W. Greenland. 21st Int. Geol. Congress, Norden. Pt. XXVI, p.69.

Bridgwater, D., 1965: Isotropic age determination from South Greenland and their geological setting. Medd. om Grønland, Bd. 179, Nr. 4.

Buddington, A.F., 1959: Granite emplacement with special reference to North America. Bull. geol. Soc. Amer., v. 70, p.671.

Chayes, F., 1965; Titania and alumina content of oceanic and circumoceanic basalts. Min. Mag. v.34, p.126.

Chinner, G.A., 1960: Pelitic gneisses with varying ferrous/ ferric ratios from Glen Clova, Angus, Scotland. J. Pet., v.1, p.178.

————., 1961: The origin of sillimanite in Glen Clova, Angus. J. Pet. v.2, p.312.

————., 1962: Almandine in thermal aureoles. J. Petrology, v.3, p.316.

Davies, O.L., 1957: Statistical methods in research and production. London.

Deer, W.A., Howie, R.A., and Zussman, J., 1962: Rock-forming minerals. London.

Engel, A.E.J., and Engel, C.G., 1962a: Progressive metamorphism of amphibolite, northwest Adirondack Mountains, New York, Petrologic Studies. In Geol Soc. Amer. Buddington vol., p.37.

——————., 1962b: Hornblendes formed during progressive metamorphism of amphibolites, northwest Adirondack Mountains, New York, Bull. geol. Soc. Amer. v.73, p.1499

Evans, B.W., and Leake, B.E., 1960: The composition and origin of the striped amphibolites of Connemara, Ireland, J. Pet., v.1, p.337.

Fyfe, W.S., 1967: Metamorphism in mobile belts. Trans. Leicester Literary and Philosophical Soc., v. LXI, p.36.

Fyfe, W.S., Turner, F.J. and Verhoogen, J., 1958: Metamorphic reactions and metamorphic facies. Geol. Soc. Amer. Mem. 73.

Harker, A., 1939: Metamorphism, London

Hietanen, Anna, 1951: Metamorphic and igneous rocks of Merrima Area, Plumas National Forest, California. Bull. geol. Soc. Amer., v.62, p.565.

——————, 1956: Kyanite, andalusite, and sillimanite in the schist in Boehls Butte quadrangle, Idaho, Amer. Mineralogist, v. 41, p.1.

——————, 1967: On the facies series in various types of metamorphism. J. Geol. v. 75, p.187.

Higgins, A.K., and Bondesen, E., 1966: Supracrustals of pre-Ketilidian age (the Tartoq Group) and their relationships with Ketilidian supracrustals in the I vigtut region, South-West Greenland. Rapp. Grønlands geol. Unders., Nr. 8.

Holland, J.G. and Brindle, D.W., 1966: A self-consistent mass absorption correction for silicate analysis by X-ray fluorescence. Spectrochim. Acta., v. 22, p.2083.

Kalsbeek, F., 1965: On the origin of some banded amphibolites and gneisses in the Belledonne Massif (French Alps). N. Jb. Miner. Abh., Bd. 192p. 177.

Kalsbeek, F., 1967: The pattern of folding in an area of migmatites between Neria and Qasigialik Fjords, South-West Greenland. Medd. om. Grønland, Bd.175,Nr.4

—————., and Zwart, H.J., 1967: Zircons from some gneisses and granites in the Central and Eastern Pyrenees. Geolgie en Mynbouw, 46e jaargang, p.457.

Korzhinskii., D.S., 1959: Physicochemical basis of the analysis of the paragenesis of minerals. New York.

Kuno, M., 1960: High-alumina basalt. J. Pet. v.1, p.121.

Larsen, O., 1966: K/Ar age determinations from Western Greenland. Rapp. Grønlands geol. Under., Nr. 11.p.57

Leake, B.E., 1964: The chemical distinction between ortho- and para-amphibolites. J. Pet., v.5, p.238.

McGregor, V.R., 1966: Migmatization and deformation in Ameralik, Godthåb Commune, West Greenland, and their effect on a swarm of basic dykes. Rapp. Grønlands geol. Unders., Nr. 11, p.29.

Misch., P., 1965: Radial epidote glomeroblasts formed under conditions of synkinematic metamorphism - a new mechanism of collective crystalloblastesis. Geol. Rdsch., Bd. 54, p.944.

Miyashiro, A., 1958: Regional metamorphism of Gosaisyo-Takanuki District in the Central Abukuma Plateau J. Fac. Sci., Univ. Tokyo, Sec.II, v.2, p.219.

—————, 1961: Evolution of metamorphic belts. J.Pet. v. 2, p.277

Newton, R.C., 1966: Kyanite-sillimanite equilibrium at 750°C., Sci., v. 151, p.1222.

O'Hara, M.J., 1961: Zoned ultra-basic and basic gneiss masses in the early Lewisian metamorphic complex at Scourie, Sutherland, J. Pet., v.2, p.248.

Pettijohn, F.J., 1957: Sedimentary rocks. New York,

Poldervaart, A., 1955: Zircon in rocks, I, Sedimentary rocks. Amer. J. Sci., v. 253, p.433.

—————., 1956: Zircon in rocks. II, Igneous rocks. Amer. J. Sci., v. 254, p.521.

Ramsay, J.G., 1967: Folding and fracturing of rocks. New York.

Rast, N., 1964: Nucleation and growth of metamorphic minerals. In Controls of metamorphism. Ed. by Pitcher and Flinn. Edinburgh.

Read, H.H., 1927: The igneous and metamorphic history of Cromar, Deeside, Aberdeenshire. Trans. R. Soc., Edinburgh., v. 55, p.317.

———., 1952: Metamorphis and migmatisation of the Ythan Valley, Aberdeenshire. Trans. Edinb., geol. Soc., v. 15, p.265.

Schreyer, W., and Yoder, Jr., H.S., 1964: The system Mg-cordierite - H_2O and related rocks. Neues Jahrb. Miner. Abh., Bd.101, p.271.

Schido, F., 1958: Plutonic and metamorphic rocks of the Nakoso and Iritono districts in the Centrak Abukuma Plateau. J. Fac. Sci., Univ. Tokyo, Sec. II, v.11, p.131.

Slemmons, D.B., 1962: Determination of volcanic and plutonic plagiocalses. Geol. Soc. Amer. Sp. Paper, Nr.69.

Sutton, J., and Watson, J.V., 1951: The pre-Torridonian metamorphic history of the Loch Torridon and Scourie areas in the North-west Highlands, and its bearing on the chronoligal classification of the Lewisian. Quart. J. geol. Soc. Lond., v.106, p.241.

Sørensen, H., 1953: The ultrabasic rocks at Touqussoq., West Greenland. Medd. om. Grønland, Bd.136, Nr.4

———., 1954: The border relations of the dunite at Siorarssuit, Sukkertoppen District, West Greenland. Medd. om. Grønland, Bd. 135, Nr. 4

Tobi., A.C., 1961: Pattern of plagioclase twinning as a significant rock property. Proc. Kon. Nederl. Ak. Wetenschappen. Ser. B., v.64, p.576.

Turner, F.J., and Verhoogen, J., 1960: Igneous and metamorphic petrology. New York

Turner, F.J. and Weiss, L.E., 1963: Structural analysis of metamorphic tectonites. New York.

Tuttle, O.F., and Bowen, N.L., 1958: Origin of granites in the light of experimental studies in the system $N_aAlSi_3O_8 - KAlSi_3O_8 - S_iO_2 - H_2O$. Mem. geol. Soc. Amer. 74.

Vance, J.A., 1961: Polysynthetic twinning in plagioclse. Amer. Min. v. 46, p.1097.

de Waard, D., 1965: The occurrence of garnet in the granulite-facies Terrane of the Adirondack Highlands, J. Pet., v.6, p.165.

Walton, B., 1966: The relationship between relic pillow structures and zoned calc-silicate skarns, and the significance of talc balls in gneisses south of Frederikshåb . Rapp. Grønlands geol. Unders., Nr. 11.p.38

Watterson, J., 1965: Plutonic development of the Ilordeq area, South Greenland, Pt.I. Medd. om Grønland Bd. 172, Nr.7.

———————., 1967: Homogeneous deformation of the gneisses of Vesterland, South-West Greenland. Medd. om Grønland, Bd. 175, Nr. 6.

Wegmann, C.E., 1938: Geological investigations in Southern Greenland, Part I: On the structural divisions of Southern Greenland. Medd. om Grønland, Bd. 113, Nr. 2

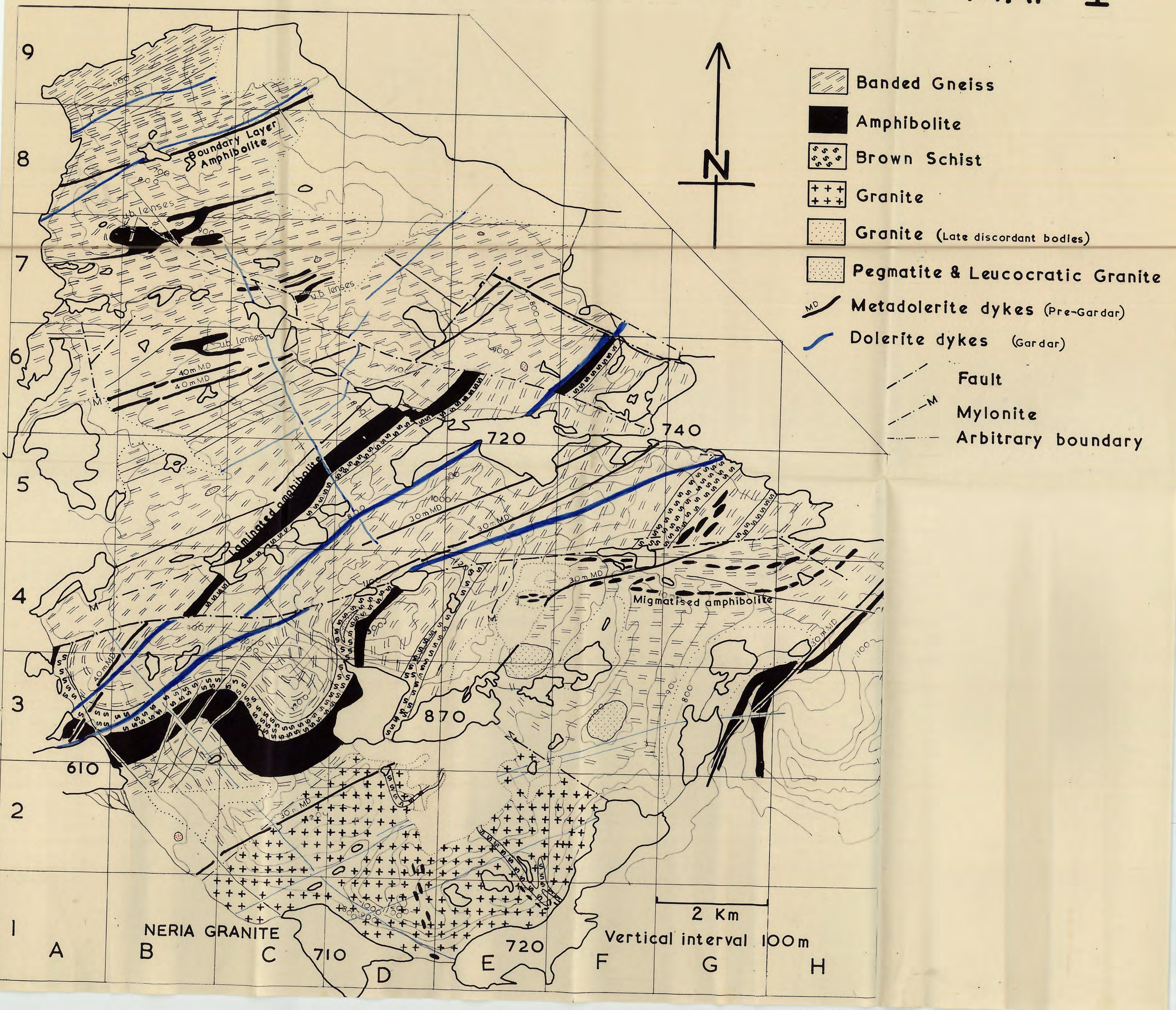
Windley, B.F., Henriksen, N., Higgins, A.K., Bondesen, E. and Jensen, S.B., 1966: Some border relations between supracrustal and infracrustal rocks in South-West Greenland. Rapp. Grønlands geol. Unders., Nr. 9.

Winkler, H.G.F., 1967: Petrogenesis of metamorphic rocks. Berlin.

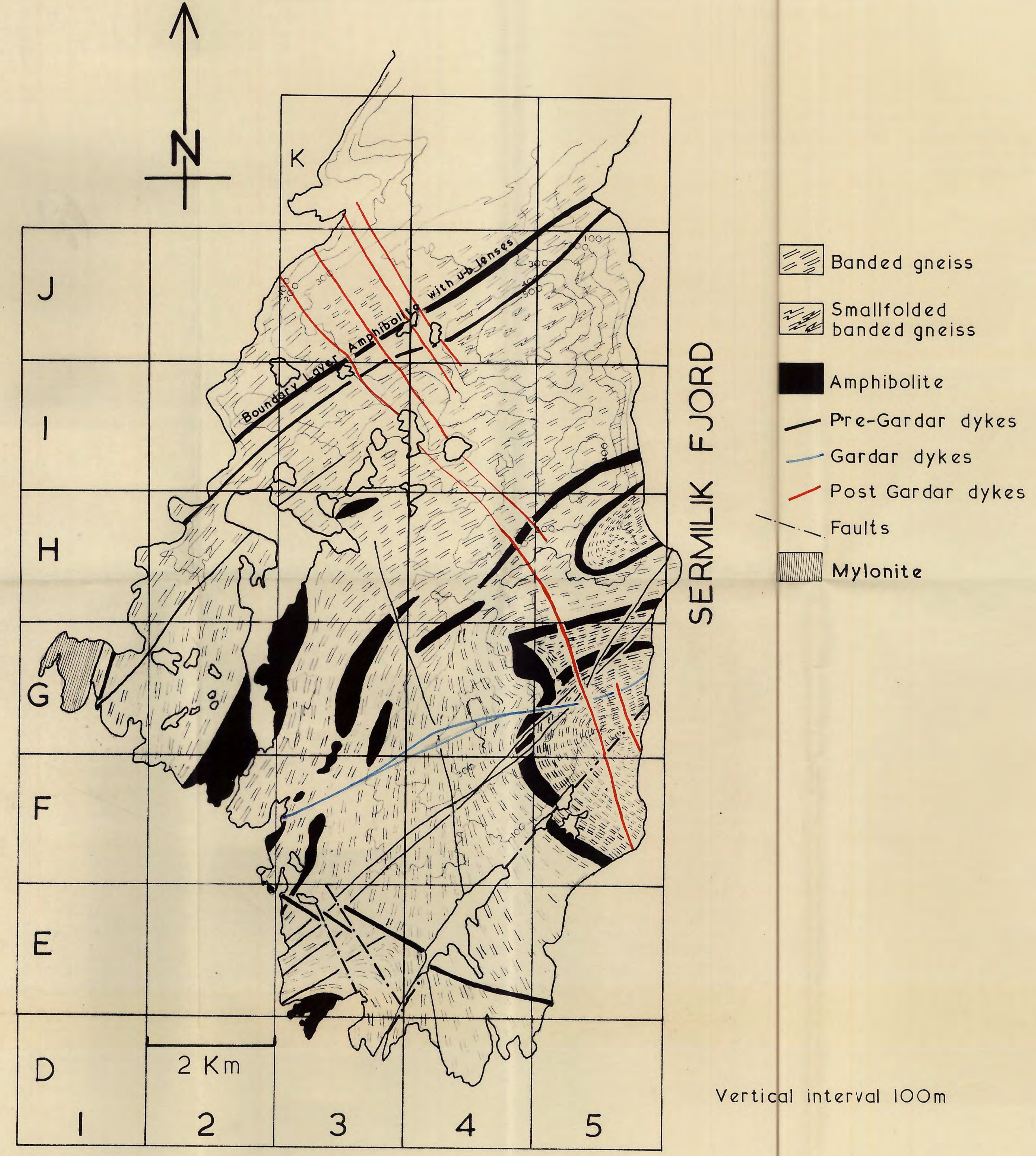
Zwart, H.J., 1958: Regional metamorphism and related granitization in the Valle de Aran. Geologie en Mÿnbouw, N.S., v.20, p.18.

—————., 1962: On the determination of polymetamorphic mineral associations and its application to the Bosost area (Central Pyrenees). Geol. Rdsch., v. 52, p.38

GEOLOGY OF N.E. NERIA, S.W. GREENLAND MAP I



MAP II


N
K
J
I
H
G
F
E
D
1
2
3
4
5
Boundary Layer Amphibolite with u-b lenses
100
200
300
400
500
SERMILIK FJORD
Banded gneiss
Smallfolded banded gneiss
Amphibolite
Pre-Gardar dykes
Gardar dykes
Post Gardar dykes
Faults
Mylonite
2 Km
Vertical interval 100m