Studies in the Structure and
Metamorphic Petrology of the
Eo-Cambrian Rocks of
Eastern Seiland,
North Norway.

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Abstract

The rocks exposed on the eastern part of the island of Seiland are the metasedimentary envelope to the large basic and ultrabasic plutons that make up the western part of the island and the neighbouring island of Stjernøy. These metasediments have been intensely deformed and metamorphosed, thus all sedimentary structures have been obliterated. The metasedimentary sequence is, therefore, a structural one.

The lowest group is the psammitic Komagnes Group. This is followed by the Eidvågeid Schist Group which is followed by the relatively thin psammitic Trollvann Group. Structurally above this is the pelitic Olderbugten Group and finally the dominantly psammitic Olderfjord Group.

Broadly speaking two major fold-forming deformations have been recognised; F.1 and F.2. The latter part of the first phase, which was responsible for tight isoclinal folds and a penetrative schistosity, was accompanied by intrusion of sheets of basic material parallel to the axial planes of the early folds.

During the static interval separating the two deformations the maximum grade of metamorphism was achieved, this, however, varies across the area and a sequence of metamorphic isograds have been recognised. The highest grade occurs in the west and is marked by a kyanite-sillimanite porphyroblastesis, migmatisation and intrusion of adamellitic sheets. The lowest grade in the east is characterised by albite, biotite, epidote, hornblende assemblages in the psammites of the Lower Komagnes Group. These isograds reflect a contemporaneous lateral change in metamorphic grade.

Following the development of these high grade assemblages in the west, there was a phase of intense flattening leading to the development of mylonitic textures in the rocks. It is suggested that this deformation phase is related to a rising basic asthenolith.

The majority of folds on the area are attributed to F.2.
By the onset of F.2. the metamorphic grade appears to have waned to
sub-garnet grade conditions.   The folds have a very variable style.
In the east, they have intensely attenuated long-limbs with a number
of vertically-stacked folds in the short-limbs.   In the west, the
limbs are of more equal length.   This change in style is related
to the different states of competence of the rocks at the onset of
F.2.

A ubiquitous feature of F.2. folds on both the major and
minor scale, is the curvature of their axial-lines.

In the east a number of oblique boudins have been recorded
which post-date F.2.   They are closely associated with rotated
tension-gashes and monoclinal folds.   It is suggested that all these
structures were formed in response to a progressive deformation
sequence.   This sequence was also responsible for the development
of the late Caledonian thrusts on the mainland.

In the north of the area there is a pronounced swing in
strike.   A number of open folds are associated with this swing.
These folds have been designated F.3.

Monoclinal warps have been recorded in the east of the area.
Their temporal relationship to the F.3. folds in the north is not known.

The final phase of movement in the area led to the development
of joints, faults and locally kink-folds.   The prevailing metamorphic
grade during this phase of deformation, and indeed all the phases
subsequent to F.2., was the Quartz-Albite-Muscovite-Chlorite Sub-facies
of the Greenschist Facies.

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CHAPTER I

Introduction

A)   Location

The island of Seiland is situated some 3 km. S.W. of the town

of Hammerfest Lat. 70° 40'N, Long. 23° 80'E in the western part of

the 'county' or fylke of Finnmark. . The island, which is 22 km.

in a N - S direction by about 30 km. in an E - W direction, is

separated from the Norwegian mainland by a long straight sound known

as Vargsund which is 3 km. wide.

The area mapped consists of some 140 sq.km. in the eastern

part of the island.   The work was carried out in the summers of

1966 - 1968 using aerial photographs on an approximate scale of

1:20,000.   These were flown in July of 1966 and were subsequently

used to make an overlap map of the area.   There are no accurate

topographic maps of the region.   The contoured A.M.S. maps proved

to be rather inaccurate in such detail as positioning of lakes,

rivers and topographic features and were therefore of little use

in the work.   However, altitudes which are quoted in the text are

taken from this series.   Norwegian Admiralty charts give an

accurate outline of the coastline and were used to estimate the

scale of the aerial photographs.

FIG 1
LOCATION MAP

N

50 KM

SØRØY

HAMMERFEST

SEILAND

STJERNØY

LOPPA

ALTA

O TROMSØ

BASIC ROCKS

TECTONIC WINDOWS

AREA MAPPED

70°N

60°N
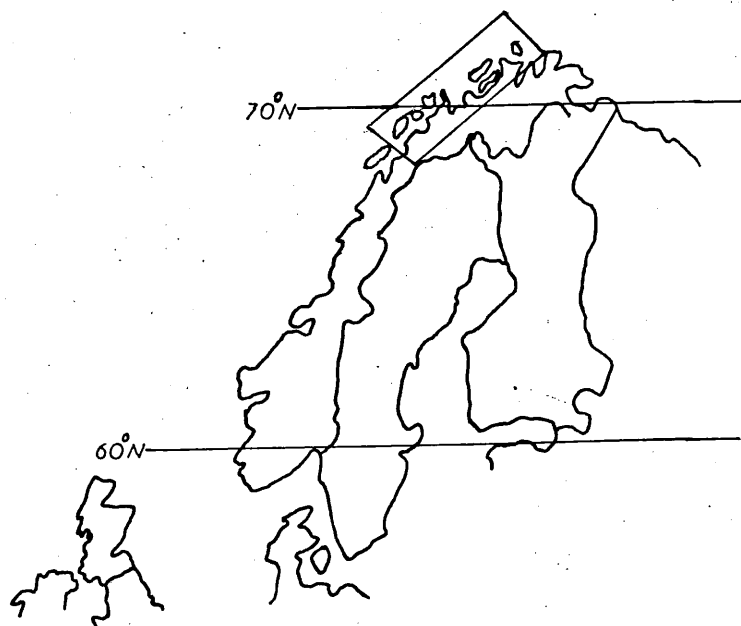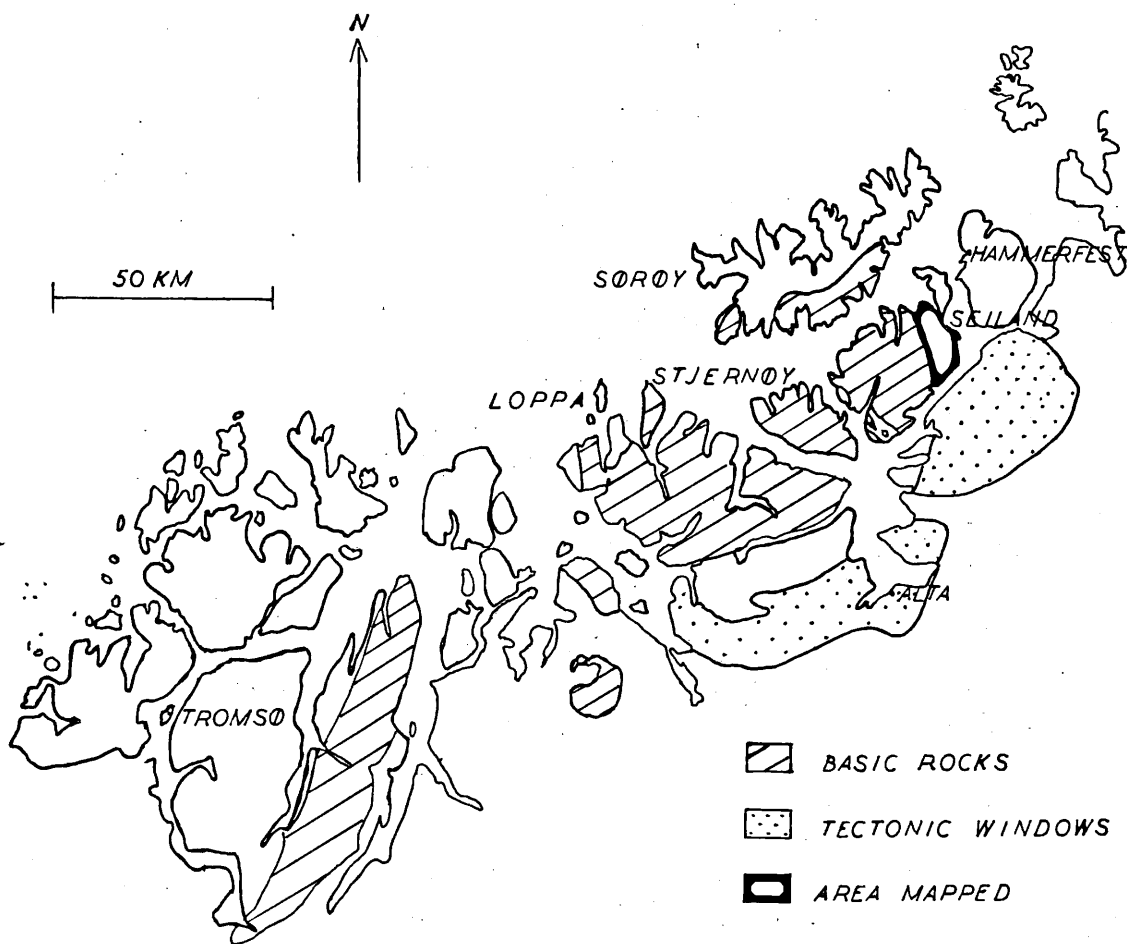
B) Regional Geology

Seiland is situated some 60 - 70 km. North-West of the main

Caledonian thrust front,where the strongly deformed and highly

metamorphosed Caledonides are thrust over both Pre-Cambrian basement

and younger less-deformed rocks of Eo-Cambrian to Tremadocian age.

In the south and west the Caledonides are thrust over

autochthonous Cambrian deposits which rest on the Archaen basement.

In the N. and N.E. of Finnmark, however, there occurs a wedge of

rocks of late Pre-Cambrian to Tremadocian age.   These are again

separated from the overlying Caledonides by a thrust contact, and

appear to rest with a tectonic boundary on the Pre-Cambrian gneissic

basement.   The relationship between the relatively unaltered

sediments that lie beneath the main thrust plane in the N.E. and

those in the south and west is not known.

On the mainland, to the S.E. of Seiland there occur a number

of tectonic windows in which Pre-Cambrian rocks are exposed (see

Fig. 1); these are known as the Raipas windows (Dahll 1891).

In the Komagfjord windows which occur on the mainland adjoining

Seiland, (Reitan 1963) records a sequence of shales, slates, volcanics,

calcareous rocks and sandstones which have been deformed probably by

the Pre-Cambrian Karelic orogeny.   The grade of metamorphism is

generally low, the maximum being attached only locally in the region

of basic intrusions.   This association,suggests Reitan, may be

fortuitous since this area also corresponds to an anticlinal crest,

which may have acted as a zone of weakness allowing intrusion
of the basic material.   The highest grade is, in fact, the Quartz
Staurolite - Sub-facies of the Almandine Amphibolite Facies.   These
Pre-Cambrian rocks are separated from the overlying high-grade
Caledonides by a thrust contact.   The general sense of translation
of this thrust appears to be in a south-easterly direction.   Since
the Komagfjord window is bounded on the north-west by Vargsund, and
high-grade Caledonides occur in Seiland on the other side of the
sound, it is assumed that the thrust passes beneath Vargsund.

Previous Work on the Caledonides of W. Finnmark

Most of the detailed structural work on the Caledonides of
W. Finnmark has been carried out since 1959 on the neighbouring
island of Sørøy by B.A. Sturt and D.M. Ramsay and a number of research
students working under their direction (Roberts 1965, 1968, Speedyman 1968).
As a result of their investigations a complex series of structural,
igneous and metamorphic events have been recognised.

Broadly speaking, two major fold phases are apparent, the earlier
phase giving rise to recumbent isoclinal folds of considerable amplitude
(approx. = 10 km.);   the anticlinal nappes close towards the south-east
and the corresponding synclinal nappes towards the north-west (Ramsay
and Sturt 1963).   This phase of folding was also responsible for the
development of a strong schistosity.   The later part of this deformation
was accompanied by the intrusion of sheets of gabbroic material, parallel

to the axial planes of the folds (Speedyman 1968). A peculiarity of the F.1. folds is that their axes are arcuate, both on a major and minor scale.

In the static interval separating the two fold phases other gabbro and diorite were intruded. The maximum grade of metamorphism was also achieved forming porphyroblasts of sillimanite, kyanite, staurolite indicative of the upper Almandine Amphibolite facies. This was accompanied by granitisation and migmatisation.

The second phase of deformation of Sørøy is generally characterised by large scale folds which have a tendency towards monoclinic symmetry. The associated minor structures show considerable variation in style. The symmetry remains however monoclinic, except in certain areas where it becomes orthorhombic. This change appears to be related to orthogonal strike swings (Roberts op.cit.).

Further intrusions of gabbroic and dioritic material occurred during the latter part of the F.2. folding (Speedyman op.cit.). Sturt and Ramsay have also recorded the extensive development of alkaline rocks of similar age in western Sørøy. They include carbonatites and nepheline syenites (Sturt and Ramsay 1965).

All the previous work on the island of Seiland has been restricted to studies on the layered intrusions that make up the western part of the island (Barth 1927 1953, OOsterom 1953). The term, Seiland Petrographic Province, was coined by Barth as a result

of this work, to describe the extensive belt of basic and ultra-
basic intrusions which constitutes an important part of the Geology
of the coastal regions of West Finnmark and North Troms (see Fig.1.).

Numerous studies of this Province have been made. Papers by
Barth (1927 and 1953) deal with the complex intrusion history of
numerous pegmatites (1927) and the layered igneous rocks that make
up a large part of western Seiland (1953).

In north-western Seiland Barth (1953) describes a layered
complex which apparently has a gradational contact with a series of
paragneisses, showing successively higher stages of metamorphic and
anatectic transformation when traced towards the gabbro, which
Barth considers eventually grades into layered gabbro. He concludes
from this that the layered gabbro is the result of high grade
metamorphic anatectic transformation of the paragneisses.

This idea has found considerable support among other workers,
notably Krauskopf (1954), Oosterom (1954, 1963), Heier (1961).
Krauskopf (op.cit.), working in the Øksfjord area, describes a
sequence of gabbro-gheisses, ultramafites, syenites, anorthosites,
amphibolites and garnet-biotite-gneisses. He recognises three
varieties of gabbro-gneiss, all of which show variable amounts of
banding. Within these gneisses, he finds layers of meta-limstone
which are apparently concordant with the layering of the country
rocks. He also describes layers of syenitic rock which are also
concordant. The layering in these rocks is marked by differences
in composition and/or grain size. The layers are very variable in
thickness and length and sometimes pinch out. Contacts are partly
gradational and the layering is parallel to a planar foliation.

These structures, which are described repeatedly by a number
of workers, are similar in many respects to those developed in
such layered complexes as Skaergaard, Duluth, Bushveld and
Stillwater.   It must be pointed out, however, that the rocks of
the Seiland Province are syntectonic relative to the Caledonian
erogeny, whereas the other well-known intrusions developed in
non-tectonic environments.

To explain the layering, shown by the Øksfjord gabbro-gneisses,
Krauskopf discusses a number of hypotheses, including parallel
injection of sills.   He suggests that such features as absence of
linear orientation of prismatic crystals, absence of rhythm in the
layers and no consistent upward transition of mafic to less mafic
material in individual layers are not compatible with an origin by
fractional crystallisation.   He also considers that, since the meta-
limstone layers are concordant with the surrounding structures, they
are not likely to be roof pendants.

Finally, he appears to favour Barth's hypothesis of the meta-
morphic and anatectic transformation of a supracrustal series of lavas,
tuffs and sediments.   Planar foliation may thus be explained as
recrystallisation parallel to bedding planes.   The syenitic layers
are explained by the transformation of trachytic lavas and tuffs.
However, he finds difficulty in explaining the presence of anorthosite
and pyroxene magnetite layers, since such lavas and tuffs of this
composition are rare if not non-existent.   He concludes that they may
possibly be explained by metamorphic differentiation.   Krauskopf
considers that the peridotites and massive gabbros, together with the
massive syenite are of intrusive origin.

Heier (1961) in a study mostly concerned with the nepheline
syenites and carbonatites on the island of Stjernøy, which lies to
the north of Seiland, places the various metamorphic and plutonic
events in a time sequence.

Youngest. e3) Nepheline syenite

e2) Hornblendite metasomatic probably associated with
carbonatites

e1) Carbonatites

d) Faulting

c) Period of Metamorphism of Granulite facies possibly
pre-dates the peridotites and even the layered gabbros

b) Peridotites

Oldest a) The gabbro gneiss complex. Age relationships
very difficult to ascertain

Heier recognises three gabbro-gneisses in the complex:-

Gabbro gneisses I he considers to be equivalent to Krauskopf's
gabbro gneisses I. In this body syenitic and quartz-rich gneisses
occur as layers parallel to the foliation. Heier considers that this
rock-type is the result of the transformation of a supracrustal series
of lavas, tuffs and sediments.

The gabbro-gneiss II is foliated but not banded. It has a
metamorphic texture. The gabbro gneiss III and the layered gabbro,
which Heier implies are cognate, have a layering produced by con-
centrations of dark minerals. Often these have indistinct and
gradational contacts with the adjacent leucocratic bands. The
layers vary in thickness from $\frac{1}{2}$ cm. to several metres. The dark

bands are on average 10 - 20 cms thick. Heier (op.cit.) cites a statement by Barth concerning the Seiland gabbro. Barth says that this "is different from the Skaergaard in that it has no definite side walls, but seems to extend without a clear break into the contiguous and analagously layered amphibolite-gneiss complex". Heier states that, in general, this is also true of the layered gabbro in Stjernøy. Thus, he implies his agreement with the supposition that the layered gabbro was also produced by a 'process of metamorphic and anatectic' transormation, as proposed by Barth for north-west Seiland.

The peridotites are apparently younger than the layered gabbro sequence since xenoliths of the latter occur in the former.

The carbonatites and hornblendites seem to be genetically associated; all gradations between the hornblendite and the gabbro-gneiss occur. This seems to indicate that the hornblendite results from the metasomatic replacement of the gabbro-gneiss and also of the peridotite.

M.G. Oosterom (1963), also working on the island of Stjernøy recognises a sequence of metamorphic and igneous events which is broadly in agreement with that of Heier.

Youngest

	(Albite Nepheline Pegmatites
Metasomatic Suite	(Nepheline Syenite
	(Carbonatite
	(Hornblendite
Metamorphism	
	(Gabbro anorthosite

Ultramafic	(Olivine meta-gabbro
Sequence	(Pyroxene Peridotite
	(Dunite, dykes in gabbro and gneiss
Metamorphism	
	(Syenite gneiss-syenite
Metamorphic	(Metalimestone
Complex	(Granulite
	(Gabbro gneiss

Oldest

In the gabbro of the ultramafic sequence Oosterom records many of the layering structures which Krauskopf was unable to reconcile with an igneous origin. For example, there appears to be no regularity in the upward transition of melanocratic to leucocratic gabbro, or any regularity in the thickness of successive layers. He says that the body has many features in common with layered intrusions of the cryptic type. Monomineralic bands he explains by diffusion of interstitial liquid between earlier-formed crystals. The centimetre scale banding, which is analagous to that found by Hess at Stillwater, Oosterom believed has a tectonic origin. He noted the absence of contact metamorphism in the Lille-Kufjord gabbro on Seiland which is part of this suite. This he explains by the fact that any contact aerole that developed, was obscured by later dynamo-thermal metamorphic processes or alternatively that the temperature differences between the host and the intrusive were negligible.

Regarding the metamorphic complex Oosterom agrees with Krauskopf that the calc-silicate bands which occur in the gabbro-

gneiss are evidence against an igneous origin for the latter.
However, he points out that Krauskopfs Gabbro-Gneiss II is the same
as his layered gabbro, which he regards as being intrusive. However,
concerning the other gabbro gneisses, Oosterom states "that most
mafic rocks could very well be of igneous origin from the mineral-
ogical and chemical point of view, but in the regions where the
calc-silicate lenses, acidgranulite and syenite gneiss layers occur
within the gabbro gneiss, Krauskopf's hypothesis of a gabbroization
of an original bedded volcanic sequence seems inescapable".

Thus, although the magmatic origin of many of the later intrusions
has been demonstrated, ideas concerning the origin of the early gabbro-
gneisses are dominated by the notion that they were formed by extreme
metamorphism of a sequence of bedded lavas, tuffs and sediments.

The work that has been carried out in the Loppen district by
Ball et. al (1963) and by Ramsay and Sturt, Roberts and Speedyman on
Sørøy and more recently on Seiland is very important in an understanding
of the environment in which these basic bodies occur. This is so because
it is in these areas that other intrusive rocks, which may be the temporal
equivalents of the Stjernøy Øksfjord suites, can be related quite
definitely to the various phases of deformation which have occurred in
the region.

In the Loppen area Ball et. al. (1963) have recognised a number
of intrusive basic rocks. In particular, a hypersthene gabbro occurs
as a sheet concordantly interlayered with the country rocks and shows
a marginal chilled-facies. This body transgresses F.1. folds and has
a foliation parallel to the axial planes of F.2 folds. A granulated

hypersthene gabbro is believed to have been formed by penetrative
deformation of the sub-ophitic hyperthene gabbro, since a textural
gradation is apparent between the two facies. It is considered by
the present author that this phase of deformation may be coeval with
a similar phase recognised on Eastern Seiland (see P.220). An olivine-
gabbro has also been recognised which is tentatively related to ultra-
basic rocks occurring as lenses and dykes in the area.

An interesting feature of the deformation pattern shown in this
area is the existence of a phase of folding between the F.1. and F.2
phases recognised on Sørøy, (Ramsay and Sturt 1963, Roberts 1965, 1968),
which has been designated F.2 in the Loppen area. This phase of folding
is only sporadically and weakly developed on Sørøy.

On western Sørøy two large gabbro masses have been recognised,
(Stumpfl and Sturt 1965). One is the Storelv gabbro which was
emplaced during the later part of the F.1 deformation and locally
shows a contact aureole with the surrounding metasediments, (Sturt and
Taylor 1971). The second is the Brevikbotn gabbro which also has a
contact aureole and cuts sharply through many F.1 structures, though
it bears a late F.1 foliation. This body is somewhat more variable
in composition than the Storelv gabbro and contains numerous limestone
xenoliths which have been converted to skarn. Both gabbros were
affected by the peak of the regional metamorphism.

Sturt (1969) has also mapped the Hasvik gabbro which was
intruded during the acme of the regional metamorphism and between
the two major phases of deformation F.1 and F.2. The gabbro, which

has a concordant sheet-like form, contains numerous xenoliths of sedimentary material and shows evidence of having wedged these off from the surrounding country rock. This is strongly suggestive of a mobile magma. The gabbro is also sporadically layered.

The work of Speedyman (1968) on the Husfjord area of south-eastern Sørøy is especially relevant to the problem of the calc-silicate bands seen in the gabbro-gneisses of Stjernøy. In the Husfjord area Speedyman has recognised a complex series of meta-morphic and intrusive events:

Youngest

      8)   Late basic dykes and nepheline syenite pegmatite

      7)   Perthosite

      6)   Slåtten and Vatna Gabbros

      5)   Emplacement of quartz diorite

Late (4)   Havnefjord Diorite

F.2 (3)   Kobberfjord Norite

F.2 (2)   Early Diorite intrusions into Husfjord meta gabbro

Late 1)   Husfjord gabbro
F.1

F.1

Oldest

The Husfjord metagabbro is of particular interest. This is a basic sheet which shows no regular layering, but ocassionally develops bands of troctolite. It was intruded into the upper limb of a large F.1 nappe structure and thus has a slightly discordant relationship to the limbs of the fold but is strongly discordant at the hinge. The intrusion contains many metasedimentary inclusions especially metalimestone rafts which have suffered contact metamorphism. An

important feature is that the limestone rafts do not seem to have been greatly disturbed from their original position. This is indicated by the fact that their strike is parallel with that of a limestone belt which lies outside the complex. The term stromatolith has been proposed by Foye (1916) for this sort of complex consisting of many alternating layers of igneous and sedimentary material in a sill-like relationship.

Speedyman (op.cit.) postulates that before the intrusion of the fairly homogeneous Havnefjord Diorite, a lens of meta-sediment existed in the Husfjord metagabbro complex. Compression along the layers of this body caused a tendency for the layering to be wedged apart allowing permissive intrusion of the Havnefjord Diorite. Meta-limestone inclusions are also found in this body, again preserving the trend of the external stratigraphy. This testifies to very permissive intrusion of a very fluid magma. There is also ample evidence of contact metamorphism and other intrusive relationships associated with this body.

Thus the work of Speedyman appears to indicate that an interbanded metalimestone gabbro complex can be produced by a process of permissive intrusion. It is probable that the calc-silicate in the gabbro gneisses on Stjernøy are of similar origin but have possibly undergone more intense flattening.

Thus the work on Eastern Seiland and Sørøy is of particular importance. In these areas the temporal relationships between the earlier intrusions and the metamorphic and structural events can be clearly demonstrated. Moreover these intrusions may well be the equivalent of the gabbro gneisses of Krauskopf (1954) Oosterom

(1954 1963) and Heier (1961). In this respect the formations of

Eastern Seiland are particularly important for they constitute the meta-

sedimentary envelope to the huge basic and ultrabasic plutons of

Western Seiland which can be traced to the neighbouring island of

Stjernøy. The Hønseby and Hammeren gabbros, the latter of which

has a stromatolith-like form, can be shown on Eastern Seiland to

have been intruded late in the local F.1 deformation. These bodies

can be traced westwards towards the plutons of Western Seiland,

where they become hornfelsed (B.A. Sturt personal communication).

Similarly the Lille Kufjord gabbro and the Melkvann ultramafite

body of South Seiland (B. Robbins personal communication) have

clearly been intruded after the local F.2 phase of deformation.

Thus the elucidation of the structural metamorphic and igneous

history of Eastern Seiland, which is the subject of this thesis, is

highly relevant to the understanding of the basic and ultrabasic

plutons of Seiland and Stjernøy.

It has been suggested that the basic rocks of the Seiland

Province are comagmatic. This however seems rather unlikely since

the intrusions span a considerable period of time, during which

there were a number of structural and metamorphic events. Age

dates on nepheline bearing rocks, which post-date the local F.2

fold phase on Seiland and Sørøy, give a late Cambrian to early

Ordovician age for the later phases of the metamorphism. On the

island of Magerøy, to the north of Seiland there is a gabbro body

which intrudes fossiliferous Silurian rocks. Thus a very considerable

period of time is involved in the intrusion history of the Seiland

Province; this would seem to rule out any ideas concerning the

comagmatic nature of these bodies.

C) Geomorphology

Physiologically the island is in sharp contrast to the rather rolling upland or 'vidda' of the mainland; a difference that is in some measure a reflection of the geology.

The western part of the island is made up of a complex of basic and ultra-basic intrusions. Here occur the greatest elevations of just over 1,000 metres. The area also has two permanent ice-caps both of which show much evidence of wastage in recent times. The eastern and largest, Seilandsjøkelen, covers about 30 km^2, while the western Normandsfjelljøkelen covers about 15 km^2 (Barth 1953).

Eastern Seiland consists largely of the meta-sedimentary envelope to these intrusions; a general westward sheet dip and variable resistance to results in a step-like profile of north-easterly facing scarps with south-westerly 'dip slopes'. Where dip slopes meet scarps a line of lakes is often seen.

Drainage

In general the rivers follow an approximate NW - SE trend of glacially modified valleys. There are some deviations from this pattern.

For example, direct breaching of an escarpment by glacial action is evident where SW of Vasbugtvann, the river, after following the normal SE - NW trend, suddenly takes a sharp turn to the NE to flow through a steep-sided breach in the escarpment of the Olderbugten group.

The profile of many rivers is suddenly sharpened by a break in

slope often marked by a waterfall. These breaks in slope are un-
doubtedly knick-points caused by post-glacial rejuvenation.

Lakes are common in the area. They range from small rock hollows
to lakes of considerable dimensions, such as Trollvann which is 4 km.
long. They also appear to vary considerably in depth. Trollvann
itself does not appear to be very deep, though the large lake above
Olderfjord shelves off very steeply from its shores.

Lakes occur in three main settings:-

a) Rock hollows related to rocks with structural fractures or fissility,
   e.g. the schists of the Olderbugten Group which abounds in small lakes.

b) Belts of major structural weakness e.g. the fault running NW - SE
   just NE of Hønsebyfjord which has been etched out by erosion into a
   long valley marked by a series of small lakes.

c) Larger hollows excavated by glacial action. Trollvann is an example.
   In the scarp west of the lake there are a numberof cirque-like features
   which were undoubtedly the source of some minor glaciers. These must
   have been responsible for the excavation of the hollow now occupied by
   Trollvann. Many lakes of this type are floored with rounded boulders
   and cobbles which appear to have been smoothed in the soles of glaciers.

Glacial Phenomena

There is ample evidence of the action of ice on Seiland. The fjords
generally show the typical 'u' profile of glacially deepened valleys, e.g.
Hønsebyfjord and Eidvaagenfjord. Ice striations in both fjords seem to
indicate an approximately north-westerly direction of ice transport.
Outcrops of the schists in Eidvaagenfjord show much evidence of polishing
as do the outcrops of gabbro in Hønsebyfjord.

Plate 1. View looking south-east down Vargsund from the
          southern end of the area. Note the raised beaches.

There is considerable evidence of post-glacial uplift both in the occurrence of knick-points and in the presence of numerous raised beach levels. On the mainland a strand line at approximately 60 metres A.M.S.L. can be clearly seen in some steep cliffs. On Eastern Seiland there is not so much evidence of this level though two other lower levels are clearly seen in wave-cut notches at approximately 10 metres and 14 metres (Plate 1).

Of further interest is the existence on the mainland of a very extensive plateau or 'vidda' with summit heights of some 5-600 metres A.M.S.L. The average heights of the mountains on E. Seiland are of this order. Here the summit tops tend to be flat plateau-like features covered in frost-heaved 'felsenmeer'. It is only in Western Seiland that the mountains are higher, up to 1000 metres but probably averaging 700-800 metres. Here the rocks are the more resistant gabbro and ultra-basic types. It is probable that these concordant summits represent a pre-glaciation erosion surface that has been strongly dissected.

Human Geography

Settlement on Seiland is confined to the coast-line. There are numerous small hamlets, the largest of which is Hønseby in the NW. The economy is based mainly on fishing. Each family has a small boat which is used to catch cod, coalfish and in the summer, salmon. The cod and coalfish are usually dried on racks out in the open air and sold to West African countries as a source of protein. The highest quality fish goes to Italy.

Land use is generally confined to growing hay and very occasionally potatoes, to support cattle and sheep during the winter, when the sun is not seen from November to February. During the summer it is light from May to mid-August.

Plate 2. Panorama looking approximately north-east showing Hønsebyfjord and Hønsebyfjell.

CHAPTER 2

Stratigraphy

A) Regional Aspects

According to Holtedahl's Geology of Norway the sedimentary rocks of Finnmark are chiefly of psammitic type and can, on good grounds be correlated with the Sparagmites of Southern Norway. Generally, these rocks can be divided into two groups:

1) The strongly deformed and metamorphosed Caledonides in the W and NW which have been thrust over

2) The relatively unmetamorphosed rocks in the S and E.

On the neighbouring island of Sørøy Sturt and Ramsey (1963), Roberts (1965, 1967, 1968) and Speedyman (1968) have recognised a conformable sequence of metasediments:-

                    Hellefjord Schist Group

                    Aafjord Pelite Group

                    Falkenes Marble Group

                    Storelv Schist Group

              (Transitional Group

              (Upper Psammite Group

Klubben   (Upper Pelite
Psammite
Group   (Middle Psammite

              (Lower Pelite

              (Lower Psammite

The strong deformation and high grades of metamorphism indicated by these rocks generally precludes the preservation of fossils.

However, specimens of Archaeocyatha were found in the Falkenes Marble
by Dr. B.A. Sturt.

These fossils were identified by Professor C.H. Holland as being
of a type restricted to the upper Lower Cambrian or lower Middle Cambrian
(Holland and Sturt 1970).

In Eastern Finnmark especially in the Digermul Peninsula area,
Reading (1965) and Reading and Walker (1966) have recognised a sequence
of slates ortho-quartzites and greywacke sandstones. There are two
tillite Horizons, at the base.

Reading divides these sediments into two groups:-

      a) The Digermul Group

      b) The Vestatana Group

The rocks of the Vestatana Group are entirely Eo Cambrian in age.
Only the lower part of the Digermul Group, however, is of comparable
age. 220 metres from the base the group passes up into the Cambrian.
The top of the group is Tremadocian and fossils occur at several horizons.

The conditions of sedimentation indicated by these rocks, are
according to Reading, shallow water. These conditions deepened temp-
orarily to allow sedimentation by turbidity currents.

On palaeontological grounds it seems that the Falkenes Marble
Group on Sørøy can be correlated with the lower part of the Digermul
Group of Reading. The Falkenes Marble Group also shows indications of
having accumulated in a shallow water environment, as does most of the
succession on Sørøy. Roberts (1968) puts forward evidence that the
Hellefjord Schist Group on Sørøy is a turbidite formation. This suggests
deepening of the basin of sedimentation. He regards the group as being
the last stage in the local preorogenic, geosynclinal depositional cycle.

Thus it appears that up to the end of Falkenes Marble times a shallow shelf sea, subject to intermittent tectonic instability existed in the area now occupied by Sørøy. The Hellefjord Schist lithology represents the deepening of the geosynclinal basin, heralding the oncoming orogeny.

Roberts (1968) claims that the succession seen on Sørøy, can be correlated with the Loppen Andsnes area, some 50 KM to the SW. · He therefore suggests that conditions of sedimentation in the geosynclinal basin were relatively uniform over a wide area.

Further, Sturt, Miller and Fitch (1967) determined the age of nepheline and biotite concentrates from some alkaline rocks on Sørøy (one specimen from Seiland) which post-date the major fold episodes. The determinations indicate an early Ordovician age for the latter phases of the metamorphism (480-491 my). Further work was carried out by Pringle and Sturt (1969) on an anatectic vein associated with the Hasvik gabbro on South Western Sørøy. This was intruded during the peak of the regional metamorphism. The results indicate an age of 520 ± 35my. Thus, the main structural and metamorphic events in the Sørøy area occurred during Upper Cambrian to early Ordovician times. The sediments, therefore, are probably of Eo-Cambrian to Upper-Cambrian age.

B) Stratigraphy of Seiland

1) General

On Eastern Seiland a conformable sequence of metasedimentary rocks can be recognised. The major lithological types include quartzites and psammites which are sandwiched between thick pelitic units (Fig2). In the absence of sedimentary structures it had not been possible to demonstrate the order of deposition, the succession discussed is,

FIG 2

GENERALISED STRATIGRAPHY

FLAGGY PSMTE-QZT-THIN CALC-SILICATES

SEMI-PELITE

OLDERFJORD GROUP

QZT BANDS

PURPLE SCHIST

OLDERBUGTEN GROUP

PSAMMITES

TROLLVAN PSAMMITE GROUP

QZT BANDS

BROWN SCHIST

EIDVÅGEID SCHIST GROUP

SCHIST BANDS

FLAGGY PSMTE-QZT

UPPER

PINK PSMTE
INTERBEDDED PSMTE-SCHIST
CALC-SILICATE SCHIST

KOMAGNES GROUP

FLAGGY PSMTE

LOWER

SCHIST BANDS
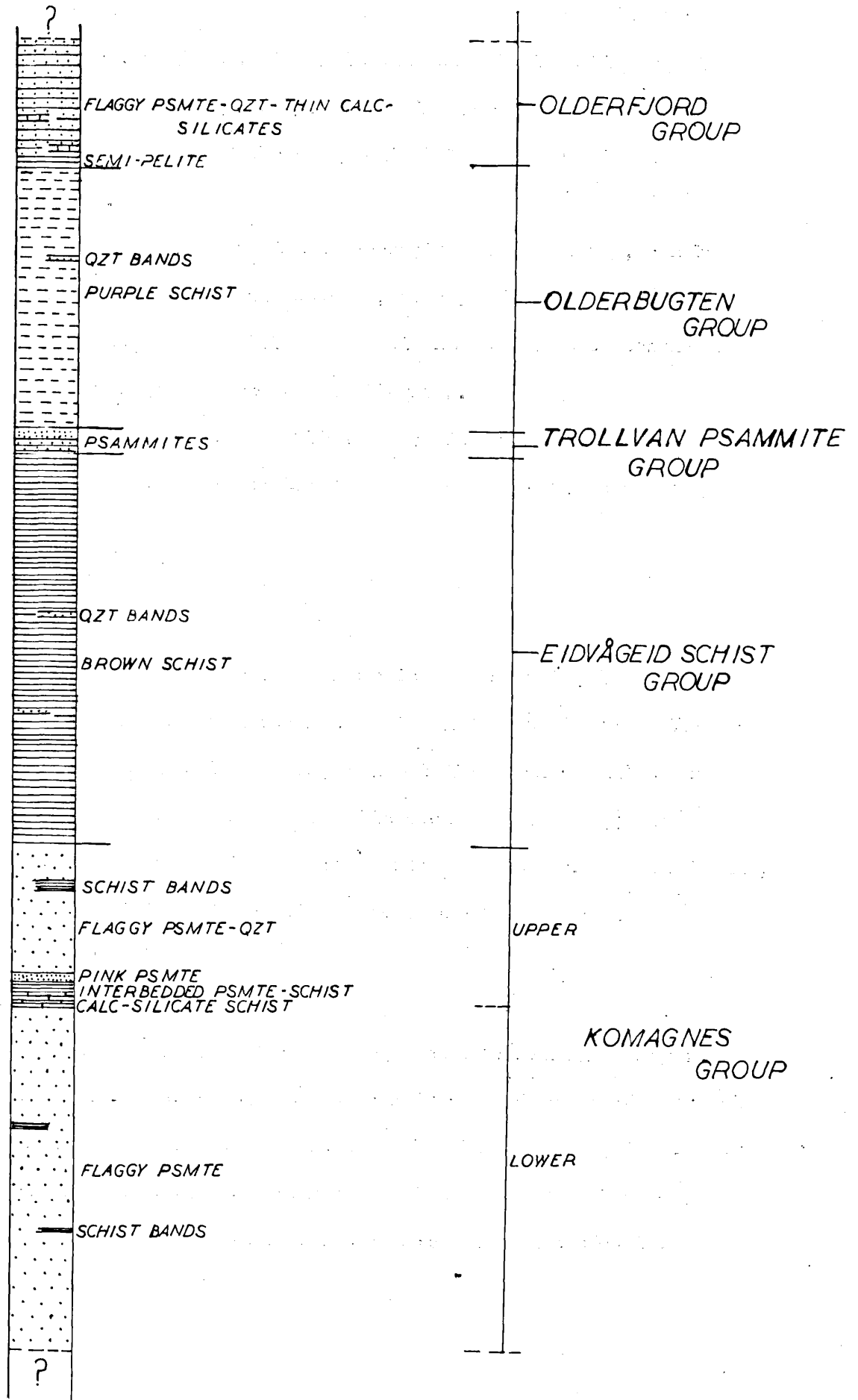
Plate 3. View looking south from the tip of Eidvaagtinn.
Trollvann is in the left centre of the picture.

Plate 4. F_2 fold in flaggy psammites at Komagnes.

therefore, a structural one.

The stratigraphy of these metasedimentary rocks is superficially rather simple, at least in terms of the variety of lithological types. This simplicity is,however, in part due to the intense extensional deformation to which the rocks have been subjected. This type of deformation, is largely responsible for the lateral discontinuity of many of the metasedimentary horizons. In this event original thickness estimations become highly conjectural, if not meaningless.

In the western part of the area, difficulty has been encountered in tracing some of the major lithological boundaries. This is largely due to the metamorphic convergence produced by migmatisation and granitisation.

There appears to be some discrepancy in the literature concerning the usage of terms to describe metamorphosed sandstones. For the purposes of this account, the following scheme is adopted:

1) Quartzite

   a pure massive meta-sandstone

2) Psammite

   a more micaceous and therefore flaggy-sandstone

3) Semi-pelite

   a metamorphosed impure sandstone, generally distinguished on Eastern Seiland by the darker colour imparted by relative abundance of phyllosilicates and amphiboles

2) The Komagnes Group

   a) The Lower Flaggy Psammites

   The structurally lowest member of the metasedimentary succession is well exposed on the coast to the north of Komagnes. It is a complex

of grey flaggy psammites. These rocks show evidence of considerable deformation by the second fold phase.

F.1. isoclines can occasionally be seen lying within the flaggy layering of these rocks (plate 5). They are particularly well-seen in the more quartzitic horizons. They presumably fold the original sedimentary banding. The enclosing micaceous psammites have a strong schistosity which is axial-planar to the folds. This flaggy layering is therefore, interpreted as an axial-plane structure to the early folds (See Fig. 4 (1) after page 43).

These early folds also have a coarse rib-like lineation parallel to their axes. This lination is also strongly developed in the layering surfaces of the psammites. Thin section examination reveals it to be formed by coarse stringers of quartz (See Plate 149 after P. 177).

Within the psammites occur a number of schistose pelite bands (Plate 7), which are never thick enough to be mappable. Their lateral discontinuity is probably due to tectonic sliding. Generally they contain strongly linated amphibole crystals which impart a dark greenish-black aspect to the rock. Other paler green varieties are actinolitic.

There are also a number of dark coloured, strongly schistose basic sheets within the psammites. The basic material appears to have been intruded along the early schistosity. These bodies were subsequently strongly foliated parallel to their margins, amphibolitised and boudinaged.

The rocks on the peninsula at Komagnes are separated from the rocks to the west by a thrust. The thrusting has produced a zone of flagginess in the psammites. To the east of this structure, the rocks are, in

Plate 5. F_1 isocline deformed by later folds, Komagnes.

Plate 6. F_1 schistosity in a pelite band.

general, similar to those to the west. On the north-east coast of the
peninsula, however, the psammites have been intensely felspathised.
This felspathisation imparts to the rocks a very massive granitoid
aspect (See Plates 188, 189 and discussion P. 2o8).

b) The Calc-Silicate Schist

This horizon can be seen in beach exposures in the Jernelv area.
It can be traced for some considerable distance in a north-westerly
direction, until its outcrop thins rapidly and is finally lost on the
fjell above Russelven. This is attributed to tectonic sliding.

In general the lithology is not very well exposed compared with
the enclosing psammites. Typically it is a dense fine-grained dark
green schist. On close examination the rock can be seen to be quite finely
schistose. Fresh surfaces are slightly lustrous, owing to the presence
of very small actinolite amphibole crystals, which define a lineation.
Despite the fine nature of the schistose laminae, when viewed from a
distance the rock often has a somewhat flaggy appearance. Frequently
quartz segregations occur along the layering.

Thin section examination of this lithology indicates that it
consists in general of an actinolite-chlorite-biotite assemblage.
Occasionally however, with epidote and calcite are present. The
presence of these minerals indicates a high lime content, but some
facies, which are calcite free, may more correctly be termed basic
schists. Generally, the rock type is intensely deformed internally,
in a disharmonic fashion relative to the psammites. Both second fold
phase structures and by a series of later chevron-type folds are
present. The latter folds are expressed in other lithologies as
kinks.

Plate 7. Pelite band in the Lower Komagnes Group, Komagnes.

Plate 8. Contact of the calc silicate schist with the
underlying psammites, north of Jernelv.

The contact of this lithology with the psammite group structurally below is very sharp. There is no evidence of any sedimentary trasition between them. It seems likely that this contact is a tectonic one (plate 8).

c) Mixed Pelitic and Psammitic Unit

The lithology which occurs structurally above the calcsilicate schist is rather variable. It consists mainly of a close interbanding of highly schistose pink psammites with dark biotite and amphibole schists. There are also some quite pure quartzite bands. These often display intense boundinage. There are also a number of dense, dark amphibolite sheets which are strongly foliated. The original basic material was obviously intruded along the schistose layering.

The thickness of the layering is generally quite variable. The laminae are usually thin, of the order of 2 - 4 cms. The layering has been very strongly deformed by the second phase of folding and there is considerable variation in thickness of the more competent psammite bands as they are traced round fold closures.

The pink colour of the psammite bands is imparted by the presence of pink porphyroblasts. These are frequently strongly augened. This porphyroblastesis is also in evidence in the schists horizons, where the augening is even stronger. In areas where this phenomenon is more intensely developed a rather coarse massive granitoid is produced. The porphyroblasts may be anything up to 3 cms. in diameter; this however is rare. The rock generally appears as a pink schistose psammite studded with small pophyroblasts. Thin-section study of these porphyroblasts indicates that they are albitie

Plate 9. F$_2$ folds in the Mixed Psammite and Pelite Unit, Jernelv.

Plate 10. F$_2$ folds in the Pink Psammite, Jernelv.

in composition (An 5). They contain large numbers of small inclusions of both muscovite and epidote. It seems doubtful that these minerals could impart to the felspar its rather uniform pink colour. It may well be that the colour is due to included haematite.

Towards the structural top of this unit the pelitic members become of less importance. The rocks become more uniform flaggy pink psammites (Plate 9).

Reference to the map indicates that both the Calc-Silicate Schist and the Pink Psammite, which respectively underlie and overlie this unit, are laterally discontinuous. The absence of these two marker horizons precludes the possibility of mapping the Mixed Pelitic and Psammitic unit further along the strike.

d) The Pink Psammite

The greater abundance of sandy material occurring towards the top of the unit discussed above has led to the development of a mappable pink psammite unit. This is a rather flaggy lithology consisting chiefly of quite pure pinkish quartzite bands approximately 2-5cms. in thickness (Plate 10). These are interbedded with thin slightly more pelitic horizons which appear to adopt in places an F.2 foliation. This is inclined at an angle to the quartzite layering, a relationship that frequently results in a rather hackly lineation on some layering surfaces (Plate 10).

The rocks are in places rather intensely deformed by the second phase folds; a structure of this age can be mapped on the fjell just north of Rastabynes. The pink colour does not appear to have been imparted by any felspar porphyroblastesis, as in the unit below. It is probably due to the original haematite-rich nature of the

sediment.

e)   The Upper Flaggy Psammites

The lithology occurring above the pink psammite unit is a complex of rock types which show a general tendency towards an increase of pelitic material towards the structural top of the succession.

Directly above the pink psammite, a series of whitish flaggy micaceous psammitic rocks occur.   These pass upwards into rather more massive psammites, which are grey in aspect which are interlayered with dense black foliated amphibolite of igneous origin.   Structurally upwards a series of grey flaggy psammitic rock occur, which display, in beach exposures at Rastaby, some rather beautiful fold structure of both generations.   The psammites which are studded with small pinkish garnets, bear a very strong lineation.   This is apparently due to microfolding during the second fold phase.

These lithologies give way to a complex of finely interlayered quartzites, psammites, semi-pelites and pelites.   Although the effects of deformation cannot be gauged, the following is quoted as an example of the scale of this banding in a single specimen.

1.   Grey semi-pelite average thickness of bands   5mm.

Intercalated with pelite average thickness   3mm.

2.   Black Garnetiferous pelite   2cms.

3.   White banded quartzite with some semi-pelite 2cms.

4.   Black garnetiferous pelite with grey semi-
pelite 2cms.

5.   White quartzite   8mm.

This type of sedimentary intercalation often coarsens considerably giving rise to moderately thick banded flaggy quartzites and brown schists horizons.   Some of these are thick enough to map.   The

Plate 11. F2 folds in the flaggy quartzites of the
Upper Komagnes Group, Rastaby Beach.

Plate 12. Basic sheets showing multiple folding, Rastaby.

quartzites are often deformed into rather beautiful folds, which tend to have a parallel geometry (Plate 11=) The flags are of the order of 5cms. thick and their surfaces are frequently marked by a hackly lineation. The interlayered pelite horizons are generally strongly schistose, dark in colour and studded with pink garnets. Very thin (2mm) quartzite bands are deformed into little puckers again with a tendency towards a parallel geometry. A strain-slip cleavage in the enclosing schists is axial-planar to these little folds. On layering surfaces the folds often appear as periclinal structures, with strongly sinuous axes.

Near the contact with the overlying Eidvågeid schist Group a rather persistent horizon of schistose psammite occurs which is rather closely interlayered with a dense dark amphibolite; this amphibolite shows evidence of having been deformed by both F.1 and F.2 (Plate 12).

Thin section examination reveals the rock to consist almost entirely of hornblendic amphibole, felspar and some sphene; quartz is apparently absent. This assemblage is characteristic of metamorphosed basic igneous material. These sheets may well be of intrusive origin. Due to the persistence of their outcrop however, the possibility that they are metamorphosed basic tuff horizons cannot be discounted. This very early position in the chronology is confirmed by the fact that they have undergone deformation by both fold phases.

A Series of psammitic rocks are exposed on the peninsula at Eidvågeid. They are apparently in a similar structural position relative to the Eidvågeid Schist as the upper psammities at Rastaby. Unfortunately at Eidvågeid the contact between the two Groups is not

Plate 13. Folded psammites on the peninsula at Eidvågeid.

Plate 14. F₁ folds in psammites at Eidvågeid.

exposed. These rocks are cosiderably more massive and pure than those in a similar structural position at Rastaby (Plates 13 and 14). They do not show the intimate interlayering of psammite, semi-pelite and pelite, characteristic of the Rastaby types. They are flaggy pink or white micaceous psammites, more akin to the rocks at Komagnes. Felspathisation has locally imparted to the rocks asomewhat massive pink appearance. There are also bands of a brown semi-pelite which are of appable dimensions.

They do however, have one feature in common with the rocks at Rastaby. This is the presence of dark amphibolite bands (Plate 15) which are again apparently folded by both F1 and F2. The fact that these bands occur in a similar stratigraphic position relative to the Eidvågeid Schist as the amphibolites at Rastaby lends some credence to the idea that there are metamorphosed basic tuffs.

3) The Eidvågeid Schist Group

Occuring structurally above the Komagnes group is a sequence of red-brown weathering schists. The outcrop of these rocks can be traced for practically the whole length of the area. It is generally $2\frac{1}{2}$ - 3 Km. in width. Geomorphologically, the schists form a rather monotonous rolling moorland, with summit heights of 200 - 250 metres.

Thickness estimations for this Group must be highly conjectural since the schists show evidence of very considerable flattening. In addition the intercalated quartzite bands suggest the presence of an early isoclinal fold in the lithology. In general, the schist does note seem to deform into folds in response to F.2. except in the presemce of basic sheets. These apparently serve to stiffen up the

Plate 15. F$_2$ folds in psammite, Eidvågeid peninsula.

Plate 16. Eidvågeid schist showing boudinage.

rock sufficiently to allow fold formation.

Study of the early folds in this lithology indicates that the foliation in the schist is axial-planar to them. It is deduced therefore, that the schistosity of the rock, which dips rather constantly west at some 25 - 30°, is an axial plane structure related to early recumbent folds. Where F.2 folds are seen they fold this schistosity.

The Eidvågeid Schist is in general a rather monotonous flaggy schist which weathers to a rusty or buff-brown colour. Within the schist there are a number of pale coloured quartzite horizons. These are laterally discontinuous. On fresh surfaces the schist is purple in colour and contains variable amounts of porphyroblastic kyanite, garnet and felspar.

It has been possible in the field to recognise four lithological types within the Group:

1) The structurally lowest unit of the schist is a buff-coloured semi-pelitic schist. On schistosity surfaces the rock has a lustrous brown appearance. The most distinctive characteristic of the rock is the absence of porphyroblastic minerals.

2) On the south eastern shore of Eidvaagenfjord the schist has a massive blocky appearance. The rock is studded with small garnet and kyanite crystals. This facies of the schist is rather more quartz-rich, and may be termed a semi-pelitic schist.

3) The most highly pelitic facies of the schist is a dense purple rock. In response to weathering, rather smooth rounded surfaces are produced. Blockiness is characteristically absent. The rock type has a high proportion of porphyroblastic garnet, kyanite and felspar. The garnets are often lensoid and up to 1.5 cms. in length. The kyanite

crystals show a considerable variation in size. Examples up to 6 cms. in length have been recorded. The random orientation of their long axes of these crystals on schistosity surfaces, indicates that they have grown mimetically. In the field these crystals have a yellowish weathering crust, which when broken reveala white lustrous cleavage surfaces. Thin section examination indicates that the kyanite has been very largely converted to white mica. The third porphyroblastic mineral growing in the schist in felspar. The mineral occurs as small white crystals which tend to be equidimensional. All the porphyroblastics are strongly augened.

4) The massive quartzite bands form the chief variation of facies within the Group. They are always laterally discontinuous, it is probable that is in part due to tectonic agencies. It has been noted however, where exposure is good, that some of these horizons split up laterally into a number of thinner units intercalated with schist. This may well be an original sedimentary feature.

Lithologically, the rocks are white to cream massive quartzites with variable amounts of felspar. They are frequently banded. The bands are generally thin, of the order of $\frac{1}{2}$cm. and are purple to grey in colour. This colour is imparted by the relative abundance of biotite and pink garnet.

The thickness of these massive quartzites is very variable. An example along the coast of Vasbugt is $4\frac{1}{2}$ metres thick. It is proceeded structurally lower, by bands of quartzite 15 cms. thick intercalated with schists. Lower again, are quartzite bands 2 cms. thick, with intercalated schists. The thinnest bands may be 5 mms. thick, though of course the effect of deformation on their original thicknesses

Plate 17. Eidvågeid schist showing boudinage.

Plate 18. Basic material in the Eidvågeid schist.

is indeterminable.

The banding surfaces of these quartzites often bears a strong lination and occasionally tight intrafolial folds are seen lying within the banding (Plate 36 after P.43). These folds are interpretated as F.1. structures. The foliation in the schist is sensibly parallel to the banding surfaces of the quartzite. F.2. folds fold this banding and the foliation in the schist.

The stratigraphic significance of these various divisions of the pelitic facies of the Eidvågeid schist is not clear. Certainly no pattern has been mapped out indicating a regular disposition of the various sub-divisions. Indeed, in the field this would be very difficult to do. In sedimentary terms they quite clearly represent variations in the proportion of sandy and muddy material.

Finally, within the Eidvågeid Schist there are a number of dark coloured basic sheets. It would seem that the original basic material was intruded along the early schistosity. These sheets were foliated parallel to their margins by subsequent deformation. They have also undergone strong boudinage (Plates 16 and 17). Exposures of schistosity surfaces reveal that many of these boudinage are discoidal in shape. Occasionally the boudin pods show evidence of folding by F.2. (Plates 18).

4)   The Trollvann Psammite Group

_   The Trollvann Psammite Group forms a continuous outcrop that can be traced for the whole length of the area. This Group provides an interesting variation from the thick pelitic groups that sandwich it on both sides.

The outcrop of the psammites generally forms quite a sharp

Plate 19. Flaggy psammite in the Trollvann Psammite Group.

Plate 20. Psammite in the Trollvann Group.

feature above the rolling moorland of the Eidvågeid schist. The

vagaries of glacial erosion, however in the Trollvann area have

reduced the coutcrop to the level of the schist.

Lithologically the group can be divided into three units which

seem to be recognisable for the whole length of outcrop. The

boundaries between the groups are sonewhat transitional. The section

from the base to top (in the structural sense) however, indicates an

increase in purity and differentiation of the psammitic content

relative to the pelitic content.

<u>Structural Top</u>

3)	Buffish Banded Psammites	12 metres
2)	Grey Flaggy Psammite	12 metres
1)	Transition Group	3.5 metres

There is considerable evidence of folding within the lithology

especially in the unit 3) so the thicknesses, which are the maximum

seen, have little meaning in the sedimentary sense.

The transition Group represents a transition from the dominantly

pelitic facies of the Eidvågeid Schist into the psammitic facies of

the Trollvann Group. Lithologically the unit is a flaggy semi-pelite;

purple in colour. Garnet is ubiquitous. This lithology grades

upwards into the <u>Flaggy Grey Psammite Unit</u>, formed of rather impure dark

grey rocks which have a flaggy banding. The flags are generally of

the order of 2 cms. thick. Fresh surfaces often have a purple hue.

This lithology has been somewhat felspathised. Evidence of this can

be seen in the form of white porphyroblasts augened in the layering.

These porphyroblasts impart a rather knotty appearance to the layering

surfaces (Plate 19).

This lithology grades upwards by a progressive lightening in colour into a buff coloured rather _flaggy psammite_. This is closely interlayered with darker pelite units, these being of the order of 2 mm. - 2 cms. thick. This interlayering imparts to the rock a banded rather schistose appearance. (Plate 20).

Marked thickening of the psammitic units is occasionally responsible for the formation of quite thick saccharoidal quartzites. This lithology characteristically shows some well-developed fold structures of both generations. Occasionally tight early isoclines lying in the banding, can be seen to be refolded by the second generation more open structures.

5) The Olderbugten Group

This group is so-called because it occurs in good coastal exposures in the south of the area at Olderbugten. It may be traced for practically the whole length of the area to Eidvaagenfjord in the north. The boundary representing the structural top of the group has proved rather difficult to map especially in the central and northern parts of the area. This is largely due to complications produced by intense felspathisation and migmatisation.

These phenomena have also complicated the lithologies within the group to such an extent that only broad statements are possible about variations due to sedimentation. A more comprehensive description of the group will, therefore, appear in the section dealing with felspathisation and related phenomena.

Broadly speaking the group is dominantly pelitic. The most distinctive lithology is a rather massive fine grained purple rock, rather prominently jointed. The rock does not have a very well

Plate 21. Migmatised schists of the Olderbugten Group.

Plate 22. Migmatised schist of the Olderbugten Group.

developed schistosity. It has largely been obliterated by a strong annealing recrystallisation. In homogeneous lithologies this has resulted in the formation of a massive purple rock with a blocky jointing. Where the rock has been felspathised and subsequently deformed a streaky white foliation is produced (Plate 21).

Small pink garnets are ubiquitous in the group. They are generally up to about $\frac{1}{2}$ cm in size and are often in equidimensional. Fairly frequent occurrences in the more felspathised facies are large garnets often of the order of 2 cms in diameter. They are character-istically surrounded by a lensoid halo of felspathic material (Plate 22).

More psammitic horizons are found within the Group. They occur as thin ribs of flaggy quartzite interbanded with semi-pelitic material (Plate 23).

In the less felspathised facies, the incipient development of the felspathisation can be seen in the occurrence of white felspar crystals often showing carlsbad twinnin. All gradations occur between this type of rock and the other extreme, where the original sedimentary origin of the rock has been completely obliterated. Intermediate types have been recorded where the felspar porphyroblasts occur as layers in the schist. Fairly characteristic are the development of augened felspathic bodies with large garnets in the cores of the lenses. Saccharoidal pegmatites also occur (see photographs).

Generally the massive unfelspathised schist does not show any folds except where small basic sheets occur. These basic sheets which have apparently been intruded along the early foliation are strongly boudined. They occur as disjointed pods within the schist. This is also true in the case of granitic sheets within the schist. The

Plate 23. A more psammitic facies of the Olderbugten Group.

Plate 24. Migmatised Olderbugten schists, Olderbugten.

banding within the rock curves strongly around these bodies.

The more felspathised facies of the schist, on the other hand, shows much evidence of folding as well as boudinage (Plate 24).

6) The Olderfjord Group

The Olderfjord Group has been named after an uninhabited fjord lying at the mouth of the long north-south trending valley which marks the western boundary of the area. The Group characteristically forms a steep scarp in the south at Olderbugten and generally occupies the high ground of the area. This may be due however, in some measure to the included basic material. On the fjell between Olderbugten and Olderfjord a large second phase fold can be mapped out in the Group. On the mainland the limbs of this fold stand out clearly marked by the lighter-coloured psammitic rocks.

The comments made during discussion of the Olderbugten Group concerning complications caused by felspathisation, also apply in considerable measure to this Group. There is however, the added complication of an injection complex of basic material. This was introduced into the rocks during the latter part of the first deformation.

In addition, the rocks have locally been intensely mylonitised. The phase of deformation responsible for this was followed by a prolonged annealing recrystallisation which has made the original sedimentary affinities of the rock extremely difficult to recognise. In the north, to the east of the Hønesby area, due to a combination of migmatisation and mylonitisation the rocks are now migmatitic gneisses. These changes have made the accurate tracing of the lithological boundary between this Group and the Olderbugten Group

virtually impossible.

The Olderfjord Group is distinctly much more psammitic than the underlying Olderbugten Group, though within it a number of lithological types may be recognised.

At Olderbugten, where the contact between the Olderfjord Group and the pelitic Olderbugten Group is best seen, the structurally lowest lithologies are massive purple rocks which have been strongly deformed and annealed. They have semi-pelitic affinities and are in strong contrast to the pelitic lithologies structurally below. These rocks are generally flaggy in appearance, especially in inland exposures. They frequently contain large white felspar porphyroblasts which are strongly augened. They are also studded, often densely, with pink garnets. The strong recrystallisiation mentioned above locally imparts to the rocks a rather massive flinty appearance.

One of the interesting features of the Group is the ubiquitous occurrence of strongly deformed calc-silicate bands. These are always very thin and never occur as mappable units. They frequently have the form of intensely drawn-out isoclines. This effect is accentuated by a bilateral symmetry produced by metamorphic differentiation. The cores of the lenses are pink and garnet-rich, the selvages are green due to a high epidote and diopside content. Despite this resemblance to tight folds, they have almost certainly been produced by intense boudinage of the relatively incompetent calc-silicate material. Frequently the lenses are folded by F.2. (See Plate 97 after P. 17). In thin section these lenses contain an anorthitic plagioclase (An.71), diopsidic pyroxene, epidote, calcite and some scapolite, together with garnet.

Plate 25. Flaggy psammite of the Olderfjord Group.

Plate 26. Flaggy psammite in the Olderfjord Group,
between Olderfjord and Olderbugten.

The flaggy purple lithology described above, passes structurally upwards into a sequence of flaggy grey and purple psammites. These rocks are very well exposed on the beach between Olderbugten and Olderfjord (Plate 25). The rocks, which show a host of excellent minor structures, contain interbanded semi-pelite horizons which show evidence of greater susceptibility to felspathisation and mylonit- isation. Felspathisation takes place in the psammites preferentially along the layering and the felspathic segregations which results are frequently augened during subsequent deformation (Plate 26). Sheets of basic material which have been intruded along the banding are strongly boudinaged.

Occurring interbanded with these are more massive pale-coloured quartzites generally very intesely jointed, giving them a rather shattered appearance. They also frequently contain fine minor structures (Plate 27).

The psammites contain some garnet. In general the crystals are of the order of 1 mm. in size. Larger garnets however, occur sporadieally; examples of up to $1\frac{1}{2}$ cms. in size have been recorded. They are always strongly augened.

In the west of the area, the Olderfjord Group has been intruded by an injection complex. The psammites are interlayered to various degrees with sheets of basic material which appear to have gently wedged the layering apart. The boundaries between pure psammite and pure basic rock are nearly always gradational and marked by a decrease in the amount of included psammite or vice versa. Boundaries therefore have to be placed on the basis of the proportions of the basic rock psammite. This phenomenon can be seen on the larger scale
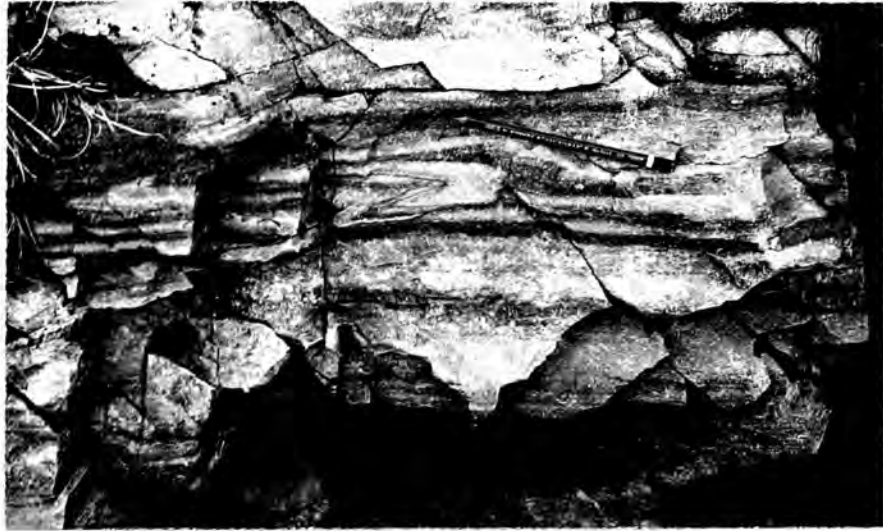
Plate 27. F2 folds in a quartzite of the Olderfjord
Group, Olderfjord beach.

Plate 28. Refolded F1 fold in the Upper Komagnes Group,
Rastaby beach.

by the inclusion of screens of psammitic rock in the basic rock
outcrop. (See Map).

Summary

The metasediments on Eastern Seiland comprise a sequence of thick
interbedded psammitic and pelitic Groups. The rocks have been sub-
jected to intense extensional deformation metamorphism and polyphase
recrystallisation. Estimates of original sedimentary thicknesses in
this event become highly conjectural. Simi larly sedimentary
structures which would have been of use in interpreting the sequence
and environment of deposition, have been obliterated.

A shallow water depositional environment is however suggested
by a consideration of the general facies of the rocks. The thick
psammitic and pelitic groups, together with the occasional lime-
rich horizons are consistent with this thesis.

Correlations with other Areas

Roberts (1968) has suggested a correlation of the metasediment-
ary sequence on Sørøy with that seen in the Loppen area 50Km to the
south-west (Ball et.al) (see Table 1). The bulk of the metasediments
in both areas show evidence of having accumulated under shallow water
conditions.

It has also been possible, on palaeontological grounds to correlate
the sequence on Sørøy with the series of unmetamorphosed sediments in
Eastern Finnmark. (Reading 1965, Roberts 1968). The sediments in
Eastern Finnmark in general also show evidence of deposition under
shallow water conditions.

Thus it would appear that the area now occupied by Finnmark was
covered with a shallow shelf-sea in Eo-Cambrian and early Cambrian

COMPARATIVE SEDIMENTARY SUCCESSIONS IN WEST FINNMARK

LOPPEN BALL ET AL 1967	S.W. SØRØY RAMSAY AND STURT 1970	E. SEILAND WORTHING 1971
	HELLEFJORD SCHIST GROUP	
PELITIC SCHISTS	AAFJORD PELITE GROUP	
CALCAREOUS SERIES WITH META-LSTN	FALKENES MARBLE GROUP	OLDERFJORD GROUP
PELITIC SCHIST		OLDERBUGTEN GROUP
IMPURE QZTS	STORELV SCHIST GROUP	TROLLVANN GROUP
		EIDVÅGEID SCHIST GROUP
QZTS AND AMPHIBOLITE	KLUBBEN PSAMMITE GROUP	KOMAGNES GROUP
FLAGGY QZTS		
MASSIVE AND/OR FOLDED QZTS		

TABLE I

times. Roberts (1968) suggests that sedimentary conditions in this basin where quite stable over a wide area. He bases this on the similarity in facies between the Loppen and Sørøy areas.

It is not possible to make a definitive correlation of the succession on Eastern Seiland with the other areas for two reasons; firstly, the younging direction of the Seiland succession is unknown and secondly, intense deformation has greatly distorted the original sedimentary relationships. Thus, attempts at correlation can only be based on the criterion of general facies similarity. On this basis, the correlation on Table 1 is suggested. It must be emphasised that this correlation is very tentative. The succession seen on Seiland is possibly older than the successions been on Sørøy and Loppen. It can hardly be younger since the Hellefjord Schist Group, the youngest member of the Sørøy succession, apparently accummulated under turbidite conditions. Roberts (1968) suggests therefore, that this development of turbidite conditions, heralded the oncoming orogeny and the cessation of sedimentation of shallow water sediments.

In conclusion it must be pointed out, that it is by no means certain that the structural relationships between the various Groups on Seiland are a reflection of the original stratigraphic relationships. The intense extension that the rocks have undergone, may mean that whole stratigraphic units may have been removed by sliding.

CHAPTER 3

Structural Aspects

Introduction

A discussion will now follow on the structure of Eastern Seiland. This will begin with a description of the minor structures together with deductions concerning possible modes of origin. This will be followed by a description of the major structures and the relationships between these and their associated minor structures.

a)   Minor Structures

There is a considerable variation in the style of minor structures across the area. This variation appears to be closely related to lithology. It is felt expedient therefore, to describe the minor structures within a stratigraphical framework.

1.   Fold Structures

Broadly speaking it has been possible to recognise two major phases of fold formation on Eastern Seiland. These are designated in these thesis as being F.1 for the earlier phase and F.2 for the later phase. Subsequent minor phases are designated F.3 etc. The two major fold phases were separated by an interval during which the maximum grade of metamorphism was achieved.

The style of the folds attributed to F.1 is uniformly isoclinal. This is not necessarily due entirely to any inherent property of the F.1 deformation, but partly to the very intense subsequent flattening which the rocks have undergone. In the west of the area where this flattening is most intense F.1 folds are very rare. They have mostly been obliterated.

The greater majority of the folds seen on Eastern Seiland are of F.2 age, the style of these folds is considerably more variable

the folds have been largely unaffected by any subsequent deformation.
They may, however, show a tight isoclinal style. This has lead in some
cases to a confusion in the temporal recognition of these structures.
The following criteria have, therefore, been adopted where style
considerations are confusing. F.2 folds are indicated:

a) By a refold relationship with F.1 (Plates 28, 29, 30, 31)

b) By the folding of a very strong foliation in the meta-
sediments, the latter being attributed to F.1.

a) F.1 Minor Folds

There is some evidence that there may be more than one phase of
fold formation within the first major period of deformation, the
evidence for this in the sediments of Eastern Seiland is by no means
conclusive. However, there are definite indications of the complex
nature of this phase of tectonic activity.

This evidence lies chiefly in the relationships seen especially
at the structural top of the Komagnes Group, between early amphibolite
sheets and the surrounding metasediments. Regarding these amphibolite
sheets, it was tentatively suggested in the preceding chapter that
they were metamorphised basic-tuff horizons. This was borne out by
the relative constancy of their stratigraphic position. This
explanation seems less likely in the case of similar amphibolite
sheets found sporadically within the Eidvågeid schist Group. Here
there is no such constancy of stratigraphic horizon or density of
occurrence. It seems more likely therefore that they are, in fact,
basic sheets which were intruded along the foliation in the schist.
The early nature of these basic sheets in the local geological history
is indicated by the fact that they have apparently been folded by both
F.1 and F.2 (See Plates 12, 32, 33). Thus, it is possible that folds

Plate 29. Refolded F_1 fold in the Upper Komagnes Group, north of Rastabyvann.

Plate 30. Refolded F_1 fold, flaggy psammite, Rastaby beach.

Plate 31. Refolded F$_1$ fold, Komagnes Group, east of Rastabyvann.

Plate 32. F$_1$ fold with warped axial plane, Eidvågeid peninsula.

Plate 33. Basic material showing two phases of folding,
Eidvågeid Schist Group, Vargsund.

Plate 34. F₁ fold in the Komagnes Group, above Vasbugt.

recognised as F.1 may, in fact, have been generated at slightly different times. The earlier phase giving rise to a foliation which acted as a plane of weakness for the intrusion of the early amphibolite sheets. These were subsequently folded by the second phase of folding within F.1. It is relevant that B. Robbins (personal communication), working in an area to the west has recorded two phases of folding within F.1 these separated by dyke intrusions. Thus the evidence from both areas may be summarised as follows:-

Basic tuffs in the sediments

$F1^1$ Foliation

Intrusion of Basic Material

$F1^2$ Foliation

However, it has not been possible to separate these two phases on Eastern Seiland. All early folds are therefore designated as F.1 though they may, in fact, belong to either $F1^1$ or $F1^2$.

Fold structures which, on the criteria mentioned above are of F.1 age fold a banding which is probably the original sedimentary banding, the absence of sedimentary structure precludes a definitive statement on this matter. In pelitic and semi-pelitic intercalations a strong schistosity develops which is in general parallel to this banding. This is interpreted as an axial plane structure related to the early folds. This foliation is designated as S1. * Intense flattening has resulted in the production of a flaggy banding with the early folds lying within it. This is refolded by F.2.

*   Note

In this thesis SS is taken as the bedding

             S1 is the axial plane schistosity to F.1

             S2 is the axial plane schistosity to F.2

FIG 3
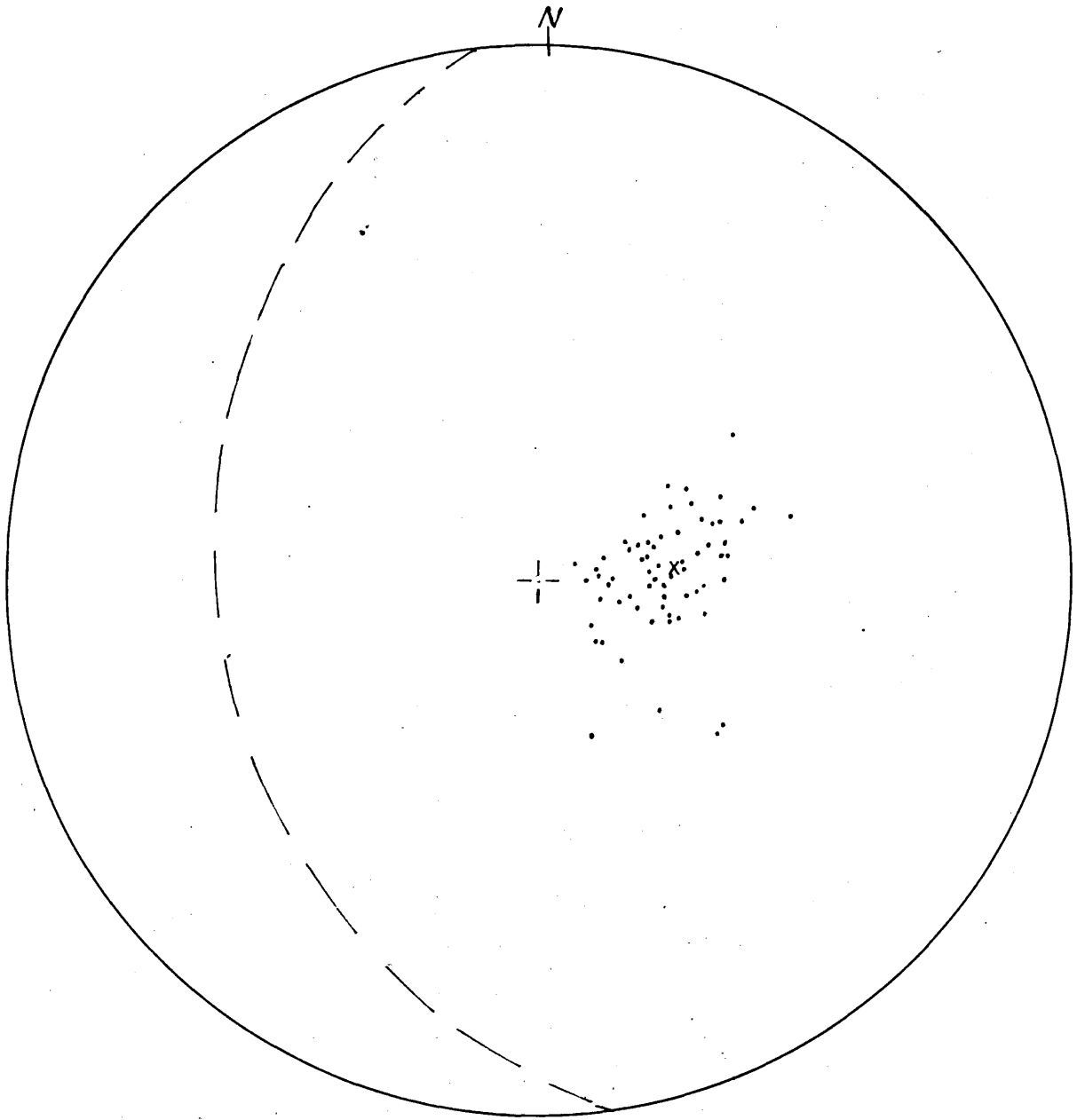
POLES   TO   SCHISTOSITY   IN   THE   EIDVÅGEID   SCHIST

ALL STRUCTURAL   STEREOGRAMS   ARE   PLOTTED   ON   THE
LOWER HEMISPHERE   OF A WULFF NET

In the Eidvågeid Schist Group a very small number of tight folds
have been recorded with the strong schistosity axial plane to them.
This schistosity is sensibly parallel to the banding of the massive
quartzite horizons. If poles to the schistosity planes (S.1) in the
schist are plotted on a stereogram a strong point maximum is obtained.
The best-fit great circle to this maximum has a strike of 352° and a
dip of $28^{\circ}W$ (Fig. 3).

The variability of form of the early folds within certain well-
defined limits, may partially be explained by their position relative
to other structural phenomena. In the west of the area they have been
very intensely flattened, in the east their morphology is more
dependant on their position relative to F.2 folds. Those occurring
on the long limbs of such folds have suffered more stretching than
these occurring on the short limbs. Thus, it is possible to say in
general, that the style of F.1 folds as observed in the field may bear
little resemblence to the original style of the F.1 deformation. The
following plates indicate the increase in the deformation of F.1 folds
from east to west.

Komagnes Group	Plates 34 and 35
Eidvågeid Schist Group	Plate 36
Trollvann Psammites	Plates 37 and 38
Olderfjord and Olderbugten Groups	Plates 39 and 40

Geometrically many of the folds approach a similar type geometry,
though they probably in reality mostly fall into Class 1C of Ramsays
classification (1967). Where flattening has been very intense, some
of the folds may be of Class 3, that is the curvature of the outer arc
exceeds that of the inner arc. The structures are therefore rootless
intrafolial type folds. Generally, however, thickening in the hinge and

Plate 35. F$_1$ fold, Komagnes Group, Eidvågeid peninsula.

Plate 36. F$_1$ fold in a massive quartzite in the Eidvågeid
Schist Group, Vasbugt.

Plate 37. F$_1$ fold in the Trollvann Psammite Group.

Plate 38. F$_1$ fold in the Trollvann Psammite Group, Vargsund.

Plate 39. F1 fold in the Olderbugten Group. Head of
Eidvaagenfjord.

Plate 40. Calc-silicate folded by F2 Olderfjord Group
east of head of Hønsebyvann.

FIG 4

Fi FOLDS FROM THE KOMAGNES GROUP

1

FLAGGY BANDING IN THE KOMAGNES GROUP

SEMI-PELITE

PSAMMITE

1.CM

2

GARNET

PSAMMITE

PELITE

1.CM
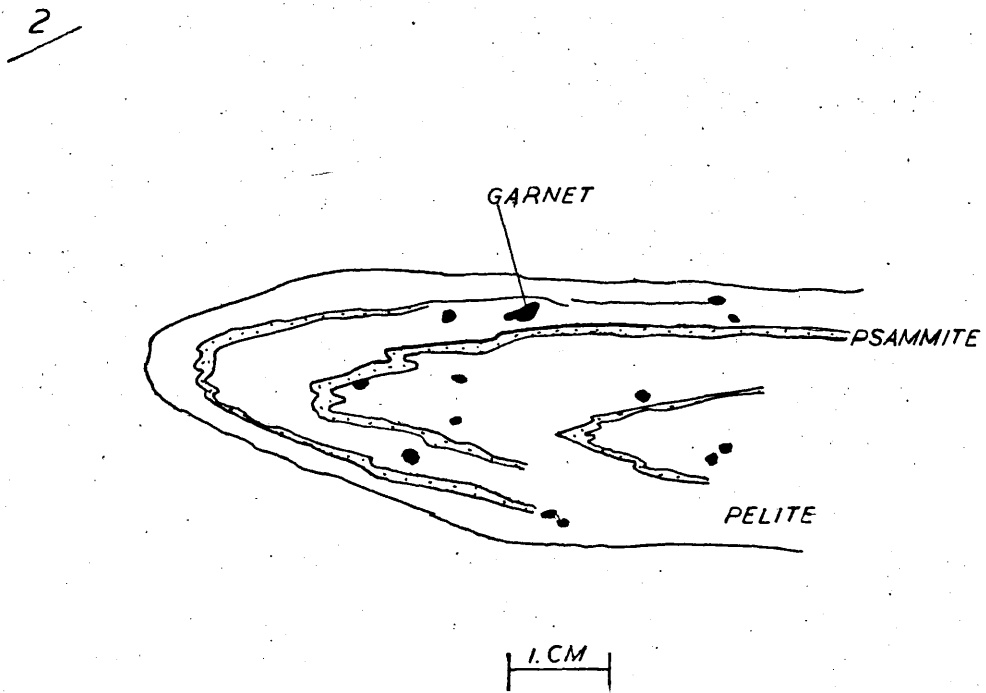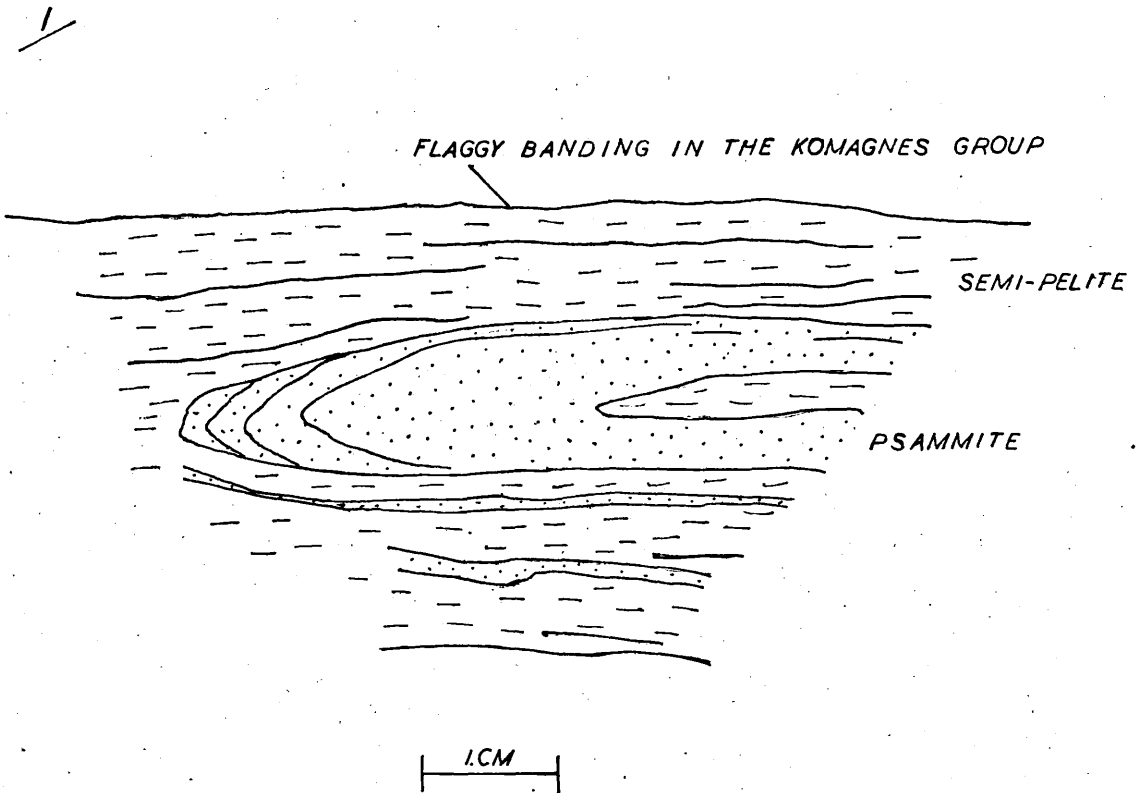
attenuation on the limbs are ubiquitous features (see Plates). This strong attentuation in S.1 seems to be at its maximum in interbanded pelite, semi-pelitic and psammitic lithologies (see Fig. 4). Due to the difficulty in distinguishing between the long and short limbs of the F.1 folds arising out of this attenuation, it has not been possible to work out the sense of vergence of the folds with some notable exceptions e.g. Plates 36 and 38.

There is some evidence to suggest that curvature of hinge lines develop as a primary feature of F.1 folds. It would seem that this tendency towards curvature is to some extent controlled by lithology. For example, in the lower part of the Komagnes Group where the psammites are often fairly massive. F.1 folds seem to be sensibly cylindroidal. This is to some degree confirmed by Fig.5A which tends to indicate a linear maximum rather than a planar distribution for the fold axes. This having a low plunge just west of north.

However, among the relatively few examples of F.1 folds which have been recorded there are a number of important examples of curvature of hinge line. A good example of this phenomenon occurs at the beach exposures at Rastaby in the south of the area. Here at the structural top of the Komagnes Group the rocks are interbanded pelites, semi-pelites and psammites. The specimen concerned is depicted in Fig.6 and Plate 41. It consists of two antiformal closures separated by a very tight synformal closure. Most of the infolded pelite has been squeezed out of this synform. The lower antiformal closure is almost cylindroidal and is associated with a quite coarse lineation which is almost coaxial with the fold hinge. The upper antiformal closure, however, is intensely curved. A chord drawn across the arc of the curvature i.e. indicating the average

FIG 5

PLOT OF F.I. FOLDS(○)AND LINEATIONS(•)

A
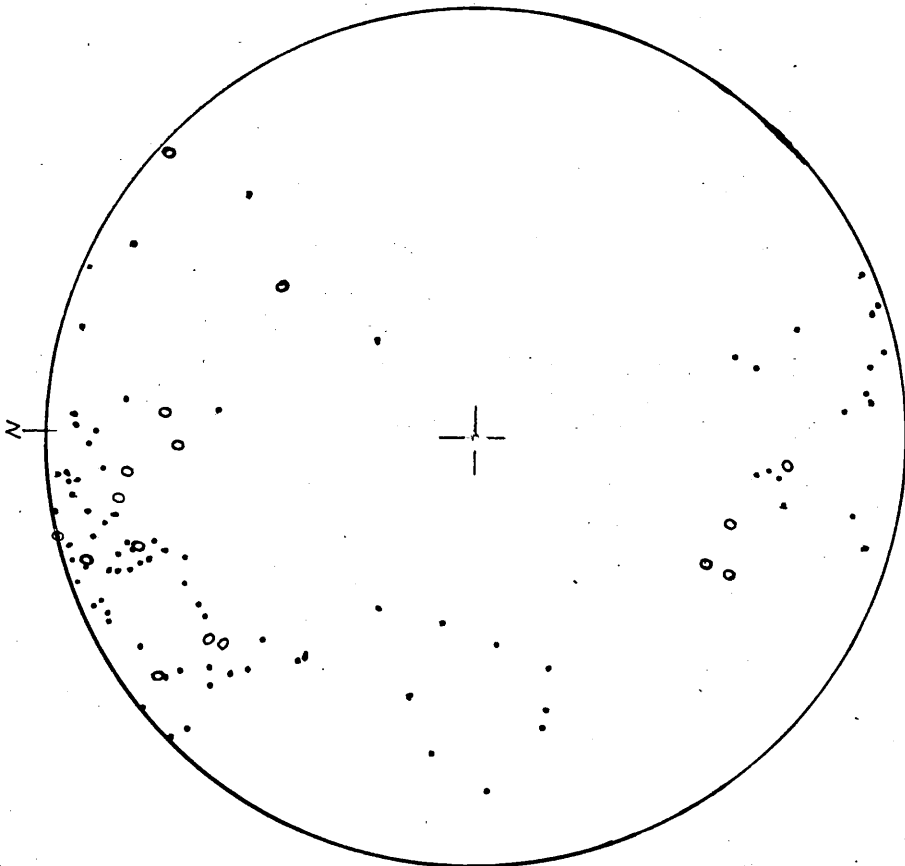
THE KOMAGNES, EIDVÅGEID SCHIST
AND TROLLVANN GROUPS

B
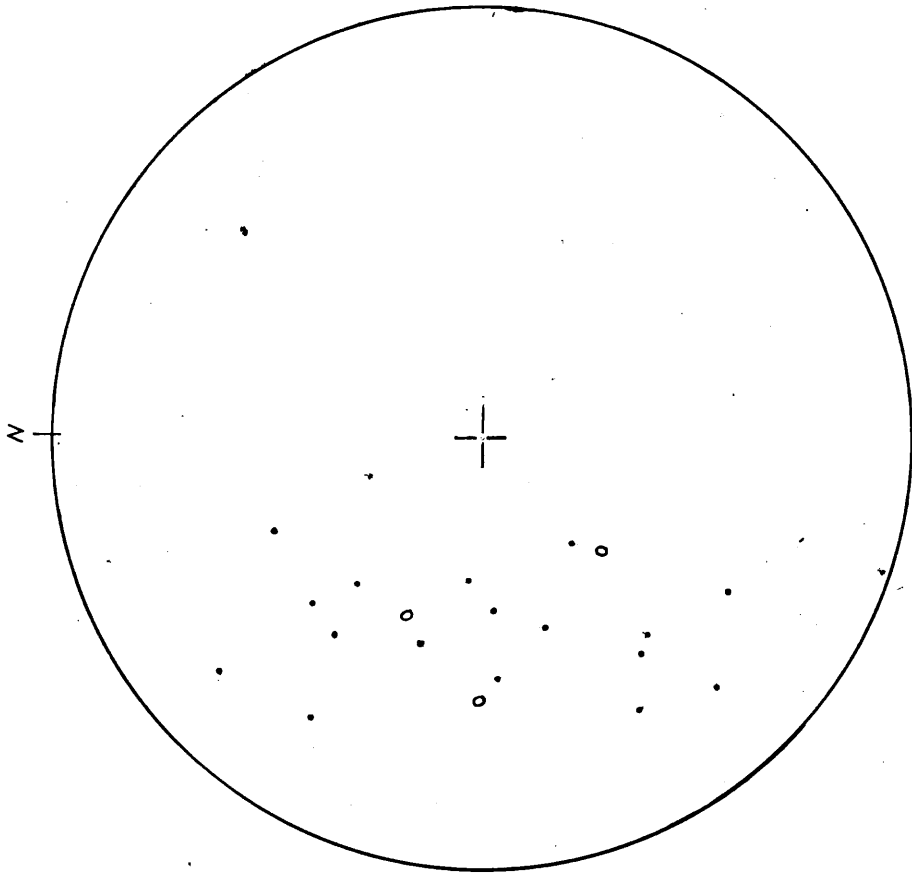
OLDERFJORD AND OLDERBUGTEN GROUPS

trend of the axis, is almost at right angles to the trend of the lower
antiformal hinge. The upper surface of the specimen which shows a skim
of biotite-rich pelitic material, bears a very coarse sinuous lineation.
This curves into and disappears at the plane separating the lower and
upper antiformal folds (see Fig.6). Cut and polished sections taken
at a number of intervals round the respective fold closures indicate
a number of features:-

1) a) The coarse lineation is a micro-fold lineation. These
        microfolds on the upper surface of the specimen forma series
        of arcs which are non-concentric. The greatest curvature of
        these arcs occur near the fold hinge. Away from this hinge
        the lineation may be almost straight.

    b) The microfolds at depth in the specimen always have the
        correct sense of facing to the major fold hinge; at the
        surface, however, the sense of facing may be reversed.
        Indeed, along the axis of any single microfold, the sense
        of facing may change from being congruent to the major
        fold hinge to being incongruent. (See Fig.6).
        These features suggest that the distortion of this lineation
        is a surface phenomenon produced by later movements
        concentrated along this surface, which represents the
        boundary between the pelitic and psammitic lithologies.

2)   The style of the major folds as revealed in the series of
        sections is relatively constant. The area of greater
        curvature is not marked by any fundamental change in
        style.

    It is worthy of mention that at this same locality, in similar
lithology there is an F.1 fold of similar style which is sensibly
cylindroidal (Plate 30). This fold has been coaxially refolded by
F.2.

FIG 6

CURVED AXIS F.1 FOLD

A-F REFER TO SLIDES IN PETROFABRIC SECTION

1.CM

LOWER FOLD HINGE

MICROFOLDS OFTEN CHANGE SENSE OF FACING ALONG THEIR AXES

VERY COARSE MICROFOLD LINEATION

UPPER FOLD HINGE

LINEATION SUB-PARALLEL TO FOLD-HINGE

LINEATION CURVES INTO FOLD HINGE

A

B

C

D

E

F
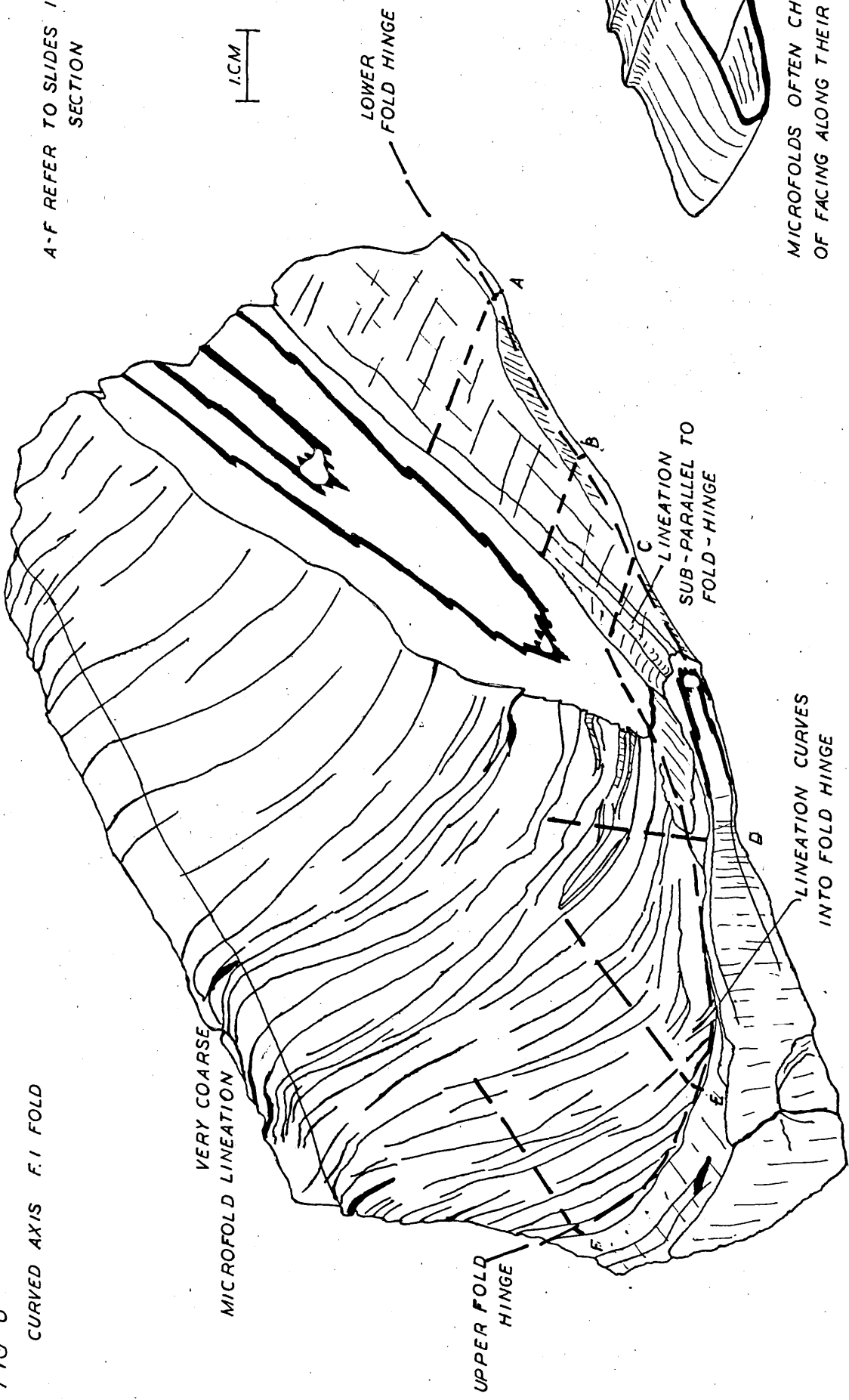
There is a considerable amount of evidence in the Upper Komagnes
Group that curvature of axial-line is a primary feature of many F.2 folds.
It seems possible therefore, that the curvature exhibited by the F.1 fold
depicted in Fig.6 may have been produced by F.2 movements. For the
following reasons this is considered unlikely:-

1)   The distortion of the specimen induced by later deformation
as exemplified by the microfold lineation appears to be
relatively slight.

2)   The general trends of the upper and lower antiformal hinges
are approximately at right angles. The trend of the microfold
lineation appears to be roughly related to each of these hinges.

A process of simple shear acting parallel to the axial-plane of
the fold and normal to its general trend, could be responsible for the
curvature. If this was the case however, it might be expected that
the arcs of the microfold lineation would be concentric, which they
are not.

Further examples of curvature of F.1 hinges have been noted from
other localities. Plate 42 shows an eye-shaped structure in migmatised
Olderfjord Group psammite. It is possible here, however, that the
curvature was produced by distortion of an originally cylindroidal
fold during the high temperatures concomitant with migmatisation.

In summary, therefore, the evidence seems to point quite strongly
towards the fact that curvature is a primary feature of some F.1
folds. This evidence is particularly strong in the case of the fold
described from Rastaby. The existence of an F.1 fold on a similar
lithology at the same exposure which is sensibly cylindroidal indicates
that the phenomenon is not ubiquitously developed in any single
lithological type, though it has not however been recorded at all in
the more massive psammitee of the Lower Komagnes Group. It is
possible that the curved axis fold described above may be of $F.1^2$ age.

Plate 41.   Curved axis F1 fold, Rastaby beach.

Plate 42.   Eyed-fold in migmatitic Olderfjord Group,
      east of Storvann.

There is no evidence at all to support this hypothesis, though it is suggested to explain the variation in geometry of folds of apparently similar age in similar lithologies.

Generally on Eastern Seiland the phenomenon must be regarded as the exception rather than the rule for F.1 folds. This however, may be due in part to the comparative rarity, deformed nature and poorness of exposure of these folds. It is obvious that good exposure of individual folds is needed to determine the true nature of their hinge lines, this is rarely present. The slightly conflicting evidence may solely be a function of this inadequacy of exposure. It might be said, however, that curvature of F.1 fold hinges has been recorded extensively on the neighbouring island of Sørøy by Ramsay and Sturt (personal communication) and Roberts (1965). Here on the flanks of larger F.1 folds the average trend of curved minor folds may be up to $90°$ from the average trend of the major folds. The minor folds however tend to approach congruence near the hinge of the major structure.

b) Linear Structure related to F.1

Linear structures which can be ascribed to F.1 are confined largely to the psammitic Komagnes and Trollvann Groups. Frequently in these lithologies the coaxial nature of F.2 refolds makes recognition of F.1 lineations difficult. However, the following types have been recognised:-

1) Fold lineations including the microfold puckering described in association with the curved axis fold.

2) Mineral lineation.

1) Folds which have been ascribed to F.1 have been measured and plotted on the stereograms (Fig. 5A and B), each stereogram representing the F.1 folds within the lithological groups indicated.

On the stereogram representing the Komagnes Group there is

something of a scatter of fold axes, the points generally have a low

plunge approximately either to the north or to the south. This scatter

may perhaps be partially explained by modifications produced by F.2

movements, though it again may be due to this tendency for curvature

of axial lines within F.1.

It is impossible to make any definite statement on the distribution

of F.1 folds in the Olderbugten and Olderfjord Groups (Fig.5B) since

there are only three points on the diagram.

2)   The other linear element that can be ascribed to F.1 is a very

strong mineral lineation that occurs particularly in the Komagne Group

and also in the quartzite bands in the Eidvågeid Schist Group. This

appears typically in the quartzites as a strong fine striping on layering

surfaces, and in the psammites of the Komagnes Group, as a rather rib-

like structure which parallel the F.2 fold axes, particularly at

Komagnes. The latter observation would seem to indicate that the

lineation is related in fact to F.2 (Plate 43). Indeed it seems

probable that the structure has been much accentuated by F.2 movement

in places where it is coaxial with F.2 folds. However, at Komagnes

there is some evidence of a slight discordance of this F.1 lineation

to F.2 fold axes (see Plate 44). Similarly on the peninsula at

Eidvågeid where F.2 folds have rather erratic trends, examples have

been recorded of these folds folding a lineation identical to the

one seen at Komagnes (see Plate 45).

This lineation has two forms:-

1)   In the psammite it consists of elongate flattened quartz

        segregations.

2)   In the basic schist bands and the green calc-silicate bands

        it consists of an alignment of tiny amphibole crystals.

        This lineation can in many cases within the Komagnes Group

        be seen to be refolded by F.2.

Plate 43. F$_2$ fold showing strong coaxial F$_1$ lineation, Komagnes.

Plate 44. F$_1$ lineation oblique to F$_2$ fold axis (lineation
parallels the hammer-shaft). Komagnes Group, Vasbugt.

If the distribution of these lineations is examined on the stereograms (Fig.5A) it can be seen that in the case of the Komagnes Group there is a concentration of points having a low plunge approximately either to the north or to the south. This is broadly in agreement with the distribution of the F.1 fold axes. Deviations may perhaps again be explained by modifications produced during the F.2 deformation. The lineation is therefore a 'b' lineation relative to F.1. An examination of the lineations seen in the Olderfjord and Olderbugten Group (Fig. 5B) shows a rather wider scatter of points, which tend to lie on a great circle. It is thought likely that the intense deformation which has occurred subsequently within these groups has served to obliterate any F.1 lineation. These structures may have a rather different origin. (See P.70).

It has not been possible to identify any boudinage structure which may be related to the F.1 deformation.

2)   Minor Fold Structures associated with the F.2 deformation.

The great majority of folds observed on Eastern Seiland are of F.2 age. They have a considerably variable morphology and frequently approach the geometric form of some F.1 folds. For this reason the criteria mentioned previously were used to assign any particular fold to one or other of the major fold - forming deformations.

The F.2 folds fold a flaggy banding in the psammitic groups which is in general attributed to the F.1 deformation. F.1 folds are occasionally seen to lie within this flaggy banding. Sometimes lineations which are regarded as of F.1 age are folded by the later folds though often these folds are apparently coaxial with the lineations.

One of the most interesting features of the F.2 deformation is that it apparently produces a primary curvature in the axes of some

FIG. 7.

F.2. FOLDS IN A CALC-SILICATE SCHIST BAND FROM THE LOWER KOMAGNES GROUP

1 CM

QUARTZ

of the fold structures. This tendency towards curvature of hinge lines
and the general style of the minor structures appears to be related to
lithological type. It is thought expedient therefore to describe the
F.2 folds as they occur within the various stratigraphic divisions.

a)   The Lower Komagnes Group

The style of folding within the lower part of the Komagnes Group
is somewhat variable. The folds may approach an isoclinal style or
may be rather more open asymmetric structures. The dip in the lower
part of the Komagnes Group is fairly uniformly to the west. The
outcrop is characterised by areas where there is a great paucity of
F.2 folds and other areas where the structures occur in abundance.
In the latter it is noticeable that the axial planes of these folds
are in general parallel to the sheet dip of the rocks. This explains
the often rather monotonous westerly dip in the psammites here, and
indeed on the rest of Eastern Seiland.

One of the important features noted in areas where there is
an abundance of F.2 structures is that they appear to be stacked one
upon the other in a vertical sense. Their axial planes have a shallow
dip to the west. In these areas the style of the individual folds is
generally isoclinal. These observations, when combined with the
fact that large tracts of country occur where there is a relative
dearth of F.2 folds, is interpreted as meaning that the vertically
stacked areas are the short limbs of F.2 structures. The latter
areas are the long limbs, which have in general undergone much
attenuation. Plates 46, 47, 48 and 49 show the vertically stacked
short limbs. On the scale of individual folds the style is
isoclinal but on a larger more generalised scale the style is of a
more open asymmetric to monoclinal type (Fig. 8).

Plate 47. Short limbs of F_2 folds. Komagnes Group, Vasbugt.

Plate 48. Short limbs of F_2 folds. Komagnes Group,
north of Rastaby.
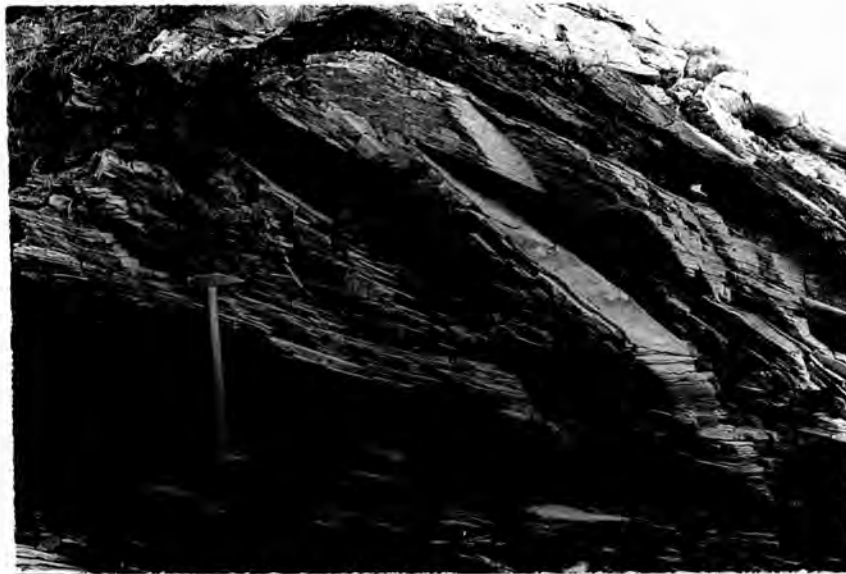
Plate 49. F$_2$ fold, Komagnes Group, Vasbugt.

Plate 50. Chevron F$_2$ folds in a schist band. Komagnes Group, Komagnes beach.

FIG 8

GENERALISED STYLE OF FOLDING

_A_

STRONGLY ATTENUATED
LONG - LIMBS

SHORT - LIMBS
STACKED FOLDS WITH
GENTLY DIPPING
AXIAL PLANES

_B_

INDIVIDUAL FOLDS
IN SHORT LIMBS
ISOCLINAL

_C_

GENERALISED STYLE
IS
MONOCLINAL
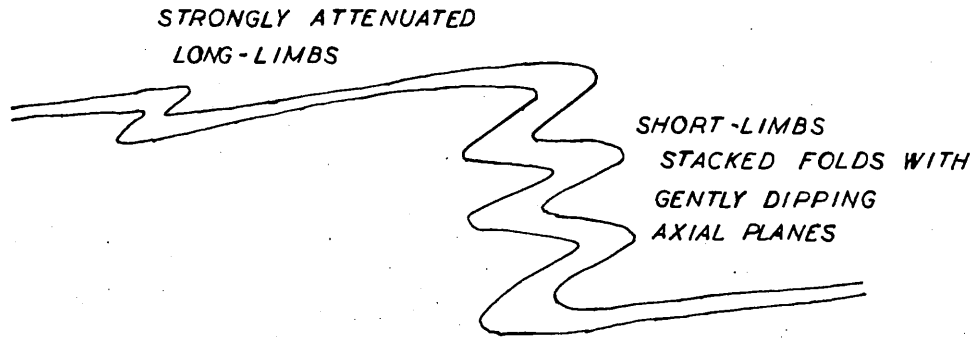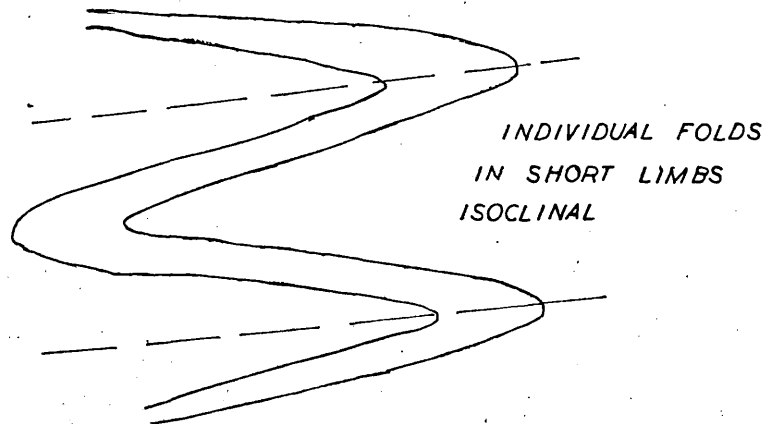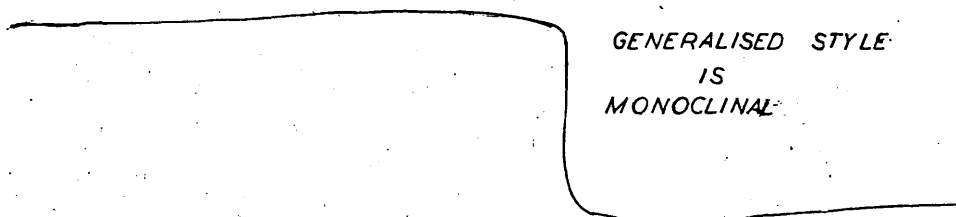
The detailed geometry of individual folds within the group is somewhat variable. However, the folds appear to show some thickening at the hinges though often this is not very marked, together with relative attenuation on the limbs. The slightly less competent hornblende schist bands often tend to show a rather more chevron-type of fold geometry. (Plate 50, also Fig. 7). In the more massive psammites the folds are rather more open asymmetric structures tending towards a parallel-type geometry (e.g. Plates 51 and 52). In the latter Plate the change in the style of the fold when traced up the axial plane seems to indicate that some sort of decollement has taken place along the layer above the base of the pencil. In the psammites where the layering is rather more flaggy the folds tend to be somewhat tighter, (Plates 53, 54 and 55), though often this does not appear to be accompanied by any marked thickening at hinge zones, (e.g. Plates 56 and 57). These tend to maintain a parallel-type geometry. The presence of pelitic intercalations greatly facilitates the tightening up of this type of fold, because this material is easily squeezed out of fold cores.

Plate 58 shows some monoclinal F.2 folds with a parallel geometry. At the top of the photograph the beds in a similar lithology appear to be unfolded. This has been accomplished by slip along a plane which runs slightly obliquely across the top of the photograph from right to left. The plane appears as a zone of flagginess, this is probably a further example of a plane of decollement.

On the scale of the outcrop it appears that the F.2 folds of the Lower Komagnes Group are sensibly cylindroidal. Examination of Fig. 9A to some degree corroborates this by the appearance of a concentration of points just west of north. These points have a low plunge. The scatter shown on this diagram can partly be explained

Plate 51.   F$_2$ folds, Komagnes Group, Komagnes.

Plate 52.   F$_2$ fold, Komagnes Group, Komagnes.

Plate 53. Tight F_2 folds. Komagnes Group, southern Vasbugt.

Plate 54. F_2 folds in flaggy psammite. Komagnes
Group, north of Jernelv.

Plate 55. F$_2$ folds in interbanded psammite and pelites.
            Komagnes Group, north of Jernelv.

Plate 56. F$_2$ folds in flaggy psammite. Komagnes Group,
            Jernelv.

Plate 57. F$_2$ folds in flaggy psammite. Komagnes Group, Jernelv (detail to left of 56).

Plate 58. F$_2$ folds in flaggy psammite. Komagnes Group, Jernelv. (Note plane of décollement at top of plate.)

FIG. 9. STRUCTURAL DATA FROM THE KOMAGNES AND EIDVÅGEID SCHIST GROUPS

A

B
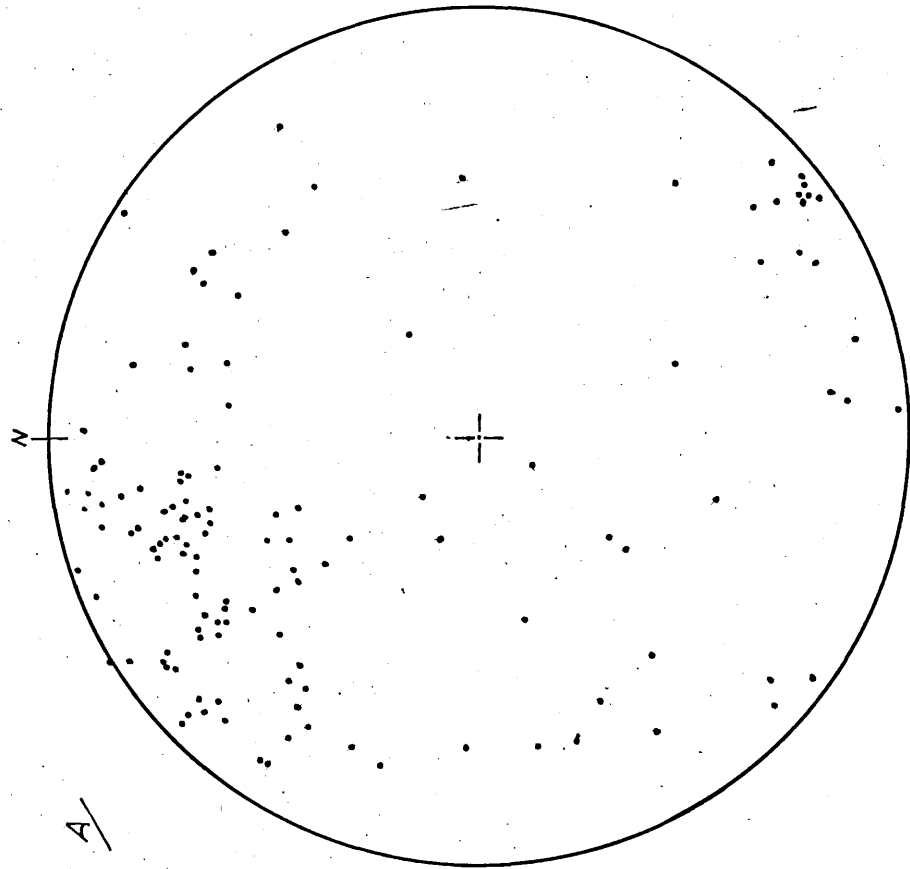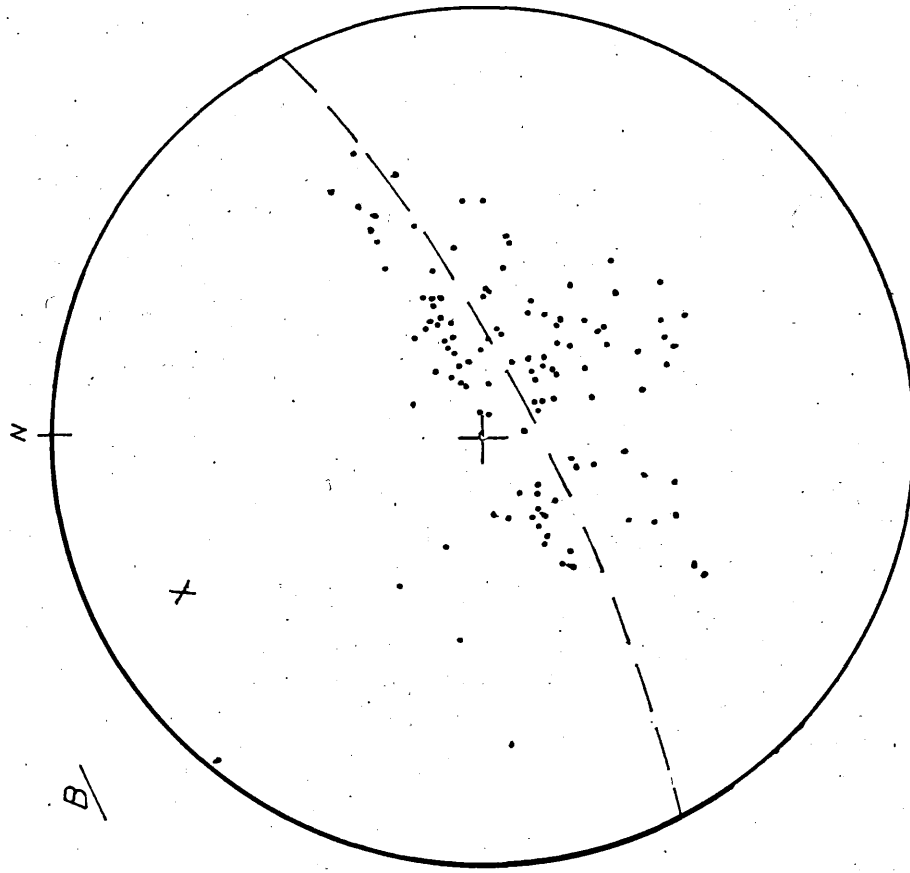
POLES TO AXIAL PLANES OF F.2.FOLDS

F.2. FOLD AXES

by the influence of a local F.3 phase of deformation. This warps F.2 folds such that their axial planes are dipping E.N.E. (Fig. 9B). The axis of refolding is apparently coaxial with the mean axis of F.2 in this group of psammites (see Fig. 9A and B).

Thus in general, F.2 folds in these psammites are sensibly cylindroidal and have a remarkably constant trend and plunge. However at a beach exposure of calc-silicate schist in the Lower Komagnes Group (see Fig.7) a number of very tight folds occur which refold a lineation defined by amphibole crystals. These folds have trends which are rather incongruous to the general F.2 trend described above. Indeed within the hand specimen individual minor folds have substantially different trends.

It would seem from these observations that lithology is a factor which greatly influences not only the genetic form but also the trend of minor folds.

b) The Calc-Silicate Schist

Generally this lithology is not very well exposed. It tends to lie at the base of cliff features formed by the more psammitic lithologies lying structurally above. It is thus much obscured by scree and vegetation. Where exposed, however, the lithology generally has a well-defined banding which is frequently considerably deformed by F.2 folds.

The style of the folds developing within this lithology is rather variable. Folds with a chevron-type geometry have been recorded, where the layering is fissile. Generally however a more flaggy layering seems to favour the development of folds with an almost parallel-type geometry, though with a tendency towards angular hinges (e.g. Plates 59 and 60). The chevron-type geometry may, however, be combined with the concentric-type geometry in any particular fold.

Plate 59. Concentric fold in calc-silicate schist,
           north of Jernelv.

Plate 60. Chevron folds in calc-silicate schist,
           north of Jernelv.

This is true of the fold depicted in Plate 59. The outer arc of this structure has an excellent concentric geometry. In the core however, the chevron geometry predominates. De Sitter (1964, P.277) shows that the maximum amount of flattening that a concentric fold can undergo before it becomes modified is 36%. The development of chevron geometry in the case of the above fold is probably due to flattening in excess of this value. The quartzite band which approximately separates the two geometric types shows evidence of thickening at the hinge and attenuation on the limbs. This would indicate some measure of flowage within this layer. This is also true of other quartzite layers nearer the core of the structure. A further adjustment to the flattening is seen in the tendency to form a box-like geometry in the layer just above the continuous quartzose horizon. It is only the outer arc of the structure that is maintaining a concentric form.

Plate 60 again shows the phenomenon already described in the Lower Komagnes Group, namely the vertical stacking of minor folds in the short limbs of more major F.2 structures. The axial planes of these folds again have a fairly shallow dip to the west, approximately parallel to the sheet dip of the psammites as a whole.

Examination of the limited exposures that occur of this lithology indicates that it has folded disharmonically relative to the enclosing psammite.

c) Mixed Pelitic and Psammitic Group

This is a mixed lithology consisting of closely interlayered pink psammitic units with biotite and amphibole schists. There is also some basic material present in the form of amphibolite sheets.

The style of the folds is generally quite tight tending towards isoclinal. There is often not a great deal of thickening of the psammitic ribs at the hinge of the folds. This is especially so in the case of the thicker units i.e. those greater than $\frac{1}{2}$ cm. in

thickness. The ribs thinner than this, however, often show very intense attenuation on the limbs of the folds. This may give rise to the development of boudins and occasionally slides, (see Fig. 10).

Plate 61 shows an example of an F.2 fold in this lithology. In the field these folds clearly fold a strong schistosity, defined by biotite in the pelitic units.

The fold structure depicted in Plate 61 makes a very interesting study. Fig. 10A is a detailed drawing of the central part of this photograph. The layer XY in the diagram shows considerable thickening on the lower inverted limb of the fold structure. This inverted limb is abruptly truncated by the layer MN. It seems very likely that the layers XY and MN were at one time continuous. During the latter part of the folding, rupturing occurred probably at the hinge of the complementary synform to the antiform shown in Fig. 10B. The antiform was thus thrust along the top of the layer MN. The rock adjacent to this plane shows some development of vein quartz and chloritic material and there is some indication of contortion of the layer just above the plane of thrusting.

Fig. 10 shows a section along the line AB or 10A. It would appear that the contact between the outer-pelitic and psammitic material (the layer PG) has acted as a plane of decollement. The psammite at the hinge of the fold has been pinched into a tight isocline. The pelitic material however has maintained continuity of style with the rest of the fold. This implies detachment and slip along the plane separating the two layers. Both the phenomena described here viz. the thrusting and the decollement are probably of late F.2 origin.

Examination of some of the minor folds in this lithology seems to indicate that many of them have curved axes. This is true of a number of folds from a beach exposure at Jernelv. Although in many cases

Plate 61. Décollement in the Mixed Psammite and Pelite
           unit, Jernelv.

Plate 62. Curved axis F2 folds. Upper Komagnes Group,
           west of Rastabyvann.

FIG 10
DECOLLEMENT IN THE KOMAGNES GROUP

A/

A.   B.

X

M   Y   N

|— J.CMS —| PLANE OF DECOLLEMENT

B/
/ EXPLANATION OF A

1/   2/
X   X
Y   N   Y
M   M   N

C/

P
← PLANE OF DECOLLEMENT

DETACHMENT
HERE

Q
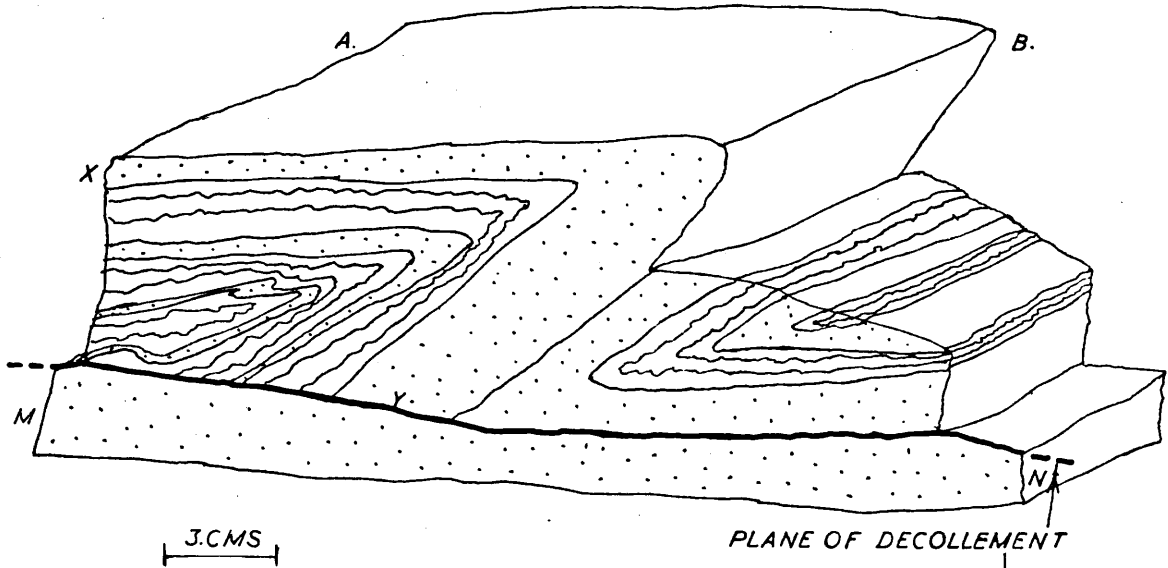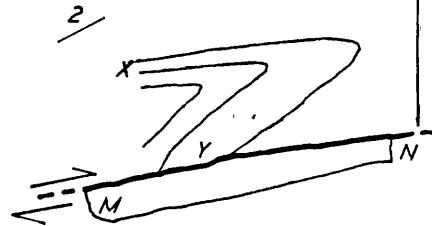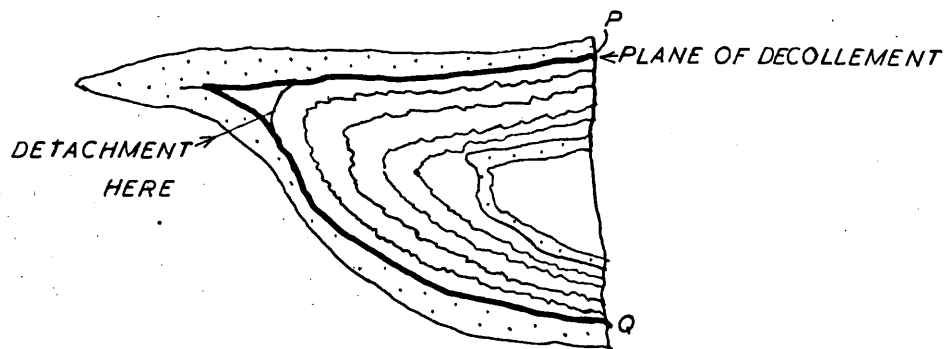
[::] PSAMMITE

[≈] PELITE |— J'CMS —|

individual folds cannot be seen fully in three dimensions, measurement
and plotting of their axes indicates some inconstancy of trend. The
phenomenon of curvature of axial line has been observed in some hand
specimens. Thus this slight divergence of axial trend is attributed
to the two dimensional 'cut effect' of the exposure, intersecting with
the curved axes folds at a number of random points on their axes.
Fig. 11 shows this phenomenon; the axial planes of the folds have a
relatively attitude, whereas the fold axes tend to show a distribution
of trend within this axial plane. Although there are only a small number
of points on this diagram this is a phenomenon that has been noted
extensively elsewhere.

d)   The Pink Psammite

The lithology is not highly folded internally. However, its outcrop
pattern indicates the presence of a fold of moderate dimensions to the
south of Jernelv. This fold, as can be seen from the map, conforms to
the pattern outlined for the Lower Komagnes Group, namely the short limb
consists of a number of fairly tight folds stacked one upon the other
with axial planes dipping towards the west. The long limbs are
characterised by a dearth of fold structures and considerable attenuation.
This attenuation is expressed in the pink psammite by the thinning and
eventual disappearance of the outcrop WSW of Russelven. The apparent
thickening of the outcrop just west of Jernelv is attributed to the
local easterly dip imparted to the horizon by bending associated with
the fault.

e)   The Upper Flaggy Psammite with Schist Bands

This unit is lithologically very variable. Generally it consists
of units of flaggy or schistose psammite rocks intercalated with pelitic
and semi-pelitic horizons. The most distinctive feature of this

FIG 11
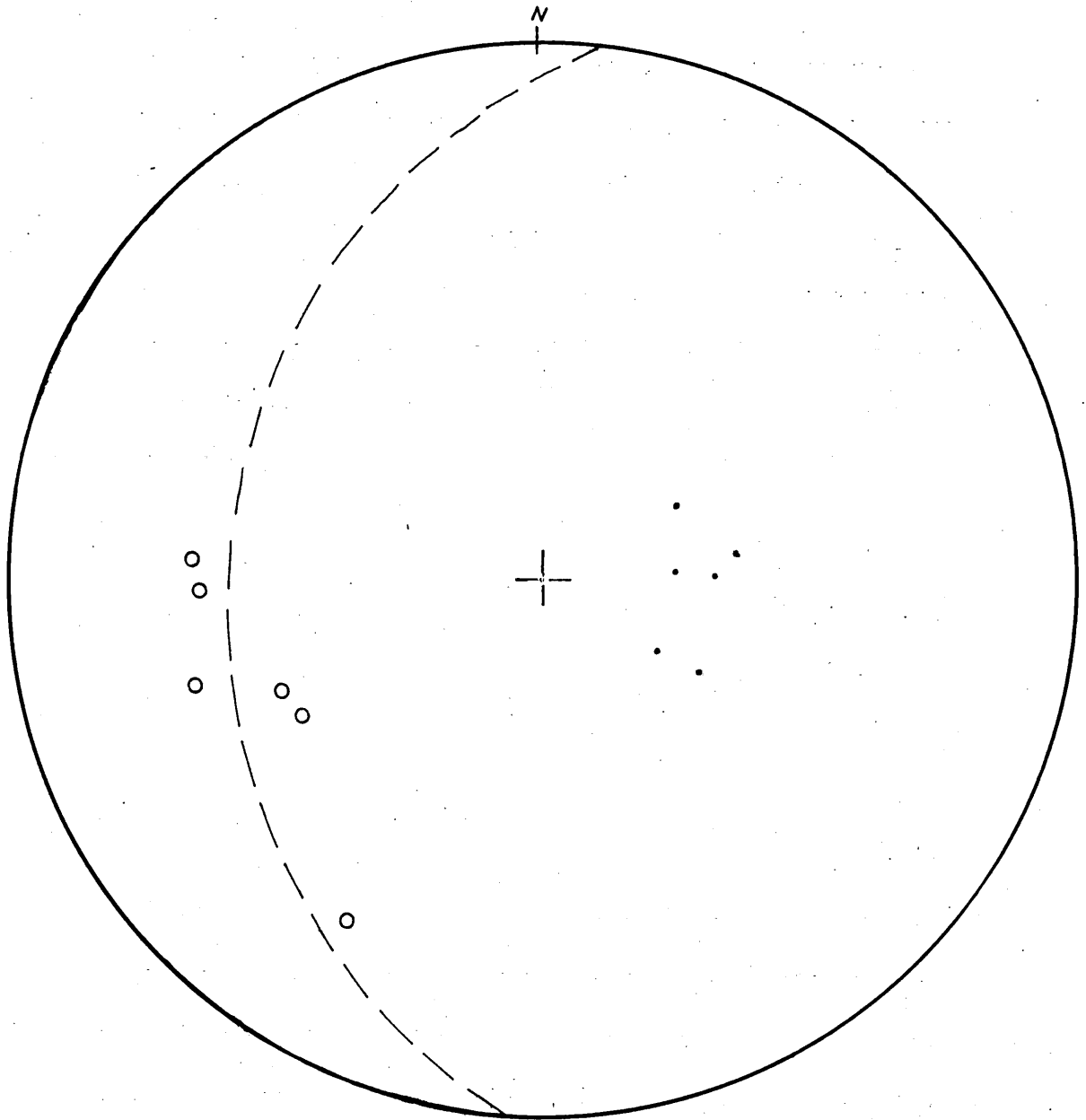
PLOT OF AXES(O)AND POLES TO AXIAL PLANES(·) IN THE MIXED
PSAMMITIC AND PELITIC UNIT. JERNELV BEACH.

Plate 63. F$_2$ folds with irregular trends. Komagnes
Group, north of Rastaby.

Plate 64. Hackly F$_2$ microfold lineation. Komagnes
Group, Rastaby beach.

FIG 12

CURVED AXIS FOLDS FROM THE UPPER KOMAGNES GROUP

A

GARNET

ELASTICUS

1.CM

B
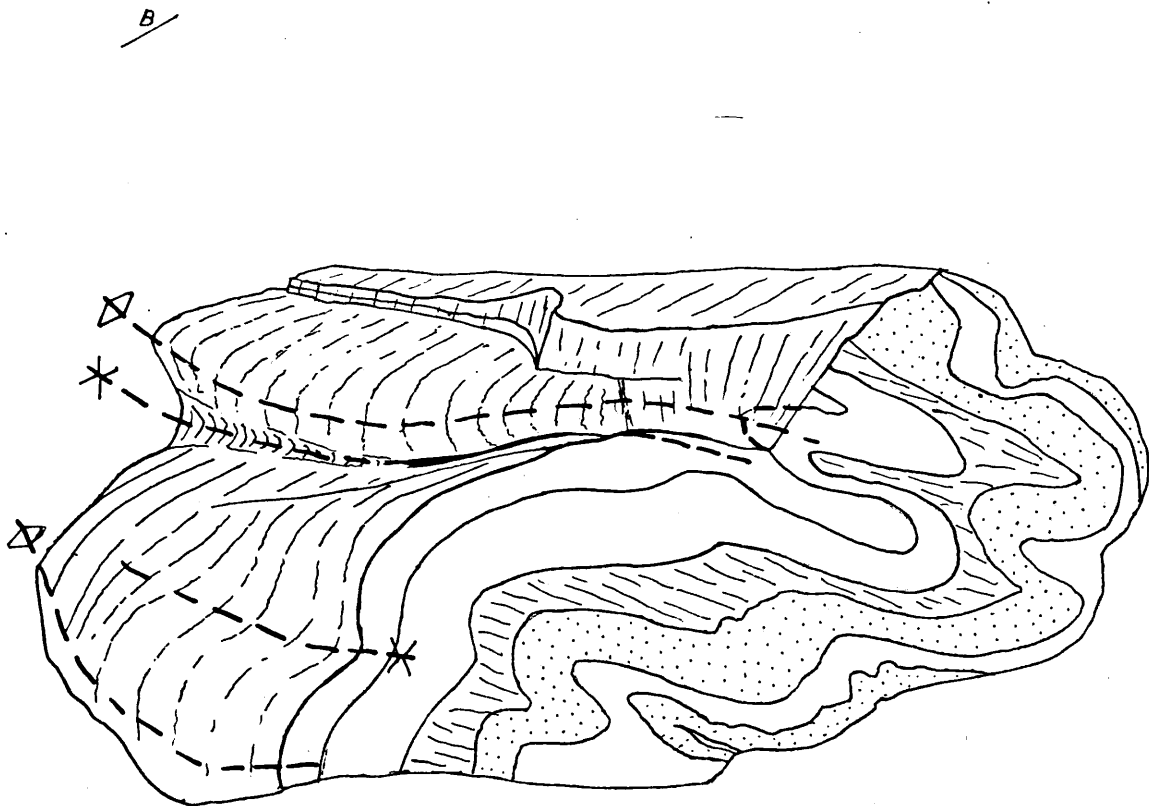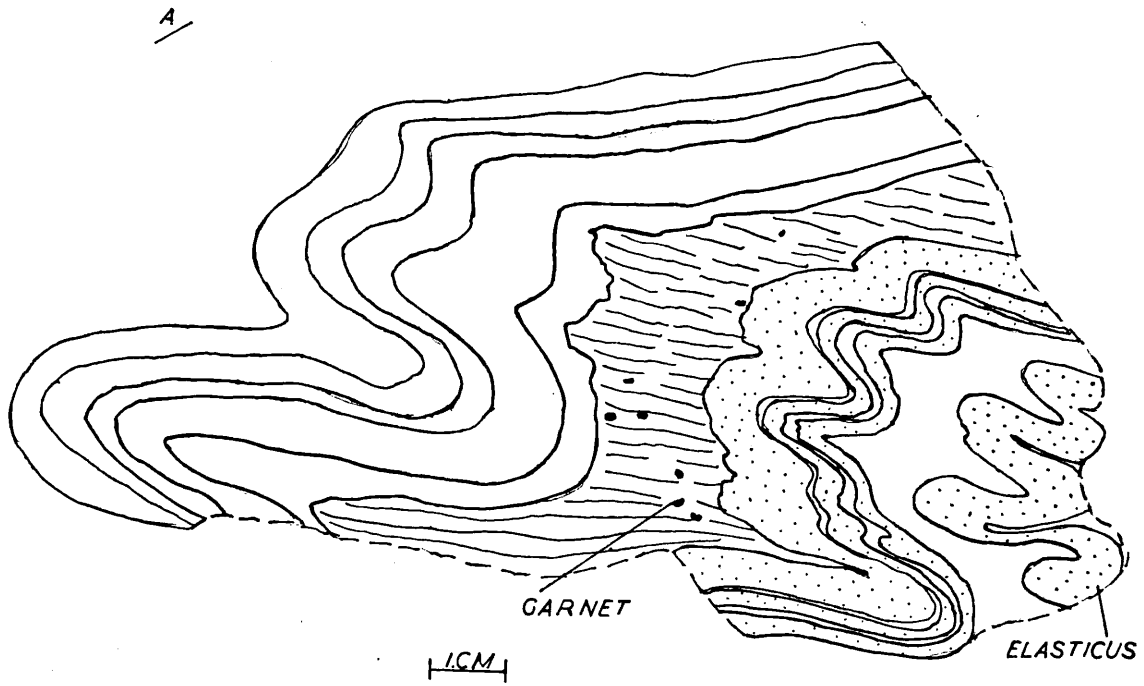
QUARTZITE

SCHIST

BANDED PSAMMITE

1.CM

lithology is the presence of F.2 folds within it which have quite strongly curving axes. The most striking examples of these structures come from those lithologies consisting of a close interlayering of psammite, semi-pelite and pelite. In these folds the pelite units have a strong schistosity which is axial planar to the folds. Sketches of these structures are given in Fig. 12. The intercalated pelitic material behaves incompetantly and is squeezed out of the cores of the tight * symformal folds. Hinges are generally quite rounded and the axes of successive folds curve quite strongly, frequently out of phase with each other. Fig. 12A shows an elasticus developing in a competent psammite unit in the core of this fold.

Plates 62 and 63 show three dimensional exposure of these curved axis structures. A fairly considerable change in the angle of pitch of the structures is indicated. The folds appear often to generate laterally from unfolded layering into asymmetric periclinal structures, which then die out again. This is well shown by SP N16 18 depicted in Fig. 13. Here the lateral development of the fold can be seen in a series of serial sections of a single layer, taken at random points along the curved axis. Also included in the diagram are graphs representing the change in orthogonal thickness of the layer in each section (Graphs A - F). It can be seen that the greatest thickness in each section is achieved at the hinge line of the fold. Similarly, the greatest thickness achieved in the layer as a whole is at the hinge line of the fold, when it reaches its minimum interlimb angle, i.e. at point 5 in graph D. The progressive increase in orthogonal thickness at the hinge lines is recorded in Graph G.

* Note

> The terms synformal and antiformal are used in the usual sense, i.e. an antiform is a fold that closes upwards, a synform closes downwards.

FIG 13 - SERIAL SECTIONS OF A SINGLE LAYER IN
A CURVED-AXIS FOLD
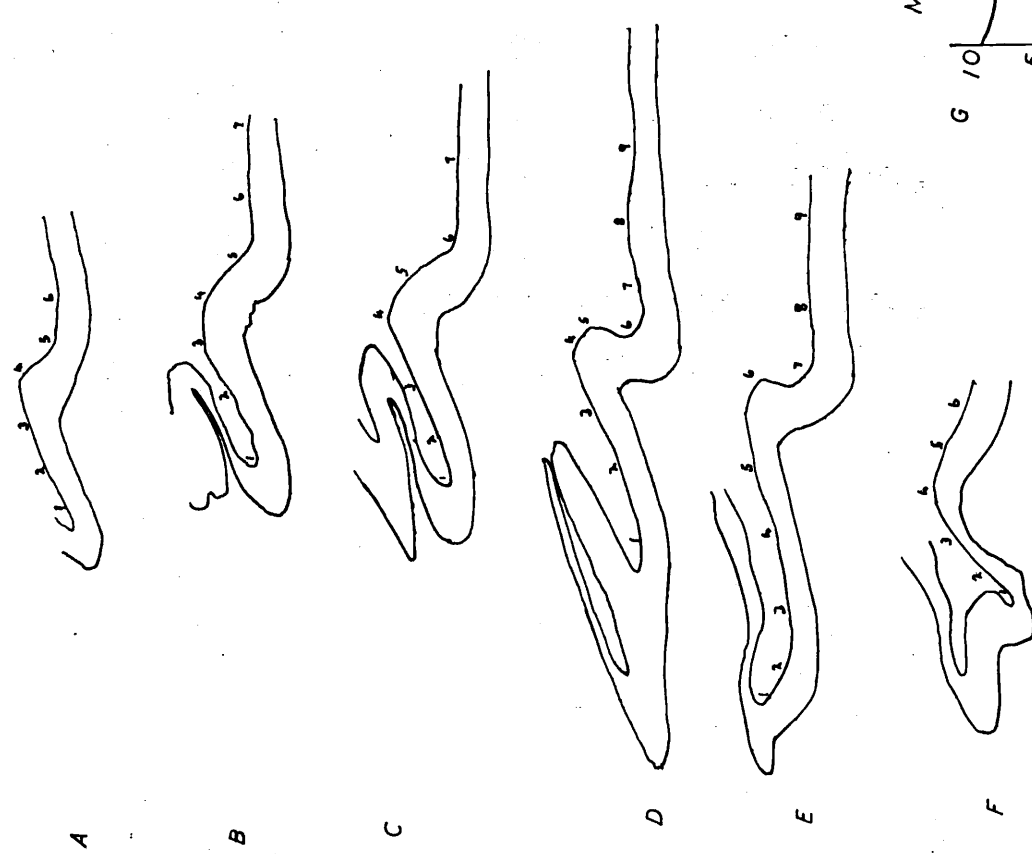
PLOT OF ORTHOGONAL THICKNESS AT VARIOUS
POINTS ALONG THE LAYER
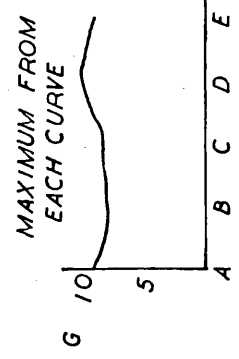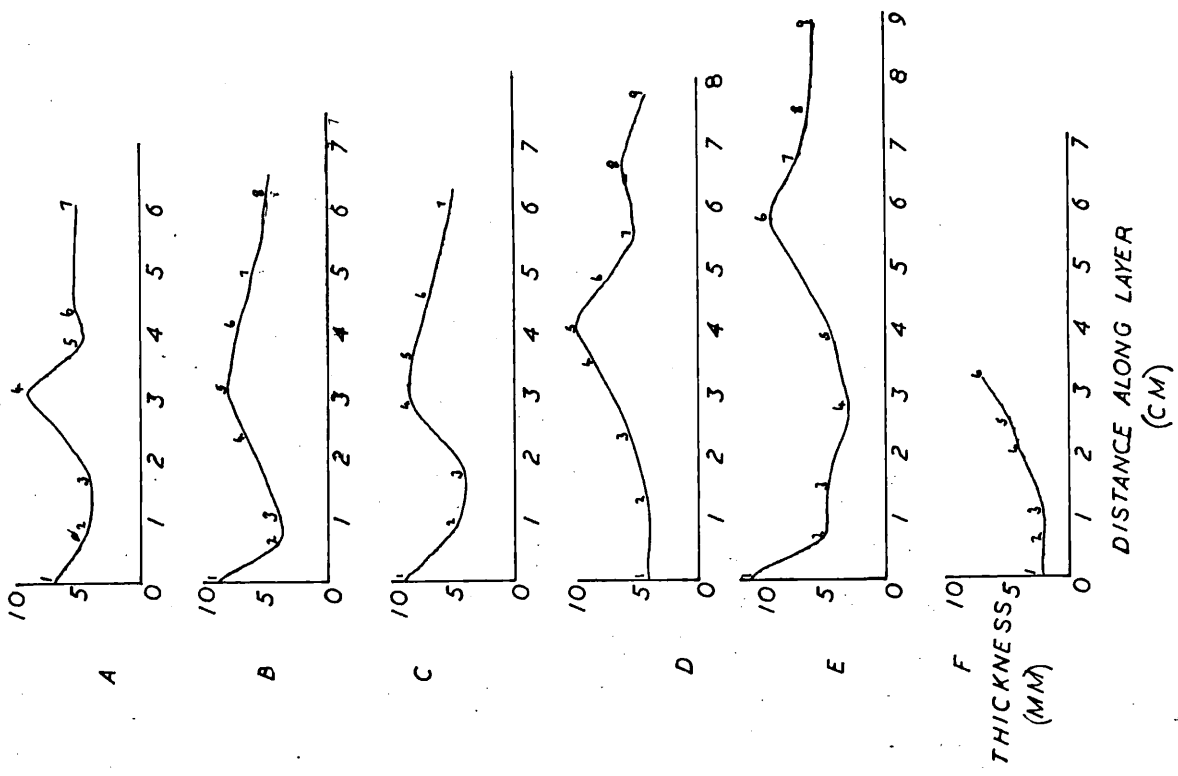
It is obvious that the folds have been subjected to considerable flattening; the presence of elasticus-type structures is clear evidence of this. It could be argued that flattening has produced the curvature of hinge lines by late inhomogeneous movement within the axial planes of originally cylindroidal folds. Study of the three-dimensional form of microfolds within these rocks would seem to indicate to the contrary. Fig. 14A shows a series of thin competent layers embedded in a pelitic incompetent matrix. The competent layers deform into fairly open structures, which are often symmetrical in cross-section. In three-dimensions the folds have a periclinal form. An E2 strain-slip cleavage develops in the incompetent bands. Due to the rather open nature of these folds it would seem unlikely that a late flattening process could have been responsible for the curvature of their axes. If, therefore, these microfolds are regarded as being incipient curved-axis folds, it would seem that curvature begins very early in the history of fold formation.

This is further substantiated by the fold depicted in Fig. 13. Such a fold, which develops laterally from unfolded layering into the overturned asymmetric structure of section D, cannot, at any period in its history have been anything but periclinal. It seems certain however that strong flattening has tightened up these folds, and considerably accentuated the curvature of their axes.

Plates 64 and 65 show a rather wave-like lineation developed on a layering surface at Rastaby. The rock concerned is a fairly competent banded psammate. Cut and polished sections of this rock show that the wave-like lineation is, in fact, a micro-fold lineation caused by small periclinal rucks. Fig. 14B shows a cross-section of one of these structures. It can be seen that the style of the folds changes with depth, from the rather simple wave-like cross section in the centre

Plate 65. Hackly F_2 microfold lineation. Komagnes
Group, Rastaby beach.

Plate 66. F_2 folds in massive psammite. Upper Komagnes
Group, beach north of Rastaby.

FIG. 14

A / PERICLINAL F.2. MICROFOLDS

PERICLINAL RUCKS

SEMI-PELITE

S.I.

SCHIST

S.2.

QUARTZITE

1.CM
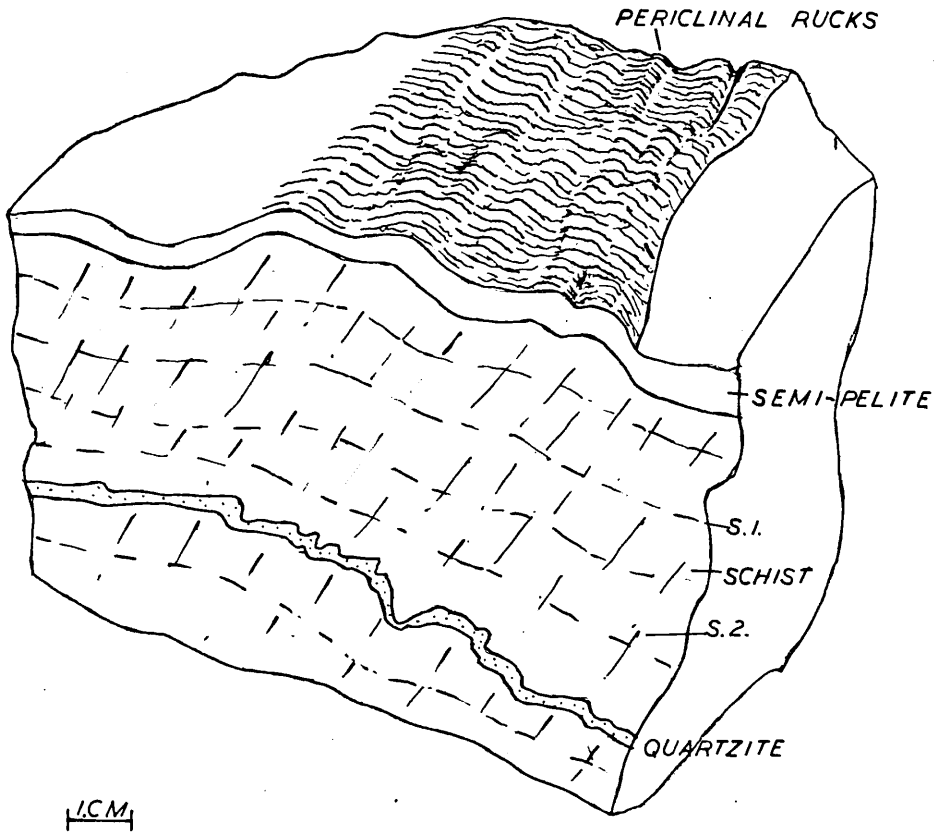
B / CROSS-SECTION OF HACKLY MICROFOLD LINEATION

BANDED FLAGGY
QUARTZITE

1.CM
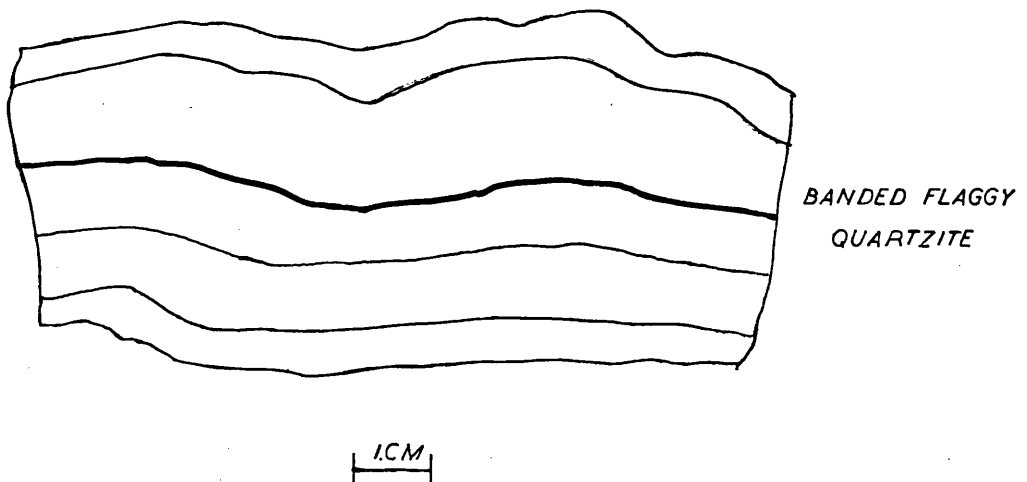
of the specimen to the rather more complex geometry which gives

expression to the folds at the surface. This variation is partially

controlled by differences in thickness and competence of the layers,

thus giving rise to a shorter wavelength in the thinner horizons.

Plate 95 (P⠀⠀) shows a fold which occurs on the beach at

Rastaby. The lithology concerned is a flaggy banded quartzite. The

fold shows a parallel type of geometry. Microfolds, which are clearly

related to the more major structures, are of the periclinal type

described above. Pelitic and semi-pelitic material is almost

completely absent. Such pelitic layers as do exist appear to have

acted as planes of decollement. The slip which has occurred along

these planes has allowed the fold to change its style with depth.

For example, the fold structures beneath the hammer head are relatively

open compared with those above the hammer head. This process has

allowed the fold to tighten up beyond the limit permitted by pure

concentric folding. The pelitic material along the planes of

decollement has been squeezed into the voids which have tended to form

between the successive disharmonically folded layers. An example of

this occurs just below the centre of the photograph, where the pelitic

material can be seen as dark areas between the buckled quartzite

layers. Examination of the photograph indicates that there is not a

great deal of thickening of layers around the hinge zones of these

folds. A parallel type geometry appears to prevail, in contrast to

the interlayered psammitic and pelitic lithologies where flowage

towards the hinges seems to be a more general occurence.

The minor flexures on this particular fold were measured and the

results are given in Fig.15. The minor folds show a fairly large

diversity of trend and plunge from the trend of the major fold, which

is circled, poles to axial-plane also show a fairly wide distribution.

FIG 15

PLOT OF FOLD AXES(O) AND POLES TO AXIAL PLANES(·) OF THE FOLD
ON RASTABY BEACH (PLATE 95)
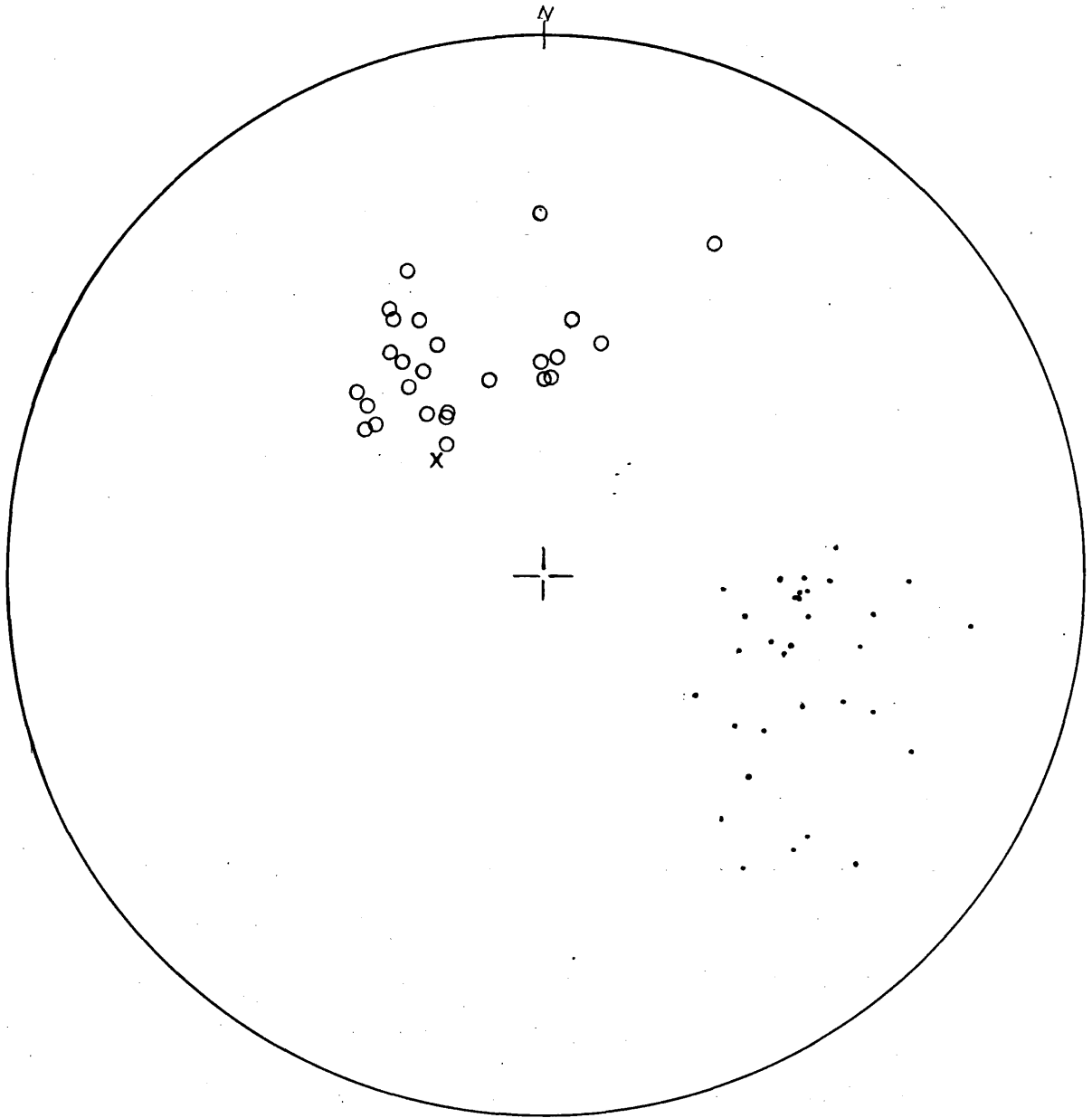
X PLOT OF MAJOR FOLD

This further indicates the considerable diversity in trend seen in F.2 folds, especially in this upper part of the Komagnes Group.

Further evidence of this change in plunge of F.2 folds was afforded by study of the beach section at Rastaby. Here a slight variation in lithology allows the geometry of a single layer to be constructed. This is shown in Fig.16. It can be seen from this diagram that there is a considerable diversity of plunge of the minor folds associated with the more major structure. This is probably due to an inherent curvature of the axes of the folds involved.

This phenomenon of curvature is not entirely ubiquitous in the upper part of the Komagnes Group. F.2 folds have been recorded at the beach exposures at Rastaby, which are on the scale of the outcrop sensibly cylindroidal. These folds are associated with a very strong microfold lineation which parallels their hinge lines exactly. Such a fold is depicted in Plate 30 (P.41.). Here cylindroidal F.2 folds coaxially refold an F.1 fold, which itself appears to be cylindroidal. These folds are approximately symmetrical in cross section.

These folds differ from other F.2 folds in two respects:

1)   The very strong microfold lineation which parallels their axes, such a lineation is absent in the curved axis structures.

2)   The cylindroidal symmetrical geometry.

The association of a cylindroidal F.1 fold with this a typical F.2 fold cannot be fortuitous. It seems likely therefore that the presence of the F.1 hinge has controlled the development of the F.2 structure. This is further suggested by the fact that in similar lithologies close by, curved axis F.2 folds are seen to occur. This would seem to preclude a lithological control in the development of this structure.

Plates 66 and 67 show F.2 folds in lithologies which are

FIG. 16. RECONSTRUCTION OF THE GEOMETRY OF PSAMMITE LAYERS ON RASTABY BEACH.

N

5 METRES

29

27

73

23

55

83

49

70

37

29

46

20CM

PLATE 28.
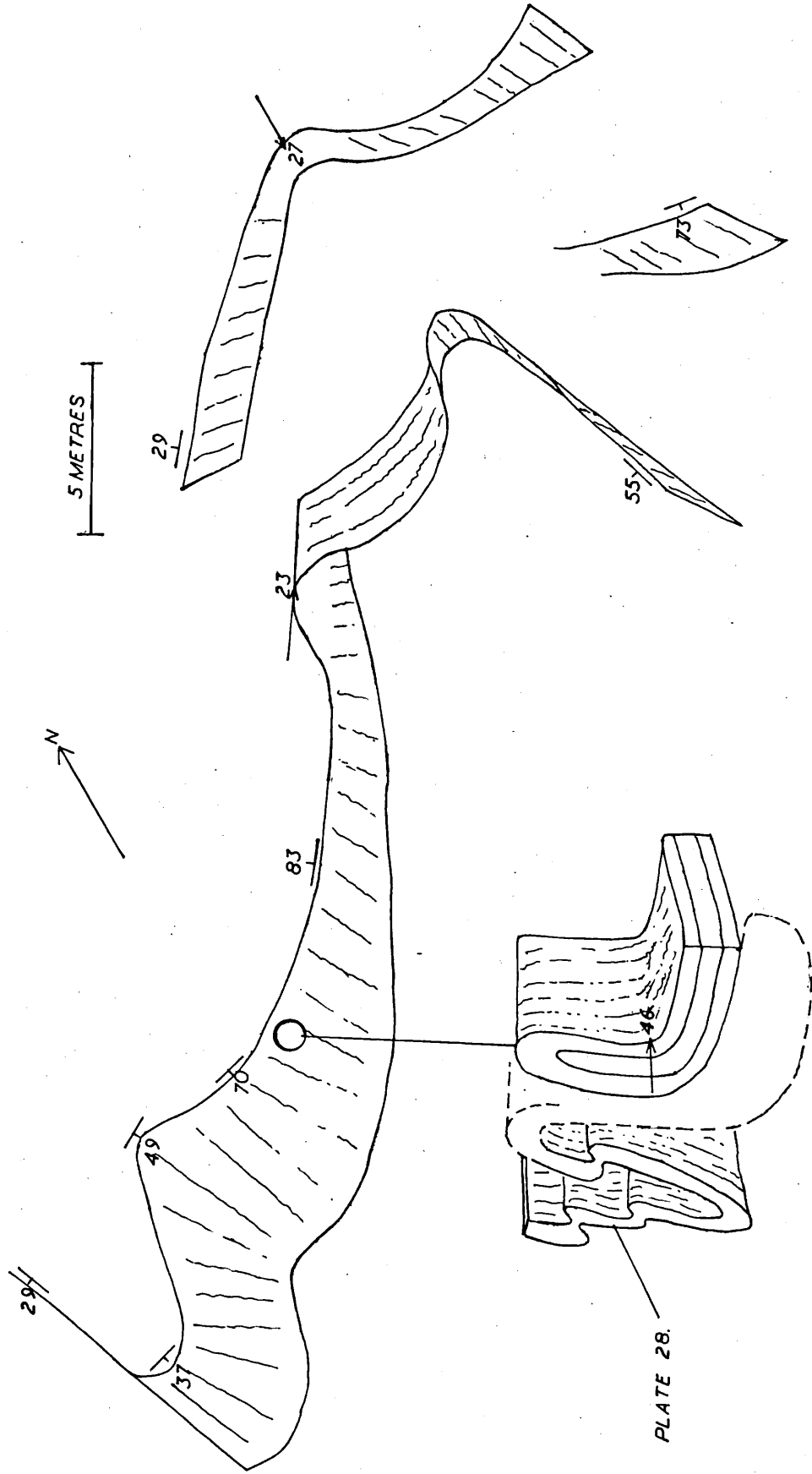
Plate 67. F2 folds in psammites with interbedded
          quartzite. Upper Komagnes Group, north of Rastaby.

Plate 68. F2 folds in flaggy psammite. Upper Komagnes
          Group, north of Rastaby.

dominantly psammitic, though in (66) interbanded amphibolitic material seems to be associated with some thickening at hinge lines and attenuation on the limbs. The folds also have a fairly acute style. Plate 67 shows an interbanded quartzite in a flaggy psammite, the quartzite shows evidence of the development of incipient pinch-and-swell structure. The folds in general have a rather robust form.

Plates 68, 69, 70 and 71 show F.2 folds in rather more semi-pelitic lithologies. The styles are quite variable but the hinge-lines tend to be angular, except in 71. Here the boudinaged amphibolite has alomst certainly exercised considerable influence over the style of fold. There is some attenuation in the limb of the folds, often this is more marked in sub-layers within the large layers. For example in Plate 69 the layer just below the pencil does not show any large thicknesses change along its length, but individual layers within this layer often show quite strong attenuation and thickening. The change in style of the folds with depth is also worthy of note. Above the pencil, the layers are rather thin and the lithology relatively homogeneous, here the folds are rather angular. Below the pencil, however, the lithology is not so homogeneous and the style of the fold is more open with the development of rounded hinges. The dark area between these two zones consists of pelitic material which has acted as a plane of decollement to accommodate the two styles of folding. These folds have strongly curving axes.

Eidvågeid

In the psammitic rocks on the ness at Eidvågeid in the north, a number of F.2 folds are seen. These rocks, although they are structurally in the same position as those at Rastaby are rather purer psammites. However, here again F.2 folds are seen with rather irregular trends. These are often very tight isoclinal structures.

Plate 69. F₂ buckles, Upper Komagnes Group. North
of Rastaby.

Plate 70. F₂ folds, Upper Komagnes Group. North
of Rastaby.

Fig. 17A shows a plot of some F.2 fold axes from the ness at
Eidvågeid, it is apparent that they have a somewhat irregular trend.

Synthesis of Data

Figs. 17B and 17C show a synthesis of all the fold axes measured
within the Upper Komagnes Group. That is all the lithologies above
and including the Calc-Silicate Schist. Two things emerge from these
diagrams:-

1) The fold axes have an irregular spread of trends; they tend
to define a great circle.

2) The pole to this great circle approximately coincides with
a pole to the plot of axial planes of the same folds. This
also approximately coincides with the poles to foliation
planes in the Eidvågeid schist (representing S.1) Fig. 3.

This great circle distribution of F.2 fold axes is attributed to
an inherent curvature of the axes of the folds concerned. The diagrams
represent a synthesis of curved-axis folds measured in random two
dimensional exposures in which the curvature of axis was not apparent.
This is adequately corroborated by the presence of hand specimens of
folds which show the phenomenon of curvature often to a considerable
degree.

f) The Eidvågeid Schist Group

F.2 folds are very rare in the Eidvågeid Schist Group. They
appear only to occur under two conditions:-

1) In the presence of sheets of amphibolite within the
schist proper.

2) In the massive quartzite bands.

Fig. 3 (P . 42) gives an indication of the consistancy of dip
within the dominant schist member of this Group. It would appear that
F.2 fold structures do not form on the schist unless the above

FIG 17

A/ F.2 FOLDS EIDVÅGEID AREA
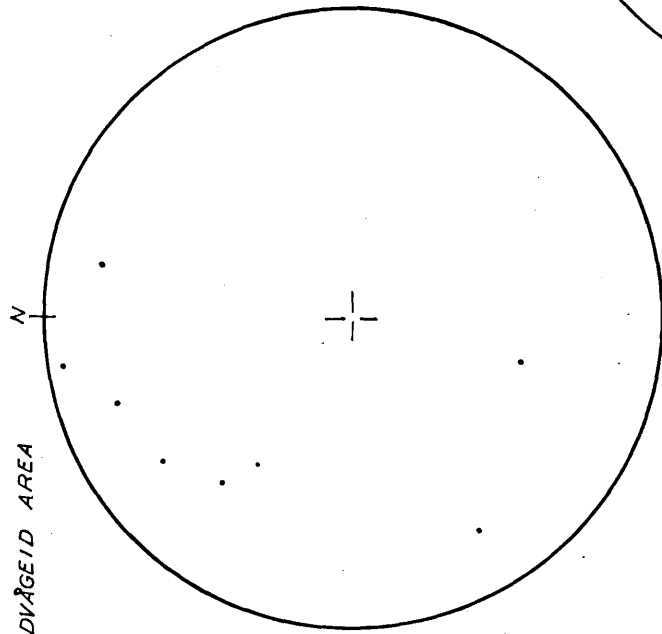
B/ F.2 FOLDS RASTABY AREA
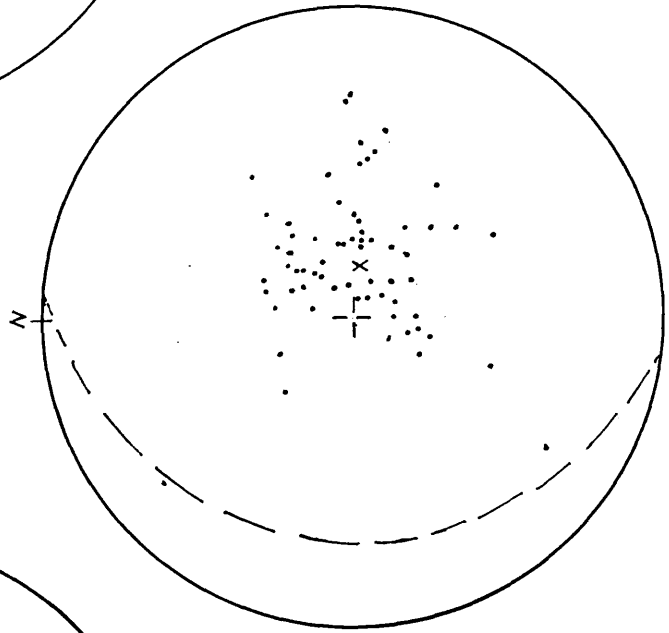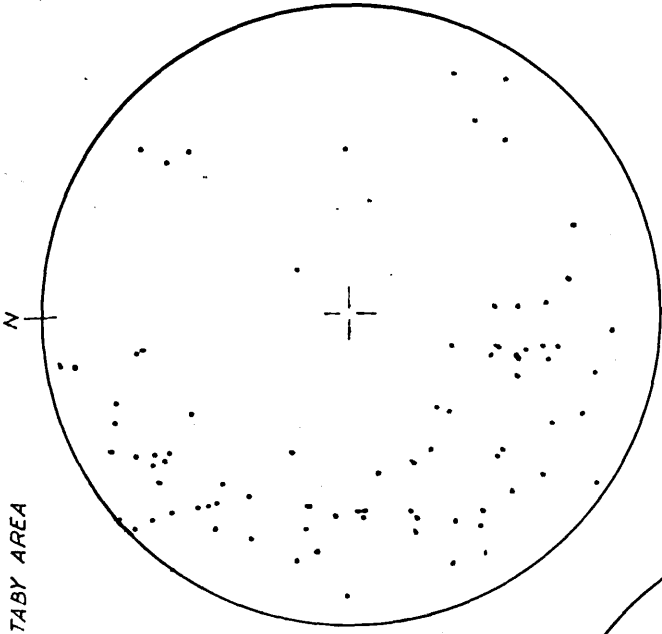
C/ POLES TO AXIAL PLANES OF THE ABOVE FOLDS

conditions 1) is met. Examples of the types of fold associated with the amphibolite sheets are seen in Plates 33 (P.41.) and 72. In all cases the folds appear to be tight structures in which the foliation of the schist is folded round the fold nose. The fold depicted in Plate 72 occurs in a rather psammitic facies of the schist, the sheet of amphibolite, which can be seen just above the pencil, bears a foliation which is also folded round the fold hinge. It is possibly one of the very early amphibolite sheets discussed previously.

Because of the absence of folds in the pelitic facies it seems likely that the F.2 deformation is taken up partly by pure flattening and or partly by interlayer slip. Examination of thin sections of the schist show that pre-F.2 porphyoblasts are strongly augened, whether this augening is entirely due to syn-F.2 flattening is indeterminable.

Similarly growth in garnet porphyoblasts is not synkinematic relative to F.2. It is not possible, therefore, to detect any rotational fabric within these garnets which could have been ascribed to F.2 movements, thus giving some clue as to the manner in which this deformation was taken up.

Fold structures within the massive quartzite horizons are also not very common, this may be a function of a number of factors.

1) The deformation style which may be similar to that already outlined for the Komagnes Group, i.e. the presence of unfolded long limbs and the rather intensely folded short limbs. The mapping out of this sort of pattern is hindered by the lateral discontinuity of these horizons.

2) The thinness and relative poorness of exposure of these horizons. Although often they do form prominent ridge-like features, this is not always the case and they may be obscured by the stunted birch which tends to form on the lower slopes of the Eidvågeid schist outcrop.

Plate 71. Basic sheet folded and boudinaged during F_2, Rastaby beach.
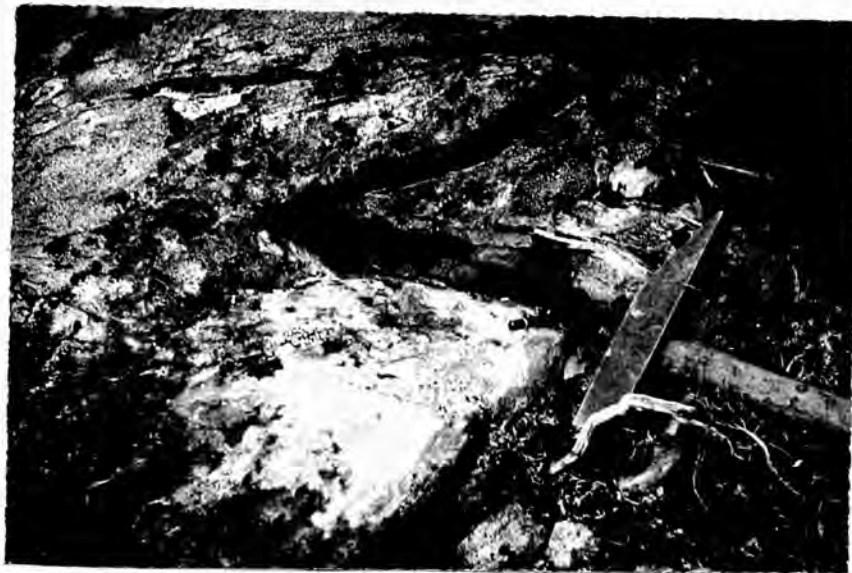
Plate 72. F_2 fold in the Eidvågeid schist inland from northern Vasbugt.

FIG. 18

FOLD STRUCTURES IN THE EIDVÅGEID SCHIST GROUP

F.1 FOLD

QUARTZITE

SCHIST WITH SCHISTOSITY AXIAL PLANAR
TO F.1 FOLD

30 CMS
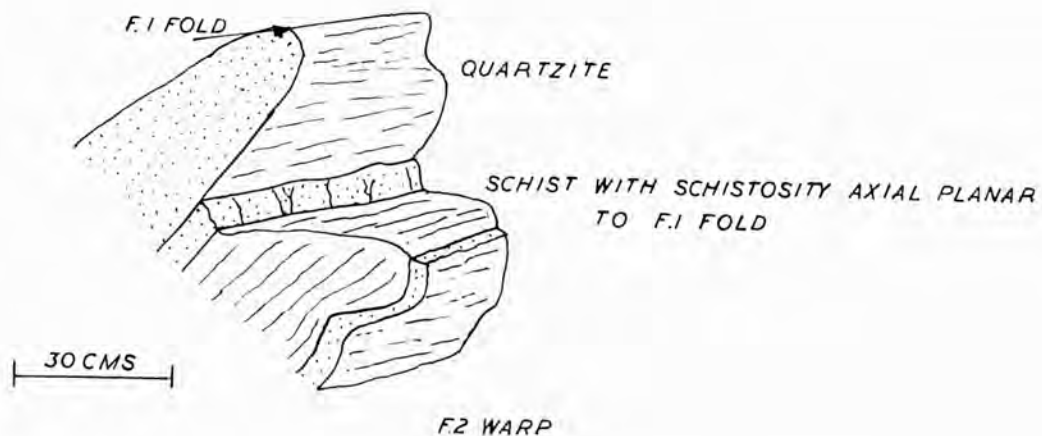
F.2 WARP

F.2 WARPS IN AN INTERBEDDED SCHIST QUARTZITE LITHOLOGY

10 CMS
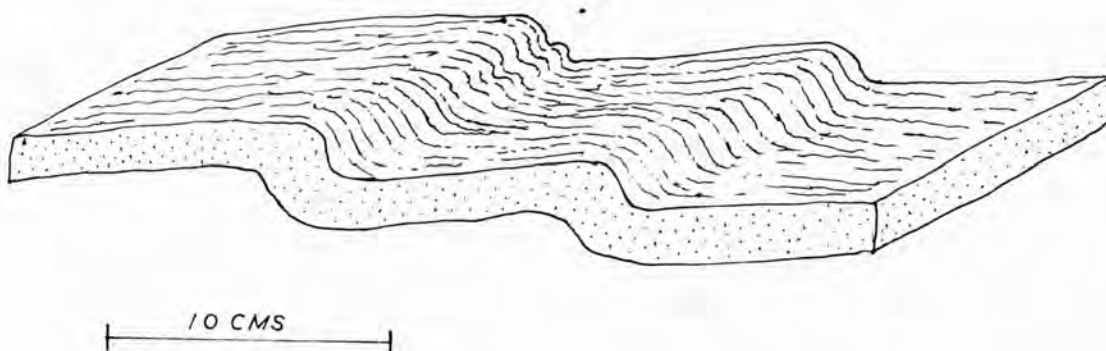
F.2 FOLDS IN MASSIVE QUARTZITE
PLATE 73

The rather massive lithology of the quartzites tends to deform into relatively open structure. This is in contrast to the tight isoclinal which form in response to the F.1 deformation (see Plate 56 P. 43). Examples of these F.2 structures are shown in Plate 73 and Fig. 10.

Plate 73 shows folds which tend towards a parallel-type geometry. The two sketches in Fig. 10 show the rather gentle warps associated with F.2 structures in interbedded schist-quartzite lithologies.

Fig. 10A shows the distribution of F.2 folds within the Eidvågeid schist. As can be seen they are coaxial with the mean F.2 axis in the Lower Komagnes Group, see Fig. 9A and also with the mean trend of F.1 lineations in the Komagnes Group, (see Fig. 5A P. 44). Indeed the folds discussed above can often be seen to be coaxial with a very intense lineation defined by elongate quartz crystals in the massive quartzites.

There does not appear to be any marked tendency towards curvature of axis. The slight spread seen in the distribution of pole to axial planes may be explained by the tendency in the massive quartzites for axial planes to be rather more steeply dipping than these of the folded amphibolite sheets. This is presumably due to the decreased facility of rotation in the thick competent quartzite bands, thus resulting in the formation of a more open style of fold.

g)   The Trollvann Psammite Group

It is apparent from the unfolded nature of the contact with the enclosing schist groups that there are no major folds involving these three lithologies, at least on the scale of the area. The deformation appears to have been taken up internally within the psammite. The following stratigraphic units were recognised:-

F/G.19   PLOT OF F.2 FOLD AXES(o) AND POLES TO AXIAL PLANES(·)
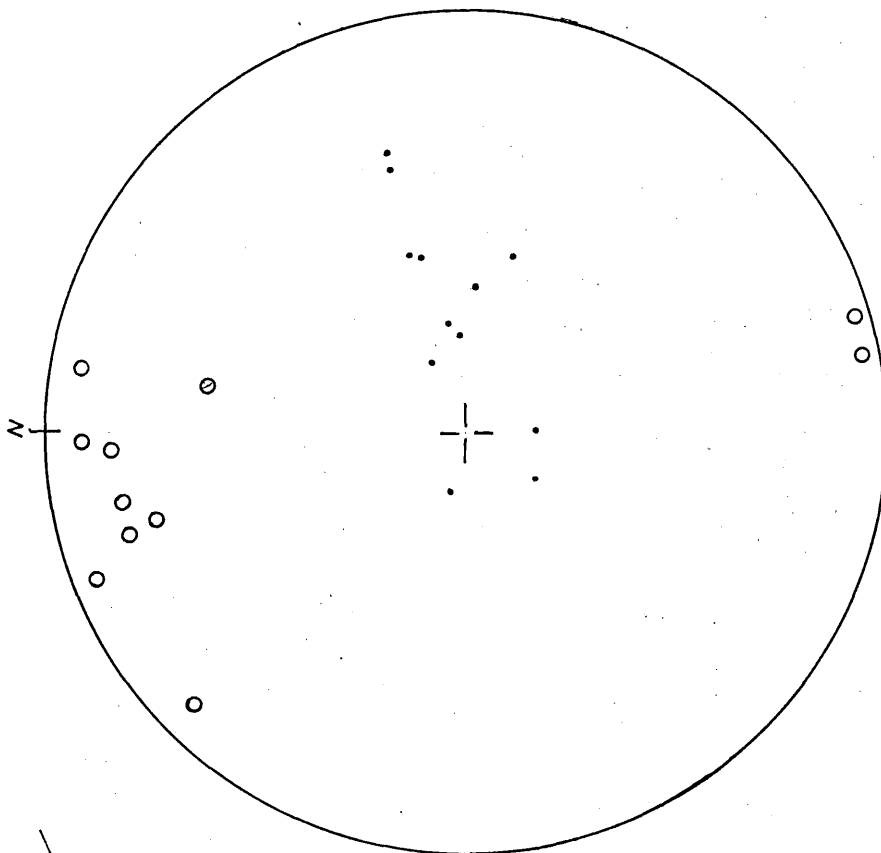
TROLLVANN PSAMMITE GROUP
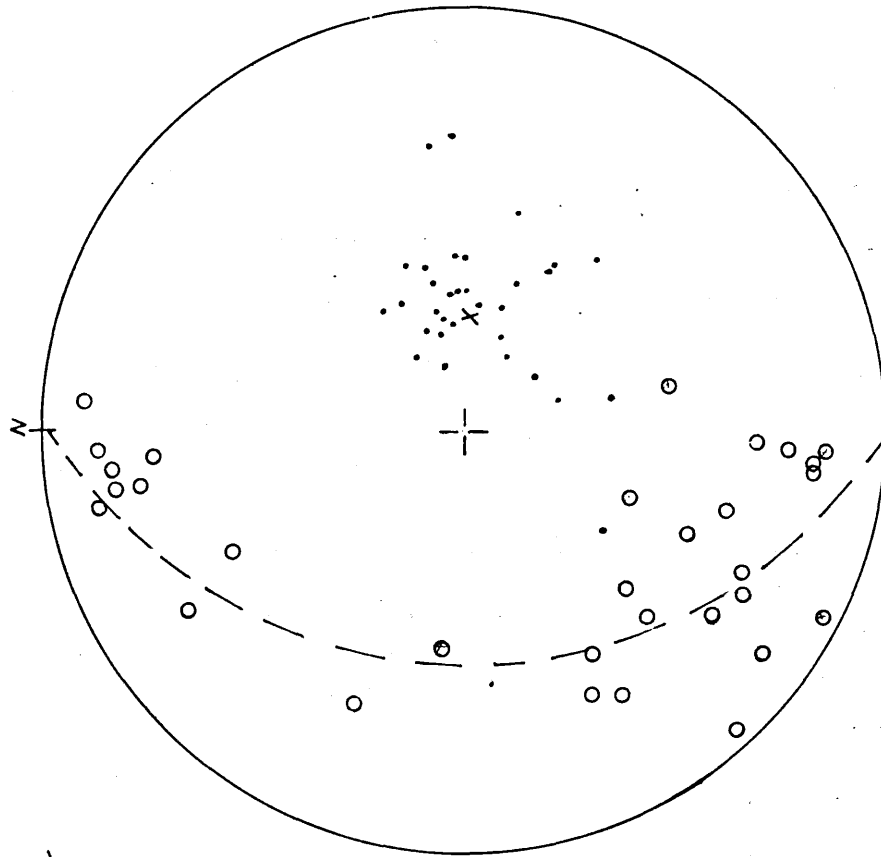
EIDVÅGEID SCHIST GROUP

Structural Top

3)   Buffish banded Psammite with pelitic intercalations and

quartzites.

2)   Grey Flaggy Psammite

1)   Transition Group.

Structural Base

No second phase fold structures have been recognised within the

Transition Group. The grey flaggy psammite unit also does not appear

to show many fold structures.

Most of the F.2 structures seem to occur within unit 3; the

buffish banded psammite with pelitic intercalations.

Examples of folds developing in this unit are given in Plate 74,

75 and 76. It can be seen from these folds appressed often with quite

considerable thickening at fold hinges and attenuation on the limbs.

Plate 75 shows once again, the tendency for fold structure to be

stacked one upon the other with westerly dipping axial planes. Plate 76

shows a folded and boudined quartzite band in a more semi-pelitic matrix.

Plates 77, 78 and 79 show examples of F.2 structures in slightly

more quartzite lithologies. Many of these folds tend to be somewhat

more gentle in style with a tendency towards a concentric type geometry.

They are rather more akin to the folds in the quartzite bands within

the Eidvågeid schist group. This applies particularly to Plate 77.

Plates 78 and 79 show rather more acute structures in flaggy psammitic

lithologies.

Figs. 19B shows the distribution of F.2 fold axes within this

lithology, and the attitude of F.2 axial planes. Once again the

tendency for F.2 axes to define a great circle is seen. F.2 folds have

been observed in these lithologies with curved axes. The great circle

Plate 74. F$_2$ fold, Trollvann Psammite Group, Trollvann.

Plate 75. F2 folds, Trollvann Psammite Group,
north of Vargsund.

Plate 76. F2 fold in Upper Flaggy Psammite of Trollvann
Group, north of Vargsund.

Plate 77. Concentric fold in Flaggy Quartzite.
Trollvann Group, west of Vasbugt.

Plate 78. F$_2$ folds in flaggy psammite, east of
Eidvaagenfjord.

distribution is therefore attributed to this phenomenon. The plot of poles to axial planes indicates that, as elsewhere, the axial planes have a reasonable constancy of attitude, with a moderate dip to the west. The great circle which represents the statistical average of the plot of poles to axial planes coincides with the great circle defining the locus of fold axes.

Thus, in the Trollvann Psammite Group we see a somewhat similar position to that seen in the psammites of the Komagnes Group.

1) Curving F.2 fold axes with a relative constancy of attitude of F.2 axial planes.

2) Fold structures with a geometry approaching that of the concentric model in the more massive quartzite horizons, in the psammitic and semi-pelitic lithologies the fold structures are rather more acute in style.

h) The Olderbugten Group

As in the Eidvågeid Schist Group fold structures do not seem to form extensively in this Group unless there is some rather more competent material present.

This may take the form of:-

1) Psammitic or semi-pelitic material within the schist.

2) Basic sheets, intruded generally parellel to the schistosity.

3) Granitic veins and sheets which have been intruded along the schistosity of the rock.

1) Plates 80 and 81 show examples of folds developed in semi-pelitic facies of the schist and facies containing distinct psammitic bands. Plate 80 shows a fold with a fairly gentle style. Plate 81 on the other hand shows some very intensely drawn out psammitic bands which have

Plate 79. F2 folds in flaggy psammite, east of
Eidvaagenfjord.

Plate 80. F_2 folds in the Olderbugten Group.

been folded by the second deformation. In the top centre right of the photograph, a fold is seen with what is apparently a conjugate-type geometry. This geometry is probably the result of a section cut parallel to the mean trend of a fold with a curved axis.

2)   Plate 82 shows an F.2 fold developing in the schist in the presence of a basic sheet. In the more pelitic facies of the schist where the lithology has not been felspathised, basic sheets appear to have the same effect as they do in the Eidvågeid Schist Group i.e. they stiffen up the schist sufficiently to allow it to fold in response to F.2.

3)   Plate 83 and 84 show a number of F.2 folds of granitic sheets which were formed at the acme of the regional metamorphism. These again stiffen up the schist sufficiently to allow it to fold in response to F.2. The structures are somewhat variable in their geometry though often they might be described as ptygmatic. The folds in Plate 83 show considerable thickness changes along the length of felspathic vein and a general tendency towards thickening in hinge zones. The veins before folding, however, may not have been of the same thickness along their lengths, so in this sense they are not reliable indices of the amount of deformation. It is apparent however, that in all cases there is a systematic variation of thickness along the veins, the thickest parts being the hinge zones and the thinnest, the limbs.

Plate 85 shows in an interesting way, the relationship between fold wavelength and layer thickness. The thick vein on which the hammer is resting has a rather large fold wavelength. The thinner vein, to the right of the photograph, has a rather smaller fold wavelength and the thinner vein again which passes off the photograph at the right centre has an even smaller fold wavelength. It is also interesting to note that the thin veins which are immediately adjacent to the thick one on the left, have taken on the dominant wavelength of

Plate 81. F$_2$ folds in the Olderbugten Group. Note
quartzite intercalations. Olderbugten.

Plate 82. F$_2$ fold in Olderbugten schist, flanks
of Eidvaagtinn.

Plate 83. F$_2$ folds in migmatised Olderbugten schists,
        Olderbugten.

Plate 84. F$_2$ folds in migmatitic Olderbugten schists,
        Olderbugten.

Plate 85. F$_2$ folds in migmatitic Olderbugten schists, Olderbugten.

Plate 86. F$_2$ folds in banded quartzite Olderfjord Group, Olderfjord beach.

the thick vein. This is presumably because they lie within the zone of contact strain of the thicker vein. In detail, however, many of them are contorted into little rucks having a very small wavelength. The wavelength of the thick vein has presumably been superimposed on the smaller rucks.

See later for the stereographic distribution of fold axes within the group.

i)   The Olderfjord Group

There is a considerable diversity of fold style within the psammites of the Olderfjord Group. It is very difficult to generalise. The folds shown in Plates 86, 87 and 88 which occur in a rather massive banded quartzite, are quite tight structures which show considerable thickness variations round individual layers. A characteristic of these structures is that frequently the tight isoclinal folds occur in zones between layers which show no folding at all. This is shown in Plate 86. It was originally thought in the field that these particular folds were of F.1 age, this being based solely on their general tight isoclinal nature. However, on close examination they could be seen to be refolding eyed-fold structures obviously of the first generation. Plate 89 shows some moderately tight folds in a rather similar lithology from the north of the area. In all the examples the layering was probably not a prominent feature during the folding, that is, the massive nature of the lithology would seem to suggest that interlayer contacts were, in fact, welded, precluding the possibility of interlayer slip.

A similar phenomenon to that mentioned above is seen in Plate 90. Here however, in strongly layered rocks. The fold involves a small amphibolite sheet, together with felspathised psammite. The layers

Plate 87. F_2 folds in quartzite. Olderfjord Group, Olderfjord.

Plate 88. F_2 folds in quartzite. Olderfjord Group, Olderfjord.

Plate 89. F2 folds in quartzite. Olderfjord Group, east of Honseby.

Plate 90. F2 folds in psammites interbedded with basic sheets. Olderfjord Group, east of Olderfjord.

above and below the fold appear to be completely unfolded.

Plate 91 and 92 show F.2 folds in other lithologies within the Group. Plate 92 shows a pegmatite vein, folded in with a thinly layered psammite. It would appear that the pegmatite was intruded, in the lower half of the photograph, transgressive to the layering, and in the upper half of the photograph parallel to the layering. This was followed by the F.2 folding.

Fig. 20A and 20B show the distribution of F.2 fold axes (20A) and F.2 axial planes (20B) within both the Olderbugten and Olderfjord groups. The pattern appears to be the same as that seen in the other groups, namely, a great circle distribution of fold axes together with a relative constancy of attitude of axial planes. These two Figs. refer to the area south of the strike swing which occurs in the north of the area. Reference to F.2 folds within the area of the strike swing will be made later. This great circle distribution of F.2 fold axes is once again attributed to curvature of the fold axes.

3) F.2 Linear and Planar Structure

These comprise

1) The fold lineations which have been discussed in the preceding section.

2) Shape and crystallographic alignment of minerals including planar orientation of micas.

For the purposes of convenience 2) will be discussed in two geologically defined areas.

a) The Trollvann Psammite and Structurally Lower Groups

The strong F.1 lineation, defined by elongate quartz segregations, has already been described in relation to the folds of the Komagnes Group. Frequently associated with this lineation, on banding surfaces,
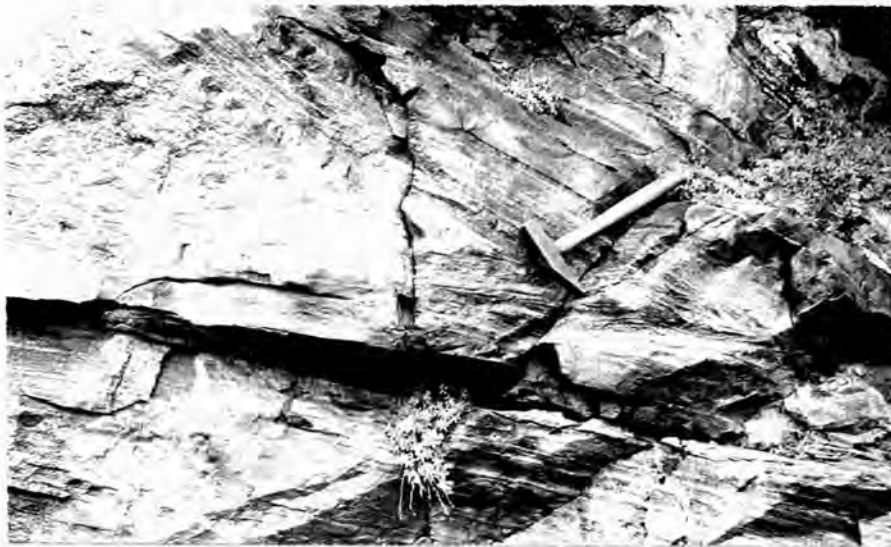
Plate 91. F$_2$ fold in the Olderfjord Group, east
of Hønseby.

Plate 92. F$_2$ fold in pegmatite and psammite.
Olderfjord Group, east of Storvann.

FIG 20   STRUCTURAL DATA FROM THE OLDERBUGTEN AND OLDERFJORD GROUPS

A/

FOLD AXES

B/
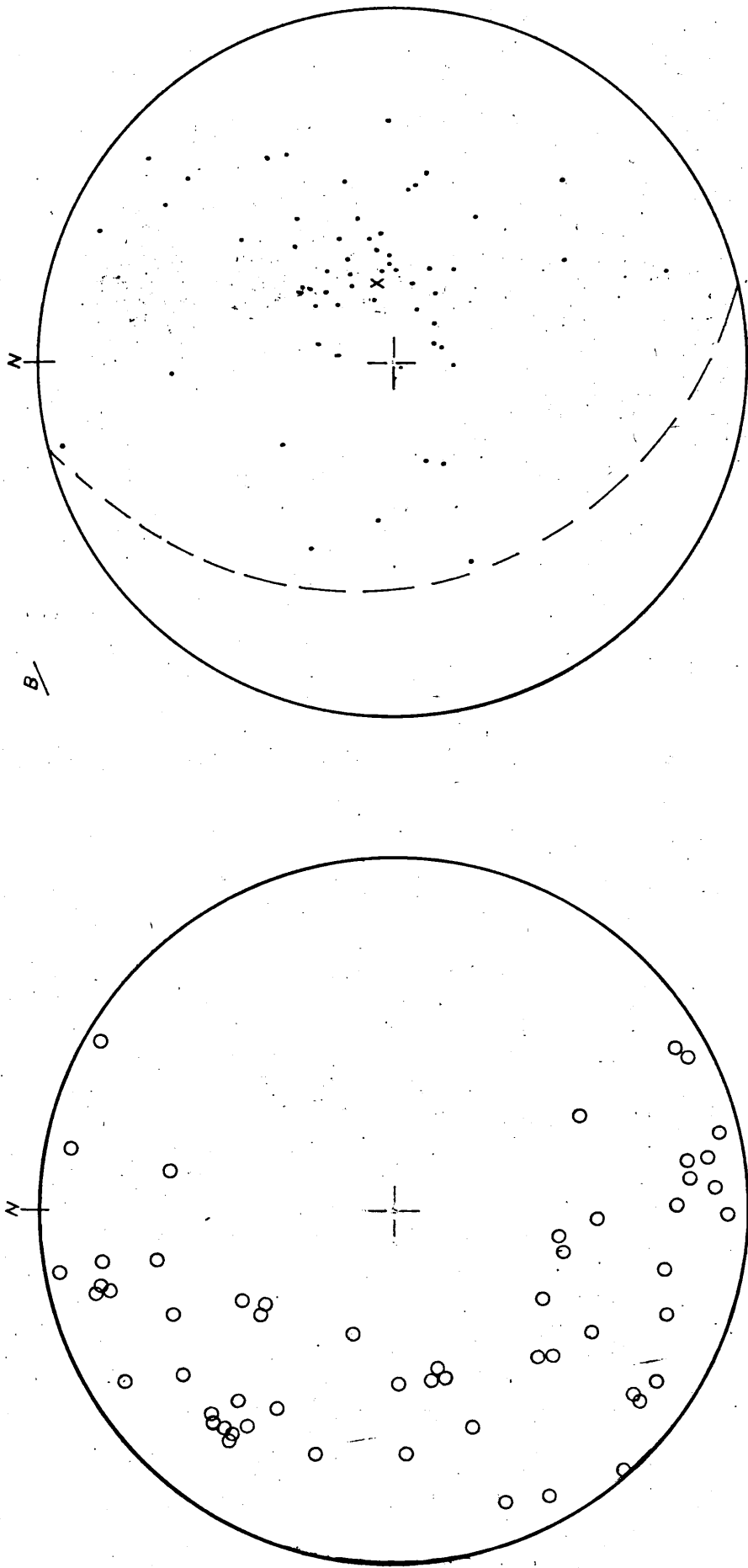
POLES TO AXIAL PLANES

are flexed biotites. Initially it was thought that this flexing was due to microfolding within F.2. However, close examination under a binocular microscope indicates that this flexing is due to bowing round the elongate quartz segregations, which have an elliptical cross-section. This effect could very well have been produced by flattening and need not have been associated with the F.2 folding.

Thin section and hand specimen examination of F.2 folds within the Komagnes group, particularly the structural top of the succession, indicates that the F.2 deformation was accompanied by some crystallisation of phyllosilicates parallel to the axial planes of the folds. Frequently a strain-slip cleavage develops involving the micas which are oriented axial-planar to the F.1 folds. However, it is clear that there is considerable crystallisation of mica axial planar to the F.2 folds. This mica can be seen in many cases to be responsible for the development of a lineation on the banding surfaces which are being folded. Since the axes of the folds curve, the lineation also appears to curve since it represents the intersection of a planar surface; the F.2 schistosity, with a curvi-planar surface; the banding surfaces folded by F.2. In Fig. 5A (P. 44) the tendency for a great circle distribution to occur in the orientation of lineations may be due in part to this phenomenon.

In the flaggy psammites, the intersection of these structural elements often finds expression in a rather hackly lineation which curves with the non-linear fold axes. It would seem that the curvature of the fold axis develops within the axial plane which itself appears to have a rather constant orientation.

On a regional scale this F.2 schistosity does not appear to be penetrative. In the Eidvågeid Schist Group, the small number of F.2

folds which have been recorded appear to fold the F.1 schistosity with very little development of an F.2 schistosity. It seems likely, therefore, that the F.2 deformation in the schist was taken up by slip along the F.1 schistosity, or by pure flattening, or both.

b)   The Olderfjord and Olderbugten Group

It has already been noted that there is an increase in the intensity of deformation from east to west in the area. This increase is related to two deformation phases other than F.2. Firstly a very intense pre F.2 flattening and secondly the tightening up of F.2 folds produced by the superincumbent weight of the plutons of western Seiland.

Examination of Fig.21 indicates a partial great circle distribution for the lineations measured in the two groups. This can be explained in one of two ways.

a)   The lineations are F.1 lineations which have been distorted by the subsequent deformation.

b)   They are related to a deformation phase other than F.1 i.e. the F.1 lineations have been obliterated and subsequent deformation has produced a new set of linear structures.

Examination of tight F.2 folds from the Olderfjord Group psammites, indicates that there is an axial planar schistosity associated with many of them. This is defined both by biotite laths and by flattened and elongate quartz crystals which seem to have undergone considerable extension in 'a'. This schistosity gives rise to a lineation on banding surfaces. It is, therefore, an intersection lineation related to the F.2 deformation. It has also been stated that the F.2 fold axes have a great circle distribution in the Olderfjord and Olderbugten Groups due to the non-linear nature of the fold axes. The intersection of this axial-planar schistosity with the layering folded by F.2 has given rise

FIG. 21

PLOT OF LINEATIONS FROM THE OLDERBUGTEN
AND OLDERFJORD GROUPS
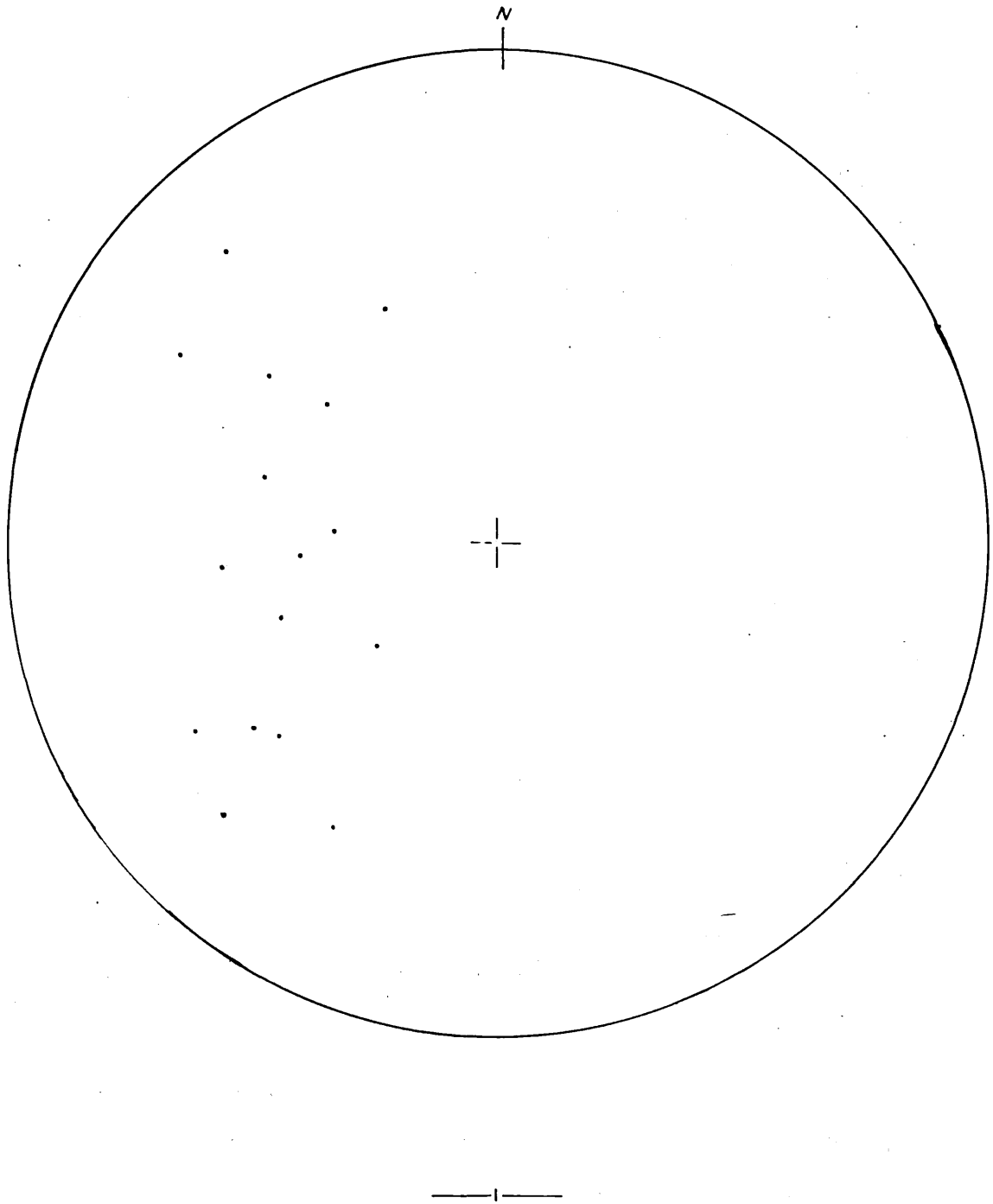
to the great circle distribution in Fig.21. Thus, the alternative b) is favoured overleaf for the origin of the lineations. The F.1 lineations were almost certainly obliterated by the intense pre-F.2 flattening.

Summary

1) In the lithologies structurally below and including the Trollvann Psammite Group, the dominant lineation is considered to be of F.1 origin. This structure has a generally constant plunge just west of north (Fig. 5A. P.44). The great circle distribution which tends to occur, is caused by the influence of a faint F.2 intersection lineation related to the curved axis folds.

2) In the Olderfjord and Olderbugten Groups F.2 and subsequent deformation has been stronger, giving rise to an axial planar schistosity in the psammites. This produces an intersection lineation.

4) Thrusting

Occurring within the psammites of the Komagnes Group are two thrust structures. The first and best exposed occurs across the neck of the peninsular at Komagnes (see map). The second outcrops on the beach near Jernelv.

The Komagnes thrust dips to the west roughly parallel to the average dip of the layering in the psammites. As the thrust area is approached the rocks become progressively more flaggy, and F.2 folds show evidence of flattening, see Plate 93. In the actual thrust zone, the psammite are extremely platey though they still show a relic of the strong local F.1 lineation. The zone of most intense deformation is some 2 metres thick, pelitic material along this zone, which has probably been utilised to some degree as a plane of easy slip, is very intensely deformed and boudinaged. (Plate 94). The long axes of these boudins appears to plunge down the dip of the psammites. Despite the obvious intense deformation, examination of thin sections of a platey psammite

Plate 93. F$_2$ fold near the thrust at Komagnes.
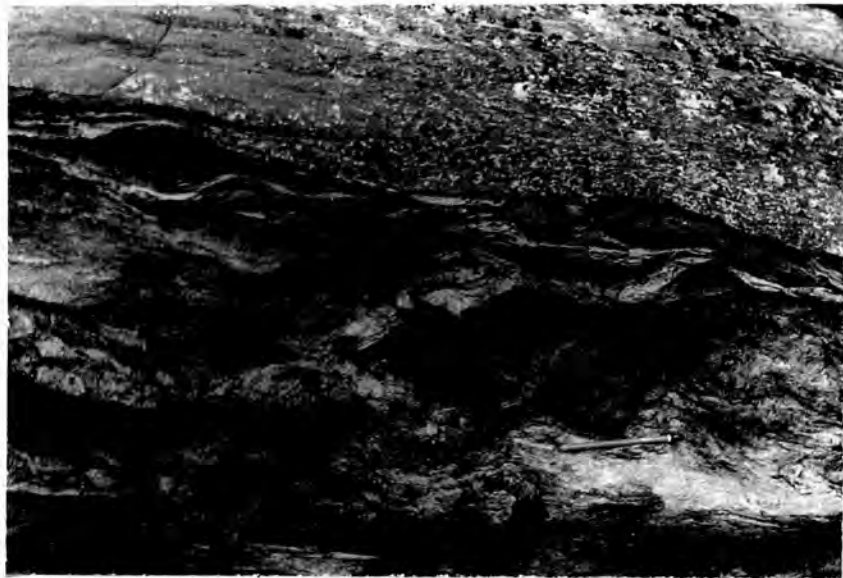
Plate 94. Intense boudinage in the thrust zone at Komagnes.

from the thrust zone do not show any cataclastic effects; the rock appears to have totally recrystallised. Neither is there evidence of mineral assemblages of lower metamorphic grade than those occurring in the psammites away from the thrust zone. It would, in fact, appear that prolonged recrystallisation took place under Middle Greenschist Facies conditions, post-thrusting.

To the east of the thrust zone occurs a concordant sheet of highly altered ultrabasic material. This is identical in composition and aspect to other bodies in southern Vasbugt and inland from Komagnes. This material becomes rather sheared close to the supposed thrust plane. It is not felt however, that it is genetically associated with the thrusting, although the highly chloritised lithology may have been utilised as a plane of easy slip.

It is rather difficult to gauge the sense and amount movement that has taken place on this structure. The rather thick zone of deformation suggests perhaps, that this has been substantial. Comparison of the rock types on either side of the thrust is helpful. On the eastern side a small zone of very strongly albitised metasediment is developed. The rocks in consequence are rather massive. The felspathisation phenomena recorded in thin sections of these rocks, are identical with those seen in the rest of the Lower Komagnes Group, though elsewhere they are not developed to such a strong degree. Apart from this one area, however, the rest of the rocks to the east of the thrust are in every way identical to those to the west. There is a slight difference in the strike across the thrust, this may perhaps be explained by rotational movements associated with the thrusting, or to the different response of the very massive felspathised rocks to the east, during F.2. The direction of movement on the thrust was probably approximately

east to west, this being normal to the strike of the plane.

The second thrust plane, which occurs in the Upper Komagnes Group at Jernelv, is marked by a rather thinner zone of deformation. Again the structure appears to be parallel to the average dip of the layering. It is almost certainly of similar age to the Komagnes thrust.

Regarding the age of this episode of thrusting, several lines of evidence seem to point towards it being a late F.2 phenomenon.

1) The total recrystallisation of the quartz fabric within the rock suggest an early date, similarly the absence of diapthoretic mineral assemblages. This recrystallisation is further suggested by a petrofabric diagram prepared for a platey psammite from the thrust zone (Fig. 22). This shows a tendency towards the development of an orthorhombic quartz fabric, identical to those seen in F.2 folds in the psammites from outside the thrust zone (e.g. III A-G and IV A-C in the petrofabric section). Thus, the rocks in the thrust zone appear to have recrystallised under an F.2 stress regime.

2) The layering of the Komagnes Group, including the platey layering in the psammites close to the thrust plane, is affected by a number of rotated boudins. It will be suggested later that these are related to the main later Caledonian thrusts which separate the high grade Caledonides from the Pre-Cambrian basement on the adjacent mainland (P.129). Thus, the thrusts discussed above appear to predate the main late Caledonian thrusting.

3) Evidence of thrusting on the minor scale has been recorded in association with F.2 folds elsewhere (see Fig. 10 P.54.). Here thrusting has occurred of the upper limb of an F.2 fold, along the layering of the lower limb.

It is suggested from these lines of evidence that this thrusting is a late F.2 phenomenon. It seems likely, therefore, that there has

FIG 22

POLES TO 0001 OF QUARTZ IN A PSAMMITE FROM THE THRUST
ZONE AT KOMAGNES

200 POINTS
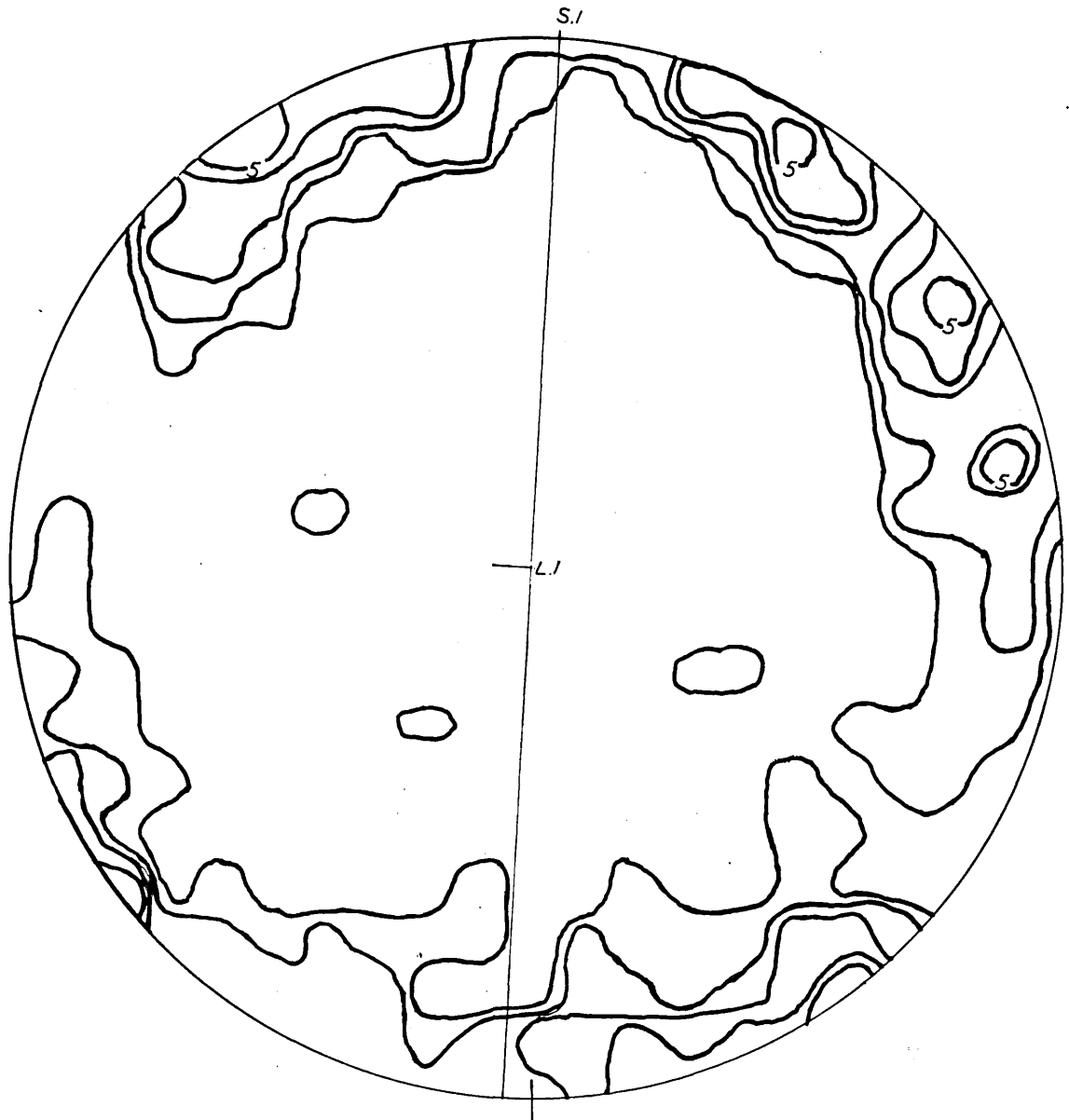
CONTOURS 5.4.3.2.1 % PER 1% AREA

SCHMIDT NET LOWER HEMISPHERE

L.I = F.I LINEATION

not been a very great deal of movement along the thrust planes since they are probably locally related to F.2 folds.

5)   Geometric Studies of folds.

Ramsay (1967) has suggested a method of classifying folds by measuring the thickness changes that occur in a single layer of any fold structure.

For example, in Fig. 23A the hinge thickness is measured; (to) then a series of tangents at varying angles of dip (a) to a tangent at the fold hinge, are constructed. The orthogonal thickness between tangents (ta) is measured for each successive angle of dip, thus a series of values are obtained which can graphically be used to represent the changes of thickness round the fold. On such a graph certain well defined special cases will occur, for example, for purely concentric folds the value (ta) will remain equal to (to) for all values of (a). For a similar fold (ta) is related to (to) as follows:

$$T = to \qquad T = \text{thickness parallel to AP of fold}$$
$$ta = Cos\ aT$$
$$\therefore\ ta = Cos\ ato$$

Thus, if a graph is plotted of $\frac{ta}{to}$ against (a) the special cases occur as well defined curves which can be used to delimit various fold classes.

FIG 23

A

t_0

T

t_d

α

α

T = THICKNESS PARALLEL TO AXIAL PLANE

B

A

t_0'

α

t_d

Y

X

t_d^2

t_0^2

C

B

C

A

B

a
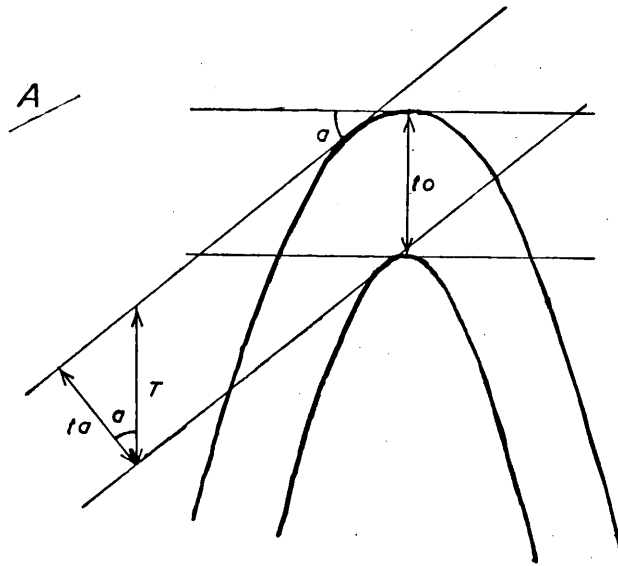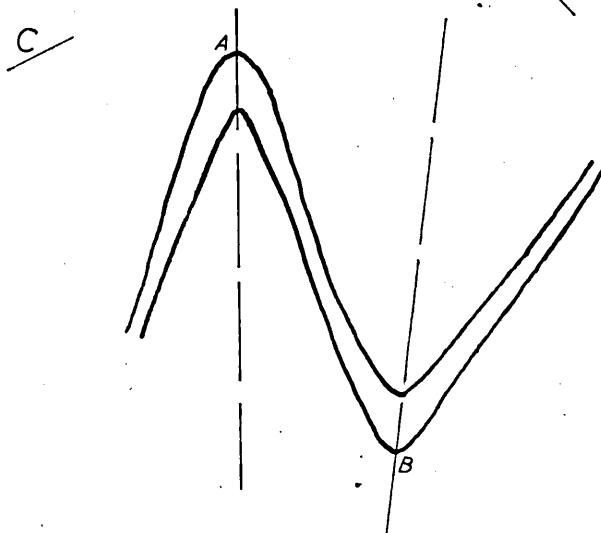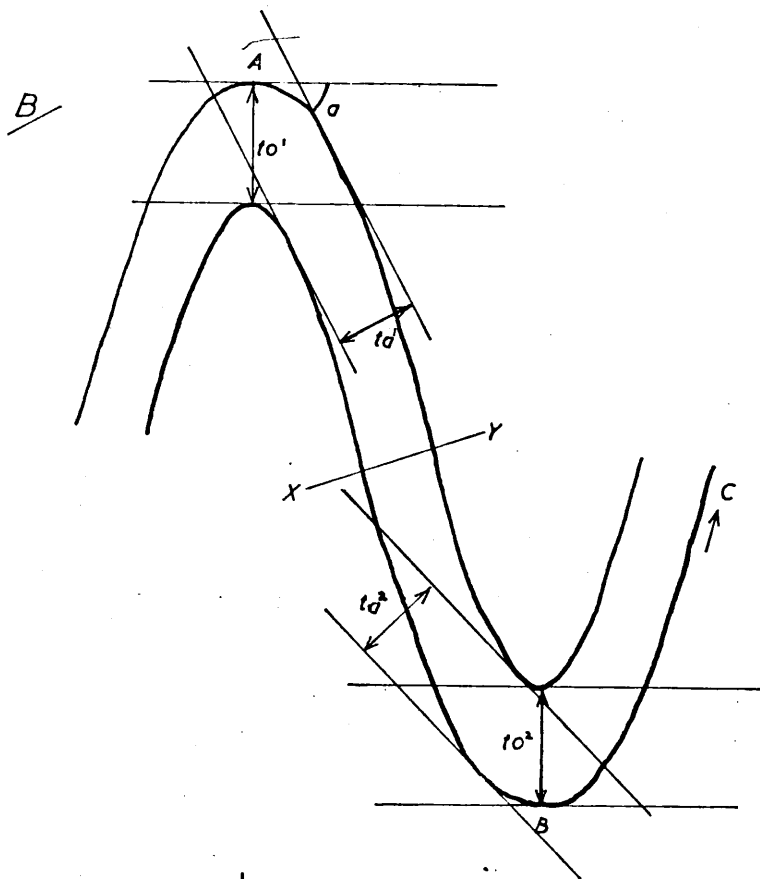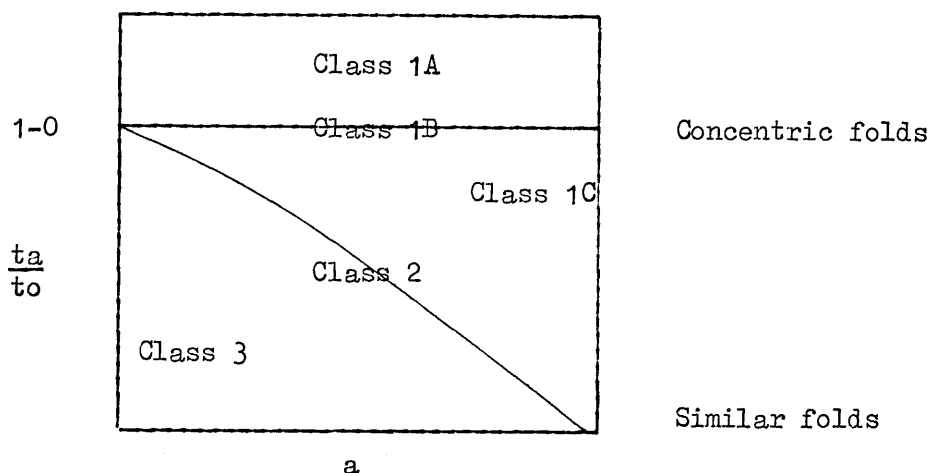
See Ramsay (1967).

Using this method, however, it is impossible to represent more then one hinge and adjacent limbs on a single graph because plotting the ratio $\frac{ta}{to}$ does not allow any relative thickening or thinning of adjacent hinge lines to show up, neither does it allow curves representing thickness changes between adjacent fold hinges to join up.

The author, therefore, has devised a system of plotting the absolute thickness (to) against the angle of dip (a). Consider Fig. 23B. A series of measurements of (ta) are made beginning at the hinge (A) for various values of (a). At some stage a maximum value of (a) is obtained. Measurement in a similar fashion, taking values of (a) at 10° intervals is then begun from hinge (B) back towards hinge (A). This process can be repeated from hinge (B) up towards hinge (C) and from (A) backwards towards the hinge to the left of (A). In this way a number of values representing thickness changes for a complete wavelength may be obtained and more if desired. These values may then be plotted on a graph.

Fig. 24 is an example of such a plot. The fold which represented in the accompanying diagram is a cylindroidal F.2 fold from Rastaby in the south of the area. The plot begins at point (A) on the left of the fold, point (B) represents the antiformal hinge at which the maximum thickness is obtained (a value = o), the plot then proceeds down the short limb of the fold until a maximum value of (a) is found (a max). Then plotting is resumed from the synformal hinge, (C) back up towards (B), the point (X) represents the junction of these two plots and gives the interlimb angle.
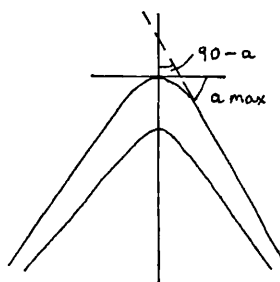
Assuming the axial plane, which contains (to), divides the fold into two halves. Since (a max) represents the maximum dip on the fold limb

$$\text{Interlimb t} = 2 \times (90 - a \text{ max})$$

Drawn on this diagram are two curves:-

1) Represents an ideal concentric fold. For this, values of (t) will always be the same as (to) for all values of (a); it is a straight line.

2) Represents an ideal similar fold. The curve for this is calculated quite easily. In Fig. 23A the value (T) represents the thickness measured parallel to the axial plane of the fold, between the tangents constructed for various values of (a). In a similar fold the value will always equal (to); the hinge thickness

FIG. 24

FOLD PROFILE

IDEAL CONCENTRIC FOLD

IDEAL SIMILAR FOLD

ANGLE OF DIP °

ORTHOGONAL THICKNESS t mm

From 23A

Thus, T = to

ta = Cos a x T

∴ ta = Cos a x to
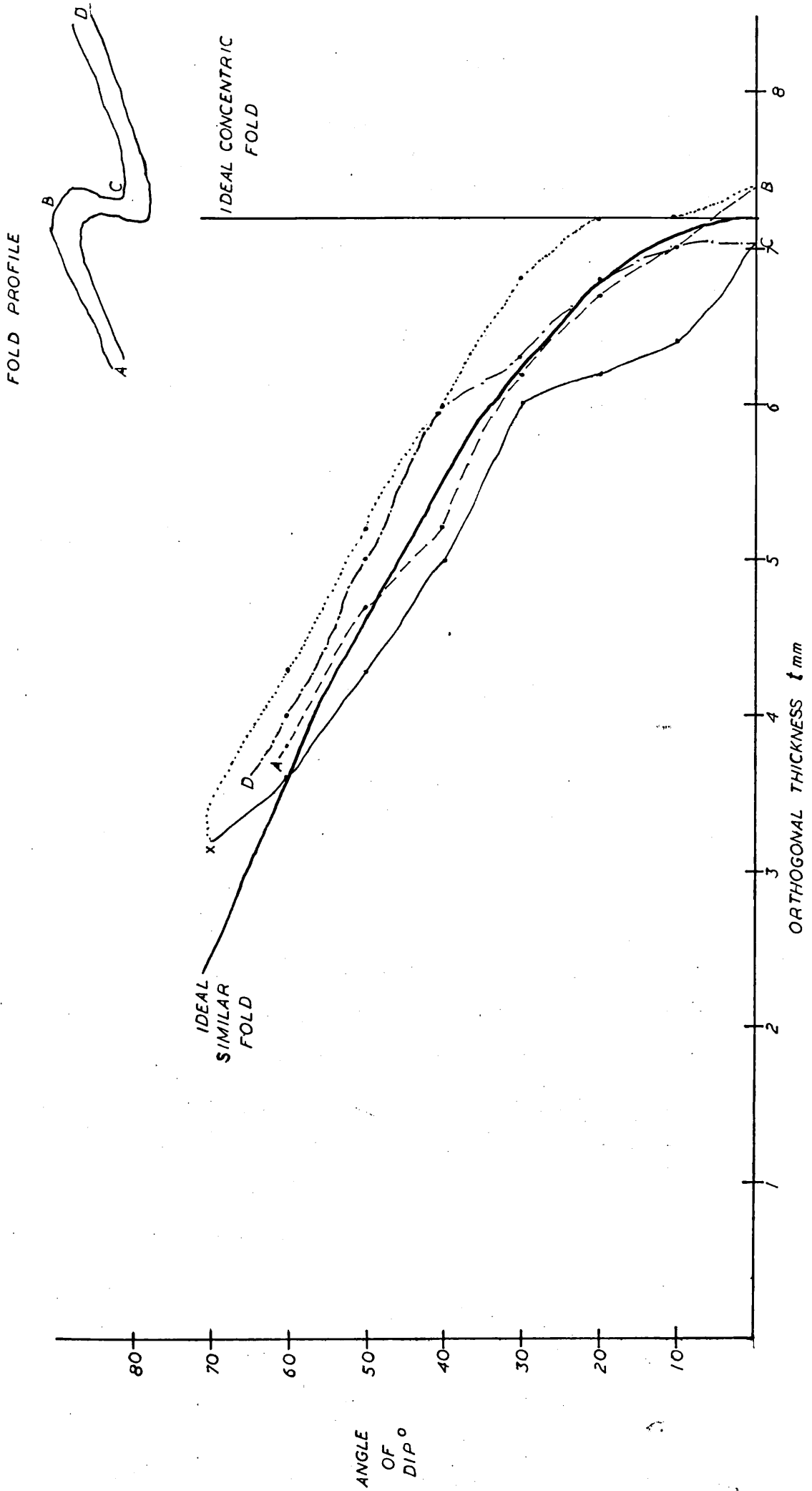
Thus, various values of (ta) can be calculated using values of
(Cos a) and a known hinge thickness (to).

It is obvious that the curves can only be drawn in the graph from
the hinge thickness values (to) i.e. on Fig. 24 point (B) or (C), but
since the plotted thickness curves (A-D) are close together, the ideal
curves are plotted from a point between the two values (B) and (C).

Examination of Fig.24 indicates that all the curves approach
quite closely to the similar fold model. Though, in fact, they lie
to both sides of this curve indicating that the limbs may fall into
either Class 1C or Class 3 of Ramsay's Classification, though only just
into either of these fields.

In Fig. 24 it can be seen that the antiformal hinge is slightly
thicker than the synformal hinge.

The interlimb t = $2 \times (90 - a\ max)^{\circ}$

= $2 \times 20^{\circ}$

= 40°

Discussion of the method

Advantages

1) Plotting absolute thickness rather than the ration $\frac{ta}{to}$ allows:-

a) Relative thickening or thinning of adjacent hinge lines
to show up on the plot.

b) The thickness changes round a number of hinges to be recorded
as a continuous line.

2) Two or more hinge lines can be represented on the graph in this way.

Disadvantages

1) The folds which have been measured in this study mostly have
thicknesses which are measured in millimetres. This presents problems
of measurement and accuracy.

2) With folds with fairly sharp hinges the first few readings e.g. for
(a) values of 0 - 40°, are all made in the hinge zone of the fold.
This is a problem that occurs in all methods of this kind.

3) Only fold systems may be measured which have the axial planes of
adjacent hinges parallel. This may be illustrated by considering
Fig. 23C.

Measurement is begun at hinge (A) down towards hinge (B) until a
maximum value of (a) is obtained. Then measurement is commenced from
hinge (B) up towards (A) but since the tangents to hinge lines are
non-parallel the values obtained from (B) are only apparent thicknesses
if one considers hinge (A) as a datum. The curves cannot, therefore,
join. This is probably the most serious drawback of the method.

Measurement of folds with curved axes

It has been mentioned that many F.2 fold axes curve in the hand
specimen and the curvature of adjacent hinges is often out of phase.
Folds of this type have been measured using the above method. It may
be thought that if adjacent hinges are curving out of phase any cut
which may be normal to one hinge may not be normal to the adjacent
hinge, therefore apparent thickening at hinge lines may invalidate the
plot. Consider Fig. 25A and B. These sketches represent sections of
folds in which the axes of adjacent hinges curve out of phase.

In   XYZ

$$\cos e = \frac{t^1}{tx}$$

$$\therefore tx = \frac{t^1}{\cos e}$$

Assuming a true hinge thickness of $t^1 = 1$

$$tx = \frac{1}{\cos e}$$

FIG 25

PLAN

LINE OF CUT

FOLD HINGE A

FOLD HINGE
B

$\neq 90$

t APPARENT $\cdot tx$

SECTION

A

t

B

tx
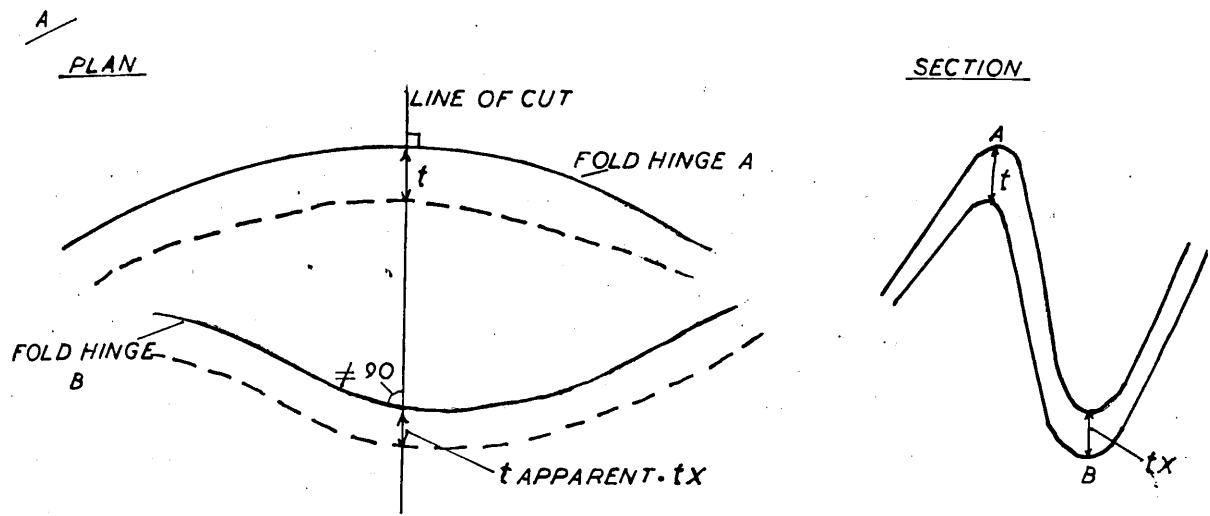
DETAIL OF HINGE B ALONG THE LINE OF CUT

LINE OF CUT

PLAN

HINGE B

t' θ tx

θ

m

n

SECTION

m

tx

n
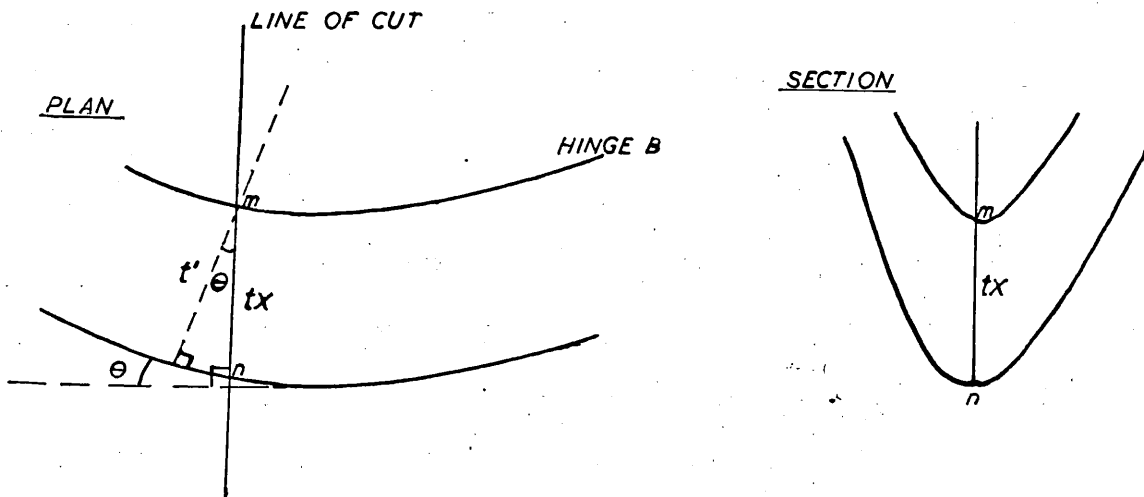
θ = DIFFERENCE IN ANGLE OF DIP BETWEEN HINGE A AND B

Let θ be for example 20°

$$tx = \frac{1}{Cos\ 20} = 1.064$$

Increase of thickness due to a 20° difference of hinge inclination = 0.064

$$\frac{0.064}{1} \times 100\% = 6.4\%$$

Thus, for a 20° difference only a 6.4% increase of thickness occurs. This can probably be ignored in any case values of (θ) are likely to be less than 20°.

Ramsay states (1967) that - "Similar folds have been envisaged in mathematical terms as parallel folds which have been subjected to an infinite compressive strain." In other words, of an originally concentric structure is subjected to flattening it will pass through, theoretically that is, a number of stages of modification until it achieves a similar-type geometry. Ramsay (1962) using actual measurements on naturally occurring folds, points out certain of the salient differences that these measurements indicate for folds of the so-called modified or flattened buckle fold model and the true similar fold model. For example, in the modified buckle fold the hinge thickness (to) is always the maximum thickness measured for that particular hinge and its associated limbs, and the thickness measured parallel to the axial - plane (Ta) is generally a minimum value for the modified buckle type. Conversely in the similar fold model the (T) value tends to be the maximum value obtained in the course of measurement.

Although the fold depicted in Fig.24 approaches quite close to the similar fold model, there is a general tendency for it to conform more to the flattened buckle fold model with a maximum (to) value and a minimum (T) value (see Table 2).

Table of T value for Fig.24   (Table 2)

(T) = thickness measured parallel to axial plane of fold.

a°	B-A (mm)	B-C (mm)	C-B (mm)	C-D (mm)
0	(T) 7.4	(T) 7.4	(T) 7.0	(T) 7.0
10	7.1	7.3	6.5	7.1
20	7.2	7.7	6.6	7.1
30	7.3	7.8	6.9	7.2
40	7.8	7.8	6.5	6.8
50	7.8	8.0	6.7	7.3
60	8.0	8.6	7.2	7.6
70		9.9	9.3	
80				
90				

The fact that 70° values of (a) were not recorded in (B-A) and (C-D), indicates that the fold is not perfectly symmetrical i.e. that the axial plane as drawn does not completely bisect the fold. This difference, however, is probably within acceptable limits. It is also notable that in two cases (C-B) and (C-D), the (T) values do not increase smoothly, the general trend however seems clear (see later).

Ramsay (1967) indicates a way in which the amount of flattening a previously concentric fold has undergone may be estimated. Consider Fig. 26. This is a plot of $\frac{ta}{to}$ against (a) the value λ 1 is the principal quadratic extension which the fold has undergone parallel to its axial plane direction and λ 2 that which it has undergone normal to this direction.

FIG 26
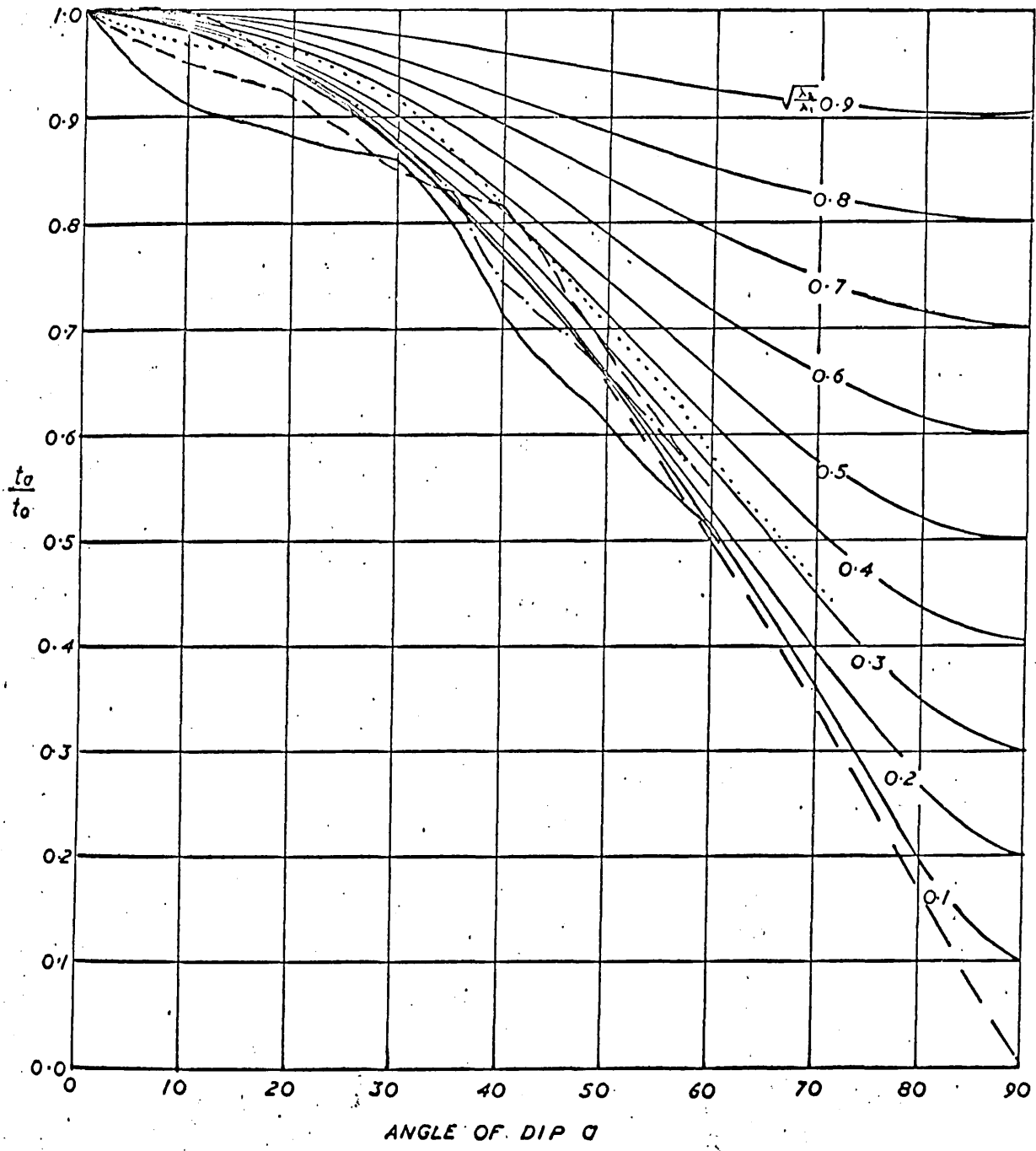
AFTER RAMSAY 1967

B—A — — — —
B—C ....
C—B ——————
C—D —·—·—·—

When λ_2 equals λ_1 the fold is a concentric structure. As λ_2 decreases the fold becomes closer to the similar fold model. Theoretically it finally achieves this at $\sqrt{\dfrac{\lambda_2}{\lambda_1}} = 0$. The point of the diagram is that it indicates the amount of flattening that the original buckle has undergone. For a more complete mathematical treatment the reader is referred to Ramsay (1967).

Fig. 26 is a plot of the limbs of the fold depicted in Fig. 24. If this idea is valid it can be seen that the fold has undergone considerable flattening, since it occurs very near to the similar fold curve $\sqrt{\lambda_2 / \lambda_1} = 0$. It is clear the lines are rather sinuous but once again the trend is clear.

Let us now consider some other examples:-

Fig. 27

This structure is once again an F.2 fold from Rasteby in the south of the area, the lithology is a pink psammite. The structure has a weak axial planar foliation defined by muscovites. This gives rise to an intersection lineation on the banding surface. The fold axes are curved so therefore is the lineation which is rather hackly. The fold measured is shown as profile in the sketch accompanying Fig.27. The diagram shows:-

1) The synformal hinge is considerably thicker than the antiformal hinge.

2) The lines (II-I) and (II-III) conform very closely to ideal similar fold curves which have been drawn in this case from the synformal hinge value. The lines (I-II) and (II-III) also approximate to the similar fold model, though ideal similar fold curves are not present. On Ramsay's classification the folds are generally of Class IC.

3) Interlimb L approx. $= 2 \times (90 - a\ max)^{\circ}$

$" \qquad = 2 \times (90 - 80)^{\circ}$

$" \qquad = 20^{\circ}$

FIG 27

ORTHOGONAL THICKNESS t_{mm}

$a°$

IDEAL CONCENTRIC FOLD

IDEAL SIMILAR FOLD

FOLD PROFILE
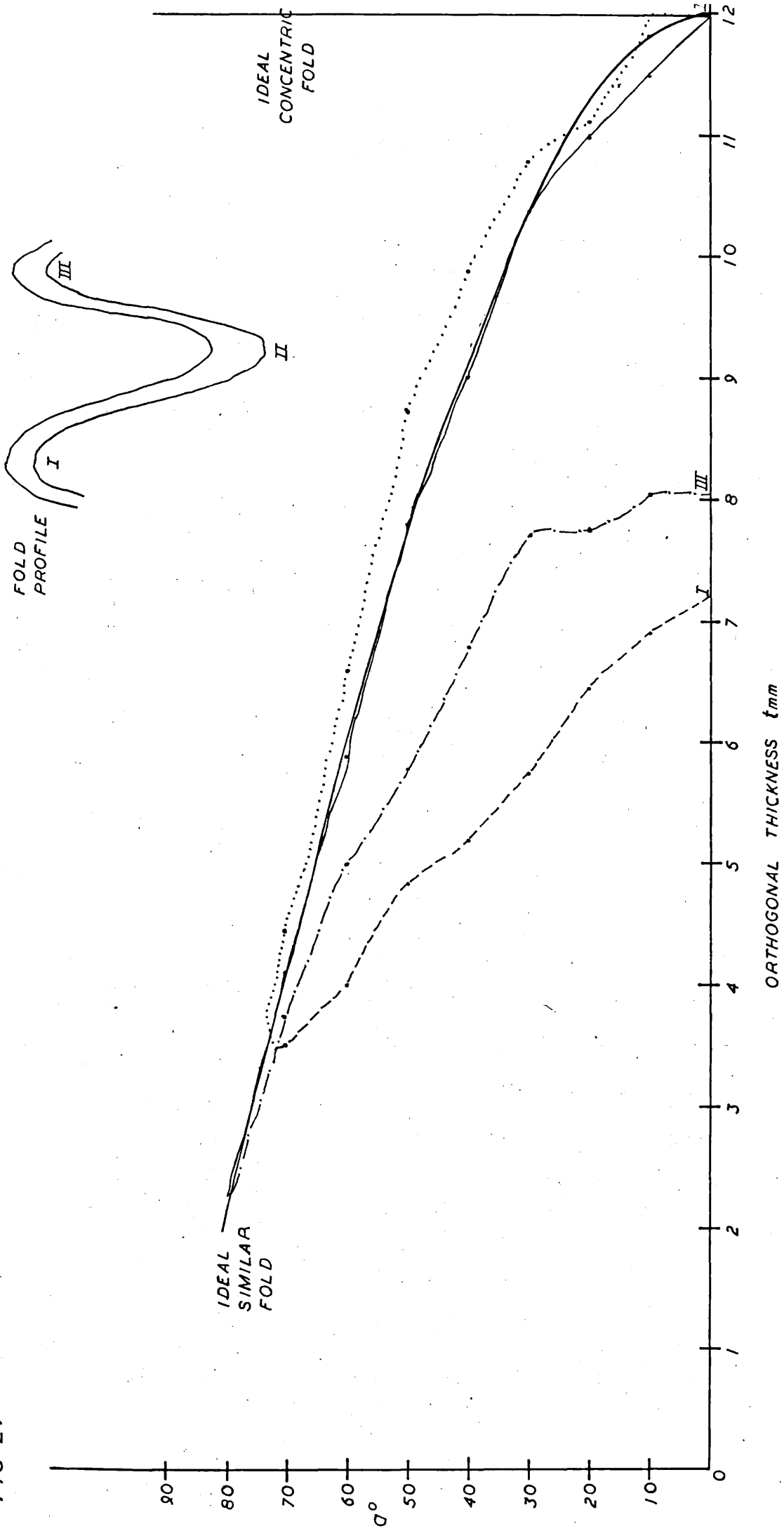
Table 3

T value for fold depicted in Fig. 27

a^o	Section (I-II) (mm)	(II-I)(mm)	(II-III)(mm)	(III-II)(mm)
0	(T) = 7.2	(T) = 12.0	(T) = 12.0	(T) = 8.1
10	7.0	12.2	11.7	8.2
20	6.9	11.8	11.7	8.3
30	6.6	12.5	12.0	9.0
40	6.8	12.9	11.8	8.9
50	7.6	13.6	12.2	9.0
60	8.0	13.2	11.8	10.0
70	10.2	13.2	12.0	11.0
80			13.0	13.0
90				

These figures show in a more consistent manner a trend which was seen in the (T) value for the fold in Fig. 24, i.e. an initial decrease in the value of T from the hinge line, then a pronounced increase of (T) to a maximum value at (a) = 70 or 80°. This means that initially for a value of say $0-20^o$ the fold is of Class 3. Then at about a = $30-40^o$ the fold changes to the Class IC model. The combination of these two geometric types probably produces the closeness to the similar fold model. This change is reflected in Fig. 27 by the change in orientation of the curve (II-III); for low value of (a) it is below the similar fold curve, i.e. Class 3. Then at a = 30° it crosses the line and from thence remains very close to the similar fold curve. The trend is seen more clearly for the curve (I-II) which after crossing the similar fold curve at (a) = 22° remains above this curve, i.e. it conforms to a Class IC geometry.

Fig. 28 shows a plot of the various curves on a plot of t/to

FIG 28

AFTER RAMSAY 1967

ANGLE OF DIP σ

$\frac{t_a}{t_o}$

$\sqrt{\frac{\lambda_1}{\lambda_3}}$ 0·9
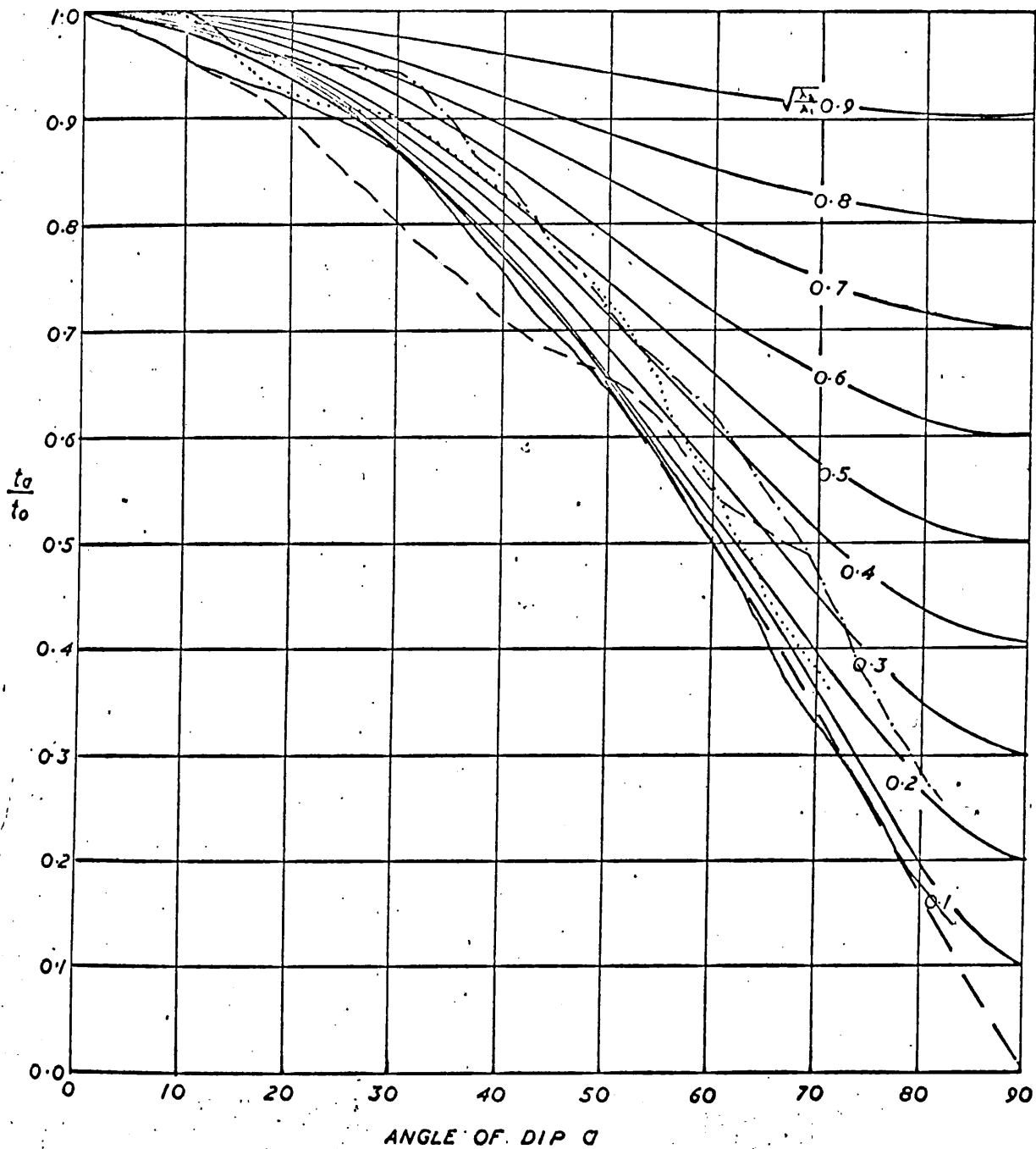
I–II	——
II–I	·········
II–III	— — —
III–II	—·—·—

against (a) diagram. Here the Class 3 geometry is again reflected

for low value of (a), by the tendency for the lines to be situated

below the similar fold curve of $\sqrt{\dfrac{\lambda_2}{\lambda_1}} = 0$. This tendency is also

shown in Fig. 26. When the lines cross the similar fold curve

they indicate that a considerable amount of flattening has taken

place, if the folds are assumed to have been originally concentric.

Fig. 29

This fold is an F.2 fold from Rastaby. The lithology is mixed,

it consists of an interbanding of quarztitic layers, semi-pelitic

layers and pelitic layers. The pelitic layers bear a strong axial

planar foliation. In the core of the fold concerned an elasticas-

type structure occurs in a quartzite band. The fold axes are curved

and bear a faint lineation. The inset in Fig. 29 shows the fold

concerned.

Examination of Fig. 29 shows:-

1) Once again the synformal hinge is considerably thicker than

   the antiformal hinge.

2) The lines (II-I) and (II-III) are probably close to the similar

   fold model whereas the curves (I-II) and (III-II) are not, in

   fact, these latter two curves are very irregular in thickness.

This irregularity can be attributed in part to the uneven nature of

the banding surface (see the fold profile); the curves are Class IC,

the other curves are also Class IC, but approach more closely to

the similar fold model.

3) The interlimb angle approx. =

$$\text{"} \quad = 2 \times (90 - max)$$

$$\text{"} \quad = 2 \times (90 - 70)$$

$$\text{"} \quad = 40^{\circ}$$

FIG 29

ORTHOGONAL THICKNESS t mm
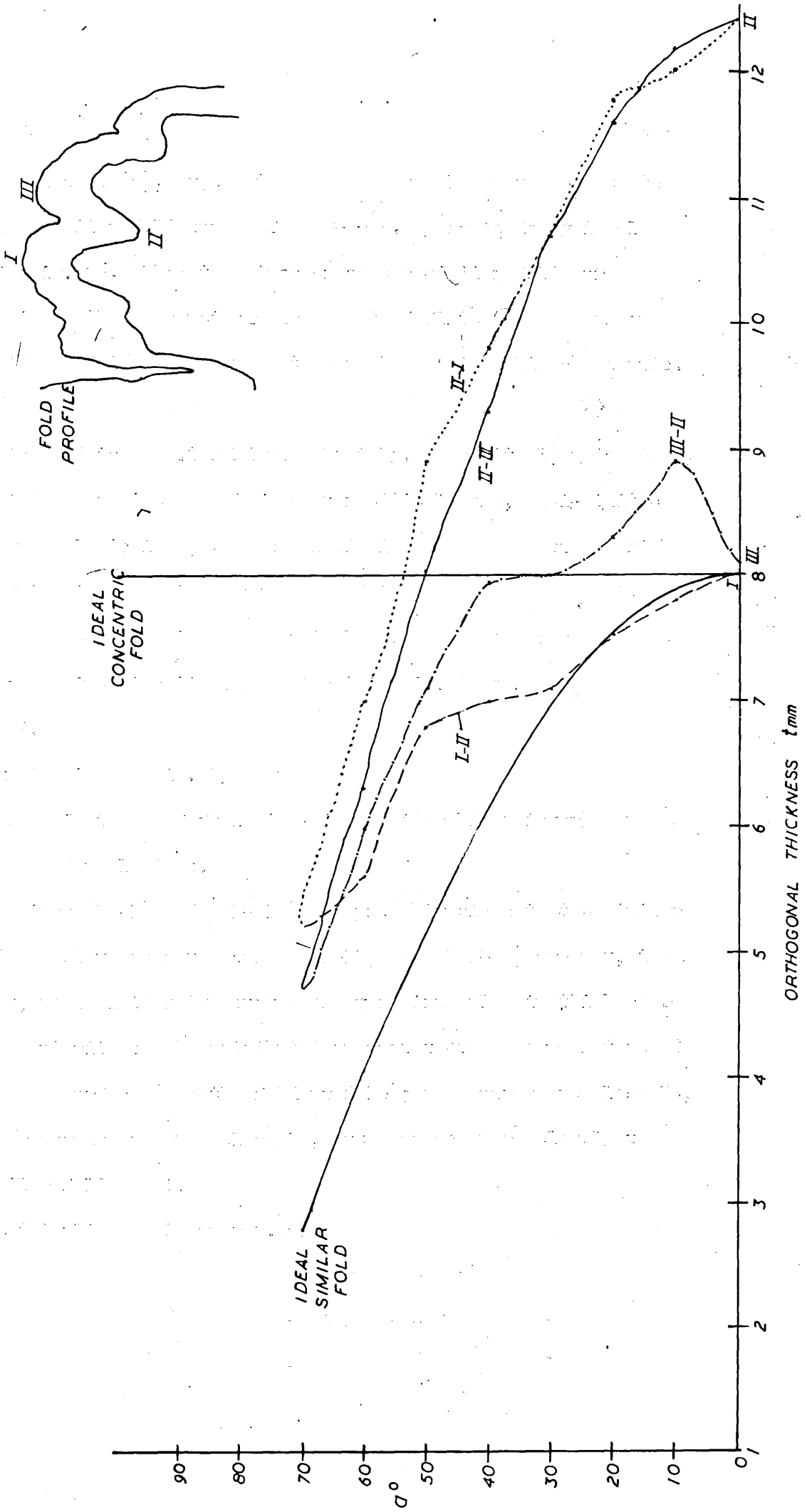
Table 4

(T) values for fold in Fig. 29

a^o	Section (I-II)	(II-I)	(II-III)	(III-II)
0	8.0	12.4	12.4	8.1
10	7.9	12.2	12.4	9.0
20	8.0	12.6	12.3	8.8
30	8.4	12.4	12.4	9.2
40	9.1	12.8	12.1	10.4
50	10.6	13.9	12.4	11.0
60	11.2	14.0	12.6	12.0
70	15.2	15.2	13.7	14.0
80				
90				

These figures show once again a slight tendency for the folds to be of Class 3 near the hinge. However, away from the zone the folds are more clearly of Class IC. The curve (III-II) shows this clearly. Here there is no tendency for a decrease then an increase in the (T) value to occur.

The curves (II-I) and (II-III) can be seen, however, to be close to the similar fold model. Though there is an overall increase in the (T) value it is less marked than in the other two curves.

These facts are again seen in the Fig. 30. It would seem from this diagram that the various parts of the fold have undergone

FIG 30
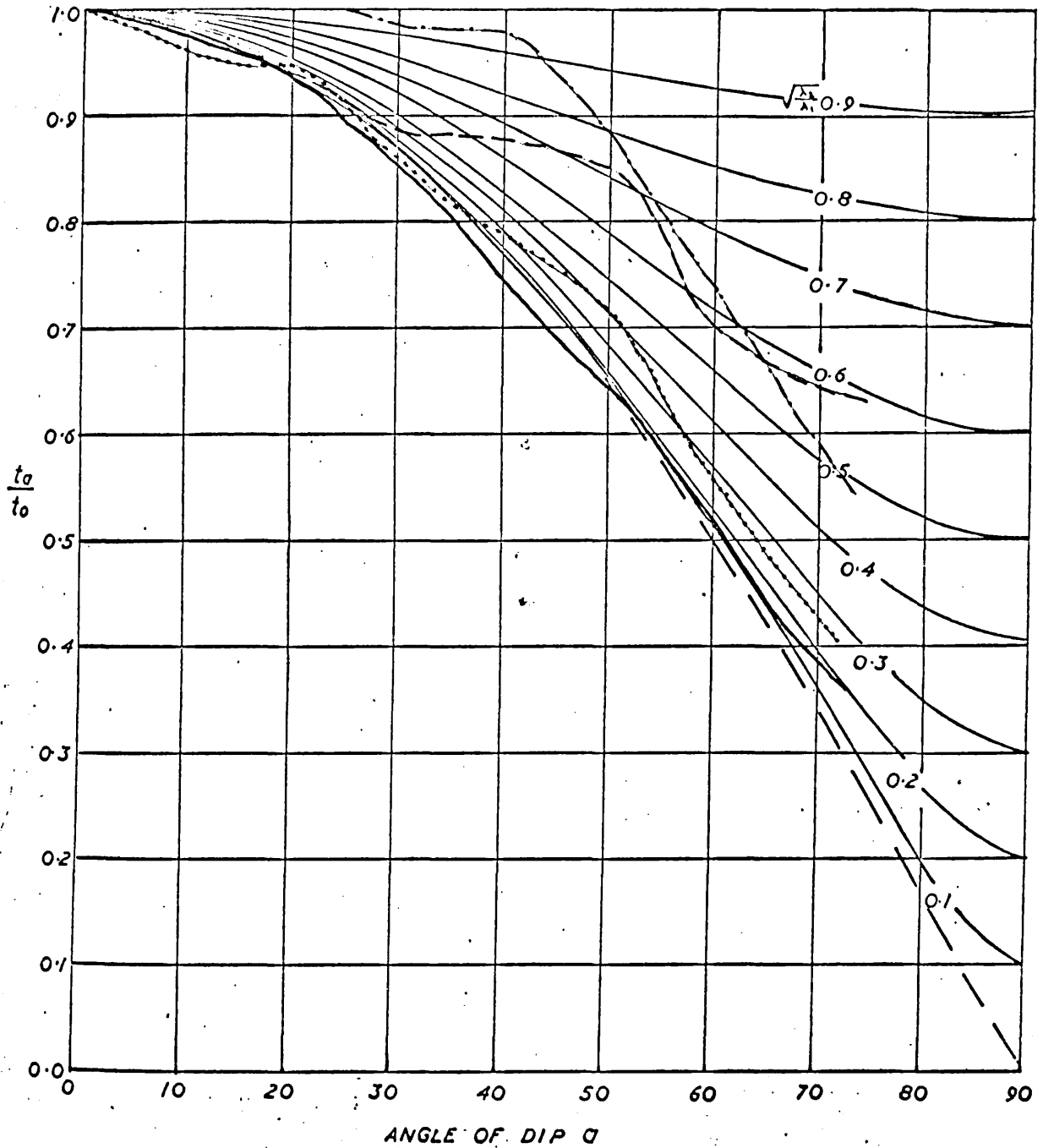
AFTER RAMSAY 1967

different amounts of flattening. It is notable that the curves
(I-II) and (III-II) appear to have undergone the least flattening,
they represent analagous parts of the fold, i.e. the plot from each
antiformal crest downwards towards the central synform. The plots
from the central synform upwards towards each adjacent anticline
however, appear to have undergone the most flattening. This reflects
a greater extension in 'a' * for the synformal hinge than for the
antiformal hinge. Examination of the rest of the fold, however,
indicates that this is not a consistent feature of the fold as a
whole. This extension in 'a' must have been accompanied by some
flowage of material from the flanks of the fold into the synformal
hinge zone, this to account for the increased thickness of this area.
In other words, the attenuation on the limbs and the relative thicken-
ing at the hinge zone is not due to pure extension in 'a' but must
also have been accompanied by some flowage within the layers. This
statement makes the assumption that:-

1) The original thickness of the layer measured was equal along
   its length.

2) There was little differential extension between anticlines and
   synclines in 'b' which could possibly account for the increased
   thickness at the synformal hinge, i.e. to greater flowage of
   material out of the anticlinal hinges (see later).

---

* Use of the term 'a' and 'b' above. In this sense 'b' is the
direction parallel to the fold axis and normal to the plane of
the diagram. 'a' is the direction normal to this lying in the
axial plane of the fold.

Fig. 31

The fold depicted in this graph is an F.1. fold from a psammite rib within the Hønseby gabbro; the fold has a strong axial planar foliation defined by biotite. In this case it was only possible to plot one limb of the structure. The graph shows that the fold has a consistent Class IC geometry, though this approaches quite closely to the similar fold model. (T) values show an almost consistent increase in value to a maximum at (a max).

Despite the very strong axial planar foliation and the superficial geometry of this structure, which might lead one to suppose that it was a similar fold, it clearly has a Class IC geometry.

Table 5

(T) values for Fig. 31

$(a)^o$	(T) value m.m.
0	8.3
10	8.4
20	8.5
30	8.4
40	9.1
50	9.3
60	10.8
70	11.7
80	
90	

Summary and Discussion

a) Many of the folds measured in this study have a Class 3 geometry

FIG 31

IDEAL CONCENTRIC FOLD

FOLD PROFILE

IDEAL SIMILAR FOLD

ORTHOGONAL THICKNESS $t\,mm$

a^0
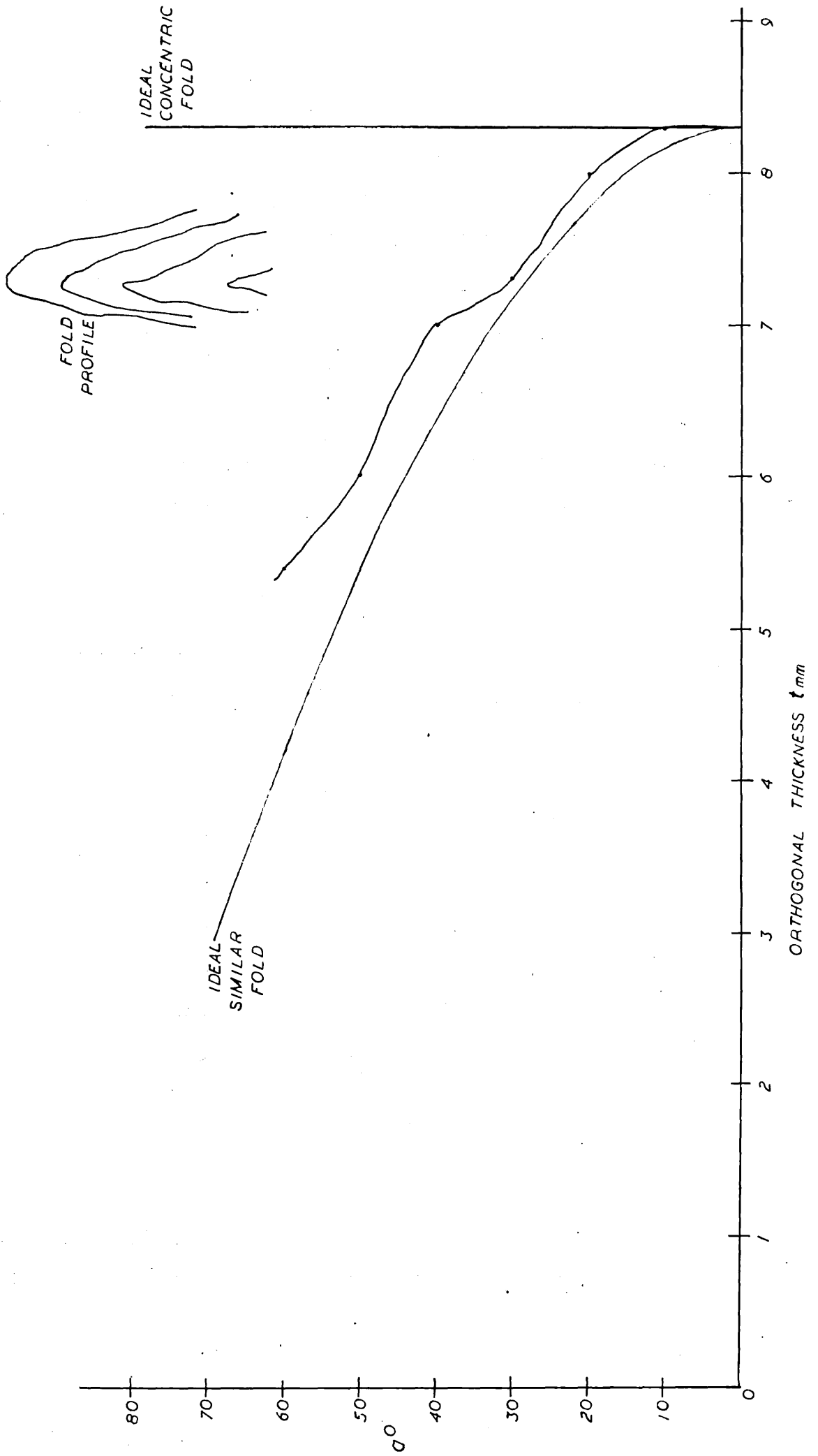
in the hinge zone. This changes for higher value of (a) to a Class IC geometry. This change of geometry probably produces the closeness to the similar fold model.

b) If the folds are treated as flattened buckle folds they appear in most cases to have undergone a considerable degree of flattening. This may not, however, always be consistent within the fold, e.g. the fold depicted in Fig. 29, here different parts of the fold have apparently undergone different degrees of flattening. From these observations it was suggested that both extensions in 'a' and flow-age within the layers was responsible for the fold form. Since relative thickening of synformal hinges is a common feature, this process was probably operative in all the folds measured.

c) The F.1 fold depicted in Fig. 31 although it superficially appears to be of similar type, it has in fact a Class IC geometry. This underlines one of the dangers of the use of the term "similar fold". Because of the strong axial planar foliation and superficial geometry of the F.1 structure in Fig. 31, there is a tendency to say that the fold is a similar fold, implying a mechanism of formation. This mistake must frequently be made in the field. Similar folds sensu-stricto must be extremely rare in nature, the term therefore should have a strict geometric definition with no undertones of mechanism implied. On this basis the most frequently used description would be that "folds approach a similar geometry". Mechanisms would have to be discussed using other criteria.

6) Petrofabric Studies

Petrofabric diagrams of poles to 0001 of quartz were prepared for a number of folds of both generations (F.1 and F.2). In all

cases the diagrams were prepared from plots on the Lower Hemisphere of a Schmidt net.

Specimen I. N.17 34D.

This is the F.1 fold with the strongly curving axis depicted in Fig. 6. From this specimen a number of slides were taken (see Fig. 6) each normal to the axis of the fold at that point.

Examination of each section indicates that the quartz is generally quite equidimensional, very little tendency for elongation in the schistosity is shown. The mineral is only very slightly strained. The equidimensional nature of the quartz is almost certainly due to recrystallisation after the F.1 deformation. The slight straining may well be a later effect. The rock has a schistosity (S.1.) defined by biotite crystals which show a general tendency to be axial planar to the fold. The orientation of these biotites has been disrupted to some degree by recrystallisation, though an axial planar fabric is still clear. There is a little garnet in the rock which overgrows this schistosity and also appears to overgrow the recrystallised quartz fabric. Later flattening has produced slight bowing of the schistosity round these garnet porphytoblasts.

The contoured petrofabric diagrams are given in Fig. 32 A-F. They show a number of features.

1) Cleft 'ac' girdles (the term 'ac' is in a sense inappropriate here since due to the curvature of the fold axis the 'a' directions represented in each slide are non-parallel). These 'ac' girdles have an asymmetric distribution of maxima within them.

2) There is a general tendency for a weak cross-girdle to develop,

FIG 32   0001 QUARTZ. EACH DIAGRAM 200 POINTS   (SEE FIG 6)
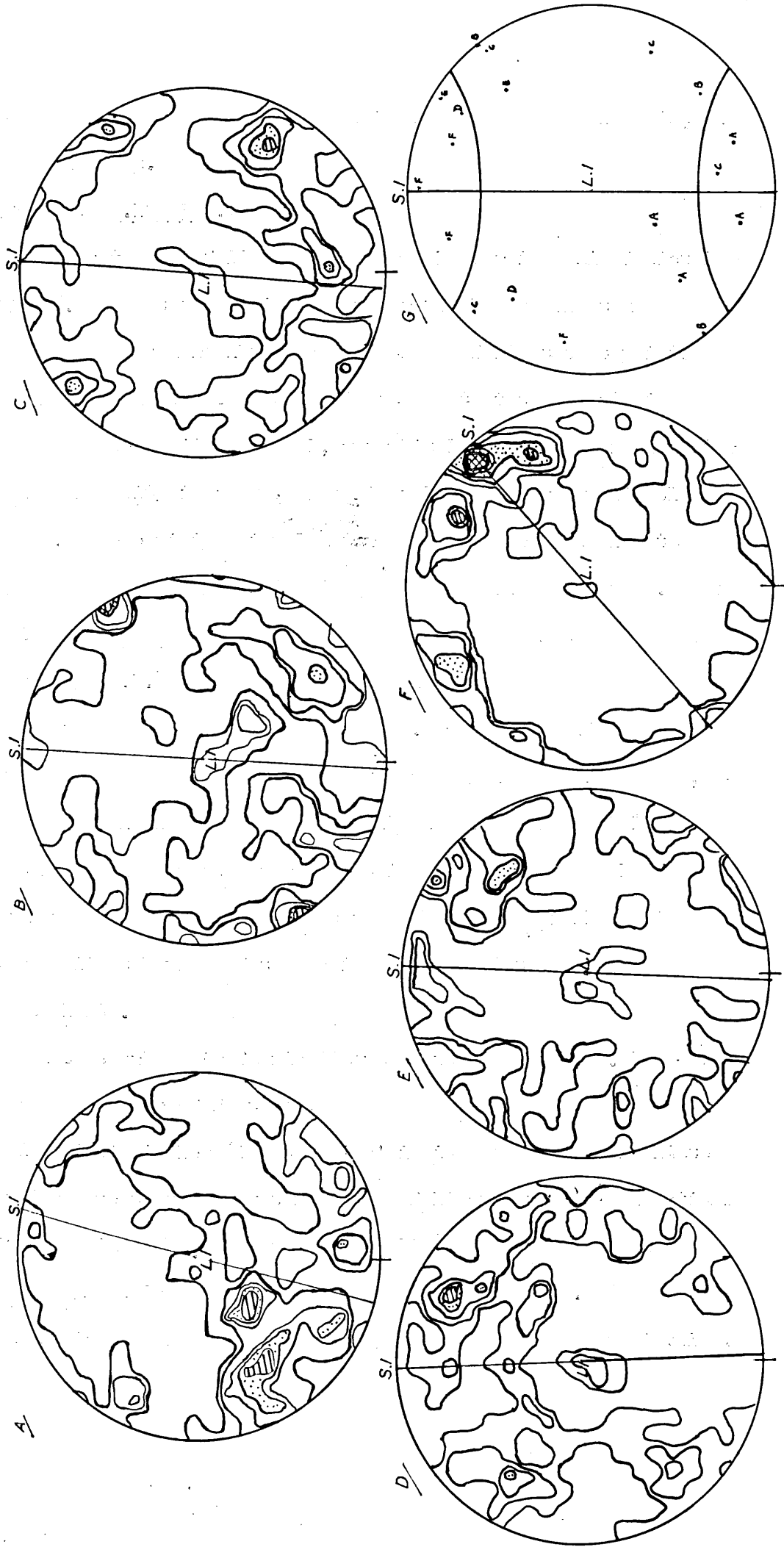
CONTOURS 7 ⬚ 6 ■ 5 ▨ 4 ▦ 3.2.1 % FER 1 % AREA   SCHMIDT NET LOWER HEMISPHERE

the axis of which is at a high angle to the axis of the peripheral

girdle. They are B^B tectonites. In some cases, e.g. B there is

a tendency to develop more than one cross-girdle.

3) The overall symmetry is triclinc though if the cross-girdles

are ignored there is a tendency for orthohombic symmetry to develop

with Sl as a symmetry plane, e.g. (B C and E).

4) (F) shows a relatively very strong concentration (70%) aligned

in (Sl). This slide also has a slightly greater preferred dimen-

sional orientation of the quartz grains than the others. The grains

however do not seem to be any more highly strained than those in the

other slide. Although this particular orientation pattern occurs

near the point of inflection of the curved axis, the absence of any

semblance of this type of orientation in (D) and (F) seems to preclude

the argument that this axial planar orientation of quartz 'c' axes

is due to shear folding (Gangopadhyay and Johnson (1962, see later).

The author thinks that this orientation pattern is a localised

phenomenon due possibly to the fact that this particular slide was

taken from a position close to the boundary between the upper and

lower antiforms. This plane has a skin of chloritic material on it

and has been a plane of some movement subsequent to F.l.

Fig.(32.G) shows a synthesis of the position of all 4, 5, 6 and

7% maxima relative to a constant plane, i.e. the plane of the 'ac'

girdles. 40° small circles have been added for the purposes of

orientation.

Some interesting features emerge from the diagram:

a) There is a tendency for the maxima from Diagrams (A, B & C), i.e.

these form the lower antiform (Fig. 6) to be clustered at the bottom

of the diagram whereas these from the upper antiform represented

by diagrams (D, E and F) are clustered at the top of the diagram.

b) The 40° small circles are the approximate locus of the points

representing these maxima.

These features are extremely difficult to explain. A survey

of the literature seems to indicate some confusion concerning the

origin of 'ac' girdles. Similarly, the invoking of certain crystal-

lographic directions in quartz as planes of glide or fracture has

elicited the disapproval of Turner and Weiss (1963). The author

feels, therefore, that unwarranted speculation on the basis of this

orientation data is inappropriate.

Certain interesting relationships do, however, emerge from this

study. The persistence of the orientation of the 'ac' girdles is

notable. In each case the axis of the fold, in any particular

slide, is the axis of the girdle. This strongly suggests that the

quartz fabric in this fold was formed at the same time as the fold

itself and was not imposed at a later date. It might be suggested

that the cross-girdles which tend to develop are in fact related to

F.2. This seems unlikely because the orientation of these cross-

girdles seems to have a consistent orientation relative to the 'ac'

girdles. If in fact the cross-girdles were related to F.2, it

might be expected that they would change their orientation in each

slide, since the slides are non-parallel.

There must have been a considerably annealing recrystallisation

of the quartz in these rocks following the F.1 deformation and

preceding the F.2 deformation. It would seem, however, that this

has not disoriented the fabric imposed in the rock by the F.1 folding.

Specimen II (N16 2A)

This is an F.2 fold from Rastaby. It has an almost parallel geometry with an angular hinge line (Fig. 33A-F inset). The quartz is equidimensional and only slightly strained, often with slightly irregular outlines but tending towards a polygonal shape. There is a crude schistosity (S.1) which has tended to be obliterated by the recrystallisation the rock has undergone, this presumably taking place between the two deformations. The schistosity S.1. parallels the layering which defines the fold. There is also a tendency for mica crystals to grow parallel to the axial plane of the fold; this planar structure is designated S.2. In a pelitic layer within the fold, an F.2 strain-slip cleavage can be seen to have developed. The rock contains some garnet and zoisite and the biotite which defines S.1 is frequently partially replaced by muscovite.

The location of each slide is given in the inset; in each case the slide is cut normal to the axis of the fold.

The results of the study are given in 33 A-D. The following results are apparent from these diagrams.

1) In (A,B) and (D) there is a tendency for a small circle distribution to occur, the axes of these small circles are at some moderate angle to (S.1). It appears that these axes do not lie in the primitive plane of the diagrams. This accounts for the relative weakness of development of the small circles on one side of the diagram, e.g. Diagram (A). Diagram (B) shows the development of the two small circles on either side of the plot, though they are as might be expected, asymmetrically placed relative to the centre.

2) This small circle distribution tends to become rather confused

FIG 33   0001 QUARTZ FOR SPECIMEN II   SCHMIDT NET LOWER HEMISPHERE   CONTOURS 7 ▨ 6 ■ 5 ▨ 4 ▢ 3.2.1 % PER 1% AREA
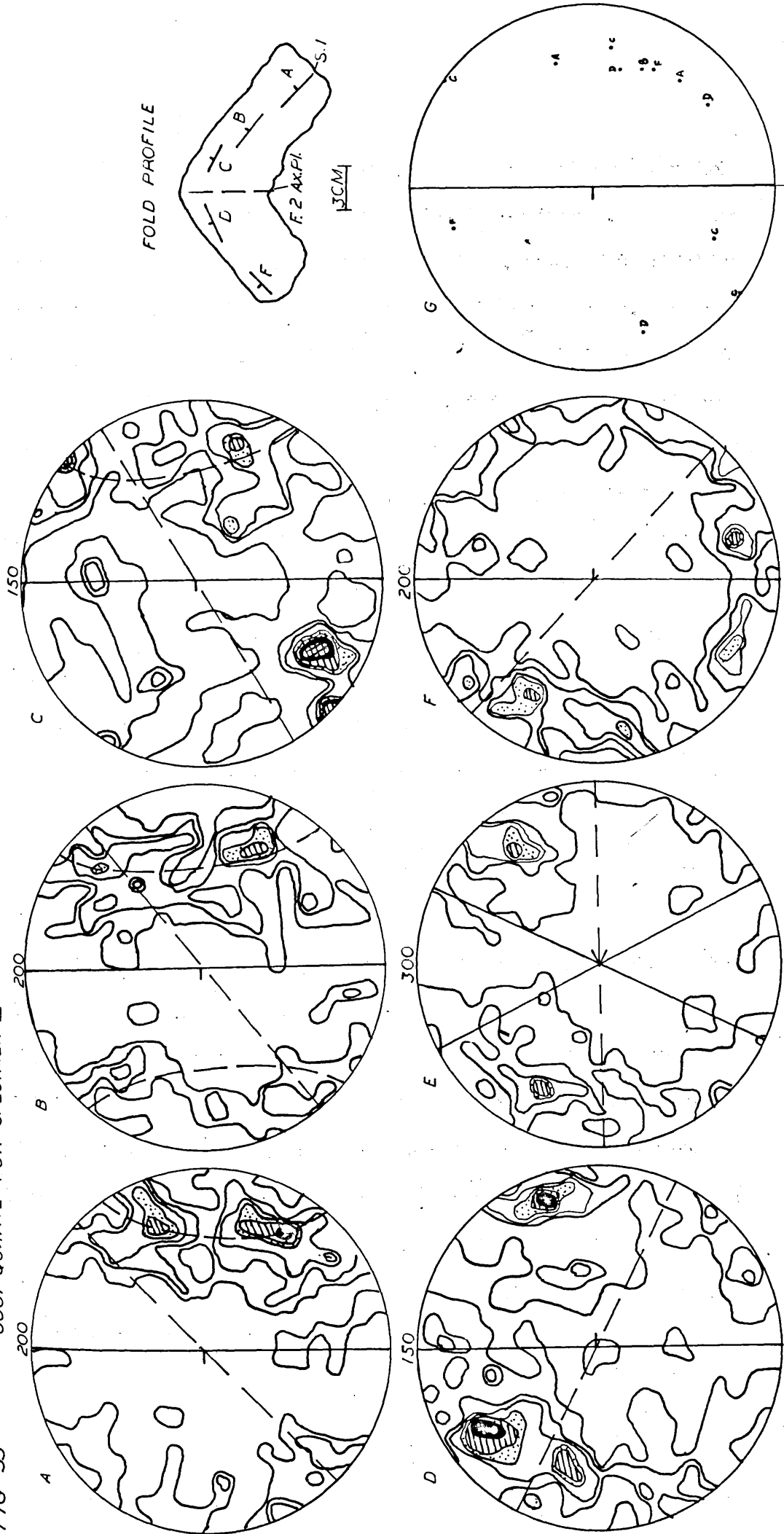
FIGURES REFER TO NUMBER OF POINTS.   DIAG E=C+C

in (C) and (D). These slides are taken from the hinge zone of the fold. It seems likely that this would be the area where the quartz would be most likely to respond to the F.2 deformation.

3) Diagram 3 shows a cleft girdle distribution with an asymmetric distribution of maxima. The diagram however, does show the greatest distribution of maxima in the area to the southwest of (S.1). This is a consistent feature of these diagrams, though in the other cases these maxima can be approximately assigned to small circles.

4) The overall symmetry of the diagrams is triclinic, though in some cases they may approach monoclinic symmetry, with the symmetry plane approximately normal to (S.1). The symmetry is most obviously triclinic in (C) and (D), the slides are from the hinge zone.

Diagram (G) shows the position of all 5, 6 and 7% maxima with the (S.1) plane of each slide rotated into parallelism. This process effectively unfolds the fold. The diagram reveals that the maxima becomes clustered in an area to the right of the mean position of S.1. The three points which most strongly depart from this cluster significantly come from slides (C) and (D). This in summary:

1) There is a tendency for maxima to lie on small circles, the axes of which are at moderate angles to the plane of S.1.

2) This distribution is most confused in the hinge zone of the fold.

3) Unfolding the fold tends to cluster the maxima. From these observations it is possible to say that the dominant fabric in these diagrams is, in fact, a pre-F.2 fabric. This becomes overprinted especially in the hinge zone of the F.2 fold due to the relative concentration of strain in this area.

It has been noted that the fabric represented by these diagrams

is pre-F.2. It differs considerably from the F.1 fabric indicated

in Specimen (I). In this specimen the fabric diagrams are almost

all 'ac' girdles. In Specimen (II), however, only diagram (F)

approaches this geometry. The more typical fabric is that of a

small circle distributions. Carter, Christie and Griggs (1964)

point out that in both naturally and experimentally deformed quart-

zites, deformation lamellae are commonly aligned sub-parallel to the

basal plane of quartz. They also point out that poles to these

lamellae, which will approximately correspond to 'c' axes, define

small circles inclined at approximately 45° to the E.W. diameter of

an equal area net. In their experiments the axes to their small

circles corresponded to the principal applied stress σ 1. The

stresses σ 2 and σ 3 were equal. This sort of distribution must

be thought of, therefore as being characteristic of a flattening

deformation.

It is possible, therefore, that the small circle distributions

seen in these diagrams are due to pre-F.2 pure flattening. Since

the axes of the small circles appear to be oblique to the plane of

(S.1) it seems likely that σ 1 was not normal to (S.1). Such a

period of flattening may have been related to late F.1 or to a

subsequent pre-F.2 deformation which will be discussed later. It

is difficult to know, however, why this flattening deformation has

not affected Specimen (I) which comes from a locality close by.

Specimen III (K5)

This F.2 fold is depicted in Fig. 34. It was taken from a beach

exposure at Komagnes. Slides were cut at intervals round the

structure, in each case normal to the fold axis, which is paralleled

FIG 34 . PROFILE OF THE F2 FOLD FROM WHICH THE PETROFABRIC DATA OF FIG. 35 WAS TAKEN K.5.
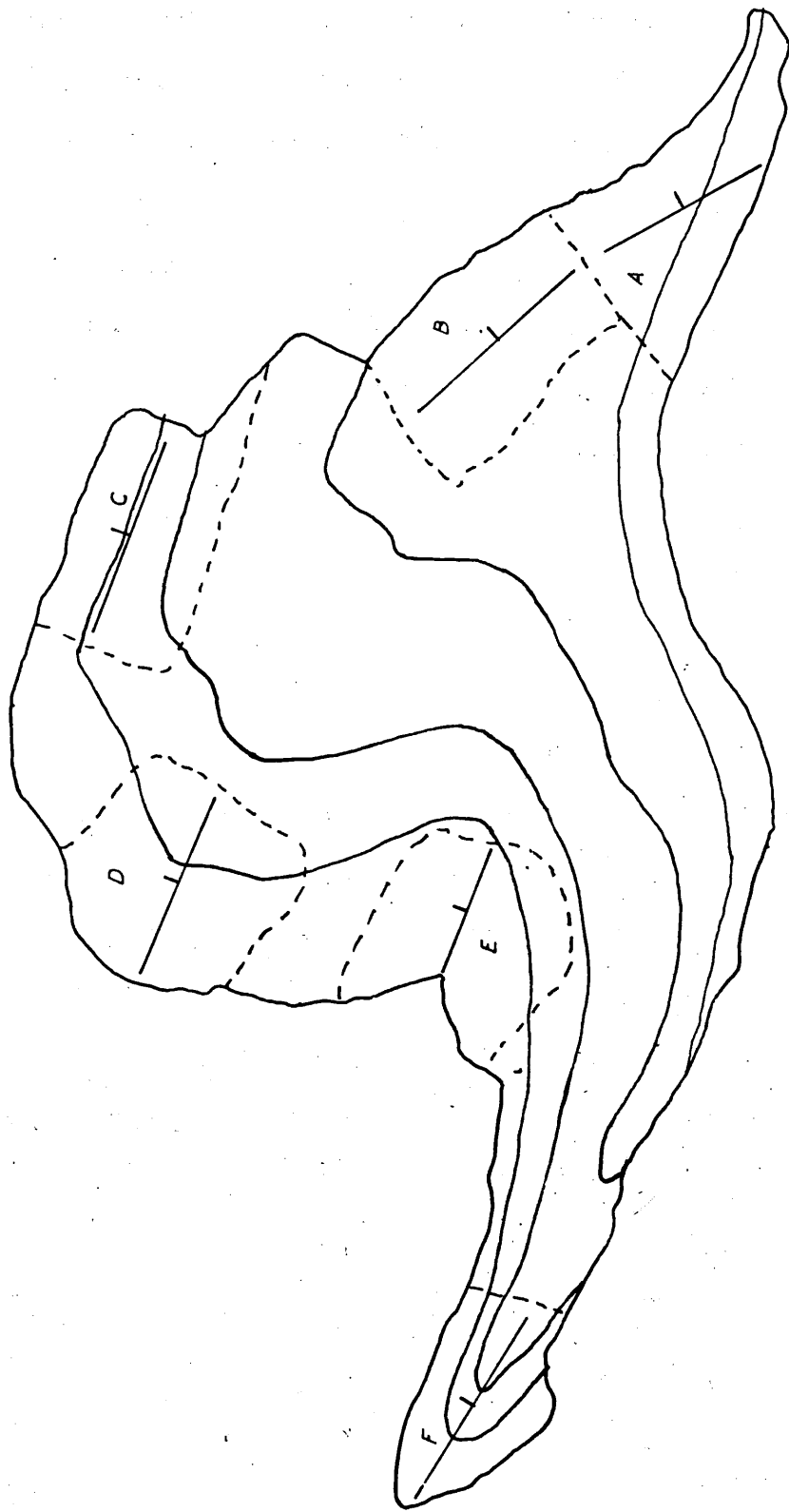
ORIENTATION MARKS ON SLIDES

1CM

by a strong lineation. This lineation is in fact an F.1 lineation which here parallels the F.2 fold axes.

Examination of the thin section indicates that two morphologics of quartz are present.

1) A coarsely crystalline variety which occurs in distinct layers in the rock; the individual quartz grains making up these layers show very little dimensional orientation. Perhaps a slight elongation parallel to the layers in sections cut normal to the fold axes. Sections cut parallel to the fold axes show again the stringers of quartz. In these cuts, however, they are rather more elongate and are undoubtedly responsible for the strong lineation. These layers of coarsely crystalline quartz are quite strongly strained and are folded by the fold.

2) The groundmass quartz is much finer-grained. It shows some straining and is generally equidimensional with a tendency towards polygonal grain boundaries.

There is a schistosity in the rock defined by mica, mostly muscovite. This schistosity is being folded by the fold. In the groundmass areas this schistosity is tending to be destroyed by recrystallisation. In slides where micro-folds are visible there is a tendency for a foliation to develop axial-planar to the saond fold (S2).

The contoured diagrams are given in Fig. 35 (A-F). It was found that the coarse grained quartz did not have any significantly different orientation from the quartz in the matrix. The lines which appear on the stereogram are purely orientation marks and do not have any structural significance. In slides (A, B and C and F) they

FIG 35   OOOI QUARTZ   ALL DIAGRAMS 200 POINTS SPECIMEN III
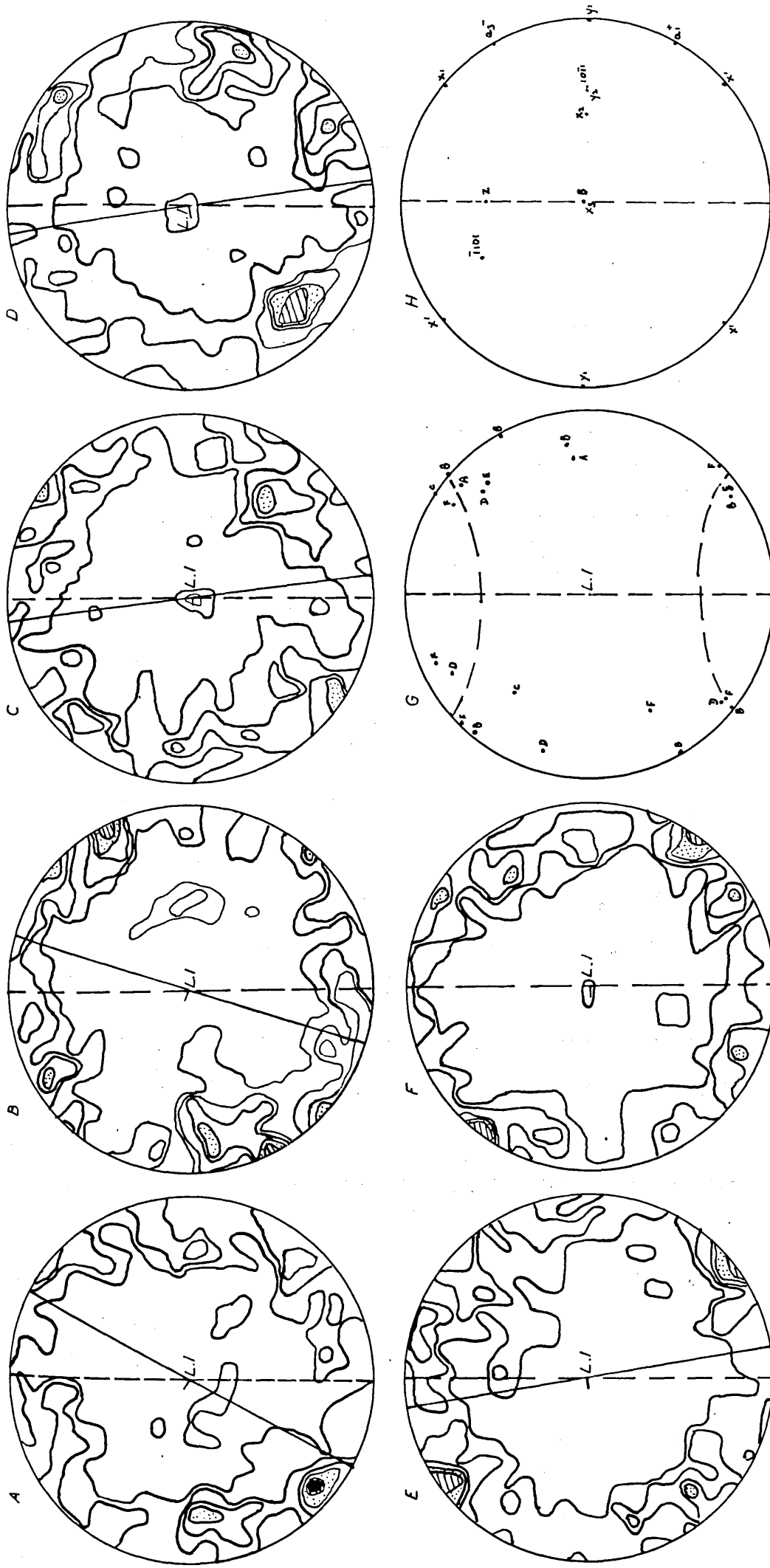
SCHMIDT NET LOWER HEMISPHERE   ———— F.2 AXIAL PLANE

CONTOURS 6 ▪ 5 ▨ 4 ▨ 3.2.1 ▦ PER 1% AREA   – – – – L.I = F.I LINEATION

approximate to (S.1). Their position relative to the fold can be
ascertained by reference to Fig. 34. The line marked F.2 axial plane
on the diagram represents a line originally drawn on the fold
specimen representing the geometrical axial plane and then trans-
ferred to each diagram. The following features emerge from these
diagrams:

1) The 'c' axes tend to be distributed in 'ac' girdles with a
fairly symmetrical distribution of maxima. The F.2 axial plane
represents an approximate plane of symmetry. This relationship is
clearer in some diagrams than in others. A vertical plane normal
to the F.2 axial plane is also frequently approximately a plane of
symmetry. So it is possible to say that the symmetry tends towards
orthorhombic.

2) Using the constructed axial plane as a datum, all 4, 5 and 6%
maxima were placed on a separate stereogram (35G). This clearly
shows the tendency towards orthorhombic symmetry, with the axial
plane and a plane normal to this as planes of symmetry.

Let us consider in (35H) that the distribution of 'c' axes that
occur in (35G) is caused by the alignment of a crystallographic
plane of quartz in the axial plane of the fold. The pole to this
plane will therefore be γ 1. The points γ 1 represent the
statistical maxima of the concentrations of maxima shown by (35G).
By rotation in the stereogram let us bring these 'c' axes concen-
trations into the centre of the projection. The point X_1 rotates
round a small circle to X_2. The point γ 1 does not change its
orientation since rotation is taking place about that point. The
point X_2 is then rotated to the centre of the projection along the

E-W diameter. This displaces the point γ 1 to γ 2, the 'c' axis

maxima now effectively occupy the centre of the projection.

Assuming that the a1 and the a3 crystallographic axes of quartz are

placed in the positions indicated, the unit thombohedron ($10\overline{1}1$)

can be plotted (Berry and Mason 1959). From the diagram it can be

seen that this is less than 2° from γ 2, the pole to the hypo-

thetical plane aligned in the axial plane.

According to the fracture hypothesis of Griggs and Bell (1938)

quartz when subjected to high stresses under great confining

pressures tends to split into needles, parallel to certain crystal-

lographic edges. The bounding planes of these needles are frequently

common crystallographic planes. One of the commonest planes to

develop is the unit rhombohedron with the long axis, the needles

aligned parallel to the edge formed by the intersection of the two

rhombohedra. Griggs and Bell suggest that in natural tectonites,

this process may be operative with the long axes of the needles

aligned in the 'a' direction of the tectonite and one of the bounding

planes becoming aligned in a plane of slip of the fabric.

The orientation data from Spec. III can be explained in this

way. Consider (35H) again; γ 1 is the pole to a rhombohedral

plane supposedly aligned in the axial plane of the fold. In order

to account for the positions of the 'c' axis maxima it is necessary

that the 'c' axes of the quartz be aligned in the plane of the

diagram, i.e. the 'ac' plane. In this situation the edge separating

the two rhobohedra, i.e. the long axes of the needles, must be

tilted relative to the plane of the diagram. The position of this

edge is easily found by locating the intersection between the N-S

diameter, which represents one rhombohedral face and the great circle representing any other. This latter great circle may be found by reversing the manipulation described overleaf, starting with the point (1101). The edge thus found, is represented by point (Z) on (35H). If it is true that the long axes of needles become aligned in the 'a' direction of the tectonite, it would seem that the 'a' direction here represented by point (Z), is not normal to the 'b' direction of the fold which lies at the centre of the projection. Usually it is impossible without deformed lineations to locate the 'a' direction of a fold. In this case although the lineation is of F.1 age and the fold is of F.2 age the former is not deformed by F2 because the 'ab' plane of F.2 contains the F.1 lineation.

The girdles which accompany the maxima can possibly be regarded as being caused by these grains which due to the fact that they were initially incorrectly oriented to the stress field for rhombohedron bounded needles to form, split into needles bounded by other crystallographic planes. The 'c' axes of such needles would obviously take up different orientations relative to 'a'. However, it would seem that needles bounded by rhombohedra are the commonest, assuming that the fracture hypothesis is true.

Regarding the fracture hypothesis, Turner and Weiss (1963 p.430) state "It has been suggested that passive rotation of splinters of this kind would ultimately align them in planes and directions of flow in the tectonite fabric and by this means it has been possible to explain some of the well known relationships between foliation and quartz patterns. We tentatively reject this explanation for two reasons. First there is no proof that the supposed slip planes

of the fabric actually exist. Secondly, some of the patterns so explained are exceedingly well developed in coarse-grained meta cherts whose fabrics must have been influenced far more by recrystallisation than by any direct componental movement."

More recent work by Bloss (1957) is relevant to the argument. He found that by crushing up crystals of rock crystal, rose-quartz and milky quartz and examining optically the resulting fragments, the most common bounding planes were the unit rhombohedra (r) and (z). This he attributes to the presence of a cleavage parallel to these crystallographic planes. The next most common bounding plane was the unit prism (m). Fragments parallel to the basal plane accounted for only ½% of the sample. This at least supports the finding of Griggs and Bell concerning the brittle fracture of quartz.

Another possible way of explaining these orientation patterns is the Translation hypothesis of Schmidt (1927). He suggested that rhombohedra (r) basal pinacoids (c) and first order prisms (m) can be translation planes in quartz, with the vertical edge (m.m.). The horizontal edge (c:m) or the rhombohedral edge (r:r) as translation directions. He accounts for the characteristic maxima (I,II,III, and IV) of Sander by assuming that either (c) (m) or (r) become oriented parallel to a single 's' surface with any one of the three translation directions oriented parallel to 'a'. The orientation data in Specimen III can be explained by assuming that rhombohedral facies are the planes of translation and the edge (r:r) (point Z on 35H) is the direction of translation. It is perhaps notable that once again 'a' is not normal to the 'b' direction of the fold.

Turner and Weiss (1963) once again do not appear to be favourably disposed to this type of orienting mechanism mainly due to lack of evidence. However, they do say that there is now experimental evidence suggesting that one of the effective glide mechanisms in quartz is translation on 'c'.

Bailey, Bell and Penge (1958) in an X-ray study on quartz in naturally deformed rocks conclude that this mineral deforms plastically by bend gliding. According to them one of the crystallographic 'a' axes is always the axis of bending. In looking for a possible glide direction they state that the most likely place for this is in the plane of the unit rhombohedron, though they were unable to establish this experimentally. More recently, Baeta and Ashbee (1969) have indicated that there are many directions of slip within quartz subjected to compression along various directions at elevated temperatures. Among the planes of slip recorded is the rhombohedral plane. The directions of slip derived from their experiments however, do not tie up with the Miller index of the point Z. However, in order to maintain the orientation of the 'c' axes in the primitive plane in Fig.(35H) it is obviously necessary that this slip direction lies somewhere in the rhombohedral plane, though not necessarily at (Z).

Other workers, (Christie, Griggs and Carter (1964) have experimentally determined a basal slip plane in quartz. This type of experimental evidence seems to suggest strongly that quartz in nature deforms plastically rather than in the brittle fashion envisaged by Griggs and Bell. The fabrics shown by specimen III seem to indicate participation of the rhombohedral planes of quartz in the genesis of the tectonite fabric.

In Specimen I the fabric appeared to be of F.1 age. In

Specimen III however, the fabric appears to be completely of F.2

origin. Possibly some of the scattered maxima can be explained as

representing a residual pre-F.1 influence. It would seem, therefore,

that the F.2 deformation in the Komagnes area has been sufficiently

strong to totally reorient any previously existing fabric.

Specimen IV (K2A)

This series of specimens comes from the large F.2 fold

depicted in Plate 51. This occurs on the beach at Komagnes quite

close to the locality from which Specimen (III) was taken. Fig. (36E)

gives the location of the specimens from which the slides were cut.

The fold axis is parallel by a strong lineation which is again an

F.1 lineation. In each case the slide is cut normal to the fold axis.

Examination of the thin section indicates that the stringers

of quartz which characterise Specimen III are here absent. The rock

has a fairly consistent grain size. There is no dimensional

orientation of quartz in the plane of the slide, the grains often

have irregular outlines and are slightly strained. There is again

a tendency for a polygonal fabric to develop. There is a crude

foliation in the rock which parallels the banding; this is (S1).

It has been partially obliterated by recrystallisation. A second

foliation (S2) is tending to develop axial planar to the F.2 fold.

This is defined by muscovite. There is some epidote in the rock.

The contoured stereograms are given in Fig. (36A-C). From

these diagrams the following facts are apparent:

1) In (36A) it would appear that (S1) and a vertical plane normal

to it, approximate to planes of symmetry, suggesting that the fabric

FIG 36   0001 QUARTZ SPECIMEN IV   ALL DIAGRAMS 200 POINTS

ATTITUDE OF SPECIMENS

A—350° 64°W
B—344° 25°E
C—269° 8°N

FOLD ATTITUDE

FOLD PROFILE

SCHMIDT NET LOWER HEMISPHERE

CONTOURS 5 4 32.1% PER 1% AREA
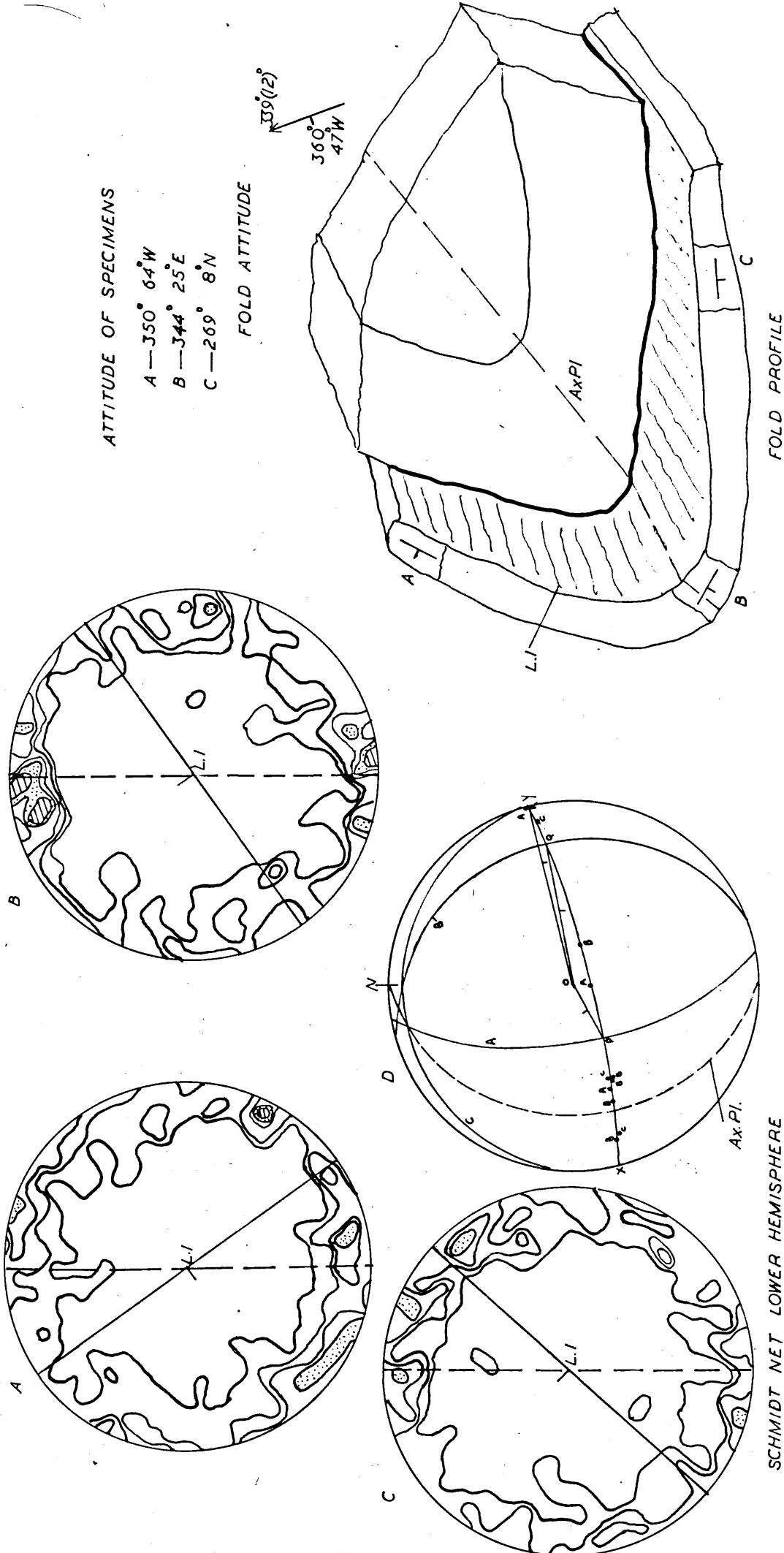
bears some relationship to F.1 though this is not clear. In (B) and (C), however, the F.2 axial plane and a vertical plane normal to this are more closely planes of symmetry. Thus, all three slides appear to approach orthorhombic symmetry though with different orientation of the symmetry planes.

2) Fig. (36D) presents a synthesis of 4 and 5% maxima from (A, B and C). This diagram was prepared as follows:- the great circles marked (A, B and C) represent the limbs of the fold as measured in the field, the intersection of great circles (A) and (C) represents the axis of the fold. A theoretical axial plane was then drawn on the diagram by bisecting the arc between (A) and (C). The great circle (XY) represents a plane normal to the fold axis, i.e. the plane containing the slides. The lines (OP, OQ and OR) are the orientation marks on the slides. Using these lines as datums the position of all 4 and 5% maxima are placed on the diagram. This procedure aligns all the orientation data to the position it had in the field.

Examination of the diagram indicates a tendency towards a concentration of maxima in the region of the constructed axial plane of the fold, This tendency is shown most strongly in slide (B) which comes from the hinge zone of the structure. Gangopadhyay and Johnson (1962) record a similar concentration of 'c' axes in the axial planes of second phase folds from Scotland. These folds appear to have been formed by shear along a direction parallel to the fold axial plane. Also the trend of the 'c' axes maxima appear to be parallel to the 'c' direction of the fold, which was determined using Ramsay's method of plotting deformed lineations. In the case

of (36A-E), however, there are some significant differences in the microscopic fabric of the quartz. In the specimens described by Gangopadhyay and Johnson (op. cit) the quartz is strongly dimensionally oriented in the fold axial plane. In (36A-E) this is not the case, the quartz shows very little dimensional orientation. In the same paper, measurement of first phase folds by Gangopadhyay and Johnson in which the quartz also shows a strong dimensional orientation in the fold axial planes, yield diagrams in which the maxima are concentrated in planes inclined at some 40-60° to the axial planes. This situation is more akin to that seen in Fig. (35A-E) though again the dimensional orientation of quartz in this specimen is absent. Gangopadway and Johnson explain the fabrics in their first folds by saying that, although shear along the axial planes must have taken place, gliding along planes 40-60° from the 'c' axes might also have been operative (these presumably aligned in the axial plane).

Thus, the two specimens (III) and (IV) present us with the slightly anomalous situation of two fold structure in similar lithologies from localities very close to each other, giving two distinctly different quartz orientation patterns. Possibly the solution to this anomaly is to be found in the slightly different lithological characteristics of the two rocks seen in thin section. Specimen (III) contains stringers of quartz which are quite strongly strained, these are absent in Specimen (IV). The groundmass quartz of the two rocks are very similar. It is possible that the stringers in (III) which must represent areas of slightly different competance, influence the development of the fabric during some initial phase

of flexural folding, this additional influence would have been
absent in the development of fold IV, although the two folds almost
certainly were subjected to a similar sort of deformation regime.
Thus, both examples (III) and (IV) would seem to indicate that
the axial planes of the folds have at some stage in the evolution
of the structures, been planes of slip or translation along which
some crystallographic direction of quartz became aligned.
(Gangopadhyay and Johnson op. cit).

Specimen V (M16 - 18)

This is the F. 2 fold from Rastaby with the strongly curving
axis depicted in Fig. 13. It is apparent from the serial sections
(P.56) that the fold generates along its axis from layering which
is virtually unfolded into an asymmetric antiform (serial section D).
In this petrofabric study a slide (A) was taken from the point where
the layering is virtually by the fold. In fact the slide was taken
from a position where the layering is even flatter than serial
section (A) of Fig. 13. A second slide (B) was taken from the
section (D). In each case the slide was cut normal to the fold axis
at that point, thus the slides are non-parallel in space.

Examination of the thin section (A) indicates that the rock
has a crude schistosity which is tending to obliterated by the
recrystallisation which presumably took place between F.1 and F.2.
In the second slide this crude schistosity is being folded by the
zone of the fold there is a tendency for mica crystals to grow
parallel to the axial plane of the F.2 fold. In the area between
the hinge zone and the limbs, the micas are clear and unstrained,
which means that post-F.2 recrystallisation must also have taken

place. The quartz in the whole rock displays slight undulose extinction and the development of deformation bands. These, however, may be late features of the deformation history. The contoured diagrams are given in Fig. (A) and (B).

Examination of these diagrams indicates that in Fig. (A) there is a small circle distribution of quartz axes. The approximate axis of this small circle is marked as (X1). This is a similar situation to that recorded in Specimen (1), a small circle distribution with the axis inclined at a high angle to (S.1). The line marked (S.1) in the diagram is an approximation to this direction since the real (S.1) was impossible to locate properly on the slightly folded surface. Diagram (B), however, indicates that there are two great circles inclined at a high angle to each other, the orientation of F.2 axial plane is given on the diagram. This diagram is, therefore, a B∧B tectonite. It is of considerable significance to note that in this diagram, which represents the fold at its maximum development, the fabric is at its most complex. It is in sharp contrast with the simplicity of the fabric in (A). This strongly suggests that the fabric in (A) is a pre-F.2 structure which becomes considerably modified in the area of the more intense F.2 folding represented by (B).

Fig. (37C) is an attempt to relate the 'c' axis girdles and small circles in both diagrams. Since the fold axis of this specimen is curved it follows that the two slides are non-parallel in space, It was founded by measurement of the hand specimen that the two slides make an angle of 23° with each other. The line (S1) in (37C) represents a plane which is common to both slides. It

FIG. 37   0001 QUARTZ   BOTH DIAGRAMS 220 POINTS   SPECIMEN II

SCHMIDT NET

LOWER HEMISPHERE

CONTOURS

6 ■ 5 ▨ 4 ▦ 3.2.1.% PER 1% AREA

B.2 = F.2 FOLD AXIS
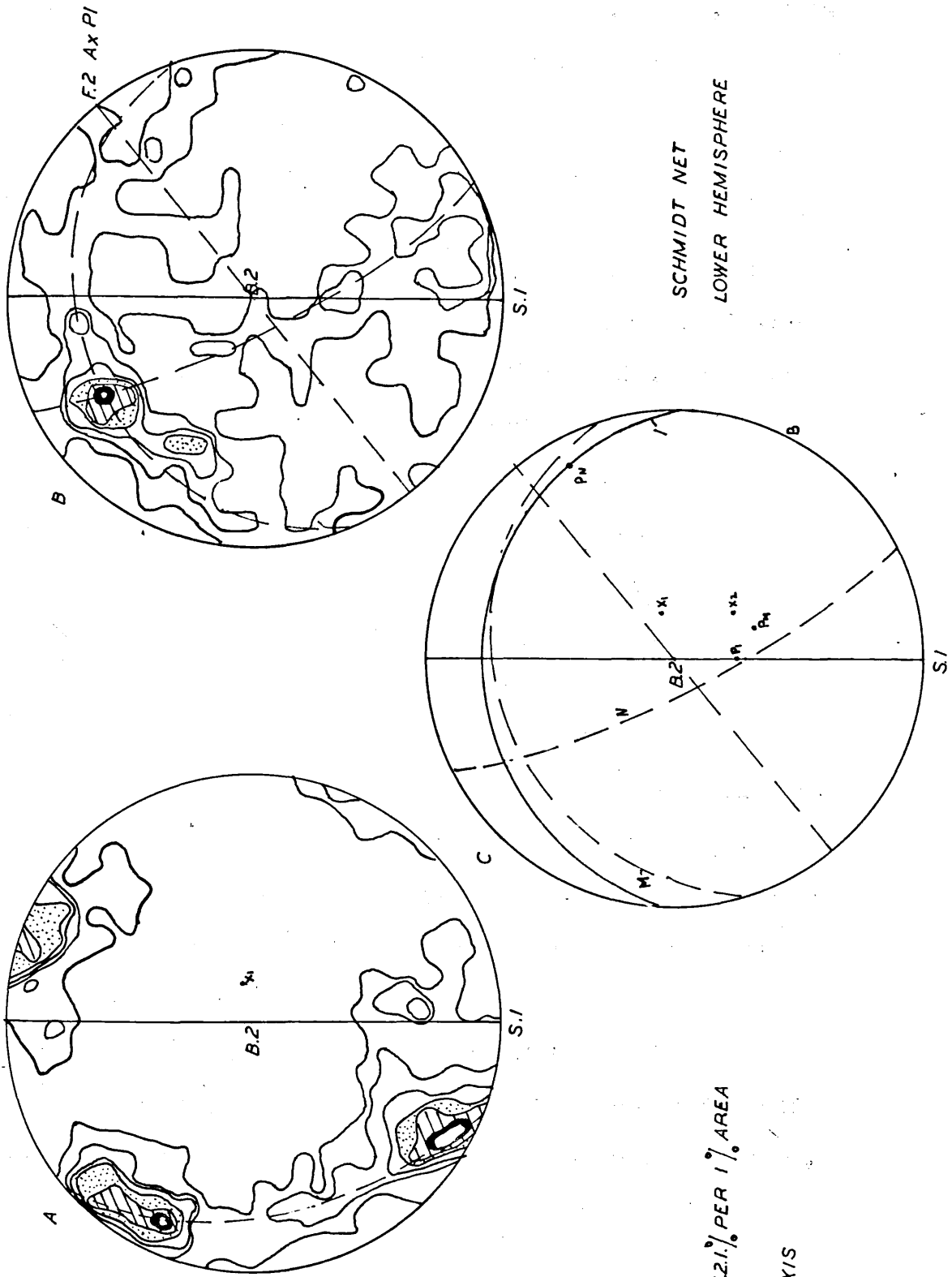
corresponds with (S.1) in Slide (A). Now assuming the primitive

of the diagram to represent the plane of slide (A), the point (X1)

is marked in the diagram. However, we know that the two slides are

23° out of parallelism with each other; thus great circle (1)

represents the plane of slide (A) relative to slide (B) which now

becomes the primitive plane (B in the diagram). The pole to this

great circle is (P.1). Since the plane of (A) is now tilted relative

to (B), the point (X1) which represents the axis of the small circle

in slide (A) moves to (X2). The great circles (M) and (N) represent

the girdles of slide (B). The poles to these are marked (PM) and

(PN) repectively. It is to be noted from this that the pole to (M),

the great circles representing the strongest concentration in slide

(B), is very close to the point (X1) which represents the axis of

the small circle in (A). It is possible, therefore, that the small

circle in (A) and the great circle girdle represented by (M) are

related.

Considerable care is needed here since:-

1) (X1) represents a small circle (PM) a great circle.

2) The positioning of the line (S1.) in both slides is open

to considerable inaccuracies as is the location of the

axis of the small circle in (A).

However, even if the two circles are not related it seems

possible to suggest that the fabric represented in slide (A) is

earlier than that represented in slide (B), since in the area of

the most intense folding the simple fabric in (A) is disrupted to

form the more complex B $\overset{\Lambda}{B}$ tectonite of (B). It is suggested there-

fore, that slide (A) has a pre-F.2 fabric. In the sense that this

fabric is distributed on a small circle whose axis is inclined at a moderate angle to S.1, it resembles the pre-F.2 fabric seen in Specimen (II) where the pre-F.1 fabric appeared to have been folded by the F.2 fold with little modification.

Summary and Discussion

I   F.1 isocline with curved axis, girdle fabrics with the axis of girdle parallel to the fold axis.

II F.2 open symmetrical fold with almost parallel geometry, the fabric tends to be aligned on small circles, the axes of which are at a moderate angle to S.1. The fabric is pre-F.2 and is folded by F.2.

III F.2 fold with moderately appressed style. Fabric entirely F.2 with 'c' axes aligned symmetrically either side of the axial plane.

IV F.2 fold moderately appressed 'c' axes tend to be in the axial plane.

V   Curved axis F.2 fold pre-F.2 fabric where S.1 is unfolded,   .
    F.2 fabric where the F.2 fold reaches its maximum development.

Recrystallisation

The slides show evidence of a strong recrystallisation between F.1 and F.2; this tends to disorient the mica fabric defining S.1, e.g. Specimens (I) and (II). Where there has been a more complete reorientation of the quartz in response to F.2, e.g. Specimen (III) and (IV) although the disoriented S.1 is still visible, there is evidence of unstrained mica growing axial planar to the second folds. The quartz in these rocks shows evidence of slight straining in the form of undulose extinction and deformation bands. These,

however, may be a later feature of the deformation. It seems clear, therefore, that there has been recrystallisation of the quartz, tending to form polygonal grain fabrics both between F.1 and F.2 and post-F.2.

Specimen (I) seems to indicate an F.1 fabric which has survived the recrystallisation which the rocks have undergone, II shows evidence of a fabric which may be due to flattening (Carter Christie and Griggs 1964). It must be admitted however, that the features on which this interpretation is based are not common to all the slides. In this specimen a tendency was noted in the hinge zone for an F.2 fabric to develop, this presumably being due to the concentration of strain in this area. This tendency for an F.2 fabric to develop in an area of stronger F.2 reorientation is seen again in (V). The specimens (III) and (IV) seem to be strongly F.2 oriented, though possibly odd maxima which do not fit the scheme, may be explained as pre-F.2 fabrics. In these latter two specimens (III) and (IV), the axial planes of the folds appear to have been zones of alignment of either rhombohedral planes of the quartz lattice (III) or of actual 'C' axes (IV). It is suggested, therefore, by analogy with other workers (Gangopadhyay and Johnson op. cit), that there has been a component of shear folding in the genesis of this structure. This is also true of the specimen (III) because of the obvious alignment of rhombohedral planes in the axial plane of the structure. These planes may be explained either as the bounding planes of the needles of Griggs and Bell, or as the translation planes of Schmidt.

A survey of the more recent literature on the experimental deformation of quartz seems to indicate that quartz deforms

plastically by bend gliding and slip on certain well-defined
crystallographic planes. One of the commonest such slip planes is
the basal pinacoid of quartz. The presence of a slip direction on
this crystallographic plane is evidenced by the occurrence of
deformation lamellae parallel or sub-parallel to it (Carter, Christie
and Griggs op. cit). More recent work by Baeta and Ashbee (op. cit)
reveals the occurrence of a number of other slip directions in quartz,
notably the rhombohedral plane. Slip on rhombohedral planes may be
responsible for the minor concentration of deformation lamellae
poles making an angle of 20-50° with the 'c' axes recorded by Carter,
Christie and Griggs (op. cit). The occurrence of identical structures
in naturally deformed rocks is strongly suggestive that they have
deformed by similar mechanisms. Evidence of such deformation lamellae
may be removed by later recrystallisation leaving the oriented fabric.
The general absence of evidence for the cataclastic textures neces-
sary for the Griggs-Bell Hypothesis seems to indicate that this is
an unlikely deformation mechanism in nature.

The petrofabric data described above poses the following
questions:-

1) Since there are so many combinations of bounding planes to
needles and planes of slip, according to the respective
hypotheses, (although rhombohedra appear to be among the
commonest detected experimentally) why should rhombohedron
have been chosen in the case of (III) and 'c' axes align-
ment in (IV)?

2) How were these planes and directions rotated into the axial
plane of the fold?

3) Why do two different orientation patterns develop from

   similar localities, i.e. (III) and (IV).

These questions are extremely difficult to answer. It is possible that further experimental works will reveal the mechanisms of alignment of the various crystallographic planes.

The author feels that in many ways this study is incomplete particularly more work could be done on the curved axis folds. This sort of study would perhaps reveal the progressive change of a pre-F.1 fabric as it was modified in a curved axis fold. Also the work could be integrated with mica diagrams.

7)   Mechanisms of Fold Formation

a)   F.1

It is very difficult to make many meaningful comments about the
mode of formation of F.1 folds for two reasons; firstly not all that
many of the folds have been found and secondly, they have certainly
been modified by subsequent deformations.   However, the following
facts are apparent:

1)   The folds have a tight isoclinal style; one such fold measured
(Fig. 31) fell into Class Ic of Ramsay's classification although it
approached quite close to the similar fold model (Class 2).

2)   The folds have a strong axial planar foliation though this has
tended to be destroyed by subsequent recrystallisation.

3)   Evidence was put forward that one F.1 fold from the beach at
Rastaby had a curved axis.   This was considered to be a primary
feature of the F.1 deformation.   This curvature, however, at least
on the scale of the outcrop, is not an ubiquitous feature.

4)   Petrofabric diagrams constructed from a number of slides cut at
intervals round a curved axis.   F.1. fold gave 'ac' girdles, the axes
of these girdles were in each case parallel to the fold axis at that
point.

Examination of the curved axis fold described above gives a strong
impression that the material involved was highly mobile at the time of
the fold formation.   Such folds have been recorded extensively on the
neighbouring island of Sørøy by Ramsay and Sturt (personal communication)

and Roberts (1965). For some folds of this type in marble on Sørøy Roberts (op.cit) favours the explanation of Nicholson (1963) advanced for comparable folds in the Sokumfjell Marble, i.e. uniaxial compression causing differential movement within the axial plane of the structures. Ramsay (1967), in a discussion of similar folds, suggests that variation in compressive strain along the fold axial surfaces may be responsible for some of the variations in plunge noted in certain similar-type folds. However, application of the parameter of Gangopadhyay and Johnson (op.cit.) viz the alignment of quartz 'c' axes in the axial plane of a fold as an indication of a similar-type mechanism of fold formation, suggests that such a mechanism has not been operative in the case of this structure.

The author is inclined to think that they are flow-folds, the apparent plasticity of the material may perhaps be explained as being due to the relatively unconsolidated state of the sediments during the early part of the F.1 deformation, this initiating the curving fold axes. As the metamorphic grade rose, the quartz fabric and axial plane schistosity became imposed on the folds.

This type of deformation mechanism is consistent with that proposed for the F.1 nappes mapped on Sørøy by Ramsay and Sturt (1963) and Roberts (1965) and also by Hooper and Gronow (1970) further south in the Sandland peninsula. The latter two authors have proposed an anticlinal rise in the Sandland area, this based on two criteria; firstly the presence of garnet-gneiss in the area of the proposed rise, this representing deeper crusal material. Secondly, and more important, is the fact that the large F.1 recumbent folds either side of the proposed rise have opposed senses of transport. To the east, on Sørøy they close eastwards and to the west, they close westwards. Possibly this,

therefore, represents a basement culmination with nappe transport by gravity gliding away from this zone. Sturt and Ramsay have questioned this hypothesis on two counts: firstly they say that the regional ubiquity of garnet-gneiss invalidates its use as a criterion for the presence of an anticlinal rise; secondly, elsewhere Gronow has stated that the regional metamorphism of the garnet-gneiss is post the local F.2 folding. Despite these objections the opposed senses of transport remain.

B.A. Sturt (personal communication) has said that on Sørøy the axes of F.1 minor fold axes on the limbs of the major folds, have a considerable degree of curvature. In the hinge zones of these folds however, the folds approach a cylindroidal geometry. This is consistent with the flow hypothesis, since in the hinge zone of an advancing recumbent fold, minor folds would tend to be rotated into parallelisim with the hinge line, whereas on the limbs, they would be freer to take up other more random attitudes, dependent on local flowage gradients.

It has not been possible on E. Seiland to locate any major F.1 folds and hence relate the position of the folds described to such structures. However, it seems likely that the Seiland folds are comparable to those on Sørøy.

b)   The Second Fold forming Deformation F.2

The conditions under which F.2 folds formed differed considerably from those prevailing during the F.1 folding. Previous to F.2 the rocks had undergone high grade regional metamorphism, and intense deformation. The metamorphic grade had waned to sub-garnet grade conditions by the commencement of F.2. A summary of the various features of F.2 folds

will now be given:

1) There is considerable evidence of a lithological control of fold styles. Uniform massive psammite lithologies tend to favour a concentric cylindroidal geometry, whereas mixed pelitic, semi-pelitic and psammitic units favour a more acute fold style with the formation of folds with strongly curving axes. These latter types are also characterised by elasticus structures and evidence of decollement along the contacts of the pelitic and psammite units.

2) Evidence of curvature is provided by:-

    a) The great circle distribution of fold axes on stereograms.

    b) Hand specimens of folds with strongly curving axes.

3) This curvature appears to begin early in the history of fold formation. The following factors indicate this:

    a) Symmetrical minor buckles, which represent incipient curved-axis folds, are periclinal.

    b) Curved-axis folds which develop laterally from unfolded layering can never have been at any stage, cylindroidal.

4) Lineations associated with the curved-axis folds are generally of the intersection type. Consequently they curve with the fold hinges.

5) Geometric studies indicate:

    a) Most of the folds conform closely to the flattened buckle model of Ramsay (1967).

b) Many have a Class 3 geometry in the hinge zone and a Class
   1C geometry on the limbs. This combination produces
   closeness to the similar fold model.

c) Intra-layer flowage is suggested by the constant increased
   thickness of synforms relative to antiforms. This feature
   cannot be explained solely by differential extension in 'a'.

6) Petrofabric diagrams for F.2 folds give slightly contradictory
results. However, in general:

1) F.2 destroys the pre-F.2 quartz fabrics to various degrees.
   The most complete destruction is seen in the cylindroidal
   F.2 folds from Komagnes. These show evidence of:-
   a) Shear folding by the alignment of quartz 'c' axes in the
      axial planes of the folds. Gangopadhyay and Johnson
      (op.cit).
   b) Alignment of rhombohedral planes of quartz in the fold
      axial planes, possibly due the use of these as trans-
      lation planes.

Discussion

Voll (1960) states that "during the advanced stage of B.1 folding
the axes of monoclines and their parasitic folds begin to curve. For
large monoclines in competant rocks this can be the case from the
beginning of folding". He states that this curvature" is particularly
pronounced in zones of strong deformation, it particularly affects the
parasitic folds of the overturned and steep limbs of monoclines, whose
axes may remain horizontal." He gives four possible reasons for the
development of this phenomenon:

1) They curve owing to unequal stretching along the strike of (S.1) (internal rotation).

2) The folds form with normal axes, the limbs are then thinned and detached and the hinges are rotated passively between surrounding (S.1) planes (external rotation).

3) (SS) is buckled before (S.1) forms and the buckles are intersected by straight (S.1) planes.

4) (SS) planes are intersected by (S.1) planes of different attitudes.

For reasons already discussed it would seem that the curvature of these structures on Seiland is initiated from the very beginning of the folding process. It would seem necessary, therefore, to look for some deformation process capable of producing this effect.

Campbell (1958) quotes and enlarges on some of the ideas of Tokuda (1926-27) concerning the origin of periclinal fold systems. Tokuda glued a piece of rice paper to a board and induced fold formation by pushing the rice paper with his thumb. The forward movement of the paper induced by the thumb produces secondary tensional stresses either side of the thumb. A combination of these two stresses produces a periclinal fold system. Campbell enlarges on this; he envisages the application of a shear couple F-F to rock material. (The direction of shear presumably lying in the layering). He states:- "the initial result is the development of a fold caused by F-F. At an early stage, the induced (oblique) forces S-S develop sufficient strength to dominate the detail of foldformation". This combination produces en-echelon periclines, the axial traces of which are normal to the shear directions F-F.

It would appear that this explanation is not viable for the folds on E. Seiland, since it is difficult to see how such a regional shear couple could be produced, and in any case such a stress field is not easily reconcilable with the development of S.2, the F.2 axial planar schistosity.

It seems to the author that the problem is not how the folds continue to develop curved axes, but how the initial periclinal buckle is developed. There does not at the present time seem to be any rational explanation for this. It seems only possible to say that the curvature is initiated at the beginning of folding by some form of differential movement within the axial plane of the structure.

Once this curvature has been initiated there are a number of ways in which it can develop. Voll claims that all the possibilities he advances have been observed with the possible exception of 1).

Fig. 13 gives some clue to the movements that have taken place in this fold structure. Consider the following table:-

Table 6 from Fig. 13 (P)

Section	Minimum Limb Thickness (m.m.)	Maximum Hinge Thickness (m.m.)
A	4	9
B	3.7	8
C	4	8.5
D	4	10
E	3	9
F	X	X

There are relatively few readings here but it is possible to pick out a trend. There is a relative constancy of limb thickness at around 4 mm. In the hinge zones, however, there is a progressive increase in thickness, until the maximum is reached where the fold reaches its maximum development. It has already been noted that in one fold, differential flowage of material appeared to have taken place. It was proposed that this had taken place from the limbs to the hinge zone. The above figures suggest that there has also been movement of material along the hinge line from the depressions to the culminations of the curved axis folds. This was presumably accompanied by stretching in the 'a' direction, but this stretching cannot account completely for the variations in thickness, which must be partially due to flowage. These facts are expressed diagrammatically in Fig. 38A.

Another factor which may help the curved axis folds to decrease the radius of curvature of their hinge line is the decollement effect between pelitic and psammitic units. This effect may be to some degree responsible for the strong development of curved axis folds in mixed pelitic and psammitic units. The possibility of flowage of material towards the culmination has already been mentioned for psammitic units. This flowage probably occurs in greater measures in the incompetent pelitic layers. A greater volume of pelitic material in the area of a culmination will greatly facilitate the sliding of successive psammitic bands over each other using the pelite-psammite contact on a plane of decollement. This is illustrated in (Fig. 38B. and C.)

This sort of effect has already been discussed in (Fig. 10. P. 54.) In this example, however, actual detachment of the pelite-psammite boundary took place. Similarly, detachment at psammite-pelite boundaries in Plate 95 has allowed a change of style of minor folds with

FIG 38   FLOWAGE IN CURVED AXIS FOLDS

A

[::::] PSAMMITE
[ ] PELITE

B

C

ADVANCE OF CULMINATION

⟹ FLATTENING

⟶ SENSE OF FLOWAGE
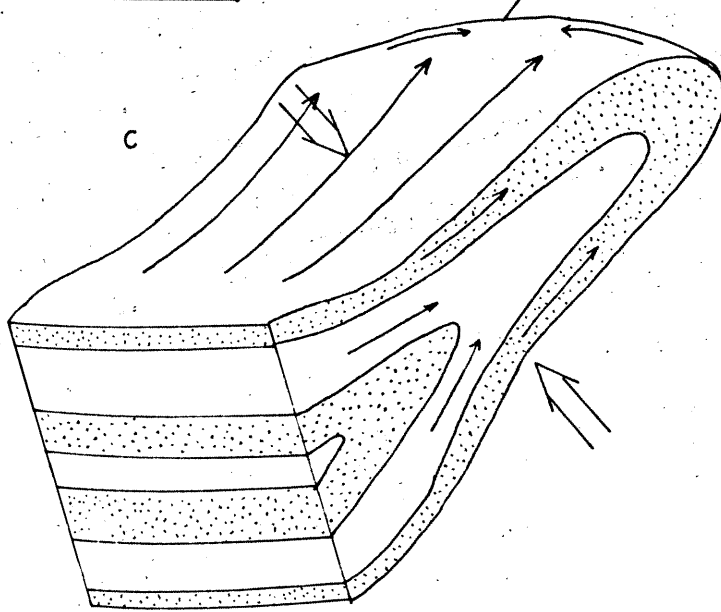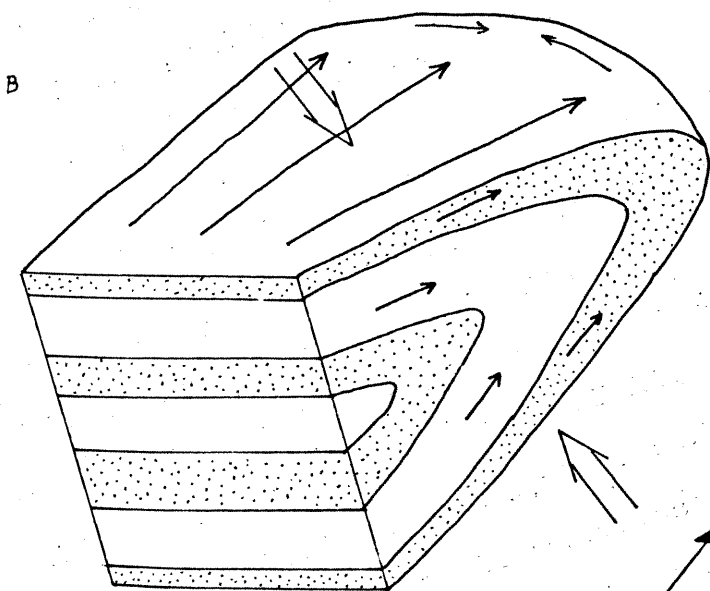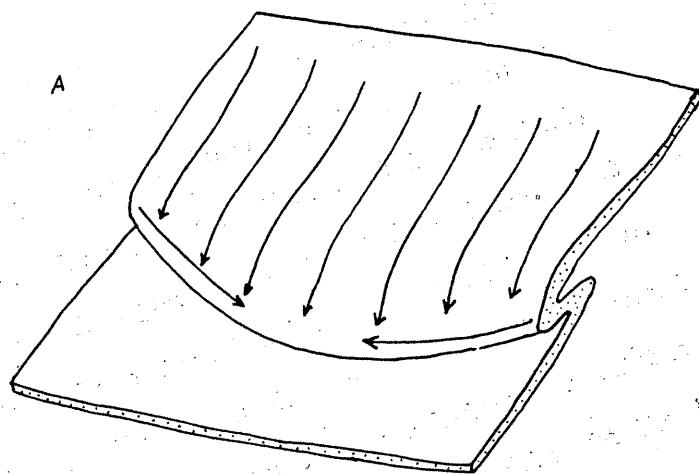
Plate 95. F$_2$ buckles in banded quartzite, Upper Komagnes
Group, Rastaby.

depth to occur in the mere major structure.

In this particular structure there is some variation in the trend of the minor folds as indicated by Figs. 15A and B. These minor folds were also observed to be periclinal in nature. This implies a varying degree of psammite-pelite detachment along the axis of the fold, which would seem to indicate that the decollement effect does considerably facilitate the development of these curved axis structures.

It seems obvious, therefore, that flattening considerably accentuates the curvature of fold axes. This is greatly helped by the processes described above. B. A. Sturt (personal communication) has indicated that curved axis F.2 folds on Søroy have been noted to develop the maximum degree of curvature in the short limbs of F.2 folds, i.e. these area that have undergone the maximum degree of flattening. This must also be the case on E. Seiland, since it has already been noted that most F.2 folds are located in the steep limbs of the more major structure. The long limbs have a relative dearth of fold due to the stretching they have undergone. These observations are in line with those of Voll (op.cit.), who has said that the folds particularly affected by curvature are the "parasitic folds of overturned and steep limbs of monoclines". The complete evidence from Seiland, however, would seem to differ slightly from Voll's in that on Seiland the curvature is initiated at the very beginning of folding.

Thus, a picture emerges of the initiation of periclinal buckles and the subsequent modification of these structures to produce the curved axis folds. This process can be envisaged as modifying the sensibly concentric buckles initially formed, through a number of stages until they approach quite closely to the similar fold model. This process of

modification would appear to take place, in response to flattening, by flowage within the layering coupled with some extension in the 'a' direction. It is also greatly facilitated by the decollement effect.

In seeking for some clue to the mechanism of rock flow involved, the petrofabric studies for the curved axis folds are not very helpful, particularly Specimen (V). There does not appear to be any rational alignment of quartz 'c' axes or any other crystallographic directions. It can be said that the generation of the fold along its axis seems to be marked by an increase in the complexity of the fabric. The elements of this fabric do not seem to be directly related to either the fold axis or the axial plane. Probably this complexity is a reflection of the three-dimensional flowage that has taken place within the fold.

Despite the fact that there is a strong axial-planar foliation in the pelite bands associated with the curved axis folds, it does not seem that this plane has been the main plane of flowage. Layer boundaries between psammite and pelites are sharp and distinct. This would seem to preclude flowage across the layers. It would appear therefore that flowage is contained within the layer boundaries, with the pelitic layers accommodating to the fold shape taken on by the more competent psammite.

The ideas of Donath and Parker (1964) are particularly relevant to the study of these structures. They assign folds to four fields of folding:

a) Quasi Flexural
b) Passive
c) Flexural Flow
d) Flexural Slip

The mechanism of folding that will operate in any particular case will depend on the mean ductility and the ductility contract of the deforming strata. The mean ductility represents the behaviour of the most abundant rock type, the ductility contrast represents the contrasts in ductility between the various layers present. Where the rock is homogeneous the ductility contract will be low. Thus, to some degree this parameter reflects the presence or absence of layers of different lithologic type. Both these quantities are dependent on environmental conditions i.e. pressure and temperature, which may change during the course of deformation.

The curved axis folds discussed above would seem to fall into the flexural flow group, which according to Donath and Parker (op.cit) form under conditions where the mean ductility is moderate and the ductility c contrast is low-high. They say:- "Material moves towards or away from hinges in flexural flow folding; the redistribution of material within the layers is most commonly reflected by relative thickening at the hinges and thinning on the limbs. Where there is high ductility contrast and appreciable flow in the hinge, the more ductile layers may show development of cleavage. Commonly flow occurs to differing degrees within different layers but it is always restricted to layer boundaries and the presence of layering is clearly essential for flexural flow folding".

These characteristics accord very well with the ideas discussed above.

The Lower Komagnes Group

The folds in the Lower Komagnes Group differ in a number of

important respects from these of the higher units of the Komagnes Group.

1) They do not show such a strong tendency to develop curved axes.

2) Fabric diagrams of folds from the beach locality at Komagnes show the orientation of either 'c' axes or rhombohedral planes of quartz within the axial planes of the fold.

No geometric studies have been carried out on these structures. Superficially, however, they show many of the characteristics of the folds from the Upper Komagnes Group. Also it must be admitted that two fabric diagrams do not constitute sufficient evidence to make a detailed assessment of the folding mechanisms operative in the whole of the Lower Komagnes Group; nevertheless the differences in both lithology and fabric remain.

One of the principal differences between the Lower Komagnes Group and the rest of the Group is the lithology; the lower Group consists of rather uniform flaggy micaceous psammites with thin interbedded schist bands. In general, the lithologies above this consist of a much closer interbanding of psammitic and pelitic units. It seems probable that this difference may be responsible for the difference in fabric. It was mentioned above that, superficially at least, folds on the Lower Komagnes Group resemble geometrically the folds of the Upper Group except in the respect that they do not have curved axes. It seems likely that they are also of Class IC, i.e. modified buckle folds. They frequently show thickening at hinges and attenuation on the limbs and some decollement features have been noted.

These observations point towards an origin by buckling. It seems

likely that subsequent to their initial origin as buckles an element
of shear folding has entered into their development. This inter-
pretation is based on the alignment of quartz 'c' axes and rhombohedral
planes in the axial plane of the folds. Gangopadhyay and Johnson (op.cit.)

It seems possible that the absence of a well-defined lithologic
banding favours the development of this type of process in preference to
the intra-layer flowage noted in the curved axis folds, where a well-
defined intercalation of pelitic and psammitic units is apparent. In
this latter type of situation each successive band has a distinctly
different competence when compared to its neighbours and hence such
adjustments as take place tend to occur within that layer. Where
the lithologies are uniform, however, the layering no longer has much
significance and can begin to react more passively.

Explained in the terms of Donath and Parker (op.cit.) the rocks
of the Upper Komagnes Group would have had a high ductility contrast
and a moderate mean ductility, whereas the psammites of the Lower
Komagnes Group would have had a low ductility contrast and a moderate
mean ductility. This latter combination may favour the development
of passive folding later in the deformation process.

8)   Boudinage on a Minor Scale

Introduction

The rocks of Eastern Seiland show evidence of having undergone
very considerable boudinage. The rock types most frequently involved
in these structures are as follows:

1) Amphibolite in schist or psammite.

2) Granite gneiss or pegmatite in schist or psammite.

3) Psammite in schist or semi-pelite.

4) Calc-silicate bands in schist.

Boudinage is recognisable on all scales and in many cases it has been of such intensity that individual pods cannot be related to comparable pods in the immediate vicinity.

It has not always been possible to relate individual boudins to any one of the two fold-folding deformations.   It is very probable that much of the boudinage is related to the pre-F2 flattening deformation which will be discussed later (P.220).   It is clear, however, from many minor folds with boudins on their limbs, that these have been formed by stretching during the F.2 folding.

However, where a boudin pod occurs in isolation and does not appear to be related to any particular fold locally it could have arisen in one of two ways.

1) By stretching on the long limbs of large F.2 folds.

2) By compression perpendicular or sub-perpendicular to the layering in response to the above-mentioned flattening deformation.

In this respect the trends of individual boudins are of no assistance.   Where they can be observed in three-dimensions on banding or schistosity surfaces they often appear to be almost equidimensional. Fig. 39 shows this; it is a plot of the long-axes of individual boudins measured most frequently in two dimensional exposures.   In such exposures, it is often difficult to obtain a clear idea as to

FIG 39

BOUDIN LONG AXES

FIG 39
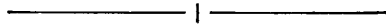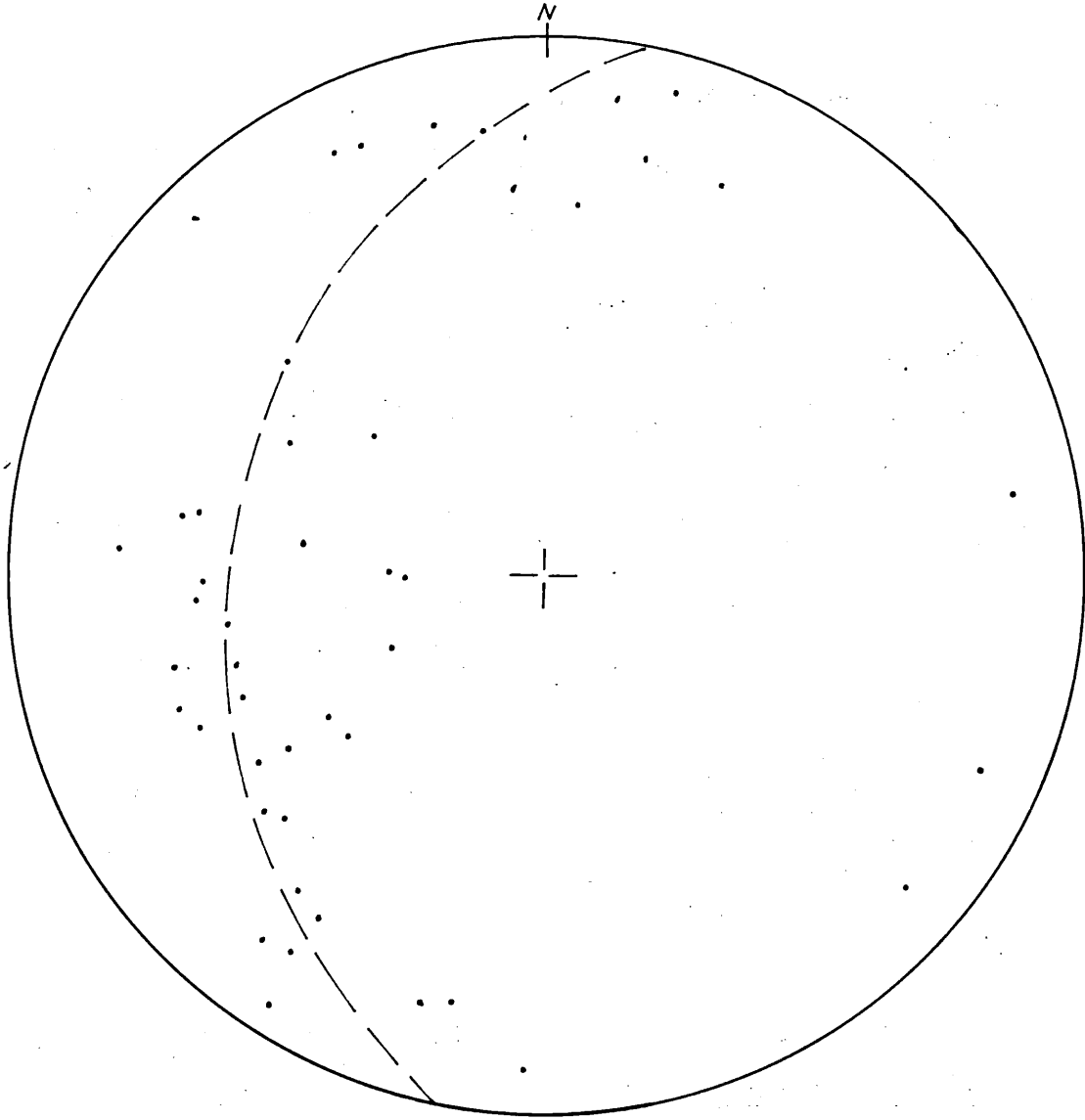
what is the true long axis of any boudin pod. The great circle

distribution shown by the stereogram is undoubtedly caused by a

combination of these factors. This great circle represents the

average disposition of the layering.

Rast (1956), in a paper on the boudinage of Central Perthshire

distinguishes between what he calls tectonic inclusions such as

relict fold closures, which he regards as being of compressional

origin, and true boudinage which is generally manifested by imperfect

oval or lozenge-shaped bodies of amphibolite. These he regards as

being of tensional origin. He recognises two varieties of boudinage

in the Perthshire area.

1) Barrel-shaped variety

In this variety the internal schistosity of the pod is parallel

to the outer surface, at the edges it is sharply pinched in. Any

two boudins are separated by a quartz vein which usually sends off-

shoots into the adjacent boudins.

2) Lozenge-shaped variety

These are formed by rotation of lozenge-shaped joint blocks, using

the joint planes as surfaces of slip.

The differences between these types Rast says can be explained as

being due to differences in the behaviour of the materials involved.

When the boudined material is effectively rigid relative to the

incompetent material, it will develop joints at an early stage of the

deformation. If these joints are not at right angles to the layering,

subsequent shearing stress will rotate each segment as the extension

proceeds.

The barrel-shaped boudins Rast suggests, are formed by a process known as necking which proceeds as follows:

a) Extension of the competent layer accompanied by plastic deformation or necking.

b) Ultimate fracture when extension surpasses the plastic limit.

c) Separation of the individual segments and formation of tectonic inclusions.

Ramberg(1955) and Ramsay (1967 p.106), see the shape of the boudin pods as being a function of the competence difference between the materials involved. If there is a large competence difference between the boudin pods and the enclosing rocks, the pods will tend to have a rectangular shape due to the absence of plastic necking previous to separation. The space between individual pods is thus usually filled with a quartz or calcite vein or with the incompetent host. When the competence difference is small pinch-and-swell structure develops by plastic eleongation. This may ultimately form lenzoid boudin pods. The incompetent material in this case flowing into the space between boudins. There is every gradation between these types. The barrel-shaped variety of Rast represents a type in which there has been some plastic flowage within the boudin pod itself, caused probably by the frictional drag exerted by the incompetent material flowing towards the boudin nodes.

Ramberg (op.cit.) appears to differ from Rast concerning the mechanism of boudin formation. He points to the frictional drag and

flowage effects seen in the incompetent material surrounding boudin

pods as being evidence that they are most frequently caused by a

compressive stress acting perpendicular or at obtuse angle to the

layering, i.e. they are of compressional origin. Coe (1959) points

out that in order to obtain boudinage in this way beds must be

isoclinally folded or the boudinage must be formed by extension

parallel to fold axes during cleavage folding. He says that the

most commonly described boudinage appears to be due to extension on

fold limbs. It seems to the author that boudinage could quite

easily be produced in the way suggested by Ramberg in non-isoclinally

folded beds, though the boudin pods would tend to be offset relative

to one another in an en-echelon fashion.

Ramberg notes that many boudins are more or less equidimensional

in shape when viewed in the plane of the schistosity, this he says is

consistent with a two-dimensional expansion in the plane of the

schistosity; in other cases boudins are rod-shaped bodies oriented

perpendicular to mineral elongation in the adjacent schists and

gneisses. These, he says, are formed by one-dimensional expansion.

The isolated pods seem in Eastern Seiland would seem to indicate that the

rocks have undergone strong two-dimensional expansion in the plane of

the schistosity.

a) Pre and Syn F.2 Boudinage

It has not been possible to relate any of the boudinage to the

F.1 deformation, the examples discussed are, therefore, associated

with the F.2 deformation. Where pods of material are arranged

systematically round a fold structure, it seems probable that their

disposition is due to the stretching that must occur, especially on

Plate 96. Boudinage associated with an F_2 fold.
Trollvann Group, east of Vasbugt.

the long limbs of folds during the folding process.

Examples of this type of structure are shown in Plates 71 (p. 62) and 96. In Plate 71 the long limbs of the folds are marked by strong attentuation and boudinage. The short limbs of the folds are relatively much thicknened. Fig. 40 A and B show some similar features. It is difficult to say in the case of 40B how much of the boudinage is due to stretching during folding and how much is due to post folding flattening; probably both have played their part in the development of this structure. Plate 97 shows one of the calc-silicate bands in the Olderbugten Group here involved in an F.2 fold in semi-pelite. It is obvious from the Plate that the calc-silicate has undergone some boudinage during the actual folding. However, these calc-silicate bands are quite common within the Olderbugten Group and they always occur as highly elongated lenses. This close association with an F.2 minor fold is the exception rather than the rule for their occurrence.

The Olderbugten Group shows much evidence of having been strongly effected by the flattening deformation. It seems likely that much of the attenuation associated with these calc-silicate bands is due to this pre-F.2 deformation. Some credence is lent to this idea by Plate 98. This shows the typical lithology seen within the shear zone; the blasto-mylonite banding and the strongly augened felspar porphyroblasts, together with the lensoid boudin of basic material.

Both Ramberg (op.cit) and Ramsay (op.cit.) attribute the development of lensoid boudin pods to a slight difference in competence between the boudin and the host rock. Pinch-and-swell structure (Plate 99 and 100) represents incipient boudinage under conditions where the extension is accompanied by plastic necking (Rast op.cit.) Further

FIG 40

FIELD SKETCHES OF BOUDINS

A

2 METRES
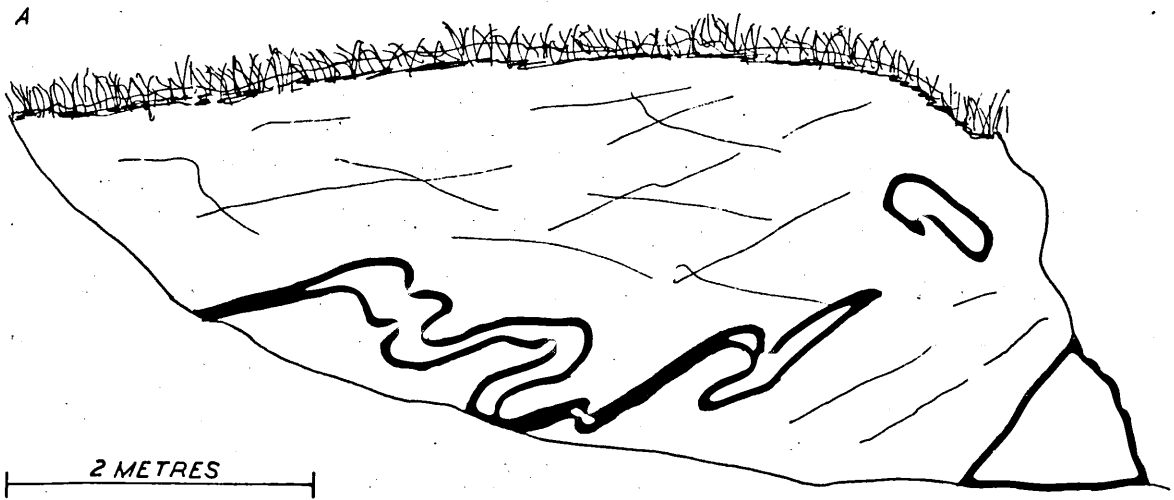
BOUDINAGED BASIC MATERIAL IN THE MASSIVE QUARTZITE
OF THE EIDVÅGEID SCHIST GROUP.

B

FELSPATHISED
PSAMMITE

1 METRE

SCHIST
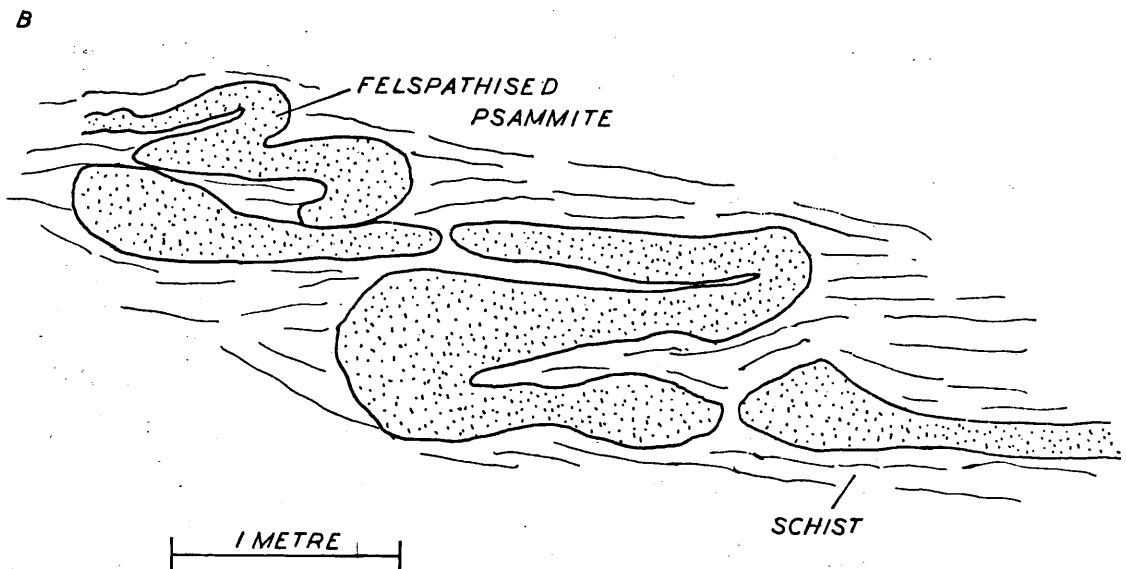
BOUDINAGE IN THE MIXED PSAMMITE AND PELITE UNIT

Plate 97. Folded and boudinaged calc-silicate band.
Olderfjord Group, east of Olderfjord.

Plate 98. Boudinaged basic material, note the augening of
the felspar porphyroblasts. Olderfjord Group,
north-east of Hønsebyvann.
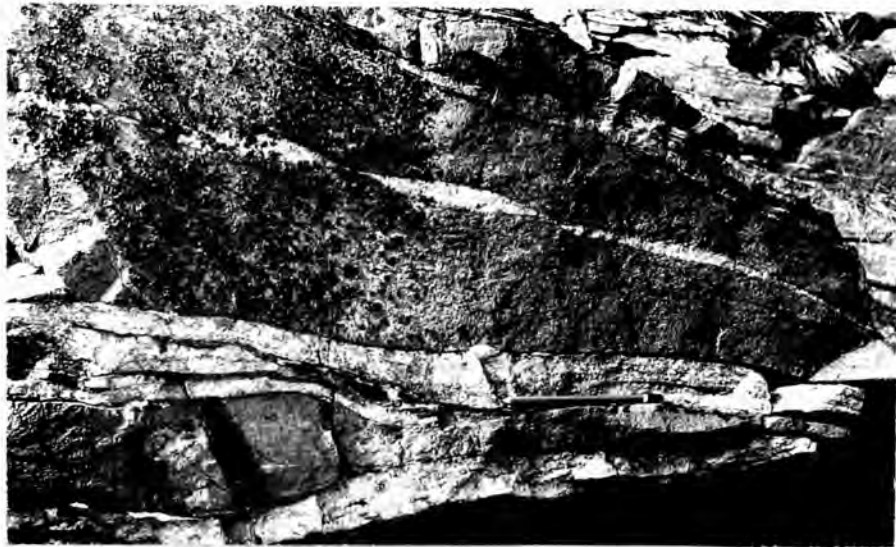
Plate 99. Pinch-and-swell in the Olderfjord Group,
         east of Olderfjord.

Plate 100. Boudinaged basic material in the Olderfjord
         Group psammites. East of Olderfjord.

extension results in separation at the nodes to give discrete pods, (e.g. Plates 101, and 102). The latter plate shows a small fold at the node of boudin pod. This was probably formed by the frictional drag exerted by flowage of the incompetent host adjacent to the competent pod towards the boudin node (Ramberg, op.cit.) Plate 103 shows a boudined felspathised psammite band within the Olderbugten Group. Here the competence difference between host and boudin may have been slightly greater because although the pods tend to maintain a lensoid form there is evidence of a quartz-filled crack at the node of the structure. It is interesting also to note at the top of the photograph the presence of a detached F.1 Fold. Rast, in his paper, says that such structures are of a compressional origin whereas true boudinage is of tensional origin. The occurrence of these two structures together seems to favour Ramberg's (op.cit.) idea that most boudinage is of a compressional origin. Plates 104 and 105 show boudinage developing in lithologies with a slightly greater competence difference. Plate 104 shows the development of some quartz segregation in the node. Plate 105 shows a classic barrel-shaped boudin. The pinching in at the node of the structure has undoubtedly been caused by plastic flowage within the boudin pod itself in response to the frict-ional drag exerted by the relatively incompetent psammite flowing towards the node. Both these examples occur in the Hammeren gabbro which is a gabbro body intimately interlayered with the psammitic country rock.

There are a number of examples of boudins which do not conform to the lensoid shape described above, they may tend towards a more equidimensional shape, or the long axis of the structure may be aligned perpendicular to the layering. Such a structure is seen in

Plate 101. Boudinaged basic material in the Olderfjord
Group, north of Storvann.

Plate 102. Boudinage in the Olderfjord Group, east
of Olderfjord.

Plate 103. Boudinage in the Olderfjord Group, east
of Olderfjord.

Plate 104. Boudinage in psammite and interbedded
Hammeren Gabbro. Olderfjord Group, east of Storvann.

Plate 105. Barrel-shaped boudin in Olderfjord Group
Psammites and Hammeren Gabbro, east of Storvann.

Plate 106. Boudin pod of gabbro in granite gneiss,
south-east of Hønsebyvann.

Plate 106. Here the disposition of the layering in the granite-gneiss host can be seen at the top left hand corner of the plates, the boudined pod clearly has its long axis transverse to this. As a general rule, the basic material contained within the meta-sediments is concordant with the layering, however, there is some evidence that this is not always the case. It seems likely that the irregular shape of this pod of basic material is due to boudinage of a basic sheet which was originally transverse to the layering. This will be discussed further in the section on major boudinage. Other examples of boudins with shapes tending towards the equidimensional are given in Plates 107 and 108. This bulbous shape, which is also recorded by Roberts (1965) on Sørøy, only seems to occur in boudined basic material. It must again be partly due to competence differences combined with possible original variation in shape of the basic material involved.

If we accept the idea that the shape of boudin pods will depend on the competence difference between the materials involved and this seems to be reasonable, it must be true that this competence difference will vary with metamorphic grades. At low grade the differences will have their maximum affect, at higher grades however, they will tend to converge. Thus at different metamorphic grades the same rock will tend to form boudins of different shape and separation. The extreme separation and general lensoid nature of the boudins on Seiland is indicative of a fairly high metamorphic grade.

b) Post F.2 Boudinage

Within the Komagnes Group occur a series of very distinctive boudinage

Plate 107. Boudin pods of gabbro in migmatitic Olderbugten
schists. Shore of Eidvaagenfjord.

Plate 108. Boudin pod of gabbro in adamellitic gneiss.

structures which are strongly discordant to the local F.2 axial trend. They appear to have been formed in a somewhat more brittle environment than is apparent for the F.2 fold structure. They are particularly well exposed on the coast section from Komagnes to Vastbugt, though a number have been recorded in the less well-exposed ground inland.

They may be described as oblique boudins in which successive boudin pods are offset relative to each other (see Plates). The nodal areas thus, have the form of asymmetric folds. These areas are frequently filled with vein quartz. This feature coupled with the frequent blunt-ended shape of the pods, gives the impression that the rocks were rather brittle during this phase of deformation. Occasionally amphibolite sheets are involved in the boudinage (Plate 109) though usually the structures occur in what appears to be relatively uniform flaggy psammite, often it is difficult to see why boudinage should have taken place at all, since the competence differences which must be present are not reflected in any distinctive changes in lithology. Plates 110, 111, 112, 113 and 114 show examples of these structures. In some cases there is evidence of a certain amount of plastic necking in the pods e.g. Plate 110 and 111, this tending to give a somewhat lensoid-shape to the pods, In other cases the pods are somewhat more rectangular, e.g. Plates 112, 113. Plate 114 illustrates the frequent angularity of the nodal area when viewed on banding surfaces. It can also be seen how this may die out rather rapidly along the axial line. Plate 115 shows a plan view of this nodal area and illustrates the fact that successive boudin pods may not have parallel alignment of their respective long axes.

A number of measurements of trend and disposition of axial plane

Plate 109. Oblique boudin in the Komagnes Group, Vasbugt.

Plate 110. Oblique boudin, Komagnes Group, Vasbugt.

Plate 111. Oblique boudin with quartz in the nodal area.
Lower Komagnes Group, Komagnes.

Plate 112. Oblique boudin in flaggy psammite of the
Komagnes Group, Vasbugt.

Plate 113. Oblique boudin, Upper Komagnes Group,
         east of Rastabyvann.

Plate 114. Oblique boudin in the Komagnes Group, Vasbugt.

were made on the asymmetric folds that develop in the nodal areas of these structures. A plot of these measurements is given in Fig. 41 A and B. It can be seen that these structures have a general westerly trend with a variable plunge, this variation in plunge is attributed to the fact that the structures have been imposed upon a layering with a slightly variable dip, this being due to the presence of F.2 folds. The axial planes of the nodal areas likewise have a somewhat variable orientation, though an average dip towards the north-west is apparent. Also plotted in Fig. 41B are a number of quartz-filled gashes which occur in the rocks, frequently they are not obviously associated with boudins. These were treated, for the purposes of measurement, as planar structures, and it can be seen that they tend to have a similar orientation to the axial plane. It is apparent that 5 out of 8 of these readings are close to the average disposition of the axial planes. These gashes are illustrated in Plates 116, 117, 118 and 119. They achieve quite large dimensions and are often rhomboid in form, (e.g. Plate 116), or may simply be quartz veins aligned across the layering. Frequently they are associated with much quartz veining along the layering. Plate 119 shows an en-echelon arrangement of gashes as does 118, though the sense of off-set of the echelons is different in these two cases. There does not seem to be any constant sense of offset in their development, it seems only possible to say that in general they have a tendency to dip to the north-west at moderate angles. There seems little doubt that these are tension gashes; the forces responsible for their development will be discussed presently.

Examination of the Plates indicates that the gashes are associated with rather gentle asymmetric folds of the same type as those seen

FIG 41

STRUCTURAL DATA FROM THE OBLIQUE BOUDINS AT KOMAGNES

A

B
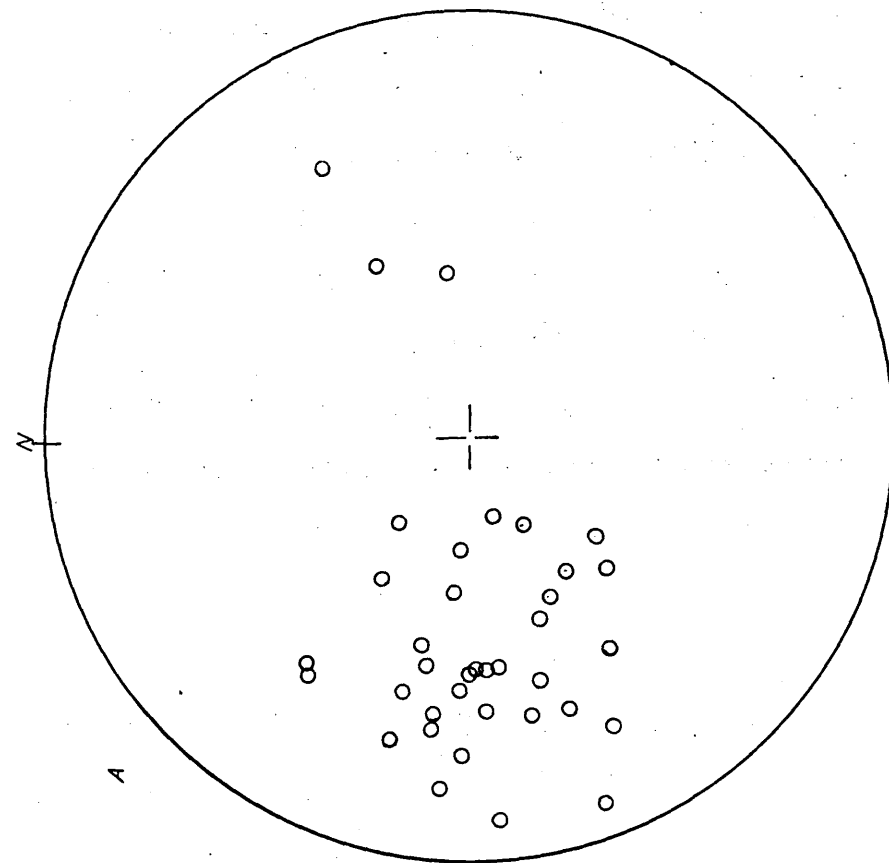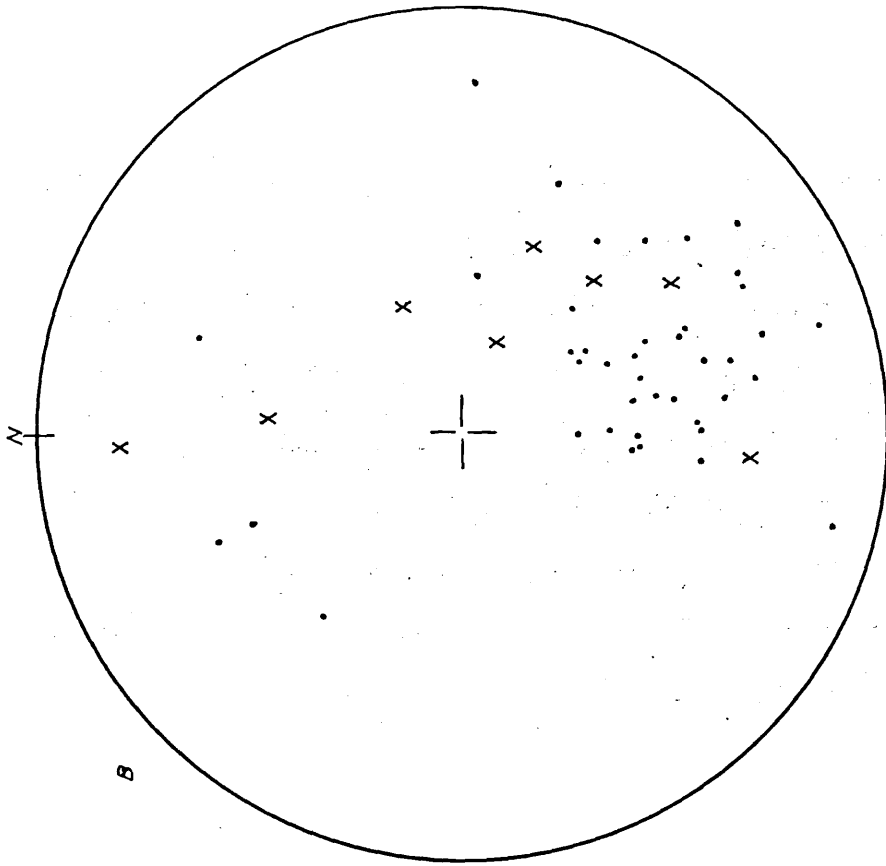
• POLES TO AXIAL PLANES

x POLES TO QUARTZ FILLED GASHES

o AXIAL TRENDS

Plate 115. Plan view of the nodal area of an oblique
boudin. Komagnes Group, Komagnes.

Plate 116. Quartz-filled tension gash, Komagnes Group, Komagnes.

Plate 117. Contorted quartz-filled tension gash,
            Komagnes Group, Komagnes.

Plate 118. Quartz-filled tension gashes, Komagnes Group, Komagnes.

Plate 119. En-echelon quartz-filled gashes, Komagnes
          Group, Komagnes.

Plate 120. Incipient boudinage developing in association with
           a quartz-filled tension gash. Komagnes Group, Komagnes.

in the boudin nodes. Plate 117 affords evidence that these folds post-date the development of the gashes. Just right of bottom centre of the plate, associated with the left hand diagonal quartz vein there is a little contiguous vein occuring as an offshoot parallel to the layering. This little vein has been folded into a small monoclinal flexure. This strongly suggests that the whole vein was emplaced before folding took place and that the gashes have been rotated by the folding.

Plate 120 shows a very similar quartz-filled tension gash associated with an incipient boudin. This boudin is of the barrel-shaped type of Rast. Here again it is rather difficult to see why boudinage should have been initiated in this lithology.

In summary it would appear that those rocks have been subjected to a period of tension, which led to the development of tension gashes and probably initiated the boudinage.

Along the beach exposures discussed above a number of folds have been observed which have similar trends and styles to the folds associated with the boudin nodes. The structures, however, are not obviously related to boudin nodes neither are they related to tension gashes. Examples of these structures are given in Plates 121, 122, and 123. It could be argued in the case of Plates 122 and 123, that the boudin pods are not seen but occur beneath the surface. This, however, cannot be true for Plate 121. Further evidence of the independent development of these folds is provided by Plate 124. This shows a boudin pod which has undergone folding by a structure with a similar trend and disposition of axial plane as the

Plate 121. Fold associated with the oblique boudins, Komagnes Group, Komagnes.

Plate 122. Monoclinal fold associated with the oblique boudins. Komagnes Group, Komagnes.

Plate 123. Monoclinal fold associated with the oblique boudins.
Note the vein quartz in the axial plane. Komagnes
Group, Komagnes.

Plate 124. Folded boudin, Komagnes Group, Komagnes.

folds in the nodes of the boudins.

Mode of Formation of Oblique Boudins and Associated Structure.

From the above observations it is possible to recognise a series of structural events which may be summarized as follows:-

1) Development of tension gashes plus probable initiation of boudinage.

2) Rotation of boudins and tension gashes leading also to the development of folds.

These events can be regarded as being part of a progressive deformation sequence. Their temporal arrangement is almost certainly due to the change in orientation of the stress-field to which the rocks were subjected. Ramsay (1967) has argued that a shearing couple is not necessary for the formation of oblique boudins. His experiments indicate that these structures can be formed by an asymmetric arrangement of the layering within the strain ellipsoid. This is undoubtedly true for the modelwhich Ramsay discusses however, in the case of the Komagnes boudins, the sequence of structural events which have been discussed above, are not easily reconciled with the simple oblique compressive stress proposed by Ramsay, particularly the later formed folds are difficult to explain using this model. The whole deformation sequence seems to be easily explained by the development of a shear couple.

The earliest feature of this deformation sequence appears to be the development of the tension gashes; it was suggested that these had probably been rotated by the subsequent folding such that

they dipped towards the north-west, to what degree this rotation has occurred is very difficult to estimate. However, it seems likely that they were initiated at a high angle to the layering, possibly with a slight to moderate dip towards the north-west. This type of orientation is consistent with a compressive stress either normal or at a high angle to the layering, since tension gashes are said to form normal to the direction of maximum extension of the strain ellipsoid. This compressive stress was almost certainly responsible for the initiation of the boudinage as well. The rest of the deformation, the rotation of boudins and folding, can be explained as being due to a shear couple acting along the layering in an approximately south-easterly direction. This latter statement is based upon the fact that the axial planes of the boudin nodes dip towards the north-west, thus the sense of transport, the 'a' direction is towards the south-east. This hypothesis is doubly attractive since the sense of translation on the thrusts which separate the Pre-Cambrian from the overlying Caledonides on the adjacent mainland is towards the south-east. (Reitan 1963). If the Komagnes boudins can be linked with these major thrusts it would appear that initially the rocks were subjected to a period of compression approximately normal to the layering, a shear couple then developed as a result of the thrusting; this was responsible for the rotation of the boudins and the development of folds which may be regarded as being drag folds. Boudins and tension gashes were especially susceptible to these rotational forces since they represented disruptions of the planar aspect of the layering, thus inhibiting interlayer slip.

It is of considerable interest that in a pelitic layer occurring within the psammites on the peninsula at Komagnes, a number of tight folds occur. The average trend of their axes is 282° with a plunge

of 32°. The axial planes strike 033° and dip 44°N. This is roughly the attitude of the rotated boudins. It seems likely that these are true drag folds resulting from the rotational shear couple discussed above.

The deformation sequence outlined above can probably be regarded as being continuous, the initial compression across the layering was perhaps due to the building up of the stresses prior to thrusting. When the yield point of the material was reached rupture and thrusting occurred giving rise to the rotational shear couple. Thus, this sequence of structural events appears to have been caused by a change in the position of the strain ellipsoid relative to the layering rather than a change in the orientation of the layering within the strain ellipsoid of the type discussed by Flinn (1962) and Ramsay (1967).

A statement of Reitan (1963) concerning the Pre-Cambria rocks of the Komagfjord tectonic window on the adjacent mainland is relevant to the arguments stated above. "The general antiforms of the Pre-Cambrian rocks constituted an obstruction to the overriding Caledonides with the result that the foremost (most north-westerly) portions of the Pre-Cambrian rocks were dragged along by the Caledonides being pulled up the curving thrust planes along which rupture occurred". It seems likely that this dragging mechanism could well be responsible for the Komagnes boudins. The obvious low metamorphic grade prevailing during the formation of these structures is consistent with the late brittle nature of the mainland thrusts.

9) Boudinage on a Major Scale

The intense boudinage to which this part of Seiland has been

subjected is further underlined by an examination of the map

(see wallet). One of the striking features is the disjointed

lensoid nature of many of the lithological groups. This

applies particularly to the various gabbro bodies, especially

the Hammeren gabbro in the west of the area. The outcrop of this

body is dominated by a series of lenses which are generally

elongated in a north-south direction, in the field the dip always

steepens markedly at the contract of these bodies. This type of

outcrop pattern can be interpreted in one of three ways:

1) The pods are periclinal fold cores

2) They are boudin pods.

3) They are both.

Examination of the smaller pods of gabbro that occur in the

fjell above Eidvaagenfjord indicates that they are certainly

boudin pods; the banding in the enclosing schist can be seen to

be sweeping round the gabbro in typical boudin fashion. This

interpretation also seems to be the most likely one for the larger

pods of Hammeren gabbro seen in the west of the area.

Just south of the large outcrop of the Eidvaagtinn adamellite

there are two large gabbro pods with anomalous trends. In contrast

to the other large pods they have an approximate east-west trend.

This may be explained satisfactorily by the thesis that the original

gabbro intrusions were discordant to the layering, whereas most of the gabbroic material was intruded as sheets concordant to the layering. This type of discordance has been observed elsewhere in the fjell to the west of Vasbugtvann. The forces responsible for the boudinage acting on such a cross-cutting body would lead to the development of the anomalous trends described above.

Further evidence of the intense boudinage of basic sheets is seen in the Komagnes Group, within the outcrop of the psammites in southern Vasbugt occurs a lens of highly altered basic rock. Inland of Komagnes occurs another lens of an identical rock type. These lenses are some 2 kilometres apart. If they were once joined this separation would seem to indicate very intense stretching; it is of course hypothetical to suggest that they were once joined since there may be intermediate pods which are not exposed.

The development of these large scale boudins is not confined to the gabbro sheets. It can be seen that the granite outcrop above Eidvaagenfjord has a rather fringe-like form. The dips in the enclosing Olderbugten Group are quaquaversal to this outcrop pattern. Thus, these bodies of granite, several tens of metres in a real extent, are in fact large boudins pods. From the map it would seem that the stretching direction for these pods is approximately north-south, this being normal to the plunge of the pod as discerned from the map. This is, of course, consistent with the direction of elongation of the gabbro pods. However, it is not possible to estimate the degree of stretching that has occurred parallel to the plunge of the pods. Evidence provided by the minor boudins suggested that this stretching was two dimensional, leading to the development of equidimensional pods Ramberg (op.cit).

This is probably the case for the large pods as well.

Thus a picture emerges on Eastern Seiland of a number of phases of very intense boudinage, a great deal of this probablypreceded F.2 though F.2 itself was undoubtedly responsible for the development of some boudinage. A further phase of boudinage occurred after F.2 resulting in the formation of oblique boudin structures which are frequently associated with folds and tension gashes. It seems likely that this phase is related to the late Caledonian thrusting developed on the adjacent mainland.

10) Kink Bands

In the psammites of the Upper Komagnes Group in the Rastaby-Jernelv area a number of structures occur which completely post-date the F.2 folding. They appear to be mainly restricted to the coastal exposures. Morphologically they are somewhat variable, some resembling joint drags, (Flinn 1952) and others are more akin to kink folds. For the purposes of this discussion the general term kink fold will be used. Examples of their structures are given in Plates 125 and 126.

In Plate 125 the structures are very obviously related to a set of joints which cut the layering in the psammite at a high angle. These joints have divided the rock into a number of prisms and movement has apparently occurred along the joint planes. This movement is expressed in the rocks by a stepping of the layering with an almost consistent throw along joints from right to left of the plate. This stepping is often marked by rotation of the layering to give little monoclinal flexures. Occasionally little horst and graben-

Plate 125. Kink folds, Komagnes Group, Rastaby.

Plate 126. Kink folds, Komagnes Group, north of Jernelv.

like structures develop on the layering where conjugate joint sets occur, as in the centre of the photograph. Generally however, at any one exposure one joint plane predominates and the sense of stepping is constant.

Fig. 42A illustrates a specimen in a banded pink psammite in which a conjugate set of these pink folds are developed. The kinked segments are no more than 2-3 mm wide and they tend to be slightly sinuous. One set of kinks is developed strongly in preference to the other which occurs as only one or two striations on the banding surfaces. The axes of the two sets diverge and their sense of offset is opposite. On fracture surfaces viewed normal to the kink axes it can be seen that they are associated with very fine joints which do not completelysplit the rock (see Fig. 42A). These joints which have obviously been healed are marked by a slight dark green discolouration of the rocks, caused by low grade mineral assemblages.

Plate 126 shows another set of these kink folds. They appear to be rather angular monoclinal folds with no distinct plane of fracture marking the limits of the kinked zone. Close examination in the hand specimen, however, reveals the presence of brittle fracture zones within the folds together with tension cracks approximately normal to the kink plane. (see Fig. 42A, also later). The fold axes on banding surfaces are marked by a lineation which is the surface expression of the joints forming the kink zones. The differences in appearance of the two sets represented in the Plates is to some degree a function of the lithological differences and probably also of the slightly different stress field to which each set was subjected. This seems likely because each type forms a distinct maxima on a stereogram.

FIG 42

A/ CONJUGATE KINKS

MINOR SET

DOMINANT
SET

DISCOLOURATION
ALONG JOINTS

|— 2CMS —|

B/

JOINTS

JOINTS

①

③

②

● TENSION VOIDS

|— 2CMS —|
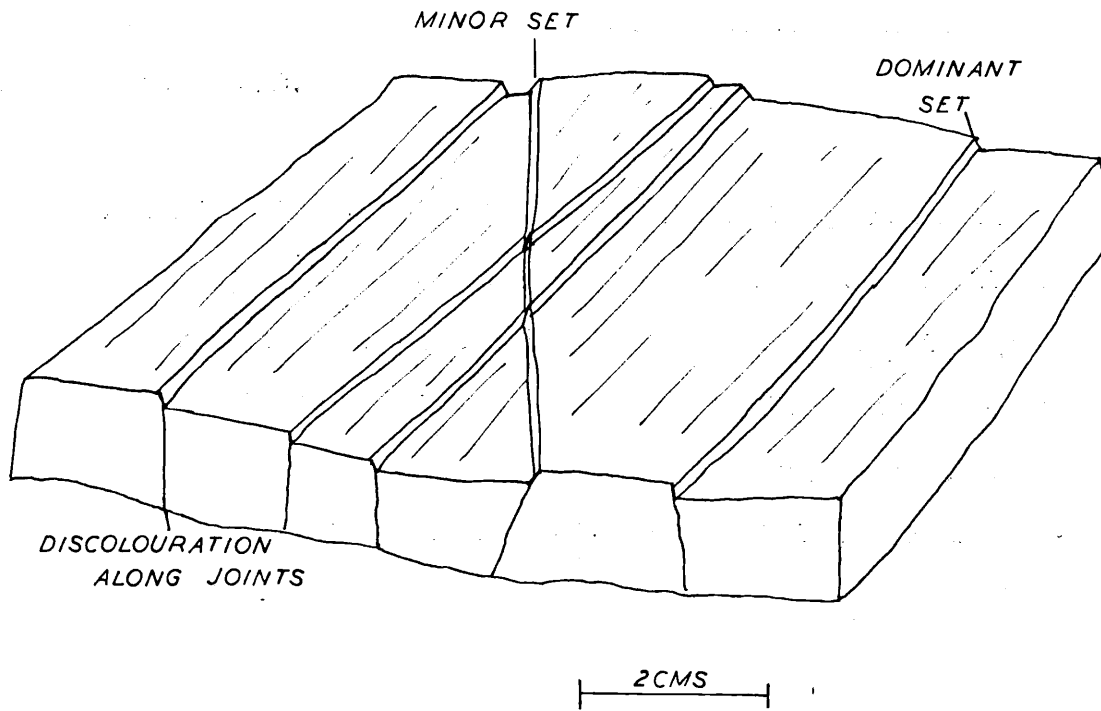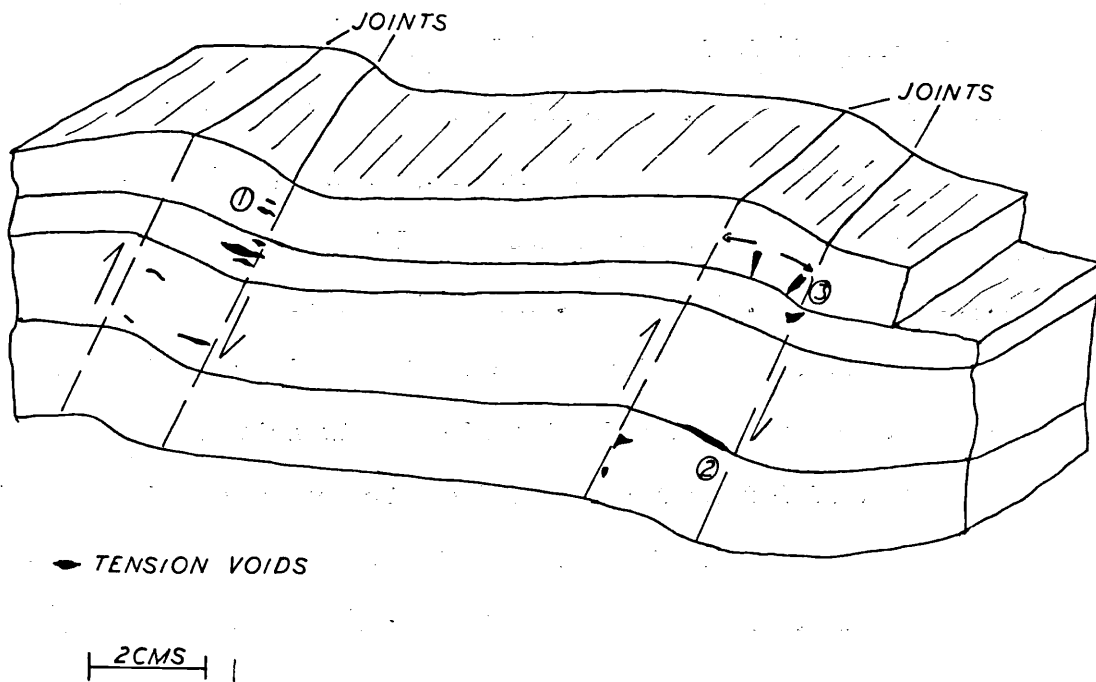
The Fig. 43A, B and C show the orientation of the two sets of kink folds. In Fig. 43A the set plunging approximately west (set a) are represented (Plate 125). The set plunging just east of south (set b) are represented by Plate 126. Fig. 43B shows for (set (a) the poles to the joints with which the structures are associated and for (set b) the poles to the axial planes of the kink folds. The average planes statistically estimated from the poles of Fig. 43B are represented in Fig. 43B together with the mean position of the kink axes.

Thus, it is clear from these observations that we have a conjugate set of kink folds which are occasionally seen in combination in a single exposure (Fig. 42A), but are generally not seen together.

Discussion
_____

Most of the literature (Johnson 1956, Ramsay 1962, Ramsay DM and Sturt 1963, Dewey 1965) seems agreed that conjugate kink zones result from the operation of a couple developing on a set of conjugate shear planes. Ramsay, in a later work (1967) questions this hypothesis on the basis that in nature where primary conjugate shears develop, the maximum compressive stress always bisects the acute angle between the shear planes. Naturally occurring kink zones, however, mostly have the principal compressive stress bisecting the obtuse angle between the kink plane. From this Ramsay goes on to suggest (p.453) that the probable mechanism of deformation is the development of an initial buckle in response to the stresses. This initial buckle imposes on the layering a specific limiting dip value which does not change during subsuequent deformation. The kinked zone then grows laterally by migration of its bounding planes.

Dewey (1965) advances a hypothesis of initial bend gliding followed by kinking on planes normal to the glide direction (i.e. the layering) to explain kinks where the maximum compressive stress bisects the acute angle between the kink surfaces. The Seiland kink-folds do not present any problems concerning the orientation of the stress field, since it is apparent that the maximum compressive stress does bisect the acute angle between the kink-planes (79°). A close study of the structures yields a lot of information about their mechanism of formation which is relevant to the theories put forward by other authors. Consider 42 B. This is a sketch of the specimen taken from the locality of Plate 126. There are two features of interest; at (1) there are a number of tension gashes with their long axes approximately normal to the planes bounding the kink zones. These gashes are not filled with vein quartz. Thewe structures are entirely consistent with the development of a shear couple in the sense shown by the arrows. Likewise at (2) the banding surfaces of the psammite have been forced apart by a similar process. Secondly, at (3) there are two small wedge-shaped fractures aligned at a high angle to the banding. These are consistent with the development of tension in the extrados of a fold undergoing flexural folding. This flexural process is apparent in the rather gentle rounded nature of the fold hinges. If the formation of these zones were caused entirely by the development of shears bounding the kink zones it might be expected:-

a) For the rock to be transected completely by shear planes.

b) For the hinges to be rather more angular.

Thus in this specimen of banded psammite the kinks have been formed by flexural folding followed by movement in response to a shear

FIG 43
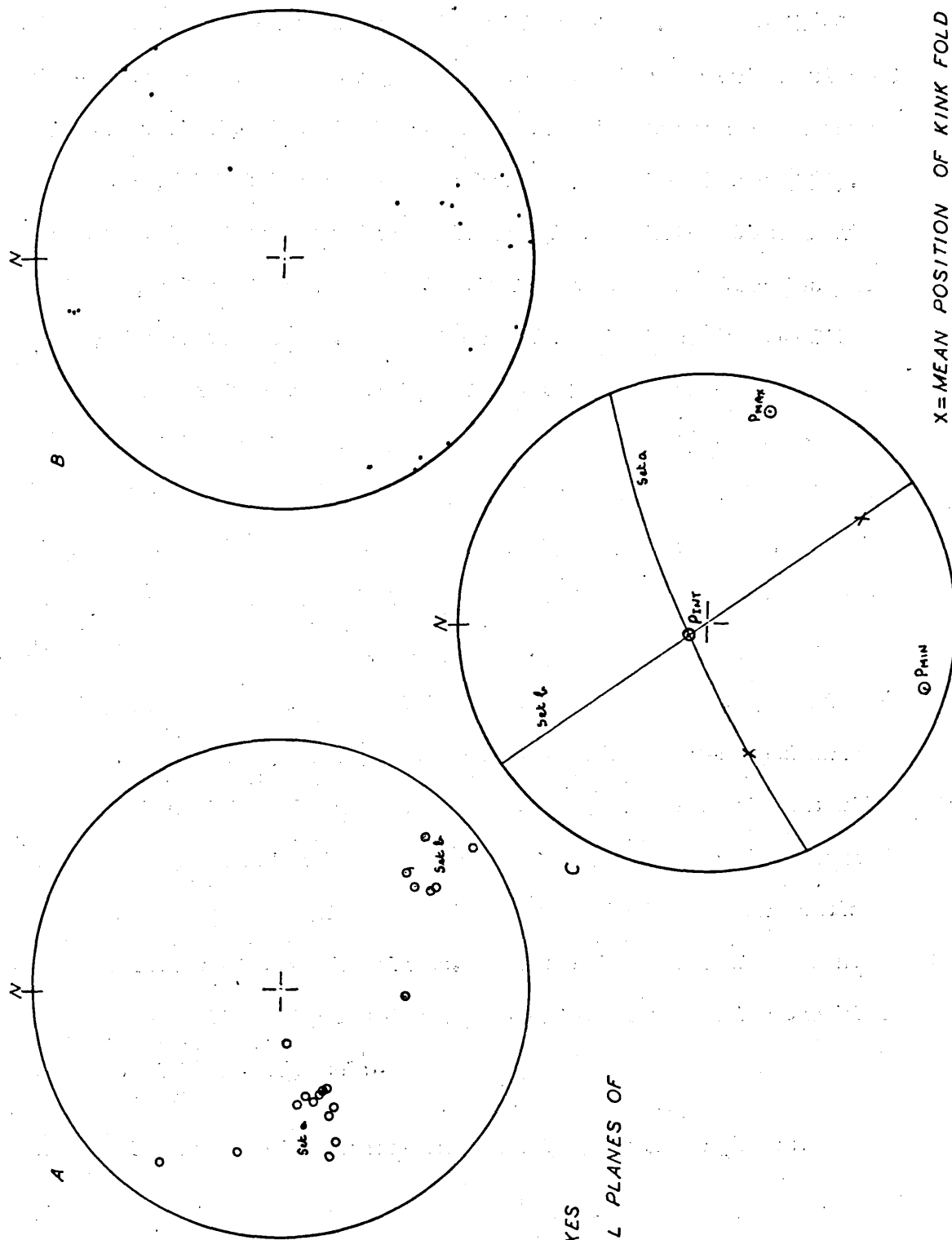
A = KINK FOLD AXES

B = POLES TO AXIAL PLANES OF
    KINK FOLDS

X = MEAN POSITION OF KINK FOLD AXES

couple.   This is probably also true of the rock type shown in Plate

125;   here, however, the brittle fracture planes are much more obvious.

Ramsay (1962) indicates a method for positioning the stress axes

of conjugate kink zones;   the intersection of the two kink planes give

the position of the intermediate stress axis (P.Int).   The maximum

compressive stress ($P.Max$) is the bisector of the acute angle between

the kink planes and ($P.Min$) is at right angles to the other two axes.

This has been done in Fig. 43C.   It can be seen that the kinematic 'b'

axis represented by (P.Int), has a very steep plunge and is thus

inclined at a high angle to the layering, this orientation explains the

diverging kink-fold axes described above and in Fig.42A.   It is very

difficult to obtain a great circle representing the average inclination

of the layering since this has been disturbed by F.2 and is, therefore,

not statistically planar.   However, it seems likely that in general none

of the stress axes lie within the layering, the overall symmetry of the

kinks, is therefore, triclinic.   The two sets have not developed

together with sufficient frequency and strength to prove this by

observation.

Regarding the stresses responsible for the formation of these

structures a number of possibilities are worth considering.   Set (a)

particularly resembles the joint drag structure described by Flinn (1952)

from Shetland.   In these structures Flinn notes two senses of movement,

one in the manner of a normal fault, i.e. down the dip of the joint and

a second movement along the strike of the joint.   The second movement

he regards as being the most important.   He suggest that the structure

developed as a conjugate set to a fault which occurs in a narrow sound

between the main island and a small offshore island.   This fault, he

suggests, has a slight horizontal component of movement, though its main

throw is vertical.

Knill (1961) describing joint drags from Mid-Argyllshire suggests that they are conjugate to the Great Glen fault. Both these authors, however, record only one set of joint drags these being conjugate to fault systems with some transcurrent component. On Seiland, however, there are two sets developed as described above. These forming a conjugate pair with diverging axis. A number of possibilities will now be considered for their origin:-

1) It will be noted from the map that a number of small faults occur in the Rastaby-Jernely area, the most southerly of these faults has a fairly moderate throw to the north. The strike of the joints associated with set (a) closely approximates to the strike of this fault. The strike of set (b) approximates to the trend of Vargsund, which separates the coast of Seiland from the mainland. It is possible that there is a large fault along the axis of this sound. The throw of this fault, as deduced from the sense of downthrow on the axial plane of set (b), would be to the east. The major drawback of this hypothesis is that the sense of throw on the joints of set (a) is opposite to that of the faults seen on the map.

2) The second possibility is that the conjugate shear planes are associated with the late Caledonian thrusting described previously. The sense of asymmetry of set (a) is the same as that of the oblique boudins at Komagnes, the axial trends and disposition of axial planes are also roughly coicident. The (P Max) postulated in Fig.43C is, however, only just south of east, whereas the thrusting direction postulated by Reitan (1963) is towards the SE.

Consideration of all the factors relating to these structures would seem to indicate that the first hypothesis is the most satisfactory for the following reasons.

1) Comparison of Fig. 43C and Fig. 44C which represents the average disposition of the major joint planes, and rose diagram Fig. 45, which reprsents the distribution of major lineaments taken from the aerial photographs and fault planes measured in the field, would seem to indicate a reasonably good agreement in trend.

2) The same metamorphic assemblages viz calcite epidote and chlorite occur both in the joints associated with the kink folds and in the retrogressed rocks adjacent to fault planes (see Post-F.2 metamorphism. P.197). This would seem to be good evidence of the temporal association of joints, faults and kink zones.

11) Joints and Faults

Jointing is developed to various degrees in the different rock types; it tends to be best developed in the rather massive lithologies particularly the massive quartzites within the Eidvogeid schist and in the Olderbugten Group where it forms very massive blocks which are the most prominent linear features on the aerial photographs. The Eidvogeid schist is not very extensively jointed, neither in general are the igneous rocks.

Fig. 44A shows the distribution of joints within the Komagnes, Eidvageid Schist and Trollvann Groups. Fig. 44B, the Olderfjord and Olderbugten Groups. Both have girdles with disinct maxima. The average planes represented by these maxima are given in Fig. 44C.

FIG 44   POLES TO JOINTS
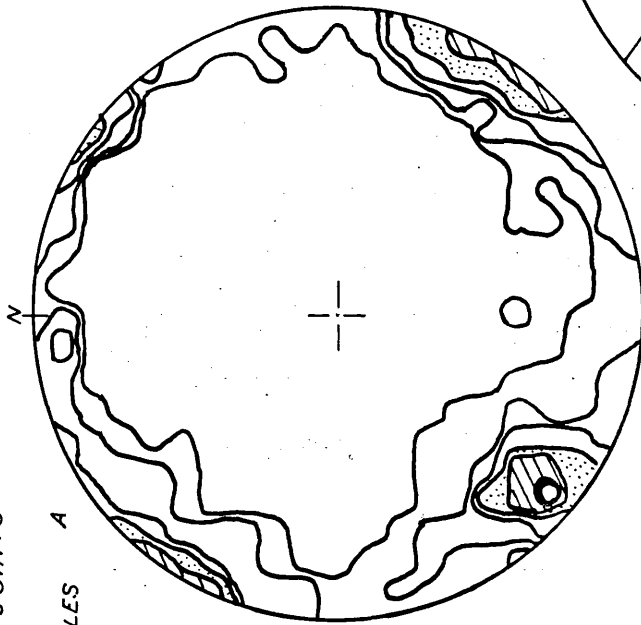
413 POLES   A

594 POLES   B

CONTOURS 6 ■ 5 ▨ 4 ▨ 3,2,1 % PER 1 % AREA
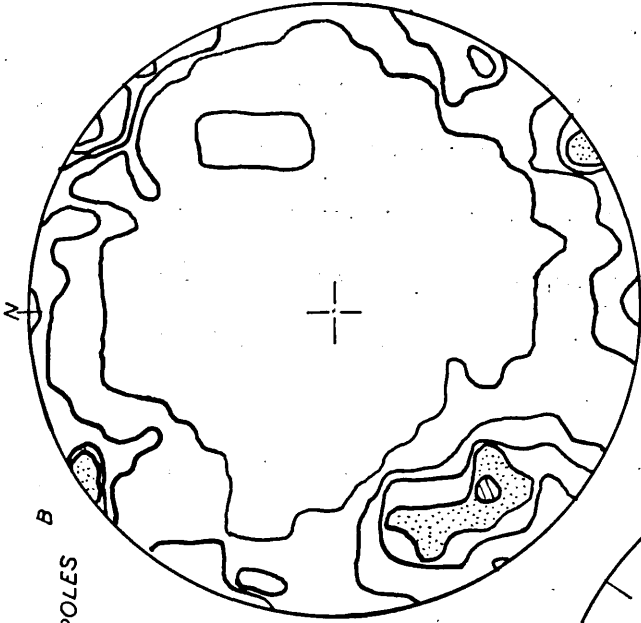
SCHMIDT NET LOWER HEMISPHERE

A—THE KOMAGNES, EIDVÅGEID SCHIST.
AND TROLLVANN GROUPS
--- IN C

B—THE OLDERBUGTEN AND OLDERFJORD
GROUPS
—— IN C

C—AVERAGE MAJOR JOINT-PLANES
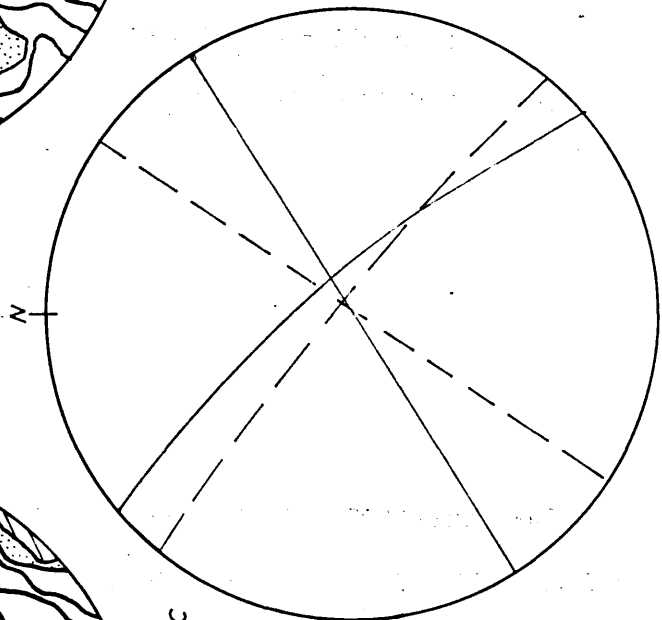
Itcan be seen that in both cases the planes are approximately orthogonal, though for the two areas the respective planes do not coincide exactly.

A number of other interesting features emerge from a study of the faults and a system of strongly developed linear features which are very conspicuous on the aerial photographs. These latter features appear in the ground as trench-like straight lineaments. Often the rocks within the trenches are brecciated to some degree but where litholog- ical control is present, there is often no evidence of displacement so they may more correctly be termed shatter-belts. A striking example of this type of phenomenon is seen to the east of Vasbugtvann; slightly less than a kilometre after leaving the lake the river takes a sudden turn to the north to occupy a straight trench which is very striking on the aerial photographs. In the rose Fig. 45 it can be seen that there is a very conspicuous maxima trendly roughly NE-SW, this corresponds reasonably well to the joint set maxima with a similar trend seen on Fig. 44C. It seems likely that these lineaments are master joints along which there has been some shattering.

There are not a large number offaults in the area, those that have been mapped are generally quite small structures.

1) Two small faults displace the outcrop of the pink psammite at Jernelv and just to the north. The most southerly of the faults is associated with a bending of the pink psammite to the north of the fault plane. This is clearly seen on the map since the dips in this horizon become locally eastwards. Downthrow to the North.

2) A fault just to the west of Vasbugtvann displacing the outcrop of the Trollvann Psammite Group. This occupies one of the trench-like features. Downthrow N.W.

3) Fault almost 2 kilometres south-west of the Vasbugtvann displacing the Hammeren Gabbro outcrop. Since this displaces a little anti-form-synform system, it probably has a slight dextral strike-slip component. Downthrow S.E.

4) Small fault to the west of the point where the Trollvann psammite passes out into Eidvaagenfjord. This fault truncates a gabbro boudin. Throw unknown.

5a) The largest fault in the area cuts the outcrop of the Hønseby Gabbro, it can be traced for more than 4 kilometres along its strike. It is very conspicuous in the aerial photograph by virtue of the fact that it is marked by a shatter belt which has eroded out into both a trench and a scarp at various places along its length. The downthrow is slight, to the south-west.

5b) Another small fault just to the north-east of 5a) along the southerly contact of the gabbro. This has a throw to the NE.

6) A small fault on the eastern side of Storvann displaces the outcrop of the Hammeren gabbro. Downthrow SW.

7) A number of small faults displaces the outcrops of massive quartzite within the Eidvågeid schist in Northern Vasbugt. Adjacent to one such fault Plate 127 was taken, it can be seen that a well developed conjugate joint set is developed in the massive quartzite. One of the joint directions corresponds very closely to the trend of the fault plane, the two, therefore, seem to be genetically related.
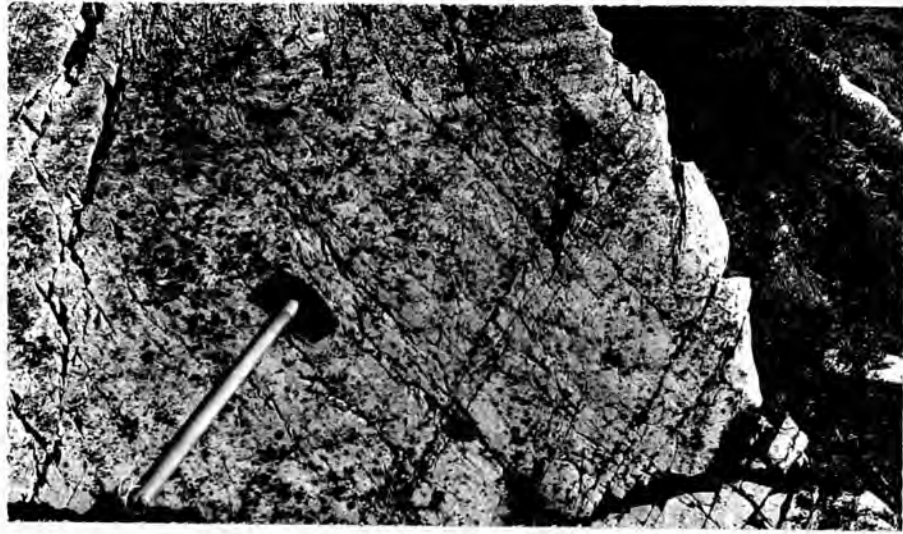
Plate 127. Conjugate joints in a massive quartzite band
of the Eidvågeid Schist Group, northern Vasbugt.

Plate 128. Joint face markings in the Ultra Basic Mass, Vasbugt.

In all the cases described above it was not possible to determine the attitude of the fault planes.

From Fig. 45 it can be seen the trend of the faults agrees broadly with the trend of the major shatter belts. The major fault (5a) is marked (X). This corresponds to a very minor maxima in the shatter belt trends. This direction (318°) however, corresponds well with the other orthogonal maxima of Fig. 44C.

Thus a picture emerges of faults, shatter-belts and joints tending to fall in trend, into one of two orthogonal maxima. This strongly suggests that all three features are genetically related.

It has already been suggested that there is a close similarity of trend between the joints, faults and the planes bounding kink folds at Rastaby. (Compare Figs. 43C, 44C and 45). The meta-morphic assemblages developing in association with these structures are also very similar. It seems clear therefore, that the forces responsiblefor their development are the same.

With regard to the joint planes themselves, they are generally smooth surfaces which slice through the various lithologies in-discriminately. More detailed analysis of the Figs. 44A and B indicated that the joint directions were not significantly different in the various metasedimentary groups. This type of jointing appears to accord with that recorded in a discussion by De Sitter (1964, p.105) on the work of Parker (1942) in Central and Northern New York and Northern Pennslyvania. Here the following is recorded. "The joints of the compound set (1) are remarkably plane, they slice clearly through hard concretions in weak strata and pass without deviation

FIG 45

ROSE DIAGRAM OF 58 MAJOR LINEAMENTS TAKEN FROM THE AERIAL PHOTOGRAPHS
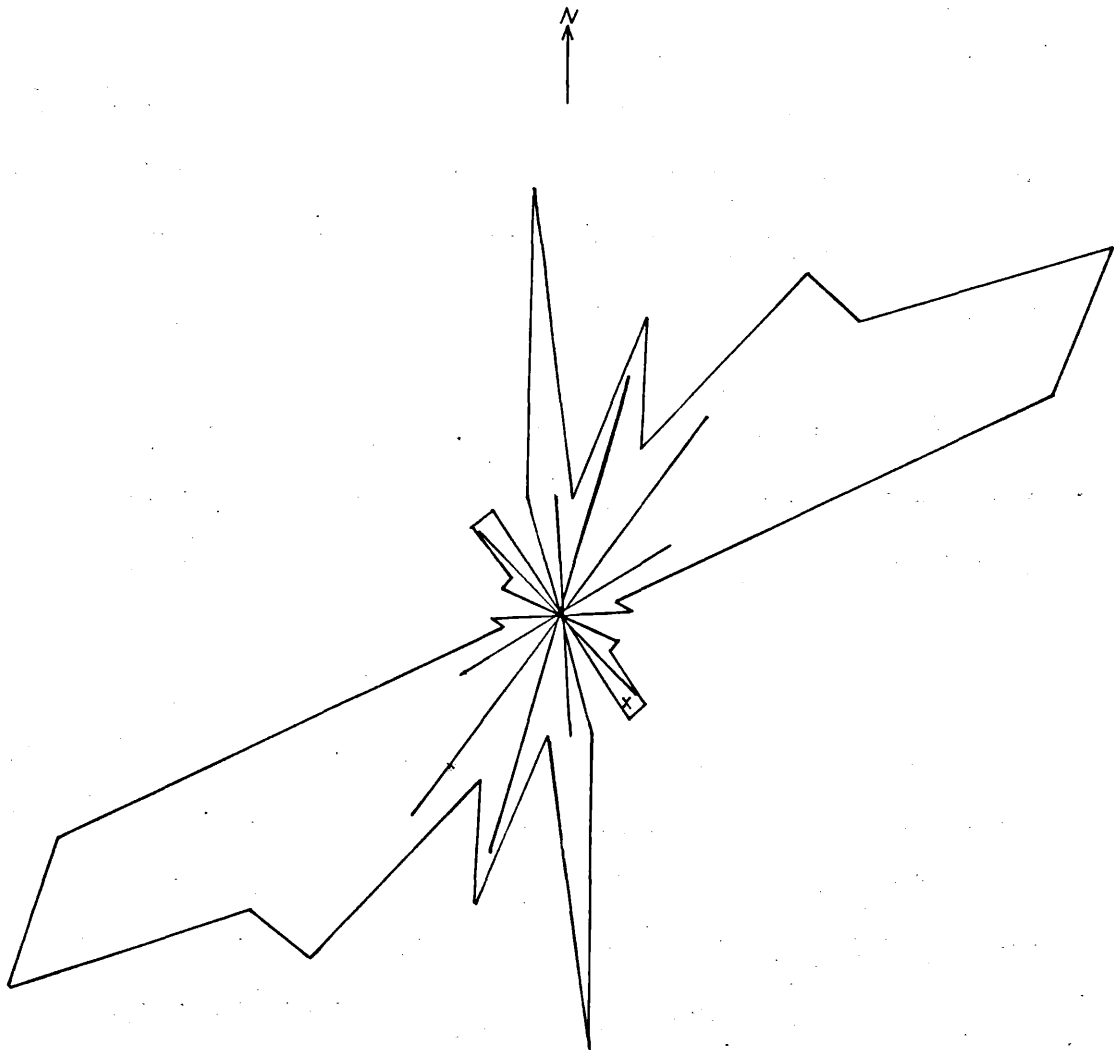
SINGLE LINES ARE FAULTS

⊢————⊣ = 1 FAULT (TOTAL OF 7)

through extreme cross-bedding etc. They are obviously shear joints".
Tension joints on the other hand are described as "curved and irregular
with a rough torn appearance".

The joints on Seiland in general appear to conform more to the
first type based on the above criteria, i.e. they are shear joints.
If this is correct, the compressive stress responsible for their
formation was presumably either approximately north-south or east-west,
since these vectors bisect the angles between the major joint planes.
It is of interest that the (P Max) derived for the formation of the kink
zones was postulated on trending just south of east. It was suggested
that these structures were perhaps related to a major fault system out
in Vargsund.

These general observations, would seem to cross-cut the suggestion
of Price (1959) that joints are post-tectonic phenomena related to
lateral expansion during uplift. They may, however, he says be
related to the movement picture because their orientation is determined
by the residual stresses. The Seiland structures, however, appear to
have formed in response to well defined stresses which gave rise not
only to joints but also to faults and kink folds during a definite
phase of hydrothermal metamorphism.

Finally a joint-face phenomena recorded in the ultrabasic mass
of southern Vasbugt is of interest. One of the joint faces in this mass
bears a number of concentric ridges, (Plate 128). These ridges are
slightly asymmetric, facing away from the centre of curvature, their
wavelength is somewhat variable, as is their amplitude. Crossing the
ridges there are faint slickensides. The relative age of these
structures is difficult to tell. Solomen and Hill (1962) have recorded
similar structures in the Renison Bell area of Tasmania, though on a

smaller scale. They point out the similarity between these structures and those in glass caused by local tensional stresses following point impact. They suggest that these tensional stresses could either have been tectonically produced or they may be due to point impact fracture caused by local blasting. It is perhaps significant that on Seiland, close by this locality there has been some blasting to excavate sites for telegraph poles.

B) Major Structure

Introduction

Attempts to elucidate the major structure of Eastern Seiland are rendered difficult by a number of factors:

I) Isoclinal folding and the relative constancy of dip makes major closures inconspicuous.

2) The absence of distinctive marker horizons.

3) Anomalous evidence of fold vergence.

Despite these difficulties it has been possible to work out a reasonably clear picture of the major structure of the area.

One of the most striking features of this structure of Eastern Seiland is the apparent ability of each of the stratigraphic groups to fold as a discrete unit, with little interfolding taking place between the various groups. It is possible that this is simply a matter of scale, but it is true to say at least in the scale of the area, that the rocks have responded to the various deformation phases as discrete entitites.

1) Major Structure due to F1.

It has been possible to distinguish only one large fold of F.1 age
on the area. This is the structure defined by massive quartzite bands
within the Eidvågeid Schist Group. The schistosity in the schist is
sensibly parallel to the banding in the quartzites, and is therefore
interpreted as an axial plane structure related to the early fold.
The fold has an isoclinal style; this may well be due however, to
subsequent deformation.

F.1 folds and lineations within the Komagnes, Eidvågeid Schist and
Trollvann Groups tend to concentrate in a maximum plunging just west of
north. Again the effects of F.2 on the original disposition of these
lineations is indeterminable; their relative constancy over a wide
area suggests however, that this has been minimal. Roberts (1965)
quotes the average F.1 trend on north-east Sørøy as being 010°. This
value is of the same order as that recorded on Seiland, the discrepancy
may in any case, be due to the arcuate nature of the axes of F.1 nappes.

2) Major Folds due to F.2

Because of the tendency for each lithological group to fold as a
unit, it is thought expedient to describe the F.2 major folds as they
occur within the stratigraphic divisions.

a) The Komagnes Group

The folds within the lower Komagnes Group are characterised by
attenuated long limbs and short limbs consisting of vertically stacked
tight folds. This picture is common to the Komagnes Group as a whole
but the Lower Komagnes Group presents some difficulties of inter-
pretation.

1) It is very difficult to determine the true sense of vergence of the folds even though they are frequently well-exposed.

2) The absence of well-defined lithological differences precludes the actual mapping of folds.

The Upper Komagnes Groups presents fewer problems because within it there are a number of well-defined lithologies, which can be mapped to elucidate the structure. In this Group the vergence of both the minor folds and the major folds is consistent with the structural pattern shown in Section 5 (see wallet). This may be illustrated by reference to the outcrop pattern of the fold in the Pink Psammite, west of Jernelv. The strongly attenuated long limb and the stacked short limb are readily apparent, the syn-F.2 boudinage. It was perhaps also responsible for the separation of the ultra-basic pods in the middle of the outcrop of the Komagnes Group.

Given this clear structural picture it is possible to make some reasoned extrapolations of the structures within the Lower Komagnes Group; there does not seem to be any good reason to postulate any change in vergence of the folds within the Lower Komagnes Group. The suggested structure of the Komagnes Group as a whole is, therefore, indicated in Section 3. There seems no reason to doubt that by analogy with the minor folds, the major folds in the Komagnes Group have arcuate axes.

b) The Eidvågeid Schist Group

There are no major folds of F.2 ages within this Group. A small number of F.2 minor folds fold the F.1 schistosity. Most of the deformation was probably taken up either by slip or by homogeneous

deformation within the schist.

c) The Trollvann Psammite Group

Although F.2 minor folds abound within this Group there does not seem to be a consistent sense of vergence that might give some clue to the overall structure within the Group. There appears to be very little infolding with the enclosing schist groups.

The structural pattern observed within the Komagnes Group, viz the vertically stacked short limbs and attenuated long limbs, is also seen within this group.

d) The Olderbugten Group

In the absent of any more competent material to stiffen-up this lithology it would seem that folds on any scale, do not form extensively. In this respect the lithology behaved in a similar manner to the Eidvågeid schist. It has, however, been possible to recognise within the schist a number of large folds. These can be mapped using gabbro and adamellite sheets as marker bands. A good example of this is seen on Eidvaagtinn, where relatively broad folds can be mapped using the adamellite as the marker band. The geometry of these folds is illustrated in section.2. As the folds are traced southwards into the schist, they appear to die out. No trace of structures is seen in the transgressive gabbro pods, neither can they be traced by variations in the attitude of the foliation in the schist. Similarly, study of the northern boundary of the adamellite, which represents the base of the schist, reveals that here also the folds have died out. As well as this lateral discontinuity of axis it seems likely that the folds also die out with depth, since at no stage can the fold pattern be traced

down into the structurally lower Trollvann Group. An analagous pattern of folding can also be seen just to the west of the Eidvaagtinn folds, involving adamellite and gabbro sheets. (Section 2).

Thus, the folds appear to be periclinal structures which are both laterally and vertically discontinuous. This pattern conforms quite well to the periclinal nature of the minor folds already described.

e) The Olderfjord Group

The large number of sheets of basic material which occur within the Olderfjord Group have made the mapping of a number major F.2 folds possible. There are however two principal difficulties:

1) The lenses of gabbro may be interpreted either as boudins or periclinal fold cores.

2) The isoclinal nature of many of the folds, espedially in the metasediments, makes the tracing of the axial zones difficult.

One common factor that does emerge from a study of the large folds, however, is their periclinal nature. This is particularly well illustrated in the block diagram of section 4. This shows the out-crop pattern that occurs in the interlayered gabbro-metasediment complex just west of Vasbugtvann. Two antiformal structures folding metasediment with gabbro are visible in the field. Their profiles indicate that they have a moderately apressed style. If the outcrop of this gabbro sheet is traced to its southern extremity, it can be seen from the straight nature of the boundary, that the folds have died out.

This is also true of the northern boundary of the complex. The folds are thus dying out along their plunges, i.e. they are periclinal structures. To what extent this lateral discontinuity is controlled by the wedging out of the gabbro sheet is difficult to say. Suffice it to say that a similar pattern has been observed in the Eidvågtinn folds and indeed in all the minor folds of Eastern Seiland.

Sections 2 and 3 show further examples of the structures occurring within the Olderfjord Group. It can be seen that the style of these structures is rather variable, ranging from the tight isoclines seen in the area south of Storvann in Section 3 to the rather more open structure on the same section just to the east. This difference in style may perahps be explained by variations in thickness of the gabbro sheets; the thinner sheets tending to form the tightest folds. The most easterly gabbro pod seen in section 3 may well be a core belonging to the same fold system as the other cores to the west. However, its rather pinched-out nature seen in a cliff at the southern end of the outcrop, suggests that in fact it is a boudin core.

A further example of tight F.2 folds is seen in the boundary of the Hønseby gabbro just to the north of the head of Hønsebyfjord in the north of the area. There is no doubt that this is, in fact, a fold and not a lens of metasediment, since a number of minor folds can be observed which have the correct vergence to the more major structure.

The largest fold that occurs on the area can be seen in the cliff between Olderfjord and Olderbugten. This fold involves the psammites of the Olderfjord Group with the schists of the Olderbugten Group in the core. The structure is the exception to the generally observed fact that folding does not occur between the various Groups. This suggests that the statement that this may merely be a matter of scale,

is in fact true. The fold is isoclinal in style and the author at
one stage, considered it to be of F.1 age. However, to the west, B.
Robbins (personal communication) has observed that it folds round the
axial planes of F.1 minor folds. It is therefore, of F.2 age. The
style is certainly comparable with other F.2 folds described above.
The presumed geometry is given in Section 5. The Olderbugten schists
were not observed again inland despite the fact that a steep-sided valley
cuts the presumed axial trend of the fold. It seems likely therefore,
that the structure has a moderate to steep plunge; and by analogy
with the other structures is probably periclinal.

Summary

A number of important facts emerge from this study:

1) The fold structures observed appear to be confined within each lith-
ological group though this may be simply a matter of scale. In relation
to the wider region.

2) Major folds where they can be mapped generally have a periclinal
aspect in common with the minor folds of the area. This lateral
discontinuity is matched by a discontinuity with depth.

3) There is a change in style across the areas. The Komagnes Group
is characterised by folds with attenuated long limbs and a tendency
for the short limbs to consist of a number of vertically stacked
moderately appressed folds. The Olderfjord and Olderbugten Groups,
however, tend to have folds where the limbs are of rather more equal length
and the vertical stacking in short limbs is absent.

Perhaps the most significant feature that emerges from this
study is the change in style that takes place across the area.
Despite this the whole area has one feature in common.   That is
the presence of both major and minor folds with curved axes and
axial planes lying within the layering.

In seeking for an explanation of this change of style a
number of facts are significant.   The rocks above the Komagnes
Group suffered, previous to the F.2 deformation, a strong flattening
deformation which resulted locally in the formation of very flinty
annealed blasto-mylonitic rocks of considerably increased competence
from the parent schists and psammites.   This coupled with the inter-
calation of a large number of highly competent gabbro and adamellite
sheets together with strong felspathisation, resulted in the rocks
of the Olderfjord and the Olderbugten Groups being in a very different
state of competence, at the onset of F.2, to the psammite and schists
of the Komagnes Group.   This must, to a considerable degree, be
responsible for the change in style.   There are, however, two
other facts which undoubtedly exerted influence.   Firstly a change
in the metamorphic grade takes place across the area.   This will
be discussed in greater detail later, but basically the grade
increases from east to west, though by the onset of F.2 the meta-
morphic grade over the whole area had waned to the same level.
Secondly, the fold styles developed in the Olderfjord and Olderbugten
Groups were subsequently modified by the superincumbent weight of
the vast post - F.2. basic and ultra-basic plutons of Western
Seiland.   This inevitably led to a general tightening up of the fold
style in the Olderfjord and Olderbugteh Groups.   It would appear
that this modification decreases with structural depth since the
Komagnes Group seems to be least affected.

Therefore, the rather more massive nature of the rocks of the two upper Groups derived chiefly from the first two factors discussed above, seems to have precluded the development of the fold styles described in the Komagnes Group. This is mirrored in the minor folds elsewhere. For example, the massive quartzites within the Eidvågeid Schist Group tend to deform into rather concentric styled folds though still with curved axes. This situation on the major scale, subsequently modified by the weight of the vast plutons, could easily have given rise to the structural picture as we see it today. Finally, the arguments put forward in the section on mechanisms of fold formation concerning the mode of origin of curved axis folds, probably apply in no less measures to the major folds of the area.

3) The F.3 Deformation

Introduction

If an imaginary line is taken from the central part of Storvann in the west, to Vasbugt, in the east, it will be noted that north of this line there is a prounounced swing in strike in the Olderfjord Group. To the south the strike is on average some 10-20° west of north, to the north, however, it is just east of north. This type of strike swing is a feature that has been recorded elsewhere in the region, e.g. by Ramsay and Sturt (1963) and Roberts (1965) on the island of Sørøy. Here the swings give rise to north - south, and east - west strike belts. The swing recorded by Roberts (1968) on N.E. Sørøy is accompanied by a change in the symmetry of the F.2 folds. The F.2 folds in the north-south belts have a monoclinic symmetry whereas those in the east-west belt have an orthorhombic symmetry. From these observations Roberts (1968) states that the swings must be regarded as a feature of either the first

or second episode of deformation on Sørøy. He goes on later to say "the evidence to date could suggest that the swing is an inherent property of the F.1 episode".

On Seiland the strike swing is not accompanied by any change in symmetry of the F.2 folds, indeed, evidence will be put forward to show that F.2 folds are refolded within the swing belt.

The F.3 folds

There are a number of major folds which are associated with the strike swing.

1) The southern contact of the Hønseby gabbro to the west of Hønseby is folded into two steeply plunging monoclinal structures.

2) A similar monoclinal buckle is shown in the screen of psammite within the Hønseby gabbro to the north of 1).

3) There is a swing of strike in the screens of psammite within the Hammeren Gabbro 2 Km. south of Storvann.

4) A large fold system that can be detected by a study of the attitude of the foliation in the gneissic rocks to the north of Hønsebyvann. This fold is depicted in Section 1. One of the peculiarities of this structure is the way in which just north of Hønsebyvann the dips suddenly rapidly steepen in the psammite across a particular line. This line, which represents the boundary between the Olderbugteh schists and the overlying Olderfjord psammites is peculiar; the dip and strike of the underlying Olderbugten schists is quite normal right to the point where the lithology is lost under

Eidvaågenfjord.   The overlying psammites however, have the typical

strike-swing attitude.

The strongest evidence that this strike swing is a post-F.2

phenomenon is provided by Figs. 46A and B.   46A shows the distribution

of presumed F.2 folds in the area of the strike swing.   This compared

with the plots of F.2 folds outside the area of the strike swing has

a rather random attitude.   Perhaps more significant, however, is

46B.   This shows the attitude of the axial planes of the folds in

46A, clearly they have been folded about an axis trending 234° and

plunging at 10° since they define a great circle of which this point is

the pole.   The rather random arrangement of F.2 axes in 46A may be

explained by the refolding in this manner, of the typical non-cylind-

roidal F.2 folds.

Thus, in this northern part of the area there seems to have

taken place a rotation in a west to east sense, of the Olderfjord

psammites, using the planes between this group and the underlying

Olderbugten schists as a plane of decollement.   This rotational

movemeht gave rise to the large folds described above.   The

monoclinal folds in the Honseby gabbro contact possibly represent

large wrinkles that developed on the approximate axis of bending.

The folds depicted in Plate 129 and 130 are very probably

F.3 minor folds, Plate 129 comes from the north of Eidvagenfjord while

the second 130 comes from just south of Hønsebyvann.

These have been marked on 46A and apparently they do not bear

any clear relationship to the axis of refolding.   This, however,

might be expected because of their wide separation in space and the

FIG 46   STRUCTURAL DATA FROM THE OLDERFJORD GROUP IN THE STRIKE–SWING AREA

POLES TO AXIAL PLANES OF F.2
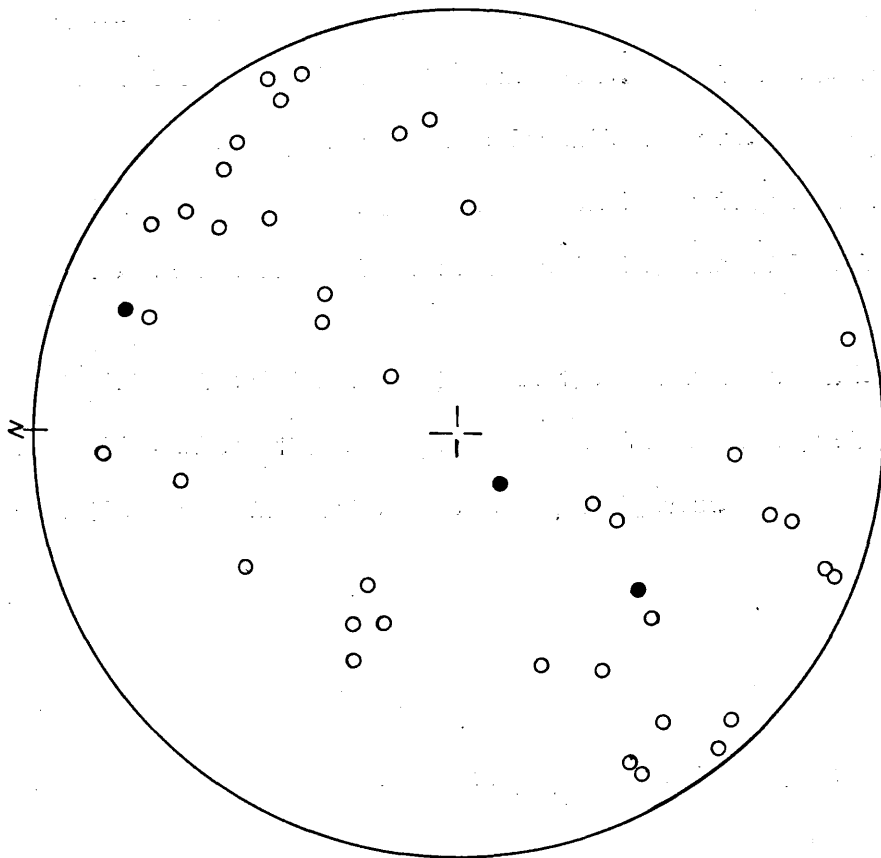
○ F.2 FOLD AXES
● F.3 FOLD AXES
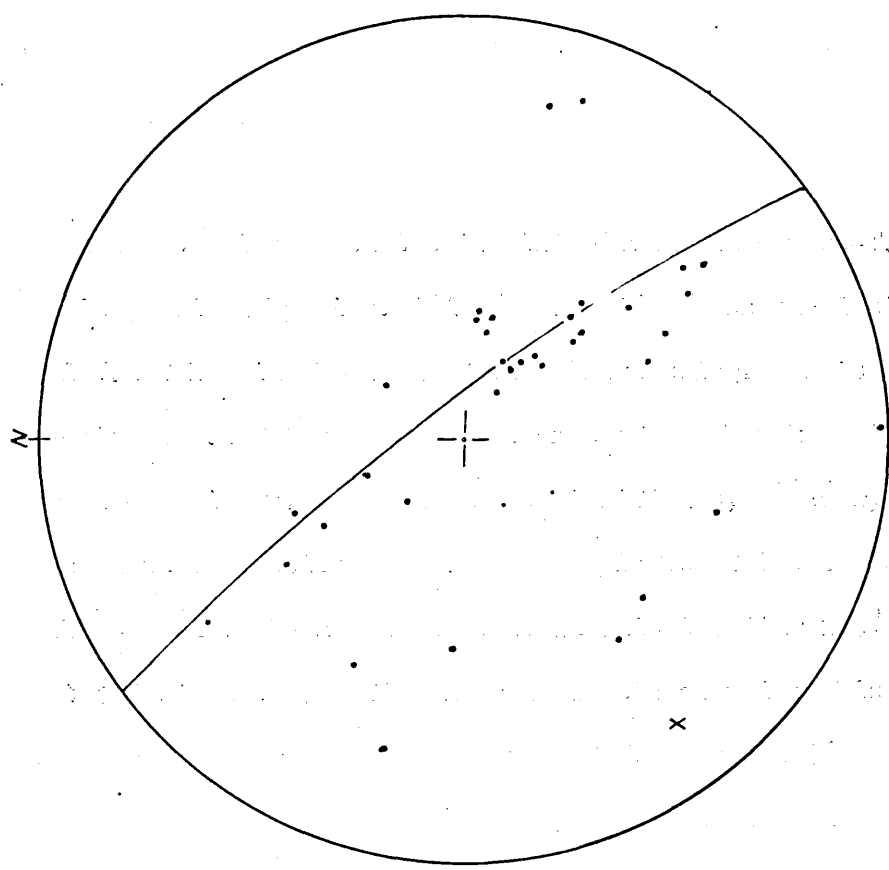
Plate 129. F₃ fold in the Olderfjord Group, shore of Eidvaagenfjord.

Plate 130. F₃ warp, Olderfjord Group, south-west of Hønsebyvann.

complexity of both the lithologies and the forces responsible for their development.

A study of the proposed plane of decollement does not yield very much useful information. The plane across which the described changes in dip occur is very distinct, however, the author has failed to find any evidence of brecciation along this zone and the under-lying Olderbugten schist appears to be normal.

The swing in strike 3) above is rather odd, it occurs well outside the main strike-swing belt. The average dip pf the layering of the psammite screens is comparable to that of the short limbs of the monoclines at the Hønseby gabbro contact. It seems to be most conveniently explained by the F.3 movements described above.

Regarding the forces responsible for these structures, preliminary mapping of the area immediately to the west by J.B. Jackson (personal communication) has revealed other swings of a similar nature, giving a rather sigmoidal aspect to the regional outcrop pattern. In the absence of further information unwarranted speculation is undesirable, however, it is possible that these swings are in some way related to the intrusion of the great plutons of Western Seiland.

The F.3[1] Folds of Komagnes

In the Komagnes area a number of folds have been observed which post-date F.2.

Plate 131. F₃ warp in the Komagnes Group, Vasbugt.

It is clear also that they post-date the oblique boudins described from the Komagnes area, since they can be seen to be deforming the layering containing these structures. They are apparently sensibly coaxial with the local F.2 trend in the psammites of the Lower Komagnes Group.

Geometrically, these folds are monoclinal in style. (Plate 131). Many F.2 folds near or in the steep limbs of these structures have their axial planes warped into an easterly dipping attitude. This tendency is clearly seen in Fig. 9B. (P.51.) Here the plot of F.2 axial planes is tending to show a great circle distribution. The axis of this great circle is 333° with a plunge of 14°.

The relationship of these structures to the F.3 folds of the strike swing belt is not known. Clearly they are very late in the structural evolution of the area. The question must remain open as to whether they are coeval with F.3 in the north For this reason they will be designated as $F.3^{1}$ folds.

A summary of the structural and metamorphic events is given in Fig. 52.

Chapter 4

Metamorphism and Petrology

Introduction

The complex series of structural events that have occurred on
Eastern Seiland have left an indelible mark on the various mineral
assemblages that have arisen during the concomitant phases of meta-
morphism.   A sequence of crystallisation of metamorphic minerals can
therefore be established in relation to the minerals that have grown
during the deformation phases.   This sequehce of metamorphic
crystallisation will be considered under the following headings:

1) The Syn-F.1 metamorphism

2) The Static period between F.1 and F.2.

3) The Syn-F.2 metamorphism.

4) The Post-F.2 metamorphism.

1) The Syn-F.1 Metamorphism

It is impossible to gauge how much subsequent metamorphism has
modified the effects of the syn F.1 metamorphism.   It is clear however
from this section study of F.1 folds that the micaceous minerals
maintain a good parallelism to the axial-planes of the folds.   The
structure defined by these micas is therefore designated as (S.1.)
It is possible however, that the mica is in part mimetic to F.1
(Plate 132)
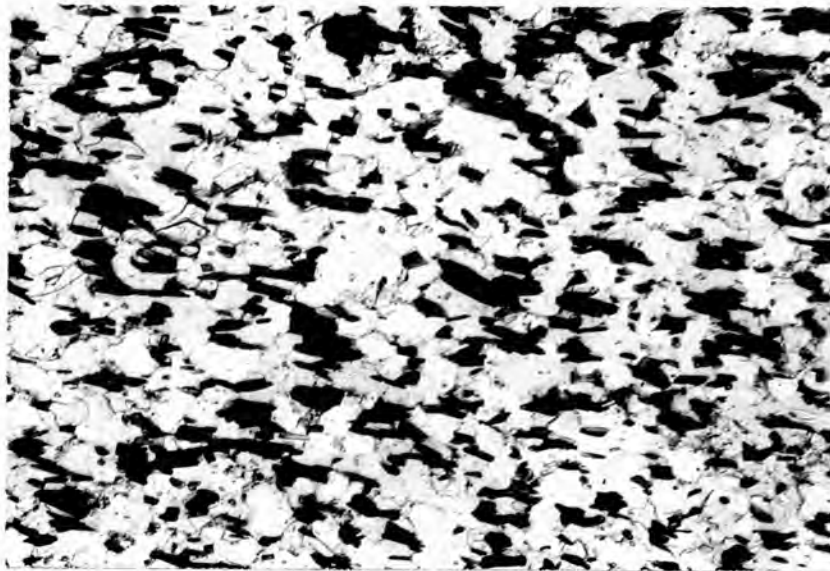
Plate 132. Biotite defining S.I. Upper Komagnes
Group, Rastaby. x 40, P.P.L.

In the Upper Komagnes Group and the metasedimentary lithologies structually above, this schistosity is clearly overgrown by large post-F.1 garnet porphyroblasts. The inclusion fabric of these porphyroblasts is recrystallised to a coarse-grained mosaic containing large biotite flakes which are obviously post-F.1 in origin.

A further clue to the grade of metamorphism during F.1 can be obtained by study of the minerals which parallel F.1. lineations in the schists. A specimen of a basic schist from the Lower Komagnes Group contains acicular actinolite crystals which define a lineation. This lineation is folded by very tight folds of F.2. age. The amphibole, which makes up about 70% of the rock is embedded in a granular aggregate of quartz and untwinned felspar with yellowish-green epidote overgrowing the amphibole (see Plate 133)* The optical properties of the amphibole are:

$$Opt \quad - \quad ve \quad 2V = \quad 75 \quad - \quad 80^{\circ}$$

$$\overset{\wedge}{Z\ C} = 19^{\circ} \quad Y = b$$

Birefringence = o.o37

Pleochroic Scheme $\quad$ Z = Apple green

$\qquad\qquad\qquad\qquad$ X = Pale Yellow

* In the following sections all optical data, unless stated otherwise, was obtained with a standard petrological microscope.

a,b,c, $\quad$ refer to the crystallographic axes

X,Y,Z $\quad$ refer to the three principal refractive

$\qquad$ indices of the optical indicatrix.

$\qquad$ (Hartshorne and Stuart 1969)
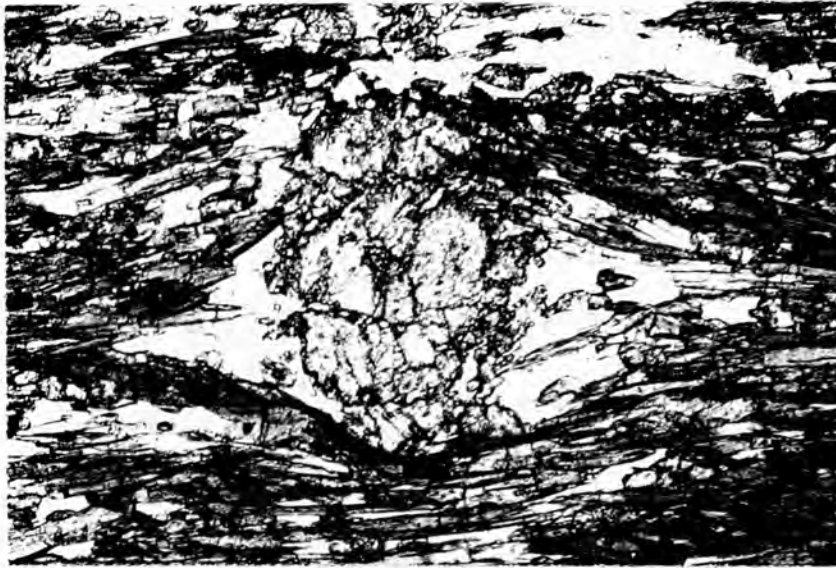
Plate 133. Epidote porphyroblast overgrowing syn F_1 actinolite. Lower Komagnes Group, Vasbugt. x 40, P.P.L.
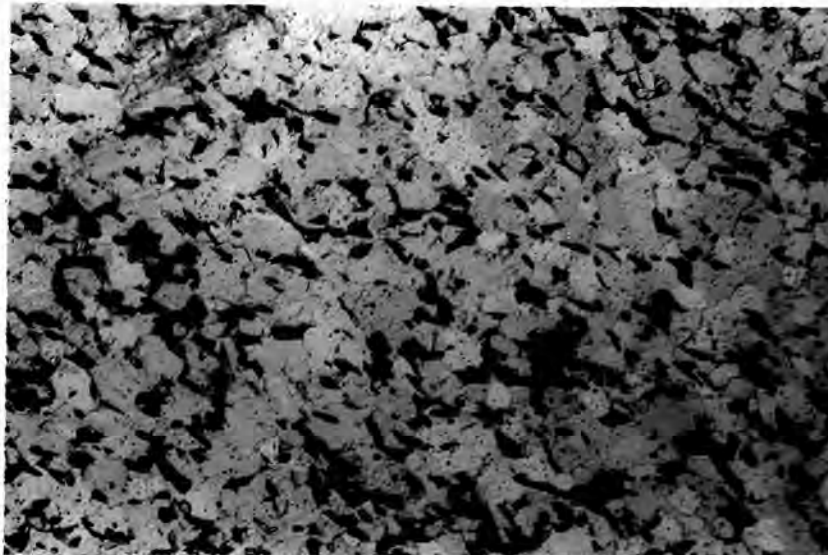
Plate 134. Decussate biotite, Upper Komagnes Group, Rastaby. x 40, P.P.L.

These properties plus the acicular habit indicate that the mineral is actinolitic rather than hornblendic. Winkler (1967, p.100) states that above the Quartz-Albite-Epidote-Biotite Sub-facies of the Green Schist Facies, actinolite is replaced by more aluminous hornblende. Although this rock almost certainly reached a higher grade subsequent to F.1 this transformation may not have taken place here due to paucity of aluminium in the rock. Elsewhere where hornblendic amphibole has been observed it is generally post-F.1, unlineated, rather poikiloblastic and contains inclusions of earlier formed mineral phases.

This paragenesis actinolite-biotite-muscovite observed in the various rock types seems to indicate that F.1 took place under middle Green Schist Facies conditions, namely in the Quartz-Albite Epidote Biotite Sub-facies. This is consistent with the observations of Roberts (1965) on Sørøy, who bases his conclusion on the presence of inclusions of biotite withinpost-F.1 garnets. On Sørøy, however, the subsequent picture is very much complicated by a very pervasive F.2 schistosity which obliterates all earlier fabrics. On Seiland the F.2 schistosity is not nearly as pervasive and therefore the dominant schistosity is an F.1 structure defined by the growth of mica. There is no evidence of epidote growth during F.1.

2) The Static Period between F.1 and F.2

During this phase of metamorphism significant differences in metamorphic grade are apparent in the various metasedimentary groups. These differences will, therefore, be considered under the following sub-headings :

A) The Komagnes Group

B) The Eidvågeid Schist Group

The Trollvann Group

The Olderbugten Group

The Olderfjord Group

The metasedimentary groups structurally overlying the Komagnes Group will be treated together since the observable variations of metamorphic grade within them are of a minor nature.

a)   The Komagnes Group

Most of the porphyroblastic mineral phases that occur in the rocks of the Komagnes Group date from this phase of metamorphism, Study of these various minerals indicates that slightly different grades of metamorphism were achieved in the Lower and Upper Komagnes Groups during this period.   This is based on two observations; firstly, garnet occurs only very sporadically in the Lower Group, that is the psammite up to and including the calc-silicate schist.   In the Upper group, however, it is relatively abundant.   This may partly be due to compositional differences in the original sediments though it is not easy to see why the mineral could not have developed in the more pelitic horizons of the lower division.

The second and more important reason is that there is a change in the felspar composition across approximately the same line. In the Lower Group the felspar is albitic and takes the form of large poikiloblastic porphyroblasts, in the Upper Group these porphyroblasts are absent and the felspar is in the oligoclase-andesine range.   These changes will be discussed in greater detail later.   The other minerals, the epidotes, amphiboles and micas

are common to both groups. Each mineral phase will now be considered in detail, as far as possible in their temporal order of occurrence.

Biotite

It is apparent that biotite crystallised right through this period of metamorphism. Rocks which show evidence of recrystallisation under these static conditions i.e. those with polygonal quartz-felspar mosaics generally contain biotite which tends to be decussate and largely unstrained.

It would seem that the presence of mica exercises considerable control over the development of these polygonal textures since in the mica-free quartzose horizons the quartz takes on a coarse lenticular form; the long axes of the lenticles being parallel to S.1.

Plate 134 shows a portion of a slide of an F.1 fold (Specimen I) described in the petrofabric section (P. 87.). This study revealed that the quartz fabric in the rock was essentially F.1 oriented though a number of cross-girdles in the diagrams could not be adequately explained. In general, in these slides the micas show a good parallelism with the fold axial plane though in places they tend to be rather decussate. This decussate fabric was almost certainly produced by post-F.1. annealing under conditions conducive to the growth of biotite.

Some of the amphibole-bearing schists show a close association between hornblendic amphibole and butite. Plate 135 shows such an occurrence. In this slide the long axes of the amphiboles are randomly arranged within the plane of the schistosity. The biotite shows no dimensional orientation and is presumbd to be an alteration

Plate 135. Decussate horneblende altering to biotite.
Lower Komagnes Group, Komagnes. x 40, P.P.L.
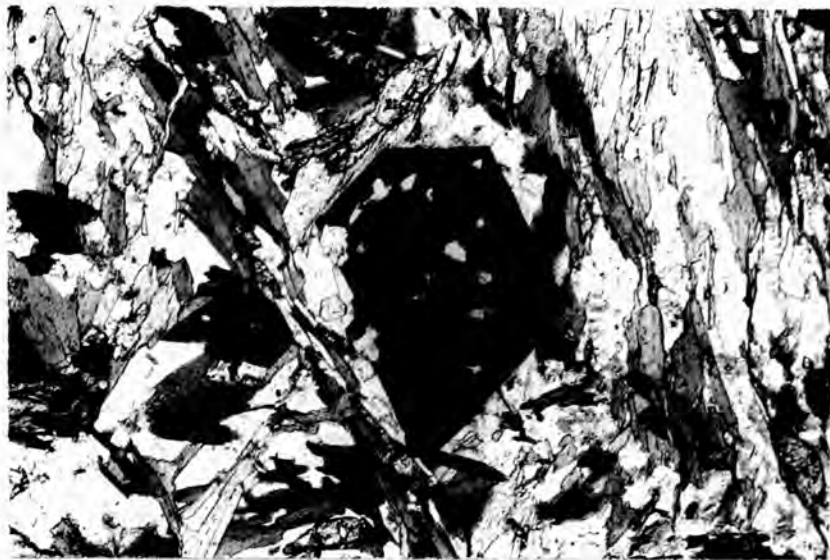
Plate 136. Euhedral Iron Ore in the Upper Komagnes Group.
x 40, P.P.L.

product of the amphibole. This type retrogressive change is probably related to the wane in metamorphic grade preceding F.2. This seems more likely than the thesis that the retrogression may be a post -F.2 phenomenon since where evidence is forthcoming concerning post-F.2 metamorphism, the characteristic assemblages are chlorite not biotite-bearing.

The pleochroism of the biotite in these rocks is generally confined to shades of red-brown or chocolate-brown. However, a slide from the Mixed Pelitic and Psammitic unit contains biotite in which Z is lincoln green. This presumably indicates a high ferric-iron to titanium ratio. (Deer, Howie and Zussman 1967, P.213). The abundance of iron in the rock is further suggested by the presence of large amounts of ore including large (1mm) euhedral crystals (Plate 136).

Thus, there is evidence of the crystallisation of biotite during the static phase and the formation of the mineral as a breakdown product of amphibole, probably during the wane of metamorphic grade previous to F.2.

Muscovite

Muscovite is quite common in the metasediments; it frequently occurs in association with biotite, defining S.1 and in the same way as biotite, occasionally shows a decussate form evincing recrystallisation during the static period. The two minerals also occur as inclusions in late albite porphyroblasts. There are some interesting relationships between the two micas. Some crystals may be composed, one half of biotite and the other of muscovite.

The two micas also occur as lenses and bands in each other -
(Plate 137). This type of relationship is very difficult to
interpret as it is not always clear that there is a reaction
relationship between the two minerals. Deer, Howie and Zussman,
(Vol. 3 P.73) state that textural relations suggest that the in-
crease of biotite in the biotite zone is coincident with a
decrease of chlorite and especially muscovite. On this basis it
would be possible to explain the association as being due to the
progressive increase of biotite at the expense of muscovite
during the rise in metamorphic grade subsequent to F.1. A number
of lines of evidence contradict this view. Firstly the muscovite
in the rocks frequently abuts against garnet crystals (Plate 138)
which in itself suggests that the muscovite is pre-garnet.
However, although biotites have been recorded as inclusions in
garnet, muscovite has not. This strongly suggests that in
most cases the muscovite is post-garnet. The abutting relationship
can be explained by a process of post-garnet muscovitisation of
biotites which had themselves been overgrown by garnet. This
muscovitisation process is further suggested by the frequent
occurrence of blebs and streaks of ore aligned along muscovite
cleavages, (Plate 139). Ore is also common in the groundmass of
(R.1), a slide which shows particularly strong development of
muscovite and also in post-F.2 muscovite porphyroblasts, (Plate 140).
This suggests that the process of muscovitisation was quite
extended in time, beginning after the phase of garnet growth
and continuing through into post-F.2. times.

There is also, however, evidence that muscovite crystallised
as a primary phase. This is suggested by the obvious control of
whole-rock composition exerted over the formation of the mineral;

Plate 137. Intergrowth of biotite and muscovite. Lower
Komagnes Group, Vasbugt. x 40, P.P.L.
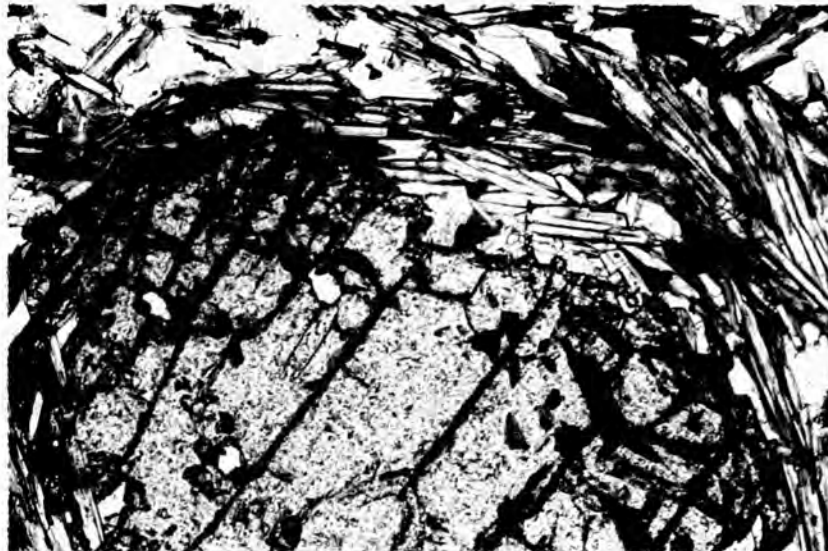
Plate 138. Muscovite abutting against garnet. Upper
Komagnes Group, Rastaby. x 40, P.P.L.
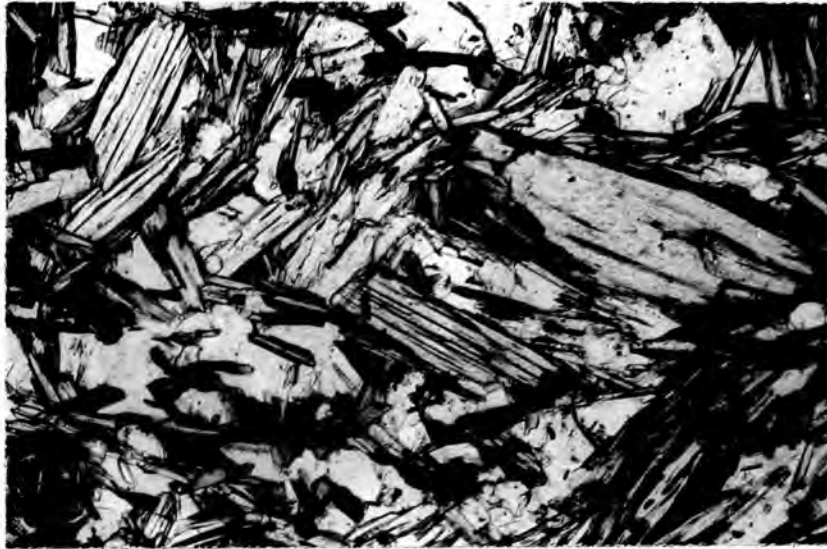
Plate 139. Muscovite with blebs and streaks of ore along
         the cleavages. Upper Komagnes Group, Rastaby.
         x 40, P.P.L.
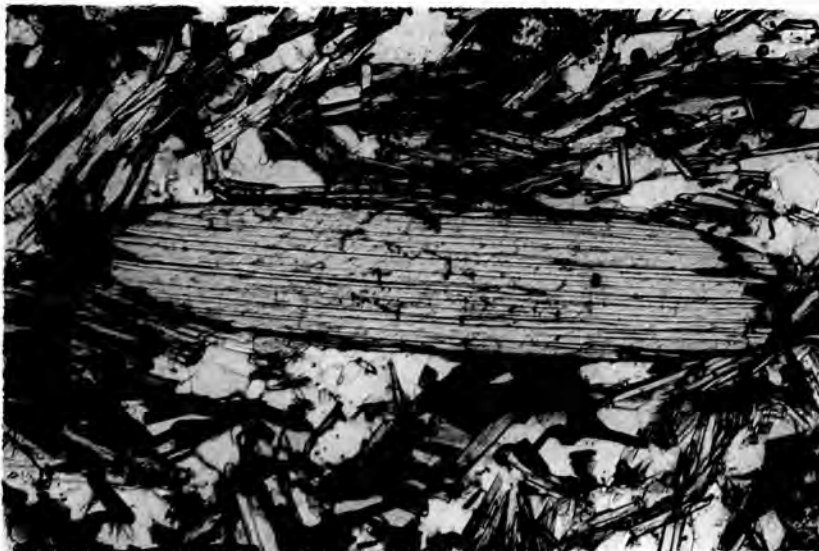
Plate 140. Large muscovite porphyroblast with blebs of ore
         along cleavage. Upper Komagnes Group, Rastaby.
         x 40, P.P.L.

it occurs most abundantly in the purer psammitic and semi-pelitic rocks.
It is here that the confusing intergrowths occur most commonly.   The
calc-silicate schists on the other hand, contain an abundance of horn-
blende, epidote and biotite and very little muscovite.   The mineral has
been recorded in association with hornblende in only two thin sections;
in one it is sericitic and obviously retrogressive relative to biotite,
and in the other it has the similar intergrowth relationship with
muscovite as described previously.   The obvious conclusion to this is
that muscovite develops as a primary phase in rocks in which there is
a relative dearth of iron, i.e. the psammites.   Thus many of the
observed biotite-muscovite intergrowths may, in fact, be equilibrium
assemblages caused by shortage of iron.

Thus, a picture emerges of growth of muscovite, probably throughout
the static phase, in the psammites and semi-pelites alongside biotite.
At some stage after the acme of regional metamorphism there began a
process of muscovisation of the biotite.   This was a retrogressive
effect undoubtely due to the greater stability range of muscovite.
It is, of course, well known that muscovite is stable in the lowest sub-
facies of the Green Schist Facies, whereas biotite is not.

Amphibole

Hornblende is a very common constituent of the calc-silicate and
basic schists especially within the Lower Komagnes Group.   The mineral
is easily distinguished from the actinolitic amphibole by its optics
and different mode of occurrence.   The actinolite always defines an
F.1 lineation and has an acicular habit, whereas the hornblende is
characterised by a rosette-like appearance on schistosity surfaces and
a stumpier habit.   In some instances the individual hornblendes are up
to 1 cm. in length.

In the actinolitic rocks there is some evidence of post-F.1
amphibole overgrowing the lineated actinolite (Plate 141). These later
amphibole crystals have, however, similar optical properties to the
actinolite. This lends further credence to the idea that the
actinolite did not invert to the more aluminous hornblende with rising
metamorphic grade, due to a paucity of alumina in the rock. The rather
random rosette-like nature of the later hornblende prisms is an indication
that they grow under static conditions; growth is, however, generally
confined to the plane of the schistosity S.1.

In thin section the hornblende is rather variable, both in its optical
properties and grain size. The mineral is frequently poikiloblastic,
relative to quartz, biotite and epidote, (Plate 142), It is often
difficult, however, to establish any clear temporal relationship between
these inclusions and the host hornblende.

There is some evidence that albitisation of the metasediments has
slightly modified the chemical composition of the hornblende.
In the strongly albitised rocks the hornblendes have the pleochroic scheme:-

$$Z = \text{Deep turquoise blue}$$

$$Y = \text{Deep green}$$

$$X = \text{Pale yellow}$$

$$Z > Y > X$$

In sections which show scant or no evidence of albitisation the
hornblende lacks the turquoise colour for the Z vibration, e.g.:

$$Z = \text{Pale Green}$$

$$X = \text{Pale yellow}$$

$$Z > X$$
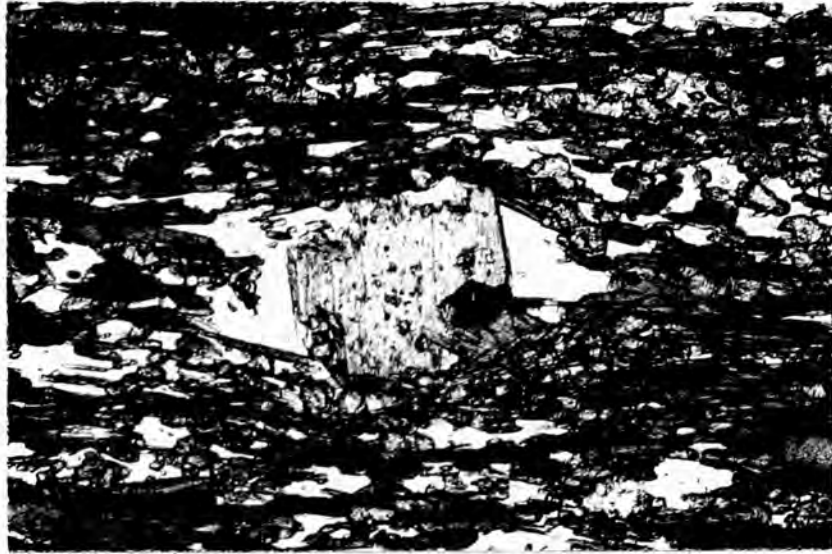
Plate 141. Horneblende porphyroblast overgrowing syn-F$_1$
    actinolite. Lower Komagnes Group, Vasbugt.
    x 40, P.P.L.

Plate 142. Laths of horneblende with epidote inclusions.
    Upper Komagnes Group, Rastaby. x 40, P.P.L.

was recorded from a zoisite-hornblende schist in which albitisation effects were absent. This contrast is probably due to the increased content of sodium in the amphiboles of albitised rocks. Where this albitisation is particularly intense it tends to destroy the (S.1) schistosity and in response to F.2 flattening, the amphiboles become decussate. F.2 movements were also responsible for some straining and augening of amphibole. Plate 143 shows a hornblende with strain extinction and bowing of a contiguous amphibole.

A list of the optical properties of some typical amphiboles is appended in Table. 7.

Table 7

Optical Properties of Hornblendes

1. Specimen M20 - 2A

Albitised hornblende-biotite-epidote schist

Opt   - ve   2V approx. = 60° - 70°

$\widehat{Z}$ C = 16° Y = b

Birefringence   0.017

Pleochroic scheme Z = Turquoise Green

Y = Deep green

X = Pale yellow   Z > Y > X

2. Specimen N17 - 47

Strongly albitised hornblende-biotite-epidote semi-pelite.

Opt   - ve   2V approx. = 70°

$\widehat{Z}$ C 15° Y = b

Birefringence   0.019

Pleochroic Scheme Z = Deep turquoise blue
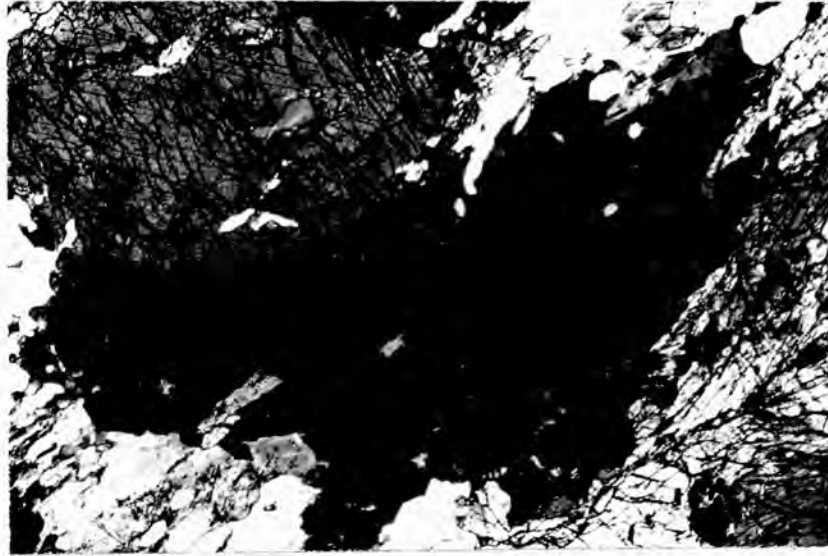
Plate 143. Horneblende showing strain due to F_2 movement.
Upper Komagnes Group, west of Komagnes.   x 40, P.P.L.
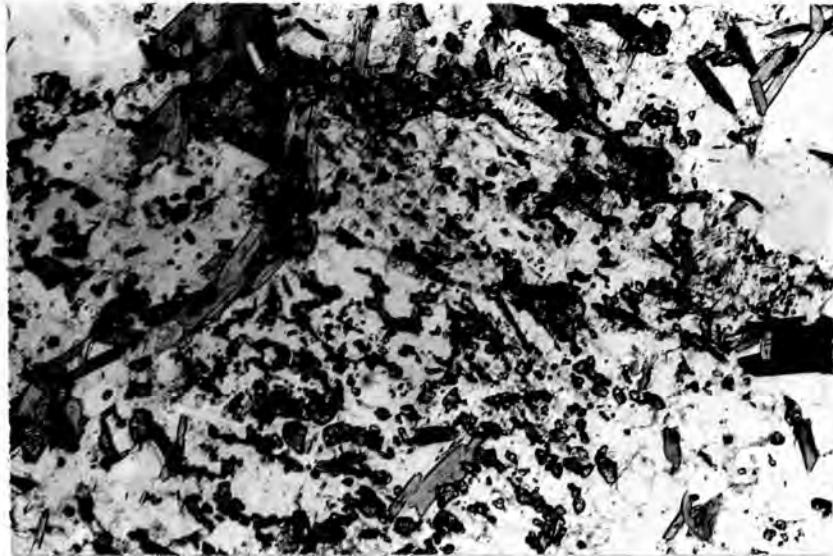
Plate 144. Granular epidote, Lower Komagnes Group,
Vasbugt.   x 40, P.P.L.

Y = Deep green

X = Pale yellow   Z > Y > X

3.   Specimen M20 - 50B

Unalbitised zoisite-hornblende-biotite schist.

Opt   - ve   ZV approx. = 60°

$\widehat{Z}$ C = 18°   Y = b

Birefringence   0.030

Pleochroic Scheme   Z = Pale green

Y = V. Pale yellow

X = Colourless   Z > Y > X

4.   Specimen N17 - 20

Unalbitised hornblende biotite epidote schist

Opt   - ve   ZV approx. = 80°

$\widehat{Z}$ C = 14°   Y = b

Birefringence   0.029

Pleochroic Scheme   Z = Pale Green

Y = Pale Green

X = Colourless   Z > Y > X

Epidote

Minerals of the epidote type are ubiquitous in the metasediments of the Komagnes Group.   The commonest, however, is epidote itself, which occurs in a variety of forms and in widely varying amounts. The mineral frequently has a greenish-yellow colour is slightly pleochroic, optically negative and has a high birefringence (0.045 has

been recorded).

The most frequent mode of occurrence is as a granular aggregate in the groundmass of the rocks (Plate 144). Occasionally, however, it occurs as porphyroblasts which clearly overgrow earlier-formed biotites and amphiboles. Plate 133 shows such a porphyroblast overgrowing F.1 actinolites. The crystal has been strongly augened by F.2 movements and is, in consequence, rather strained and granulated. Further evidence for this pre-F.2 epidote growth is provided by the existence of prisms of the mineral that have been strained on the limbs of F.2 microfolds.

In some sections normal epidote can be seen to be rimming a brown isotropic substance which sometimes has a euhedral outline. Occasionally within these compound grains there are particles of a brownish birefringent mineral which can sometimes be seen to be rimming the pseudomorphs. Optical study of this mineral yields the following pleochroic scheme:

$$Z = \text{Chocolate brown}$$
$$X = \text{Pale yellow} \quad Z > X$$

There seems little doubt that this brown mineral is allanite which in places has broken down by radioactive decay of included rare-earth elements to yield the metamict pseudomorphs (Deer Howie and Zussman 1967, P.68).

Clinozoisite also formed during this static phase and its formation appears to have been controlled by the original composition of the sediment. This is strongly suggested by the fact that the mineral has only been recorded in slides which contain no iron ore. Presumably

with a high iron content ferriferous epidote would have formed. This
paucity of iron is further suggested by the very pale shades of green
of the accompanying amphiboles (Winchill 1951, Pt II, p.436).

Textural evidence indicates that the clinozoisite post-dates the
amphibole. Most of the available iron in the rock is, therefore,
probably locked up in the amphibole.

Plate 145 shows poikiloblastic clinozoisite prisms overgrowing
earlier formed amphibole and biotite. The mineral has a low bire-
fringence (0.007), is optically positive, has a 2V of approximately
10° and most sections show straight extinction.

This mineral, which has grown with the long axes of the prisms
lying in S.1 has been strongly strained and bowed by F.2 movements.

Garnet:

Garnet has a rather distinctive distribution in the Komagnes
Group. In the metasediments of the Lower division the mineral is quite
rare and has only been recorded in one thin section. In the Upper
division i.e. in the rocks structurally above the approximate level of the
Calc-Silicate Schist, it is relatively abundant.

Textural studies suggest that the mineral overgrows the S.1
schistosity defined by both biotite and muscovite and also epidote and
hornblende (see Plate 146) though this latter relationship is seen only
in one garnet from the Lower Komagnes Group (Plate 146), elsewhere it
is equivocal. This growth of garnet therefore, probably represents
the acme of regional metamorphism in the Komagnes Group.
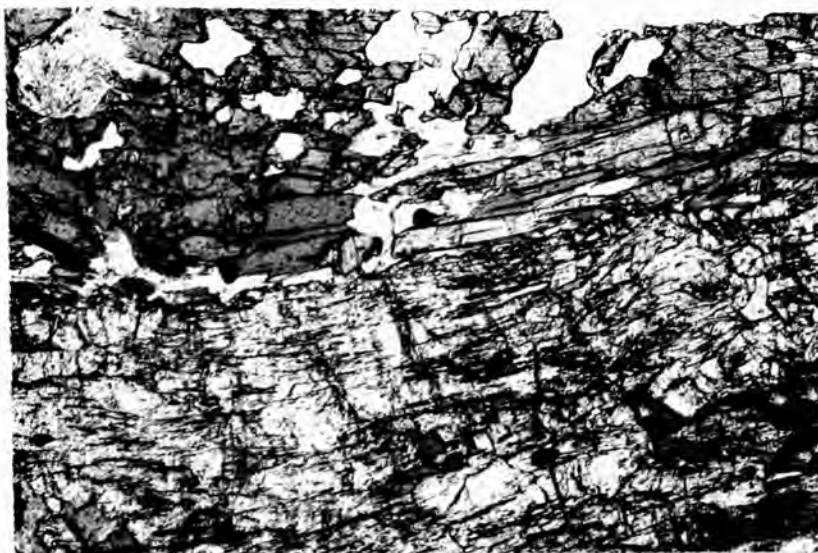
Plate 145. Clinozoisite overgrowing biotite and horneblende.
Upper Komagnes Group, west of Komagnes. x 40, P.P.L.

Plate 146. Garnet overgrowing epidote and horneblende.
Lower Komagnes Group, Komagnes. x 40, P.P.L.

The mineral is widely variable in form and size; a garnet porphyroblast 3 cms across has been recorded in a semi-pelite from Rastaby though this is exceptional. The most frequent occurrence is as skeletal crystals which have a very faint pink hue and are perfectly isotropic. Sometimes, however, the mineral occurs as good poikiloblastic euhedra. The inclusions are quartz, felspar and disoriented biotite flakes.

An interesting relationship is seen in Plate 147. Here a garnet porphyroblast occurs in the core of a small F.2 fold. It is clear that the garnet has behaved as a rigid kernel during the deformation, such that the quartzo-felspathic band which defines the fold has been flattened round the garnet giving a rather bulbous appearance to the structure. All the garnets are strongly augened by F.2 movements and there is no evidence of growth during F.2.

Determination of the 'a' cell edge of a garnet from a semi-pelite at Rastaby, using X-ray diffraction methods and the refractive index, yields a composition of Alm 18, Pyr 7, Spess 75, using the tables of Sriramadas (1957).

$$RI = 1.80$$
$$'a' = 11.59$$

Sturt (1962) has observed that with increasing grade of regional metamorphism the amounts of Ca and Mn in garnets decreases with an accompanying increase in the amounts of the smaller ions Mg and Fe. Atherton (1965) although agreeing with the general trend, points out that there may be a strong whole rock compositional control in the chemistry of garnet. For example, if a rock is very rich in Mn,
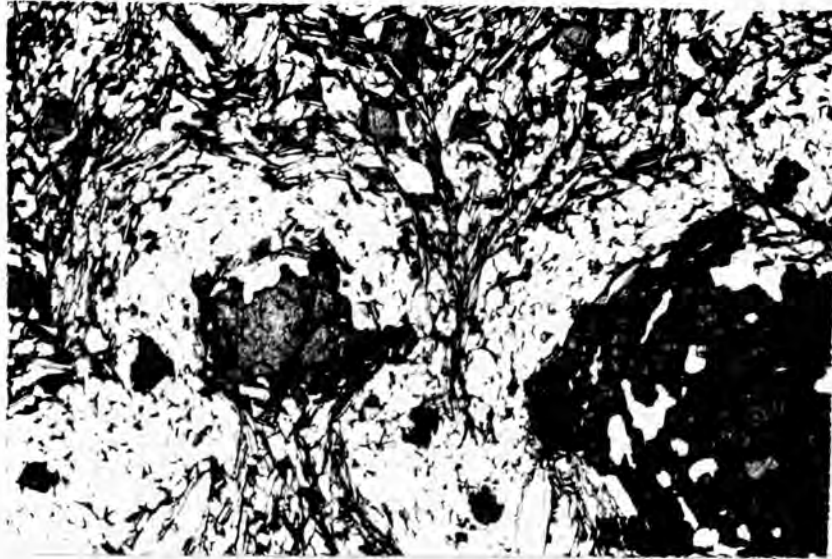
Plate 147. Garnet in the core of an F_2 buckle. Upper
Komagnes Group, Rastaby. x 10, P.P.L.
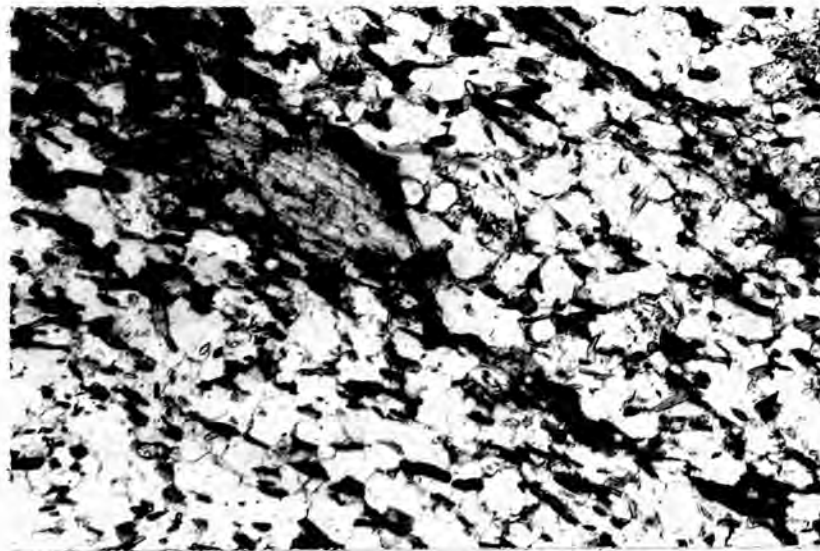
Plate 148. Calcite segregation, Upper Komagnes Group,
Rastaby. x 40, P.P.L.

garnet may be the only phase capable of taking this up, thus even at relatively high grade a garnet may be rich in Mno.   A whole rock analysis of the rock in question has not been obtained.   However, further X-ray work on garnets from the Eidvågeid schist group and the Olderbugten Group reveals them to be rich in the Almandine molecule, i.e. Fe rich.   These observations, which in the light of Atherton's work must be regarded tentatively, are certainly consistent with the observed change in grade that occurs across the area.

Calcite

Calcite is a common accessory constituent of the rocks.   Generally it occurs as granular anhedra showing the typical extreme birefringence together with a negative uniaxial sign.   Its relationship to other mineral phases is often difficult to determine.   It seems certain that most of it is associated with a post-F.2 hyrothermal metamorphism. However, some calcite has been recorded as inclusions in pre-F.2 garnet. It seems likely that it is, therefore, a primary constituent of the rock. Plate 148 shows a calcite porphyroblast overgrowing a biotite segregation aligned along S.1.   This type of calcite growth is easily distinguished from the growth occuring in the post-F.2 veins and cracks where it is accompanied by chlorite and epidote.

Quartz and Felspar

A considerable amount has already been written on the orientation and habit of the quartz of the Komagnes Group in the petrofabric section. In summary, from this study the following emerged:

1)   In psammites quartz generally has two morphologies.

a)   A fairly coarsely crystallise variety which occurs as
     stringers in the rocks.   Within these stringers which
     generally define S.1, the quartz show little or no
     dimensional orientation but is frequently strained.
     These layers are mica-free and give rise on banding
     surfaces to a strong F.1 lineation.   (Plate 149).

b)   A rather finer-grained equidimensional fraction which is
     often a mixture of quartz and felspar.   These areas tend
     to show a rather polygonal outline obviously due to post-F.1
     recrystallisation.   It is this recrystallisation that has
     tended to disrupt the S.1 fabric defined by mica, which
     occurs commonly in these areas.   The difference in fabric
     between (1a) and (1B) is obviously a function of the presence
     or absence of felspathic constituents and more important, of
     micaceous constituents.   The reasons for this will be discussed
     more fully later.

2)   The optical orientations of the two fractions (1a) and (1b) were
not significantly different in any one slide.

3)   In the Komagnes area the quartz fabric appeared to be more
controlled by F.2 whereas in the Rastaby the fabrics were dominantly
pre-F.2.

Felspar

     It is the felspathic constituents of these rocks that constitute
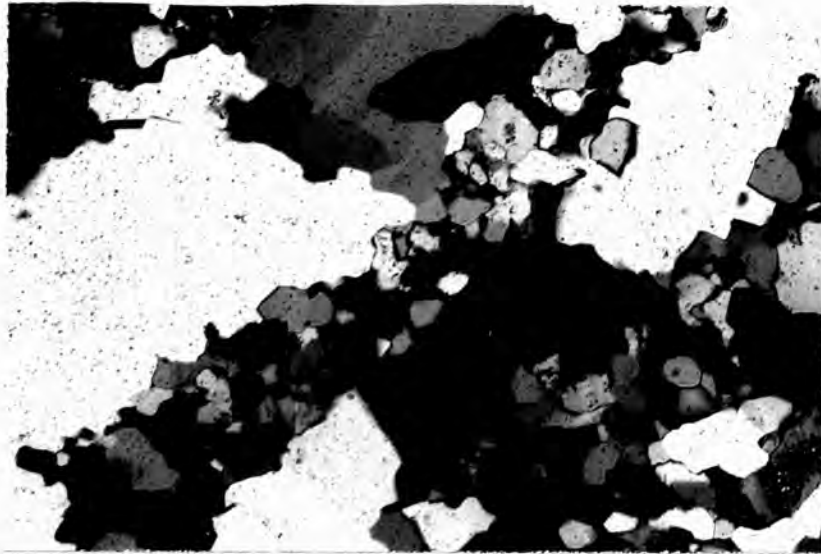
Plate 149. Stringers of quartz, Lower Komagnes Group,
Komagnes. x 40, X Pls.
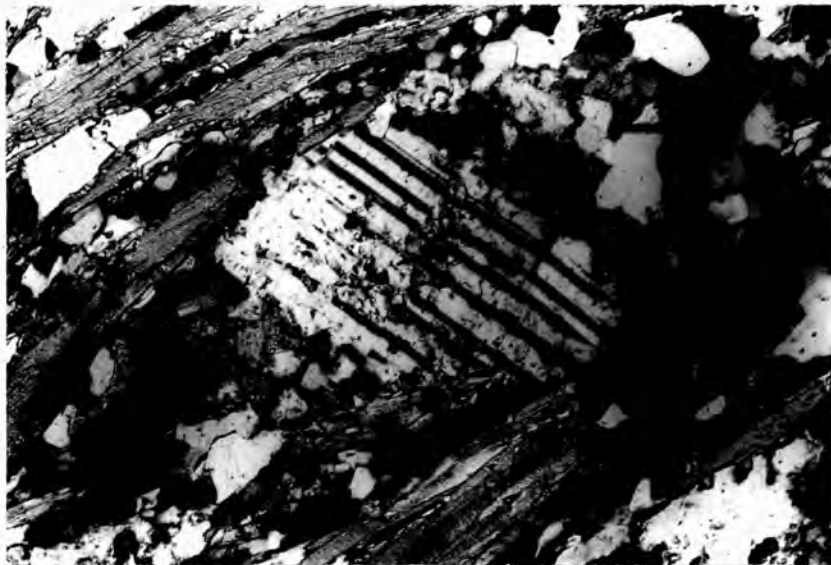
Plate 150. Poikiloblastic albite porphyroblast, Lower
Komagnes Group, Komagnes. x 40, X Pls.

the most important evidence of a change in metamorphic grade.

The Lower Komagnes Group

This consists of the structurally lowest metasediments up to
the approximate level of the Mixed Psammite and Pelite unit. This
is characterised by the abundant occurrence of large porphyroblasts
of albite. (Plate 150). Measurement of albite twins yielded a
maximum extinction angle of 15°, this coupled with the positive sign
and refractive index slightly less than balsam gives a composition of
(An_7). This albitic composition is confirmed by X-ray diffraction
study; a whole rock crush of an albitised psammite yielded good-
albite peaks. This albite is demonstrably of metasomatic origin.
Felspar porphyoblastesis is a common feature in the area and it will
be demonstrated later that there is a systematic change in the
composition of these porphyroblasts with metamorphic grade. In the
other metasedimentary groups it has often been possible to make a
distinction between the composition of the groundmass felspar and the
porphyroblasts. In the case of the Lower Komagnes Group this has not
been possible, albite appears to be the only felspar phase present.
There may be two reasons for this:-

1) Any earlier felspar, perhaps indicative of a higher metamorphic
grade, broke down during the albitisation, which took place at a
slightly lower grade.

2) The previous felspar phase was also albitic and merely recrystallised
during the albitisation.

Since there is no evidence of any retrogression of other mineral phases and garnet is notably sparse in the group, the alternative 2) is favoured i.e. that albite is the felspar representative of the maximum grade of metamorphism achieved within the Lower Komagnes Group.

The Upper Komagnes Group

In lithologies above and including the Pink Psammite albite is absent. The felspar is a plagioclase which occurs as a polygonal mosaic with quartz, (Plate 151). Lack of good twinning makes identification difficult, though refractive indices are always greater than balsam. The small number of twins that have been measured indicate a composition of (An 38), more basic schists however, have yielded (An 54).

Summary and Discussion

The rocks of the Lower Komagnes Group contain assemblages consisting of albite, hornblende, epidote, biotite, muscovite with sparse garnet. These minerals grew during the static phase of metamorphism between F.1 and F.2. Thus, the maximum grade of metamorphism achieved during this phase was the Quartz-Albite Epidote Almandine Sub-facies of the Green schist facies, Winkler (1967).
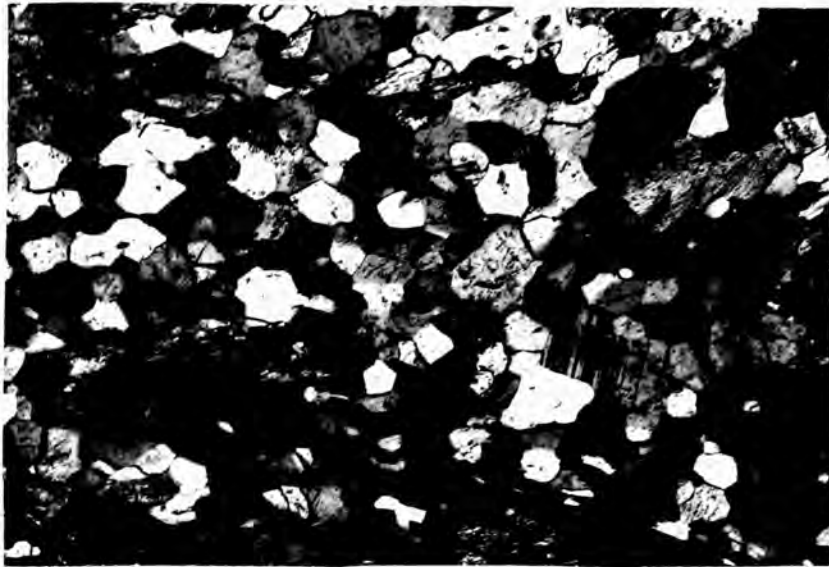
Plate 151. Polygonal quartz felspar mosaic. Upper
          Komagnes Group, Rastaby. x 40, X Pls.

The Upper Komagnes Group during the same phase, acquired a slightly higher grade of metamorphism characterised by plagioclase, horneblende, epidote, biotite, muscovite and abundant garnet. This indicates metamorphism in the Staurolite Almandine sub-facies, Winkler (op.cit.) The absence of staurolite is perhaps due to a compositional control.

Perhaps the most significant indicator of this change in grade is the change albite to plagioclase. De Waard (1959) records a very similar change in a metamorphic belt in Indonesia. He notes a sudden jump in felspar composition across a very well defined line. In the zone of lowest metamorphic grade the compositions are of the order (An0-10), on the other side they are in the An 20+) range. Very few rocks contain An contents between these values. De Waard (op.cit.) concludes that the line across which this jump occurs, represents the isograd between the greenschist facies and the Almandine Amphibolite facies. The line across which this change occurs on Seiland is indicated in Fig. 49 (P.220).

b) The Eidvågeid Schist and Structurally Higher Groups

Introduction

During the static phase of metamorphism the highest grade conditions represented by a kyanite-sillimanite porphyroblastesis were achieved in the metasedimentary groups. The following account is a description of these minerals as far as possible in their temporal order of occurrence.

Biotite

As a general rule biotite defines a good schistosity which appears
to be overgrown by the later porphyroblasts. This schistosity is
generally axial-planar to early folds and sensibly parallel to the
lithologic banding. For this reason the structure has been termed
(S.1.). These are, however, indications that more than one phase of
deformation may have been involved in its formation. This will be
discussed in greater detail later.

In thin section the mineral appears in various shades of brown
and red-brown defining the schistosity which sweeps round the strongly-
augened porphyroblasts.

Garnet

It has been possible to recognise two different types of garnet
growth. These two growth phases are separated by an episode of
mylonitisation. For the purposes of this account the earliest phase
will be termed Garnet I and the later phase Garnet II.

Garnet I

This phase of garnet growth produced very large (several
millimetres) garnet porphyroblasts. These are completely ubiquitous
in the pelitic and semi-pelitic metasedimentary Groups. Generally
they tend to have a near-circular cross section, though occasional
euhedral examples have been recorded showing a rhombdodecahedral form.
These garnets frequently have a truncating relationship with the
schistosity defined by biotite (Plate 152). Inside the garnets,
however, there is a coarse inclusion fabric consisting of polygonal
quartz and descussate biotite (Plate 153). One example of a shimmered
kyanite crystal has also been recorded. These two observations; the
truncating relationship with the schistosity and the coarse inclusion
fabric present two rather conflicting pieces of evidence. The
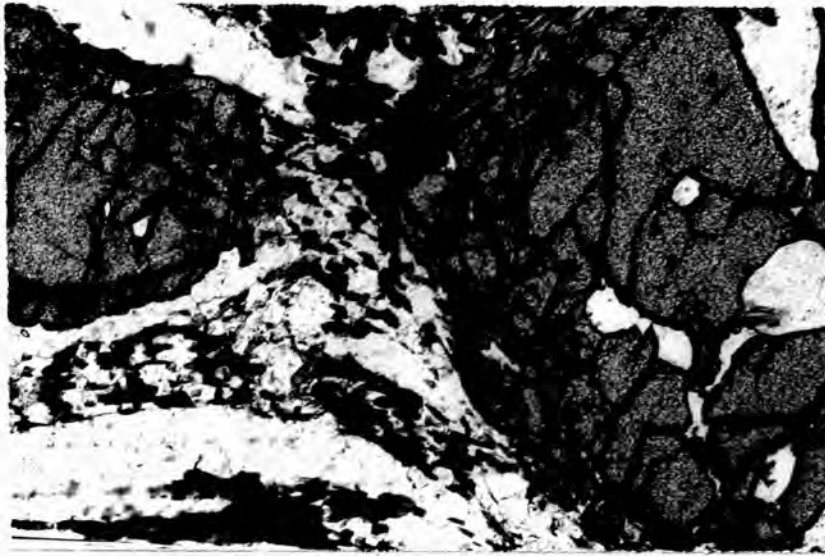
Plate 152. Garnet I porphyroblast truncating S.I.
Eidvågeid schist. x 40, P.P.L.
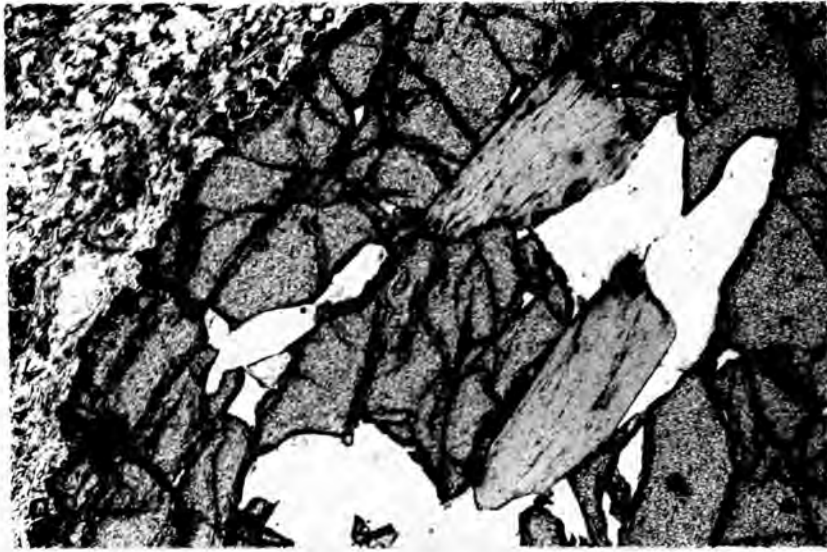
Plate 153. Garnet I porphyroblast with coarse inclusions of biotite, quartz and felspar. Eidvågeid Schist Group. x 40, P.P.L.

Plate 154. Garnet I showing atoll structure. Olderbugten Schists west of Olderbugten. x 10, X Pls.

truncating relationship suggests that the garnets have overgrown an earlier formed schistosity in which they have been augened by subsequent flattening, whereas the coarse inclusion fabric suggests that the garnets have overgrown an earlier annealed fabric and that the schistosity, previously denoted (S.1.), is in fact a post-garnet structure.

Consider Plates 154 and 155. These show large felsphar porphyroblasts with atoll rims of garnet, within these felspars occur large decussate biotites. These are similar in many ways to the atoll structure described by Rast (1964 p.94) from the Highlands of Scotland. These have been regarded, by many authors, as dissolution phenomena caused by growth of felspar and other constituents at the expense of the garnet. The dissolution of the centre portions of the garnet is due to an original zonation of the mineral. Rast states (op.cit.) "Zoned garnets which occur in the Central Highlands of Scotland and possess zones with different refractive indices, have grown during two different periods of metamorphism (M1 and M2). It is suggested that the atoll garnets represent the remnant M2 rims while their inner cavity, filled by felspar, mica and quartz, was originally the M1 garnet which is normally protected by the rim. If, however, the rim is mechanically breached, the unstable inner zone becomes replaced and the replacing felspars become nucleated at the interface of the two zones of contact"

The existence of this type of compositional zoning of garnet is siggested by a garnet from Eidvageid schist adjacent to a post-F.2 fault at Eidvageid. This shows a core that has been selectively chloritised by the retrogressive metamorphism accompanying faulting (Plate 173).
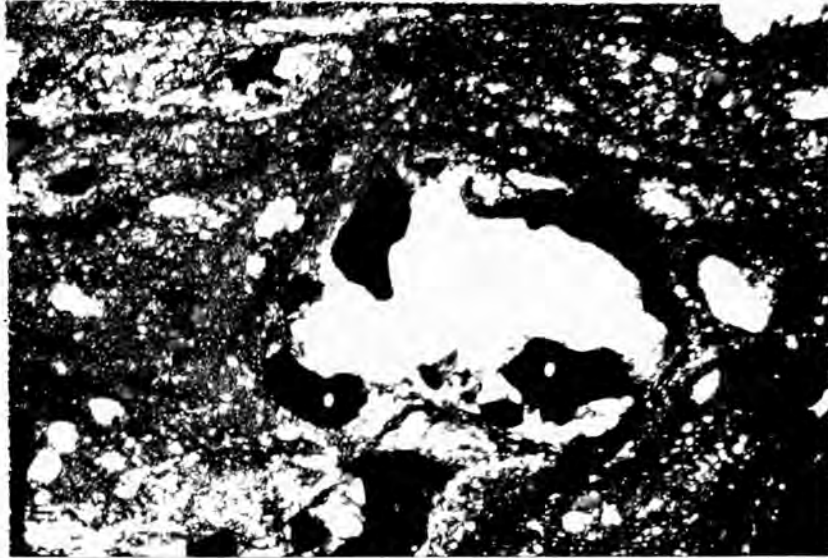
Plate 155. Garnet I with breached atoll structure.
Olderbugten Schists, west of Olderbugten. x 10, X Pls.
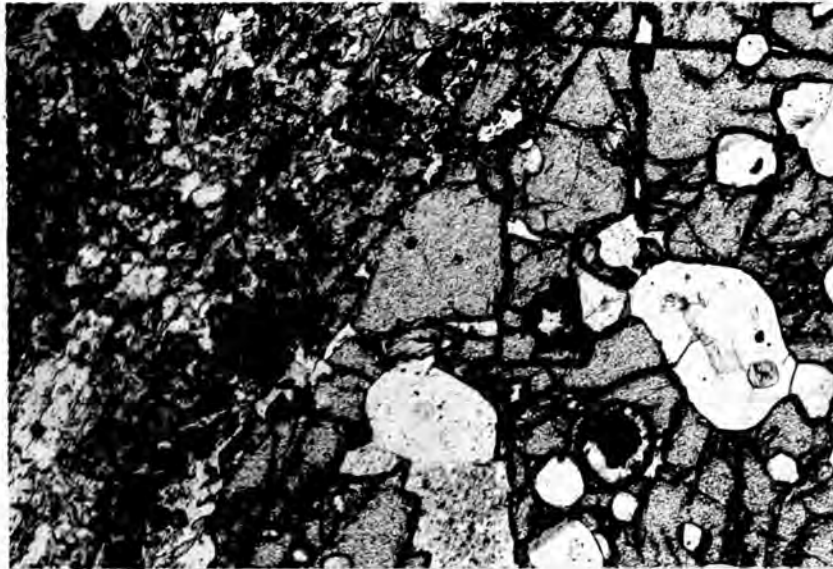
Plate 156. Overgrowth of garnet II on garnet I. Note the
discrete garnet II in the groundmass. Eidvågeid
Schist. x 40, P.P.L.

It seems likely, therefore, that the coarse polygonal fabric developed inside these garnets is the result of dissolution and subsequent recrystallisation. The atoll garnets represent an advanced stage of this process. This is further suggested by the fact that rims and discrete porphyroblasts of garnet II never show these dissolution phenomena. It can be demonstrated that these grew after the maximum grade of metamorphism and the mylonitisation. It would seem, from the relationship displayed in Plate 154 and 155 that this dissolution took place at about the same time as the felspar porphyroblastesis. It can be shown that this was coeval with the maximum grade of metamorphism.

D. Powell (personal communication) working in the Moines of N.W. Scotland, has noted garnets in pelitic rocks overgrowing an external foliation. The internal fabric of these garnets is obviously continuous with the external fabric. When, however, these pelites are traced into a zone of migmatisation, the internal fabric of the garnet recr;stallises in a manner very similar to that seen in Seiland. This means that the idea that the internal fabric of garnet represents a frozen relic of a previous phase of metamorphism must be applied with care, particularly where the inclusion fabric is coarse-grained as on Seiland.

Garnet II

The second growth of garnet post-dates the mylonitisation and main kyanite growth. It has two modes of occurrence:

1. As overgrowth on Garnet I, Plate 156.
2. As discrete grains within the groundmass.

The mineral clearly overgrows the schistosity and the mylonitic fabric, and is not so strongly augened as Garnet I. Generally, it is Euhedral and contains a large number of inclusions of the groundmass biotite; it can also be seen to be lobed round kyanite crystals. Occasionally dense granular concentrations of sub-hedral grains occur (Plate 157), these overgrowing an annealed mylonitic fabric.

Generally the selvage of garnet II on garnet I is rather thin, though this discrete garnet II porphyroblasts may be anything up to 1/2 mm. in size. The very characteristic inclusion fabric of both types makes them easily distinguishable. This difference in inclusion fabric indicates that the two types probably do not represent a continuous phase of garnet growth, rather they represent two phases of growth, separated by the mylonitisation during which growth was inhibited. This is further suggested by the fact that during the maximum grade of metamorphism marked by the felspar porphyroblastesis garnet I appears to have been undergoing dissolution.

In no case have curved inclusion traces been observed in examples of garnet cut normal to the external schistosity, this suggests, at least in the case of garnet II that growth took place during the static phase, or a phase of irrotational strain. One section, however, cut parallel to (S.1.) reveals curved inclusion trails in granular garnet II, (Plate 158). It seems highly unlikely that rotation could have taken place about an axis normal to the schistosity. It is probable, therefore, that these spiral inclusions are a reflection of the manner in which the garnet has grown. Kingery (1960) mentions such spiral growth from dislocations in crystal structures.
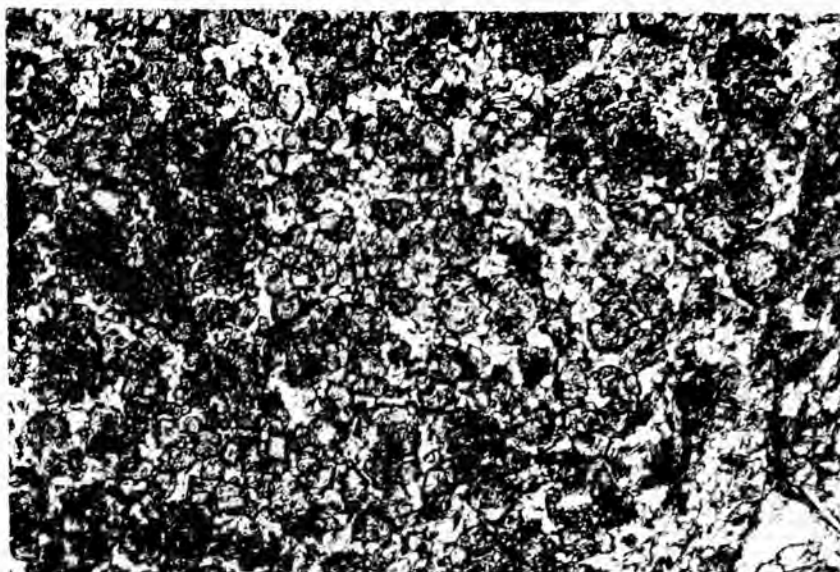
Plate 157. Concentration of subhedral garnet II, Eidvågeid Schist. x 40, P.P.L.
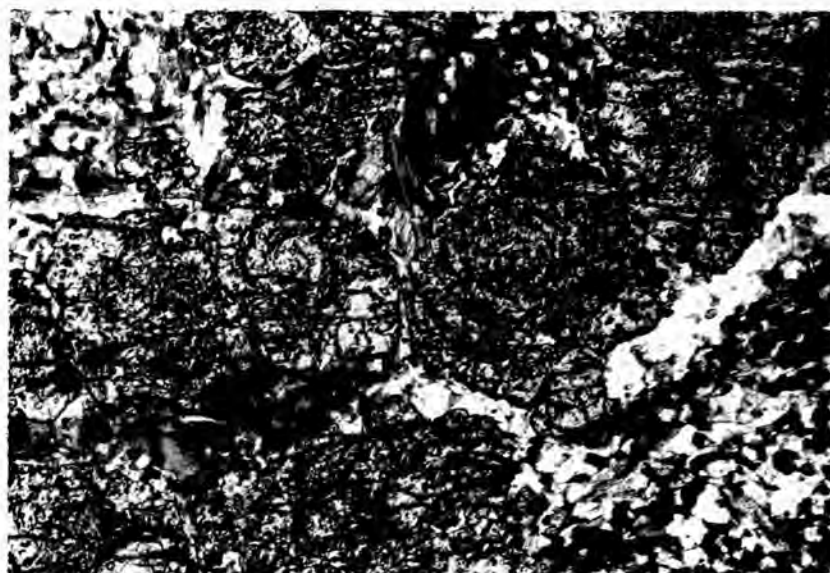
Plate 158. Spiral inclusions in garnet II, Eidvågeid Schist. x 40, P.P.L.

Determination of the "a" cell edge using X-ray diffraction methods and the refractive index of a garnet I from the Olderfjord Group yielded the following composition using the tables of Sriramadas (1957). Alm 77, Pyr 10 Spess 13.

$$RI = 1.815$$
$$"a" = 11.53$$

It will be remembered that a garnet from the Upper Komagnes Group was very rich in the spessartine molecule. Although both of these garnets almost certainly grew before the acme of regional metamorphism, the observed increase in the Almandine molecule is consistent with a rise in metamorphic grade across the area, Sturt (1962).

The Significance of S.1.

The following observations have been made above concerning the relationships between the various phases of garnet growth and the schistosity.

1. Garnet I frequently truncates the schistosity, though it is very strongly augened.

2. Garnet II clearly overgrows the schistosity and it is not so strongly augened.

From these observations it was suggested that garnet I was overgrowing the F.1. axial-planar schistosity (S.1.). Certain evidence, however, suggests that the situation is more complex than this. In rocks which have been strongly mylonitised, thorough recrystallisation produces

an annealed fabric with a schistosity defined by biotite. This schistosity must represent the result of post-mylonitisation possibly mimentic recrystallisation; thus although it is parallel to the anial plane of F.1. folds it cannot represent (S.1.). This situation arises because the plane of intense flattening, which developed during the mylonitisation was co-plar with (S.1.). Thus, the biotite foliation represents the product of two deformations, F.1. and the mylonitisation. (S.1.) itself is probably represented in the strain-shadow areas around garnet I where the truncating relationship is preserved.

Kyanite and Sillimanite

These polymorphs are widespread though they have a fairly distinctive distribution. Porphyroblastic kyanite occurs commonly in the Eidvågeid Schist, occasionally with fibrolitic sillimanite. In the Olderfjord and Olderbugten Group however, both porphyroblastic, kyanite and sillimanite occur together with fibrolite.

In the Eidvågeid Schist the kyanite occurs as megascopic prismatic crystal (examples up to 6 cms. have been recorded). The long axes of these crystals tend to lie within the foliation of the schist. This suggests strongly that they have grownmimetically in the schistosity. The long axes of these crystals were measured in the field and the result is given in Fig. 47. It can be seen that the locus of their distribution is a great circle corresponding to the foliation in the schist (see Fig.3).

FIG 47

KYANITE LONG AXES FROM THE EIDVÅGEID SCHIST GROUP
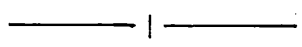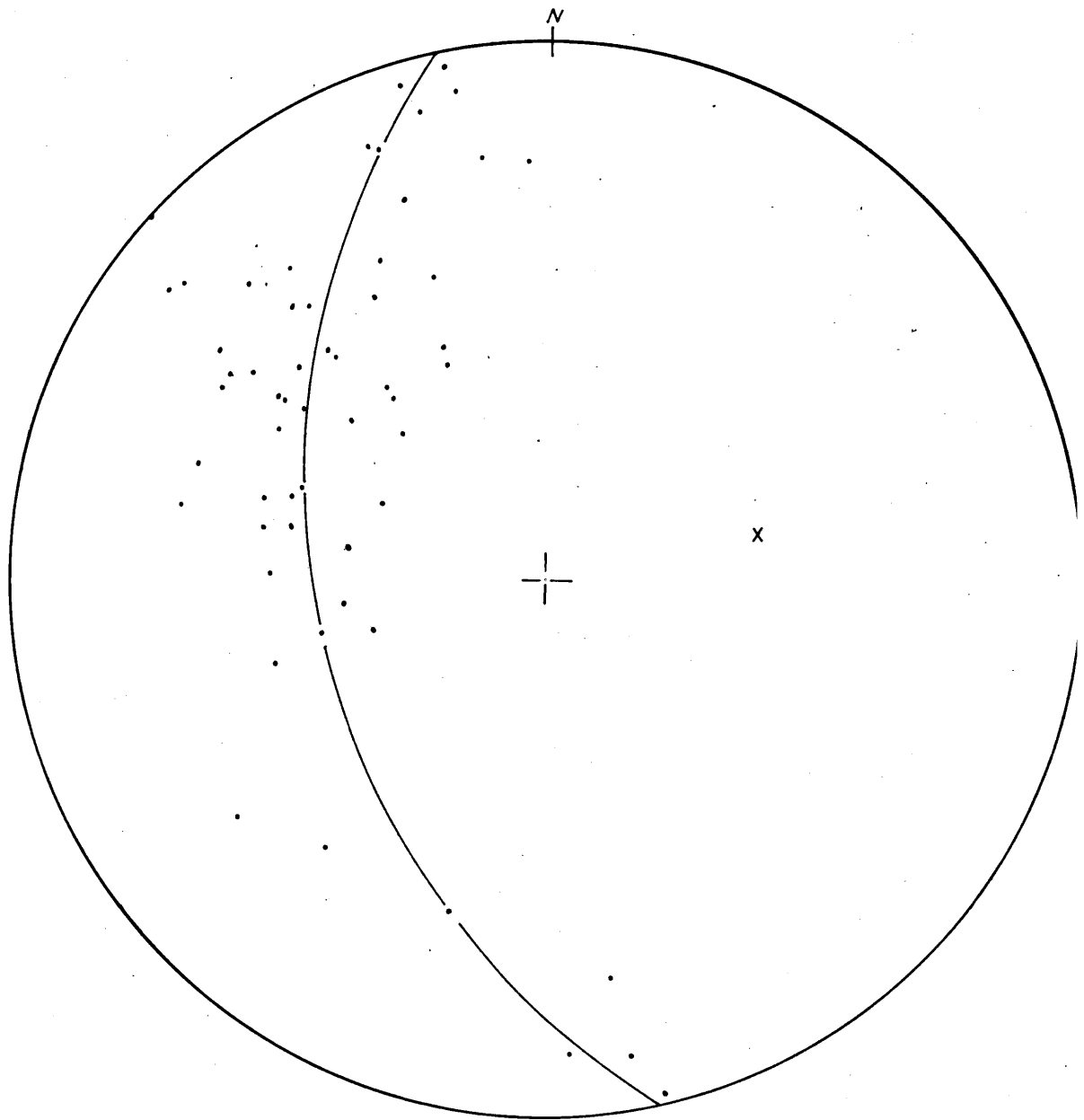
In this section these large kyanites can be seen to be strongly
shimmered by muscovite (see Plate 159). Occasionally this pseudomorphing
is complete, resulting in a lens of white mica. Where solid kyanite
cores exist, however, they can be seen to be strongly strained and
augened in the foliation. Often the pseudomorphs have a lenticular form.
Frequently, the partially pseudomorphed areas are prismatic in form
and there seems no doubt that they were once single crystals, the effect
of flattening has been to split the kyanites into a number of optically
discontinuous sub-grains (Plate 160).

Kyanite crystals are frequently overgrown by Garnet II. This may
either contain the kyanite as an inclusion or it may be lobed around the
kyanite (Plate 161). This gives a very good idea of the chronology of
kyanite growth. One example has been recorded of a kyanite inclusion in
a large alkali felspar porphyroblast. So it would appear that kyanite
growth was pre-felspathisation and pre-garnet II, also pre-mylonitisation,
which was undoubtedly responsible together with F.2. for the augening.

It has already been mentioned that both kyanite and sillimanite
occur within the Olderfjord and Olderbugten groups, if the distribution
of these two minerals is plotted on the map, no clear pattern emerges.
Although it may be possible that F.2. has folded the isograds giving
rise to an apparently irrational distribution, this cannot entirely
explain the relationship seen in thin section.

In a number of this sections both kyanite and sillimanite have been
recorded apparently in equilibrium. In other sections where kyanite alone
occurs there is often some evidence that it is a pseudomorph after
sillimanite. Plate 162 shows a characteristic sillimanite prism. In end
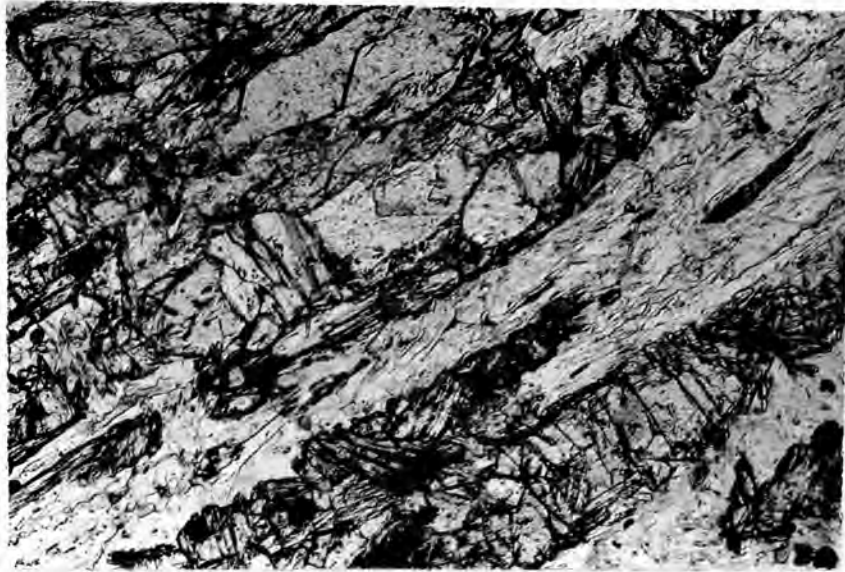
Plate 159. Shimmered kyanite, Eidvågeid Schist. x 40, P.P.L.

Plate 160. Highly strained and recrystallised kyanite,
Eidvågeid Schist. x 40, X Pls.

Plate 161. Kyanite overgrown by garnet II, Eidvågeid
Schist. x 40, X Pls.
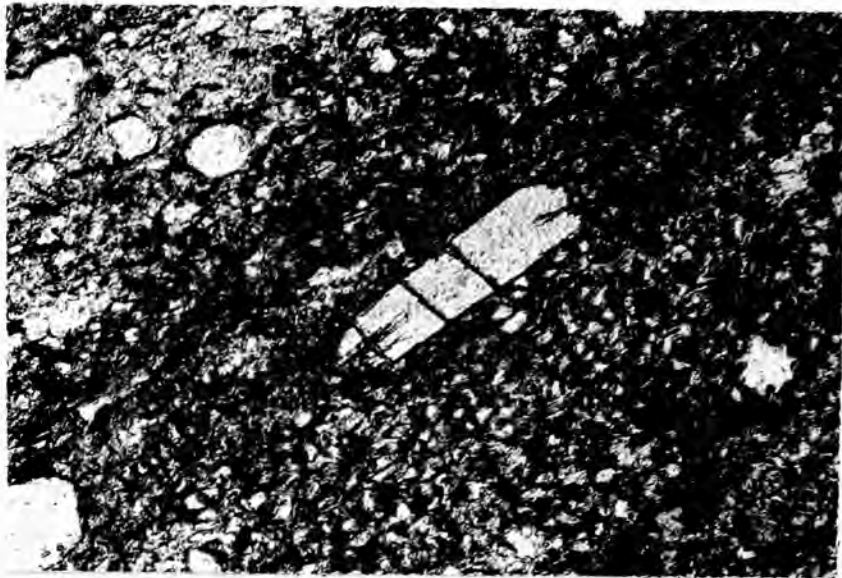
Plate 162. Sillimanite prism, Olderfjord Group, Hønseby.
x 40, P.P.L.

section, the mineral tends to be square with one prominent cleavage
(Plate 165). In rocks within the Olderfjord and Olderbugten Groups
where only porphyroblastic kyanite occurs the mineral tends to take on
an identical prismatic shape to the sillimanite. An important feature
is the fact that the cleavage planes within the kyanite appear to bear
a rather random relationship to the outline of these prisms
(Plates 163 and 164).

Table 8 shows some of the relationships that exist between the
cleavage planes of the kyanite and the bounding prisms. Where it was
possible to determine the optic sign this was always negative.

Table 8

Extinction angle to bounding crystal	Cleavage	Extinction	Orientation sign of ray nearest to crystal length	Interference colours
33	Abs	St to Cl	Slow	Grey
12	X	St to Cl	Slow	White
20	X	St to Cl	Slow	White
13	X	St to Cl	Slow	White-yellow
42	X	St to Cl	Slow	White
40	Abs	Obl to Pr	Fast	White
12	Cross fracture	Obl to Pr	Slow	Grey-white
45	X	St	Slow	Yellow
0	X	St	Slow	Yellow
12	X	Obl to Cl 31°	Slow	Grey
0	X	St	Slow	Grey
32	X	St to Cl	Slow	White
34	X	St to Cl	Slow	Yellow
9	Cross fracture	Obl to Pr	Slow	Grey

St to Cl	=	straight to cleavage
Obl to Pr	=	oblique to prismatic outline of crystal
St	=	straight
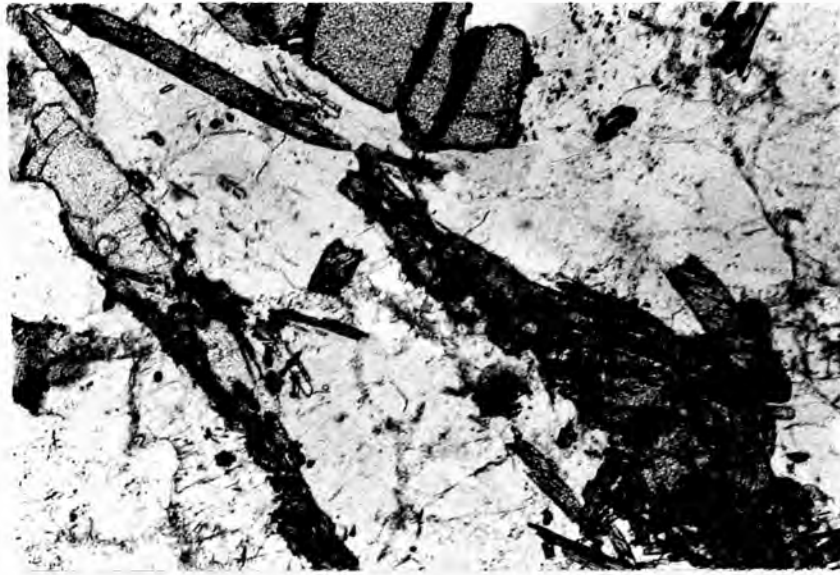
Plate 163. Kyanite pseudomorphs after sillimanite.
Olderfjord Group, Hønseby. x 40, P.P.L.

Plate 164. Kyanite pseudomorph after sillimanite. Note the
anomalous relationship between the cleavage and the
prism outlines. Olderfjord Group, Hønseby. x 40, P.P.L.

The strong similarity in shape between the bounding crystal and the sillimanite prisms, together with the apparently non-crystallographic relationship of the internal mineral to this prism shape, strongly suggests that the kyanite is a pseudomorph after sillimanite. Hietanan (1956) has recorded the transformation sillimanite to kyanite in the Boehls Butte Quadrangle. She attributes this to a regional metamorphic event following the thermal event which produced the sillimanite. On Seiland the occurrence of these pseudomorphs does not bear any close relationship to the outcrop of late F.1. igneous rocks such as the Hønseby gabbro which could possibly have been responsible for the growth of sillimanite. Similarly, there are good reasons for believing that not all of the sillimanite is due to contact metamorphism by the syn and post-F.2. plutons of Western Seiland. Firstly, many of the sillimanite porphyroblasts have similar age relationships to the kyanite of the Eidvageid schist, i.e. the crystals are augened and strained by the mylonitisation and flattening. Secondly, fibrolite can be seen to be nucleating on sillimanite porphroblasts and there is evidence that this fibrolite is overgrown by garnet II.

Thus, it would appear that both kyanite and sillimanite developed at the peak of regional metamorphism, occasionally the two polymorphs are found to stably co-exist. Sillimanite often occurs as the stable polymorph. In examples where kyanite occurs above there is however, evidence that it may have been formed by inversion of pre-existing sillimanite.

These relationships would seem to indicate that the temperature and pressure conditions operative during this phase of metamorphism were close to the kyanite-sillimanite phase boundary. The aluminium silicate triple-point was determined by Althaus (1967) as lying at $595 \pm 10^{\circ}$C at 6.5 ± 5 Kb. Richardson (et.al, 1969) in a similar

determination using slightly purer starting material, obtains a polygon

of experimental uncertainty, the centre of which has the co-ordinates

$622^{\circ}C$ and 5.5 Kb. The triple point, they say lies somewhere within this

polygon, which does not contain the Althaus determination. Althaus,

in a later paper (1969), suggests that the kyanite-sillimanite phase

boundary is at least bivariant. In a further experiment using both his

own starting materials and those of Richardsons (et.al.) in the same

bomb, he found after five days that in his own material the kyanite was

converted into sillimanite and in the Richardson sample, the sillimanite

was converted to kyanite. The chief difference between the two starting

materials was a high content of Fe_2O_3 in the Althaus sample. He

tentatively suggests, therefore, that impurities such as Fe_2O_3 may alter

the position of the phase boundary.

Since most natural systems are likely to contain large amounts of

impurity, it is felt that the Althaus determination probably most closely

replicates the natural system. This will, therefore, be accepted in the

following discussion.

The curve representing the onset of anatexis of a paragneiss with

a felspar composition in the oligoclase-andesine range, cuts the Althaus

equilibrum curve at approximately $670^{\circ}C$ and 9.15 Kb, (Winkler 1965, p.177).

Since there is evidence that partial melting of the gneisses took place

during the acme of regional metamorphism, the temperature and pressure

conditions prevailing in the Olderfjord and Olderbugten Groups must have

been something of this order.

Finally, brief mention may be made here of the occurrence of

fibrolite, this mineral occurs most commonly in the Olderfjord and

Olderbugten Groups, though it has been recorded in the top of the

Eidvågeid Schist Group. The mineral appears to nucleate at the grain boundaries of the other mineral phases, both on sillimanite, kyanite, garnet, quartz and felspar. Evidence will be presented later than the mineral appears to have developed wither synchronous with or slightly later than the mylonitisation.

Muscovite

During the steady decline in temperature that must have followed the acme of regional metamorphism there is evidence of progressive muscovitisation of the kyanite and biotite. This may partially have overlapped into F.2. The results of this process are manifested most intensely in the Eidvågeid Schist Group. In the structurally higher groups, the process does occur, but not to such an intense degree. The muscovite takes on a number of morphologies.

(1) As decussate flakes pseudomorphing kyanite (Plate 165). Harker (1939, P.349) regards this so-called shimmer aggregate as being due to late stage cooling when only very limited diffusion is possible. In this context it can be shown that most of the shimmer development took place pre-F.2. since the pseudomorphs are augened in the axial planes of F.2. folds.

(2a) As very large porphyroblasts overgrowing the schistosity but rather intensely strained (Plate 167). These frequently show ore segregations along their cleavage planes, which suggests that they have grown at the expense of the groundmass biotite. Their porphyroblastic form makes this very distinctive when compared to the shimmer aggregate.
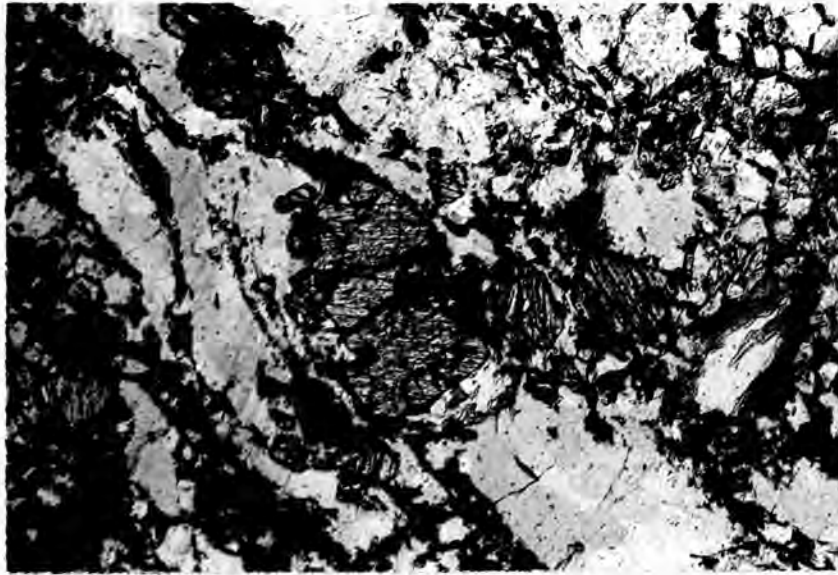
Plate 165. Basal section of sillimanite, Olderbugten Group,
west of Olderbugten. x 40, P.P.L.
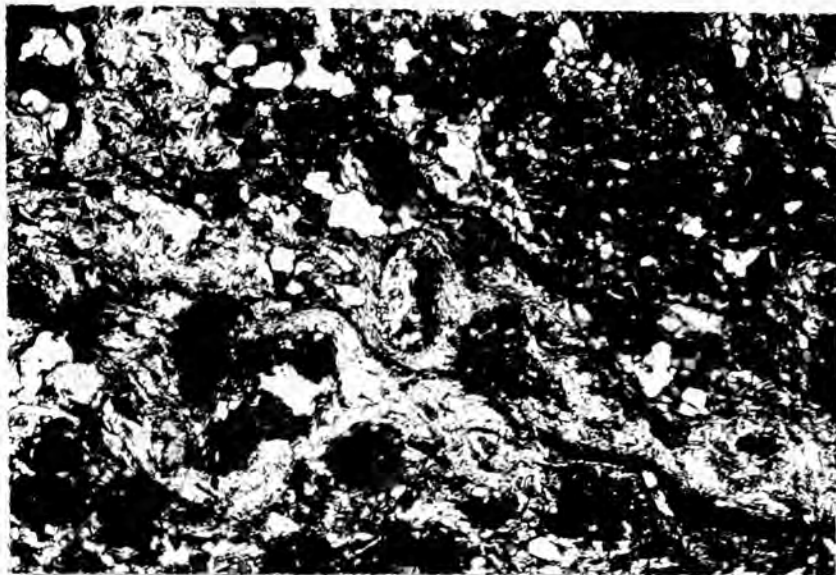
Plate 166. Decussate muscovite pseudomorphing kyanite.
Eidvågeid Schist. x 40, X Pls.
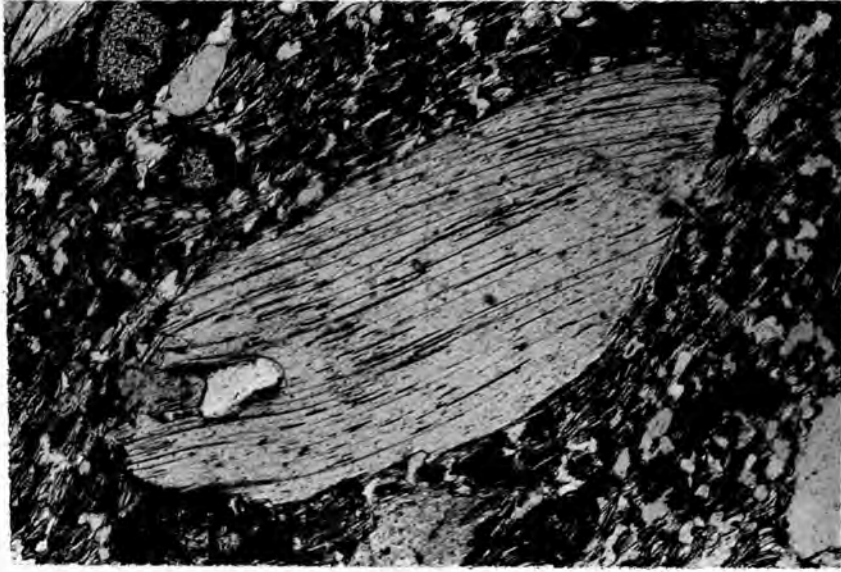
Plate 167. Deformed muscovite porphyroblast. Note the ore
grains aligned along cleavages. Eidvågeid Schist.
x 40, P.P.L.
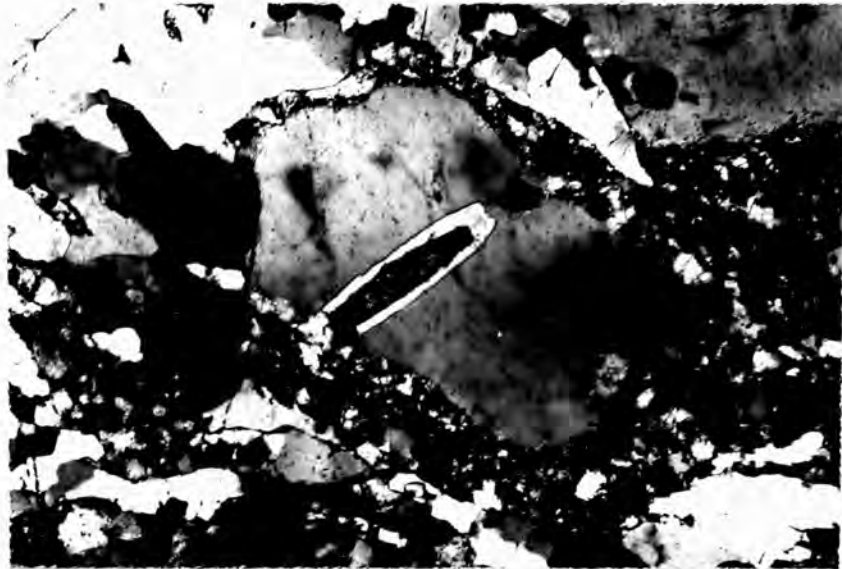
Plate 168. Sillimanite prism enclosed in felspar. Olderbugten
Group, west of Olderbugten. x 40, X Pls.

(2b) As small flakes in the groundmass formed by conversion of biotite. The relative lack of muscovitisation in the structurally higher groups is probably due to a dearth of water. Possibly the water responsible for the muscovitisation in the Eidvageid Schist and Komagnes Groups was driven out of the overlying Olderfjord and Olderbugten Groups.

The Calc-Silicate Bands

The mineralogy of these bands is very well shown by the specimen H2852. This consists of one such layer interfolded with a slightly more basic schist. The calc-silicate component consists of a very granular aggregate of diopsidic pyroxene which gives the following optical properties.

$$\text{Optically +ve} \quad 2V \text{ approx.} \quad = \quad 60^\circ \quad \overset{\wedge}{ZC} \quad = \quad 32^\circ$$

Birefringence   0.033

This occurs together with granular garnet and clinozoisite which frequently shows polysynthetic twinning. The felspar in this rock is very basis (An 80). There is also a little scapolite. The mineral relationships observed in this slide indicate the following sequence of crystallisation:

(1)   Pyroxene

(2)   Clinozoisite

(3)   Garnet

No determination of the garnet has been made but it is almost certainly a lime-rich variety.

The basic rock in the same slide contains very little garnet. The plagioclase, however, appears also to be (An 80). There is also diopsidic pyroxene present. The chief difference between this component and the calc-silicate band lies in the abundance of hornblendic amphibole, in the former. This gives the following optical properties :

$$\text{Opt} \quad \text{-ve} \quad 2V \text{ approx.} \quad = \quad 70^{\circ}$$

$$\overset{\wedge}{ZC} \quad = \quad 14^{\circ} \quad Y = b \quad \text{Biref} = 0.030$$

$$Z \quad = \quad \text{brownish green}$$

$$Y \quad = \quad \text{pale yellow}$$

$$X \quad = \quad \text{pale yellow} \qquad ZY = X$$

Quartz and some clinozoisite are also present.

The presence of a very basic felspar in these rocks indicates a compositional control in the formation of the mineral. The felspar of the pelitic rocks are always in the andesine range.

Quartz and Felspar

The felspar phases in these rocks have two characteristic modes of occurrence.

    (1) Groundmass felspar

    (2) Porphyroblastic felspar

(1) In the pelitic and semi-pelitic schists Michel Levy determination of plagioclase compositions yields an average of about (An 34). In the Eidvageid Schist and structurally higher groups, the observed textures are the result of the recrystallisation of a mylonitic fabric, the felspar occurs generally as a fine grained mosaic (see later for further details).

(2) The porphyroblastic felspar will also be discussed in greater detail later.

The quartz in the pelitic and semi-pelitic rocks occurs either as polygonised grains in the foundmass or as composite stringers. In the psammites the quartz occurs as irregular grains, often strained with a tendency for them to be elongated in the schistosity. These stringers probably represent original quartz-rich layers in the metasediments which became strongly flattened during the mylonitisation and subsequently recrystallised.

Summary

During the static phase the following assemblages arose in the EIDVÅGEID SCHIST GROUP :

Quartz, Andesine, Biotite, Almandine, Garnet (2 growth stages), Kyanite, Fibrolite, Muscovite.

The pelitic rocks of the OLDERFJORD AND OLDERBUGTEN GROUPS contain :

Quartz, Andesine, Biotite, Almandine, Garnet (2 growth stages), Kyanite, Sillimanite, Fibrolite and a little muscovite.

The Calc-Silicate rocks contain :

An-rich felspar, Epidote, Diopside, Garnet (probably lime-rich).

The Basic Schists :

An-rich felspar, Epidote, Diopside and Hornblende.

These facts indicate that the grade of metamorphism was slightly higher in the Olderfjord and Olderbugten Groups since porphyroblastic sillimanite is present as well as kyanite. There is evidence of inversion of sillimanite to kyanite suggesting that the temperature and

pressure conditions were close to the kyanite-sillimanite equilibrum curve.
Only fibrolitic sillimanite appears in the Eidvageid Schist, evidence will
be presented later that this developed later than the porphyroblastic
aluminosilicates; coeval with or slightly later than the mylonitisation.

The maximum grade of regional metamorphism was followed by a period
of slow cooling which led to muscovitisation of kyanite and biotite in
the Eidvageid Schist group. This process does not seem to have operated
extensively in the higher metasedimentary groups.

Thus, during the static phase the maximum grade achieved, appears
to have been the Kyanite Almandine Muscovite Sub-facies in the Eidvågeid
Schist Group and locally the Sillimanite Almandine-Orthoclase Sub-facies
in the Olderfjord and Olderbugten Group. The distribution of these facies
is given in Fig. 49 and the temporal order of growth in Fig. 62.

(3) The Syn. F.2. Metamorphism

Examination of the relationships of the various metamorphic minerals
to structures that arise during the F.2. folding, indicate that the grade
of metamorphism during this deformation phase had, over the whole area,
waned to about the same level.

Study of thin-sections slides from the late F.2. thrust at Komagnes
indicates that the rock had thoroughly recrystallised under conditions
favourable for the growth of epidote biotite, and muscovite, chlorite,
appears to be absent. Similarly, F.2. folds from within the Komagnes
Group show development of axial planar muscovite and biotite, garnet
becomes augened in this newly formed schistosity (S.2.) and in no case
has been observed to overgrow S.2. (see Plate 147). This F.2. foliation

is not penetrative except in some of the pelitic bands in tight F.2.
folds, even here vestiges of S.1. can be seen in the form of strain-slip
cleavages.

Examination of F.2. folds from the Eidvageid Schist Group and the
other metasedimentary groups reveal similar relationships; (S.1.) is
folded by the second phase folds and biotite and muscovite develop axial
planar to F.2., kyanite, garnets I and II, silliminate and other
porphyroblasts become augened in the anial planes of these structures.

B. Robins (personal communication), working on an adjacent area to
the west, has recorded sillimanite growing axial-planar to F.2. folds.
In all the thin-sections examined by the author, this type of
relationship has not been observed. In the pelitic Groups, however, such
as the Olderbugten Group, where syn or post-F.2. porphyroblastic
sillimanite is most likely to develop, it is very difficult to date the
growth of this mineral in relation to F.2., especially in the absence of
F.2. folds, which tend not to form in this lithology. Thus, it is not
possible to say definitely that there is no syn-F.2. sillimanite. Certain
facts are however, clear; at least some of the sillimanite porphyroblasts
have selvages of fibrolite which elsewhere can be dated as pre-Garnet II.
This implies that the sillimanite is also pre-Garnet II. Plate 168 shows
a sillimanite porphyroblast enclosed by alkali felspar, which can quite
definitely be dated as pre-mylonitisation. Where, however, such clear
relationships are not apparent, the position of sillimanite porphyroblasts
must remain equivocal. Examples of possible syn or post F.2.
sillimanite have been recorded; however, such examples tend to be rather
poikiloblastic and appear to be overgrowing a fine-grained biotite fabric
of the typed formed during mylonitisation. This would imply a post-
mylonitisation origin for the sillimanite.

This type of relationship means that the author cannot rule out the occurrence of syn-F.2. sillimanite on Eastern Seiland. Indeed, the mineral is to be expected, especially in the west where it may form as a contact phase to the late and post-F.2. basic plutóns. The occurrence described by Robins is almost certainly of this type. It may, however, be due to mimetic growth relative to F.2.

Thus, it seems that during F.2. times the metamorphic grade appears to have dropped down to Middle Greenschist Facies condition. The characteristic biotite, muscovite, epidote assemblages indicate conditions in the Quartz-Albite-Epidote - Biotite Sub-facies, with possible elevation by contact metamorpism to the upper part of the K-felspar Cordierite Hornfels Facies, Winkler (1965) during late and post F.2. times, this only occurring in the western part of the area.

(4) The Post F.2. Metamorphism

There is evidence of a post-F.2. hydrothermal metamorphism which gave rise to retrogressive changes in the body of the rocks and mineralisation along late brittle fractures.

The low grade assemblages which occur are quite widely developed in the Komagnes Group, though they tend to be most intensely developed adjacent to faults and kink zones. The Eidvågeid Schist Group and the other metasediments are not widely retrogressed though the low grade assemblages again occur in the region of faults.

In the body of the rocks the retrogressive changes are manifested by the chloritisation of biotite, amphibole and clinozoisite. The chlorite occurs an irregular felts often with relict biotite inclusions. This type of retrogression of green biotite yields a green chlorite and iron-ore which would seem to indicate a less ferriferous composition for the chlorite (Plate 169). This is also good evidence that the green

Plate 169. Chloritised green biotite. Note the iron ore
produced by the reaction. Upper Komagnes Group,
Rastaby. x 40, P.P.L.
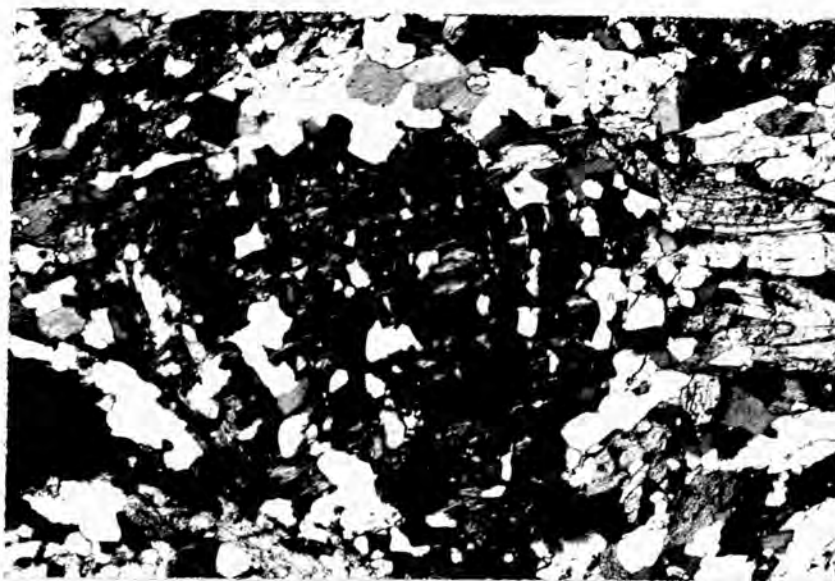
Plate 170. Chloritised garnet, Upper Komagnes Group,
Jernelv. x 40, X Pls.

colour is due to a high iron content.

Chloritisation of the very sporadic garnet that occurs in the Lower Komagnes Group has also been recorded; Plate 170 shows such a garnet which occurs in a rock specimen taken adjacent to the fault zone at Jernelv. Relicts of garnet can be seen within the chlorite pseudomorph which itself reflects the original skeletal form of the garnet.

Perhaps the most interesting manifestation of this hydrothermal metamorphism is to be found in the infills of late brittle cracks which transect the layering at a high angle within the Komagnes Group. The infills along these cracks are responsible for the discolouration occuring along the bounding surfaces of kink folds and joint drags (P 139). A fine example of the association of this phase of metamorphism with the kinking is seen in Plate 171. Here an albite crystal can be seen in which a conjugate set of kink zones is developed. One of the joint planes of this structure is infilled with epidote. If this crack is traced through the rock other minerals present within it are calcite and chlorite. The body of the rock also contains much chloritised biotite. A smilar occurrence parallel to the axial plane of a kink-fold from the Calc-Silicate schist is depicted in Plate 172, this consists of a coarse-grained aggregate of chlorite, epidote and calcite.

Further evidence of the association of this hydrothermal metamorphism with faulting is seen in samples of the Eidvågeid Schist taken from the northern corner of the head of Eidvaagenfjord at Eidvageid. It is clear from the strong brecciation occurring in the
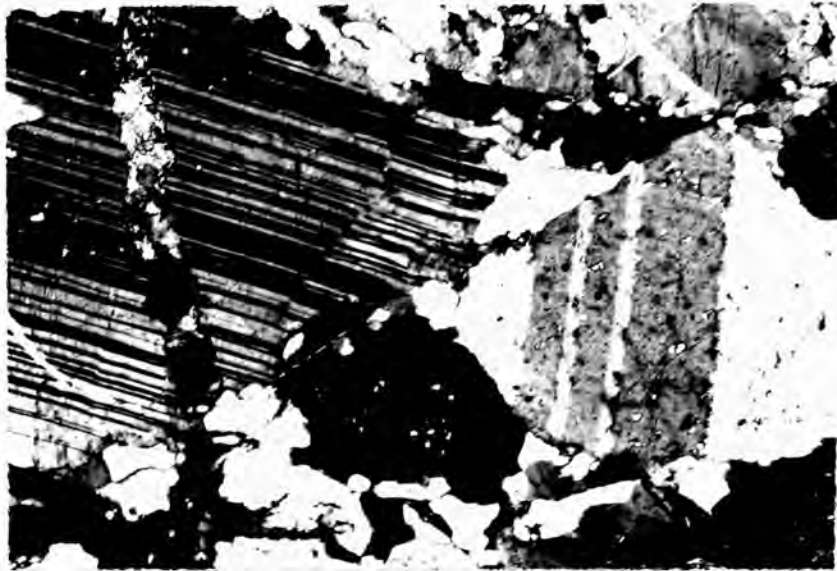
Plate 171. Conjugate kinks in albite; one kink plane contains
calcite and epidote. Lower Komagnes Group, north
of Jernelv. x 40, X Pls.
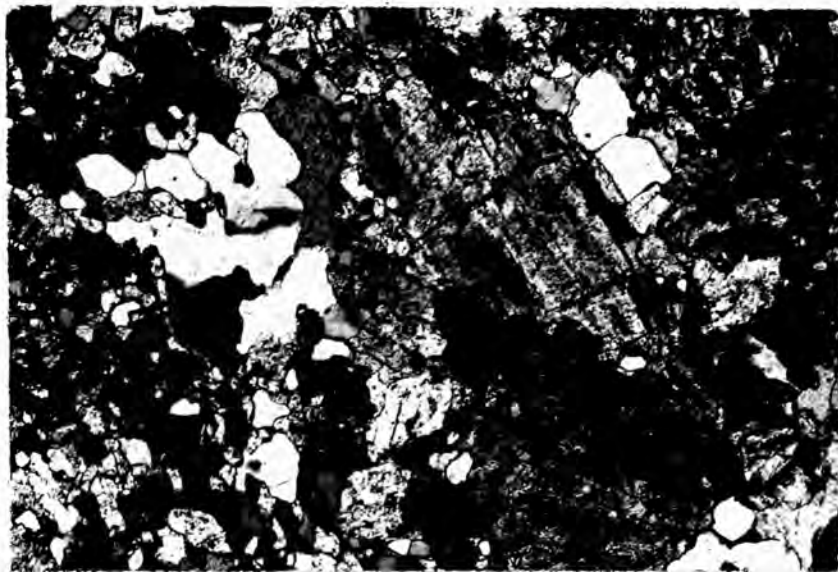
Plate 172. Coarse calcite, epidote and chlorite segregation
in a kink fold. Upper Komagnes Group, Jernelv.
x 40, X Pls.

thin neck of land that separate Eidvaagenfjord from Vasbugt, that there

is a fault present. This has been mapped by D. Powell. The fault is

obviously responsible for the local chloritisation of biotite and garnet

in the Eidvågeid Schist Group; Plate 173 shows an interesting example of

a garnet whose core has been selectively chloritised. This observation

lends further credence to the argument that stoll structure is due to

selective dissolution of zoned garnet, put forward in the section on the

static metamorphism.

Thus, during the phase of faulting, jointing and kink-fold formation,

the metamorphic assemblages calcite, epidote, chlorite were formed along

the structural planes formed by these deformations. These assemblages

are characteristic of the Quartz-Albite-Muscovite Chlorite Sub-facies

of the Greenschist Facies.

The observations described above are further evidence of the

temporal association of joints, faults and kink folds.

A summary of the sequences of mineral growth during metamorphism

is given in Fig. 53.

(A) Granitic Rocks Migmatites and Felspathisation

Introduction

The peak of regional metamorphism indicated by the kyanite

sillimanite porphyroblastesis in the metasediments was accompanied by

extensive migmatisation, felspathisation and intrusions of

adamellitic sheets.
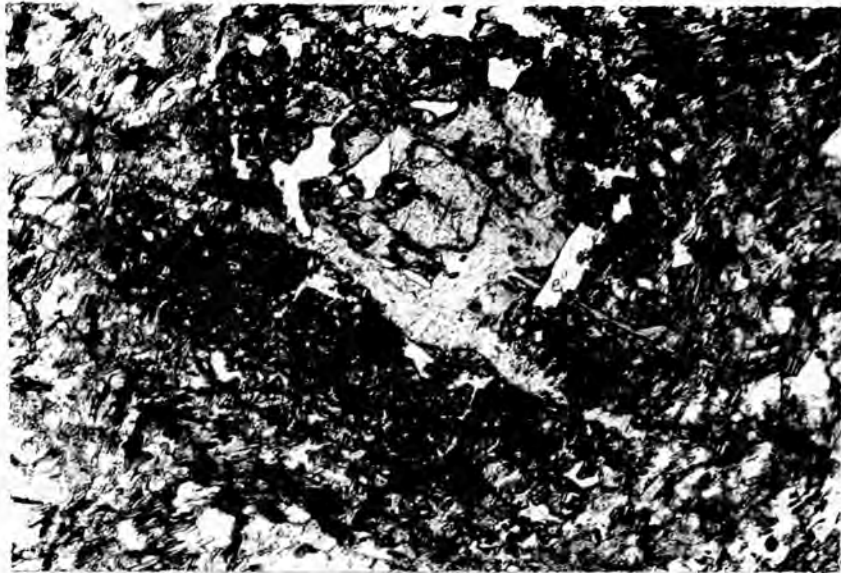
Plate 173. Garnet with a selectively chloritised core.
Eidvågeid Schist. x 40, P.P.L.

Plate 174. Migmatitic Olderfjord Group striking into
porphyritic Eidvaagtinn adamellite. Eidvaagtinn.

(1) The Migmatites

The development of migmatites occurs exclusively in the Olderfjord

and particularly the Olderbugten Groups. The migmatisation appears to

have developed pre-felspathisation, the latter process is closely

associated with the Eidvaagtinn adamellite. These relationships are

illustrated by Plate 174 which shows migmatitic psammites of the

Olderfjord Group striking into strongly porphyroblastic adamellite.

The migmatites are most easily studied on the coastal exposures

of the Olderbugten Group, both in the area of Olderbugten itself and

on the shores of Eidvaagenfjord, in the north. It would appear that the

schists and semi-pelitic rocks of this group are the most susceptible

to migmatisation.

In the field the migmatites may best be described as arterites.

They consist of a leucosome comprising of generally parallel quartz-

felspathic bands or streaks which are often ptygmatic (Plates 175, 176

and 177). The general attitude of these bands is parallel to the sheet

dip of the metasediments. Often the leucosomes are lenticular in shape

with large pink garnets in the cores which are intimately penetrated

by the leuosome material. Similarly, the thicker leucosomes are studded

with small pink garnets. Deformation and recrystallisation imparts a

flinty streaked-out aspect to these rocks (Plate 178).

The melanosome of these migmatites is generally a purple

garnetiferous schist which does not differ greatly in appearance from

the schist seen in areas which are not extensively migmatised.

Plate 175. Migmatitic Olderbugten schist, Olderbugten shore.

Plate 176. Migmatitic Olderfjord Group, Eidvaagenfjord.

Plate 177. Migmatitic Olderfjord Group, Eidvaagenfjord.

Plate 178. Deformed migmatitic Olderfjord Psammites,
north of Olderbugten.

Petrography

*Modes of two of the leucosomes are appended in Table 9.

Table 9

	Example 1 Lensoid leucosome Olderbugten schists.	Example 2 Thick leucosome vein Olderbugten schists.
Potash felspar	42.25	44.30
Plagioclase	26.45 (An 33)	17.75 (An 41)
Quartz	16.45	34.27
Biotite	10.70	1.12
Garnet	3.35	
Sillimanite		2.52
Others	0.80	0.04
Total	100.00	100.00

From these modes it is clear that Example 1 is adamellitic and Example 2 is granitic.

It is unfortunate that in almost every case the original grain boundary relationships between the minerals in the leucosomes, have been disrupted to various degrees by the mylonitisation. Example 2, however, appears to show only slight granulation at grain boundaries. The rock has no foliation. The potash felspar crystals are sub-hedral to anhedral, some have a long dimension of 7 mm. They are untwinned and generally strained.

---

* Note

These modes were obtained by staining the slides with sodium cobaltini-trite solution after etching with hydrofluoric acid, this produces a yellow stain on the potash felspar. Example 1 was also stained for plagioclase with potassium rhodizonate solution, which imparts a pink stain to the mineral.

The plagioclase crystals are of the same order of size and shape, they show some polysynthetic and carlsbad twinning, the twin lamellae are frequently bent. Potash felspar-potash felspar grain boundaries and potash felspar-plagioclase grain boundaries are frequently marked by a thin zone of granulation along which dense sheaves of fibrolite have grown. The quartz occurs as irregular pockets between the felspar grains and shows some evidence of polygonisation. It can be seen as inclusions in the latter. The small amount of biotite present is decussate and shows red-brown pleochroism. The sillimanite is acicular and the crystals may reach 1/2 mm. in length. Plate 179 illustrates some of these features.

The smaller lensoid leucosomes generally show rather more deformation, though it is obvious that they have basically a similar mineralogy with variations notably, in the amounts of garnet and biotite present (e.g. Example 1). This depending on the degree of admixture with the melanosome. The dense concentrations of biotite which occur in these admixed zones are clearly overgrown by Garnet II (Plate 180).

The origin of these leucosomes will be discussed in greater detail later, but it is believed that they are probably products of partial melting of the metasediments.

(2) The Eidvaagtinn Adamellite

This body occurs as two laterally discontinous sheets. The most westerly of the two sheets can be traced northwards from the western flanks of Eidvaagtinn to the northern corner of Hønsebyvann. In the region of Hønesbyvann the sheet has been boudinaged into a number of pods
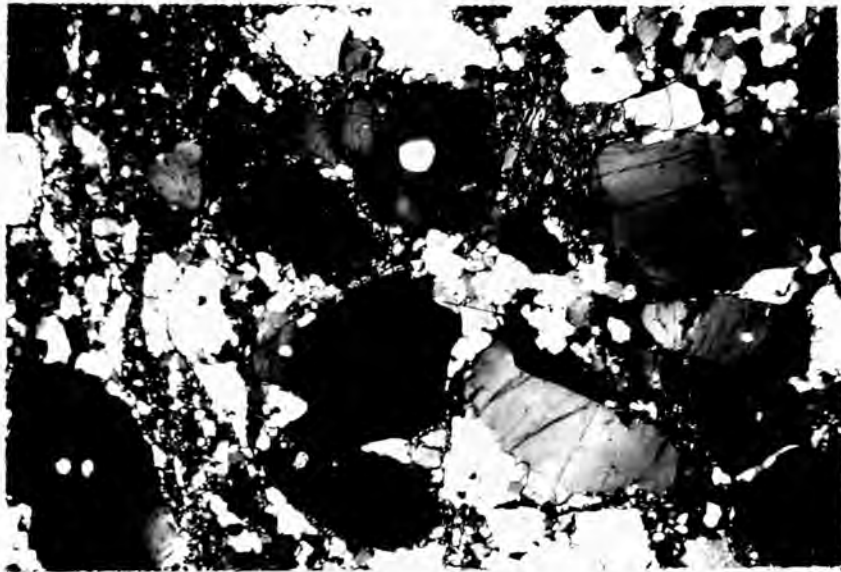
Plate 179. Deformed migmatitic leucosome. Olderbugten
Group, Olderbugten. x 10, X Pls.

Plate 180. Garnet II overgrowing the biotite of a migmatitic
melanosome. Olderbugten Group, Eidvaagtinn.
x 40, P.P.L.

which can bemapped in the field. The easterly sheet which is folded (Section 2) has an arcuate outcrop pattern which is caused by topographic·variations.

It is generally impossible in the field to map a definitive contact to the adamellite, since the associated felspathisation makes the passage metasediment to adamellite imperceptible. The adamellite is generally gneissose in appearance and it is obvious that it has undergone considerable deformation. Large pink alkali-felspar ovoids which make up a large proportion of the rock survive only as "porphyroclasts" in a fine-grained streaky groundmass.

Petrography

A mode of the adamellite is appended below.

Table 10

Potash felspar	30.29
Plagioclose	39.97
Quartz	28.77
Ore	0.37
Others	0.60
	100.00

This shows quite good agreement with the norm of the adamellite (Specimen 5, Table II). The rock in question has been strongly mylonitised and is now as augen-gneiss. Staining with sodium cobaltinitrite and potassium rhodizonate for alkali felspar and plagioclase respectively, allowed very satisfactory distinction between the two minerals. The disadvantage of the staining, however, lies in the fact that it obscures the biotite, which occurs in small amounts in the rock.

This section study of the rock shows that the original texture has been almost completely obliterated. The potash felspars occur generally on large porphroblasts (12 mm) in a fine-grained recrystallised groundmass (Plate 181). These surviving crystals and the groundmass grains show carlsbad and gridiron twinning (Plate 182). X-ray diffraction studies indicate an affinity with intermediate microcline. How much this structural state is an original feature is impossible to gauge. Some of the larger porphyroblasts show development of bleb and hair perthites (Plate 183).

The plagioclase, which makes up 39% of the mode, is generally untwinned making determination very difficult. Some myrmekite develops at the junction between the plagioclase and the microcline. The quartz in the rock occurs both on granules in recrystallised quartz-felspathic zones and as stringers which define the rock foliation with biotite. The latter mineral in the most recrystallised rocks occurs as laths with a chocolate brown pleochroism which define the foliation. In other facies where recrystallisation has not been so complete, the biotite has a rather granular aspect, probably due to the effect of the mylonitisation. Plate 184 shows this type of biotite overgrown by Garnet II. Other accessories are allanite and green amphibole.

The Origin of Adamellite

The gradational contacts found in association with the adamillite suggest a possible origin by metasomatism. Certain lines of evidence, however, point towards a magmatic origin. Plates 185 and 186 show examples of basic xenoliths in the granite. These display sharp igneous contacts; (185) shows little granitic veins penetrating the xenolith and (186) shows a similar situation with the veins deformed, probably by F.2. Elsewhere post-adamellite pegmatites and aplites have been recorded. On S/r/y, Taylor (personal communication) has mapped very
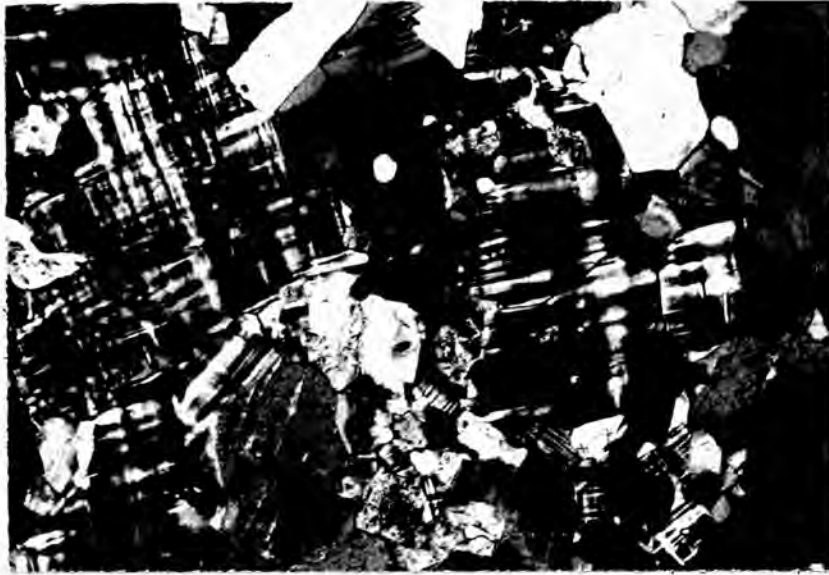
Plate 181. Granular mosaic of microcline. Olderfjord
Group, west of Hønseby. x 40, X Pls.
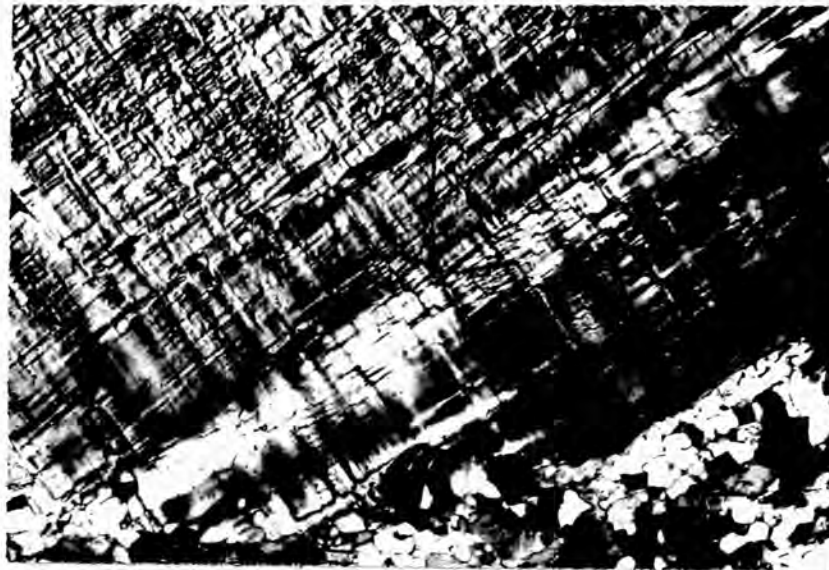
Plate 182. Gridiron twinning in microcline. Eidvaagtinn
adamellite. x 40, X Pls.
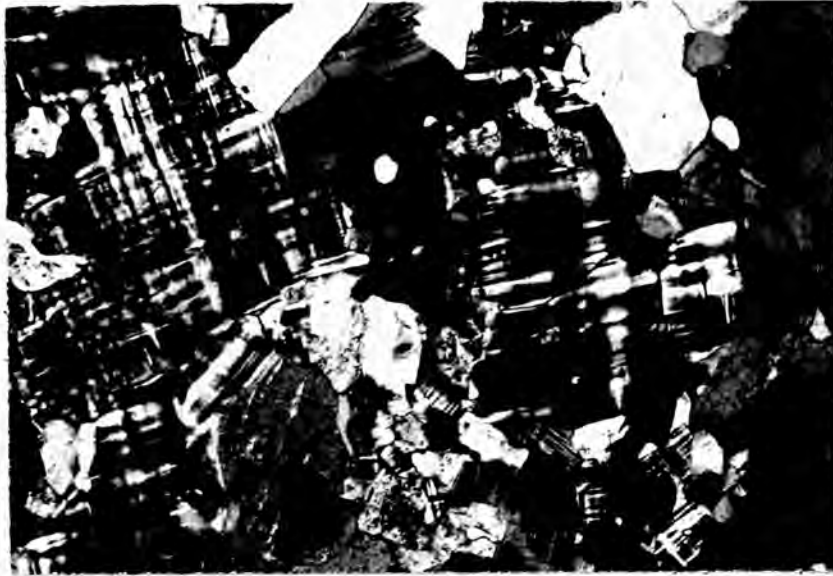
Plate 181. Granular mosaic of microcline. Olderfjord
Group, west of Hønseby. x 40, X Pls.

Plate 182. Gridiron twinning in microcline. Eidvaagtinn
adamellite. x 40, X Pls.
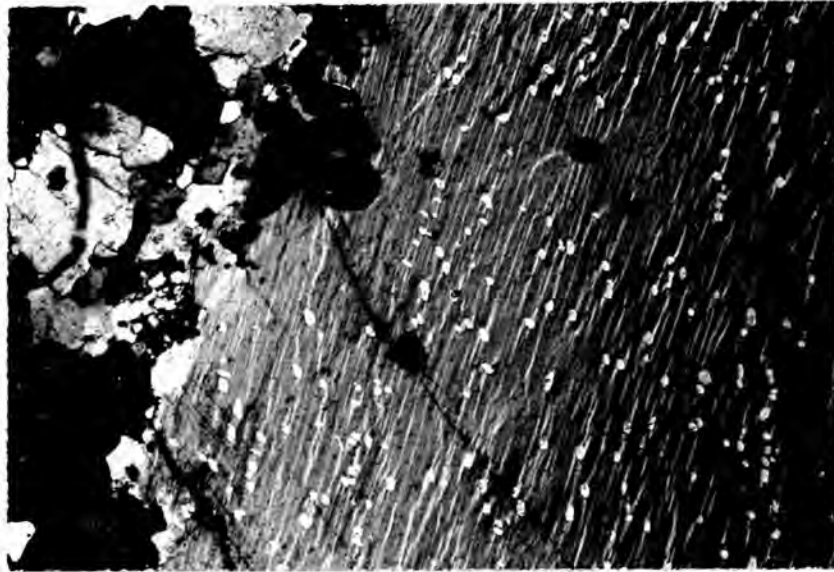
Plate 183. Bleb and hair perthite in potash felspar.
Eidvaagtinn adamellite. x 40, X Pls.
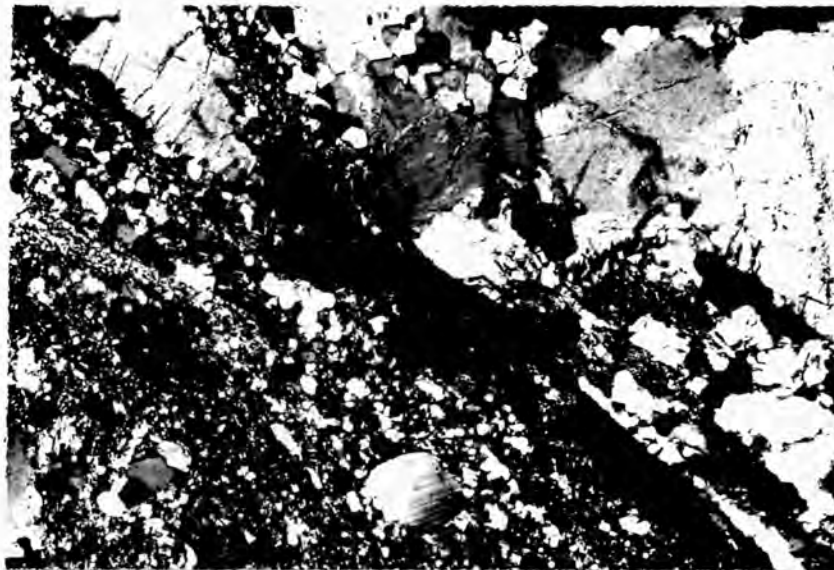
Plate 184. Garnet II overgrowing fine-grained biotite in the
Eidvaagtinn adamellite. Note the recrystallised
mylonitic fabric. x 40, X Pls.

Plate 185. Basic sheet xenolith in the Eidvaagtinn adamellite.

Plate 186. Basic sheet xenolith in the Eidvaagtinn adamellite.
Note the folded adamellite veins.

very similar adamellitic sheets which are unequivocally of magmatic origin. All these factors lead to the conclusion that the adamellite was intruded as a series of sheets, which were capable of producing intense felspathisation in the adjacent metasediments.

The origin of granitic rocks of this type has been for many years the subject of considerable controversy. Recent experimental work by Winkler (1967) and Von Platen (1965) has suggested that migmatitic rocks may be explained by partial melting of crustal rocks at high temperatures and pressures. Workers such as King (1965) have attacked these conclusions as being too facile and ignoring petrographic and field evidence. King (op.cit.) believes that in many cases the apparent mobilisation features seen in migmatites may be due to the fact that they have initially acquired a granitic composition by a process of metasomatism and it is this that has made them mobile, rather than a process of partial melting acting on a rock inherently containing the elements of a granitic melt. One of his reasons for believing this is that many rocks do not contain enough of the granitic elements to produce a reasonable quantity of granitic liquid.

This controversy is obviously not going to be entirely resolved here but the author proposes, with a rather limited amount of data, to test the validity of the ideas of Winkler and Von Platen.

Winkler (1967) has proposed that the composition of the minimum temperature melt produced by partial melting of a metamorphic rock, is controlled by the ratio of normative albite to normative anorthite, the Ab/An ratio. In his book, Winkler (op.cit.) gives a series of points for this composition on a projection of the system Quartz-Albite-Anorthite-

Orthoclase at PH_2O-2Kb, see fig. 48. From this it is clear that the effect of changing the Ab/An ratio from 7.8 - 1.8 is to shift the composition of the minimum temperature melt (M.T.M.) closer to the Qtz-Or join.

Consider Specimen I in Table II (P. 214). This was chosen in the field since it appeared to be showing a minimum of felspar porphyroblastesis and no migmatisation. It probably, therefore, fairly closely represents the composition of the metasediment, previous to the migmatisation. The calculated Ab/An ratio of this rock is 0.97.

Extrapolating the data of Winkler in Fig. 48, the author has estimated the composition of the M.T.M. of a rock with an Ab/An ratio approximately equal to 1 (Specimen 1).

Winkler also states that the effect of pressure on this system is to displace the M.T.M. compositions towards the Ab-Or join, e.g. a rock of Ab/An = 2.9 has its M.T.M. displaced to point (A) on Fig. 48, if the pressure is raised to 7Kb. It will be remembered that the coexistence of kyanite and sillimanite in the Olderfjord and Olderbugten Groups plus the presence of partial melting, suggested that the pressure operative during the maximum grade of metamorphism was approximately 9 Kb. Using a similar pressure extrapolation, to the one giving point (A) on the rock with Ab/An = 1, the point (B) is obtained. This must approximately correspond to the M.T.M. of a rock with a Ab/An ratio equal to 1, partially melted at 9Kb.

If the approximate cotectic lines are now sketched in for this M.T.M. it will be noted that the composition of the adamellite (Specimen 5, Table II) lies close to this cotectic. This suggests that

FIG 48

PROJECTION OF THE SYSTEM Qtz-Ab-An-Or-H₂O
AFTER WINKLER AND VON PLATEN

● ROCK ANALYSES 1 AND 5

9Kb COTECTIC

Qtz

Or

Ab
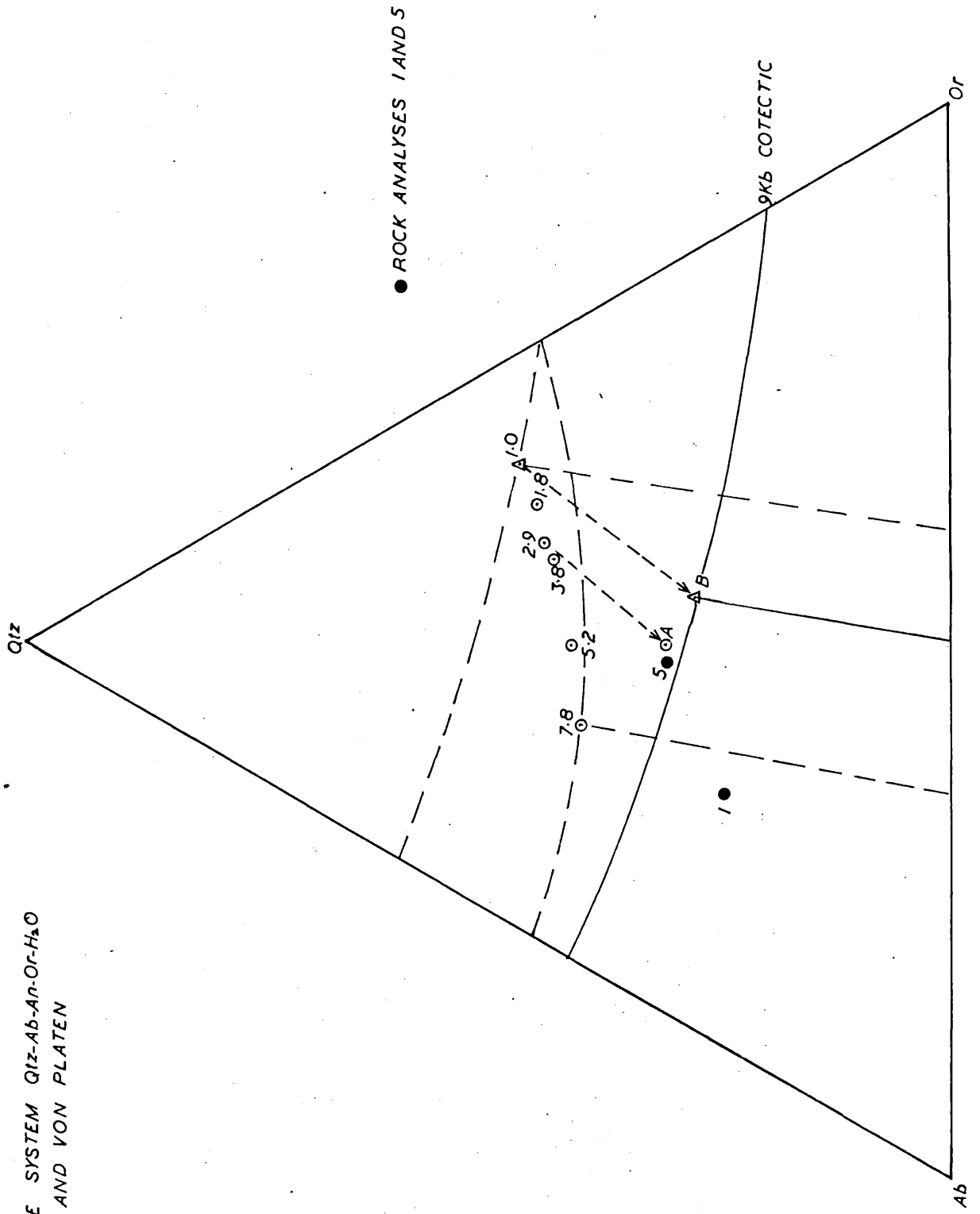
partial melting of a metasediment with an Ab/An ratio of 1, at 9Kb

could produce an adamellite of the composition of the Eidvaagtinn

adamellite.

There are, of course, many criticisms that can be made about this

line of argument.

(1) The extrapolations are subject to error.

(2) The Ab/An ratio of 1 is taken from only one chemical analysis.

There are undoubtedly variations over the whole area. This, however,

will not effect the argument unduly if the Ab/An ratio is low, since

the point representing the M.T.M. will still be shifted in such a manner

as to allow the adamellite to lie close to the cotectic.

(3) The value 9Kb is only approximate. It represents the pressure

during the coeval growth of kyanite and sillimanite which can be dated

as being before the felspathisation produced by the adamellite. The

pressure, however, must have been of this order.

Weill and Kudo (1968) have attacked the work of Winkler and Von

Platen, claiming that it is based on a misunderstanding of Tuttle and

Bowens work on the granite system. They contend that there is no unique

melting point neither is there a unique M.T.M. composition in systems

with constant Ab/An ratio, rather it is the composition of both the

alkali felspar and the plagioclase phases that determines at what

temperature a mixture of quartz plus these two phases will begin to melt.

They explain the expirically determined correlation between

Ab/An ratio and position of the cotectic minimum as found by Von Platen

(1965) as being due to the fact that the tie-lines joining

co-existing felspars in the system Q-Ab-An-Or-H_2O are sub-parallel

to the lines representing equal Ab/An ratios. Thus it only appears

that the Ab/An ratio controls melting, whereas it is the composition
of the co-existing felspars that actually controls melting.

The case put forward by Weill and Kudo (op.cit.) is a strong one,
however, the extrapolations executed in Fig. 48 must still be
approximately correct, since the planes of equal Ab/An ratio in the
system Q-Ab-An Or-H$_2$O jare, according to Weill and Kudo, sub-parallel
to the tie-lines joining the co-existing felspars, which actually control
melting. Thus, the position of the 9Kb cotectic must be approximately
correct.

Thus, it seems feasible that partial melting of a series of
metasediments with Ab/An ratio approximating to 1 at about 9Kb could
produce an adamellite magma of the composition of the Eidvaagtunn
adamellite. The very limited chemical data available suggests that the
Ab/An ratio was of this order.

The migmatites described earlier, it is suggested, are also products
of partial melting. In support of this there is definite evidence of
the magmatic mobility of this type of material. Plate 187 shows pale-
coloured granitic material invading foliated Hønseby gabbro. It is
certainly feasible that this material could have been squeezed out of the
migmatites as a liquid. These facts strongly suggests, therefore, that
the Eidvaagtinn adamellite represents a concentration of migmatic
leucosome material derived by partial melting within the body of the
Olderfjord and Olderbugten Groups. The intrusion of this liquid into
its present position was accompanied by intense felspathisation. It is
obvious that much more work needs to be done on this topic; a possible
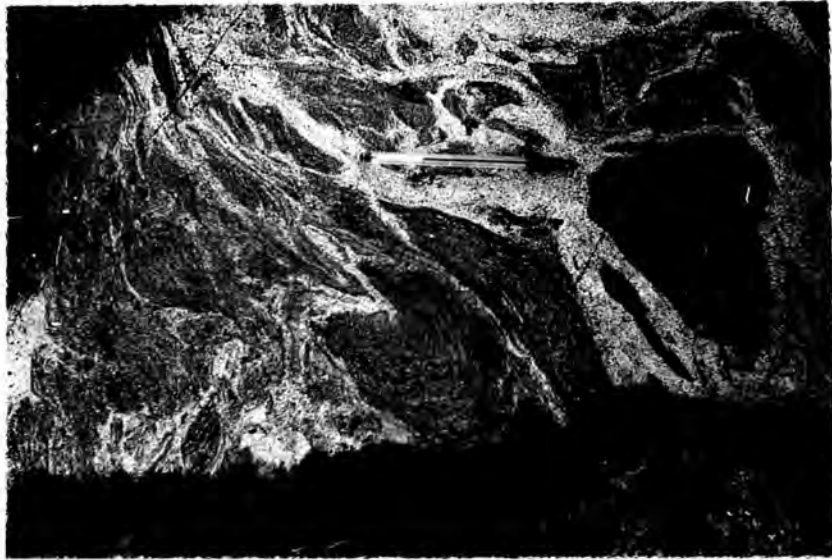
Plate 187. Leucocratic quartzo-felspathic material invading
foliated Hønseby gabbro. Eastern side of Hønsebyfjord.

Plate 188. Strongly albitised psammite of the Lower Komagnes
Group. Northern coast of the peninsula at Komagnes.

line of approach might be to examine the chemical composition of migmatitic leucosomes in relation to the cotectic lines of Fig. 48. It may even prove possible to define a cotectic using this method.

(3) Felspathisation Features

Study of the compositions of megascopic felspar porphyroblasts in the metasediments reveals a systematic change which is closely related to the changes in metamorphic grade indicated by the other porphyroblasts.

(a) The Lower Komagnes Group

This group appears to have been affected by an intense albitisation. In the hand specimen the albite occurs as segregations, veinlets and distinct porphyroblasts. These segregations are folded by F.2. and frequently have a micaceous selvage which is bowed by F.2. movements. These are a number of lines of evidence which suggest that these albites have been produced by a process of metasomatism.

(1) Bands of the characteristic flaggy psammite, when traced along the strike can be seen to be becoming progressively more felspathised, until a very massive gneissose rock is produced which contains relict folds and boudins, (Plates 188 and 189). This type of progression is best seen in a traverse north along the eastern coast of the peninsular at Komagnes. At the southern end of the traverse, the psammites are the normal flaggy variety, at the northern end, the rocks depicted in the two plates occur.

At no other locality within the Lower Komagnes Group has such intense albitisation as this been recorded. Normally in this section the psammites are studded with a number of albites. In rocks from the northern end of the peninsula, however, albite makes up 50-60% of the

Plate 189. Relict F_1 folds in albitised psammite of the
Lower Komagnes Group. Northern coast of the
peninsula at Komagnes.
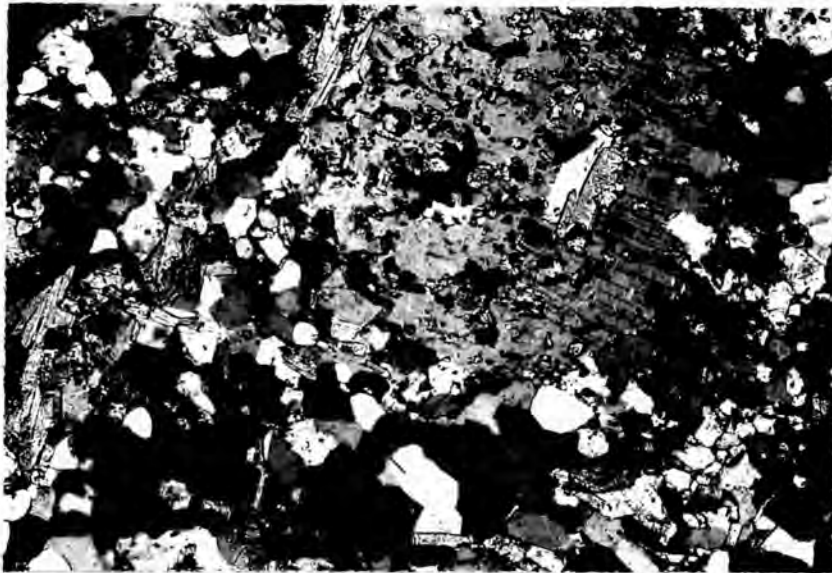
Plate 190. Ragged albite overgrowing the groundmass fabric
of the psammites of the Lower Komagnes Group.
Vasbugt. x 40, X Pls.

rock mode. There seems no doubt, therefore, that this area was very close to a centre of intense soda-metasomatism.

(2) In thin section there is ample evidence of the replacive nature of much of the felspar. Porphyroblasts may be observed at various stages of overprinting the groundmass fabric. A ubiquitous feature is the presence of large numbers of inclusions of groundmass epidote, biotite, muscovite and quartz. Often these can be seen at various stages of digestion by the albite. The earliest stage of growth is typified by Plate 190. Here a rather ragged albite can be seen with a large number of inclusions of quartz and epidote, although the crystal is transected by a strip of polygonised quartz the two halves are in optical continuity. Plate 191 shows an anhedral albite overgrowing sheaves of muscovite. The latter has been partially digested at the edge of the albite. This partial digestion of groundmass minerals can also be seen in Plate 192, where quartz appears to be undergoing the process; the undigested quartz remains as irregular bulbous blebs. All of these stages may be developed in a single slide.

This progressive albitisation of the psammite leads to the destruction of the foliation and the production of a granular granitoid rock.

(b) The Upper Komagne Group

There is no strong development of porphyroblastic felspar in the Upper Komagnes Group. Small white ovoids have been observed, these are mostly untwinned, optically negative with a large 2V. They are almost certainly in the oligoclase-andesine range.
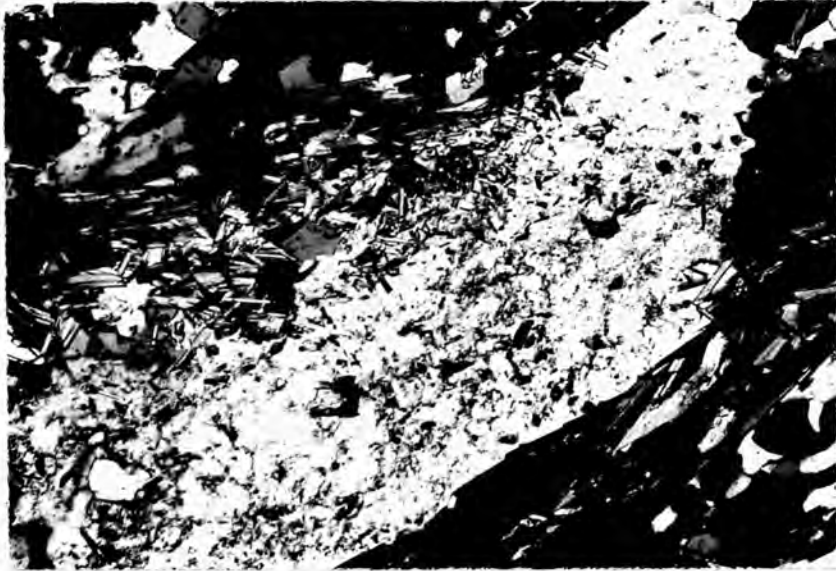
Plate 191. Albite containing partially digested muscovite.
Lower Komagnes Group, Vasbugt. x 40, X Pls.
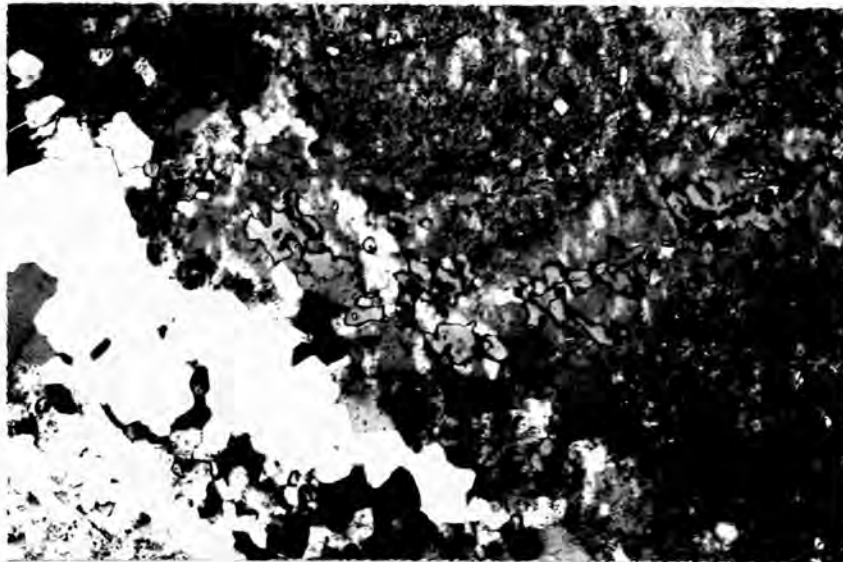
Plate 192. Albite with quartz inclusions. Lower Komagnes
Group, Vasbugt. x 40, X Pls.

(c) <u>The Eidvageid Schist Group</u>

The composition of the felspar porphyroblasts varies within the
Eidvågeid Schist group. A distinctive change takes place across the
line indicated in Fig. 49, this approximately corresponds to the first
appearance of fibrolitic sillimanite. To the east of the line, the
porphyroblasts are plagioclase (An 29). They are characteristically
ovoid or lenticular in shape and are occasionally twinned. They
frequently contain inclusions of mica, (Plate 193), and their irregular
edges give the impression that they have arisen by replacement of the
groundmass minerals in a manner similar to the albites of the Lower
Komagnes Group. The accompanying aluminosilicate is porphyroblastic
kyanite.

To the west of the line indicated in Fig. 49 the felspar
porphyroblasts consist of both plagioclase and alkali felspar together
with fibrolitic sillimanite.

(d) <u>The Alkali Felspar Porphyroblastesis</u>

This occurs with varying degrees of intensity to the west of the
line mentioned above. It is at its most intense in the Olderbugten
Group in association with the Eidvaagtinn adamellite. As this body is
approached across the strike, it is apparent that the amount of
porphyroblastic felspar is increasing markedly. The adamellite appears
to have been, therefore, the centre responsible for this
porphyroblastesis.

In the field the felspathisation is marked by a studding of the
metasediments especially in the schistose groups, with large (2 cm.)
ovoidal or tabular felspars; they are generally white in colour and
have been deformed to varying degrees by the later mylonitisation,
which causes them to become streaked-out and augened.

FIG 49

MAP OF METAMORPHIC ISOGRADS FOR THE STATIC PHASE BETWEEN
F.1 AND F.2

Ab. Epid. Hbde. Biot. Musc

Plag. Epid. Hbde. Biot. Musc.
Garnet.

Plag. Ky. Biot. Musc.
Garnet.

K-Felsp. Plag. Ky. Biot. Musc. Fibr.
Garnet.

K-Felsp. Ky. Sill. Biot. Fibr.
Garnet.

2KM
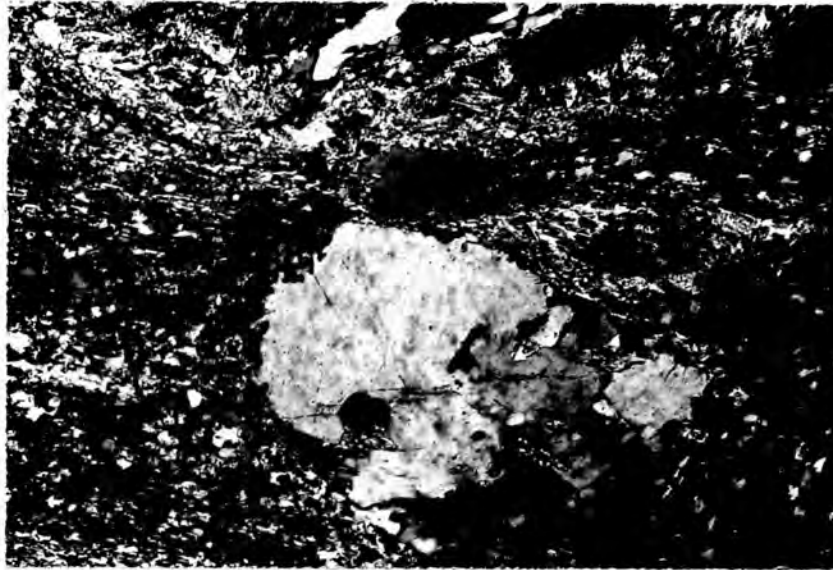
Plate 193. Felspar porphyroblast with mica inclusion.
Eidvågeid Schist. x 40, X Pls.
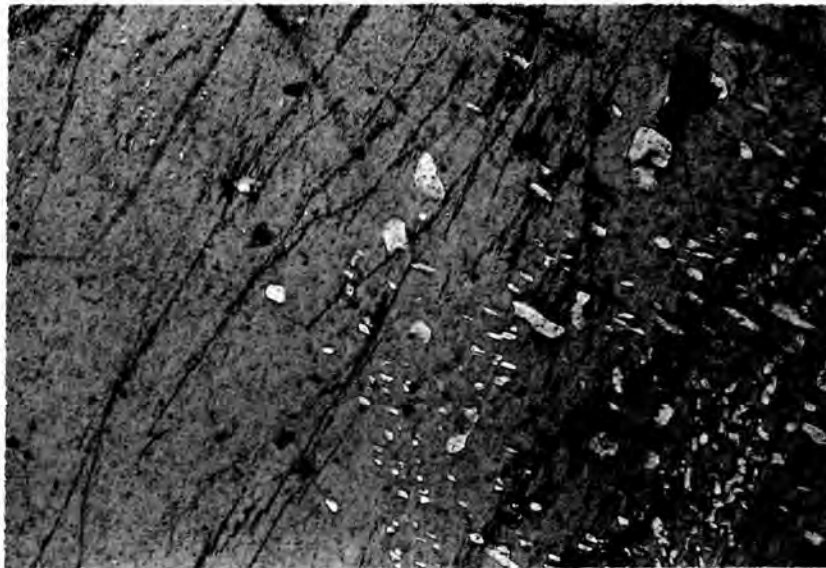
Plate 194. Bleb perthites in a potash felspar porphyroblast.
Olderbugten schists, Hønsebyvann. x 40, X Pls.

Plate 195. Large potash felspar porphyroblast with inclusions.
Olderbugten schists, Hønsebyvann. x 40, X Pls.

Plate 196. Myrmekite in a migmatitic leucosome. Olderbugten
schists. x 40, X Pls.

In thin section it is apparent that these porphyroblasts are
of alkali felspar; plagioclase has only been recorded in association
with alkali felspar in the Upper Eidvågeid Schist. Carlsbad and
gridiron twinning are very common, occasionally the crystals are
perthitic, the intergrowth being of the blab-type (Plate 194). The
porphyroblasts are almost always strongly poikiloblastic; they contain
inclusions of all the groundmass minerals including kyanite and garnet
(Plate 195). Occasionally myrmekite develops at the junction of the
porphyroblasts and the groundmass (Plate 196). This growth is lobed into
the body of the alkali-felspars.

Geochemical Changes Accompanying Felspathisation

In order to ascertain the chemical changes involved during the
alkali felspar felspathisation, a series of four specimens were taken
along the strike of a band within the Olderbugten Schists, the total
distance between the first and the last specimen is about 10 metres.
Each specimen was chosen in the field because it seemed to indicate a
visible increase in the presence of felspar porphyroblasts over its
predecessor. Plates 197, 198 and 199 show the sites at which three of the
specimens were taken.

Chemical analyses of these specimens are given in Table II
(Specimens 1-4) together with calculated norms and Niggli values. Also
included in an analysis of the Eidvaagtinn adamellite (Specimen 5).

Figs. 50A and B shows plots of Niggli values for this
felspathisation series. The trends seen may be summarised as follows :

Plate 197. Slightly felspathised schist of the Olderbugten Group. South-west flanks of Eidvaagtinn.
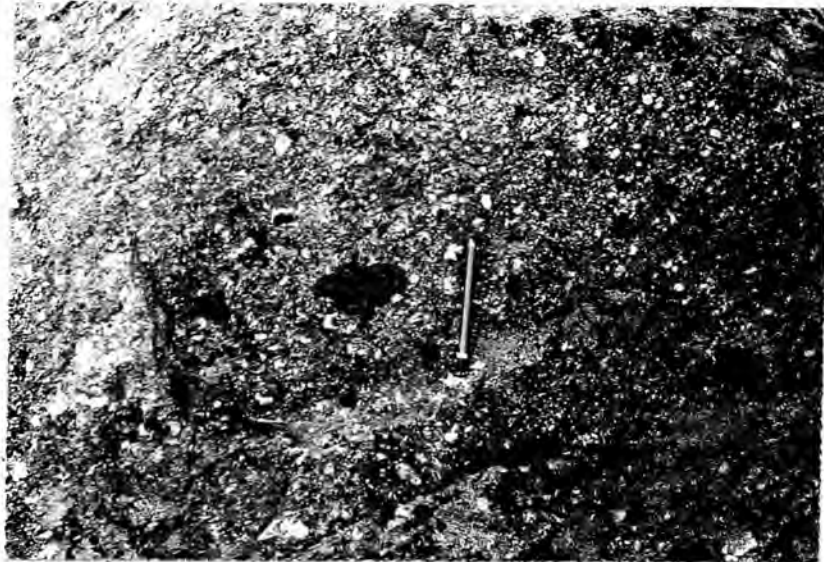
Plate 198. Felspathised schist of the Olderbugten Group. South-west flanks of Eidvaagtinn.

Plate 199. Highly felspathised schist of the Olderbugten
             Group, south-west flanks of Eidvaagtinn.

FIG 50

PLOT OF NIGGLI PARAMETERS (N.P) FOR THE GRANITISATION SERIES
IN THE OLDERBUGTEN GROUP
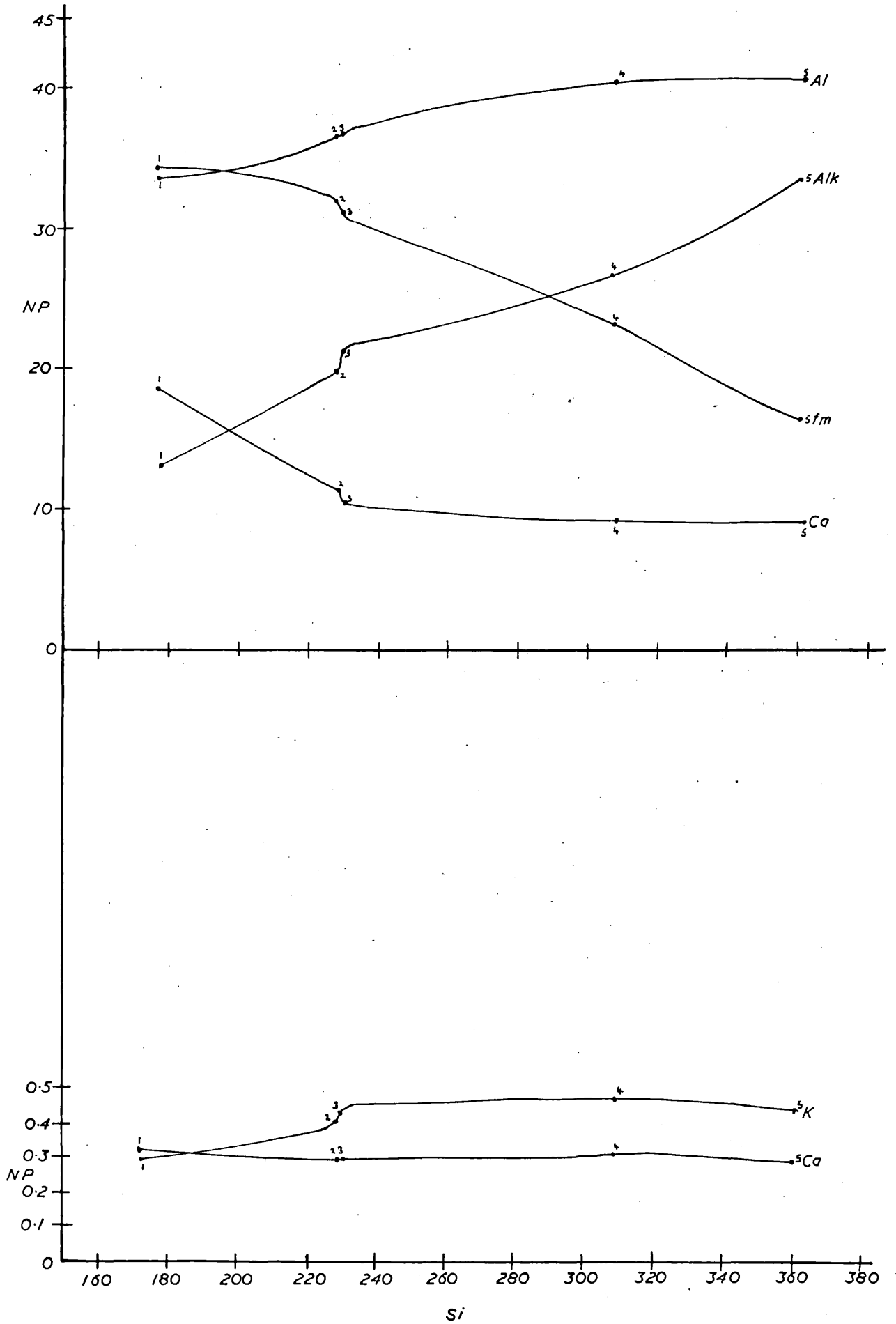
FIG 51

PLOT OF NORMS FOR THE GRANITISATION. SERIES
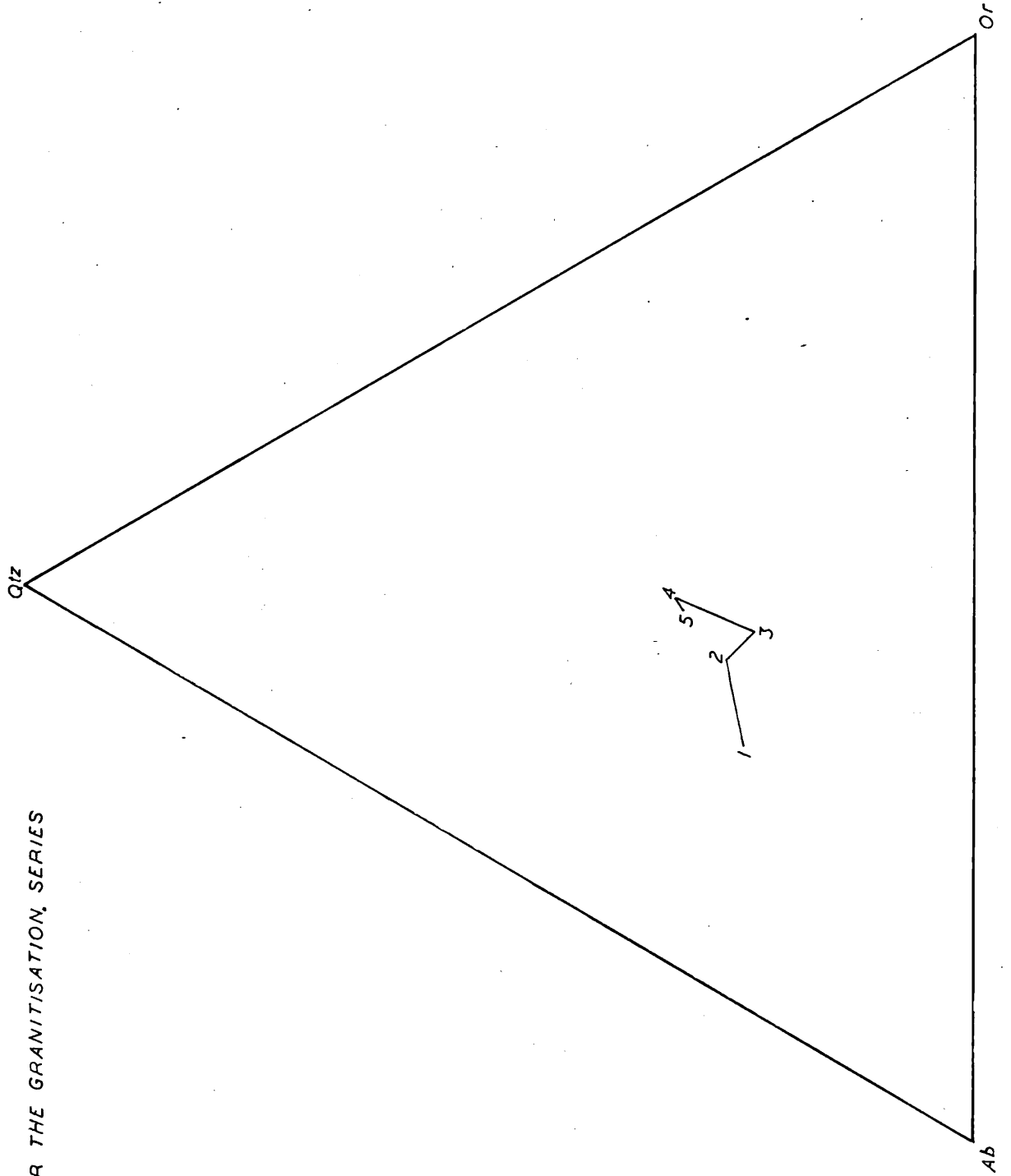
With increasing Sl : (a) Al rises   (d) Ca decreases

                     (b) Alk rises   (e) K shows a slight overall
                                              increase.

                     (c) fm decreases   (f) Mg remains about the same.

These changes indicate an enrichment in potassium, sodium and silica, with a concomitant depletion in iron, calcium, magnesium and manganese. The Al parameter shows an increase with progressive grantisation, though the actual weight per cent of alumina shows a decrease. This implies that the increase in Al is only relative, undoubtedly due to the depletion in the fm and Ca constituents.

Fig. 51 shows the change in normative quartz and felspar, associated with the granitisation.

Table II Felspathisation Series

Geochemical Data

Oxide Wt. %	1	2	3	4	5
S_1O_2	56.96	62.64	62.74	68.50	71.66
T_1O_2	0.3.	0.84	0.79	0.63	0.46
Al_2O_3	18.31	17.08	17.17	15.34	13.72
Fe_2O_3	2.33	0.35	0.92	0.58	0.97
FeO	6.63	6.58	6.23	3.70	1.87
MnO	0.25	0.13	0.13	0.07	0.04
MgO	2.39	1.71	1.66	1.09	0.64
CaO	5.67	2.94	2.71	1.91	1.71
Na_2O	3.07	3.42	3.40	3.28	3.76
K_2O	1.98	3.35	3.95	4.39	4.77
H_2O+	0.99	0.53	0.59	0.47	0.52
H_2O-	0.24	0.07	0.10	0.11	0.14
P_2O_5	0.22	0.27	0.22	0.04	0.12
Total	99.85	100.41	100.61	100.11	100.38

Analyst H. Lloyd

Trace Elements PPM

	1	2	3	4	5
Nb	21	24	24	11	8
Rb	163	244	275	255	292
Sr	151	193	192	110	70
Zn	127	173	164	102	49
Zr	240	461	408	326	264
Y	58	110	93	61	41

Percentage Norms

	1	2	3	4	5
Quartz	12.164	17.179	15.807	24.555	26.634
Orthoclase	11.700	19.796	23.341	25.941	28.187
Albite	25.977	28.939	28.770	27.754	31.816
Anorthite	26.692	12.821	12.007	9.214	6.471
Corundum	1.334	3.129	2.901	1.816	-
Magnetite	3.378	1.232	1.334	0.841	1.406
Ilmenite	1.538	1.595	1.500	1.196	0.874
Apatite	0.510	0.626	0.510	0.093	0.278
Water	1.230	0.600	0.690	0.580	0.660
	99.850	100.410	100.61	100.11	100.38

Niggli
Parameters

Al	33.53	36.64	37.03	40.57	40.66
fm	34.41	32.04	31.06	23.40	16.50
Ca	18.88	11.47	10.63	9.18	9.21
Alk	13.17	19.85	21.28	26.84	33.63
Si	177.03	228.04	229.62	307.44	360.35
Ti	1.89	2.30	2.17	2.13	1.74
P	0.29	0.42	0.34	0.08	0.26
H	12.75	7.29	8.42	8.68	11.07
K	0.30	0.39	0.43	0.47	0.45
Mg	0.32	0.29	0.29	0.31	0.29
Qz	24.33	48.65	44.48	100.09	125.83

===

1 - 4 Felspathisation series K24, 32A, B, C1 and C2.
    5 Eidvaagtinn adamellite.

These geochemical changes indicate that the metasomatism which
is obviously associated with the intrusion of the Eidvaagtinn adamellite
tends to change the composition of the schists such that they become more
like the adamellite.

Summary and Discussion

Associated with the intrusion of the Eidvaagtinn adamellite, which
was produced by partial melting within the Olderbugten and Olderfhord
groups, a zone of intense alkali-metasomatism develops. When traced
outwards from the adamellite core, there appears to be a zonal distribution
of the felspar perphyroblasts which is accompanied by a change in the
composition of the other porphyroblastic minerals.

(1) In the Olderfjord and Olderbugten Groups the assemblages consist of
K - felspar, porphyroblastic kyanite and sillimanite together with
fibrolite.

(2) In the structural top of the Eidvågeid Schist Group, K-felspar,
plagioclase with kyanite and fibrolite.

(3) In the Lower Eidvågeid Schist plagioclase and kyanite.

(4) The Upper Komagnes Group contains plagioclase and no aluminium
silicates.

(5) The Lower Komagnes Group albitic felspar with no aluminium silicates.

The complete mineral paragenesis for each zone is given in Fig. 49.
It is obvious that these changes represent a lateral change in
metamorphic grade.

In many ways this picture is similar to that recorded by
Goldschmidt (1921) in the Stavager area and by Reid (1927) and Reynolds
(1942) in the Highlands of Scotland and Antrim. In the Stavanger area
Goldschmidt records a Trondhjemitic intrusion emplaced along the boundary
plane between garnet-bearing phyllites and green schists. Between the
Trondhjemite and the phyllites there is a zone of migmatite beyond which
porphyroblasts of sodic-plagioclase develop in the phyllites. Goldschmidt

concludes that the Si and Na were derived from the Trondhjemite while the Ca was derived from non-magmatic solutions circulating in the contact zone.

In the Dalradians of Scotland, Albite schists which occur in the nose-zone of the recumbent Carrick Castle fold, are regarded by Reynolds (1942) as representing a zone of soda-enrichment driven forward in advance of a syntectonic igneous intrusion emplaced along the axial zone of the fold. This was subsequently removed by erosion leaving the albite schists as the only evidence of its existence. Likewise Read (1927) in his description of ;the migmatitic "Older Granites" of Deeside notes that these rocks are largely oligoclase-biotite gneissee which marginally grade outwards through a zone of oligoclase porphyroblast schist. Read records the Trondhjemitic composition of the migmatite leucosomes and draws an analogy between the mode of formation of the oligoclase porphyroblast schists and Goldschmidts albite schists.

Reynolds (op.cit.) notes that the albite schists occur in the south-west Highlands in the biotite zone whereas further to the north-east, in Reads area, the oligoclase porphyroblasts schists occur in the kyanite and sillimanite zones. This variation in felspar content he implies is due to a thermal gradient, with the highest temperature in the north-east. Indeed, Read (1940) suggested that the Barrovian zones are not only thermal zones but also metasomatic zones through which material from the heart of granitisation migrated, changing in composition and temperature as it advanced from the core.

The picture on Eastern Seiland is in many ways a condensed version of the situation in Scotland as described by Read and Reynolds (op.cit.).

The chief difference lies in the composition of the adamellite which
is not analagous to the Trondhjemitic leucosomes of Reads area. The
difference, of course, explains the extra K-felspar zone which does not
exist in the areas described above. The origin of the adamellite has
already been explained in terms of partial melting at 9Kb and $660^{\circ}C$.

A slight anomaly in the distribution of these felspar zones round
the adamellite is the existence of the zone of most intense
albitisation furthest away from the adamellite. It is also noteworthy
that Weight per cent Na2O in the granitisation sequence described above
remains fairly constant, rising only slightly. This suggests that sodium
was not quantitatively important during the felspathisation associated
with the adamellite and could hardly be responsible for the zone of
intense soda metasomatism seen at Komagnes. This raises the possibility
that these were two centres from which migration occurs; one the
Eidvaagtinn adamellite and the other possibly Trondhjemitic, now lost
under Vargsund. The difference in composition is considered to be a
function of the variation in the metamorphic grade prevailing in the two
areas. Such an intrusion would be analagous to that postulated by
Reynolds (op.cit.) in the core of the Carrick Castle fold.

With regard to the chronology of these two felspathisations, all
textural evidence seems to suggest that all the porphyroblasts felspar
phases of Eastern Seiland developed late in the static interval between
F.1. and F.2. Thus, within the limits of accuracy imposed by a
timescale erected on the relationships between the various metamorphic
minerals, the felspathisations must be regarded as being sensibly
synchronous, even if they did emanate from more than one source.
Therefore, the zones defined by the change in composition of both the
felspathic and other mineral assemblages reflect a contemporaneous

lateral change of metamorphic grade. These facts support Reads

contention (1940) based on the Scottish Highlands, that metamorphic zones

represent not only thermal and pressure zones but also metasomatic zones.

Chapter 5

The Mylonite Belt

Introduction

After the main phase of porphyroblast growth and the migmatisation
and granitisation, there ensued a phase of intense deformation which
fundamentally changed the texture within the Eidvågeid schist and
structurally higher metasedimentary groups. The development and
recrystallisation of these textures is post-dated by the second phase
of garnet growth.

The term which comes closest to describing the texture of these
highly deformed rocks is mylonite. Waters and Campbell (1935) give a
survey of all the currently used mylonitic terminology. All of these
terms are inadequate to describe the textures developed in Eastern
Seiland. This is because they all imply a degree of physical milling
produced by translative movements parallel to the banding of the mylonites,
this being followed by recrystallisation to various degrees. It will be
demonstrated later that the movements producing the texture on Eastern
Seiland were not translative, the various terms used to describe
mylonite are therefore not strictly applicable. In the absence of a
suitable term "mylonite" will be used in a purely descriptive non genetic
sense.

If the effects of the F.2. folding are ignored, the zone of rock in
which these mylonitic texture occur is somewhat in excess of 4,000
metres in thick mess and extends along the strike at its most intense
in the Olderfjord and Olderbugten Groups and to a lesser degree in the
Eidvågeid Schist Group (Fig. 52). The Komagnes Group appears to have

FIG 52.
MAP OF THE DISTRIBUTION OF MYLONITIC ROCKS

N

☐ .MYLONITE

▨ UN-MYLONITISED

⊢ 2KM ⊣

largely escaped the deformation though some effects are apparent, as will be shown later. In the field the rocks are rather variable, this being a function of their pre-mylonitisation aspect. The rocks are generally fine-grained and rather massive, with a blocky jointing. Porphyroblasts and migmatitic segregations are streaked-out to various degrees imparting a strong banding to the rocks which is sensibly parallel to the original lithological banding. The Eidvaagtinn adamellite has a very strong deformational banding, marked by flattened and streaked-out porphyroblasts. Sections cut parallel to the mylonitic banding show no trace of any lineation or preferred orientation of streaked-out porphyroblasts. Intense recrystallisation in many cases, particularly in the psammitic rock type, has produced extreme flintiness. These flinty rock types are studded with relict porphyroblasts and pink garnets. Plates 200 to 203 show some of the field characteristics of the rocks.

The mylonitic banding shown in the photograph is folded by F.2. (Plate 204).

(A) <u>Textural Relationships</u>

The entire range of textures seen in the quartzo-felspathic minerals within the mylonitic belt can be explained in terms of a progressive reduction in grain-size due to recrystallisation along zones of high strain energy. A sequence of changes can be recognised, particularly in the felspathic porphyroblasts, which although they have been frozen into the rock at different stages, must represent the temporal order of development of the mylonitic texture.

The first stage appears to be the breakdown of the large porphyroblasts into irregular lensoid sub-grains. This appears to
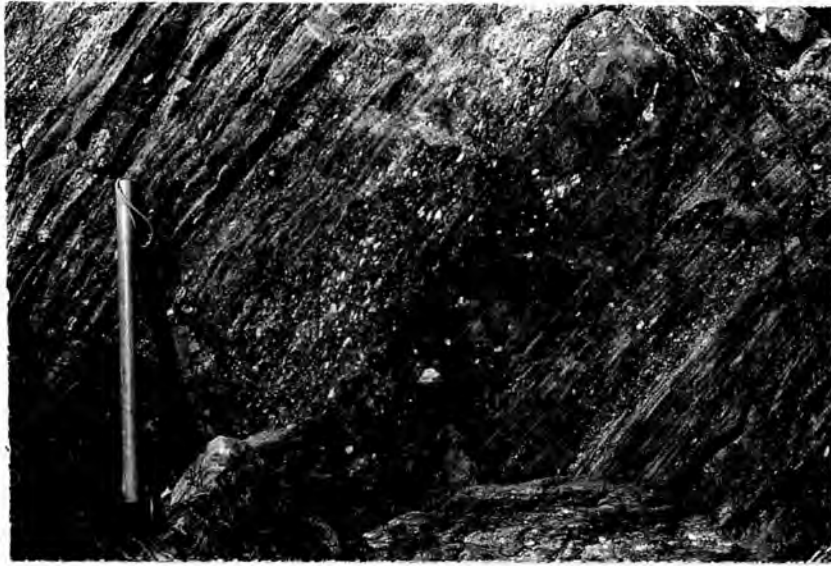
Plate 200. Mylonitised facies of the Olderfjord Group,
south-west of Hønsebyfjord.

Plate 201. Felspathised and mylonitised facies of the
Olderfjord Group, north of Hønsebyvann.
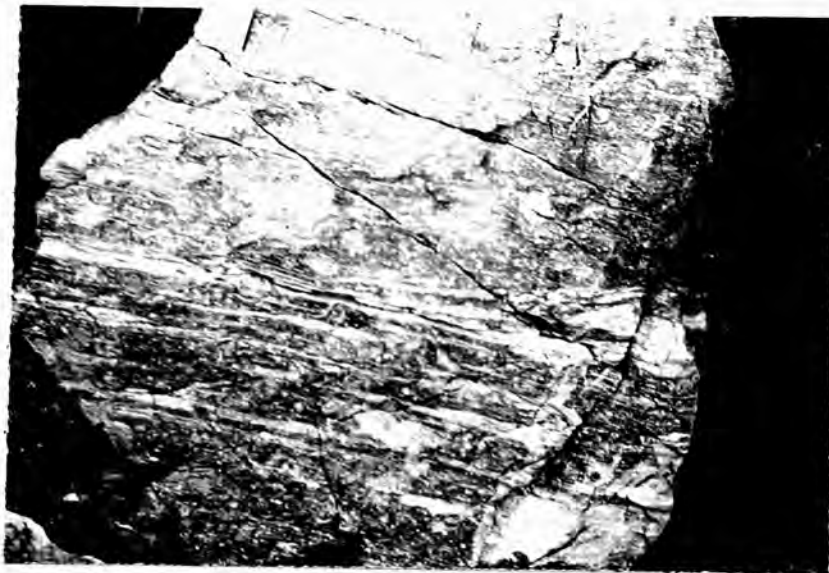
Plate 202. Mylonitised Olderfjord Group, Storvann.

Plate 203. Mylonitised Olderfjord Group, Storvann.

Plate 204. F₂ fold deforming the mylonitic banding in the
Olderfjord Group. North of Hønsebyvann.

involve in many cases, actual physical rotation of the lattice, since

features such as twinning and exsolution lamellae are disoriented in

the resulting texture (see Plate 205). This type of breakdown may occur

at the edges of porphyroblasts or in zones transecting the porphyroblasts.

Generally the boundaries between the comminuted zones and the porphyroblasts

are sharp. The surviving relics show evidence of strain extinction. An

example of a zone of comminution cutting a microcline porphyroblast is

given in Plate 206, within the zone of breakdown, the twin lamellae can

be seen to be disoriented.

Further reduction in grain size is accomplished by recrystallisation

of sub-grains along the margins of the coarser lenticles formed by the

initial breakdown of primary porphyroblasts (Plate 207). These zones

are undoubtedly sites of concentration of strain energy, this

concentration giving impetus to the recrystallisation. Repetition of this

process may result in the complete reduction of porphyroblasts and

groundmass to a fine-grained aggregate (Plate 208).

Under the influence of the high temperatures and pressures

prevailing in the rocks, grain growth and recrystallisation took place

in the rocks. The degree of recrystallisation appears to be controlled

by two factors:-

(1) Local concentrations of strain energy. This factor has its greatest

influence in the region of felspar porphyroblasts, which appear to have

acted as relatively rigid bodies during the deformation. This strain

energy may have concentrated around them. This is well seen in the

relatively coarse polygonal fabrics that develop at the sites previously
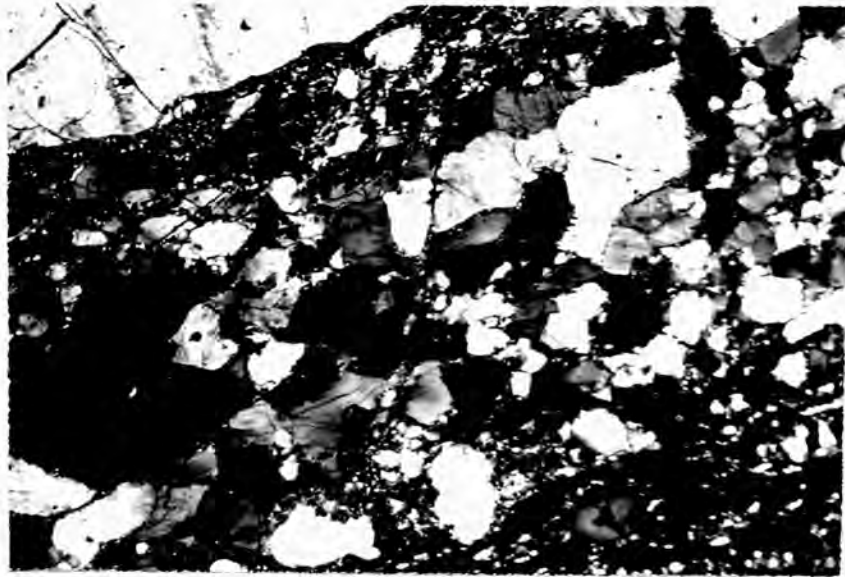
Plate 205. Lensoid sub-grains illustrating an early phase
in the development of the mylonitic fabric.
Eidvaagtinn adamellite. x 40, X Pls.
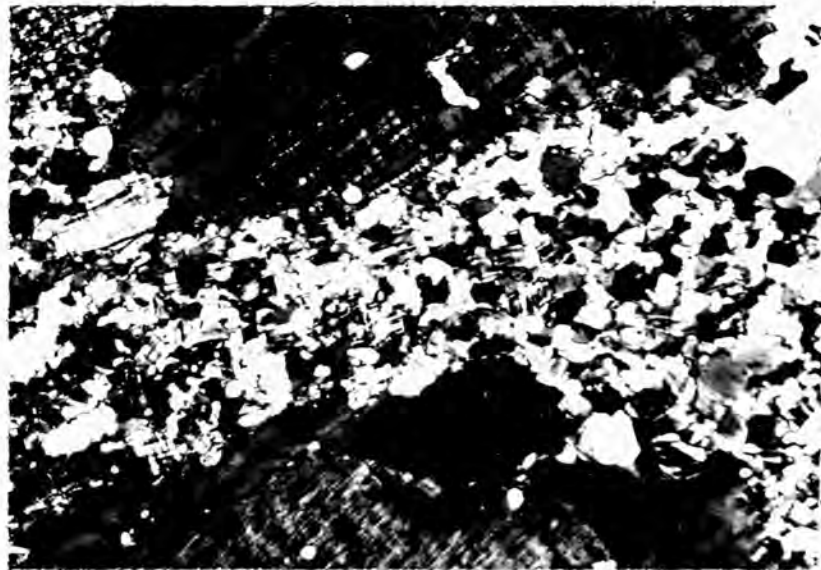
Plate 206. Zone of comminution cutting a microcline
porphyroblast. Eidvaagtinn adamellite. x 40, X Pls.

occupied by felspar porphyroblasts see Plates 208 and 209.

(2) The intensity of deformation undergone by any rock type. This is particularly well illustrated by the case of the Eidvaagtinn adamellite, which in polished sections can be seen to be intensely deformed; relict pink porphyroblasts have contiguous trails of very fine-grained pink material streaked-out either side of the relic. These trails obviously represent comminuted felspar. The whole rock has the appearance of an augen-gneiss with the pink areas interspersed, with parker ferro-magnesium rich zones. In thin section however, the rock can be seen to be completely recrystallised to a polygonal mosaic with a foliation defined by biotite. Relict microcline crystals appear as isolated augen (Plate 210). This concentration of strain in the adamellite probably took place because previous to the mylonitisation the rock had a relatively isotropic fabric in contrast to the enclosing schists. This may have led to a concentration of strain within the adamellite since in the initial stages of the deformation it was not so able to yield by ductile flow.

The other minerals in these rocks behave in a variety of different ways. The early garnet porphyroblasts appear to be totally unaffected by the deformation. Obviously their optical isotropism does not allow evidence of strain to show, but certainly there is no evidence of granulation. The fine-grained mylonitic fabric continues right to the margins of the garnet, the internal fabric remains coarsely crystalline, though there is evidence of slight strain extinction. Thus, as far as can be ascertained the garnet behaves as an incompressible kernel in the rock.

Plate 207. Sub-grains developing at the margins of the
           lensoid grains illustrated in Plate 205.
           Eidvaagtinn adamellite. x 120, X Pls.
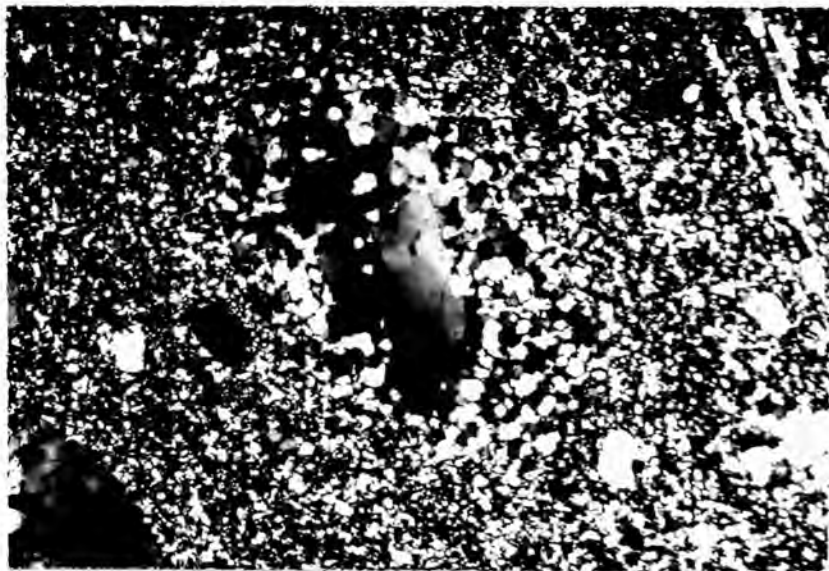
Plate 208. Felspar porphyroblast recrystallisation after
           mylonitisation. Olderbugten schists, Vargsund.
           x 40, X Pls.
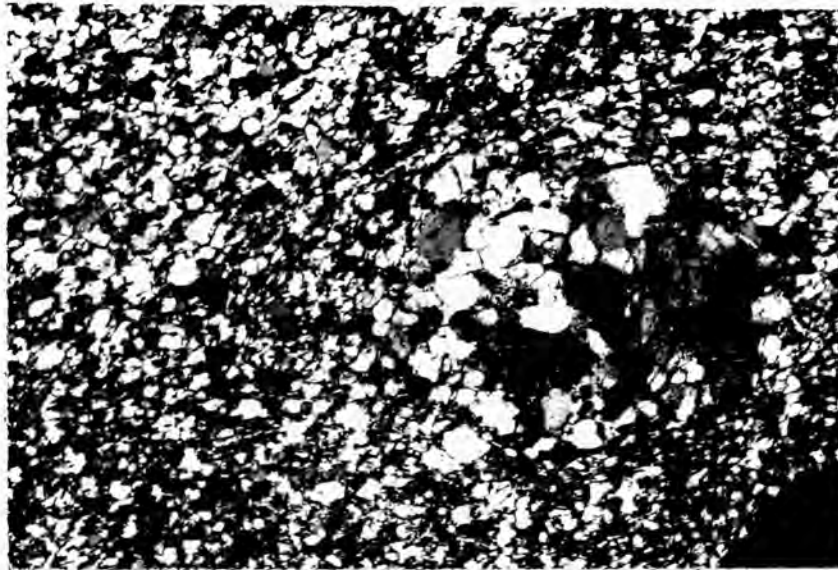
Plate 209. Relict felspar porphyroblast showing recrystallised
fabric. Olderbugten schists, south-west of Hønsebyvann.
x 40, X Pls.
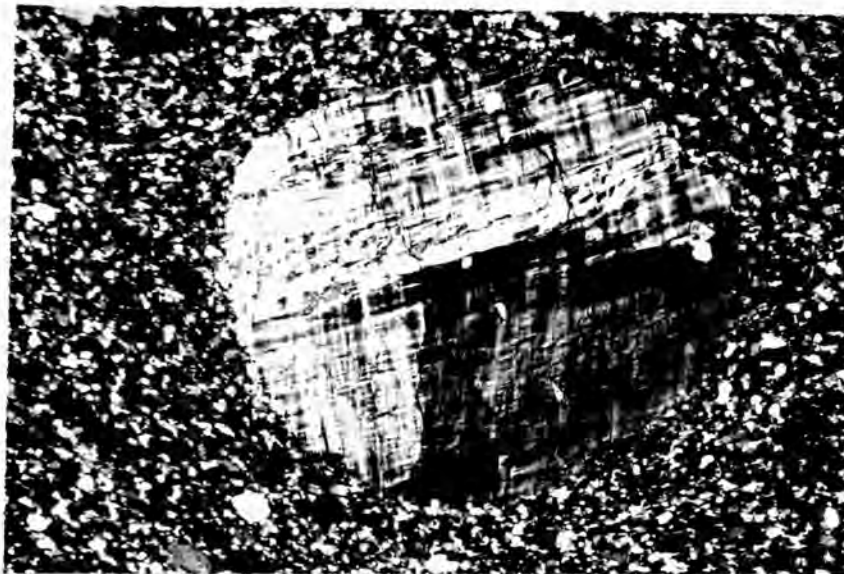
Plate 210. Microcline porphyroblast in recrystallised
groundmass. Eidvaagtinn adamellite. x 40, X Pls.

The aluminosilicate grains tend to break down into a series of optically discontinuous sub-grains (see Plate 160 and discussion). The original continuity of these crystals can sometimes be seen by the euhedral zones within which the sub-grains occur. Generally these crystals are strongly augened in the mylonitic banding.

The occurrence of fibrolite within these rocks provides an interesting clue to their metamorphic relationships. The mineral can be seen to be nucleating at a number of sites.

(1) As overgrowths on kyanite and sillimanite (Plate 211).

(2) At the boundaries of sub-grains produced by the breakdown of both the plagioclase and alkali felspar (Plates 212 and 213).

(3) At garnet-felspar grain boundaries (Plate 214).

The fibres are usually randomly arranged across the boundary between grains. A clue to the metamorphic relationships of this mineral is provided by one thin section which shows Garnet II overgrowing the fibrolite. It is thus possible to say that the mineral nucleated on the sub-grain boundaries formed during mylonitisation and is post-dated by Garnet II. Therefore, both the kyanite, sillimanite and fibrolite were not mineralogically unstable during the mylonitisation (Fig. 52).

A number of authors have described similar occurrences of fibrolite, e.g. Theodore (1970) in a mylonite belt in Southern California and Sturt (1970) records fibrolite with very similar grain boundary relationships relative to felspar from the aerole of the syntectonic hasvik gabbro. He explains the fibrolite as being due to exsolution of an excess of Al_2O_3 and S_1O_2 over the formula requirements of both

Plate 211.   Fibrolite nucleating in kyanite crystals.
Olderbugten schists.   x 160, P.P.L.
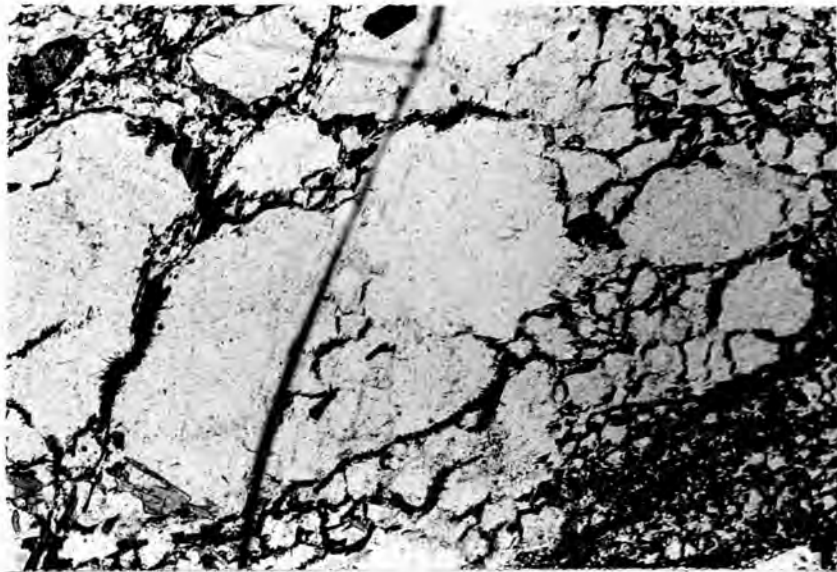
Plate 212.   Dense sheaves of fibrolite nucleating at the
grain boundaries of plagioclase.   Olderbugten
schists.   x 40, P.P.L.
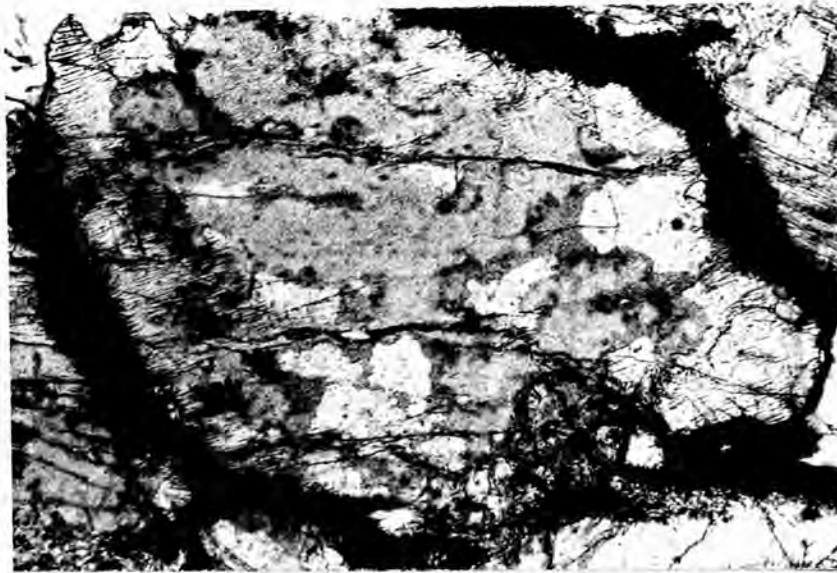
Plate 213. Dense sheaves of fibrolite nucleating on potash felspar grain boundaries. Olderbugten schists. x 40, P.P.L.
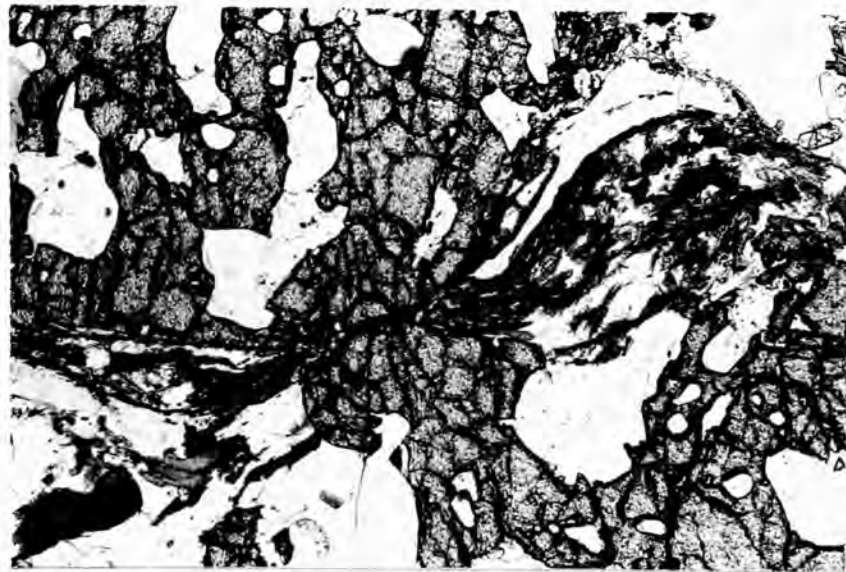
Plate 214. Fibrolite developing in the zone of compression between two garnets that have been subjected to flattening. Olderbugten schists. x 40, P.P.L.

felspar and garnet. He goes on to describe the control exerted over the sites where exsolution occurs, by the local thermal strain environment. Briefly, he concludes that exsolution will occur from a mineral, if the exsolved phase has a lower density than the host and the latter is capable of expanding relative to the enclosing minerals, due to its larger coefficient of expansion. Similarly exsolution will also occur from a mineral which has a lower density than the exsolved phase if the host is under relative compression.

It has already been noted that fibrolite nucleates at the boundaries of sub-grains formed during the mylonitisation; these are zones of high strain energy. Similarly the fibrolite seen in Plate 214 occurs in a zone of strong compression caused by the flattening of the relatively rigid garnet. Fibrolite also tends to nucleate around the kyanite and sillimanite which also behaved in a relatively rigid fashion. It seems likely, therefore, that exsolution of fibrolite may be favoured by both thermal strain as suggested by Sturt, and by tectonically induced strains acting on minerals with different compressibilities.

It is difficult to ascertain how biotite reacts, since the mineral is almost all recrystallised. All the mylonitic rocks have a foliation defined by biotites which would suggest that the recrystallisation described above took place under some sort of directed stress.

The quartz in the rocks appears to react in two ways :

(1) In the rather felspathic areas the mineral breaks down with the felspar and recrystallises into the typical polygonal mosaic.

(2) In the more quartz-rich, felspar-poor areas it takes the form of lenticles and stringers which are themselves made up of a number of elongate sub-grains.

(B) 1. Summary

The textural relationships described above indicate a phase of intense deformation preceded by growth of porphyroblastic kyanite and sillimanite and by granitisation and migmatisation. The aluminosilicates do not seem to have undergone any diapthoretic changes during the derformation. Fibrolitic sillimanite tend to crystallise possibly as an exsolution product, along the sub-grain boundaries produced by the deformation. The deformation texture subsequently underwent recrystallisation and grain growth, which tended to produce a polygonal mosaic studded with relict porphyroblasts. A foliation marked by the mylonitic banding and by preferred orientation of biotite suggests that recrystallisation took place under directed stress. The relationship observed in the aluminosilicate phases suggest that deformation and recrystallisation took place under upper amphibolite facies conditions. Finally, the recrystallised texture is overgrown by the second phase of garnet growth.

2. Discussion

A number of other authors have recorded textures which are very similar to thos described above. Theodore (1970) in particular, in a mylonite belt in Southern California, describes some very similar relationships for fibrolite. His photographs of deformed quartzo-felspathic mosaics are very similar to those occurring in rocks from Eastern Seiland. He concludes that mylonitisation took place within the stability field of sillimanite and that the deformation was probably related to "a rise to higher crustal levels of magma deep within the Southern Californian Batholith - magma not necessarily restricted spatially to the mylonite zone itself".

Sutton and Watson (1959) in a description of mylonite belts in Tanganyika note the occurrence in thin section of two textural types :

(1) Textures which are typically cataclastic "certain minerals are more or less completely granulated while others more resistent, escape all but marginal granulation. These remain as porphyroblasts in a streaked-out laminated and microbrecciated matrix whose composition varies with that of the original rock".

(2) This group varies from types where the matrix is only a little coarser than the powdery groundmass, to varieties whose average grain-size is about 0.5 mm. The finer-grained varieties have a texture resembling a fine hornfels made up of a mosaic of quartz, felspar, biotite and epidote. The larger mineral species embedded in this groundmass are the same as the porphyroblasts in the mylonites (1) above. In many of the rocks without cataclastic texture some of the grains are enclosed in a shell of the same mineral growing in optical continuity, the garnet commonly occurs in this way. Sutton and Watson points to this as evidence of regrowth after mylonitisation, they also point out that all the rocks of the mylonite belts are remarkably clean looking. The products of dislocation metamorphism are absent. They conclude "the prevalence of this almost hornfelsic texture together with the fact that the large mineral grains appear to be built up around ovoid cores resembling the typical mylonites suggest that a period of mechanical granulation may have been followed by recrystallisation whose final stages took place under more or less static conditions. This phase of crystallisation did not affect any rocks outside the mylonite belts and it seems to have been very closely connected with the formation of

these belts; it may perhaps have been caused by heat generated during the movement". They believe that these steeply dipping mylonitic zones are conjugate shears resulting from the operation of a roughly east-west trending maximum compressive stress, producing transcurrent movements at depth.

Similarly, Sturt (1969) records, in a series of small wrench-faults cutting the hornfelses of the syntectonic Hasvik gabbro, evidence of cataclasis and subsequent grain growth under amphibolite facies conditions; comminuted garnets display later overgrowths and some examples of fibrolite growing within the recrystallising fabric have been noted. He relates the wrench faults again to a set of conjugate shears.

From this brief survey it is clear that similar textures to those developed in Eastern Seiland can be achieved by translative movements, as in the case of Sturt and Sutton and Watson. Theodore, however, does not commit himself about the nature of the movements in Southern California, though he seems to be in doubt as to whether the process of simple shear is an adequate explanation (p. 435).

For a number of reasons which will now be discussed it is believed that the process of simple shear is inadequate to explain the texture developed in Eastern Seiland. In this context the experimental work of Carter (et.al) 1964, is of considerable relevance. They have produced syntectonic recrystallisation in samples of flint and quartzite deformed at temperatures ranging from $900^{\circ}C$ to $1750^{\circ}C$ and confining pressures between 15-28 Kb. The textures they produce in this apparatus are identical

to those observed by Theodore (op.cit.) and the present author. They
also point out the similarity between the textures observed in the
experimentally deformed Eureka quartzite and those in a Lower Cambrian
orthoquartzite from the Moine Thrust, and go on to suggest (p. 724)
that these textures are produced purely by recrystallisation and not by
mechanical crushing.

One of the features of quartz-bearing rocks subject to this type
of compression ($\sigma_1 > \sigma_2 = \sigma_3$) is the small circle
orientation of deformation lamellae (these are usually sub-parallel
to the basal plane of quartz) about the axis of maximum compressive stress.
It will be recalled from the petrofabric section that Specimen II
showed precisely this type of orientation, (P.91.) with the axis of the
small circles at a high angle to (S.1.). It was suggested that this
orientation had been produced pre-F.2. since the axes of the small
circles bore the same relationship to (S.1.) on both limbs of the fold.
This specimen is a quartzite from the Upper Komagnes Group occurring
very near the Eidvågeid Schist Group boundary. It is suggested that the
deforming stresses that produced the mylonitic fabric in the higher
metasedimentary groups were also the cause of this pre-F.2.
orientation of the quartz fabric. If this is the case these stresses,
were by analogy with the experimental work, of the pure flattening type.
There are two other lines of evidence which suggest that this is the case :

(1) At no time have lineations been observed in the strongly mylonitised
rocks. Sections cut parallel to the foliation in the Eidvaagtinn
adamellite reveal no trace of the preferred orientation of either the
relict porphyroblasts or the comminuted trails. If there had been some
translative movements a lineation would almost certainly have been formed.

(2) In the section on boudinage it was noted that many of the boudinaged basic sheets within the schist groups were sub-circular to elliptical, when viewed in the plane of the schistosity. It was also noted that there was no overall preferred orientation of boudin long axes (Fig. 39). These features, it was suggested, were caused by two dimensional expansion in the plane of the schistosity, produced by pre-F.2. flattening.

It is deduced, therefore, from these various observations that the stresses which produced the mylonite fabric and the pre-F.2. boudinage may be represented by a deformation ellipsoid of the type k = 0 (pure flattening) with the XY plane lying in the plane of the mylonitic banding and the axis of maximum compressive stress (Z) normal to this banding. This is consistent with the suggestion of Johnson (1967) that many mylonite belts have been produced by intense flattening. Johnson regards the Moine Thrust zone as an example; he regards it as having been produced by a deformation ellipsoid of the type 0 less than or equal to k is greater than 1 modified in local domains to 1 is greater than k less than or equal to infinity, he is of the opinion that large scale translative movements were not in operation during the development of the present mylonitic fabric.

These conclusions present a certain dichotomy; it would seem that identical textures can be produced in zones of translation, with the production of physical milling (Sutton and Watson and Sturt) and also in zones where no milling has occurred. In the latter case pure flattening appears to have been the deformation regime. This is quite simply due to the fact that in both cases the observed textures are recrystallisation texture, ratherthan essentially cataclastic features. Thus, two different processes; i.e., translation and pure flattening, under very high strain rates, may produce the same end product as the result of recrystallisation at high metamorphic grades.

Origin of the Deforming Stresses

Brooks (1970) in a geophysical survey of the coastal areas of
West Finnmark, notes the presence of a positive anomaly of 100 milligals
centred under the island of Sørøy. This anamaly implies the existence
of a large zone of high density material in the crust. In his interpretation
of this zone, Brooks describes the ideas of Belousov (1966) who regards
the heat necessary for metamorphic reactions and granitisation as being
derived from basic asthenoliths which are produced in the Upper Mantle.
These bodies are considered by Belousov to force their way up through the
crust in a manner analagous to salt diapirs. As they rise through the
crust they become according to Belousov, progressively more acid by
assimilation of crustal material and are obviously responsible for strong
vertical movements within the crust.

Belousov also suggest that the so-called basalt layer beneath the
Conrad discontinuity is in fact, a zone of degranitised granulite.
Brooks (op.cit.) suggests that if such a zone became admixed with large
amounts of basic magma, of asthenolithic origin, a zone of high density
would be formed of a type consistent with the observed anomaly in West
Finnmark.

Now it is clear from the very large basic and ultrabasic plutons
of Western Seiland that a considerable body of basic magma was introduced
into the crust. Most of the plutons were, however, explaced in post-
F.2. times. It seems possible that this body of magma was, in pre-F.2.
times, rising through the crust probably in a series of pulses, as a
basic asthenolith. Such a body could very well have been responsible
for the strong flattening exemplified by the mylonite zone and could also
have been the source of heat responsible for the regional metamorphism
and granitisation. Further upward movement of the asthenolith would

also have produced the F.2. fold phase and the subsequent flattening
which exaggerated the curvature of fold axes. Finally in post-F.2.
times the plutons were emplaced and upward movement ceased. In conclusion
a fair analogy may be drawn here between the ideas outlined above and
Theodore's (1970) suggestion that his mylonite were caused by a "rise
to higher crustal levels of magma deep within the Southern Californian
Batholith".

FIG 53 SUMMARY OF STRUCTURAL AND METAMORPHIC EVENTS
SEE FIG 49

		F.1.	STATIC PHASE	MYLON-ITE	F.2.	Oblique Boudin?	F.3.	Joints Faults

LOWER AND UPPER KOMAGNES GROUP (L AND U)

Biot	U
	L
Musc	U
	L
Chlor	U
	L
Garnet	U
	L
Actin	U
	L
Hbde	U
	L
Epid	U
	L
Calcite	U
	L

OLDERBUGTEN AND OLDERFJORD GROUPS (O.G.) EIDVÅGEID SCHIST GROUP (E.G.)

Biot	OG
	EG
Musc	OG
	EG
Chlor	OG
	EG
Garnet	OG
	EG
Kyanite	OG
	EG
Sillim	OG
	EG
Fibrolite	OG
	EG
Pyrox	OG
	EG
Hbde	OG
	EG

GABBRO INTRUSION

MIGMATISATION ADAMELLITE

CHAPTER 6

Summary and Conclusions

A summary of the geological history of the area will now be given.

1) The metasedimentary rocks mapped on Eastern Seiland have a general westerly sheet dip. They have been subjected to intense extensional deformation and high grade regional metamorphism. These agencies have considerably distorted the original sedimentary thicknesses and obliterated all sedimentary structures. The sequence discussed is therefore a structural one. The rocks comprise a series of thick alternating psammitic and pelitic groups. There are some minor lime-rich horizons, especially in the structurally highest Olderfjord Group. The general facies of the rocks would seem to indicate deposition in a shallow water environment. A tentative correlation with metasedimentary sequences on other areas is suggested.

2) The earliest phase of deformation on the area (F.1.) is represented by a number of isoclinal folds. The styles of the original folds must however, have been considerably modified by subsequent deformation. The degree of modification increases markedly from east to west. Deductions concerning the original overall style of F.1., folds must there be speculative. There is however, some evidence that curvature of axial-line is a primary feature of many F.1. folds.

In the psammites, particularly of the structurally lowest groups, there occurs an intense rib-like lineation on the layering surfaces of the rocks. This is a F.1. lineation and can occasionally be seen

to parallel the axes of F.1. minor folds.

The only F.1. fold of major proportions on the area, occurs within the Eidvågeid Schist Group. It has been mapped using the massive quartzite horizons as marker bands. The schistosity in the Eidvågeid Group is sensibly parallel to the banding in the quartzites, and is regarded therefore, as an axial plane structure to F.1.

The parallelism of biotite to the axial planes of F.1. minor folds and the parallelism of actinolite crystals to F.1. fold axes indicates that the F.1. deformation took place in the Quartz-Albite-Epidote-Biotite Sub-Facies of the Greenschist facies.

The later part of the F.1. deformation was accompanied by intrusion of gabbroic sheets parallel to the axial-planes of the early folds.

3) Following F.1. a phase of static metamorphism ensued, during which the maximum grade of regional metamorphism was achieved. There appears however, to be a variation in grade with the area. The lowest grades, characterised by albitic felspar, hornblende, biotite and epidote assemblages, were achieved in the east. The highest grades, characterised by andesine, kyanite and sillimanite, were achieved in the west. A sequence of metamorphic isograds delimiting the various zones is described. The attainment of the highest grades in the west, was accompanied by migmatisation, granitisation and intrusion of adamellitic sheets. The latter are believed to be products of partial melting.

4) Following the migmatisation and granitisation, a phase of intense mylonitisation occured. This deformation phase fundamentally modified both the fabrics and the thicknesses of the metasediments.

Textural studies indicate that kyanite and sillimanite were stable
during the deformation; and the development of the mylonitic fabric
was immediately post-dated by growth of fibrolitic sillimanite.
The absence of pronounced linear structures in the mylonites suggests
that translative movements were not involved in their formation.
It is believed therefore, that the mylonitic fabrics developed in
response to intense flattening, produced by a rising basic asthenolith.
This intense flattening was also undoubtedly responsible for a con-
siderable amount of the boudinage developed on the area.

5) The majority of folds recorded on the area are attributed
to the second fold-forming deformation (F.2.) By the onset of F.2.
the metamorphic grade appears to have waved to sub-garnet grade
conditions; it is apparent that garnet porphyroblasts are always
augened in S.2. The folds have a very variable style, this variation
appears to be closely related to lithology, though there also appears
to be a more general change in style across the area. In the east,
the folds have intensely attenuated long-limbs with a number of
vertically-stacked folds in the short-limbs. In the west the limbs
of the folds tend to be of rather more equal length. The change in
style is related to the different states of competence of the rocks
at the onset of F.2.

A ubiquitous feature of F.2. folds on both the major and the
minor scale, is the curvature of their axial lines. The curvature
appears to begin very early in the fold forming process. It is however,
considerably accentuated by later flattening. The lineations associated
with these folds are of the intersection type.

6) In the Komagnes area a number of oblique boudins have been

recorded. They are closely associated with rotated tension-gashes and monoclinal folds. These structures are strongly discordant to the local F.2 trend and clearly post-date the F.2. deformation. It is believed that all these structures were formed in response to a progressive deformation sequence. This deformation sequence was also responsible for the development of the late Calledonian thrusts which outcrop on the adjacent mainland.

7) In the north of the area there is a pronounced swing in strike. This swing is due to rotation in a west-east sense of the Olderfjord Group, using the Olderbugten Group as a plane of decollement. A number of large folds developed in association with these movements. They have been designated F.3.

F.3 folds have been recorded in the Komagnes area, they are monoclinal warps. There temporal relationship to the F.3. folds in the north, is not known.

8) The final phase of movement in the area led to the development of joints, faults and locally, kink-folds. The prevailing metamorphic grade during this phase of deformation, and indeed all the phases subsequent to F.2. was the Quartz-Albite-Muscovite-Chlorite Sub-facies of the Greenschist Facies.

Discussion

One of the most striking features of the geology of EasternSeiland is the intense extension which the rocks appear to have undergone. Much of this extension has been produced by the intense flattening responsible for the development of the mylonites. This deformation manifests itself on the major scale, in the lensoid aspect of many of the lithological groups, particularly the gabbros in the west. On the

minor scale boudins appear to have no axis of elongation. Three-dimensional exposures often reveal that individual pods are discoidal in shape; extreme stretching is suggested by the absence of contigous pods.

Further evidence of this extension is provided by the attenuation of the metamorphic zones. These zones represent a contemporaneous lateral change in metamorphic grade from the middle Greenschist Facies to the upper Amphibolite Facies. This transition is accomplished in a vertical thickness of rock of approximately 2,700 metres. An analaoug transition is accomplished in the Dalradians of north-eastern Scotland over a distance of approximately 10 miles. (Winkler 1967), This distance however, represents a traverse across the metamorphic zones at their thinnest. Clearly, if the zones in Seiland where of comparable thickness, subsequent extreme attenuation is implied.

Comparisons of this nature are clearly open to question and although considerable attenuation of the zones has obviously occured, there may be other factors which have contributed to their relative thinness.

The Falkenes Marble Group, one of the youngest members of the Sørøy succession, has been assigned on palaeontological grounds to the upper Lower Cambrian or the lower Middle Cambrian (Holland and Stuart 1970). The Hellefjord Schist Group, the youngest member of the succession of Sørøy shows evidence of deposition under turbidite conditions. Roberts (1968) claims that this formation represents the deepening of the sedimentary basin heralding the oncoming orogeny.

Isotopic age dates indicate that the main structural and meta-morphic events in the region took place in upper Cambrian to lower

Ordivician times. This means that the sediments on Sørøy could not
have been buried under any considerable overburden before the onset
of the orogeny. This implies that the structural and metamorphic
events took place at a fairly high level in the crust. Such a high
level environment could well be locally favourable to the development
of very steep thermal gradients. It is possible that the attenuated
metamorphic zones on Seiland, reflect a t least in part, such a steep
thermal gradient.

Regional Setting

Roberts (1968) notes the occurrence on Sørøy of N-S and E-W
strike belts. In the N-S belts, F.2. folds have a monoclinic symmetry
and a sense of transport to the east, away from the central parts of
the orogen. He regards these structures as being folds in 'b'
(where 'b' is parallel to the margins of the orogen). The elongation
of boudins parallel to 'b' indicates that this direction was an axis
of stretching during the F.2. folding. Similarly F.2. folds show a
strong point maximum distribution on stereograms.

In the E-W strike belts the folds have an orthorhombic symmetry
and are regarded as being folds in 'a' in the regional sense.

Lindström (1955, 1957, 1958), records a similar picture from an
area in Swedish Lapland. This area represents the marginal thrust
zone of the orogenic belt. He records a N-S or NE-SW set of folds
having a constant sense of overturning towards the E or S-E. Another
transverse set of folds are overturned to the N-S or S-W, with no
clear preference for either direction. The first set are regarded
as folds in 'b' and the second set folds in 'a'. The distribution
of these two sets is however, not confined to strike belts, but appears
to be random.

Rutland (1959) working in the Nodland area of central Norway notes the absence of pronounced linear structures in 'a'. Roberts (1968) has suggested therefore that Lindstrom's area represents the marginal thrust zone of the orogenic belt and Rutland's, the central part. He suggests that Sørøy contains elements of the structural pattern of both areas.

Seiland does not appear to fit into this picture. The outcrop of the late Caledonian thrusts on the mainland adjacent to Seiland, implies that these structures exist at no great depth beneath the island, in this respect Seiland is analagous to Lindstrom's area. The structural picture on Seiland is however, distinctively different. The F.2. folds are strictly speaking, folds in neither 'a' or 'b' since they appear to have arisen by pure flattening. This is indicated by such features as curvature of fold hinge-lines and absence of elongation of boudins. It is this predominance of deformation by pure flattening that is the most distinctive feature of the geology of Seiland. The author feels that this is very closely related to the uprise of the basic asthenolith, later responsible for intrusion of the plutons of Western Seiland. This phenomenon has exerted a fundamental control over the structural and metamorphic history of the area.

<u>Appendix</u>

Due to shortage of space the basic rocks on Eastern Seiland are included as an appendix. They are at present being studied in detail by J.B. Jackson of Bedford College.

(A) <u>The Hønseby Gabbro</u>

This was intruded as a concordant sheet after the F.1. folding, evidence of the continued presence of deforming stresses is, however, provided by the presence of a foliation in the gabbro which is parallel to the margins of the sheet, the foliation in the enclosing metasediments is also parallel to the banding which contains tight folds of F.1. age. There is no evidence of lineation.

A number of thin screens of psammite can be found within the gabbro as well as thin calc-silicate lenses. The thickest of these screens is seen on the western side of Hønsebyfjord.

The gabbro itself has been folded by F.2., as witnessed by the large fold that can be mapped in the contact on the eastern side of Hønsebyfjord, Plate 215 shows foliated Hønseby gabbro deformed by a tight F.2. fold. The large monoclinal structures in the contact and the psammite screens also indicate deformation by F.3.

In the field the gabbro is rather variable in facies occasionally it is mafic and flaggy as in Plate 215. Elsewhere it is more leucocratic. It is sporadically layered with dark chocolate brown mafic layers alternating with rather more leucocratic layers. The gabbro is frequently garnetiferous particuarly in the mafic layers. Some of the garnets are of golf-ball proportions. The foliation is generally parallel to the layering.

Plate 215. F$_2$ fold deforming foliated Hønseby gabbro.
Eastern side of Hønsebyfjord.

Plate 216. Anatectic veins in the contact zone of the
Hønseby gabbro. Northern contact, eastern side
of Hønsebyfjord.

Petrography

No traces of any original igneous texture are visible in the gabbro. The felspar has recrystallised completely to a granular often polygonised mosaic. Michel Levy determinations on albite twins yields a value of An_{71}. The gabbro contains two pyroxenes :

(1) A green clinopyrozenes with the following optical properties :

Opt   +ve   2V approx. $= 70^{\circ}$

$\hat{ZC}$ $= 44^{\circ}$

Birefringence 0.034

According to J.B. Jackson (personal communication) who has analysed this pyroxene, it is an augite.

(2) A pleochroic orthorhombic pyroxene (hypersthene)

Opt - ve   2V approx. $= 60^{\circ}$

Straight extinction

Birefringence 0.011

X = pink

Z = pale green

These pyroxenes occur together in elongate clots which define the foliation in the rock. There is no lineation of prismatic crystals. The clots may be formed of a number of single crystals of pyroxene or they may consist of an aggregate of polygonised grains. They are generally surrounded by a selvage of green amphibole which often contains vermicules of quartz. This clearly indicates that the amphibole is a reaction product after the pyroxene. The amphibole has the following properties :

Opt -ve   2V approx. = 50°

$\hat{ZC}$ = 18°

Birefringence = 0.022

Pleochroic scheme X = light yellow brown

                   Y = dark lincoln green

                   Z = lincoln green

Z less than or equal to Y is greater than X

The amphibole itself in places appears to be altering to a red-brown biotite which is again occasionally vermicular, the final product of the metamorphism appears to be granular garnet, which clearly over-grow the earlier-formed biotite and amphibole.

The Hornfelses

Later recrystallisation has entirely obliterated all traces of hornfelsic texture in the adjacent metasediments. There is, however, clear evidence that the intrusion of the gabbro caused mobilisation of metasediment in the immediate contact zone. Plate 216 shows a calc-silicate band at the northerly contact of gabbro on the eastern shore of Hønsebyfjord. The rock is shot-through by a large number of leucrocratic veins which are undoubtedly of anatectic origin. These consist of stringers and blebs of quartz which often show inclusions of rutile in the Widmanstatten figure. There are two felspars, one a highly or granular plagioclase with composition (An_{65}) and the other an alkali felspar, myrmetite develops at the junction between the two. Also occurring within the veins is a granular orthorhombic pyroxene which is often intimately bassociated with vermicular biotite, apparently in reaction relationship to the pyroxene. Finally the biotite is overgrown by a granular garnet. The body of the rock into which these veins penetrate, consists of diopsidic pyroxene, orthorhombic pyroxene and labradorite intermixed in a granular aggregate.

This assemblage Diopside-Orthopyroxene-Anorthitil felspar would seem to indicate that the metasediments adjacent to the gabbro underwent metamorphism under the Orthopyroxene sub-facies of the K-felspar Corderite Hornfels Facies, Winkler (1967).

(B)   The Hammeren Gabbro

This gabbro mass is rather different in facies and mode of intrusion from the Honseby gabbro. It has not been possible to ascertain the age relationships between the two masses, but B. Robins (personal communication) has evidence that the gabbro was intruded just after the F.1. folding, i.e. at about the same time as the Hønseby gabbro. In the field the gabbro is very fine-grained and is generally interlayered with the psammite of the Olderfjord Group. This interlaying occurs on all scales from the scale of a single exposure (Plate 217), up to the scale of the map, where it is apparent that the gabbro contains a large number of psammite screens.

Foye (1916) has proposed the term stromatolith for this type of interlayering of sedimentary and igneous material in sill-like relationships. Speedyman records (1968) the same situation on Sørøy in relation to the late F.1. Eusfjord gabbro and the later Havnefjord Diorite. He attributes it to permissive intrusion of magma along layering subject to compression parallel to its length. Whether or not this explanation can be applied to Seiland is difficult to say, the lateral continuity of the psammite bands, however, testifies to very permissive intrusion.

The stromatolithic complex has subsequently been intensely boudinaged during the flattening accompanying mylonitisation and then folded by F.2. (Plate 218).

Plate 217. Psammites of the Olderfjord Group interlayered
with Hammeren gabbro. North of Olderfjord.

Plate 218. F_2 fold deforming interlayered psammites of the
Olderfjord Group and Hammeren gabbro.

Petrography

As well as the different intrusive relationships, the gabbro is also petrographically different from the Hønseby gabbro. In the field it is generally rather fine-grained and chocolate in colour whereas the Hønseby gabbro tends to be rather coarse and buff in colour. In thin section the Hammerengabbro is rather more amphibolitised, and the amphibole is not the green horneblende seen in the Hønseby gabbro. Rather it is a member of the cummingtonite-grunerite series. The optical properties are :

$$Opt \quad -ve \quad\quad 2V \text{ approx.} = 80\text{-}90^{\circ}$$

$$\widehat{ZC} = 19^{\circ} \quad X = b$$

Birefringence 0.038

Z = very pale apple green

Y = very pale apple green

X = very pale yellow

Z = Y greater than X

The negative sign indicates that it is gruneritic. The mineral occurs in quite large (2.1/2 mm) crystals which show evidence of strain extinction. The pccasional presence of vermicles of quartz suggest that the mineral is an alteration product after pyroxene, with which it is intimately associated. There are again two pyroxenes present, a clinop;roxene with the following properties :

$$Opt \quad +ve \quad\quad 2V \text{ approx.} = 70^{\circ}$$

$$\widehat{XC} = 44^{\circ}$$

Birefringence $= 0.035$

and an orthopyroxene

$$- ve \quad 2V = 60^{\circ}$$

Straight extinction

Birefringence 0.017

The pyroxenes occur intimately mixed in a granular aggregate, frequently rimmed by the amphibole.

The felspar is a granular aggregate. Michel Levys determinations yielded a composition of (An_{71}), twin lamellae are frequently bent, there has been a considerable amount of recrystallisation.

Finally biotite occurs as descussate flakes which appear to be forming by breakdown of the amphibole.

The author has been unable to recognise any certain indications of a hornfelsic texture in the immediate environs of the Hammeren gabbro. This is considered to be the result of the pervasive recrystallisation.

Discussion

In the introduction it was stated that workers such as Barth, Heier, Oosterom and Krauskopf believed that many of the early gabbro-gneisses seen on the adjacent island of Stjernøy and on Western Seiland were the result of the ;metamorphism and anatectic transformation of a series of basic lavas and metasediments. This type of explanation was invoked to explain such features as the presence of metasedimentary layers parallel to the banding in the igneous rocks, and the general gradational contact of the complex with the enclosing metasediments.

Study of the Hønseby and Hammeren gabbros, which have many features in common with the gabbro-gneisses on Stjernøy and are almost certainly their temporal equivalents, reveals that such bodies can be formed by purely magmatic processes as evidenced by the mobilised hornfelses. This

conclusion is in agreement with the findings of Speedyman (1968) with regard to the Husfjord gabbro and the later Havnefjord diorite.

(C) The Ultra Basic Lenses

Within the Komagnes Group there are four lense-shaped outcrops of a very dense, dark green rock. Two of these lenses occur on the shore of Southern Vasbugt; along the strike from these outcrops is a third lens, above Komagnes. The fourth occurs as a sheet across the neck of the peninsula at Komagnes. These bodies appear to be concordant with the layering of the psammites. They are generally sheared at the margins though away from this zone the rock appears to have almost no planar structure at all. Some facies are dotted with black segregations of iron ore.

Petrography

This section examination indicates that they have undergone a rather complex series of metamorphic events. The earliest formed mineral phases in these rocks appear to be two series of amphibole :

(a)   Opt + ve $2V = 50\text{-}60^{\circ}$

$\widehat{ZC} = 19^{\circ}$ $Y = b$

Birefringence 0.045

Non-pleochroic

This amphibole belongs to the cummingtonite-grunerite series.

(b)   Opt - ve $2V$ $70\text{-}80^{\circ}$

$\widehat{ZC}$ unable to measure $Y = b$

Non-pleochroic

This amphibole is tremolitic.

Both amphiboles appears to be breaking down to chlorite.
Possibly as follows :

Tremolite   --> calcite + magnetite + MgFe chlorite

Gruneritic amphibole --> FeMg chlorite

The groundmass of the rocks is speckled with calcite and iron ore.

The chlorite itself is completely decussate and appears to be
interfingering along amphibole cleavages.

The properties are :

Opt -ve   2V approx. = 20°

Pleochroic X = pale green

Z = buff   Z greater than X

The best fit for these properties is ripidolite, an FeMg chlorite,
Winchell (1951, Vol II, p.383).

The final product of the metamprphism is talc, which appears to be
growing as small anhedra at the expense of the chlorite. The mineral
shows straight extinction and high polarisation colours.

The assemblage is characteristic of metamorphism of an ultrabsic
rock under Middle Greenschist Facies conditions. The form of these bodies
of rock strongly resembles the Alpine peridotites and serpentinites
described in Turner and Verhoogen (1960). They note (p.310) the
constant association of the Alpine peridotite with geosynclinal
sediments and say that many writers have considered that the bodies have
been emplaced "during the earliest stages of folding which terminates
sedimentation and ushers in orogeny". This may well be true of the

Seiland bodies which have obviously been strongly boudinaged and
metamorphosed. No more definite evidence of their chronology is
available.

REFERENCES

Althaus, E.   The Triple Point Andalusite-Sillimanite-Kyanite. An Experimental Study. Contr. to Min. and Pet., no.16, 1967, pp.29-44.

_____   Experimental Evidence that the Reaction of Kyanite to Form Sillimanite is a Least Bivariant. Amer.J.Sci., vol.267, 1969, pp.273-277.

Anderson, E.M.   On Lineation and Petrofabric Structure and the Shearing Movement by which they have been produced. Q.J.G.S., vol.104, 1948, pp.99-132.

Ash, R.P.   The Geology of Skjervøy, North Troms, Norway. N.G.U. Årbok, Nr. 255, 1967.

Atherton, M.P.   The Composition of Garnet in Regionally Metamorphosed Rocks. Controls of Metamorphism, Oliver and Boyd, 1965, pp.281-290.

Baëta, R.D. and Ashbee, K.H.G.   Slip Systems in Quartz: I Experiments. Amer.Miner., vol.54, 1969, pp.1551-1573.

Bailey, S.W., Bell, R.A. and Peng, C.J.   Plastic Deformation of Quartz in Nature. Bull.Geol.Soc.Amer., vol.69, 1958, pp.1443-1466.

Ball, T.K., Gunn, C.B., Hooper, R.R. and Lewis, D.   A Preliminary Geological Survey of the Loppen District, West Finnmark. N.G.T., vol.43, pt.2, 1963, pp.215-246.

Barth, T.F.W.,   The Layered Gabbro Series at Seiland, N. Norway. N.G.U., Nr. 184, 1953, pp.191-200.

Belousov, V.V.   Modern concepts of the structure and development of the earth's crust. Q.J.G.S., vol.122, 1966, pp.293-314.

Bloss, F.D.   Anisotropy of Fracture in Quartz. Amer.J.Sci., vol.255, 1957, pp.214-225.

Brooks, M.   A Gravity Survey of the Coastal Areas of W. Finnmark, Northern Norway. Q.J.G.S., vol.125, 1970, pp.171-192.

Burke, J.E.   Grain Growth in Ceramics. Kinetics of High Temperature Processes, ed. W.D. Kingery, Tech.Press, M.I.T., John Wiley, 1959.

Campbell, J.D.   En Echelon Folding. Econ.Geol., vol.53, 1958, pp.440-472.

Carter, N.L., Christie, J.M. and Griggs, D.T.   Experimental Deformation and Recrystallisation of Quartz. Journ.Geol., vol.72, 1964, pp.587-733.

Coe, K.   Boudinage Structure in West Cork, Ireland. Geol.Mag., vol.96, 1959, pp.191-200.

Currie, J.B., Patnode, H.W. and Trump, R.P.   Development of Folds in Sedimentary Strata, Bull.Geol.Soc.Amer., vol.73, 1962, pp.655-673.

Dalziell, I.W.D. and Bailey, S.W.   Deformed Garnets in a Mylonitic Rock from the Grenville Front and their Tectonic Significance. Amer.J.Sci., vol.266, 1968, no.7.

Dearman, W.R. and Freshney, E.C.   Repeated Folding at Boscastle, North
        Cornwall, England. Proc.Geol.Assoc., vol.77, pt.2, 1966,
        pp.199-215.

_____, Shiells, K.A.G. and Larwood, G.P. Refolded Folds in the
        Silurian Rocks of Eyemouth, Berwickshire. Proc.Yorks.
        Geol.Soc., vol.33, 1962, pp.273-286.

Deer, W.A., Howie, R.A. and Zussman, J.   An Introduction to the Rock
        Forming Minerals. Longmans, 1967.

De Sitter, L.U.   Structural Geology. McGraw-Hill, 1964.

Den Tex, E. and Floor, P.   A Blastomylonitic and Polymetamorphic Graben
        in W. Galicia, N.W. Spain. Etages Tectoniques, Neuchatel
        University Colloquium, 18-21 April, 1966.

De Waard, D.   Anorthite Content of Plagioclase in Basic and Pelitic
        Crystalline Schists as related to Metamorphic Zones in the
        Usu Massif, Timor. Amer.J.Sci., vol.257,1959,pp.553-562.

Dewey, J.F.   Nature and Origin of Kink Bands. Tectonophysics, vol.1,
        1965, pp.459-494.

_____   Kink Bands in Lower Carboniferous Slates of Rush, Co.
        Dublin. Geol.Mag., vol.103, 1966, pp.138-142.

Dhonau, T.J.   The Relation between the Quartz Fabric and Fold Movements
        in a Folded Calc-silicate Rock from the Moines of Scotland.
        Geol.Mag., vol.98, 1961, pp.313-316.

Donath, F.A. and Parker, R.B.   Folds and Folding. Bull.Geol.Soc.Amer.,
        vol.75, 1964, pp.45-62.

Fairbairn, H.W.   Elongation in Deformed Rocks. Journ.Geol., vol.44,
        1936, pp.670-680.

_____   Hypotheses of Quartz Orientation in Tectonites.
        Bull. Geol.Soc.Amer., vol.50, 1939, pp.1475-1491.

_____   Structural Petrology of Deformed Rocks. Addison Wesley
        Press Inc., Camb.Mass., 1949.

Flinn, D.   A Tectonic Analysis of the Muness Phyllite Block of Unst
        and Uyen, Shetland. Geol.Mag.,vol.89,1952,pp.263-272.

_____   On Folding During Three Dimensional Progressive Deform-
        ation. Q.J.G.S., vol.118, 1962, pp.385-433.

Foye, W.G.   Are the "Batholiths" of the Haliburton Bancroft Area,
        Ontario, correctly named? Journ.Geol., vol.24, 1916,
        pp.783-791.

Gangopadhyay, P.K. and Johnson, M.R.W.   A Study of Quartz Orientation and
        its Relation to Movement in Shear Folds. Geol.Mag.,
        vol.99, 1962, pp.69-84.

Goldschmidt, V.M.   Die Injectionsmetamorphose im Stavanger Gebiete.
        Vidensk.Skrift 1, Mar-Nat.K1. No.10, 1921.

Griggs, D. and Bell, J.F. Experiments Bearing on the Orientation of
Quartz in Deformed Rocks. Bull.Geol.Soc.Amer.,
vol.49, 1938, no.10-12, pp.1723-1746.

Hansen, E. Slip Folding. Carnegie Inst. Geophys.Lab. Year Book,
published 1967, pp.536-539.

Harker, A. Metamorphism. Methuen, 1939.

Harland, W.B. Tectonic Facies, Orientation, Sequence, Style and Date.
Geol.Mag., vol.93, 1956, pp.111-120.

Harris, A.L. and Rast, N. Oriented Quartz Inclusions in Garnet.
Nature, vol.185, 1960, p.448.

Hartshorne, N.H., and Stuart, A. Practical Optical Crystallography 1969.
Edward Arnold Ltd.

Heier, K.S. Layered Gabbro, Hornblendite, Carbonatite and Nepheline
Syenite on Stjernøy, North Norway. N.G.T., 41, 1961,
pp.109-155.

Hietanen, A. Kyanite, Andalusite and Sillimanite in the Schists in
Boehls Butte Quadrangle, Idaho. Amer.Miner., vol.41,
no.1, 1956, pp.1-27.

Higazy, R.A. The Trace Elements of the Plutonic Complex of Loch Doon
(Southern Scotland) and their Petrogenetic Significance.
Journ.Geol., vol.42, no.2, 1954, pp.172-181.

Hills, E.S. Examples of the Interpretation of Folding. Journ.Geol.,
vol.53, 1945, pp.47-57.

Hobbs, B.E. Microfabric of Tectonites from the Wyangala Dam area, New
South Wales, Australia. Bull.Geol.Soc.Amer., vol.77,
pp.685-706.

Holland, C.H. and Sturt, B.A. On the Occurrence of Archaeocyathids in
the Caledonian Metamorphic Rocks of Sørøy and their
Stratigraphic Significance. N.G.T., vol.50, 1970,
pp.341-355.

Holtedahl, O. (ed.) The Geology of Norway. N.G.U., Nr. 208, 1960.

Hooper, P.R. The 'A' Lineation and the Trend of the Caledonides of
N.Norway. N.G.T., vol.48,1968, nr.4, pp.261-268.

_____ and Gronow, C.W. The regional significance of the Caledonian
structures of the Sandland Peninsula, West. Finnmark,
northern Norway. Q.J.G.S., vol.125, 1970, pp.193-217.

Jenkins, G. Deformation Features in Cataclastic Quartz. Geol.Mag.,
vol.97, 1960, pp.70-84.

Johnson, M.R.W. Conjugate Fold Systems in the Moine Thrust Zone.
Geol.Mag., vol.93, 1956, pp.345-350.

_____ Mylonite Zones and Mylonite Banding. Nature, vol.213,
1967, pp.246-247.

Kerr, P.F.   Optical Mineralogy. McGraw-Hill, 1959.

King, B.C.   The Nature and Origin of Migmatites: Metasomatism or
                    Anatexis. Controls of Metamorphism, Oliver and Boyd,
                    1965.

Kingery, W.D.   Introduction to Ceramics. John Wiley & Sons, 1960.

Knill, J.L.   A Classification of Cleavages. 21st Inst.Geol.Congress,
                    1960, 18, pp.317-325.

_____   Joint Drags in Mid-Argyllshire. Proc.Geol.Assoc., vol.72,
                    1961, pp.13-19.

_____ and Knill, D.C. Some Discordant Fold Structures from the
                    Dalradian of Craignish, Argyll, and Rosguill, Co.Donegal.
                    Geol.Mag., vol.95, 1958, pp.497-510.

Krauskopf, K.B.   Igneous and Metamorphic Rocks of the Øksfjord area,
                    Vest-Finnmark. N.G.U. Årbok, Nr. 188, 1953, pp.29-50.

Lindström M.   Structural Geology of a small area in the Caledonides of
                    Arctic Sweden. Lunds Univ. Årsskr., 51, 1955, NR 15.,
                    pp.1-31.

_____   Tectonics of the area between Mt. Keron and Lake Allesjaure
                    in the Caledonides of Swedish Lapland. Lunds Univ.
                    Årsskr., 53, 1957, NR 11, pp.1-33.

_____   Tectonic Transports in Three Small Areas in the Caledonides
                    of Swedish Lapland. Lunds Univ. Årsskr., 54, 1958, NR 3;
                    pp1-85.

MacColl, R.S.   Geochemical and Structural Studies in Batholithic Rocks
                    of S. California. Bull.Geol.Soc.Amer.,vol.75, pp.805-822.

Mukhopadhyay, D.   Some Aspects of Coaxial Refolding. Geologische Rundschau,
                    Band 33, Heft 3, 1966, pp.819-825.

Nicholson, R.   Eyed Folds and Interference Patterns in the Sokumfjell
                    Marble Group, N. Norway. Geol.Mag., vol.100, 1963,pp.59-68.

Oosterom, M.G.   Some notes on the Lille-Kufjord Layered Gabbro, Seiland,
                    Finnmark, N. Norway. N.G.U., Nr. 195, 1955, pp.73-87.

_____   The Ultramafite and Layered Gabbro Sequence in the
                    Granulite Facies Rocks on Stjernøy (Finnmark, Norway).
                    Leidse Geol.Medd., vol.28, 1963, p.176.

Padget, P.   The Geology of the Caledonides of the Birtavarre Region,
                    Troms, N. Norway. N.G.U., Nr. 192, 1955.

Pitcher, W.S.   The Aluminium Silicate Polymorphs. Controls of
                    Metamorphism, Oliver and Boyd, 1965, pp.327-341.

Poldervaart, A. and Hess, H.H.   Pyroxene in the Crystallisation of
                    Basaltic Magmas.   Journ. Geol., vol.59, 1951,
                    pp.472-489.

Powell, D. and Treagus, J.E.   On the Geometry of 'S-shaped' Inclusions
                    in Garnet Porphyroblasts.   Min.Mag., vol.36, no.279,
                    1967, pp.453-456.

Price, N.J.   Mechanics of Jointing in Rocks.   Geol.Mag., vol.96,
                    1959, pp.149-167.

Pringle, I.R. and Sturt, B.A.   New Evidence for the Age of the
                    Caledonian Orogeny in West Finnmark, Northern Norway.
                    N.G.T., vol.49, 1969.

Ramberg, H.   Natural and Experimental Boudinage and Pinch and
                    Swell Structure.   Journ. Geol., vol. 63, 1955,
                    pp.512-526.

_____   Strain Distribution and Geometry of Folds.   Bull.
                    Geol. Inst. Univ. Uppsala, vol.42, 1963.

_____   Evolution of Drag Folds.   Geol.Mag., Vol.100, no.2,
                    1963, pp.97-106.

Ramsay, D.M. and Sturt, B.A.   A Study of Fold Styles, the Associations
                    and Symmetry Relationships from Sørøy, N. Norway.
                    N.G.T., vol.43, 1963, pp.411-430.

_____   The Alkaline Complex of the Brevikbotn
                    area, Sørøy, N. Norway, N.G.U., Nr. 231, 1965.

Ramsay, J.G.   Geometry and Mechanism of Formation of Similar Type
                    Folds.   Journ.Geol., vol.70, 1962, pp.309-327.

_____   The Geometry of Conjugate Fold Systems.   Geol.Mag.,
                    vol.99, 1962, pp. 516-526.

_____   Folding and Fracturing of Rocks.   McGraw Hill, 1967.

_____   The Deformation of Early Linear Structures in Areas
                    of Repeated Folding.   Journ.Geol., vol.68, 1968,
                    pp.75-93.

Rast, N.   The Origins and Significance of Boudinage.   Geol.Mag.,
                    vol.93, 1956, pp.401-408.

_____   Morphology and Interpretation of Folds - A Critical
                    Essay.   Geol.J., vol.4, 1964, pp.177-185.

Rast, N.   Nucleation and Growth of Metamorphic Minerals. Controls of Metamorphism, Oliver and Boyd, 1965, pp.73-102.

_____ and Sturt, B.A. Crystallographic and Geological Factors in the Growth of Garnets from Central Perthshire. Nature, vol. 179, 1957, p.215.

Read, H.H.   The Igneous and Metamorphic History of Cromar, Deeside, Aberdeenshire. Trans.Roy.Soc.Edin., 55, 1927, Pt.II, pp.317-353.

_____   Metamorphism and Igneous Action. The Advancement of Science (Rep.Brit.Assoc. 1939-1940) 1, pp.223-251.

Reading, H.G.   Eocambrian and Lower Palaeozoic Geology of the Digermul Peninsula, Tanafjord, Finnmark. N.G.U. Nr.234, 1965, pp.167-191.

_____   and Walker, R.G. Sedimentation of Eocambrian Tillite and Associated Sediments in Finnmark, Northern Norway. Palaeogeog., Palaeochem., Palaeoecol., vol.2, 1966, pp. 177-212.

Reitan, P.H.   The Geology of the Komagfjord Tectonic Window of the Raipas Suite, Finnmark, Norway. N.G.U. nr.221, 1963.

Reynolds, D.L.   The Albite Schists of Antrim and their Petrogenetic Relationship to Caledonian Orogenesis. Proc.Roy.Irish Acad. 48B No.3, 1942, pp.43-66.

_____   The Sequence of Geochemical Changes leading to Granitization. Q.J.G.S., vol.102, 1946, pp.389-446.

_____   On the Relationship between "Fronts" of Regional Metamorphism and "Fronts" of Granitisation. Geol.Mag., vol.84, 1947, pp.106-109.

Richardson, S.W., Gilbert, M.L. and Bell, P.M. Experimental Determination of Kyanite-Andalusite and Andalusite Sillimanite Equilibria; the Aluminium Silicate Triple Point. Amer. J.Sci., vol.267, 1967, pp.259-272.

Roberts, D.   The Structure and Metamorphic Petrology of the Eo-Cambrian Rocks of the Langstrand Area, Sørøy, N. Norway. Unpublished Ph.D. thesis, 1965 (University of London).

_____   Hellefjord Schist Group - A probable Turbidite Formation from the Cambrian of Sørøy, W. Finnmark. N.G.T., vol.48, no.4, 1968.

_____   The Structural and Metamorphic History of the Langstrand-Finfjord Area, Sørøy, Northern Norway. N.G.U., Nr. 253, 1968.

Schmidt, W.   Untersuchungen über die Regelung des Quartz Gefüges etc. Fort.Miner.Krist.Pet., 11, 29.

Shainin, V.E.   Conjugate Sets of En Echelon Tension Fractures in the Athens Limestone at Riverton, Virginia. Bull.Geol.Soc. Amer., vol.61, 1950, pp.509-517.

Sherbon Hills, E.   Elements of Structural Geology. Methuen, 1966.

Skevington, D. and Sturt, B.A.   Faunal Evidence bearing on the Age of Late Cambrian to Early Ordovician Metamorphism in Britain and Norway. Nature, vol.215, pp.608-609.

Soloman, M. and Hill, P.A.   Rib and Hackle Marks on Joint Faces at Renison Bell, Tasmania: a Preliminary Note. Journ.Geol., vol.70, 1962, pp.493-496.

Speedyman, D.L.   Igneous and Metamorphic Geology of the Husfjord Area, Sørøy, Northern Norway. Unpublished Ph.D. thesis, 1968 (University of London).

Strand, T.   The Scandinavian Caledonides: A Review. Amer.J.S^{c}i.,vol.259, 1961, pp.161-172.

Stumpfl, E.F. and Sturt, B.A.   A Preliminary Account of the Geochemistry and Ore Mineral Parageneses of some Caledonian Basic Igneous Rocks, from Sørøy, Northern Norway. N.G.U., Nr.234, 1965, pp.196-230.

Sturt, B.A.   The Composition of Garnets from Pelitic Schists in relation to the Grade of Regional Metamorphism. Journ. Petrology, vol.3, 1962, pp.181-191.

_____   Wrench Fault Deformation and Annealing Recrystallisation during Almandine Amphibolite Facies Regional Metamorphism. Journ.Geol., vol.77, 1969, pp.319-332.

_____   Exsolution during Metamorphism with Particular Reference to Feldspar Solid Solutions. Min.Mag., vol.37, 1970, pp.815-832.

_____ and Ramsay, D.M.   The Alkaline Complex of the Brevikbotn Area, Sørøy, Northern Norway. N.G.U., Nr. 231, 1965.

_____, Miller, J.A. and Fitch, F.J.   The age of Alkaline Rocks from W. Finnmark, N. Norway, and their Bearing on the Date of the Caledonian Orogeny.   N.G.U., vol.47, 1967, pp.255-273.

Sutton, J. and Watson, J.   Metamorphism in Deep Seated Zones of Transcurrent Movement at Kingwe Bay, Tanganyika Territory. Journ. Geol., vol.67, 1959, pp.1-13.

Sylvester, A.G. and Christie, J.M.   The Origin of Crossed-Girdle Orientations of Dyke Axes in Deformed Quartzites. Journ. Geol., vol.76, 1968, pp.571-580.

Theodore, T.G.   Petrogenesis of Mylonites of High Metamorphic Grade in the Peninsular Ranges of Southern California. Bull. Geol. Soc.Amer., vol.81, 1970, pp.435-450.

Tilley, C.E.   The Role of Kyanite in the Hornfels Zone of the Carn Cuinneag Granite, Ross-shire. Min.Mag., vol.24, 1935, no.149.

Tokuda, S.   Japanese Journ.Geol.Geog., vol.5, 1926-27, pp.41-76.

Turner, F.J. and Verhoogen, J. Igneous and Metamorphic Petrology.
McGraw Hill, 1960.

————————— and Weiss, L.E. Structural Analysis of Metamorphic Tectonite.
McGraw Hill, 1963.

Voll, G.   New Work on Petrofabrics. Liv.Manch.Geol.Journ.,
vol.2, 1960, pp.503-565.

Von Platen, H.   Experimental Anatexis and Genesis of Migmatites.
Controls of Metamorphism, 1965, Oliver and Boyd.

Waters, A.C. and Campbell, C.D. Mylonites from the San Andreas Fault Zone.
Amer.J.Sci., vol.29, 1935, pp.473-503.

Weill, D.F. and Kudo, A.H. Initial Melting in Alkali Feldspar - Plagioclase -
Quartz Systems. Geol.Mag., vol.105, 1968, pp.325-337.

Wilson, G.   The Relationship of Slatey Cleavage and Kindred
Structures to Tectonics. Proc.Geol.Assoc., vol.57,
1946, pp.263-302.

Winchell, A.N. and Winchell, H. Elements of Optical Mineralogy. 3 Parts.
John Wiley & Sons, 1951.

Windley, B.F.   Superposed Deformation of the Ketilidian Gneisses in
Sardlag area, S. Greenland. Medd. om Grøn., Bd. 179,
no.3.

Winkler, H.G.F.   Petrogenesis of Metamorphic Rocks. Springer Verlag, 1967.

Wynne-Edwards, H.R. Flow Folding. Amer.J.Sci., vol.261, 1963, pp.793-814.
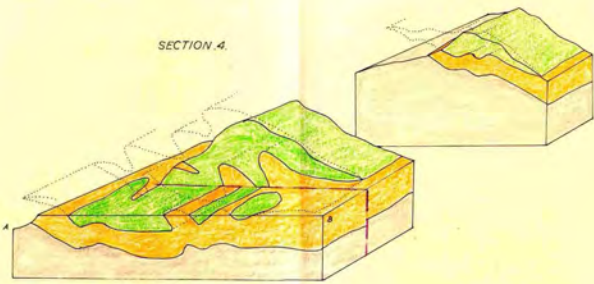
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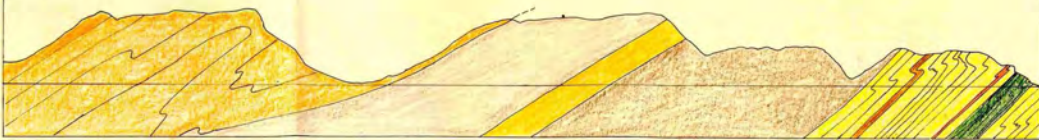SECTION. 2.
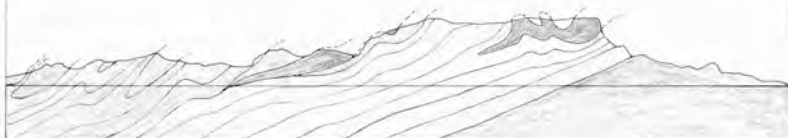
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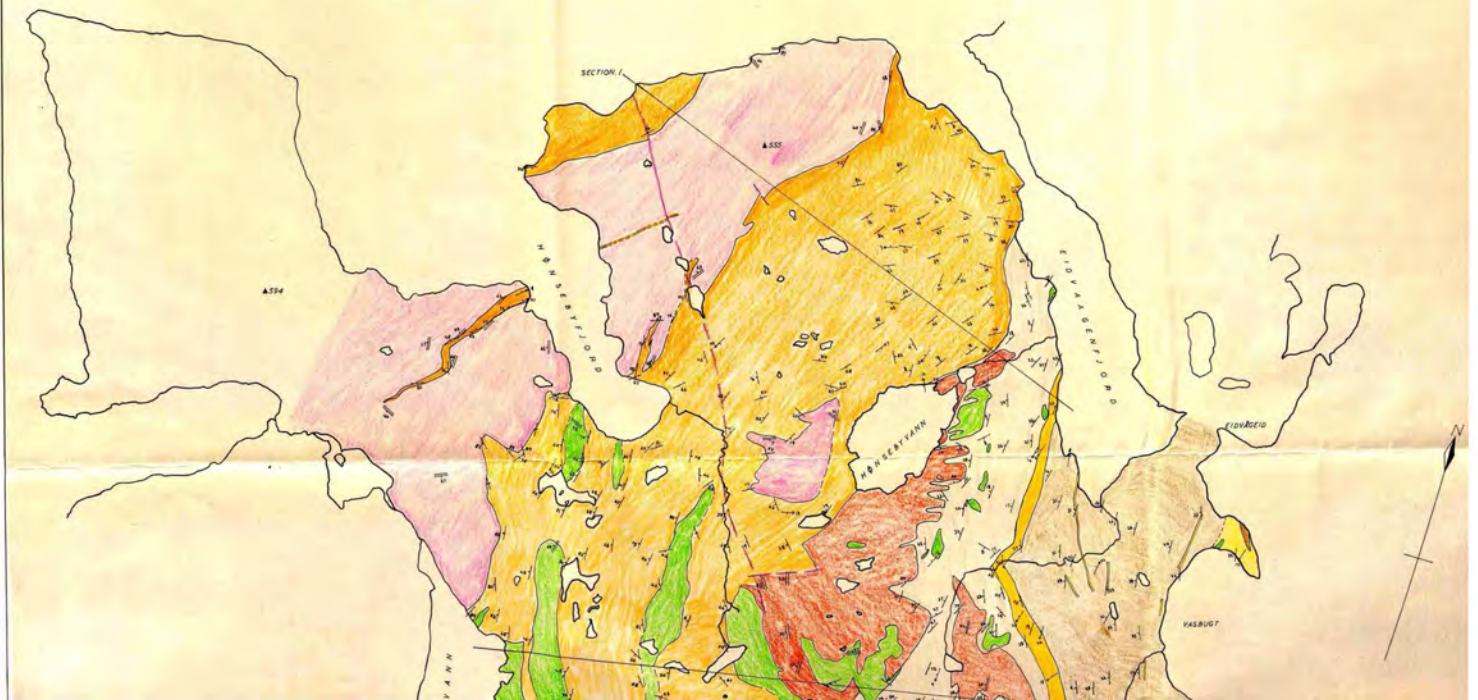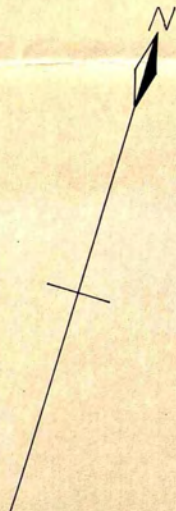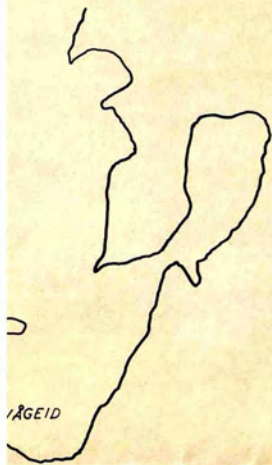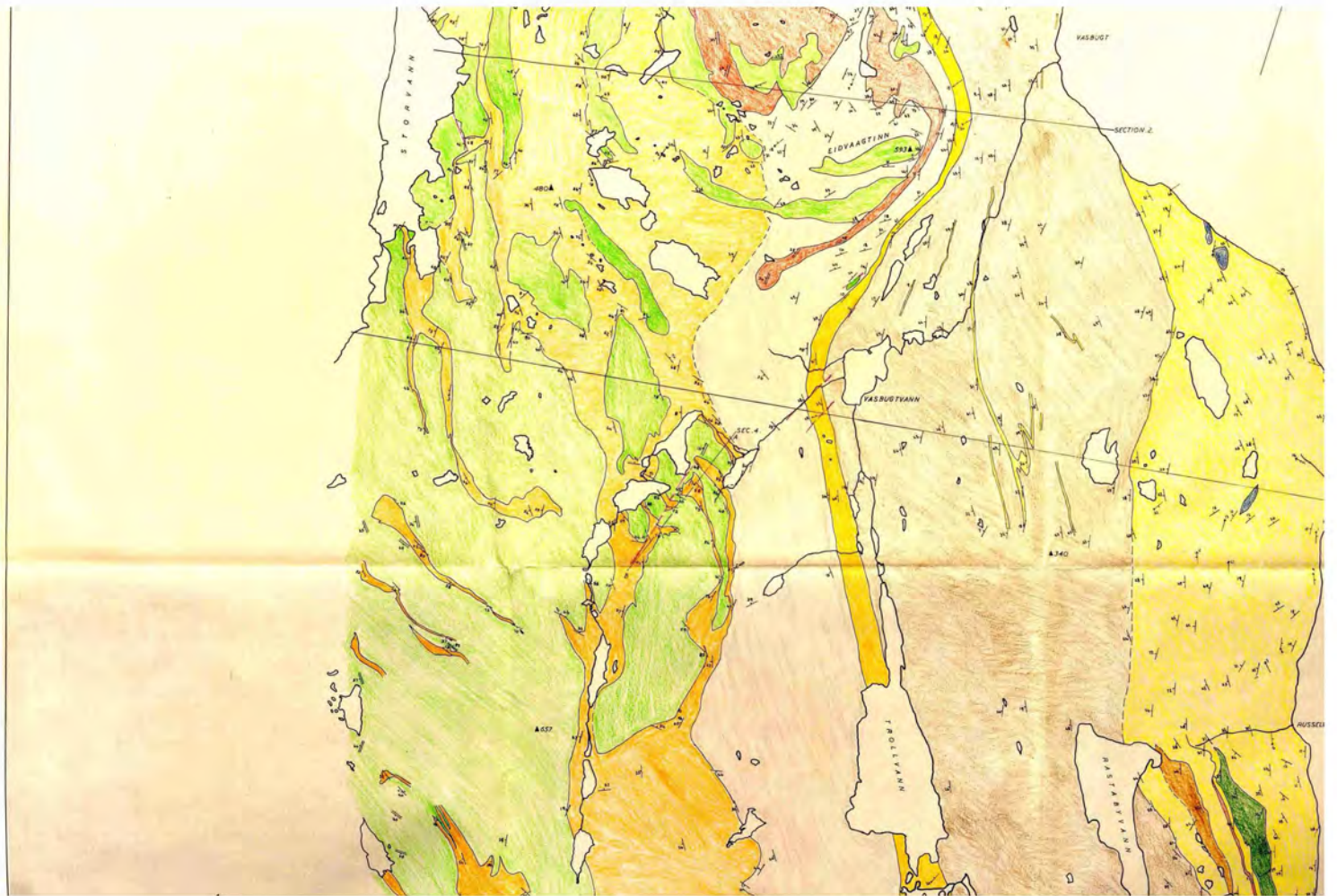
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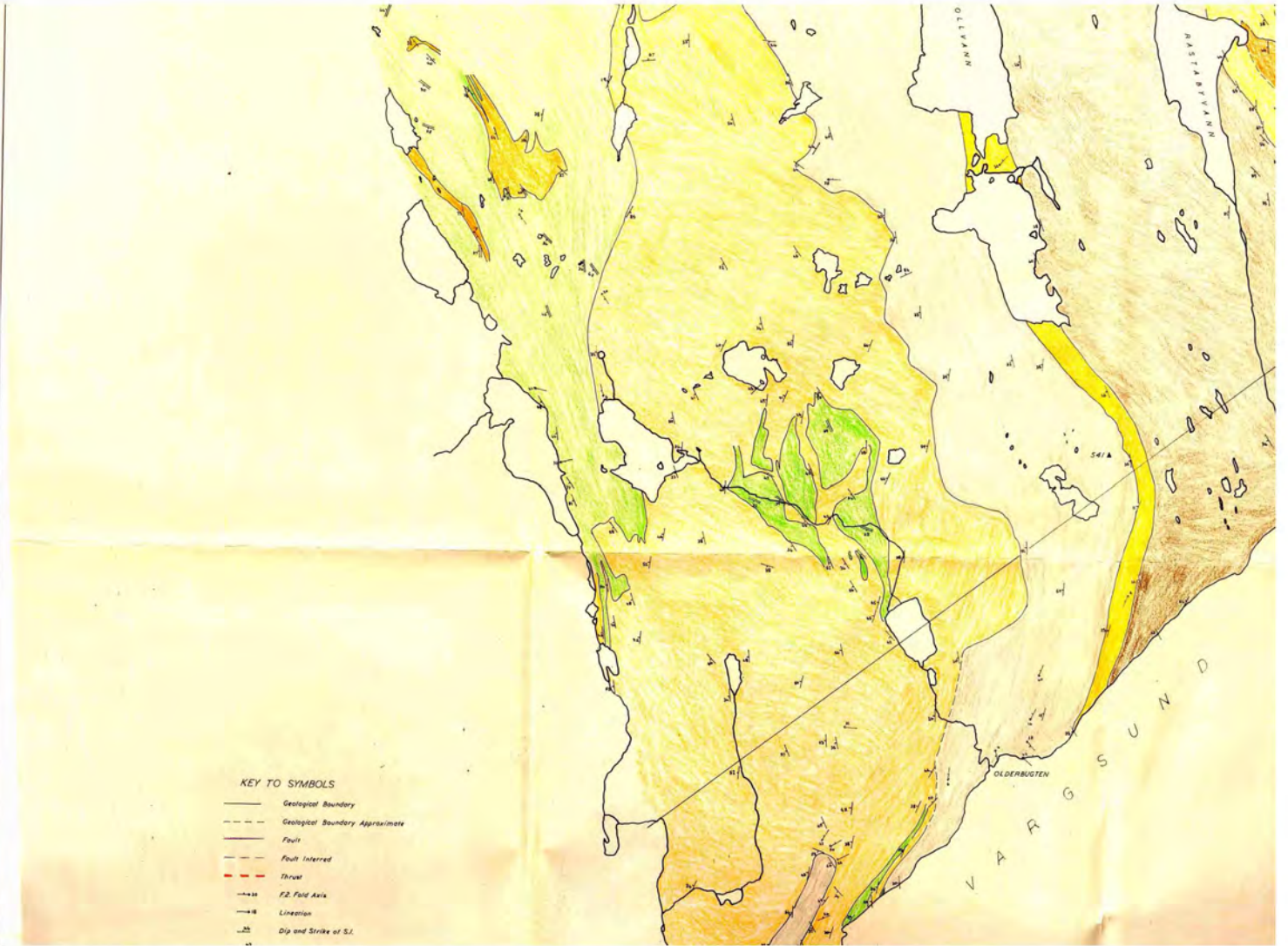
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SECTION .5.

GEOLOGICAL MAP OF EASTERN SEILA

SEILAND
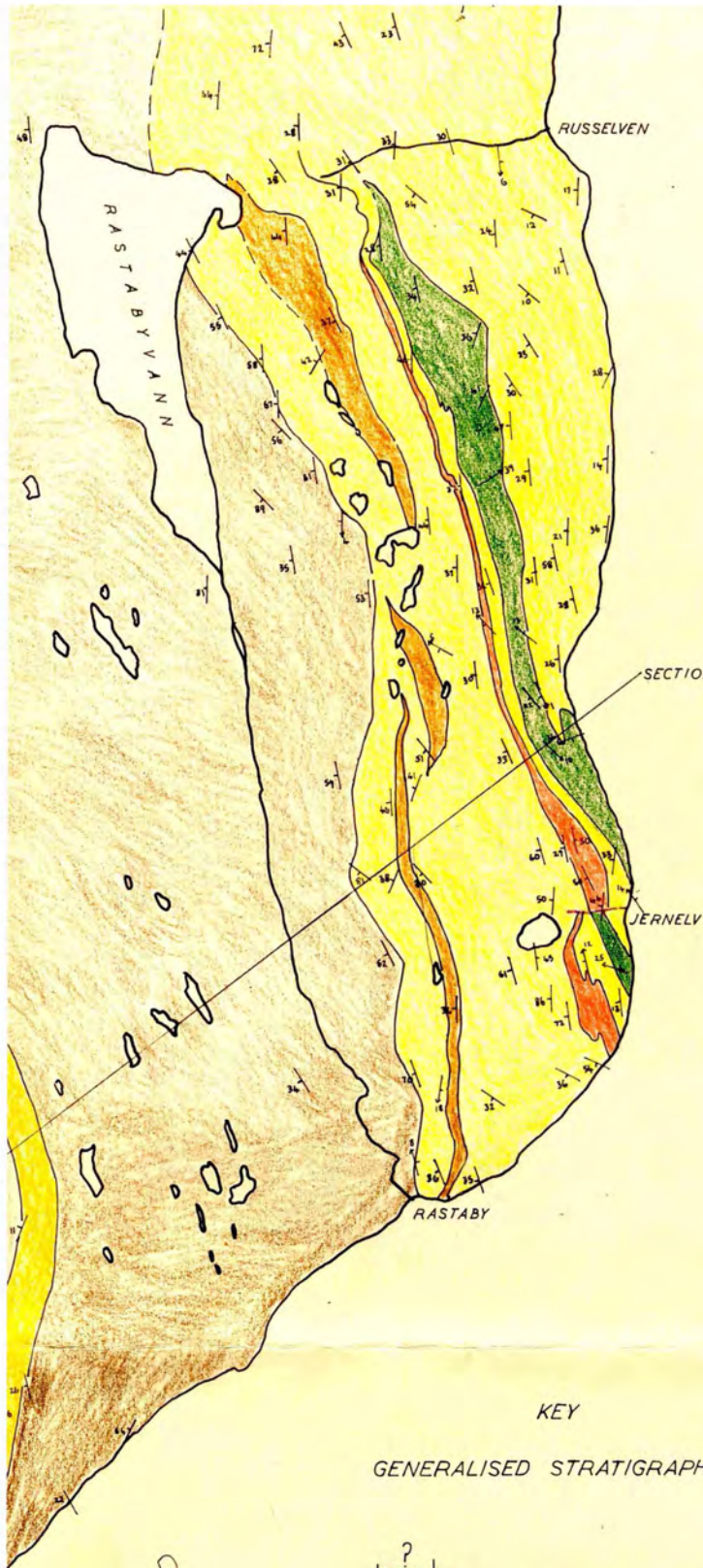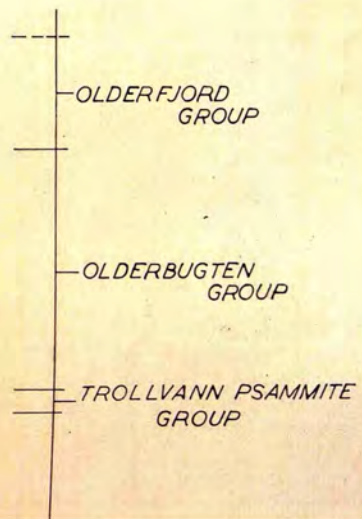
ÅGEID

N

SECTION. 3.

KOMAGNES

RUSSELVEN

KEY TO SYMBOLS

Geological Boundary
Geological Boundary Approximate
Fault
Fault Inferred
Thrust
F2. Fold Axis
Lineation
Dip and Strike of S.I.

RUSSELVEN

RASTABYVANN

SECTION. S.

JERNELV

RASTABY

KEY

GENERALISED STRATIGRAPHY

?

FLAGGY PSMTE-QZT THIN CALC-
SILICATES

SEMI-PELITE

QZT BANDS
PURPLE SCHIST

PSAMMITES

OLDERFJORD
GROUP

OLDERBUGTEN
GROUP

TROLLVANN PSAMMITE
GROUP

KEY TO SYMBOLS

———	Geological Boundary
— — —	Geological Boundary Approximate
———	Fault
— — —	Fault Inferred
– – –	Thrust
●→₃₅	F.2 Fold Axis
●→₁₅	Lineation
³⁴	Dip and Strike of S.I.
⁴⁵	Dip and Strike of Schistosity in Igneous Rocks
▲553	Altitudes in Metres

SCALE APPROXIMATELY 1:20,000

OLDERBUGTEN

V A R G S

OLDERFJORD

IGNEOUS ROCKS

EIDVAAGTINN ADAMELLITE

HAMMEREN GABBRO

HØNSEBY GABBRO

ULTRA-BASIC PODS

KEY

GENERALISED STRATIGRAPHY

?

FLAGGY PSMTE-QZT THIN CALC-
SILICATES
SEMI-PELITE

QZT BANDS
PURPLE SCHIST

PSAMMITES

QZT BANDS

BROWN SCHIST

SCHIST BANDS

FLAGGY PSMTE-QZT

PINK PSAMMITE
INTERBEDDED PSMTE AND SCHIST
CALC-SILICATE SCHIST

FLAGGY PSAMMITE

SCHIST BANDS

?

OLDERFJORD
GROUP

OLDERBUGTEN
GROUP

TROLLVANN PSAMMITE
GROUP

EIDVÅGEID SCHIST
GROUP

KOMAGNES
GROUP