Bergvesenet

Postboks 3021, 7002 Trondheim

Rapportarkivet

Bergvesenet rapport nr	Intern Journal nr	Internt arkiv nr	Rapport lokalisering	Gradering
BV 2135				Fortrolig
Kommer fra ..arkiv	**Ekstern rapport nr**	**Oversendt fra**	**Fortrolig pga**	**Fortrolig fra dato:**
Sulitjelma Bergverk A/S	"529200004"			

Tittel

Structure and petrology of granites and associated metasedimentary rocks in a region north of ùrnes.

Forfatter	Dato	Bedrift
HOLMES M.	1960	Sulitjelma Gruber A/S

Kommune	Fylke	Bergdistrikt	1: 50 000 kartblad	1: 250 000 kartblad

Fagområde	Dokument type	Forekomster
Råstofftype	**Emneord**	

Sammendrag

Et storre omrade rundt Glomfjord er undersokt med tanke pa strukturgeologi og petrologi i granitter og metasediment. Tre foldefaser er observert, og metamorfosegradene relateres til disse forskjellige foldefasene. Granittene har gjennomgatt en liknende strukturell historie som metasedimenta. Granitt.

STRUCTURE AND PETROLOGY OF GRANITES AND ASSOCIATED METASEDIMENTARY ROCKS IN A REGION NORTH OF ØRNES, NORDLAND, NORWAY

Michael Holmes

Thesis submitted for the Degree of Ph.D. in the University of London

M.Holmes
University College,London
November 1960

Statement of Thesis

The area studied consists of about 150 square miles of metasediments and granites situated towards the centre of the Caledonian orogenic belt. The granites occur both as structurally basal masses, with concordant envelopes of metasediments, and as concordant intercalations within the metasedimentary sequence. In every case, the schist-granite contents are sharply defined.

Three periods of deformation have been recognised, termed F_1, F_2 and F_3 respectively, the latter two possibly representing overlapping phases of a single major fold episode. Each of the periods produced distinctive structures from completely isoclinal first folds, through overturned and recumbent second folds to broad open folds of the latest deformation.

The first folding was responsible for the development of a regional schistosity and also of numerous slides which are sensibly coincident with the bedding. Cross-cutting slides are also present, and are correlated with the F_2 folding.

The metamorphic history of the metasediments can be related to the various fold periods. Thus fine-grained biotite and granular quartz developed mainly during F_1 times, while garnets formed subsequently, and enclose the earlier fabrics. Rotation of the garnet porphyroblasts by F_2 deformation is commonly seen. A later age of metamorphism is preserved in calcareous rocks where undeformed crystals of considerable size are present. Late stage, probably F_3, deformation is indicated by localised fracture of various minerals.

The granite masses are shown to have undergone a similar structural history to the metasediments. Reasons are given for referring them to the base of the succession, and for regarding them as having been in existence as granitic rocks before subsequent sedimentation. Localised felspathisation of the overlying sediments can be shown to be related to the underlying granite. The concordant sheet granites are mainly regarded as representing disconnected portions of the basal granites, isolated in F_1 times.

Contents

v

Page

List of Figures

List of Tables

List of diagrams in back pocket

General Description of the Area

The area with which this thesis is concerned is situated in the Nordland region of N. Norway, just north of the Arctic Circle. Ørnes is a village in the SW of the area and is the most important centre for the whole of the region. Situated some 50 km N of Ørnes is Bodø, an important market town of Nordland with a population of some 10,000.

The country is mountainous; the highest peaks, Scetertind and Galtskart, are just above 1000 metres, and are covered by perpetual snow. Access to the heart of the region is afforded by the deep glacial valleys, which predominantly have an E-W alignment and are often occupied along some of their length by glacially dammed lakes. Glacial débris covers much of the valley floors forming marshy waterlogged country, while thick woods cover most of the valley sides. Mountain ridges between the valleys are intermittently grass-covered, but rock exposures above the tree line are generally very good.

Towards the east, the general height of land increases towards an extensive dissected plateau around Scetertind and Galtskart. This area, which is wholly above the tree line, is an irregular northern extension of the Glombraeen icecap, and forms the watershed for rivers flowing W towards Ørnes, and those flowing E towards the Sokumvand region (fig.2). Situated between the high ground around Scetertind and Galtskart, at a height of 350 metres, is Lysvand, the largest lake of the area.

North of Inner Galtskart is a further area of high ground, consisting of an extensive glacial corrie, now partially occupied by three lakes and surrounded by mountains between 600 and 900 metres in altitude.

The dominant E-W topographical 'grain' of the area is determined by the regional southerly dip of the rocks. This also accounts for the steepness of the N-facing scarp slopes, some of which are vertical. Only in the extreme NE of the region do the structural and topographic features trend N-S.

Exposure in the region is variable. In the W, exposure is confined to coastal and stream sections, with more extensive tracts on the mountain ridges, but to the E there is a gradual increase of exposure as the height of the ground increases, and vegetation is less prolific, until in the Steffodalen region there is nearly 100% exposure.

Glaciation

The whole of the region shows evidence of extreme glaciation with over-deepened and U-shaped valleys, commonly occupied by glacial lakes. Most of the mountain summits, which vary between 600 and 1000 metres, probably represent the remnants of the pre-glacial peneplain. Apart from local and recent screes, there is little loose material on the high ground. A single end moraine, present as a well-defined ridge some 20 or 30 metres high, occurs on the N face of Galtskart. Irregular glacial and fluvo-glacial deposits cover many of the valley floors, consisting of unsorted boulder clay, or local developments of varved clays. Peat covers much of the glacial deposits, and produces waterlogged country, particularly in the extensive Markvand valley, and the area S of Storviken. The lower part of the Markvand valley, and the whole of the valley SE of Bolden, is covered by a uniform and regular white sand. This is banked up on the eastern side of Skraaven (fig.5), but its thickness in the valley floor is unknown. In a cutting N of Reipaa, at least 10 metres of the sand is exposed. The grains are subangular to sub-rounded, and well sorted, ranging from 0.1 to 0.3 mm. diameter. Most of the sand is composed of quartz and felspar, but a whole series of coloured silicates are present indicating local derivation from a calcareous environment. A particle size analysis, performed by Mr. J. Konig, shows similarities with glacial outwash material. Part of a shell of an ostracod, identified by Dr. E. Robinson as a member of the sub-family Cytherinae, indicates a marine origin of the sands.

Methods Used

Three field seasons, each of about ten weeks' duration, were spent in mapping an area of approximately 200 sq.km. Aerial photographs, on a scale of 1 : 16,000, were available for the eastern half of the region, and the majority of the time was devoted to detailed mapping of this area, using transparent 'ethylon' overlays to the photographs. For the region W of a line from Ornes to Grimstad, no aerial photographs exist, and enlargements of the Norwegian 1 : 100,000 topographic maps had to be used. Exposure in this area is poor, and a total of about four weeks was spent in mapping the region to determine the major structural elements and extend the more important boundaries westwards.

A base map was constructed from the aerial photographs, on a scale of 1 : 16,000, by the slotted template method (American Manual of Photogrammetry 2nd ed.). The accuracy of the map is determined by the precise location of control points, taken from the 1 : 100,000 map. Since many of the control points used are situated on the coast, they can be relied upon as being accurately located.

Acknowledgements

The research was performed during the tenure of a D.S.I.R. award. Part of the travelling expenses have been defrayed by the central research fund of the University of London and part by D.S.I.R.

Thanks are due to Professor S.E. Hollingworth for introducing the author to this field of research, and for his encouragement both during fieldwork and laboratory studies. Dr. M.K. Wells has supervised the research, and the writer is grateful to him for information relating to adjacent areas, for frequent discussions and for help in the preparation of the thesis. To other members of the research group, particularly Mr. R. Bradshaw, Dr. R.W.R. Rutland, Mr. P. Strogen and Mr. M.A. Jones, the author is indebted for fruitful discussion. The granite analyses were performed by Miss D.S. Sutherland. Thin sections have been cut by Mr. M.C. Kennard, often from very difficult material. Photographic work has been carried out by Miss S. Jackson. Mr. A. Mroz instructed the author on the methods of construction of a base map from aerial photographs, which was completed, by kind permission of Professor E.H. Thompson, in the Photogrammetry Department of University College, London.

The kindness and hospitality of many Norwegians is gratefully acknowledged, particularly that of Herr Schulz in providing accommodation in the schoolhouse at Ørnes during each field season.

Structural Setting of the Ørnes Region in the Norwegian Caledonides

The region lies well within the intensely deformed part of the Caledonian orogenic belt, approximately 200 km W of the Caledonian 'Front'. This can be traced for many hundreds of kilometres in Central and Northern Sweden, where Caledonian sediments of Eo-Cambrian to Silurian age locally rest with unconformity on the basement rocks of the eastern foreland. Relatively undisturbed and unmetamorphosed Caledonian sediments form only a very narrow and impersistent strip, and in general they are overthrust from the W by progressively more metamorphosed schists of a similar age.

In the Sulitelma region, which is 100 km E of Ørnes, and thus in an intermediate position between the latter and the autochthonous margin of the orogenic belt, the rocks are still in a sufficiently low metamorphic state for faunal remains to be preserved. Kautsky (1953) has shown that Upper Ordovician fossils are present and that successions from the Eo-Cambrian to Upper Ordovician or Silurian can be recognised. It is probable that the much more intensely deformed rocks of the present region are broadly of a similar age. The rocks of Sulitelma are regarded by Kautsky as belonging to a large-scale thrust mass - the Seve Nappe - that has travelled an undetermined distance from the west (Kautsky 1947, 1953). Within the major nappe smaller scale thrusting has operated which has locally isolated small amounts of basement granite within the metasediments. The basement over which thrusting has taken place

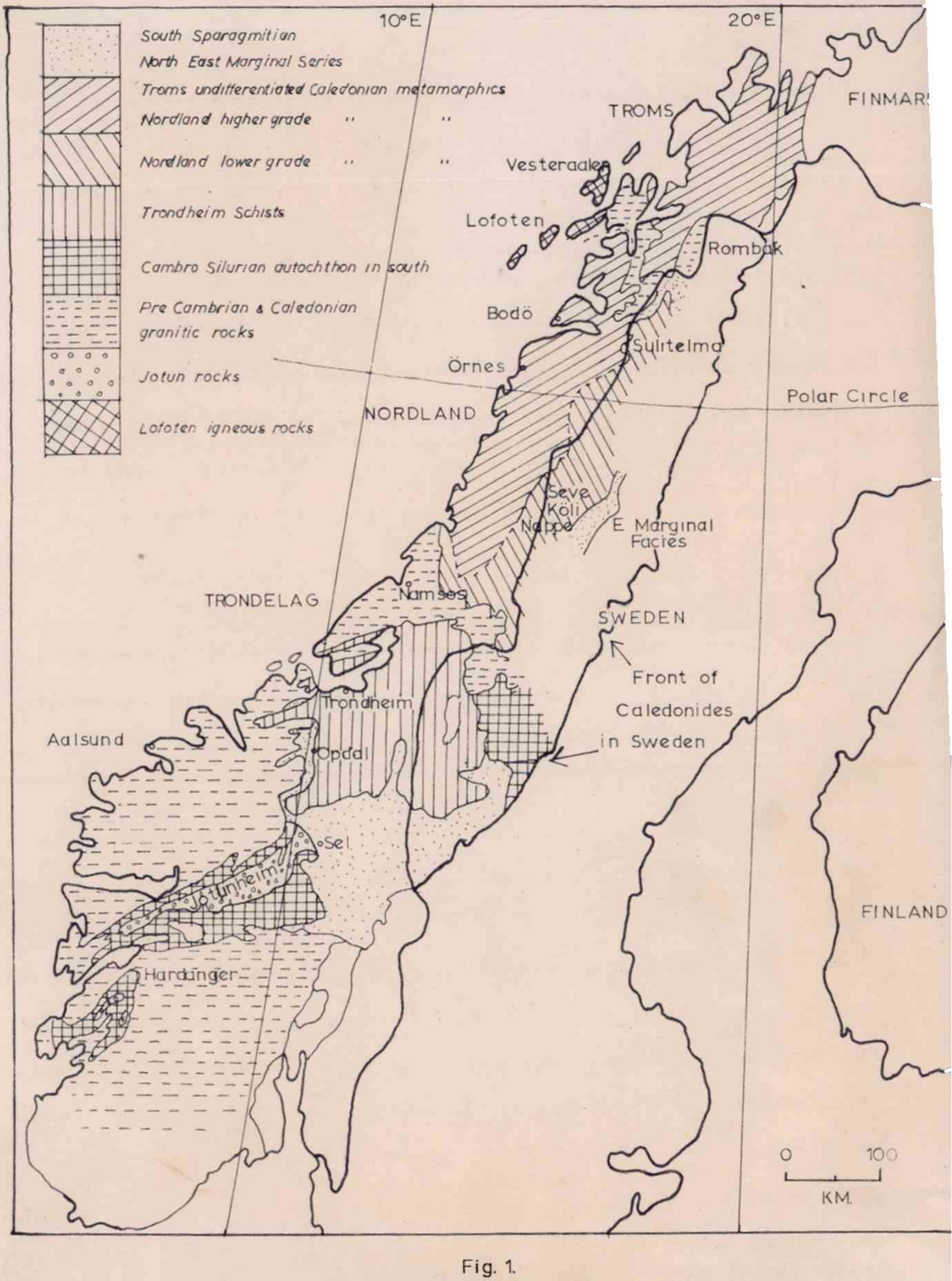
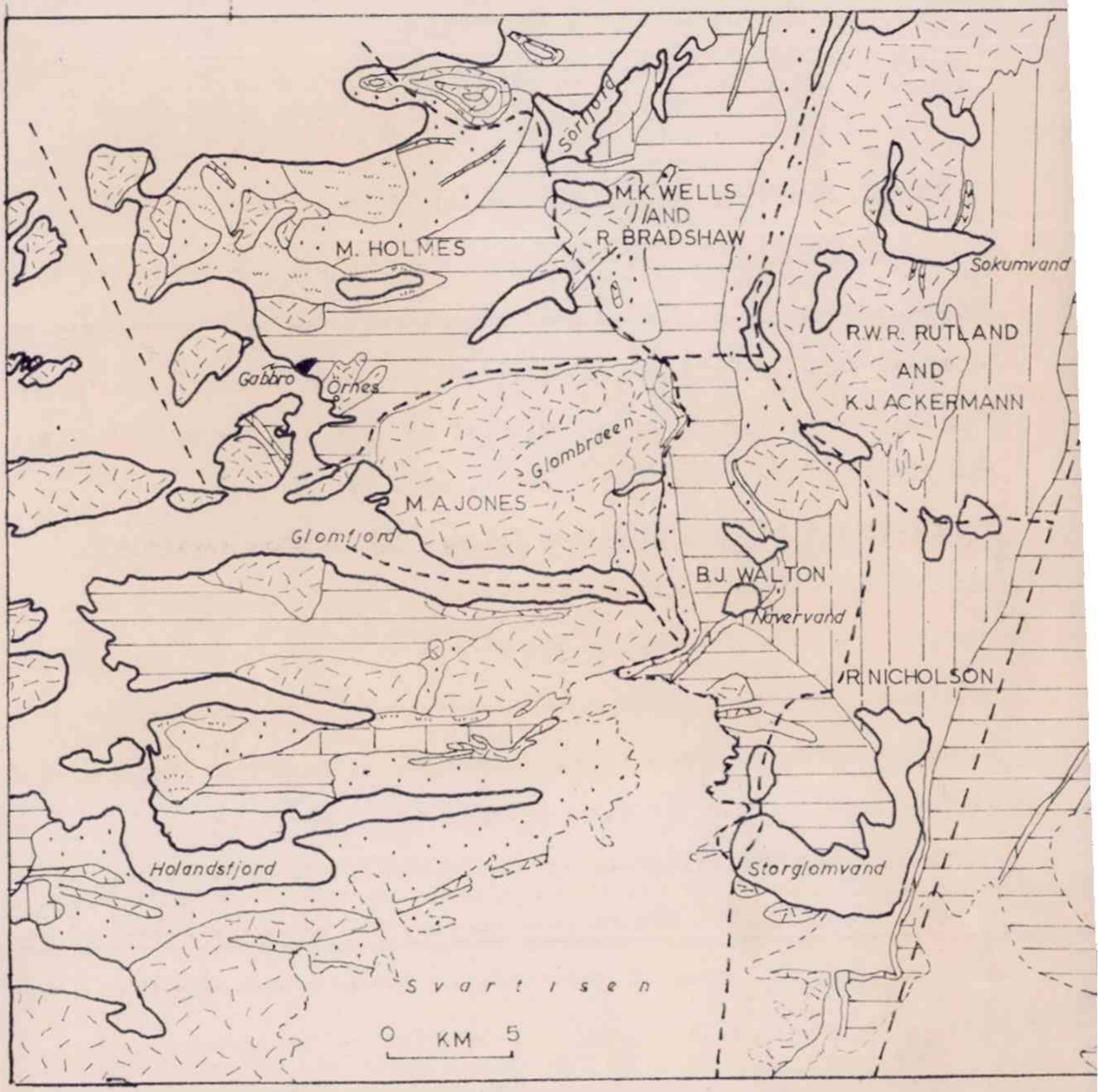

10°E
20°E
South Sparagmitian
North East Marginal Series
Troms undifferentiated Caledonian metamorphics
Nordland higher grade " "
Nordland lower grade " "
Trondheim Schists
Cambro Silurian autochthon in south
Pre Cambrian & Caledonian granitic rocks
Jotun rocks
Lofoten igneous rocks
TROMS
Vesteraalen
Lofoten
Rombak
Bodö
Sulitelma
Örnes
Polar Circle
NORDLAND
Seve Köli Nappe
E Marginal Facies
TRONDELAG
Namsos
SWEDEN
Front of Caledonides in Sweden
Trondheim
Aalsund
Opdal
Sel
Jotunheim
Hardanger
FINLAND
0 100
KM.

Fig. 1.

is recognised, and belongs to the eastern cratogenic block.

The rocks in Nordland as a whole, including those of the Ørnes region, consist principally of granitic masses, sandwiched between and enveloped by thick groups of garnet mica schists and marbles. On the most recently published survey map (O. Holtedahl and J. Dons, 1953), two types of granites are recognised, the 'Basal Granites' of Nordland and granitic igneous rocks of assumed Caledonian intrusive origin. The granites of the Ørnes region, and all of those mapped by the Department, are included by Holtedahl within this latter group, though such an interpretation is very much open to question. In a discussion of the Caledonian Mountain Chain (1944) Holtedahl suggests that the Caledonian granites were injected during the height of tectonic processes, which probably also involved extensive movement over the associated metasediments. The sections he gives across the Caledonides indicate the Western Nordland granites to be situated in the core of a tectonic depression of Caledonoid trend - the Nordland Synclinorium - resting above the Cambro-Silurian metasediments. Holtedahl's view concerning the tectonic position of the granites is not substantiated by the present study.

In view of the location of the Ørnes region, it is to be expected that basement rocks (including granites) would have become involved in the Caledonian folding. The concordant nature of the granite masses of Glomfjord and Bjellåtind are tentatively regarded as being of this character. Both the granite masses and the overlying metasediments show evidence of more than one period of folding and perhaps repeated metamorphism. The locations of several minor, and possibly occasional major thrust zones have been identified.



SIMPLIFIED GEOLOGY OF THE GLOMFJORD REGION

(After Holmsen 1932)

Areas mapped by members of the University College research group included

	Granite		Schist		Diorite
	Gneiss		Marble		Drift

Fig 2

General Character of Folding in the Ørnes Region

The relationship between the Ørnes area and the wider region mapped by members of the Geology Department, University College, is shown in fig.2. From the evidence of this whole region* it is now established that folding has occurred in at least two major episodes. The first of these was very intense, with the development of recumbent and tightly isoclinal folds. It is this period of deformation that has had the greatest effect on the original stratigraphy, as large-scale repetitions of distinctive sedimentary sequences have been produced; but as the folding is completely isoclinal, such repetitions are sometimes difficult to identify. Contemporaneous with the first folding there was large-scale sliding, causing elimination of parts of the succession in some areas. Sliding of this nature is not always easy to observe, and sometimes must be postulated on the basis of non-matching successions. Associated with the first-folding is a regional schistosity which is now the most pronounced planar structure the rocks possess. For the most part, it is coincident, or very nearly so, with the compositional banding, but at the closures of the isoclines, it can be seen to be an axial plane schistosity cutting right across the original bedding. Attenuation on the limbs of the isoclines can be shown in places to have been intense, and was accompanied by a corresponding flow of material into

* The complete area so far covered by the mapping of members of the Geology Department, University College, has been named after the most important fjord penetrating the area, Glomfjord. In future, references to this complete area will be called the Glomfjord area.

the cores. This indicates that even in groups of rocks that show no evidence of large-scale closures, there has probably been a high degree of bedding plane slip.

One of the best examples of an early large-scale isoclinal fold is the Krokvatn-Rebenfjell fold, occurring to the E of the Glomfjord Granite. The axial plane of this fold is parallel to the limbs, dipping towards the E, and the plunges of the southward-directed fold closures are inclined down the dip (Hollingworth et al 1960, Walton 1959). In the northerly part of its outcrop, around Krokvatn, the rocks in the core of the fold are a varied group of calcareous metasediments with well-defined boundaries. Even though there is complicated minor folding within the sediments, individual closures can be readily mapped due to the marked contrast of lithologies. An axial plane schistosity is the dominant planar structure (in rocks of appropriate composition) and demonstrably cuts across the banding in the core of the fold; while in the limbs it is coincident with the bedding. As the axial region of the fold is traced further S, towards Rebenfjell, a thick group of pelitic schists is encountered, dominated completely by axial plane schistosity. Due to the uniformity of composition, no examples of bedding cutting schistosity can be seen. One has to travel for about 15 km from the prominent closures in the N along the axis of this remarkable fold, before meeting the next obvious closures to the S. In the intervening belt, the rocks give the misleading impression of belonging to an unrepeated succession. A further important characteristic of the fold is the extreme amount of flow into the core, coupled with attenuation on the limbs. Bordering the fold on either side are important slides isolating it from the Glomfjord Granite and associated metasediments on the W, and from the Sokumfjell marble group on the E. The positions of each of these slide zones cannot be observed directly in the field, and have to be inferred from considerations of the succession.

When folds of this character are looked for in the Ørnes region, one meets the difficulty of poor exposure (p. 2). There are several instances where the successions of rock types are repeated and where isoclinal folding must be postulated, even though no fold closures can be identified. Detailed stratigraphic correlation of local successions, and the recognition of distinctive sedimentary sequences, is essential, and reliance must be placed on experience gained from the adjoining areas.

In the Glomfjord area, the first fold slides have had important effects on the major stratigraphy (map 4). The eastern margin to the region is marked by a slide zone separating the Vegdal Schists from the Sokumfjell Marbles (Hollingworth et al, Rutland, Nicholson 1960). The latter are separated from the uppermost Sokumvand diorites (the Harefjell Group of Rutland) by another slide, isolating the diorites in a synformal structure. To the W of the Sokumvand synform, the Meloy Schist Group is in close proximity, separated by a thin band of Sokumfjell marble. The junctions between all the groups are thus all characterised by the presence of important slides. In no cases can the exact positions of the slides be located by direct observation of mylonites, slickensides or other related

Fig.3(a). View of the south side of Lysvand from Inner Galtskart

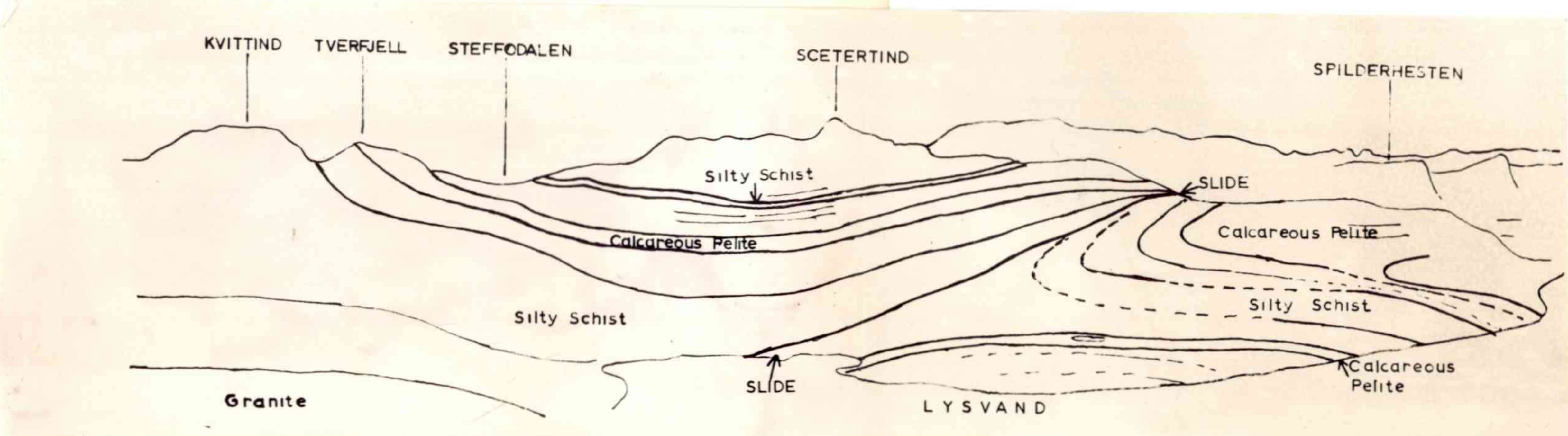


Fig.3(b). Geological boundaries visible in fig.3(a) phenomena. The exact age relations between the slides are unknown, but as they are exactly parallel to the regional schistosity - a first fold structure - and as they are commonly bent by the second folding, they are assigned to the early period of deformation.

Evidence for the existence of similar major slides in the Ørnes region is generally not so satisfactory. The most important of the inferred slides has a NE-SW alignment, and separates the Bjellåtind and Glomfjord Granites with their associated metasediments in the S from the big group of pelitic schists and sheet granites in the N. Other major slides have been identified from considerations of stratigraphic successions.

Besides this large-scale sliding, small-scale deformation between individual beds can be shown to have occurred. Isoclinal folds can often be traced in beds of one lithology that have quite simple boundaries with their neighbours. Although such boundaries have every appearance of being due to original sedimentation, their parallelism to schistosity and to the isoclinal drag folds developed during the first folding demonstrate that sliding on every scale from unimportant bedding-plane-slip to major thrusts has occurred throughout the Glomfjord area, and it is not surprising that no primary sedimentary structures have been positively identified.

Evidence of a later period of folding superimposed on the early isoclines, is widespread throughout the area. It is variable in style from broad open folds to recumbent overfolds, and may be of two phases. In contrast to the early isoclines, axial plane schistosity is rarely developed, and the folds deform the regional schistosity. In addition to the schistosity, all the major slides are folded by the second fold period. Associated with the second folding is a series of minor structures with very different characteristics from those of the early fold period. The minor folds are typically open and often angular in style with few of the flow characteristics of the early isoclines. Spasmodically developed is a well-defined strain-slip cleavage, that deforms the early schistosity and produces a planar structure which may be almost at right angles to the lithological banding. These second fold features tend to dominate the Ørnes region, and in the N, intense penetrative linear structures and strain-slip cleavage obliterate all traces of the early folds.

Double folding of the type outlined above has been described from a number of areas, notably the Scottish Highlands. In the Loch Leven area (Weiss and McIntyre 1957) large recumbent overfolds with NW axes are superimposed by varied folds with subvertical NE-trending axial planes. The Loch Morar early folds are asymmetric, and have an E-W trend (Ramsay 1957) and are superimposed by NE-SW fold. In both the areas, the geometrical relationships of superimposed structures are described in some detail.

Examination of the fabrics of the rocks in the field is important in deducing their metamorphic history. Thus on the closures of many of the first folds, an axial plane schistosity with unbent mica crystals cutting across the bedding indicates post first-fold recrystallisation. Moreover, in regions showing strain-slip cleavage, a second-fold product, even though the rocks are highly deformed on a minor scale, the mica crystals are unbent, proving that recrystallisation occurred after the later phase of deformation. Observations of this type correlate the structural history with the processes of metamorphism, a study that has been supplemented to a large extent by microfabric analysis.

Relationship of Folding to Banding

Reference to the effects of the intense deformation of the first folds on the bedding has already been made, but as it is such an important and striking feature of the region, it is considered in more detail in the ensuing discussion.

Continuous exposure above the contact of the Bjellåtind Granite (fig.4,28) on the mountain of Galtskart,* affords excellent opportunities of studying in detail what appears to be a simply bedded succession. From a vantage point on the N side of Bjellåtind the rapidly alternating sequence of metasediments can be seen to be overlying with apparent conformity the mass of the Bjellåtind Granite. The immediate contact rocks appear to occupy a normal sedimentary position above the granite which itself is finely banded parallel to its upper surface. As far as can be seen in continuous exposure even the beds which are only a metre or so thick persist to the limits of exposure, a distance of more than a kilometre. No minor structures are visible in the immediate contact metasediments to the granite, but away from this region, examples of isoclinal minor folds can be seen, particularly in the more competent but finely-banded beds, such as siliceous schists. Here, the axial planes of the isoclines (fig.31) are parallel to the schistosity and 'bedding', showing that apparently

* In references to Inner Galtskart, the more important of the two mountains named Galtskart, the prefix 'Inner' is omitted; when distinction between inner and outer Galtskart is required, it is termed Inner Galtskart.

Fig.4(a). View of the whole of Inner Galtskart from Tvertind, looking south

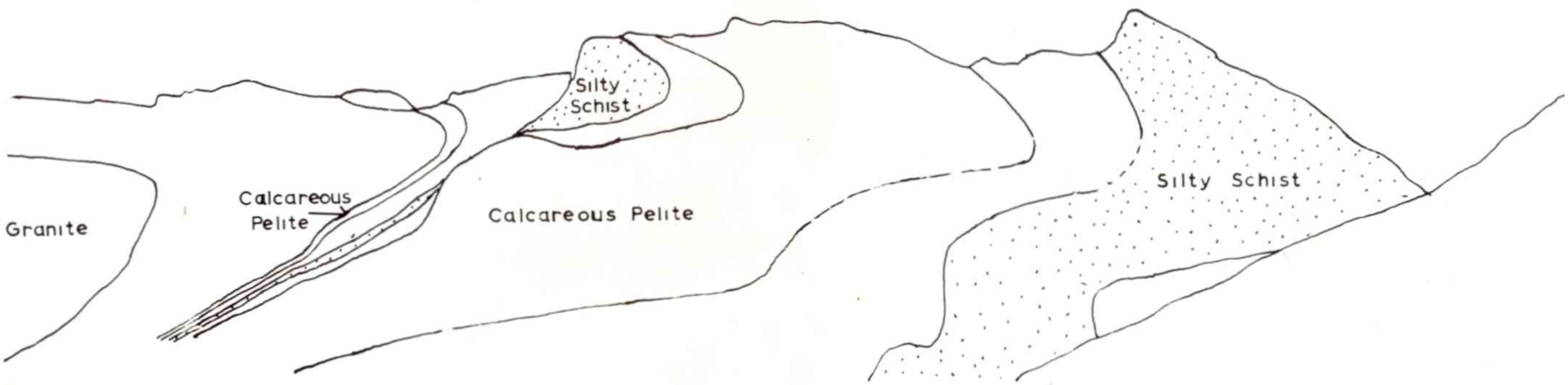


Fig.4(b). Geological boundaries visible in fig.4(a)

Fig.5. View of Skraaven from the east, showing deposit of glacial sand

Figs 6 and 7. Views of Inner Galtskart and Bjellåtind from the south, showing concordant envelope of metasediments over the Bjellåtind granite simply-bedded rocks are intensely deformed on a minor scale, and indicating that a great deal of movement must have occurred parallel to the 'bedding' planes. To a large extent, the individual beds must have acted as tectonic units in first-fold times, maintaining regular boundaries with the neighbouring rocks, and yet folding on a very intense scale internally. In these sediments that do not show such deformation styles, there is no reason to suppose that they have escaped the folding that the nearby rocks have undergone. It is more likely that they have the wrong properties for preserving such a fold style. Contrasting styles of deformation of the type outlined above have been described from the Scottish Highlands (Sutton and Watson 1952). The visible 'bedding' planes although superficially simple in origin are therefore interpreted as schistosity planes that were formed during first fold times, and along which sliding has occurred to a greater or lesser extent. Similar phenomena have been described from other highly folded terrains, e.g. the Scottish Highlands (King and Rast 1955).

An important conclusion from this is that the observed granite contact and its internal banding, which have the same appearance as the 'bedding' in the sediments above, is also a first-fold structure. This means that the granite may be separated from its metasedimentary cover by a slide of unknown importance, and may itself have undergone the intense first-fold deformation. Such a conclusion is supported by the probable existence of an elongated isoclinal fold observed by the writer in the granite of the N face of Bjellåtind (fig.12).

This fold is quite inaccessible, but appears to be defined by the convergence of two thin bands of dark biotite schist. Within the core so determinated, other bands also appear to close. Interpretation of the structure is not entirely unambiguous, but the dark bands are very clear-cut and persistent, and diverge in the correct sense for the presumed plunge of the fold over a distance of 2 km or more.

It is clear, therefore, that the observed regional schistosity is a first-fold structure, and only in restricted cases can it be seen to cut across the bedding. For this reason, a repeated succession that shows a simply-bedded nature can represent an early first-fold isocline whose axial region has been so attenuated as to mask any of the original closures (see p. 50). Such is thought to be the case for the Western Galtskart Succession, where a large-scale repetition involving five distinctive bands has been observed.

One of the most striking features of the Ørnes region is the finely-banded nature of the sediments which have the appearance of being folded into a gentle dome over the Bjellåtind Granite. This feature is representative of only the latest phase of deformation, and a closer study reveals the internal complexities of this consistently banded succession.

Terminology of Rock Groups

During the initial stages of field mapping, rock units were identified and named largely according to their mineral content. Terms such as hornblende-garnet-biotite-schist, siliceous garnet-mica-schist etc. were employed. In addition, certain distinctive rocks acquired names based on some distinctive feature of their lithology. Outstanding among these are the 'hornblende rock', a green calcareous gneiss, with prominent hornblende crystals; and 'silty schist', a massive grey quartzo-felspathic schist. These familiar terms are reserved for formations that are sufficiently unique to serve as reliable markers in stratigraphic and structural correlations. Finally, it should be noted that the successions include many examples of thin and rapidly alternating schists of various kinds which cannot be shown separately on the map, but must be grouped according to their dominant chemical or sedimentational character. In this way, groups of calcareous schists have been mapped which include rocks containing a wide range of calcareous minerals, sometimes alternating with thin marbles.

The following rock groups and mineral assemblages are recognised:-

Pelitic rocks. These contain typical assemblages such as:

(1) Plagioclase-muscovite-biotite-quartz

(2) " " " " almandine

(3) Plagioclase-muscovite-biotite-quartz - kyanite - staurolite.

Rocks of diverse appearance fall in this group, but in general they are schistose or foliated due to the mica, and weather to irregular brown or rusty-coloured slabs. Relics of original sedimentary structures are rarely seen, and irregular porphyroblastic garnets may be prominent. Kyanite and staurolite, although only rarely present, are usually easily identified.

Semi-Pelitic Rocks (Psammitic rocks, quartzo-felspathic rocks).

The name is adopted to suggest gradation towards original sandstones or quartzites from pelitic material (shales or clays). Typical pelitic minerals, such as kyanite and staurolite, are lacking, though there may be very subordinate amounts of garnet occasionally; while the proportion of total mica, particularly biotite, is reduced. A complementary increase of quartz and felspar is the consequence, and these two minerals often make up more than 80% of the rock.

One distinctive quartzo-felspathic formation has acquired the field name 'silty schist', and is a massively weathering grey rock, not unlike some of the grey granites when viewed from a distance. Occasionally the rock is very finely banded, and may preserve evidence of isoclinal folding (fig.32). Two distinct layers of silty schist are present in the Galtskart succession, and can be traced towards the E into the area mapped by Wells and Bradshaw, where the term was first applied.

The remainder of the semi-pelitic schist groups generally weather to more flaggy grey or brown rocks, which are normally readily differentiated from true pelitic schists. Certain thinly banded rocks - given the name Rusty Schist on account of the weathering of biotite to hematite - are often intermediate between pelitic and semi-pelitic schists, and therefore difficult to classify.

Calcareous Pelitic Rocks

This group, which forms very distinctive bands in the Örnes Region, is probably a unique stratigraphic horizon (see p. 176). It consists of a pelitic schist with a variable but normally small proportion of calcareous minerals, such as hornblende, diopside and scapolite. A little calcite, normally less than 5%, is always present.

It is a rock-type of uniform mineral content and appearance. It is always light brown or green-brown in colour, generally banded on a scale of about 10 cms, due to the inequalities of the percentages of the calcareous minerals present, and often showing careous weathering due to irregularities in the distribution of calcite. Occasional foliation surfaces, which are always rough and uneven, contain abundant green hornblende crystals, which have a broad and stumpy crystal habit and may be oriented parallel to an F_2 linear direction.

Calc Schists, Calc-silicate Rocks and Amphibolites

Members of this group are massive and green in colour and generally occur in thin bands. In the field, subdivision is often difficult, though on a petrographic basis it is desirable to distinguish the three kinds of rocks as follows:

(i) Calc Schists: derivatives of impure marbles often interbedded with pure calcite marbles. They normally contain both diopside and hornblende, together with quartz, calcite, biotite, scapolite etc. A very varied mineral assemblage is characteristic of the group.

(ii) Calc-silicate Rocks. When the proportion of calc-silicate minerals, such as hornblende or diopside, exceeds about 80%, the term calc-silicate rock is applied. Members of the group may be of sedimentary or igneous origin, and are sparsely but widely distributed in the Ørnes region.

(iii) Amphibolites. This group consists essentially of hornblende-plagioclase combinations, but quartz and biotite may also be present, and garnet may be an important constituent. Diopside may accompany hornblende in certain amphibolites. Two further subdivisions are generally recognised (cf. Turner, Williams and Gilbert, 1954).

(a) Igneous amphibolites: these consist principally of hornblende and plagioclase, with small amounts of quartz and biotite. Typical accessory minerals are magnetite and sphene. Garnet may be abundant.

(b) Sedimentary amphibolites: the percentage of quartz and biotite is, in general, higher than in igneous amphibolites, while diopside may accompany or even substitute for hornblende. Small amounts of calcite, scapolite and epidote are also found, while in the rocks from the Ørnes region garnet is lacking.

A considerable amount of overlap exists between the two groups, and commonly it is difficult to assign rocks to a particular subdivision with certainty.

Marbles

Both grey and brown varieties are found, as well-differentiated and generally thin layers. The grey variety is often a nearly pure calcite rock, the calcite occurring in large interlocking crystals; while the brown variety contains a series of accessory minerals such as phlogopite, tremolite etc. Grey marbles are the more common. As the marbles usually weather more rapidly than neighbouring rocks, they form easily traced mapping horizons.

Migmatites

When a pelitic or semi-pelitic schist contains well-defined lenticular patches of felspathic material, intimately associated with the schistose part of the rock, the term migmatite is applied. They are generally associated with the major granite masses, and the most well-defined occurs in a narrow belt near the Glomfjord granite.

Granites and related rocks

Rocks belonging to this group are distinctive in appearance, and even when finely banded, as in the case of some of the Storviken rocks, are readily recognisable. They consist of massive pink or grey variably banded gneisses, with a low percentage of mica and other ferromagnesian minerals. The latter are generally well oriented to give a well-defined foliation. In certain cases the grey 'granites' approach some of the more massive silty schists in appearance. However, the latter are always somewhat more schistose, with a smaller grain size, and generally a more granular quartzo-felspathic fraction.

Investigation shows that many of the granites recognised in the field are really adamellites, while others are close to quartzo-felspathic schists. In the case of the latter group, the percentage of felspar is always higher than in the undoubtedly sedimentary quartzo-felspathic schists of the area.

Quartz Dioritic Gneisses

The name is applied to banded gneisses containing plagioclase, biotite and hornblende, with other minerals, e.g. quartz and microcline, in variable but generally subordinate amount. They are of restricted occurrence, and the largest outcrop is in the vicinity of Ørnes. More calcareous-rich gneisses are found spasmodically elsewhere, the dark bands being composed of hornblende, diopside and biotite.

Ultrabasic Rock

A single lenticular mass is found to the W of Ørnes. The rocks are variably green or brown, massive in the centre of the complex, and schistose on the margins. Relict pyroxenes and olivines are common in the central portions, but secondary amphiboles and chlorite make up a large percentage, particularly towards the margins. Large rosettes of amphiboles are distinctive in certain facies of the complex.

Major Structure

Summary of Major Structures

To assist in the structural interpretation of the Örnes Region, equal area projections of poles to bedding and schistosity (π S diagrams) and linear structures (B diagrams)* were plotted (plate 1). Nine separate sub-areas were recognised, each dominated by a single major structure, or series of related major structures. When all the diagrams are combined (fig.8) it can be seen that the poles to the bedding and schistosity lie on a single great circle, within which are two distinct maxima, corresponding to two mean surfaces. The point of intersection of the two surfaces, which represent separate limbs of related major folds, intersect within the maximum of the B diagram for the whole area. This indicates that for most of the area the majority of the measured linear structures were produced at the same time as the major structures now visible and that the region approaches homo-axiality with a WSW plunging axis. The main exception to this is in the SE, where, as shown in the diagrams for the separate sub-areas (plate 1), the linear structures mostly trend south-eastwards. From the $\widehat{\pi}$ S diagram it can be seen that the axial planes of the folds are either nearly horizontal or nearly vertical.

* Observation in the field shows that all the linear structures present in the Örnes region are parallel to the b axes of the major folds. No a lineations, of the type described by Kvale (1953) are present.

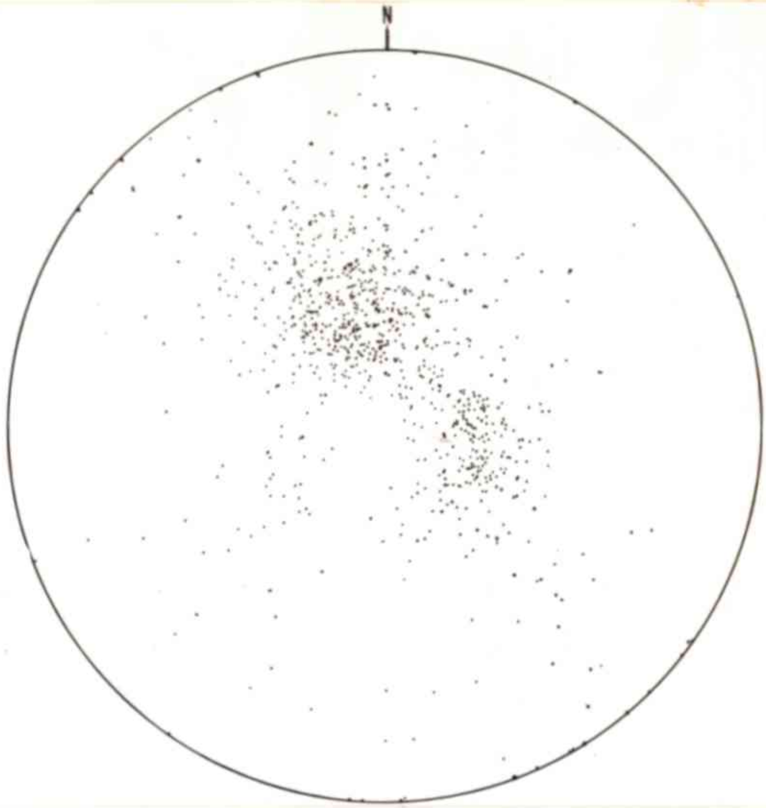


Fig.8(a). Measurements of 950 poles to bedding and schistosity for the whole region

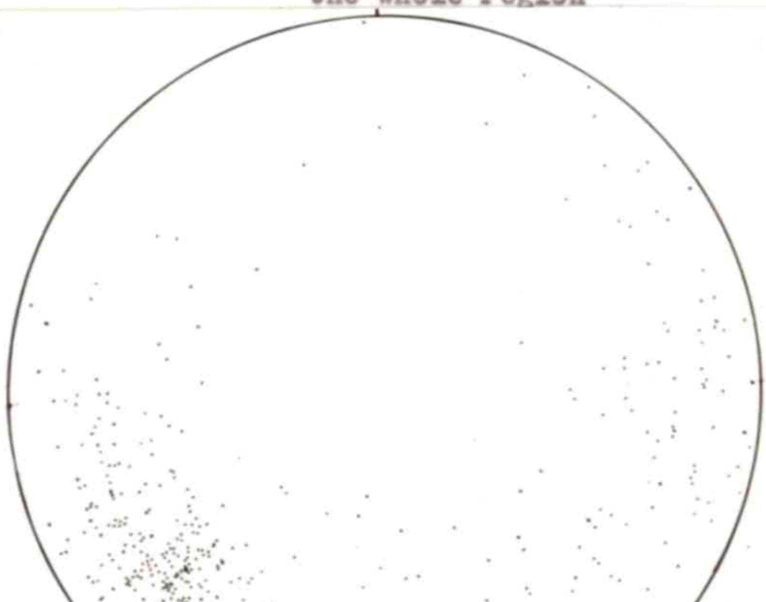


Fig.8(b). Measurements of 450 linear structures for the whole region

Due to the mountainous topography of the Ørnes region, many of the changes in trend of outcrop have no structural significance. Thus the V-shaped outcrops on Galtskart and Bjellåtind are due wholly to topographic control. In certain parts of the region, notably Blaatind and Storviken, it is difficult to project the boundaries across regions of unexposed ground because the dips are low; large-scale correlations are then uncertain. To overcome these difficulties, the outcrops of the more important beds were projected on to a plane normal to the direction of regional plunge. The method depends upon the folds maintaining their direction and amount of plunge indefinitely, and does not take into account variations in thickness of beds along the direction of plunge. In order to cover the whole of the region, the most distant points on the boundaries had to be projected about 18 km, and it is these that are likely to be the most inaccurate. Throughout the region except the S and NW, the regional linear structures are fairly constant on bearings between 210° and 240°. The average, as determined from the B diagram for the whole region, is 230°, with a dip of 17°. Boundaries from the whole of the Ørnes Region were plotted on to a plane at right angles to this (fig.9). It is immediately apparent that the most important structure of the whole of the Ørnes Region is a large-scale overfold, closing southwards, and with an axial plane that is nearly horizontal. The two limbs of the fold are at 60° to one another, a figure that corresponds to the angular difference of

TRANSVERSE PROFILE OF THE KJEIPEN FOLD

(projection normal to the regional fold axes, 230°, plunge 17°)

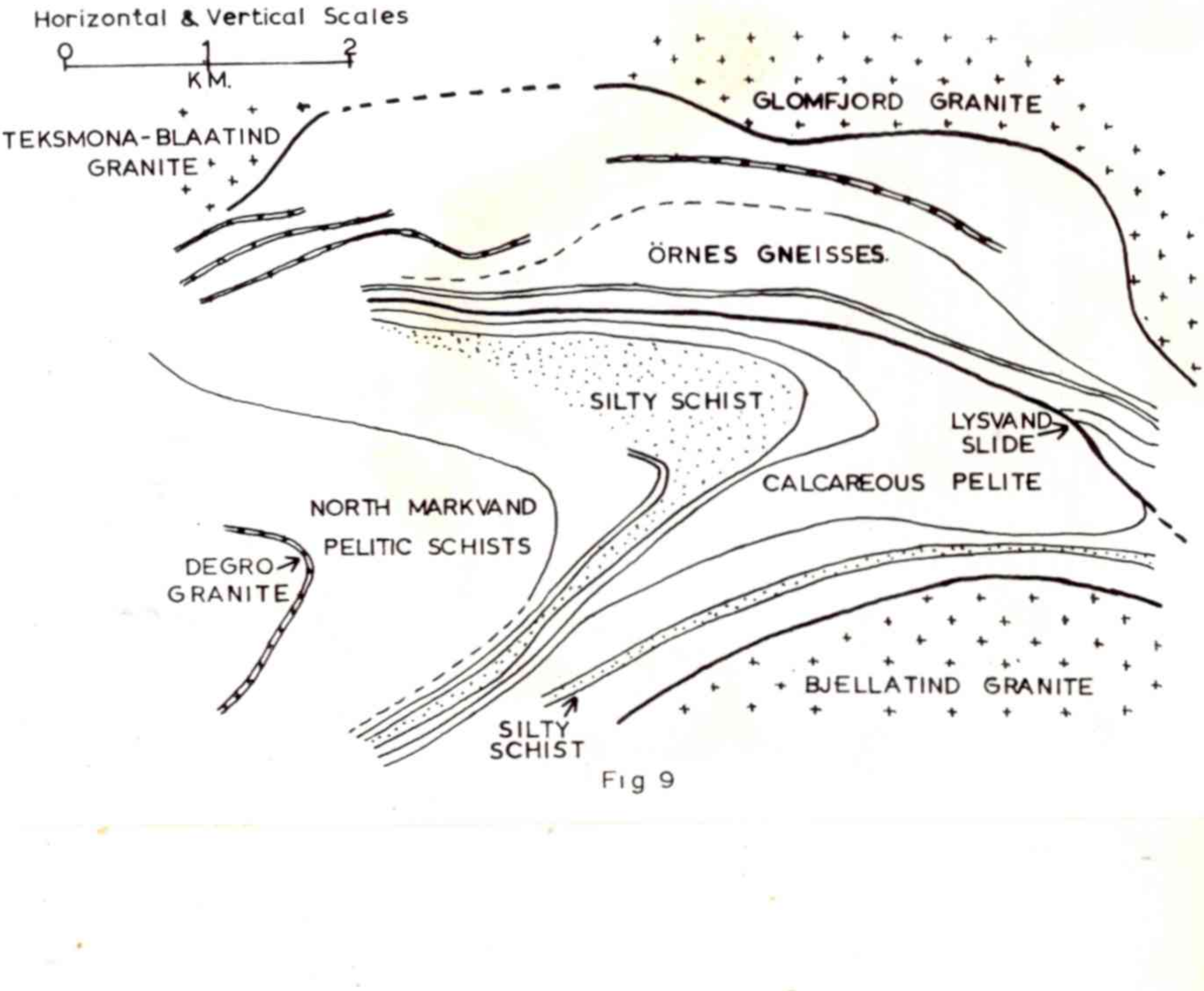


Fig 9

maxima of the π S diagram. The fold, called the Kjeipen Fold involves, and can be identified in all the rocks from Storvikvand to Lysvand (see map 3). Beneath the fold the Bjellåtind Granite occurs in the core of the Bjellåtind Antiform, a structure that is later than the Kjeipen Fold as it deforms the axial plane of the latter into a gentle arc.

Southwards from Lysvand, the metasediments dip constantly under the Glomfjord Granite, whose northern boundary projects above the Kjeipen Fold, suggesting the granite to lie structurally above the whole of the metasedimentary sequence. Moreover, the boundaries of the Mesöen and Teksmona-Blaatind granites plot as a natural continuation of the Glomfjord granite boundary, indicating the probability of correlation of these granites. In each case, the lithologies of the granites and the associated metasediments, which dip beneath them, are similar.

On the island of Mesöen and on the mainland at Bugten, linear structures with easterly dips predominate, suggesting that a true representation of the regional structure may be different, in this area, from that indicated in fig. 9. Similarly, the projection does not cover the Fykan granite and neighbouring rocks, because of the different plunge direction. In the NW of the Ørnes Region, the projected outcrops of the Skjeggen granite on fig.9 show very sinuous forms unrepresentative of the true structure as determined from direct field observation and are not included. The anomaly in this case is due to fluctuations in the WSW plunges as they are crossed by the

SSE-trending fold of North Skjeggen.

In order to obtain a true representation of the structural relationships of the southern group of rocks, where E-W linear structures are dominant, the outcrops of the major units were projected on to a plane normal to the regional linear direction (fig.10, on a plane normal to 100°, with a dip of 10°). This shows the Glomfjord granite to be situated in an important overfold, the Spilderdalen Fold, closing northwards in the opposite direction to the Kjeipen fold. Within the upper limb of the Spilderdalen Fold, the Fykan Granite is situated, itself forming the core of an isoclinal fold, the Steffodalen Fold. Above the main Spilderdalen fold, part of a major isoclinal fold, the Krokvand Fold is indicated, one of the largest isoclinal folds identified in the Glomfjord Region, and mapped by Messrs. Wells, Bradshaw and M.A. Jones. Part of the Bjellåtind antiform is again indicated on this diagram.

In both of the diagrams the overall picture produced is not an exact representation of the structural relations of the rocks, due to inherent limitations of the method of projection such as variations in the directions of regional plunge and effects of later refolding. Other major structures can be determined from field observations but their properties are such that they appear on neither of the diagrams. The equivalence of style of the Kjeipen and Spilderdalen folds suggests that they are of the same general age, and the composite diagram (fig.11) shows their

TRANSVERSE PROFILE OF THE SPILDERDALEN FOLD

(projection normal to the regional fold axes, 100° plunge 10°)

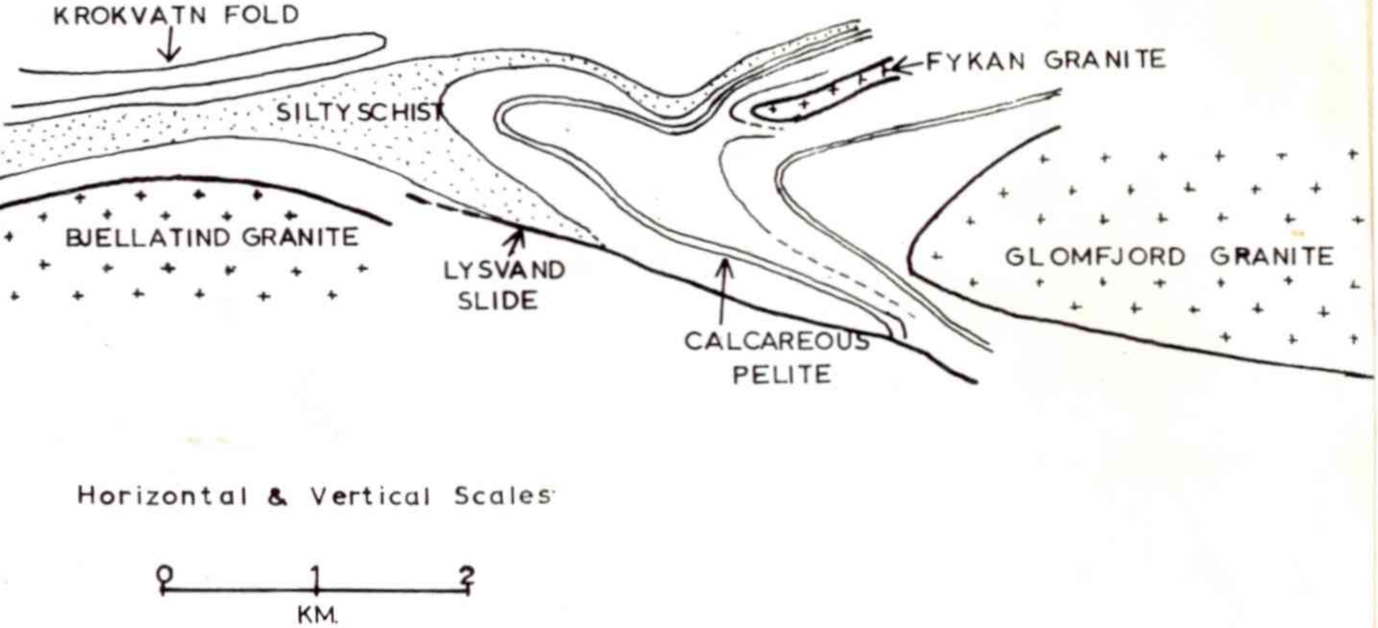


Fig 10

COMPOSITE TRANSVERSE PROFILE OF THE ÖRNES REGION

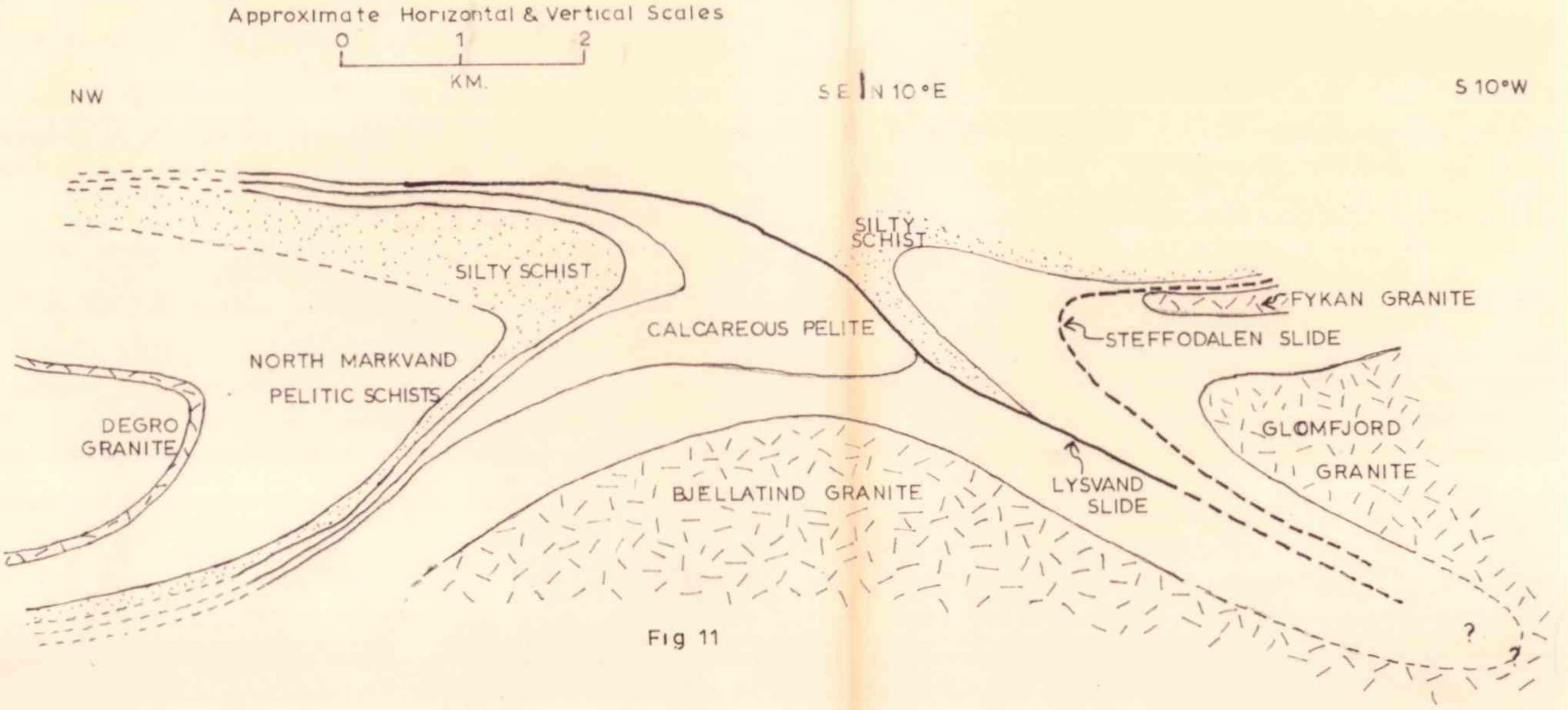


Fig 11

broad interrelations. These structures are seen to refold isoclinal folds (notably the Fykan granite fold) while they themselves are also refolded by structures of a later age (notably the Bjellåtind Antiform. This indicates three main periods of deformation, termed F_1, F_2 and F_3 respectively. In the section dealing with minor and micro structures it is shown that the styles of these phases of deformation are quite distinct. Despite this, it is possible that F_3 may represent simply a phase of F_2 folding, probably of relatively late age.

Folds of F_1 age:- In addition to the Steffodalen Fold noted above, other structures of this earliest period of folding can be recognised. Large-scale lithological repetition, involving the whole of the upper part of the Galtskart succession, is clear from field observation (see p. 50). Throughout the limits of outcrop, no closure can be seen, but the individual components of the succession do become extremely attenuated northwards towards Sörfjord. This is believed to be due to the existence of a completely isoclinal fold, the Galtskart Fold, which is dominated by an axial plane schistosity, and in which closures are no longer visible. It is significant that analogous repetitions of the succession occur above the eastern margin of the Bjallåtind granite; but in this case, there is abundant evidence of isoclinal fold closures. This structure, mapped by Mr. Bradshaw, is here termed the Kvittind Fold.

Within the granite masses of Bjellåtind and Glomfjord are probable isoclinal folds, termed the Laksaadalsvand and Spilderhesten

Folds respectively (see p. 52 and figs. 12, 13).

Other folds of F_2 age. When the Spilderdalen fold is traced westwards from Steffodalen, the uniform southerly dips represent the under limb of the structure. However, in the region of gneissic rocks to the E of Ørnes, the dips are high and locally reversed. As the correlation of basal Glomfjord granite with Teksmona-Blaatind granite is suggested from fig.9, it is probable that the steep dips in the gneissic rocks represent part of a monoclinal flexure, the Djupvikfjell Fold.

In the successions of Skjeggen and Markvand, two well-defined zones of steep dips occur, namely on the S side of Skraaven and to the NW of Markvand. These are related to two monoclinal flexures termed the Skjeggen and Breitind Folds respectively, and are thought to represent major drag folds on the upper limb of the Kjeipen Fold.

On the northern margin of the Bjellåtind Granite, in the ground mapped by Wells and Bradshaw, a zone of E-W verticality is encountered. This either represents a fold analogous to the Spilderdalen overfold, or a tight synform, and is called the Laksaadalen Fold.

A major structure of either F_2 or F_3 age involves the succession to the N of Skjeggen and forms a large-scale antiform, the Bolden Antiform, whose axis corresponds to the trend of the valley running SE from Bolden.

Other folds of F_3 age. Certain gentle worps, which are not intense enough to show up on figs. 9 and 10, can be recognised

Fig.12. The north face of Bjellåtind showing probable isoclinal fold, indicated by the convergence of two dark-coloured bands to the left hand side of the main face

Fig.13. The north face of Spilderhesten, showing probable isoclinal fold, closing towards the left, in the lower slopes

directly in the field. Although of a relatively gentle nature, these folds often determine to a large extent the shapes of the outcrops. Their axial planes are normally steep and often nearly vertical. The most important of these, the Storvik Synform, is a complementary structure to the Bjellåtind Antiform, and accounts for the reversal of dip in the Storvikvand-Storviken region.

In the region between these two important F_3 folds, smaller structures, representing minor puckers, can be identified. On the NW of Outer Galtskart, the Markvand Synform accounts for the regional change of NW dips on Inner Galtskart to the SE dips in south Skromdalsvand. When traced both towards the NE and SW, its identity is lost. A similar structure on almost the same axis can be identified on Blaatind, the Blaatind Synform.

Two other F_3 structures are established in the SW part of the region, by considerations of structural and stratigraphic evidence. A large part of the region is covered by water or is unexposed, and many of the conclusions must be tentative. Correlation of the Glomfjord-Mesöen granite with that of Teksmona-Blaatind has already been suggested, and the two must connect by an antiform, plunging west, to account for the observed structural arrangements. This structure, the Ørnes Antiform, has a steep axial plane and an easterly trending axis. Although the Skjeggen granite is regarded as representing a reappearance of the basal granite, it is thought to be separated from the Teksmona-Blaatind granite by an F_1 slide. Thus, a synformal structure is inferred from the arrangement of regional dips, the Reipaa Synform, which separates the two granites, and is the complement to the Ørnes Antiform.

Examination of the linear structures in the southern region (plate 1) shows that in the E the undoubted F_2 lineations dip E, while in the W the reverse is the case. The culmination, termed the Spildervand Culmination, is situated around Spildervand, and probably has a north-south axis.

The Slide Zones

Sliding of varied amplitude took place contemporaneously with the fold periods, particularly the first but occasionally the second. For the most part, this was spread fairly evenly throughout the rocks, producing the present schistosity and also the common small-scale dislocations which can be observed in many outcrops. Occasionally, however, sliding was concentrated in narrow, well-defined zones. Later metamorphism has completely obliterated any direct evidence of these slide zones, and they can only be identified by considerations of successions and thicknesses of marker bands etc.

Throughout the region, the granites and associated metasediments are interfolded by F_2 and F_3 movements, and in one case by F_1 movements (the Steffodalen fold). Elsewhere, in isolated regions, the two groups can be seen to be folded isoclinally but independently of one another, suggesting that the two are separated by slide zones of some importance. Stratigraphic considerations, however, indicate that the lowermost 30-40 metres of the metasedimentary sequence have probably undergone only slight movement in relation to the granite, and that the Basal Slide is probably situated above this (see also p. 177). Due to the lack of information on F_1 structures, the significance of this slide zone, which is probably present above both the Bjellåtind and Glomfjord successions is unknown.

Northwards from Galtskart, the succession becomes extremely attenuated, and many of the distinctive marker bands are eliminated. A similar thinning occurs in the North Markvand pelitic rocks. Although part of the attenuation may be due to the Kjeipen Fold, which is probably a shear structure, this does not appear to be adequate to explain the total amount of thinning observed. The latter is regarded as having been caused by sliding along the upper limb of the Galtskart Fold, contemporaneous with this first phase of deformation. The Galtskart Slide follows the outcrop of the thick marble-quartzite band in the valley between Inner and Outer Galtskart (see fig. 23A). The projected outcrop of the slide is shown in map 3 ; westwards it projects to the south of Markvand and through Reipaa, while northwards it probably reappears in Storviken due to the Storvik Synform.

On the S side of Lysvand, a closure within calcareous pelitic schists (probably a continuation of the Kjeipen fold) is cut off by a cross-cutting slide, the Lysvand Slide (fig.3). Over some distance of its outcrop, distinctive rock types are eliminated against the slide. The structure separates the Kjeipen fold, with its dominant south-westerly plunging linear structures from the E-W Spilderdalen fold. The amount of movement along the slide may have been relatively slight, as the regional successions on either side are similar. This slide, which is F_2 in age, is the only large-scale discontinuity of this age in the Ørnes Region, and is unique amongst the slides in that it can be seen cutting obliquely across major folds.

Isolated with its local envelope of schists, in the upper limb of the Spilderdalen Fold is the Fykan Granite. It occurs in the core of a completely isoclinal fold, and must be separated from the remainder of the succession, which does not show an equivalent fold, by slide zones developed at the time of folding. The considerable structural complexity of the Steffodalen region is probably to be accounted for by these Steffodalen Slides, as suggested in fig. 14 and map 3.

Lithological similarities between the Glomfjord and Bjellåtind granites and their associated metasediments, and also the Skjeggen granite associated with metasediments to the N, are discussed in section 2. They are thought to represent stratigraphic equivalents that were separated by sliding during F_1 times. The Skjeggen Slide follows the southern boundary of the Skjeggen granite. It probably accounts for the thinning of the granite eastwards, and the elimination of successive members of the North Markvand pelitic series against the granite.

Details of Major Structure

F_1 Folds.

These belong to the earliest phase of deformation and are completely isoclinal in character. An associated and often pronounced axial plane schistosity is normally developed.

(1) The Steffodalen Fold and associated structures

Situated within the upper limb of the Spilderdalen fold, in the east of Stiffodalen, is the Fykan granite. Towards the north, the latter is seen to form the core of a northward-closing isoclinal fold, which also involves a local envelope of schists (see fig. 10). It is clear that a slide of considerable importance must separate this structure, termed the Steffodalen Fold, from the schist sequence to the E. The westerly extension of this slide cannot be determined with certainty, but probably follows the major trend of the schist sequence of Steffodalen, as shown in map 3. Close comparison exists between this structure and parts of the Montagne Noire gliding nappes (de Sitter and Trümpy 1952). A series of sections of the development of these nappes, reproduced from de Sitter 1956, together with a comparison with the Steffodalen region, is shown in fig.14. Although geometrically similar it cannot be determined whether or not the Steffodalen structures developed by the same mechanism.

The Fykan granite is lithologically similar to the Glomfjord granite, and it is possible that the two granites connect on the south side of Glomfjord by an isoclinal fold, the complement to the Steffodalen fold. No repetition can be established in the

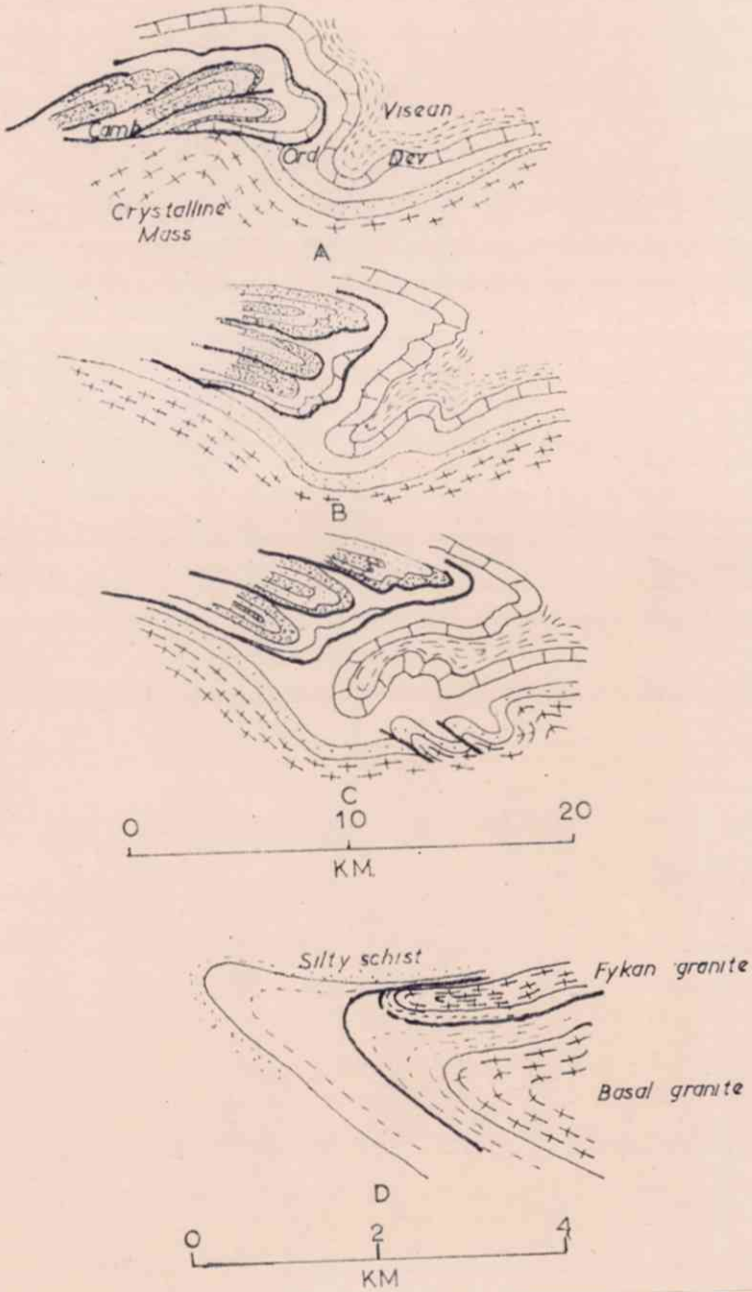


Fig.14. A-C Development of the Montagne Noire gliding nappes, after de Sitter and Trümpy (de Sitter 1956).

D The Steffodalen region, showing the Fykan granite nappe forming part of the limb of a major overfold around the basal Glomfjord granite. Close geometric symmetry with parts of fig.C is apparent.

Fig.15. View of Högnakken from the south, showing the abrupt termination of one of the sheet granites.

Fig.16. Banded sheet granites on Grötset, with Storvikvand in the foreground. The gentle synform is the Storvik synform.

schist sequence between the two granites in the area examined on the east of Glombraeen (M.A. Jones: personal communication) but a large amount of sliding would clearly accompany the development of such a structure and the symmetry of lithologies could well be destroyed.

Close comparison exists between the Steffodalen and Krokvand-Rebenfjell folds (Walton 1959, Hollingworth et al 1960), although the foliation of the outer margin of the Fykan granite follows the arrangement of the fold, inside it is wholly axial plane. It is probable that these two major structures were developed in the same general period of deformation.

The recognition of the Steffodalen fold is important in the understanding of the structural history of other granites of the Ørnes region. Except for its northern termination, the Fykan granite has every appearance of representing a sheet granite in an unrepeated succession, and is exactly analogous to the sheet granites of Blaatind and Skjeggen. Thus it is possible that some or all of these other sheet granites represent isolated fractions of basal granite. Such an explanation is suggested in the case of the Blaatind granites which are lithologically very similar to the Glomfjord and Bjallåtind granites. Some of the other sheet granites, notably those of Skjeggen and Storvik are probably thrust sheets, emplaced during F_1 times, and in the occasional instances where terminations of the granites are visible, they do not appear to occupy isoclinal fold closures (fig.15).

One other major isoclinal fold has been recognised in the Steffodalen region, and is indicated by the westerly termination of a thick marble band. Its characteristics are broadly similar to those of the Steffodalen fold.

(2) The North Kvittind Folds

Beneath the silty schist of Kvittind, a duplication in the succession around a distinctive pelitic schist has been identified by Wells and Bradshaw as caused by an isoclinal fold, closing westwards. The schistosity cuts across the lithological banding at the nose of the fold, and is an axial plane structure; the axis of the fold plunges east. On the main geological map, all the information of this region, and that on the eastern margin of the Bjellåtind granite, was kindly supplied by Dr. Wells. On the shores of Lysvand, the lower limb of the fold is no longer seen, due to the action of a slide developed at the same time as the folding.

Rusty bands within the silty schist on the north face of Scetertind show evidence of a tight, nearly isoclinal, fold, closing towards the east. It appears to be of equivalent style to the fold in the north of Kvittind, but due to the inaccessibility of the cliffs, its properties are not known in detail. Slides probably separate this from the isocline in pelitic schists described above. Towards the west, the silty schist is eliminated by the Lysvand Slide (fig.3), but eastwards, the schist continues and has a very distinctive outcrop (see map 4). Its northern boundary follows that of the Bjellåtind granite into the area E of Sörfjord, where it closes northwards due to an isoclinal fold on a north-south axis, and forms part of the lower limb of the Krokvand-Rebenfjell fold. (Wells: personal communication). Its southern boundary, however, becomes involved in the Spilderdalen fold, and follows the schist envelope of the Glomfjord granite, again on the lower limb of the Krokvand-Rebenfjell fold. Thus the silty schist has a trifurcating outcrop centred on Kvittind, analogous to the outcrop of the west Lysvand calcareous pelitic schist shown in fig.26. One branch continues northwards and represents the core of a north-south isocline, one southwards, and again probably is repeated by folding, and one westwards into Scetertind where the easterly closing isocline already described occurs. This last branch is also involved in the Kjeipen fold, and reappears on Galtskart due to the Bjellåtind Antiform.

(3) <u>The Galtskart Fold and associated structures</u>

One of the most controversial aspects of the structural interpretation of the Örnes region concerns the possibility of isoclinal folds of major importance affecting apparently simply bedded successions. The most striking of these involves the upper part of the Galtskart succession, illustrated in figs. 4, 58, where lithological repetition involving a 5-part sequence is admirably displayed. No major closures have been identified in a series of traverses across the area, and the schistosity is everywhere completely parallel to the lithological banding. All the minor structures seen in this part of the succession belong to either the F_2 or F_3 period of deformation, except for occasional small-scale isoclinal folds in the upper silty schist (fig.31). These are only very occasionally visible, and have their axial planes parallel to the dominant schistosity. This is, therefore, evidence of the rocks having undergone F_1 deformation.

When traced northwards from Galtskart, the succession becomes extremely attenuated, and on western Sörfjord, it can be shown that many of the distinctive marker bands, such as the silty schists, have been eliminated, or so thinned as to be unrecognisable. This attenuation is probably to be accounted for by the action of the northerly closing Galtskart fold, but there are other possibilities. Thinning on the limbs of the Kjeipen fold is illustrated in fig.9; this is largely to be accounted for by the thickening of the west Lysvand calcareous pelitic schist in the Suppevand region by a series of F_3 minor folds, situated in the core of the Kjeipen fold, and is most probably a separate phenomenon from the thinning from Galtskart northwards. In addition, the Galtskart slide, which probably formed at the same time as the Galtskart fold, may account for some of the attenuation (see p. 94).

Projection of the North Kvittind folds from Kvittind to Galtskart is likely, ~~and is probably of the same general phase of deformation~~. Thus, although the Galtskart succession has every appearance of being a normal one above the Bjellåtind granite, there is probably successive repetition caused by major isoclinal folds.

(4) The Laksaadalsvand and Spilderhesten Folds

Both these structures occur in the basal granites, in the Bjellåtind and Glomfjord granites respectively (figs 12 and 13). That in the Bjellåtind granite is determined by the convergence and closure of a dark band high up in the N face of Bjellåtind, and although it is not clear in the photograph, the writer is convinced of its existence due to detailed examination of the region through binoculars. The bands are probably dark biotite schists that are fairly commonly seen in the Glomfjord and Blaatind granites; none are exposed in accessible parts of the Bjellåtind mass. The closure of the presumed structure is eastwards, and the two bands continue to diverge in the correct manner when traced on to the NW face of the mountain; the probable axis of the fold, from the observed divergence of the bands, is aligned east-west.

The fold on the north side of Spilderhesten is a more convincing structure involving three distinctive bands, probably of mica schist, which are visible from a distance in steep cliffs. The closure of the fold appears to be eastwards, and the axis of the fold, which probably runs nearly parallel with the north face of Spilderhesten is SW.

In both cases, the areas where the presumed folds outcrop are completely inaccessible, and it is not possible to be entirely certain whether or not the structures exist. If their existence is accepted, it indicates that the basal granites, together with the schists, have undergone isoclinal F_1 deformation even though contemporaneous sliding has generally separated the two.

R. Nicholson (1960) has photographed similar isoclinal folds in the Svartisen granite, a mass which he regards as basal, and the stratigraphic equivalent of the Glomfjord granite. There can be no doubt of the existence of these folds, which are beautifully exposed, and afford confirmatory evidence for the less clearly displayed structures in the Ørnes region. Nicholson regards the Svartisen granite folds as representing F_1 isoclines and indicates the variance of this view with that of Skjeseth and Sörensen (1952) who interpret the western equivalent of the Svartisen granite which outcrops in Holandsfjord as an allochthanous sheet emplaced during the early folding.

F_2 Folds

These are the most widespread and important structures of the Ørnes region, and have the greatest effect upon the trends of outcrop. They consist generally of broad-scale recumbent overfolds with gently inclined axial planes.

(1) Kjeipen Fold

This structure involves the whole of the succession from Storvikvand to Lysvand, and successive closures can be traced in all the major lithological units. Because the axial plane of the fold is nearly horizontal, its outcrop on the surface is sinuous, and is determined largely by relief. Prediction of its position in unexposed ground is therefore difficult, and is complicated

Fig.17. View of Scetertind from the north-west, showing closure of the Kjeipen fold in calcareous pelitic schists in the foreground (see also fig.19B)

Fig.18. Closure in calcareous pelitic schist on Kvittind. This is the most northerly closure of the Steffodalen fold identified

by the fact that it has been refolded in the F_3 period of deformation. Direct observations of successive closures are only possible in isolated regions. In order to fix the position of the axial plane with some degree of accuracy, the successive points of verticality on fig.9 were projected back to their position on the base map. Good agreement between observed and calculated zones of verticality was obtained for well-exposed regions and by joining all the calculated points, the trace given in map 3 was obtained. The location of the axial plane in the north is uncertain, and north of Degro, where the last exposure of vertical rocks occurs, the trace is based entirely on projection. Although there are local variations in the direction of dip of the axial plane, its mean inclination is to the south-west, bent by the Bjellåtind Antiform and Storvik Synform. Southwards, the fold is cut out by the Lysvand Slide.

Reference to map 3 shows that the lower limb of the fold is exposed to the north of Lysvand and Galtskart, while the whole of the region to the west of the trace of the axial plane is within the upper limb.

The clearest demonstration of the fold is in the Galtskart-Suppevand region. On Galtskart itself, the regional dip is westwards, forming the western limb of the Bjellåtind Antiform, but when traced towards the south-west, the succession becomes involved in a whole series of complicated minor folds, and is finally overturned to dip south-east in the region to the west of Suppevand.

Rusty bands at the base of the silty schist that outcrops on the north of Kjeipen show a whole series of asymmetric drag folds with gently inclined axial planes (fig. 19A). The drag folds are dextral, in the correct sense for the major closure of the fold on Kjeipen, and the whole of the massive north face of the mountain is regarded as being situated in the core of the fold. When the boundary of the silty schist with the underlying pelitic schists is traced south-westwards it becomes involved in a complicated series of minor folds, and then finally dips constantly south-east, and is situated on the upper limb of the main fold. Marble bands in the pelitic schists show extremely complicated minor folds with horizontal axial planes and a common development of small-scale slides (fig.21).

Unlike the silty and pelitic schists that, en masse, show a relatively simple arrangement, the calcareous pelitic and associated thin pelitic schists are complexly folded in the Suppevand region. The group as a whole follows the same structural pattern as the associated silty and pelitic bands, but instead of the regional dip changing through a well-defined zone where the dips are vertical, the schists are folded into a series of persistent symmetric folds with a varied amplitude and vertical axial planes. A detailed map of this region is shown in fig.20, and a cross-section in fig.22. The northern boundary of the calcareous pelitic schist is disharmonic, as indicated by the

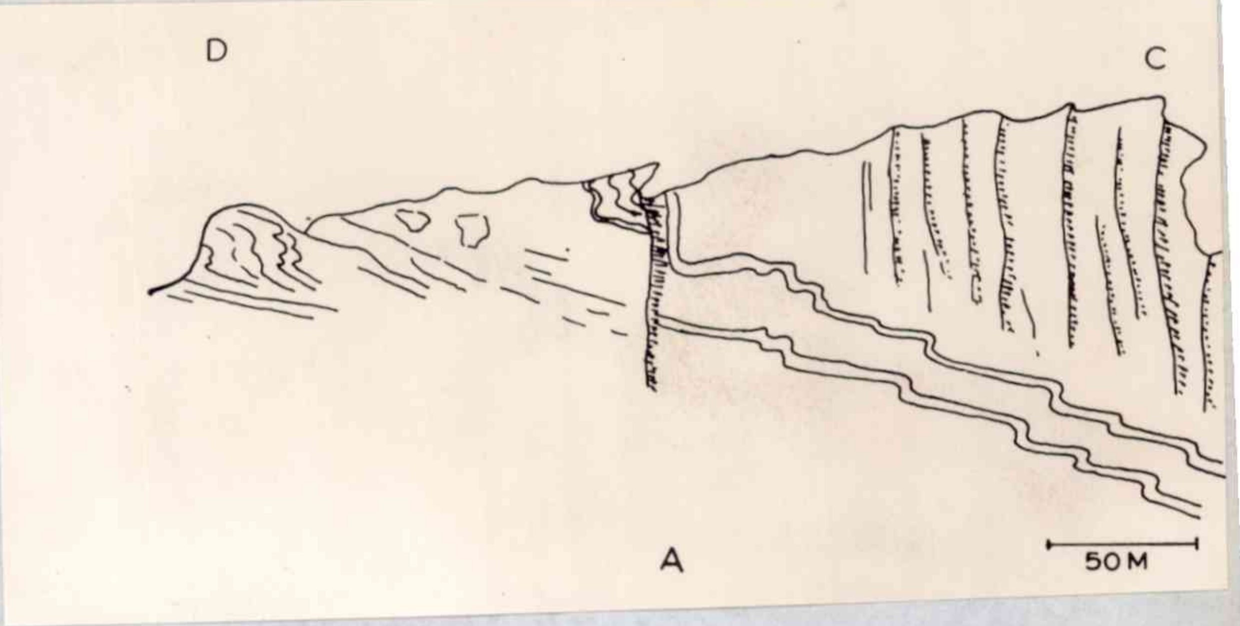


Fig.19A. The summit of Kjeipen from the north, showing drag folds in rusty schists, situated on the lower limb of the Kjeipen fold. C-D refers to the location of the section on the map, fig.20.

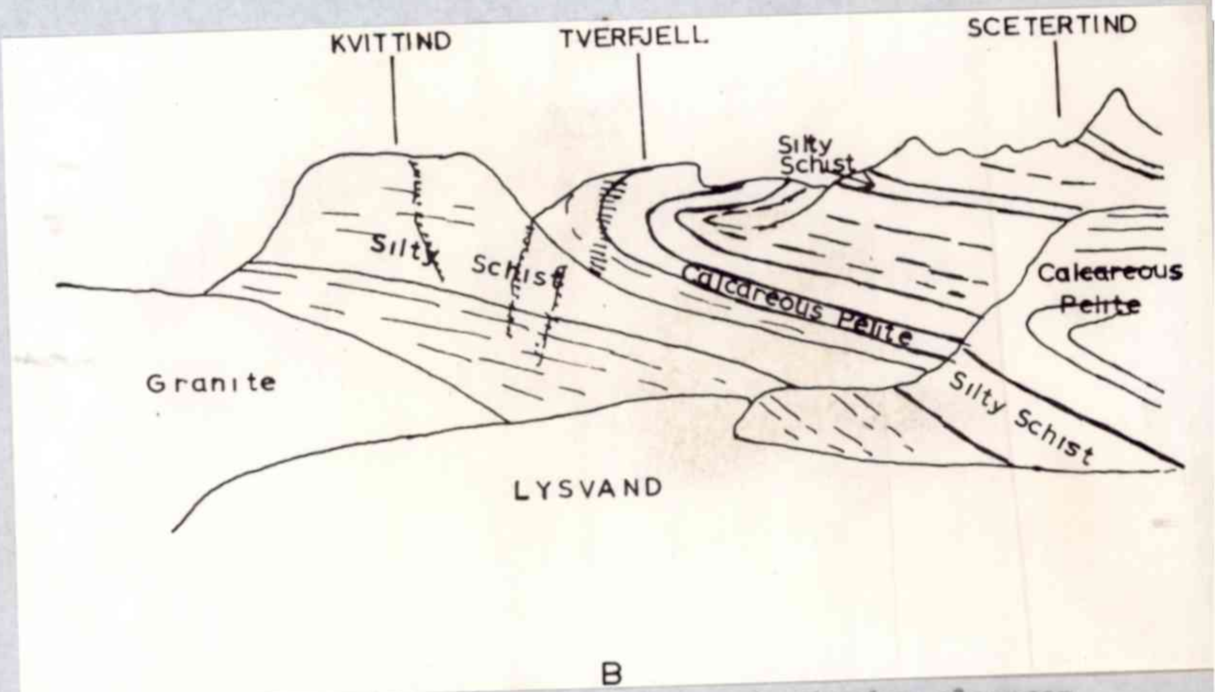


Fig.19B. View of the eastern end of Lysvand, showing closures belonging to the Steffodalen fold (on Tverfjell) and to the Kjeipen fold (right foreground). The location of the Lysvand slide, which separates the two structures, cannot be seen from this viewpoint. (See also fig.17).

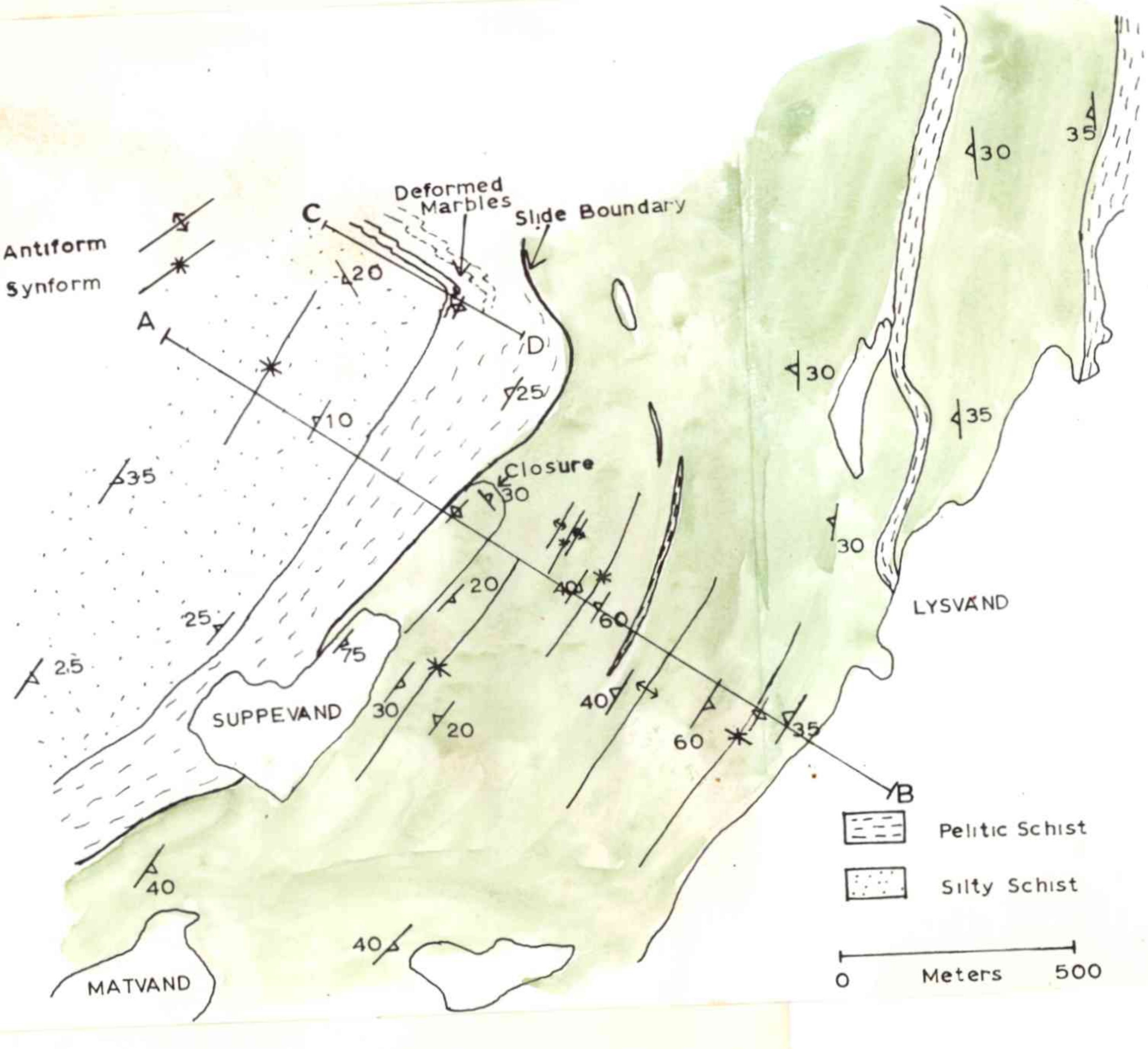


Fig.20. Details of the fold pattern around Suppevand, an area situated in the axial regions of the Kjeipen fold. Calcareous pelitic schists are coloured green. Section A-B is illustrated in fig.22, and C-D in fig.19A.

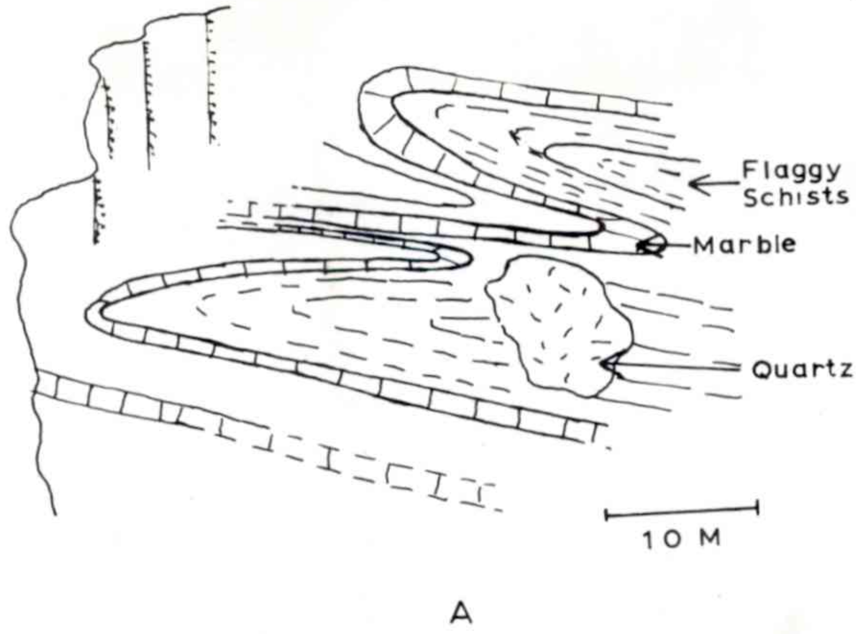


Fig.21. The fold pattern in marbles and schists on the north face of Kjeipen. The age of the folds is unknown.

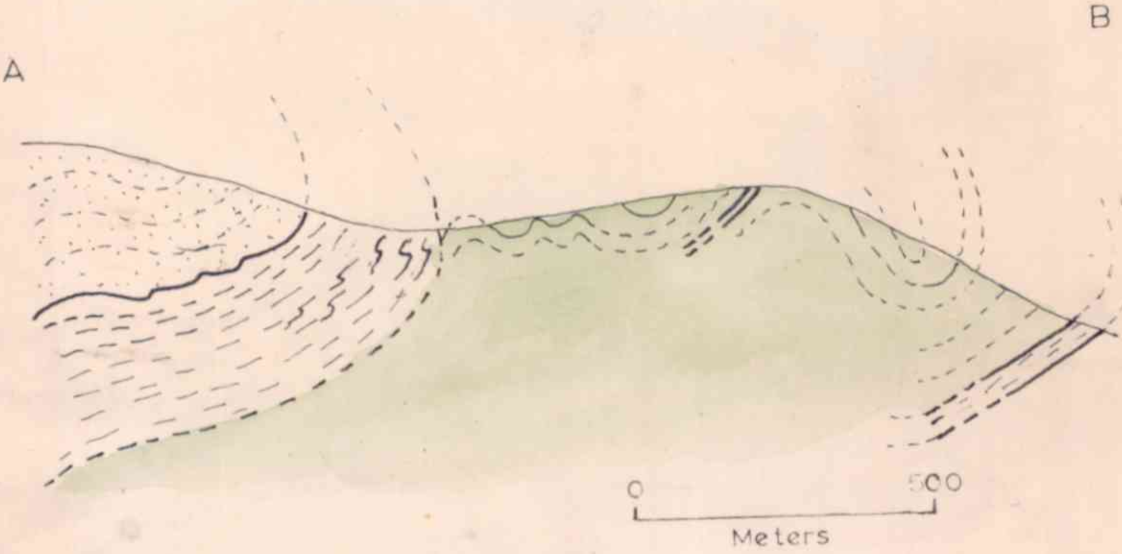
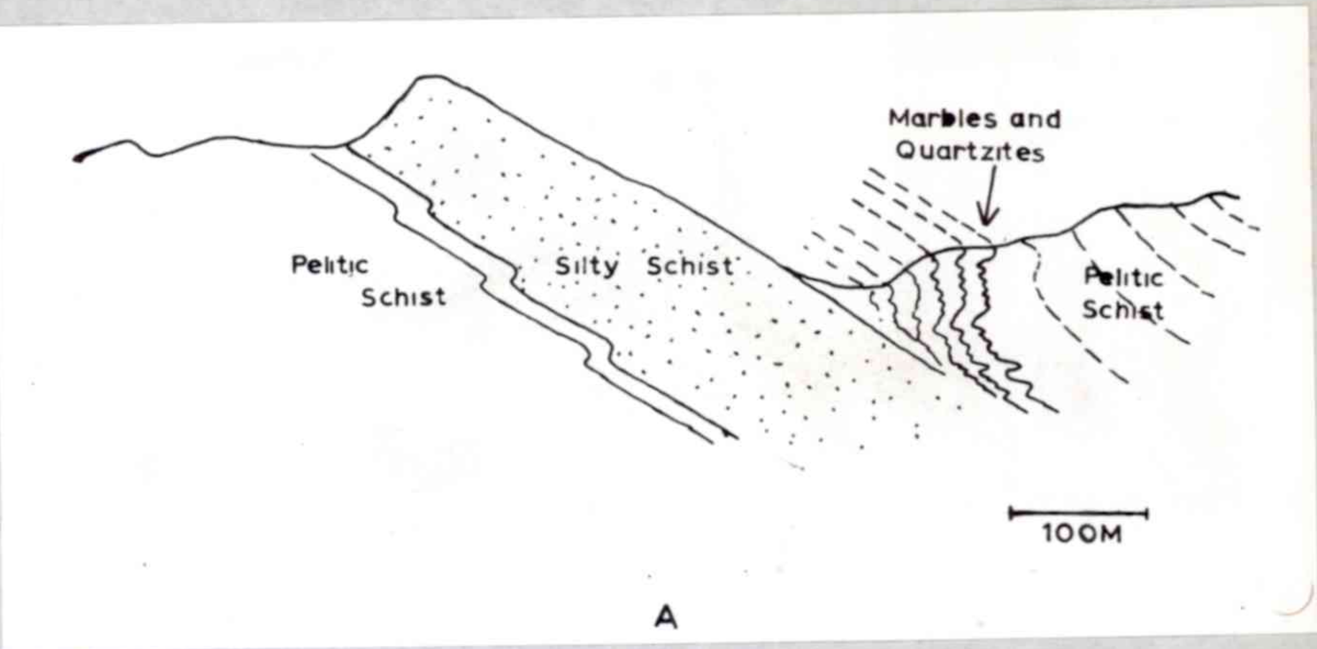


Fig.22. Section across the nose of the Kjeipen fold at Suppevand. The line of section is indicated on the map, fig.20.

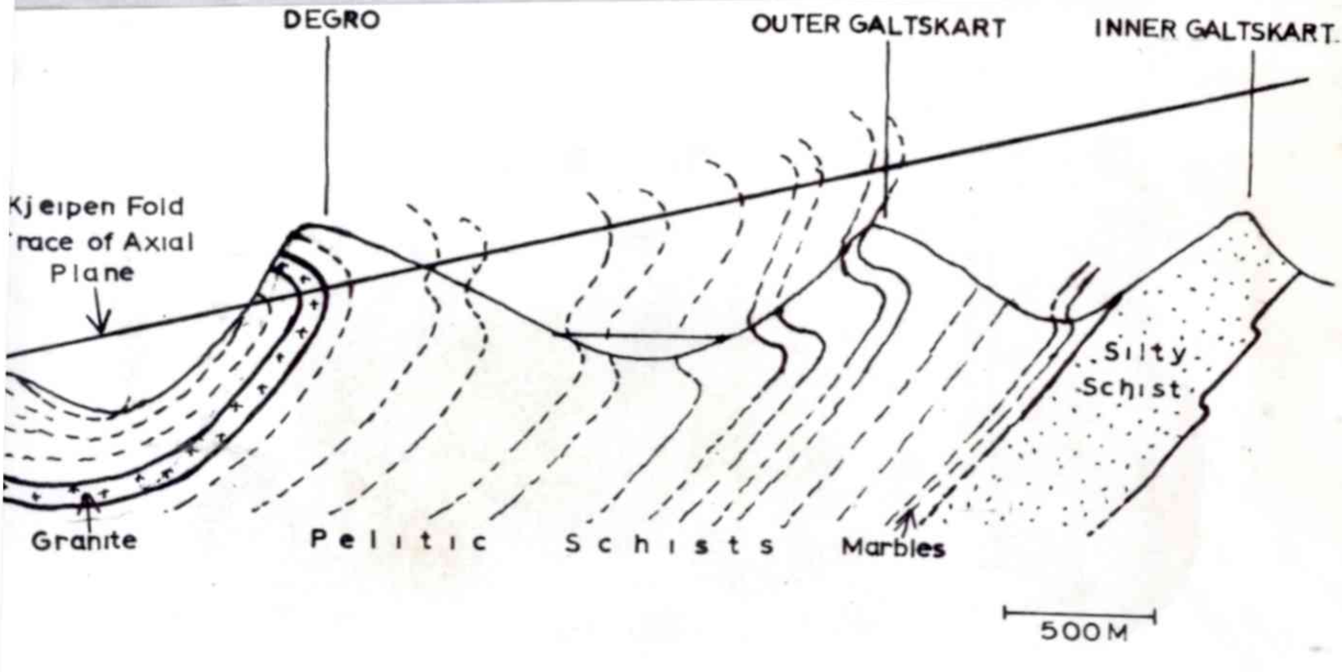
closure in the calcareous pelite which is not followed by the adjoining pelitic schist. From considerations of the regional structure, the calcareous pelitic schists of the Suppevand region must be situated within the core of the main Kjeipen fold. The symmetrical folds now seen cannot represent drag folds produced at the same time as the Kjeipen fold, because they have vertical axial planes, while that of the major fold is horizontal. They are therefore correlated with the F_3 period of deformation, and are thought to represent minor puckers on the flank of the Bjellåtind antiform, caused by flexuring of the relatively soft calcareous pelitic schists as the blunt nose of the silty schist moved slightly towards the south during the formation of the major F_3 fold. Just to the north of Kjeipen, minor overfolds with horizontal axial planes in calcareous pelitic schists are common (see fig. 45). These are much more intense than the structures with vertical axial planes seen to the south, and may represent structures produced at the same time as the Kjeipen Fold. Small-scale slides are commonly associated with the overfolds.

Other drag folds on the lower limb of the Kjeipen Fold, and in the correct sense for the observed closure are common in the Galtskart succession (fig.23A). In accordance with the major structure, they have gently dipping axial planes. The most intense of these folds occur in marbles and quartzites of west Galtskart (fig.48). All the drag folds observed in the Galtskart



A

Fig.23A. View of the valley between Inner and Outer Galtskart from the north. Drag folds with gently inclined axial planes, and situated on the lower limb of the Kjeipen fold, are prominent.



B

Fig.23B. Section from Degro to Inner Galtskart showing proximity to the axial plane of the Kjeipen fold. The monoclinal flexures on Outer Galtskart are interpreted as representing drag folds of considerable amplitude rocks are in the wrong sense and are too intense to be associated with the broad Bjellåtind Antiform. Moreover, the inclination of their axial planes indicates that they cannot be associated with the F_3 period of deformation.

The π S diagram of bedding and schistisity for 150 measurements for the Suppevand area (plate 1) which incorporates the closure of the silty schist on Kjeipen, has two well-defined maxima. These correspond to the two limbs of the fold, with dips to the W at 30° and the SE at 30° respectively. The latter is shown on the diagram as the much larger of the two concentrations, due principally to the location of most of the Suppevand region within the upper limb of the fold. From the diagram it is seen that the angular distance between the two limbs is 50°, a result close to that obtained from fig.9. Field evidence shows that the axial plane of the fold is nearly horizontal, and by construction from the π S diagram it can be seen to dip to the south-west at about 15°.

Even though there is a slight scatter in the B diagram of 43 linear structures, the maximum clearly falls as a pole to the great circle through the maxima of the π S diagram, indicating that the majority of the lineations are a consequence of the Kjeipen Fold. The reasons for the scatter of linear structures in this and all the other sub-regions is discussed in section 2p99.

In the North Markvand pelitic series, from outer Galtskart to Storvikvand, detailed mapping is difficult due to lack of continuous distinctive marker bands. On the north side of Storvikvand, a well-defined sheet granite outcrops, and probably continues southwards across the lake to connect with another sheet granite visible in the north face of Degro (fig.24). Here it becomes involved in an important overfold, closing towards the south-east, and with a gently dipping axial plane; the associated schists are similarly deformed. Parts of the schistose succession have a well-developed fracture cleavage, whose axial planes are approximately parallel with the major fold, and which was probably formed at the same time as the main fold. The direction of the closure of the fold on Degro is the same as the closure in the silty schist on Kjeipen, and their probable correlation is indicated in fig.9.

Southwards from Degro, towards Galtskart, the strike of the beds is constantly NE-SW, but the dips are high and variable in direction. On outer Galtskart two large-scale dextral drag folds are revealed by marble and calc schist marker bands within monotonous pelitic schists (fig. 23B). They have nearly horizontal axial planes, and are probably drag folds on the under limb of the main fold. The proximity to the axial region of the major Kjeipen Fold is shown in map 3. Similar structures probably exist in the Skromdalsvand region, and the rapid changes of dip may be due to folds of this nature, structures which are difficult to identify in relatively flat ground due to the attitude of their axial planes.

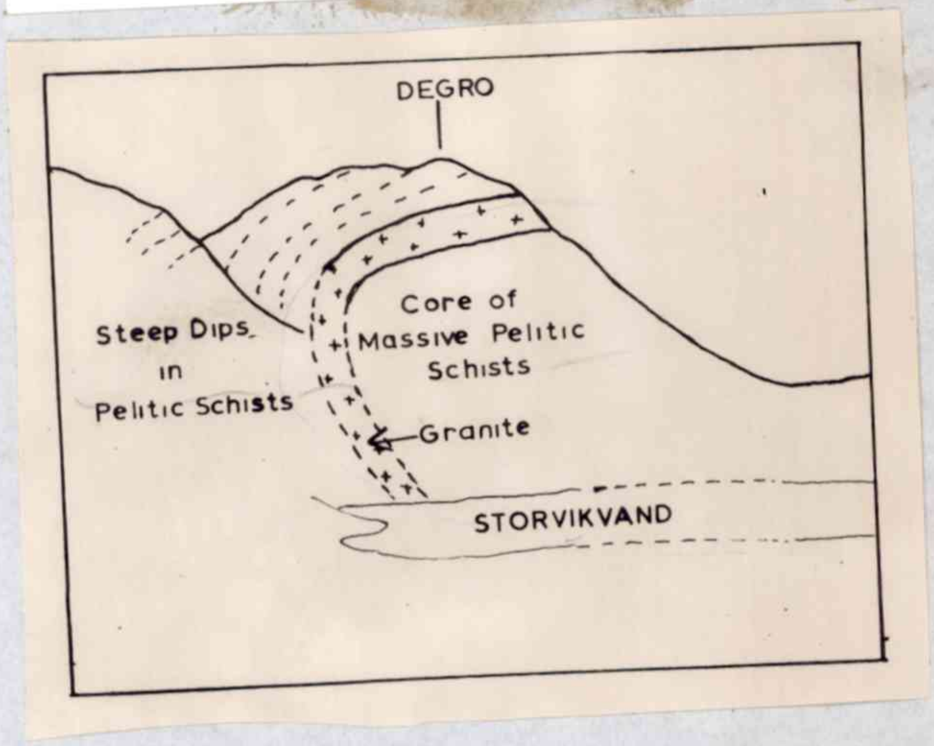
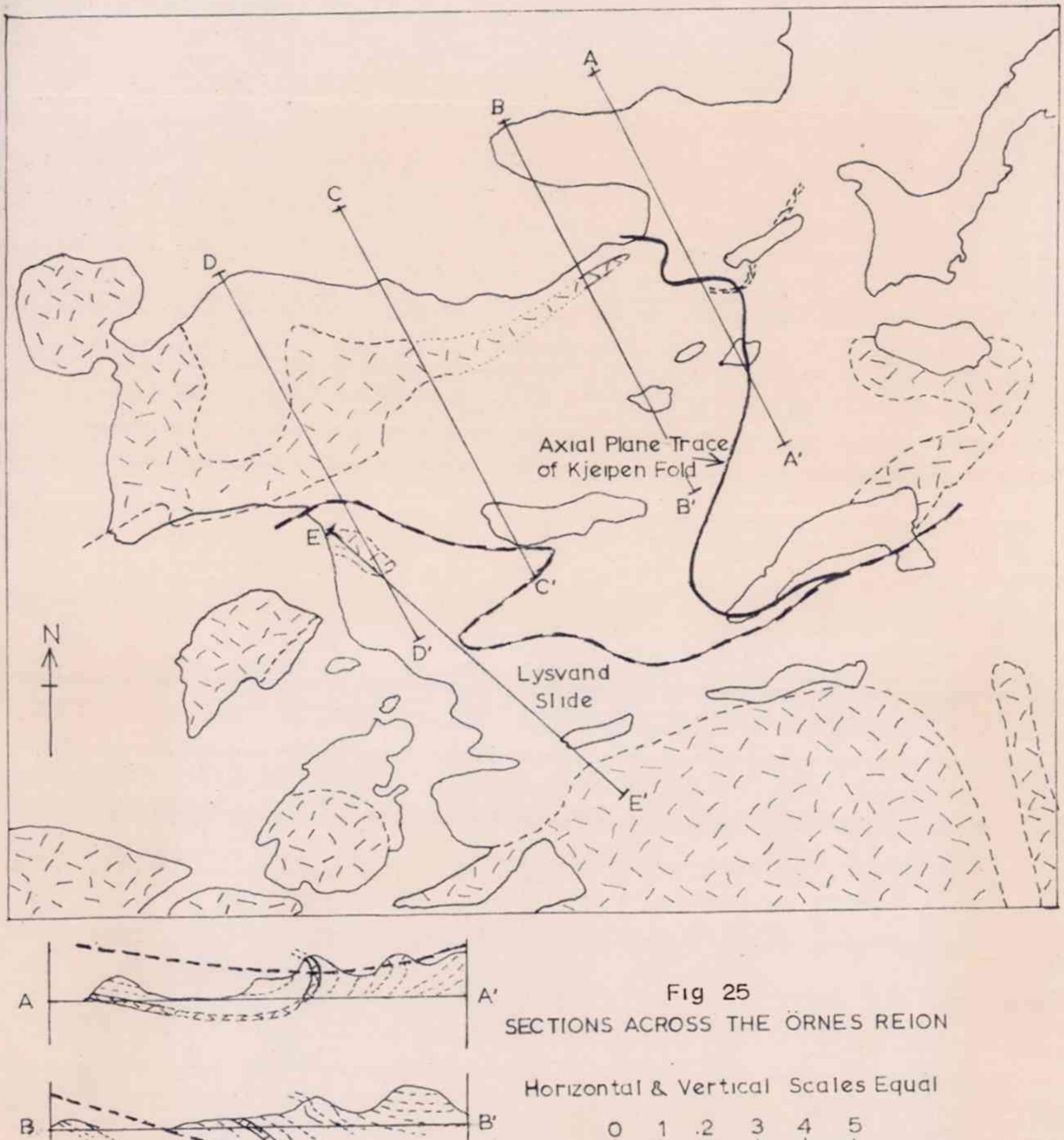


Fig.24. Degro from the north east, showing the most northerly closure of the Kjeipen fold identified. The sheet granite is conformable with the associated metasediments

Northwards from Storvikvand, there is no evidence of closures similar to those described above, and it appears that the whole of the Högnakken-Grötset massif is situated within the lower limb of the fold. The axial plane must then be above ground level.

When the North Markvand pelitic schists are traced SW from Skromdalsvand, the dip, although variable in amount, is constant towards the SE. Only in the SE of the sub-region does the Markvand synform lead to local reversals in the direction of dip (see p. 86). The reason for this constancy in dip direction is the location of the rocks on the upper limb of the Kjeipen fold. The axial plane of the Kjeipen fold which outcrops to the east, has a general south-westerly dip and will therefore project underneath this region. As the linear structures of F_2 age are undeflected in the region north of Markvand, there is no reason to suppose that later refolding has altered the position of the projected axial plane. In central Markvand, it will be some 1000 metres beneath the surface (see fig.25).

The π S diagram of 160 measurements from the whole of the North Markvand Region (plate 1) shows two maxima representing dips to the south at 40° and to the NW at 30°. Moreover, the extension of the maxima is on a great circle, the pole to which is represented by the maximum to linear structures, also by chance, based on 160 measurements, indicating the connection between the major fold and the measured linear structures. The more important of the two maxima of the π S diagram is that representing the



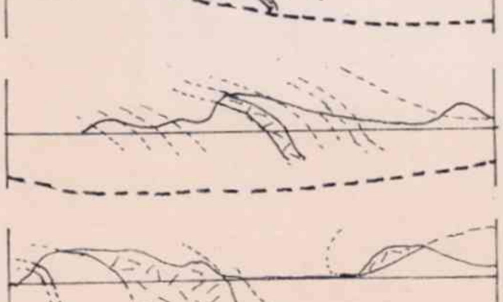
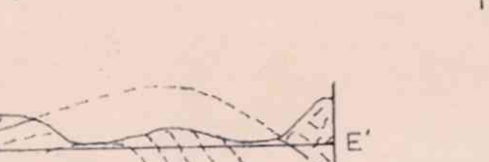


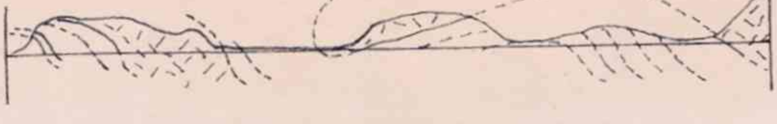
Fig 25
SECTIONS ACROSS THE ÖRNES REION

Horizontal & Vertical Scales Equal

0 1 .2 3 4 5
KM.

Trace of axial plane of Kjeipen fold





central and western part of the region situated on the axial regions and on the upper limb of the Kjeipen fold. However, dips from certain parts of the lower limb of the fold, notably around Storvik, fall in the same maximum due to refolding by the Storvik Synform. The other maximum represents dips from the eastern part of the region on the lower limb of the fold. Field observation shows that the axial plane of the major fold has only a gentle dip, and by plotting its position on the π S diagram, it is seen to dip westwards at about 15^{o}. This value is probably correct only in a limited region, as the Storvik synform must have some effect upon its inclination. In fact, if the position of the axial plane is constructed using this value, it outcrops once again on the Storviken peninsula, which can be seen in the field to be incorrect.

Within the region W of Skromdalsvand, zones of steep dips are occasionally encountered. The region is situated wholly on the upper limb of the Kjeipen Fold, and the structures probably represent monoclinal drag folds on one limb of the major fold. All are in the correct sense for this interpretation. The two most well defined of these belts occur on the north of Markvand (the Breitind Fold), and another on south Skraaven (the Skjeggen Fold).

A series of sections indicating the broad structural relationships of this region are shown in fig.25, in each of which is included the position of the projected axial plane of the

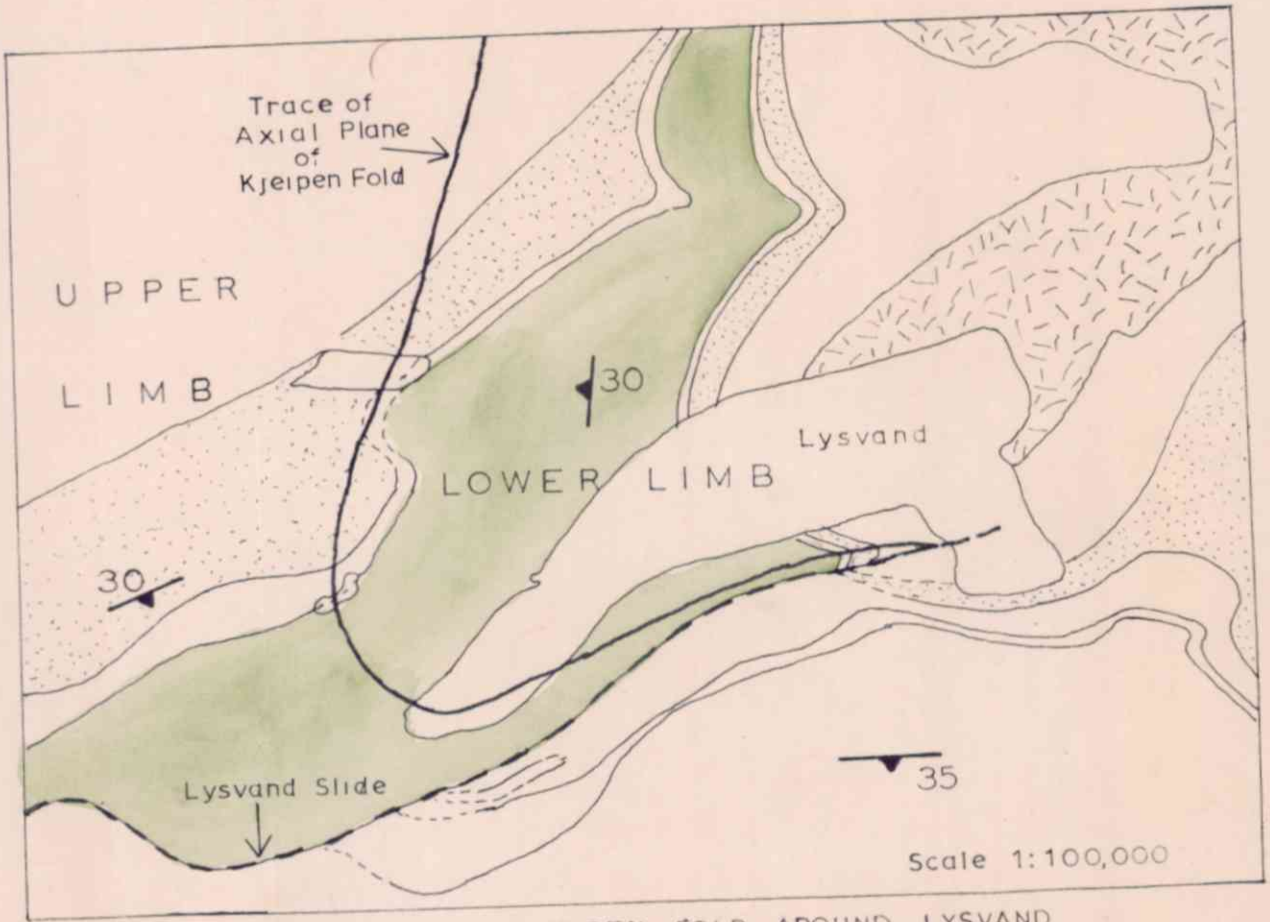
Kjeipen Fold. In the same way as the Galtskart succession is overturned in the Suppevand region, so the North Markvand pelitic series from the SE of Storvikvand is overturned on the north of Markvand.

Correlation of the lowest rocks of the Storviken peninsula with those of the Galtskart succession has already been mentioned in connection with the projection of the Galtskart Slide (see p. 43). If this is the case, the overlying granite-semi-pelitic schist sequence must correlate with the lowermost part of the North Markvand pelitic schists of Outer Galtskart. It has already been shown that the Skjeggen granite and schist sequence to the north must be wholly within the upper limb of the Kjeipen fold and separated from the Galtskart succession by the North Markvand pelitic schists. Therefore, the calcareous sequence of North Skjeggen cannot correlate directly with that of West Storviken; the former must close within the envelope of the latter, and the closure probably projects above Högnakken and Storvikvand.

The πS and B diagrams for the Skjeggen region show a more complicated pattern than that seen in the area to the east, due principally to the presence of major folds on N-S axes. There is a single maximum to the 125 measurements of poles to bedding and schistosity, representing a southerly dip of 40° and the complicated scatter around the maximum is due to these N-S structures. A similar relationship exists for the diagram of 46 linear structures, and again the greatest maximum is still due to structures associated with the Kjeipen fold, on bearings around 235°, and a mean dip of 25°.

Southwards from the Suppevand region, the influence of the Kjeipen fold quickly disappears, due to the action of the Lysvand Slide. Spanning the western end of Lysvand is a thick calcareous pelitic schist whose northern boundary, as already described, becomes involved in the Kjeipen fold. Convergence of the southern and eastern boundaries of the same group, which has a roughly triangular outcrop, occurs on the north face of Scetertind, and is shown in fig.26. The exact position where the two boundaries meet is inaccessible, and the nature of the junction is unknown, but when plotted on fig.9 it appears to represent a further closure of the Kjeipen fold. Within the calcareous pelitic schist in the region of convergence is clear evidence of a major closure (fig.3) in the same direction as the main Kjeipen fold, and which is truncated by the Lysvand Slide. The reason for the distinctive shape of the outcrop of the calcareous pelitic schist is due to later refolding of the structure around the Bjellåtind Antiform. Thus, the eastern and southern margins of the calcareous pelitic schist are correlated, the former situated on the lower limb of the fold, and the latter on the upper limb.

No further closures of the Kjeipen fold can be recognised south of this band of calcareous pelitic schist, and the outcrop of the rocks of the southern region is determined wholly by the Spilderdalen fold, and structures of the F_3 period of deformation. Separating the two regions is the Lysvand Slide, to be



RELATIONSHIPS OF KJEIPEN FOLD AROUND LYSVAND

Fig 26

Fig.26. The calcareous pelitic schist of West Lysvand is coloured green. Its distinctive outcrop is due to refolding of the Kjeipen fold around the Bjellåtind antiform.

described in detail later, whose general outcrop follows the southern boundary of the calcareous pelitic schist of Western Lysvand.

(2) Spilderdalen Fold

In the south of the Ørnes Region, where the structural trend is generally E-W, the dominant structure is an overfold closing northwards of the same style, and probably the same age, as the Kjeipen fold. It is separated from the latter by the Lysvand Slide. Like the Kjeipen fold, the structure has a gently inclined axial plane, which has been refolded by the later F_3 period of deformation. The position of the axial plane was plotted on map 3 in the same way as was that of the Kjeipen fold, namely by relating the points of verticality of fig.10 to their geographical position on the base map. Again, the sinuous nature of the axial plane is largely topographically controlled. Over most of the Ørnes region, only the lower limb of the structure is seen. Its upper limb outcrops in the eastern part of Steffodalen, and forms the whole of the eastern boundary of the Glomfjord granite.

The clearest demonstration of the fold occurs at the head of Spilderdalen, where the granite boundary, which dips gently eastwards under Glombraaen, becomes vertical, and then dips moderately steeply southwards along the whole of its northern outcrop (map. 1). Fig.27, which shows a view of the head of Spilderdalen, indicates that the metasediments associated with

Fig.27. The head of Spilderdalen from the west. The region is situated in the core of the Spilderdalen fold; the latter accounts for the pattern of the features visible

Fig.28. The conformable contact of the Bjellåtind granite with the overlying metasediments on the north face of Bjellåtind. The contact is sharp and occurs below the lowermost snow patches. Banding within the granite is clearly displayed

Fig.29. The contact of the Glomfjord granite with its associated metasedimentsin the nose of the Spilderdalen fold at Glombraaen. View from the north

the granite also follow the same structural pattern, with steep dips in the Steffodalen area. An alternative explanation is that the regional convergence of dips, represented by the north-easterly dips of the contact sediments to the Glomfjord granite and the southerly dips in the pelitic schists of Scetertind, represent a tight and symmetrical synform with a steeply dipping axial plane. However, in the presumed axial regions of this suggested fold, the sheet dip is everywhere vertical or nearly so.

Towards the north of Steffodalen, a series of distinctive marker bands occur on the north of the pelitic group, and show a series of major closures with E-W trends of verticality, and easterly plunging axes (see figs 18, 19B). They close in exactly the same way as does the Glomfjord granite margin, and when traced both westwards and southwards, form part of the regional succession away from the granite. The closures are therefore interpreted as a northerly continuation of the Spilderdalen fold, the direction of the axes of each successive closure being identical while the steep sheet-dips of central Steffodalen are regarded as being situated in the nose of the fold. The southern boundary of the main silty schist of Kvittind also follows the pattern of the Spilderdalen fold, but its northern margin is remarkably elongated northwards into the area mapped by Wells and Bradshaw, and becomes involved in a fold of a very different nature (see map 4).

Within the boundaries of the observed Spilderdalen fold, which cannot be traced north of Scetertind due to the action of the

Lysvand Slide, the detailed relationships are often complicated. This is particularly the case in the east of the Steffodalen region, where the Steffodalen Fold, an earlier structure, is encountered. Here minor folds of isoclinal character are abundant and lead to extreme complications (figs 36-40, p.). On the south side of Scetertind a series of large-scale isoclinal sinistral drag folds in marbles are visible (see p. , fig.30B) in the wrong sense for the Spilderdalen fold. They are associated with the F_1 schistosity in interbanded pelitic schists, which is parallel to their axial planes, a direction that is markedly divergent from the axial plane direction of the main Spilderdalen fold. Occasional dextral monoclinal flexures, with amplitudes of about 10 metres are visible on the southern flanks of Scetertind. They have gently inclined axial planes and are of an equivalent style to the major fold, and probably represent drag folds on its lower limb. Similar structures are visible in the schists in contact with the granite below Glombraaen.

The π S diagram for 80 measurements in the Steffodalen region shows two distinct maxima, corresponding to the two limbs of the major overfold. In the N, the mean dip is southerly at about 40°, while in the S and E it is towards the NE at about 35°. The scatter of the maxima is on a well-defined great circle whose pole corresponds with the maxima to 30 linear structures on a bearing of 110, with a 10° dip. This is the axis of the major fold, and suggests that the linear structures were developed at the same time. Construction of the axial plane from the π S diagram indicates that it dips to the SE at about 20°.

Towards the W, in the North Spilderdalsvand region, only the lower limb of the fold is exposed, due to the inclination of the axial plane of the fold, and the dip is everywhere S, with the granite structurally above the metasediments. The rocks consist principally of the Steffodalen pelitic series and exposure is not good; little is known of the details of minor structures. One isolated overfold in massive pelitic schists has the same style of deformation as the Spilderdalen fold and probably represents part of a dextral drag fold whose upper limb has been removed by erosion. Its axial plane has a similar inclination to that of the major fold (fig.44C). Most of the other minor folds are earlier structures, much more intense than that described above, and on different axes.

Structural data diagrams for the North Spilderdalsvand area indicate the position of the region on the lower limb of the Spilderdalen fold. 106 measurements to bedding and schistosity form a single concentration, within which the maximum is fairly diffuse. The mean dip is just to the E of S, at 40°, but variations in dip from 20° to 70° are common through an arc of about 60°. No obvious trends in the scatter of the π S diagram can be seen, but the slight elongation along a great circle whose strike is just E of N does correspond to the great circle in the Steffodalen region. The 40 measurements of the B diagram show the largest scatter in any single sub-region, due principally to refolding of earlier F_1 lineations. However, the greatest single maximum is almost due east with a 15° dip, and is a consequence of the Spilderdalen fold. A slight maximum with a westerly dip may be due to the proximity of the Spildervand culmination (see p. 91). Comparison with the B diagram for the Steffodalen region, shows that the maximum has moved from a bearing of 110 to 090. The reason for this is not clear but may be due either to original variation or later refolding.

On the coast south of Ørnes and on the island of Mesøen the rocks still dip constantly under the Glomfjord-Mesøen granite, even though the amount of dip is variable. Exposure is not good, and detailed observations have not been possible.

Variations in the amount of dip from 25° to 60° are shown in the π S diagram based on 56 measurements for the South Ørnes region. On the mainland there is only a slight variation from the mean southerly direction of dip. As for the North Spilderdalsvand region, there is a marked scatter in direction of the 20 linear structures measured, due principally to the identification of early F_1 linear structures. An easterly concentration is due to F_2 structures directly related to the major overfold and are mainly to be found in the east of the region. On Mesøen all the linear structures measured have westerly plunges, and again are probably related to the Spilderdalen Fold.

(3) Djupvikfjell Fold

The recognition of this fold and its significance in the regional structure depends on the stratigraphic correlations of the Ørnes and Blaatind regions. Lowermost in the succession on Blaatind, an area that is completely isolated by drift, is a calcareous pelitic schist. It is of identical lithology to the calcareous pelite of North Djupvikfjell, and although the regional dips of the two schists are nearly at right angles (southwards on Djupvikfjell and almost westwards on Blaatind) they appear to connect up across the valley NE of Mosvold. Such a projection is illustrated in fig.9, and is suggested by a reasonable correlation of stratigraphic successions (see section 2 b 170 and fig.60). In each case there is a schist sequence of comparable lithologies lying underneath a granite mass. The granites themselves are closely comparable.

An approximate prediction of the position of the axial plane of the Spilderdalen fold when projected westwards may be made, knowing its inclination in Steffodalen and allowing for the effect of the Spildervand culmination. Assuming the axis to maintain its bearing of 110 and dip of 15° on Steffodalen, the position of closure of the granite will be some 2000 metres above ground level at Djupvikfjell. The effect of the Spildervand culmination will be to bring the axial plane down towards ground level, and assuming that the observed 15-20° dip of regional lineation is representative of the major fold, it should reappear on the Blaatind-Teksmona massif. Thus, the Teksmona-Blaatind granite is interpreted as representing the reappearance of basal granite in the lower limb of the Spilderdalen fold, due partly to the effect of the Spildervand culmination, but also to other F_3 folds on different axes. The nose of the fold is interpreted as being situated in the unexposed ground around Reipaa. (Fig 25)

Once these broad relationships described above have been recognised, it is possible to give a rational explanation of the steep dips observed in the Ørnes gneisses. These comprise a steep limb of an essentially monoclinal fold, which has affected part of the under limb of the Spilderdalen fold. The monocline is asymmetric, with gentle dips in the north where the Blaatind rocks occur and a steeper southerly limb in the Bugten region.

The B diagrams for both the Blaatind and East Ørnes regions, which are affected by the Djupvikfjell and Spilderdalen folds each show an easterly plunging maximum, with a 15° dip, which corresponds with the constructed axis of the Djupvikfjell fold. The main concentration from 50 measurements in the πS diagram for the Blaatind region represents the gentle upper limb of the monocline. That of the East Ørnes region (based on 25 measurements) indicates a southerly dip of 80° and is located on the steep limb of the monocline. The planar structures in this region are mainly caused by the arrangement of pegmatites in the long limbs of F_1 isoclines, and by the gneissic banding which is parallel to these. That they are representative of the regional 'bedding and schistosity' is shown by the conformable contact with adjacent schists on South Djupvikfjell. Finally, the southern limb of the monocline is represented by the rocks in the South Ørnes region, where the mean dip as calculated from 35 measurements, is towards the S at 40°.

(4) F_2 folds in the Skjeggen-North Markvand region

A series of monoclinal flexures, representing drag folds on the upper limb of the Kjeipen fold, have already been described in the section on the Kjeipen fold (p. 67).

(4) Bolden Fold and associated structures

In the metasediments on the N of Skjeggen, linear structures on bearings between 160° and 190° are seen. These are generally associated with sheet dips to the SW, varying in amount from 60 to 90°, an arrangement that is quite unlike that on the mountain itself. Little time for detailed mapping of the region was available, and as the area is not covered by aerial photographs, and is poorly exposed, only the broadest features could be determined. The tentative correlations of the region are shown on the main geological map, where the connection between the steep south-westerly dips of the coastal succession with the more gentle south-easterly dips of the Skjeggen massif is thought to be due to an antiform. That this structure continues southwards is shown by a calcareous pelitic schist that can be traced on the west side of Skjeggen and crosses over the valley, to reappear on

Skraaven; the dips indicate an antiform plunging southwards. Similarly, the northern boundary of the Skjeggen granite crosses the valley via a steeply dipping zone to reappear on Skraaven. These steep dips are part of the Skjeggen fold on a different axis, and possibly of a different age from the Bolden antiform, and it is in this region that the two structures converge.

Exposures of granite on the coast north of Bolden dip steeply eastwards and a zone of steep dips can be identified southwards from this, on the west side of Skraaven. The only other structural information on this region comes from the west of the Kunna peninsula, where the rocks dip gently eastwards.

There are two possible explanations to account for this arrangement:-

(a) the structure is an asymmetric synform, complementary to the Bolden antiform,

(b) the rocks at Kunna are overturned with respect to those at Skjeggen, and the total structure is a major overfold, with a gently inclined axial plane, of equivalent style to the Kjeipen and Spilderdalen folds.

It is the latter alternative the writer favours.

The Skjeggen granite is regarded as basal granite detached during F_1 times (p. 95). It structurally overlies metasediments, suggesting that it is overturned. On the second explanation, therefore, the rocks at Kunna are the right way up, and it is the Bolden overfold which accounts for this inversion. All the islands to the W and SW of Kunna are mapped by the Survey as granite, and as they cover a very large area, the most likely explanation is that they represent basal granites. Thus, the balance of evidence appears to be in favour of interpreting the structure as a major overfold, with the rocks of Kunna and all the islands to the W forming the lower limb. The axis of the fold is south-easterly, on a very different trend from the Kjeipen fold. Detailed relationships between the two structures are unknown, due to the poorly exposed nature of the terrain.

(6) Laksaadalen Fold

On the northern shores of Laksaadalsvand, vertical quartzites with an E-W trend are encountered. When traced towards the N, the sheet dip is consistently southwards, although exposure is poor and the details obscured. This region has been mapped principally by Wells and Bradshaw, and interpretation as to its structural arrangement can only be determined with reference to the areas N and E. It appears, however, that there are two possible explanations to account for the zone of steep dips:

(i) the existence of a tight synform on the northern margin of the granite,

(ii) a major overfold with similar characteristics to the Spilderdalen fold.

Regardless of which of these is the correct interpretation, it appears that the fold is of F_2 age, with an easterly plunging axis, and a well-developed associated lineation.

F_3 Folds

Superimposed upon the earlier structures are broad-scale and gentle warps with steeply inclined axial planes. Certain of these can be recognised directly, the most important being the Bjellåtind antiform, but in the less well exposed regions, particularly in the west, the positions of the axial planes have to be inferred. In every case where folds with gently and steeply inclined axial planes are closely associated, it can be shown that those with steep axial planes are later than or refold those with gently inclined axial planes. This affords a method of distinction between F_2 and F_3 structures.

(1) The Bjellåtind Antiform

The way in which the Kjeipen fold is domed over the Bjellåtind granite is shown in fig.9, and the structure is termed the Bjellåtind antiform. Only a small part of the structure is visible in the Ørnes region, and it is responsible for the appearance of the Bjellåtind granite, and the culmination which extends into the Sörfinnset region. Here it is termed the Sörfinnset anticline, and has a N-S axis. It accounts for the westerly sheet dips of the Grötset-Galtskart region, and the eastward dips of the rocks W of Sörfjord. When the arrangement of the part of the fold that occurs in the Ørnes region is constructed (from the 30° westerly dips of Galtskart, and the 30° SW to S dips of South Lysvand) it is seen to have a SW axis, plunging at 20°, and a near vertical axial plane. This axial direction is displaced from that of the main Sörfinnset Anticline, and indicates the major structure to be periclinal.

Towards the SW of the Lysvand region, the antiform rapidly loses its identity, and the regional dip is determined wholly by the major F_2 overfolds. The series of symmetrical folds in the calcareous pelitic schist of the Suppevand region which have vertical axial planes are correlated with the Bjellåtind antiform, and are thought to have developed by crumpling of the relatively soft calcareous pelites in contact with the massive core of the Kjeipen silty schist (see also p. 60 and fig. 22). The direction of axes of the folds corresponds with that of the Bjellåtind antiform.

A possible continuation of the same structure is represented by the Örnes fold (p. 87), the structural details of which are less well known. In addition, the Spildervand culmination (p. 91) with an essentially N-S axis, is probably of the same general age, and may represent a contemporaneous cross-fold (King and Rost 1955).

(2) The Storvik Synform

On the W side of the Sörfjord the sheet dip is towards the NW or W, due to the situation on the lower limb of the Kjeipen fold. When traced north-westwards, however, on to the Högnakken-Grötset region, the beds are almost horizontal (fig.16), and further to the NW, on the Storviken peninsula, dip gently southwards. This gentle synform, with an axis which plunges to the SW at 15°, has an axial plane that is nearly vertical, and is comparable in style to the Bjellåtind antiform. It is the complementary synform and was probably formed during the same general period of deformation.

Towards the SW, the fold cannot be traced for any great distance. The rocks here dip consistently to the SE and are situated wholly on the lower limb of the Kjeipen fold.

The lowest rocks seen in the W of Finneskua, consist of calcareous pelitic schists, lithologically identical to those in the Galtskart succession. It appears that they are lateral equivalents, and that their reappearance is due to the synform, a view that is supported by the mapping of Wells on the east of Grötset (map 4). The structural importance for this correlation is discussed elsewhere (p. 68).

(3) Markvand and Blaatind Synforms

Between the Bjellåtind antiform and Storvik synform, two gentle F_3 structures are developed. On the Outer Galtskart-Molndalstind ridge, a zone where the beds are horizontal, accounts for the change in sheet dips (map 1). The structure is synformal, and has an axial plane that dips steeply to the SE, with a SW trending axis. The structure is not a simple synform, and a whole series of minor ripples arranged en échelon can be identified in the axial region. Towards the SW, a thick marble band marks the core of the fold. The structure is termed the Markvand synform.

In an analagous way to the Storvik synform, the Markvand synform only occurs over a limited region, probably due to its situation on the pre-existing Kjeipen fold. Towards the NE, it dies out where the steep dips of the South Skromdalsvand region in the closure of the Kjeipen fold are encountered, while to the SW it probably also disappears when the regional dip is wholly south-eastwards.

Along the projected axis of the Markvand synform, another similar structure is seen in the Blaatind massif. The arrangements of dips on the mountain indicates a very gentle synform with a steep axial plane and an axis which plunges towards the SW at 18°.

(4) The Ørnes Fold

The structural interpretation of this region is based upon mapping that was considerably less detailed than the regions to the E, using an enlargement of the 1 : 100,000 Norwegian topographic map; large tracts of unexposed ground complicates matters. Any conclusions as to the regional structural arrangement are therefore highly tentative.

In the section on the Spilderdalen fold, the westerly projection of the latter was discussed (p. 77), and it was concluded that the granite could reappear in the Teksmona-Blaatind massif. A cross-section across this region (fig.25) compares closely with that of Vogt in 1912 and the correlation of metasediments dipping underneath granites has already been outlined.

The structural arrangement of the Blaatind and South Ørnes regions indicates an antiform, the Ørnes antiform, with an axial plane that dips steeply to the NNW with an axis plunging at 19°.

On this interpretation, the island of Mesøen should be situated in the nose of the antiform, and there should be some indication that the granite and schist sequence does connect up with that on Blaatind. Only a single day was available for mapping the east coast section, principally to determine the correlations of the succession with that on the mainland, and to fix the position of outcrop of the Ørnes Gneisses. In the south of the island, the granite boundary is a direct continuation of that of the Glomfjord granite seen on the mainland only 2 km away; its outcrop on the west of the island is taken from the existing survey map. The bend in outcrop is wholly topographic. The succeeding group of metasediments is broadly similar to that on the mainland (see p. 169) and the dip is still towards the south at about 40°. A little further N, however, the strike changes rapidly, until the rocks dip towards the SW, in a fairly well defined zone aligned NW-SE across the island. This is exactly the direction to be expected on the present interpretation, and connects up with local strike directions in the Teksmona granite. However, on the north of Mesøen the strike changes back to almost due E-W, and quartzitic rocks project directly into gneisses on the mainland. The reasons for both these phenomena are not clear, but it suggests that the Ørnes fold is not a simple antiform. In addition, a series of isoclinal folds have been identified in the quartzites of Mesøen, probably of the same age as those in the gneisses.

(5) The Reipaa Synform

In the same way as the recognition of the Ørnes Antiform depends on a correct interpretation of the local stratigraphy, so the establishment of this fold is entirely dependent on the extrapolations through the unexposed ground west of Markvand. Correlation of the Djupvikfjell succession with that of lower Blaatind has already been mentioned (p. 79), and it is the western extrapolation of the succession that is of prime importance. Two possibilities exist:-

(i) the succession continues with an E-W strike, and disappears out to sea at Reipaa. It is this alternative the writer favours.

(ii) the succession connects with that of South Breitind, by means of a major overfold, correlating the Blaatind with the Skjeggen granite.

The regional dip for the critical region to the W of Markvand is steep and towards the south, except for very occasional measurements on the southern flanks of Breitind. These may have been recorded from blocks that were not in situ, but are anyway of very local significance. Thus there is little structural indication of a major fold closure, and as the succession is very varied lithologically and no bedding cutting across the

the measured schistosity has been observed, the possibility of there being an F_1 isoclinal fold with associated axial plane schistisity appears to be remote.

Moreover, the northern boundary of the Skjeggen granite has been shown to connect with the granite on Skraaven by a steep NE, SW belt, while a small band of pelitic schists, of similar general lithology to the south Breitind pelitic schists has been identified on the extreme south-west tip of Skraaven, again with a NE-SW alignment. Finally, the schist sequence of south Breitind, on the contact with the Skjeggen granite, is quite unlike that on Blaatind.

The two granites are therefore interpreted as being separated by a band of schists which outcrops along Markelven.

From the arrangement of the steep southerly dips of West Markvand, and the more gentle westerly dips of Blaatind, the rocks of the latter must overlie those of the former (see fig.25), and the structure must be synformal. Using these two regional dips as representing the limbs of the fold, its axis is seen to plunge SW at 18°, and its axial plane to dip steeply NW; these are comparable properties to the Ørnes antiform to which it is the complement.

Both the Reipaa synform and Ørnes antiform are upside down structures if the basal stratigraphic position of the Glomfjord granite is accepted. The minor Blaatind synform, on the common limb to the two structures has already been mentioned, and a suggested correlation made with the Markvand synform. Reference to map 3 shows how the axis of the Ørnes antiform corresponds almost exactly with that of the Bjellåtind antiform, while the Reipaa synform corresponds fairly well with that of the Storvik synform. The structures are all of the same age, and correlation of the two sets of structures is suggested.

(6) F_3 Folds in the Steffodalen Region

No large-scale F_3 structures are seen in the region, but a series of small symmetrical folds with vertical or steeply inclined axial planes, analogous to the folds in calcareous pelitic schists of the Suppevand region are common (fig.49D). Like the Suppevand folds, they are remarkably persistent, and can sometimes be traced for distances up to a kilometre. Their direction is constant, and is just S of E, parallel with the axial direction of the major Spilderdalen fold.

(7) Spildervand Culmination

The linear structures in the S of the Ørnes region have a general E-W trend. However, in the W, the predominant trend is westwards, and in the E eastwards. The culmination of linear structures is not well-defined but must have an approximate N-S axis passing through Spildervand. It is probably responsible for the reappearance of the basal granite in Blaatind (see p. 79), and may represent F_3 cross folding produced at the same time as the Ørnes and Bjellåtind folds. This phenomenon is seen in other complexly folded belts and has been described from the

Scottish Highlands by King and Rast (1955) and is regarded as being due to accommodation by constriction of the material when folded at depth. Sutton and Watson (1954) have described a series of oblique subsidiary folds on the limbs of major folds. They have shown that these oblique folds are due to variations in original plunge of the major fold, and are therefore coeval with the major folds. Whether or not the Spildervand culmination is an F_2 structure, comparable to those described by Sutton and Watson, or an F_3 cross-fold, of the same age as the Ørnes and Bjellåtind folds cannot be determined on the information available.

The Slide Zones

Two main groups of slides can be established, according to the age of their development.

(1) During F_1 times, a large amount of thrusting took place, distributed rather evenly through the succession. Concentration of movement occurred along certain boundaries of lithological contrast, and such zones can be postulated from structural and stratigraphic considerations.

(2) A single example of an F_2 slide has been identified separating the Kjeipen and Spilderdalen folds. The stratigraphic significance of the slide is probably slight, as the lithological groups on either side are identical.

F_1 Slides

(1) Basal Slide

In both the groups, later metamorphism has completely obliterated indications of structural dislocation, and uniform schistosity extends across all the postulated slide.

The recognition of isoclinal folds in the basal granites has already been outlined; similar isoclines can be identified in the overlying metasediments but it is only rarely that the two are interfolded on an isoclinal style. It may be inferred therefore that separating the two broad groups is a slide zone of some importance. The basal 30 to 40 metres of the metasedimentary succession above the Bjellåtind granite shows no evidence of isoclinal folding, and is not repeated on a major scale higher up in the Galtskart succession. It is remarkably constant even in detail over the limits of outcrop of 3 to 4 km, and is regarded as having remained approximately in contact with the granite throughout all periods of deformation. The major basal slide probably occurs above this zone, and accounts for the elimination of the isoclinal folds of Galtskart, and the lowermost fold on Kvittind.

(2) Galtskart Slide

Just as the basal slide separates the Galtskart isoclinal folds from the basal granite so has the Galtskart Slide eliminated the continued repetition of the succession upwards. In the valley between Inner and Outer Galtskart, a thick marble and quartzite

band

band separates the Galtskart succession from that of the North Markvand pelitic group. Both lithologically and structurally, the two series are different (see also pp. 131, 173). The lithological differences are described in section 2 b 170 while the main structural difference is the presence of a series of major isoclinal folds in the Galtskart succession and the lack of large-scale repetition by isoclinal folding in the North Markvand pelitic rocks.

The marble-quartzite sequence, which itself is very highly deformed often with an intense fracturing of the quartzite members, is taken to represent the position of the slide zone, although its exact location is unknown. On map 3 the slide is represented by the outcrop of this distinctive group. The situation of the slide on the upper limb of the Galtskart fold is also shown in map 1 which indicates the extreme thinning of the successions both northwards towards Sörfjord, and westwards towards Markvand. Both these regions are situated on the limbs of the Kjeipen overfold , but as described earlier (p. 51), the structure does not appear to be of the style to produce such thinning, particularly in massive bands such as the silty schist. It is suggested, therefore, that thinning in both these regions is a consequence of the F_1 slides, and in the Galtskart region was associated with the development of the Galtskart fold. A similar thinning and partial elimination can be seen in the case of the North Markvand pelitics, and in this group, on the north of Outer Galtskart, extreme F_1 sliding is indicated by irregular lènticular masses of marbles and amphibolites.

(3) Steffodalen Slides

These structures account for the isolation of the Steffodalen fold in the upper limb of the Spilderdalen overfold (fig.14) and are described on p. 45 .

(4) Skjeggen Slide

The main reason for the recognition of this slide is the stratigraphic similarity between the Bjellåtind granite and the metasediments on Galtskart, and the Skjeggen granite with the associated metasediments to the north (fig.64). A partial elimination of the North Markvand pelitic series against the southern boundary of the granite, which represents the position of the slide, is shown in map 3. As the boundary is not well exposed, it is impossible to observe the nature of this elimination. The Skjeggen granite, when projected eastwards, connects up with the much thinner Stortind granite; the latter is entirely eliminated a little further east. This is regarded as being controlled by the Skjeggen slide.

The slide itself is affected by the Kjeipen fold, indicating that it is of F_1 age. Other similar structures probably account for the southerly elimination of sheet granites of the Storvik region and may account for the presence of the granites in the metasedimentary sequence (see also p. 48).

The Lysvand Slide

The diagram of linear structures for the whole of the Ørnes region shows two distinct maxima (fig. 8B), corresponding to south-westerly trending axes in the majority of the area, and E-W trends in the southern margin of the Ørnes region. Considerations of the major structures show that each of these regions is dominated by a single major overfold, the Kjeipen and Spilderdalen folds respectively, which from examination of the attitudes of axial planes and style of deformation are regarded as having developed at the same time. The region of convergence of these two folds is a well defined zone which follows the southern boundary of the calcareous pelitic schist of West Lysvand, and corresponds almost exactly with the change in trend of the associated linear structures. Detailed information is available only for a small part of the zone, from North Kvittind, to East Scetertind, as the region of projected outcrop on to Djupvikfjell and Blaatind is only poorly exposed.

The clearest demonstration of the structure is on North Scètertind, where the closure in calcareous pelitic schists is truncated by a massive band of silty schist (fig.3). The latter is quickly eliminated towards the west, and the slide must follow the lower boundary of the silty schist. Eastwards, its effect is no longer visible, but it accounts for the absence of the fold on Kvittind; here the slide is conformable with the banding.

South-westwards from Scetertind, the southern boundary of the west Lysvand calcareous pelite is not well exposed. However, a further distinctive band of calcareous pelitic schist, which outcrops on North Scetertind, and follows the Spilderdalen fold is eliminated westwards in the poorly exposed ground north of Digermulen. There are two explanations to account for this:-

(a) the calcareous pelite forms the core of an isoclinal fold, which closes westwards,

(b) it is eliminated against the Lysvand slide in a similar manner to the silty schist further to the west.

On the southern side of the North Scetertind calcareous pelite is a distinctive group of rocks of undoubted stratigraphic significance (fig.60). These are not seen on the north of the same band, and if an isoclinal fold does exist, the succession must have been eliminated by sliding. In addition, the same group of distinctive rocks is seen further to the west, this time bordering upon the West Lysvand calcareous pelitic schist, suggesting stratigraphic correlation (see also p. 175). Because of these factors the writer favours the interpretation of the western termination of the calcareous pelitic schist as due to the action of the Lysvand slide, in a similar way to the silty schist of North Scetertind. On this interpretation, the observed change in strike of the calcareous pelitic schist near its westerly termination, may represent a drag fold produced during the formation of the slide, while the rapid way in which the fold dies out south-westwards would then be expected. The geometrical relationships on North Scetertind suggest a movement of the southern block towards the west, and the direction of the fold in the calcareous pelite supports such a contention.

In the region of Ørnes the gneissic rocks are thick but north-westwards and eastwards their presumed lateral equivalents are considerably thinner. It is possible that the Lysvand Slide has had some influence upon the variation in thickness of the group (see also p. 284).

Westwards from Ørnes, the Blaatind region is again dominated by east-west linear directions and is also within the influence of the Spilderdalen fold. For this reason, the westerly continuation of the Lysvand slide is regarded as following the western boundary of the calcareous pelitic schist of East Blaatind.

The Lysvand slide is the only large-scale structure that in one place at least is demonstrably cross-cutting. It is associated with two major F_2 folds and is probably of the same general age.

Effects of Refolding on Linear Structures

The majority of linear structures in the Örnes region were developed during the F_2 period of deformation, and their orientation depends upon whether they occur in regions dominated by the Kjeipen or Spilderdalen folds. Later folding has been on axes almost the same as those of the F_2 major folds, and it is only possible to assign the linear structures to the F_2 phase on considerations of style and intensity. Slight scatter of the F_2 linear structures does exist, however, and can be caused by one or both of two factors:-

(i) original variation

(ii) later refolding.

The effects of refolding of earlier structures have been described previously (Weiss 1959, Ramsay 1960). Both these authors have outlined a method to explain the effects of refolding of linear structures. According to whether the later folding is concentric or similar in style, the early linear structures will lie on a small circle or a great circle around the new axial direction respectively. In the majority of cases, the F_2 linear structures are refolded by F_3 folds on almost identical axes, and the scatter may be difficult to identify, or to separate from original variation. Occasionally, F_1 linear structures are refolded by F_2 folds on very different axes, and the effects are easier to observe. These are described in some detail at the end of this section.

(a) Sub-regions affected principally by F_2 linear structures

(1) Blaatind Region. Two maxima are seen in the B diagram, corresponding to plunges of 262 at 16° and 220 at 25°, the first of which is by far the more important. The lesser of these, consisting of only five measurements, represents a restricted area in the south. When the axis of the Blaatind synform, an F_3 structure, is plotted, it falls in the extreme southern part of the major maximum, and bears no simple relation to the arrangement of F_2 linear structures. For this reason, the observed scatter is regarded as representing original variation that appears to have no aerial control.

(2) North Markvand Region. A single well-defined maximum to 160 measurements of the B diagram is seen, and the axes of the later F_3 folds are on bearings that are almost identical. No aerial control to the slight scatter could be determined, and the few measurements in the north-east quadrant are evenly distributed through the whole area. As the F_3 folds are on the same bearings as the F_2 linear structures, no scatter due to refolding is to be expected, and the total slight variation observed is probably original.

(3) East Örnes Region. Lack of detailed structural information prevents any conclusions from being drawn. The single maximum is in an easterly direction, and does not appear to have been affected by refolding.

(4) Suppevand Region. The B diagram shows a marked scatter

which can be related to separate geographical regions. As a generality, those that fall on the more easterly part of the maximum are confined to the W of the region, and vice versa, while the single sub-maximum is caused by linear structures from a restricted area on the N of Kjeipen. The nearest large-scale F_3 structure is the Bjellåtind antiform, whose axis falls in the centre of the main maximum. As the axes of both F_2 and F_3 folds are very nearly equivalent, the scatter observed cannot be due to refolding. The aerial control on the scatter must therefore be due to original variation in the F_2 linear directions.

(5) Lysvand Region. A similar systematic variation in the orientation of the B diagram can be shown to occur in this region. Thus, the linear structures on bearings around 250° predominate in the South Lysvand region; those on 230° correspond to the rocks of East Galtskart where dips are low, and those on 210° to the steeper dips of West Galtskart. When the position of the later F_3 Bjellåtind antiform is plotted, it falls midway between the 210 and 230° maxima, and a small circle between the two, using the F_3 axis as centre, can be constructed. The 250 maximum is a long distance from this small circle, suggesting original variation, while it is possible that the observed scatter on Galtskart is a direct consequence of the Bjellåtind antiform.

(6) Steffodalen Region. Two distinct linear directions prevail on bearings of 110 and 125°. The former predominate in the N and W of the region, and the latter in the SE, particularly

around the Fykan granite. The style of the linear structures is constant, and all probably belong to the F_2 period of deformation. The constructed axis of the Spilderdalen fold falls in between the two maxima, and as there are no major F_3 folds, the scatter is regarded as being due to original variation in axial direction of the linear structures.

In the remaining sub-regions the linear structures do not show such a simple arrangement. For two of these, the complications are due to F_1 linear structures on variable axes, but in the North Skjeggen region, they are caused by F_2 structures on varied axes.

(7) Skjeggen Region. Four distinct maxima can be recognised in the pattern of the B diagram, one of which is the dominant. Just under half the total of linear structures measured are grouped around a mean of 230°, corresponding to the areas influenced by the Skjeggen and Kjeipen folds. Another maximum, around 150°, is caused by linear structures in the North Skjeggen region and is related to the Bolden fold. Only a few structures on bearings around this value are represented, but due to the rapidity of mapping, this is unrepresentative. The two remaining maxima, one plunging E and the other W, occur in the Breitind-N.W.Markvand region, and are related to the monoclinal flexures on the upper limb of the Kjeipen fold. The reason for the variation of this trend from that of the regional linear direction is not known.

If great circles are drawn on the π S diagram for each of the recognised folds, their poles correspond accurately to the maximum of the associated minor structures. In each case, the folds have one limb represented by the main maximum in the π S diagram, and consequently, the great circles for each of the folds pass through approximately the same point.

(b) Sub-regions affected by both F_1 and F_2 linear structures

(1) South Ørnes Region. Few linear structures have been recorded from this region, but those that have been show a marked scatter. The two most important maxima are easterly and westerly plunging respectively, and are of F_2 age. Their variation is due to the Spildervand culmination, and in the W of the region, particularly in Mesøen, the linear structures dip W, while on the mainland and in the E of the region they dip E. The amount of separation of the maxima is 25°, and is probably a measure of the intensity of the culmination.

The remainder of the scatter is due to structures in the region to the E of Bugten, where southerly trending linear directions are spasmodically developed. Some of these are of F_1 age, but some are very gentle puckers on nearly N-S axes and may be correlated with the Spildervand culmination. No trend in the F_1 structures can be observed.

(2) North Spilderdalsvand. It is this area that shows the largest scatter of linear structures of any of the sub-regions. The π S diagram, on the other hand, shows a single maximum due to the position on the lower limb of the Spilderdalen fold; the scatter within this one maximum is considerable, but cannot be

related to separate geographical areas.

The largest maximum on the B diagram represents easterly plunging linear structures, related directly to the Spilderdalen fold, which are found throughout the region. Occasional westerly plunging axes of the same age are found, principally in the W of the region, and may represent an approach to the Spildervand culmination whose main axis is some distance to the W. Linear structures on bearings around 070 are found principally in the N part of the region, towards the Lysvand slide.

An important series of linear structures are found in the S of the region, on axes varying from 130° to 190° and generally with high dips. They consist principally of axial directions of intense F_1 minor folds in marbles and calc schists that have been refolded by the Spilderdalen fold, and exhibit a well-defined trend. To account for this trend, comparison of this region with that of Steffodalen was made, where the approximate direction of the F_1 Steffodalen fold is ESE. When this is plotted stereographically, together with the maximum of F_1 structures in the North Spilderdalsvand region, a great circle may be drawn between the two points. The observed trend of F_1 linear directions fall almost exactly on this great circle, suggesting the scatter to be a consequence of refolding of the early F_1 lineations about the Spilderdalen fold. If this is the case, the refolding should be of similar style, a view supported by field evidence.

By unfolding around the Spilderdalen fold, it can be seen that the present NS F_1 linear structures of the Spilderdalsvand region become E-W, in conformity with all the other linear structures and major folds of this age to the E of the Glomfjord granite, corroborative evidence for regarding the metasediments of the South Ørnes region to be upside down.

Summary of Age Relationships of Folds

The geometric relationships of the major structures indicate directly the influence of three major fold episodes, termed F_1, F_2 and F_3 respectively, each of lessening intensity, and each with typical characteristics. It has been shown that the whole of the Ørnes region has undergone deformation during each of the fold periods.

F_1 Folds. The earliest folds are entirely isoclinal in character, often with a pronounced axial plane schistisity. In the Ørnes region, the Steffodalen fold is the only major F_1 structure to be identified with certainty, but a series of probable isoclinal folds can be established in the Galtskart succession on the basis of lithological repetition. Linear structures are rarely preserved, except in restricted regions, e.g. North Spilderdalsvand.

Contemporaneous with this fold period, a large amount of sliding took place, occasionally concentrated into well-defined zones. They may be identified on structural and stratigraphic evidence.

F_2 Folds. It is these folds that have the greatest effect on the geometrical arrangement of the rocks of the Ørnes region; they consist of recumbent overfolds with gently inclined axial planes.

Two major F_2 folds have been recognised, namely the Kjeipen and Spilderdalen folds, with SW and E plunging axes respectively. Other less important F_2 structures have been recognised, mainly monoclinal flexures on the flanks of the larger folds. In the NW of the region the Bolden antiform is probably of a similar age, and may also have a gently inclined axial plane. It probably represents a contemporaneous cross-fold. The majority of the linear structures were developed at the same time as the F_2 folds, and are consequently concentrated in two general directions, corresponding to the two most important F_2 folds. Field evidence suggests the F_2 folds to be similar in style, and the trend of the refolded linear structures in North Spilderdalsvand corroborates this view.

Separating the two major folds is the Lysvand slide, the only large-scale F_2 structure of this nature seen in the Ørnes region. It also separates the two regional linear directions showing them to be associates of the F_2 folds.

<u>F_3 Folds.</u> These consist of symmetrical folds, generally of a gentle nature, and of large amplitude. Their axial planes are vertical or very steeply inclined, on axes identical, or very nearly so, to the F_2 folds. The style of the folds is concentric, and although they are demonstrably superimposed upon the F_2 folds, it is probable that they represent a later phase of the same general period of deformation.

Similar conclusions as to the styles and characteristics of the three periods of deformation may be deduced from a study of minor structures (see section 2 h III).

General conclusions and comparison with the remainder of the Glomfjord region

Correlation of fold styles is possible with the remainder of the Glomfjord region, which has been shown to have undergone a similar structural history. Thus the F_1 period of deformation is displayed to advantage in the Krokvand-Rebenfjell fold (Hollingworth et al 1960, Walton 1959) and other major isoclinal folds have been described by Rutland (1958) and Nicholson (1960). Although these are comparable in style with the F_1 folds of the Ørnes region, it is impossible to be certain whether they all developed at the same time. In addition the isoclinal folds in the basal granites may belong to a still earlier phase of deformation. F_2 folds are also found throughout the Glomfjord region. The largest of these in the east of the region is the Sokumvand synform, a N-S structure that is in part overturned. It appears that the F_2 folds of the western part of the Glomfjord region are more intense structures, with more pronounced overturning than those in the east.

The F_3 structures are only spasmodically developed in the Ørnes region, and show interesting relationships to the earlier structures and to one another.

A marked constancy of all the linear structures in any one area, together with parallelism to the axial directions of the major folds is a characteristic of the Ørnes region. However, it is not necessarily the case that the two structures are products of the same period of deformation. Thus, in the NE part of the region, the predominant lineation is an F_2 structure and developed at the same time as the Kjeipen fold, whereas all the readily identified major folds have a similar orientation, but are probably of F_3 age.

When a wider region is considered, however, different relationships are found to exist. Thus, in the N of the Glomfjord region, around Sörfjord, the latest folds of presumed F_3 age have a N-S alignment, markedly divergent from the F_2 fold directions. It appears probable that the basal granites influence the geometry of the F_3 folds to a marked extent. The Sörfinnset antiform, which involves the rocks around Sörfjord, has a N-S axis. When the Laksaadalen fold is encountered, the fold is no longer present. However, the Bjellåtind antiform appears to be a re-emergence of the same structure, but here it has a SW plunging axis, in conformity with most of the other F_3 folds of the Örnes region. In the Steffodalen region, the F_3 folds are on a very nearly easterly plunging axes. Thus, the alignment of the F_3 folds appears to be affected by the proximity of the Glomfjord granite, the largest mass of basal granite in the whole of the Glomfjord region. Such an influence of basal rocks upon later structures has been noted elsewhere. In the Kettleman hills the direction of alignment of domelike structures is dependent upon the shape of the basin border. Some authorities, e.g. Glangeaud 1949 (quoted from de Sitter) suggest that folds of the pre-Alpine Western Alps depend upon the basement structure. Fault patterns are also often dependent upon pre-existing faults. In general, however, the alignment of previous structures, when refolded by a later organic phase, have little effect upon the directions of the later folds. In these cases the consolidated basement tends to move in blocks, in the manner suggested by Argand. It is not suggested that the separate phases of deformation of the Ørnes and Glomfjord regions represent different orogeneses. Indeed the progressive change in style from F_1 intense isoclinal folding to the F_3 gentle doming suggests the development of a single orogenesis. As this is the case, the individual phases of deformation are probably partly continuous, and there is no reason to suppose, particularly in the case of F_2 and F_3 folds, that a period of quiescence intervened. The basement was therefore in all probability continuously active, with a constant influence on the arrangement of the folds.

Certain parts of the Ørnes region show complex F_2 and F_3 interrelationships. Thus, the Skjeggen region is situated on the convergence of axes trending S and SW, while at Spildervand, a probably N-S culmination of presumed F_3 age crosses the SW plunging Ørnes fold. In both these cases, the structures may represent contemporaneous cross-folds (King and Rast 1955). In both cases the proximity of basal rocks may be an important factor.

Comparison exists between the F_3 structures associated with granites with certain of the mantled gneiss domes of Finland (Eskola 1948). Here, rejuvenation and tectonic injection of granites into their metasedimentary covers can be demonstrated, a mechanism which may explain the way in which the Bjellåtind antiform dies out south-westwards. It is also probably indicative of the largeness of the granite mass, as it has a pronounced effect upon the regional structure; this is further evidence for regarding the granite as basal.

Henderson (1943) has described the structure and metamorphism of pre-Cambrian rocks in the Great Salt Lake district, where granitic batholiths have invaded the already deformed country rock. The batholiths are concordant with the structures in the sediments and may have caused their development. This shows similarities with the Ørnes region, except that in the Great Salt Lake area, the granites have caused extensive thermal metamorphism of the surrounding sediments, while in the Ørnes region it is suggested that the granites are older than the metasediments.

Minor and Micro Fold Deformation

In accordance with the pattern of major folding already described, minor and micro folds* belonging to each of the three periods of deformation have been identified throughout the whole of the Ørnes region. The minor folds are the most abundant; some of these occur as drag folds on the limbs of major folds, and can then be dated readily, but others occur as isolated examples, and can only be assigned to the appropriate fold period from considerations of style, and the attitude of the geometrical components of the structures. Good agreement between major and minor structures is obtained, and an identical tectonic history may be deduced from a study of the minor folds, to that indicated from the relationships of the major structures.

F_1 Deformation

The most important structural feature developed at this time was the regional schistosity. For the most part, this is coincident, or nearly so, with the bedding, but occasionally where small-scale overfolds or isoclinal folds are present, it is seen to be an axial plane structure. Examples of this are clearly displayed in the pelitic schists where the regional schistosity is most pronounced. When interbanded material of a different nature is present in the pelitic schists, the form of deformation is displayed (fig.30 A-D); similar structures are probably of

* The term "minor fold" is here used to describe structures from a few centimetres to several metres in amplitude, and corresponds to the mesoscopic division of Weiss (1955). Micro folds are those that can only be observed on microscopic scales.

Fig.30. Folds belonging to the F_1 period of deformation

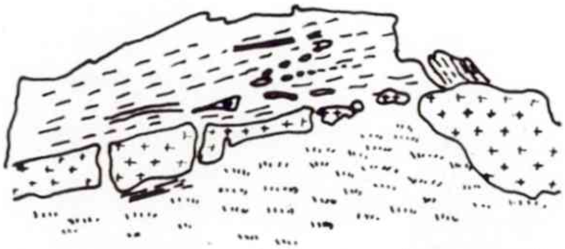
A. Fractured quartzite in pelitic schist, South Galtskart, showing intense folds. The schistosity is axial plane to the observed folds

B. Asymmetric drag folds in marbles within pelitic schists, South Scetertind

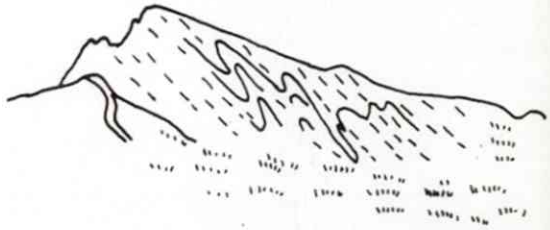
C and D. Isoclinal folds in interbanded quartzites and mica schists west of Lysvand, showing dominance of axial plane schistosity in the schists.

E. Isoclinal fold in calcareous pelitic schists, South Lysvand.

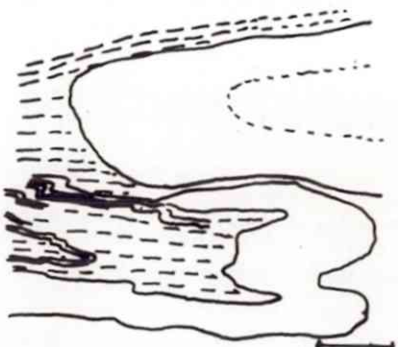
F. Isoclinal folds with associated thrusts in calcareous pelitic schists, North Lysvand



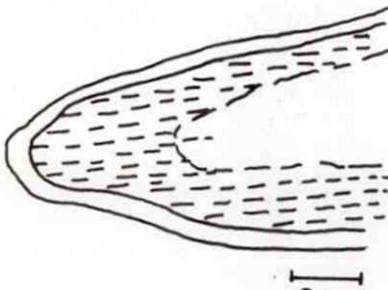
A 50M.



B 100M.

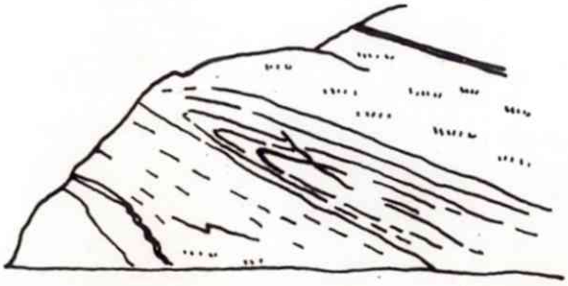


C 20cms



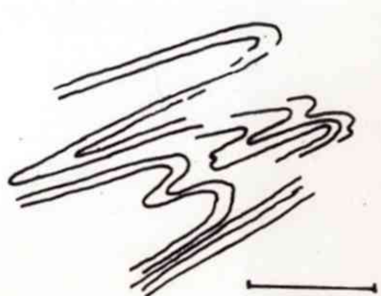
2cms

D



50M.

E



10M

F

widespread occurrence but cannot be detected in the more uniform pelitic rocks. Spasmodic development of F_1 axial plane schistisity is displayed in a number of examples involving interbanded pelitic schists and quartzites (figs 30 C and D and fig.31).

Microscopic examination of the pelitic schists shows that recrystallisation has occurred after F_1 times and has largely masked the early folds. An early and relict schistosity, sometimes at right angles to the present structure, is frequently seen in the orientation of trails of inclusions in porphyroblastic garnet crystals.

Due to the diverse mineral assemblage in calcareous pelitic schists, the regional schistosity is never as well defined as in the pelitic schists (see also p. 25). Evidence of F_1 deformation is consequently sparse, but intense folds have been identified in scattered localities (fig. 30 E and F) and do show an incipient development of axial plane schistosity. Associated with some of these folds are small-scale slides, and as in the case of the major slide zones, are frequently only visible over a restricted outcrop.

Microscopic evidence of F_1 folds in calcareous pelitic schists is lacking and most of the minerals probably date in their crystallisation from post-F_1 times. The mineral paragenesis of the group, probably covered a large time-range.

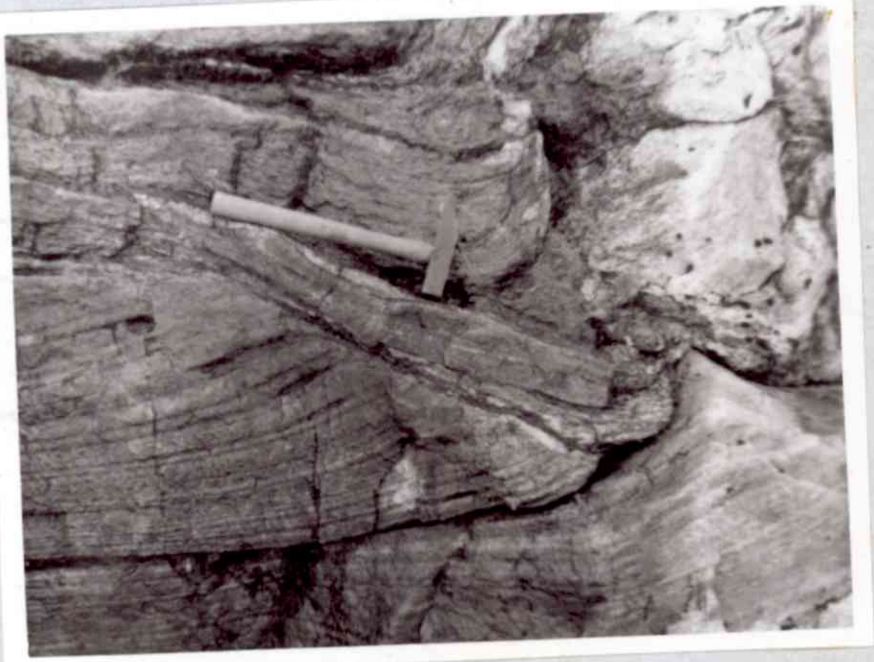
Because of their consistently banded character and relatively low mica content, the semi-pelitic schists often show F_1 structures

Fig.31. The differences of F_1 deformation between micaceous pelitic schists and quartzites, Steffodalen

Fig.32. Intense F_1 isoclinal folds, with evidence of flow towards the axial regions, in silty schists, North Galtskart

to advantage. Good examples of F_1 overfolds and isoclinal folds are present in the semi-pelitic schists of Galtskart, with occasional small-scale slides also developed (figs 32-35). In one of the outcrops (fig.35) an F_1 overfold has been refolded by a much more gentle structure, indicating directly the effects of more than one period of movement.

In the series of banded semi-pelitic and pelitic rocks of the Steffodalen region, good examples of superimposed folding are developed, and structures belonging to all three periods of deformation have been identified (figs 36-40). Superimposition of F_2 upon F_1 folds, and F_3 upon F_2 can be seen, and in each case the folds possess similar characteristics to the equivalent major folds. Thus, the F_1 folds show an incipient axial plane schistosity, are normally isoclinal or nearly so in character and commonly show a pronounced thickening towards the axial regions. The F_2 folds are less intense, with no axial plane schistosity, but with a common development of strain-slip or fracture cleavage; they have developed by shearing processes, and the beds maintain constant thicknesses around the folds. F_3 folds are gentle warps and have developed by concentric folding. Occasionally, as in fig.40, the F_3 folds are superimposed upon a series of beds that show convergence in one direction due to earlier folding. It appears that the inclination of the axial plane of the F_3 fold changes across the earlier fold due to this convergence (see fig.41). Essential requirements for this arrangement appear to be a competent band within less competent



Figs 33 and 34. Folds in banded semi-pelitic schists, North Galtskart, showing a common development of minor thrusts

Fig.35. Refolded F_1 isoclinal folds in semi-pelitic schists, North Galtskart

Fig.36. Accordion folds of ?F_2 age, refolded by gentle warps, in banded semi-pelitic schists, Steffodalen

Fig.37. Isoclinal F_1 folds in banded semi-pelitic and pelitic schists, refolded by F_2 shear-folds, Steffodalen. A high degree of plasticity in F_1 times is indicated by the disharmony of the folding. Incipient fracture cleavage is associated with the F_2 folds

Fig.38. F_1 isoclinal folds in pelitic schists refolded by gentle ?F_3 folds, Steffodalen

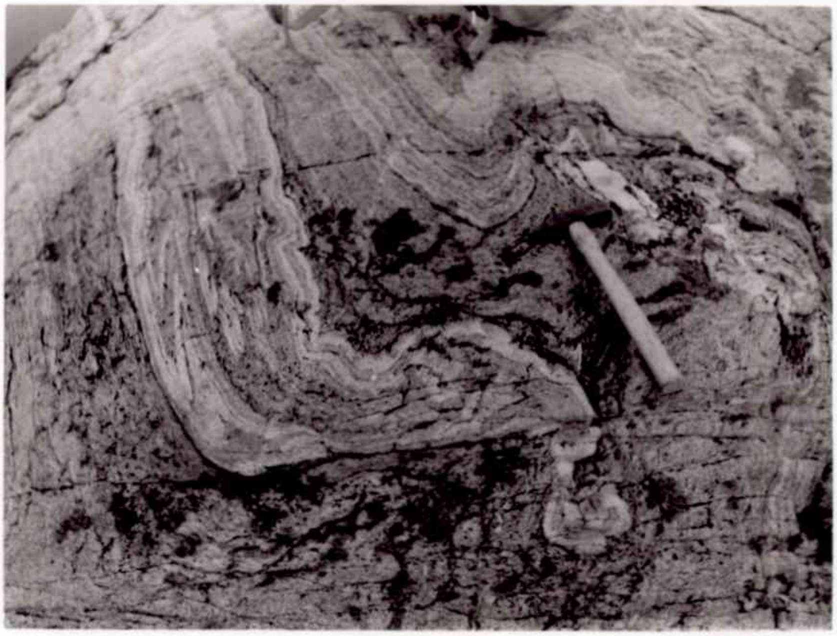
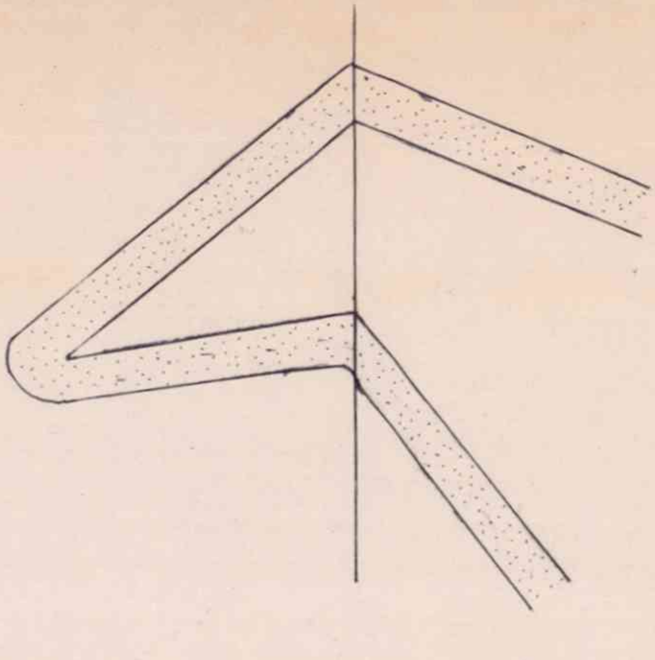
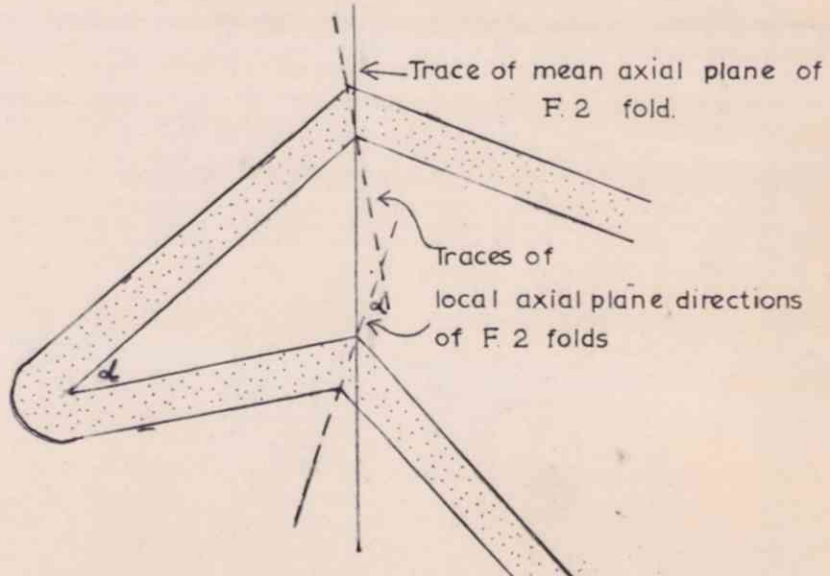


Fig.39. Refolded F_1 isoclinal fold in semi-pelitic schists, Steffodalen. Note the two sets of drag folds, one within the isoclinal fold and of F_1 age, and the other on its margins and of F_2 age. The inclination of the axial planes of the later folds seen in the two light-coloured bands is variable (see fig.41). Moreover the axial plane directions of the small F_2 drag folds appear to fan around the axial plane of the larger folds.

Fig.40. Disharmonic F_1 folds between banded semi-pelitic schists and amphibolites, Steffodalen



(A) Overfold refolded by shear folds. The axial plane trace of the F.2 fold is undeviated across the overfold.



(B) Overfold refolded by a concentric fold. The presence of a competent band within the sucession may cause local deviation of the direction of the mean axial plane of the F2 fold.

Fig 41.

material around which folding is initiated, and a concentric style to the folds. This may form a means of differentiating the style of superimposed folds, as if the later folds are similar in style, the inclination of the axial plane remains constant (see also Ramsay 1957). Moreover, if the refolding is on approximately the same axis as the earlier fold the angular relationship between the axial plane traces of the later folds is almost the same as that between the limbs of the earlier fold (see fig.41).

F_1 folds on a microscopic scale have not been observed, and as in the case of the pelitic schists syn- or post-F_1 crystallisation has been operative throughout.

In the calc-schists and amphibolites, evidence of F_1 folds is rarely seen, unless deformation has involved interbanded rocks of a different nature. This is due mainly to the unbanded nature of this group, and the general absence of S-planes of any description. Occasional examples of F_1 folds involving calc schists or amphibolites with other rocks are spasmodically developed; in some, the calcareous rocks deform in the same manner as their associates, but elsewhere, and notably in fig.39, marked disharmony between the two is apparent. Rarely F_1 overfolds within a single band of a calc schist or amphibolite can be seen. They are normally exhibited by the arrangement of thin, and often very persistent, quartz partings whose origin is uncertain. It is suggested that the quartz is of secondary origin, having been segregated along incipient S-surfaces.

Micro-folds of F_1 age have not been identified in this group of rocks, and a large amount of late-stage, probably post-F_3, recrystallisation now masks most of the earlier structures.

Rocks belonging to the Gneissic Group are found mainly in the immediate environs of Ørnes, but thin bands are also present in the Galtskart rocks. They consist of massive creamy-white quartzo-felspathic rocks, with large platy biotite and hornblende crystals arranged in thin bands or in discrete clots orientated en échelon. Due to the outcrop of the main group of gneisses in heavily vegetated country at sea level, detailed three-dimensional examination of the rocks is not possible. However, the majority of the gneissic rocks contain abundant thin pegmatites, arranged in elongated isoclinal folds (fig.79); the schistosity of the gneisses, determined by the orientation of the platy minerals is rarely well-defined, but as far as can be seen, is axial plane to the folds in the pegmatites.

In one of the thinner gneissic bands on the coast at Sörfjord, development of a predominant schistosity normal to the trend of the junction with an adjoining quartzite is well displayed (fig.42A). The boundary shows a series of tight minor folds whose axial planes are parellel to the schistosity of the gneisses.

Occasional examples of possible F_1 micro-folds are seen, determined by the arrangement of individual biotite crystals. Subsequent recrystallisation has occurred and the crystals are unbent.

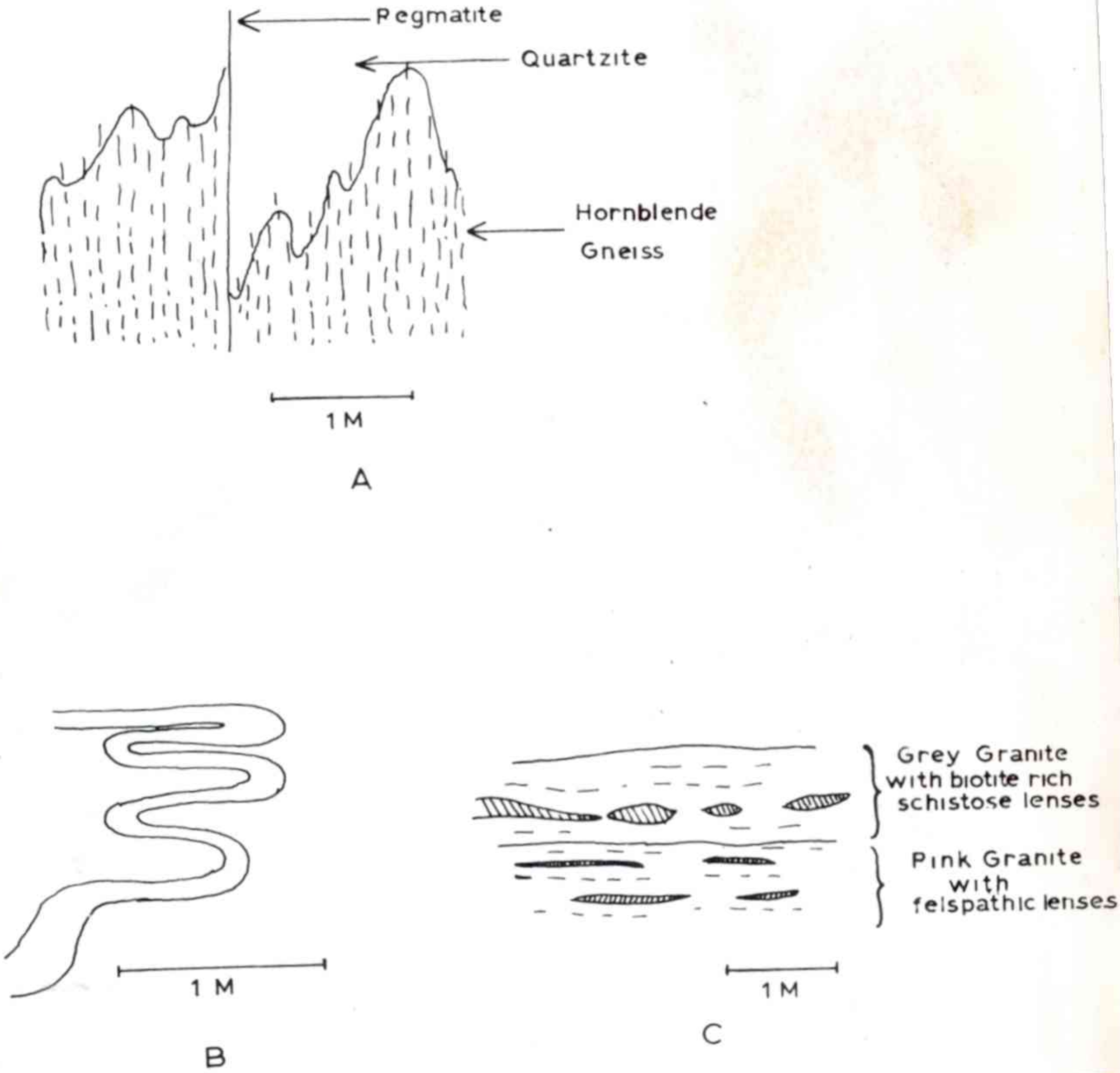


Fig.42A. A pegmatite developed along the direction of F_1 schistosity discernable in the hornblende gneiss. The junction of the gneiss with quartzites shows F_1 minor folds

B. F_1 flow folds in a banded sheet granite, North Storviken

C. Part of the north-west Storvikvand sheet granite, showing sheared layers produced probably during F_1 times

Due to their distinctive lithologies the Marbles often show the style of folding to advantage. Thus, in the pelitic schists of South Scetertind, a whole series of dextral drag folds can be identified by the pattern of thin interbanded marbles (fig.30A). Other examples of intense F_1 folds involving interbanded marbles and schists are seen in the rocks north of Spilderdalen (fig.43 A-D). In some of these (A-C) the individual marble bands, although highly deformed, maintain constant thickness over large distances of outcrop. The pattern of folding is very irregular and was probably accomplished by flow. The other example (D) shows marked disharmony between the folds in a brown and grey marble within associated schists. The boundaries of the brown marble are sinuous, and indicate local injection during folding. In contrast, the grey marble shows a much more simple pattern, and in one place has deformed by fracturing rather than folding.

An unusual fold involving marbles and calc schists is present to the N of Digermalen (fig.43E). The upper surface of the fold is clearly diapiric into the overlying rusty-weathering schists, while the lower boundary appears to be a plane of detachment. Thus, the overall form of the structure simulates a concentric fold, and good agreement between the observed plane of décollement, and that calculated from the arrangement of the upper surface of the fold is obtained (see de Sitter 1956). However, the patterns of minor folding on the limbs and within the core of the fold are unlike those normally associated with concentric folds. Moreover, the axes of the drag folds are on variable axes,

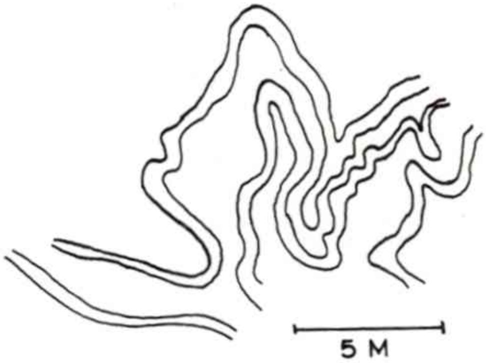
Fig.43. F_1 folds in marbles

A-B. Interbanded marbles and calc-schists, South Scetertind, showing a good development of flow-folds.

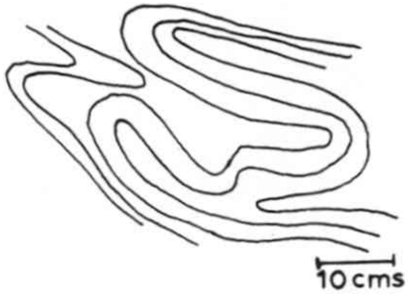
C. Irregular folds in marbles within pelitic schists, South Scetertind. Schistosity dominates the schists.

D. Variable styles of deformation in grey and brown marbles, South Scetertind. Localised flow of the brown marble is suggested by its sinuous contact with the associated schists.

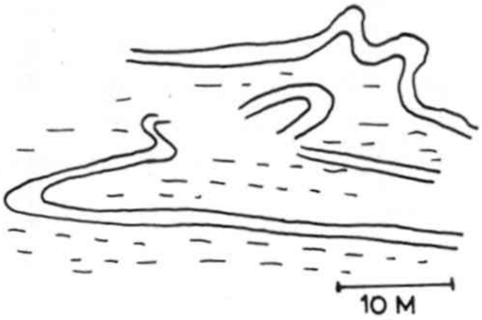
E. An unusual 'concentric' fold in brown marbles, North Digermulen. The observed plane of detachment corresponds closely with that calculated from the amplitude of the fold (de Sitter 1956, p.189) but the internal complexities exhibited by thin calc-schists probably indicate intense deformation at considerable depths.


5 M

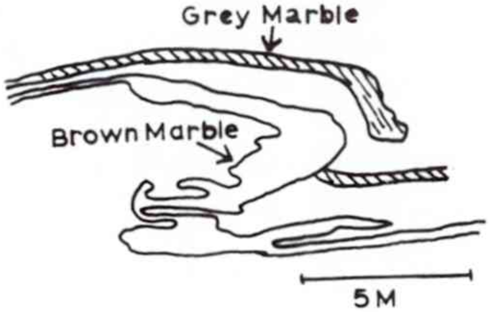
A


10 cms

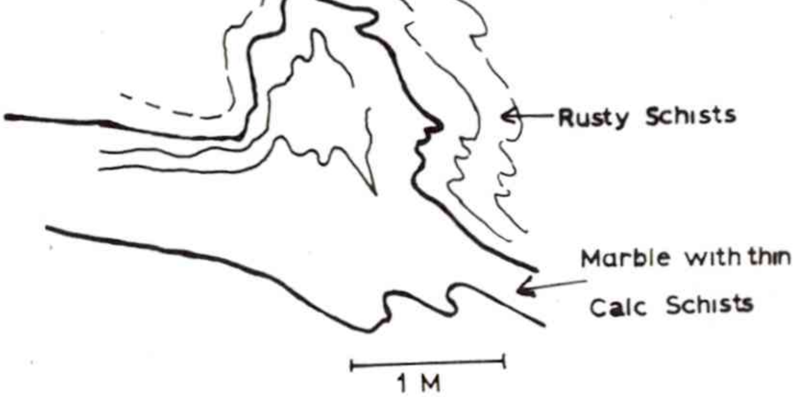
B


10 M

C


Grey Marble
Brown Marble
5 M

D


Rusty Schists
Marble with thin
Calc Schists
1 M

E

indicating further complications, and the possibility of more than one period of folding. Not enough measurements were taken to determine the reasons for this variation in axial direction. It is therefore suggested that the equivalence of the fold to a concentric structure is coincidental, and that it was produced in F_1 times, probably at considerable depths. The geometrical arrangement may be due to the effects of refolding.

Late-stage, probably post-F_3, recrystallisation has effectively obliterated any evidence of F_1 micro-folds; in general, the marbles are composed of large calcite crystals with an average grain size of several mm, which occasionally show a slight amount of bending.

Evidence of minor structures in the <u>Ultrabasic Mass</u> of Torsvik are completely lacking. The original fabric of the rock has been almost completely obliterated by late-stage crystallisation of rosettes of amphiboles and chlorite, and only rarely can the original igneous minerals be identified (see also p. 291). It is clear from petrographic examination that the observed metamorphism is late-stage, possibly post-F_2.

Throughout the areas of <u>Granite</u> in the Ørnes region, evidence of folds belonging to any of the fold periods is uncommon. Two probable major F_1 folds in the Glomfjord and Bjellåtind granites have already been described (see p. 52), while the only definite isoclinal minor folds, with typical F_1 features such as axial plane schistosity are present within the Glomfjord Granite (M.A. Jones, personal communication).

The only probable F_1 minor structure observed in the granites of the Ørnes region is present within one of the banded sheet granites of Storviken (fig.42B). Here, a massive pink granite about 15 cm thick, and interbanded with coarse-grained biotite schists and grey granite, is folded into a series of irregular recumbent isoclines. The style compares closely with deep-seated folding in other areas (e.g. Weiss and McIntyre 1957, where quartzite bands in schists deformed by early folds show comparable structures). Later recrystallisation has been widespread, particularly in the biotite schists whose present fabric is unrelated to the fold.

Irregular schlieren of felspar-rich or biotite-rich lenses are common in parts of the North Storvikvand Granite (fig.42C) and probably owe their origin to shearing processes during F_1 times.

Crystallisation of the granites, particularly the felspathic fraction, is late-stage, and evidence is given elsewhere (p. 221) for regarding the bulk of the crystallisation to have been accomplished during or later than F_2 times. Thus, evidence of F_1 micro-folds are unlikely to be preserved, and has never been seen. The present schistosity, caused by the parallel arrangement of biotite crystals, is everywhere completely planar, and parallel to that in the overlying schists. In both groups it probably has a common origin in syn- or post-F_1 crystallisation.

Conclusions on Style and Conditions of F_1 Deformation

A general correspondence between major and minor folds produced in the earliest period of deformation exists, and conclusions as to the conditions of formation are applicable to folds on all scales.

(1) Minor structures produced in F_1 times are overfolds or isoclinal folds. The amount of shear of material into the noses of the folds is very variable, and while some show pronounced variation in thickness, others approach accordion folds in their geometrical arrangement. Both of these types of folds are regarded as having been formed by shearing processes. Occasional examples of folds that were probably formed by flow of material are seen, notably in the case of the recumbent isoclinal folds in marbles. The single example of a fold with a concentric style (see p. 125), is probably a much more complicated structure.

(2) Associated with many of the F_1 folds is a pronounced axial plane schistosity in appropriate rocks, and in the cores of the folds this can be seen to cut across the bedding. Axial plane schistosity is most highly developed in pelitic schists where the percentage of micaceous material is high.

(3) The present fabric of the rocks was produced during or after F_1 times, and microscopic study indicates that in a large number metamorphism has been prolonged.

(4) Superimposed upon these, the earliest and most intense folds, are more gentle structures belonging to the F_2 and F_3 periods of deformation. The latter possess very different properties and were probably displaced in time and space.

F_2 Deformation

On a major scale, the F_2 folds are recognised as open overfolds with horizontal or gently inclined axial planes. On a minor scale, the F_2 structures have similar properties, and can often be directly related as drag folds on the limbs of the major folds. A true axial plane schistosity is never developed, although in certain areas small-scale puckers of highly micaceous material, formed by strain-slip cleavage, is the most prominent structure the rocks possess.

Isolated examples of F_2 drag folds, with gently inclined axial planes, are present in Pelitic Schists, principally in the south of the Ørnes region (figs 44 B and C). They are generally exactly analogous to the major folds upon whose limbs they are developed and are similar in style. The regional schistosity is bent around the folds, and no new S-surface is developed parallel to the axial plane of the F_2 folds.

In addition, some of the regions of pelitic schists, notably north of Markvand, are dominated by F_2 microfolds. These consist of puckering of the micaceous laminae by a series of closely-spaced cleavage planes, generally about 1 cm apart. The amount of movement along each of these planes has been slight, and the micas now occupy the form of symmetrical cleavage folds (fig.44A). Analogous structures illustrated by Billings (1942) and de Sitter (1956) have been termed shear cleavage and fracture cleavage respectively. Other authorities use the term strain-slip cleavage. de Sitter regards slaty cleavage as a development from fracture cleavage by increased movement and recrystallisation, and recognises the difference between the two in that fracture cleavage fans upwards towards the top of an anticline, and slaty cleavage fans downwards (see also G. Wilson 1946). Moreover, fracture cleavage tends to fan around the folds to a greater extent than slaty cleavage. Such characteristics have not been observed in the North Markvand pelitic schists, as the micro-folds are developed upon a single limb of the Kjeipen fold (see p. 65), and as far as can be seen in the limits of outcrop, maintain a constant direction. Fig.47 shows the clearest example of fracture cleavage micro-folds in the whole of the Ørnes region, but similar structures are seen spasmodically throughout the whole of this group of pelitic schists. In the sole example where measurements of the inclination of the axial plane of the micro-folds was possible, it was found to be equivalent to that of the major Kjeipen fold in the same area, which supports the view that the micro- and major structures are coeval.

Extensive syn- or post-F_2 recrystallisation of the pelitic schists is admirably displayed in these rocks that show fracture-cleavage folds. In all the examples examined in hand specimen and thin section, the micas are uniformly recrystallised and unbent, even on the hinges of the folds (see also figs 103).

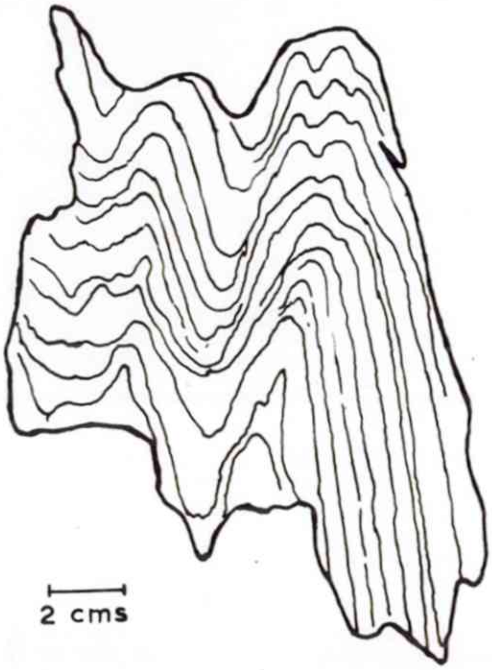
The age of certain of the minor folds in Calcareous Pelitic schists is sometimes difficult to establish. Thus, a series of overfolds in the rocks north of Kjeipen have gently inclined axial planes, in conformity with the F_2 major structures, but have associated small-scale thrusts, a typical F_1 phenomenon (fig. 45).

Fig.44. F_2 folds in pelitic schists

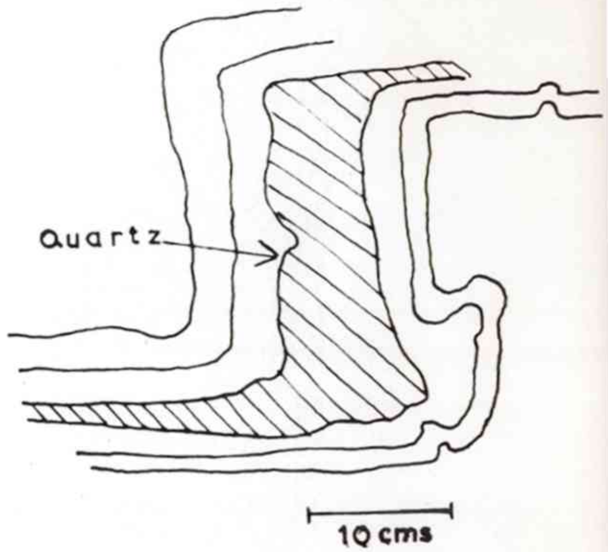
A. Symmetrical shear folds in highly micaceous schists, South Spantind

B. Segregation of quartz into the axial regions of a monoclinal drag-fold in semi-pelitic schists, Spantind

C. Overfold in pelitic schists, South Scetertind. The fold probably represents part of a drag-fold on the lower limb of the Spilderdalen fold



A



B

C

Fig.45. Overfolds in calcareous pelitic schists, North Kjeipen. They are probably of F_2 age

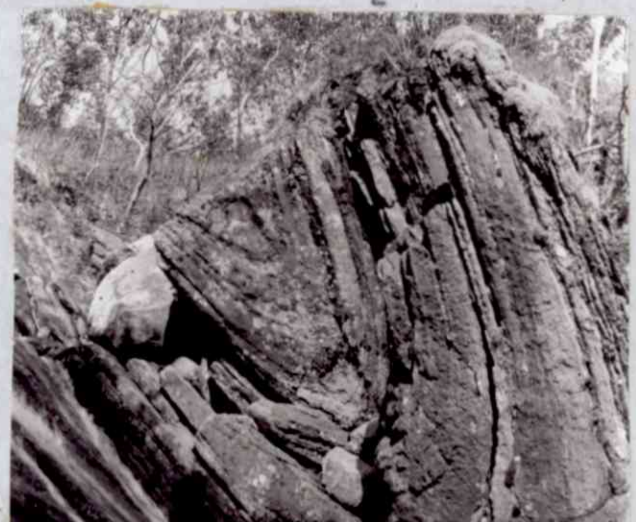


Fig.46. F_2 overfold in calcareous pelitic schists, West Lysvand

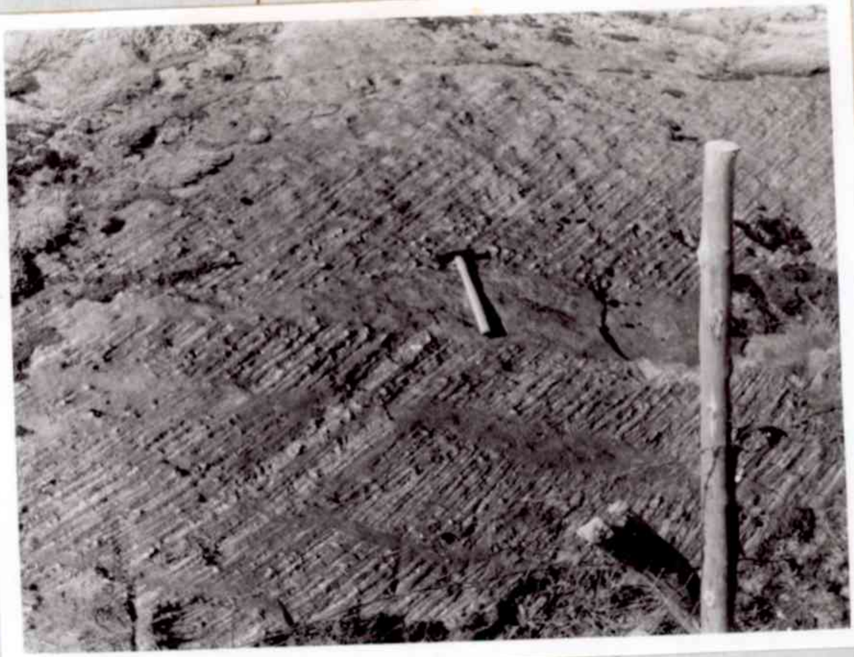


Fig.47. F_2 fracture cleavage in pelitic schists, Markelven. The lithological banding is inclined gently towards the bottom right hand corner of the photograph.

Fig.48. F_2 shear folds in marbles and quartzites in the valley between Inner and Outer Galtskart

An F_2 age to the folds is favoured as their axial directions correspond to that of the major F_2 fold, and the other undoubted F_1 folds observed in this group of rocks are very much more intense.

Another isolated F_2 overfold is shown in fig.46 and is associated with F_1 accordion folds with a very different style and axial direction. As far as can be seen, the succession involving the accordion folds is repeated about the overfolds.

A loose block of calcareous pelitic schist on the north of Skjeggen shows how a symmetrical fold is eliminated by an axial plane pegmatite (fig.76). Because the block is not in situ, it is impossible to be certain of its regional relationships and the age of the fold. The style of deformation is not diagnostic, but as the pegmatite is developed parallel to the axial plane of the fold it is suggested that the structure is of F_2 age. Elsewhere in the Ørnes region the F_3 pegmatites are cross-cutting bodies which are not governed by the attitude of pre-existing folds.

The <u>Semi-Pelitic Schist</u> group is the only one in the Ørnes region that shows directly the effects of all three periods of deformation. Examples from Steffodalen show refolding of earlier structures to advantage, and from the style of the folds it is possible to determine their relative ages (see also p. 116). Figs 36-40 show examples of all three types of folds.

Microscopic examples of F_2 folds can sometimes be identified by the orientation of mica crystals; in the majority of cases later recrystallisation has been operative, and the micas are rarely bent. (Fig 103)

In the Calc-Schist and Amphibolite group, F_2 minor deformation is not commonly developed. For the most part, it consists of scattered development of boudinage whose axes are parallel to the other regional F_2 linear structures.

Crystallisation of the constituent minerals of the group has been largely accomplished since F_2 times, and no examples of microfolds of any description have been identified.

Similar relationships as found in the preceding group are also present in the Gneisses. Here the only evidence of F_2 deformation is the localised development of boudinage in the rocks on the SE of Blaatind. Segregation of calcite, and its recrystallisation in perfect rhombs up to 3 or 4 mm in length has occurred in the regions of low stress.

F_2 folds in Marbles are particularly abundant in the thick band of alternating marbles and quartzites on the W of Galtskart (figs 48, 49E). In the quartzites angular accordion folds affect the discontinuous layers, while the intervening marbles show a more complicated fold pattern. This variation in fold style is clearly dependent upon the differences in physical properties between the quartzites which were liable to fracture, and the marbles which more easily flowed. It is the former which determined the major pattern of folding.

An unusual F_2 structure involving marbles and calc-schists

Fig.49 A. Disharmonic folding, West Sörfjord

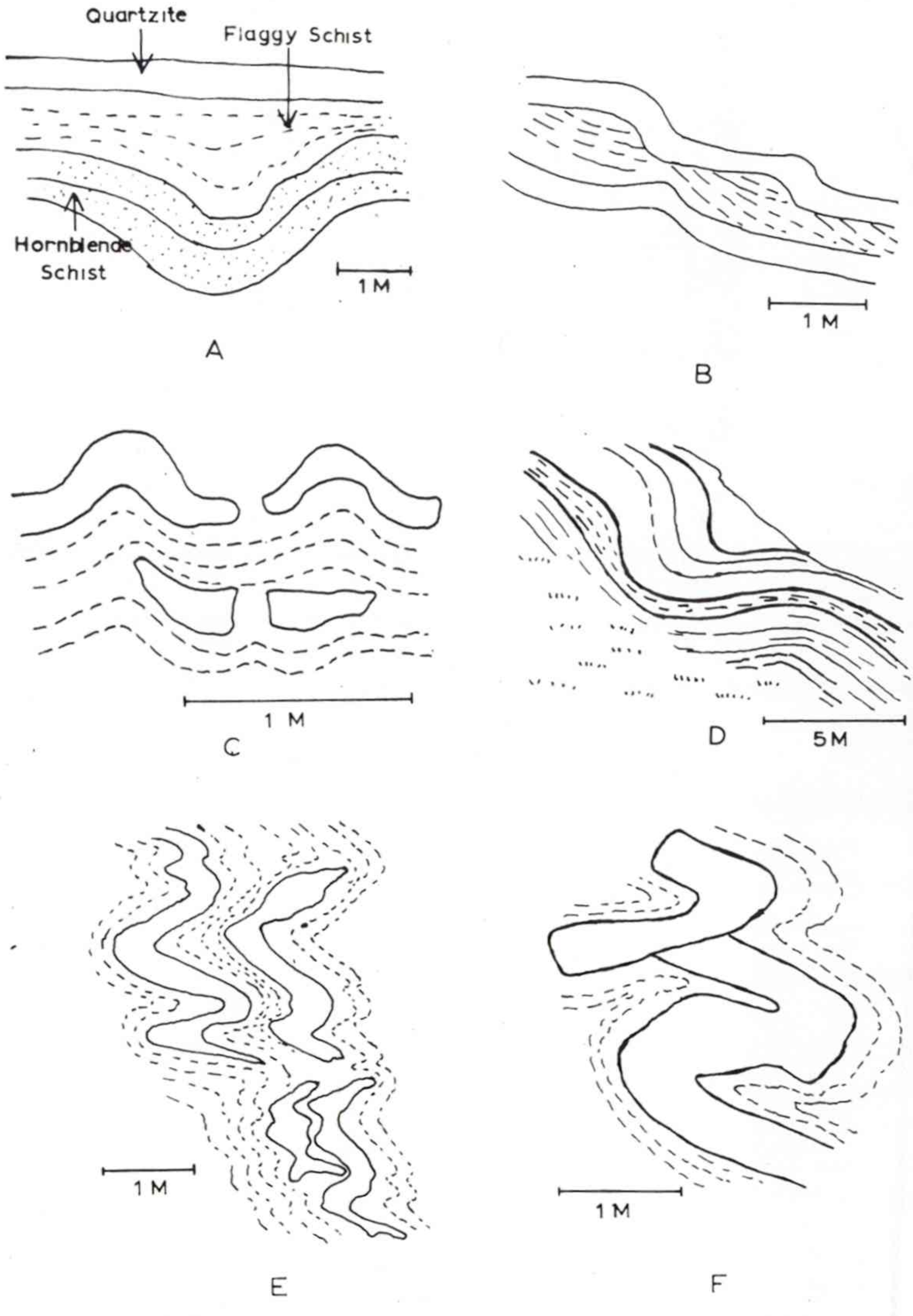
B. Disharmonic folds in quartzites and amphibolites, West Sörfjord

C. Fractured quartzites in pelitic schists, showing symmetrical F_3 concentric folds, Breitind

D. Gentle F_3 folds in massive pelitic schists, South Scetertind

E. Fractured quartzites in brown marbles in the valley between Inner and Outer Galtskart, showing F_2 shear-folds

F. Angular F_2 box-folds in a grey marble, South Scetertind. The small-scale thrust is not visible in associated calc-schists, indicating the latter to have recrystallised subsequent to the period of folding


Quartzite
Flaggy Schist
Hornblende Schist
1 M
A
1 M
B
1 M
C
D
5M
1 M
E
1 M
F

occurs on the S of Scetertind (fig.49F). A grey marble about 50 cm thick is deformed into angular box-folds, with local development of minor thrusts. The associated calc-schist has a visible schistisity which is not interrupted by the thrust in the marble band, and indicates crystallisation subsequent to the thrusting and folding. Thus, although the style of the fold is suggestive of F_2 deformation, it has some associated F_1 features. Nearby F_1 folds in other marble bands are very much more intense, and for this reason the fold is assigned to the F_2 fold period.

As mentioned previously, late-stage crystallisation of the marbles has occurred throughout, and the slight amount of deformation of the calcite crystals occasionally seen is probably to be correlated with F_3 folds' deformation.

F_2 structures in the Ultrabasic Mass of Torsvik are completely lacking (see also p. 128).

As in the case of F_1 folds, F_2 minor and micro-folds are not commonly seen in the granites. However, in all of the finely banded sheet granites of the N of the region, minor folds of presumed F_2 age are developed spasmodically (figs 50, 51). Commonly, the folds have a marked development of long and short limbs, all in the same direction. Similar structures are seen in the granite on Teksmona where the sense of the drag folds is the same as that in the underlying quartzites, indicating them to have a common age. The contemporaneous development of pegmatites with small-scale drag folds is seen within the granite on the N of the same island ~~(fig.~~). F_2 minor deformation is entirely lacking in the

Figs 50 and 51. F_2 minor folds in finely banded sheet granites, North Storviken

Bjellåtind Granite. This is due probably to the massive nature of the latter, combined with the position on the under limb of the Kjeipen fold.

The majority of crystallisation of the granites occurred after the F_2 fold period, and no examples of F_2 micro-folds have been identified.

Conclusions on Style and Conditions of F_2 deformation

(1) The structures produced in this phase of deformation were either overfolds or closely-spaced shear folds, analogous to fracture cleavage. The style of the folding was variable, sometimes as symmetrical similar folds, with a varied amount of thickening towards the axial regions, and sometimes as accordion folds with angular hinge-lines.

(2) Except for the development of fracture cleavage which can be readily distinguished, no new planar structures were developed.

(3) In the majority of cases, the axial planes of the folds are only gently inclined, a characteristic that serves to distinguish the folds from F_3 structures.

(4) The style of the folds is considerably less intense than the F_1 structures upon which they are superimposed, suggesting that they formed at a higher tectonic level, probably at a considerably later date in the history of the rocks.

(5) Metamorphism was operative during and after this phase of deformation, and varies according to the character of the rock groups.

F_3 Deformation

The separation of this phase of deformation from F_2 folds was accomplished principally by examination of the major structure (see fig.9). Nevertheless, it is possible to identify F_3 minor folds, even though their style is sometimes equivalent to F_2 folds. As a generality, F_3 folds are concentric in style, with vertical or steeply dipping axial planes.

Examples of F_3 minor folds are very scattered, and some of the lithological groups contain none.

The most widespread examples of F_3 minor folds are seen in the semi-pelitic schists of Steffodalen (fig. 49D), and are recognised by the gentle nature and concentricity of the warps. They are noteworthy in sometimes possessing horizontal axial planes in marked contrast to the remainder of the F_3 folds of the Ørnes region.

Scattered examples of other F_3 minor folds are shown in figs. 49 A-C, where the fold pattern is determined by the existence of semi-pelitic layers within softer pelitic material. Again the style of folds is concentric. Some of the structures, particularly on the S of Scetertind are remarkably constant, and can occasionally be traced over several hundred metres.

A whole series of symmetrical open folds with amplitudes of around 10 metres are present in the calcareous pelitic schists of West Lysvand, and are probably due to the situation of the rocks in the nose of the Kjeipen fold (see p. 60).

Occasional gentle warps and monoclinal flexures are developed throughout the Ørnes region, and probably owe their existence to F_3 deformation. Commonly, the axes of these warps are very different from the regional linear directions; for example, on North Galtskart, gentle folds in calcareous pelitic schists are on a N-S axes, a direction which corresponds to the major F_3 Sörfinnset Antiform.

A large amount of the slight microscopic deformation of micas and plagioclase crystals which is common to many of the rocks may be due to this phase of deformation. It cannot be determined with certainty whether or not the late-stage crystallisation of marbles and calc-schists etc. is post-F_3, because the occurrences of F_3 folds are so scattered.

Conclusions on Style and Conditions of F_3 deformation

(1) Structures belonging to this phase of deformation are concentric folds and minor warps with vertical or steeply inclined axial planes.

(2) No new planar structures of any description are associated with the folds.

(3) The style of the folds is indicative of their formation at high tectonic levels, and it is possible that they represent the dying phase of the more intense F_2 deformation.

General Conclusions on Structural History, from a study of minor and micro deformation

Three main phases of deformation can be recognised from this study, and may be correlated directly with major folds of equivalent styles. These phases, termed F_1, F_2 and F_3 respectively, represent a systematic change from deep-seated conditions where deformation was intense to high level environments when folding was relatively gentle. The largest changes in physical conditions occurred between F_1 and F_2 times, and the two later fold periods may be regarded as sub-divisions of a single major phase. In restricted regions, superimposed folds have been identified, indicating directly the time sequence of folding; this is much more commonly observed on a major scale.

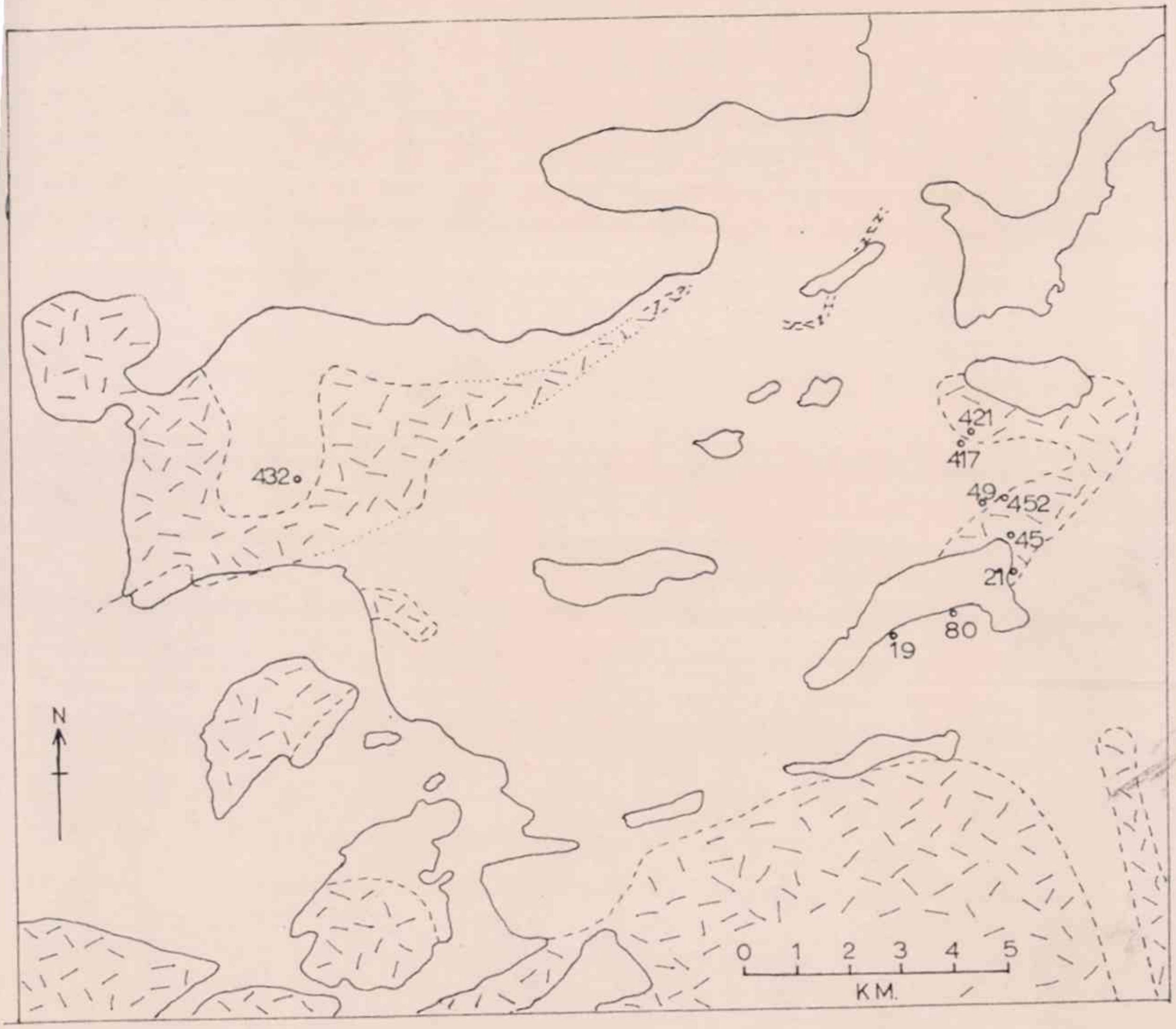
Microscopic studies indicate that the rocks have undergone polymetamorphism, and that the height of metamorphic processes varies in different groups. This is dealt with more thoroughly in section 3 .

Petrofabric Analyses

Analyses of the orientation of the cleavage directions of mica crystals, and of the 'c' axes of quartz crystals have been completed for 9 rocks (figs 52-55), principally with a view to comparison of the types of fabric present in the basal Bjellåtind granite and the immediate contact metasediments. In each of the diagrams the schistosity or foliation plane is indicated. The crosses represent the constructed F_2 linear direction, as determined from the local direction of the appropriate region. None of the rocks contains a visible lineation.

(a) Mica Fabrics

The mica crystals throughout, show good preferred orientations which lie within the macroscopic schistosity. Maxima are commonly around 15%, but vary from 6% to 30% in different rocks. Few complications exist in the spread of the maxima, probably indicating crystallisation during a single period of deformation; subsequent incipient S-planes have not developed. The most pronounced maximum, of 30%, is found in a quartzite immediately above the Bjellåtind granite (fig.54, H 49). Here abundant small muscovite crystals are evenly distributed throughout the rock, and commonly occur as inclusions within quartz. It appears that they crystallised prior to development of the present quartz crystals. The two biotite diagrams of H 452 represent separate bands of a migmatitic rock on the immediate contact of the Bjellåtind granite



SPECIMEN LOCATION MAP OF PETROFABRIC ANALYSES

Fig 52

with the overlying metasediments (fig.54). The left hand diagram, for 100 crystals, comes from a granular quartzo-felspathic band, where the biotites are small and uncommon. The right hand diagram, for 200 crystals, is from a biotite-rich band, where the biotites occur as larger crystals, often arranged in clots, and showing a less pronounced orientation. Large microcline crystals are associated with the biotite, and quartz is almost entirely lacking. It appears that the reason for the less well-developed mica orientation is due to recrystallisation of the micas at a later date than those in the more granular fraction of the rock.

(b) Quartz Fabrics

A variety of patterns are developed in the orientation of the c axes of the quartz crystals, and by comparison with the schematic diagrams of Fairbairn (1949) they can be divided into two broad groups.

(1) A single maximum falling in the a tectonic direction. This is particularly pronounced in the case of the two quartzites measured, but similar diagrams are also found in some of the granites.

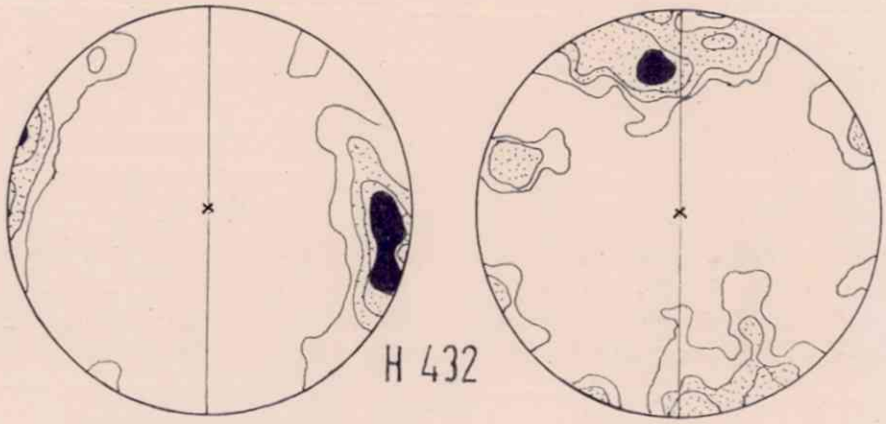
(2) a girdle around the b direction, commonly with separate maxima within the girdle. This is particularly well developed in one of the granites.

Certain of the rocks, notably H 21, show no obvious pattern to the fabric produced.

Figs 53-55. Petrofabric analyses of the specimens shown in fig.52. For H 452 (fig.54) the left hand diagram for 100 biotites is derived from the felspathic fraction of the rock, while the right hand diagram, for 200 biotites, is from the more pelitic bands

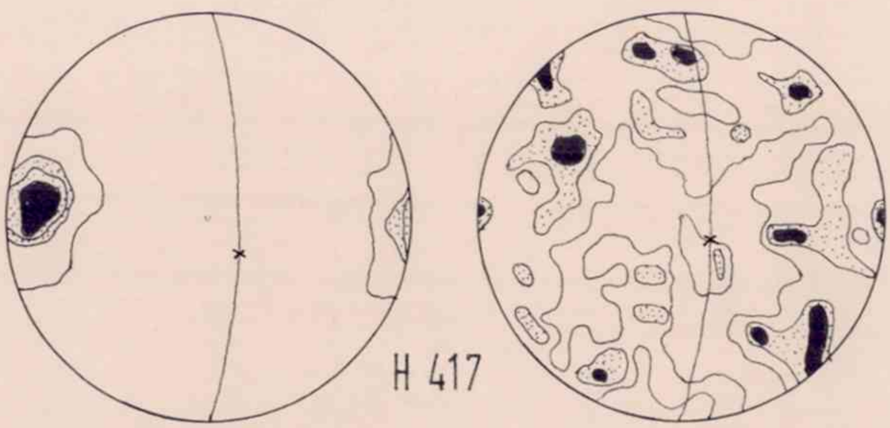
H 432	Quartzite
H 417	Quartzo-felspathic schists
H 421	Granite
H 45	Granite
H 49	Quartzite
H 452	Banded granite
H 21	Granite
H 80	Silty schist
H 19	Quartzo-felspathic schist

Crosses represent local linear directions



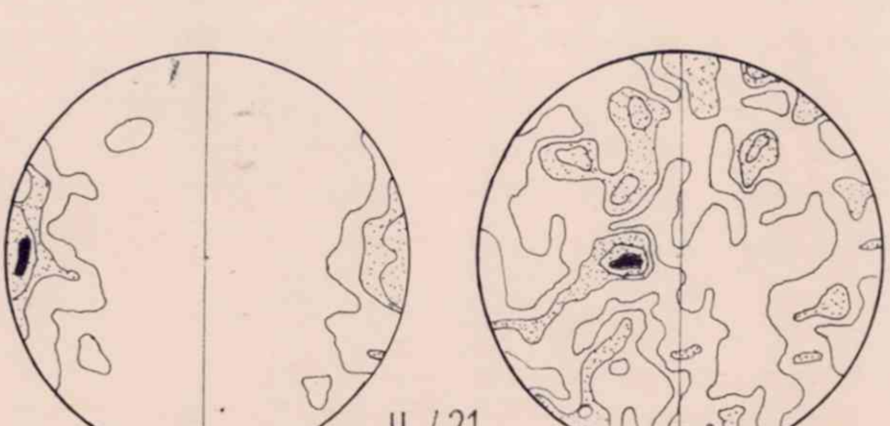
100 Muscovite
Contour 1,4,8,16 %

150 Quartz
Contour 1,2,3,7 %



100 Biotite
Contour 1,4,12,16 %

200 Quartz
Contour 1,2,3 %



100 Biotite
Contour 1,4,8,16 %

200 Quartz
Contour 1,2,3,4 %

Fig 53

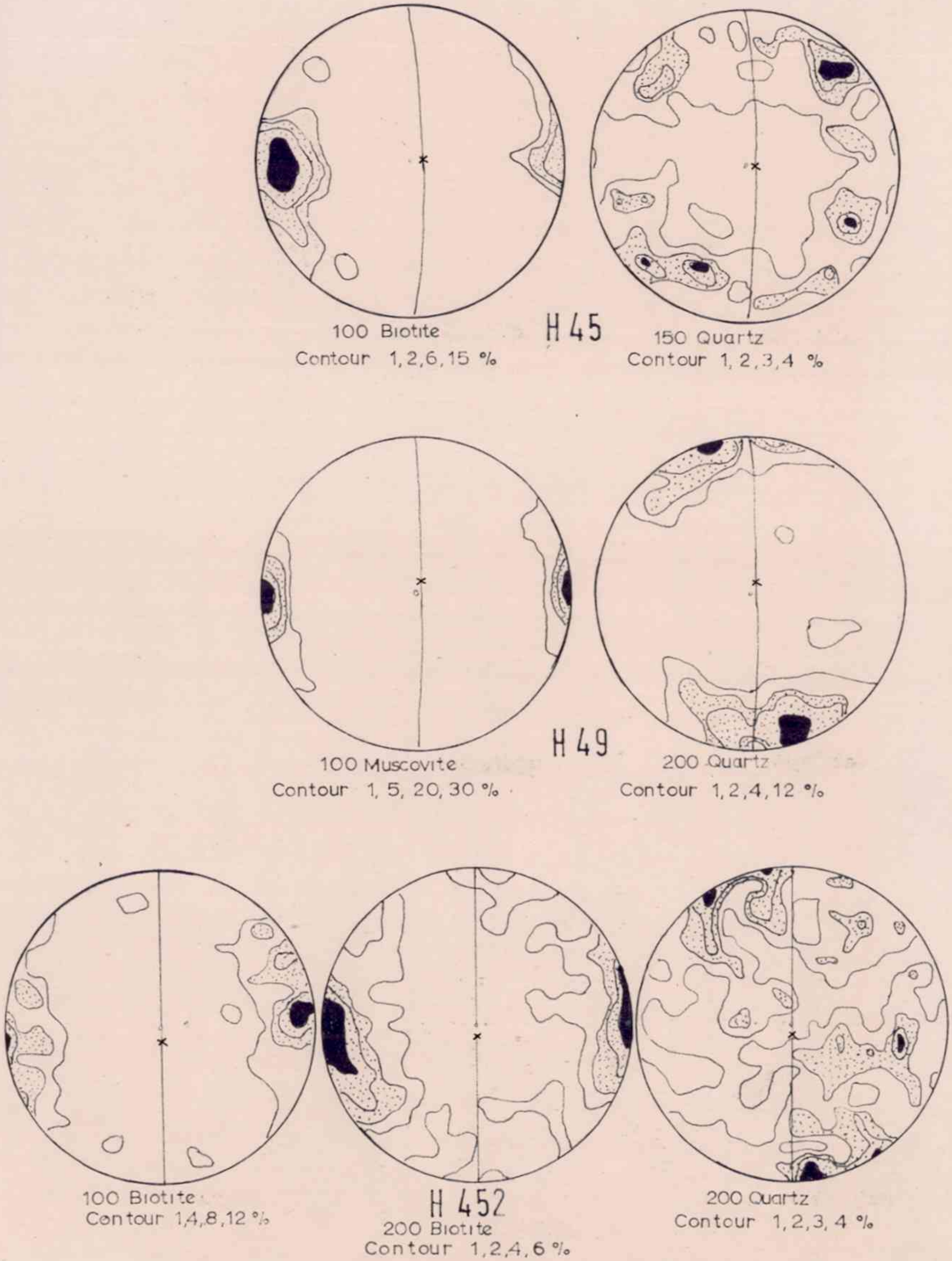
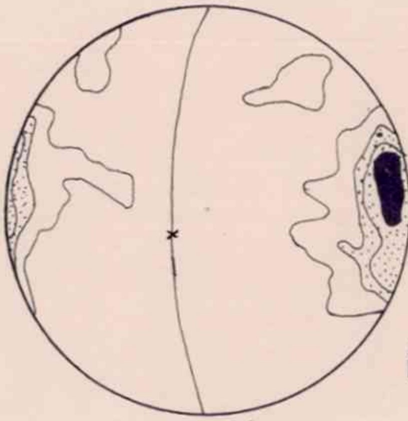
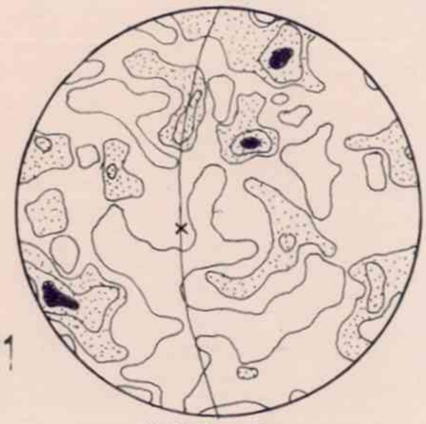

100 Biotite
Contour 1,2,6,15 %
H 45
150 Quartz
Contour 1, 2, 3, 4 %
100 Muscovite
Contour 1, 5, 20, 30 %
H 49
200 Quartz
Contour 1, 2, 4, 12 %
100 Biotite
Contour 1,4,8,12 %
H 452
200 Biotite
Contour 1,2,4,6 %
200 Quartz
Contour 1, 2, 3, 4 %

Fig 54

H 21

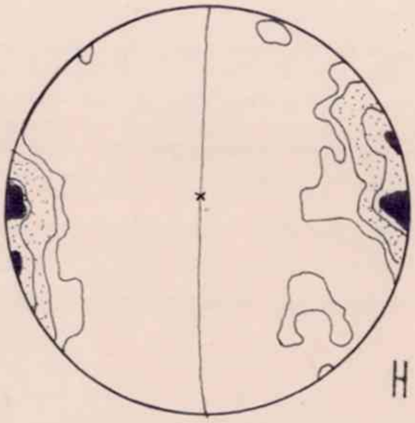


100 Biotite
Contour 1,4,8,12%

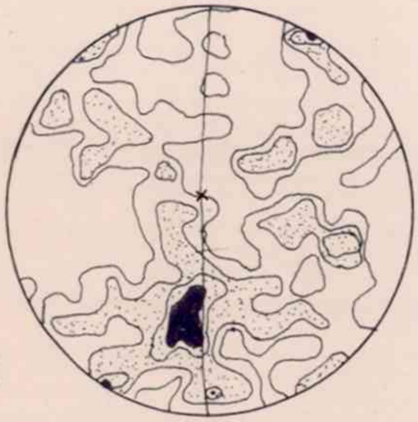


200 Quartz
Contour 1,2,3,4 %

H 80

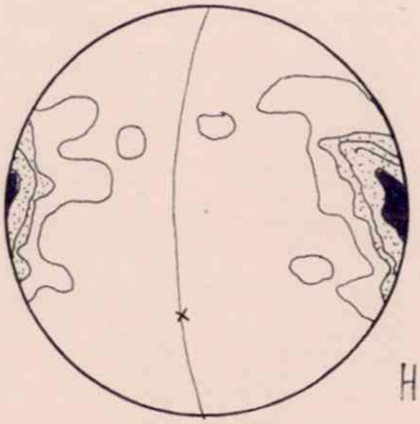


100 Biotite
Contour 1,2,4,12%

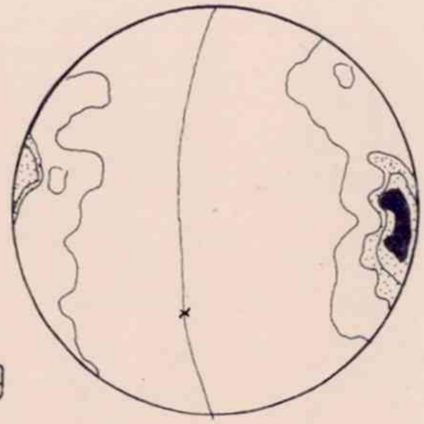


200 Quartz
Contour 1,2,3,4 %

H 19



100 Biotite
Contour 1,2,5,20%



100 Muscovite
Contour 1,5,10,15%

Fig 55

Single Maximum Diagrams

The most pronounced of these are found in the quartzites, one of which shows a 12% maximum. H 452, a migmatitic rock on the exact contact of the Bjellåtind granite, and close to one of the quartzites, shows a similar but less pronounced single maximum. It has an incomplete ac girdle in association with the single maximum. Other rocks which also possess this form of diagram are H 421, another granite, and H 80, a silty schist from South Lysvand. Neither of these show as good concentrations as the quartzites; in H 421 the scatter occurs along a broad girdle sub-parallel to the ab plane, while in H 80 the scatter may possibly be related to two broad girdles, intersecting in the maximum, and both oblique to the ab plane. It may be significant that in all the examples of a single point maximum, the maximum is always situated at a small distance from the 'a' direction. The reason for this arrangement is unknown.

Microscopic examination of this group of rocks shows that the quartz crystals, in general, have a uniform appearance. This is most clearly displayed in the case of the quartzites ~~(fig.)~~, where the interlocking crystals are seen to have a form elongation parallel to the c axes. A similar, but less pronounced, phenomenon can be observed in the other rocks, where the quartz is usually concentrated into small veins, the component crystals of which are again roughly elongated parallel to the c axes.

Girdle Diagrams.

The most pronounced of these is from the Bjellåtind granite (H 45) which has an ac girdle with four distinct maxima arranged symmetrically about the foliation plane. H 417, a contact schist to the Bjellåtind granite, shows a less well developed girdle of the same general arrangement. In each case the maxima are between 3 and 4%.

In contrast with the rocks showing point maxima, the quartz ~~crystals~~ from these two rocks generally occurs as separate equigranular crystals. If it is concentrated into small veins, the component crystals show no form orientation.

One of the rocks (H 21), a granite from the Bjellåtind mass, shows no readily distinguished pattern, although there may be a broad girdle at an angle of about 30° to the foliation plane. The quartz is present in the rock in small veins.

Significance of Quartz Fabrics

Single maxima diagrams, with a concentration in 'a' are common, from slikenside zones (Fairbairn 1949). They are regarded by some authorities as having developed by recrystallisation of needles of fractured quartz that are elongated parallel to a (Sander, Griggs and Bell, quoted in Fairbairn). Girdle diagrams are common in many rocks from folded terrain, e.g. in many of the fabrics from the Moine rocks (F.C. Phillips 1945). They are often regarded as having developed by rotational movement about 'b' during forward movement of the crystals. There is no experimental evidence to suggest that either of these processes operates in nature, and the reasons for the differences in patterns observed is unknown.

As found in the rocks of other folded belts, the quartzites of the Ørnes region show a much more well-defined pattern than the quartzo-felspathic schists and granites. There is no visible difference in the two latter groups, and it appears that the felspathic fraction of the rock prohibits orientation of the c axes of quartz to some extent. This is well displayed in the case of H 49 and H 452, which come from close to one another. Each show a similar fabric pattern, but that in the quartzite is much more well-defined. There appear to be three possibilities for this poorness of fabric development:-

(1) The felspathic fraction of the rocks partially prohibits the formation of the fabrics.

(2) Later recrystallisation, possibly an associate of the formation of microcline, has partly destroyed the fabric; in the quartzites, this phase of recrystallisation is not represented.

(3) The felspathic rocks contain the remnants of an earlier fabric that has been entirely obliterated in the pure quartzites. The incomplete and poorly developed girdles seen in some of the felspathic rocks that bear no obvious relationship to the F_2 folds may represent relict F_1 fabrics. Due to complete lack of information as to the orientation of F_1 axial directions, this cannot be determined.

Which of the three explanations is the correct one cannot be established on the information available. A late-stage to the crystallisation of the quartz crystals in the quartzites is suggested, by the way in which they contain inclusions of muscovite crystals. It is probable that they crystallised during or after F_2 times, although the mechanism of their orientation is unknown. Similarly, much of the quartz in the granites and quartzo-felspathic schists appears to be of relatively late-stage crystallisation. Therefore, it is suggested that the age of crystallisation of both quartzites and quartzo-felspathic rocks is probably similar and the reason for their difference of quartz fabric is due to either (1) or (3). More comprehensive analysis, particularly in areas where the orientation of F_1 folds is known, is necessary before more light can be thrown upon the subject.

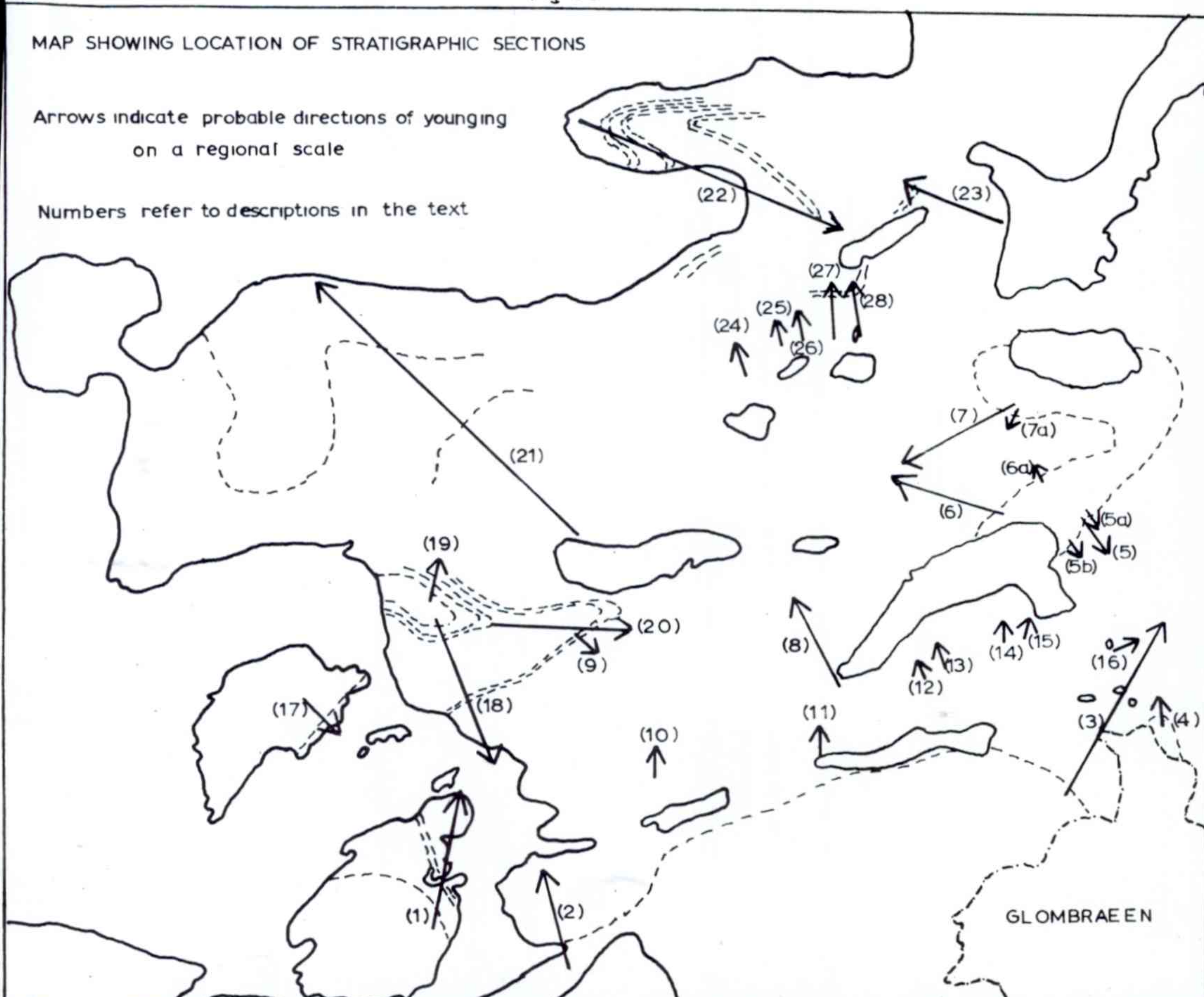
The present study does indicate that a large proportion of the quartz material has probably recrystallised during or later than the F_2 period of deformation.

Successions and Stratigraphy

Once the major structure has been established, it is possible to consider original stratigraphic relationships. Effects of F_2 and F_3 folds upon the observed tectonic successions are relatively easy to eliminate, but the isoclinal F_1 folds may be difficult or impossible to recognise. Certain of the F_1 isoclines may be recognised by lithological repetition even though deformation has been so intense as to render the cores invisible. However syntectonic slides may have eliminated part of the fold, and lithological repetition will then no longer be present (e.g. in the isoclinal folds on the north of Kvittind, p. 49).

The descriptions of successions with probable stratigraphic significance that follow are therefore tentative, and although the effects of all identified important slides and isoclinal folds are taken into account, there may be others which have not been discovered. As far as possible, each of the sections illustrated has continuity of field exposure combined with a tectonic unity. For example, although the Galtskart succession has at least two important F_1 isoclines, it was thought preferable to represent it as a continuous section, with indications of the folds rather than an overcomplicated description with a series of smaller sections. In the field, the section is a tectonic unit, and superficial examination suggests that it represents a unique stratigraphic succession.

Fig 56


MAP SHOWING LOCATION OF STRATIGRAPHIC SECTIONS
Arrows indicate probable directions of younging on a regional scale
Numbers refer to descriptions in the text
(1)
(2)
(3)
(4)
(5)
(5a)
(5b)
(6)
(6a)
(7)
(7a)
(8)
(9)
(10)
(11)
(12)
(13)
(14)
(15)
(16)
(17)
(18)
(19)
(20)
(21)
(22)
(23)
(24)
(25)
(26)
(27)
(28)
GLOMBRAEEN

Each of the stratigraphic sections has a vertical scale included which represents the approximate thicknesses along the measured section: this may bear little resemblance to the original sedimentary thickness. For example, the Suppevand section (fig.58) is measured across the nose of the Kjeipen Fold, and is clearly very much thicker than its equivalent on Galtskart. Consequently all variations in thickness indicated in figs. 58-63. may be caused by a combination of structural and stratigraphic controls, and the main significance is the order in which the beds occur.

Exact recognition of unique successions has been of prime importance to enable structural correlation across the large tracts of unexposed ground, particularly in the west of the region. Several distinctive rock types were useful in this connection, notably the Calcareous Pelitic Schist, the Silty Schist, the Hornblende Rock (a calcareous gneiss) and a Euhedral Garnet Kyanite Schist. The latter is almost certainly a unique stratigraphic band, consisting of plentiful small euhedral pink garnets and spasmodic poorly oriented kyanite crystals set in a micaceous groundmass. It can be traced in a band about 5 metres thick over a distance of 15 km, and forms part of the most distinctive stratigraphic succession of the Ørnes region. Other characteristic rock types are also useful for correlation, notably marbles, calc-schists and amphibolites, granites and quartzites. The more abundant pelitic and semi-pelitic material is only useful in conjunction with these distinctive bands.

On each of the stratigraphic sections the varied lithologies have been subdivided in the same fashion as on the main geological map. The ornament, shown in fig.57, is consistent throughout, while certain mixed groups, e.g. calc-schists and marbles, or semi-pelitic and pelitic schists are also indicated.

Structural evidence suggests the Glomfjord Granite may be regarded as situated at the base of the succession (see ~~p. and~~ fig. 11), and it is also regarded as representing the lowermost stratigraphic unit (see Hollingworth et al 1960, Walton 1959, Nicholson 1960). From considerations of the major structure, correlation is made between the Glomfjord and Bjellåtind granites, and it is upon these assumptions that the whole of the succeeding descriptions are based. In fig.56 the directions of regional younging are indicated on the basis of this relationship, Local reversals due to F_1 isoclines may be present.

Regionally, the illustrated sections can be divided into several groups that from structural evidence may be correlated directly:

(a) Three sections away from the Glomfjord Granite, three from the Bjellåtind and one from the Fykan granite form one group.

(b) Three sections across the Blaatind Massif, and one from Teksmona form another group, and probably correlate with those of section (a).

(c) In the NW of the region, three widely separated sections are described.

In addition to these, three groups of local sections, whose stratigraphic significance cannot be disputed,are illustrated:-

(d) Eight sections along the strike from Blaatind to Tverfjell are correlated. Two of these are separated structurally from the remainder, due probably to the action of F_1 slides; the exactness of correlation and the absence of any equivalent succession elsewhere strongly suggests the uniqueness of this group.

(e) Four sections in the metasediments immediately above the Bjellåtind granite are included to show the details of variation in the contact rocks. No equivalent sections could be studied in the case of the Glomfjord granite, due to lack of exposure.

(f) Five sections giving details of part of the North Markvand series are described, to show the persistence of thin bands over large distances of outcrop.

(a) Sections 1-8 away from basal granites (fig.58).

The successions that have been examined in most detail are those above the Bjellåtind granite, on Galtskart (sections 6 and 7) where the rocks are excellently exposed. Section 6, from South Galtskart is used as a standard to which the remainder are related. Five good marker bands are present, each of which persists northwards to Sörfjord and southwards to the Suppevand-South Lysvand region. The recurrence of such distinctive lithologies as the silty schist and calcareous pelitic schist suggests strongly the possibility of major F_1 isoclines, the most important of which is the Galtskart fold whose axial plane is situated within the upper calcareous pelite (see p. 50). Another isoclinal fold centred about a

KEY TO ROCK TYPES OF STRATIGRAPHIC SECTIONS

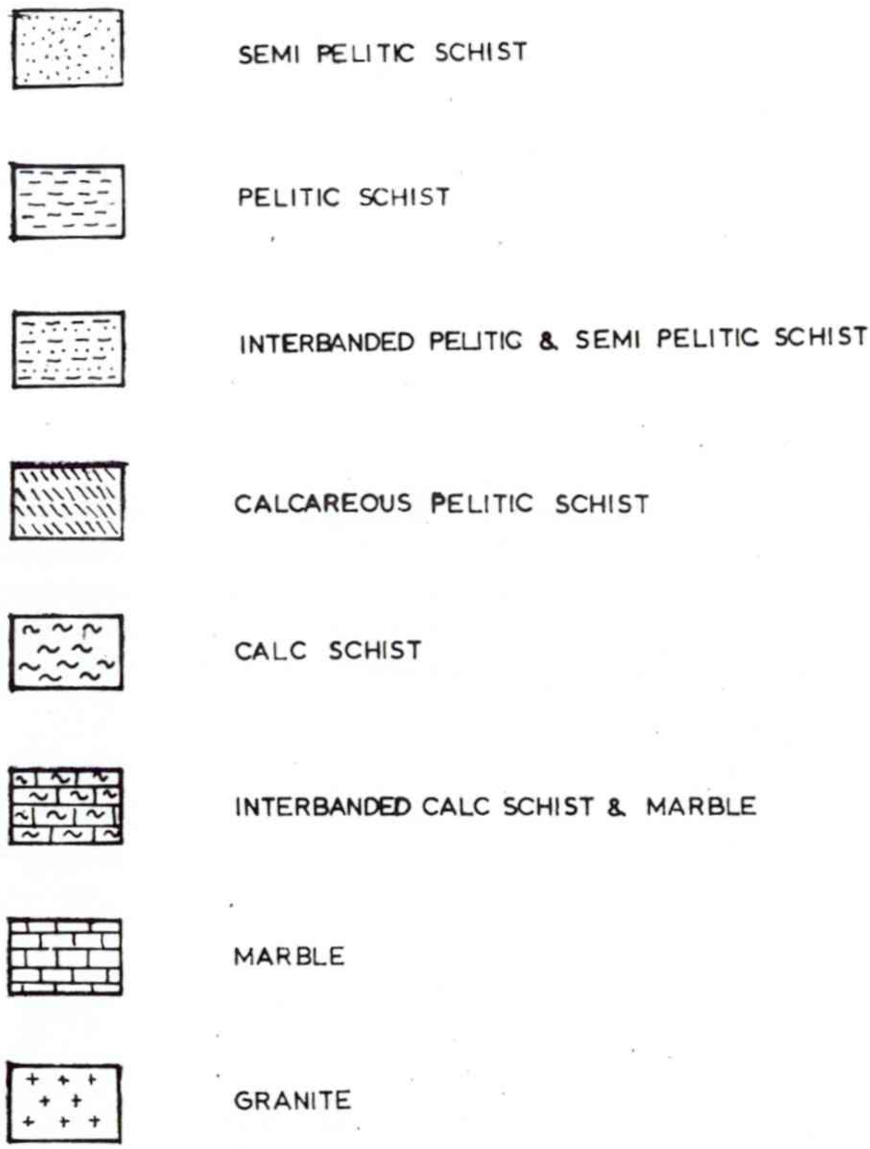
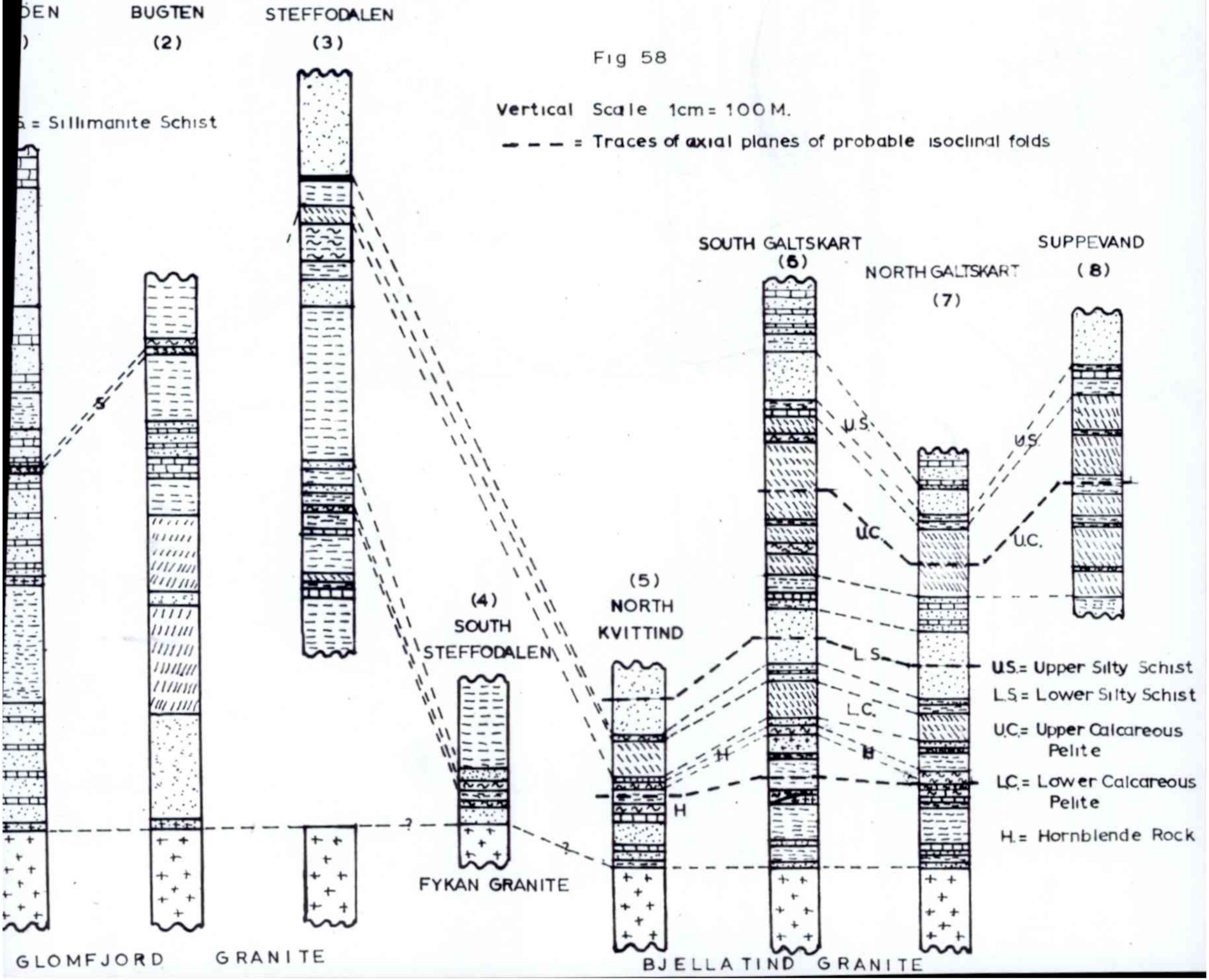


Fig 57

Figs 58-63. Stratigraphic sections, numbered according to their locations in fig.56


ÖEN
)
BUGTEN
(2)
STEFFODALEN
(3)
Fig 58
Vertical Scale 1cm = 100 M.
= Traces of axial planes of probable isoclinal folds
S = Sillimanite Schist
SOUTH GALTSKART
(6)
NORTH GALTSKART
(7)
SUPPEVAND
(8)
(4)
SOUTH
STEFFODALEN
(5)
NORTH
KVITTIND
U.S.
U.C.
L.S.
L.C.
H
U.S. = Upper Silty Schist
L.S. = Lower Silty Schist
U.C. = Upper Calcareous Pelite
L.C. = Lower Calcareous Pelite
H. = Hornblende Rock
FYKAN GRANITE
GLOMFJORD GRANITE
BJELLATIND GRANITE

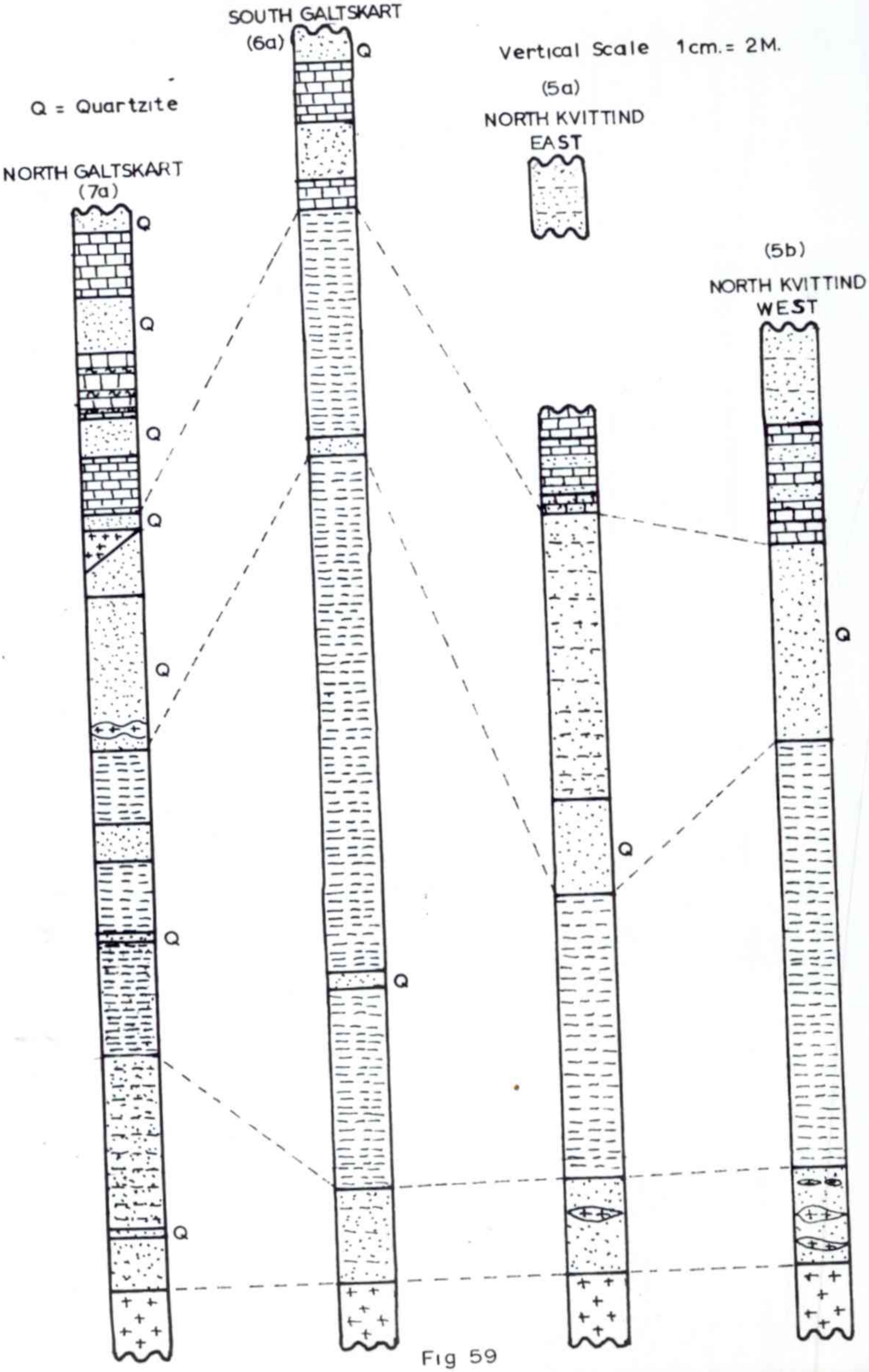

SOUTH GALTSKART
(6a)
Q
Vertical Scale 1cm.= 2M.
(5a)
NORTH KVITTIND
EAST
Q = Quartzite
NORTH GALTSKART
(7a)
Q
Q
Q
Q
Q
Q
Q
(5b)
NORTH KVITTIND
WEST
Q
Q
Q

Fig 59

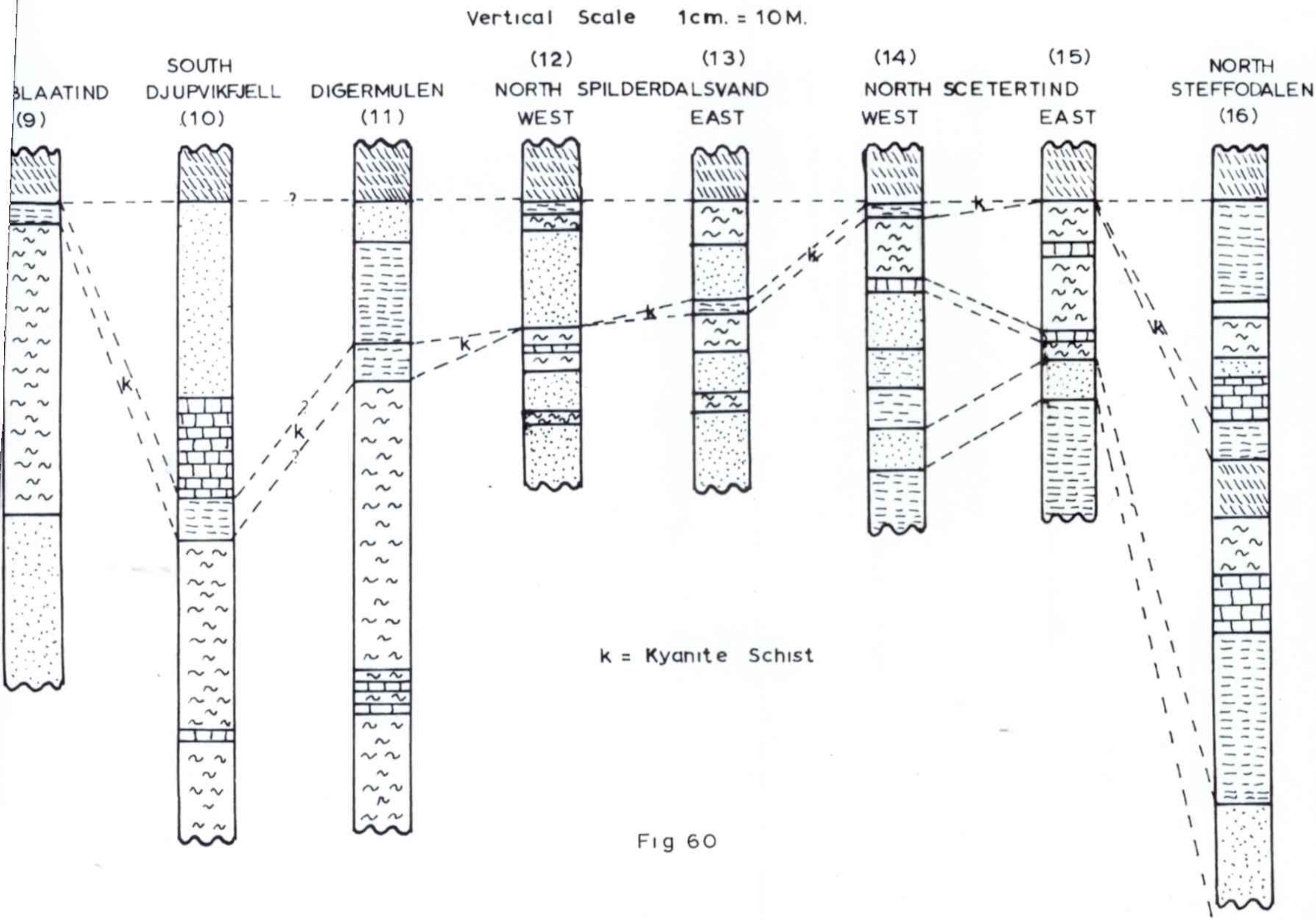

Vertical Scale 1cm. = 10M.
BLAATIND (9)
SOUTH DJUPVIKFJELL (10)
DIGERMULEN (11)
(12) NORTH SPILDERDALSVAND WEST
(13) EAST
(14) NORTH SCETERTIND WEST
(15) EAST
NORTH STEFFODALEN (16)
k
?
k = Kyanite Schist

Fig 60

coarse-grained mica schist has been identified on North Kvittind by Wells and Bradshaw. Two distinct layers of hornblende rock are present, the lowermost of which is eliminated northwards and westwards by the action of a slide. The same slide probably also accounts for the absence of repetition beyond the hornblende rock. A further isoclinal fold also probably exists within the silty schist of North Kvittind, to project across Lysvand and reappear on Galtskart.

Thus, the Galtskart succession which possesses the appearance of a simply bedded succession is probably repeated by a series of isoclinal folds which have increased the thickness of the original stratigraphic succession by a factor of three or four, to the present 1100 metres.

The thick marble-quartzite band at the top of sections 6 and 7 represents the position of the Galtskart slide and separates the Galtskart succession from that of North Markvand.

Structural considerations indicate that repetition between the Bjellåtind and Glomfjord granites occurs about the silty schist of North Kvittind (fig.11), which is taken to represent the youngest rock, stratigraphically, in the southern part of the Ørnes region. In the Steffodalen region the immediate underlying rocks to the silty schists correlate closely with those of North Kvittind (sections 3 and 5), but succeeding these is a thick group of banded semi-pelitic and pelitic schists not seen in the sections over the Bjellåtind granite. Due to the lack of exposure, details of possible repetition by isoclinal folding cannot be determined.

although certain isoclines, notably in a marble band of West Steffodalen, have been identified. It appears that a large part of these pelitic and semi-pelitic rocks underlie, and are stratigraphically older than, the metasediments on Galtskart, their elimination northwards being caused by an F_1 slide.

Continuous exposure to the Glomfjord granite contact over 1000 metres is visible at Bugten (section 2). The lithologies encountered are very similar to those in the Galtskart-Steffodalen region, but detailed correlation is impossible. Indeed, correspondence with the neighbouring island of Mesøen (section 1) is inexact except for the upper part of the succession, where a staurolite-bearing schist is particularly useful for correlation. This lack of conformity between the various successions is probably mainly caused by isoclinal folding. Continuous exposure along the strike is never adequate to show how the various bands are eliminated laterally.

The Fykan granite, regarded as the equivalent of the Glomfjord granite, has a metasedimentary cover of similar lithologies (fig. 58). A tentative correlation with the Steffodalen rocks is shown.

In each case a similar group of metasediments is encountered above the presumed basal granites. Sometimes, correlation between measured sections is good (e.g. sections 5, 6 and 7) but elsewhere cannot be demonstrated with such certainty. Nevertheless, the sediments are regarded as belonging to a single stratigraphic unit, the lower part of which is composed of the banded semi-pelitic and pelitic schists of Steffodalen, and the upper part of the mixed calcareous pelitic and silty schist group of Galtskart.

(b) The Blaatind and Teksmona Sections (17-20, fig. 61)

Structural equivalence of the Glomfjord and Teksmona-Blaatind granites has already been suggested (p. 87). The majority of the associated metasediments consist of interbanded semi-pelitic and pelitic schists, marbles, together with sheet granites, and correspond closely to the rocks of Masöen. The two sheet granites are lithologically identically to the basal granites and show a general increase in thickness towards the NW. One of them is represented by a thin band on Mesöen, but neither is seen in the Bugten succession. The lenticular nature of the Ultrabasic mass, only seen in section 18, is indicated. Contacts of the Ultrabasic mass with the surrounding metasediments are always poorly exposed. Towards the top of each of the sections, a calc-schist of varied thickness is indicated, the probable lateral equivalent of the Ørnes Gneisses (see also p. 283); in section 20, the uppermost calcareous pelitic schist correlates with the upper calcareous pelite of Galtskart (section 6).

The present arrangement of basal granite overlying metasediments is due to F_2 inversion, and has been corrected for in sections 17-20.

(c) North Markvand Sections 21-23 (fig.62)

The Galtskart Slide, whose position is shown on map 3, separates the rocks of groups (a) and (b) from those of group (c). For the

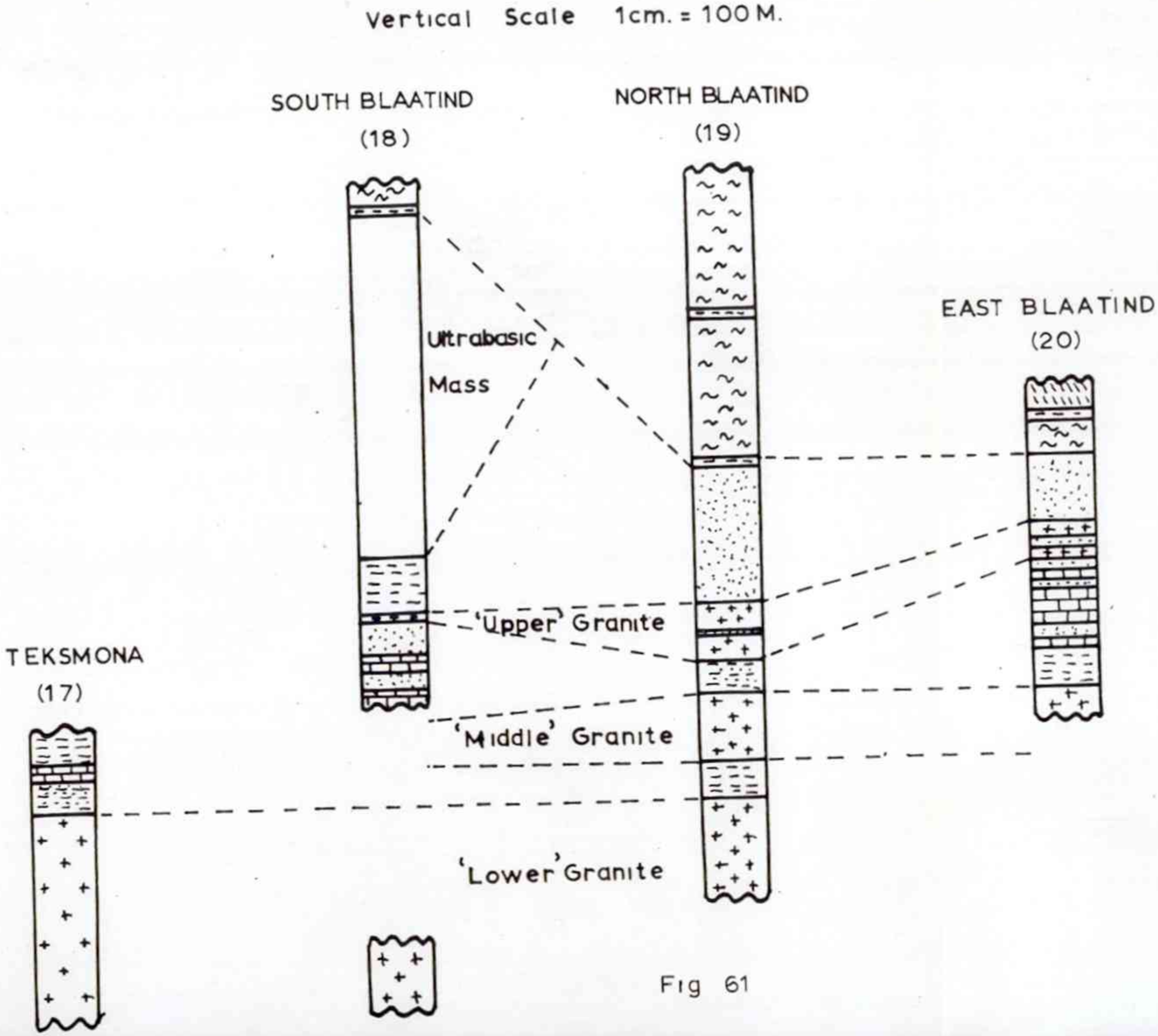
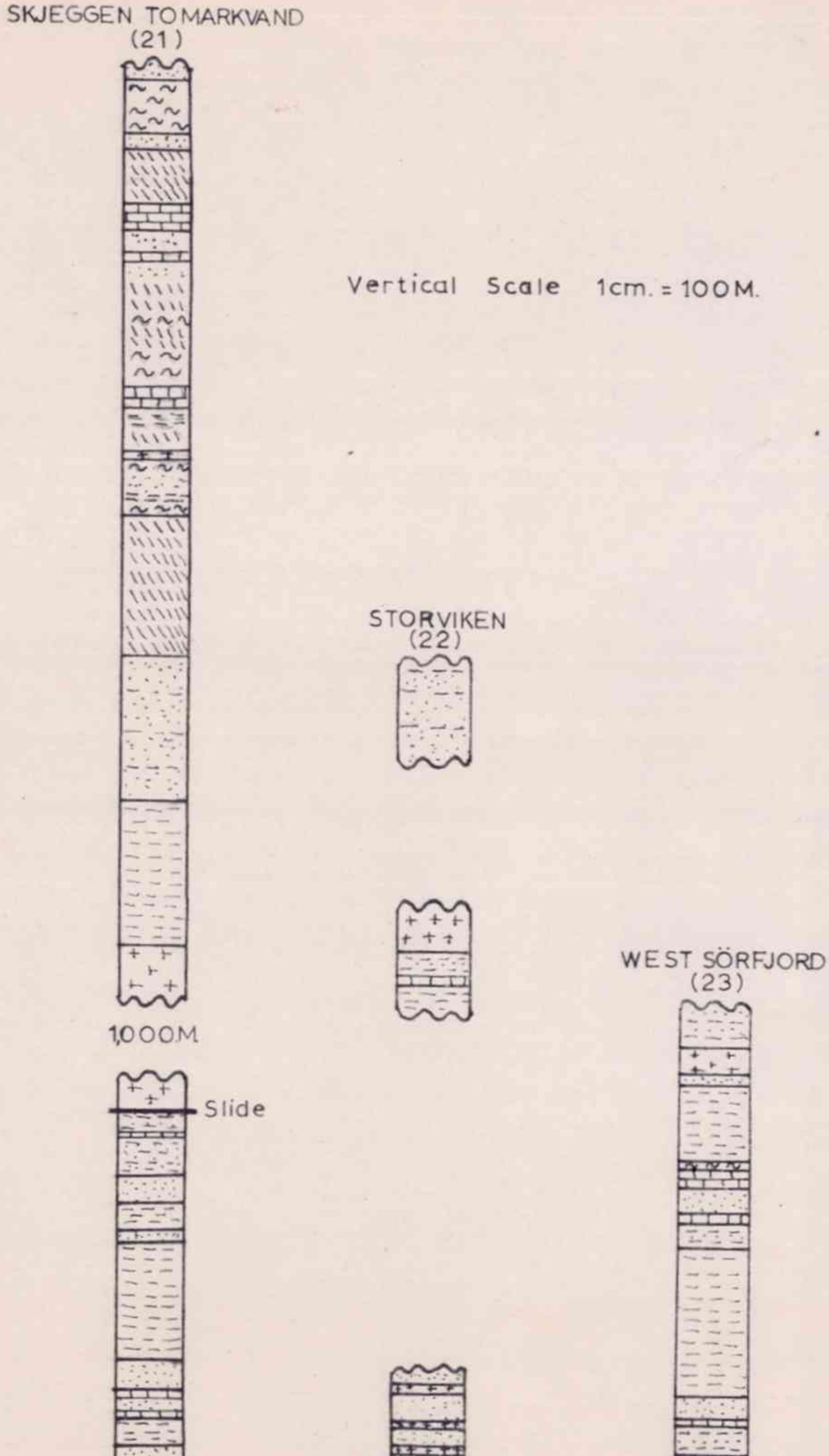

Vertical Scale 1cm. = 100 M.
SOUTH BLAATIND (18)
NORTH BLAATIND (19)
EAST BLAATIND (20)
TEKSMONA (17)
Ultrabasic Mass
'Upper' Granite
'Middle' Granite
'Lower' Granite

Fig 61


SKJEGGEN TOMARKVAND
(21)
1,000M.
Slide
Vertical Scale 1cm. = 100M.
STORVIKEN
(22)
WEST SÖRFJORD
(23)

most part, the North Markvand rocks consist of interbanded pelitic and subordinate semi-pelitic schists, with occasional marbles and amphibolites. Calcareous pelitic schists and calc schists are rare, except in the extreme NW, on the N of the Skjeggen granite (map 1). The presence of calcareous pelitic schists in sections 22 and 23 is due to the reappearance of the uppermost part of the Galtskart succession. Major lithological repetition by F_1 isoclines in the North Markvand rocks is absent.

The most complete succession is that from Skjeggen to Markvand (21). The rocks are younging continually northwards, due to the position on the upper limb of the Kjeipen fold. From Markvand to the Skjeggen granite are 600 metres of interbanded pelitic and semi-pelitic schists with occasional marbles; this represents the lateral equivalent of a series of similar rocks which eastwards, in the Skromdalen region, are about 1000 metres thick. The variation is probably due to the differing situation on the Kjeipen fold (fig.9).

The succeeding Skjeggen Granite, which is here over 1000 metres, is lithologically similar to the basal Glomfjord granite and is regarded as its lateral equivalent. Its southern boundary with the North Markvand pelitic schists is probably a slide of some importance (see p. 95), while the metasediments to the N are regarded as the equivalents of the Steffodalen and Galtskart successions. It is significant that the rocks nearest the granite consist of pelitic and semi-pelitic schists similar to the Steffodalen rocks, while the succeeding calcareous group has the same appearance as many of the members of the Galtskart group. The order in which they occur is therefore the same as that inferred in the case of the metasediments in the southern part of the Ørnes region.

At the base of section 22 is a calcareous pelitic schist that is correlated on structural evidence with the upper calcareous pelitic schist of Galtskart (sections 6 and 7). Lying above it is a series of marbles and flaggy schists, correlated with the marble-quartzite band of West Galtskart, and succeeding this is a group of thin sheet granites and siliceous schists. The sheet granites are very finely banded and have a 'sedimentary' appearance (see fig. 50, 51). These represent the lowest part of the North Markvand series; the granites are not found in an equivalent position elsewhere in the region, and in the Storviken peninsula itself are seen to be extremely impersistent. A large zone of unexposed ground, covered by the Storviken bay is followed by a group of flaggy schists overlaid by a further sheet granite. The upper boundary of the latter is not exposed and the thickness of the granite is unknown. However, it is probably a maximum of 200 metres, and as it is almost certainly the lateral equivalent of the Skjeggen granite, it is clear that the latter thins eastwards. In an equivalent position east of section 22, no granite is visible, probably due to the influence of the Skjeggen slide. In the inaccessible cliffs of West Høgnakken a series of sheet granites is clearly visible. The rapid elimination of at least one of these is shown in fig. 15.

The lower part of section 23 represents the uppermost rocks of the Galtskart succession, and the calcareous pelitic schist indicated probably correlates with that at the base of section 22. In the succeeding group of marbles and quartzites is the probable northerly projection of the Galtskart slide, and following this is the North Markvand pelitic series. The latter corresponds closely with the equivalent rocks of section 21. Towards the top of the section is a sheet granite, which outcrops on the northern shore of Storvikvand. This granite dies out westwards on the N face of Degro.

(d). Detailed sections Tverfjell-Blaatind (Sections 9-16) (fig.60)

These can be divided into two distinct groups that are now structurally separate. All the sections have a calcareous pelitic schist at the top, but whereas in sections 9 and 10 it is the West Lysvand calcareous pelite, in sections 11-16 it is that of North Scetertind. Structural considerations suggest that the two layers are stratigraphic equivalents, now separated by the Lysvand Slide (see p. 97), and the present correlations are therefore justified.

Very similar rock types are present in all the sections, the most outstanding of which is the euhedral garnet kyanite schist. The remainder of the succession beneath the calcareous pelitic schist consists of varied calc schists, marbles, and semi-pelitic schists.

Sections 9-12 are worthy of closer examination with regard to the nature of the calc-schists present. In section 9, taken from

East Blaatind, they consist of massive, finely-banded green hornblendic rocks which make up the main calc-schist of Blaatind (see also p. 283). That of section 10 is of similar lithology, and represents the marginal facies of the thick Ørnes Gneisses; the latter continue southwards over a large outcrop, and are at least 4-500 metres thick. The calc schist of section 11 is again of a similar nature; the amount of its southerly extent is unknown due to the location of Spilderdalsvand. In both sections 10 and 11 thin marbles in equivalent positions occur within the calc schists. When traced further to the E the calc schists are very much thinner, and contain a higher proportion of calcareous minerals. In many of the sections, interbanded marbles are present. It therefore appears probable that the Ørnes Gneisses are lenticular in outcrop, and are represented both eastwards and westwards by much thinner calc schists (see also p. 285).

The suggested correlation of the two separate bands of calcareous pelitic schist has already been noted, and it is necessary to account for the lack of this associated distinctive stratigraphic section along the remainder of the southern boundary of the West Lysvand calcareous pelitic schist. The principal reason is probably the occurrence of the Lysvand Slide which follows this boundary, and indeed a tectonic lens of material very similar to the calc schist is seen in the cliffs on the southern shore of Lysvand. Similar rocks are also present near the lower calcareous pelitic schist of Galtskart, supporting the suggested equivalence of this with the upper band of calcareous pelitic schist. In no case is a complete stratigraphic section comparable with that of sections 9-16 obtained, probably for tectonic reasons.

(e) Bjellåtind Granite Sections (5a, 5b, 6a, 7a (fig.59).

Each of these sections shows in detail the lowermost 40 to 50 metres of the metasedimentary sequence, and are included to show the variation of the immediate cover to the basal granite over three km of outcrop. Each distinctive band that is more than about 20 cm thick is represented, and the sections probably give an exact representation of the original stratigraphy; the basal slide separating the granite and its immediate cover of schists is thought to be situated above these sections (see also p. 42). Correspondence between the various sections is good, and a four-fold division can be established:-

(1) 3-4 metres of flaggy semi-pelitic schists in contact with the granite, containing tourmaline-bearing pegmatites in sections 5a and 5b. Lenticular quartz masses are also common.

(2) A varied group pf pelitic schists, predominantly garnet-bearing, and with occasional thin quartzite or semi-pelitic schist layers. Quartz segregations are common in all but section 7a. The thickness of the group varies from a maximum of 30 metres in South Galtskart, to a minimum of 10 metres in North Kvittind (5a).

(3) In every section except 6a, the pelitic group is followed by a series of quartzites or siliceous schists, about 6 metres thick, and forming the base to the succeeding group.

(4) A series of marbles and flaggy schists with occasional

quartzites forms the top of the detailed sections. Those on Galtskart consist of differentiated grey marbles and quartzites with a few calc-schists in section 7a, while those on North Kvittind contain impure marbles and flaggy schists.

An even more detailed section over 3 metres of the basal metasediments in shown in fig. 66 , and its petrography described separately (p. 233).

(f) Skromdalsvand Sections (24-28, fig.63).

These sections, taken over a distance of about 3 km of outcrop, show details of the upper part of the North Markvand pelitic series. Very good correlation exists of bands sometimes less than 5 metres thick; the most distinctive of these is a big garnet schist.

The Degro granite shown in sections 27 and 28 can be seen to die out abruptly westwards on the north face of Degro. In an equivalent position on Spantind is a series of pelitic schists with thin interbanded amphibolites; further west, in section 21, similar rocks are much nearer the Skjeggen granite. This is due to the action of the Skjeggen slide (p. 95). Northwards from the Degro granite, in sections 27 and 28, a thick group of pelitic and semi-pelitic schists is present before the eastern continuation of the Skjeggen granite is reached. These are not represented further to the W, again due to the slide.

General Conclusions and Regional Stratigraphy

Considerations of the separate successions described above

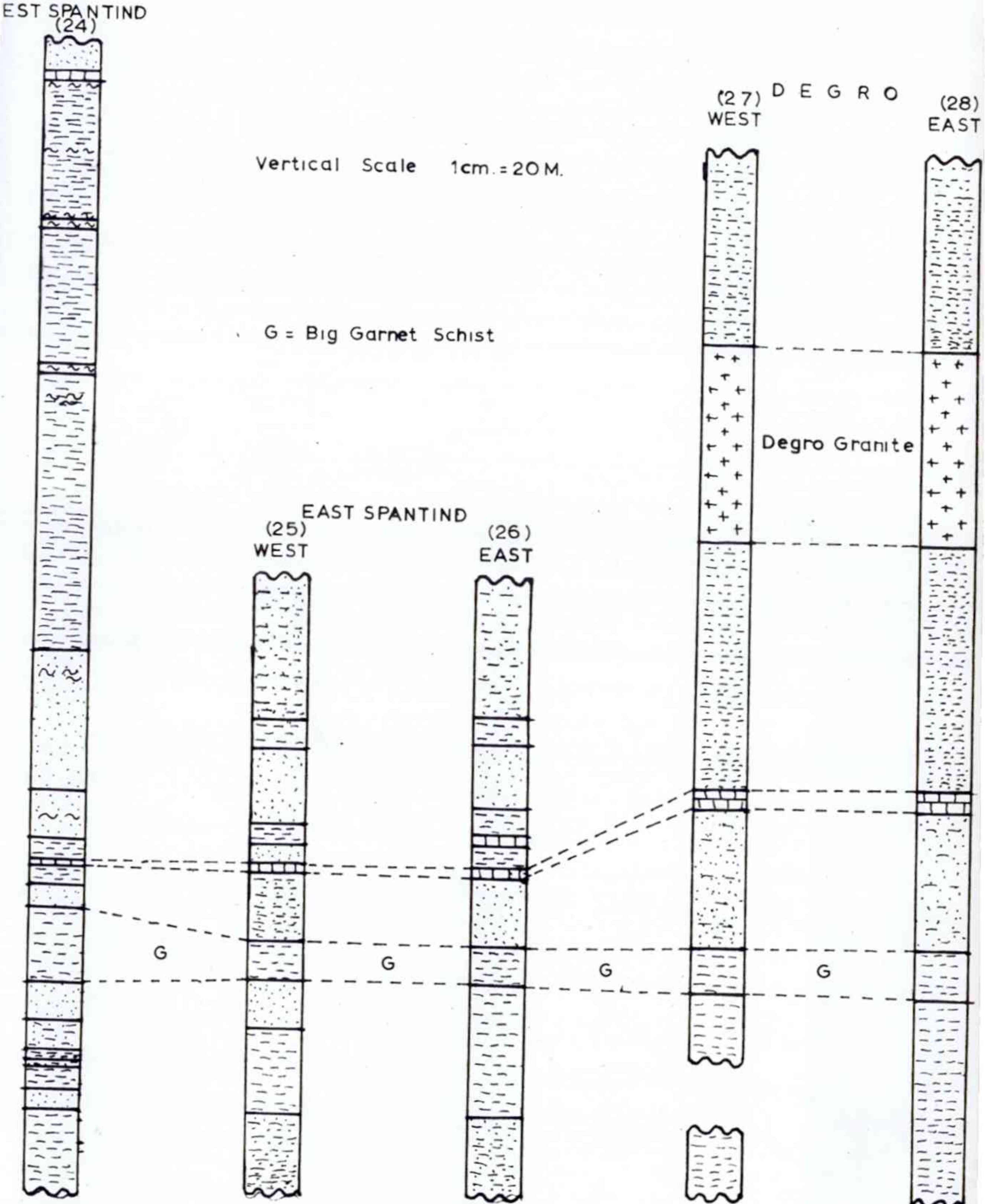

EST SPANTIND
(24)
Vertical Scale 1cm.=20 M.
G = Big Garnet Schist
EAST SPANTIND
(25)
WEST
(26)
EAST
DEGRO
(27)
WEST
(28)
EAST
Degro Granite
G
G
G
G

Fig 63

allow the establishment of a broad stratigraphy, on the assumption that the major granites are situated at the base. A series of distinct groups may be recognised (summarised in fig. 64):-

(1) Basal Granite - probably at least 1000 metres thick; the base is never seen.

(2) Basal Succession - lowermost 40 metres of the metasedimentary sequence, described in detail above the Bjellåtind Granite.

(e) Steffodalen Succession - interbanded pelitic and semi-pelitic schists. The thickness is unknown, but is probably several hundred metres.

(4) Galtskart Succession - a varied group of silty schists, calcareous pelitic schists etc., probably several hundred metres thick.

(5) North Markvand Succession - principally interbanded pelitic and subordinate semi-pelitic schists with occasional marbles and amphibolites. Sheet granites are also present in the northern part of the outcrop. Thickness unknown, probably 4-500 metres.

(6) Skjeggen Granite - the thickest of the 'sheet' granites and very similar lithologically to the basal granites. It probably represents the reappearance of the basal granites. At least 1000 metres thick.

(7) Skjeggen Succession - lithologically very similar to groups 3 and 4 and probably represents their equivalent. Several hundred metres thick.

TENTATIVE STRATIGRAPHIC RELATIONSHIPS OF THE ÖRNES REGION

(Scales of thicknesses are unknown)

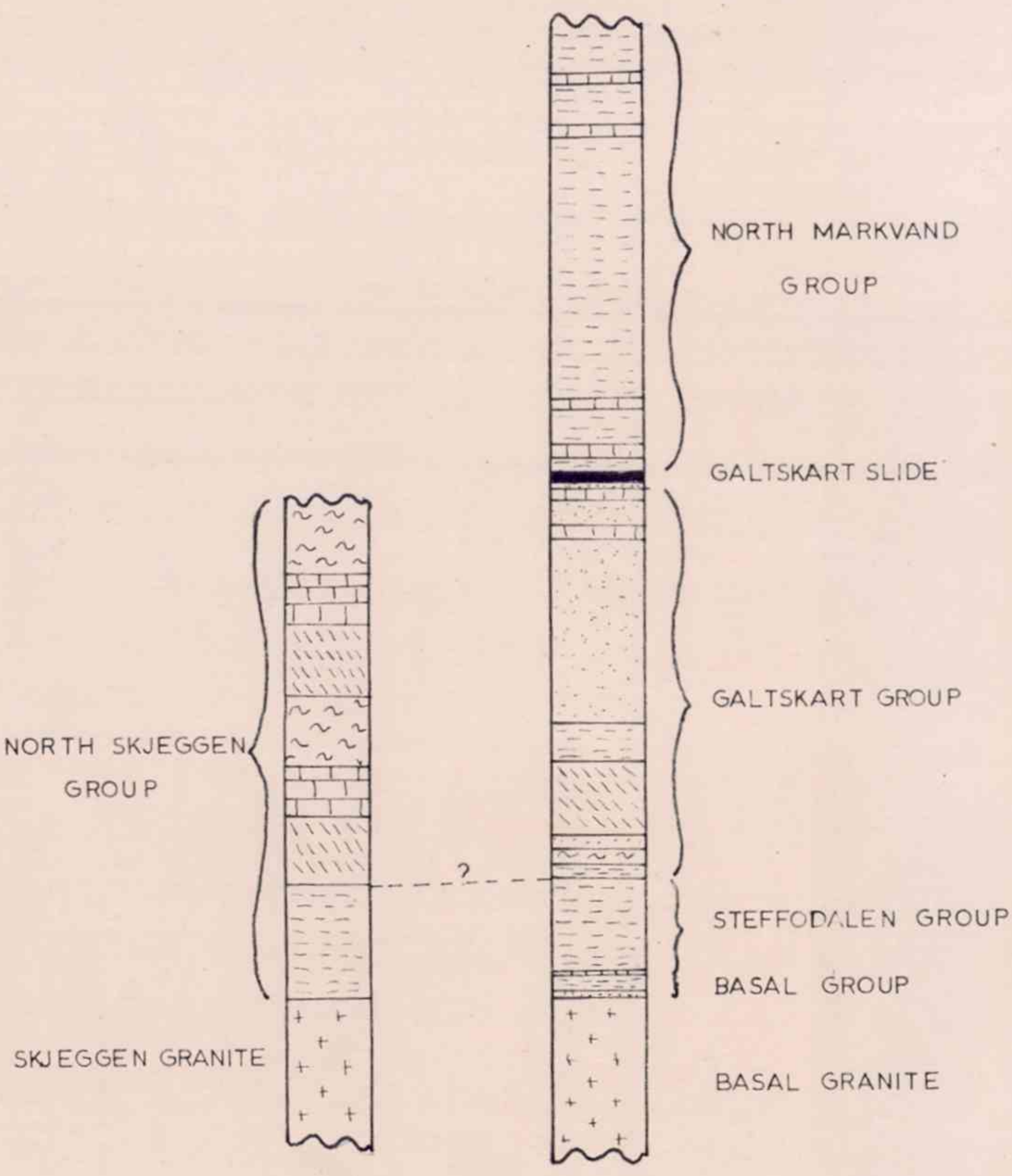


Fig 64

The basal granites are regarded as representing granitic basements before sedimentation took place (see also p. 232). Thus, the local succession represents the initial transgressive deposits, followed by the possible flysch deposits of the Steffodalen and Galtskart groups. The North Markvand group, which probably constitutes the youngest rocks stratigraphically of the Ørnes region, represents continued geosynclinal conditions. The thin layers of amphibolite which are common in this group may indicate associated igneous activity.

Comparison of the stratigraphy of similar areas in other parts of Norway is included:-

(a) Holtedahl (1938) gives a three-fold division to the rocks of the Opdal Region:

(1) Basal Gneisses - fine-grained augen gneisses and banded gneisses with evidence of recumbent folding. Their general lithology is similar to that of the Glomfjord granite, particularly its inner portions (M.A. Jones: personal communication).

(2) Sparagmite - consists of flaggy metamorphosed sandstone with layers of mica schist, lying conformably on the basal gneisses. The boundary between the two groups is diffuse, and the sparagmite is highly deformed. The general character of the sparagmite is similar to that of the silty schists.

(3) Trondheim Schists - dark, often hornblende-bearing rocks that

grade upwards into coarse mica schists. The hornblende-bearing rocks are regarded as volcanics.

(b) In Southern Norway Holtedahl has described the character of the unmetamorphosed sparagmites which comprise the Eo-Cambrian sediments lying above the basement pre-Cambrian, but beneath the lower Cambrian Holmia shales (A.J.S. 1922). Here, they consist of alternating grey or red arkosic sandstones with associated thick limestones (the Biri Limestone) and thin shales. A good horizon of tillite has been identified.

(c) The Palaeozoic formations of Finnmark (Holtedahl 1919) show a complete section of unmetamorphosed rocks overlying the basal pre-Cambrian, which briefly consist of:-

(1) Lower Cambrian - alternating shales and sandstones.

(2) Dolomite-bearing sandstone and shales (basal Ordovician) followed by a varied group of sandstones.

(3) Tillite-bearing sandstone of Middle Ordovician age.

(d) The Sulitelma District (Kautsky 1952) has been divided into two major units, viz:

(1) Archean rocks of eastern kratogenic block, mainly granitic in character, are overlain by unmetamorphosed Palaeozoic sediments.

(2) Thrust on top of these is a large nappe, the Seve Nappe, consisting of highly metamorphosed sediments, Cambro-Silurian in age, with complicated internal disturbances. Several good

marker horizons can be identified, such as the Juron Quartzite, and the Pieske Limestone. Volcanic rocks are common. A comprehensive stratigraphy is difficult to evaluate, because of the subdivision of the main nappe ihto four smaller units, but the following is suggested:-

(1) Juron Quartzite - Eo-Cambrian to Cambrian

(2) Pieske Limestone, volcanics towards base - Lower Ordovician.

(3) Conglomerate-sandstone series - Upper Ordovician.

(e) Skjeseth and Sorensen (1952) have described the rocks in a region <u>south of Holandsfjord</u>, some 20 km south of Ørnes. The rocks involved are granites and metasediments of very similar type to those of the Ørnes region, and they regard the granites as representing large thrust sheets, tectonically overlying the schists. The whole mass was then folded together. This view of the origin of the granites is fundamentally different from that suggested in the case of the granites of the Ørnes region.

(f) Walton (1959) has mapped an area some 15 km to the east of the Ørnes region (fig. 2). The stratigraphic succession, which he tentatively correlates with that of the Opdal region is:-

(1) Glomfjord Granite - partly Pre-Cambrian

(2) Psammitic Schists - comparable with the Sparagmite, and equivalents of the Silty Schist group - Eo-Cambrian.

(3) Pelitic and Calcareous Pelitic Schists - Cambrian

(4) Sokumfjell Marbles - Lower Ordovician

(g) Nicholson (1960) in an area to the south of that mapped by Walton recognises a four-fold division to the series :-

(1) Basal Granite

(2) Holmvand Group - varied pelitics and marbles, with a siliceous schist directly above the basal granite.

(3) Sokumfjell Marbles - Ordovician.

(4) Vegdal Group - very micaceous pelitic schists - Middle Ordovician.

Comparison of the stratigraphy developed in the Ørnes region with that seen in the near-by areas mapped by Walton and Nicholson, shows that important differences exist. Basal psammitic rocks above the basal granite in the Ørnes region are only very thin, and rocks of equivalent lithology, the silty schist group, are found some distance from the granite boundary (see maps 1 and 4). In parts of the area mapped by Nicholson, however, the basal psammitic group is very reduced in thickness. Elsewhere in the Caledonides of Norway, a basal Eo-Cambrian sparagmite is almost ubiquitous, and the absence of such a zone in parts of the Glomfjord region appears to be worthy of mention.

The remainder of the metasedimentary sequence in the Ørnes region is similar in generalities to that of the rest of the Glomfjord region, except in the case of the North Markvand series. This group of pelites which are predominantly coarse-grained garnetiferous rocks, do not appear to have an analogue elsewhere in the Glomfjord region. A possible correlation with the Vegdal group appears to be unlikely, as a series of typical thin sections of the latter, kindly supplied by Dr. Nicholson, and examined by the writer, are quite unlike the North Markvand rocks. The occurrence of well differentiated sheet granites appears to be restricted to the northerly outcrops of the North Markvand group, and extend NE into the region mapped by Wells.

Whether or not the Sokumfjell Marble Group is the stratigraphic equivalent of the marbles around Sörfinnset, which presumably lie between the North Markvand and Galtskart successions, is unknown at the present stage of the research project. Not until the area further to the N has been mapped, when subsequent correlations may prove possible, can this be determined. Consequently, the place of the North Markvand succession within the regional stratigraphy is unknown. Its age can range from Upper Cambrian to Middle or even Upper Ordovician, although the writer favours a restricted time-span to the group.

<table>
<tr><th>Opdal Region</th><th>Finnmark Region</th><th>Sulitelma Region</th><th>Navervand Region</th><th>Storglomvand Region</th><th>Ørnes Region</th><th>Age</th></tr>
<tr><td></td><td></td><td></td><td></td><td></td><td></td><td>Upper Ordovician</td></tr>
<tr><td></td><td>Tillite-bearing sandstone</td><td>Conglomerate sandstone</td><td></td><td>Vegdal Group</td><td rowspan="2"></td><td>Middle Ordovician</td></tr>
<tr><td></td><td>Varied sandstones
Dolomite-bearing sandstones and shales</td><td>Pieske limestone</td><td>Sokumfjell Marbles</td><td>Sokumfjell Marbles</td><td>Lower Ordovician</td></tr>
<tr><td rowspan="2">Trondheim schists</td><td></td><td></td><td rowspan="2">Pelitic and calcareous pelitic schists</td><td rowspan="2">Holmvand Group</td><td>N.Markvand Succession
Galtskart Succession</td><td>Upper Cambrian</td></tr>
<tr><td></td><td rowspan="2">Juron Quartzite</td><td>Steffodalen Succession</td><td>Lower Cambrian</td></tr>
<tr><td>Sparagmite</td><td>Shales and sandstones</td><td>Psammitic schists</td><td>Psammitic schists</td><td>Basal Succession</td><td>Eo-Cambrian</td></tr>
<tr><td>Basal Gneisses</td><td></td><td>Granitic Basement</td><td>Glomfjord Granite</td><td>Glomfjord Granite</td><td>Basal Granites</td><td>Pre-Cambrian</td></tr>
</table>

Field Characters of Quartz Veins and Segregations

Im most of the rocks of the Ornes region, quartz veins and segregations both parallel to and cross-cutting the lithological banding are present. The percentage of quartz in the meta-sediments is very variable; in places where there is abundant quartz, it may be related to the proximity of a granitic body but elsewhere it is less common and appears to be due to segregation during metamorphism. By the variation in mode of occurrence of the quartz veins, and relationships to the rocks in which they occur, it is possible in a general way to correlate their age of formation with the main periods of deformation.

The generally accepted view as to the formation of quartz veins and rocks is that they are secreted from the country rock during metamorphism, and, depending upon the relationships of the processes of secretion to the regional folding, will depend the form adopted by the veins (e.g. G. Wilson 1953). Wilson, working on the Sutherland rocks, recognises segregation into three varieties of planes of weakness:

(1) along secondary cleavage or foliation

(2) along original bedding

(3) along fissures oblique to these.

Examples of all these varieties can be seen in the Ornes region, and close comparison exists between many of the illustrations he gives.

Certain of the quartz veins follow theregional bedding or schistosity, and are shown in figs 67, 70. Occasional zones of

Fig.65. Semi-pelitic schists overlying the Bjellåtind granite, North Kvittind.

Siliceous mica schist with quartz and tourmaline H 25, H 31

Lenticular pegmatite H 29, H 30

Lenticular pegmatite H 28

Flaggy semi-pelitic schist H 27

Pegmatite H 26

Fig.66. The lowermost 3 metres of the metasediments above the Bjellåtind granite on North Kvittind

extreme sinuosity of the vein systems (as at point A in fig.67 A-B) suggest that the apparently simply-bedded succession may be highly deformed.

Good examples of secretion during the fold periods are shown in fig. 68 A and D. Both syn-F_1 and syn-F_2 quartz veins are represented. In the majority of cases, due to lack of three-dimensional exposure, it cannot be determined whether or not the veins are rodded in 'b' in the manner indicated by Wilson.

Occasional examples show the effects of folding upon pre-existing quartz material (fig.68, C and E) and indicate the fracture and detachment of once continuous veins. This occurs in association with both F_1 and F_2 structures.

Secretion of quartz into oblique fissures is indicated in fig. 69 , and once again the age of the processes of secretion is probably variable. Sometimes, the quartz fills late-stage and undeformed gashes and joints (figs 69A, 71) but elsewhere occurs in deformed tension fractures (fig.69C).

Examples from the Ørnes Region therefore indicate that the secretion of quartz material spans a large time period of the metamorphic history of the rocks. Separate varieties of veins do not appear to be restricted to any one particular rock type or to any one area, but the profuse quartz veins in the immediate contact sediments of the Bjellåtind granite, which are probably derived from the granite itself, are all of presumed F_2 age.

Fig.67. Quartz vein systems in the metasediments

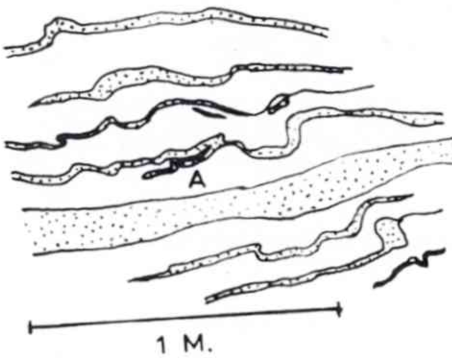
A-B. Quartz veins in amphibolites, West Lysvand. They are generally parallel to the schistosity, but occasionally, as in the regions marked 'A', are transcurrent.

C. Irregular development of quartz in pelitic schists, South Scetertind. There is a general control on the alignment of the veins by the F_1 schistosity.

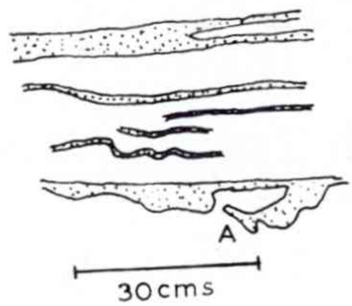
D. An enlargement of part of one of the veins shown in C, illustrating the form of the quartz-schist junction.

E. Fractured quartz veins in pelitic schists, North Galtskart.

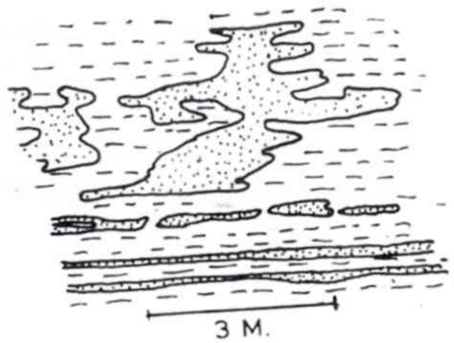
F. An F_1 overfold preserved in the pattern of a fractured quartz vein, South Galtskart



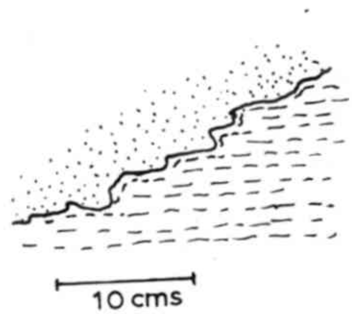
A



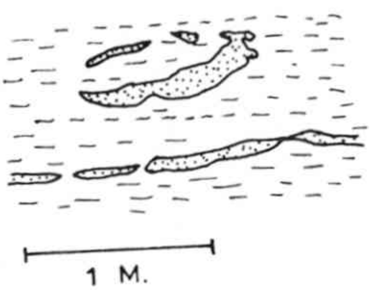
B



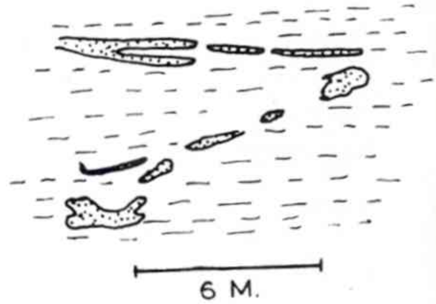
C



D



E



F

Fig.68. <u>Quartz vein systems in pelitic and semi-pelitic schists</u>

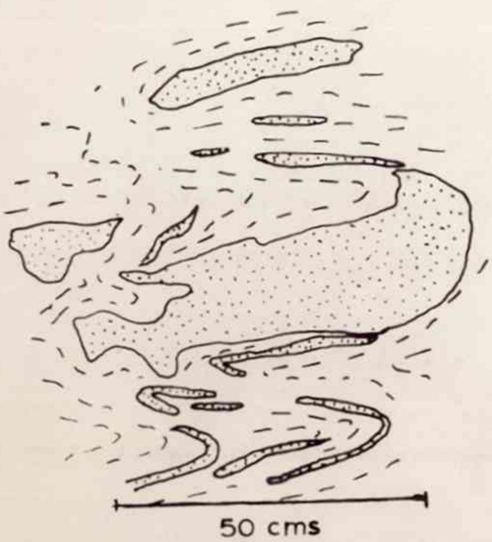
A. Segregated quartz occurring in F_1 isoclinal folds, within pelitic schists, Suppevand

B. Quartz veins in pelitic schists, Bugten. Their alignment is largely controlled by F_1 schistosity

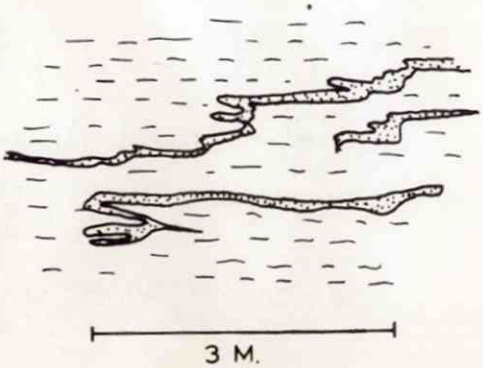
C. Quartz veins in pelitic schist, North Bören, showing F_2 minor folds

D. Segregation of quartz into an F_2 minor fold in pelitic schists, Breitind

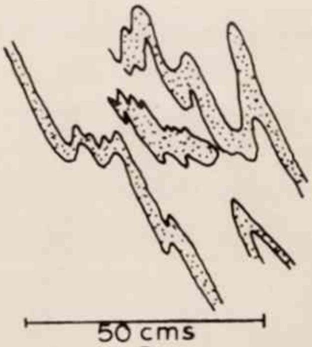
E. Irregularly folded quartz veins in semi-pelitic schist, West Sörfjord



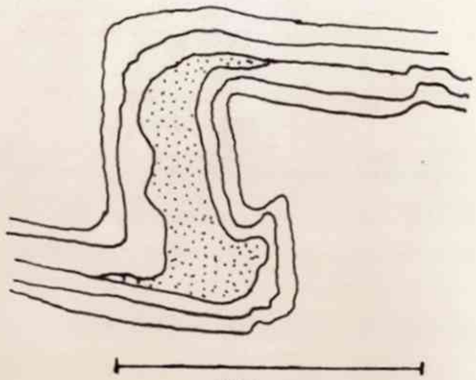
A



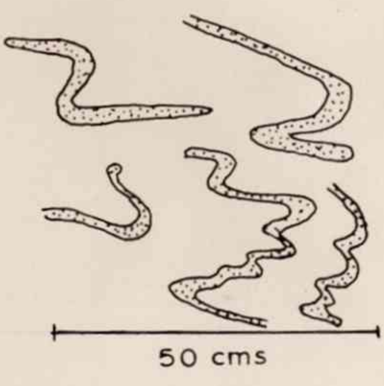
B



C



D



E

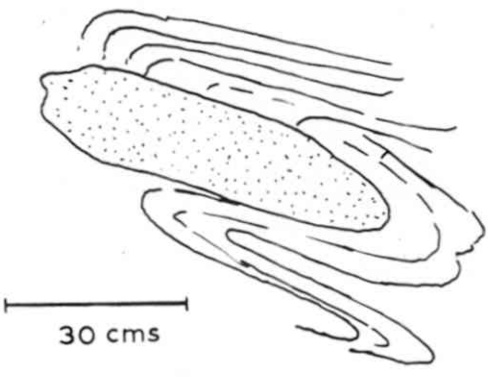
Fig.69.

A. Segregation of quartz into the nose of an isoclinal fold in pelitic schists, Suppevand

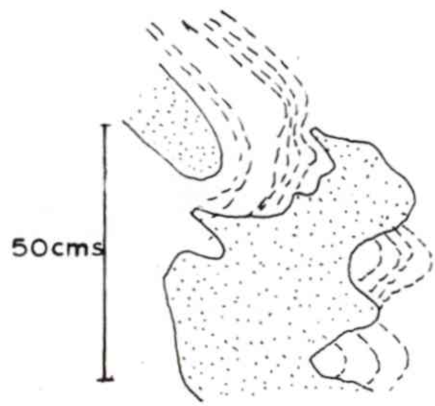
B. Irregular transcurrent quartz patches in pelitic schists, West Sörfjord

C. Quartz filling deformed tension gashes in pelitic schist, Bören (see G. Wilson 1952)

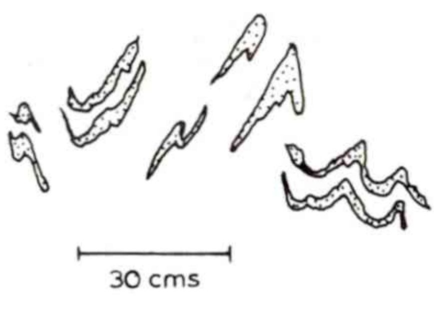
D. An irregular development of quartz in amphibolites, North Skjeggen



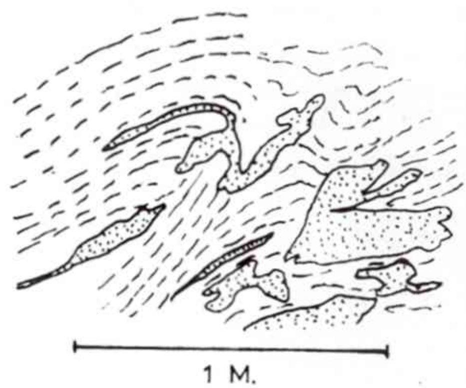
A



B



C



D

Fig.70. Quartz veins in amphibolites, West Lysvand

Fig.71. Lenticular pegmatites and cross-cutting quartz veins in the metasediments immediately above the Bjellåtind granite, North Kvittind

Field Characters of Felspathic Pegmatites

Pegmatites varying from sharply-defined cross-cutting bodies to diffuse irregular lenses are seen spasmodically throughout the Ørnes region. The largest concentration occurs within the basal granites and in the nearby metasediments, while elsewhere they occur only occasionally, and are generally well-differentiated and cross-cutting.

Certain of the pegmatites occupy complexly folded veins, analogous to ptygmatic structures. This term was first used by Sederholm (1926) to describe contorted quartzo-felspathic veins in migmatites or gneisses. He regarded them as originally planar layers that were made mobile by partial fusion, and deformed in an irregular manner by flexional movements. Read (1927), from a study of ptygmatic folding in the Sutherland granite complex, concludes that the veins were never planar, and that they are true igneous injections. Their tortuous form is caused by the resistance of the country rock to planar deformation. G. Wilson (1952) explains the highly folded nature of the veins by likening the host rock to a mass of jelly, and the pegmatite to a mass of putty intruded by migmatic sources. de Sitter (1956) regards the structures as being due to the different way in which the vein and host rock react under the same external stress field. Many of the veins he envisages as forming by segregation from the host rock during migmatisation or granitisation. Such segregation, according to de Sitter, need not necessarily occur along planar directions in the rock, and therefore may not represent the total amount of deformation the rock has undergone.

Forcible injection of externally derived material appears to have occurred in some of the examples (figs 72, 73) and these are closely comparable with illustrations given by Read for the Sutherland veins. None of the examples from the Ørnes region occur within migmatitic rocks, and no exact analogues to the illustrations of G. Wilson have been found. The highly contorted pegmatites of the gneissic region around Ørnes (fig.79) were probably derived from the host rock by secretion during the height of metamorphic processes. The mechanism for their formation is therefore similar to that described by de Sitter. However, the pegmatites are controlled by the planar arrangement of the gneisses and have the appearance of being caused by tectonic forces. They therefore probably reflect the total amount of deformation of the host rock.

In marked distinction to this is the arrangement shown in fig.72 of a complicated vein system, of more than one age, occurring in amphibolites. A well-defined straight pegmatite that slightly transgresses the banding of the amphibolite, is in places cut, but not displaced, by a quartz vein, while elsewhere it interrupts the course of a second quartz vein. Both the quartz veins are highly folded, sometimes ptygmatically, and are occasionally fractured. They have the appearance of being of the same age or possibly later than the felspathic pegmatite, which

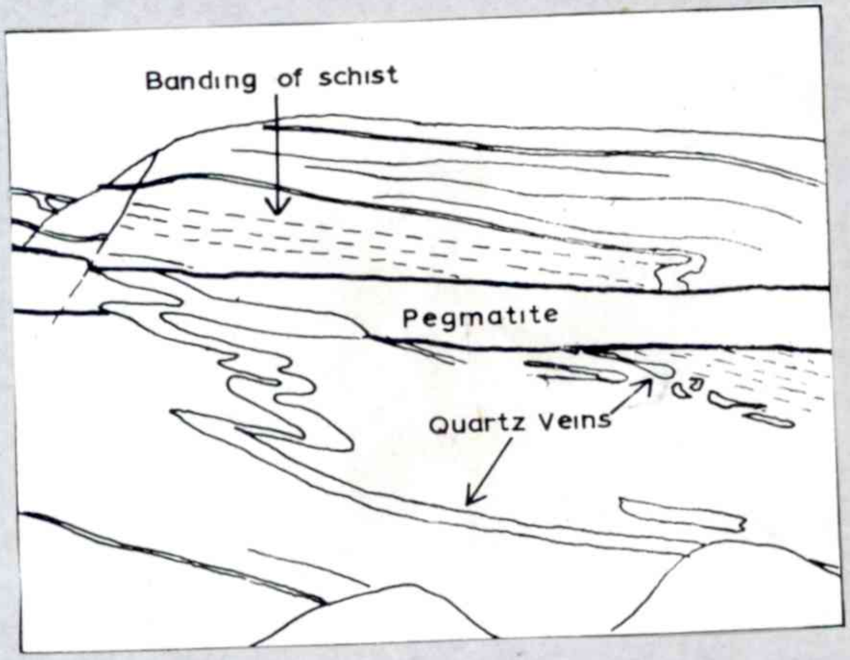


Fig.72. Quartz veins and pegmatite in amphibolites, North Skjeggen

demonstrates that the folds they occupy cannot be tectonic. This suggests forcible injection of material from an external source.

Occasional examples (fig 72 A and D) also indicate forcible injection, by the way in which distinctive marker bands in the country rock are offset. Other instances of straight-sided, cross-cutting pegmatites whose detailed field relationships are obscured by unexposed ground, probably have similar characteristics.

Two remarkably persistent and straight pegmatite dykes cut the rocks of the Steffodalen region. They are each from one to two metres wide, and can be traced over several km. Their orientation is probably controlled by a master joint direction which, as determined from aerial photographs, is most strongly defined in this direction. An origin from the Glomfjord granite is suggested, to which the dykes converge, while the undeformed nature of the pegmatites indicates them to have formed subsequent to the F_3 deformation.

Other less persistent cross-cutting pegmatites occur spasmodically in the Galtskart succession, and again appear to represent material derived from the Bjellåtind granite. In this succession, also, the largest granitic pegmatite from the whole of the Ørnes region is encountered (fig.30A). It is only seen on the south side of Galtskart where it is a band about 15 metres thick, conformable within the bedded succession; the pegmatite is terminated abruptly northwards. The reasons for this termination are not clear, but it is possible that the pegmatite follows the trace of an F_1 slide, and was introduced at a relatively late stage; the coarseness of grain size and lack of mineral cataclosis supports this postulate. Details of the character of the northerly termination of the pegmatite are obscured by unexposed ground.

The isolated example of fig. 73 F shows thin felspathic pegmatites that have undergone deformation, together with the rocks in which they occur. Another example from the same region (fig.73) illustrates lenticular coarse-grained felspathic schlieren that show no internal evidence of shearing. They may have formed by disintegration of pre-existing continuous veins, with subsequent recrystallisation.

Less regular sinuous pegmatites occur occasionally (fig.73 B and C) and generally show an overall elongation parallel or sub-parallel to some well-defined S-surface. They are comparable with some of the ptygmatic folds illustrated by Read.

Further isolated examples show pegmatites that are probably associated with one of the phases of deformation (fig.74). One of these shows net-like veining of highly deformed marbles and calc-schists (fig.74 A) and occurs just above the thick layered granitic pegmatite of South Galtskart. It is suggested that the veining took place subsequent to the deformation of the sediments, and is connected with the development of the underlying layered pegmatite. Fig.74 B shows a pegmatite that occupies the core of an isoclinal fold, with margins everywhere parallel to the

Fig.73. <u>Felspathic pegmatites in the metasediments</u>

A. Cross-cutting pegmatite in calc schists and flaggy pelitic schists, South Scetertind

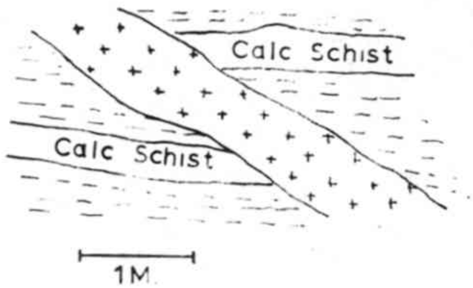
B. Net-veining of pelitic schists by thin pegmatites, Bugten

C. Irregular pegmatites in hornblende schists, South Breitind

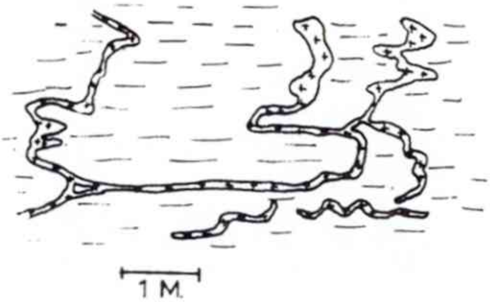
D. Cross-cutting pegmatites in calcareous pelitic schists, West Skjeggen. The pegmatites offset distinctive marker beds in the schists

E. Lenticular microcline-rich pegmatites in massive pelitic schists, Storvik

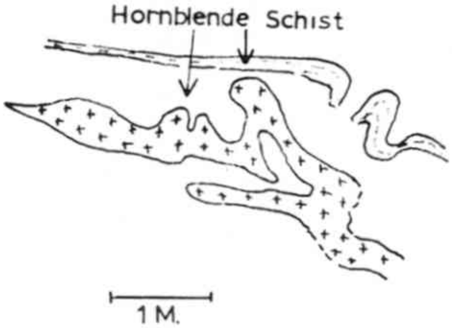
F. Layered pegmatites showing F_2 minor folds in massive pelitic schists, Storvik


Calc Schist
Calc Schist
1M.

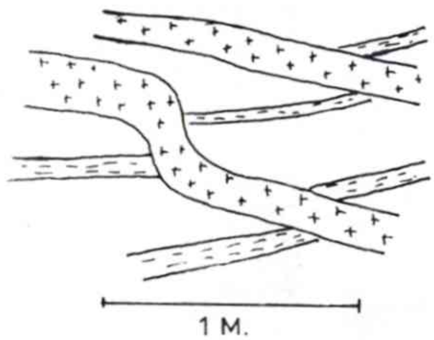
A


1 M.

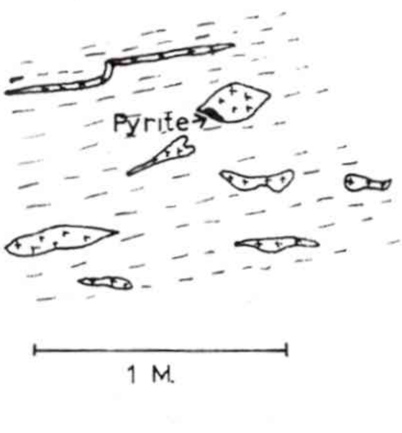
B


Hornblende Schist
1 M.

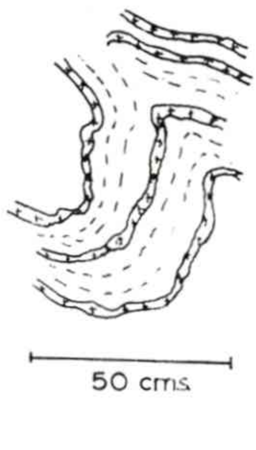
C


1 M.

D


Pyrite
1 M.

E


50 cms.

F

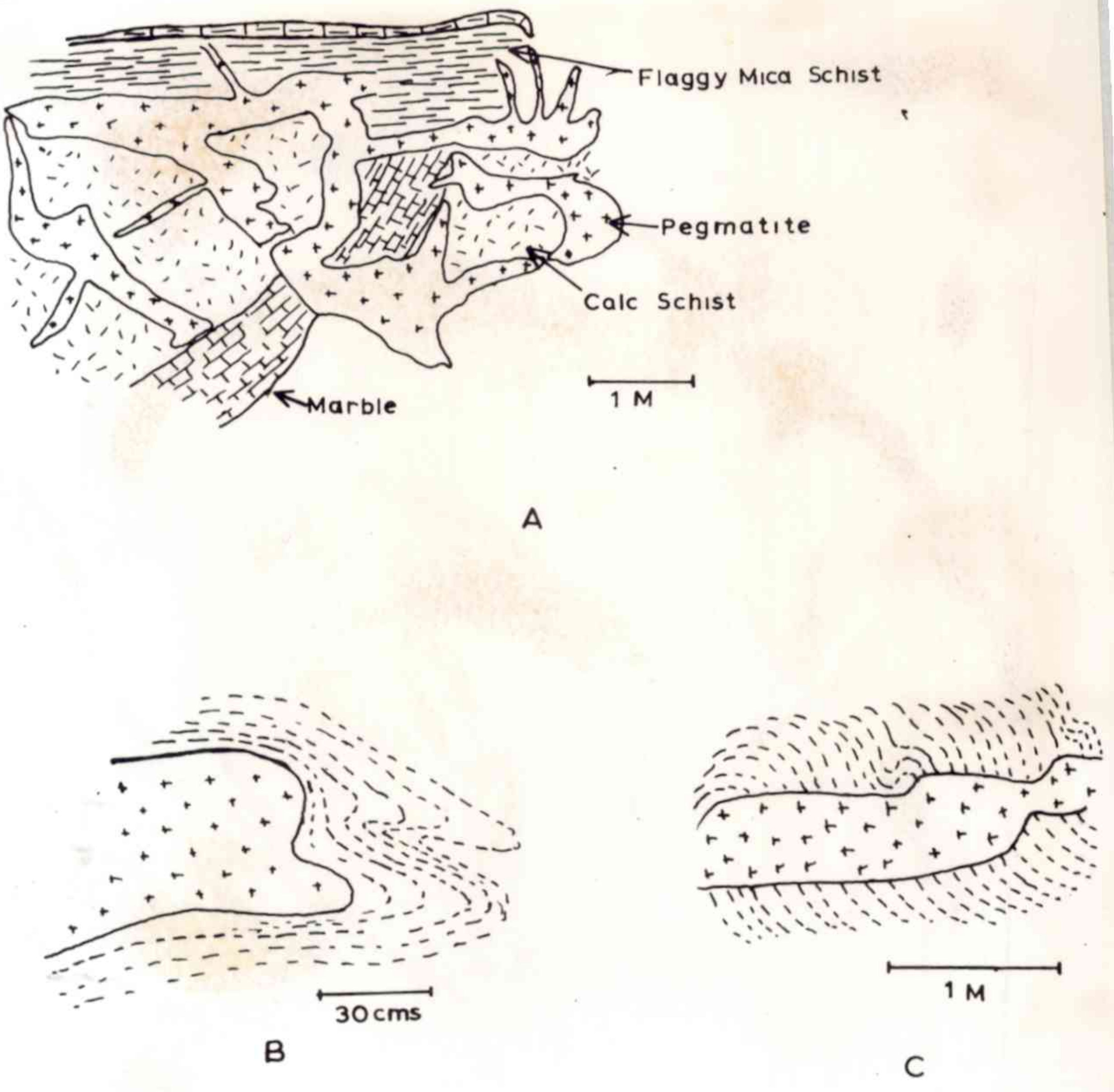


Fig.74 A. Net veining of mica schists, calc schists and marbles by pegmatites, South Galtskart

B. Pegmatite occupying the core of an F_1 isoclinal fold in pelitic schists, Bugten

C. Cross-cutting pegmatite in pelitic schists, Torsvik schistosity of the rocks in which it occurs. Contemporaneous development of the pegmatite with folding is suggested.

An extremely unusual pegmatite complex occurs in the ultrabasic mass of Torsvik (fig.75). Three distinct layered units can be identified, namely (i) a thin, coarse-grained quartz vein, (ii) a dark-green micaceous pegmatite with diffuse coarse-grained elongated felspathic patches arranged normal to the dyke wall, (iii) a coarse-grained quartzo-felspathic pegmatite. The whole is about 6 metres thick, and no marginal facies to the ultrabasic mass are developed. The origin of this composite dyke is unknown, and no connection with the surrounding schists can be seen. A discussion of the petrography of the dyke is given in section 3 .

Diffuse and lenticular pegmatites are concentrated in the contact rocks of the Bjellåtind granite. Their origin is clearly related to the granite, and the lenticular form they possess is probably due to F_2 deformation.

In addition to the irregular and late-stage pegmatites in the Ørnes Gneisses (fig.78) are the layered pegmatites that show isoclinal folds, that have already been described (p. 199).

The diversity of mode of occurrence of the pegmatite bodies indicates various modes and ages of formation. They span a time range from syn-F_1 to post-F_3, and as a general rule, it appears that the early pegmatites were formed by secretion from the host rock, while the later pegmatites were derived from external sources and were forcibly injected.

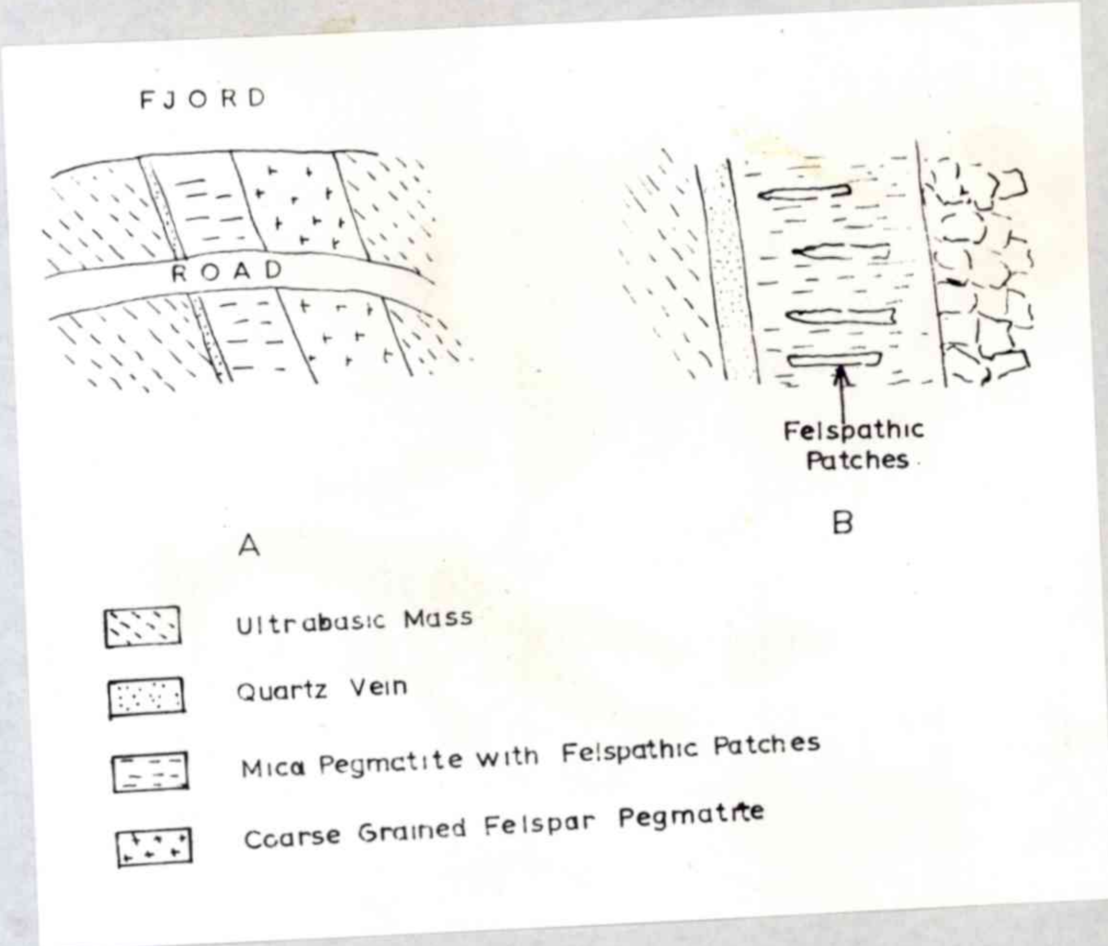


Fig.75 A. Plan of a composite pegmatite dyke in the ultrabasic mass of Torsvik, showing the arrangement of the individual components. The dyke is vertical, and about 3 metres wide

B. An enlarged view of A

Fig.76. An axial plane pegmatite in folded calcareous pelitic schists, North Skjeggen. The fold is of F_2 age

Fig.77. An overfold in quartzites, North Skjeggen. One limb of the fold is completely granitised

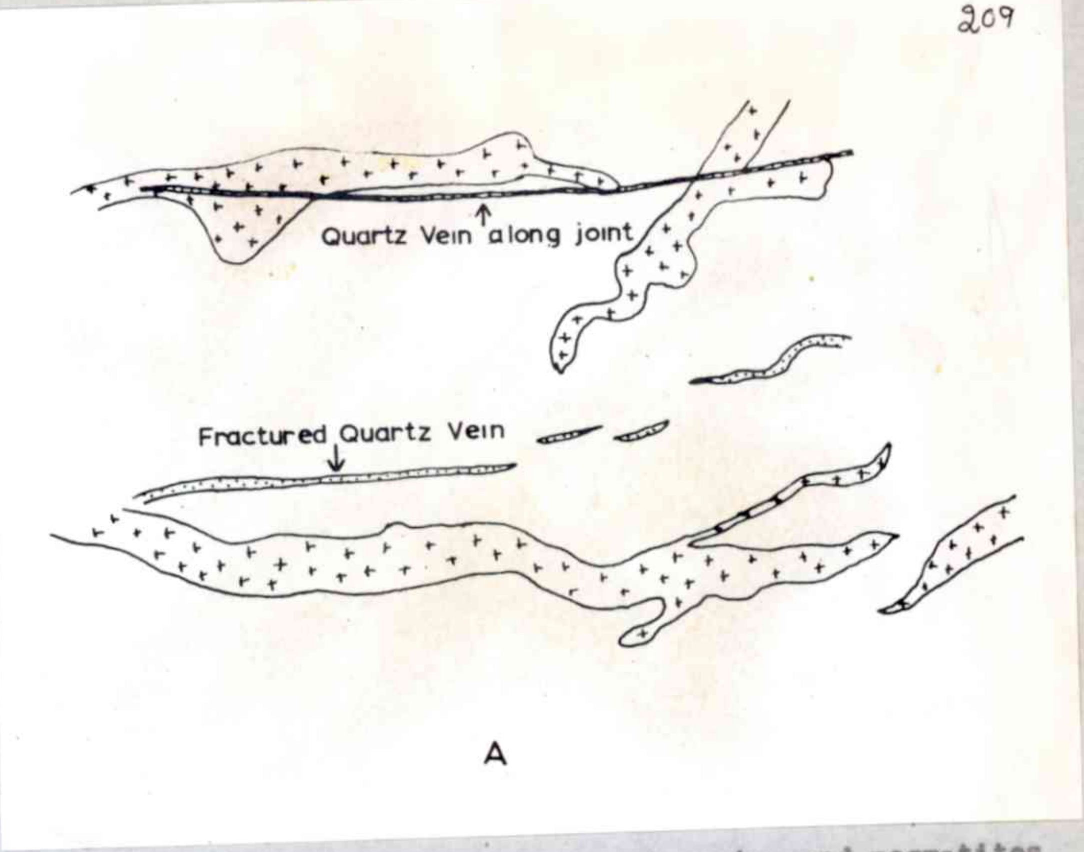


Fig.78.A Cross-cutting and late-stage quartz veins and pegmatites in the gneissic rocks, East Ørnes

Fig.78.B F_1 pegmatites occupying isoclinal folds in gneissic rocks, East Ørnes

Petrographic Descriptions

For purposes of convenience, the descriptions are divided into the stratigraphic groups established in Chapter 2 . Within each group, further subdivision into the various rock types is made.

(1) Basal Granite

Stratigraphic correlation is made between the Glomfjord, Bjellåtind, Skjeggen and Blaatind granites (see p. 161) which are regarded as the oldest rocks seen in the Ørnes region. The Glomfjord granite is being studied by Mr. M.A. Jones, and is not dealt with here. Each of the remaining granites is described separately below.

(a) Bjellåtind Granite

That part of the Bjellåtind granite exposed in the Ørnes region, and examined in some detail by the writer, occurs to the E of Lysvand. Only the uppermost 20 or 30 metres of the mass are exposed in this area, and the deeper parts of the granite have been examined further to the E by Wells. In the E.Lysvand region, the rock is a uniform pink gneissose granite, with thin irregular felspathic schlieren separated by impersistent laminae of biotite and occasional hornblende crystals. The latter give the rock a well-defined foliation. Spasmodic magnetite crystals, up to 2 mm. in diameter, are unevenly distributed, and occasionally are concentrated in coarse-grained lenticular felspathic pegmatites. Other pegmatites, of variable characteristics, are also found within the granite.

Table 1. The Bjellåtind Granite

(1) Mineral Assemblages

	Quartz	Micro-cline	Plagio-clase	Bio-tite	Horn-blende	Sphene	Orth-ite	Magne-ttite	Apa-tite
H 21	X	X	X	X		X	X	X	X
H 22	X	X	X	X		X	X	X	X
H 33	X	X	X	X		X	X	X	
H 45	X	X	X	X	X	X		X	X
H 47	X	X	X	X	X	X	X	X	X
H 421	X	X	X	X		X		X	X
H 423	X	X	X	X	X	X	X	X	X

(2) Plagioclase Compositions

H 21	An 6	(7 determinations)
H 22	An 8	(7 determinations)
H 33	An 11	(5 determinations)
H 45	An 5	(5 determinations)
H 47	An 9	(5 determinations)
H 421	An 10	(3 determinations)

Average from whole of granite An 8

(3) Modal Analyses

	Quartz	Micro-cline	Plagio-clase	Bio-tite	Horn-blende	Sphene	Orth-ite	Magne-tite	Total
H 33	17.7	35.9	39.8	3.8		2.1	.2	.5	100.0
H 47	16.2	29.3	46.2	3.2	3.4	1.0	.1	.1	99.5

(4) Chemical Composition

H 33

	(a) from modal analysis	(b) by rapid analytical methods
SiO_2	68.7	68.72
Al_2O_3	15.6	16.19
Fe_2O_3	.1	1.39
FeO	.5	1.36
MgO	1.1	.19
CaO	1.6	1.17
Na_2O	4.2	4.60
K_2O	6.2	6.20
TiO_2	1.6	.62
P_2O_5	-	.07
MnO	-	.08
Total		100.59

The rock type on the NE face of Galtskart, some 30 metres below the contact metasediments, is a more coarsely crystalline variety than on the E of Lysvand, with a patchy distribution of the biotite crystals. Further varieties doubtless exist, as indicated by the bands of ? mica schist in the granite, seen in the inaccessible north face of Bjellåtind.

The granites consist principally of quartz and felspar, with variable amounts of biotite and occasional hornblende. A varied suite of accessory minerals may be present, including sphene, orthite, magnetite and apatite. Modal analyses for two of the granites are shown in Table I, indicating that the total felspar comprises about 75% of the rocks, with plagioclase dominant; quartz makes up about 20%, and the remaining 5% is biotite, or biotite and hornblende. In other thin sections, for which modal analyses are not available, microcline exceeds plagioclase. Plagioclase determinations* of six of the rocks (Table I) show on average a composition of An 8. Thus, the rocks have a mineral content equivalent to alkali granites, while both sodic and potassic types are represented. The chemical analysis of part of the Bjellåtind granite is shown in Table I. It corresponds most closely to an adamellite, as is indicated in a modal analysis of the same rock. The composition of the plagioclase is sodic andesine. This analysis may not be representative of the remainder of the Bjellåtind granite.

* A review of the various methods used, and their suitability, appears in Section 3L316.

The appearance of a typical granite in thin section is shown in Fig.79. A green-brown variety of biotite occurs in small ragged unbent crystals, showing strong preferred orientation and often concentrated into irregular layers. Associated with the biotite are small crystals of sphene, which are often concentrated into clots. The amphibole is a highly-coloured variety with extreme pleochroism from deep olive-green to light grass-green. Due to the marked body-colour, detailed optical measurements cannot be made, but it is possible that the amphibole is a soda-rich hornblende. In mode of occurrence, the amphibole varies from small ragged fragments, in association with biotite and sphene, to large discrete crystals. Some degree of corrosion is always visible, and inclusions of biotite, iron ore and quartz are common.

Felspars make up 75% of the rock, and plagioclase is considerably more common than microcline. Some of the felspar interrelations are discussed on p. 215 Both felspars form irregular crystals of varied size, but the microcline tend to be the larger; inclusions of dark minerals and quartz are common in plagioclase, and of quartz and plagioclase in microcline. The plagioclase is finely twinned on the albite law, and often shows strain and fracture of the twin lamellae. Separate patches of plagioclase in optic continuity are sometimes seen, surrounded by microcline. The latter are also found in larger porphyroblastic crystals. Occasionally, the plagioclase shows progressive zoning, a phenomenon which appears to be due to local reaction between the plagioclase and

Fig.79. Typical texture of the dark minerals in a hornblende-bearing variety of the Bjellåtind granite. Small sphene crystals are associated with the hornblende, and are also found in thin irregular lenticles. x 40.

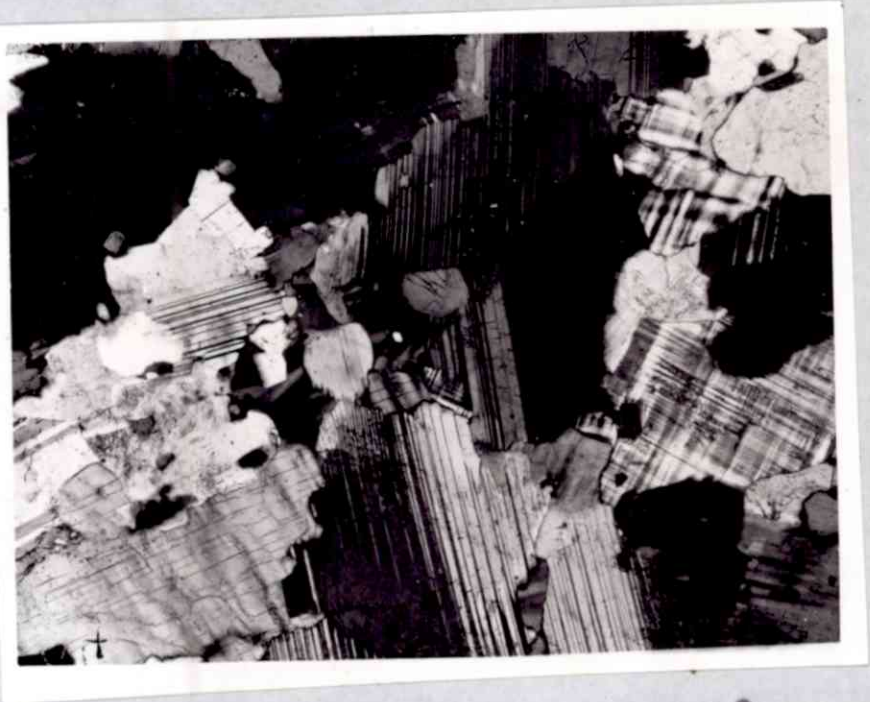


Fig.80. Texture of the felspathic part of the Bjellåtind granite. The large dark crystal at the top left is microcline and when in conjunction with a slightly fractured plagioclase crystal, towards the right, the latter shows the development of a sodic rim. Variability of microcline twinning is well displayed. x 40.

neighbouring amphibole. Both felspars show faint dust-like inclusions, probably of iron-ore; this is more common in microcline where it is probably related to a crystallographic direction.

Two generations of quartz are present, the earlier as inclusions in the felspars and in small granular clots, and the later as larger crystals in irregular elongated patches arranged parallel to the foliation. Both varieties show some degree of strain.

Thin sections of other specimens of the Bjellåtind granite show similar characteristics. A few of the rocks contain large sub-hedral iron-ore crystals, up to 3 mm. in diameter. Occasionally, these are rimmed by small amounts of sphene, indicating that the ore may contain TiO_2. Sphene is sometimes also present in much larger crystals, up to 3 mm. across. Certain of the rocks contain small grains of yellow orthite, normally associated with the dark minerals, and sometimes causing zones of discolouration with adjacent biotite crystals. Minute orthite grains are enclosed in some of the micas.

Characteristics and Interrelationships of the Felspars

The potash felspar present is microcline, but the degree of twinning developed is very variable; some crystals show very fine cross-hatching, while others show only a rudimentary or patchy development of twinning. In the latter, the only visible lamellae are twinned on the albite law. Both kinds of twinning may be present in a single rock, and there appears to be no correlation between size or mode of occurrence of the microcline crystals, and the twin lawsdeveloped. Absence of twinning or the development of a single set of twin lamellae in microcline has been regarded as indicative of crystallisation at low temperatures, with an absence of soda in the microcline atomic structure (Eskola 1951, Rutland 1958). No analyses of the distinct types of microcline are available for the rocks of the Ørnes region, but petrographic evidence suggests that the variability of twinning is not controlled by composition (see p. 219).

Examination of the plagioclase crystals from a series of specimens of the granite shows that crystals finely twinned on the albite law are dominant (e.g. fig.80, 81). This contrasts markedly with those from the metasediments, where combined albite-pericline twins are common. Deformation of the plagioclase crystals is indicated by fracturing and bending of the twin lamellae. Sometimes separate fragments of a fractured plagioclase crystal that were clearly at one time in optical continuity are surrounded by poikiloblastic and undeformed microcline, suggesting the latter to be the later felspar. The converse texture is never seen, indicating the unlikelihood of the phenomenon being a three-dimensional effect of simple intergrowth.

Occasionally clear albitic rims occur on those margins of plagioclase crystals that are in contact with microcline (fig.80). Here, it is evident that the rim-albite is in equilibrium with the microcline, and the sodic rims are probably caused by ex-solution of albite from the structure of the microcline and subsequent

Fig.81. Irregular development of myrmekite in part of the Bjellåtind granite. That in the twinned plagioclase is probably true myrmekite, while that in the black untwinned crystal at the centre of the photograph may be para-myrmekite. x 40.

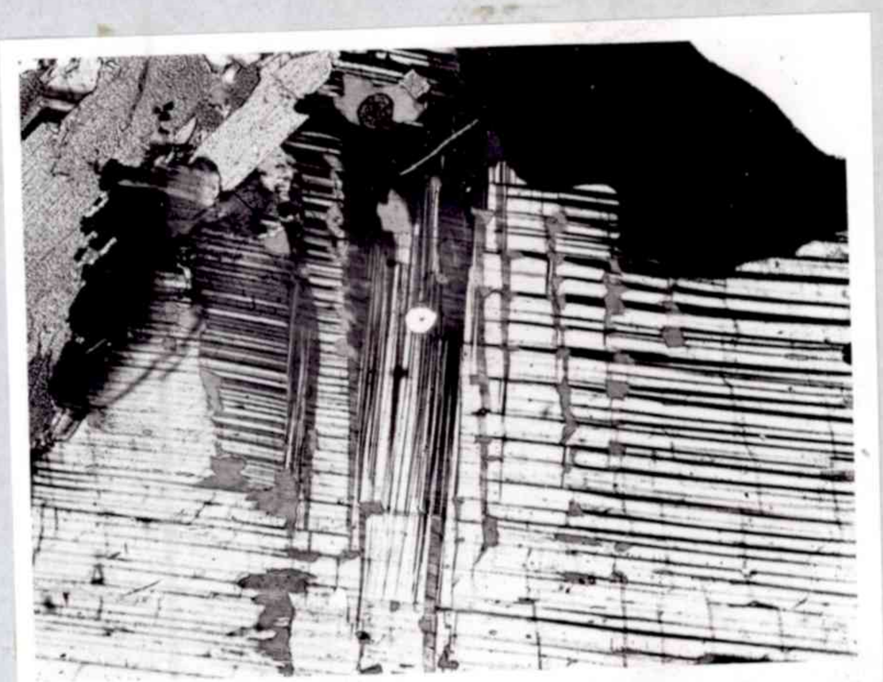


Fig.82. Combined albite-pericline twin in plagioclase from a migmatite, Blaatind. The twin lamellae are impersistent and spindle-shaped; partial fracture of the crystal is visible on the left hand side of the photograph. x 30.

crystallisation upon the pre-existing plagioclase (Tuttle 1952). Those microcline crystals that have yielded soda to adjacent plagioclase should contain less soda than those which have not. It may be expected, therefore, that differences in twinning of microcline occur, due to variability of soda content. Such variation is observed, but appears to be independent of the distribution of sodic rims to plagioclase crystals; this may indicate the inability of the microcline to recrystallise subsequent to ex-solution of the soda. Chemical analyses may yield an answer to the problem.

Emmons (1953) describes similar albitic rims to plagioclase, without the associated microcline. He regards them as being due to internal redistribution of albite, consequent upon the twinning of the plagioclase.

Vermicular quartz in plagioclase on the junction with microcline is occasionally seen (fig.81, 83) and indicates instability between the two felspars. As in the rocks of the Sokumfjell district, described by Rutland (1958), there appear to be two types of myrmekite present (a) in plagioclase that is earlier than the microcline, and has a similar appearance to the bulk of the plagioclase of the rock. This is termed para-myrmekite.

(b) in plagioclase that is later and lobes into the microcline, termed true myrmekite.

(a) Para-myrmekite

Edelman (1949) has described similar structures, and regards the quartz as released from the plagioclase by basification

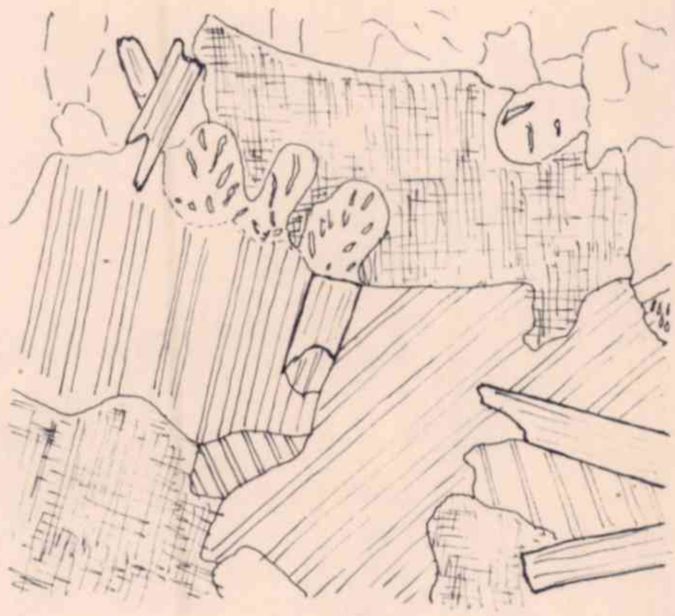


Fig.83. Development of true myrmekite lobing into microcline, granite gneiss Blaatind.

consequent upon replacement by microcline. Lime is released during replacement, as the microcline structure can more readily accept soda, and this lime causes the basification. Rutland (1958, p.262) outlines objections to the postulates of Edelman, principally on the evidence that the plagioclase containing para-myrmekite is, in the Sokumfjell rocks, of equivalent composition to the remainder of the plagioclase, and not more basic as is required. He concludes that if the microcline originally contained more soda than at present, later ex-solution may have reverted the plagioclase back to its original composition. In the rocks of the Ørnes region, there is no evidence that this has occurred, as the demonstrable ex-solution of soda from microcline produces thin albitic rims to the plagioclase crystals, not wholesale albitisation. Moreover, it appears fortuitous that the plagioclase should revert to its exact original composition, and that the vermicular quartz should not be utilised in the reversion. Detailed determinations of the composition of para-myrmekite plagioclase crystals are not available for these granites, and it is not known if their composition is different from the surrounding crystals. The postulates of Edelman are therefore accepted by the writer in the case of the Bjellåtind granite.

(b) True Myrmekite

These compare closely with the true myrmekites of Sederholm (1916) which he regarded as being caused by the replacement of potash felspar by plagioclase with the consequent ex-solution of silica. The plagioclase in which this type of myrmekite is seen in the Ørnes rocks is untwinned, and lobes into microcline crystals. Two possibilities exist for the derivation of this late-stage plagioclase: (1) from external sources, as a metasomatic product subsequent to the formation of microcline,

(2) from redistribution of soda, that was originally present, in more sodic microcline. Evidence for widespread late-stage metasomatism is not seen in these rocks, but secondary redistribution of soda from microcline, with the development of sodic rims to plagioclase crystals is commonly observed. The writer therefore favours the second interpretation, and regards the development of true myrmekite as a localised concentration of soda exsolved from microcline.

From these general deductions as to the course of crystallisation of the felspars, several conclusions may be drawn regarding the petrogenesis of the granites:

(1) Polysynthetic twinning of plagioclase on the albite law is common, and is regarded by Emmons (1953) as a consequence of strain on originally untwinned or zoned felspars. Later bending and fracturing of the crystals is clearly indicated in most of the rocks and is probably to be correlated with the F_3 phase of deformation.

(2) The crystallisation of microcline is later than that of plagioclase, although reasons are given elsewhere (p. 231) for regarding the majority of the potash felspar, at least, as having been present in the original rock.

(3) The formation of sodic rims to some of the pre-existing plagioclase crystals depended upon inter-action between the two felspars, and is probably of a later age than the microcline. The occasional patches of true myrmekite are probably of a similar age.

(b) Skjeggen Granite

The granite is more variable than the upper parts of the Bjellåtind granite. Again the rocks are predominantly pink in colour, but many do not show the pronounced felspathic banding characteristic of the Bjellåtind mass. In addition, the degree of layering of the dark minerals is less pronounced. On the N of Bolden banded pink and grey granites are prominent, with occasional lenticular masses of biotite schist. A similar arrangement is observed on the outer margins of the Glomfjord Granite (M.A.Jones, personal communication).

Mineral assemblages are similar to those from the Bjellåtind granite, except for the occurrence of muscovite in two of the rocks (H 431, H 435), the unique occurrence of almandine in H 435, and the total absence of hornblende. Modal analyses of two of the rocks are shown in Table 2, and indicate extreme variations in the percentages of all the major minerals. One of them (H 435) corresponds approximately to a potassic leucogranite, except for its rather basic plagioclase (An 12) while the other lies closer to a quartz monzonite except for the low An content of the plagioclase (10%). Whether or not this is representative of the range of

Table 2. The Skjeggen Granite

(1) Mineral Assemblages

	Quartz	Microcline	Plagioclase	Biotite	Chlorite	Muscovite	Sphene	Orthite	Magnetite	Apatite	Garnet	Tourmaline
H 276	X	X	X	X			X	X		X		
H 286	X	X	X	X	X		X		X	X		X
H 431	X	X	X	X	X	X	X		X			
H 435	X	X	X	X		X		X	X		X	

H 434 (Biotite schist) Quartz, Plagioclase, Biotite, Chlorite, Hornblende, Sphene, Magnetite, Apatite, Epidote.

(2) Plagioclase Compositions

H 276 An 10 (from 3 determinations, only approximate)
H 435 An 12 (from 2 determinations, only approximate)

(3) Modal Analyses

	Quartz	Microcline	Plagioclase	Biotite	Muscovite	Orthite	Apatite	Garnet	Magnetite	Total
H 276	7.4	33.8	47.0	11.2		.6				100.0
H 435	29.7	49.8	18.3	.8	.8			.6		100.0

Chemical Analysis (by rapid analytical methods)

H 433		
	SiO_2	75.32
	Al_2O_3	12.72
	Fe_2O_3	0.48
	FeO	0.84
	MgO	0.28
	CaO	0.48
	Na_2O	3.54
	K_2O	5.75
	TiO_2	0.25
	P_2O_5	0.02
	MnO	0.03
	Total	99.69

variation of the whole granite is unknown.

Certain petrographic differences from the Bjellåtind granite can be shown to exist, particularly in the non quartzo-felspathic fraction of the rocks. Biotite crystals are generally small and ragged, green in colour, and often poorly oriented; partial chloritisation is common. Other dark minerals are rare, and consist mainly of small sphene and magnetite crystals. The latter are occasionally much larger, up to 2 or 3 mm. in diameter, as in the Bjellåtind granite.

The textures and relationships of the quartzo-felspathic part of the rocks are generally similar to that seen in the Bjellåtind mass. Occasionally, patches of remarkably coarse-grained and undeformed plagioclase crystals occur. Others of the rocks are more finely granular than any seen in the Bjellåtind granite, even though the mineral inter-relationships are still the same. A small amount of muscovite is sometimes present, generally in small crystals and probably developed from the felspars. This is a noteworthy difference from the Bjellåtind granite.

A single specimen of a biotite schist from the N of Bolden shows very distinctive features (fig.83). Biotite and hornblende in roughly equal proportions constitute about 50% of the rock, the former being greenish-brown and sometimes slightly bent, while the latter is highly pleochroic from light grass-green to deep blue-green. Both form large and ragged crystals. A large part of the remainder of the rock is composed of variably twinned and

Fig.84. Biotite schist from within the Skjeggen Granite at Bolden. The biotite crystals are slightly bent, while the plagioclase shows zoning dependent possibly upon interaction with hornblende. x 40.

zoned plagioclase of later crystallisation than the hornblende and biotite. Zoning is more pronounced when the plagioclase is adjacent to hornblende, and appears to be due to inter-action between the two minerals. A variation of 6% An was recorded from one crystal, with the outside more calcic than the inside, suggesting basification of the plagioclase by hornblende. A small amount of quartz is present, while accessory minerals include sphene, epidote, magnetite and apatite, all as small crystals. Their early origin is indicated by their occurrence as inclusions in biotite and hornblende. Occasional crystals of orthite, surrounded by pistacite, are seen.

The mineral assemblage is indicative of an amphibolite, within the staurolite-kyanite sub-facies (cf. Turner and Verhoogen 1951). It is impossible to be certain whether the schist is of igneous or sedimentary origin, as the mineral assemblage is not diagnostic. The absence of microcline, which is abundant in the surrounding granite, indicates low K_2O, which is suggestive of an igneous amphibolite. In similar rocks from the Glomfjord granite, equivalent assemblages are found, except for the rare occurrence of epidote, and the presence of occasional grains of microcline (M.A. Jones, personal communication).

(c) Blaatind Granites

The specimens of the Blaatind granites examined are similar to the other presumed basal granites. Those that occur as sheets within the metasedimentary cover in E. Blaatind are therefore regarded as representing parts of the basal granites, detatched in F_1 times.

Table 3. The Blaatind Granites

Mineral Assemblages

	Quartz	Plagio-clase	Micro-cline	Bio-tite	Mus-co-vite	Horn-blende	Mag-ne-tite	Apa-tite	Sphene	Epi-dote	Chlor-ite
H 187	X	X	X	X	X			X			
H 194	X	X	X	X		X	X	X	X	X	
H 196	X	X	X	X					X	X	
H 366	X	X	X	X	X		X		X		
H 426	X	X	X	X	X		X		X		X
H 427	X	X	X	X				X	X		

Schist within granite

	Quartz	Plagio-clase	Micro-cline	Bio-tite	Mus-co-vite	Horn-blende	Mag-ne-tite	Apa-tite	Sphene	Epi-dote	Chlor-ite
H 425	X	X		X			X	X	X		

Plagioclase Composition

H 187	An 13	from 4 determinations
H 194	An 14	from 3 determinations
H 196	An 10	from 4 determinations
H 366	An 14	from 3 determinations

Modal Analysis

	Quartz	Micro-cline	Plagio-clase	Bio-tite	Musco-vite	Apa-tite	Magne-tite	Total
H 366	41.6	31.1	21.2	5.6	.3	.1	.1	100.0

In general, the minerals present and their inter-relationships are similar to those in the other basal granites. A migmatitic variety is seen in the easternmost sheet granite, and contains a high percentage of primary muscovite. A coarse-grained variety of the middle sheet granite shows interesting felspar relations (fig. 82). The plagioclase crystals often possess discontinuous and spindle-shaped twin lamellae. Small inclusions of microcline appear to be of later development, since the microcline may seal the fractures in broken plagioclase crystals. The latter are frequently bent and fractured.

Mineral assemblages in the basal granites

Stable assemblages of quartz, plagioclase, microcline, biotite and occasional hornblende, or quartz, plagioclase, microcline, biotite and occasional muscovite are indicative of the staurolite-kyanite sub-facies of the amphibolite facies (cf. Turner and Verhoogen 1951). The biotite schists within the granites can be assigned to a similar sub-facies. No relicts of a higher grade of metamorphism have been found in either granites or metasediments, and it is probable that this is the highest grade of regional metamorphism that the rocks of the area attained. In the case of the basal granites, the peak of regional metamorphism was reached during, or just after, the F_2 phase of deformation.

Later diaphthoresis is occasionally seen in the Skjeggen granite, where there is partial chloritisation of biotite; this is never present in the Bjellåtind granite. Slight clouding of the felspars, particularly the plagioclase, by secondary mica, occurs in all the basal granites. Both these effects are slight and localised, and do not affect the main fabric of the rocks.

During the height of metamorphism, the granites occupied their present tectonic positions, which may account for some of the differences between them. Thus, the lack of development of microcline in the contact metasediments of the Skjeggen granite compared with those above in the Bjellåtind granite, may be due to the differences in thickness between the two after F_1 deformation.

Metamorphic History and Age of the basal granites

Two felspars granites to which the granites belong are grouped with the low-temperature sub-solvus variety of Tuttle and Bowen (1958). The felspars are normally regarded as having formed by unmixing from a homogeneous high-temperature phase, a process that has been shown, experimentally, to be greatly accelerated by the presence of water vapour. In these conditions micas are stable and amphiboles unstable - the latter are only associated with hypersolvus granites. The marked water vapour content of the granite, associated with unmixing of the felspars, should cause extensive granitisation of the surrounding metasediments. The immediate contact rocks do contain a higher percentage of microcline than those some distance from the granite, but the contact is always sharp (figs 28, 65) and the overlying rocks are still recognisably sedimentary even though they may approach a granite in composition. This lack of granitisation may suggest that the granites possessed their present mineralogy before subsequent sedimentation (cf. Nicholson 1960).

It is suggested, from the size and lack of extensive cataclasis of the constituent minerals of the basal granites, that the crystallisation is syn- or post-F_2. Structural evidence indicates the probability of the granites and schists having been in contact during or before the F_1 deformation (p. 21). Thus, the lack of extensive granitisation of the metasediments is either indicative of the previous existence of a two-felspar granite in F_1 times, or the formation of the felspars by a different mechanism from that indicated by Tuttle and Bowen. The presence of hornblende, a mineral theoretically only found in hypersolvus granites, suggests that the granites of this region did not form in the same manner as those described by Tuttle and Bowen.

Other lines of evidence may be used to suggest that the granites are fundamentally different from the schists, and owe their present appearance to original compositional characteristics, not to extensive metasomatism:-

(1) Comparison of the compositions of plagioclase in schists and granites indicates a persistent difference (see also p. 241). Those in the granites are usually albitic, with between 6-10% An. Those in the schists are very variable, and dependent on the chemical composition of the rock. Even in the quartzo-felspathic schists, however, with a mineralogy similar to some of the granites, the plagioclase always has a composition of around An 15 to An 20.

This difference indicates the probability that the plagioclase is original in both granites and schists, and has not been introduced metasomatically.

(2) In the metasediments, the plagioclase is evenly distributed and granular. It has every appearance of having developed by the normal processes of metamorphism.

(3) Microcline in both granites and schists is later than plagioclase, and in the schists occurs in more or less well-defined layers, that often appear to owe their existence to metasomatic introduction from external sources. As microcline shows a general increase in the metasediments towards the granite, it appears that the latter is the source of at least some of the potash felspar. The association of abundant microcline with albitic plagioclase in the basal granites indicates the probability of the former having been in existence before the phase of crystallisation now represented. Thus, the microcline probably recrystallised from pre-existing material, and localised migration during this phase may have produced the restricted microclinisation of the metasediments.

(4) The persistence of bands of biotite schist and amphibolite within the basal granite, with little or no microcline, indicates either that there has been little migration of potash felspar, or that the rocks are resistors to granitisation (e.g. Reynolds 1946, Pitcher 1952).

All the evidence points to the conclusion that the basal granites possessed a composition closely similar to that now seen, from the earliest times onwards. Except for localised metasomatic introduction of microcline into the contact metasediments, there has been little migration of felspathic material. These conclusions, together with the stratigraphic and structural evidence (p. 161), indicate the probability that the granites are at the base of the succession, and are logically regarded as having been in existence before subsequent sedimentation. They are tentatively assigned to the Pre-Cambrian basement, although their history prior to formation of the overlying sediments, is now entirely masked by later metamorphism.

The highly banded nature of parts of the basal granites may be due to one of three factors:-

(1) Original sedimentary differences that have survived a pre-Cambrian granitisation.

(2) Flow banding of a pre-Cambrian igneous granite.

(3) Differentiation of an originally homogeneous granite during metamorphism.

At the present stage of research it is impossible to be certain which of these explanations is the correct one. Comprehensive chemical analyses of all the varieties of granites may help in the problem. However, the banding is remarkably persistent over large outcrops, and is exactly analogous to the sedimentary banding of some of the siliceous schists. Pending future research, therefore, it is suggested that the banding is due to the first of the factors stated above.

(2) Schists adjoining the basal Bjellåtind granite (Basal Succession)

Descriptions of the successions shown in fig.59 are included to indicate the petrographic variation of the sediments in immediate contact with the basal Bjellåtind granite.

(a) Details of the lowermost three metres of the North Kvittind succession are shown in fig.66, and it consists of alternating foliated semi-pelitic schists and thin lenticular and layered tourmaline-bearing pegmatites. The schists are rusty weathering, with a planar schistosity devoid of puckering and undisturbed by projecting porphyroblasts. Thin laminae of microcline are common. These rocks are closely comparable with the contact rocks of the Glomfjord granite (M.A. Jones, personal communication). In addition to the three prominent layers of tourmaline-bearing pegmatites indicated in fig.66, the schists above H 29 have a common development of thin felspathic tourmaline-bearing schlieren. It is noteworthy that these are absent in the 2 metres of metasediments below this.

The mineralogy of the schists is similar to the underlying granite except for the spasmodic development of muscovite. However, a modal analysis of the lowest of the schists (H 27) indicates significant differences from the basal granite (Table 4). The micas are arranged in laminae, with muscovite subordinate to biotite. In contrast to the common development of green-brown biotites in the basal granite, those from the schists are always brown. Granular quartz and felspar constitute the remainder of the rocks, the

Table 4. Basal Succession

(a) North Kvittind

Mineral Assemblages

(1) schists

	Quartz	Micro-cline	Plagio-clase	Biotite	Musco-vite	Magne-tite	Tourma-line	Apatite
H 27	X	X	X	X		X	X	
H 29	X	X	X	X	X	X		
H 31	X	X	X	X	X			
H 457	X	X	X	X	X	X	X	X
H 459	X	X	X	X	X	X		

(2) pegmatites

	Quartz	Micro-cline	Plagio-clase	Biotite	Musco-vite	Magne-tite	Tourma-line	Apatite
H 26	X	X	X	X	X			X
H 28	X	X	X	X	X			
H 29	X	X	X	X	X	X	X	X
H 30	X	X	X		X	X	X	
H 31	X	X	X			X	X	

Plagioclase Compositions

(1) schists

H 27	An 24	from 5 determinations
H 29	An 26	from 3 determinations
H 31	An 30	from 4 determinations
H 457	An 26	from 3 determinations

(2) pegmatites

H 26	An 5	from 6 determinations
H 28	An 17	from 6 determinations
H 29	An 3	from 4 determinations
H 31	An 23	from 3 determinations (approx.)

Modal Analysis

	Quartz	Microcline	Plagioclase	Biotite	Magnetite	Total
H 27	48.2	24.0	16.6	10.6	0.6	100.0

(b) North Galtskart

Mineral Assemblages

(1) Quartzo-felspathic schists

	Quartz	Plagioclase	Microcline	Biotite	Muscovite	Chlorite	Magnetite	Tourmaline	Apatite	Sphene	Garnet
H 401	X	X	X	X	X		X	X	X		
H 404	X	X	X	X	X	X			X		
H 405	X	X		X	X		X	X			
H 413	X	X	X	X	X	Epidote			X		
H 414	X	X	X	X	X		X		X		
H 415	X		X	X	X		X		X		
H 416	X	X	X	X			X	X	X		X
H 417	X	X	X	X			X		X		
H 419	X	X	X	X			X	X	X	X	
(2) Pegmatite											
H 403		X	X				X				
(3) Marble											
H 400	X	X	X	X	X	Chlorite			X	X	Calcite
(4) Amphibolite											
H 402	X	X	Hornblende	X	Diopside	X	Epidote				X

Plagioclase Compositions

(1) Quartzo-felspathic schists

H 401 An 27 from 3 determinations

H 404 An 5 (inside of crystal)
An 20 (outside of crystal)

H 405 An 0 from 3 determinations

H 413 An 16 from 3 determinations

H 414 An 12 from 3 determinations

H 415 An 32 from 4 determinations

H 417 An 16 from 2 determinations

H 419 An 13 from 4 determinations

(2) Marble

H 400 An 33 from 3 determinations

(3) Amphibolite

H 402 An 34 from 3 determinations
An 34 (outside of crystal)
An 24 (inside of crystal)

(c) <u>South Galtskart</u>

<u>Mineral Assemblages</u>

	Quartz	Micro-cline	Pla-gio-clase	Bio-tite	Chlor-ite	Musco-vite	Garnet	Apa-tite	Sphene	Magne-tite
H 452	X	X	X	X		X		X	X	X
H 453	X		X	X		X		X		X
H 454	X		X	X		X		X		X
H 455	X		X	X	X	X	X			

<u>Plagioclase Composition</u>

H 452 An 20 from 2 determinations

An 17 (inside of crystal)

An 21 (outside of crystal)

H 454 An 36 from 3 determinations

An 31 (inside of crystal)

An 36 (outside of crystal)

latter also occurring as larger crystals in irregular laminae 2-3 mm. wide. Occasional lenticular quartz segregations, composed of large crystals, are also present.

Felspar relations are similar to those in the underlying granite. Fractured plagioclases commonly show combined albite-pericline twins, sometimes patchy in development. Their average composition of An 28 (Table 4) is considerably more basic than the majority of the associated pegmatite plagioclases and all of those from the underlying granite. Microcline, which often occurs in thin laminae, is both finely and coarsely twinned, and is less deformed and fractured than plagioclase. Interaction between the felspars with the development of sodic rims to plagioclase when in contact with microcline is clearly displayed.

The three pegmatite layers shown in fig.66 are each distinctive.

(i) H 26, a grey granitic and foliated pegmatite similar in appearance to the basal granite, is probably situated only a few cms above the latter. Granular patches of quartz and felspar are separated by porphyroblasts of variable size of microcline, plagioclase and quartz. Small muscovite crystals, probably formed from the felspars, are widespread. Microcline shows a patchy development of twinning and is more abundant than plagioclase; the latter is often fractured. Secondary alteration of the felspars is much more widespread than either the underlying granite or the overlying metasediments.

(ii) H 28 is a coarse-grained tourmaline-bearing pegmatite about 1 metre above H 26. The coarseness of texture (grain size of 3-4 mm) and the complicated inter-relationships of the constituent minerals are distinctive (fig.85). Large unstrained quartz crystals are poikiloblastic towards the felspars. Plagioclase, finely twinned on the albite law, commonly shows a small amount of fracturing, and is clouded by secondary mica. Coarsely twinned microcline is unfractured and generally free of secondary alteration. Muscovite and biotite, the former in large crystals and the latter in less abundant and smaller crystals, have ragged terminations, and do not appear to be in equilibrium with the felspars. The lack of widespread cataclasis and the largeness of the average grain size indicates the probability of post-F_3 crystallisation. However, the lenticular nature of the body suggests that it has probably undergone some previous deformation.

(iii) H 29, H 30. These tourmaline-bearing pegmatites occur some 2 metres above the granite contact, and are the lowermost rocks in the metasediments which contain tourmaline. Tourmaline occurs as large disoriented and zoned crystals, pleochroic in yellows and browns; it is probably schorlite. Quartz is abundant as interlocking crystals of varied size. Felspars occur in patches which show extreme secondary alteration, with only occasional faint twinning of plagioclase visible. One of the rocks shows a contact with schist, and the felspars of the latter are completely free of alteration.

Thin pegmatitic schlieren in the schists above H 29 and H 30 show similar characteristics, except for the occurrence of large

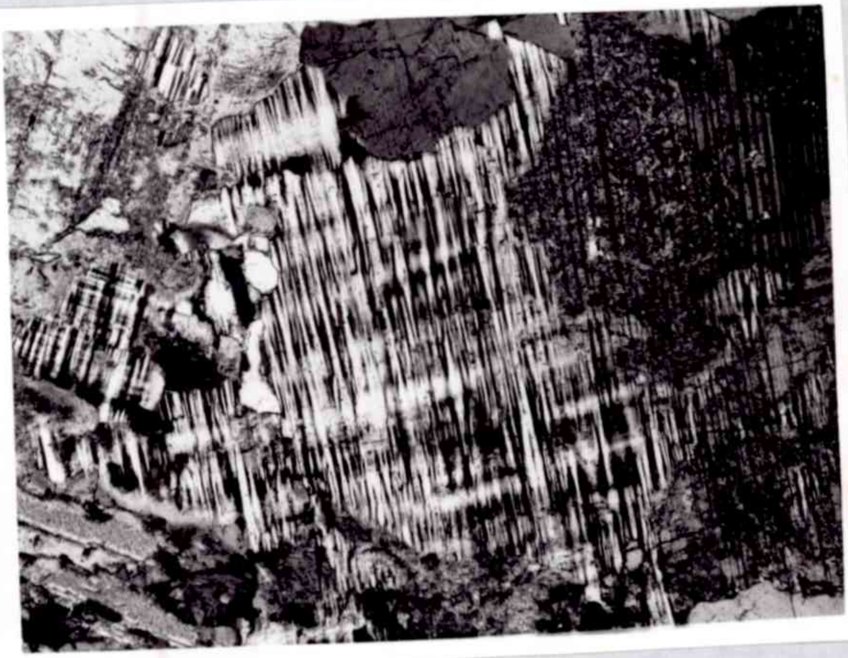


Fig.85. Interrelationships of the felspars from a pegmatite of the basal series, North Kvittind. The plagioclase is finely twinned and partially fractured, and shows a high degree of secondary alteration. The microcline, on the other hand, is free of alteration and unfractured; it appears that microcline is in the process of replacing plagioclase. x 30.

microcline porphyroblasts and rutile crystals up to 1 mm across rimmed by ilmenite. This association of rutile, tourmaline and ilmenite is uncommon in the Glomfjord region, and the only occurrence recorded from near Ørnes.

Remainder of the North Kvittind succession (fig.59, sections 5a and 5b).

Immediately overlying the rocks described above is a group of coarse well-foliated schists with abundant muscovite and subordinate biotite wrapped around occasional garnet porphyroblasts up to 1 cm in diameter. Fragmented quartz laminae separate micaceous layers, and there is an occasional development of irregular quartz veins. Localised chloritisation of biotite is developed.

Higher up in the succession, about 20 metres above the basal granite, is a group of quartzo-felspathic schists similar to the Galtskart silty-schists. They are fine-grained and granular, with a high proportion of microcline. Inclusions of orthite in biotite cause zones of discolouration; separate crystals of orthite are also present. Muscovite is of late-stage crystallisation, and cuts across the earlier fabric.

Summary and Conclusions. Although there is no systematic variation in the proportion of microcline in the metasediments above the basal granite, there is a general decrease away from the granite. Thus the semi-pelitic schists in immediate contact contain abundant microcline-rich laminae, while in the silty schists some 20 metres above, microcline is more granular and less common. The reason for the absence of microcline in the intervening garnet-bearing rocks is either due to incompatibility of microcline with garnet, or more likely, preferential introduction of the mineral into schists of particular composition. In both metasediments and basal granite, the microcline is of analogous appearance, and in each case its relationships with plagioclase, probably of earlier crystallisation, are similar. Thus, the microcline in the metasediments was probably derived from the basal granite where abundant microcline is encountered. This suggests localised granitisation of the metasediments (see also p. 231).

Plagioclases in the schists have a variable composition (Table 4) but one that is consistently more basic than those from the basal granites. In addition, the plagioclases from the schists are granular and often slightly fractured, and do not show evidence of a metasomatic origin. This evidence is used to suggest a primary difference between the metasediments and the granites (see also p. 231). The composition of the plagioclase crystals tends to become more basic away from the granite. The reasons for the extreme variation of plagioclase compositions in the pegmatites are unknown. The twin laws of the schist and pegmatite plagioclases are different. Whereas those from the pegmatites are invariably twinned on the albite law, combined albite-pericline twins are common in the schists.

A consistent difference in the biotites of the basal granites, compared with the overlying metasediments, is the green-brown colour of the former and the brown or red-brown colour of the latter. This may be due to variation in content of iron in the biotites.

Muscovite, present throughout the schist succession, is never seen in the basal granite, probably because of original variations in composition. Tourmaline is concentrated in the schists some distance from the granite, but even so it is clearly related to the late-stage pegmatites derived from the latter.

(b) The detailed part of the North Galtskart succession (fig.59, section 7a) consists of about 40 metres of flaggy micaceous schists, with marbles and calc schists towards the top. Two types of mica schists are present, each of which is interbanded with the other, and each occurring in roughly equal proportions (see fig.59):-

(a) The pelitic schists consist of well-foliated micaceous rocks, with muscovite occurring in thin laminae together with subordinate biotite. Slight bending of the muscovite is probably due to F_3 deformation. Green-brown or brown biotite is found in different rocks. Granular quartz and felspar separates the micaceous layers, while occasional quartz veins are present. Superficial alteration of plagioclase is common, and the felspar relationships are the same as are seen in the basal granites.

(b) Flaggy semi-pelitic schists are similar to equivalent rocks from the North Kvittind succession, described above (p. 233). Again microcline tends to occur in thin laminae, and there is an overall general decrease of the mineral away from the granite. Plagioclase is patchily sericitised and often shows combined albite-pericline twins, associated with zoning. The latter is sometimes well picked out by sericitised cores and clear rims to the crystals and is a separate phenomenon from sodic rims along microcline junctions, which are also developed. Both the types of myrmekite described on p. are seen. In one of the rocks muscovite is abundant, while microcline is subordinate; this may be due to reaction of the potash felspar with garnet, small crystals of which are present.

A coarse-grained pegmatite cutting the schists consists principally of large sericitised microcline crystals several cms across, which contain irregular veins of both highly altered and unaltered, untwinned felspar. These veins may represent perthitic albite. Lack of cataclasis indicates post-F_3 crystallisation.

Towards the top of the succession an unusual quartzo-felspathic schist occurs, associated with marbles. It contains a considerable proportion of phlogopitic biotite, and large , irregularly distributed calcite crystals. A thin band of amphibolite within the marbles is composed principally of dark green hornblende. Diopside occurs in occasional large crystals, partially replaced by pale green, probably actinolitic, amphibole. A similar variety of amphibole is found along the margins of a quartz vein. Plagioclase is present as small, strained, and sometimes zoned, crystals, of an earlier age of crystallisation than the hornblende.

(c) The South Galtskart succession (fig. 59, section 6a) contains rocks that are more uniformly pelitic than those on the N of Galtskart, with the spasmodic development of large garnets and a general lack of felspathisation. The rocks are generally well foliated with micas arranged in distinct layers which are wrapped around the occasional garnet porphyroblasts. S-shaped trains of inclusions of granular quartz within the latter are sometimes arranged almost at right angles to the present schistosity indicating rotation of the garnets, probably during F_2 times. The preservation of an early schistosity of this nature is common elsewhere in the Galtskart pelitic schists, and has also been described from other parts of the Glomfjord region (e.g. Ackermann 1960, Ackermann et al 1960). In the 'tails' of the garnet porphyroblasts the biotite crystals are considerably larger than elsewhere, and are partially chloritised. Zoned plagioclase, with comparable properties to that elsewhere in the basal successions, is present. Determinations show that the central portions of the zoned crystals are more acid than the rims (An 31 and An 37 respectively in one case).

Summary and Conclusions. The successions of North and South Galtskart are broadly similar to that on the N of Kvittind. In the quartzo-felspathic schists there is a trend towards more basic plagioclases away from the granites. The percentage of combined albite-pericline twins tend to increase away from the granite, and are very rarely seen in the granite itself. Irregular zoning of the plagioclase crystals commonly seen in this succession is of unknown significance. It may be a consequence of strain, associated with the complicated twins commonly developed.

Mineral assemblages are in accordance with those present in the kyanite-staurolite sub-facies, the main difference from the basal granite being the occurrence of muscovite. The reasons for the absence of calc-silicate minerals, such as grossularite or diopside, in the calcite-bearing quartzo-felspathic schist in the N. Galtskart succession are not clear, although the calcic nature of the plagioclase (An 33) and the presence of phlogopite is presumably a consequence of the calcareous affinities of the rock.

Rocks of comparable lithologies throughout the remainder of the Ørnes region contain similar assemblages, and there is no increase in metamorphic grade towards the basal granites.

(3) <u>Steffodalen and Galtskart Successions</u>

From structural considerations, the large group of interbedded pelitic and semi-pelitic schists of Steffodalen are regarded as being older than the metasediments overlying the Bjellåtind granite, and constituting the mountain of Galtskart (p. 168). However, it is difficult to trace the Steffodalen rocks westwards due to lack of exposure. Within the Steffodalen group are thin bands of calcareous pelitic schists, marbles etc., rocks which are prominent in the Galtskart succession, and conversely in the Galtskart succession are rocks of similar composition to the main mass of the Steffodalen group. For this reason the two successions are described together, while the pelitic rocks of the Steffodalen group are dealt with separately below.

Pelitic Schists

Most of the rocks belonging to this group are well foliated and show rusty weathering with prominent micaceous laminae. Porphyroblastic garnets are commonly distinctive. A characteristic series of finely interbedded pelitic and semi-pelitic schists forms part of the Steffodalen group. The individual bands are from ½ cm. to several cms across, and consist of alternate layers of quartzo-felspathic and micaceous material; the latter contain garnet. A sedimentary origin is probable, as the bands are remarkably continuous, and show the pattern of even the earliest folds (figs 36, 40).

(1) Steffodalen Pelitic Schists

The easterly outcrop of the schists in Steffodalen itself is separated from the probable westerly equivalents in the Blaatind-South Ørnes region.

In Steffodalen the rocks are predominantly biotite-rich and well foliated, with occasional ragged garnets. The latter contain abundant inclusions of quartz, and sometimes rotation of the porphyroblasts is indicated by the orientation of the trails of inclusions (fig.86E). Biotite forms crystals of variable size which wrap around the garnet porphyroblasts. Gentle F_2 or F_3 microfolds are sometimes visible, but in general biotite has recrystallised subsequently and is unbent. Equigranular quartz and plagioclase separates the micaceous laminae, and the felspar is usually slightly bent and fractured. Occasional quartz veins are also present.

Table 5. Pelitic Schists of the Steffodalen Group

(a) In Steffodalen itself. 10 specimens examined.

The predominant assemblage is quartz, plagioclase, biotite, muscovite, garnet. Microcline is present in only two of the rocks. Apatite and magnetite are present with occasional tourmaline and epidote (pistacite).

Plagioclase composition

H 245 An 47 from 3 determinations

Mineral properties

H 121 Biotite Ny = 1.620 Pleochroism: X = very pale brown
Y = Z = red-brown

Corresponds to iron-rich variety, probably siderophyllite

H 249 Biotite Ny = 1.625 Pleochroism: X = very pale yellow-brown
Y = Z = very dark brown

probably siderophyllite

(b) In Blaatind-South Ørnes. 9 specimens examined

The predominant assemblage is quartz-plagioclase-biotite-garnet. Occasional rocks contain muscovite or one or more of kyanite, staurolite and sillimanite. Apatite and magnetite are common with occasional sphene and pistacite. Chlorite is sometimes present.

Plagioclase composition

H 146 An 37 from 3 determinations H 265 An 15 from 3 determinations

H 148 An 22 from 3 determinations H 463 An 38 from 4 determinations
An 25 (inside of crystal)
An 20 (outside of crystal)

Mineral properties

H 146 Sillimanite 2V = 16° Nx = 1.653 Nz = 1.667

Modal analysis

	Quartz	Plagio-clase	Biotite	Garnet	Magnetite	Apatite	Total
H 142	19.7	56.8	22.4	.3	.3	.5	100

Fig.86. The form of garnet and kyanite porphyroblasts in pelitic schists, from the Steffodalen and Galtskart successions.

A. Inclusions of magnetite that have been incorporated in a garnet crystal of later development, Bugten.

B. Granular quartz of F_1 age within a garnet porphyroblast. The texture outside the garnet is considerably coarser than that inside. Barfjell.

C. A garnet porphyroblast growing across isoclinal microfolds preserved in biotite crystals. South Galtskart.

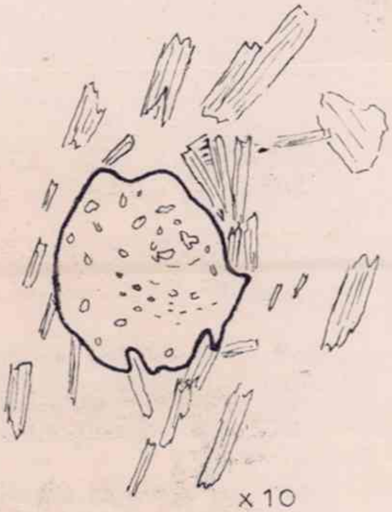
D. Fine-grained magnetite and quartz inclusions within a garnet crystal. Sphene, which is stippled, is concentrated in the tails of the porphyroblast. Lath-like hornblende crystals are of a later stage of crystallisation than the garnet. Spilderdalen.

E. S-shaped inclusions in garnet, indicating F_2 rotation of the crystal during growth. Steffodalen. Later crystallisation may be indicated by the clear rim to the crystal.

F. Disoriented late-stage kyanite crystals, probably post-F_2 in age, growing across earlier biotite crystals, South Galtskart.

x 15

A

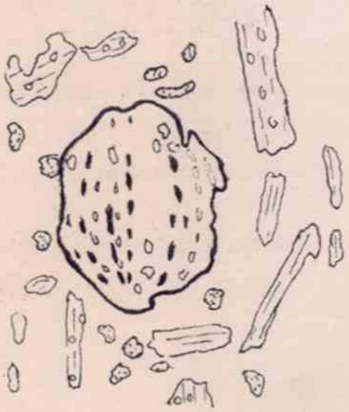


x 10

B

x 15

C

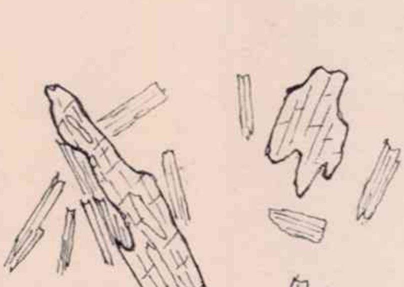


x10

D

x 15

E



x 10

F

The pelitic schists of the Blaatind-South Ørnes region are interbanded with other rocks of variable character, and constitute only a small part of the total succession. Abundant brown and ragged biotite define a schistosity that is often interrupted by garnet porphyroblasts. Discolouration around minute inclusions of unknown composition is a common feature in the biotite. The variation in euhedrism and outline of the garnet porphyroblasts is shown in fig.86 (A, B). Preservation of an early schistosity and the subsequent rotation of the garnets is clearly demonstrated by the orientation of granular quartz inclusions. Subsequent renewed crystallisation may be indicated by the occasional occurrence of inclusion-free rims. Partial chloritisation of biotite appears to be concentrated near garnet crystals. The small amount of kyanite present in some of the rocks is poikiloblastic and probably crystallised relatively late. Much of the plagioclase shows albite and pericline twinning and is often fractured. Local zoning around inclusions of quartz and other minerals is sometimes observed.

A distinctive sillimanite and occasionally staurolite-bearing schist has been identified in a thin stratigraphic band between Bugten and Mesöen. At Bugten the rock consists of large plagioclase crystals, separated by zones of granular quartz and cordierite; associated sillimanite occurs in large, roughly oriented needles (fig.87). Trails of magnetite grains are present

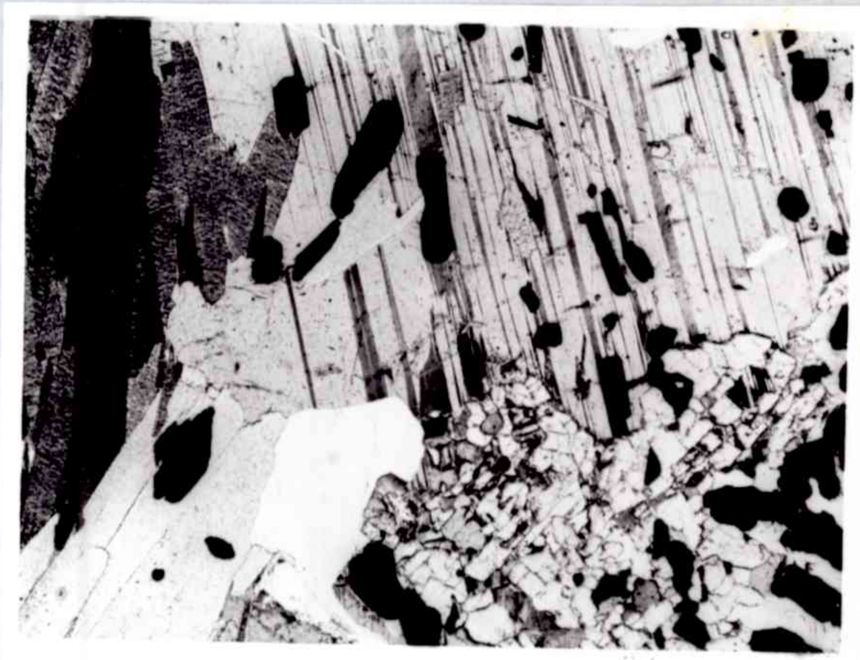


Fig.87. A large plagioclase crystal, showing irregular twin lamellae, and containing inclusions of magnetite, in staurolite schist, Mesöen. The light-coloured material with high relief in the bottom right hand corner is a single, large, sieve-like porphyroblast of staurolite, with common inclusions of magnetite plagioclase etc. x 30.

Fig.2. A large [illegible] crystal, [illegible]

throughout the rocks; their average size is smaller inside the plagioclase than elsewhere. Clinozoizite is found as small grains in thin bands within some of the plagioclase crystals. A few ragged crystals of biotite occur. Very occasionally small euhedral staurolite crystals are present. The stratigraphic equivalent on Mesøen has a somewhat different petrography. Plagioclase forms large crystals often twinned on the albite-pericline law, and commonly zoned around inclusions of magnetite, quartz and biotite. The twin lamellae of the plagioclases are usually spindle-shaped, and the crystals are sometimes fractured. Large green-brown biotites are characteristic, as are the occasional sieve-like poikiloblastic crystals of staurolite up to 3 or 4 mm in diameter. No sillimanite or cordierite has been identified.

The mineral assemblages in the pelitic schists of the Steffodalen group are generally indicative of the kyanite-staurolite sub-facies. However, unusual assemblages occur in the staurolite-bearing schists, and are worthy of further comment. Staurolite appears to be stable over a narrow temperature range and is commonly substituted for by kyanite and almandine (Turner and Verhoogen 1951). Associations of kyanite and almandine are common in the Ørnes region, but staurolite is rare. The absence in the Mesøen staurolite schist of any potash-bearing minerals except biotite is probably due to the high percentage of iron and magnesium; all the available potash may then have been used in the formation of biotite. In the other assemblage, where sillimanite, staurolite and cordierite co-exist in apparent equilibrium, type minerals from three subdivisions of the amphibolite facies are present in a single rock (cf. Turner and Verhoogen 1951). Cordierite may have developed preferentially to biotite due to lack of available water - in this case, there would be no other mineral capable of accommodating MgO. The general anhydrous nature of the rock supports this view.

Galtskart Pelitic Schists

Both well-foliated, finely-banded micaceous schists, and coarsely crystalline garnet- or kyanite-schists occur. In the garnet-free rocks, biotite is the dominant mica, is always brown in colour, and often arranged in closely-spaced laminae.

Semi-Pelitic Schists (Quartzo-Felspathic Schists)

Distinction was made in the field between 'silty' schists - massive grey and variably banded rocks, and flaggy semi-pelitic schists (see p. 24).

Silty Schists

Quartz, plagioclase and biotite are always present, while muscovite is found in all but two of the rocks sectioned. Microcline, in variable amount, is common, but not ubiquitous (Table 7). Garnet is widespread, but only present in small amounts.

Biotite is normally found as small brown or green-brown crystals, generally well oriented, and sometimes bent by slight F_3 micro-folds. Occasional clots of larger crystals are also

Table 6. Pelitic Schists from the Galtskart Succession

20 specimens examined.

The predominant assemblage is quartz, plagioclase and biotite. Muscovite is present in 11 of the rocks and garnet in 14. Microcline occurs in two of the rocks, and is incompatible with garnet. Kyanite is found in 5 of the rocks and staurolite in 1. Apatite and magnetite are present with occasional tourmaline. Secondary chlorite is quite common.

Plagioclase composition

H 61 An 30 from 3 determinations

H 70 An 25 from 1 determination

H 74 An 35 from 4 determinations

H 88 An 33 from 5 determinations

H 117 An 28 from 2 determinations

H 213 An 37 from 3 determinations

H 313 An 28 from 2 determinations

H 363 An 39 from 3 determinations

Mineral properties

H 70 Kyanite Ny = 1.719 Nz = 1.725

Modal Analysis

	Quartz	Plagioclase	Biotite	Muscovite	Kyanite	Garnet	Apatite	Magnetite	Tourmaline	Sphene	Chlorite	Total
H 61	64.9	8.7	13.7	8.9	1.1	2.2	.5	1.1				99.[illegible]
H 70	61.8	8.1	19.8	.2	9.8		.2			.1		100.[illegible]
H 207	34.8	19.1	27.2		3.5	14.3	.3	.4	.2		.4	100.[illegible]
H 363	20.5	33.5	25.9		1.2	17.8	.4	.4	.4			100.[illegible]

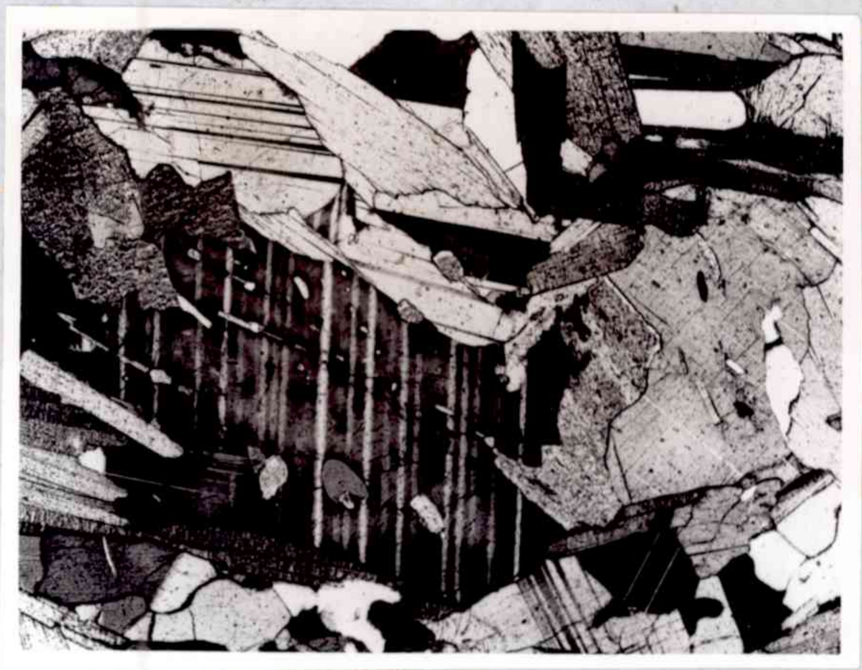


Fig.88. A coarse-grained pelitic schist from Blaatind, showing inclusions in plagioclase crystals, and a slight irregularity of some of the twin lamellae. x 30.

Table 7. Semi-pelitic Schists of the Galtskart Succession

(a) Silty Schists. 12 specimens examined.

The predominant assemblage is quartz, plagioclase, microcline (in 9 rocks), biotite, muscovite. Garnet is present in 5 rocks, and secondary chlorite in 6. The most important accessory mineral is apatite, but sphene, magnetite, epidote (pistacite and clinozoizite) and tourmaline occasionally occur.

Plagioclase Composition

H 62	An 19 from 5 determinations	H 90	An 24 from 5 determinations
H 72	An 26 from 5 determinations	H 217	An 32 from 4 determinations
H 73	An 12 from 2 determinations	H 245	An 47 from 3 determinations
H 80	An 28 from 3 determinations		

Modal Analysis

	Quartz	Plagio-clase	Micro-cline	Bio-tite	Musco-vite	Garnet	Apa-tite	Total
H 62	56.2	4.9	6.9	17.4	14.0	.2	.4	100.0
H 73	32.3	35.8	6.7	22.0	2.0	.7	.5	100.0

(b) Flaggy Semi-pelitic Schists. 13 specimens examined.

The predominant assemblage is quartz, plagioclase, microcline (7 rocks), biotite and muscovite. Garnet is present in 4 of the rocks. Secondary chlorite is present in 6 of the rocks. Spasmodic apatite, sphene, magnetite, tourmaline and pistacite occur.

Plagioclase Composition

H 254 An 18 from 2 determinations

Modal Analysis

	Quartz	Plagio-clase	Musco-vite	Bio-tite	Garnet	Apa-tite	Magne-tite	Total
H 310	45.0	25.2	15.6	13.3	.5	.2	.2	100.0

present. Discolouration around small inclusions of orthite, and rarely pistacite, are seen, and the biotite crystals are commonly partially altered to pale green chlorite. Muscovite is usually subordinate to biotite, forming late cross-cutting crystals, but when it is abundant it has the same type of occurrence as biotite.

The quartzo-felspathic material is granular. Plagioclase is usually the dominant felspar and often exhibits slight bending and fracturing. Superficial sericitisation is commonly observed. Microcline, present both in compact and small poikoloblastic crystals, is probably of a later age than plagioclase. Myrmekite is developed along some of the plagioclase-microcline boundaries.

Apatite is the most widespread accessory mineral, generally as small grains but in one of the rocks (H 395) in rounded crystals up to 2 mm in diameter. The remaining accessories (Table 7) are only found in very small amount.

Mineral assemblages belong to the potash-excess field of the staurolite-kyanite sub-facies (Turner and Verhoogen, 1951). The occurrence of garnet, micas and microcline in the same rock is indicative of disequilibrium conditions, as normally reaction would take place between the potash, felspar and garnet to form mica. The lateness of development of microcline is probably the reason for this association, while the small and ragged garnets may be of a relict nature.

Some of the silty schists approach a granitic composition, a fact which may be apparent from their field characters. However, microcline is never as abundant as in the basal granites, and generally occurs in smaller granular crystals. The common occurrence of green-brown biotite crystals, which are found in the basal granites, but in few others among the metasediments also links the two rock types. In other neighbouring regions (Walton 1959) it has been suggested that the basal granites may have been derived by felspathisation of a rock type closely similar to the 'silty' schist, a view that is supported by the present study (see also p. 313).

Flaggy Semi-Pelitic Schists

The assemblages present in this group are similar to those of the silty schists, with microcline present in a high percentage of the rocks. Garnet is again only a subordinate mineral, occurring always in small crystals. Close comparison exists petrographically with the silty schist group, the principal difference being the ubiquitous occurrence of brown biotite, and the smaller percentage of felspathic material in the flaggy semi-pelites. The garnets are usually more obviously skeletal.

Calcareous Pelitic Schists

A characteristic of this group is the abundance of the mineral species present, often up to ten or twelve in a single thin section. Constancy of assemblage from rock to rock is also noteworthy, even to the accessories. The most important minerals are quartz, plagioclase, scapolite, biotite, hornblende, diopside and calcite, all of which, with rare exceptions, occur in every

Table 8. Calcareous Pelitic Schists, Galtskart Succession

14 specimens examined.

The predominant assemblage, which is remarkably persistent, is quartz, plagioclase, biotite, diopside, hornblende, calcite, scapolite (12 rocks), garnet (8 rocks). Muscovite is found in 3 of the rocks. Sphene and magnetite are common, with occasional epidote (pistacite or clinozoizite) and apatite. Chlorite occurs in 2 of the rocks.

Plagioclase Composition

H 15 An 36 from 3 determinations
H 39 An 34 from 1 determination
H 55 An 38 from 4 determinations

Mineral Properties

Amphibole 2V = 68½ optically negative

Pleochroism X = very pale yellow-green
Y = green
Z = dark green

Modal Analysis

	Quartz	Plagioclase	Biotite	Scapolite	Diopside	Hornblende	Calcite	Magnetite	Garnet	Sphene	Total
H 39	56.5	4.0	12.0	15.6	4.9	3.0	3.7	.2			100.0
H 56	39.9	1.9	19.0	24.4	7.2	2.1	2.7	.3	2.3	.2	100.0
H 64	30.1	17.0	19.0	12.9	10.0	6.5	3.4			1.1	100.0

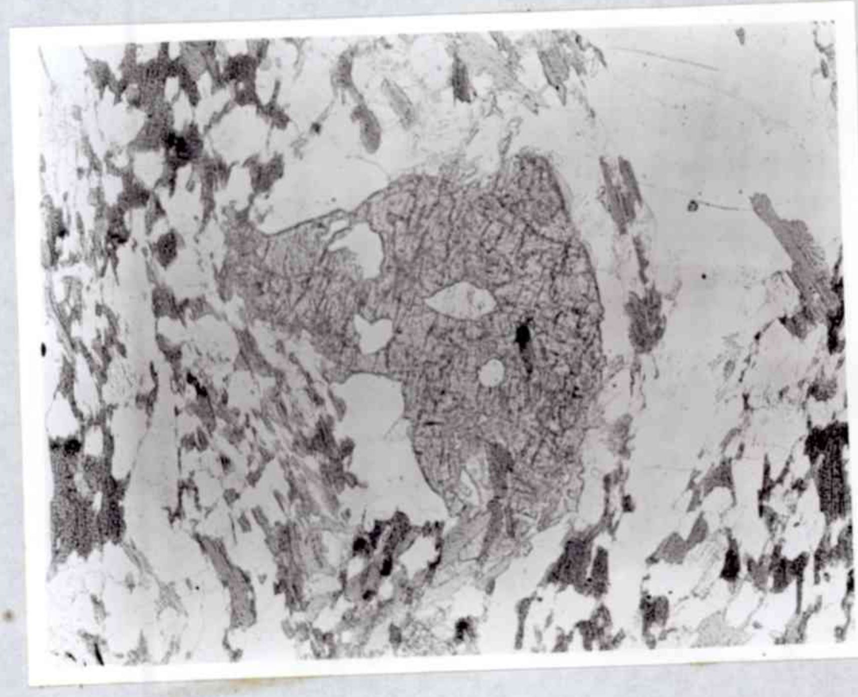


Fig.89. An undeformed porphyroblastic diopside crystal in a calcareous pelitic schist, North Digermalen. The crystal developed after the gentle microfolds is visible in ragged biotites. x 30.

rock (Table 8).

Petrographically, the rocks consist of pelitic schists with a variable percentage of calcareous minerals added (normally about 30%). Small ragged brown biotites occur in rough layers, and are sometimes slightly bent. Pleochroic haloes around minute inclusions are sometimes common. Quartz is present in granular crystals and occasional late veins. Small porphyroblastic plagioclases, often showing combined albite-pericline twins, are commonly fractured; the intricate twinning may be a consequence of this deformation. Some of the crystals are zoned.

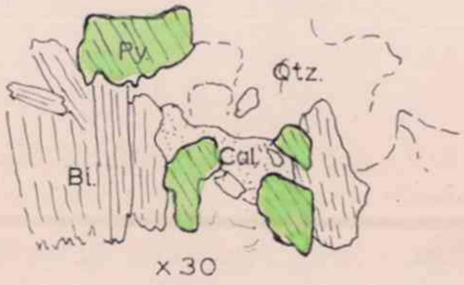
The mode of occurrence of the calcareous minerals is variable, and some of the complex interrelationships are shown in Fig.90. Diopside is found commonly in sieve-like porphyroblasts up to 2 mm in diameter. Lateness of development is indicated by lack of deformation, and the way the crystals grow across the earlier fabric. Hornblende is often associated with the diopside, as an alteration product, but discrete crystals of the mineral are also seen. Rounded and granular scapolite crystals are commonly associated with diopside and hornblende, to form distinct bands a few mm thick. Scapolite occurs in some of the quartz veins. Small amounts of calcite are also of late origin.

Ragged garnets, with many inclusions, sometimes form small irregular porphyroblasts. Associated biotite, hornblende, and magnetite are possibly alteration products of the garnet. Magnetite is associated with biotite, and sometimes enclosed by the latter. The epidote, which occurs as occasional small grains

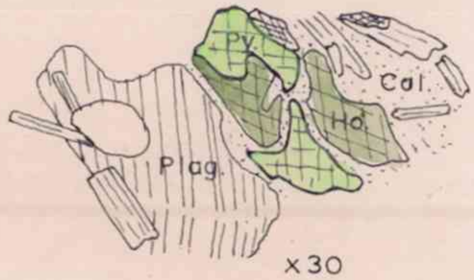
Fig.90. Textures in calcareous pelitic schists of the Galtskart succession.

Py.	=	pyroxene (probably diopside)
Bi.	=	biotite
Cal.	=	calcite
Qtz.	=	quartz
Ho.	=	hornblende
Plag.	=	plagioclase
Scap.	=	scapolite
Ga.	=	garnet

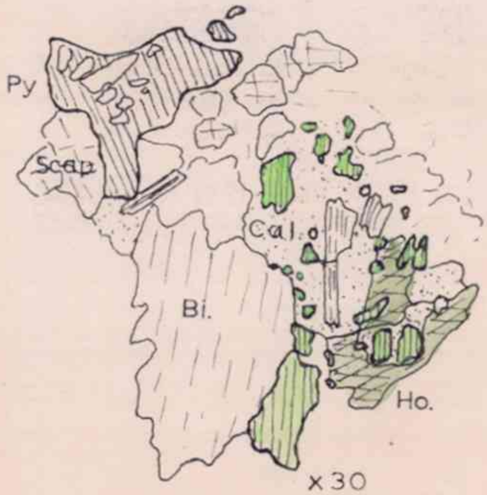
In each of the diagrams, the fragments of hornblende or diopside that are coloured similarly are in optical continuity.


Py.
Qtz.
Cal.
Bi.
x 30

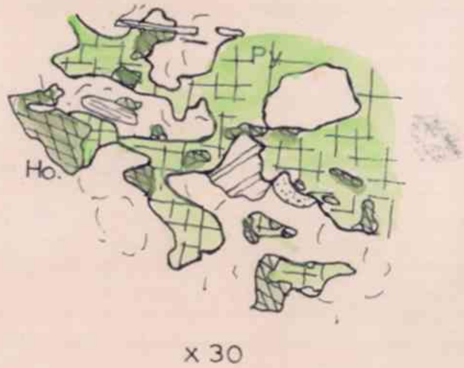
A


Py.
Cal.
Ho.
Plag.
x 30

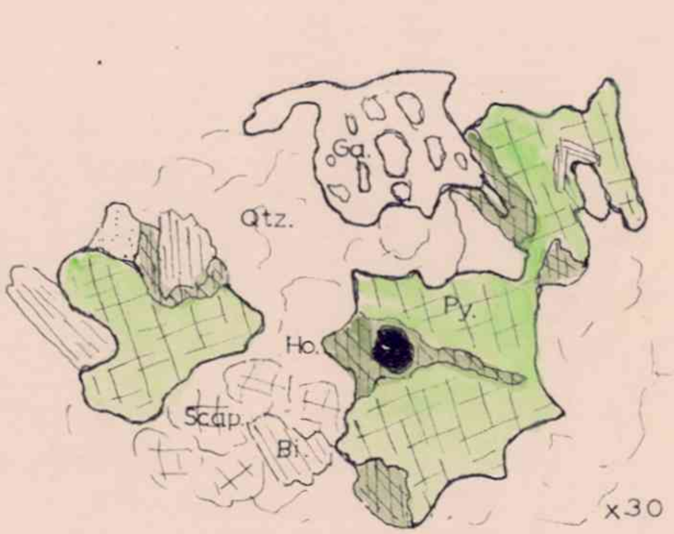
B


Py
Scap.
Cal.
Bi.
Ho.
x 30

C


Py.
Ho.
x 30

D


Ga.
Qtz.
Py.
Ho.
Scap.
Bi.
x 30

E

can be either clinozoizite or pistacite.

A thin band within the West Lysvand calcareous pelitic schist, is composed principally of large porphyroblastic garnets 3-4 mm in diameter with a normal pelitic matrix which also contains granular crystals of scapolite and a few small ragged hornblendes.

As the mineral assemblages observed are so distinctive, their significance in relation to metamorphic facies is worthy of discussion. The stable assemblage, almandine-diopside-hornblende, is only developed in the sub-facies of that name (Turner and Verhoogen, 1951) a different sub-facies from that to which the remainder of the rocks of the Ørnes region belong. It is clear that there is a lithological control to the assemblage, and the lack of equilibrium is indicated by the relict nature of the garnet, and the lateness of development of diopside and hornblende. Thus, the assemblages probably conform to the potash-excess field of the kyanite-staurolite sub-facies, although the co-existence of diopside and biotite is anomalous. The presence of calcite is compatible with excess silica due to the inability of wollastonite to form under these conditions. Absence of microcline and the rarity of muscovite is probably due to original composition.

The occurrence of abundant scapolite requires further explanation. D.M.Shaw (1960), in dealing with the geochemistry of scapolite, reviews the existing literature on the petrography of all forms of the mineral. For regionally developed scapolite, of similar mode of occurrence to that in the Ørnes region, a variety of explanations have been proposed. In some cases, it can be related directly to igneous sources (e.g. Barlow, 1910, Carlson 1957, Barth, 1930), but elsewhere its origin is less clear. Thus Sundius (1915), in describing the development of scapolite from the Kiruna district, where it is associated with iron, invokes two causes (a) contact pneumatolysis (metasomatism), (b) remobilisation of pre-existing material. He believes the latter to be less important, as there are no impure marbles for the production of the necessary CO_2, and there is inadequate Cl in the associated sediments. Von Eckerman (1922) describes the development of scapolite in calcareous rocks with associated skarn minerals under conditions of deep-seated metamorphism. A similar mode of development is suggested by White (1959) for a group of scapolite-pyroxene rocks in South Australia. Edwards and Baker (1953), in describing scapolite-bearing Pre-Cambrian rocks of Queensland, Australia, regard its formation as due to Na-metasomatism. Under normal conditions, soda metasomatism leads to the production of albite, but here, due to the existence of calcareous shales, scapolite forms preferentially, due to the abundance of CO_2 and Cl.

Somewhat similar conditions to those described by Edwards and Baker appear to be applicable to the Ørnes region, although evidence of Na-metasomatism is slight in the calcareous pelitic schists. Here the scapolite is of late development. If scapolite were formed by metasomatism there should be a close correlation between its occurrence in the calcareous rocks with that of plagioclase in pelitic and semi-pelitic rocks. Within the calcareous pelitic schists bent and fractured plagioclase (presumably of relatively early formation) occurs in the same rock as scapolite. Interaction between the two minerals is rarely seen due to the restriction of the scapolite to the late quartz veins; but when the two minerals are seen in contact, there is no evidence of scapolitisation of the plagioclase. Moreover, there is no evidence of extensive metasomatic development of plagioclase as a late product in the remaining metasediments, and the composition of the plagioclase varies according to the schists in which they are found. Because of these factors, the writer considers that the scapolite developed by internal redistribution of pre-existing material (CO_2, Cl etc). This process was probably auto-metasomatic, developed at the end of regional metamorphism and probably coeval with the quartz veins.

The composition of the scapolite in the calcareous pelitic schists is unknown, but as the birefringence is generally high, it probably contains a considerable proportion of the Ca end-member meionite. This is compatible with its occurrence in calcareous sediments.

Calc Schists

Thin layers of banded or unbandeddark green calc-schists occur in association with, or separate from, marble bands. Both diopside and hornblende are normally present, and often make up a considerable proportion of the rock. Quartz and plagioclase are present with rare exceptions, while microcline is also common. Calcite is

Table 9. Calc Schists, Amphibolites and Calc Silicate Rocks, Galtskart Succession

(a) Calc Schists. 7 specimens examined.

The group is very variable, particularly in the relative proportions of the various minerals, and generalisations are difficult. However, the predominant mineral assemblage is quartz, plagioclase, microcline, hornblende, diopside, calcite (5 rocks), scapolite. Sphene, biotite, epidote (pistacite or clinozoizite) and occasional magnetite, tourmaline and apatite are also present.

Plagioclase composition

		(i)	(ii)	
H 444	An	55	60	inside of crystals
	An	40	40	outside of crystals

		(i)	(ii)	(iii)	
H 447	An	30	37	29	inside of crystals
	An	35	39	40	outside of crystals

(b) Amphibolites

(1) Sedimentary amphibolites - 8 specimens examined.

The predominant assemblage is quartz, plagioclase, hornblende, biotite (6 rocks), diopside (3 rocks), calcite (3 rocks), scapolite (3 rocks), garnet (3 rocks), microcline (2 rocks). Sphene is widespread, with epidote (pistacite or clinozoizite), magnetite and apatite less commonly developed.

Plagioclase composition

H 6	An 40 from 6 determinations	H 135	An 35 from 3 determinations
H 84	An 31 from 3 determinations	H 216	An 37 from 4 determinations

Mineral Properties

H 135 Amphibole $2V = 80^\circ$ optically negative $Z\ c = 12^\circ$

Pleochroism X = very light yellow-green
Y = dark grass-green
Z = dark blue-green

Table 9 (cont.)

(2) Igneous Amphibolites. 9 specimens examined.

The predominant assemblage is quartz, plagioclase, hornblende, biotite (4 rocks). Garnet is present in 3 of the rocks. Magnetite and sphene are widespread, with occasional epidote (pistacite or clinozoizite) and apatite. Chlorite is sometimes present.

Plagioclase composition

H 16 An 28 from 3 determinations
H 40 An 26 from 6 determinations
H 52 An 33 from 3 determinations
H 94 An 26 from 2 determinations
H 205 An 50 from 1 determination

Mineral Properties

H 6 Amphibole $2V = 78^\circ, 74^\circ$, optically negative
Pleochroism: X = nearly colourless
Y = grass-green
Z = very dark green

H 16 Amphibole $2V = 84^\circ, 81^\circ$, optically negative. $Zc = 17^\circ, 20^\circ$
Pleochroism: X = colourless
Y = yellow-green
Z = dark green
$Ny = 1.654$, $Nz = 1.668$

H 205 Amphibole $2V = 55^\circ$, optically negative $Zc = 12\frac{1}{2}^\circ$
Pleochroism: X = Very pale green
Y = Z = dark green

(c) Calc-Silicate Rocks

In this group an amphibole or diopside makes up more than 70% of the total. Most of the calc-silicate rocks have sedimentary affinities, and therefore contain small amounts of minerals characteristic of that group.

Mineral Properties

H 118 Colourless to pale green amphibole $2V = 88^\circ$ optically negative
$Zc = 5^\circ$
$Nz = 1.63, 1.624$
Component of Nx and Ny = 1.618, 1.617
Probably Tremolite-Actinolite

often seen, sometimes in association with microcline. Scapolite and a number of accessory minerals, the most important of which are sphene and epidote, also occur. Garnet is present in only three of the rocks examined. Two broad groups of calc schist may be recognised: (a) fine-grained rocks showing mineral banding, (b) coarse-grained unbanded types with diopside and biotite forming large porphyroblasts.

(a) Fine-grained varieties. Ragged hornblende and diopside crystals, often with inclusions of quartz and microcline occur together. The hornblende is usually highly coloured, although in a few instances it may be accompanied by a very pale hornblende. Diopside is usually markedly green, and probably Fe-rich. Biotite, when present, always forms ragged, often slightly bent, flakes, commonly partially chloritised. Discolouration around small allanite crystals is sometimes seen in both hornblende and biotite.

Plagioclase and quartz usually form fairly small granular crystals, with the occasional development of late quartz veins composed of crystals several mm in diameter. The plagioclase, commonly twinned on the albite-pericline law, is often zoned and sometimes shows bent twin lamellae.

Late scapolite in rounded crystals of varied size is often concentrated near quartz veins. No interaction with plagioclase was observed, but common boundaries between the two minerals are rare.

Microcline occurs as small irregular crystals in variable amounts. Some rocks contain larger porphyroblasts clearly of late development. Calcite is uncommon, and usually forms late crystals of small size. The epidote present is normally clinozoizite, occurring as irregular grains; in one of the rocks it is associated with a quartz vein of late development.

(b) <u>Coarse-grained variety</u>. Large diopside porphyroblasts, several mm across and commonly enclosing many small crystals of quartz, biotite etc., are characteristic of this group, which is less abundant than (a). Small pale-green hornblende crystals appear to have formed from diopside. Bands of disoriented biotites occur in one of the rocks, the crystals showing discolouration around allanite inclusions.

The remaining minerals have a similar mode of occurrence as in group (a) except for a general increase of grain size. Interaction between the two felspars with the production of true myrmekite is well displayed in one of the rocks (Fig.91).

Mineral assemblages developed in the calc schist group conform to the staurolite-kyanite sub-facies. The absence of grossularite is probably due to the ubiquitous occurrence of diopside and hornblende, in place of grossularite-diopside. In one of the rocks (H 444) it appears that microcline is later than scapolite, but whether this is a generalisation that can be applied to the remainder is unknown. The genesis of scapolite is regarded as being governed by the same factors as in the case of the calcareous pelitic schists (p. 265), while the microcline is thought to have

Fig.91. Development of true myrmekite in a microcline-rich zone of a calc-schist, North Steffodalen. x 40.

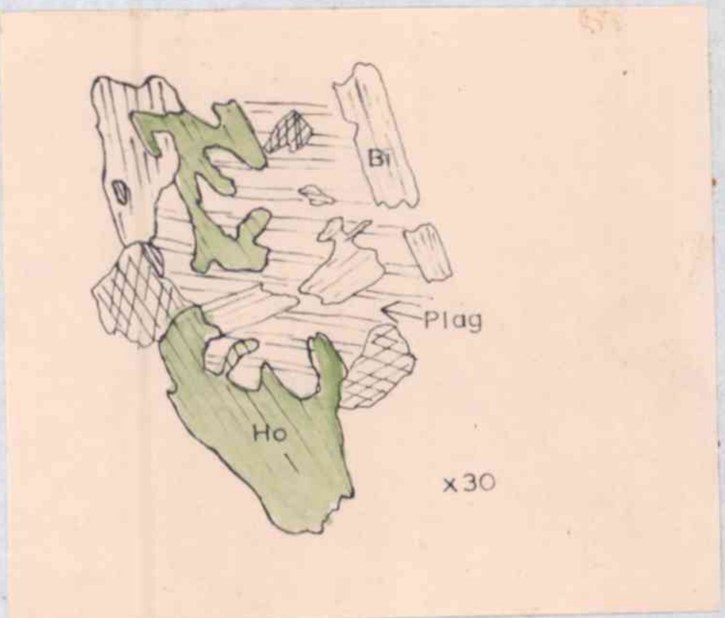
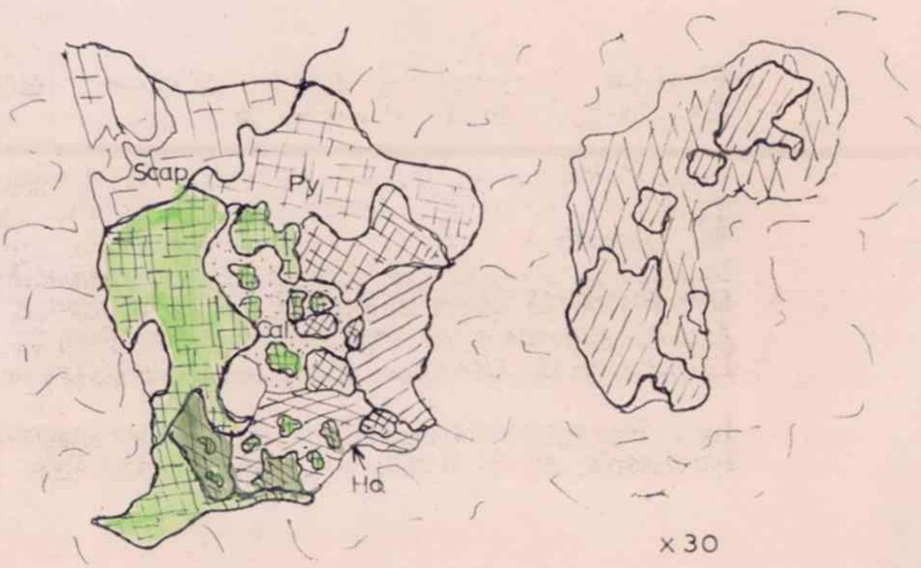


Fig.92. Relationships between hornblende, biotite and plagioclase from a calc schist, South Galtskart. The coloured hornblende is all in optic continuity. Plagioclase appears to be later in development than hornblende and biotite.

Fig.93. Textures in calc-schists and impure marbles from the Galtskart Succession.

A. Textures of calcareous minerals surrounded by quartz in a coarse-grained calc-schist. Complicated interrelations exist between hornblende and pyroxene (probably diopside). Fragments of individual crystals are coloured similarly, while the cleavage indicates either pyroxene or amphibole. Calcite is probably of later crystallisation than both pyroxene and amphibole.

B. Two generations of tremolite in impure marble. Fragments of Tr 1 are coloured similarly.



A

B

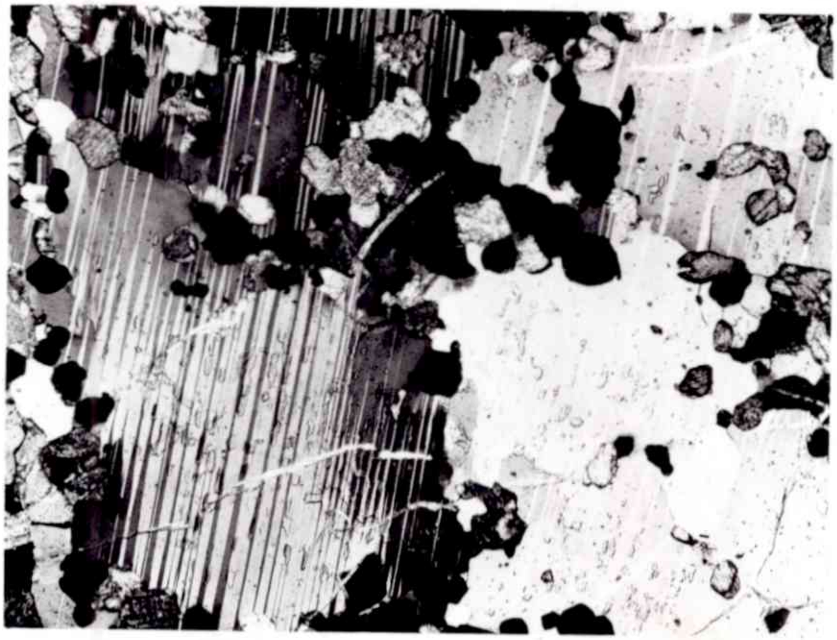


Fig.94. Large plagioclase crystals, showing spindle-shaped twin lamellae and inclusions of epidote, diopside and garnet from an amphibolite, South Galtskart. x 30.

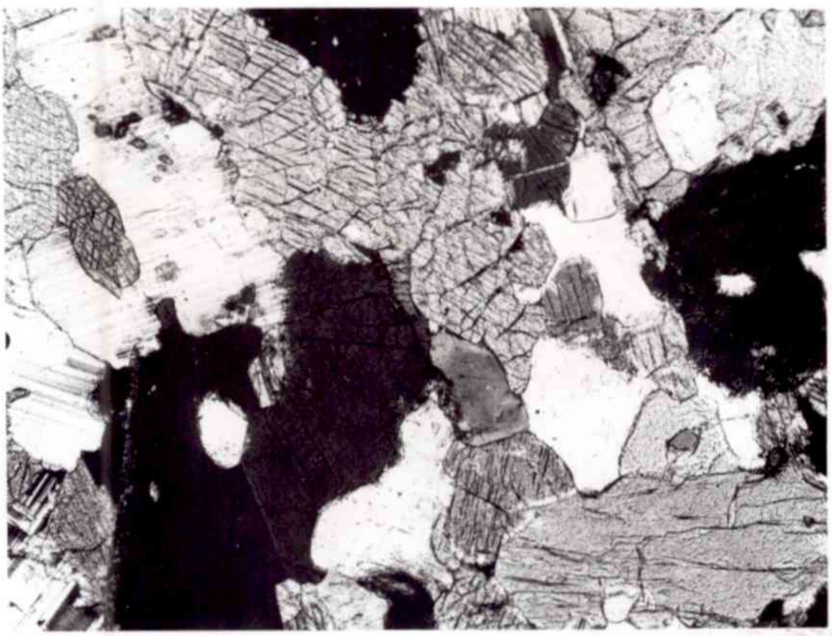


Fig.95. General appearance of coarse-grained amphibolite, North Spildervand. The hornblende crystals are ragged and poorly oriented. Plagioclase sometimes shows zoning, as exhibited by the small crystal almost in the centre of the photograph. x 40.

been introduced metasomatically.

An unusual granular epidote-diopside-quartz rock, with an average and remarkably constant grain size of 1/5 mm, occurs as a thin band within a pelitic schist lens in the Ørnes gneisses. The epidote is pistacite (Table 9) and is equal in amount to diopside. Together, they form 50% of the rock, the remainder being principally quartz, with very occasional crystals of sphene. The association diopside-epidote-quartz, with no accompanying felspar or grossularite, is unusual. It appears unlikely that the rock is a product of metamorphic differentiation, as its fabric is fine-grained, homogeneous and granular. In the field it has every appearance of representing an original band.

Amphibolites

Two subdivisions are recognised according to whether the parent rock is of presumed igneous or sedimentary origin.

(i) Igneous amphibolites. This group is composed principally of hornblende, plagioclase and quartz, with occasional small and ragged biotites. Magnetite and sphene are prominent accessory minerals, and in some of the rocks make up as much as 10% of the total. Large, irregularly shaped garnets, several mm in diameter, are abundant in some of the rocks.

In the garnet-free amphibolites, the most common mineral is highly-coloured hornblende, contrasting with the paler varieties in the calc-schist group, and constituting more than 50% of the total. They are very variable in size in different rocks, from large disoriented poikiloblastic crystals several mm long to small and well-oriented crystals. Biotite is uncommon, and occurs sometimes as ragged crystals, and elsewhere as extensive flakes which penetrate the hornblende.

Plagioclase often shows complex twinning and some degree of fracturing (Fig.94). Zoning around inclusions of hornblende indicates local interaction between the two minerals. Quartz occurs both in granular crystals and in late veins.

Sphene and magnetite are usually associated, the former generally in clots composed of a multitude of individual grains, and the latter as granular crystals. In the rock where sphene is most abundant (H 87) the hornblende is blue-green.

In the garnet-bearing amphibolites, the garnets are porphyroblastic with abundant inclusions, some of which indicate rotation during growth. Hornblende is usually ragged and poikiloblastic, and in most cases has formed subsequent to the garnet. All the garnets of this group are markedly pink and are probably grossularite.

(ii) Sedimentary amphibolites. Distinction from the presumed igneous amphibolites is made on the occurrence of biotite or diopside as an important constituent, together with a higher percentage of quartz.

Hornblende is generally of a similar appearance to that in the igneous amphibolites, and always forms very ragged crystals. The biotites are red-brown, and sometimes indicate a relict schistosity, largely masked by later development of hornblende.

Diopside is usually subordinate in amount to hornblende, and may be missing altogether. It usually forms granular crystals. One of the rocks (H 6) contains abundant euhedral pistacite and clinozoizite, up to 3 mm long, concentrated in a quartz vein. Garnet is very rare, having been found in only one of the rocks, as small, rounded crystals surrounded by bent biotites (H 134).

The mineral assemblages in both igneous and sedimentary amphibolites conform to the staurolite-kyanite sub-facies. Contrasting markedly with the calc-schists is the absence of scapolite and microcline. In the case of scapolite this is probably due to differences in original composition, while for microcline, regarded as being derived metasomatically, the amphibolites may have been chemically resistant to the invasion of microcline-bearing solutions.

Calc Silicate Rocks

These rocks, composed of 80% or more of calc-silicate minerals, have affinities with both amphibolites and calc-schists. The majority are obviously of sedimentary origin, and consist principally of diopside and amphibole in very variable proportions. Small amounts of calcite, scapolite, sphene etc. are commonly present. Diopside may form large crystals, probably several cms in diameter, and often partially altered to colourless amphibole. Association with pale-green hornblende is shown in Fig.93A. The relationship between tremolite and calcite is shown in Fig.93B. It is this group of rocks which shows the most complicated textures while the lack of deformation throughout indicates late, probably post-F_3, crystallisation. Only one rock in this group is believed to have affinities with igneous amphibolites: it is made up largely of small well-oriented and highly-coloured hornblende crystals.

Mineral assemblages of the calc-silicate rocks are similar to those of the calc-schists and amphibolites: only the proportions of the minerals differ. The original rocks may, in most cases, have been very impure dolomitic rocks. The association diopside-tremolite-calcite corresponds to one of the products of reaction in Bowen's series for the progressive metamorphism of silica-deficient marbles. This falls in the cordierite-anthophyllite sub-facies, and equivalent assemblages in the staurolite-kyanite sub-facies should contain grossularite or forsterite. It is possible, therefore, that the composition of these rocks favoured more complete reaction to the late retrogressive metamorphism than was the case for rocks of other kinds described above. Calcareous assemblages are anyway notoriously unreliable facies indicators.

Hornblende Gneisses and Related Rocks (Table 10).

In the immediate hinterland of Ørnes a group of white or grey gneissic rocks, with bands of quartzo-felspathic material separated by thin layers of dark lustrous biotite and hornblende crystals, covers a wide outcrop. The layering of the dark minerals is very variable, and sometimes biotites and hornblendes

Table 10. Hornblende Gneisses and Related Rocks from the Galtskart Group

17 specimens examined.

The predominant mineral assemblage is quartz, plagioclase, microcline, biotite and hornblende. Calcite is present in 3 of the rocks, scapolite in 3, and diopside in 4. Small amounts of sphene, epidote (pistacite, clinozoizite or orthite) and apatite are widespread, while magnetite is sometimes present. Chlorite occurs in a few of the rocks.

Plagioclase Composition

H 67	An 41 from 6 determinations	H 102	An 38 from 8 determinations
H 97	An 37 from 4 determinations		Nz = 1.550) Corresponds to
H 98	An 45 from 5 determinations		Ny = 1.549) An 36
H 101	An 37 from 4 determinations	H 103	An 36 from 7 determinations
		H 317	An 36 from 2 determinations

Mineral Properties

H 98	Amphibole	2V = 78° optically negative Z c = 20°	
		Pleochroism	X = light green
			Y = dark yellow-green
			Z = dark green
	Biotite	Ny = 1.639	
		Pleochroism	X = very pale yellow
			Y = Z = very dark brown
H 249	Biotite	Ny = 1.625	
		Pleochroism	X = very pale yellow-brown
			Y = Z = very dark brown
H 317	Amphibole	2V = 80° optically negative. Z c = 11°	
		Pleochroism	X = light yellow-green
			Y = Z = dark grass-green

Table 10 (cont.)

Modal Analysis

	Quartz	Plagioclase	Microcline	Biotite	Muscovite	Hornblende	Diopside	Sphene	Epidote	Magnetite	Apatite	Calcite	Scapolite
H 98	24.1	53.0		12.0		8.7		.6	1.6				
H101	31.0	50.8		11.4	5.6					.6	.6		
H102	19.5	57.8		21.5	.8				.3		.2		
H147	28.2	6.2	4.7	.2		11.9	19.0	1.4				7.6	20.8
H281	20.8	25.2	10.0	.6		37.4	1.7	.6			.6	2.7	

Fig.96. Typical texture of the Ornes gneisses. The large crystal in the centre of the photograph is plagioclase and contains small inclusions of microcline. On the left of the photograph are two large hornblende porphyroblasts. x 30.

are found in small clots. When traced westwards the gneisses disappear into unexposed ground, and their projected equivalents along the "line of outcrop" consist of finely-banded green rocks, very similar to many of the calc-schists of the remainder of the region. Eastwards, the gneisses can be traced to Spildervand, but from then on occur merely as thin bands of calc-schists within pelitic schists (see also p. 175).

Thin bands of hornblende-rich gneiss (hornblende rock) are found spasmodically throughout the Galtskart succession.

(a) Main group of Ørnes Gneisses. Biotite, plagioclase and quartz make up the majority of the rocks, while dark-green hornblende is often present in small amounts. Ragged brown or green-brown biotites, either in layers or disoriented clots (Fig.96), are from less than 1 mm to 3 or 4 mm in length; associated crystals of pistacite occur as inclusions or as separate grains. Sometimes, the pistacite crystals contain a core of albanite.

Highly-coloured hornblendes are generally associated with biotite, and commonly contain inclusions of quartz and albanite. Discolouration around crystals of the latter is often observed. Optical measurements suggest that the hornblende crystals are rich in Ca-Mg rather than Na (Table 10).

Large crystals of plagioclase show combined twins, often with spindle-shaped lamellae, and showing evidence of deformation. Irregular zoning around inclusions, particularly of epidote, is common. In some of the rocks patches of microcline within plagioclase occur, apparently quite independent from the crystallographic orientation of the latter. In other cases, however, the microcline twin lamellae are parallel to the albite lamellae of the enclosing plagioclase. Separate crystals of microcline are also present. Sometimes development of true myrmekite is visible.

In some of the hornblende-free gneisses, late muscovite is present. Small granular crystals of pink garnet occur in some of the rocks, but only in very subordinate amount.

(b) Marginal Facies of the Gneisses and their Western Equivalents. These rocks are banded calc-schists which show exactly analagous relationships to the main group of calc-schists already described (p. 266). The hornblende is highly-coloured, while diopside is markedly green and probably iron-rich. Both microcline and scapolite are relatively later minerals.

(c) Easterly Equivalents of Gneisses. The principal difference from the main group of gneisses is the more granular nature of the quartz, which tends to occur in aggregates, separated by large crystals of plagioclase.

A thin layer of banded calc-schists occurs in the northern part of the Bugten succession, and close to the unexposed southern boundary of the main group of gneisses. The main difference from the calc-schists of group (b) described above is the greater abundance of calcite and scapolite, and the presence of large diopside porphyroblasts.

It appears, therefore, that the main group of gneissic rocks has a lenticular outcrop, with a skin of calc schists. Whether or not there is graduation between the central portions and the skin is unknown. Both to the E and W the gneisses are represented by banded calc schists of similar appearance (Fig.97).

Mineral assemblages in the main group of gneisses are equivalent to many of those present in the basal granites, and correspond to the kyanite-staurolite sub-facies. Again muscovite and hornblende are mutually exclusive. The average composition of the plagioclase is An 37, i.e. intermediate andesine, and together with the available modal analyses (Table 10) indicate approximation of the rocks to quartz diorites.

The origin of the gneisses is difficult to determine, as evidence of their original character is largely obscured by later metamorphism. However, thin lenticles of pelitic schists and marbles are present well within the gneisses, and in places are separated from the latter by gradational lithologies similar to those which separate the main gneissic outcrop from the surrounding schists. It is suggested that these gradational facies are due to sedimentary causes and that the gneisses have always been fundamentally different from the schists.

In the remainder of the Glomfjord region, the rocks most closely comparable to the Ørnes gneisses are in the Sokumvatn district (Rutland 1958). They consist of dioritic and monzonitic rocks with occasional metasedimentary intercalations. Rutland regards them as being meta-volcanics and has recognised occasional relict igneous textures in the plagioclases.

MAP SHOWING OUTCROP OF ÖRNES GNEISSES

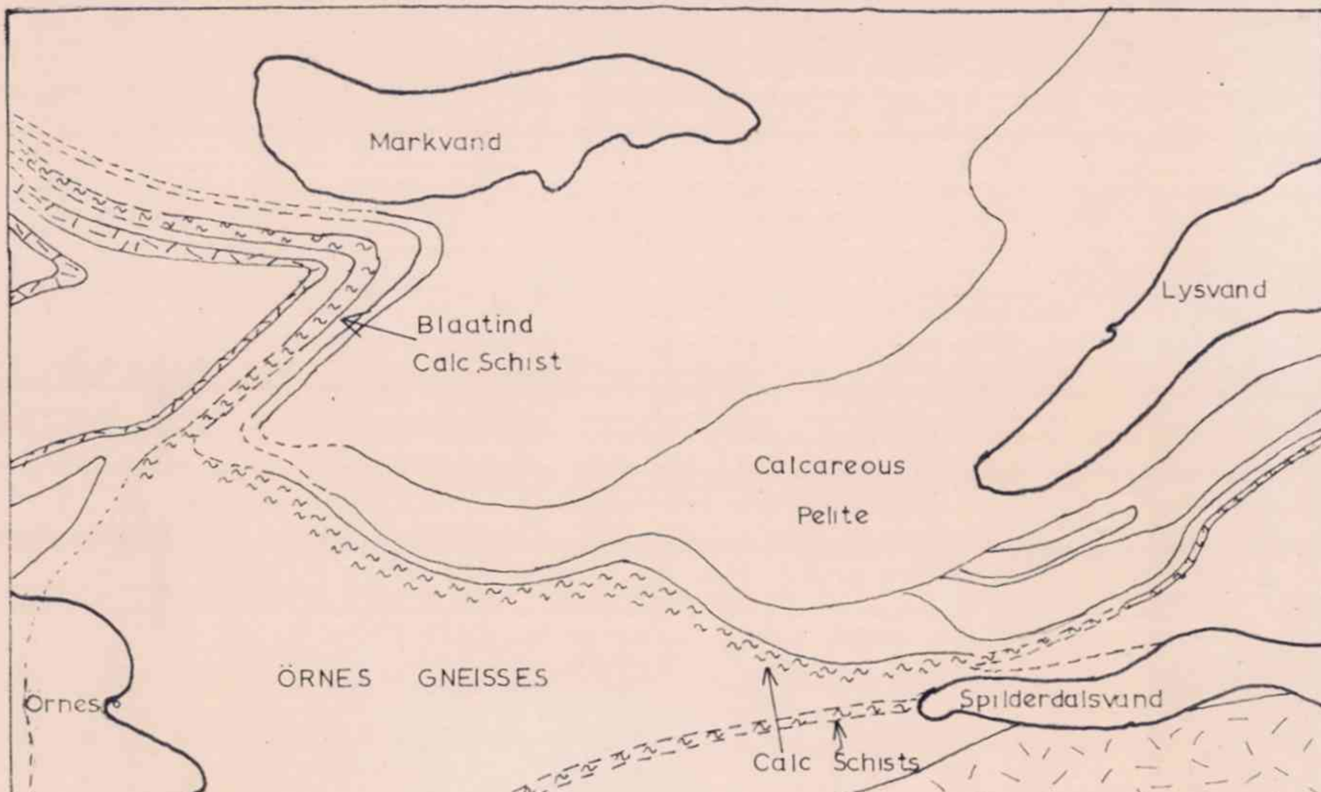


Fig 97

The crystallisation of the felspathic fraction of the Ürnes gneisses is demonstrably post-F_2, as indicated by the coarseness of grain size and the general lack of mineral cataclasis. However, the presence of early pegmatites within the gneisses, occupying isoclinal folds (Fig.78) indicates that the rocks were in existence before F_1 times. Their present composition is different from the surrounding metasediments, principally in the abundance of intermediate plagioclase. Moreover, the plagioclase in the metasediments is almost certainly of an earlier generation than that in the gneisses. This indicates the unlikelihood of the gneisses having developed by metasomatic introduction of felspathic material, and supports the view that they were originally distinct from the associated metasediments. Their present composition approximates to quartz diorites, and their field relationships suggest that they may represent a volcanic sequence, with occasional sedimentary intercalations.

Marbles

Most of the marble bands in the Galtskart succession consist of large interlocking crystals of calcite, often 4-5 mm in diameter. Occasional slight deformation of the twin lamellae is visible, and some recrystallisation is indicated by rims of untwinned material around crystals possessing bent twin lamellae. In general, the crystallisation of the marbles is late, probably post-F_3. Tests on occasional specimens using Lemberg's solution indicated that they were pure calcite marbles, and no examples of dolomitic rocks

Table 11. Marbles from the Galtskart Succession

5 specimens examined.

Some of the specimens are almost pure calcite marbles, with occasional crystals of quartz, phlogopite, tremolite, diopside or magnetite, but others contain considerable proportions of plagioclase, muscovite etc. and are very impure marbles.

Plagioclase composition

H 63 An 40 from 7 determinations
H 66 An 30 from 3 determinations

Table 12. Pegmatites from the Galtskart Succession

17 specimens examined.

Quartz and plagioclase are ubiquitous, while microcline is present in 10 of the rocks, muscovite in 13 and biotite in 9. Garnet is found in 3 of the rocks examined. Small amounts of magnetite, apatite and epidote are occasionally found. Tourmaline is abundant in one particular group of pegmatites.

Plagioclase composition

H 37 An 6 from 6 determinations
H 52 An 32 from 3 determinations
H 154 An 2 from 5 determinations
H 216 An 27 from 1 determination
H 268 An 35 from 2 determinations

have been recorded. Occasional crystals of phlogopite, tremolite, muscovite and iron ore form a small percentage of some of the rocks.

One of the rocks contains only 70% calcite, the remainder being composed mainly of quartz, plagioclase and muscovite. The reason for the absence of calcareous minerals such as diopside or grossularite is uncertain, but evidence that the assemblage is not one of equilibrium is afforded by the irregularity of extinction of plagioclase and the lateness of formation of muscovite.

Pegmatites

The variable modes of occurrence of pegmatites have already been discussed (p. 198): this is coupled with variation in mineral interrelationships. The pegmatites are generally of a relatively simple mineralogy.

A straight-sided cross-cutting pegmatite, from S. Suppevand, is composed mainly of large microcline porphyroblasts which are occasionally perthitic, with less common and smaller fractured and sericitised plagioclase crystals. A few myrmekitic patches are developed, and late muscovite is common. Poikiloblastic zoned tourmaline crystals up to 3 mm in diameter are present, and occasional ragged garnets also occur.

The thick layered pegmatite of S. Galtskart (Fig.30A) consists of large microcline, quartz and garnet porphyroblasts, separated by finely-twinned and slightly fractured plagioclase. Sericitisation of the plagioclase is intense and patches of the rock appear to have undergone granulation. A little late muscovite is present.

A series of cross-cutting pegmatites from the Ørnes gneisses consist principally of large crystals of felspar. Plagioclase is commonly slightly strained and in one of the rocks is perthitic (patch perthite). The twinning of the microcline is often poorly developed, and interaction between the two felspars is indicated by the development of myrmekite and the presence of sodic rims to some of the plagioclase crystals (Fig.98). Occasional clots of large, slightly bent, biotite flakes are present, while small amounts of late muscovite are common.

Several examples of pegmatite-schist contacts have been studied in thin section. In some, the contact is normal to the schistosity of the rock, and the pegmatite consists of granular quartz and felspar. In others, the pegmatite consists of large porphyroblastic microcline crystals 3-4 mm in diameter, separated by zones of granular quartz, microcline and clouded plagioclase. Sometimes, plagioclase forms larger, finely twinned, porphyroblasts, that are occasionally perthitic. Junctions with microcline often exhibit the development of myrmekite. Within certain of the granulated zones biotite of similar appearance to that in the schist is present, suggesting permeation and development by metasomatic processes. Late muscovites of small size are common. Junctions with the schists may be conformable or cross-cutting and are always sharp.

On the eastern margin of the Torsvik ultrabasic mass a felspathic pegmatite, cutting both the ultrabasic mass and the neighbouring schists is present. Large quartz crystals show an

Fig.98. Large finely-twinned plagioclase crystals with microcline inclusions, in a pegmatite, South Galtskart. The margins of the largest plagioclase crystal shows a good development of myrmekite (probably paramyrmekite). x 40.

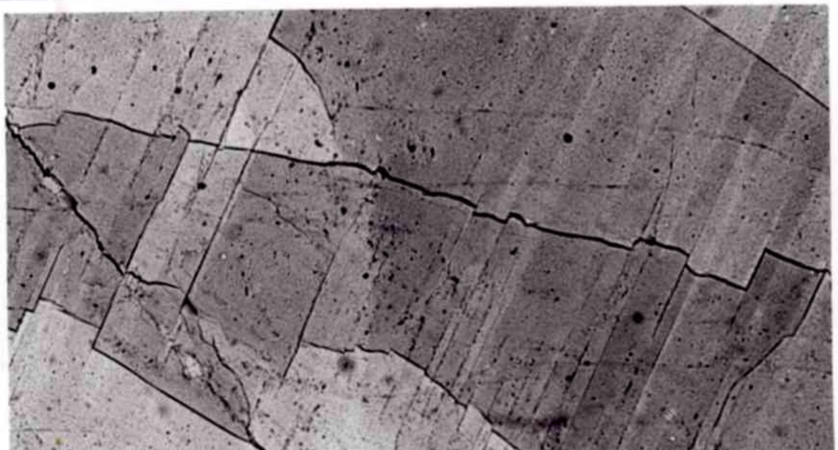


Fig.99. Development of parallel-sided fractures in a large, partially strained quartz crystal, from a pegmatite, North Torsvik. x 40.

unusual development of parallel-sided strain lamellae (Fig.99) and are separated by highly altered plagioclase crystals and large late muscovite crystals.

Certain generalisations as to the mineral parageneses may be made. Plagioclase is generally more highly altered and fractured than the microcline and is probably the earlier of the two felspars; interaction between the felspars, of a similar form to that seen in the basal granites, is spasmodically developed. Examples of both perthite and antiperthite are seen, and from the patch-like nature of the former it may be tentatively concluded that the felspars crystallised at a lower temperature than that prevalent in the staurolite-kyanite sub-facies. According to Rosenquist (1952) a series of separate types of perthite may be recognised according to the temperatures prevailing. Patch perthite corresponds to the green schist facies.

In most of the rocks the late muscovite has probably been derived from the felspars. The occurrence of coloured silicates and ore minerals is rare.

It is significant that in most of the pegmatites examined, the incidence of deformation is slight and often entirely lacking, indicating a general late, post-F_2 or post-F_3 crystallisation. This is borne out by the field relations in most cases.

Ultrabasic Rocks (Table 13)

A single section through the Torsvik ultrabasic mass is

Table 13. Ultrabasic Rocks

Mineral Assemblages

	Amphibole	Pyroxene	Olivine	Phlogopite	Chlorite	Carbonate	Muscovite	Epidote	Magnetite
H 104	X	X							X
H 105	X	X		X	X				X
H 106	X	X	X	X	X	X			X
H 110	X	X	X	X	X	X	X		X
H 182	X				X				X
H 269	X						X	X	X

Pegmatite from Ultrabasic

	Quartz	Plagioclase	Microcline	Biotite
H 107	X	X	X	X
H 108	X	X	X	

Mineral Properties

H 104 Colourless Amphibole $2V = 69, 74^\circ$ optically negative $Z\ c = 15, 18^\circ$
Component of Nx and $Ny = 1.630$
$Nz = 1.637$
Probably actinolite

Pyroxene $2V = 62^\circ$ optically negative
Possibly hypersthene

H 105 Amphiboles (a) colourless to very pale green $2V = 61^\circ, 63^\circ$ optically negative
$Z\ c = 12^\circ$
Possibly ferroedenite
(b) colourless $2V = 87, 85^\circ$ optically negative. $Z\ c=18^\circ$
tremolite

Pyroxene $2V = 44^\circ$ optically positive. Diallagic augite

Chlorite Colourless to very pale green $2V = 24^\circ$, optically positive. Possibly penninite

H 106 Amphibole (fibrous) $2V = 79, 81^\circ$ optically positive. $Z\ c = 0^\circ$
Probably anthophyllite

Olivine $2V = 84^\circ$ optically positive
Probably forsterite

H 110 Amphibole $2V = 89^\circ$ optically positive $Z\ c = 0$
Probably cummingtonite

accessible for examination along the coastal road NW of Ørnes. The marginal rocks consist of a light green schistose facies, while the main part of the mass consists of green or brown massive rocks, with no preferred schistosity. Large rosettes of amphiboles are often present, while in most of the rocks there is also relict olivine and pyroxene. Many variable secondary minerals also occur, notably chlorites and carbonate minerals.

The marginal facies is made up principally of small crystals of tremolite-actinolite (Fig.102, H 104) that show a good orientation. Occasional crystals of the same material are late and cut across the schistosity. A few large crystals of hypersthene appear to be in process of replacement by amphibole, but do not show evidence of deformation. Disseminated magnetite grains are common.

Within the ultrabasic mass three distinctive rock types are found, the relative abundance and interrelationships of which are unknown:

(a) H 105. Large equidimensional crystals of diallagic augite are partially replaced by tremolite-actinolite crystals of varied size (Fig.100). Patches of chlorite (probably penninite) appear to be later than the amphibole. Magnetite is concentrated in the regions away from the relict pyroxene, suggesting that it is a by-product of the conversion to amphibole.

(b) A dark green rock containing prominent amphibole rosettes is composed largely of rounded, irregular and undeformed olivine

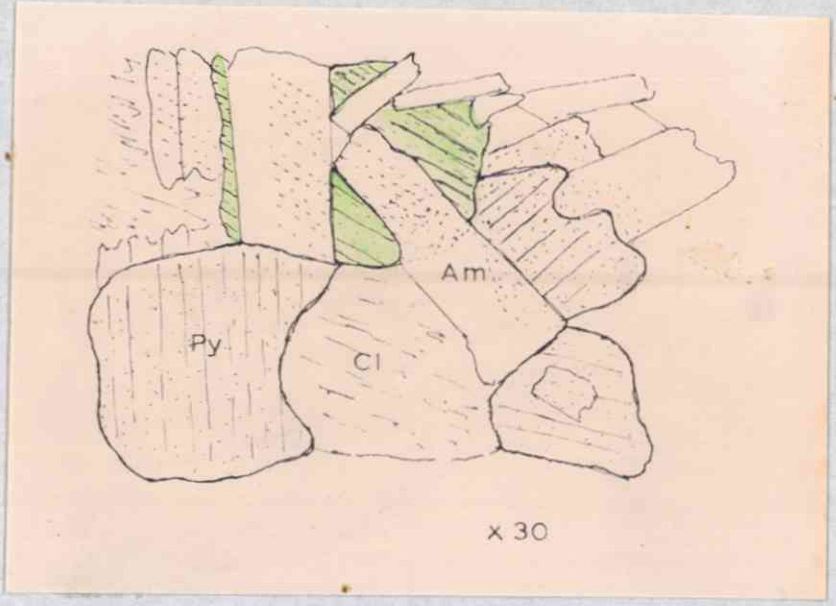


Fig.100. Textures in part of Torsvik ultrabasic mass. The pyroxene is probably diallagic augite, and contains trains of inclusions of magnetite arranged along the cleavage. These are preserved in the crystals of amphibole (tremolite-actinolite) which are clearly of later development. Py = pyroxene, Am = amphibole, Cl = Chlorite.

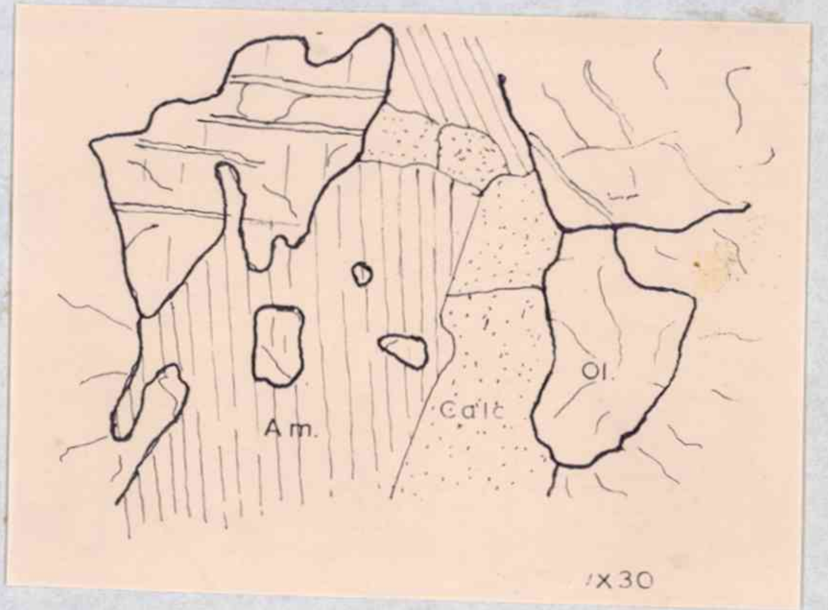


Fig.101. Large crystals of olivine veined by light green secondary serpentine, and replaced by amphibole. The calcite is probably of still later development. Ol = olivine, Am = amphibole, Calc = calcite.

Fig.102. A single large relict pyroxene crystal in a tremolite schist, the marginal facies to the ultrabasic mass of Torsvik. x 30.

crystals (forsterite) traversed by cracks filled with secondary carbonates, and partially replaced by amphibole (probably anthophyllite) (H 106). The carbonate is recrystallised into large crystals in association with the amphibole (Fig.101). Small amounts of pyroxene on the margins of the olivine crystals probably developed before the anthophyllite. Colourless chlorite, again a probable alteration product of olivine, occurs as inclusions within the latter and is associated with amphibole. Granular crystals of iron ore in olivine are widespread.

(c) This is similar in appearance to (b) and probably represents a more complete alteration product of the latter. Rounded and corroded crystals of olivine (forsterite), pyroxene, amphibole (cummingtonite), phlogopite and carbonate occur in a fine-grained groundmass, a large proportion of which consists of poikiloblastic carbonate crystals (H 107). Near the single large olivine crystal, which is about 4 mm in diameter, the matrix is more coarsely crystalline, consisting principally of carbonate and amphibole. Some of the phlogopites are slightly bent.

Towards the NW margin of the ultrabasic body is a dark green layer of amphibolite, consisting of small crystals of highly coloured hornblende and pistacite, with a little iron ore and plagioclase. Concentration of hornblende in one part of the slide (H 182 A) causes layering of the rock. Plagioclase is rare and forms small poikiloblastic crystals.

Cutting the ultrabasic mass is a composite dyke whose field relations are described elsewhere (p. 206). A section of the fine-grained part of the complex consists of granular quartz and plagioclase crystals with abundant small green biotites, which show no preferred orientation. The occasional felspathic patches, visible in the field, are composed of plagioclase, while their margins contain small crystals of microcline. Myrmekite patches on the boundaries of the two felspars are common. The coarse-grained felspathic pegmatite is made of large crystals of finely twinned plagioclase, and strained quartz, with subordinate small microcline crystals. Interaction between the two felspars, with the development of sodic rims to the plagioclase,is well displayed , and it appears that microcline is the later felspar. A little late muscovite is developed within the plagioclase. Lack of cataclasis in the pegmatites indicates a late - probably post-F_3 - formation. Reasons for the existence of such an acidic pegmatite dyke within an ultrabasic body are unknown, although an external source is clearly indicated. The pronounced differentiation (see Fig.75) is also due to unknown causes.

Lack of cataclasis of minerals throughout the ultrabasic mass, with the preservation of primary undeformed olivine crystals up to 1 cm in diameter suggest that the emplacement of the mass occurred late in the tectonic history of the area (probably post-F_2). This correlates with the incompleteness of the metamorphism and the survival of primary minerals, indicating the probability that regional metamorphism of the surrounding rocks, which reached a peak during or immediately after, F_2 times, occurred before the injection of the ultrabasic rocks. The development of the non-aluminous amphiboles and chlorite, minerals of general lower grade, and common in the epidote-amphibolite facies, may then be correlated with the later stages of metamorphism of the surrounding rocks. Explanations for the occurrence of aluminous hornblendes in the thin layer of amphibolite described above, a rock which has clearly undergone complete metamorphism to the kyanite-staurolite sub-facies, are not clear. It may represent a band that was already in existence before emplacement of the ultrabasic mass. The marginal facies, composed principally of tremolite schists, appear to have undergone more complete metamorphism than the internal portions of the mass.

Developed along many of the joint faces of the ultrabasic mass are large and disoriented crystals of talc, which represent a hydrothermal alteration product of the amphiboles.

The mineral assemblage present in the ultrabasic mass indicates that originally it consisted almost entirely of augite and olivine, with probably a little iron ore, but with no plagioclase. It was therefore composed of pyroxenites and peridotites in its unmetamorphosed form. These bodies are known to be intruded at relatively low temperatures, which would account for the lack of contact metamorphism in the surrounding schists. Their association with orogenic belts is widespread, and Hess (1938) envisages formation of a peridotite magma by the localised impact of a thickened granitic layer on a peridotite layer within orogenic belts. Experimental evidence (Bowen and Tuttle, 1949) indicates the impossibility of generation of a magma of this type at less than 1000°C. Bowen (1928) suggests intrusion of olivine crystals which originally separated from a basaltic magma, the intrusion aided by magmatic liquid or even water vapour. The associated basic magmas, so commonly seen with peridotites elsewhere (e.g. Otago, New Zealand) may be represented by amphibolites in the Ornes region. It is noteworthy that the emplacement of the ultrabasic mass of the Ornes region probably occurred at a late stage in the orogenic history, while the production of basic lavas and peridotites is usually associated with the earliest stages of folding.

Chlorite-bearing Schists

A slight amount of chloritisation of biotites is commonly seen in the rocks of the whole region, but occasionally complete chloritisation is observed. These chlorite-bearing schists are restricted to an indefinite band that follows the southern boundary of the S. Scetertind calcareous pelitic schist. Only a few of the rocks in this zone show this marked development of chlorite, but such a restriction is significant and may be structurally controlled. The nearest large-scale structure is the Lysvand slide, believed to be some 500 metres N of the band.

The affected rocks cover a range of lithologies, including semi-pelitic and pelitic schists and amphibolites. In some, original biotite is still present, but in others complete chloritisation has occurred, and the chlorite is sometimes present in late fibrous masses. The remainder of the rocks are normally equivalent to their non chlorite-bearing counterparts. However, in three of the rocks, a large amount of granulation has been operative, and a very high percentage of magnetite is also present. The latter does not appear to be a consequence of chloritisation.

It therefore appears that there is a regional control, possibly caused by localised retrogression associated with late movements within the slide zone.

4. North Markvand Group

Petrographically, this group consists of interbanded pelitic and semi-pelitic schists, with occasional thin bands of marbles and amphibolites. An important difference from the Galtskart group is the virtual absence of calcareous pelitic schists and calc-schists, and the presence of sheet granites in the northern parts of the outcrop.

Pelitic Schists

The petrographic relationships that are seen in the Galtskart pelitic schists are also present in these rocks. Both garnetiferous and non-garnetiferous pelitic schists are present, and the forms adopted by the garnet porphyroblasts are similar to those in the Galtskart and Steffodalen successions (cf. Fig.86). Kyanite,

Table 14. North Markvand Group

(1) Pelitic Schists. 15 specimens examined

The predominant mineral assemblage is quartz, plagioclase, biotite, muscovite, garnet (12 rocks) and kyanite (7 rocks). Sillimanite is present in three of the rocks, and chlorite in 4. Small amounts of magnetite and apatite are common and occasional rocks contain tourmaline.

Plagioclase composition

H 335 An 30 from 2 determinations
H 357 An 27 from 2 determinations
H 358 An 24 from 3 determinations

(2) Semi-pelitic Schists. 10 specimens examined

The predominant assemblage is quartz, plagioclase, biotite and muscovite. Microcline is present in 4 of the rocks, and does not occur in those rocks that contain garnet (5 rocks). Magnetite and apatite are ubiquitous in small amount, and a little chlorite and sphene are sometimes present.

Plagioclase composition

H 288 An 28 from 3 determinations
H 299 An 30 from 3 determinations
H 328 An 32 from 2 determinations

Modal Analysis

	Quartz	Plagio-clase	Micro-cline	Bio-tite	Musco-vite	Apa-tite	Magne-tite	Total
H 299	47.7	26.3		2.5	23.2		.3	100.0
H 300	43.7	30.5	9.6	13.7	2.3	.2		100.0

(3) Calcareous pelitic schists. 2 specimens examined

The assemblages present are the same as in equivalent rocks from the Galtskart succession (Table 8, p. 259).

Table 14 (cont.)

Plagioclase Composition

H 369 An 37 from 5 determinations

Modal Analysis

	Quartz	Plagio-clase	Bio-tite	Diop-side	Horn-blende	Scap-olite	Magne-tite	Total
H 369	40.5	24.5	20.8	7.4	.9	3.6	.6	100.0

(4) Calc Schists. 2 specimens examined

Mineral assemblages are the same as in equivalent rocks of the Galtskart succession (Table 9, p. 267).

Mineral properties

H 199 Amphibole $2V = 62^{\circ}$, optically negative Z c = 21°
Pleochroism: X = very pale green
Y = patchy yellow-green
Z = grass green

(5) Amphibolites

(a) Sedimentary amphibolites. 2 specimens examined

The mineral assemblages are the same as in equivalent rocks of the Galtskart succession (Table 9, p. 267).

Plagioclase Composition

H 284 An 38 from 3 determinations

(b) Igneous amphibolites. 5 specimens examined.

The mineral assemblages are the same as in equivalent rocks of the Galtskart succession (Table 9, p. 268).

Mineral properties

H 278 Amphibole $2V = 71, 67^{\circ}$ optically negative Z c = 12°
Pleochroism X = very light green
Y = yellow green
Z = dark green

predominantly as small, ragged crystals, but occasionally as late-formed cross-cutting crystals of larger size, is widely but sparsely developed. The mineral appears to be more common than in the Galtskart rocks. A few of the rocks show a variable but usually slight development of late fibrolite from biotite. In one example the fibrolite now occupies an F_2 microfold that was originally defined by a biotite crystal.

The main interest of the group is the widespread development of micro-foldings (see also p. 131). Unfortunately, only a few examples of these highly contorted schists have been studied. In some examples (e.g. H 298) thin micaceous bands, in which the majority of the crystals are unbent, are separated by zones of granular quartz, the whole being deformed into a series of symmetrical similar folds. The folds are regarded as being of F_2 age, indicating a considerable amount of post-F_2 crystallisation. Another type contains poikiloblastic kyanite crystals growing across the micro-folds, indicating a post-F_2 stage of crystallisation.

Semi-pelitic Schists

These are considerably less abundant than the pelitic schists, and generally occur as indistinct bands within the latter. Either biotite or muscovite may predominate, and the micas are usually arranged in bands or clots. Microcline, present as small granular crystals, is found only in those rocks where biotite predominates. The quartz-plagioclase fraction of the rocks shows similar relationships to their equivalents in the Galtskart

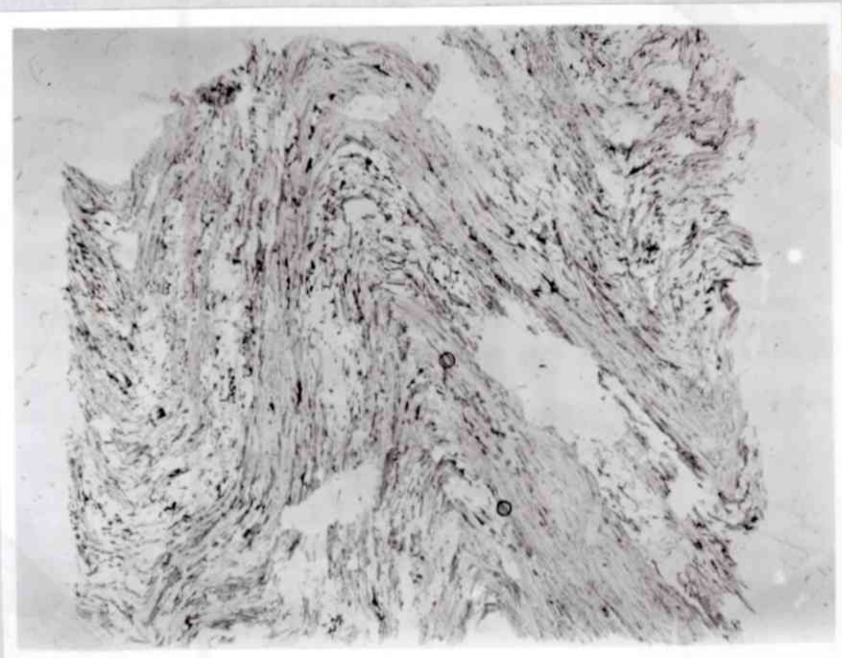


Fig.103. F_2 microfold in mica-schist, South Spantind. The fold is a shear structure and analogous to the fracture-cleavage folds seen throughout the North Markvand pelitic series. In general, the micas are unbent. x 5.

succession. A few rocks, mainly in the biotite-rich group, contain small and ragged garnets, and there is mutual exclusion of garnet and microcline. It is evident that the biotite-rich group is allied to the pelitic schists,and indeed in certain circumstances it is difficult to differentiate specimens from garnet-free pelitic schists. The group in which muscovite is dominant has a lower content of total mica, and more nearly approaches pure sandstone in composition.

In all important respects the group is similar to its counterpart in the Galtskart succession.

Calcareous Pelitic Schists

Thin impersistent bands of calcareous pelitic schists are found, particularly in the rocks to the NW of Markvand. They are somewhat more flaggy than their equivalents in the Galtskart succession, but the mineral relationships are identical. Small biotites, sometimes affected by F_2 microfolds, define an approximate schistosity, across which pyroxene crystals up to 2 or 3 mm in diameter have developed. Secondary hornblende is present in small amount, as an alteration product of diopside, while late scapolite and calcite is also found. Granular quartz and plagioclase, the latter showing a variable amount of fracturing,occur, and occasional coarsely crystalline quartz veins traverse the rocks.

Calc Schists

These are uncommon in the N.Markvand group, and only two rocks, both associated with marbles, were identified. One of these (H 199) consists of large pale green diopside crystals up to 2 mm in diameter, partially replaced by calcite, scapolite and hornblende. Late quartz and scapolite occur in irregular veins, and a little biotite, in small bent crystals, is also present. The other rock (H 294) consists of patches of diopside, hornblende and biotite, separated by quartzo-felspathic material. Diopside and hornblende form sieve-like porphyroblasts up to 4 mm long, and sometimes the diopside crystals are partially altered to hornblende along their cleavage directions. Biotite forms irregular, corroded crystals of an earlier age than the diopside and hornblende. Microcline is the dominant felspar, and there is an abundant development of true myrmekite, often forming complete crystals, on almost every plagioclase junction. The earlier plagioclase occurs as deformed crystals, sometimes possessing spindle-shaped twin lamellae.

Amphibolites

Thin bands of amphibolite within the pelitic schists are a common feature of the N.Markvand rocks. Those of presumed igneous origin are the most common.

(i) Igneous Amphibolites. Highly-coloured hornblende forms between 50% and 90% of the rocks, and varies from small, ragged, roughly oriented crystals to large poikiloblastic crystals up to 2 mm across. Inclusions of the other minerals are common. Biotite is very variable in quantity, and forms small, red-brown, bent flakes. Both biotites and hornblende are sometimes chloritised. Plagioclase often shows complicated twinning (Fig.104) with spindle-shaped

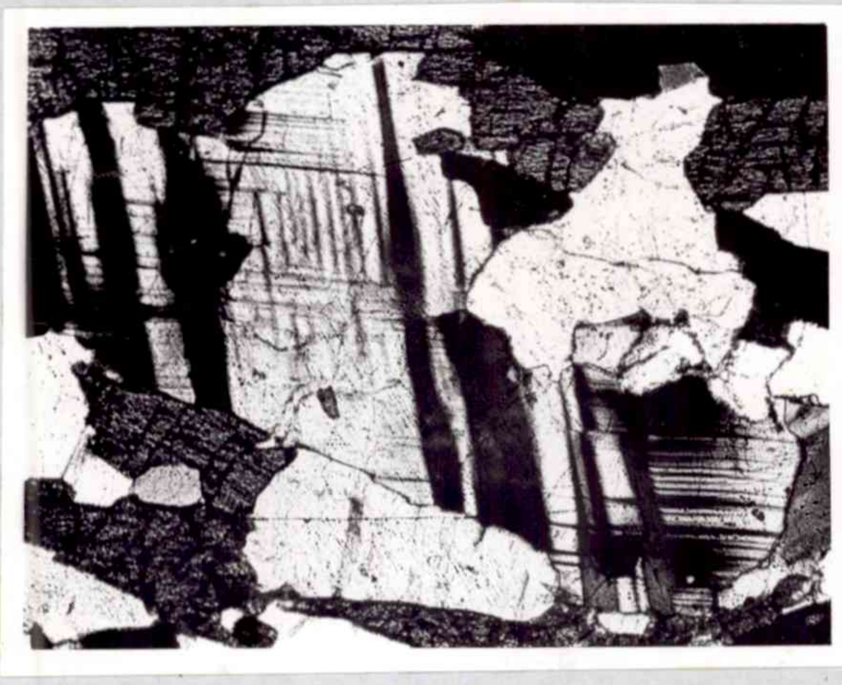


Fig.104. Irregular albite-pincline twinning of plagioclase crystals from an amphibolite, S.W. Breitind, showing a good development of spindle-shaped twin-lamellae. x 40.

lamellae, and zoning which is dependent upon proximity of other calcareous minerals. Quartz is always subordinate in amount, in small granular crystals. A little late calcite is present in some of the rocks.

The mineral assemblages, and their relative proportions suggest that the rocks have an igneous parentage, and they are similar to the equivalent rocks in the Galtskart succession.

(ii) Sedimentary amphibolites. The two presumed sedimentary amphibolites identified are both closely associated with marble bands. One of them (H 284) consists of ragged biotites and hornblendes, with occasional large poikiloblastic diopside crystals. Granular quartz and plagioclase is also present. In the other rock (H 330) hornblende is dominant, with subordinate small, rounded diopside crystals. Relict F_1 folds are shown by the arrangement of bent and fractured biotites.

Marbles

In contrast with the pure calcite marbles of the Galtskart succession, those in the N. Markvand group generally contain a high proportion of phlogopite, tremolite etc., and are usually markedly brown in colour. None of the rocks has been studied in detail.

Pegmatites

Felspathic pegmatites are entirely lacking from the N. Markvand rocks, probably due to the distance from the basal

Glomfjord and Bjellåtind granites.

Sheet Granites

These bodies, present in the northerly parts of the outcrop of the group, are highly banded with a 'sedimentary' appearance (Figs 50, 51). A series of specimens taken across the granite of N.E. Storvikvand (a section no longer exposed, due to recent roadworks) shows that two varieties can be distinguished, within which considerable variation exists:- (a) grey, well-foliated granites, often with thin pink felspathic layers, (b) pink, more homogeneous and less well-banded granites. Modal analyses of the granites, together with the chemical composition of one of the rocks, are shown in Table 5.

(a) Grey granites: small, green-brown biotites are generally evenly scattered, and are sometimes partially chloritised. Late muscovite is only occasionally present. The quartzo-felspathic fraction of the rocks is similar to that in the basal granites (p.) except for the generally more granular nature in the sheet granites. Occasional large microcline crystals up to 3 mm in diameter are seen. No myrmekite has been recorded. Clouding of the felspars is restricted to the rocks in which muscovite is found.

(b) Pink granites: the principal difference from the grey variety is the occurrence of biotite, with associated sphene and magnetite in irregular clots or layers. In addition, the quartzo-felspathic content tends to be more coarse-grained. The general appearance of this group of rocks is closer to the Bjellåtind granite than the grey varieties.

Table 15. Granites from the North Markvand Group

Mineral Assemblages.

	Quartz	Plagioclase	Microcline	Biotite	Chlorite	Muscovite	Magnetite	Epidote	Apatite	Sphene	Tourmaline
H 343	X	X	X	X		X	X	X	X		
H 344	X	X	X	X			X		X	X	X
H 345	X	X	X	X			X	X	X	X	
H 346	X	X	X	X		X	X	X		X	
H 347	X	X	X	X	X		X		X	X	
H 348	X	X	X	X			X		X	X	X
H 354	X	X	X	X	X	X	X	X		X	
H 408	X	X	X	X		X	X	X	X		
H 411	X	X	X	X		X	X	X		X	

Biotite schist within Finneskua granite

	Quartz	Plagioclase	Microcline	Biotite	Chlorite	Muscovite	Magnetite	Epidote	Apatite	Sphene	Tourmaline
H 409	X	X	X	X		X	X		X		

Plagioclase Determinations

H 343 An 7 from 3 determinations
H 344 An 5 from 3 determinations
H 345 An 5 from 3 determinations
H 346 An 10 from 3 determinations
H 347 An 5 from 3 determinations
H 348 An 10 from 3 determinations
H 409 An 16 from 3 determinations

Modal Analysis

	Quartz	Plagioclase	Microcline	Biotite	Muscovite	Chlorite	Sphene	Apatite	Magnetite	Epidote	Total
H 343	23.4	40.5	26.8	4.8	3.7	.3			.5		100.0
H 344	27.6	28.3	35.6	7.2			1.0	.3			100.0
H 345	18.3	35.0	39.8	5.3			.7		.9		100.0
H 346	8.6	32.5	48.0	9.9				.2	.5	.3	100.0
H 347	35.1	31.8	28.9	3.6			.5		.1		100.0
H 348	27.8	32.0	35.8	3.7	.2				.5		100.0
H 354	2.7	58.3	36.2	1.9					.9		100.0
H 408	25.5	24.8	43.7	4.6	.7				.7		100.0

Table 15 (cont.)

Chemical Composition (determined from modal analysis)

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O
H 343	71.4	16.0	.1	.2	1.5	.9	4.2	5.6	.3
H 344	75.6	13.7	.2	.2	2.1	.6	3.0	6.5	.1
H 345	68.5	15.6	1.1	.7	1.5	.7	3.7	7.0	.1
H 346	65.1	17.4	.9	.6	2.8	.7	3.4	8.9	.2
H 347	75.8	12.7	.1	.1	1.0	.6	3.3	5.1	.1
H 348	73.3	14.0	.8	.4	1.0	.7	3.4	6.9	.1
Average	71.6	14.9	.5	.4	1.6	.7	3.5	6.7	.1

Chemical Composition (by rapid analytical methods)

H 408		
	SiO_2	70.45
	Al_2O_3	14.89
	Fe_2O_3	1.33
	FeO	0.96
	MgO	0.33
	CaO	0.34
	Na_2O	3.66
	K_2O	6.95
	TiO_2	0.38
	P_2O_5	0.07
	MnO	0.02
	Total	99.38

Fig.105. A sheet granite from N.E. Storvikvand, showing development of microcline porphyroblasts, and partially fractured plagioclase. x 30.

The projected equivalent of the Storvikvand granite on Degro is comparable with the pink variety of the Storvikvand granite.

Fine-grained and highly banded pink granites fom the majority of the sheets on the Finneskua peninsula. The noteworthy differences from the Storvikvand granite are the highly banded nature and the common occurrence of muscovite. Small green-brown biotites are generally arranged in bands.

A biotite schist within one of the Finneskua sheet granites is different from those of the basal granites in the absence of hornblende. It consists of clots of green-brown biotite flakes several mm long, separated by plagioclase and quartz. A little late muscovite is also present. All the minerals except muscovite show some evidence of fracturing.

The principal differences of the sheet granites compared with the basal granites are the smaller size of all the component minerals of the former, the common occurrence of biotite as evenly scattered crystals, and the highly banded character and remarkably persistent nature of the layering. The rocks are regarded as representing part of the sedimentary sequence and were probably arkosic sediments before metamorphism. An increase in felspathic content would lead to convergence towards the basal granites, and it is suggested that the parent material to the latter was similar to the sheet granites in composition.

The positions on the ternary diagram quartz-orthoclase-albite of the three analysed granites from the Ørnes region are shown in Fig.107. Also included as a contoured zone are analyses of granitic rocks from all parts of the world (quoted from Tuttle and Bowen, 1958). Although the analysis of the Skjeggen granite falls in the centre of this zone, those from the Bjellåtind and Finneskua granites are some distance away, indicating the probability of considerable chemical variation of the granites as a whole. This is confirmatory evidence for regarding the granites as representing a heterogeneous felspathised sedimentary sequence. Analyses of some of the typical 'silty' schists, the possible unfelspathised equivalents of the granites, will be available in the near future, and it will be interesting to determine the relationships on the ternary diagram between them and the various granitic bodies.

5. North Skjeggen Group (Table 16)

Due to the rapidity with which the area had to be mapped, few specimens were collected. The principal aim was comparison with equivalent rocks from the Galtskart Group, the probable stratigraphic equivalents. In every case, such a comparison was found to exist, and the rocks appear to have undergone an identical metamorphic history.

Table 16. North Skjeggen Group

Only very occasional specimens were collected and sectioned from this group, but throughout they show analogous characteristics to their counterparts in the Galtskart succession.

Amphibolites

Plagioclase composition

H 273 An 56 from 3 determinations

H 430 An 65 from 3 determinations

Mineral properties

H 273 Amphibole $2V = 66, 69^{\circ}$ optically negative $Z\ c = 18^{\circ}, 14^{\circ}$

Pleochroism: X = nearly colourless

Y = yellow-green

Z = grass green

Felspar Determination and Characteristics

A considerable number of determinations of plagioclase compositions have been completed for most of the rock groups present in the Ørnes region (tables 1-15). Universal stage methods described by Rittmann, Turner (1947) and Smith (1958) have been used with varying degrees of success. In each of the methods, the accuracy of determination of An content appears to be between 5 and 10%.

(1) Rittmann Method. The main advantage of this method is the rapidity with which measurements may be made. Albite or pericline twins must be identified, and the individual crystals must show cleavage. Most of the twins present in the rocks of the Ørnes Region are albite or pericline, or combinations of the two, but frequently a cleavage direction cannot be identified. The method is particularly useful in crystals from the granitic rocks, where the separate lamellae are often very narrow, and prohibit measurement by Turner's method. Rutland (1958) remarks that the presence of vicinal composition faces in albite twins causes a complex twin reaction on identification of the twin law and, like Barber (1936), prefers the use of Turner's method. Experience from the Ørnes region, however, shows that for these rocks at least, even if the twinned crystal does yield a complex twin law, the composition, as determined by the Rittmann method, corresponds closely to that given by Turner's method. The graph of extinction angle with composition is linear, provided the inclination of the

composition plane to the cleavage can be recognised. The greatest disadvantage of the method is that it does not provide any reliable information on twins other than albite or pericline.

(2) Turner's Method. This, theoretically, is the best method available for the determination of the properties of plagioclase by Universal stage measurement. Not only is the composition of the crystal determined but in addition the twin law can be identified with certainty. In metamorphic plagioclases, however, there are many practical difficulties to be overcome. Commonly, the crystals are small, and the twin lamellae so fine as to make measurements on them difficult or impossible. Sometimes, the triangle of error, produced by the intersection of XX' etc. is so large that it is impossible to be certain where it is in relation to the composition plane. In such cases, it is clear that identification of the twin law cannot be made with certainty. On one occasion, two separate determinations were performed on the same crystal, and although individual measurements were not displaced by more than 2-3°, one of the determinations indicated a complex twin, and the other a simple twin. An added disadvantage is that in acid plagioclases, X and X' fall extremely close to one another, making an exact determination of the great circle joining them impossible.

In addition to this difficulty of correct identification of the twin law, is the fact that the curves giving content of An of the plagioclase crystals, are not always linear, and over the range An 0 to An 30 there may be two possible results. However, for basic plagioclases, with wide lamellae, the method is admirable, and has been used with high degrees of accuracy on plagioclase crystals from amphibolites and calcareous gneisses. In no case have twins other than albite or pericline or combinations of the two been identified in the Ørnes Region. Occasional dubious complex twins have been found, but in no case were the measurements of the required degree of accuracy to be entirely certain.

(3) Smith's Method (1958). The method is only applicable to pericline twins, and yields the composition of the plagioclase by relating X, Y and Z to the inclination of the pericline composition plane. Slight differences exist in the curves for high and low temperature forms, and although the majority of determinations from rocks of the Ørnes region fell near the low temperature curve, the scatter was too great to permit any conclusions being drawn. The main advantage of the method is the linear nature of the curves giving composition.

Plagioclase characteristics

Certain generalities about the twinning of the plagioclase crystals can be made. In the granites there is a large preponderance of albite twins, and indeed in the sections available of the Bjellåtind granite at the E of Lysvand, pericline twins are entirely absent. As in the plagioclase crystals from many other areas, those from the Ørnes region show a general increase in coarseness of the twin lamellae with increase in An content.

Combined albite-pericline twins and pericline twins are common throughout the metasediments. The view of Köhler (1941) that such combined twins develop by adjustment of the crystal lattice on inversion from high to low temperature form, cannot be demonstrated from these rocks. It is clear that the plagioclase crystals in the metasediments have developed by normal metamorphic processes. Rutland (1958) notes that the plagioclase crystals from the gneissic rocks of the Sokumvand region show alternate broad and narrow twin lamellae, while those from the schists have twin lamellae of equal width. He correlates the difference as being due to variation in mode of development. The gneissic plagioclase crystals he regards as forming by inversion from an originally high temperature structural state, while those from the schists must have developed by progressive strain under regional metamorphism. A similar relationship exists in the Ørnes region where the plagioclase crystals from the hornblendic gneisses of Ørnes, and from the presumed igneous amphibolites show alternate broad and narrow lamellae (fig.96) in marked contrast to the schists. In the case of the Ørnes gneisses there is no further direct evidence for regarding the rocks as having been volcanic (but see also p. 284).

Plagioclase-microcline inter-relationships are described in the section on basal granites (p. 215), and are not dealt with here.

Fig.106. Felspar characteristics.

A. Deformation of plagioclase twin lamellae by ? F_3 movements, in a tourmaline pegmatite, South Suppevand.

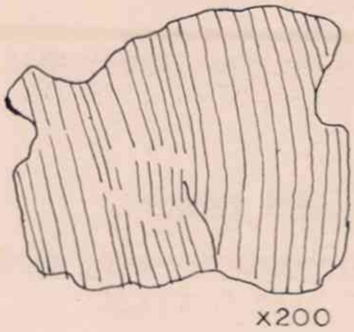
B. Perthite in Bjellåtind granite.

C. Complicated albite-pericline twin showing extreme irregularity of twin lamellae.

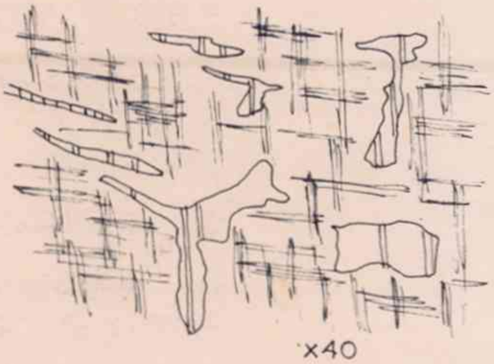
D. Combined albite pericline twin in silty schist, South Lysvand, showing deformed wedge-shaped twin lamellae.

E. Plagioclase-microcline relationships in a sheet granite from N.W. Storvikvand. The plagioclase shows deformed twin lamellae, while the microcline is undeformed and probably of later crystallisation. A little myrmekite is present.

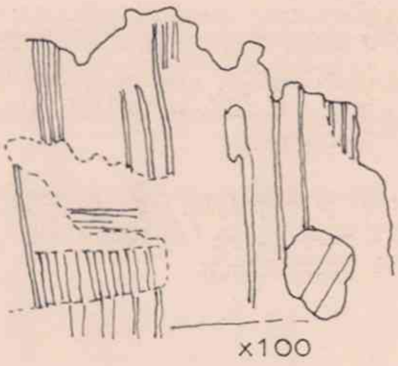
F. A single fractured crystal of plagioclase replaced by undeformed microcline in a biotite schist within a sheet granite, North Blaatind.



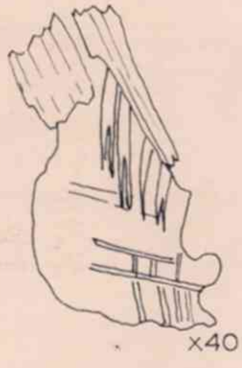
A



B

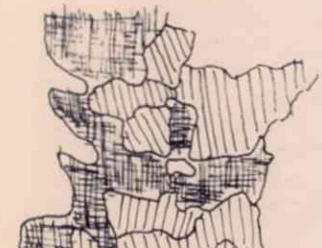


C



D

E



F

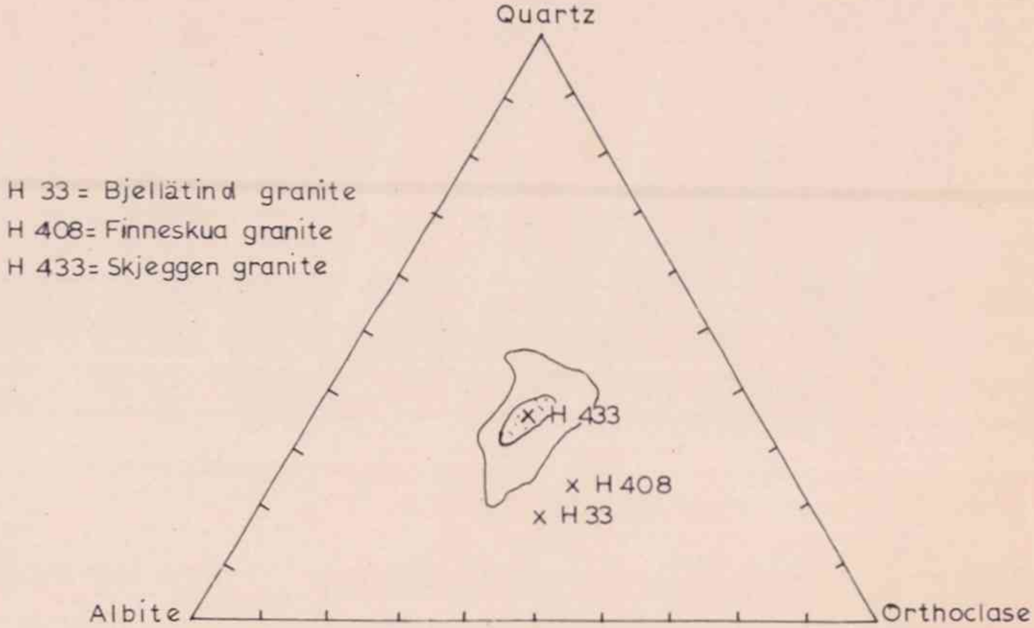


Fig. 107. Ternary diagram for quartz-orthoclase-albite showing analyses of three of the granites of the Örnes region. Contoured zone represents analysed plutonic granites quoted in Washington's tables (Tuttle & Bowen 1958) Contours at 2 & 6%

Diagrams and photographs showing some of the more interesting plagioclase characteristics are shown in fig.106. Deformation of the twinned crystals is observed in almost every rock. Emmons (1953) regards twinning to be a consequence of strain. Wedge-shaped twin lamellae are sometimes visible (e.g. figs 94, 104), and in certain circumstances appear to be related to zones of deformation within the crystal. The presence of such wedge-shaped lamellae clearly demands the occurrence of vicinal faces, and although Barber (1936) regards vicinal faces as being indicative of primary twins, it appears that in some of the rocks from the Ørnes region, they are to be regarded as secondary, and a consequence of strain.

Bibliography

Ackermann, K.J. 1960 Mineral paragenesis of the metasediments invaded by granite in the area around Sokumvatn, Gildeskål, Norway. Ph.D. thesis, University of London.

Ackermann, K.J., Nicholson, R. and Walton, B.J. 1960. Mineral development and deformation in metasedimentary rocks in the Glomfjord region, Northern Norway. Int.Geol.Congr. 21st session, Copenhagen.

Alling, H.L. 1921 The mineralogy of the felspars. Jour. of Geol. 29, p.193.

Barber, 1936 The Tertiary igneous rocks of the Pakokku District, Burma, with special reference to determinations of the felspars by the Federoff method. Geol.Surv.India Mem.68, pt.II, p.221.

Barth, T. 1939 The progressive metamorphism of the Sparagmite of Southern Norway. Nors.Geol. tids. 18, p.54.

Billings, M.P. 1942 Structural Geology, New York.

Bott, M.P.H. 1956 A Geophysical Study of the Granite Problem. Quart.Jour.Geol.Soc.Lond. 112, p.45.

Cheng Yu Chi 1944 The migmatite area around Bettyhill, Sutherland. Quart.Jour.Geol.Soc.Lond. 99, p.107.

Crowder, D.F. 1959 Granitisation, Migmitisation and Fusion in the Northern Entiat Mountains, Washington. Bull.Geol.Soc.Am. 70, p.827.

Davidson, C.F. 1943 The Archean rocks of the Rodil district, South Harris, Outer Hebrides. Trans.Roy. Soc.Edin.61, p.71.

Edelman, N. 1949 Microcline porphyroblasts with myrmekite rims. Bull.Comm.Geol.Finlande 80, p.73.

Emmons et al. 1953 Selected petrogenetic relations of plagioclase felspars. Geol.Soc.Am. Mem.52.

Eskola, P. 1948 The nature of metasomatism in the processes of granitisation. 18th Int.Geol.Congr.London.

1949 The problem of mantled gneiss domes. Quart. Jour.Geol.Soc.Lond. 104, p.461.

Fairbairn, H.W. 1949 Structural Petrology of deformed rocks. Cambridge, Mass.U.S.A.

Fyfe, W.S., Turner, F.J. and Verhogen, J. 1958. Metamorphic reactions and metamorphic facies. Geol.Soc. Am. Mem.73.

Gorai, M. 1951 Petrologic studies on plagioclase twins. Am. Miner. 36, p.884.

Grout, F.F. Origin of granite. Geol.Soc.Am. Mem.28, p.45.

Harker, R.I. 1954 The occurrence of orthoclase and microcline in the granitic gneisses of the Charn-Inchbae Complex, E.Ross-shire. Geol.Mag. 41, p.129.

Harpum 1954 The formation of epidote in Tanganyika. Bull.Geol.Soc.Am. 65, p.1075.

Henderson 1943 The structure and metamorphism of Pre-Cambrian rocks, Great Slave Lake. Am.Jour.Sci. 241, p.430.

Higazy, R.A.H. 1949 Petrogenesis of Perthite pegmatites in the Black Hills, South Dakota. Jour.of Geol. 57, p.555.

Hollingworth, S.E., Wells, M.K. and Bradshaw, R. 1960. The Geology and structure of the Glomfjord region, Northern Norway. Int.Geol.Congr. 21st session, Copenhagen.

Holmsen, G. 1932 Rana. Norges Geol.Unders. 136.

Hietanen, A. 1956 Kyanite, andalusite and sillimanite in the schist in Boehls Butte Quadrangle, Idaho. Am.Min. 41, nos 1-2, p.1.

Holtedahl, O. 1919 The Palaeozoic Formations of Finnmark. Am.Jour.Sci.

1920 The Scandinavian Mountain Problem. Quart.Jour.Geol.Soc.Lond. 67, p.387.

1922 Tillite-like conglomerate in the Eocambrian sparagmite of Southern Norway. Am.Jour.Sci. 4, p.165.

1938 Geological Observations in the Opdal-Sunndal-Trollheim District. Nors.geol.tids. 19, p.326.

1944 On the Caledonides of Norway. Nors.Videns. Akad.i Oslo Mat-Naturv. Klasse 4.

1948 Deep-seated crystal deformations in the north-west part of the Caledonides of Southern Norway. Int.Geol.Congr.Lond., pt.13.

Holtedahl, O. and Dons, J.A. 1953

Berggrunskart over Norge.Bilag til Norges Geol.Unders. 164.

Kautsky 1947 Neue Gesichtspunkte zu einigen Nordskandinavischen Gebirgsproblem. Geol.Fören Stockholm Förh. 69, p.108.

1953 Der geologische Bau des Sulitelma-Salojauregebietes in den Nordskandinavischen Kaledoniden. Sver.Geol.Unders.Arbok 46 (for 1952) 4. English summary.

King, B.C. and Rast, N. 1955. Tectonic styles in the Dabradians and Moines of parts of the central Highlands of Scotland. Proc.Geol.Assoc. 66, p.243.

Köhler, A. 1949 Recent results of investigations of the felspars. Jour.of Geol. 57, p.592.

Kvale, A. 1953 Linear structures and relation to movement in the Caledonides of Scandinavia and Scotland. Quart. Jour. Geol.Soc.Lond. 109, p.51.

Rosenqvist, L. 1941 The Lönset Anticline in the Opdal area. Nors.Geol.tids. 21, p.25.

1941 Metamorphism and Metasomatism in the Opdal area. Nors.Geol.tids 22, p.106.

1952 The metamorphic facies and the felspar minerals. Univ.Bergens Årbok. Naturv. Rekke 4.

Rutland, R.W.R. 1958 The structural geology of the Sokumvatn area Gildeskål, Northern Norway, and some related petrogenetic problems. Ph.D. thesis, University of London.

1959 Structural Geology of the Sokumvatn area, North Norway, Nors.geol.tids. 39, p.287.

Rutland, R.W.R., Holmes , M and Jones M.A. 1960. The granites of the Glomfjord area, Northern Norway. Int.Geol.Congr. 21st session, Copenhagen.

Sederholm, J.J. 1916 On Synantetic minerals and related phenomena. Bull.Comm.Geol.Finlande, 48.

1926 On migmatites and associated Pre-Cambrian rocks of south-western Finland. Part II. Bull.Comm.geol.Finlande, No.77, p.69.

Shaw, D.N. 1960 The Geochemistry of Scapolite, pt.I. Jour.of Pet. Vol.I, p.218.

de Sitter, L.V. 1956 Structural Geology, New York.

Skjeseth and Sörensen, H. 1952. An example of granitisation in the central zone of the Caledonides of Northern Norway. Norsk.Geol.Unders. 184, p.154.

Smith, J.V. 1958 The effect of composition and structural state on the rhombic section and pericline twins of plagioclase felspars. Min.Mag.31, p.914.

Sörensen, H. 1954 A petrographical and structural history of the rocks at Engenbraeen, Holandsfjord. Norsk. Geol.Unders. 191, p.71.

Misch, P.	1949	Metasomatic granitisation of batholithic dimensions. Part I. Am.Jour.Sci. 247, 209-245, 372-406, 673-705.
Nicholson, R.	1960	The metamorphic rocks of the Storglomvatn area, Northern Norway. Ph.D. thesis, University of London.
Pettijohn, F.J.	1949 (2nd ed. 1957)	Sedimentary Rocks
Phillips, F.C.	1945	Micro-fabrics of the Moine schists. Geol.Mag. 82, p.205.
Pitcher, W.S.	1952	The migmatitic older granodiorite of the Thorr district Co.Donegal. Quart.Jour.Geol. Soc.Lond., 108, p.413.
Ramsay, J.G.	1957	Superimposed folding at Loch Morar. Quart. Jour.Geol.Soc.Lond., 113, p.271.
	1957	Moine-Lewisian relations at Glenelg. Quart. Jour.Geol.Soc.Lond., 113, p.487.
	1960	Deformation of early linear structures in areas of repeated folding. Jour.Geol. 68, p.75.
Read, H.H.	1927	A note on ptygmatic folding in the Sutherland Granite complex. Summary of progress. Geol.Survey G.B. part II.
	1943	Meditations on Granite. Proc.Geol.Assoc. 54, p.64.
	1943	Granites and granites. Geol.Soc. Am. Mem.28, p.1.
	1948	A Commentary on Place in Plutonism. Quart. Jour.Geol.Soc.Lond., 104, p.155.
	1949	Contemplation of Time in Plutonism. Quart. Jour.Geol.Soc.Lond., 105, p.101.
Reynolds, O.	1946	The Sequence of Geochemical Changes leading to Granitisation. Quart.Jour.Geol.Soc.Lond., 102, p.389.
	1952	The Difference in Optics between Volcanic and Plutonic Plagioclases and its bearing on the Granite Problem. Geol.Mag. 89, p.233.

Sutton, J. and Watson, J. 1952. The Scardroy Inlier, Ross-shire. Quart.Jour.Geol.Soc.Lond., 108, p.99.

1954 The Moines of Fannick Forest. Quart.Jour. Geol.Soc.Lond., 110, p.21.

Turner, F.J. 1947 Determination of plagioclase with the 4-axis universal stage. Am.Miner.32, p.389.

1948 The Mineralogical and Structural Evolution of the Metamorphic Rocks. Mem.Geol.Soc.Amer. 30.

Turner, F.J. and Verhoogen, J. 1951. Igneous and Metamorphic Petrology. McGraw Hill.

Turner, F.J., Williams, H. and Gilbert, C.M. 1954. Petrography.

Tuttle, O.F. 1952 Origin of the contrasting mineralogy of extrusive and plutonic salic rocks. Jour.Geol. 60, p.107.

Tuttle, O.F. and Bowen, N.L. 1958. Origin of granite in the light of experimental studies in the system NaAl SiO_3-$KAlSi_3O_8$-SiO_2-H_2O. Geol.Soc.Am. Mem.74.

Walton, B.J. 1959 The metamorphic and structural geology of the Navervand region, near Glomfjord, Northern Norway. Ph.D. thesis, University of London.

Weiss, L.E. 1959 The geometry of superposed folding. Bull.Geol. Soc.Am. 70, p.91.

Weiss, L.E. and McIntyre 1957. Structural Geometry of Dabradian rocks, Loch Leven. Jour.Geol.65, p.

Wilson, G. 1946 The relationship of slaty cleavage and kindred structures to tectonics. Proc.Geol.Assoc. 57, p.263.

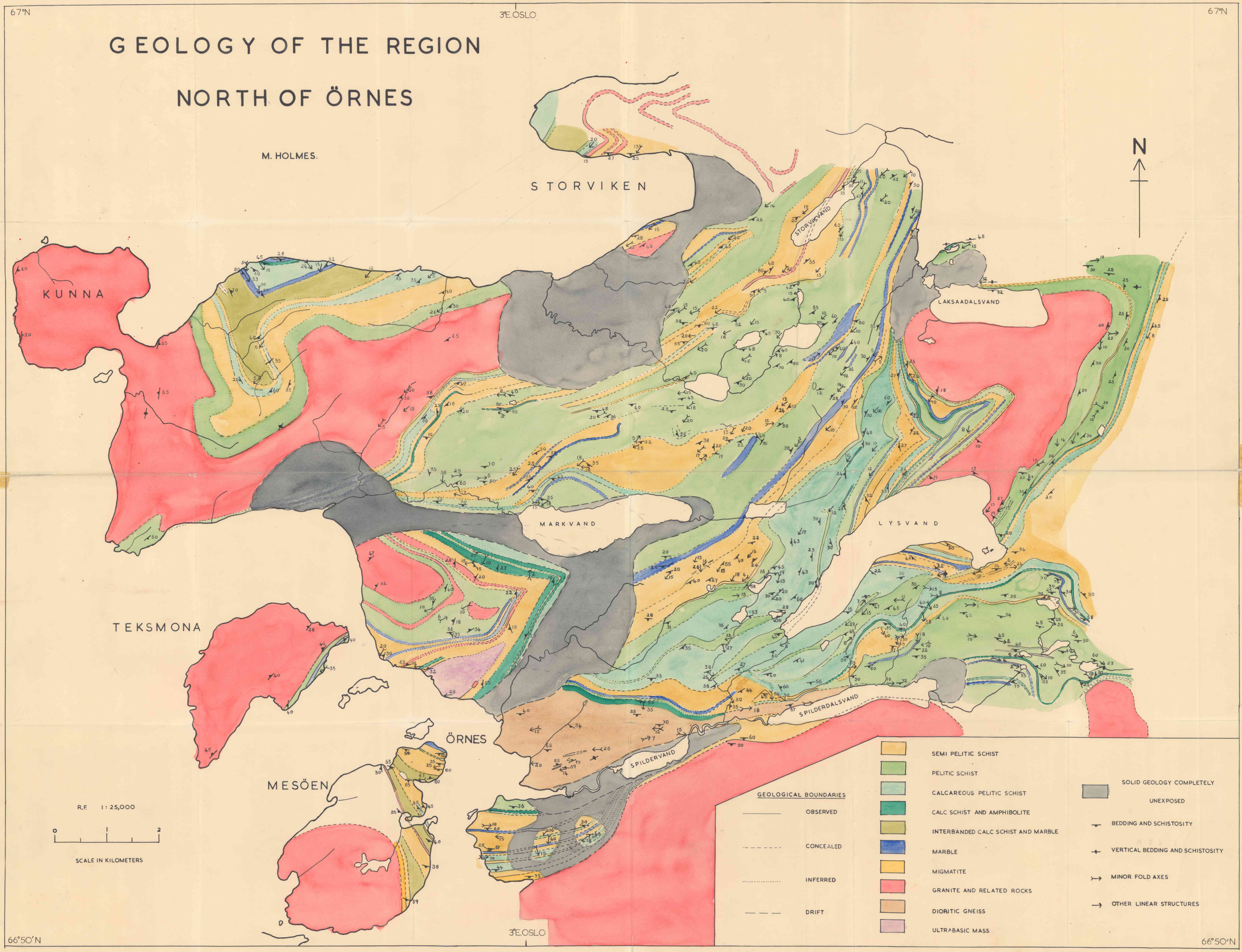
1952 Ptygmatic structures and their formation. Geol.Mag. 89, p.1.

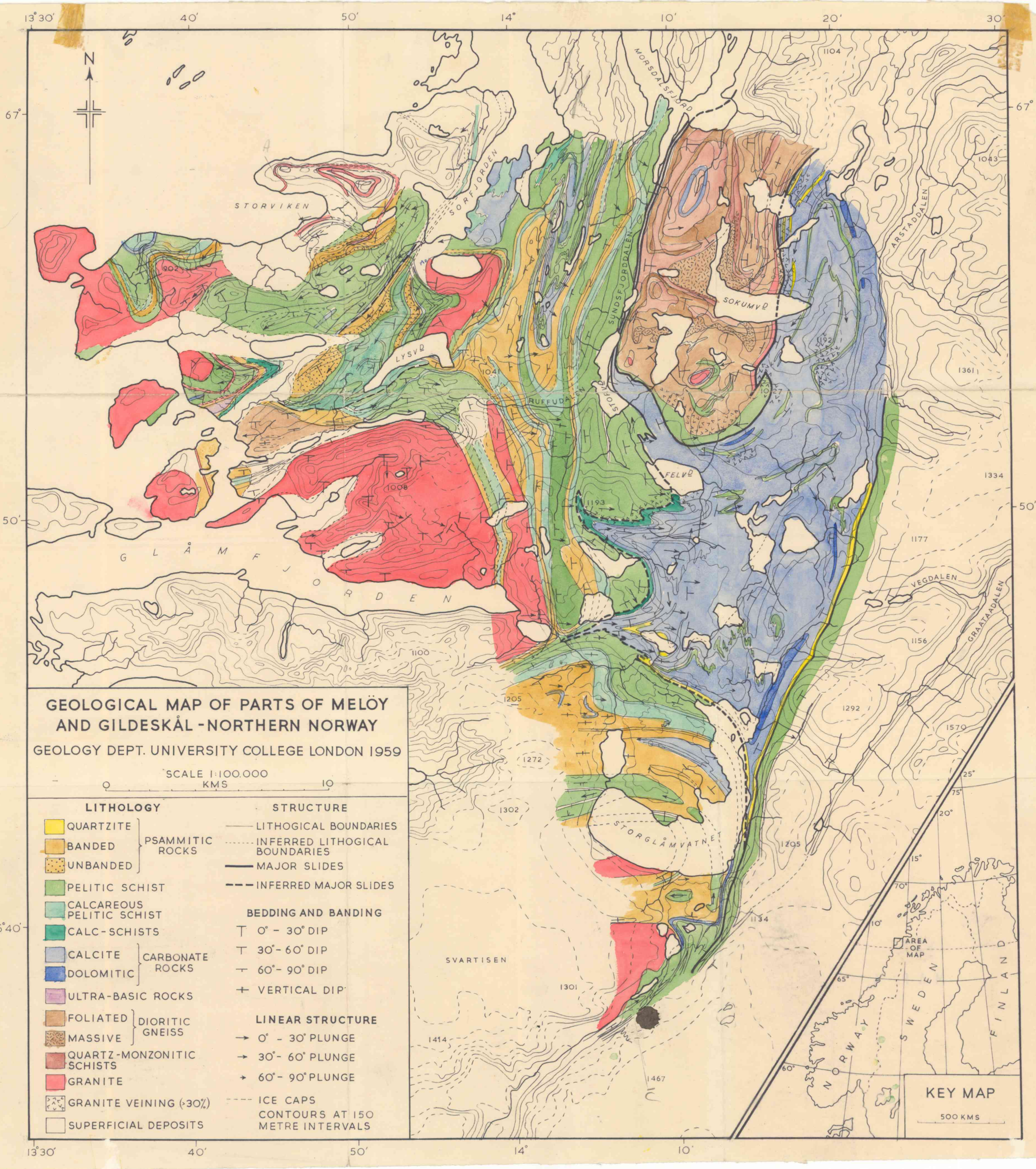
1952 Quartz vein system in the Moine Series, Melness. Geol.Mag. 89, p.141.

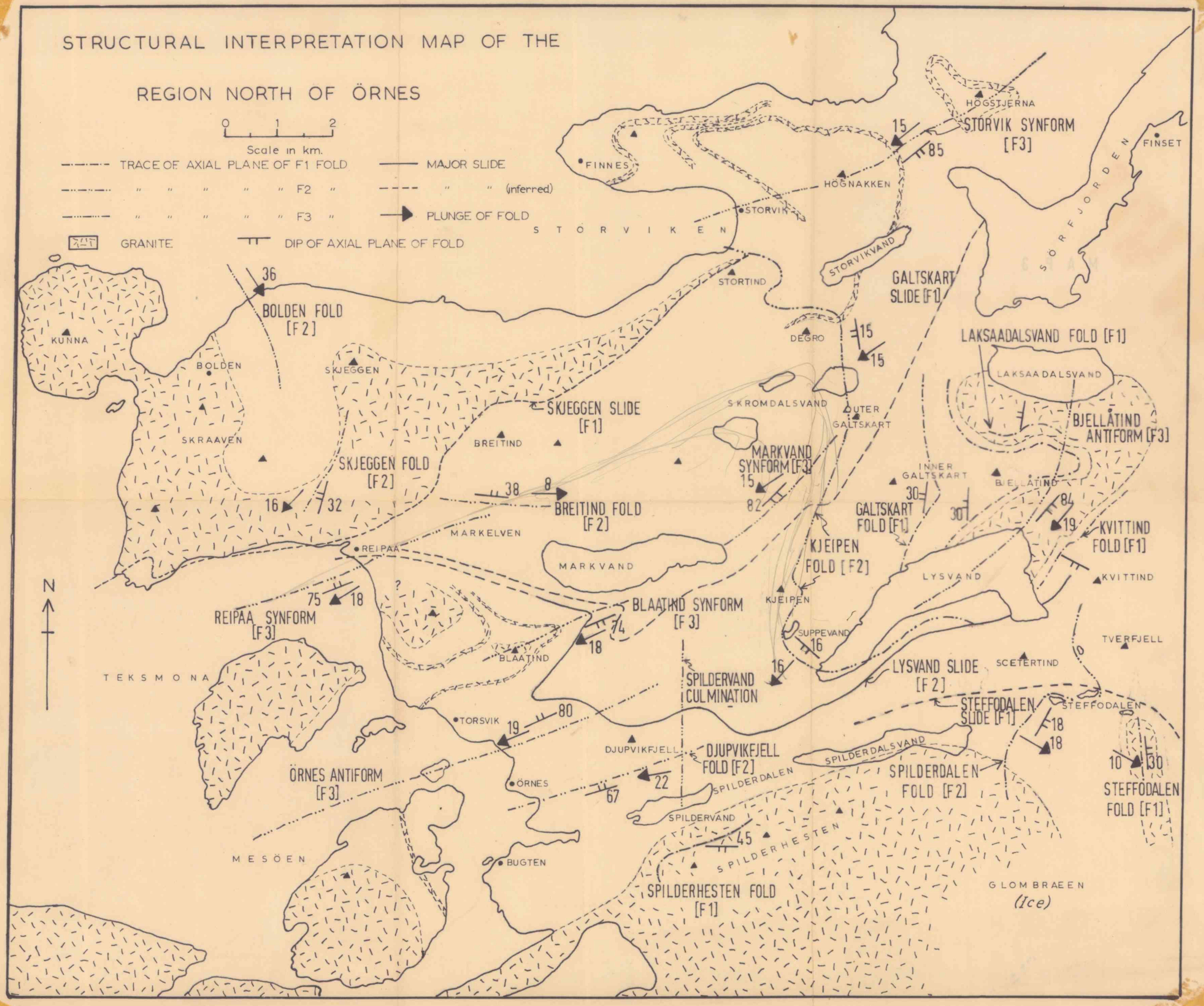
Wilson, G. 1953 Mullion and rodding structures in the Moine Series of Scotland. Proc.Geol.Assoc. 64, p.118.

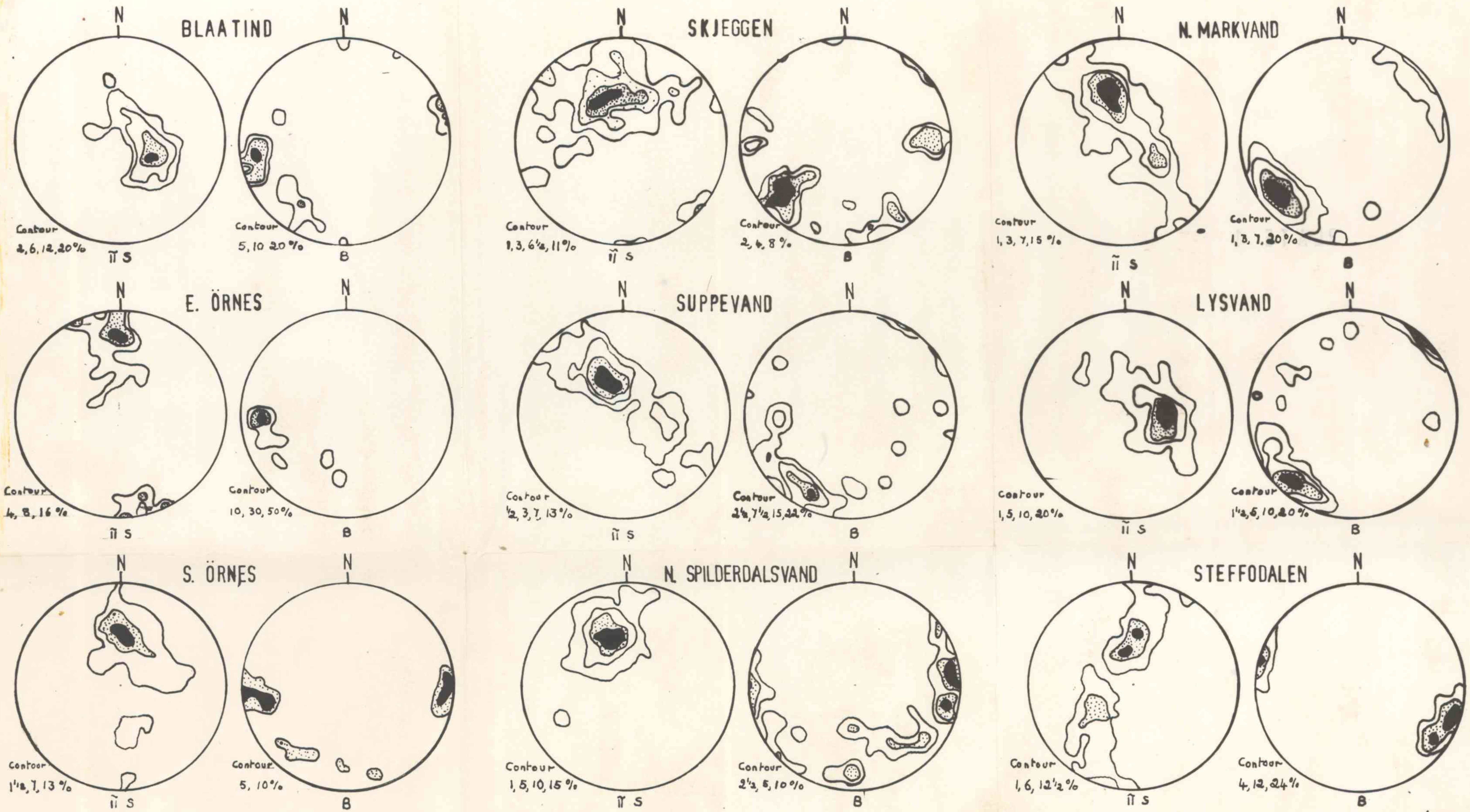
Winchell, A.N. and Winchell, H. 1951. Elements of Optical Mineralogy.

Wiseman, J.D.H. 1934 The central and south west Highland Epidiorites: a study in progressive metamorphism. Quart.Jour.Geol.Soc. Lond. 90, p.354.

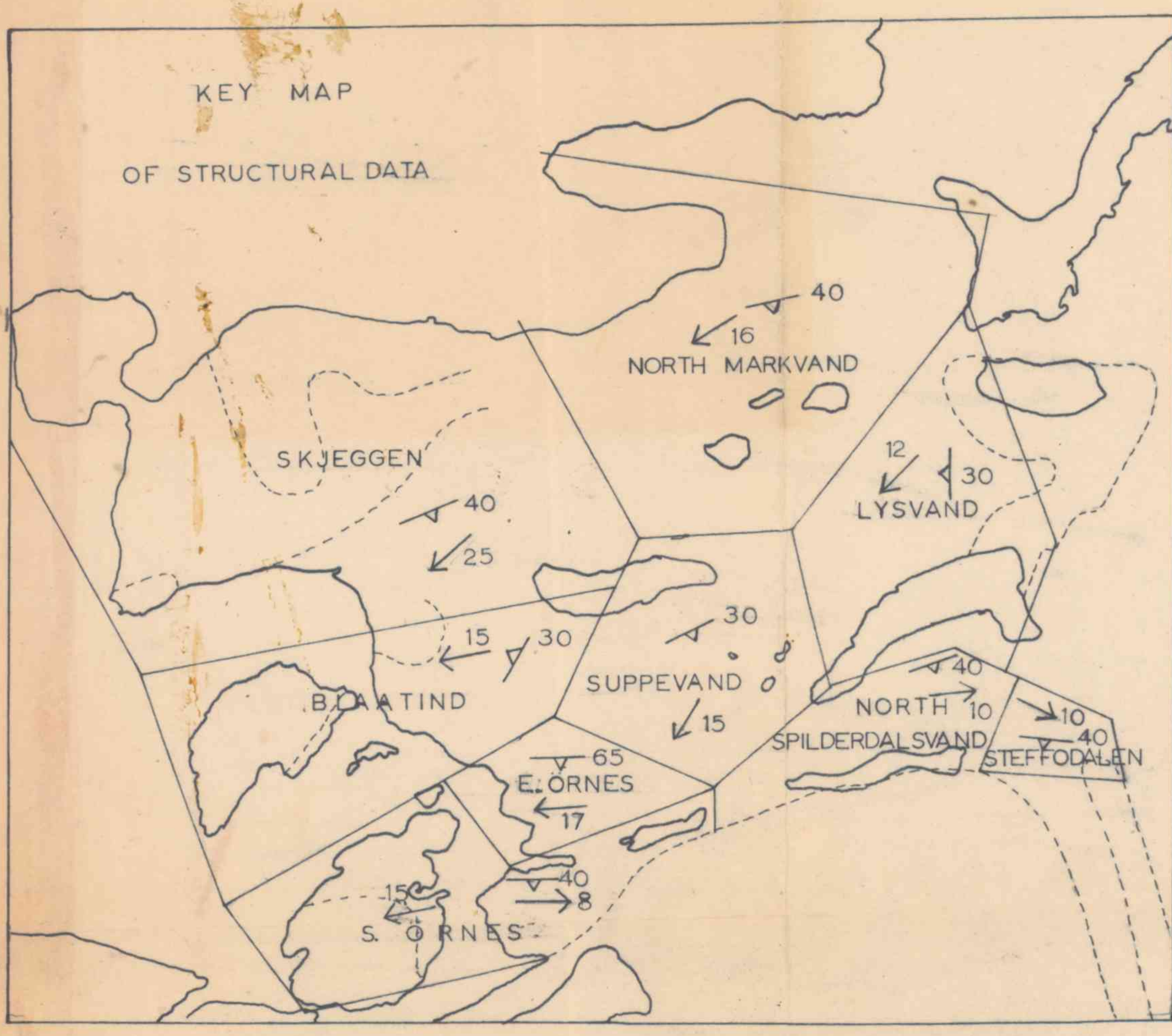

GEOLOGY OF THE REGION
NORTH OF ÖRNES
M. HOLMES.
67°N
3°E. OSLO
67°N
66°50'N
3°E. OSLO
66°50'N
N
STORVIKEN
STORVIKVAND
KUNNA
LAKSAADALSVAND
MARKVAND
LYSVAND
TEKSMONA
ÖRNES
SPILDERDALSVAND
SPILDERVAND
MESÖEN
R.F. 1 : 25,000
0 1 2
SCALE IN KILOMETERS
GEOLOGICAL BOUNDARIES
OBSERVED
CONCEALED
INFERRED
DRIFT
SEMI PELITIC SCHIST
PELITIC SCHIST
CALCAREOUS PELITIC SCHIST
CALC SCHIST AND AMPHIBOLITE
INTERBANDED CALC SCHIST AND MARBLE
MARBLE
MIGMATITE
GRANITE AND RELATED ROCKS
DIORITIC GNEISS
ULTRABASIC MASS
SOLID GEOLOGY COMPLETELY UNEXPOSED
BEDDING AND SCHISTOSITY
VERTICAL BEDDING AND SCHISTOSITY
MINOR FOLD AXES
OTHER LINEAR STRUCTURES

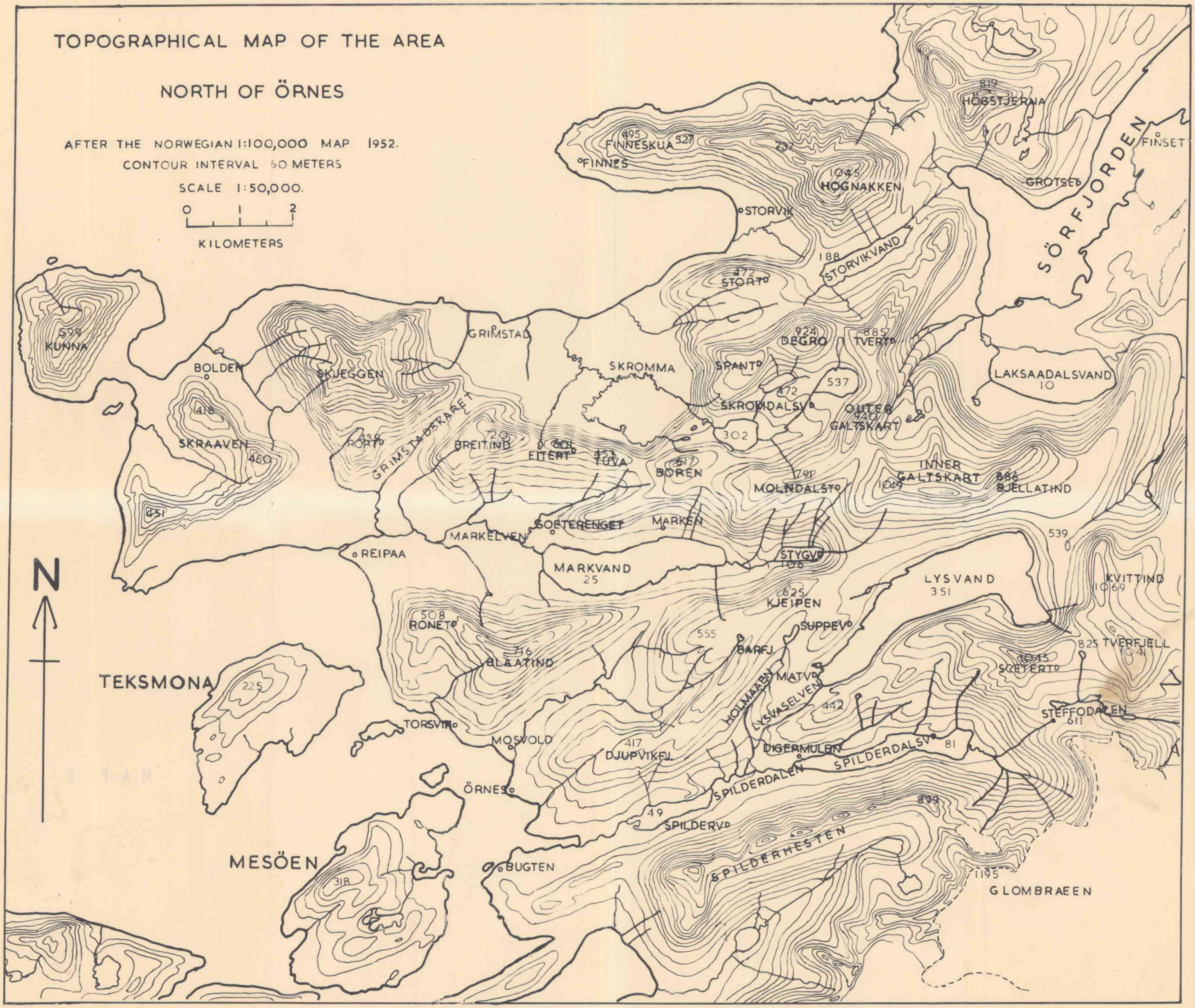

GEOLOGICAL MAP OF PARTS OF MELÖY
AND GILDESKÅL - NORTHERN NORWAY
GEOLOGY DEPT. UNIVERSITY COLLEGE LONDON 1959
SCALE 1:100.000
0 KMS 10
LITHOLOGY
QUARTZITE
BANDED
UNBANDED
PSAMMITIC ROCKS
PELITIC SCHIST
CALCAREOUS PELITIC SCHIST
CALC-SCHISTS
CALCITE
DOLOMITIC
CARBONATE ROCKS
ULTRA-BASIC ROCKS
FOLIATED
MASSIVE
DIORITIC GNEISS
QUARTZ-MONZONITIC SCHISTS
GRANITE
GRANITE VEINING (>30%)
SUPERFICIAL DEPOSITS
STRUCTURE
LITHOGICAL BOUNDARIES
INFERRED LITHOGICAL BOUNDARIES
MAJOR SLIDES
INFERRED MAJOR SLIDES
BEDDING AND BANDING
0° - 30° DIP
30° - 60° DIP
60° - 90° DIP
VERTICAL DIP
LINEAR STRUCTURE
0° - 30° PLUNGE
30° - 60° PLUNGE
60° - 90° PLUNGE
ICE CAPS
CONTOURS AT 150 METRE INTERVALS
KEY MAP
500 KMS
AREA OF MAP
NORWAY
SWEDEN
FINLAND
N
STORVIKEN
SÖRFJORDEN
MORSDALSFJORD
SUNDSFJORDDALEN
SOKUMVD
LYSVD
RUFFUDALEN
STORÅ
FELVD
GLÅMFJORDEN
STORGLÅMVATNET
SVARTISEN
ARSTADDALEN
VEGDALEN
GRAATAADALEN
13°30'
40'
50'
14°
10'
20'
30'
67°
50'
66°40'


STRUCTURAL INTERPRETATION MAP OF THE
REGION NORTH OF ÖRNES
0 1 2
Scale in km.
TRACE OF AXIAL PLANE OF F1 FOLD
" " " " " F2 "
" " " " " F3 "
GRANITE
MAJOR SLIDE
" " (inferred)
PLUNGE OF FOLD
DIP OF AXIAL PLANE OF FOLD
S T O R V I K E N
FINNES
STORVIK
HÖGNAKKEN
HÖGSTJERNA
STORVIK SYNFORM
[F3]
15
85
FINSET
S Ö R F J O R D E N
STORVIKVAND
STORTIND
GALTSKART
SLIDE [F1]
DEGRO
15
15
LAKSAADALSVAND FOLD [F1]
LAKSAADALSVAND
KUNNA
36
BOLDEN FOLD
[F2]
BOLDEN
SKJEGGEN
SKJEGGEN SLIDE
[F1]
SKRAAVEN
BREITIND
SKROMDALSVAND
YTTER
GALTSKART
BJELLÅTIND
ANTIFORM [F3]
MARKVAND
SYNFORM [F3]
15
82
INNER
GALTSKART
BJELLÅTIND
SKJEGGEN FOLD
[F2]
16
32
38
8
BREITIND FOLD
[F2]
30
30
GALTSKART
FOLD [F1]
84
19
KVITTIND
FOLD [F1]
KJEIPEN
FOLD [F2]
MARKELVEN
REIPAA
MARKVAND
LYSVAND
KVITTIND
KJEIPEN
N
75
18
REIPAA SYNFORM
[F3]
?
BLAATIND SYNFORM
[F 3]
74
18
SUPPEVAND
16
TVERFJELL
BLAATIND
TEKSMONA
16
LYSVAND SLIDE
[F2]
SCETERTIND
SPILDERVAND
CULMINATION
STEFFODALEN
SLIDE [F1]
STEFFODALEN
18
18
TORSVIK
19
80
DJUPVIKFJELL
DJUPVIKFJELL
FOLD [F2]
SPILDERDALSVAND
10
30
ÖRNES ANTIFORM
[F3]
22
ÖRNES
67
SPILDERDALEN
SPILDERDALEN
FOLD [F2]
STEFFODALEN
FOLD [F1]
SPILDERVAND
45
SPILDERHESTEN
MESÖEN
BUGTEN
SPILDERHESTEN FOLD
[F1]
GLOMBRAEEN
(Ice)



BLAATIND
N
N
Contour
2, 6, 12, 20 %
π S
Contour
5, 10 20 %
B
SKJEGGEN
N
N
Contour
1, 3, 6½, 11 %
π S
Contour
2, 4, 8 %
B
N. MARKVAND
N
N
Contour
1, 3, 7, 15 %
π S
Contour
1, 3, 7, 20 %
B
E. ÖRNES
N
N
Contour
4, 8, 16 %
π S
Contour
10, 30, 50 %
B
SUPPEVAND
N
N
Contour
½, 3, 7, 13 %
π S
Contour
2½, 7½, 15, 22 %
B
LYSVAND
N
N
Contour
1, 5, 10, 20 %
π S
Contour
1½, 5, 10, 20 %
B
S. ÖRNES
N
N
Contour
1½, 7, 13 %
π S
Contour
5, 10 %
B
N. SPILDERDALSVAND
N
N
Contour
1, 5, 10, 15 %
π S
Contour
2½, 5, 10 %
B
STEFFODALEN
N
N
Contour
1, 6, 12½ %
π S
Contour
4, 12, 24 %
B


KEY MAP
OF STRUCTURAL DATA
40
16
NORTH MARKVAND
12
30
LYSVAND
SKJEGGEN
40
25
30
15
30
BLAATIND
SUPPEVAND
15
40
NORTH
10
10
40
SPILDERDALSVAND
STEFFODALEN
65
E. ÖRNES
17
40
8
15
S. ÖRNES


TOPOGRAPHICAL MAP OF THE AREA
NORTH OF ÖRNES
AFTER THE NORWEGIAN 1:100,000 MAP 1952.
CONTOUR INTERVAL 50 METERS
SCALE 1:50,000.
0 1 2
KILOMETERS
N
FINNESKUA
FINNES
STORVIK
HÖGNAKKEN
HÖGSTJERNA
GROTSET
SÖRFJORDEN
FINSET
STORVIKVAND
STORT
DEGRO
TVERT
SPANT
LAKSAADALSVAND
KUNNA
BOLDEN
SKJEGGEN
GRIMSTAD
SKROMMA
SKROMDALSV
OUTER GALTSKART
SKRAAVEN
PORT
GRIMSTADSKARET
BREITIND
SOL
FITERT
TUVA
BOREN
MOLNDALST
INNER GALTSKART
BJELLATIND
SOETERENGET
MARKEN
MARKELVEN
REIPAA
MARKVAND
STYGVD
LYSVAND
KVITTIND
KJEIPEN
SUPPEV
RONET
BLAATIND
BARFJ
TVERFJELL
SÆTERT
TEKSMONA
HOLMAABN
MATV
LYSVASELVEN
STEFFODALEN
TORSVIK
MOSVOLD
DJUPVIKFJ
DIGERMULEN
SPILDERDALSV
SPILDERDALEN
ÖRNES
SPILDERV
SPILDERHESTEN
MESÖEN
BUGTEN
GLOMBRAEEN